THE JORDAN RIFT VALLEY

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In memory of Leo (Yehuda) Picard (3.6.1900–4.4.1997), who first introduced me to the complexities of the Jordan Valley.

Contents

Preface

The Jordan Rift Valley has attracted increasing scientific interest from numerous scholars, since the beginning of the 19th century and even before. The puzzling geology, the variety of unique environments with their specific, partly endemic faunas and floras, together with the great antiquity of prehistoric and archaeological sites, all contributed to a worldwide interest for a long time, which continues in the present day. Scientific expeditions from Great Britain and other European countries, as well as the United States, had visited and studied the region from the end of the 18th century onward, studies usually culminated with the publication of lengthy volumes. The Jordan Rift Valley is a narrow, elongated depression, which separates Israel and Palestine to the west, the Kingdom of Jordan and Syria to the east, along almost 400 kilometers. It houses the lowest point on Earth, the Dead Sea, which is more than 400 meters below sea level, serving as a terminal drainage base level for the entire Jordan Valley. The Valley is bounded by faults on both flanks, but not continuously along its entire length. Internal faults are also quite common, thus forming an intricate and complicated rift valley system. It is certainly a continuation of the Red Sea spreading system, but opinions vary considerably as to the tectonic nature of this lineage.

The environment is Mediterranean in the northern Jordan Valley, grading to a bare, extremely arid desert to the south. The morphology makes for relatively high temperatures, and the border faults are responsible for a wealth of springs. The combination of warmth and water created unique subtropical microenvironments along the depression, which became a refuge for animals and plants of various provenances. In times of wetter climates throughout the Quaternary, even the southern Jordan Valley became habitable to man, and prehistoric sites are thus common. It is within the Jordan Rift Valley that the earliest hominid sites outside Africa were discovered.

The combination of favorable environments, ease of movement along the Valley, Africa on one side and Eurasia on the other, all made the Jordan Rift Valley a most suitable gateway for early hominids on their way to settle the world. This process commenced at least two million years ago, maybe even earlier, an age attributed to the earliest artifacts found in the Jordan Rift Valley.

The complicated geological and structural processes, accompanied by climate changes, had created ever changing, unique natural environments along the Jordan Rift Valley, which during many periods resembled those of subtropical Africa. Thus it was not altogether surprising that the common thought is that hominids chose this route for migration. Truly, it was probably the only continental route, since the rest of the region is hilly, or covered by sand dunes and marshes. Had the Jordan Rift Valley not been there, the history of mankind may have looked somewhat different, some people assume. It is of prime importance to analyze and understand Man's migration, which can only be done by combining the geological, climatic and anthropological evidence. It is the basis of the multidisciplinary approach adopted in this book and the rationale for writing it.

The book is divided into 12 chapters. An introduction, delineating the singularities of the Jordan Rift Valley, its uniqueness as a probable oceanic feature on land, its possible role in human migration, the terminology used and the concepts that led to writing the book. Chapter 2 deals with the history of research, both geological and anthropological–archaeological. Chapter 3 describes the present-day characteristics of the Jordan Rift Valley. The pre-Rift regional geology is dealt with by A. Flexer in Chapter 4, while Chapter 5 describes the lithostratigraphy of the Embryonic and Eritrean stages, periods during which the initial development of the Jordan Valley took place. Chapter 6 discusses the late Cenozoic palynostratigraphy and climates, including presentation of climatic models, based on detailed pollen analyses of numerous boreholes and outcrops.

Chapter 7 discusses relative and absolute datings, the paleogeography and development of natural environments during the Embryonic and Eritrean stages in the Rift's evolution. In Chapter 8, A. Ginzburg and Z. Ben-Avraham focus on results of geophysical studies, presenting conclusions regarding the structure, particularly of the deeper layers. Chapter 9 discusses the structural history of the Jordan Rift Valley throughout time, as part of the Red Sea spreading system, throughout the Embryonic, Eritrean and Levantine stages, from the Oligocene up to the present day. Chapter 10 presents the debate on tectonics and evolution of the Jordan Rift Valley, and a new model is proposed to explain its peculiarities, as part of the Red Sea spreading system. A different but common view, which regards the Rift as a leaky transform, is advanced by Z. Garfunkel at the end of the book, as an Appendix.

The last two chapters are concerned with human habitation in the Jordan Valley and vicinity. Chapter 11 is devoted to the stratigraphy and dating of the Quaternary in-Rift formations deposited during the Levantine stage, when the Rift became an endoreic system, including the significant human sites. Chapter 12 is dedicated to the paleoecology of Man, his relations with the natural environments and his food and water supplies; as well as various data and theories (unfortunately more of the latter!) that may pertain to human migration from Africa through the Levant, on the way to settle the world.

The book is intended, because of its essentially multidisciplinary nature, for a variety of students and scholars: Anthropologists, who would want to follow the first steps of Man out of Africa and the evolution of his cultures in this part of the world thereafter. Archeologists, for whom the Jordan Valley as part of the Levant is crucial for the development of agriculture, commerce and urbanism (Jericho, the oldest city in the world, is located in the Jordan Valley). Geologists would want to become familiar with the evolution of this singular phenomenon on Earth. Geomorphologists could be interested in landscape-shaping processes in a rift valley environment, with a steep environmental gradient, under constantly changing climates and base levels. Geophysicists, for whom the Jordan Rift Valley is a unique locality to study crustal movements similar to those occurring beneath the oceans, can only here see such processes with their own eyes, and not solely with the aid of remote sensing instruments, sophisticated as they may be. Stratigraphers, palynologists and paleoclimatologists will find interest in the longest available continuous continental sequences, spanning the entire late Cenozoic. Evolutionists, botanists, zoologists and physiologists will marvel at the adaptation and evolution of plant and animal communities in constantly changing environments, which on the other hand offer shelters nearby even in the worst of times. Historians of thought and sciences will find interest in the heated debate on the tectonic nature of the Jordan Rift Valley, a debate that started some time in the mid-19th century and has not ended until the present day.

It is my humble hope that this book will contribute to the understanding of Man and our environment. For we are advancing into a new era in the history of Man, a time when we feel no more dependent on the environment, but could rather ruin it at will. The harmony of the past could then turn to a disaster of the near future.

Aharon Horowitz
Ramat Aviv

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Although in most cases nothing personal was involved, I want first to thank the numerous researchers who studied the Jordan Rift Valley, its surroundings and problems concerned with its history and development. Many of them are mentioned in the reference list, but my gratitude is extended to all whose efforts provided a firm foundation to my writings.

Some 30 years ago I moved from the Geological Survey of Israel, Jerusalem, to Tel Aviv University, but the Survey remained a second home for me ever since, where I enjoyed all kinds of help and cooperation (not to mention amounts of coffee!). I am indebted to Gideon Steinitz, Director, and Varda Arad, Librarian, for their great help in obtaining access to all necessary literature. The database made by the Survey, in the form of geological and other computerized maps, satellite and shaded relief DTM images (Hall 1996, 2000, Sneh et al. 1998a), provided indispensable foundations for numerous considerations. Parts of these maps and images are reproduced by permission in this book, for which I am deeply indebted to the Survey. Particular thanks are due to John K. Hall and Amihai Sneh for their considerable help, cooperation and patience, good will and smiles. Besides the geological maps, hours on end of fruitful discussions and days in the field with Amihai Sneh helped clear up so many points. His help is considerably appreciated.

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CHAPTER 1

Introduction: the singularity of the Jordan Rift Valley

My first encounter with the Jordan Rift Valley and its peculiar geological and anthropological characteristics was back in 1960, when I was studying second year geology at the Hebrew University of Jerusalem. I was hired as a driver and assistant for one of the first excavation seasons at Ubeidiya, a site that became famous some time before, due to some finds of a few fragments of bones attributed to *Homo erectus* (Tobias 1966). Unfortunately, to general dismay, no further such remains have ever been found since, despite many years of subsequent excavations. But the mere occurrence of this site prompted considerable further interest in the region and its occupation by hominids.

The site of Ubeidiya (Figs 11.2.2 and 12.1.4.1) is found in a complex of tilted and contorted fluvio-lacustrine beds, rich in artifacts, bones and mollusks. The excavations, which continued for many years, involved detailed geological studies not only at the site itself but all along the Jordan Valley, resulting in a wealth of reported information. The first stages of research had been led by the late Leo Picard (geology) and Moshe Stekelis (prehistory), both from the Hebrew University of Jerusalem. The work is being continued until the present day, by their students and followers.

The natural thing for me, after studying palynology at the State University of Leiden, the Netherlands, with Thomas van der Hammen, was to do my dissertation on this region. Ever since, I have been almost continuously involved with studies of the Jordan Rift Valley, divided into two main stages. The first comprised studies of outcrops, sites and shallow boreholes, comparing the results with other Quaternary formations in the country, resulting some 20 years later in the publication of a synthesis of *The Quaternary of Israel* (Horowitz 1979). The second stage followed rising interest in the search for oil and natural gas within the Rift Valley deposits, which resulted in many deep boreholes sunk into the Rift Valley fill, totaling some 40 km in thickness of the penetrated formations. I had analyzed pollen grains, which turned to be the only fossils by which the stratigraphy of the fill could be worked out, in most of these boreholes. New techniques and concepts developed for this project, and others in the region, were summed up in a book on *Palynology of arid lands* (Horowitz 1992a). The resulting palynostratigraphy and comparisons made with outcrops, as well as with drill sequences of marine sediments, gave a sound basis for understanding the history and evolution of the Jordan Rift Valley through time.

Numerous studies, in both Israel and the Kingdom of Jordan, encompassing all branches of geology, prehistory and archaeology, have been carried out in the Jordan Rift Valley during the last 20 years. These considerably widened our knowledge, and somewhat changed the ideas presented back in my 1979 and 1992 books, thus constituting the basis for this one.

1.1 A GEOGRAPHICALLY, GEOLOGICALLY AND ENVIRONMENTALLY UNIQUE ENTITY

1.1.1 Geography

The Jordan Rift Valley, also known as "The Jordan–Arava Rift Valley", "The Jordan–Dead Sea Rift", "The Dead Sea Rift" or "The Dead Sea Transform" (the last two terms are widely used by geologists, to include almost the entire longitudinal rift system of the Levant), is a long, narrow, approximately north–south oriented depression, separating Israel and Palestine to the west from the Kingdom of Jordan and Syria to the east (Fig. 1.1). Historically, ever since pre-biblical times, the Jordan Valley had served as a natural boundary between peoples occupying the Land of Cana'an to the west, mostly practicing sedentary agricultural economy, and the nomadic tribes of Transjordan to the east, mainly shepherds. This conflict, which to a certain degree still persists today, is diminishing now with the introduction of modern agricultural technologies.

The southern Levant is longitudinally divided into four main morphotectonic units: the coastal plain bordering the eastern Mediterranean, the western highlands, the Jordan Rift depression and its southward and northward extensions, and the eastern highlands of Transjordan and southern Syria. The Jordan Valley (Brawer 1957, pp. 40–41), which occupies some 5,650 km^2, is subdivided into four major parts:

(1) the northern sector, down from an approximate elevation of 100 m above sea level, with an area of 535 km^2, extends from the Lebanese–Syrian border to the northern end of the central Jordan Valley, including two lakes: the Hula, 14 km^2, now artificially drained, and Lake Kinneret (Sea of Galilee), some 165 km^2, now somewhat shrunken due to excess exploitation of its waters;

(2) the central sector, between Lake Kinneret and the Dead Sea, 1,100 km^2 (465 west of the Jordan River, 635 to the east);

(3) the Dead Sea had occupied some 50 years ago an area of 1,050 km^2, but this has been considerably reduced ever since, particularly its southern basin, due to human activities such as potash production from its waters and extensive use of its freshwater supplies to the north;

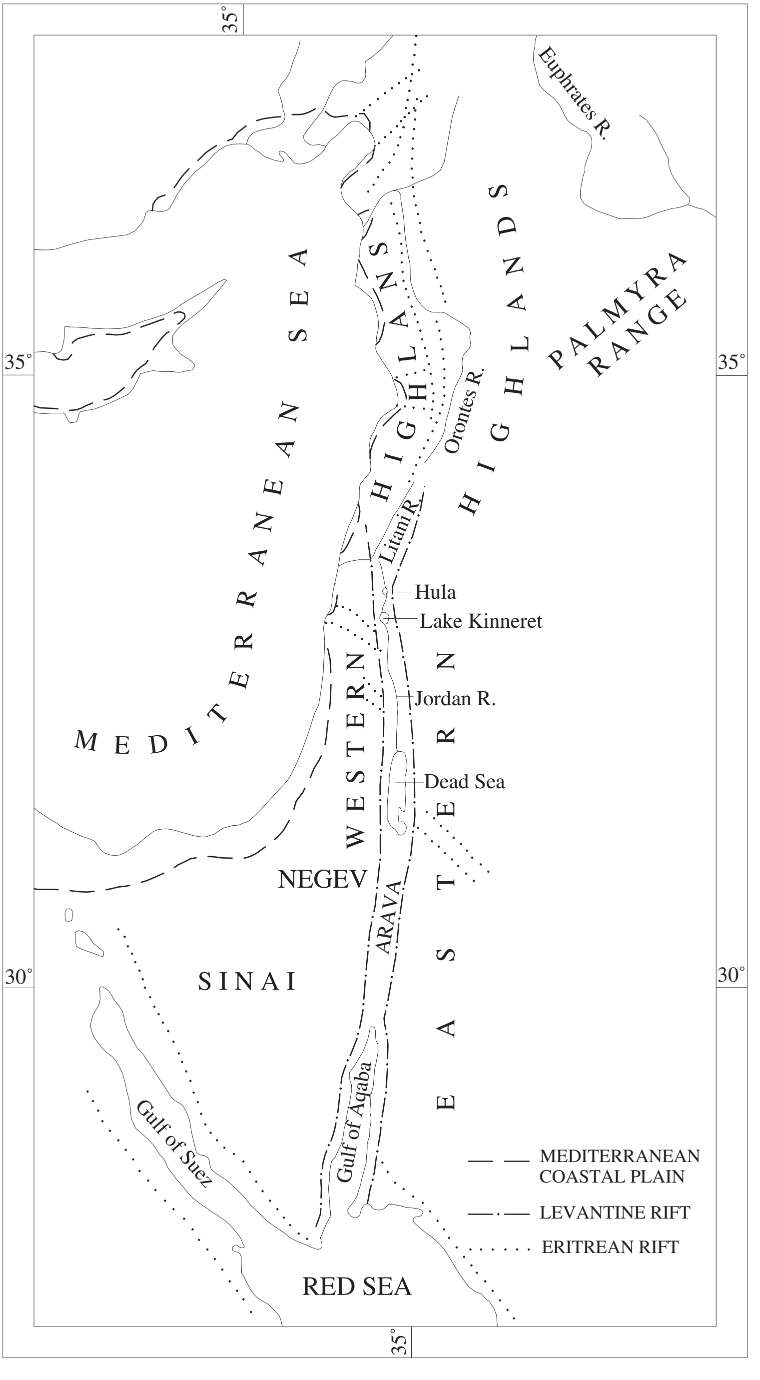


Figure 1.1. The Levant, general location map. R.: River.

(4) the Arava, from the Dead Sea southward to the Gulf of Aqaba (Elat), stretches over some 3,000 km^2.

The present book deals with the part of the Rift which is drained into the Dead Sea (Fig. 1.2), bordered by two internal watersheds: to the north, the one separating the Lebanese drainage system going to the Mediterranean from the internal Jordan River drainage; to the south, the watershed between the Dead Sea drainage, running north, and the southward system leading to the Gulf of Aqaba. The western highlands watershed separates the Dead Sea drainage from that leading to the Mediterranean; the eastern highlands watershed divides between the Dead Sea drainage and the eastward system, flowing either to the Persian Gulf or toward endoreic internal depressions.

The Jordan River runs along the Valley southward, from southern Lebanon and Syria, passes two intermediate lakes on its way, the Hula and Lake Kinneret, and pours its waters into its terminal, endoreic, hypersaline lake, the Dead Sea. The major part of the water comes from the north, but several rivers and springs help on the way, the more conspicuous ones arriving from the east. South of the Dead Sea the drainage system, leading generally to the north, is well developed but dry, except for occasional floods in wintertime. This wadi system extends in some places over distances of several hundred kilometers.

The three lakes in this system are considerably different: the Hula, 70 m above sea level, was only a few meters deep before drying up, following its land reclamation project in the 1950s, and apparently not much different throughout the last two million years, ever since its formation (Horowitz 1973); Lake Kinneret, some 200 m below sea level, attains up to 40 m in water depth but is very young, having been subsiding to its present shape over the last 18,000 years (Neev & Emery 1967); while the Dead Sea, almost 400 m below sea level, has two distinct basins, the northern which is about 380 m deep, and the southern, which is very shallow, at only 2–3 m, and is now practically dry. Interestingly, the southern basin is much older, being formed at least two million years ago, but commencing slow subsidence much before that, as early as the Oligocene; while the northern basin had formed within the same process that also created Lake Kinneret (Horowitz 1996a, 1997), during the late Pleistocene.

The Jordan Rift is bordered on both sides by highlands (Fig. 1.3): the Galilee, Samaria, Judea and the Negev to the west, the Golan and Transjordan plateaus to the east. These highlands were elevated in the process of formation of the Jordan Rift Valley depression during the last two million years, and are therefore viewed here, following Hull (1886, p. 108), Picard (1943, p. 41) and Horowitz (1979, pp. 53, 61), as an integral structural part of the Rift system of the Levant. Indeed, it seems that recognition of the uplifting processes as an integral part of the Syrian–African Rift system was to some degree underrated by many scholars, which, in my view, resulted in several invalid explanations for the tectonic history and development of this major lineament.

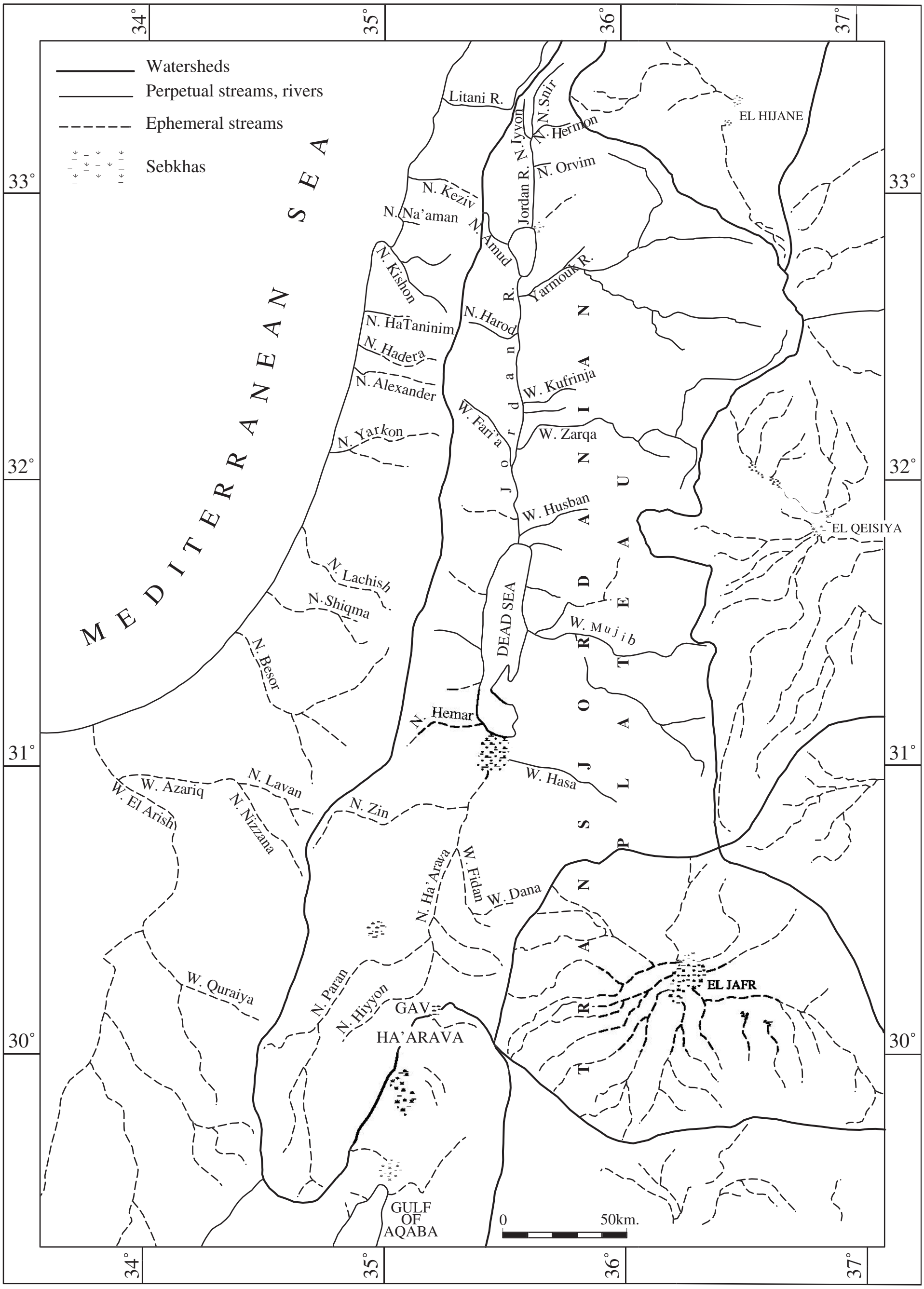


Figure 1.2. The Jordan Valley drainage pattern. W.: Wadi (ephemeral stream), R.: River, N.: Nahal (Hebrew for stream or wadi).

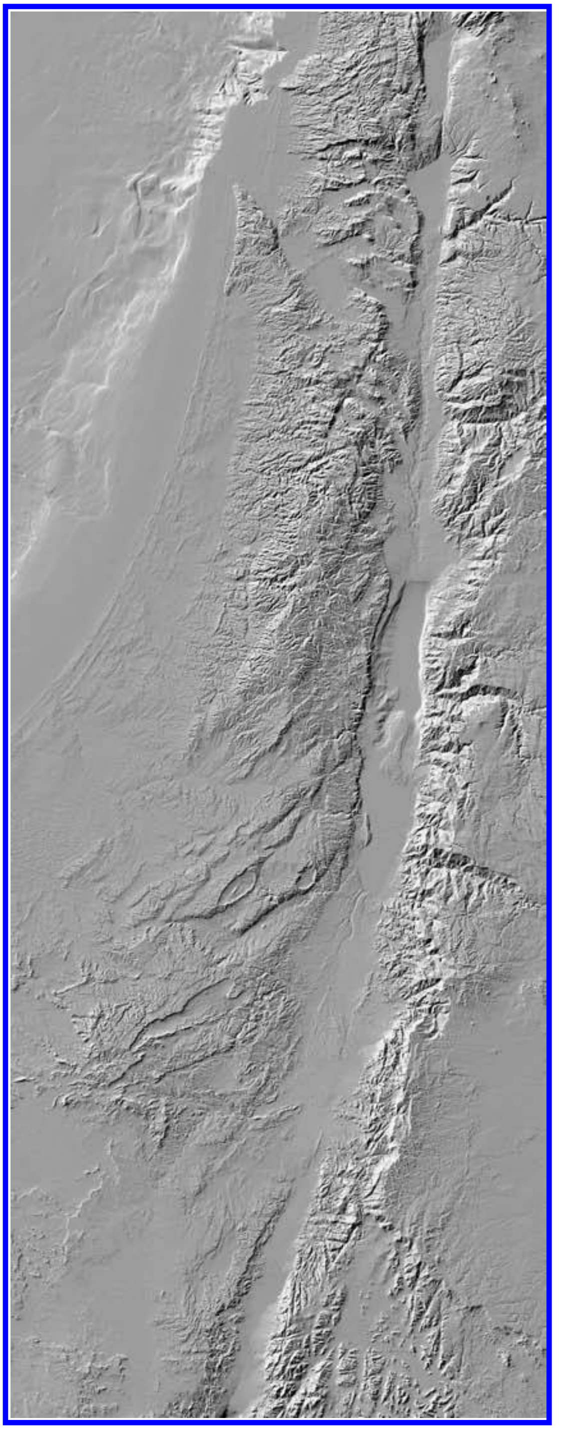


Figure 1.3A. Image of the southern Levant, based on the 25 m Digital Terrain Map (DTM) by Hall (1996). Coordinates 880–300N; 80–240E; 420 × 160 km. By permission of the Geological Survey of Israel.

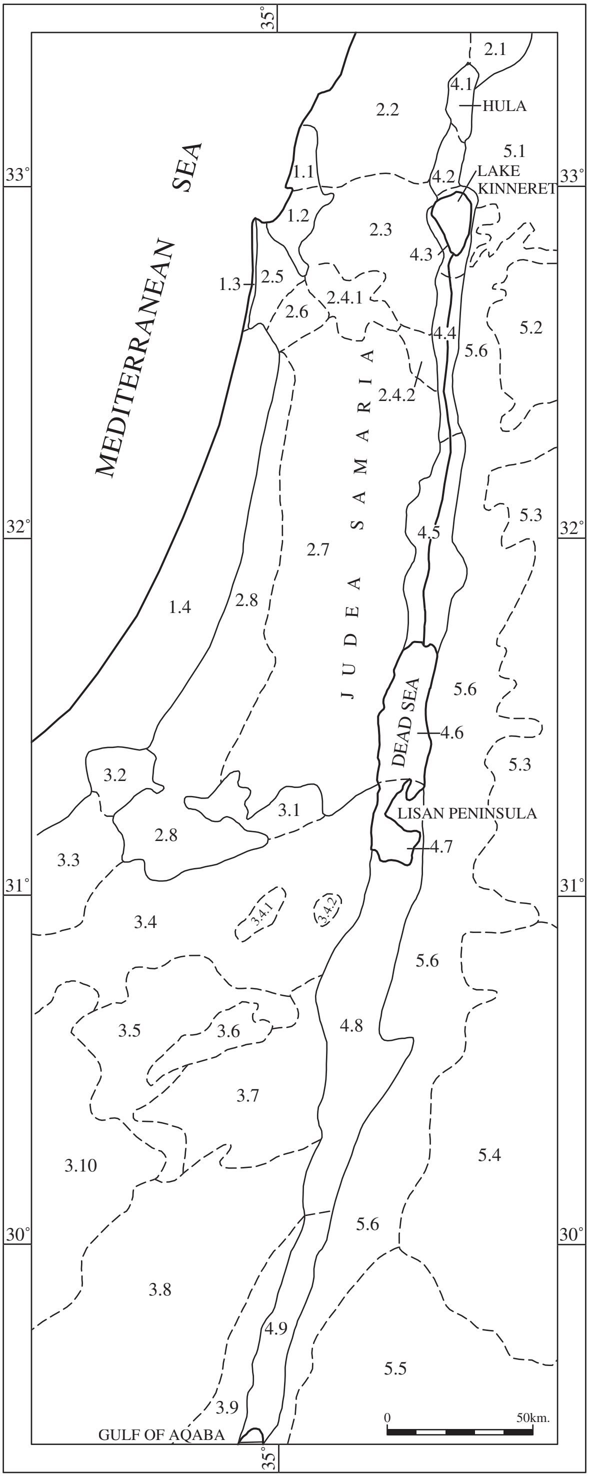


Figure 1.3B. Morpho-structural regions of the southern Levant: (1) Coastal plain: (1.1) western Galilee coastal plain, (1.2) Zevulun Valley, (1.3) Carmel coastal plain, (1.4) central and southern coastal plain; (2) western highlands and transversal valleys: (2.1) Mount Hermon and the Anti Lebanon, (2.2) upper Galilee and southern Lebanon mountains, (2.3) lower Galilee mountains, (2.4.1) Yizre'el Valley, (2.4.2) Bet She'an Valley, (2.5) Mount Carmel, (2.6) Menashe highlands, (2.7) Judea and Samaria hills (occasionally referred to as "mountains"), (2.8) Shefela hills; (3) Negev: (3.1) Be'er Sheva Valley, (3.2) Loess basin of Nahal Besor, (3.3) Haluza and Agur sand dunes, (3.4) northern and northwestern Negev fold belt, (3.4.1) Hatira anticline and erosion cirque, (3.4.2) Hazera anticline and erosion cirque, (3.5) central Negev highlands, (3.6) Ramon anticline and erosion cirque, (3.7) central Negev fold and fault zone, (3.8) southern Negev erosion plains of Nahal Paran and Nahal Hiyyon, (3.9) southern Arava and Elat bordering highlands, (3.10) Wadi el Arish basin; (4) Jordan Valley: (4.1) Hula Valley, (4.2) Korazim block, (4.3) Biq'at Kinarot (Lake Kinneret and surrounding lowlands), (4.4) central Jordan Valley, (4.5) southern Jordan Valley, (4.6) northern Dead Sea basin, (4.7) southern Dead Sea basin, (4.8) northern Arava, (4.9) southern Arava; (5) eastern highlands: (5.1) Golan plateau, (5.2–5.5) Transjordanian plateau: (5.2) Gil'ad plateau, (5.3) Moav plateau, (5.4) Edom plateau, (5.5) endoreic basins of the southern Transjordanian plateau, (5.6) western slopes of the eastern highlands.

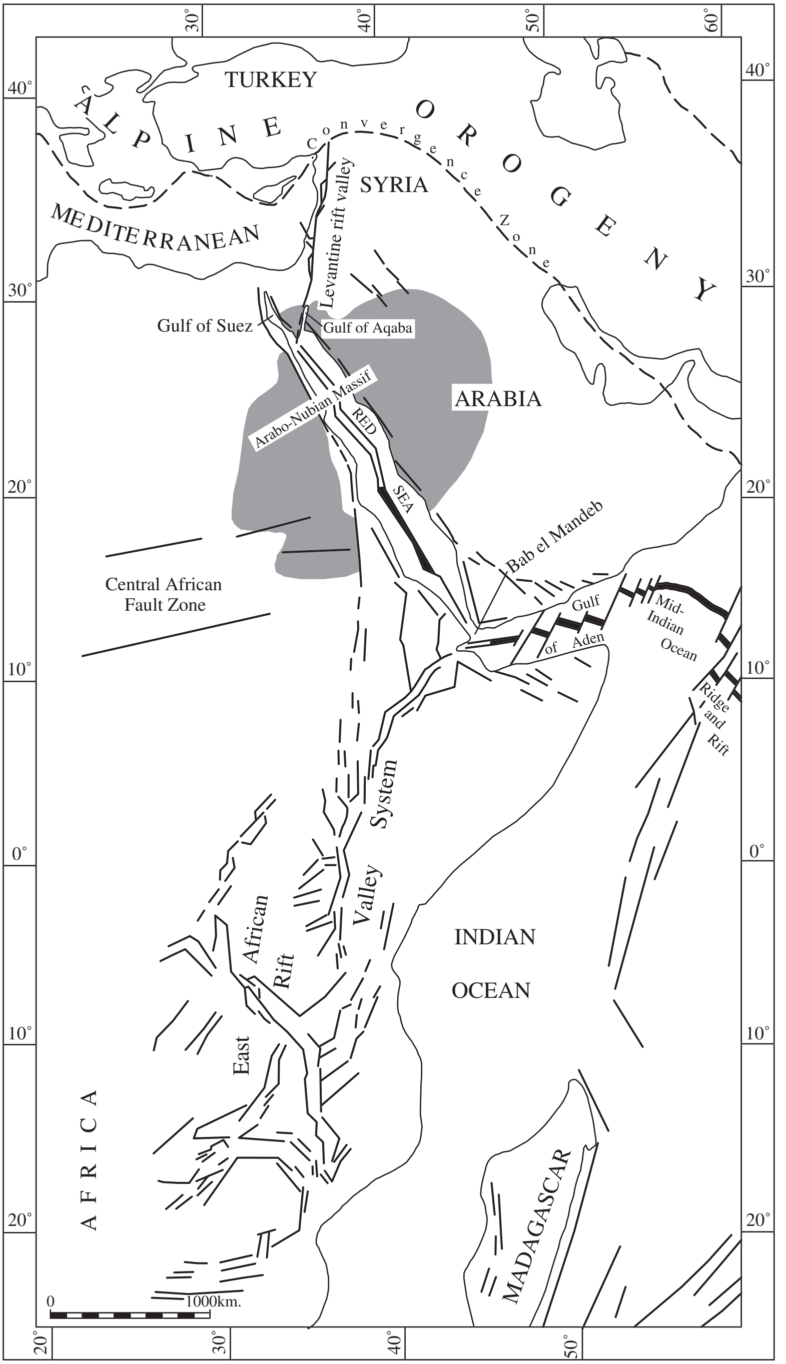


Figure 1.4. Schematic configuration of the Syrian–African Rift system.

The Jordan–Dead Sea depression is part of the Syrian–African Rift system (Fig. 1.4), running all the way from southern Africa through East Africa and the Red Sea to southern Turkey. As such, it had served as a passageway throughout time for people, fauna and flora from Africa to the Middle East, Asia, Europe (and possibly back), a process which involved (and still does!) struggles for food supplies and living space, which more often than not resulted in extinction of species on the one hand, while others consequently replaced them and flourished, at least for some time. Animals and plants fought over subsistence resources, while men fought for any possible reason, which they still do today.

This unique geographical location, a corridor between Africa, Europe and Asia for at least the last two million years, made the Jordan Rift Valley an important factor in the development, migration and settlement of people, animals and plants. These processes shaped the culture and economics of mankind, consequently resulting in the development of agriculture and modern way of life in this part of the world from where, it seems, they spread across the Earth.

Later in history, probably as early as Neolithic times, when seafaring and navigation were developed, the region became a connecting hub for commerce between the east and west, being the shortest land route for goods from the Red Sea and the Indian Ocean, through the Gulf of Aqaba to the Mediterranean coast. This, naturally, also gave rise to many clashes, wars and occupations of the region.

1.1.2 Geology

The combination of considerable rifting (the Dead Sea is the lowest point on Earth!), ensuing large-scale faults and the dry climate, which leaves most of the area barren of vegetation and poor in pedogenic processes, is a Godsent gift to geologists, who can study most of the rock formations without much hassle, since almost all are excellently exposed (see figures throughout this book, particularly in Chapters 5 and 11). The relatively easy access to the outcrops along the Rift and the leading wadis makes the region a paradise for studying its geology. Indeed most early expeditions to the southern Levant, particularly during the 19th century, commenced their research, however preliminary, in this area (Lynch 1849, Lartet 1869, Hull 1886 and many others mentioned in Chapter 2).

Rock units from the Precambrian onward (Figs 1.5 and 3.2.1–3.2.4) are exposed along the flanks of the Jordan Rift Valley. These represent the major stages in the geological history of the Near East, both pre-rifting and during the complex rift formation processes. The Precambrian basement rocks are exposed mainly on the eastern flanks of the Jordan Rift, along the southern end of the Dead Sea and almost continuously down to the Red Sea; on the western flanks these rocks crop out from the Elat region southward. The rocks, both magmatic and metamorphic, comprise the northernmost tip of the Arabo-Nubian Massif, surrounded by large-scale molasses which accompanied mountain building

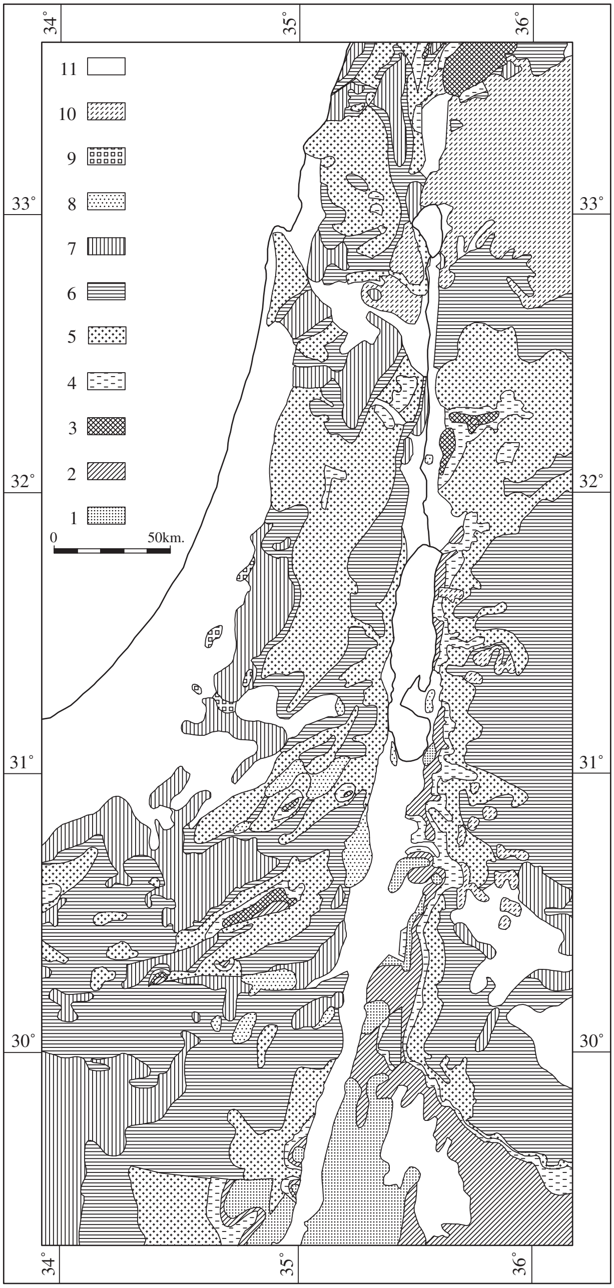


Figure 1.5. General geological map of the southern Levant: (1) Precambrian crystalline basement, (2) Paleozoic and early Mesozoic, mostly continental, (3) Jurassic, mostly shallow marine, (4) early Cretaceous, shallow marine and continental, (5) Cenomanian–Turonian, shallow marine, (6) Senonian–Maastrichtian, deeper marine, (7) Eocene–Oligocene, marine, (8) Neogene, continental, (9) Neogene, shallow marine, (10) Tertiary–Quaternary volcanics, (11) Quaternary, continental.

(Picard 1943, p. 43), representing the last stages of the Pan-African (late Algonkian) orogeny.

Peneplanation followed at the transition to the Paleozoic (Picard's "Lipalian Interval", 1943, p. 44), and its outcome can be seen especially along the eastern flanks of the Rift; this regional peneplain is covered by extensive sandstone layers of continental origin, the Nubian Sandstones, testifying that the country was generally above sea level, at least throughout the Paleozoic, Triassic, and intermittently during the Jurassic and early Cretaceous, comprising vast plains drained by huge rivers. The entire complex is generally referred to as the "Lower Clastic Division" (Shaw 1948).

The "Middle Calcareous Division" which follows, cropping out on both flanks of the Rift along almost its entire length, represents a shallow sea which submerged the Near East from the Cenomanian onward, until the Eocene, almost continuously. The lower part of this suite, the Judea Group, is mainly built of limestones and dolomites deposited in shallow, warm, usually littoral environments, very rich in mollusks and other megafossils. The middle part, the Mount Scopus Group of Senonian–Maastrichtian age, was deposited in a deeper sea, typified by chalks rich in planktonic foraminifera. During this time the region was gently folded, thus thicknesses are greater in the ensuing synclines and different facies appear on the higher, shallower anticlines. The folding continued also during the third stage of the "Middle Calcareous", accompanied by slow uplifting, resulting in deposition of the chalks and limestones of the Avedat Group, mainly in the synclines. For details see Chapter 4.

Further emergence of the land gave rise to sedimentation of the clastic units, mostly of continental origin, of the "Upper Clastic Division". These had been deposited in two major environments: rivers following the retreating sea, and lakes forming in places where subsidence was accelerated, mainly along the line of the future rift valley, from the Oligocene through the beginning of the Pleistocene. The Jordan Valley Rift was only formed in its present-day configuration, as an endoreic system, some two million years ago (Horowitz 1992a, Chapter 9). The rock units testifying to all these processes are found both in the deeper parts (known from boreholes) and around the Rift, cropping out. From the Oligocene onward volcanics are also known to have accompanied subsidence and later rifting, which became a great help for radiometric datings of the various evolutionary stages.

Sediments deposited within the Rift proper during the last two million years, of the Hula, Jordan and Dead Sea groups, comprise both lacustrine and fluviatile formations, occasionally accompanied by spring deposits and volcanics. These are mainly known from numerous boreholes drilled in the Rift, sometimes penetrating the entire rift valley fill to a depth of several kilometers (Horowitz 1997). Outcrops of these formations also occur on the flanks, however in much less developed form, in limited thicknesses and occasionally severely eroded. These, which crop out due to either high lake stands or subsequent structural disturbances, had served to reconstruct the history of the Rift well before borehole material was available (Picard 1943, Horowitz 1979).

The rare combination of all these features in a relatively small area, easy to get to, with excellent exposures, made the Jordan Rift Valley an almost ideal natural laboratory for geological studies ever since the beginning of the 19th century, when expeditions from various parts of the world (see Chapter 2) arrived in the region, continuing to the present day, resulting in a wealth of publications (Inbar et al. 1989, Arad & Bartov 1994, Qummou et al. 1997, Arad et al. 1997, 1998).

1.1.3 Environments

Situated on the boundary of the Saharan and Mediterranean regimes, the natural environments along the Jordan Rift Valley are extremely variable, ranging from Mediterranean to the north, with average annual rainfall of almost 500 mm, to extremely arid, Saharan, in the Arava south of the Dead Sea and down to the Gulf of Aqaba, with hardly any rainfall at all. A detailed description of the various natural environments follows in Chapter 3. This extremely steep environmental gradient persisted in the region at least for the last two million years. During this period the climate changed considerably by the alternation of dry (as at present) and humid periods, following the retreat and advent of glaciers in Europe (Horowitz 1992a, pp. 373–382). The gradient, however, remained steep for the entire time span.

The steep environmental gradient along the Jordan Valley is further accompanied by a wealth of local natural habitats along its flanks. These are affected by topography, relief and the direction the slopes face sun and rains, thus giving place to a variety of environments that, in turn, are also changing according to the environmental gradient along the region. Perennial water sources such as rivers, springs and lakes again add to the variety of biotopes as do, in a different manner, the intermittent wadis, only occasionally flooded by winter rains, or the saline environments.

Such a variety of natural environments in a relatively small area, where migrations of animals, plants and people are quite easy due to the short distances involved, gave rise to many different refuge niches. These provided shelter, water and food supplies even during the driest climate phases of the Quaternary, resulting in a richly varied fauna and flora which, in turn, enabled human settlement in the Jordan Valley almost continuously ever since our ancestors, probably in the form of *Homo erectus*, first arrived from Africa on the scene, approximately two million years ago (Horowitz 1979, p. 296; 1995; 1996b).

1.2 AN OCEANIC FEATURE ON LAND?

There is hardly any doubt that the Jordan Rift is a natural continuation of the Red Sea spreading system, an observation already made during the mid-19th century by almost all early travelers. As such, it is regarded by all students as an integral part of the Syrian–African Rift system. The clear and pronounced dissimilarity of rock formations on both flanks of the Rift is also reported by all investigators.

At about this point disagreement begins and a heated discussion as to the "true" tectonic nature and origin of this structural mega-lineament begins.

Early views (for details see Chapter 2) regarded the Jordan Rift as a typical continental feature, either as a rift ("sillon" of Lartet 1865), compressional "ramp valley" (Willis 1928), or as a tensional graben, induced by regional upwarping, forming a series of "troughs of disruption" which do not make a continuous rift valley (Picard 1943, p. 116). Other investigators, such as Dubertret (1932, but dismissing the idea later), Quennell (1959), Freund et al. (1970) and many others, suggested that the Jordan Rift is a major sinistral wrench fault, along which the eastern flank moved more than 100 km northward. This view also connected the Rift with the Red Sea system, regarding the Levantine rift as a major transform fault of the Red Sea mid-ocean type opening (Garfunkel 1981).

A different opinion on the tectonic connection of the Jordan Rift Valley with the Red Sea system, expressed by Horowitz (1979, pp. 60–62), Mart & Horowitz (1981) and others, views the Valley as a direct continuation of the Red Sea spreading center. These authors do not agree with the conclusion defining the Jordan Rift as a wrench fault, but rather return to Picard's (op. cit.) opinion of a series of discontinuous troughs, bordered by elevated shoulders (a detailed discussion of these opinions is advanced in Chapter 10, along with a modified model accounting for all the newly acquired data). The combination of an uplifted region dissected by a central rift system reminds one much more of a mid-ocean ridge and rift complex, rather than of a transform fault.

In both of these opinions, however different, the Jordan Valley is regarded as a continuation of the Red Sea embryonic ocean. The question only remains whether the Rift would indeed open sufficiently for the Red Sea waters to invade, as the Gulf of Aqaba already did. Naturally, even such invasion would not prove any of the parties' point, unless it is accompanied (or not) by a mid-ocean type ridge and rift complex and magmatic intrusion. Since the solution to this problem is not expected until several million years from now, it remains only a philosophical speculation. However, those who want to study, according to their belief, either a transform fault characterized by lateral displacement, or a spreading center typified by perpendicular opening, both representative oceanic features, can do so on land in the Jordan Rift Valley, with considerable convenience.

There is of course still another possibility which we have to face: Earth can still play us a trick and revert things altogether. This has already happened at least once with the Red Sea system: the Gulf of Suez is the natural straight ahead continuing lineament of the Red Sea, both spreading concurrently until the late Miocene. Ever since, the Gulf of Suez is being filled up with sediments, but shows hardly any significant subsidence. The Gulf of Aqaba and the Jordan Rift, northward up to southern Turkey, which had subsided together with the Red Sea and the Suez, continue to subside until the present day, at an accelerated rate. Should another change of direction or preference occur in the future, all our dreams of a new ocean will change. However, fortunately we do not have to worry about that right now.

1.3 A GATEWAY IN THE HISTORY OF HUMAN MIGRATION FROM AFRICA

The southern Levant, located at the junction of the three continents, Africa, Asia and Europe, had no other option but to be the gateway through which early Man left Africa to settle elsewhere on Earth, simply because no other continental route existed at that time, two to three million years ago (navigation at sea seems most unlikely at this stage of human evolution). Migration and geographical spreading became the second stage in human evolution, following the emergence of the genus *Homo* in Africa, a theory which is at present a common belief (although not fully proven). It is also generally considered that migration from Africa was not restricted to a single wave, but rather to a continuous, multiphase process of "sorties" (Bar-Yosef 1994, 1998).

The southern Levant was much flatter at the beginning of the Pleistocene (Horowitz 1979, p. 332), when migrations commenced. The route could thus have been anywhere in the country, not only the usually considered coastal plain and the Jordan Valley, but including also what now constitutes the highlands. The region had been traversed by a series of broad, shallow river valleys (op. cit., pp. 115–119) which did not seem to hinder human migrations, rather serving as subsistence suppliers. It is in all three present-day morphologic belts west of the Jordan River that the earliest artifacts were discovered (Horowitz 1979, p. 296; Bar-Yosef 1994; Ronen 1996), in layers estimated to be some two million years old, or even older. It is quite plausible that if Man had occupied the present-day highlands, the chances of finding evidence are quite slim, due to the subsequent erosion that continuously affected these elevating regions, during the entire Quaternary (Horowitz 1992b, 1995, 1996b), resulting in the absence of prospective strata.

Comparing the Pleistocene environments of the Levant with the Jordan Rift, it is quite certain that the latter provided better habitats for man. During drier periods the coastal plain was covered by dunes, offering a rather hostile environment for settlement and migration. Indeed, hardly any sites of dry periods are found in these eolian sediments prior to the Neolithic, in contrast with the wetter phases, during which the coastal plain, covered by soils and marshes constituting more favorable environments, was densely settled. The central hilly region must have been rocky and barren during the dry periods, forested during the humid, in both cases not the most comfortable environments. The rivers of the coastal plain flow in a general east–west direction, thus possibly somewhat obstructing migration, but were never really large ones.

The Jordan Valley, on the contrary, has a north–south drainage pattern, with only subordinate east–west components and no conspicuous topographic barriers. The climate was always warm, even during the colder phases, thus presenting Man, when arriving from Africa, with familiar environments. Freshwater was always there, all along the Valley, since the Jordan River and particularly the fault scarp springs, all the way down to the Gulf of Aqaba, were active even during the driest periods. It is therefore plausible to assume that the Jordan Valley was the preferred route for human migration, although surely some also took place along the coastal plain and through the central parts of the country.

1.4 TERMINOLOGY

Probably the first problem concerning the terminology of the Jordan Valley is how to refer to its structure. Numerous terms had been adopted by various authors through the years, the most common being "depression", "rift", "graben", "ramp" and quite recently "transform" (see also Fig. 2.1.3). Of these, the only objective, purely descriptive term is the first, which is no longer used. The trouble with all others is that in some way or another they hint a mode of formation for the depression, and hence are adopted by various students to express opinions concerning tectonic origin. Attitudes also change through time: Billings (1954) regarded "graben" as a result of gravity faulting (p. 203), while "rift" is a consequence of strike-slip movement (p. 219); Press & Siever (1998) practically dropped the use of "graben" altogether, referring to "rift valley" (p. 661) as "A fault trough formed at a divergent plate boundary or other area of tension". This definition partly eliminates the necessity to provide genetic explanations.

The *Dictionary of geology* (Himus 1954) differentiates between "graben" and "rift" by the angles of their bordering faults. The first is defined as "A structure resulting from the subsidence of a strip of country between two normal faults heading inwards", the second should only be longer, delimited by "roughly parallel faults". These definitions did not change much, and are repeated almost exactly in the *Glossary of geology* (Jackson 1997), seemingly justifying the use of the term "rift" for the Jordan Valley. Only seemingly, since, as will be discussed in Chapters 9 and 10, parts of the morphologic depression conveniently referred to as the "Jordan Rift Valley" are not bounded by faults at all, or occasionally only by a single one, usually on the eastern side. The complicated depression is actually made from a series of faulted troughs active during different periods, connected by monoclines against a fault, or by synclines. I shall nevertheless keep using "rift", as it is the most common term for the Jordan Valley, and is thus retained throughout this book just for the sake of convenience.

1.4.1 Transliteration of geographical names

Two main problems exist when transliterating Hebrew and Arabic geographical names into Latin characters. The first concerns spelling, which changed considerably through time, so that a place name may have been written differently in earlier and later publications and maps. This problem becomes more acute since the French and German languages were extensively used by earlier researchers, but also because some Hebrew and Arabic letters and pronunciations do not have an exact Latin parallel. As far as possible, use is made of the latest editions of maps of Israel and Transjordan. These, unfortunately, also occasionally differ in spelling a name of the same place; in this case, usually the Israeli version is adopted west of the Jordan River, while the Transjordanian is used to the east. Wherever deemed necessary, the different transliterations are mentioned. However, the reader should be prepared to confront some difficulties, especially when referring to earlier works.

The second difficulty arises from the fact that many locations in the Holy Land already have well-established English names, which are not necessarily an exact transliteration of either the Hebrew or the Arabic names. Such is for example "Jerusalem" for "Yerushalayim" (Hebrew), or "Ursulim" and "El Quds" (Arabic). Usually, in such cases use is made of the common English names, rather than the local ones. However, this is not always the case, as sometimes the Hebrew name or its partial, literal translation into English is preferred. The "Sea of Galilee", which is in fact a lake, is referred to here as "Lake Kinneret".

In some instances the preference is to translate into English a part of a name, especially when the meaning would be lost for the reader by using the local one. Thus use is made of "Mount" or "Hill" rather than "Har" (Hebrew) or "Jebel" (Arabic), "River" for "Nahal" or "Nahar" (Hebrew), but not for "Wadi" (Arabic) which is retained, being internationally accepted as a well-defined geographic term. Wherever necessary, literal explanations of names are provided. It is hoped that this will not cause too many difficulties. At any rate, nearly all locations mentioned in the text appear in one or more of the figures provided.

Wherever citations of previous publications are used, the transliteration of geographical names is given in the original way. When necessary, these are accompanied (in parentheses) by the present-day common names.

1.4.2 Stratigraphic terms

It seems that there is hardly any difficulty involved with the use of pre-Oligocene stratigraphic terms, for most of these are commonly accepted by the international geological community, thus disagreements are of a subordinate nature. We do, however, run into major difficulties in using Oligocene, Neogene and particularly Quaternary stratigraphic terminologies. The main reason for this is that many stages had initially been defined only as local ones, some even in continental deposits, which could not be followed over considerable distances. Due to frequent facies changes, no global, or even wide regional distribution of the Quaternary units could be ascertained. Correlations with marine and particularly ocean sequences, which can be recognized worldwide are still, in most cases, controversial, especially for the Quaternary.

Another difficulty comes from the use of radiometric datings and magnetostratigraphy, the latter also based on such "absolute" chronology. Both methods can only be applied to limited suites of rocks, which rarely occur in most crucial localities, nor are always intimately connected with continuous continental fossil bearing sequences. On top of that, various radiometric methods occasionally yield contradictory results, even for the same rock samples. K. Heine (Department of Geography, University of Regensburg 1997, *in letteris*) quite clearly demonstrated this problem, by using different techniques for dating samples from Dieprivier, Namibia. The uppermost sample yielded a radiocarbon age of 2,325 ± 150 years, TL age of 8.2 ± 0.6 Ka and U/Th age of 153 ± 20 Ka. A lower sample yielded, using the same methods, respectively, 10,850 ± 255 years, 13.5 ± 0.8 and 73 ± 6 Ka. The lowermost sample conveyed ages of 20,090 ± 680 years, 29.5 ± 1.8 and 62 ± 3 Ka. The radiocarbon ages are referred to by Heine as having been influenced by recrystallization of carbonate; the TL ages are of transport and accumulation, while the U/Th figures are ages of calcrete formation. These explanations could be acceptable for this specific case, but the trouble is that most sites are dated by a single method only, so that discrepancies, if they exist, are not always apparent. If the date is within reason, it is usually accepted for the "absolute" age of the site or bed. Similarly, Porat et al. (1999) discuss ages obtained for Lower Paleolithic sites in Israel, indicating that TL ages in the Tabun Cave in Mount Carmel are almost twice as old as ESR ages, which are themselves older than suggested stratigraphic assignments (Mercier et al. 1995).

Considering potassium–argon ages, which is a widespread technique these days (and on which paleomagnetism dating is based), the words of Chazot et al. (1998, p. 51), relating to volcanic events of the Red Sea, speak for themselves: "In the last twenty years much of the age dating of volcanic rocks has involved whole rock K–Ar techniques ... Many of the volcanic rocks have experienced secondary processes in the form of contamination with upper and lower crustal rocks and surface alteration with groundwater. All of these factors influence the K–Ar system so that the K–Ar 'dates' bear little or no relationship to the actual primary crystallization age of the rock in question". Instead, these authors suggest that $^{40}Ar/^{39}Ar$ are much more reliable, but continue: "Even if one uses the $^{40}Ar/^{39}Ar$ data to screen the K–Ar data, such that those samples affected by secondary processes are eliminated, the K–Ar data have little meaning. Furthermore, more than 75% of the K–Ar data are shown to be in error by 2–35 Ma". An example of these difficulties is detailed in Section 5.3.9, dealing with dating of the En Yahav dike.

For various reasons, discussed in brief below, even the exact absolute date for the Pliocene–Quaternary boundary is not yet well established, with opinions differing over more than a million years, spanning an interval from 2.8 to 1.4 Ma.

A local stratigraphic terminology, based on climatostratigraphy, was developed for the late Cenozoic of Israel (Horowitz & Horowitz 1985; 1990; Horowitz 1992a, pp. 343–345), encompassing the Neogene (Miocene and Pliocene) and the Quaternary. The Neogene sequences were correlated and dated by analyzing both pollen and planktonic foraminifera in the same borehole, penetrating the

Mediterranean offshore sequence (Horowitz & Derin 1987, Horowitz 1990); the Oligocene was added to the scheme later, when a deeper borehole was drilled in the southern Dead Sea (Horowitz 1996a). The entire late Cenozoic Levantine palynostratigraphic sequence is thus directly related to foraminifera zones, as defined by Blow (1969).

The Oligocene was originally defined, conforming with many other Series in the global stratigraphy, as a large-scale transgressive cycle, comprising the Sannoisian, Stampian and Aquitanian stages (Gignoux 1955, pp. 472, 474), with the highest sea level during the Stampian (whose upper part was called Chattian). Quite soon the Aquitanian was recognized as the basal part of the Miocene, the Sannoisian and the lower part of the Stampian became known as Rupelian, while the Chattian made the upper stage of the Oligocene. Consequently, the Rupelian is referred to as "early" and the Chattian as "late" Oligocene, with the "middle" usually lost.

A detailed study of foraminifera in Mediterranean marine Oligocene sediments in Israel (Martinotti 1981) resulted in a threefold subdivision of the sequence, which is also the nomenclature adopted here. The early Oligocene, comprising Foraminifera zones P18 and P19; the middle, encompassing P20 (N1) and P21 (N2); and the upper, P22 (N3). The maximum transgression occurred during the middle Oligocene, particularly its later part, corresponding to N2 (Fig. 9.1.1).

The Miocene is subdivided into three palynozones. Ma, which had already commenced in the late Oligocene, continuing into the early part of the early Miocene; it corresponds to the later part of the Chattian, Aquitanian and early Burdigalian, from approximately 28 Ma up to some 19–18 Ma ago, encompassing Foraminifera zones N3 (P22) through N6. Palynozone Mb, of late early through middle Miocene age, up to approximately 10 Ma ago, corresponds to the late Burdigalian, Langhian and Serravallian, Foraminifera zones N7 through N14. The late Miocene Palynozone Mc corresponds to the Tortonian and Messinian, lasting up to about 5.2 Ma ago, Foraminifera zones N15 through N17. In the present state of knowledge, palynostratigraphy does not allow further subdivision of the Miocene.

The Pliocene is subdivided into two distinct palynozones, each broadly synchronous with a transgressive cycle. Pa correlates with the Tabianian (or Zanclean) Stage of the early and middle Pliocene, that commenced some 5.2 Ma ago and ended at about 3.5 Ma, corresponding to Foraminifera zones N18, N19 and the lower part of N20. Palynozone Pb corresponds to the late Pliocene Piacenzian transgression, the upper part of Zone N20 and the entire N21, from some 3.5 Ma ago up to approximately 2.65 Ma.

The last figure is not accepted by many authors in Israel (just a few examples: 1.8 Ma in Goldsmith et al. 1988 although I am one of the authors; 1.4 Ma in Bartov 1993, p. 24; 1.6 Ma in Begin & Zilberman 1997, p. 46). The situation is also similar on a global scale (see discussion in Ehlers 1996, p. 3) and is summarized by Ehlers (p. 227): "Therefore at the moment two different lower boundaries of the Quaternary are in use, 1.6 million and 2.4 million years". Ehlers seems to prefer the "longer" Quaternary, but uses a figure of 2.6 Ma for the lower boundary (p. 448). It seems however that the differences in absolute ages for the lower boundaries of both the "long" (2.8–2.4 Ma) and "short" (1.8–1.4 Ma) Quaternary may result either from inaccuracies in the radiometric methods used for datings, or from misidentification of this boundary.

For several reasons, I prefer to use the "long" Quaternary, mainly because it appears that regardless what definition this epoch is based upon, most evidence seems to indicate a date approximately 2.7–2.6 Ma ago for the lower boundary. When considering the first major cooling as the Pliocene–Quaternary boundary in both the oceans and the continental northern hemisphere (Ehlers 1996, p. 448), this is quite clear. When taking the appearance of early hominids as a definition, the boundary becomes more problematic, since it may have to be stretched even beyond the 2.65 Ma figure (op. cit., p. 153). However, since the Quaternary, above all, should be considered a geologic-stratigraphic entity, it is my opinion that its definition should conform with most other chronostratigraphic units, namely being defined as a marine transgressive cycle (Horowitz 1979, p. 5; 1991). The Tertiary–Quaternary boundary should thus be set at the maximum of the regression separating the Tabianian from the Calabrian, some 2.65 Ma ago, which roughly corresponds also to the Brunhes–Matuyama boundary. This opinion seems to have established itself with many nowadays (see for instance Suc et al. 1997).

Incidentally, the reason for the existence of long and short "Quaternaries" may lie in the use of magnetostratigraphy. Both boundaries are at a transition from normal to reverse magnetic fields and could very well be mistaken for one another. This is even more likely, since the younger, normal Olduvai Event is not always recognizable; it has been demonstrated (Horowitz 1979, p. 349; 1982; 1992a, p. 344) that in a certain case (Darlymple 1972) more than half of the samples falling within the Olduvai time range are magnetically reversed. Even in some of the continuous cores taken from the bottom of the ocean the Olduvai is not always discernible (Eicher 1968, p. 80). Confusing the two paleomagnetic events could thus easily lead to two "Quaternaries".

The Quaternary of Israel is subdivided into ten palynozones, QI through QX, of which the oldest one is further divided into three sub-palynozones, QIa through QIc. I previously accepted West's (1969, pp. 219, 235) subdivision into "Preglacial" and "Glacial" Pleistocene; but for the sake of tradition and convenience I did not accept his rejection (p. 225) of the term "Holocene". Thus the Quaternary of Israel was subdivided into three main periods, the Preglacial Pleistocene, the Glacial Pleistocene and the Holocene. Evidently, these were used as stratigraphic, rather than climatic terms, and thus could be applied also in regions where no Quaternary glaciations are recorded, such as the Levant. The Preglacial Pleistocene palynozones, QI and QII, from approximately 2.65 Ma ago to approximately 1.8 Ma, correspond to Foraminifera Zone N22, while the Glacial Pleistocene and Holocene palynozones, QIII through QX,

Table 1.4.1. Stratigraphic nomenclature for the late Cenozoic of the southern Levant.

Stratigraphy	Age	Main foraminiferas used in Israel ( ) – benthonic	Foraminifera Zone	Palyno-zone	Climate in Israel	Alpine Glacials	Mediterranean stages
Holocene				QX	Dry Med.		Versilian
	11 Ka						
LATE CENOZOIC – QUATERNARY – PLEISTOCENE				QIX	Wet Med.	Würm	
	70 Ka						
				QVIII	Dry Med.	R/W	Monastirian
	130 Ka						
				QVII	Wet Med.	Riss	
	425 Ka		N23				
				QVI	Dry Med.	M/R	Tyrrhenian
	790 Ka						
				QV	Wet Med.	Mindel	
	1.25 Ma						
				QIV	Dry Med.	G/M	Milazzian
	1.49 Ma						
		Globorotalia		QIII	Wet Med.	Günz	
	1.82 Ma	*truncatulinoides*					
Earliest				QII	Dry Med?	D/G	Sicilian
	1.99 Ma						
				QIc	Wet, cool, temperate	Donau	
	2.17 Ma						
			N22	QIb	Dry, warm, temperate	B/D	Calabrian
	2.29 Ma						
		(*Hyalinea balthica*)		QIa	Wet, cool, temperate	Biber	
	2.63 Ma						
PLIOCENE Late		*Globorotalia inflata*	N21	Pb	Dry, cool, temperate		Piacenzian
	3.5 Ma						
Middle		*Gt. crassaformis*	N20				
Early		*Gt. puncticulata*	N19	Pa	Wet, cool, temperate		Tabianian
		Gt. margaritae	N18				
	5.2 Ma						

			Gt. conomiozea	N17		Dry, hot,	Messinian
	Late		*Gt. acostaensis*	N16	Mc	Saharan	
M			*Gt. menardii*	N15		desert	Tortonian
			(*Borelis melo*)				
I		10.2 Ma					
			Gt. mayeri	N14			Serravallian
O			*Gt. mayeri & druryi*	N13			
	Middle			Hiatus		Wet, sub-	
C			*Gt. peripheroronda*	N9	Mb	tropical	Langhian
			P. glomerosa &	N8			
E			*G. des sicanus*				
		16.2 Ma	(*Miogypsina*)	N8–N9			
N			*G. des trilobus*	N7			
							Burdigalian
E	Early		*G. ita stainforthi*	N6			
			G. ita dissimilis	N5			
			G. kugleri	N4			Aquitanian
		25.2 Ma			Ma	Dry, sub-	
O			(*Lepidocyclina*)			tropical	
L	Late		*G. ciperoensis*	P22 (N3)			Chattian
I							
G	Middle	30 Ma	*G. opima opima*	P21 (N2)			
O			*G. ampliapertura*	P20 (N1)			Rupelian
C							
E	Early		*G. cipolensis*	P18/P19			
N							
E							
		36 Ma					
Late			*G. cerroazulensis*	P16/P17			
Eocene			*G. semiinvoluta*	P15			Priabonian
		39.4 Ma					

G.: *Globigerina*; *Gt.*: *Globorotalia*; *P.*: *Praeorbulina*; *G. des*: *Globigerinoides*; *G. ita*: *Globigerinita*; Med.: Mediterranean. Planktonic foraminifera zones from Blow (1969), modified by B. Derin (Consulting and Geological Services Ltd. Ramat Gan 1997, pers. comm.) and Martinotti (1981). Correlation of palynozones and foraminifera zones from Horowitz & Derin (1987). Quaternary ages from correlation with ocean oxygen isotope curve, in Horowitz (1989c).

correspond to Foraminifera Zone N23, from approximately 1.8 Ma ago until the present day.

At present I am inclined to refrain from further use of West's terminology, mainly because it seems that it has not been generally accepted by the scientific community. I also refrain from using "early", "middle", and "late" Quaternary (or Pleistocene), since these terms had been already applied in so many different ways. The Quaternary is then simply subdivided into the Pleistocene (Palynozones QI–QIX) and the Holocene (Palynozone QX), while most age references are related to the palynozones. The stratigraphic terminology, as well as suggested correlations with the Alpine glacials and circum Mediterranean eustatic stages, are detailed in Table 1.4.1.

Dating of the Quaternary palynozones is based on continuous sequences covering the late Pliocene and the Quaternary, correlated on the basis of inferred climatostratigraphy with oxygen isotope curves obtained for deep sea cores from the northern Atlantic Ocean (Horowitz 1989c; and see Section 6.6 and Fig. 6.6.2). These correlations help in securing the local paleoenvironmental reconstructions within the internationally accepted oxygen isotope curves framework. Radiometric ages, however problematic occasionally (see Chapters 7 and 11), frequently help to date points in the sequences, more often than not supporting the dated palynostratigraphic scheme (Heimann 1990).

It should be noted that not all researchers agree with the terminology used here, especially regarding the correlation of humid Quaternary phases in the Levant with glacial advances in Europe (see for instance the discussions by Butzer, Farrand and Issar in Horowitz 1977b, pp. 87–100), or the 2.65 Ma age for the Neogene–Quaternary boundary. I do not maintain that the use of this terminology and boundary age is ultimately correct; however, it seems to work quite well and did so for the last 30 years in this part of the world (see also Besançon & Sanlaville 1984 for positive correlation of the humid phases of the entire Near East with the Mediterranean low sea levels of the glacials).

It should be noted that no better systematic subdivision or terminology was ever suggested for the entire Quaternary of Israel, except for the use of such terms as "Early" (or worse, "Lower"), "Middle" and "Late" ("Upper") Pleistocene, which had never been clearly defined, and thus could occasionally be highly misleading. Therefore, the terminology based on the late Cenozoic palynostratigraphy of Israel and its correlations with marine stages is retained here.

1.5 CONCEPTS AND AIMS OF THIS BOOK

Several books dealing with the geology of the Jordan Rift Valley, particularly the Dead Sea, had already been published ever since Louis Lartet's (1877) pioneering monumental work. As well, many theories concerning the structure and origin of this conspicuous feature had been suggested, dropped, and occasionally revived throughout the last two centuries (for details see Chapters 2 and 10). However, both books and theories are based on outcrops, hardly ever on information gathered from boreholes. Also, most previous studies deal with geology, geochemistry, tectonics and many other aspects of the region, neglecting the past environments or only discussing them in a secondary way, nor paying much attention to the timing of events and chronostratigraphy.

Previously, our knowledge of the Rift's stratigraphy, and especially its southern part, the Dead Sea, as known from outcrops, was quite limited. This resulted primarily from the fast subsidence of the region, burying older formations and causing rapid erosion of their few marginal remnants. Information gathered during the last 10–15 years, from deep boreholes, particularly detailed palynostratigraphy, accompanied by numerous geophysical studies, has significantly broadened our knowledge of the Rift Valley fill, enabling a much better understanding of its history and development. The synthesis presented is based primarily on this new information, together with other studies of the region in the last 20–25 years, naturally without disregarding previously published data.

The aim of this book is therefore to tie up the ends, and to present a synthesis of the environmental and structural evolution of the Jordan Rift Valley through time. In addition, this lineament is primarily of Quaternary age, a period during which humans had been almost continuously inhabiting the Near East (for details see Chapters 11 and 12). Prehistoric and archaeological sites are common within the Rift, affected throughout time by the changing environments. The relationships of environments and people are discussed, within the framework of settlement pattern, economy and migrations.

Concerning models and theories, it is my concept that each of these could turn out to be right, even when they may seem contradictory. It is therefore my aim to present all views in (as far as I can) a balanced way. Although, naturally, I prefer my own, I only hope that this preference will not affect the contents of this book too deeply. To avoid a completely subjective approach, an Appendix is included, presenting the opposing model, by Zvi Garfunkel. Theories presented by other colleagues in Chapters 4 and 8 are, naturally, not censored.

1.6 UNPUBLISHED SOURCES

Among writers and editors, there are two prevailing approaches when referring to unpublished material, confidential reports of oil and mineral companies, hard to get reports, dissertations and the like, or studies written in such languages as Hebrew or Arabic. One group maintains that if results are not publicly and easily available, critical reading is impossible, thus such studies are not to be referred to. The opposing view realizes that by ignoring these, one loses a wealth of information already gathered. The presentation of such information, by translation or otherwise, via a "secondary" source, may naturally be hazardous, chiefly because it may only be partially brought to the reader, affected by personal biases or different interpretations. Such information could of course be cross checked by a local reader, but hardly ever by the general public, which means that it is not entirely free of possible criticism.

Numerous such studies exist concerning the Jordan Rift Valley, for several reasons. In many instances, students who finish their formal studies, writing their dissertation or thesis, choose another profession or go to the industry or other places, where publications are not necessary for their promotion. Occasionally their mentor would publish some of the study, usually in a general way, frequently as a "secondary" source, with all the above mentioned limitations. Studies made for the industry are hardly ever published, and are frequently written in the local language. Fortunately, almost all such studies are available at the libraries of the Geological Survey of Israel, Jerusalem, and the Natural Resources Authority, Amman, both acting as national archives.

I did not see any choice but to use and refer to a variety of unpublished or hard to get sources in this book. I am assuming all responsibility for doing so, in the hope that the wealth of information thus advanced will indeed enhance knowledge.

CHAPTER 2

History of ideas and research

Considered sacred by the three main religious beliefs of the western world, the Holy Land has endlessly, during the last 2,000 years and more, been a zone of dispute over who rules. This also increased curiosity about its geography and natural history, among many other topics. Thus the Near East, and especially the Holy Land, Jerusalem and the Jordan River, have for ages been the pearl in the crown of many expeditions and travelers, for obvious religious reasons. However, geological observations had hardly been made until the last decades of the 18th century, mainly because this branch of science only commenced its development at approximately this time (unless one considers the biblical story about Sodom and Gomorra a geological note).

Exceptions, in the Middle Ages, are by several travelers who mentioned geological phenomena in their diaries, such as Joinville (ca. 1255), in his *Histoire de Saint Louis*, who described the fossil fish found in Syria (now Lebanon) as one of the marvels of this country. Pierre Belon, a naturalist from Mans, described in 1547 a few fossils from the road between Jerusalem and Bethlehem, but considered them to be seeds of some sort. J. Cotovicus described in 1619 several locations in the Jordan Valley. Maraldi, in 1703, officially informed the Paris Academy of Sciences of the presence of fossil fish in the Lebanon. These were later mentioned by others, such as Corneille le Brun (Cornelis de Bruyn) in 1714 and Jonas Korte in 1741. The Dead Sea itself is described by Büsching (1766). For further details see the meticulous accounts by Goren (1999, in press).

The history of geological and archaeological research in the Jordan Rift Valley, as well as almost any other scientific field in the Near East, depended almost primarily on developments, trends and changes in local and international politics, the general European penetration, the rediscovery of Palestine in the 19th century, and the special georeligious connotations of the region. Three major phases are discerned in this respect: up to the end of World War I the entire region was part of the Ottoman Empire, when all research activities in geology, archaeology and prehistory were carried out by foreign expeditions and individuals, or by experts invited by the Turkish authorities, notably French, German, British and American. The usual outcome of these expeditions were travel books, encompassing almost every aspect of the region, with very little data or detailed studies of any particular aspect. As an example, 24 books dealing with the Dead Sea were published before 1918, while there were only ten from then until the present day (Arad et al. 1998). Exceptions to this are the detailed studies by Anderson, Lartet, Hull and Blanckenhorn, in both geology and prehistory. The "Palestine Exploration Fund" was founded in London in 1865, for the purpose of "elucidating and illustrating the Bible", giving considerable support to studies of the Holy Land, followed by the foundation of the German Society for the Study of Palestine, in 1877.

Under the Sykes-Picot agreement of 1916 and the completion of occupation in 1918, the Near East was divided between France, which had acquired a mandate over Syria (including Lebanon), while Palestine (including Transjordan) was assigned to Great Britain. The British pledged support in the Balfour Declaration of 1917 for a Jewish Homeland in Palestine, west of the Jordan River, while the area to the east became, in 1923, the independent Kingdom of Jordan, under British protection. The Jordan Rift Valley separated the two political entities, but both remained under British influence until 1948. This second stage mainly involved geological research oriented toward economic goals, primarily the search for oil, groundwater and minerals. It was carried out by British or British-directed companies such as the Iraq Petroleum Co. (IPC, Palestine) and others, resulting in a wealth of information, usually unpublished reports. The early 1930s saw also the development of geological and prehistoric studies by scientists who lived in the region, both government officials and local university people.

The third stage, following termination of the British mandate in 1948, saw in both Israel and the Kingdom of Jordan an accelerated development of universities, government institutions such as geological and archaeological surveys, as well as locally based oil, groundwater and mineral companies. Research in the Jordan Rift Valley became the goal of local scientists, who could invest much more time and effort than visiting expeditions. As a result, this phase is typified by a considerable body of data from the region. Numerous studies have been published as papers in journals, rather than books or reports, during the last two stages. Just as an example, the stratigraphic lexicon of Picard (1938) holds 102 entries, the one by Bentor (1960) contains 354, and a recent edition by Hirsch & Roded (in prep.) includes close to 1,000.

2.1 GEOLOGY

Geological studies of the Jordan Rift Valley have gone through three phases. The first involved travelers and expeditions from many countries who visited the region for a short while, whose reports (with the exception of Russegger, Anderson, Lartet and Hull) had usually been superficial, touching upon many topics but merely in a general way.

The second phase began with investigators who came to the region for a longer stay, at the invitation of the relevant authorities. The German geologist and paleontologist Max Blanckenhorn was invited toward the end of the 19th century, by the Turkish Sultan Abdul Hamid II, to conduct geological studies of Syria and Palestine. His numerous publications (almost 80, including several books and geological maps) cover the period from 1889 to 1927. Soon after the British army had taken the region from the Turks, in the 1920s, the main interests became groundwater, minerals and oil, and studies were carried out by state geologists and the IPC. Most of these investigations were summarized by Ionides & Blake (1940) in a report on *Water resources, soils, minerals and geology of Transjordan*, by Blake & Goldschmidt (1947), in a book on the *Geology and water resources of Palestine*, by Wetzel et al. (1947) and by Ionides (1951), who dealt specifically with the Jordan Valley.

The third phase actually commenced in 1924, with the pioneering studies of Leo Picard, who arrived in Palestine at this time. His first publications concerning the Jordan Rift Valley date to the early 1930s, and ever since he has been the leading figure in geological research in Israel and especially the Jordan Rift Valley, which retained his attention until his untimely death at the age of 97, on April 4th, 1997. In contrast with the preceding stages, investigations were conducted by scientists who lived in the country, which resulted in a wealth of meticulous, detailed, continuous studies. Picard had founded the Geological Laboratory in 1932, which later developed into the Department of Geology, at the Hebrew University of Jerusalem. After 1948, when Israel became independent, Picard also founded the Geological Survey of Israel, which has since been involved with research in the Jordan Rift Valley. The search for groundwater and oil resulted in extended investigations of the Valley by Tahal (Water Planning for Israel) and by several oil companies, accompanied by stratigraphical, geophysical and geochemical studies which broadened our views.

In Transjordan, besides the Amman and Yarmouk Universities, several research institutions with economic goals were founded, such as the Natural Resources Authority, the Transjordan Petroleum Co. and others. More specifically, the Jordan Valley Authority was founded to coordinate, among other things, research in this region. A synthesis of the geology of Transjordan, with considerable reference to the Jordan Rift Valley and its mode of formation, was presented by Bender (1968a and further publications).

Scientists in other universities in Israel also became interested in the Jordan Rift Valley. The international community shows great interest in the Rift, which resulted in much cooperation in the last years. Similar trends are also seen in the Kingdom of Jordan, with an increasing number of publications (Qummou et al. 1997). These trends culminated following the peace treaty between the two states, and nowadays cooperation among earth scientists gradually increases, to the benefit of both parties.

In terms of techniques, strategies and ideas, it is quite difficult to subdivide the geological research of the Jordan Valley into definite stages. Techniques have naturally improved over time and new ones have been adopted, almost as soon as they emerged. The one significant turn in research commenced in the 1950s, when deep boreholes and geophysical investigations revealed the wealth of the subsurface, showing that what we knew before about the Rift was only the tip of an iceberg.

Ideas, on the other hand, always ran wild, usually preceding the acquisition of hard data, as seems to be the universal way with geologists. As an example, strange as it may sound, several theories had been proposed for the origin of the Jordan Rift Valley even before any geological study or mapping was done there, some by people who never even visited the region.

Over the years, it was always the Dead Sea that attracted most attention from researchers, and this is still true today. The reasons are quite obvious, as the Dead Sea is the lowest point, and the most saline lake, on Earth. Add to it the biblical stories, the historical references to its strange asphalt emanations, the abundant sulfur, and one realizes why the rest of the Jordan Rift Valley seems neglected, especially by the earlier investigators.

A similar situation applies to the Lisan Formation (Figs 2.2.2, 3.5.1, 9.6.4, 11.2.7, 11.2.8, 11.3.1 and 11.4.1). Since this is one of the youngest Quaternary formations, and by far the most extensively cropping out, it received much more attention than it deserved from its place in stratigraphy (219 entries in the Geological Survey of Israel library at the end of 1997!). The Quaternary lasted for some 2.5 million years, and the Lisan represents only less than 50,000 years but is nevertheless the subject of most studies in the Jordan Valley until today.

Locations mentioned in this account are marked on Fig. 2.1.1.

2.1.1 Stratigraphy and structure

It is quite difficult to find out exactly where the first geological reference to the Jordan Rift Valley was made. It seems that what could be considered one of the initial observations was by Richard Pococke (1754), who visited the region in 1738 and mentioned fossil mollusks from the vicinity of the Dead Sea. Probably the first theory to explain the formation of the Jordan Rift Valley (with hardly any data to support it) was presented by François Chasseboeuf de Volney (1825, first published in 1787), who visited Egypt and Syria (which included Palestine at that time) in 1783 and 1785. He had suggested that the Jordan Valley and the Dead Sea had been formed as a result of the violent collapse of a region, which in ancient times led to the Mediterranean Sea. At approximately the same time Büsching (1766), also reported the Dead Sea's peculiarities.

Probably the first geological map of the Near East (or, as defined by the author, "mineralogical map") is by Guettard (1751, and see discussion in Ya'alon 1991). It took almost 100 years until a second, much more detailed map was published by Russegger (1847, and see discussion in Ginzbourg 1982).

The first years of the 19th century saw three eminent travelers to the Near East who also included the Jordan Rift Valley in their travels. U.J. Seetzen in 1806,

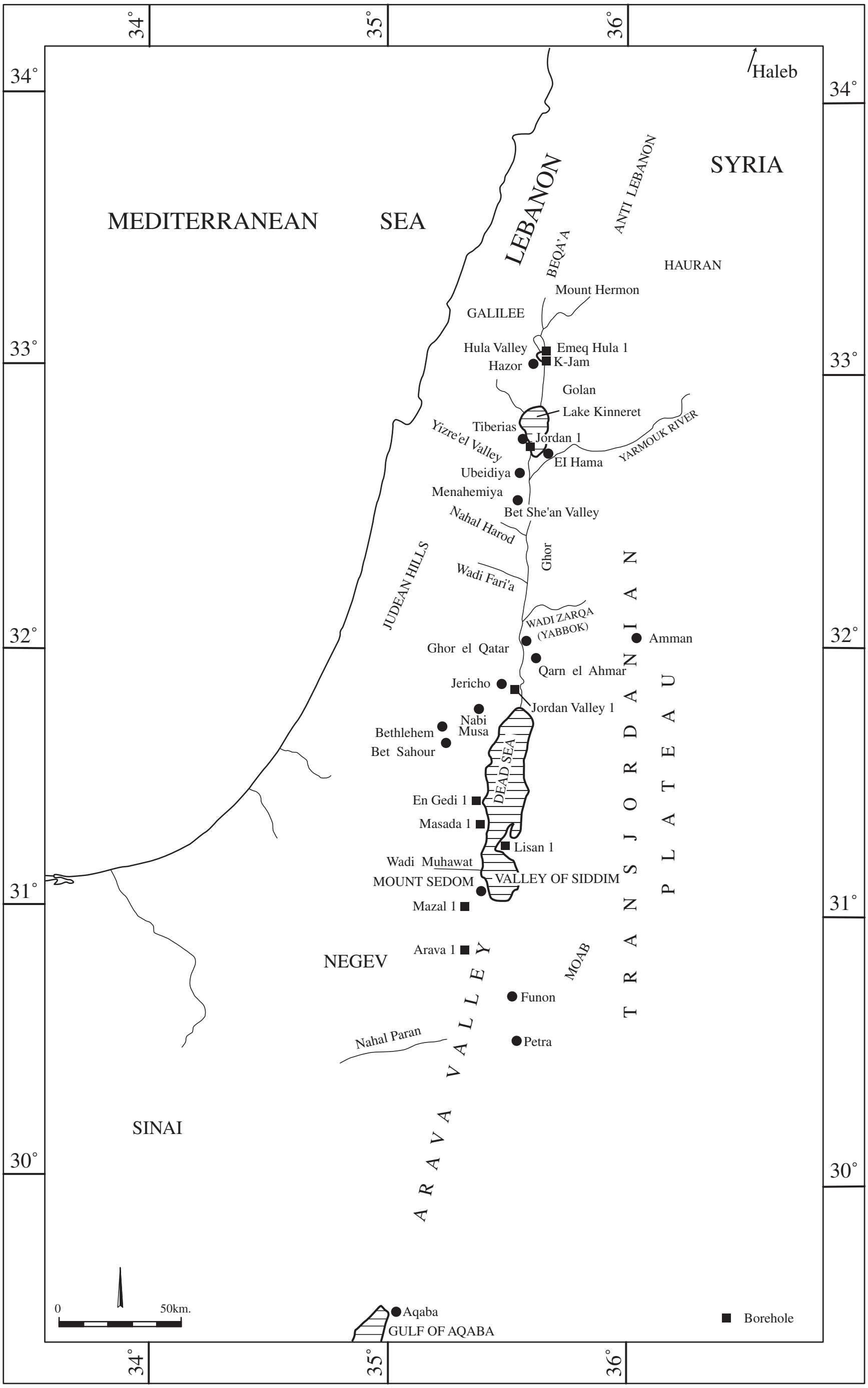


Figure 2.1.1. Location map of geological sites mentioned in this chapter.

Domingo Badia (alias *Ali Bey*) in 1807 and F. Burckhardt (*Sheik Ibrahim*) in 1810. The three of them (disguised as Arabs, for safety) made numerous geological observations, and may well be considered pioneers of (what was then) modern research in this field, as well as in other branches of natural history. Burckhardt (1822) described the Arava Valley, and thought it to have been the ancient channel of the Jordan River, formed solely by erosion, leading to the Gulf of Aqaba before the formation of the Dead Sea. It should be remembered that both Seetzen (1810) and Burckhardt thought the Dead Sea was above sea level, which made the "Long Jordan" (or the "River") theory quite plausible.

Two officers of the British Navy, Irby & Mangles (1868) visited the Near East during 1817–1818, including the regions situated east of the Ghor (the flat area of the Jordan Valley between Lake Kinneret and the Dead Sea, formed by the upper surface of the Lisan Formation) in the Jordan Valley, this time (probably the first) dressed as Europeans. They discovered megalithic remains, described in more detail in Section 2.2.

Léon de Laborde (1836) studied the topography between Aqaba and Petra in 1828, expressing the opinion that the Jordan's course to the Gulf of Aqaba was interrupted by uplifting, which provoked the destruction of Sodom and Gomorra, consequently forming the Dead Sea. Letronne (1839), who studied the hydrography of these regions in 1835, showed the existence of an elevated twin anticlinal ridge within the central Arava, which could not possibly allow the passage of the Jordan waters southward, to the Gulf of Aqaba. In spite of his protests, however, the opposite view was held until Louis Lartet, some 30 years later, following detailed geological observations, proved that Letronne was right (see below). At the same time Callier (1839) also described the depression of the Dead Sea.

A new phase of detailed geological research, including cross sections and maps, commenced in the Near East with the studies of E. Botta (1833) in the Lebanon. Shortly after, in 1837, two German naturalists, Schubert & Roth (1838/1839), visited Palestine and made numerous geological observations, including the basalt flows of the north (already mentioned by Badia).

The discovery which provoked the highest sensation concerning the Jordan Rift Valley was undoubtedly its elevation, below sea level. Moore & Beke (1837), while boiling water for their tea, noted the water temperature exceeded 212°F. They calculated that the Dead Sea lay 178 m below ocean level. Several weeks later, without knowing this, Schubert & Roth (1838/1839), considering barometric readings, found a figure of 93 toises (184 m) below ocean level. In 1838, J. Russegger (1841) and J. de Bertou (1838), both using a barometer, calculated an elevation of some 420 m below ocean level, a figure also arrived at (approximately) by Symonds (1841), using trigonometry. Other measurements followed, but the figure of 392 m below ocean level obtained by L. Vignes (1864), a member of Duc de Luynes' expedition, remained valid from then on.

Bertou (1839) studied the entire length of the Arava, confirming the observations by Letronne that its middle part is some 160 m above sea level. As Lartet

(1877, p. 14) puts it: “Ainsi tombait la théorie de l’ancienne prolongation du Jourdain jusqu’a la Mer Rouge”. At the same time a new hypothesis emerged, which attempted to explain both the negative elevation and extreme salinity of the Dead Sea, a theory which presumed an ancient connection of the Red and Dead seas, separated by consequent uplifting (Lartet 1877, p. 14, but he does not specify whose theory it was).

E. Hitchcock (1843) summarized the principal geological observations made by various scholars in western Asia, presenting them to the American Geological Association. He recognized the existence of a fault going all the way from Aqaba to the Anti Lebanon, and maybe up to Haleb (Syria), and saw this as an explanation for the formation of the Rift, following the opinion of von Buch (1841). This view was held by many at the time, and was confirmed by Lartet’s studies (1865). Hitchcock rejected Russegger’s (1841) ideas that the Dead Sea was a crater-like depression, following eruptions, but suggested (p. 369) that Mount Sedom was a volcanic product. Hitchcock also (“malheureusement”, according to Lartet 1877, p. 15) tried to explain the formation of the Dead Sea, attributing it to a disastrous breakdown. This collapse caused, simultaneously, uplift of the desolate Mount Sedom, made of rocksalt and submersion, and on the other hand, of the fertile Valley of Siddim, which is now the southern, shallow part of the Dead Sea (although some scholars now think it is the northern end of the lake). Finally, burning of bituminous sources completed this great catastrophe, in the way described in the Bible. Hitchcock did not accept, nor had Robinson & Smith (1841), and later on Gaillardot (1849), the theory that the burning of bitumen and bituminous limestones (common around the Dead Sea) was itself the sole cause of this remarkable collapse.

Tristram (1865a, p. 328) thought that Mount Sedom was “elevated from below”, an idea dismissed altogether by Hudleston (1885, p. 33), based on the supposed Cretaceous age and cross sections by Lartet (1869). Montague’s (1849) view was that the Dead Sea “lake” did not exist prior to the destruction of Sodom and Gomorra, and that up to this time the Jordan continued its southerly course toward the Gulf of Aqaba. Almost all researchers, by the middle of the 19th century, recognized that the strata east of the Valley are elevated by faulting in comparison to those to the west (cf. Hitchcock 1843, Lartet 1865, Tristram 1865a,b, Hull 1886).

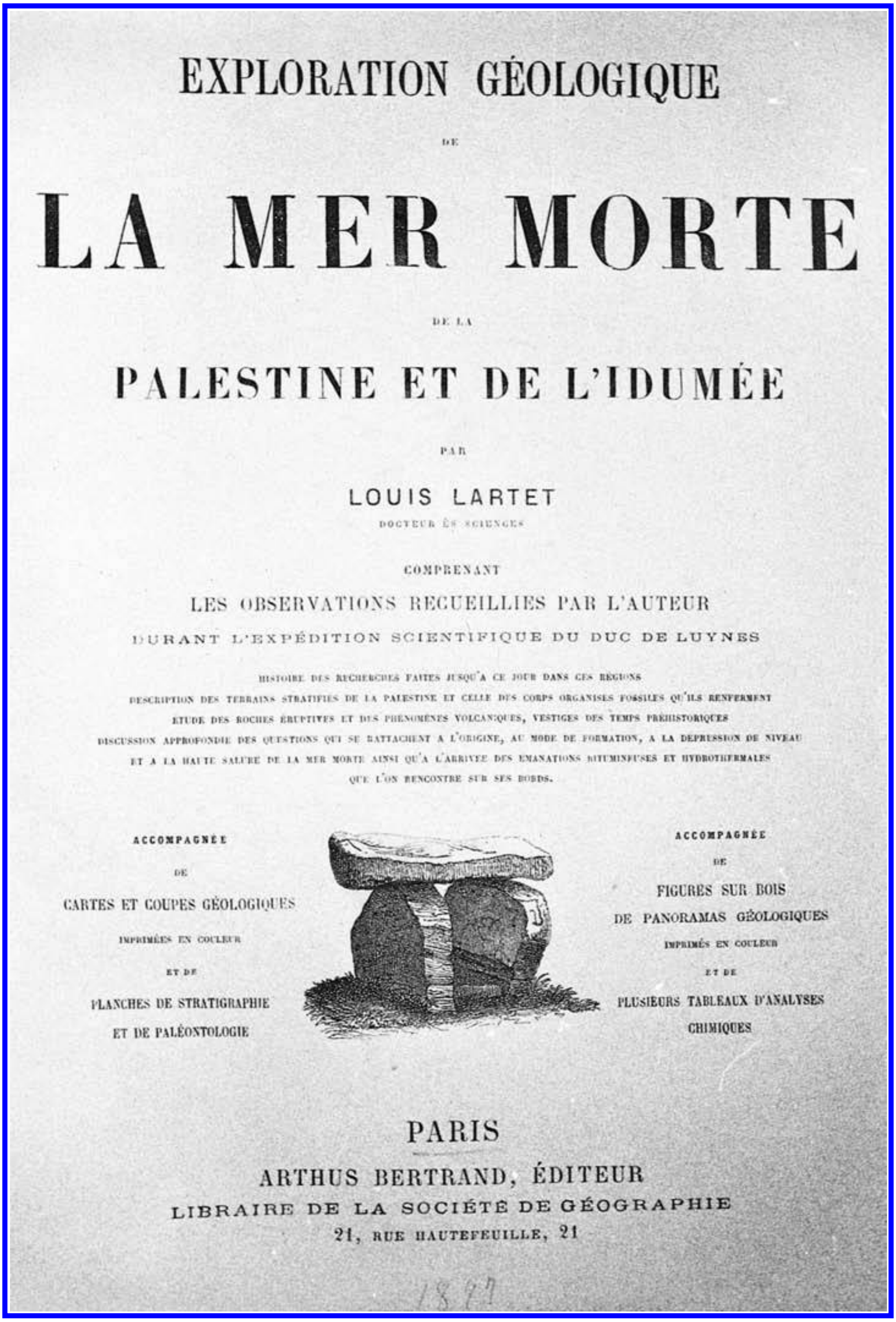
A British team headed by A. Molyneux (1848) visited the Jordan River and the Dead Sea in 1847. An American expedition of Lieutenant Lynch (1849) visited the region in 1848 with a geologist, H.J. Anderson (1852). To the dismay of Lartet (1877, p. 15), Anderson’s report was accompanied neither by cross sections nor by geological maps. On the other hand, there are numerous chemical analyses of rocks collected on the itinerary. Anderson (1852, pp. 205–206), who was very concerned with the formation processes of the Jordan Valley, believed that the great depth of the Dead Sea resulted from extended evaporation under the “tropical sun”. He presented two hypotheses “neither of them very promising: the first assumes a fissure or a series of fissures to have taken place in one of the early geological eras, and unequal denudations or partial accumulations subsequent to these to have imparted gradually the present relief". This hypothesis is basically no different from ideas previously expressed by Leopold von Buch and Hitchcock.

The second "supposes the *carina* (keel) to have once been one of a continuous descent, so that no lakes in the course of the river existed at that time. The disleveling agencies have then, in the course of geological ages, disturbed the line of the *carina* by their effects on the whole embracing platform, depressing and uplifting its various sections according to the laws of the disturbing forces. The depressed areas become the cavities which hold lakes, the elevated portions separate these completely or not, according to the depth of the connecting valleys and the still deeper subsequent excavation".

Anderson (op. cit.) further maintains that "the fissure is favored in the Jordan case, by the volcanic character of some of the neighboring districts, by the straightness of the Ghôr and by that diversity of structure, on opposite sides of the Valley, which might be expected where a fissure is modified into a fault. On the other hand it is liable to the serious objection that it cannot explain the origin of curvilinearly winding valleys, which are far more numerous than the rectilinear excavations". He brings many other objections to the second theory, the strongest being the fact that all wadis south of the Dead Sea flow northward, with no southward drainage, as would be expected if they led to the Gulf of Aqaba.

Following Lynch, a French expedition visited the Dead Sea in 1850 and 1851, headed by F. de Saulcy (1854). The reports by this expedition aroused more interest in France, and it was followed by another, led by Duc de Luynes, in 1864. Louis Lartet was the geologist, and his results were presented for the first time at the meeting of the French Geological Society in 1865, in Paris. Consequently, he published the first general, concise study on the geology of Palestine and the neighboring countries, Egypt and Arabia (1869), which served for a very long time as the basis for understanding their development. The monumental synthesis of his detailed work on the Dead Sea, including also the paleontology and prehistory of the region, accompanied by cross sections and a geologic-bathymetric map, finally appeared in 1877 (Fig. 2.1.2).

Lartet (1865, pp. 442–448) first presented the basics of his theory of the formation of the Jordan Rift Valley to the French Geological Society. He (p. 448) was probably the first to use the term "rift" (sillon) for the Jordan Valley. His thoughts were somewhat expanded and elaborated in 1877. The order of events, according to Lartet (1865, p. 447) calls for the eastern, main fault, to have already been active before the Cretaceous, to explain the dissimilarities of both flanks of the valley. Basalts (porphyres) had erupted through this fault system. The land on both sides of the valley emerged from under the Tertiary sea, and the great valley was formed, soon to be detached from the ocean. This left a series of lakes connected by the Jordan River, and also accounts for the salinity of the terminal basin, the Dead Sea. Earthquakes were responsible for the formation of hot



EXPLORATION GÉOLOGIQUE

DE

LA MER MORTE

DE LA

PALESTINE ET DE L'IDUMÉE

PAR

LOUIS LARTET

DOCTEUR ÈS SCIENCES

COMPRENANT

LES OBSERVATIONS RECUEILLIES PAR L'AUTEUR

DURANT L'EXPÉDITION SCIENTIFIQUE DU DUC DE LUYNES

HISTOIRE DES RECHERCHES FAITES JUSQU'A CE JOUR DANS CES RÉGIONS
DESCRIPTION DES TERRAINS STRATIFIÉS DE LA PALESTINE ET CELLE DES CORPS ORGANISÉS FOSSILES QU'ILS RENFERMENT
ÉTUDE DES ROCHES ÉRUPTIVES ET DES PHÉNOMÈNES VOLCANIQUES, VESTIGES DES TEMPS PRÉHISTORIQUES
DISCUSSION APPROFONDIE DES QUESTIONS QUI SE RATTACHENT A L'ORIGINE, AU MODE DE FORMATION, A LA DÉPRESSION DE NIVEAU
ET A LA HAUTE SALURE DE LA MER MORTE AINSI QU'A L'ARRIVÉE DES ÉMANATIONS BITUMINEUSES ET HYDROTHERMALES
QUE L'ON RENCONTRE SUR SES BORDS.

ACCOMPAGNÉE

DE

CARTES ET COUPES GÉOLOGIQUES

IMPRIMÉES EN COULEUR

ET DE

PLANCHES DE STRATIGRAPHIE

ET DE PALÉONTOLOGIE

ACCOMPAGNÉE

DE

FIGURES SUR BOIS

DE PANORAMAS GÉOLOGIQUES

IMPRIMÉS EN COULEUR

ET DE

PLUSIEURS TABLEAUX D'ANALYSES

CHIMIQUES

PARIS

ARTHUS BERTRAND, ÉDITEUR

LIBRAIRE DE LA SOCIÉTÉ DE GÉOGRAPHIE

21, RUE HAUTEFEUILLE, 21

1877

Figure 2.1.2. The first book dedicated to the geology of the Dead Sea, by Louis Lartet (1877).

springs, and emanations of bitumen, or asphalt. Lartet (1877, pp. 166–167, Plate III) had correctly identified the eastern large-scale faulting of the Jordan Rift Valley, along the Ghor, and thought it to be the main fault in the region, as was also previously hypothesized by von Buch, Hitchcock and others (but had not been clearly demonstrated).

Lartet maintains (p. 174) that the absence, in the Arava, of any late Cenozoic formations of marine origin proves that this region was uplifted before the depression of the Dead Sea and the Jordan Valley was formed. He therefore concludes that the Dead Sea seems to have never been connected with the ocean. However, its water level during the Lisan Lake (Fig. 2.2.2) times was considerably higher than today. Lartet also claims that the salinity of the Dead Sea is the combined result of rocksalt washed from the surrounding terrain (especially Mount Sedom, which he thought to be of Cretaceous age), and the considerable evaporation in the area. The absence of basaltic fragments and bitumen in the Lisan Formation led Lartet to believe (pp. 178–179) that this lake had existed before the volcanic activity in the Hauran and the Golan, for quite a long time.

Lartet (1877, pp. 259–260) stresses the considerable difference between rock formations on the eastern and western cliffs bounding the Dead Sea. The western comprises mainly Cretaceous and some Tertiary limestones, while the eastern consists primarily of Nubian Sandstones. He therefore concluded that a great fault ("une ligne grande de fracture") separates the two. This led him to think that the pre-faulting formations are buried at least 300 m below the bottom of the Dead Sea. The faulting is greater to the east of the Dead Sea, and since the Judean Hills are only gently folded, the Cretaceous formations should underlie the Dead Sea basin (1877, Plate III).

Liman Coleman (1867) developed and presented the idea that the Red Sea, the Gulf of Aqaba and the Jordan Valley are all parts of the same "fissure", which runs from Bab el Mandeb all the way through to the Lebanon. Coleman referred to the "graben" as a "fissure, a crevasse" formed by some "stupendous convulsion". Fraas (1867) had made detailed paleontological investigations around the Dead Sea, also describing (1880) the occurrences of sulfur in this region. Reclus (1872), an important French geographer who had not actually visited the region, following earlier reports, considered the Dead Sea "the most curious" of all inland seas. She also supposed that because the Dead Sea waters do not contain any iodine, it never constituted a part of the ocean, although the contrary hypothesis had been previously expressed.

Tristram (1884), who was mainly involved with studying the fauna and flora of Palestine, had supposed that the Dead Sea valley was once an arm of the Gulf of Aqaba, based on his observations that forms of life in the Dead Sea area are quite unlike creatures in the surrounding region, but bear a strong affinity to the Ethiopian domain, with a trace of Indian admixture.

Diener (1885) studied the Hula basin and southern Lebanon, presenting detailed geological and structural maps of this region. He maintained that north of the Jordan Valley the rift structure disappears by "virgation". E. Hull, a British geologist sent by the Palestine Exploration Fund (1886, p. 103), states that the vast region in which the Near East is located was submerged by the ocean waters throughout the Cretaceous and Tertiary. "It may be stated in general terms, that during the Miocene epoch, the general outlines and areas of land-sea were roughly marked out and determined; and, with more or less modification, have remained as such down to the present day". Hull was the first to indicate a direct connection between the uplifting of the mountainous ranges of the Near East and the formation of the "Jordan–Arabah (Arava) Valley".

Hull (p. 104) claims "that this deep depression is the direct result of a 'fault' or fissure of the crust, accompanied by displacement of the strata, the relations of the formations on opposite sides leave no room for doubt". This fact had been recognized by Leopold von Buch (1841), Hitchcock (1843), Smyth (1868), Tristram (1865b) and Lartet (1865, 1877), with whose views Hull's are "in accord". Hull again mentions the "general dissimilarity in the geological structure of the opposite sides of the Jordan–Arabah Valley throughout the greater part of its extent". Hull emphasizes that "this dissimilarity of the strata does not appear, as along the plains of Jericho and the Sea of Gennesareth (Lake Kinneret)". Hull indicates (p. 105) that "there is one leading and continuous line of dislocation along the Jordan–Arabah depression". This line is accompanied by numerous smaller faults. Hull again stresses (pp. 107–108) the horizontal strata on both sides of the Valley, extending far to the west, in Sinai, and to the east, in Arabia. He maintains (p. 108) that "the movements in the Earth's crust, causing the depression of the rift on one hand and the uplift of the mountains on the other, were not spasmodic or cataclysmal. It is far more probable that they were gradual, extending over a long period of time".

Hull (p. 108) is "disposed to think that the fracture of the Jordan–Arabah Valley, the elevation of the tableland of Edom and Moab on the east, and of Palestine on the west, were all the outcome of simultaneous operations and due to similar causes, namely, the tangential pressure of the Earth's crust due to contraction (being in its term due to the secular cooling of the crust) … Owing to powerful lateral pressure, acting from east and west directions, the strata over the whole area now under consideration were forced into a series of synclinal and anticlinal curves, at right angles to the direction of the pressure … (p. 109): The land area was gradually rising out of the sea, the tablelands of Judæa and of Arabia were more and more elevated, while the crust fell in along the western side of the Jordan–Arabah fault; and this seems to have been accompanied by much crumpling and fissuring of the strata … It is probable that the first waters of the Salt (Dead) Sea were those remaining from the ocean itself. The fauna of the Lake of Tiberias (Kinneret) may thus have had in the main a marine origin … Confining our attention now to the trough of the Salt Sea, we may suppose that as the lands on either side rose, the bed of the trough, owing to continued subsidence, became more deep over the area now occupied by the waters of the Salt Sea itself;

and into this gulf all the waters flowing from the bordering lands would naturally empty themselves".

Hull was thus the first to suggest compression for the origin of the Jordan Rift Valley, a theory that was further elaborated almost 50 years later by Willis (1928, see below), partly revived here (see Section 10.4). It is quite interesting to note that Hull, in a lecture delivered in Dublin in 1883, before visiting the Jordan Valley, strongly supported the "River Theory", that the Jordan flowed to the Red Sea (see discussion in Hudleston 1885, p. 62ff.).

Hull maintains that the drainage system, which led to the ocean before the Miocene (pp. 111–112), was divided, following the uplift at that period, into two systems, one which continued to flow through part of the earlier channels, the other forming toward the newly born basin. Hull also thought that, concurrently, the Mediterranean was formed in its present-day configuration.

Hull also gives a detailed account of his theory about the origin of the faunas of the Hula and Kinneret lakes and the Jordan River, based on observations by Roth (1858), Lortet (1880) and Tristram (1884). He considers that the composition of fauna, in particular fish and mollusks, points to a connection between the Jordan system and rivers entering the adjoining seas. He thinks that the origin lies in the sea, and maintains that saltwater animals could have adapted to freshwater, given sufficient time. These animals became enclosed and isolated in the Jordan drainage system in Miocene times, when the tectonic activity took place, separating the Valley from the sea. Thus the Jordan fish are considered descendants of those living in the Eocene sea. Hudleston (1885, p. 42), based on the Jordan system fish identifications by Günther (1864), maintains that at some time a connection must have existed with African waters, thus doubting Lartet's views (above).

Hudleston (1885, p. 63) challenged the idea that the Jordan Valley, in times of higher water level, was connected to the Gulf of Aqaba. He maintains that since the sill separating the Valley from the Mediterranean, in the Yizre'el Valley, is much lower, any possible waterway should have preferred this route. Incidentally, this idea was later confirmed (see Chapter 7). Another interesting point had troubled Hudleston (p. 72) and apparently also Hull: "since this sea has no outlet, what have become of the materials that have disappeared?", referring to all the water and debris removed by the wadis leading to the Dead Sea. His suggestion is that "the bottom of the Jordanic fissure has not always been perfectly watertight", and that there is "the possibility of some connection between the waters of this fissure and the steam power necessary for the production of the three large volcanic areas that lie to the east and northeast of the Lake of Tiberias (Lake Kinneret) … Water is necessary to a volcanic vent as it is to a steam engine". However, no solution is proposed for the sediment load brought to the basin.

The question of life in the Dead Sea had troubled many. Schubert (1840) reported life forms, but these were found close to springs or streams, where the water is not as saturated with salt as in the main water body of the lake. These were also dealt with by Ehrenberg (1849), who reported some living foraminifera, among other microscopic creatures, and also by Tristram (1873) and others (Clapp 1936). Only in the 1930s did Elazari-Volcani (1936 and later publications) discover algae and bacteria which lived in the concentrated brine itself.

Nötling (1886) contributed some ideas, as did many, on the destruction of Sodom and Gomorra. Nötling (1887), in addition to other studies in the central Jordan Valley and the surrounding regions, made a detailed survey of the hot springs in and around El Hama, southeast of Lake Kinneret, where the Yarmouk River enters the Jordan Valley. He drew a geological map of the region, studying in particular the lava flows and the stratigraphy of the Quaternary sediments.

Hume (1901) made an extensive study of the rifts of eastern Sinai, maintaining that the principal morpho-structures are a result of dislocation, rather than of erosion, an idea prevailing some time before (see above). He stresses three principal directions of faulting: the two systems of longitudinal rifts, the gulfs of Suez and Aqaba; and a third, latitudinal, which crosses the entire peninsula. The movements along these three systems are responsible for the present-day shape of Sinai.

Bonney (1904) may have been the first to challenge the term "rift", basing his objection on the fact that "rift" is not an accurate translation for "graben". He further quotes Suess (1892–1909) as stating that the Jordan Rift Valley should accurately be referred to as a "sinking graben". Bonney therefore suggested "trough valley" as a substitute applicable to this feature.

Suess (1909, pp. 304–321) suggested a continuity of the sequence of valleys extending from northern Syria to southeastern Africa, describing it as a dominant lineament of the Earth's crust. Gregory (1921) recognized the similarity between the Jordan Valley and the African rifts, and considered both as parts of a greater Levantine-African system, running from East Africa through the Red Sea, the Gulf of Aqaba and the Jordan Valley.

Willis (1928) had challenged Gregory's view, practically reversing it, proposing the term "ramp" for the Jordan Rift Valley, based on the idea that it was formed by compression. This view was accepted by Clapp (1936) and also for a while by Picard (Willis & Picard 1932). According to Willis the Dead Sea is surrounded by thrust faults, rather than normal ones. This he bases on the direction of the fault dips, presumably going downward and away from the Dead Sea. This led him to challenge Gregory in understanding the tectonic processes. Willis' view of the genesis of the "graben" requires that the faults represent thrust planes, whereas Gregory believed them to be normal faults.

Willis quotes Gregory (1921, p. 357): "the Great Rift Valley was formed by the subsidence of strips of the Earth's crust between parallel faults. This fault-formed valley is continuous from Palestine to south of the Zambezi, except for about 80 miles in southern German East Africa". Gregory (p. 358) maintains that "the floor of the rift valley consists of rocks which have been lowered between parallel faults", causing a "sunken keystone" between them. Gregory suggested a mechanism of tension for rifting, basing his idea mainly on the absence of compressional features on both sides of the entire African Rift Valley system, and

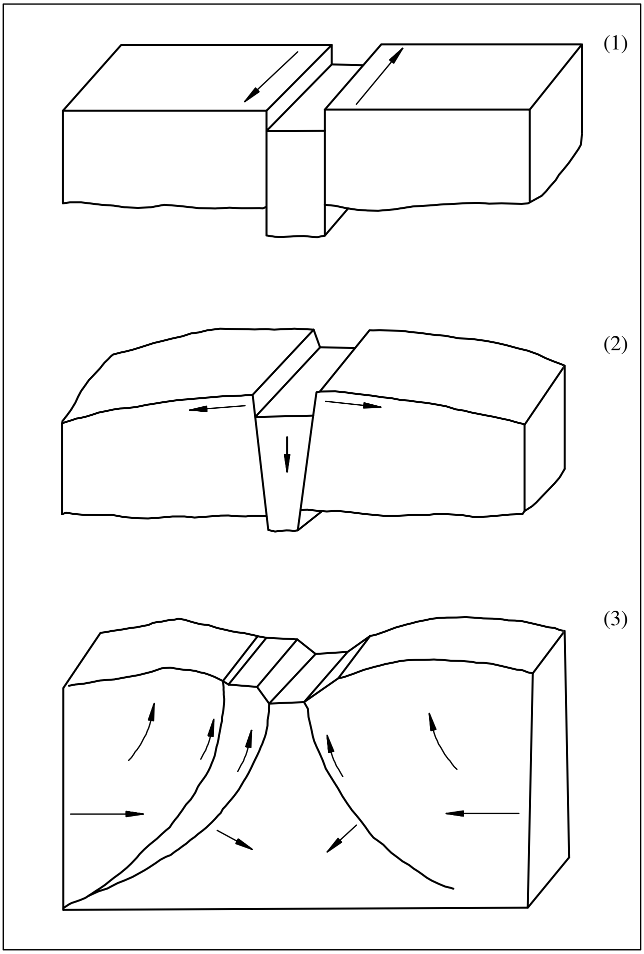


Figure 2.1.3. Structural configuration of: (1) Rift, (2) Graben, (3) Ramp. After Willis (1928), modified.

the widely distributed volcanic rocks in and close to the rift. This idea is opposed by Willis for the Dead Sea, suggesting instead a compression mechanism which lowered a "keel", termed by him "ramp" (Fig. 2.1.3). This is based on Willis' conclusions that "the bounding faults are not to be considered normal faults, but rather thrust faults".

Willis (1928, p. 496) also challenged Suess' (1909) view: "it seems probable that the continuity has been overstressed". Willis further says that "geologists, taught that subsidence along faults is a common phenomenon, have not been critical in examining the criteria by which to determine whether they had to deal with a case of a sunken block or of upthrust plateaus. The surface features of the two cases are very similar, and can be distinguished only by unbiased investigation of the related physiographic and structural phenomena". Willis (pp. 496–497) also objects to some of the structural observations and delineations of faults in Palestine, made by Blanckenhorn (1914).

It is interesting to note that Willis was probably the first to disengage the formation of the Jordan Rift from the folding of the Syrian arc (see Chapters 4 and 9). He recognized that the folding preceded the rifting, while only the uplift of the regions bordering the Rift was connected with its subsidence.

G.S. Blake was a geological adviser of the Palestine Government. His report (1928) on the "*Geology and water resources of Palestine*" summarizes the state of art at that time, containing also some discussion of the Jordan Rift Valley and a detailed geological map of Palestine and Transjordan. Blake based his stratigraphic considerations mainly on Blanckenhorn's studies. He does not readily accept the simplistic theories hitherto suggested for the origin of the Rift. Rather, he maintains (p. 26) that "The history of the rift faulting requires more study before it can be definitely unraveled". One of his more important observations, usually neglected even today, when it does not conform to fashionable theories, was that (pp. 26–27) "The rift faults are by no means continuous … North of the Dead Sea the main faults turn away from the rift and die out, starting again further north … Northward the north–south faulting is less evident … the repeated east–west faulting are evidenced". Blake confirms that a few faults in the Dead Sea region result from overthrust, but does not see these as sufficient evidence for Willis' (1928) "ramp" theory. He also recognized that the volcanic activity close to the rift borders predated the main rift movements.

Picard (1929) made a detailed geological and prehistoric study of the Bet She'an Valley, including an explicit map and cross sections, in which he paid considerable attention to the Quaternary beds and prehistoric occupations from the Mousterian, Neolithic and Bronze Age. Continuing this study (Picard 1932), he described in detail the geology of the central Jordan Valley, between Wadi el Öschsche (Nahal Harod) and the southern end of Lake Kinneret. The "Melanopsisstufe" (Ubeidiya Formation) was considered at that time as of late Pliocene age. As in the previous study, this one is also accompanied by a map and cross sections, this time concentrating almost exclusively on formations confined to the Rift Valley. Picard also presents in detail (1932, Table 24) the sedimentary and volcanic units, their lithology, facies and tectonic settings, accompanied by paleoclimatic inferences for the entire sequence, from the basal Pliocene (Miocene of today) through the Quaternary.

Blanckenhorn (1912) was under the impression that the age of the salt of Mount Sedom was Quaternary, and it constituted an uplifted horst. The intrusive nature of Mount Sedom salt was probably first recognized by Wyllie (1931). Lees (1931) wrote that "only in two places throughout the length of the Valley the Lisan beds suffered any dislocation, at Qarn al Ahmar (Zahrat el Qurein) and at Mount Sedom. In both cases the disturbance is due to intrusive bodies which have formed elongated domal structures in the Pliocene and Miocene deposits, underlying the Lisan beds. The latter had also been affected by the movement, although to a lesser degree. At Mount Sedom the Lisan has been elevated to a height of several hundred feet above its normal level in the surrounding country. The intrusive body at Qarn al Ahmar is basalt, and at Mount Sedom it is salt". The "Biblical effect" was still in power at that time, as Lees adds that it is interesting to speculate whether the story of Sodom and Gomorra might not in some way be connected with the movement of the salt dome. This effect has not entirely vanished, as the last book on the topic was published only a few years ago, by Neev & Emery (1995).

Dubertret (1932), who studied in detail the geology and structure of the Lebanon and northern Palestine, was the first to suggest the "Strike-Slip Theory" (see Chapter 10) for the Jordan Rift Valley. He postulated a 160 km southward movement of the Sinai block, between the rifts of Suez and Aqaba, and up to the Lebanon, based on Lartet's observations indicating the differences between the eastern and western flanks of the Rift, as well as the fit between the two coasts of the Red Sea which, when reconstructed, involved a necessary southward moving of the Arabian side (Lartet 1869, footnote on p. 13). Later, Dubertret (1970, 1973 *in letteris*) himself changed his mind and opposed this idea altogether.

Clapp (1936) suggested that the southern basin of the Dead Sea, where the water depth is only 2–6 m, according to soundings made by Lynch in 1849, submerged only during early historic times. In a cross section Clapp (1936, p. 884) shows the Senonian rocks underlying the southern Dead Sea quite close to the surface. This conclusion is contrary to both the cross section by Lartet (1877, Plate III) and what we know today (Fig. 2.1.4) from boreholes sunk in this area, which penetrated several thousand meters of Quaternary sediments (see Chapter 6).

Clapp (1936, p. 890) indicates that "the precise geologic (structural) term to be employed for the Dead Sea Graben has been debated". Maybe not surprisingly, nothing has changed in this respect (see Chapter 10). He also states that "all travelers to the Jordan Rift Valley seem to have recognized the unstable state of equilibrium there". The general idea was that the depression was formed by a series of movements, which apparently took place during the Neogene, probably extending into the Quaternary.

Clapp (1936) found that the dips and distribution of bounding faults confirm Willis' (1928) opinions. He (p. 893) maintains that "displacements underlie the valley as well as the highlands. Major faults do not extend throughout the entire length of the ramp as continuous bounding planes, but they coincide with one or the other mountain wall for a few miles, then curve off into the lake or the highlands, as the case maybe, giving an en échelon effect. Yet, physical consistency throughout the graben must presumably be explained by some fundamental cause. Faults exist aside from those affecting the graben itself can be seen" (in several localities mentioned).

Vaumas (1949) made a study of the Hula basin, claiming that it constitutes a syncline, which is a continuation of the Beqa'a to the north. This view is mainly based on westward dips of the basalts east of the valley which, in Vaumas' opinion, reflect the inclinations of the underlying rocks. The western escarpment of the Hula Valley is explained as a fold. Avnimelech (1956) made a study of the evolution of Lake Hula, and was probably the first to emphasize the importance of transversal faults

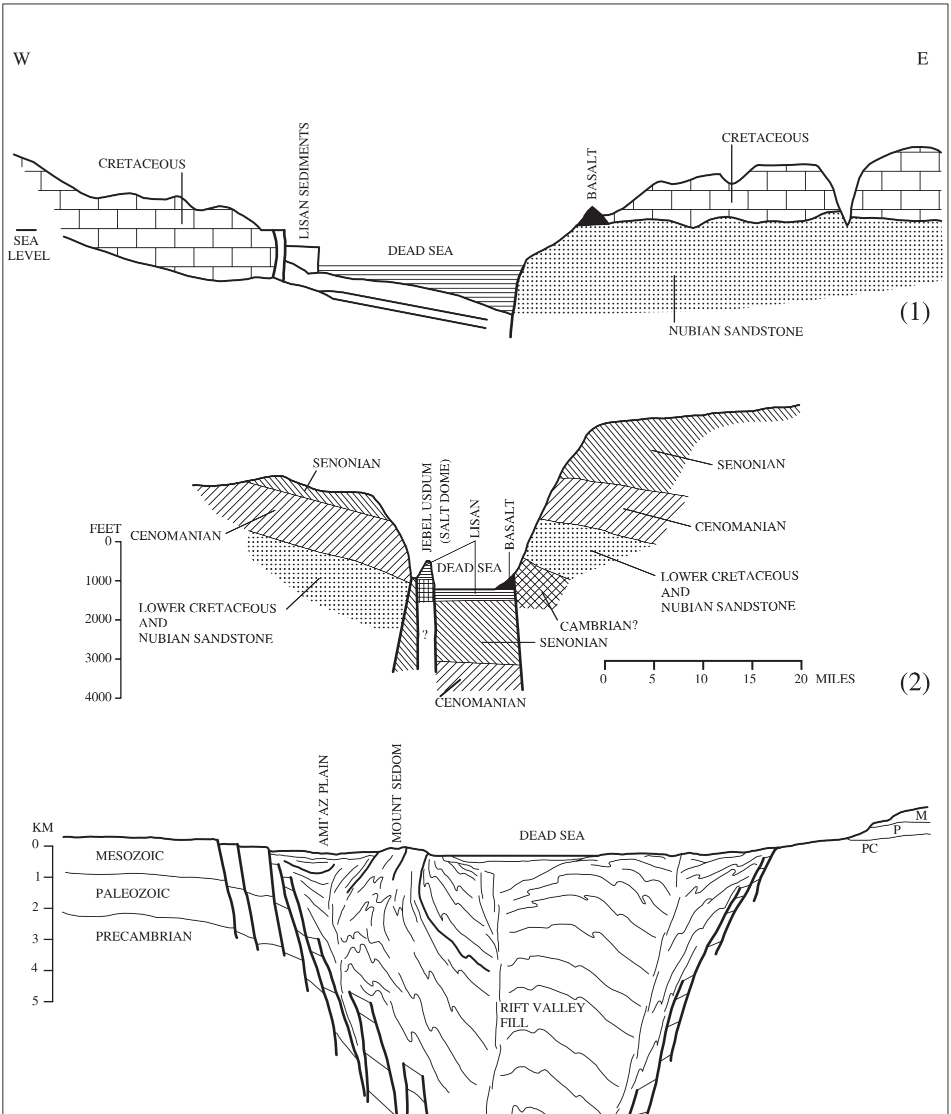


Figure 2.1.4. Cross sections of the southern Dead Sea according to: (1) Lartet (1869), (2) Clapp (1936), (3) Zak (1967).

in the development of this basin. Such faults have subsequently been described from other basins of the Jordan Rift Valley (see Section 3.1 and, Chapters 8 and 9).

Following Dubertret (1932), Quennell (1959), who worked in the Kingdom of Jordan, revived the “strike-slip” theory to explain the formation of the Jordan Rift Valley. The theory was dismissed by almost all until 1961, when R. Freund, then a young, good-looking student, had presented it to the Israel Geological Society. The extent suggested by Dubertret for the lateral movement, 160 km, was recalculated and set at 105 km (Freund et al. 1970). Ever since, this has been the leading theory, although it has its opponents (such as myself, see Horowitz 1979, pp. 56–62, and Chapter 10 below).

In addition to these fashionable theories, others were also proposed at the same time, such as the idea of Bloch & Picard (1970) that the Dead Sea is a sinkhole, caused by a collapse into the hollow formed by dissolution of the underlying salts. The volcanic disaster idea as the cause of destruction of Sodom and Gomorra also persisted until quite recently (Block 1975).

A turning point in our knowledge of the Jordan Rift Valley came in the 1960s, when interest in oil and gas led to extensive geophysical studies and drilling activities. Some 14 boreholes have each penetrated several kilometers of the rift valley fill. Palynological studies of these sequences, accompanied by radiogenic datings, resulted in detailed stratigraphy, which permitted structural and paleoclimatic reconstructions (Horowitz & Horowitz 1990, Horowitz 1992a, 1996a, and see Chapters 6, 9 and 10).

2.1.2 Paleoclimates

Historic changes in the level of the Dead Sea had been noted by many early investigators, such as Robinson & Smith (1841). Probably the first mention of the possibility of climatic changes in this part of the world is by Forbes (1861/1862), who maintained that the climate of Palestine had changed since "ancient periods". Walcott (1868) had referred to the persistence with which the water level of the Dead Sea rose throughout the last 19 centuries. The water covered a district in the southern basin at that time which was the ancient fertile Vale of Siddim, and to the north the plain of the Jordan River. Walcott, who was a missionary, not exactly a trained scientist, considered that the conversion of the Vale of Siddim to an embayment of the Dead Sea is an historic process. Hull (1886, pp. 123–127) deals also with climatic changes, particularly comparing Biblical with recent times. Historic changes in the Dead Sea's water level were later discussed by numerous researchers such as the physician and amateur scientist Masterman (1902), geologists such as Blanckenhorn (1912) and Schrötter (1924) and the archaeologist Albright (1924), and summarized in great detail by Klein (1986). Clapp (1936) suggested that the deepening of the southern basin may have resulted in part from earthquakes, and not entirely from a mere rise in the water level.

The question whether the glaciation of the ice ages had affected Israel (and in connection with the Jordan Rift Valley, its northernmost part), or whether glaciers had indeed occupied the higher mountains, had troubled several investigators. The general view, through the entire history of research, and still holding today, was that temperatures in the Levant never dropped enough to maintain any significant, large scale glaciers.

Lartet (1877, p. 169) had indicated that no signs of large-scale glaciers during the Quaternary could be observed in the Middle East. He raised a question whether, in addition to higher amounts of rainfall, a temperature drop at this time is plausible. He mentions (p. 180) that Hooker (1862) and Girard de Rialle (1868) claimed that mountains higher than 1,200 m in the Lebanon and Syria, such as the Hermon, were glaciated during the Quaternary. Lartet indicates that layers pointing to past higher lake levels are seen also around the Hula and Kinneret lakes (p. 181). He thus sees an intimate connection between the Quaternary glaciation in Europe and the high lake levels of the Levant. Lartet thought a considerable temperature drop unlikely, and attributed the lakes' greater size to higher precipitation and decreased evaporation, which seem sufficient in his opinion to explain the high levels of water in bodies such as the Lisan Lake.

Lartet (1877) thus suggested that the Ice Age is characterized in Palestine, Syria and the Lebanon by considerably higher lake levels (such as the Lisan Lake, p. 179), as compared with the present day. No doubt, in his opinion (pp. 174–175), the great extension of the Lisan Formation (Dépôts de la Liçan) sediments represents a different climate regime.

The Lisan Lake, according to Lartet, was formed at the end of the Tertiary as an endoreic basin, its boundaries dictated by an equilibrium between the amount of rainfall and evaporation. The lake's level reflects therefore the atmospheric conditions during the Quaternary. The transition from the Lisan Lake to the Dead Sea (p. 179) is attributed to a major climatic change. Lartet proceeds to say that a detailed study of the Dead Sea and the Lisan Lake could yield very important paleoclimatic information for the entire Levant, a study which is being carried out today by a large group of scientists (GIF 1997).

Lartet also mentions (p. 171) that numerous indications of a very different "ancient" climate exist in the Levant. These include a variety of calcareous gravel and breccia horizons in Judea, the Galilee and the Lebanon, and also around the Dead Sea and in the Arava. Lartet maintains (p. 172) that all these are remnants of past large rivers, and a hydrographic regime much different than that of the present day. Further, he also concludes (p. 173) that the River Nile was larger and faster than it is today.

Blanckenhorn (1910, 1920/1921) suggested that the climate in the Levant during the Pleistocene must have been temperate, precipitation falling during all seasons of the year. He referred to this type of climate as "pluvial". This is probably the first suggestion of the possibility of summer rains in this region. Loewengart (1928) explained the fact that coastal dunes are interwoven with red soils (locally known as "hamra"), as being the result of a succession of wet and dry periods.

Picard (1932, Table 24) presents in detail the paleoclimatic inferences of sedimentary units in the central Jordan Valley, for the entire sequence from the early Pliocene (Miocene of today) throughout the Quaternary. A warm, moist climate is inferred for the basal, Herod Formation, which consequently dries up and oscillates during the Pliocene. The Mediterranean climate is thought to have been initiated in this region with the Ubeidiya Formation (thought then to be of late Pliocene age). He considered the Lisan Formation to be of "Pluvial B" age (middle and upper Pleistocene), indicating a humid climate. Picard (1952) applied this subdivision to the Quaternary formations of the Hula and the central Jordan Valley.

Bergy (1932) studied prehistoric sites in the Lebanon, and considered the typical Mediterranean hamra soils as glacial mud. Picard (1937) studied the Pleistocene climate of Palestine as reflected in faunal remains recovered from prehistoric excavations. He summarizes (p. 70): "the Mediterranean climate set in either with or after the Vindobonian transgression, that is, from the time of the decisive mountain building phase of the Miocene. From paleogeographical point of view *the pluvial periods is not only a phenomenon of the Pleistocene, but of the whole post-Miocene epoch* (starting from the Pontian)" (Picard's own italics).

Picard (1943, Table in front of p. 160) further developed the "Pluvial" idea suggested by Blanckenhorn (above), dividing the Quaternary into "Pluvial A", typified by the Samra Formation gravels; "Pluvial B", characterized by the Lisan Lake; and "Pluvial C", during which terraces (the "Upper Terrace") were formed in the wadis leading to the Jordan Valley. Pluvials A and B are separated by the "Main Interpluvial", B and C by a questionable one.

Shalem (1950) summarized his ideas of the Quaternary climates of the Levant, based on numerous previous publications. He rejected altogether the pluvial theory, stating that differences in sediments and extension of lakes should only be attributed to tectonic activity and changes of base levels, rather than to any climatic change. In his own words (p. 595) "The changes in the climate of the Levant are a result of local geomorphological developments, in their widest sense; and that glacial pluvial, and planetic influences had no part in them; and if so, only to a negligible degree, so that they cannot be separately distinguished as such".

Picard (1952) maintained that the Pleistocene climate of Israel, based on the Jordan Valley sediments (and for the first time, pollen analysis) was Mediterranean, with fluctuations resulting in the three "pluvials".

Quite recently, Ashbel (1961) suggested that Mount Hermon, north of the Hula Valley, had been covered by glaciers, their moraines reaching down to the northern part of the Hula Valley (these "moraines", incidentally, are gravel beds deposited by Neogene rivers). Vaumas (1965 and former papers) tended to see many gravel beds in Cyprus, Israel and the Lebanon as having been formed by solifluction processes, possibly connected with glaciation of the higher altitudes.

Palynological analyses, which contributed considerably to our knowledge of past climates in the Jordan Rift Valley and the neighboring regions (as well as stratigraphy, see Chapter 6), were applied to the rift valley fill from the 1950s. The first were done by Fritz Brotzen and his wife Pnina in Sweden (Picard 1952, p. 148), on samples from the K-Jam borehole, drilled in the Hula Valley in 1950 to a depth of 120 m. These were followed by Lorch (1959), who further analyzed the same borehole, and by Horowitz (1971) who first drew a pollen diagram for the uppermost part of the Hula basin fill. Several samples from the Dead Sea fill were analyzed by Klaus (1965) in 1959, by Remy (unpublished), and in the 1960s by Horowitz (unpublished) and Rossignol (1969a,b). Pollen analyses of Lake Kinneret sediments were first published by Horowitz (1969, 1971). These were followed by extensive palynological studies of boreholes, outcrops and prehistoric sites all along the Jordan Rift Valley by various investigators (Horowitz 1992a, Chapters 9 and 10).

2.1.3 The search for oil

Solid and semi-solid hydrocarbons are present in the Dead Sea basin in four principal forms: as asphalt blocks floating in the water or washed ashore; in veins, seeps and as cavity fillings in Paleozoic through recent rocks (Fig. 2.1.5); as ozocerite (mineral wax) veins on the eastern shore of the lake, in association with asphalt and heavy oil; and as outcrops of bituminous rocks of Senonian age (Clapp 1936, Nissenbaum 1978).

Asphalt blocks floating on the lake's surface are the most dramatic phenomenon, thus having attracted the highest attention. The blocks are of various sizes, occasionally weighing more than 100 t. They are found floating, or subsequently cast up on the shores (Fig. 2.1.6). They usually appear following earthquakes, such as in 1834 and 1837 (Hitchcock 1843, p. 371). However, the exact source of the asphalt

Figure 2.1.5. Asphalt impregnating wadi gravel in Nahal Hemar (center), near one of the transversal faults (left). Looking eastward.

Figure 2.1.6. Asphalt, previously floating, cast up on the Dead Sea shores. Photo courtesy of A. Nissenbaum, sitting.

is unknown. Asphalt seems to have been the most important natural resource of the Dead Sea in ancient times, so that it was called "Mare Asphaltum" or "Lacus Asphaltites" by the Romans (Pliny, Strabo), or "Asphaltitis Limne" by Flavius Josephus, names which were in partial use until the beginning of the 20th century.

Hydrocarbon occurrences in and around the Dead Sea area have been known and mentioned ever since Biblical times. A detailed account of historical aspects of Dead Sea asphalt is given in Nissenbaum (1978). The earliest reference to asphalt in the Bible is its use as mortar in the construction of the Tower of Babel (Genesis 11:3). The Dead Sea asphalt is explicitly referred to in the Bible in the story of the war between the kings of the five cities near the Dead Sea (including Sodom and Gomorra) and the invading kings from the north (Genesis 14:10). The war ended in the total rout of the defenders, who fled into the Vale of Siddim (which is the southern Dead Sea; Genesis 14:11), where they fell into the "slime pits". "Slime" is the word used by Martin Luther for translating "Hemar" (Hebrew for asphalt). Ever since Biblical times (and maybe before) Dead Sea asphalt was an important commodity for trade in the region, and a source of numerous stories and legends (Nissenbaum 1977).

Historical records of the Dead Sea asphalt are numerous. It seems that the concession for asphalt collection was a good reason for many wars in the region. One of the earliest was in 312 BC, between Antigonus, King of Syria, and the Nabatean Arabs. The Syrian King was defeated, and the Nabateans continued to export the substance to Egypt, where it was an important constituent in embalming. One of the most detailed descriptions of asphalt exhalation and collection techniques is given by Diodorus Siculus (ca. 50 AD; see Nissenbaum 1978). Strabo (63 BC–20 AD) also discussed these phenomena in detail, describing the emanations of "invisible soot" (most probably H_2S) which tarnished copper and silver, preceding the rising of asphalt to the lake's surface. The asphalt occurrences are also mentioned by Josephus (37–95 AD), Tacitus (55–117 AD) and others (for details see Nissenbaum 1978). Further descriptions come to us from the Middle Ages, but fewer than before.

A detailed discussion of the ancient applications of asphalt is given in Nissenbaum (1978). It was used as a medicine, for embalming, in agriculture, as a sealant for building purposes and as a weapon, the famous "Greek fire" (Forbes 1964). Asphalt was employed as a mortar in the walls of Jericho as early as 2500–2100 BC (Abraham 1960). Josephus mentions its use for caulking boats. It was also widely utilized later in Europe as a ground for etching.

Almost every traveler visiting the Dead Sea during the 19th century (e.g. Seetzen 1810, d'Aoust 1834, Russegger 1841, Robinson & Smith 1841, Tristram 1864, Lartet 1866 and many others) mentioned and discussed the asphalt occurrences. Lartet (1866) indicates a possible connection of the Dead Sea asphalt with subterranean volcanic phenomena, as was also proposed earlier by d'Aoust (1834). Delachanal (1883) published a paper on the composition of asphalt from the Dead Sea.

Another occurrence of bituminous rocks is the Senonian "Oil Shales", which crop out west of the Dead Sea, reported by Hitchcock (1843), Anderson (1852), Lartet (1866), Blanckenhorn (1912), Blake (1928) and others, from Nabi Musa and other localities. The first analysis is given in Hitchcock (p. 364) who found some 25% bituminous substances. Anderson (p. 155) found only 13.55% organic matter, the deposit being highly variable.

The occurrences of hydrocarbons around the Dead Sea had convinced many investigators that these should be regarded as possible oil indicators. Wade (1921) and Fohs (1927) published papers on oil prospects in Palestine, where they also referred to hydrocarbon seepages in Wadi Muhawat (Nahal Hemar, Fig. 2.1.5). These were also referred to in detail by Lartet (1866), Blanckenhorn (1912), Blake (1928, 1930) and others. Clapp (1936), who came as a government adviser on oil prospects, suggested that a possible source rock for oil in the Dead Sea area is the Senonian oil shales, a view which still persists.

Geophysical investigations of the Rift Valley already commenced in the 1930s, by A. Löhnberg and A. Löwenstein, and were submitted as unpublished reports to the Palestine Mining Syndicate Ltd. More on the history of geophysical studies can be found in Chapter 8.

Toward the second half of the 20th century efforts to discover oil and gas have considerably expanded, when local and international companies entered the scene both in Israel and the Kingdom of Jordan. This resulted in numerous reports dealing with the possibilities for oil prospects (Picard 1947, 1959, Ball & Ball 1953, Gardner 1956 and others), geophysical studies, geochemical investigations and finally drillings. The first boreholes were drilled in the 1950s in the Dead Sea area, such as Mazal 1 in 1953, Masada 1 in 1954, En Gedi 1 in 1955 and Arava 1 in 1959. On the Jordanian side, Jordan Valley 1 was drilled in 1959 and Lisan 1 in 1960. Although the Dead Sea was the main target area, one of the first boreholes, Jordan 1, was drilled near Tiberias in 1957, and Emeq Hula 1 was sunk in the Hula Valley in 1969. Some of these boreholes (Picard 1959) had "excellent shows, but not yet leading to production, in all formations from Triassic to recent". Ever since, numerous boreholes were sunk into the Rift Valley fill (see Chapter 6) and into neighboring structures. Except for several small gas fields west of the Dead Sea, all the holes are practically dry.

The only area within the Rift itself which produces minor amounts of natural gas is the Hula basin. This gas is most probably liberated from the peat layers common to the region. The first borehole, K-Jam, drilled in 1950, only penetrated some 120 m into the basin's fill, but was followed by other, deeper drillings.

2.1.4 The search for minerals

Picard (1954), in his essay on the *History of mineral research in Israel*, points to the fact that until late in prehistory people had all the raw materials they needed in the Jordan Valley, since these were almost entirely limited to flint and clays. However, this situation changed when metals came into use. Of these, copper was discovered and used as early as Chalcolithic times (for details see Section 2.2). What remains obscure is the source of minor additions to copper smelting, such as antimony and arsenic. Several sources for metals such as iron and gold, in wide use in ancient times, are suggested. Ritter, who had not visited Palestine, but was supplied with the best sources (1850) and Blanckenhorn (1912), followed by Picard (1954), propose that iron was mined in Wadi Zarqa (Nahal Yabbok), a tributary of the Jordan River, where ancient mines had been found. The ore is of metasomatic origin, consisting mainly of limonite and hematite, similar to the occurrences in Nahal Paran, close to the central Arava. Iron was probably also imported from Syria and Lebanon.

Another occurrence of iron, in connection with the Jordan Rift, is sedimentary, oolithic, low grade ore, embedded within the early Cretaceous sequence, found where these rocks are exposed at the Rift's flanks east of the Hula, in Wadi Fari'a and other localities (Blake 1930). No archaeological use of these was found.

Gold is supposed (Picard 1954) to have been imported. However a small Early Islamic mine was recently found near Elat (Gilat et al. 1993). Claude (1929) described a gold mine at the Dead Sea, which was not verified.

The search for minerals follows the geological studies of the Jordan Valley ever since the late years of the 18th century, and was initiated by Napoleon. However most early expeditions, particularly those looking for metals, disregarded Palestine altogether in this respect, for reasons justified until the present day, in spite of the Biblical promise: "A land whose stones are iron, and out of whose hills you can dig copper" (Deut. 8:9). A detailed search for non-metallic mineral deposits commenced only much later (Range 1921, Blake 1930, Picard 1932, Shaw 1948), although usually based on observations by earlier researchers, such as Anderson (1852), Lartet (1877), Hull (1886) and Sachsse (1897). These included mainly phosphates, bituminous rocks, Dead Sea salts, groundwater and building stones.

It is interesting to note that in 1948, at the termination of the British mandate, seven prospective areas were listed (reprinted in Picard 1954). Of these, five are intimately connected with the Jordan Rift Valley. The Dead Sea for potash, bromine and magnesium; Mount Sedom for petroleum, bitumen and salt; Nabi Musa and the Yarmouk area for bituminous limestones; Nabi Musa for phosphates and Menahemya for gypsum. To those the Hula should be added, with peat and natural gas (Picard 1952).

2.1.5 Chemical analyses of Dead Sea waters

A detailed historical account of chemical analyses of the Dead Sea waters is given in Nissenbaum (1970). The first analysis, of a sample collected by Pococke in 1738, in a glass stoppered bottle, was made by Macquer, Lavoisier and Sage (1778), who presented their results to the French Royal Academy of Sciences in 1781. The specific gravity was found to be 1.2406 g/cc, the dissolved salts comprising ordinary marine salt with little alkaline earth, and calcium/magnesium chlorides. They had also searched for bitumen and, finding none, concluded that "it is without foundation that several authors had attributed the astringent and disagreeable taste to bitumen … this astringency is a property of … the magnesium base of Epsom salt".

Marcet (1807) found a specific gravity of 1.211 g/cc, with basically similar salts, and also analyzed water from the Jordan River. He was followed by Klaproth in 1809, Gay Lussac (1819), Hermbstädt (1822) who analyzed both Dead Sea and Jordan River water samples, and Gmelin (1827). In no reference is the place of collection indicated, but Nissenbaum assumes that the water samples had been taken from the northern end of the lake. This locality is affected by yearly rainfall changes, which may account for differences in the analyses. With these studies the pioneering era of chemical analyses of Dead Sea waters ended, and they became a routine procedure in following expeditions.

2.2 PREHISTORY AND ARCHAEOLOGY

The archaeology of the Holy Land had attracted the attention of early travelers, particularly for religious reasons. Albright (1960, p. 23) mentions a list of such people, who indicated the existence of ancient remains in the region. First was probably the Swiss Dominican Felix Schmid, who described some archaeological sites observed during his journeys to the Holy Land in 1480 and 1483 (but published only in the middle of the 16th century). The German physician Leonhard Rauchwolff (Rauwolf) visited the region in 1575 and made systematic observations of nature, especially as regards botany. Architecture and archaeology are described in drawings by the Fleming Johann Zuallart from 1586 and the Dutchman Johann van Kootwijck in the last years of the 16th century.

Interest did not fade during the 17th and 18th centuries, with reports by Quaresmius in 1639, Pietro della Valle in 1650, Michael Nau in 1679, and Henry Maundrell in 1703, with much new archaeological material. The Dutchman Adrian Reland published in 1709 a handbook entitled *Palaestina ex monumentis veteribus illustrata* (Palestine illustrated by ancient monuments), which revolutionized the scholarly approach to this part of the world. Bishop Pococke's (1754) journey in 1738 resulted in plans, drawings and copies of inscriptions never done before.

The first quarter of the 19th century saw travelers swarming into Palestine. The more significant of their reports came following visits by the German Ulrich Jasper Seetzen (1810) in 1805–1807, the Swiss Johann Ludwig Burckhardt (1822), who discovered Petra in 1801–1802, and the Englishmen Charles Leonard Irby and James Mangles (1868) during 1817–1818. A significant revolution came in 1838, with the visit of the Americans Edward Robinson and Eli Smith (1841). These two scholars correctly identified many archaeological sites in the Jordan Valley, such as Mezada (Masada, first suggested by Seetzen), Dan and others, thus opening new possibilities for future research.

The first excavations of archaeological sites in Palestine are by F. de Saulcy (1854), in 1850–1851, near Jerusalem, resulting in a collection of artifacts now at the Louvre. Subsequently, the Palestine Exploration Fund was established in 1865, and one of its first activities was sending Charles Warren to excavate Jerusalem. Incidentally, Warren also excavated several test pits in Jericho, but concluded (wrongly) that this site is not at all promising, so that it was abandoned until the beginning of the 20th century when an Austro-German expedition headed by Ernest Sellin and Carl Watzinger dug at the site from 1907 to 1909.

The scope and time scale of this book call for more attention to prehistoric archaeology, which encompasses approximately the last two million years in the Jordan Rift Valley, and somewhat less to Biblical archaeology, which is the study of human remains spanning only the last millennia, from the Early Bronze Age to the present day.

Three phases of archaeological–prehistoric research in the Jordan Rift Valley (and actually in the entire Levant) roughly correspond to general trends in the development of scholarly approaches toward human sites, communities and cultures. Locations mentioned in this chapter are marked in Fig. 2.2.1.

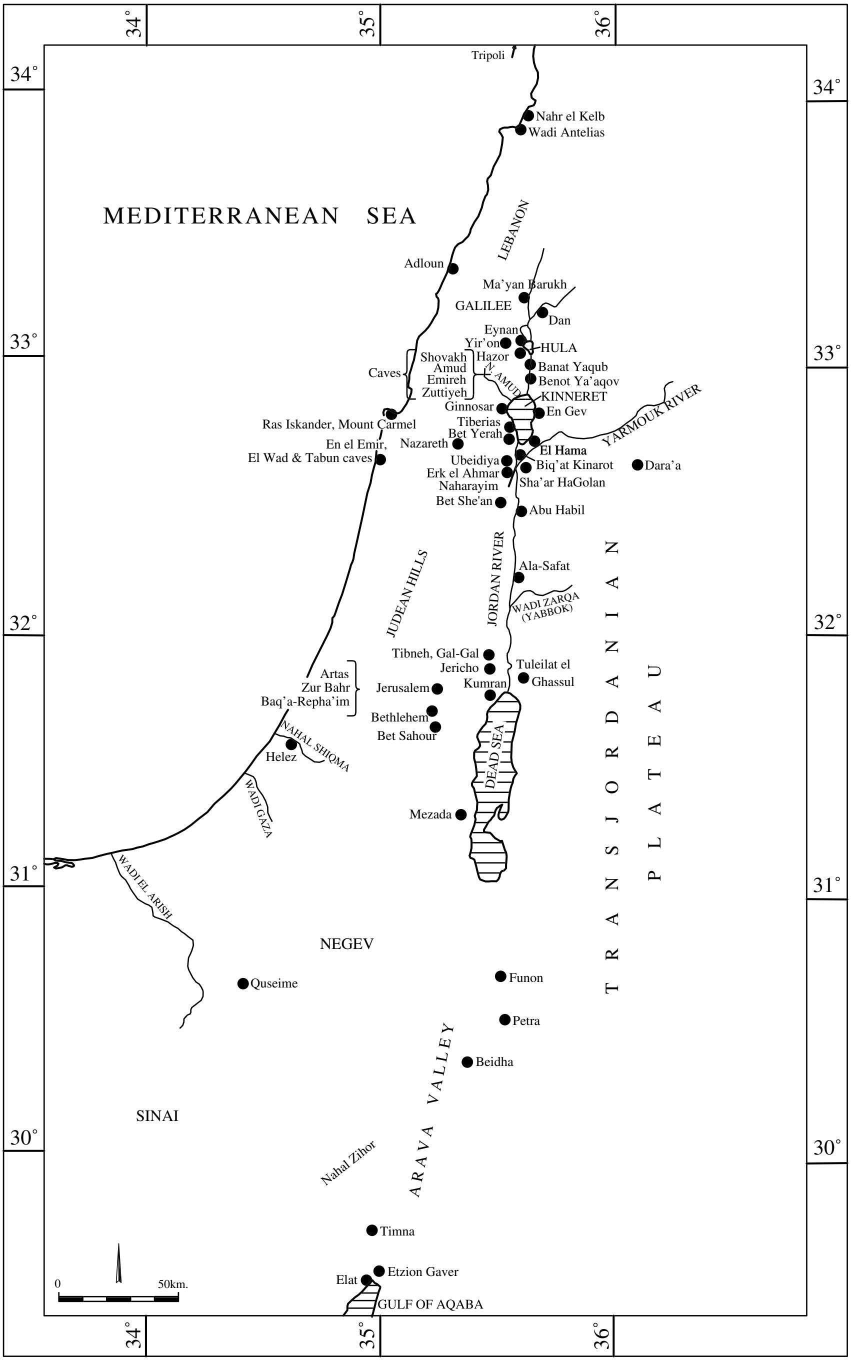


Figure 2.2.1. Location map of archaeological sites mentioned in this chapter.

2.2.1 The first phase: collections

A detailed account of field techniques, concepts and approaches in the prehistoric research of Israel can be found in a book by Bar-Yosef (in prep.), entitled *Prehistory of the Holy Land*. Much of the information put forward here is cited from the manuscript of this book, with the kind help of O. Bar-Yosef (Harvard University 1997, pers. comm.). Two recently published books on the prehistory of Jordan, by Gebel et al. (1997) and by Henry (1998), are a great help in understanding the eastern side of the Jordan Valley.

The first phase was mainly concerned with surface collections of artifacts which were only rarely reported in scientific publications. The most significant drawback of this approach involved the collection of only the nice looking pieces, while all the rest was either disregarded or dumped. Practically, the only conclusions that could be drawn from this stage of "research", was that prehistoric man indeed inhabited the region under discussion. This stage concerned the Jordan Rift Valley in only a subordinate way since, until the pioneering studies of Picard in the 1930s, only the late Pleistocene Lisan Formation was known from the region. Thus a search for older human occurrences was not conducted, as it was thought that any such indications would have been submerged under the waters of the Lisan Lake (Fig. 2.2.2).

Travelers of the late 18th and 19th centuries who visited, among other regions, Palestine quite frequently, then usually regarded as part of Arabia, had noted both archaeological and prehistoric remains. During this stage of research most scholars, who arrived from both Europe and North America, referred to their discoveries using European terminologies, and interpreted them in the context of questions pertaining to human cultural evolution in their own lands. One has to refer, at least during the initial stages of research, to findings in the Jordan Rift Valley and the rest of the Levant alike, since most travelers described the entire region as a single geographic entity.

The first prehistoric remains to attract attention were dolmens in Transjordan and the vicinity of the Dead Sea, discovered in 1817–1818 by Irby & Mangles (1868). They described dolmens from the Dead Sea region in Ala Safat, near the confluence of Wadi Zarqa and the Jordan River, and also numerous structures and large stones from the Transjordanian Plateau, both dolmens and menhirs (Fig. 2.2.3).

Louis Lartet was a geologist interested mainly in the Dead Sea, but well informed in other branches of the sciences, thus describing also prehistoric occurrences. Some of the descriptions and drawings by Irby & Mangles are reproduced in Lartet (1877, pp. 233–235), who also described, on the mountains east of the Dead Sea, besides the dolmens, a menhir 2.30 m high. Lartet visited these, and those found north of Jericho, in 1866, and suggested relating them to the ancient Hebrews (Israelites), or possibly that they are burial sites of a tribe known by the Arabs as the "Haouanet". According to Lartet, who compared the

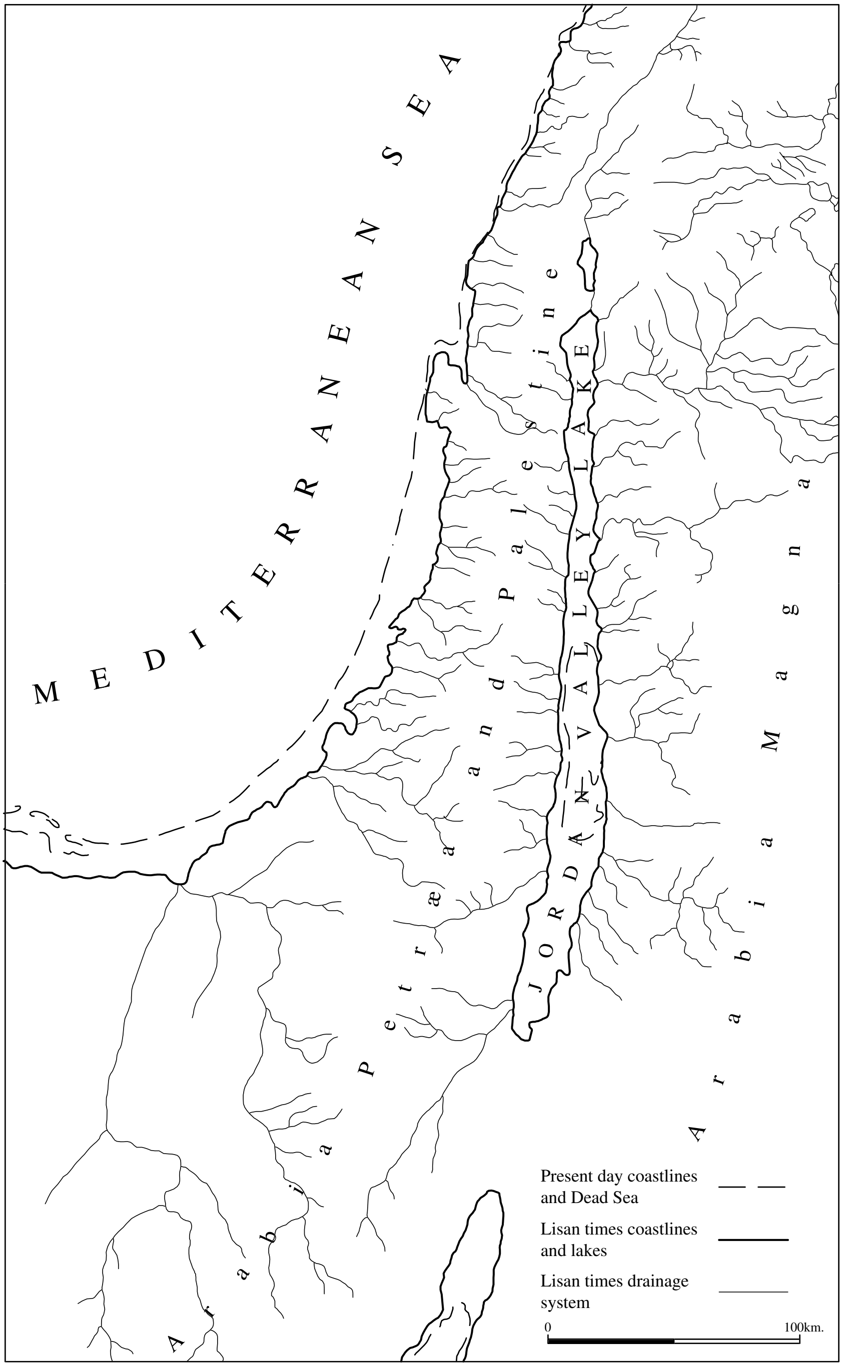


Figure 2.2.2. Paleogeography in the times of the Lisan Lake, according to Lartet (1877) and Hull (1886).

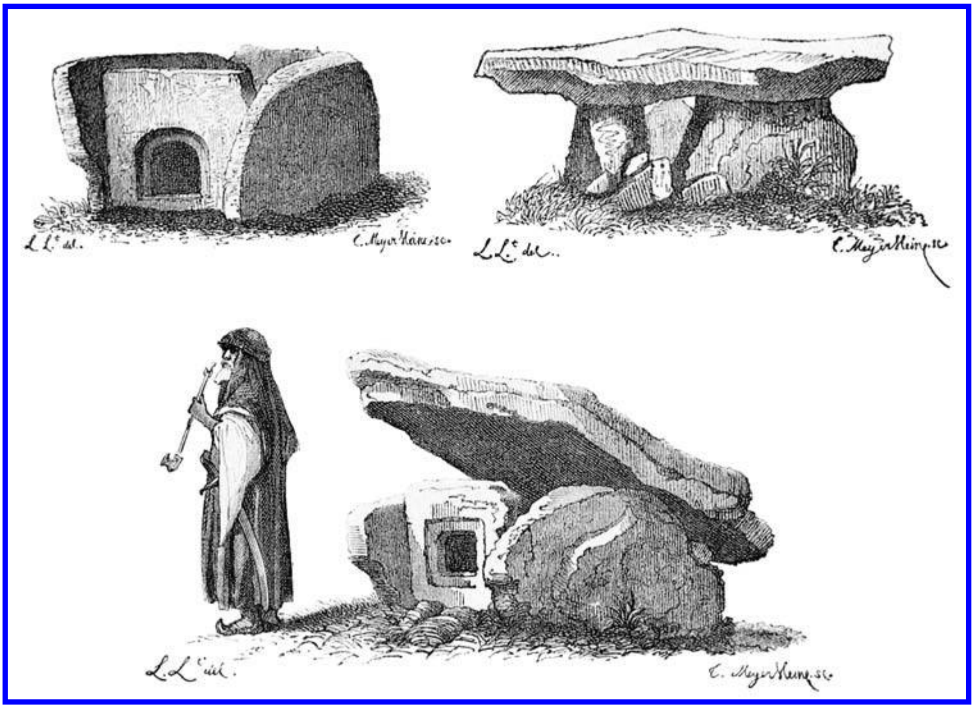


Figure 2.2.3. The first described dolmens of the region, from Lartet (1877).

dolmens with ones from Algeria (p. 237), there is no way to decide whether the megalithic structures were erected by the same race of people inhabiting the region today, or by a different group. Incidentally, Lartet reports that the story went, among the local Bedouins, that the dolmens had been used by ancient tribes as a shelter from mosquitoes. He also maintained that one should dig in the dolmens, in order to ascertain their exact age. Other travelers also mention these dolmens.

Holland (1868) and Palmer (1871) excavated a few of the dolmens in Sinai (called by the local Bedouins "Nawamis"). But it was only C.T. Currely, a member of Sir Flinders Petrie's expedition to Sinai in 1904, who, after digging several more, reported their contents and related them to the Predynastic period (Petrie & Currely 1906). Some of the megalithic structures in Transjordan were later excavated by M. Stekelis (1935, 1960/1961).

The first surface collections of flint artifacts, in the 1850s, were made by Father J. Morétain, who lived in Bet Sahour, a village close to Bethlehem, and described in a paper presented by him to the International Congress of Anthropology and History in Paris (in Buzy 1928). The assemblage comprised numerous axes, saw blades, arrowheads, knives and even a bone point. These were related by Cazaliz de Fondouce to a historic period, basing his conclusions on faunal identifications by Lartet, of bones collected from the breccias of Wadi Antelias in the Lebanon, attributed to living species such as roe deer, fallow deer, ibex, wild oxen and an equid. Later on J. Arcelin had assigned these artifacts to the Neolithic, rather than Paleolithic.

Lartet (1877, p. 225) described the artifact collections of Father Morétain from Bet Sahour, as well as (p. 224) prehistoric sites from the Lebanon, in Nahr el Kelb near Beirut, and Adloun caves to the south. He also reports (p. 227) on artifact finds by Cazalis de Fondouce in En el Emir, near Nazareth. Somewhat later Oliphant (1886), followed by Mülinen (1908), mentioned the Mount Carmel caves.

Probably the first discovery of flint implements within the Jordan Valley was by Father Richard in the 1850s, who found them at Gal-Gal and Tibneh, somewhat north of Jericho, where the tomb of Joshua was believed to be. Richard claimed, conforming with the wave attributing everything from the Holy Land to the Bible, that the artifacts were those used by Joshua for circumcising the children of Israel. Lartet (1877, pp. 227–230) argues that the artifacts, being quite similar to those he found near Nahr el Kelb (Lebanon) and similar to those regarded in Europe as prehistoric, could not really be the particular ones used for circumcision. The tomb of Joshua was identified as such by Guérin, who visited the region in 1852, in Tibneh, and de Saulcy (1854) adopted this opinion. Lartet further argues that there is neither proof for the flint implements being used for circumcision, nor for their presence at the tomb itself. However, Lartet maintains that these difficulties had not for a moment stopped Richard in his conclusions. Lartet also indicates that such flint tools had usually been found in connection with the dolmens and other megalithic constructions, both on the Transjordanian Plateau and within the Jordan Valley itself, and so should not be attributed to historic times. Lartet (p. 232) compares surface finds of artifacts from Egypt, Sinai, Palestine and the Jordan Valley, and maintains that they show similar characteristics. Nevertheless he says that there is no way of dating the superficial strata of these regions at that stage.

Thus, as O. Bar-Yosef (in prep.) states: "This early period can be summarized as a time when it was mandatory to demonstrate the antiquity of prehistoric artifacts, and to reject the idea they all could be related to Biblical events of historical importance."

Paul-Emile Botta and his partner Hedenborg identified a prehistoric breccia in Wadi Antelias, north of Beirut, in 1883 and also in Ras el Kelb, on their way to Tripoli, in Lebanon. These locations were visited and described previously, in 1865, by L. Lartet (1877, p. 221), and were later excavated by Zumoffen (1897) from 1890, the first systematic excavation of prehistoric sites in the Levant.

Père Germer-Durand was one of the most systematic and enthusiastic researchers in the last part of the 19th century. His extensive collections came mostly from the close environs of Jerusalem such as Baq'a-Repha'im, Zur Bahr and Artas. He read a paper in January 1896 at the École Biblique de St. Étiènnes, Paris, summarizing the prehistory of the Holy Land. Germer-Durand concluded, based most probably on the books published by Lartet (1869, 1877) and Hull (1886), that the Jordan Valley (and a considerable part of the coastal plain) had been submerged (Fig. 2.2.2) under the waters of the Mediterranean Sea (although both Lartet and Hull had shown a lake for the Jordan Valley). Prehistoric Man could thus inhabit only the hilly areas of both Cis- and Transjordan. Incidentally, these conclusions hindered any systematic prehistoric survey of the Jordan Rift Valley for quite a long time.

G. de Mortillet, who was the first editor of *Materiaux*, a journal dedicated to prehistoric archaeology and related topics of interest, mentions the collection of Morétain, and also quotes the presence of Chellean artifacts near Bet Sahour and, interestingly, in the vicinity of Tiberias, collected by Father Richard. It seems however that prehistoric research as such was not carried out in the Jordan Rift Valley at least until the beginning of the 20th century. Rather, prehistorians were considerably more interested in the dolmens of Transjordan and Sinai (see above). In the course of the same decades several scholars had also revealed prehistoric remains in other countries of the Levant, such as the Lebanon, Egypt, Transjordan and Arabia (Blanckenhorn 1920/1921, Bar-Yosef in prep., and others). From the beginning of the 20th century until 1914, when World War I had commenced, prehistoric research of the Holy Land was expanding considerably. However, most scholars, again as a result of the hypothesis that the Jordan Valley was submerged during most of the Quaternary, looked for prehistoric remains in the hilly parts of the country. Another reason for this preference was because most of these researchers had been staying in Jerusalem, either as clergymen or diplomats.

A geological survey of the Holy Land by Max Blanckenhorn (1913–1915) started in the last years of the 19th century, continuing until the 1930s, and was always accompanied by observations of prehistoric remains. Among Blanckenhorn's important discoveries were hand axes near Dara'a, in Transjordan and the site of Ras Iskander on Mount Carmel, which he assigned to the "Campignian" (known today as "Neolithic–Chalcolithic"). Blanckenhorn (1920/1921) also published an essay on the Stone Age of Palestine, Syria and North Africa.

P. Karge, from the University of Münster, Germany, manager of the Jerusalem office of the German-Catholic Görresgesellschaft, published a volume entitled *Repha'im* in 1917. He himself had also surveyed the valley of Ginnosar, west of Lake Kinneret, in the years 1909–1911 and discovered the important caves of Nahal Amud. The notable finds are not in the Valley proper, but in one of the major wadis leading to it, which is an integral part of the Jordan River drainage system. The book sold successfully, so that a second edition was printed in 1925.

Another area that for a long time had a fate similar to that of the Jordan Rift Valley, in terms of lack of prehistoric research, was the Negev desert. A survey by Woolley & Lawrence (1914/1915), for unknown reasons, failed to recognize the rich and varied surface occurrences of flint artifacts known today.

They concluded therefore that prehistoric Man had not inhabited the deserts during the Stone Age.

It is quite interesting to note that during World War I the collection of prehistoric artifacts was done mainly by army officers, both from the British and the German sides. However, again this did not occur in the Jordan Rift Valley. On the British side Major Ramsay found blades and flakes near Quseime in northern Sinai, while Van Riet Loewe, who later became an eminent South African prehistorian, collected hand axes near Wadi el Arish and Wadi Gaza, in the same region (Field 1956).

On the German side Wiegand (1920) also located considerable occurrences of flint artifacts near Quseime, assigning them to the Middle Paleolithic, Upper Paleolithic and Mesolithic cultures. Another German officer, J. Bayer (1918, 1922), collected hand axes and microliths in 1917 near Helez in the northwestern Negev, creating the term "Ashkelonian" in order to designate a very primitive culture, which lasted for a very long period, beyond the Paleolithic.

The state of biblical archaeology, during the first phase of research, was better as compared with prehistory. Most travelers and researchers not only collected, but also tried to identify the sites with those names known from the Bible or history. Thus Masada (Mezada) was already identified by Smith in 1838 (Robinson & Smith 1841), followed by many others that corroborated the identification, some of whom even published detailed plans of the superficial remains. Excavations at Mezada, however, did not begin before the 1950s. Most of the biblical sites along the entire Jordan Valley, including the important sites of Bet She'an, Hazor and Dan, had already been identified in the 19th century. However, except for a few which were unearthed during the first decades of the 20th century, most large-scale excavations did not begin until the 1950s.

2.2.2 The second phase: excavations

The time span separating the two World Wars (1918–1939) became a flourishing period for the archaeology, prehistory and geology of the region. The principles for all branches of archaeology, from the termination of the Stone Age through the Byzantine, were laid down in the years 1920–1935. These were also the years when many institutions involved with archaeological research were established. In Palestine, these were the governmental Department of Antiquities, the Department of Archaeology at the Hebrew University of Jerusalem, the British School of Archaeology and others, creating a most favorable academic milieu. Similar institutions were formed in the same period in the Kingdom of Jordan.

Following Lartet, Blanckenhorn and others, L. Picard, who was mainly involved in hydrogeological research, was the first to systematically study the Quaternary stratigraphy of Palestine, particularly in the Jordan Rift Valley and the coastal plain. He marked locations of numerous artifacts and reported his finds in his diverse papers (1932 and other publications). Among his many studies he described the outcrops at Ubeidiya (Picard 1932, 1943) in the central Jordan Valley (Fig. 12.1.4.1), termed "Melanopsisstufe" by Blanckenhorn, who discovered them in 1907. Blanckenhorn had assigned these layers to the late Pliocene (and at some stage, in 1914, to the "Diluvium", but back to the late Pliocene in 1931), following an earlier discovery of similar looking beds in northern Syria in 1897. Another geologist of Picard's times, N. Shalem (1937), also described many prehistoric artifacts and sites from the Galilee, Mount Carmel and the coastal plain. F. Brotzen (1928), geologist and micropaleontologist, also surveyed Mount Carmel and identified several sites there, during the years 1925–1926.

It seems that the first discovery of an Acheulian site within the limits of the Jordan Rift Valley proper was in 1924, by Major Bradock of the British Border Police, at Ma'yan Barukh, north of Hula Lake (called "Abel" at that time, after the Biblical name, or Sambariyeh in Arabic). The site was visited and reported by Maccurdy (1936), who was director of the American School of Prehistoric Research in Jerusalem. At the same time P.L.O. Guy (1924) published a short summary of prehistoric remains in the Hula Valley. This discovery seems to have opened new horizons for researchers of the Jordan Valley and its close vicinity. In 1925–1926, F. Turville-Petre (1927), who was motivated by the book *Repha'im* (Karge 1917, see above) excavated the caves of Zuttiyeh (Fig. 11.7.2) and Emireh, in Nahal Amud. He came upon a stunning discovery: a fragmentary hominid skull beneath the lowermost archaeological layer in Zuttiyeh. This is considered (Bar-Yosef, in prep.) one of the oldest hominid relics from the Near East, together with the later discovery in Ubeidiya (Tobias 1966). Another discovery of two fragments of human femur, possibly *Homo erectus*, by Geraads & Tchernov (1983), occurred while the two re-examined bone collections from the Benot Ya'aqov Formation, north of the bridge. These discoveries acted as a trigger for prehistoric studies of the Holy Land and enlivened the enthusiasm of many scholars to participate in uncovering the remote past of the region.

It is noteworthy that the first systematic excavation of a prehistoric site in Palestine, in 1925, is intimately connected with the Jordan Valley. This was the excavation of Zuttiyeh and Emireh caves in Nahal Amud, west of Lake Kinneret, by Turville-Petre. The discovery of human skull fragments in Zuttiyeh indicated that there are great possibilities for prehistoric research in this region. These were followed by excavations of sites, especially caves, in almost the entire area of Palestine.

From 1925 onward surveys and excavations frequently became separate activities, occasionally conducted by the same scholars, but also more and more by short-term visitors, including many priests and diplomats. One of the priests, the Czech J. Petrbok, published several reports on Paleolithic and Neolithic artifacts from the central Jordan Valley (Petrbok 1925, 1926a,b). Other clergymen, such as Father D. Buzy (1928) and P. Dovigneau, had been active in prehistoric surveys of the Holy Land, not in the Jordan Valley but rather in the Judean Hills and the

Negev desert, where they found, together with R. Neuville, a wealth of artifacts and sites previously unknown. Other findings of artifacts in the desert are reported by H. Field (1953) from surveys he carried out during the 1920s. At approximately the same time Lambert (1928) conducted the first excavation at El Wad Cave, on Mount Carmel. His finds and conclusions, long forgotten, have recently been revived by Weinstein-Evron (1998).

Three eminent prehistorians arrived in the Holy Land in the second half of the 1920s. R. Neuville was appointed a French consular clerk (Chancelier) to Jerusalem; M. Stekelis, expelled from the Soviet Union as a Zionist after spending three years in Siberia, emigrated to Palestine in 1927, and in 1936 founded the Laboratory of Prehistory at the Hebrew University of Jerusalem; D.A.E. Garrod, who later became known for excavating the Mount Carmel caves, came to Jerusalem in 1928, following completion of her fieldwork in Gibraltar. Of these three, only Stekelis became, in later years, intensely involved with prehistoric research in the Jordan Rift Valley. Ubeidiya in the central, and Banat Yaqub (Fig. 12.1.6.1) in the northern Jordan Valley are the two notable sites intimately connected with his name, sites which he excavated intensively with his collaborators for many years.

The discoveries by Picard of strata older than the Lisan both north and south of Lake Kinneret had attracted Stekelis' attention, and he commenced surveying these regions. Upon spotting several Lower Paleolithic artifact occurrences, the second stage of research began, with systematic, meticulous excavations. These involved detailed documentation, collection of all types of man-made materials, including debitage (and not only those nice looking pieces), collection of animal and sometimes vegetal remains, detailed geological and stratigraphic accounts; in all, the beginning of a truly multidisciplinary approach. This, naturally, involved collaboration with scientists studying different aspects of the sites, a habit that continues today in almost any prehistoric excavation or survey of the Jordan Rift Valley, to the shared benefits of all concerned. Unfortunately, such cooperation is not always the case in excavations of historic sites in the region, but this situation is constantly improving.

The systematic collections of rich and varied assemblages of artifacts enabled also comparisons with other sites and cultures elsewhere, especially in southern Europe and East Africa. The French chronological and cultural terminology was usually applied to prehistoric cultures of the Levant, with minor additions of local character.

The discovery of an important prehistoric site in the northern Jordan Valley, in the Gesher (bridge) Benot Ya'aqov area, called "Banat (or Benat) Yaqub" in Arabic, by Garrod and Gardner, occurred in 1935 while they were traveling in the region (Fig. 12.1.6.1). They spotted, in a dump created by heavy machinery while enlarging the narrow gorge of the Jordan River, some Acheulian artifacts. In the course of the same year Picard and Bates collected numerous mammalian bones at this locality, while Stekelis analyzed a section there, in an excavation of a sewage pit dug for the stables of the Border Police. In 1936 a rare opportunity occurred, when the Jordan River was dammed about 3.5 km north of the bridge at Benot Ya'aqov. Stekelis walked northward in the dried up channel, noting various concentrations of artifacts and bones. He started excavating the site of Banat Yaqub in 1937, together with the geologist P. Solomonica, but could not go on for long due to the outburst of riots in the country, followed later on by the onset of World War II.

The great enthusiasm shown by Stekelis, following Picard's studies, combined with the discovery of Banat Yaqub and his cooperation with G. Haas, who studied the fossil vertebrates, and with other collaborators, opened the way for systematic surveys and excavations in the Jordan Valley. These have been continued by later generations of their students, including the well-known names of D. Gilead, O. Bar-Yosef, E. Tchernov and N. Goren-Inbar.

This stage of research had actually commenced in the 1930s when both Neuville (published only in 1951) and Maccurdy (in Garrod & Bate 1937) realized the importance of the Levant as a crossroads between the three large continents of the Old World, Africa, Asia and Europe. Maccurdy still held the idea common in those times, that Man had originated in Asia and spread over to Europe, but he noted that the spread of knowledge of prehistory went the other way round, from west to east, being founded, as most other branches of science, in Europe. The adequate procedure for building a chrono-cultural framework for this newly revealed region was to excavate stratified sites. These were usually caves, which were thought to comprise the most important locale for shelters of prehistoric man during the Quaternary. As a result, from 1925 to 1942, 20 caves were excavated in the Levant, and only three open air sites. Since caves are not present in the Jordan Valley itself, but only occasionally in wadis leading to the depression, prehistoric research in this period mainly focused on the hilly regions of the Holy Land.

The classification of artifact assemblages into prehistoric cultures usually followed the lines applied in western Europe. Even though the French nomenclature became of common use in the Levant, several local terms had been suggested and most are still in use, for cultures with typical characteristics, such as the Natufian, Atlitian and others.

In this period of extensive excavations no conspicuous efforts were made to reconstruct the lifestyles of prehistoric man, and the main interest other than in the artifacts themselves was focused on migrations. These were reconstructed based on the assumption that lithic industries are directly identified with social units. Consequently, the geographic distribution of lithic cultural assemblages along a time trajectory was interpreted as evidence for movements of groups of people.

Detailed excavations were carried out in Bet She'an by G.M. FitzGerald in the years 1921–1933, a site occupied from the Neolithic almost continuously until the present day. Excavations of this important city are still going on. At the same time, A. Mallon (1932) excavated the important Chalcolithic site of Tuleilat el Ghassul, facing Jericho on the Jordanian side of the Rift Valley.

One of the most intriguing archaeological problems of the Arava Valley, south of the Dead Sea, concerns ancient copper mining (Glueck 1940). Indications of copper occurrences in this region are given at the end of the 19th and the beginning of the 20th century by several investigators, summarized in Blake (1937). Following their discoveries, a joint expedition of the American School of Oriental Research in Jerusalem, the Hebrew Union College in Cincinnati and the Department of Antiquities of the Government of Jordan had surveyed the Arava Valley in the spring of 1934. Relics of ancient copper mining had been found along the entire valley, from some 30 km south of the Dead Sea, all the way south to the Gulf of Aqaba. Two sites are particularly well known: Timna, some 30 km north of Elat, on the western rim of the Arava Valley; and Funon, approximately 100 km north of Timna, to the east of the valley.

The main and most widespread mining activities are known from the Iron Age, but the mines had already been exploited since Chalcolithic times almost continuously, by people inhabiting the nearby regions. Among them were the Egyptians, the Hebrews (the famous "King Solomon's Mines"), the Edomites, the Romans and later on the Arabs, especially during the Middle Ages (Glueck 1940). The Bedouins and other nomad tribes, who apparently did not possess the technique of extracting metal from the ores, used the occurrences for collecting semiprecious gem stones, principally malachite and azurite.

The expedition led by Glueck also located Etzion Gaver, on the northernmost tip of the Gulf of Aqaba, which served as the main port for King Solomon's copper trade with Arabia and the south. Excavations of copper producing settlements are still going on, on both the Israeli and Jordanian sides of the Arava.

The site of Kumran, at the northwestern tip of the Dead Sea, was discovered in 1947 when Bedouins found and sold several pieces of parchment scrolls, bearing scriptures. The discovery led to extensive excavations and surveys in the area from 1949 onward, resulting in a rich corpus of scrolls. Excavations of the important city of Hazor, in the Hula Valley, commenced in 1955. Dan, another site to the north of the Hula Lake, which was already identified by Robinson in 1838, was dug from 1963.

Archaeological surveys and excavations were also carried out in the central Jordan Valley in the early 1940s. Excavations at many sites such as Bet Yerah, an important tell at the southwestern tip of Lake Kinneret, started in these years.

2.2.3 The third phase: multidisciplinary research

Towards the end of World War II, in the early 1940s, Stekelis, together with B. Mazar and S. Yeivin, had surveyed the triangular area called "Biq'at Kinarot", limited by Lake Kinneret to the north, the Yarmouk and the Jordan rivers. Among the significant discoveries was the site of Sha'ar HaGolan, which consequently became known as the type locality for the Yarmoukian (Pottery Neolithic) Culture. Together with the geologist Solomonica, Stekelis also checked the dumps along the

Jordan River formed by stream channel enlargement, during the construction of the hydroelectric power station at Naharayim. There, they found some ancient-looking artifacts and predicted a possible future find of a very ancient site in the region, a prediction which materialized in 1959, with the discovery of human remains and artifacts at Ubeidiya.

In these years Picard (1943) published his monograph on the *Structure and evolution of Palestine*, a volume in which the Quaternary deposits of the country received considerable attention, with emphasis on those of the Jordan Rift Valley. For many years this work constituted the basis for the geochronology of prehistoric sites in the region. The last years of the 1940s saw the development of political unrest, riots and local wars, which hindered any prehistoric (or other) research in the country.

Research was resumed in 1949, following the ceasefire agreements between Israel and the Arab countries. The beginning of the 1950s involved great interest in prehistoric research, especially of Neolithic sites in the central and northern Jordan Valley. Two important excavations had been carried out in these years, Sha'ar HaGolan by Stekelis from 1949 to 1952 (Stekelis 1966a) and Eynan, a famous Natufian site ("Mallaha" in Arabic), by J. Perrot somewhat later. The latter became known due to the well-preserved structures and burial pits, including the burial of a man with his dog (Perrot 1960).

The main trend following the end of World War II involved creation of a chronological framework for the sediments in which artifacts and sites were located. For this reason, collaboration with paleontologists, stratigraphers and later on isotope scientists became indispensable. Archaeologists and prehistorians alike realized that, without an objective timing, there is hardly any significance to their conclusions regarding Man's behavior, evolution and migration. This tendency was manifested in the Jordan Rift Valley, especially in the Lower and Middle Acheulian sites of Ubeidiya and Banat Yaqub, where major efforts to study stratigraphy, vertebrate and invertebrate paleontology, palynology, paleomagnetism and radiometric dating resulted in a better understanding of the overall picture.

Fortunately, many prehistoric sites in the Valley have intimate connections with volcanic rocks, thus various radiometric and paleomagnetic methods were extensively applied in the last 30 years for their dating, which helped considerably in clarifying some crucial problems. The stratigraphic and chronological assignment of late Pleistocene–Holocene sites became much more simplified with the development and advance of radiocarbon dating.

In the same years (1954–1958) Kenyon excavated the site of Jericho, just north of the Dead Sea, unearthing a wealth of data pertaining to the early part of the Neolithic sequence. The tower, the walls, the structures, the plastered skulls, domesticated grains and animals, all illuminated in a revolutionary way the socio-economic changes of the period (Kenyon 1960). During the 1960s prehistoric research in the Jordan Valley had again been focused on older, Lower

Paleolithic sites. The accidental discovery of the site at Ubeidiya, with its wealth of artifacts, vertebrates, mollusks and some fragments of *Homo erectus* in 1959, was followed by 15 years of continuous excavations which, somewhat intermittently, are going on until the present day. The first stage of excavations, from 1960, was led by M. Stekelis (1966b), collaborating with the geologists L. Picard and U. Baida (1966a,b) and the paleozoologist G. Haas (1966). These were followed by their students O. Bar-Yosef, E. Tchernov and N. Goren-Inbar. The site of Ubeidiya is thus, at least for the present, the best-documented Lower Paleolithic occurrence in the Levant (Bar-Yosef & Goren-Inbar 1993). Huckriede (1966) reported finds of possibly similar artifacts at Abu Habil, on the eastern side of the Jordan Valley, some 30 km south of Ubeidiya (Fig. 11.2.3). Horowitz et al. (1973) reported some similar flakes found south of the Hula Valley, at Benot Ya'aqov bridge area. Neither of these was ever excavated, nor studied in detail.

It should however be stressed that Ubeidiya is not, contrary to common belief arising out of its public relations, the oldest site in the Levant or even, for that matter, in its close vicinity within the central Jordan Valley. Several sites older than Ubeidiya are known from the Lebanon (Hours 1975); three artifact occurrences are known in Israel, from Nahal Shiqma in the southern coastal plain, from the Erk el Ahmar Formation in the central Jordan Valley (Fig. 12.1.2.1), just south of Ubeidiya (Horowitz 1979, p. 296; Braun et al. 1991) and from the Galilee. A recent discovery by A. Ronen (1996) near Yir'on (Fig. 12.2), west of the Hula Valley, of several flakes in a gravel horizon underlying a basalt flow which yielded a radiogenic age of 2.4 million years, seems to stretch even further the antiquity of *Homo* in the Levant. A fourth, not yet finally proven, is from Nahal Zihor in the southern Negev (Ginat 1997) where flakes were found, possibly recovered from lake sediments (Fig. 11.5.2) correlative to the Erk el Ahmar Formation.

In the late 1960s a new and different approach to prehistoric research appeared in the region, when scientists became considerably interested in and had a much greater concern for the economic basis of human societies. The new approach also involved a wide application of quantitative methods, along with a large variety of detailed attribute analyses, in describing artifact assemblages (Bar-Yosef & Goren-Inbar 1993, among others). This strategy was also accompanied by site catchment analysis, and large-scale excavations were carried out chiefly for the retrieval of bones and plant remains. This new approach infiltrated quite quickly into prehistoric studies of the Jordan Valley, culminating at the beginning of the 1970s in the development of an overall environmental approach. This was manifested in a project in the southern Jordan Valley, north of Jericho (Bar-Yosef 1980, Schuldenrein & Goldberg 1981, and others). The "paleo-ethnological" approach, which was developed by A. Leroi-Gourhan, hardly found its way into prehistoric research in the Near East, except for an excavation of a "floor" of one of the earliest structures at Eynan, in the Hula Valley, by F. Valla (Bar-Yosef, in prep.).

The beginning of the 1960s also saw renewed interest in sites which had been studied before. D. Gilead (1970) went in Stekelis' footsteps and continued excavations at the Paleolithic sites of Banat Yaqub and Ma'yan Barukh. In Nahal Amud, just west of Lake Kinneret, S. Binford (1966) excavated the Shovakh Cave, while a Japanese expedition headed by H. Suzuki and H. Watanabe dug the Amud Cave during 1962–1964, an excavation which resulted in the discovery of a complete human skeleton (Suzuki & Takai 1970). Gisis excavated the terrace of the Emireh cave in 1973, and Zuttiyeh in 1974 (Gisis & Bar-Yosef 1974).

Since 1965 attempts had been made on the eastern side of Lake Kinneret to clarify and classify the Epipaleolithic sequence (Bar-Yosef 1975). In this framework the site of En Gev 1 was excavated by Stekelis & Bar-Yosef (1965), while sites such as En Gev 3 and others were excavated by various investigators. J. Perrot continued excavating Eynan during 1971–1975.

Various Neolithic and Chalcolithic sites have also been excavated in these years in the southern Jordan Valley, around the Dead Sea and in the Arava, to the south, including wadis leading to the basin, both in the Israeli and Jordanian parts of the region. One of the best known Neolithic sites is Beidha, excavated by D. Kirkbride (1966). Extensive work was carried out on the prehistory of the Jordanian region, starting in the late 1970s, by D. Henry (1986) and has continued ever since (Gebel et al. 1997, Henry 1998).

The foundations laid down by the pioneering studies have brought scientific activity to the present situation, when both archaeological and prehistoric excavations are being conducted all along the Jordan Rift Valley, both in the Israeli and Jordanian territories.

CHAPTER 3

The Jordan Rift Valley at present

The Jordan Rift Valley, as defined in this book, runs along some 350 km, from the northernmost point in Israel, in an almost straight north–south line, bending somewhat westward south of the Dead Sea, until the watershed at Gav Ha'Arava, some 70 km north of the Gulf of Aqaba (Fig. 1.3). It lies between longitudes 35° and 36° east, latitudes 30° and 33°20′ north. The rift structures continue both to the north and south of this region. The Dead Sea is the terminal, endoreic basin for the entire Jordan Valley, fed by numerous rivers and wadis leading to this hyper-saline lake, comprising a well-developed drainage system extending over wide areas (Fig. 1.2).

The Jordan Rift Valley is subdivided into eight morphotectonic units (Fig. 1.3), differing considerably from one another (the subdivision adopted here is not the only existing one, but seems the most favorable). The Hula Valley, approximately 24 × 7 km (Fig. 3.1) is the northernmost sector, attaining an altitude of some 70 m above sea level, bordered by mountains rising to almost 1,000 m to the east and west of the valley, some 2,800 m to the northeast, and several hundred meters to the north, serving as the main water source. The Hula Valley was, until artificially dried up quite recently, occupied by a shallow lake and extensive marshes, drained southward by the Jordan River.

The Korazim block, approximately 10 × 6 km, 300 and more meters above sea level, borders the Hula to the south (Fig. 3.2). The eastern side is incised by the Jordan River gorge, where the river descends more than 270 m over a distance of 14 km. To the west the Upper Galilee mountains rise to almost 900 m in altitude, to the east the Golan Plateau attains almost 1,000 m above sea level at its northern end, descending gradually southward into Biq'at (valley of) Kinarot, where it rises again to form the higher southern Golan.

Biq'at Kinarot, almost 30 km long, with a maximum width of 15 km but varying considerably (Fig. 3.3), lies more than 200 m below sea level, occupied by Lake Kinneret, which is up to 40 m deep. The lake is surrounded to the northeast, west and south by lowlands which, together with the lake, constitute Biq'at Kinarot. The Lower Galilee hills, several hundred meters high, border the valley to the

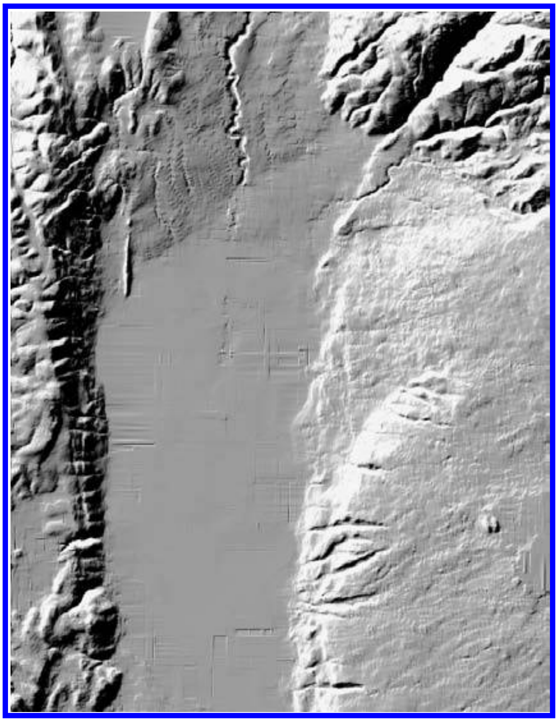


Figure 3.1. Image of the Hula Valley (coordinates 274–300N; 200–220E; 26 × 20 km). Derived from the 25 m DTM of Israel (Hall 1996). By permission of the Geological Survey of Israel.

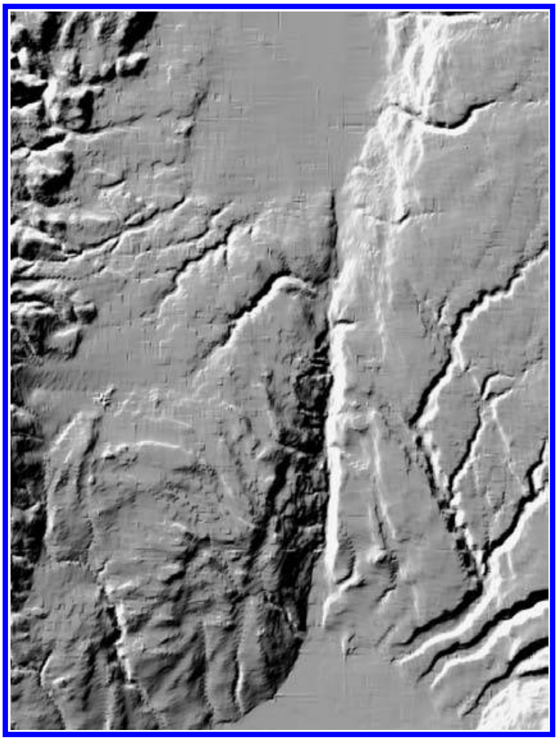


Figure 3.2. Image of the Korazim Highlands (coordinates 255–275N; 200–215E; 20 × 15 km). Derived from the 25 m DTM of Israel (Hall 1996). By permission of the Geological Survey of Israel.

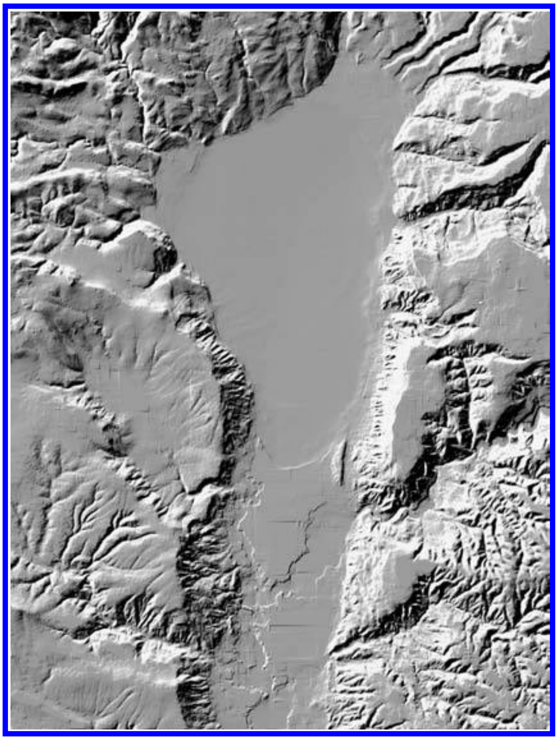


Figure 3.3. Image of Lake Kinneret and Biq'at Kinarot (coordinates 220–260N; 190–220E; 40 × 30 km). Derived from the 25 m DTM of Israel (Hall 1996). By permission of the Geological Survey of Israel.

west, but not too closely. To the east Lake Kinneret is bounded by the steep cliffs of the southern Golan Plateau, up to 500 m above the lake's level.

The lowland continues south some 35 km, in the rather narrow central Jordan Valley (Fig. 3.4). It is 6 km wide south of Biq'at Kinarot, but is joined by the somewhat elevated Bet She'an Valley about midway to its narrowest point, near Marma Feiyad, where the valley is only 3 km wide, approximately 250 m below sea level. The central Jordan Valley is bordered to the west by the Lower Galilee hills at its northern part, and by the Bet She'an Valley, some 100 m below sea level, at its southern half. To the east the Gil'ad Plateau rises gradually up to some 1,000 m above sea level. The Jordan River flows through the valley, continuing down to the Dead Sea via the southern Jordan Valley. Both parts of the Jordan Valley, flattened by deposition of the late Pleistocene Lisan Formation, are referred to by the local Arab inhabitants as the "Ghor". The river's narrow, deeper floodplain is called "Zor".

The southern Jordan Valley (Fig. 3.5) extends southward to the Dead Sea, some 60 km from Marma Feiyad, widening considerably on its way, gradually descending down to 350 m below sea level and more. It is almost 20 km wide near Jericho, the oldest, lowest city in the world. The valley is bordered westward by the Samaria (Shomeron) and Judea highlands that gradually ascend up to 900 m above sea level and eastward by the steeper, higher Moab Plateau.

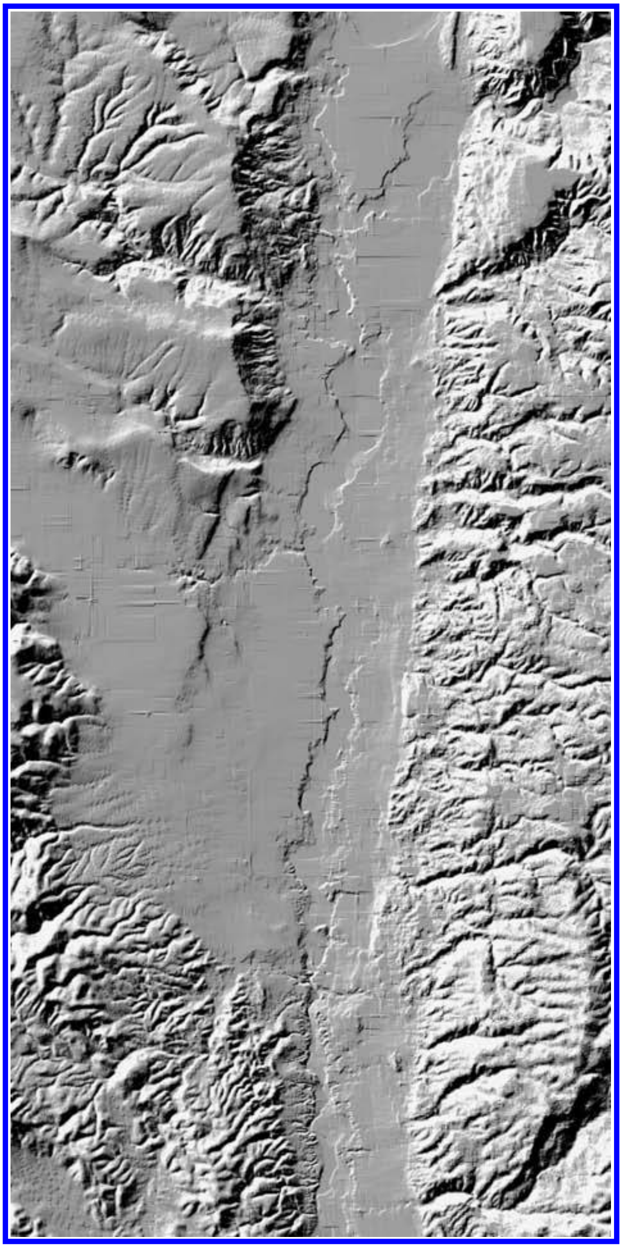


Figure 3.4. Image of the central Jordan Valley (coordinates 185–236N; 190–215E; 51 × 25 km). Derived from the 25 m DTM of Israel (Hall 1996). By permission of the Geological Survey of Israel.

The Dead Sea (Fig. 3.6) stretches some 75 km southward, its width varying from 10 to almost 15 km. Its "official" elevation is 392 m below sea level, but this figure varies annually with changing amounts of rain, as well as with water usage to the north and by the potash factories. The lake is bordered by steep fault scarps both east and west, several hundred meters in elevation. The approach from both north and south is gradual, rather flat. The Dead Sea basin itself comprises two

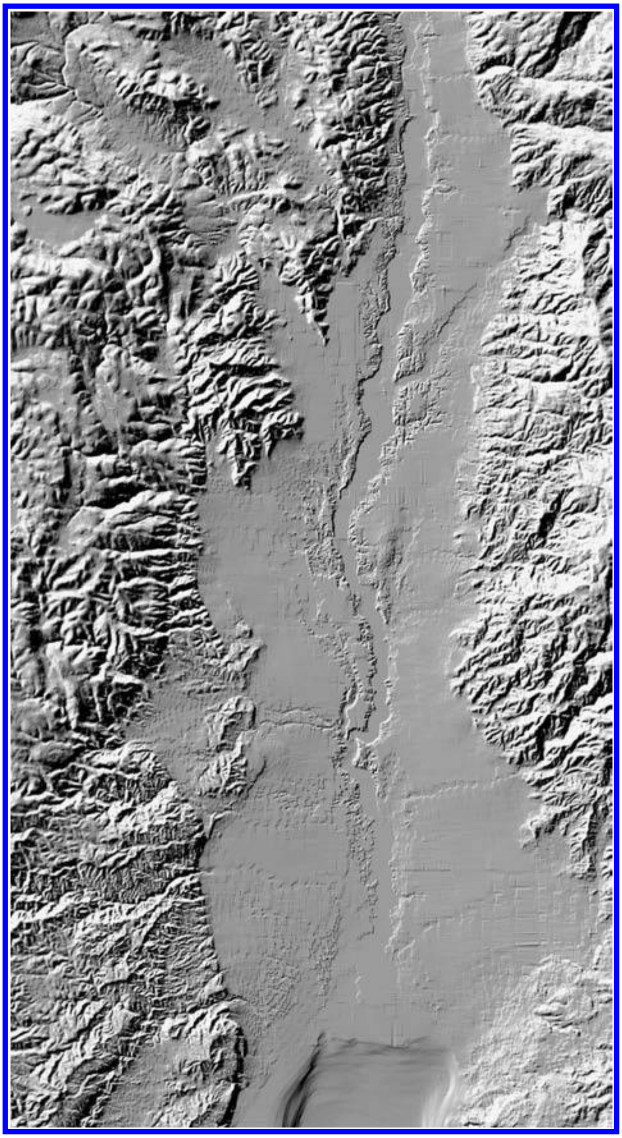


Figure 3.5. Image of the southern Jordan Valley (coordinates 125–190N; 180– 215E; 65 × 35 km). Derived from the 25 m DTM of Israel (Hall 1996). By permission of the Geological Survey of Israel.

distinct parts, the northern, whose depth is up to several hundred meters of water, and the southern, in which water depth does not exceed a few meters. The latter is actually dry today, occupied by potash evaporation ponds.

The northern Arava (Fig. 3.7), bending somewhat westward, is slightly more than 100 km long. The width varies considerably, from 15 km just south of the Dead Sea, to more than 30 km where Nahal Mashaq enters from the west and Wadi Fidan from the east, practically none at Gav Ha'Arava (back of the Arava) which is the watershed between the Dead Sea and the Gulf of Aqaba, 200 m above sea

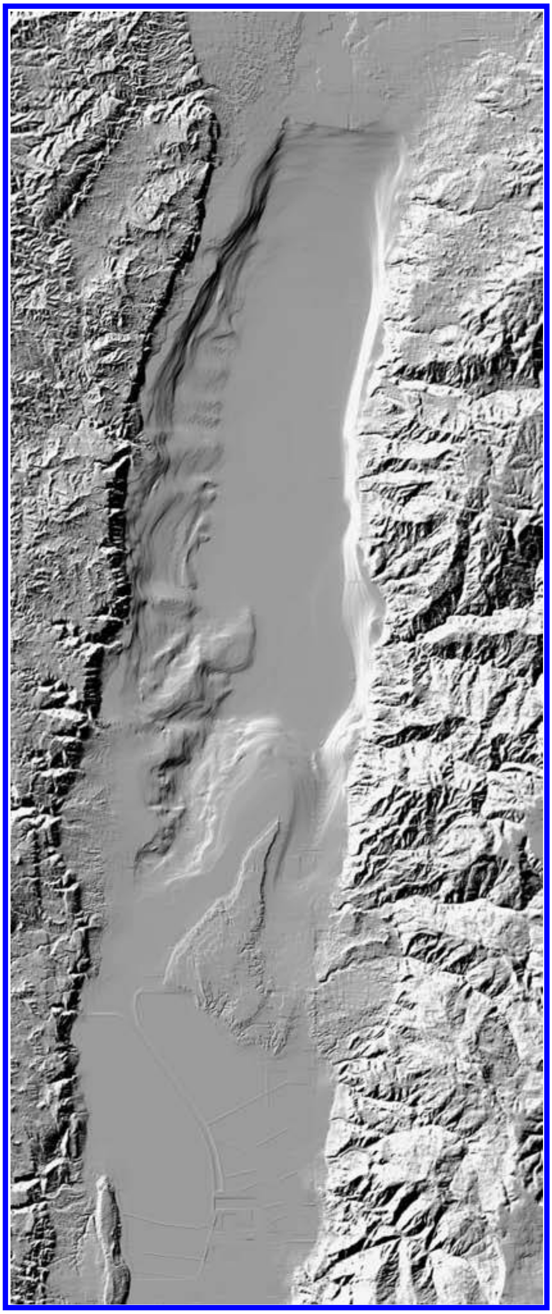


Figure 3.6. Image of the Dead Sea (coordinates 050–135N; 180–215E; 85 × 35 km). Derived from the 25 m DTM of Israel (Hall 1996). By permission of the Geological Survey of Israel.

level. The rise southward from the Dead Sea is gradual all the way, occupied by Nahal Ha'Arava (Wadi Arabah) which drains northward. The northern Arava rises slowly westward, toward the northern and central Negev fold belt, and somewhat more steeply to the east, toward the Edom Plateau.

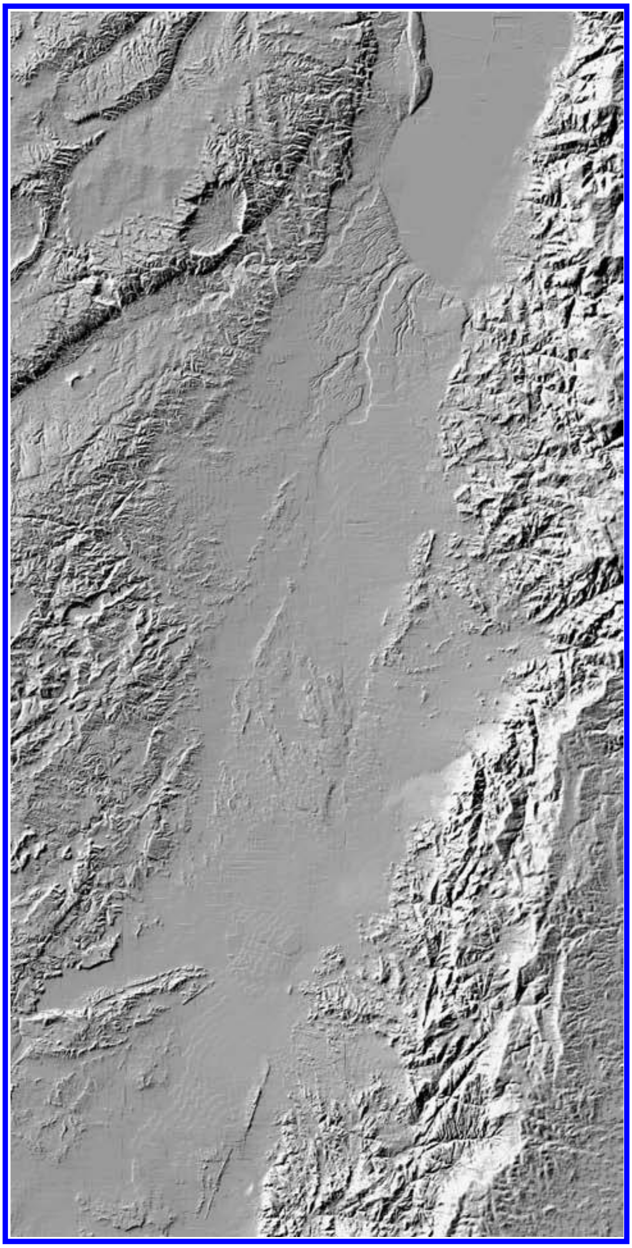


Figure 3.7. Image of the northern Arava (coordinates 950–060N; 150–205E; 110 × 55 km). Derived from the 25 m DTM of Israel (Hall 1996). By permission of the Geological Survey of Israel.

3.1 STRUCTURE

Horowitz (1979, Chapter 3), following previous authors, distinguished three main phases in the Cenozoic structural development of the Near East: the formation of the Levantine Fold Belt, which commenced some time during the late Cretaceous and was active at least until the Eocene (also known as the "Syrian Arc", see Chapter 4); the Eritrean Fault System (previously spelt "Erythrean") (or stage) of the late Miocene–Pliocene; and the Quaternary Levantine stage, which shaped the region in its present configuration, creating the Jordan Rift Valley depression, accompanied by uplifting of its shoulders both east and west of the Rift. An additional, Embryonic stage (see Chapter 9), is defined here to describe the structural changes that took place during the interval comprising the Oligocene, early and middle Miocene, typified by synclinal subsidence, predating the two principal faulting stages of the Jordan Rift.

Although frequently underestimated, uplift of the Jordan Valley shoulders parallel to its bearing, on both sides, is the main structural element responsible for morphologically shaping the predominantly synclinal depression (Horowitz 1979, p. 53, Sneh 1996). The axes of uplift are located some distance from the Valley, the latter thus acquiring the form of an elongated, narrow synclinorium, over which several downfaulted troughs are superimposed, forming the deeper basins. This combined north–south oriented uplift and down-bending affected all previous structures of both the bordering highlands and the depression, thus helping to accentuate the latter (Figs 3.1.1A and B). Besides, the Jordan Valley depression comprises four main taphrogenic elements: branching faults, bordering faults, crescentic faults and internal faults (Figs 3.1.1A and B, 10.2.1 and 10.2.5). The ages of these and their connections with the formation processes of the depression, are discussed in detail in Chapter 9.

The northern sector of the Jordan Valley depression crosses the earlier Levantine folds diagonally. In some localities synclines cross the Jordan Valley (Fig. 3.1.2), seemingly undisturbed by faulting. This is the case in the vicinity of Marma Feiyad, at the boundary of the central and southern Jordan Valley, and near Gav Ha'Arava, about midway between the Dead Sea and the Gulf of Aqaba (ten Brink et al. 1999, Bartov et al. 2000a, Horowitz 2001). It is however quite difficult to decide whether these elevated low structures are indeed Syrian Arc synclines, accentuated during later movements in the region, particularly the Embryonic stage, or an outcome of the consequent uplifting of both Rift shoulders.

It is quite apparent that the deeper basins, such as the Hula, the central Jordan Valley and the Dead Sea, where drillings revealed several kilometers of fill, occur where the Valley crosses previously elevated structures. The only exception appears to be Lake Kinneret, situated where the depression crosses a syncline. This is however misleading, because drillings in the lake only penetrated several tens of meters of fill, overlying late Cretaceous and Eocene rocks, indicating that this is, structurally, a rather shallow basin.

The branching faults (Fig. 3.1.3) are a suite of latitudinal structures, which are more pronounced along the southern part of the Jordan Valley, in the Dead Sea and the Arava Valley (Horowitz 1979, p. 53). They are usually tens of kilometers long, but some may even be longer. Their exact connection with the Levantine, north–south faulting which created the Jordan Valley in its present shape, is very much

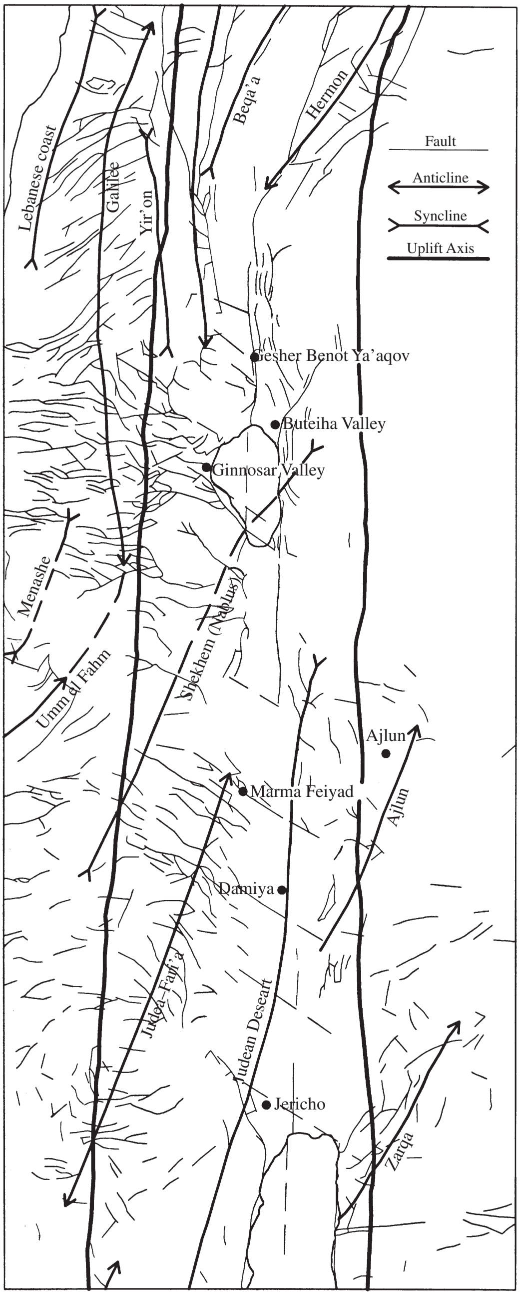


Figure 3.1.1A. Structure of the Jordan Rift Valley and neighboring regions, northern sector (compiled from: Avnimelech 1956, Bender 1974a,b, Kashai 1988, Heimann 1990, Bartov 1994, Garfunkel & Ben-Avraham 1996, Frieslander et al. 1997, Sneh et al. 1998a).

debatable and will be discussed in detail in Chapters 4, 9 and 10. The movement on these branching faults is basically horizontal (Fig. 3.1.4), dextral west of the Rift, most probably sinistral to the east, with lateral offsets in the order of up to a few kilometers (for details see Section 4.9). Consequently, some vertical displacements

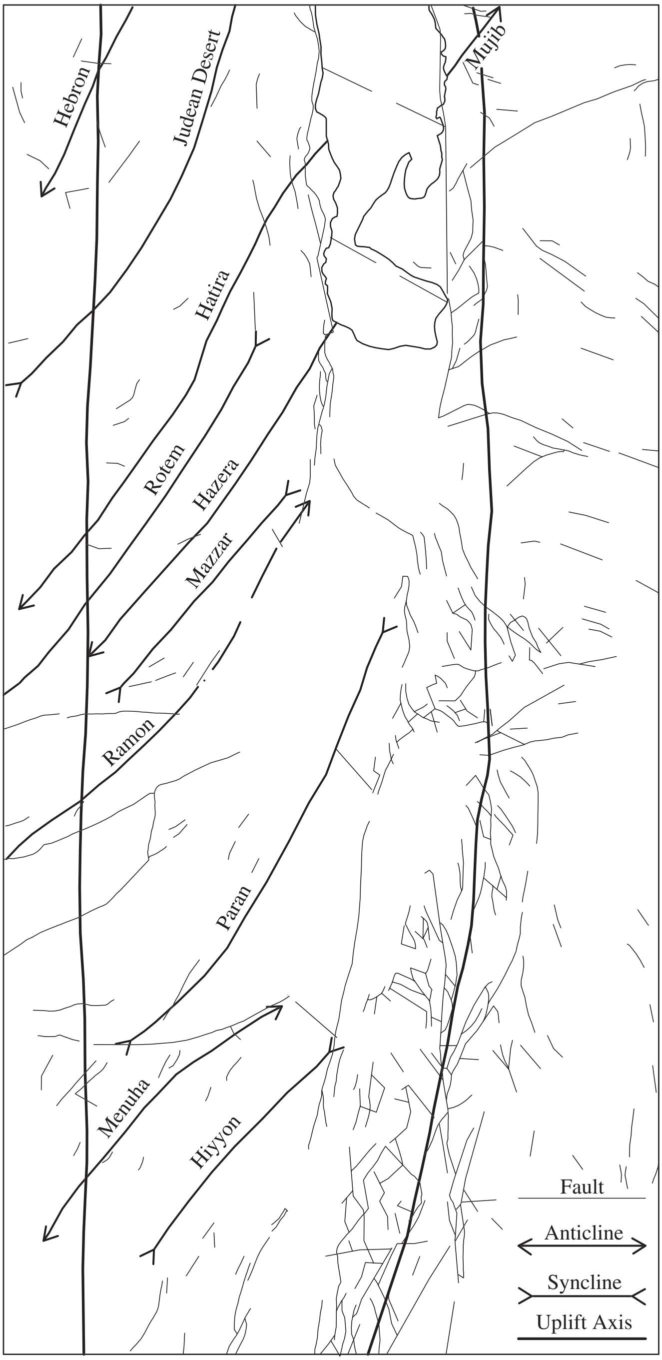


Figure 3.1.1B. Structure of the Jordan Rift Valley and neighboring regions, southern sector (compiled from: Avnimelech 1956, Bender 1974a,b, Kashai 1988, Heimann 1990, Bartov 1994, Garfunkel & Ben-Avraham 1996, Frieslander et al. 1997, Sneh et al. 1998a).

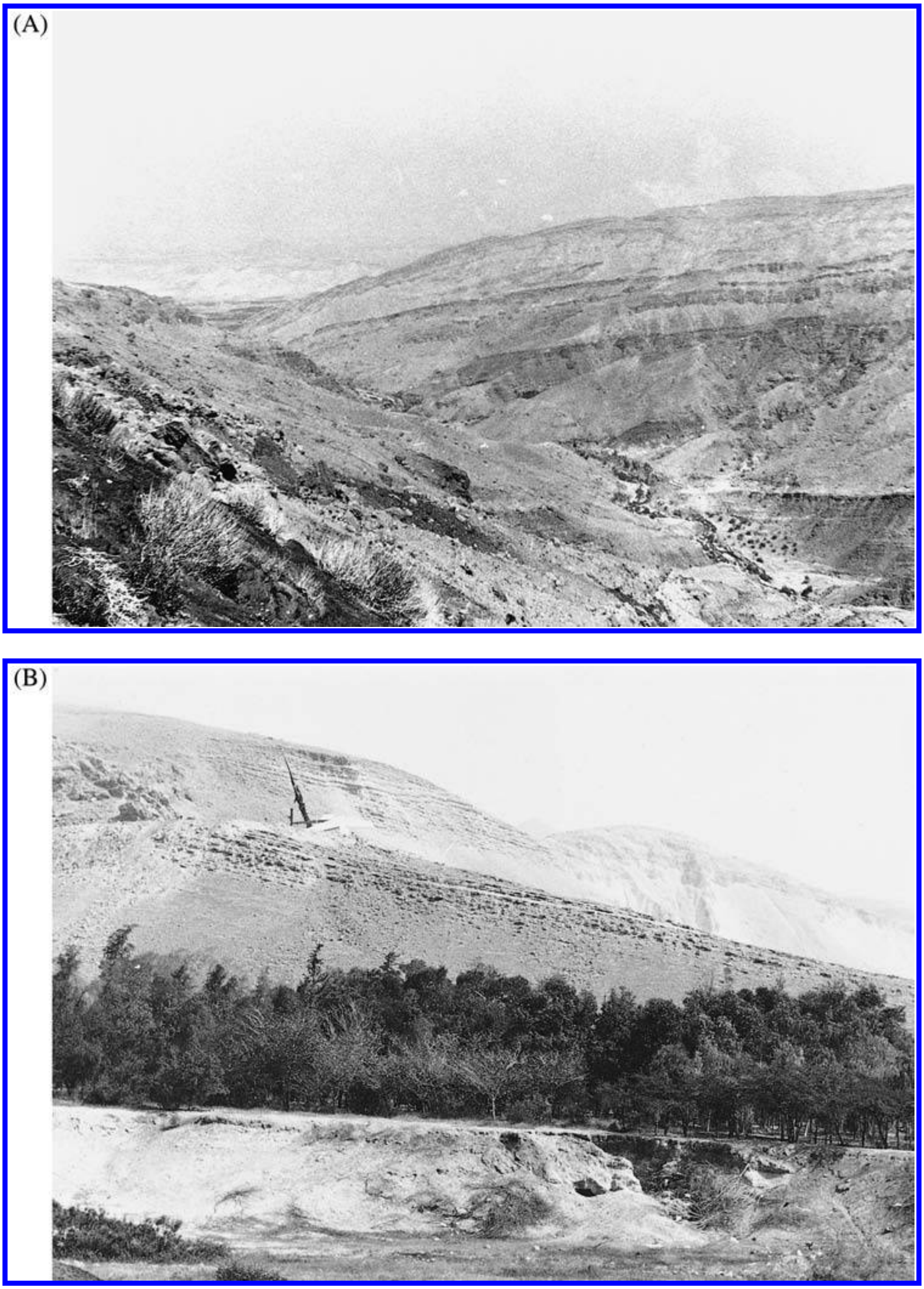


Figure 3.1.2. The southern Jordan Valley syncline north of the Dead Sea. (A) The eastern flank, dipping west. (B) The western flank, dipping east. See also Fig. 3.5.3. (C, next page) Image of the area (coordinates 170–190N; 190–220E; 30 × 20 km). Derived from the 25 m DTM of Israel (Hall 1996). By permission of the Geological Survey of Israel.

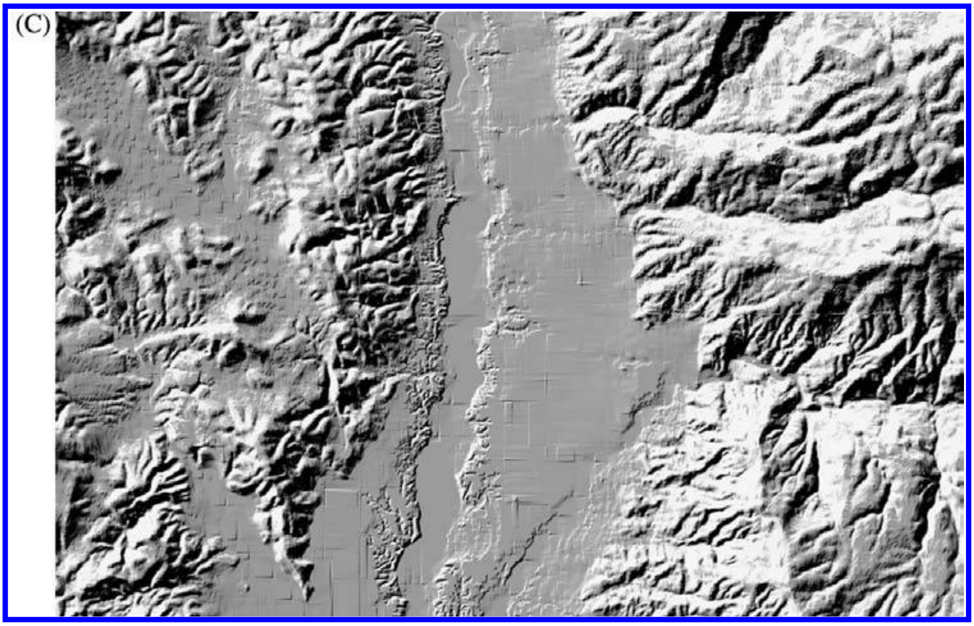


Figure 3.1.2. (*Continued*).

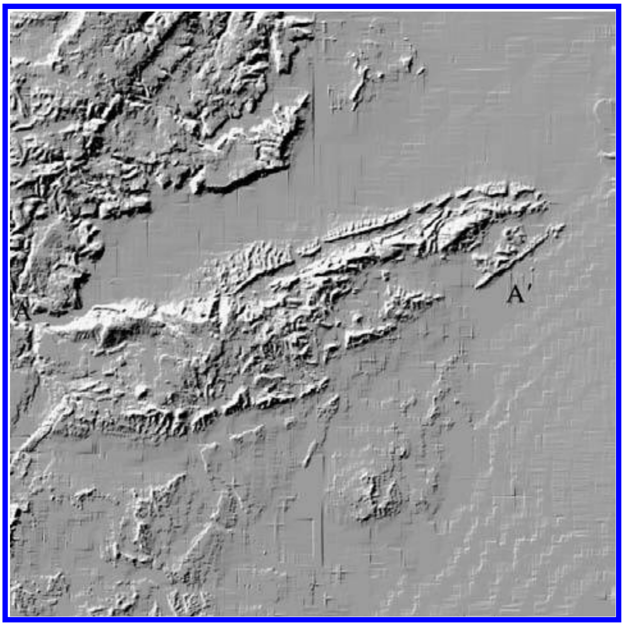
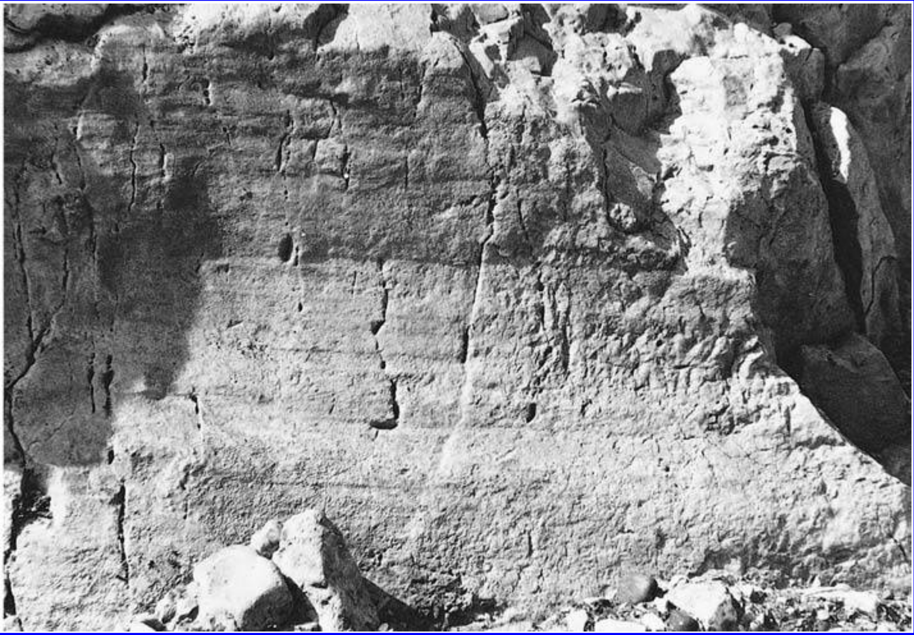


Figure 3.1.3. Image of the Paran Eritrean branching fault at the southern flank of Nahal Paran, approaching the Arava on the right. The fault runs east–west at the middle of the image (A′A′), approximately on coordinate 970N (coordinates 960–980N; 150–170E; 20 × 20 km). Derived from the 25 m DTM of Israel (Hall 1996). By permission of the Geological Survey of Israel.

(A)



(B)

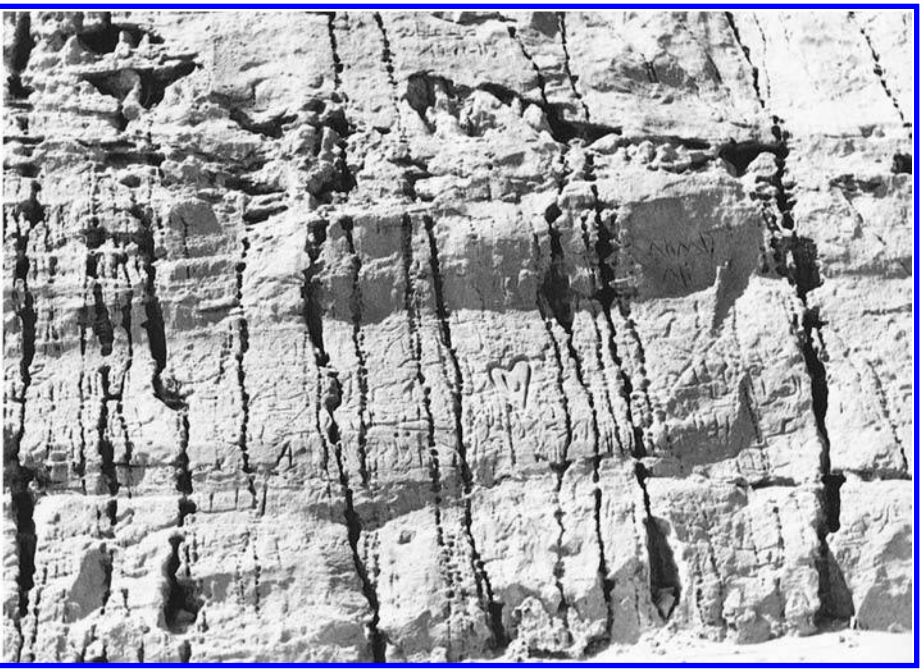


Figure 3.1.4. Horizontal slickensides on an east–west trending Levantine branching fault. (A) At Nahal Hemar, west of the Dead Sea, facing south (detail of the fault shown in Fig. 2.1.5). (B) At Wadi Gharandal, east of the central Arava, facing north.

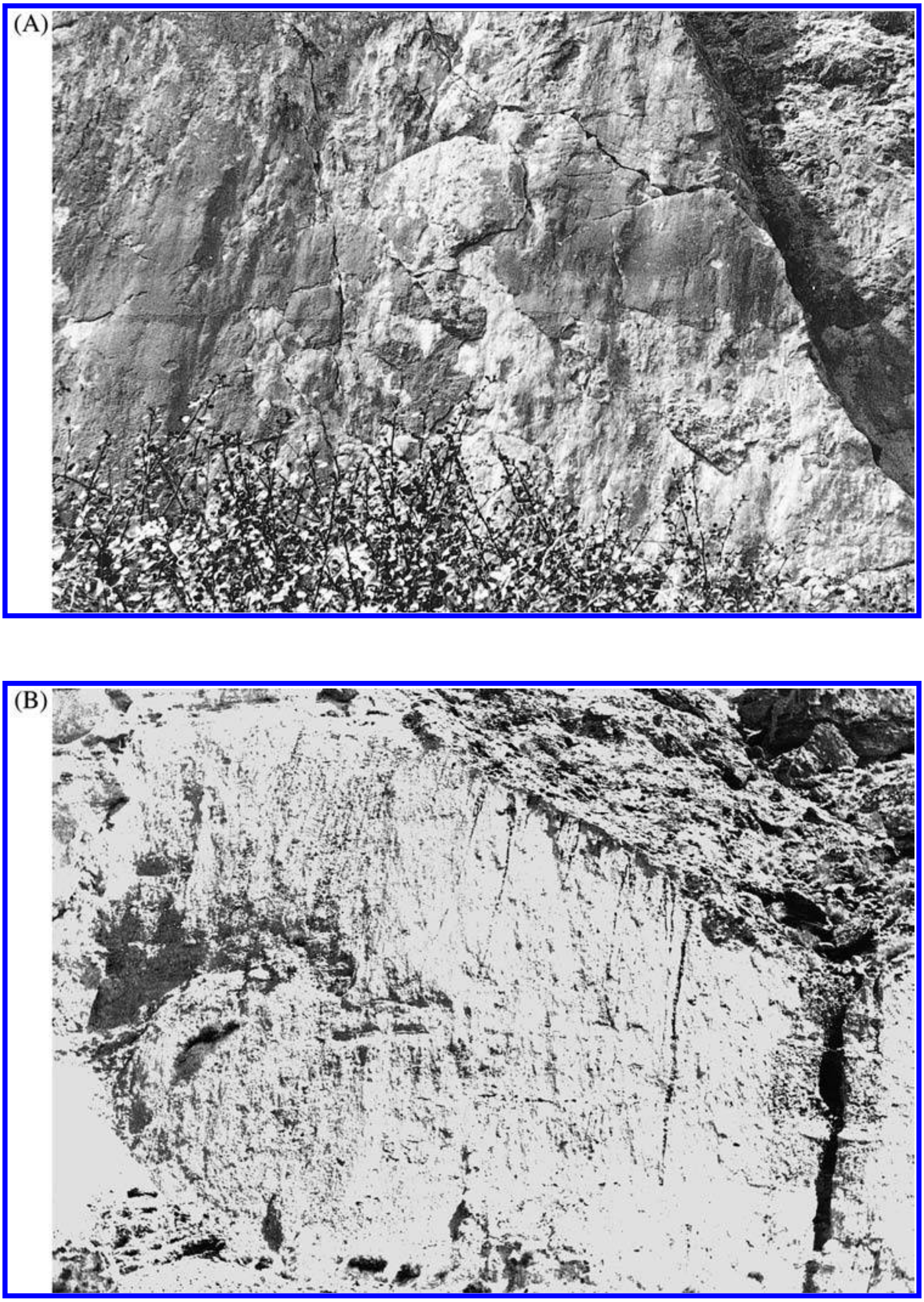


Figure 3.1.6. Vertical slickensides on the principal faults bordering the Dead Sea. (A) East facing at the northwestern end of the lake. (B) West facing at the northeastern shore.

are also evident. Steinitz & Bartov (1991) refer to these faults as the "Central Sinai-Negev Shear Zone", including in this group what they consider a continuation of this shear system into Transjordan.

Another set of branching faults, partly synchronous with the former but also showing more recent activity, accompanies the Jordan Valley, much more prominent in the sector between the Dead Sea and Lake Kinneret (Fig. 3.1.5). The general trend of this Eritrean system is NW–SE, usually constituting gravity faults, but some lateral movements can also be observed. These branching faults systems predate the Levantine faulting, which later created the north–south Rift, thus they do not form any morpho-structural elements within the Jordan Valley itself, only outside and approaching the depression.

The bordering and internal faults define and delineate the three deeper basins, the Hula, central Jordan Valley and Dead Sea. The first and last are occupied by lakes, whose extension greatly depends on water balance, so that the basins' structural limits do not exactly correspond to those of the water bodies. Lake Kinneret, as mentioned above, occupies a very shallow structural depression (although opinions vary, see Section 8.4). Typically, the southern parts of the bordering faults run in a general north–south direction (Fig. 10.2.5), while to the north they bend away from the depression, gradually dying out in the neighboring highlands. This phenomenon, termed by Picard (1931, p. 99) "crescentic faults", is almost symmetrical on both sides of the Valley. The bordering faults are not continuous along the Jordan Valley, and the longest is no more then several tens of kilometers. The throw along these faults, where the deepest parts of the basins are bordered, attains several kilometers. The fault scarps are very steep, almost vertical, and where

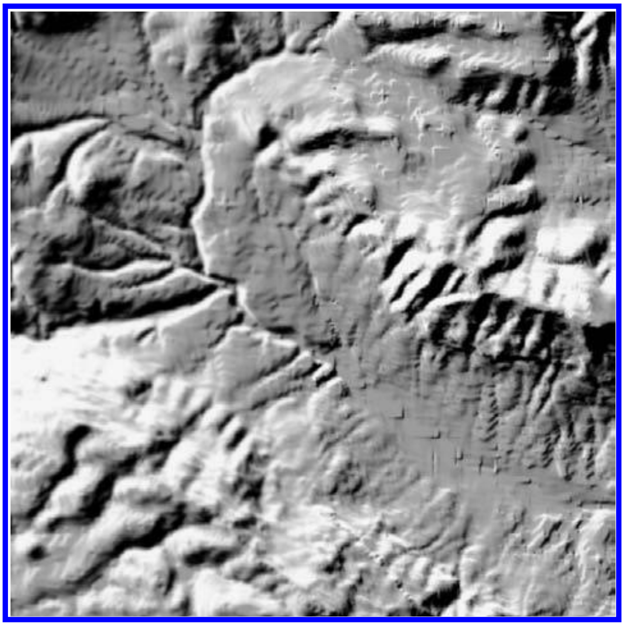


Figure 3.1.5. Image of Eritrean system faults, forming the Wadi Fari'a graben (coordinates 180–190N; 180–190E; 10×10 km). Derived from the 25 m DTM of Israel (Hall 1996). By permission of the Geological Survey of Israel.

these could be observed, display slickensides with either vertical or somewhat inclined striations (Fig. 3.1.6). Lateral offsets on the faults, wherever these could be ascertained, is usually dextral west of the Jordan Valley, sinistral to the east, in the order of hundreds of meters.

The internal faults hardly ever appear on the geological maps, since most of them are covered by young alluvium, occasionally by lacustrine sediments, sometimes by the lakes' waters. They are, however, extremely important in defining the limits of the basins (Fig. 3.1.7). The general direction of the internal faults is NW–SE, they occasionally bend, and at least several of them are considered listric by some authors (Kashai 1988, Frieslander et al. 1997), while others maintain all these faults are normal (see Chapter 8). The throw can be considerable, up to several kilometers, but small scale internal faults are also known (Heimann 1990). These faults were for a long time considered as never extending beyond the bordering ones, but a recent RADARSAT survey shows some of them to stretch out beyond the north–south oriented bordering faults which define the Rift's margins (Arkin et al. 1999a,b).

Occasionally, crescentic faults could be mistaken for internal ones, such as that bordering the southwestern shore of Lake Kinneret, since they may acquire similar strikes. To further complicate the picture, this bearing also characterizes the older, Eritrean fault system in the region, on both sides of the Jordan Valley, extending quite far into the neighboring highlands (Fig. 10.2.1), so that the crescentic faults are occasionally treated as part of the transversal fault system (see Section 4.9).

The internal faults system subdivides the Jordan Rift Valley into distinct basins, separated by elevated terrains. Some of the basins, like the Hula, Lake Kinneret

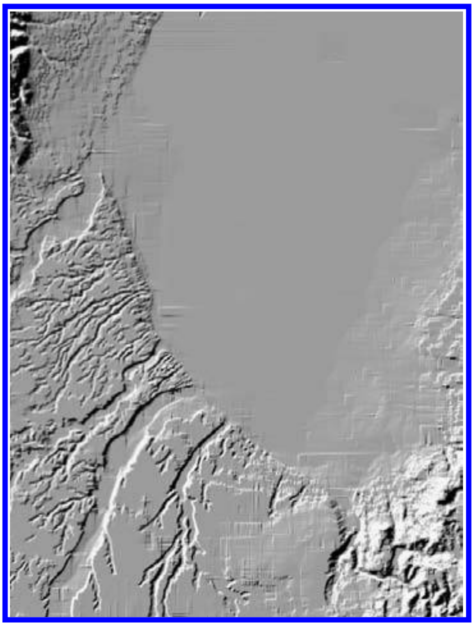


Figure 3.1.7. Image of the scarp of the internal Amazyahu fault, south of the Dead Sea (coordinates 030–050N; 180–195E; 20 × 15 km). Derived from the 25 m DTM of Israel (Hall 1996). By permission of the Geological Survey of Israel.

and the Dead Sea, are obvious from looking at the maps. Others, no less important, are known only from below the surface, from drillings and geophysical surveys. The most prominent seems to be the central Jordan Valley, where more than 4.5 km of fill were penetrated, without reaching bedrock. Similarly, the northern Arava is subdivided, based on geophysics, into the Dead Sea basin to the north, with estimated thickness of more than 2,000 m of fill, and southward into the Shezaf and Zofar basins, with diminishing sequences (Frieslander et al. 1997).

3.2 GEOLOGY

The deeper parts of the Jordan Valley, which constitute most of its extension, are covered either by lakes or by late Pleistocene–Holocene fill, of lacustrine, fluviatile or mixed origin. Several exceptions are some elevated blocks within the valley, such as Korazim, Marma Feiyad, Ghor el Qatar, Mount Sedom, the central Arava and others, where older Quaternary or Neogene formations make up the valley floor. These rock units are discussed in detail in Chapters 5 and 11.

The rock formations making the rift shoulders span almost all periods, from the Precambrian through the late Tertiary. Comparing east and west, the shoulders are quite different from one another (Figs 3.2.1–3.2.4).

The eastern flanks are progressively younger from the south, where Precambrian rocks line up the Arava and southern Dead Sea, northward where the Golan Plateau is covered by Quaternary volcanics. The western flanks display an inverted situation, where Neogene and Eocene rocks border the central and northern Arava, grading to early Cretaceous at the Hula. Superimposed on these general trends, on both sides of the rift, are the Levantine folds, which cause older rocks to crop out where anticlines approach the Jordan Valley, while younger formations typify synclines.

The Hula Basin is covered with recent sediments and soils. Its eastern flanks are chiefly made from Quaternary volcanics of the Golan Plateau, while Cenomanian and some early Cretaceous formations constitute the western cliffs. To the north, Cretaceous and Jurassic strata occur in the cores of the Lebanon and Hermon anticlines, Eocene and Pliocene in the Beqa'a syncline. The Korazim block to the south exposes some early Quaternary fluvio-lacustrine and volcanic layers, overlying Pliocene basalts.

These Pliocene basalts make up most of the Korazim highlands, which border the Golan Quaternary volcanics to the east, separated by the Jordan gorge fault. Eocene limestone and chalk dip under the Korazim basalts from the west, with no apparent major faulting. At the northeastern tip of the block, in the Benot Ya'aqov bridge area, an array of faults and very small blocks exposes almost the entire Quaternary sequence of the Hula basin, in which several important prehistoric sites are embedded.

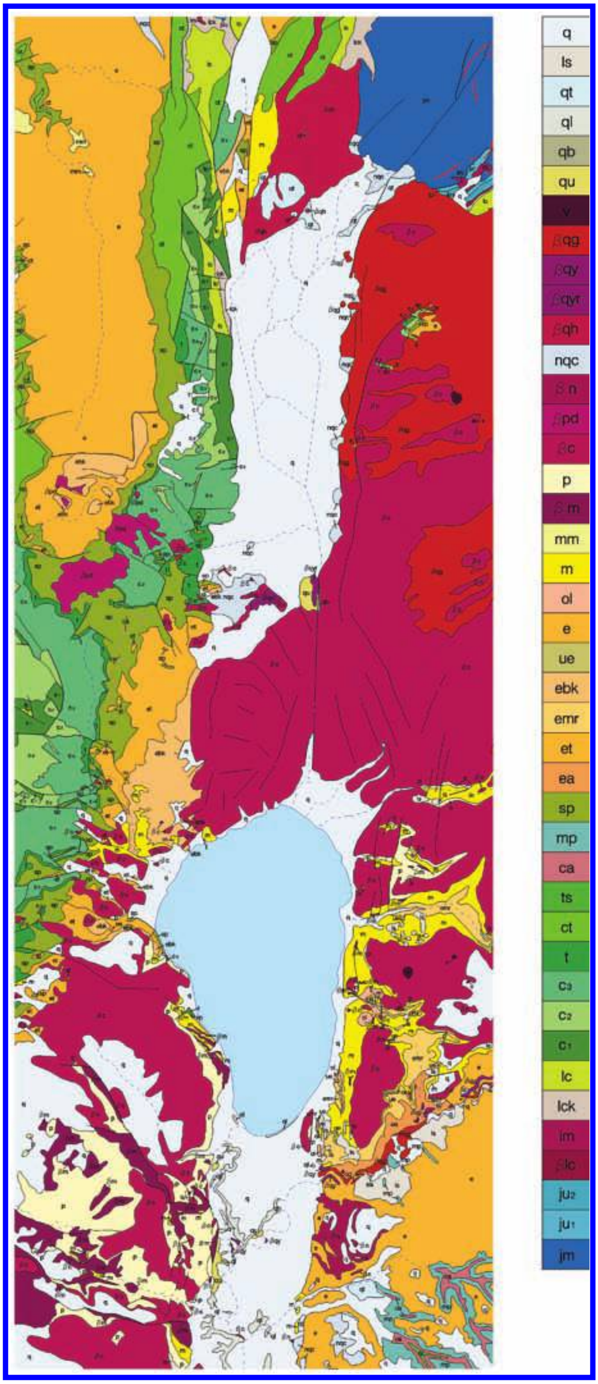


Figure 3.2.1. Geological map, northern part of the Jordan Valley (coordinates 220–305N; 190–220E; mapped area 85 × 30 km). From: Sneh et al. 1998a. The geological map of Israel, 1 : 200,000. Jerusalem: Israel Geological Survey. By permission of the Geological Survey of Israel. (q) Alluvium, Mallaha and Ashmura Fms. (Hula), Tabgha Fm. (around Lake Kinneret), Holocene; (ls) landslide, Quaternary; (qt) Dan and Kefar Yuval travertines, Quaternary; (ql) Lisan Fm., late Quaternary; (qb) Benot Ya'aqov Fm., Quaternary; (qu) Gadot and Mishmar HaYarden Fms. (Hula), Erk el Ahmar and Ubeidiya Fms. (central Jordan Valley), Quaternary; (v) Volcanic cone, Quaternary; (βqg) Golan and Raqqad basalts, Quaternary; (βqy) Yarmouk Basalt, Quaternary; (βqyr) Yarda Basalt, Quaternary; (βqh) Hasbani Basalt, Quaternary; (nqc) undivided conglomerate units, Neogene–Quaternary; (βn) undivided Neogene volcanics; (βpd) Dalton Basalt, Pliocene; (βc) Cover, Intermediate and Dalwe basalts, Neogene–Quaternary; (p) Bira and Gesher Fms., Pliocene; (βm) Lower Basalt and Late Miocene Volcanics, Miocene; (mm) marine carbonate units in Lebanon, Miocene; (m) Tel Hai Fm. and Tanur Conglomerate (Hula), Pliocene; Herod Fm. and Umm Sabune Conglomerate (central Jordan Valley), Miocene; (ol) Fiq and Susita Fms., En Gev Sands, Oligocene; (e, ue, ebk, emr, et, ea) various Eocene Fms.; (sp, mp, ca) Mount Scopus Group, Senonian–Paleocene; (ts, ct) undivided Cenomanian–Turonian–Santonian (Lebanon and Transjordan); (t) Turonian Fms. (Israel); (c3, c2, c1) Cenomanian Fms.; (lc, lck) early Cretaceous Fms.; (im, βlc) igneous rocks, Mesozoic; (ju2, ju1, jm) Jurassic Fms. solid lines represent faults.

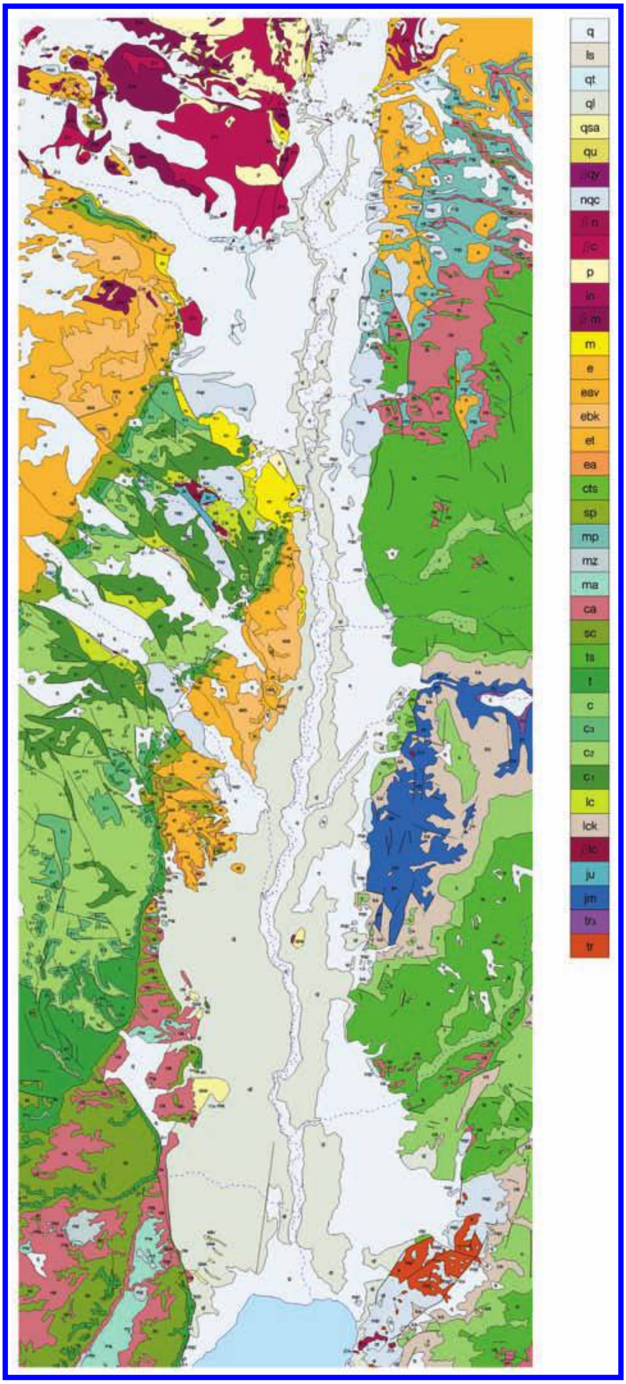


Figure 3.2.2. Geological map, central and southern Jordan Valley (coordinates 125–230N; 180–220E; mapped area 105 × 40 km). From: Sneh et al. 1998a. The geological map of Israel, 1 : 200,000. Jerusalem: Israel Geological Survey. By permission of the Geological Survey of Israel. (q) Alluvium, Fatza'el and Damiya Fms., late Quaternary; (ls) landslide, Quaternary; (qt) Bet She'an and other travertines, Quaternary; (ql) Lisan Fm., late Quaternary; (qsa) Samra and Oolitic Fms., Pliocene; (qu) Erk el Ahmar and Ubeidiya Fms., Quaternary; (βqy) Yarmouk Basalt, Quaternary; (nqc) undivided conglomerate units, Neogene–Quaternary; (βn) undivided Neogene volcanics; (βc) Cover and Intermediate basalts, Neogene–Quaternary; (p) Bira and Gesher Fms., Pliocene; (in) intrusive and pyroclastic rocks, Miocene; (βm) Lower Basalt and Late Miocene Volcanics, Miocene; (m) Hazeva and Dana Fms., Miocene; (e, eav, ebk, et, ea) various Eocene Fms.; (cts) undivided Cenomanian–Turonian– Senonian (Transjordan); (sp, mp, mz, ma, ca, sc) Mount Scopus Group, Senonian–Paleocene; (ts) undivided Turonian–Santonian (Transjordan); (t) Turonian Fms. (Israel); (c) undivided Albian–Cenomanian (Transjordan); (c3, c2, c1) Cenomanian Fms.; (lc, lck) early Cretaceous Fms.; (β1c) igneous rocks, Mesozoic; (ju, jm) Jurassic Fms.; (tr3) late Triassic (Israel); (tr) Permian and Triassic Fms. (Transjordan); solid lines represent faults.

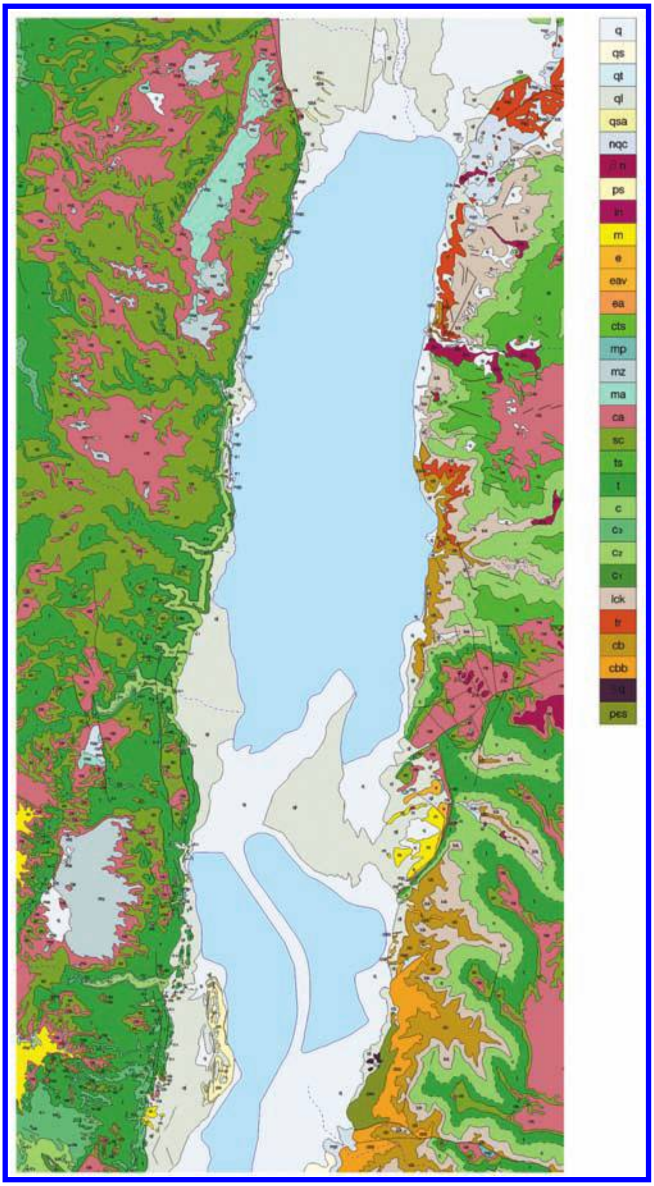


Figure 3.2.3. Geological map, Dead Sea (coordinates 045–140N; 170–215E; mapped area 95 × 45 km). From: Sneh et al. 1998a. The geological map of Israel, 1 : 200,000. Jerusalem: Israel Geological Survey. By permission of the Geological Survey of Israel. (q) alluvium, Holocene; (qs) sand dunes, Holocene; (qt) travertines, Quaternary; (ql) Lisan Fm., late Quaternary; (qsa) Samra and Oolitic Fms., Pliocene; (nqc) undivided conglomerate units, Neogene–Quaternary; (βn) undivided Neogene volcanics; (ps) Sedom and Amora Fms., Pliocene; (in) intrusive and pyroclastic rocks, Miocene; (m) Hazeva and Dana Fms., Miocene; (e, eav, ea) various Eocene Fms.; (cts) undivided Cenomanian– Turonian–Senonian (Transjordan); (mp, mz, ma, ca, sc) Mount Scopus Group, Senonian–Paleocene; (ts) Turonian–Santonian (Transjordan); (t) Turonian Fms. (Israel); (c) Albian–Cenomanian (Transjordan); (c3, c2, c1) Cenomanian Fms.; (lck) early Cretaceous Fms.; (tr) Permian and Triassic (Transjordan); (cb, cbb) Cambrian Fms. (Transjordan); (δq, pϵs) Precambrian Fms. (Transjordan); solid lines represent faults.

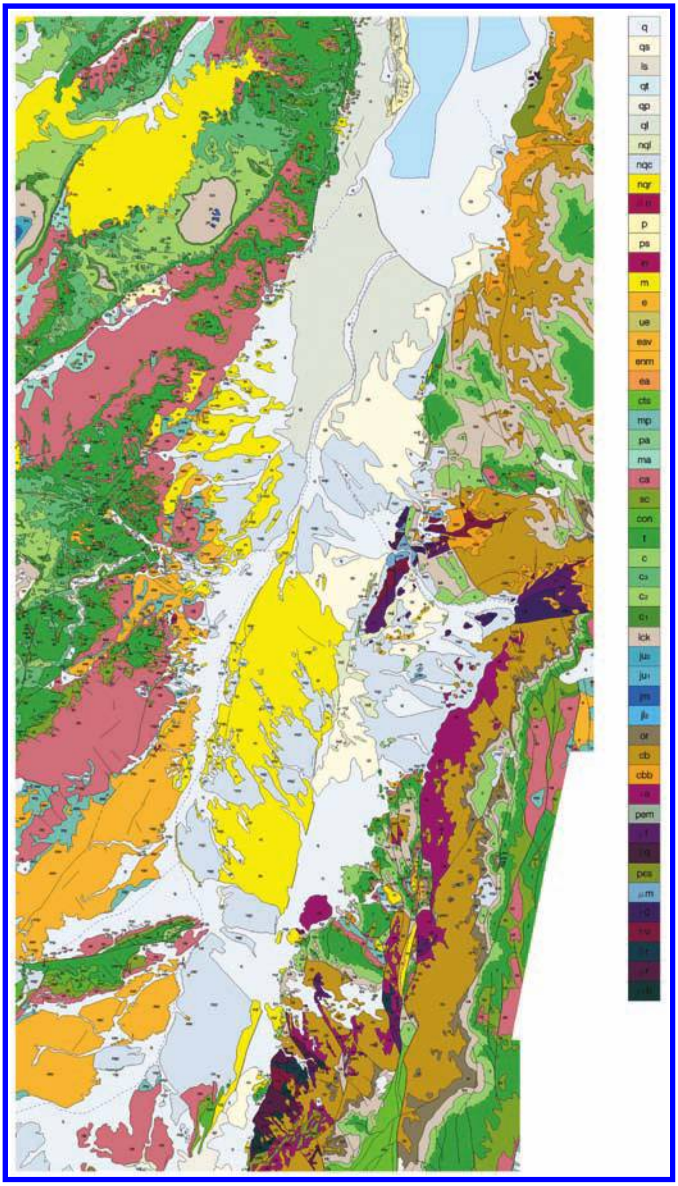


Figure 3.2.4. Geological map, central and northern Arava (coordinates 950–060N; 150–205E; mapped area 110 × 55 km). From: Sneh et al. 1998a. The geological map of Israel, 1 : 200,000. Jerusalem: Israel Geological Survey. By permission of the Geological Survey of Israel. (q) alluvium, Holocene; (qs) sand dunes, Holocene; (ls) landslide, Quaternary; (qt) travertines, Quaternary; (qp) playa deposits, Quaternary; (ql) Lisan Fm., late Quaternary; (nql) Zehiha Fm., Quaternary; (nqc) undivided conglomerate units, Neogene–Quaternary; (nqr) Ar Risha Gravel (Transjordan), Neogene–Quaternary; (βn) undivided Neogene volcanics; (p) Mazzar Fm., Pliocene; (ps) Sedom and Amora Fms., Pliocene; (in) intrusive and pyroclastic rocks, Miocene; (m) Hazeva and Dana Fms., Miocene; (e, ue, eav, enm, ea) various Eocene Fms.; (cts) undivided Cenomanian– Turonian–Senonian (Transjordan); (mp, pa, ma, ca, sc) Mount Scopus Group, Senonian– Paleocene; (con) Zihor Formation, Coniacian; (t) Turonian Fms. (Israel); (c) Albian–Cenomanian (Transjordan); (c3, c2, c1) Cenomanian Fms.; (lck) early Cretaceous Fms.; (ju2, ju1, jm, jl2) Jurassic Fms.; (or, cb, cbb) Ordovician and Cambrian Fms. (Transjordan); (νa, pϵm, γf, δq, pϵs, μm, νg, πu, δr, μr, $\gamma\nu$b) various Precambrian Fms. (Transjordan); solid lines represent faults.

Biq'at Kinarot is largely occupied by Lake Kinneret, surrounded by Oligocene, Miocene and Pliocene volcanics, fluviatile and lacustrine beds. Late Cretaceous rocks are exposed at some small-scale elevated blocks west of the lake. To the northeast the Buteiha valley, to the west the Ginnosar valley, both of subordinate size, and to the south the continuation of the Jordan Valley, are covered by late Quaternary to recent sediments and soils.

The line of small elevated blocks continues also south of Lake Kinneret, in the northern part of the central Jordan Valley, where they border the valley floor, which is covered by the late Pleistocene Lisan Formation. These blocks expose early Quaternary formations such as Ubeidiya and Erk el Ahmar, in which important prehistoric remains were discovered. Miocene and Pliocene basalts, as well as lacustrine and lagoonal sediments, border the central Jordan Valley to the west, and southward to Bet She'an Valley, which is covered by late Quaternary sediments. Further south, on the western rims, Eocene and late Tertiary rocks form the closure at Marma Feiyad. To the east this sector is bounded by cliffs of Eocene limestone and Pliocene basalts, except for the wide Yarmouk River valley, where Quaternary beds and volcanics occur.

The southern Jordan Valley, covered by the Lisan Formation, meets on its western flanks gently dipping Eocene rocks, occasionally covered by Neogene conglomerates and Quaternary lake terraces. Southward, from about midway to the Dead Sea, the Eocene is replaced by Senonian chalk, still with occasional outcrops of the younger sediments. To the east this sector borders the huge anticlinorial structure of Ajlun, where the deep Wadi Zarqa gorge exposes the entire Mesozoic sequence, from the Triassic onward. As in the west, this structure dips into the rift, but more steeply.

The western cliffs of the Dead Sea comprise late Cretaceous limestones, covered in places by various late Cenozoic conglomerates. The eastern flank consists of Paleozoic through early Cretaceous Nubian Sandstones, grading southward to Precambrian magmatic and metamorphic rocks. The restricted low lying areas around the Dead Sea which are not submerged under the lake's water are covered by the Lisan and younger lacustrine and wadi deposits.

The northern Arava is covered by Lisan and older lake and spring sediments and terraces. These become progressively older southward, where most of the area exposes the Miocene Hazeva Formation. The eastern cliffs are mainly composed of a variety of Precambrian rocks and Nubian Sandstones, while to the west gently dipping Eocene and Miocene beds vaguely define the limits of the Arava Valley.

3.3 SEISMICITY

The present-day seismicity of the Jordan Valley (Fig. 3.3.1) seems quite strange at first glance. One would expect from a recently shaped rift, or a contemporary

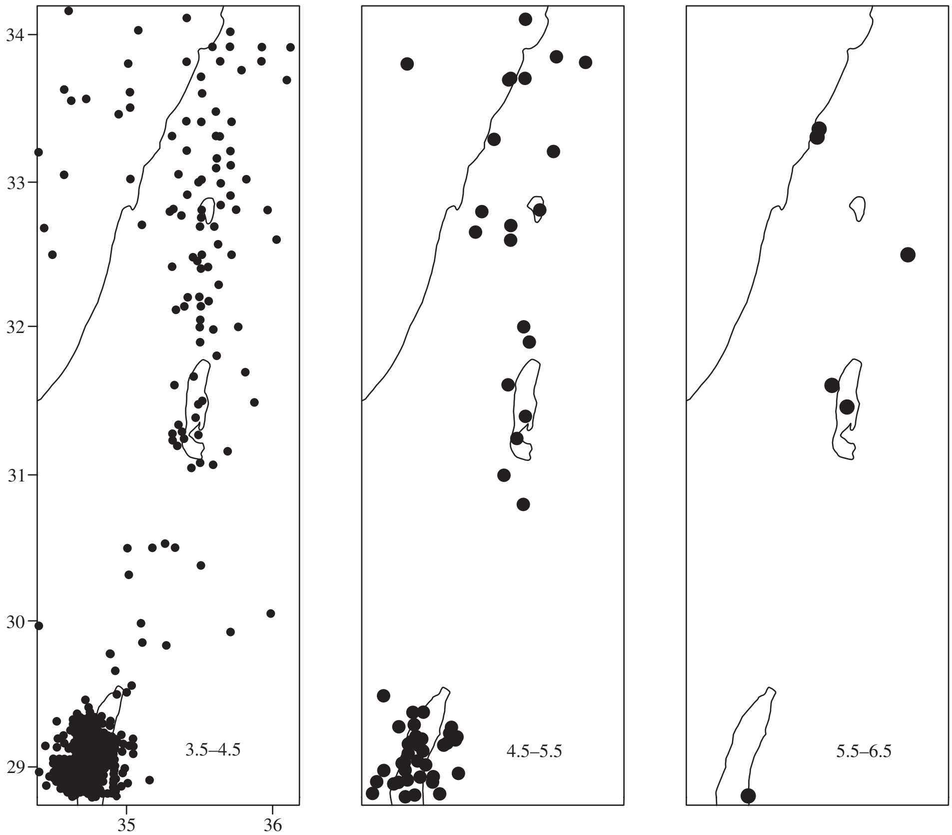


Figure 3.3.1. Seismicity of the southern Levant. Earthquakes epicenters from 1900 through 1996, according to magnitudes. After Amiran (1951), Arieh (1967, 1985), Feldman et al. (1997).

one, a much stronger seismicity, distributed along this structure, particularly considering that most students believe it to be an active transform fault system (see Appendix). Indeed, the recent tremors do not necessarily represent the entire span of seismic pattern through time, and thus maybe misleading, representing merely a momentary stage of a much longer and composite process.

The seismicity of the region is constantly monitored by the Seismology Division of the Geophysical Institute of Israel, and published in the regularly updated Seismological Bulletin of Israel. The 1997 edition is the source for all information advanced here (Feldman et al. 1997).

Weak earthquakes, of magnitude in the range 2.5–4.5, show some concentration along the Jordan Valley. On the other hand, there seems to be no direct relation between tremors stronger than 4.5 and the Rift. The strongest seismic activity of the Near East is in and around the Gulf of Aqaba, south of the region discussed. Weaker earthquakes are somewhat concentrated in the Dead Sea, Kinneret and Hula basins, the rest are scattered all over the southern Levant.

It is probably best to cite Shamir (1997), who summarized the seismic activity of the Rift (referred by him as DST – Dead Sea Transform): "A major characteristic which seems to dominate the seismic activity along the DST during this period is the spatially non-uniform distribution of earthquakes. Activity is localized at specific segments of the DST, specifically where major stepwise discontinuities in the fault structure produce broad pull-apart basins. These are the Gulf of Elat (Aqaba) and the Dead Sea, Kinneret and Hula basins. The DST segments in between these basins, namely the Arava and the lower (central and southern) Jordan Valley, are characterized by a nearly complete lack of seismic activity, including micro-seismic, during the current century." This view is also shared by others (Arieh 1985, Rotstein & Arieh 1986, Shapira 1997).

Shapira (1997), while repeating these observations, concludes that "the pull-apart model, assumed to characterize this area, cannot as yet be verified by the seismological observations, i.e. the main faults bordering the lake (Dead Sea) are less active as compared to those within the lake and so far show a dominant normal faulting mechanism, rather than the expected strike-slip". He also notes that "the Jordan Valley north of Damiya up to the Sea of Galilee (Lake Kinneret) is practically a-seismic. This part of the DST is well instrumented and it is quite unlikely that creep will go unnoticed in terms of micro-seismicity."

Shamir & Feldman (1997) studied micro-seismicity in the Hula basin, concluding that the north–south oriented bordering faults are quite inactive, while normal faulting is concluded for the internal, SW–NE faults, especially in the southern part of the Hula and the bordering Korazim block. More on the seismicity of the region can be found in Chapter 8.

Historic earthquakes are mentioned by many, summarized in Amiran et al. (1994) and others (see Chapter 8). Some of these may have been devastating events, as can be seen in archaeological sites in the Jordan Valley and the close vicinity, such as Hisham Palace in Jericho, Bet She'an, Susita east of Lake Kinneret and others (Fig. 3.3.2). However, since no accurate data can be concluded from the descriptions, particularly no locations of the centers of tremor, these historic events have little significance for understanding the seismic pattern of the Jordan Rift Valley.

3.4 CLIMATE

The Near East lies at crossroads of climatic regimes, the interplay of which is responsible for a great variability of almost all factors over very short distances. The main effects result from the European polar fronts and western winds in winter on the one hand, and the Sahara high pressures and Indian Ocean monsoon-system lows in summer, on the other. Added to these but by no means less important, is the Mediterranean Sea, which stores solar energy in summertime, to release it in winter. The ensuing results are very steep gradients in temperatures, cloudiness,

Figure 3.3.2. Sinistral movement of a north facing wall in the crusader fortress of Vadum Jacob, south of Benot Ya'aqov, between the Hula and Lake Kinneret, most probably accompanying the 1202 AD earthquake.

wind directions and velocities and rainfall, both from the north southward and from the west eastward. The low elevations of the Jordan Valley cause "pulling" northward of the desert effects of the south and east. Descending air from its bordering highlands causes higher temperatures and lower precipitation all along the rift (Fig. 3.4.1). In rare cases, particularly on clear winter nights, the descending air could be extremely cold, causing frost.

The southern Levant should actually have belonged to the planetary desert system of the Sahara, since it occupies its natural continuation toward the southern Asian dry lands. In practice, however, this is different, because of the winter rains typical of this region. Thus the Jordan Valley lies in a transition zone between the typical Mediterranean, subtropical climate to the northwest, and the bare Saharan desert to the south. Due to its unique topography, the desert boundary in the Valley is deviated to the north, as compared with the bordering highlands, but the climatic gradient remains of a similar nature.

Summer is typified by very high barometric pressure, resulting from air descending over the Sahara, which also causes the high temperatures of this season. As a result, no rains reach the region during this time of the year, which could stretch over more than 6 months. The low pressure of the Indian Ocean monsoon belt causes almost constant western to northwestern winds over the entire Near East. These change direction in the Jordan Valley and veer to the south, probably as a combined result of topography and the lower pressure over the Red Sea. Northern winds are therefore typical for the Jordan Valley, especially during the afternoons, when the synoptic conditions are fully developed.

These wind directions are occasionally reversed in the transition seasons, the spring and early fall, when the Saharan pressure is not too well developed and the Arabian desert develops a high barometric pressure, sending dry, hot air to the region. It is, however, extremely rare for monsoon rains to reach the Near East.

The winter climate is affected by a different regime, since during this season both the Saharan high and the monsoon low are less effective over the region. This, together with the expanding high pressure over the north pole, pushes the southern end of the global western winds belt into the Mediterranean. Consequently, the winter winds regime over the Levant is no different than during the summer, only for another cause. Occasionally in winter, when both the Saharan and Arabian highs are weak, polar air penetrates the Levant, causing a drastic drop in temperatures, which may adversely affect vegetation. During such events temperatures in the Jordan Valley may even drop below zero.

There only remains the question, why rains in the winter? The primary idea, still maintained by some, is that the southward deviation of the western winds belt brings cyclones from the Atlantic Ocean, all the way to the Levant. However, this simplistic view could not hold when isotopic analyses of rainwater (Gat & Dansgaard 1972) showed that most, if not all, rain being precipitated in the southern Levant originates from the Mediterranean. When satellite weather images became a common tool for meteorology (Goldreich 1993), the Mediterranean source was further corroborated.

To explain the present-day rain regime, characterized by winter thunderstorms, Horowitz & Assaf (1981) suggested the following model (see further discussion in Section 6.7, Fig. 6.7.1): the Mediterranean acts as a solar energy reservoir, acquiring most of its stored heat during the long, hot summer. The water balance of the Mediterranean is negative, thus Atlantic Ocean water makes up the balance by streaming in through the Straits of Gibraltar. The Atlantic waters are less saline than those of the Mediterranean, floating eastward. On their way they are heated by the sun and the descending Saharan air masses, evaporating, so that what remains becomes warmer and more saline, thus heavier. This causes large amounts of warm, oxygen-bearing water to sink down to the bottom, mixing the Mediterranean in such a way that its entire volume becomes an energy store.

This stored energy is released on contact with cold, polar air masses, which only arrive at the Mediterranean in wintertime. The rendezvous creates instability, which results in formation of cyclonic thunderstorms over the Mediterranean. These are carried in a general southeastward direction by the combination of northerly and westerly winds, which define the location of rainstorms in the Levant. The rain regime of the Levant is thus typical of cold fronts, with large amounts of water being poured down in a short duration. Warm fronts are quite rare in the Levant, but when they do occur, they usually do not produce more than drops. This

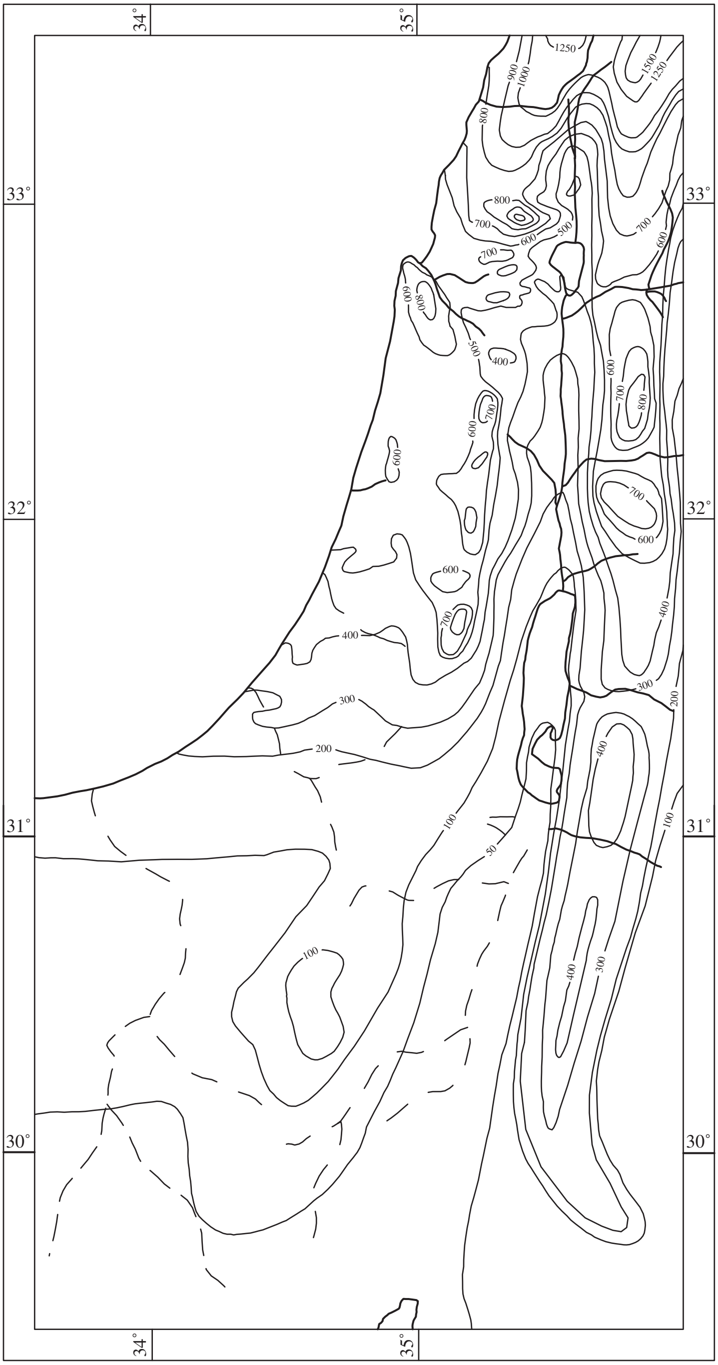


Figure 3.4.1. Average annual rainfall of the southern Levant.

rain regime has its consequences for vegetation, geomorphic processes, and in fact almost every natural aspect, as discussed below.

Dew is a very important constituent of precipitation in the Jordan Valley, especially in its central and northern parts. The high diurnal temperatures cause excessive evaporation, so that the air is quite humid. At night, when the air cools it sinks down into the deeper depression, which results in considerable quantities of dew.

The resulting Jordan Valley climate is subtropical, warm, quite moist Mediterranean to the north, especially in the Hula Valley, grading to subtropical steppe southward, toward Lake Kinneret, where savanna is developed. This stretches into the central Jordan Valley, drier as one goes further south. Desert is typical of the southern Jordan Valley, getting progressively warmer and drier toward the Arava, where a typical, extremely arid Saharan climate prevails.

3.5 GEOMORPHOLOGY, EROSION AND DEPOSITION

The Jordan Rift Valley is a relatively young tectonic feature, thus its morphology depends to a large degree on structural characteristics. The present-day tectonic and morphological setup only commenced some two million years ago, during Palynozone QII times (for further details see Chapter 9), when the Jordan system became endoreic due to a combined process of subsidence along the Valley, and uplift of both its shoulders, east and west. Subsidence was not uniform (Horowitz 1987c, 1989b), causing the formation of several deeper basins, separated by uplifted saddles (see beginning of this Chapter).

The deepest and terminal basin is the Dead Sea, where a considerable volume of Quaternary deposits was accumulated. Lacustrine fine grained sediments, usually silts rich in carbonate derived from the bordering hills, are also being laid down in the other basins, the Hula and Kinneret. Paludine peat is formed in marshes north of the Hula, and organic silts accumulate in the Buteiha marshes northeast of the Kinneret. The Jordan River developed a floodplain where it flows from Lake Kinneret to the Dead Sea, depositing gravel and silts, which occasionally moved from one locality to another, depending on the river's activity.

The Dead Sea and northern Arava are much more affected by floods than the north due to the poorer vegetation, thus characterized by deposition of coarse clastics, typically gravel, which cover all of the wadis' floors. Those are frequently relocated until reaching their final destination, the Dead Sea, where they are laid down as alluvial fans, characterizing the wadis' outlets. Such alluvial fans, quite often of considerable sizes, are also typical for the wadis' outlets to the Arava, both on the east and west, where the floods suddenly lose their energy on leaving their narrow gorges and arriving at the much wider plain.

Springs are very common in the Rift Valley, as a result of faulting which causes exposure of aquifers. Two types of sediments are typical, travertines at the outlets, which can occasionally cover wide areas (Horowitz 1979, p. 164), and highly organic deposits of the springs' pools and marshes, characteristic of the larger ones (Horowitz 1992a, p. 110). Saline and hypersaline playas, ponds and marshes are particularly abundant around and south of the Dead Sea, depositing highly organic silts, occasionally rich in gypsum and rocksalt.

Erosion, especially prevalent in the streams and wadis leading to the Jordan Rift Valley, depends primarily on relief, resulting from vertical movements, climate and vegetation. In a general way, one can say that during the last two million years the relief is continuously being accentuated, thus the rates of erosion are progressively increasing. This is true for most of the Quaternary, but distinct phases of accelerated subsidence and subsequent erosion are known for some stages of this period (Horowitz 1989b; 1992a, p. 327). The changing Quaternary climates also affected erosion differently throughout the period, and the combination of the above factors created a rather complex morphologic system.

Rates of erosion were calculated for a stream leading to the Hula Valley (Horowitz 1975b), indicating that a minimum amount of one-third of a cubic kilometer of Eocene limestones was removed from this wadi alone during the last 100 Ka. The considerable volumes removed from the late Pleistocene Lisan Formation south of the Dead Sea can be seen by the highly incised landscape of this region (Fig. 3.5.1). Considering that the thickness of Quaternary fill in the Dead

Figure 3.5.1. Erosion of the Lisan Formation at Nahal Perazim, just west of the southern Dead Sea. The top layer is only 18–15 Ka old.

Figure 3.5.2. Terraces along Nahal Fatza'el, north of the Dead Sea, formed by alternating periods of deposition and erosion. The higher corresponds to the Baq'a Conglomerate, of Palynozone QVII age, the lower to Nahshon, of QIX. The lower terrace is developed on sediments of the Lisan Formation; the base (light) is Ami'az Member, the darker top is Fatza'el Member, whose conglomeratic nature is seen in the left foreground.

Sea basin is several kilometers (see Chapter 6), these rates of erosion are not surprising.

The rates of erosion around the Jordan Rift Valley are primarily determined by its structural development. The style of erosion, on the other hand, is chiefly affected by climate. In a very general way (further, detailed discussion is given in Chapter 6), two alternating climate types are characteristic for the Quaternary of the southern Levant. One rather dry, of the present-day style, typified by a long, dry summer and a short rainy winter, during which rain is poured from thunderstorms, causing floods especially to the south. This type is not supportive of rich floras, and the combination of scarcity of plants and frequency of floods causes severe erosion in the wadis and streams, followed by considerable incision and canyon formation.

The other, wetter climate is characterized by gentle rains, mainly originating from warm fronts, with a much more uniform spread over the year, including some summer rains. The ensuing rich vegetation, combined with perennial flow in most of the water courses even to the south, results in gentle erosion and usually the accumulation of alluvial and colluvial silts and gravel in the wadis. The alternations of these two types of climate is expressed in the formation of terraces (Fig. 3.5.2)

Figure 3.5.3. Lake terraces higher and older than the Lisan Lake (foreground), south of Marma Feiyad, on the eastward dipping Umm Sabune Conglomerate west of the Jordan River.

along the wadis and streams (Horowitz 1979, p. 121), known from all over the Near East (Besançon & Sanlaville 1984).

Other terraces skirting the Jordan Valley are those left by higher lakes stands, typical of the wetter periods. Such lake shore terraces, representing the retreat of the Lisan Lake (Fig. 11.4.1), are very abundant around the Dead Sea. Terraces formed by earlier lakes are less frequent due to subsequent erosion (Fig. 3.5.3), but do occur along the entire Valley (Horowitz 1979, p. 129).

3.6 NATURAL ENVIRONMENTS AND SUBSISTENCE RESOURCES

Four major types of natural environments are represented in the Jordan Valley (Fig. 3.6.1). Its northern part, down to the southern end of Lake Kinneret, is Mediterranean (Figs 3.6.2 and 3.6.3); southward, down to the northern limit of the southern Jordan Valley, the environment is steppic, Irano-Turanian (Figs 3.6.4 and 3.6.5); while further south the Saharo-Arabian (previously termed "Saharo-Sindian") domain takes the upper hand, getting drier as one goes further south (Fig. 3.6.6). Within the latter, wherever water is plentiful, up to its northern boundary, are local

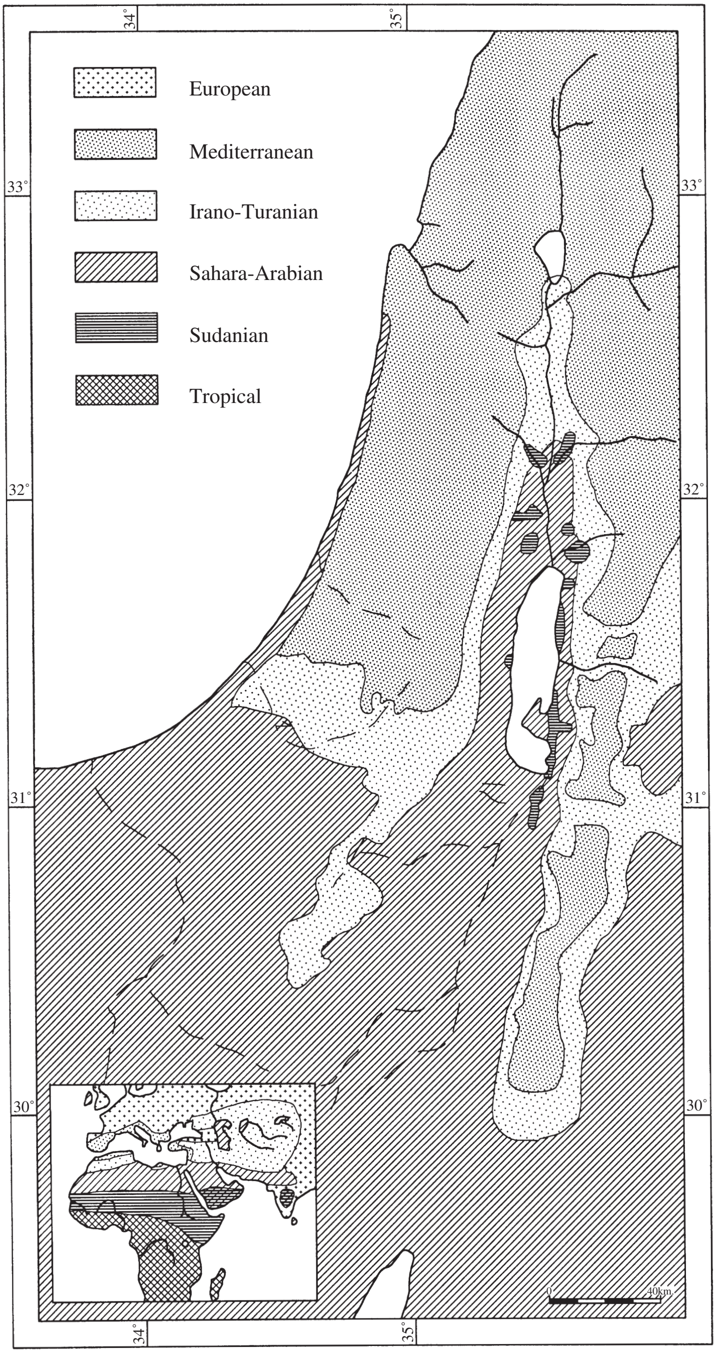


Figure 3.6.1. Natural environments of the southern Levant. Modified from Zohary (1959).

Figure 3.6.2. Mediterranean environment, *Q. calliprinos* woodland, eastern Galilee.

Figure 3.6.3. Mediterranean environment, *Q. ithaburensis* park forest, western Golan Heights.

Figure 3.6.4. Irano-Turanian environment, *Ziziphus* savanna, western Golan Heights.

Figure 3.6.5. Irano-Turanian environment, *Artemisia* steppe, southern Jordan Valley.

Figure 3.6.6. Saharo-Arabian environment, *Zygophyllum* desert, central Negev.

enclaves of Sudano-Zambesian (previously termed "Sudano-Decanian") environments (Fig. 3.6.7). These environments were defined according to phytogeographic principles by Eig (1931–1932), and were somewhat modified by later authors. Their exact boundaries are also a matter for dispute between various students, but the version presented here is based on Zohary (1959, 1973). A detailed vegetation communities map of the Jordan Valley and surroundings is presented in Fig. 3.6.8.

The limits of the Mediterranean, Irano-Turanian and Saharo-Arabian environments on both the eastern and western rift shoulders extend far southward in comparison with the depression itself. These limits extend far south on the eastern flanks as compared with the western, the latter being continually in the rain shadow. This general southward extension results from the lower elevations of the Jordan Valley, which is constantly heated up by descending air (Fig. 3.6.1).

The Mediterranean environment encircles almost the entire Mediterranean Sea, characterized by its unique climate (see above). The average annual rainfall in the northern Jordan Valley varies considerably, from some 1,500 mm to the north, on Mount Hermon, gradually diminishing to 400 mm at the southern limit of this domain, with corresponding effects on fauna, flora, soils and so on.

The Irano-Turanian environment comprises central Asia, Iran, Anatolia and some regions in north Africa, especially in the vicinity of the Atlas mountains. The climate is extremely continental, with four distinct seasons: cold, snowy or

Figure 3.6.7. Sudanian environment, *Tamarix*, *Moringa* and *Acacia*, Ain Umm Hudeib, eastern Dead Sea shore.

dry winter; temperate, rainy spring and fall; and hot, dry summer. The climate of the Irano-Turanian environments of the Levant is different (see above), so that this type was designated in our region only by the similarity and characteristics of its semi-arid flora, rather than its climate. Rainfall gradually decreases in the Irano-Turanian zone of the Jordan Valley, from averages of 400 mm to the north to 200 mm at the southern end.

The Saharo-Arabian environment extends over the Sahara and Arabia deserts, characterized by seasonality quite similar to the Mediterranean domain, only with much less rain and higher temperatures, especially during the day. The ensuing environment is thus arid to extremely arid, enjoying less than 200 mm of rain annually. A characteristic of the Saharo-Arabian rainfall is its instability. Some years may be quite rainy, others practically dry. The Sudanian enclaves are no richer in rain than the surrounding desert, but they receive a constant supply of water from springs or high level groundwater. This makes an oasis type environment, quite rich in vegetation and wildlife, rather than the typical Sudanian landscapes dominated by savannas.

Within the four major environment types are numerous local variations, resulting from either extreme temperatures, or availability of water. To these must be added the drainage that, when partly or entirely blocked, produces marshes; and the soil and water salinities, which change considerably along the Jordan Valley.

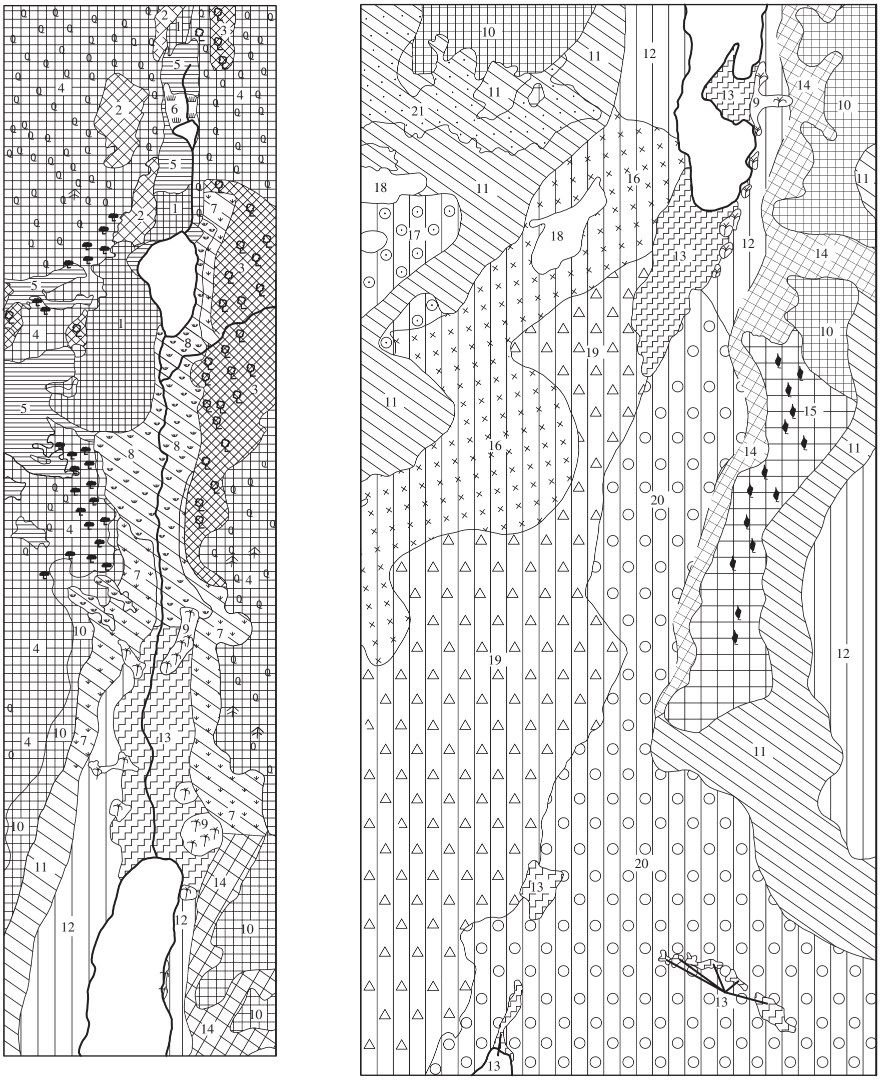


Figure 3.6.8. Vegetation map of the Jordan Valley and neighboring regions, compiled from Zohary (1959) and Waisel (1984). Left: northern sector, right: southern sector. (1) Marginal Mediterranean batha, accompanied by *Ziziphus*. (2) Mediterranean batha. (3) Mediterranean batha, with relics of *Quercus ithaburensis* forests. (4) Mediterranean batha, with relics of *Quercus calliprinos, Pistacia palaestina* and *Pinus halepensis* forests. (5) Alluvial valleys whose natural vegetation has been destroyed by agriculture. (6) Hydrophil vegetation of the Hula swamps, mainly *Phragmites* and *Papyrus*. (7) Irano-Turanian steppe, dominated by *Retama*. (8) Irano-Turanian steppe, dominated by *Ziziphus*. (9) Tropical oases. (10) Marginal Mediterranean batha. (11) Irano-Turanian steppe, dominated by *Artemisia*. (12) Saharan deserts, with *Anabasis* and *Zygophyllum*. (13) Saline sebkhas, with *Tamarix, Atriplex, Suaeda* and others. (14) Irano-Turanian steppes. (15) Dry Mediterranean maquis of *Quercus calliprinos* and *Juniperus phoenicea*. (16) Saharan desert, dominated by *Zygophyllum*. (17) Saharan desert, dominated by *Zygophyllum* and *Haloxylon*. (18) Dunes, dominated by *Artemisia* and *Lotus*. (19) Saharan hammada desert, dominated by *Anabasis* and *Acacia tortilis*, mostly in wadis. (20) Saharan hammada sandy desert, dominated by *Haloxylon*. (21) Irano-Turanian steppe of loess areas, whose natural vegetation has been destroyed by agriculture.

In some cases salinity is a product of the present-day conditions, but mostly it is inherited, from past saline water bodies, or from saline springs, dissolving buried salt accumulations of Neogene age (Horowitz 1992a, p. 66). Salinity problems are encountered from Lake Kinneret, which is somewhat saline due to emissions of salt by internal springs, southward to the Arava. The maximum salinity is undoubtedly in and around the Dead Sea, which is the most saline lake on Earth.

3.6.1 Hydrology and water supply

A glance at the hydrographic map (Fig. 1.2) readily shows the wealth of water in the Jordan Valley drainage system. The principal sources are high amounts of precipitation to the north, part of it as direct rains especially on the mountains surrounding the Hula Valley (Fig. 3.4.1), another as snow accumulating on Mount Hermon, serving as a reservoir for the dry summer. These waters are drained to the Hula Lake by three rivers, the Iyyon, Snir and Dan, merging into the Jordan River which further crosses Lake Kinneret, on its way to the Dead Sea. In addition, rain falling on the highlands on both sides of the Valley maintains several perennial rivers which join the system from both east and west. The more important ones are to the east of the Jordan, such as the Yarmouk, Zarqa, Mujib and Hasa rivers. West of the Valley, the Amud, Harod and Fari'a rivers feed the system, however with considerably smaller amounts of water as compared to the east.

A major part of the rains is accumulated as groundwater, especially west of the Jordan Valley, where the country comprises porous rocks, mainly limestone and dolomite. These are fed into the Jordan system chiefly by the numerous springs along the bordering faults, which are further enriched in water due to the common regional eastward dips of the rock formations. East of the Rift, a region which enjoys more rain since it is higher and not in the rain shadow, most of the rock layers comprising the Transjordanian Plateau dip to the west, so the plentiful springs have higher output than to the west, while some of the rainwater flows to the Valley as runoff, feeding the rivers.

South of the Dead Sea no perennial rivers occur anymore, and the flow in the numerous wadis is ephemeral, depending on floods caused by the winter rainstorms. These are irregular, but in rainy years supply considerable amounts of water. Some of the water arrives at the Dead Sea, but it seems that the major part is absorbed by the porous gravel of the wadi beds, and is thus kept as a reservoir of high level groundwater. This allows occasional perennial vegetation to grow in these ephemeral water courses (Fig. 3.6.18), acting as a secondary enrichment of the rare rainwater, which provides subsistence even in the extremely arid Saharan environment.

Salinity along the Jordan Valley increases southward. The Hula Lake and marshes are typically freshwater, since the considerable amounts of water flowing through the valley do not allow any salinity to develop. Also, it seems that no fossil salts have ever accumulated in this region in any considerable quantities. Lake Kinneret, although enjoying water quantities of the same order of magnitude as the Hula, is significantly more saline. The main reason for that is salts supplied by a suite of saline springs located on active faults in and around the lake. These springs are fed by recent meteoric freshwater which, on their way, dissolve buried Pliocene rocksalt (Horowitz 1970, Gvirtzman 1997, Nativ 1997, Flexer et al. 2000).

The Jordan River, upon leaving Lake Kinneret, is as saline as the lake's water. Flowing south its salinity increases, mainly by dissolving salts of the Lisan Formation sediments, through which it flows all the way down to the Dead Sea. Rivers, chiefly those from the east, dilute the Jordan's salinity to acceptable values. The Dead Sea, the terminal lake of the Jordan system, is extremely hypersaline, for two reasons: intensive evaporation under the strong sun during most of the year, and dissolution of fossil rocksalt bodies, notably Mount Sedom, which push their way up as diapirs (Belitzky 1996).

Salinity of shallow water bodies follows the general trend. Marshes north of the Hula Lake (Fig. 3.6.19) and those northeast of Lake Kinneret are made of freshwater, those occupying the Jordan River floodplain from Lake Kinneret southward to the Dead Sea are somewhat saline to brackish, while around and particularly south of the Dead Sea the area is covered with hypersaline playas (Fig. 3.6.17). Brackish environments, rather limited in area, are found in those localities where rivers or springs water mingle with those of the Dead Sea. These biotopes maintain unique faunas, with relics from the Pliocene such as fish of the Cyprinidae family (Ben-Tuvia & Golani 1993) and a foraminifer, *Ammonia beccarii* (Ehrenberg 1849, Almogi-Labin et al. 1991, 1992, 1995).

Besides the rare present-day rains, some of the springs and groundwater aquifers in the Arava are supplied by a unique source of late Pleistocene fossil water. These "paleowaters" show isotopic composition different from the present-day rains (Gat & Galai 1982), and radiocarbon ages indicating a late Pleistocene–early Holocene time of recharge (Kaufman et al. 1973). Groundwater level is quite high all along the Jordan Rift Valley, thanks to its low elevation and to the constant flow of water in aquifers on both sides, but more from the east (see above). Wells can easily be sunk anywhere, yielding fresh or slightly brackish water. Only in certain localities near the Dead Sea are these too saline for use.

It is thus quite clear that paucity of water is not a limiting factor for settlement of the Jordan Valley, nor for maintaining decent environmental conditions for vegetation and wildlife. The combination of wide range of temperatures and availability of water makes the place a real subtropical paradise for its inhabitants. In addition, these provide ideal refuge environments for plants and animals, expressed in their large number of species from varied provenances (Danin 1988, Tchernov 1988).

3.6.2 Fauna and flora

The Near East is a meeting point for three principal biogeographic domains, the Palearctic to the north, Oriental to the east and Ethiopian to the south (Fig. 3.6.1).

All three contributed to the evolution of the more localized Mediterranean entity, which commenced its development some time during the late Cenozoic. The flora (Fig. 3.6.8) and fauna of the Jordan Valley comprise members of all four domains, with additions of some endemic plant and animal taxa. The aquatic fauna also includes relics from the times the Valley was submerged under the Mediterranean sea water, during part of the Pliocene. Naturally, both fauna and flora are more copious and varied to the north, with the number of species and their abundance decreasing southward. The present discussion is chiefly based on a detailed account of the zoogeography of Israel in Yom-Tov & Tchernov (1988), and on vegetation studies by Zohary (1959, 1973) and Danin (1988).

Almost all animals and the majority of plants are edible (some examples are shown in Figs 3.6.9–3.6.13), depending on taste, availability and how much effort is needed to catch, collect, or prepare a necessary amount of nutrient. In the last stages of human evolution other factors were added such as agriculture, while later mainly religious restrictions and superstitions appeared on the scene. Incidentally, all these restrictions are readily overcome in times of hunger or shortage, when people would eat practically everything (Schwabe 1994). When food is plentiful, they would prefer the tastiest (which always changes according to fashion) and the largest, easy to catch or collect animals and plants.

Figure 3.6.9. *Tilapia* (St. Peter's fish) in Lake Hula, when it still existed as such.

Figure 3.6.10. *Uromastix* in the northern Arava (playing dead).

Figure 3.6.11. Partridge, nesting in the central Jordan Valley.

Figure 3.6.12. Ibex near the Dead Sea.

Figure 3.6.13. Christ thorns (*Ziziphus spina-christi*), with edible fruit, at the Lisan Peninsula.

3.6.2.1 *Fauna*

A variety of invertebrates inhabits the water bodies of the Jordan system. These include members of almost all phyla, which usually serve as nutrients in the food chain of higher organisms. The invertebrates do not seem to comprise any significant part of human food in this region, except maybe for clams, of which *Unio* and *Corbicula* are very abundant in the muddy bottoms of the Hula and Kinneret lakes, as well as some streams. However, their flesh is smelly, and their relics are never plentiful in sites. Remains of both clams are reported from several prehistoric and archaeological sites (Avnimelech 1938). Gastropods such as *Melanopsis, Theodoxus* and *Melanoides* are very abundant in springs, streams and lakes, but are too tiny to constitute any considerable edible contribution. A single species of crab, which can grow to quite a large size, *Potamon*, is copious in all freshwater bodies, but there are no indications of its use for food, although its close relatives are eaten in Europe even today.

Other invertebrates which are edible and appear in large quantities in the Jordan Valley are land snails of various species, some, like *Helix* and *Levantina*, attaining a decent size. Numerous remains of these snails, especially *Levantina*, are reported from a variety of archaeological sites, and until the late 1930s had been sold for consumption on Jerusalem markets (Avnimelech 1938). Yet another is the bee which, although not edible itself, produces honey, which is known to have been consumed throughout history, possibly even in prehistoric times.

The Jordan drainage system is quite rich in a variety of fish, all of which are edible. Of these, several large ones are particularly notable as protein sources, such as the endemic *Barbus longiceps* and *B. canis*; the African Cichlidae, of which six taxa live in the Jordan Valley, one quite well known as "St. Peter's Fish" (Fig. 3.6.9). This family is typically tropical, but its members living in the Jordan system evolved into endemic species and even genera (Ben-Tuvia & Golani 1993), probably a result of the long time since the last aquatic connections with Africa. Another edible, easy to catch fish is *Clarias gariepinus* (*C. lazera*), the local catfish, also considered of African descent. It is very abundant in the Hula and Kinneret lakes, but also in the nearby marshes and larger spring pools, since it is partly adapted to breathing air. Years ago, as kids, we used to catch plenty of *Clarias* with our bare hands, in the Hula swamps.

Others are small fish, living either in springs and streams or in the lakes (except of course the Dead Sea), or occasionally in both. Of these the only ones worth catching are those which live in schools, by the use of an appropriate net. The most abundant is *Acantobrama terraesanctae* which lives in large, dense schools in Lake Kinneret, known as the "Kinneret Sardine", still exploited today. Other small fish include Cyprinodontidae such as *Aphanius dispar* which live in springs around the Dead Sea, and Blenniidae such as *Salaria fluviatilis*. The origin of both is marine, but following the retreat of the Pliocene sea they adapted to freshwater. Similarly, several invertebrates such as the crustacean *Typhlocaris* and others, are of Pliocene marine origin (Por 1963).

The fish of marine and tropical origin had most probably colonized the Jordan Valley in Pliocene times, when the Nile was close and potent enough to provide a comfortable passageway from Africa, and the Mediterranean invaded the Valley for a considerable length of time. This process could have also extended into QI times. Apart from these, a variety of palearctic fish live in the Jordan system, whose origin lies in the Euphrates province. Krupp (1987) made a thorough study of the history of colonization, concluding that the Euphrates is the source area for the Jordan fish, which later migrated to the Orontes River. In addition, subsequent speciation in the Jordan system resulted in a considerable number of endemic forms. This colonization history conforms very well to the sequence of paleogeographic events (see Section 7.2). During the Oligocene and most of the Miocene the central and northern Jordan Valley were part of the Persian Gulf domain, closely connected to the Euphrates; during the Pliocene the northern sector was drained to what constitutes now the Orontes system; while the central and southern sectors were tied up directly to the Mediterranean. The endoreic Quaternary Jordan Valley was an ideal cradle for endemism, particularly of fish. As revealed from numerous excavations along the Jordan Valley, a variety of fish was eaten by members of almost any human culture. *Clarias* remains are however the most common (Tchernov 1988), probably because they always were the easiest to catch.

Only six species of amphibians are known from the Jordan Valley, all of a northern, Palearctic origin. Except for the marsh frog (*Rana ridibunda*), which is permanently connected to water bodies, all others need water only for laying eggs and for their tadpole stage, but live in moist microenvironments, however. *Rana* is also the only edible amphibian species in the region, but no clear indications of its use as such are available. The green toad, *Bufo viridis*, is the only one better adapted to drier environments, and is thus found also around the Dead Sea. The frogs *Hyla arborea* and *Pelobates syriacus* are restricted to the Mediterranean parts of the Jordan Valley, as are the rare salamanders and newts.

Reptiles, being cold blooded, are very abundant in this part of the world, due to the suitability of their physiology to the warm environments. More than 100 species are known from Israel alone, again representing Palearctic, Oriental and Ethiopian origins, subject to local adaptations and modifications. Important turtles in the Jordan Valley include two species, *Mauremys caspica* and *Testudo graeca*. The first lives in freshwater bodies of the central and northern Jordan Valley, the second occupies the Mediterranean terrestrial environments of the southern Levant. The Squamata are represented by a variety of geckos, lizards, skinks and snakes, which are more abundant in the southern parts of the Jordan Valley and in the Arava. Several kinds of vipers, such as *Vipera*, *Cerastes* and *Echis*, can be considered as food, being rather large and fleshy. So are other snakes, such as *Coluber jugularis*, *Malpolon monspessulanus*, and possibly also the abundant water snake, *Natrix tesselata*. Two lizards, in particular, were until recent times consumed in the region, both living in the Arava and the southern Jordan Valley. *Uromastix aegyptius* (Fig. 3.6.10), a vegetarian which can grow to a length of

75 cm, is indeed tasty, and *Varanus griseus*, occasionally attaining up to 150 cm in length, has similar virtues. At present both are quite rare (and, needless to say, protected).

Nile crocodiles (*Crocodilus niloticus*) lived in the Israeli coastal plain streams until the beginning of the 20th century. It is not clear when these creatures (again quite tasty, even though predators) disappeared from the Jordan Valley. Their remains are however quite common in many excavations.

The variety of biotopes in the Jordan Valley maintains many kinds of birds, from those connected to water bodies in the Hula, Kinneret and Jordan River floodplain (the Zor), to typical desert species of the Dead Sea and the Arava. In addition, being the main route between Europe and Africa, the region is a resting place for numerous migratory birds, as well as a nesting place for southern species and a winter resort for northern. Several bird species are endemic to the Jordan Valley, especially its southern part.

Birds are a twofold food source: their flesh when hunted, and their eggs during the nesting season, which can quite easily be collected. Water fowl such as a variety of ducks, geese, coots and frankolins would probably be preferred, as would partridges (Fig. 3.6.11), quails, pigeons and doves among the terrestrial species. Ostrich was quite common in the southern Levant, including the Arava, until the beginning of the 20th century. It was extensively hunted and its eggs collected by people of many cultures, as evidenced from the abundance of remains in excavations.

Barn owl and other nocturnal birds of prey play an important role in archaeology and environment reconstruction. Since their stomach juices are basic they do not digest bones, but rather vomit them as pellets. Analysis of these pellets yields important information about composition of the fauna, especially the small rodents. These pellets are also a good source for pollen grains (Horowitz 1992a, p. 73).

Many large mammals have disappeared from the Jordan Valley and its surroundings due to excessive hunting, especially at the end of the 19th and the beginning of the 20th century (Yom-Tov & Mendelssohn 1988). These include the bear *Ursus arctos syriacus* which lived just north of the Hula Valley, a subspecies of leopard (*Panthera pardus tulliana*), the Syrian onager (*Equus hemionus*), the Arabian oryx (*Oryx leucoryx*) and the cheetah (*Acinonyx jubatus*) which lived in the Arava, among others. Some other large mammals disappeared somewhat longer ago, but can still be regarded as components of the Jordan Valley fauna. For example, hippopotamus had vanished some 5,000 years ago, while lions are still reported from the Middle Ages. It is not known exactly when cervids, such as *Dama mesopotamica* and others, became extinct in the Jordan Valley, but their bones are plentiful in numerous prehistoric sites. Other mammals are seriously endangered, such as the otter (*Lutra lutra*), which was common in the Hula until its reclamation.

Other large mammals were not exterminated, but rather domesticated, however their wild counterparts are no longer around. These include goats, cattle and possibly sheep. Wolves and wild cats, some of the earliest domesticated animals, can still be seen along the Jordan Valley, but quite rarely.

Large mammals inhabiting the region today include very rare leopards (*Panthera pardus nimr*) in the vicinity of the Dead Sea; wild boars (*Sus scrofa*) all along the Jordan Valley, but more abundant to the north, both in the watery biotopes and the surrounding bushy highlands; hyenas (*Hyaena hyaena*) which are more abundant to the south; several species of gazelles, of which *Gazella gazella* to the north and *G. dorcas* to the south are more abundant. Ibex (*Capra ibex*) is very abundant in the southern Jordan Valley, around the Dead Sea (Fig. 3.6.12), somewhat less so in the Arava.

Other mammals of medium size make good food and are quite abundant along the Valley. These include the porcupine (*Hystrix indica*), hare (*Lepus capensis*) and possibly also the rock hyrax (*Procavia capensis*). It is not unlikely however that smaller mammals, such as the plentiful mice and rats, were also consumed when necessary. Medium sized mammals common in the Jordan Valley are foxes, jackals and Egyptian mongoose, the first more to the south, the others in the Mediterranean parts.

A factor of no lesser importance in human settlement, is a suite of invertebrates which cause diseases, of which several are typical or even endemic to the Jordan Valley. Notable are leeches, some of which only suck blood, while others transfer bilharzia parasites; malaria was, until recent times, very abundant along the entire Jordan Valley, even its dry southern part; several species of *Trypanosoma* cause diseases in cattle and man, such as leichmaniasis. All the above have direct connection with fresh and brackish water bodies, since at least part of their life cycle (or of other animals distributing the parasites) is in water.

Terrestrial invertebrates also distribute diseases, such as the cave fever transmitted by ticks (*Ornithodorus*), which may have made life miserable for prehistoric man. Rats, mice and other animals spread parasites and maladies such as rabies and so on. Scorpions, vipers, venomous spiders and insects, all quite abundant in the Jordan Valley, do not seem to help improve the quality of life.

3.6.2.2 *Flora*

A distinction must be made between the vegetation of the highlands on both sides of the Jordan Valley, and of the lowlands comprising the Valley itself, although both define similar environmental domains (Figs 3.6.1 and 3.6.8). The main reason lies in several thousand years of human occupation that wiped out most of the natural vegetation of the Jordan Valley, through the advent of agriculture, especially in the Mediterranean but also in some of the Irano-Turanian environments. The hilly and mountainous terrains are more difficult to manage, and indeed, there is no reason to start agriculture there, when the fertile valleys are around. It seems best therefore to describe the vegetation of the highlands first, and subsequently try to reconstruct the Jordan Valley flora from the few relics which are still there. Incidentally, these remained intact only in localities declared sacred for various reasons. In the rest of the area, synanthropic plants prevail. Even the Hula Valley marsh vegetation did not escape human intervention: papyrus (*Cyperus papyrus*), the most abundant plant, generally considered "the" characteristic for this biotope, was introduced to the region by the Egyptians some 4,000 years ago (Bein & Horowitz 1986).

The Mediterranean vegetation of the highlands is characterized, depending on soils and available water, by oak woodlands, park forests or maquis of different densities, accompanied by several other trees and quite rich undergrowth. As is the usual case in the southern Levant, the variety of species and the extent of vegetation cover both diminish from the north southward. As a rule, the drier terra rossa soil habitats are occupied by the evergreen Kermes oak (*Quercus calliprinos*, Fig. 3.6.2), while in the wetter ones, comprising basalt or alluvial soils, the winter deciduous Tabor oak (*Q. ithaburensis*, Fig. 3.6.3) and occasionally Cyprus oak (*Q. boissieri*) prevail. Note that the controlling factor for the latter species is perennial soil moisture, rather than the amount of rainfall, which keeps water available for these winter deciduous trees, enabling them to grow during the rainless summertime.

Various mesophytic trees accompany *Q. calliprinos* woodlands in the northern, moist parts of the highlands, such as buckthorns (*Rhamnus alaternus, R. punctatus*), trilobed sorbus (*Eriolobus trilobatus*), maple (*Acer syriacum*), hawthorns (*Crataegus azarolus, C. monogyna*), laurel (*Laurus nobilis*), wild plum (*Prunus ursina*) and others, escorted by many herbaceous species. Fruit trees such as hawthorns, wild plum, oriental strawberry tree (*Arbutus andrachne*), almond tree (*Amygdalus communis*), olive tree (*Olea europaea*), Syrian pear (*Pyrus syriacus*), carob (*Ceratonia siliqua*), or shrubs like the grapevine (*Vitis vinifera*) are exploited by both people and birds. These diminish southward, so that in the driest environments the oaks are escorted only by *Rhamnus palaestinus* and a poorer variety of undergrowth.

Q. calliprinos woodlands and park forests are common in the Upper Galilee and the northern Golan, usually accompanied by terebinth (*Pistacia*) and other trees. *Q. ithaburensis* park forest is well developed on parts of the southern Golan Plateau, where it is accompanied by *Pistacia atlantica*, also a winter deciduous tree. Other trees which come together, but in the drier microenvironments, are Christ thorns (*Ziziphus spina-christi* and *Z. lotus*, Fig. 3.6.13), with a rich annual vegetation, including wild wheat (*Triticum dicoccoides*), wild barley (*Hordeum spontaneum*) and wild oat (*Avena sterilis*). Relics of *Q. ithaburensis* park forest are known from the northern part of the Hula Valley, and are thought to have comprised the climax vegetation of the entire Valley, except for the marshes, where hydrophil vegetation prevailed.

The transition from Mediterranean woodlands to Irano-Turanian savannas can comprise, following edaphic controls, each of several plant communities. The most common are garigue (Fig. 3.6.14), characterized by shrubs of *Phlomis* and *Thymelaea*; and batha (Fig. 3.6.15), dominated by low cushions of *Sarcopoterium*

Figure 3.6.14. Mediterranean garigue, eastern Galilee.

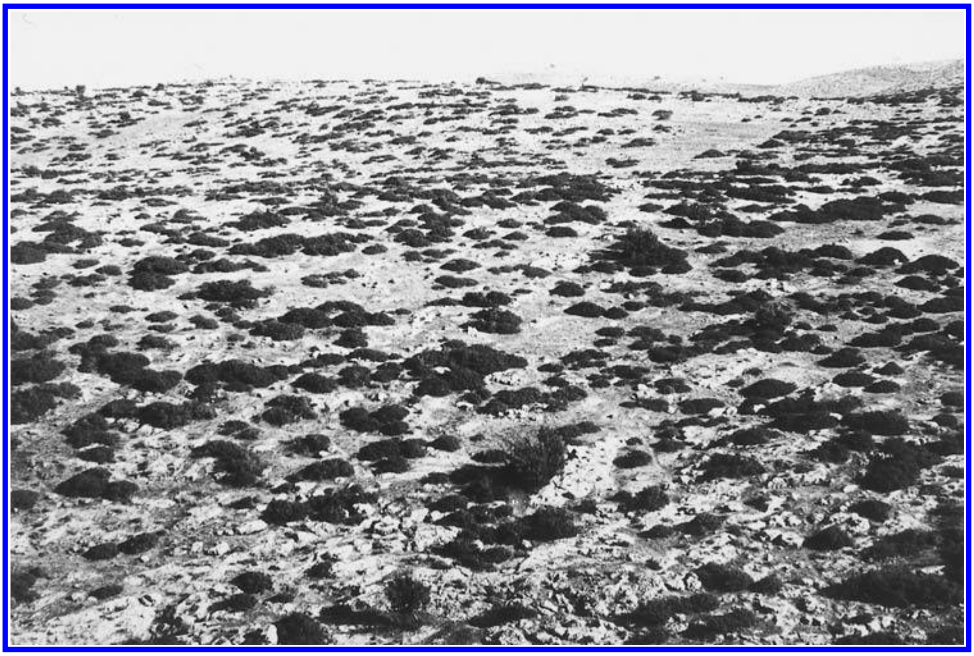


Figure 3.6.15. Mediterranean batha, Judea.

spinosum. Occasionally, edible fruit trees grow in the transition zones, such as carob, almond or olive, accompanied by the mastic tree (*Pistacia lentiscus*), also widely exploited by humans for its resin used as chewing gum.

Christ thorn trees, whose fruit is edible, begin their appearance in the southern reaches of the Mediterranean domain, but are the most important component of the Irano-Turanian steppe vegetation, southward (Fig. 3.6.4). There they form, together with a rich herbaceous flora, a savanna-like landscape, especially on the hills surrounding Lake Kinneret and the central Jordan Valley. Typical ephemeral plant families for this environment include Gramineae, Papilionaceae and Compositae. The seeds of many plants belonging to these families are edible, as are occasionally leaves, stems or roots. They also make excellent pasture for herbivores and birds of all kinds. A typical east African savanna tree which accompanies this environment in the central Jordan Valley is the white acacia (*Acacia albida*, Fig. 3.6.16).

Bordering the Irano-Turanian savannas, toward the desert areas, is a steppe characterized by the wormwood *Artemisia herba-alba* (Fig. 3.6.5). Common co-dominant shrubs are *Thymelaea hirsuta*, or *Noaea mucronata* in the Judean Desert. Trees that may occasionally occur in places enriched with groundwater, are Atlantic pistachio (*Pistacia atlantica*) and almond trees (*Amygdalus ramonensis*). Both trees are also found in the desert regions of Israel and Transjordan, wherever suitable

Figure 3.6.16. *Acacia albida* stand at Nahal Tavor, central Jordan Valley.

microenvironments exist. Incidentally, these microenvironments can carry tens of species of Mediterranean plants, which are relics from wetter and better times.

The Saharo-Arabian desert vegetation, growing in regions where annual rainfall is 80 mm or less, comprises a restricted number of species, of which the most common plants are *Zygophyllum* (Fig. 3.6.6) and *Gymnocarpos*. These are replaced, where salinity prevails, by *Suaeda, Halogeton, Salsola* and others (Fig. 3.6.17). The extremely arid environments to the south only carry plants in places of secondary enrichment in groundwater, typically wadi beds (Fig. 3.6.18) and rock crevices, common in limestones and dolomites. Trees growing in these localities include three species of acacias, *Acacia gerrardii negevensis, A. raddiana* and *A. tortilis*. Where groundwater is higher tamarisk (*Tamarix nilotica*) is the dominant tree. The principal shrubs are *Retama* and *Atriplex*. In rainy years the ephemeral winter vegetation is rich in Compositae, which serve as palynological indicators of the Saharo-Arabian environment (see Chapter 6).

Hydrophil vegetation grows in the Jordan Valley in swamps of the Hula Valley (Fig. 3.6.19) and the Buteiha, at the northeast corner of Lake Kinneret; along the Jordan (Fig. 3.6.20) and other perennial rivers; on the shores of the Hula and Kinneret lakes; and in connection with springs (Fig. 3.6.21). The most abundant plants of these biotopes are various reeds (*Phragmites* and *Arundo*) and cattails (*Typha*), accompanied by a variety of other hydrophil plants. *Papyrus*, imported

Figure 3.6.17. Sebkha near Sedom, southern Dead Sea, with *Suaeda* and *Tamarix nilotica*.

Figure 3.6.18. Extremely arid wadi bed with *Acacia*, northern Arava.

Figure 3.6.19. Papyrus in the Hula swamps.

Figure 3.6.20. Hydrophil vegetation along the Jordan River, south of Lake Kinneret.

Figure 3.6.21. Spring vegetation in the Arava, dominated by date palms.

by humans, is common in the Hula swamps (see above); raspberry (*Rubus sanguineus*) grows abundantly in almost any wet habitat along the Jordan Valley, bearing tasty berries; date palms (*Phoenix dactilifera*) accompany springs along the Jordan Valley and the Arava, and occasionally fig trees, but both may have been introduced by man. Riparian trees common for the hydrophil environments of the Jordan Valley include plane trees (*Platanus orientalis*) in the Galilee, willows (*Salix acmophylla*), poplars (*Populus euphratica*) and a variety of tamarisks (*Tamarix*) along the entire Valley.

Water plants, partly or entirely immersed, grow in the Hula and Kinneret lakes, the Jordan River and some of the larger perennial springs. Common are water milfoil (*Myriophyllum*), pondweed (*Potamogeton*) and water lilies (*Nuphar* and *Nymphaea*), accompanied by a variety of others. The Sedom sebkha, the largest continuous salt marsh which existed south of the Dead Sea until it recently dried out, housed, among others, halophil plants such as various species of *Suaeda*, *Nitraria* and tamarisk species, of which the most resistant to high salinity is *Tamarix nilotica* (Fig. 3.6.17).

The combination of a constant supply of freshwater and high temperatures in the Rift Valley springs enabled thermophilous trees of Sudanian origin to establish themselves in oases. The more saline ones support date palms and *Juncus arabicus*. The freshwater oases, of which En Gedi is famous, carry a variety of trees, such as *Calotropis procera*, *Moringa peregrina*, *Salvadora persica*, *Cordia sinensis*, *Ziziphus spina-christi*, *Ficus carica* and acacias. A lesser variety of these trees occurs in oases such as Jericho, and others northward up to the northern end of the southern Jordan Valley.

3.6.3 Soils

Pedogenic processes are hardly ever well developed in the southern Levant, nor its Mediterranean environments, due to the relative aridity which does not allow formation of deep profiles, and the rapid erosion that removes most of the soils on the highlands. The colluvial and alluvial soils do not have much chance to develop decent profiles, since in most localities they are constantly being covered by ever younger alluvium, brought by the frequent floods (see Section 3.4). For these reasons the soils of the southern Levant, including the Jordan Valley, reflect also to a great extent their parent rocks' characteristics, rather than prolonged pedogenesis, which would have better-represented climatic parameters, as is the case in other regions where mature soils are found. The brief discussion here is based mainly on the soil survey of Israel (Ravikovitch 1969).

The highlands on both sides of the Jordan Valley contribute much of the raw material for the alluvial soils filling the depression, thus it seems appropriate to mention briefly the typical hilly and mountainous soils. From the north down to the Dead Sea, in the Mediterranean domain, three types of soil associations typify the highlands: brown basaltic soils, reddish terra rossa and grayish rendzina.

Basaltic soils cover volcanic terrains of the eastern Galilee, the Golan and parts of the Transjordanian Plateau. They are relatively deep and clayey, typically with underdeveloped A(B)C profile. The internal drainage is quite poor, the organic content is usually low, both limiting their agricultural use. They contain some carbonate, occasionally up to 25%, which is quite strange when compared with terra rossa, formed on limestone and dolomite, usually altogether devoid of carbonates. Terra rossa, the typical Mediterranean soil, is developed principally on hard carbonate rocks. It is very well leached, composed mainly of clays, with little or no organic matter. This soil, as well as some areas covered by basaltic soils, is being eroded today in most places. It seems that terra rossa profiles, typically A(B)C, are not developing under present-day conditions, so it should be referred to as a relict soil (Ravikovitch 1969, p. 16) or a paleosol (Horowitz 1979, p. 167).

The characteristic soil that is being developed at present is the mountain rendzina, a gray, highly calcareous soil, occasionally quite deep. The typical profile is AC, with little or no organic material, indicating rather limited leaching typical of the present climate. These soils develop mostly on chalk, and are quite common in the lower elevations of the highlands. They are more common to the north, but occur down to the southern limit of the Mediterranean region.

The arid highlands bordering the Dead Sea and the Arava are not covered by any real soils. Rather, these regions comprise alternations of mountainous hammadas, desert stony lands and occasionally brown desert skeletal soils. All are highly calcareous, are occasionally rich in gypsum, consist of some clay but are dominated by gravel of various sizes, no organic material at all, and a characteristic (A)C profile.

Most of the Jordan Valley, as well as valleys leading into it, is covered by a variety of clayey alluvial soils (vertisols), with an AC profile, occasionally with an underdeveloped (B) horizon, usually very good for agriculture. In places where hills border the lowlands these grade into colluvial (Fig. 3.6.22).

The Hula Valley alluvial soils are quite deep, with some organic matter, calcareous to a certain degree and, due to availability of water and bad drainage, partly hydromorphic. Toward the center of the basin, where marshes were formed, peat soils are typical (Fig. 3.6.23), very rich in humus. The Korazim block is covered by brown basaltic soils, similar to the type occurring on the Golan and other volcanic lands. Lake Kinneret is surrounded by a narrow strip of colluvial–alluvial soils, widening only in the Ginnosar and Buteiha valleys.

South of Lake Kinneret extensive areas which are covered by the Lisan Formation develop the clayey–silty, gray valley rendzina soil (Fig. 3.6.24), occasionally hydromorphic, with various degrees of slight salinity. The soil is highly calcareous, very fertile, with typically A(B)C profile. Such soil also occurs on travertines of the Bet She'an Valley, where it meets the central Jordan Valley. South of the Bet She'an Valley, where the Jordan Valley becomes narrow, a mixture of colluvial–alluvial soils cover the region, except for the Jordan River floodplain in which purely alluvial soils prevail, down to the Dead Sea (Fig. 3.6.25).

Figure 3.6.22. Alluvial-colluvial wadi soil, central Jordan Valley.

Figure 3.6.23. Peat soil, Hula Valley.

Figure 3.6.24. Rendzina soil developed on Lisan Formation sediments, central Jordan Valley.

Figure 3.6.25. Jordan River floodplain alluvial soil, central Jordan Valley.

Figure 3.6.26. Coarse desert alluvium, wadi approaching the southern Dead Sea.

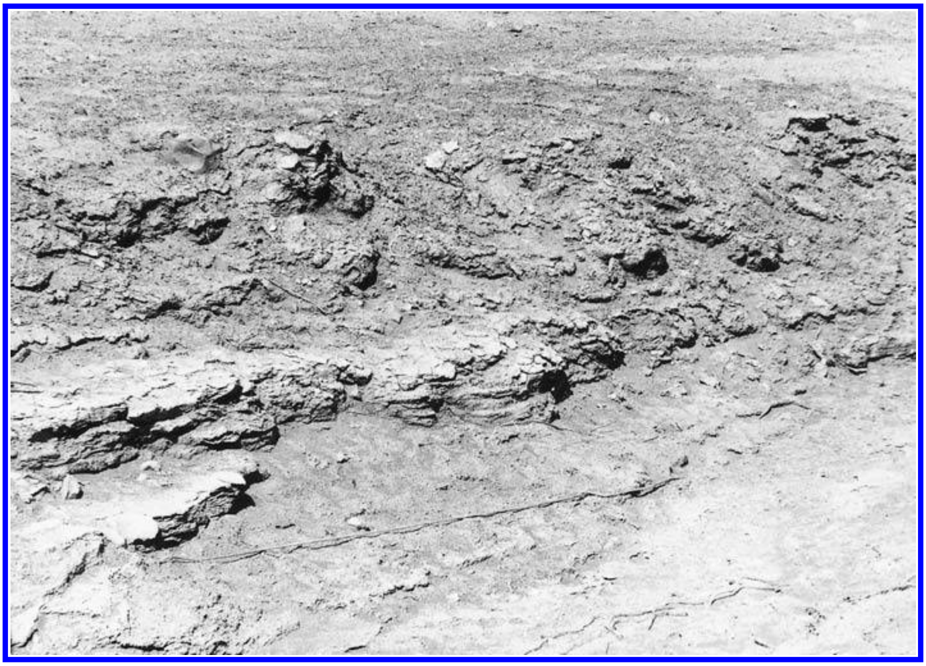


Figure 3.6.27. Solonchak, sebkha near Sedom.

Figure 3.6.28. Desert stony lithosol, northern Arava.

The southern Jordan Valley down to the northern tip of the Dead Sea is covered by desert alluvial soils. The soils are grayish–whitish, occasionally yellowish, silty, highly calcareous, devoid of organic matter and rich in gravel. The characteristic profile is (A)C, while a rudimentary (B) horizon may occasionally develop. Salinity is common in the desert alluvial soils, which are suitable for agriculture only after thorough artificial leaching. Alluvial red-brown, fertile paleosols of the late Pleistocene fluviatile Fatza'el Formation cover several valleys approaching the central and southern Jordan Valley.

The Dead Sea's eastern and western shores, where mountains approach the lake quite closely, do not allow for any significant soil accumulations, except where large wadis are present. In these localities coarse desert alluvium prevails (Fig. 3.6.26), deposited by floods. Wherever freshwater is available, agriculture is possible.

South of the Dead Sea, in the Sedom and other sebkhas (Fig. 3.6.17), some 15 km of the northern Arava are covered by hydromorphic desert alluvial soils, in some of which salinity attains very high values (Fig. 3.6.27), up to a degree of loamy alluvial solonchak (Dan et al. 1981, p. 32). Southward, the northern Arava surface consists of calcareous desert stony lithosols (Fig. 3.6.28), mingled with desert alluvium, hammadas and sandy lithosols. Hardly any profile can be defined for these, which naturally lack any organic substances. Given enough water, these substrates can yield nice crops, thus are extensively used today for winter agriculture.

CHAPTER 4

The pre-Neogene geology of the Near East

Akiva Flexer

The Near East and its adjacent areas encompass three different plates (Fig. 1.4): the African, Arabian and Anatolian. The triple junction near the city of Karaman Marash, in southern Turkey, is located at the northern termination of the Syrian–African Rift system. The Near East is distinguished by three ongoing tectonic processes along the plate boundaries, spreading or divergence along the Red Sea, collision or convergence along the Cyprean Arc and the Taurus mountain range, and lateral plate motion along the Dead Sea–Jordan Rift Valley. Nevertheless this present-day tectonic regime, which started in the Neogene, is totally different from the pre-Neogene situation.

By and large, one may say that the geology of the Near East or the Levant was shaped and controlled by the "struggle" between the Arabo-Nubian Massif to the south and the Tethys Ocean to the north. The Levant, including the eastern Mediterranean, is situated between two major regional tectonic units, the stable Arabo-Nubian Massif to the south and the Alpine mobile orogenic belt to the north. The wide area separating the Massif and the orogen comprises the stable and unstable shelf.

Four major pre-Neogene tectonic stages had affected the region during the Mesozoic and early Cenozoic:

(1) Neotethyan Triassic rifting caused by stretching and thinning of the lower and upper parts of the lithosphere (stretching factors approximately 2.5–3.0 and 1.5–2.0, respectively).

(2) Reactivation of the Neotethyan rifting which occurred in the early-mid Jurassic.

(3) Post-rift subsidence and development of a pronounced basin–margin profile in the mid-late Jurassic to late Cretaceous.

(4) Closure of the Neotethyan Ocean during the late Cretaceous–Paleogene, which converted the long-lived passive margins into active ones.

4.1 PRECAMBRIAN

The Precambrian rocks of the Elat terrain, at the northern tip of the Bay of Aqaba (Fig. 1.5), comprise the northeastern extension of the Sinai Peninsula basement. This terrain, together with the basement outcrops of southern Transjordan, is the northernmost part of the Arabo-Nubian Massif (Fig. 1.4). The evolution of this Massif has been discussed by many authors (Alsharhan & Nairn 1997, pp. 22–36, and references therein). Models suggesting growth by arc suturing and ophiolite obduction, the opening and closing of back arc basins and/or by a combination of arc accretion and continental fragments, have all been advanced.

Since its consolidation in the late Proterozoic and turning into a stable nucleus-massif or craton, the Arabo-Nubian Massif eventually controls the geological evolution of its adjacent areas in the Middle East by dictating the tectonic style, facies of sediments, thicknesses of surrounding strata and strand lines of the Phanerozoic. The Massif, as a stable nucleus, is surrounded by four sedimentary belts (Picard 1939) which have evolved during the Phanerozoic history, easily distinguished in the present-day by their morphotectonic character.

Eyal et al. (1991) pointed out that the evolution of the Arabo-Nubian Massif took place in four main stages, following Bentor (1985). These stages have been reinterpreted in plate tectonic terms by many authors (see for references Kröner et al. 1990, Eyal et al. 1991). During the first stage (~950–850 Ma), an oceanic crust composed mainly of tholeiitic pillow basalts, gabbros and some trondhjemites was formed. The second stage (~850–650 Ma) was characterized by island arc volcanism, producing mainly andesitic to dacitic rocks, their sedimentary derivatives and plutons with similar composition. The rocks of the first two stages underwent deformation and metamorphism to various degrees and, at the end of the second stage, were accreted to form the Arabo-Nubian Massif. The third stage (~650–580 Ma) represents cratonization of the shield and is characterized by widespread calc-alkaline magmatism, mainly of intermediate to acid composition. The last stage (~600–550 Ma) comprises mainly alkaline to peralkaline granites and rhyolites, representing an intracratonic–taphrogenic magmatism. Within the Elat and Transjordan terrains, only rocks of the last three stages are represented.

The total area of Precambrian outcrops in the Elat region is only about 70 km^2. Nevertheless, a surprisingly large variety of rock types are included, comprising metamorphic, plutonic, volcanic and sedimentary. This heterogeneity is only partly original; essentially it is a consequence of extensive tectonic disturbances of large magnitude, which caused different levels of the Precambrian crust to be exposed in a relatively small geographical vicinity. It is notable that almost all the Precambrian rock types exposed in the Sinai and Transjordan, less affected by faulting, can be matched by similar ones in Israel. Therefore, the area of Elat is taken here as a representative for the Precambrian of the southern Levant.

The study of the Israeli Precambrian rock sequence, in addition to the intense tectonic dissection, is also made difficult by the fact that its outcrops are not continuous. Most of the Precambrian basement was already buried in early Paleozoic times under a thick sequence of sediments. This cover was increased considerably by marine sedimentation, which in this area lasted at least from Cenomanian to late Eocene times. As a consequence of the strong faulting movements of the Jordan Rift Valley, which took place during the Tertiary and continued into the Pleistocene, parts of the old Precambrian basement were exposed again.

4.1.1 Rock assemblages

The Precambrian rock sequence of southern Israel can be divided into three units, according to field relationships (Bentor 1985): the oldest is composed of a variety of metamorphic rocks, termed the "Metamorphic Complex". The next, "Intrusive Complex", consists essentially of igneous rocks, ranging in composition from ultrabasic to acid. This group comprises also orthogneisses clearly derived from granites, granodiorites and quartz–diorites, as well as dike rocks of various ages, but older than the third rock group. Although probably more than one period of dike intrusion can be discerned, all are included in the term "Older Dike Group". Wherever contact relationships can be observed, the various rocks of the Intrusive Complex can be shown to be either intrusive into each other, or into rocks of the Metamorphic Complex. It can, therefore, be concluded that all the rocks of the second group, intrusive country rocks, dikes and their metamorphic derivatives, are younger than the Metamorphic Complex.

The third rock group is composed of volcanic rocks, lavas and associated agglomerates, tuffs and subvolcanic intrusions, their hypabyssal dike equivalents, the "Younger Dike Group", as well as by a thick conglomerate sequence. This group, designated the "Volcano-Conglomeratic Complex", overlies rocks of the two older complexes along a major unconformity. It is in turn overlain unconformably by "Nubian" sandstones of early Cambrian age.

The Metamorphic Complex rocks are confined to the two southern massifs, Elat and Shahmon (Roded). On the geological map of Elat (Bentor & Vroman 1955) two different rock formations, the Elat Schists and the Roded Schists, were distinguished. Studies by Weissbrod (1969a,b) have shown, however, that certain rock types are found in both sequences, although they are not equally frequent in the two massifs. The two sequences may, therefore, have a common origin and are thus dealt with together here. The various rocks composing the Metamorphic Complex are, in order of decreasing frequency: mica-schists, gneisses, chlorite-schists, quartzites, phyllites and epidote-schists.

The Intrusive Complex rocks are more widely distributed than those of the two older groups. They form about half of the Elat massif, the major part of the Shahmon, almost the entire Timna, but are absent from the Amram and Neshef massifs. The Intrusive Complex contains a large variety of rock types. These are, starting with the major intrusive rock masses in order of increasing acidity: gabbros, monzonites, quartz–diorites, granite–porphyry and granites, which are the dominant igneous rock type in the Elat massif.

The Volcano-Conglomeratic Complex overlies rocks of the two former complexes, along a very pronounced unconformity. It comprises relatively small plutonic masses, partly of subvolcanic character, basic and acid extrusives, lavas as well as tephra, their dike equivalents (the Younger Dike Group) and finally the Elat Conglomerate or the finer clastic series termed "Zenifim Formation". Rocks of the Volcano-Conglomeratic Complex constitute the entire Amram and Neshef massifs, occur in restricted areas in the Elat and Timna, but are absent from the Shahmon massif, with the exception of some dikes which could be attributed to this phase.

The term Infracambrian (or "Lipalian Interval" of Picard 1943, p. 43) was coined (Wolfart 1967) to describe a succession of non-metamorphosed or just barely altered sedimentary, mostly clastic sequences, overlying clearly metamorphosed or crystalline Precambrian rocks. The Volcano-Conglomeratic Complex, the Elat Conglomerate and Zenifim Formation comprise in Israel this Infracambrian succession. According to geophysical studies (Ginzburg & Folkman 1981) the Zenifim Formation increases in thickness from a few tens of meters in the south up to 3 km in central Israel. The Infracambrian is widely distributed in the adjacent countries, attaining a thickness of 2–3 km. Such is the conglomerate and graywacke series of the Saramuj Formation (Picard 1939) in Transjordan, the Derikand Sadan Redbeds and volcanics in Turkey, as well as numerous clastic shallow marine formations in Saudi Arabia, Oman, Yemen and other countries (Alsharhan & Nairn 1997, p. 84).

4.2 PALEOZOIC

The general outline advanced here is based mainly on Alsharhan & Nairn (1997, pp. 87–229). The most complete section of Paleozoic rocks is exposed in a great, thin, arcuate belt around the eastern flank of the Arabian Massif (Fig. 1.4). The rocks are also exposed in small, widely scattered inliers, within the intensely folded and thrust faulted regions of the Taurus–Zagros–Oman fold belt. During the Paleozoic, the Middle East went through a series of epeirogenic movements, which resulted in regional structural developments and caused pronounced stratigraphic breaks.

During the Paleozoic, the Middle East acted as a relatively stable, passive continental margin, over which terrestrial and shallow marine sediments had accumulated. The total thickness of the Paleozoic sediments in the Middle East is somewhat speculative, for there are many areas where there is little or no data. The greatest reported thicknesses exceed 6,250 m in Oman, more than 4,700 m in southwestern Iran, northeastern Iraq, northern Syria and southeastern Turkey. This very general isopach pattern, which ignores thinning against the Summan Platform in central Arabia, differs considerably, therefore, from the more detailed isopach maps of the early Paleozoic in the Tabuk and Widyan basins, in northern and northwestern Arabia, presented by Al Laboun (1986).

The early Paleozoic of the Middle East was a period of gentle, epeirogenic movements of large, generally rigid crustal units. Widespread transgressions and regressions crossed a shallow, epicontinental shelf, mantled with clastic sediments thickening to the north and east. The sediments source areas of that period generally lie to the south, while more open marine conditions existed to the north. Within this sequence, there are marked unconformities and considerable lateral facies changes. The sedimentary breaks are attributed to movements of several basement highs, particularly active during the late Paleozoic and early Mesozoic.

Evidence of Caledonian deformation is totally lacking in the Middle East, but the time when it occurred elsewhere maybe represented by sea level changes. Evidence of the late Ordovician glaciation is more complete, as compared with North Africa, thereby providing a better understanding of the details of that event. The Silurian Shale development, the Qusaiba Shale Member in Saudi Arabia, is economically important in North Africa and Arabia, and seems to have played the role of major source rock for oil.

The effects of the late Paleozoic Hercynian Orogeny in the Middle East are reflected by uplift, accompanied by erosion and diversification of topography, much as in North Africa. This event was brought to an end by deposition of the younger Permian Khuff carbonates, during the transgression following the clastic deposits of the late Carboniferous–early Permian Unayzah Formation.

The Paleozoic column is incompletely represented (Fig. 1.5) in southern Israel (Cambrian and Permian), in southern Transjordan (Cambrian–Silurian, Bender 1974a, pp. 39–59) and in the western Sinai–Gulf of Suez area (Cambrian and early Carboniferous, Said 1962, p. 129). The reconstruction of the Paleozoic paleogeography is rendered difficult by the incomplete sections in sparse outcrops or drill holes. Whereas the synthesis that follows is based primarily on data presented in Weissbrod (1981) and Gvirtzman & Weissbrod (1984), some of the interpretations and conclusions are not necessarily in accord with these authors.

As visualized by Wolfart (1967, pp. 6–31) and many others, the generalized regional framework upon which Paleozoic deposition began to take place was the continental crystalline core (Arabo-Nubian Massif), surrounded in arcuate fashion by a mobile shelf that included Israel and northern Sinai. The sedimentary record for the Paleozoic is rather fragmentary for the eastern Mediterranean and Persian Gulf areas. All of the Paleozoic periods have been identified in the general region, but in no locality has a complete representation been found. There were marine transgressions and regressions, but slight angularity of the unconformities has also been reported from field observations.

Detailed studies of outcrops and well data by Weissbrod (1969a,b) in Sinai and the Negev demonstrate a normal transgressive–regressive sequence of the Cambrian above the Precambrian basement. Picard's (1959) views, which are based on many field observations and studies in the region, were that "since the beginning of the Paleozoic the country was under a periodic struggle between the geosynclinal reign of the Tethys to the west and the continental influence of the Arabian shield mass to the east. Up and down movements of the shelf and continental slope, most likely with deep seated faulting, rarely raised the country above sea level."

Weissbrod (1969a,b) notes that the Nubian Sandstone Formation, which marks the beginning of the Paleozoic Era, rests unconformably on a flat peneplain surface of the crystalline basement in southwestern Sinai, resembling the basement of southern Transjordan, which is also flat. In southern Israel, on the other hand, this surface is uneven. These differences may be due to block movements at the end of the Lipalian Interval, which caused dislocation and rejuvenation of the relief in southern Israel. Weissbrod identifies sediments of Cambrian and "sandstone, dolomite and several shales sequences of pre-Carboniferous to lower Carboniferous age". Isopachs of the Paleozoic by Weissbrod show a general northeast–southwest depositional strike, with thickening toward the northwest.

The Cambrian of southern Israel and Sinai (Yam Suf Group) is represented by two sedimentary cycles. Overlying the igneous–metamorphic complex of the newly stabilized Arabian Craton (Garfunkel 1980, Bielski 1982) is a sequence of coarse to medium-grained arkoses and subarkoses. The vertical trends of the sorting and maturity parameters of these sediments are suggestive of tectonic stability. Some paleocurrent studies of the lowermost clastic unit (Amudei Shelomo Formation) suggest western to northern paleoslopes (Karcz & Key 1966, Karcz et al. 1971). This unit is topped by the Timna Formation (Garfunkel 1970), a sequence of fine-grained arkoses, shales, dolomites and laminar silts with late Georgian marine fauna (Parnes 1971). Its most distant craton-wise outcrops are known from the southernmost Negev, Sinai and southern Transjordan (Bender 1974a, pp. 39–59), but not from northwestern Saudi Arabia (Powers et al. 1966, Garfunkel 1970).

The second sediment cycle is represented by well-sorted fine-grained arkosic sandstones, overlain by quartz arenites with paleodrainage trends similar to those of the first cycle. There are various indications for an unconformity separating the middle(?) Cambrian Shehoret and the late(?) Cambrian Netafim formations (Weissbrod 1969a,b, 1981). This fluviatile section is topped in southeastern Sinai by some littoral units, of alternating siltstone and sandstone, with sedimentary structures and the trace fossils *Diplocretarion* (Weissbrod & Sneh 1983). It is quite probable that the extent of this transgressive event is not fully recognized, due to the difficulty in discriminating between distal fluvial deposits and littoral deposits of a siliciclastic sea. The facial trends throughout the Cambrian cycles indicate that the marine source was located north and west of the Negev area (Weissbrod 1981).

The cumulative thickness of the Paleozoic section in the Levant area amounts to only 2,200 m. This rather thin column suggests either low rates of sedimentation, and/or extensive erosional events. The discontinuous distribution pattern of thin, epicontinental marine intercalations of early Cambrian and early Devonian age and of fluviodeltaic complexes of late Ordovician age, on the northern margins of the Arabian Craton, can be interpreted as either due to differential tectonic movements of a block located at the northern margins of Gondwanaland (Turkey and Cyprus?), akin to the behavior of other cratonic blocks in North Africa (Garfunkel 1970, Garfunkel & Derin 1984); or alternatively, the missing section(s) (e.g. Permian epicontinental deposits overlying Infracambrian arkoses in most of the drill holes in southern Israel) can be attributed to a Carboniferous epeirogenic uplift, related to the Hercynian Orogeny in Europe (Weissbrod 1981, Gvirtzman & Weissbrod 1984).

The traditional views support the second model, that is, thick Paleozoic sedimentation over the Negev and central Israel, with increasing marine influence from the southeast, toward the northwest. Weissbrod (1981) and Segev et al. (1985) mention a Paleozoic sequence attaining a thickness of several thousand meters. On the basis of seismic prospecting, Grossman (1984) also points out the occurrence of a considerably thick Paleozoic column. However, the total absence of any evidence for early Paleozoic, Devonian and probably even Carboniferous sedimentation in the deeper drillings of the Negev and Transjordan, the absence of thick conglomerates (expected to be found if such a deep truncation had occurred), all raise the question whether sedimentation occurred at all. One can imagine the existence of an elevated block in the central Negev and the southern Coastal Plain during early Paleozoic through Carboniferous times. The block was possibly connected to the Tauride block (Garfunkel & Derin 1984). Sedimentation occurred only on the block margins and in a depressed area between this block and the Arabian Craton.

Many studies (Domzalski 1967, Flexer et al. 1984) speculate on the possibility of sedimentation controlled by deep seated basement faults. This issue cannot be satisfactorily resolved until further deep drill hole information, from central and northern Israel and its continental shelf, provides additional clues as to the missing portions of the Paleozoic sequence.

Whereas early Permian deposits in southern Israel and central Transjordan (part of the Negev Group) are of fluviatile to deltaic facies (Weissbrod 1981, Eshet 1983), very much akin to the older Paleozoic deposits in the area, the late Permian signals the onset of a fast, localized subsidence of the margins of the Arabian Craton, where prisms of epicontinental marine shales, sandstones and limestones have accumulated (Garfunkel & Derin 1984). The sedimentation pattern of the late Permian is the precursor of the Mesozoic style (Weissbrod 1976). It may reflect the transient formation of the Mesogean Sea, or the Paleotethys (Argyriadis et al. 1980, Bernoulli & Lemoine 1980), a shallow predecessor of the Tethys.

4.3 TRIASSIC

Sequences attaining as much as 900 m of sedimentary rocks of Triassic age are present throughout Arabia and the Arabian Gulf region, except for southwestern Arabia, Yemen, southern and central Oman. In the Zagros basin, the maximum thickness of deposits is some 1,200 m. In northern Iraq, northwestern Iran, northern Syria and southeastern Turkey, the Triassic totals more than 1,500 m.

The classic threefold division of the Triassic is recognized in the central and eastern regions of Saudi Arabia, where clastic lower and upper sequences bracket a mainly carbonatic middle Triassic. In the Arabian Gulf, however, this division is not obvious. In Oman and the United Arab Emirates, the early and middle Triassic were mainly times of carbonate deposition. The two contrasting paleogeographic environments can be illustrated by the spread of the early Triassic clastic-continental domain, and the distribution of Ladinian–Carnian carbonate–evaporite depositional basins.

Cox (1924, 1932) advanced the first concise description of the Triassic sequence outcrops in Transjordan, also providing faunal lists, which until very recently was the best available corpus of data. He indicated that the succession of beds wedged out southward, coming to lie on progressively older horizons, of early-middle to late Cambrian age. This basal, erosional contact is often poorly developed, contrasting with the easily identifiable upper contact with the predominantly coarse-grained, pebbly, white Cretaceous sandstone. The Triassic sediments are exposed in northern Transjordan along the Dead Sea, to Wadi Husban in the Zarqa River valley and in Wadi Na'ur. Bandel & Khoury (1981) divided the succession into nine formations. This division has been generally accepted in subsequent mapping and sedimentological studies.

Triassic sediments in the subsurface include an unconformity bounded sequence of carbonates, clastics, shales and evaporites. It is found mostly in wells drilled in the northern highlands, as well as the Risha and Hamza areas of northeast Transjordan. Andrews (1992) integrated all the available subsurface data in northwest Transjordan into a new lithostratigraphic framework of five formations, combined into what he termed the "Ramtha Group".

The successions in the Iraq–Transjordan frontier area indicate the effects of movements of the Ha'il-Rutbah Arch. In Syria, however, as seen from several deep wells, the Triassic is more complete, and despite the different names of the formations, the lithofacies sequence shows features similar to those found in the Arabian Gulf. There is no general scheme for correlating the Triassic throughout Syria. Although there are individual correlations for small areas, they do not incorporate the facies variations, making it difficult to correlate successions on a broader scale. The thickness and facies of the Triassic beds indicate that there were uplifted regions, bordering the older central Syrian and Mesopotamian depressions, throughout the Triassic.

In southeastern Turkey, the Triassic crops out in the Zap Anticline and in the Mardin and Hazro areas. On the flanks of the Zap Anticline, early and middle Triassic rocks occur, represented by up to 1,000 m of marginal marine to continental carbonates and shales, of the Cigli Group.

Triassic rocks are known in the Levant both from the surface and subsurface. Surface sections are quite fragmentary and isolated, and earlier correlation attempts made by several authors (Wetzel & Morton 1959, Daniel 1963) achieved only partial success. Deep oil drilling started in the 1960s in Sinai, Israel, Transjordan, Syria and Lebanon and enabled us to (Fig. 4.3.1) piece together the previous fragmentary data, into a more complete picture (Druckman 1974a,b, 1984, Picard & Flexer 1974, Druckman et al. 1982). The Triassic system in Sinai, the Negev and central Transjordan displays a picture of four shallow marine sedimentary cycles of the Tethys Ocean, advancing over the Arabian Craton. Druckman et al. (1982) point out that the scene of ingressions and regressions affected a relatively narrow belt, no wider than several tens of kilometers.

The first ingressive cycle of the Yamin and Zafir formations (Scythian–early Anisian) is composed of shallow marine carbonates, shales and sandstone indicating near shore environments in southern Israel. The second cycle consists mainly of carbonates of the Ra'af Formation (Anisian), terminated by continental sandstones of the lower part of the Gevanim Formation. The third cycle starts with deposition of the Gevanim fossiliferous shales, which passes into the highly fossiliferous Saharonim limestone and marls. This cycle, of late Anisian to early Carnian age, is the most extensive in its southward geographical distribution.

The third cycle is terminated by a thick sequence of evaporites, the Mohilla Formation of Carnian age. The fourth cycle, of shallow marine carbonates, overlies the evaporite sequence that may have taken place during late Carnian to Norian times. Most of the deposits of this cycle were truncated during a regional emergence period, at the Triassic–Jurassic boundary. The quite simple picture of transgressive and regressive Triassic sedimentary cycles over the margins of the Arabian Craton becomes more complicated west–northwestward, where the present-day Mediterranean Sea is located. Three deep boreholes located in the coastal plain of Israel, Heletz Deep 1, Ga'ash 2 and Atlit 1, tell of a different tectono-volcanic history.

A significant misfit is recorded in the middle Triassic between the global eustatic curve and the inferred sea levels on the margins of the northern Arabian Craton. The peak of the most extensive transgression in Israel and Transjordan occurred in the Anisian–Ladinian (Bender 1974a, pp. 60–65; Druckman 1974a,b; 1984; Parnes 1975; Druckman et al. 1982), whereas the global eustatic maximum is of Norian age. In the Heletz Deep 1 borehole, on the southern Coastal Plain of Israel, a sequence of about 600 m of breccias and polymictic conglomerates was penetrated. The gravel consists of neritic carbonate elements of late Permian to Anisian age, packed in black, argillaceous micrite containing Scythian to Anisian palynomorphs, underlain by 40 m thick shale, probably of early Triassic age

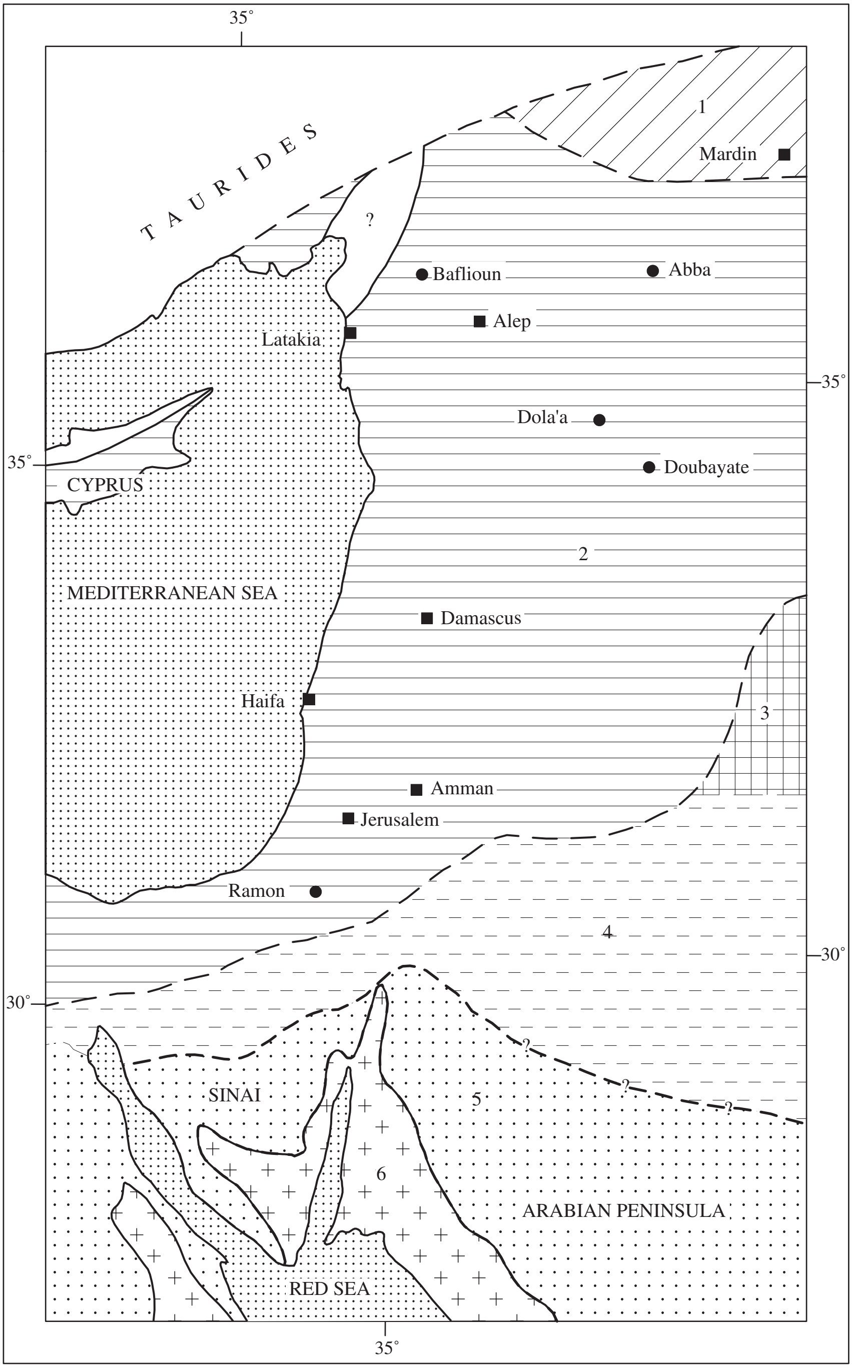


Figure 4.3.1. Triassic facies of the southern Peritethys shelf (modified from Picard & Flexer 1974): (1) Mardin high dolomitic, (2) open marine epicontinental, (3) Ga'ara high, (4) littoral continental, (5) continental and (6) Precambrian basement.

(Druckman 1984). This facies is reminiscent of some poorly sorted submarine fault scarp breccias, of late Anisian–Ladinian age from the Antalya nappes, representing an extensive diastrophic phase in the embryonic eastern Tethys (Marcoux 1978). Some pyroclastic volcanism of early Anisian age is reported from the eastern shore of the Dead Sea (Wetzel & Morton 1959). Carnian rocks from the subsurface of the northern and central Negev (and from northeastern Syria? Weber 1963, in Bender 1974a, pp. 60–65), consist of alternating thick gypsiferous and thin dolomitic strata, the facial changes apparently reflecting a system of basins with intervening elevated sills (Druckman 1974a,b).

Late Triassic rocks were encountered in northern Israel only in a couple of boreholes. In Atlit 1 in the northern (Carmel) coastal plain, 180 m of Carnian to Norian carbonates underlie a 2,500 m thick sequence of basaltic to andesitic flows, alternating with epicontinental carbonates of early Liassic age (Gvirtzman & Steinitz 1983). Ga'ash 2, about 50 km further south, penetrated 267 m of carbonates and shales rich in organic material, of Norian age. This is the only occurrence in Israel of Alpino-typic Triassic facies, different from the common Germano-typic facies (Derin & Gerry 1981, Gvirtzman 1981).

Much of this recently gathered information can be tentatively interpreted in the context of the incipient Tethys. The local transgressive maximum of the Anisian–Ladinian was probably caused by subsidence of the continental shelf (Garfunkel & Derin 1984), due to the distensional regime in what later came to be the Levantine basin. Such a distensional phase is clearly recorded from the late Anisian to late Triassic of the eastern Mediterranean (Argyriadis et al. 1980, Şengör & Yilmaz 1981, Robertson & Woodcock 1982). Consequently, the stratigraphic column of middle Triassic to Jurassic age in the northern Negev attained a configuration of continental-margin-associated prism (Druckman 1974a,b, Goldberg & Friedman 1974, Freund et al. 1975), quite different from the platform-shaped configuration of the Permian to early Triassic sequence (Weissbrod 1969a, 1981). The formation of sunken and uplifted blocks with a relative relief of about 250 m (Freund et al. 1975) controlled the isopach and facies pattern from the Carnian onward (Lapierre & Rocci 1976).

In conclusion, the Triassic sedimentation history is governed by the opening of the Tethys. The spreading of the Tethys caused a distensional regime over the area, reviving vertical basement block movements, together with volcanic activity. The thickness and facies of the Triassic sediments are controlled by the block reactivation along the newly formed passive margins of the Levant–Sinai Plate.

4.4 JURASSIC

During the Jurassic, the Middle East was, for the most part, covered by shallow seas, their level periodically fluctuating (Alsharhan & Nairn 1997, pp. 235–291).

Although the transgressions and regressions were relatively small in terms of absolute sea level change, they induced major changes in the sedimentological environments. The open Tethys (or Neotethys) Ocean lay to the north and northeast, as indicated not only by facies distribution, but also by significant changes in the recorded sedimentary thicknesses.

A progressive, worldwide eustatic rise in sea level began during the Sinemurian, the middle part of the early Jurassic, as shallow seas spread over the eastern and northeastern parts of the Arabian Craton, ending a period of regression and emergence that had characterized the latest Triassic and earliest Jurassic. The Mardin paleohigh in southeastern Turkey remained a positive feature, constituting a barrier separating the shallow open seas of the northern margin of Arabia from the region to the east, resulting in the formation of clastics and evaporites in parts of the northern and eastern margins of the Arabian Platform. Over the northern Arabian Plate in Transjordan, Israel, Iraq, Syria and southeastern Turkey, conditions were much the same as further south. By the end of the early Jurassic, a shallow marine platform formed, upon which carbonate and evaporitic facies were deposited according to sea level conditions.

This early Jurassic transgression continued into the middle Jurassic, again marked by short lived still stands and/or minor transgressions and regressions. Over the vast carbonate platform covering most of eastern Arabia, these minor events were recorded by Murris (1980) as alternations of shallow carbonate platform and open marine (slightly deeper water) limestones and subordinate clastics. The sea level fluctuations, marked by relatively small but distinctive facies changes, can be traced for considerable distances across Saudi Arabia and correlated with similar events in Iran.

Although the sea level generally continued to rise during the late Jurassic, sedimentation rates appear to have more than kept pace with, and finally exceeded, the rate of flooding, with the consequent development of extensive shoal and sebkha environments, where spacious evaporites accumulated. Because of the extraordinary great economic importance of these rocks in the Arabian Gulf region, their subsurface distribution, thicknesses and lithofacies changes are well known, and even relatively minor sea level changes can be documented and traced over great distances.

Marine and associated environments began to be widespread over the Middle East, North Africa and Europe during the Jurassic. There were still terrigenous influences evident at times in southern and southeastern Israel, but fine clastics (shales and silts) predominated over gravel. The tectonic-depositional setting generally favored carbonate formation.

Three tectonic events had an important influence on the Jurassic of Israel (Picard & Hirsch 1987): first, late Triassic–early Jurassic global and regional activity, related to the spreading and opening of the Tethys Ocean. It brought on a long late Triassic–early Jurassic emergence of the entire African–Arabian shelf (Hirsch 1984), followed by volcanism and shallow marine ingressions over a NNE and NNW oriented horst and graben system. Second, the appearance of a "hinge belt", which provided for a wide carbonate platform over most parts of Israel, with basin conditions that prevailed westward, where the present-day shelf and slope are located. Third, late Jurassic–early Cretaceous rifting and volcanism, which accentuated the picture of a north–south oriented graben along the present-day Levant coast. This possibly caused a continuous sedimentation, from the Tithonian into the Neocomian in the Gevar'am trough, in contrast with the intensive denudation of the emerged Levantine block, which existed from the Tithonian to Valanginian times (Hirsch 1984); Fig. 4.4.1 is an example for a facies map of a single, typical time increment.

The local geological history of Jurassic times can be better explained against the background of regional and global events. Therefore the following local descriptions are accompanied by the worldwide Jurassic tectonic-sedimentary framework.

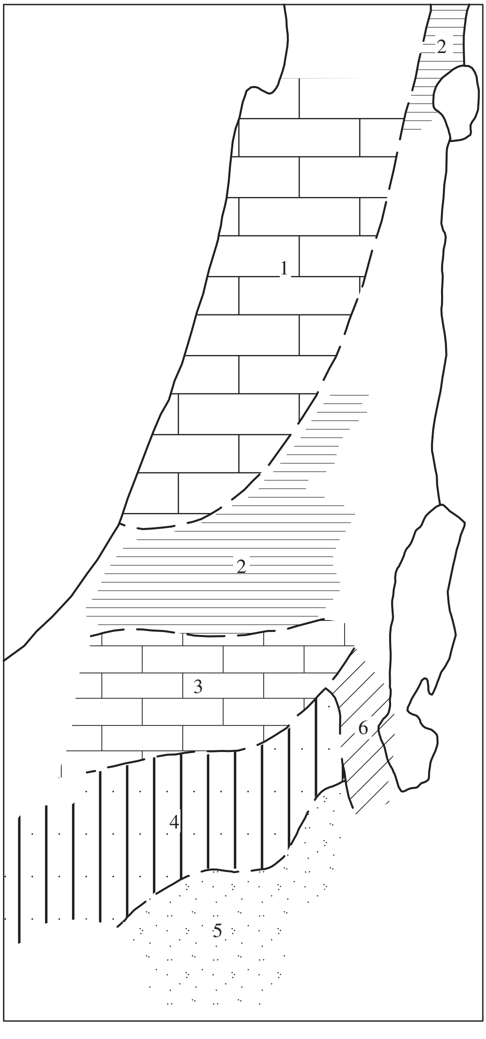


Figure 4.4.1. Late Toarcian–Aalenian facies of Israel (modified from Picard & Hirsch 1987): (1) carbonates, (2) shale dominant, (3) carbonate dominant, shale and sandstone interbeds, (4) arenite dominant, shale and limestone interbeds, (5) arenite dominant and (6) dolomitized.

The Triassic–Jurassic boundary is distinguished by a regional unconformity (Fig. 4.4.2). The global eustatic low of the Norian–early Liassic is represented on the margins of the northern Arabian Craton by a prominent break in sedimentation, whereby extensive areas were exposed, subject to fluviatile and karstic processes and to regional lateritization (Goldberg & Friedman 1974, Murris 1980, Valeton et al. 1983, Hirsch 1984). The erosional truncation into the Triassic sequence in southern Israel does not exceed 70–100 m (Druckman 1974a,b). However, the distensional regime, which started in the late Triassic, most probably extended into the Liassic as well, as can be inferred from the substantial variations in the thickness of Liassic rocks in the Negev over small distances (Freund et al. 1975, Druckman 1977). This regime is apparent from the differentiation of central Israel into a NNE oriented horst and graben system during the Liassic (Grossman 1983, Gvirtzman & Weissbrod 1984) and from the extensive volcanism in central and northern Israel, of primarily basic alkaline affinity (Bonen 1980, Druckman & Kashai 1981, Gvirtzman & Steinitz 1983, Steinitz et al. 1983). Seismic surveys clearly show Jurassic downfaulting along the coast of Israel (Ginzburg et al. 1975, Freund et al. 1975). It is quite probable also that the accumulation of the 750 m thick sequence of Liassic dolomites and anhydrites in northeastern Syria (Weber 1963) and the formation of the Lurestan Shelf basin in Mesopotamia (Murris 1980) are broadly related events, in line with extensive evidence attesting to the development of rifted margins and of shelf edge subsidence throughout the Betic to Hellenic segment of the Tethys, between the Pleinsbachian and the Toarcian (Dewey et al. 1973).

The controversy regarding the shape of the global eustatic curve in the Liassic (Hallam 1981, Vail & Todd 1981) does not interfere with the overall good correspondence between the five sedimentation cycles of Jurassic rocks in Israel (Derin 1974, Bein & Gvirtzman 1977, Hirsch 1980, Parnes 1980, 1981, Lewy 1981, 1983) and the global eustatic curve. Particularly noteworthy are the following events:

(1) The extensive distribution of micritic limestones, probably of a Toarcian age (Qeren Member of the arenaceous Inmar Formation, Nevo 1963). These carbonates are the local representative of a widespread early Jurassic transgression around the Tethys (cf. the Marrat Formation in Saudi Arabia; Powers et al. 1966), made possible by the establishment of well-defined seaways in the Tethys realm (Dewey et al. 1973).

(2) The paraconformity within the Zohar Formation in the Judean Desert, between the Brur and Kidod formations in the central coastal plain of Israel, and within the Upper Dhruma Formation in central Saudi Arabia (Lewy 1983), are in phase with a prominent eustatic minimum at the Bathonian–Callovian boundary.

An important tectonic phase, associated with a major regional unconformity and sea level drop, is recorded at the Jurassic–Cretaceous boundary. "The late Jurassic–early Cretaceous erosion, which took place primarily due to tilting of the Arabian platform, has progressively removed from NW to SE the entire Jurassic,

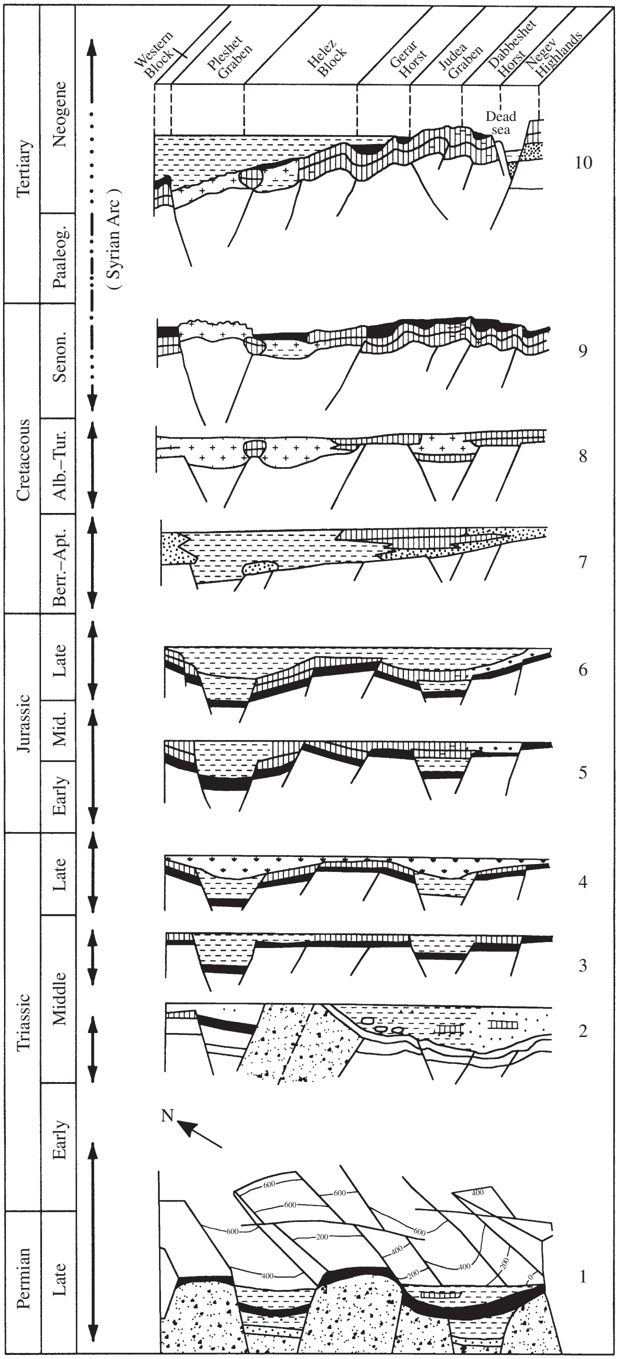


Figure 4.4.2. Tectono-sedimentary development of grabens and horsts, sag basins, tilting and vertical inversions in Israel (modified from Cohen et al. 1990): (1) interblock "passive" sags overlying NNE trending grabens (isopachs in meters), (2) landward tilt, sedimentary wedge, (3) horsts and grabens, (4) interblock "passive" sags, (5) horsts and grabens, (6) interblock "passive" sags, (7) step faulted basinward tilt, sedimentary wedge, (8) mild horst and graben influence, (9) inversion and intense folding, intrablock "active" fault-fold Senonian sags with local anoxic sedimentation and (10) Westward tilt of entire basin, sedimentary wedge, formation of Jordan Rift Valley.

Triassic and marine Upper Paleozoic, bringing in the southern Negev, the Lower Cretaceous Kurnub Sandstone in direct stratigraphic contact with Lower Paleozoic Nubian Sandstones" (Hirsch 1984).

4.5 CRETACEOUS

The Cretaceous period commenced with rifting, divergence and volcanism over the Levant countries and the eastern Mediterranean, and terminated with convergence and folding processes (Flexer et al. 1989). The period witnessed a large-scale transgression and sea level rise, known all over the world (Vail et al. 1977). There are two distinct phases in the geological evolution of the Middle East that reflect Cretaceous tectonic history. For the greater part of the Cretaceous, the depositional environment conditions of a shallow carbonate shelf persisted over the region, continuing the pattern established during the Jurassic. During this time, the subduction that occurred under the zone of the present-day Zagros had very little effect on the Arabian Platform. However, by Campanian times sediments were being deposited in a developing foredeep, which marked the closing of the Neotethys and the emplacement of nappes and ophiolites in Oman and Iran. The ophiolites are dated at about 90 Ma and extend from Iran (Kermanshah region) to Oman.

From this time onward, the latest Cretaceous and the Cenozoic which followed are characterized by gradual exhumation of the Arabian Platform and progressive restriction of the marine area, leading to the form of the present Arabian Gulf. Two main sedimentary provinces are distinctive in Israel and the adjacent areas of the Levant during the Cretaceous, a platform to the east and a basin to the west, separated by the narrow hinge belt formed during Jurassic times. A huge wide carbonate platform (Fig. 4.5.1) extended between the open Tethys to the west (WNW) and the Arabo-Nubian Massif to the east (ESE). The edge of the platform (the hinge belt) was located close to the present-day coastline of Israel and Sinai (Bein 1976, Bein & Weiler 1976, Sass & Bein 1982). The Cretaceous sediments over the platform consist of thick (1,500 m on the average) sequences of carbonates, sandstones and shales containing abundant microfossils, such as planktonic and benthonic foraminifera, ostracodes and calcareous nannoplankton, as well as layers distinctively rich in megafossils like ammonites, bivalves and gastropods.

The Cretaceous system is divided in Israel into three sedimentary cycles: the terrigenous Kurnub Group, of Neocomian to Barremian age, at the base; the Judea Group, made of marly, semi-pelagic basin sediments and platform carbonates, of Aptian to Turonian age, in the middle; and the Mount Scopus Group, very uniform, plankton rich marly chalks, with subordinate chert and phosphorite, of Senonian age, at the top. This situation is very similar in all adjacent countries.

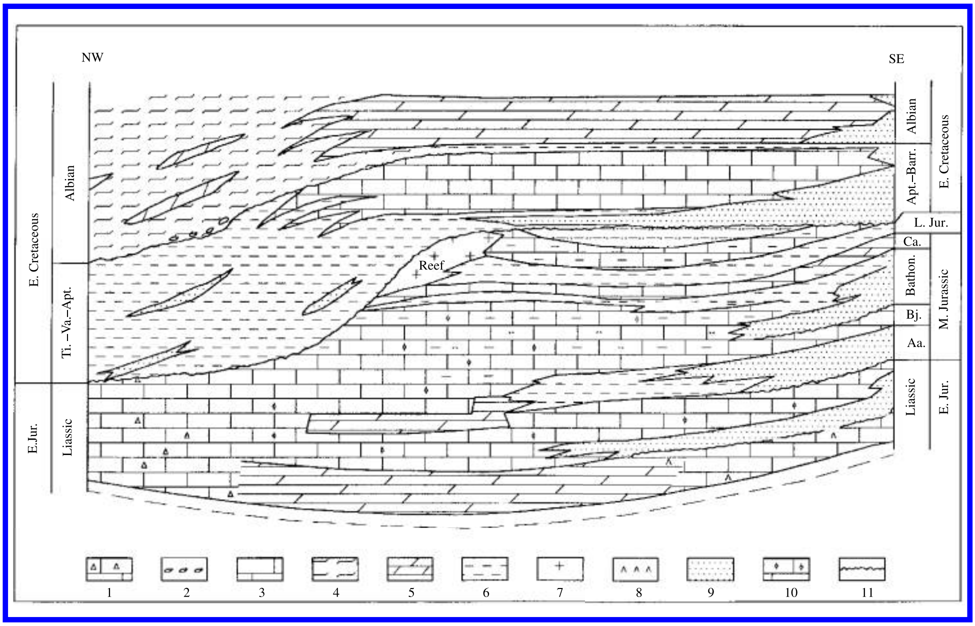


Figure 4.5.1. Jurassic–early Cretaceous stratigraphy across the southern Levant (modified from Cohen 1988): (1) argillaceous limestone, (2) conglomerate, (3) limestone, (4) marl, (5) dolomite, (6) shale, (7) reef, (8) anhydrite, (9) sandstone, (10) oolitic limestone and (11) unconformity.

Over the Arabian Platform, three regional unconformities can be recognized, underlying the Albian, Coniacian and Paleocene sediments. These unconformities subdivide the Cretaceous into lower, middle and upper divisions, rather than the internationally recognized twofold division. Sediments of the early Cretaceous (Berriasian to Aptian) cycle are referred to as the Thamama Group; the middle Cretaceous (Albian to Turonian) cycle makes the Wasia Group; while the late Cretaceous (Coniacian to Maastrichtian) cycle is known as the Aruma Group (Harris et al. 1984, Alsharhan & Nairn 1986).

The beginning of the first cycle (Neocomian) is marked by a tectonic, volcanic and erosional phase. Most of the Levant, namely Sinai, Transjordan, Israel, Syria and Lebanon, was a land covered by a considerable thickness of eolian–fluviatile sands, silts and shales, comprising the Nubian Sandstone of the Kurnub Group (Fig. 4.5.2), accumulated on the platform. Westward, on the continental slope and rise, a sequence exceeding 1 km in thickness, was deposited, of monotonous detrital black shales of the Gevar'am Formation, filling up the deeply incised erosional relief (Cohen 1971, 1976).

The differentiation into platform and off-shelf domains by the hinge line persisted during the deposition of the second cycle (Aptian–Turonian). A belt of rudistid barrier reefs occupied the hinge zone, providing much of the calcilutite–calcisiltite detritus to the sedimentary prism-shaped contourites of the Talme Yafe

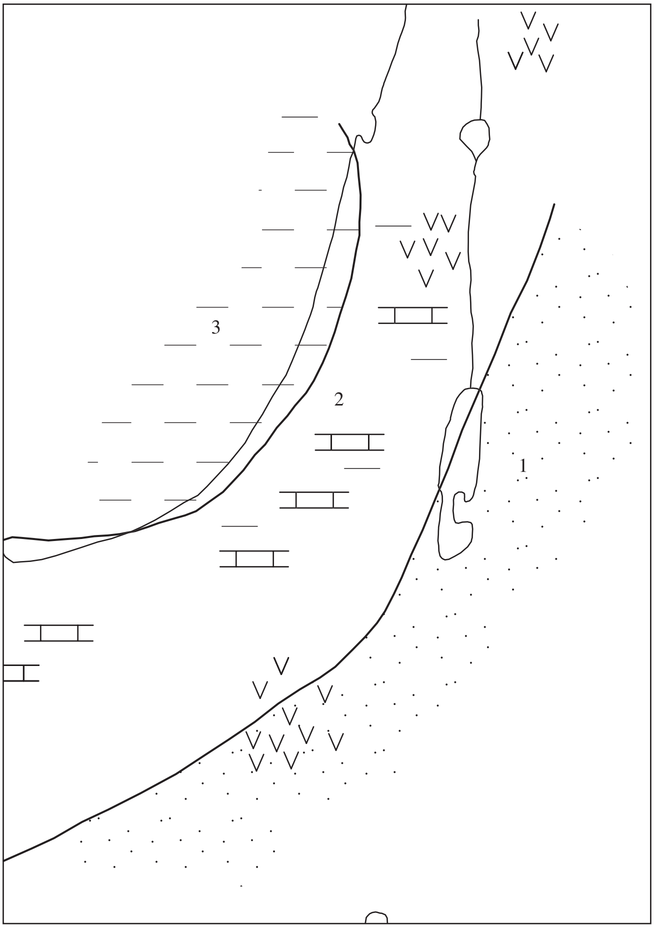


Figure 4.5.2. Early Cretaceous lithofacies belts of the southern Levant: (1) mostly continental sandstone, (2) sandstone–shale–limestone facies, (3) thick marine shaley sequence and (V) volcanics.

Formation (Bein & Weiler 1976). Simultaneously, the platform accumulated shallow marine limestones and dolomites (Fig. 4.5.3) of the Judea Group (Sass & Bein 1982). It is notable that a mirror-like picture occurs contemporaneously in the Persian Gulf (Cohen 1988). The third sedimentary cycle (Senonian) is characterized by open sea conditions, with both thickness and facies of chalks, marls, phosphorites and chert (Fig. 4.5.4) being controlled by the incipient folding of the Syrian Arc (Flexer 1968, 1971, Flexer & Honigstein 1984).

The Senonian begins with a remarkably high sea level, which established an environmental pattern entirely different from that of the preceding Cenomanian–Turonian, changing the platform into a ramp (Sass & Bein 1982). The pelagic chalk deposition persisted over the previous platform and basin areas (Mount Scopus and HaShefela groups, respectively). The increasing depth of the sea is distinguished by three steps, corresponding to the three sedimentary cycles. The steps are bounded by two major intervals, expressed in unconformities. Each step

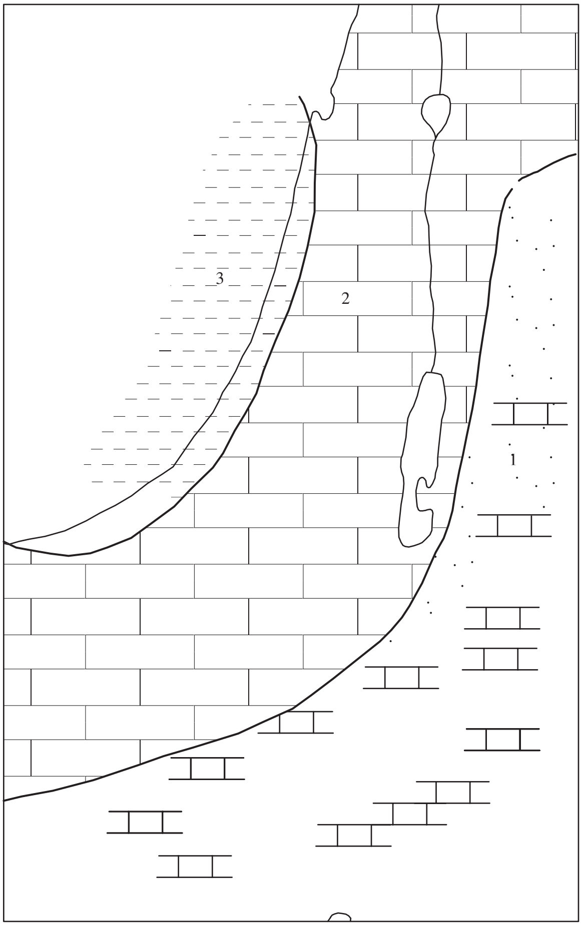


Figure 4.5.3. Mid-Cretaceous (Aptian– Turonian) lithofacies belts of the southern Levant: (1) thin carbonate sequence, interbedded with sandstone, (2) mostly carbonates (limestones and dolomites) and (3) mostly calcilutite facies.

comprises minor sea level oscillations and each major interval of unconformity can be subdivided into minor, subtle and partially local unconformities.

An important phase of distensional diastrophism, expressed in faulting and volcanism, characterized the Jurassic–Cretaceous boundary in Israel. The evidence for this tectonic phase is rather extensive, consisting of angular unconformities between Nubian Sandstone units in Sinai (Weissbrod 1969b), erosional disconformity in the Hazera erosion cirque, extensive occurrences of conglomerates and laterites in the Negev, Transjordan and Sinai (Wetzel & Morton 1959, Goldberg & Friedman 1974), development of extensive sebkha evaporites in the eastern Arabian Peninsula (Murris 1980) and incision of the Gevar'am Canyon in the continental shelf of Israel (Cohen 1976). The relative depth of this channel with respect to its shoulders, some 930 m, very significantly exceeds the eustatic sea level drop (Vail et al. 1977). Such a denudation necessitates, therefore, a considerable uplift in addition to sea level drop.

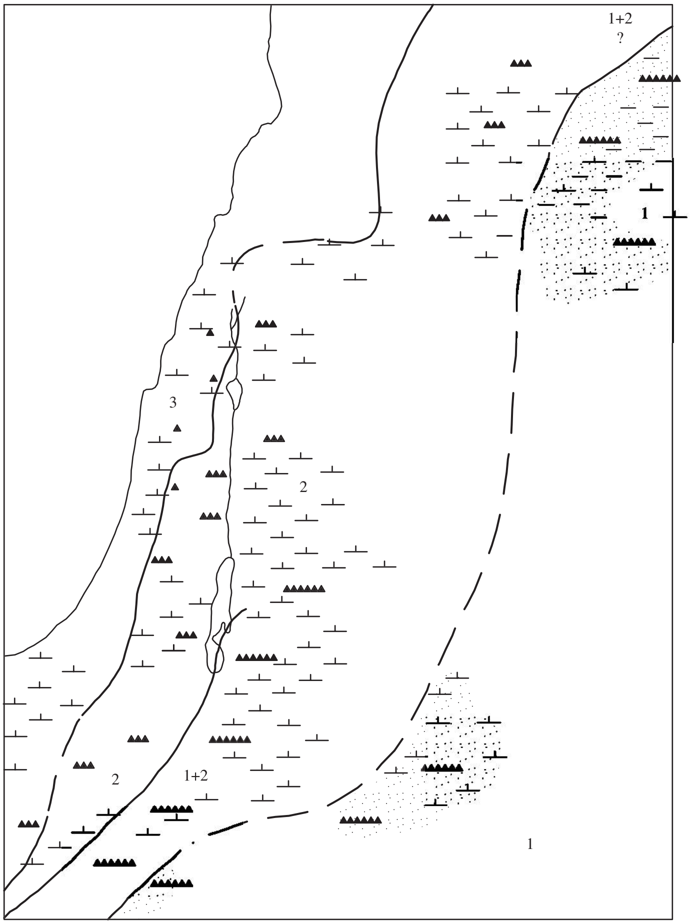


Figure 4.5.4. Senonian lithofacies belts of the southern Levant (modified from Flexer 1968): (1) sand, flint and chalk, (2) chalk, marl and flint, and (3) mostly chalk, sparse flint. Number of "teeth" represents thickness of flint.

Distensional diastrophism is further evidenced by extensive occurrences of basic-alkaline volcanism in northern Israel and the Lebanon during early Kimmeridgian to early Neocomian times (Dubertret 1954, Renouard 1955, Mimran 1972, Bonen 1980), by fault-induced thickness variations of basic flows and pyroclastics in Samaria (Mimran 1972) and of Kimmeridgian carbonates in the Haifa Bay and Carmel area (Derin 1974). Both the southward increasing depth of truncation in Israel and Transjordan (Bender 1974a, Bandel 1981, Druckman 1984) and the decreasing age of the lowermost onlapping units in the same direction (Freund et al. 1975), suggest that the uplift was asymmetrical with a hinge located somewhere in Saudi Arabia.

The litho- and biostratigraphy of the transgressive–regressive Cretaceous record in Israel is in good agreement with global eustatic trends. The early Neocomian eustatic low is represented by the development of Wealden facies in Israel, Transjordan and Lebanon (Aharoni 1964; Bender 1974a, pp. 70–83; Beydoun 1977). The extensive Aptian transgression is well recorded by a widespread blanket of carbonates, overlying the Neocomian terrigenous deposits in the coastal plain of Israel, termed the Telamim Formation (Cohen 1971), whereas only a thin representative is exposed in the erosional cirques of the Negev. The eustatic low at the late Aptian–Albian boundary is represented in Israel both on the shelf-platform and in the basin. In the former, there is a prominent phase of arenaceous and shaley intercalations between carbonate units, in the Mount Hermon area, Samaria, Judea and the coastal plain, termed the "Main Shale Break" (Bein 1974), while west of the hinge line, thick pelitic sequences of semi-pelagic facies (Gevar'am and Talme Yafe Formations) are separated by an unconformity (Bein & Gvirtzman 1977). A similar regressive phase is reported also from Saudi Arabia, where a clastic blanket covers the entire platform (Murris 1980).

A poorly documented Albian distensional phase is inferred from some recently dated hypabyssal intrusions of peralkaline basalt and syenite in the Timna and Ramon areas, dated at 107 Ma (Starinsky et al. 1980, Beyth & Segev 1983), the Terbol 1 well in northern Lebanon (Beydoun 1977), and in the southern Ansariyeh mountains in Syria (Dubertret 1937, Wolfart 1967).

The Albian to Turonian stratigraphic record of the platform carbonates in Israel is known and well documented, whereas details of the eustatic global sea level curve of the corresponding time interval were not released for publication (Vail et al. 1977). We therefore suspect that the differences between the local and global curves are not necessarily "real". The five sedimentation sub-cycles of the Albian to Turonian (Bein & Gvirtzman 1977, with chronostratigraphy after Lewy & Raab 1976) involve alternating phases of shallowing and deepening. The regressive phases are recorded by the following:

(1) The oolitic iron horizons of the Hidra Formation in the Galilee (early Albian, Rohrlich et al. 1980), or the Tammun Formation in Samaria and Judea.

(2) The fine-grained, laminated dolomites with relict gypsum nodules (Soreq and Bet Meir Formations of Judea; late Albian and early Cenomanian, respectively; Sass & Katz 1982).

(3) Coarse-grained dolomites, with strontium and ^{18}O isotopic evidence suggesting partial exposure (Sakhnin Formation of Cenomanian age).

The intervening, transgressive phases are represented by the following:

(1) Coarse-grained dolomites and skeletal limestones of Albian age (Giv'at Yearim and Kesalon Formations, Sass & Bein 1982).

(2) Cenomanian marlstone and fossiliferous limestone (Motza Formation, Sass & Oppenheim 1965).

(3) Ammonite-bearing argillaceous sequence of the early Turonian Derorim Formation (Freund 1961) and equivalent facies, recorded also in central Transjordan (Wetzel & Morton 1959) and in the coastal regions of Lebanon (Wolfart 1967).

Throughout the time span discussed the carbonate platform was fringed by a rudist biohermal belt, with lutitic carbonates being precipitated on the continental slope (Talme Yafe Formation, Bein & Weiler 1976).

The Turonian–Senonian boundary is characterized by a global eustatic fall, but without a corresponding tectonic event. Much of the top of the then youngest formations in Israel were exposed to vadose-karstic dissolution (Weiler & Sass 1972, Buchbinder et al. 1983), and a notable hiatus is recorded also in northern Transjordan (Bender 1974a, pp. 70–83), the Damascus basin and western Lebanon (Dubertret 1962, Wolfart 1967), and on structural highs in Saudi Arabia (Murris 1980). The preservation of a substantial Coniacian section in southern Israel and Sinai (Lewy 1975, Bartov & Steinitz 1977) is indicative of localized tectonic subsidence.

There is an overall good correspondence between the global Senonian eustatic cycles and those of Israel, represented by the Mount Scopus Group (Flexer 1968, Flexer & Honigstein 1984). Generally, during the Santonian–early Campanian a monotonous chalky column was deposited on the platform and west of the hinge line (Flexer 1968, Luz 1970, Bein & Gvirtzman 1977). Substantial thickness variations disclose the first folding phase of the Syrian Arc system (Bentor & Vroman 1951, Flexer 1968, 1971, Steinitz 1976). Toward the late Campanian and Maastrichtian, the progressing folding defined a facies pattern whereby in southern Israel and central Transjordan phosphorite, chalk, chert and oil shale accumulated, in partially sealed tectonic basins, with local development of hypersaline conditions (Bentor & Vroman 1951; 1960; Reiss 1962; Bender 1974a, pp. 70–83, Steinitz 1977; Nathan et al. 1979), whereas the flanks of the anticlinal highs accumulated either a thin cherty sequence or elsewhere some phosphoritic laterites (Nathan & Shahar 1977).

The isopach and facies trends cited are substantiated by a variety of geochemical studies, implying that much of the late Campanian to Maastrichtian sequence was imprinted diagenetically by meteoric water (Kolodny 1980, Kolodny et al. 1980, Amit & Bein 1982, Bein & Amit 1982). In the regional context, this folding is the reaction of the epidermal sedimentary column on the margins of the Arabian Craton to the oblique collision of Africa with Eurasia. This collision event, induced by a drastic change in the pattern of relative motion of Africa with respect to Europe (Dewey et al. 1973), is recorded in the northeastern Mediterranean by the emplacement of ophiolites along the Cyprus–Tauridic Arc and the Oman–Zagros Suture, during the early Campanian and Maastrichtian (Ricou 1971, Glennie et al. 1974, Şengör & Yilmaz 1981). It is represented in the Levant area by an "S"-shaped chain of asymmetric folds, with deep seated reverse faults (Picard 1943, p. 14; Renouard 1955; De Sitter 1962; Salem 1976; Mimran 1976).

To sum up, detailed lithological, microfaunal and biometric investigations, using relative abundance, diversity indexes and duration charts of ostracodes and foraminifera, allowed the recognition of sea level changes during the Cretaceous in Israel. Three major transgressive–regressive sedimentation cycles occurred on the northwestern margins of the Arabian Craton. These cycles are the Neocomian–Aptian, which is mostly characterized by terrigenous sediments; the Albian–Turonian, comprising basin marls and platform carbonates, and the Senonian, which is typified by uniform marly chalks. The cycles are separated by two major regional unconformities, at the Aptian–Albian and Turonian–Coniacian boundaries. The sedimentary cycles are related to regional tectonic and volcanic events and eustatic changes. The paleodepth curve illustrates the gradual rise in sea level, reaching its maximum during the late Cretaceous, with conspicuous advances during the late Aptian, late Albian–Cenomanian, early Turonian, early Santonian and early Campanian. Major lowstands occurred at the Aptian–Albian, Cenomanian–Turonian, Turonian–Coniacian and Campanian–Maastrichtian boundaries. This model for Israel agrees well with other regional and global sea level fluctuations. Four anoxic events (black shales), accompanying the transgressions, correspond to the Cretaceous oceanic record. One could hypothesize the presence of mature oil shales in the present-day eastern Mediterranean basin, close to allochthonous reef blocks detached from the Cretaceous platform.

4.6 PALEOGENE

At the early stages of the Paleogene (Paleocene, Eocene and Oligocene), a transgression re-established extensive marine conditions over most of the northern part of the Arabian shelf, persisting until the late Eocene. A widespread transgression occurred between the Paleocene and early Eocene. More restricted conditions developed as the sea level fell during the Oligocene, culminating in a late Paleogene hiatus, during which almost all the Oligocene and late Eocene beds were removed from over much of Arabia and the Levant by erosion. The Paleogene of the northern margins of the Arabian Craton is characterized by its homogeneity and uniformity of sediments in space, and by the transition in time from a transgressive regime at the beginning of the period (e.g. Paleocene to middle Eocene) to a regressive regime at the end (late Eocene and Oligocene).

Extensive deposition of marl, shale and chalk across the Middle East marks the Paleocene transgression. These soft sediments are part of the late Cretaceous sediments, and resemble them in the field, and together comprise the Mount Scopus Group (Flexer 1968). No break is discerned in the field at the Cretaceous–Tertiary boundary. Sedimentation was controlled by the tectonic relief of the incipient folds (Arkin et al. 1972, Salem 1976). The Paleocene marls (Taqiye Formation in Israel and Esna Shales in Egypt) accumulated mainly in the synclines, whereas on the anticlinal flanks they are thinner or wedge out completely.

An extensive early Eocene transgression depositing mainly chalk and flint, the Mor Formation (Braun 1967) or the Adulam Member of the Zor'a Formation (Buchbinder 1969), is terminated by the occurrence of conglomerates, extensive intraformational slumping and disharmic folding (Wetzel & Morton 1959 in Transjordan, Beydoun 1977 in Lebanon, Benjamini 1979, 1980 in Israel).

These data are suggestive of some sort of tectonic activity over the Middle East at the end of the early Eocene. The combined evidence suggests a peak in the folding activity, an event of regional extent, well expressed by orogenic movements in the Alpine–Tauride system. This orogenic phase, along the northern margins of the Syrian Arc, probably reflects the resumption of compression between the African and European plates (early Eocene), after a period of relative transcurrent movement between them, during the late Senonian–Paleocene (Dewey et al. 1973).

A new transgressive cycle begins in the middle Eocene, depositing chalk with a high content of planktonic fossils in the deeper areas to the west, while benthonic nummulitic and reef limestones were laid down over the shallow shelf areas. The southward and eastward extension of this middle Eocene rise in sea level was probably surpassed only by the late Cretaceous transgression. A new tectonic and sedimentary regime starts, following middle Eocene times (Picard 1943, p. 55).

Relics of late Eocene chalk, marl and some coral bioherms occur in the coastal plain, in the foothill areas of Israel, in the western and southern Negev and in the Arava (Bentor & Vroman 1963, Sakal et al. 1966, Gvirtzman 1970, Benjamini 1980). These rocks represent the last major transgression which reached into the interior of the Levant area from the west. Rocks of Oligocene age (Saqiye Group) are confined only to the Coastal Plain foothills area in Israel, to a depression in the Lake Kinneret–Wadi Taiyiba area (Wetzel & Morton 1959, Gvirtzman 1970, Michelson 1972) and to the deeper basin of the southern Dead Sea (Horowitz 1996a).

The present-day morphotectonic landscape of the Middle East was actually beginning to form during the late Eocene and Oligocene times. Fission track dating of apatites has established that the domal uplift of the Sinai area started in the late Oligocene, at 26.6 Ma ago (Kohn & Eyal 1981). However, there is extensive evidence that the accumulation of alkaline volcanics and coarse clastics in the Suez Graben, presumably post-updoming, was well under way in the Oligocene (Garfunkel & Bartov 1977), concomitantly with the opening of the Red Sea and the Gulf of Aden (Azzaroli 1968, Lowell & Genike 1972). Whether updoming of what would become the Jordan Rift Valley area, was the decisive factor behind the newly established sedimentation pattern, cannot be determined at present. An alternative explanation may have to do with the establishment of the ancestral Nile in the early Eocene, as a major supplier of clastic sediments to the southeastern corner of the Levant basin (Salem 1976). Not only did it affect the lithologic composition of the Saqiye Group in the coastal plain of Israel, it also elicited the accumulation of a thick sedimentary prism, which may have induced a fast subsidence of the continental margin.

To sum up, the Paleogene period commenced with a wide, extensive transgression, continued from the Cretaceous times and terminated with folding, uplift, faulting and volcanism. In the Paleogene there is in general a good correspondence between the global eustatic record and the inferred record of the northern end of the Arabian Craton. The major regressive cycle of the Paleogene in Israel starts in late Eocene times, preceding the major global eustatic fall of the early Chattian. The Mesozoic and Cenozoic unconformities are summarized in Table 4.1.

Table 4.1. Mesozoic and Cenozoic unconformities.

Age	Geographical distribution	General characteristics	Local activity	Worldwide events
Late Miocene* (Messinian)	Levant basin	Hiatus, evaporites and clastic deposits	Red Sea and Jordan Valley rifting	Mediterranean desiccation
Middle Miocene (Langhian)	Southeastern Levant basin	Erosion and angular unconformity, fault-talus breccias, fluviatile deposits	Tectonics related to the Red Sea and Gulf of Suez opening	–
Post middle Eocene,* Oligocene	Israel, Sinai, Transjordan, eastern Mediterranean	Denudation period and clastic deposition	Syrian Arc folding phase II	Global deep low-stand, Alpine Orogeny
Early-middle Eocene boundary	All over Israel, Lebanon, Transjordan	Intra-formational slumping conglomerates, hiatuses disharmonic folding	Movements in the basin floor	Initial orogenic movements Alpine–Tauride system
Cretaceous–Tertiary boundary	Israel (partially)	Paraconformity	–	Worldwide sea-level fall
Turonian–Senonian boundary*	Israel, Transjordan, Syria, Lebanon, Saudi Arabia	Early Coniacian hiatus, chalk vs. Limestone, karstic phenomena	Syrian Arc folding phase I	Global eustatic fall
Late Aptian–early Albian	Israel, Sinai, Saudi Arabia	Sedimentological break, clastics vs. Carbonates, "Main Shale Break"	Intrusive and extrusive activity	Conspicuous sea-level drop. Eastern Alps orogeny, discontinuities in ocean-floor spreading
Late Jurassic–early Cretaceous*	Widespread over all the Levant countries	Angular unconformity on a regional scale; conglomerates, laterite; deep incisions	Extensive block faulting, lava flows	Drastic lowstand, orogenic activities of Zagros and Hellenides, accelerated sea-floor spreading
Late-middle Jurassic (Callovian)	Widespread over Israel, Sinai, Saudi Arabia	Paraconformity with a small stratigraphic break	–	Global sea-level fall
Late Triassic*	Widespread: Israel, Transjordan, Sinai	Hiatus, lateritization karst, shallow incision	Faulting and lava flows	Rifting in the Tethys; global sea-level drop; early Cimmerian movements(?)
Middle Triassic	Deep boreholes in the Coastal Plain	Stratigraphic hiatus(?) and 600 m scarp conglomerate(?)	Block faulting	Initiation of the Tethys

*Major unconformities.

4.7 TECTONIC MOSAIC

As already mentioned, the Levant region, including the eastern Mediterranean, is situated between two major regional tectonic units, the stable Arabo-Nubian Massif to the south, and the Alpine mobile orogenic belt to the north. These two long-existing structural entities have, to a large extent, affected the Phanerozoic tectono-sedimentary history of the region (Fig. 1.4). The wide area between the Massif and the orogen includes the stable and the unstable shelf. This zonation has been outlined by several authors (Beydoun 1988, Cohen et al. 1988).

The stable shelf is characterized by a flat, comparatively thin sequence of slightly deformed sedimentary cover, up to 2 km made up of continental and shallow marine sediments, overlying the crystalline basement. The unstable shelf contains a fairly thick sedimentary sequence, up to 13 km, ranging in age from Paleozoic to Quaternary, including ample source and reservoir rocks. A favorable tectonic history provides the complementary ingredient that makes the unstable shelf of the Middle East a prolific oil province.

Nearly all the major faults in the Arabian Massif and surrounding zones (including Israel) are aligned in the following directions: northeast, northwest, north–south and east–west. Geological studies in various parts of the region indicate that these major fault trends may have been "etched" in the Arabian Massif since Precambrian times (Jarrige et al. 1986, Husseini 1988).

Surface and subsurface investigations in Israel confirm the existence of the northeast and northwest fault systems. Moreover, Phanerozoic differential movements along these faults are manifested in thickness and facies variations of the sediments (Cohen 1988). The effects of the north–south trending faults have mostly been felt since Neogene times, particularly in association with the formation of the Jordan Rift Valley. However, this subject is beyond the scope of the present discussion and is dealt with in Chapter 9. The east–west fault system is well expressed in the Negev and Sinai (Bartov 1974) by a shear belt, composed of six conspicuous dislocation lines. Geophysical studies indicate that several east–west faults also exist in central and northern Israel (Domzalski 1967, Klang & Gvirtzman 1983). They are concealed, however, by Tertiary sediments.

The Paleozoic and Mesozoic tectonic mosaic of the southern Levant was made up of northeast trending grabens (e.g. the Pleshet, Judea and Galilee grabens), their neighboring horsts and several northwest trending blocks (Cohen et al. 1990). These tectonic elements (Fig. 4.7.1) were bounded by regional faults. The grabens were formed in three known phases: pre-late Permian, early to middle Triassic and early to middle Jurassic. Formation of each graben resulted in a depositional sag basin, centered above the graben and extending beyond its boundaries.

Two types of inversions were active, inversion of regional tilts and inversion of vertical tectonic movements. Examples of regional tilts are the southeast, landward

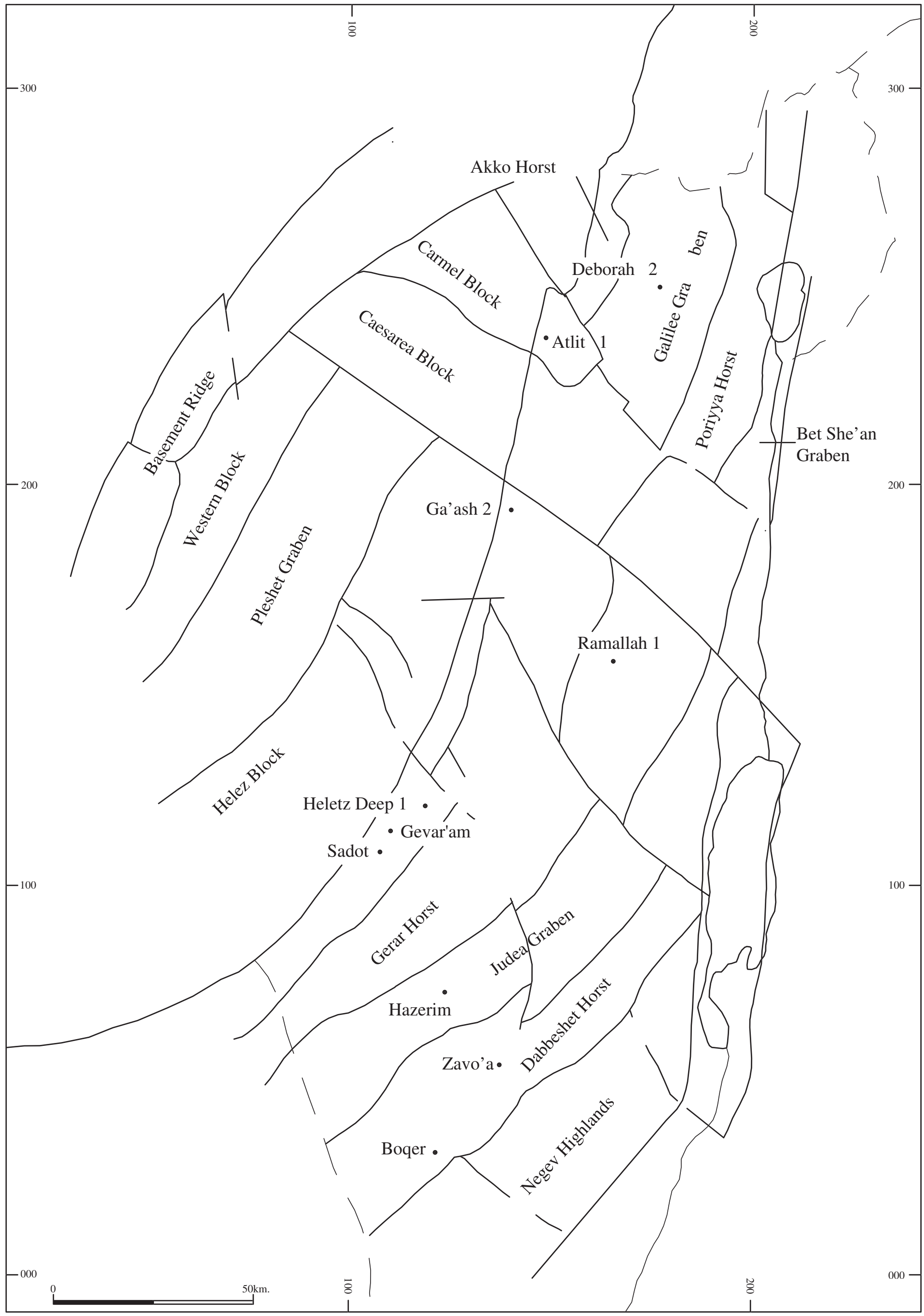


Figure 4.7.1. Tectonic pattern of the southern Levant, showing Paleozoic–Mesozoic blocks, grabens and horsts inverted in late Cretaceous times. Dead Sea, Bet She'an (central Jordan Valley) and Hula grabens formed during the late Cenozoic (modified from Cohen et al. 1990). Locations of boreholes mentioned in the text are indicated.

middle Triassic Anisian sedimentary wedge and the northwest, basinward wedge of the early Cretaceous (Berriasian to middle Aptian) sequence. A vertical inversion occurred in the late Cretaceous. The Paleozoic–Mesozoic grabens became raised blocks, whereas the horsts turned into depressed areas, in which synclinal sediments, rich in organic matter, accumulated in asymmetrical fault-fold sag basins. These events occurred as a result of intraplate stresses, and are related to major global orogenies.

The grabens, the sag basins, and also the argillaceous parts of the basinward sedimentary wedges, are considered to be sites of hydrocarbon generation. Commercial and sub-commercial production, as well as abundant oil and gas shows, confirm the presence of source rocks. Traps of various forms are widely available; in the horsts, in the grabens, in the sedimentary wedges and in the sedimentary sag basins.

4.7.1 Graben and horst block pattern

Detailed seismic interpretation in the northwestern Negev, based on 7,000 km of seismic lines, has revealed the existence of a buried Mesozoic northeast trending graben, flanked by raised blocks or horsts on either side. Tectonic events, including the intense folding and faulting that has taken place since late Cretaceous time, severely interfered with this Mesozoic pattern. As a result, the Paleozoic–Mesozoic block pattern is not expressed in the surface geology. The present seismic interpretation, however, unequivocally indicates that the late Cretaceous–Tertiary movements followed the pre-existing early Mesozoic lines of weakness, but in an opposite sense (Fig. 4.7.1). Thus, the bounding faults of the ancient graben were later reactivated, and currently delineate the limits of a recently formed horst. This tectonic analysis confirms the concept suggested by Freund et al. (1975), that a graben existed in central Israel, which is now the site of the large Judea–Samaria anticlinorium (Fig. 4.7.1).

Re-examination of existing data from various sources reveals the presence of a similar Mesozoic block pattern in other parts of the country. When put together, these elements form a discrete mosaic of blocks, grabens and horsts (Fig. 4.7.1). A map of the gross Jurassic thickness (Fig. 6 of Cohen et al. 1990) shows the effect of the graben and horst pattern on the bulk of the sediments. Steep contour gradients reflect the presence of the bounding faults. Thus, sedimentary units are thicker in the graben.

An anticlinorium with an eroded crest, flanked by Senonian beds, was located by seismic interpretation (Seismic Geocode Ltd. 1984) some 100 km west of the Judea–Samaria anticlinorium, now under the Mediterranean Sea. This anticlinorium is an inverted graben, similar to the Judea structure. It is situated at the center of the Pleshet graben (Cohen et al. 1988). A similar phenomenon is believed to exist in the Galilee highlands, making it also the product of an inverted graben (Fig. 4.7.1). A graben and horst pattern, as well as the inverted blocks, has been described by Lovelock (1984) in Syria and Iraq.

Magnetic data and a refraction survey (Ginzburg & Folkman 1981), as well as information from several deep wells (Ramallah 1, Deborah 2) show that the thickest sedimentary column in onshore Israel exists in the two graben-hilly ranges of Judea and the Galilee. According to magnetic data, the sedimentary sequence in the offshore Pleshet graben is even thicker.

Some of the blocks follow a northwest rather than northeast trend. Although the northeast bearing was more common and most of the blocks trend in that direction, northwest lineaments are in some places dominant. These lineaments seem to be part of the Najd system of faults, that are known to have been active in the Arabian Massif during the Precambrian (Jarrige et al. 1986). They have been reactivated several times in the history of the region. Movement along some of these lineaments seems to be taking place at the present time.

4.7.2 Tectonic inversions

Tectonic inversions played an important role in the geological history of the area. Two types of inversions have been observed: inversion of regional tilts, both basinward and landward, in a "see-saw" fashion; and inversion of the sense of vertical movement of blocks in a "yo-yo" fashion. The late Cretaceous vertical inversion (Syrian Arc) reactivated the pre-existing northeast bounding faults of the various blocks, as well as the smaller size normal faults within the main blocks. This resulted in the formation of folds of various sizes, that "lean" on these reactivated faults (Cohen et al. 1990).

Tilts affected the entire region and its contained blocks. They occurred in two main directions: basinward and landward (Fig. 4.4.2). The "basinward" tilt resulted in the formation of wedges which extended from the platform domain, where they were made up mainly of clastic sediments, into the deeper parts of the basin, where they comprised fine-grained deposits, such as the early Cretaceous Gevar'am shale. Another example of a basinward sedimentary wedge is the Ladinian Saharonim Formation (Picard & Flexer 1974).

The "landward" tilts, on the other hand, resulted in wedges made up mainly of clastic sediments, such as the Anisian Gevanim Formation (Picard & Flexer 1974), which is absent on the Helez block but attains a thickness of some 300 m in the northeastern Negev block. The clastic components of the Gevanim Formation were derived from the Arabian Massif to the southeast. Another source of clastic material could have been the offshore basement ridge, or a continent that extended from the ridge westward, which might have provided clastic material for the offshore Pleshet graben (Hirsch 1984, Cohen et al. 1988). There is some evidence that folding was associated with the landward tilt, due to development of a compressional stress.

The regional tilts, which strongly affected the sedimentation processes, are probably related to tectonic interplay between the Hercynian and the Alpine orogenic belts to the north, and the Arabian Craton to the south (Fig. 4.4.2). The alternating landward and basinward tilts resemble the motion of a see-saw.

The late Cretaceous inversion had a decisive effect on the present-day tectonic style. It was manifested in the reactivation of the pre-existing faults, as a result of which the grabens became elevated blocks, whereas the horsts turned into depressed ones. Similar phenomena are described in the northern Alpine foreland, resulting from compressional intraplate deformation. The elevated blocks were now subject to erosion, whereas the depressed blocks became sites for accumulation of synclinal deposits (Fig. 4.4.2).

Inversion of tectonic movements also occurred along smaller faults within the main blocks, although these reactivated faults did not everywhere pierce the entire sedimentary column. Strata closer to the surface were merely "bent" over the buried faults. A series of asymmetrical folds was thus formed, with their steep flanks draping over these buried, reverse faults (Reches et al. 1981, Gelberman & Kemmis 1987).

4.7.3 Sag basins

The geometry and nature of the sedimentary bodies which were formed in late Permian, late Triassic, late Jurassic and late Cretaceous times, reflect tectonic regimes different from the simple "graben and horst" pattern. During these time intervals, sediments were laid down in sag basins, of which two types have been identified: depositional sags, which are broad, interblock basins formed in a relatively quiet, passive tectonic environment, at the final stage of graben tectonics (cf. Harding 1984); and fault-fold sags, which are narrow, intrablock basins formed at times of upward inversion, in an active tectonic background (Fig. 4.4.2).

Late Permian, late Triassic (Fig. 4.4.2) and late Jurassic sediments were laid down in broad synclinal troughs, whose axes lay over the Judea or Galilee grabens, while their flanks extended far beyond the boundaries of the grabens, over the adjacent blocks. They seem to terminate periods of active graben tectonics. The raised areas between the sag basins, within each graben zone, reflect the presence of intervening "highs" on the graben floor.

A late Permian sag currently under study is outlined by wells drilled on the rims of the basin, as well as by seismic data in its central part. It is filled up with basal sands, overlain by shales and then carbonates, whereas clastic sediments occur close to this basin, similar to the Tabuk and Widyan depressions on the Arabian Craton, described by Al Laboun (1986) and by Sharief & Moshrif (1989).

The late Triassic sag is filled up with evaporites, mainly anhydrite and some salt, alternating with dolomite and limestone, partly reefoid closer to the edges of the basin. The late Jurassic sag is filled up with shales, overlain by carbonates and topped by a series of alternating sands, shales and carbonates. Reefs (barrier reefs, according to Derin 1974) are present on the western edge of the basin. Another sag basin may have existed further west, centered around the Pleshet graben (Fig. 4.7.1). This sag seems to be filled up with shales that were deposited in deeper waters, probably under euxinic conditions.

The distribution of Senonian sediments is related to troughs formed adjacent to the steep flanks of asymmetric anticlines. The steep flanks are caused by the draping of strata over concealed reverse faults. They were formed during the intensive tectonic regime of the late Cretaceous inversion. Many of the fault-fold sags are of relatively smaller dimensions, where they develop next to the asymmetric late Cretaceous Syrian Arc structures. They are much larger in the neighborhood of large structures. For instance, large Senonian sag basins exist on both sides of the Judea anticlinorium; the western occupies the entire width of the inverted Gerar horst, which subsided at that time; while a syncline on the eastern side of the anticlinorium occupies a large part of the inverted Dabbeshet horst (Fig. 4.7.1). A rise in level resulted in an extensive deep sea, whose bottom was pitted with these basins, wherein chalk rich in bitumen accumulated.

4.7.4 Conclusion

A tectonic mosaic existed in the area studied during the Paleozoic and Mesozoic, made up of northeast trending grabens (Pleshet, Judea and Galilee) and their neighboring horsts, as well as several northwest trending blocks. These tectonic elements were bounded by regional faults, aligned in both directions. Grabens were formed in the pre-late Permian, early to middle Triassic and early to middle Jurassic phases. Each graben formation process ended with a depositional sag basin centered above the graben and extending beyond its boundaries.

Two types of inversions were active, of regional tilts and of vertical tectonic movements. Examples of the regional tilts are the southeast landward sedimentary wedge of the middle Triassic (Anisian) Gevanim Formation, and the northwest basinward wedge of the early Cretaceous (Berriasian to middle Aptian) sequence. Vertical inversion occurred in the late Cretaceous, when most of the Paleozoic–Mesozoic grabens became elevated blocks, whereas the horsts of that age turned into depressed areas in which synclinal sediments, rich in organic matter, accumulated in asymmetrical fault-fold sag basins. The entire process was accompanied by regional tilts (landward or basinward). These tectonic events occurred as a result of intraplate stresses, related to major orogenic events of global extent.

The three Paleozoic–Mesozoic grabens, Judea, Pleshet and the Galilee, as well as the western part of the Helez microplate, were basinal areas, where source rocks, including shales and fine-grained carbonates were deposited. These units are referred to as "generative basins". There is conclusive evidence for the commercial potential of two of these provinces, the Helez block and the Pleshet graben, while oil and gas shows are known for the other two, the Judea and Galilee. Raised blocks between the grabens contain prospective areas with unconformity traps, faulted anticlinal closures of early Mesozoic age, and late Mesozoic to Cenozoic fault and anticlinal traps. Turbidity sands are expected to be present in the mouths of canyons that developed in the Helez block, particularly during the early Cretaceous and Neogene.

The depositional sag basins contain both reservoir and cap rocks in the form of clastic sediments, reefs, shales and anhydrites. Commercial and sub-commercial production, as well as widespread oil and gas shows, confirm the presence of source rocks, coupled with a favorable tectonic environment. Reservoir quality rocks are found in a variety of trap situations. Of particular interest are sectors that were never previously explored, or only very little, as well as deeper strata not penetrated by earlier wells. There are thus significantly untested basin areas and structures, both on- and offshore.

4.8 THE SYRIAN ARC FOLDING SYSTEM

The term "Syrian Arc folding system" is applied to a tectonic province in the Levant, distinguished by its huge "S"-shaped belt of folds. The term, as introduced by Krenkel (1924), has been recently revived, but "Levantides Fold Belt" is also in use. The Syrian Arc is only a segment in a larger arc of foreland folds, encircling the Arabian Craton from Egypt via Israel, Transjordan, Lebanon, Syria, Iraq, the Emirates and Saudi Arabia. The Syrian Arc folds have many common geological characteristics: northeast trend, asymmetry due to the presence of basement controlled reverse faults, and multiphase history of deformation. Ages of folding range from Turonian to Neogene (Quaternary?), whereas the deformation peaks are discerned at the late Turonian, middle Campanian, post-middle Eocene and late Neogene–early Quaternary (but see also discussion in Chapters 9 and 10).

Israel, northern Sinai, northern Transjordan, Lebanon and southern Syria are crossed by a series of surface and subsurface anticlines and synclines which strike mostly in east–northeast, northeast and north–northeast directions (Fig. 4.8.1). This folding system comprises a segment of a broader belt of folds, which continue eastward to Iraq and the northeastern parts of Saudi Arabia (Fig. 1.4). The belt is located mainly on the unstable shelf between the Arabian Craton and the Zagros–Taurus thrust zone.

Different tectonic interpretation and descriptive terms were proposed for the belt of folded mountains of the lands bordering the eastern Mediterranean. The common characteristic of this tectonic province is the presence of generally simple folds, developed in the sedimentary cover. The belt is, however, not uniform in its structural details, as some of the folds are basement controlled through deep seated reverse faults, while others are rootless, more superficial epidermis folds. Moreover, the tectonic activity is not synchronous, as some of the folds exhibit multiphase tectonic history, whereas others underwent simple deformation at different periods of time.

Yet, the Syrian Arc has many common tectonic characteristics and quite a definite time for its peak of activity. The general understanding is that this tectonic stage was initiated in the Turonian, roughly coinciding with the main Alpine stage, associated with the collision of the African–Arabian and Eurasian plates.

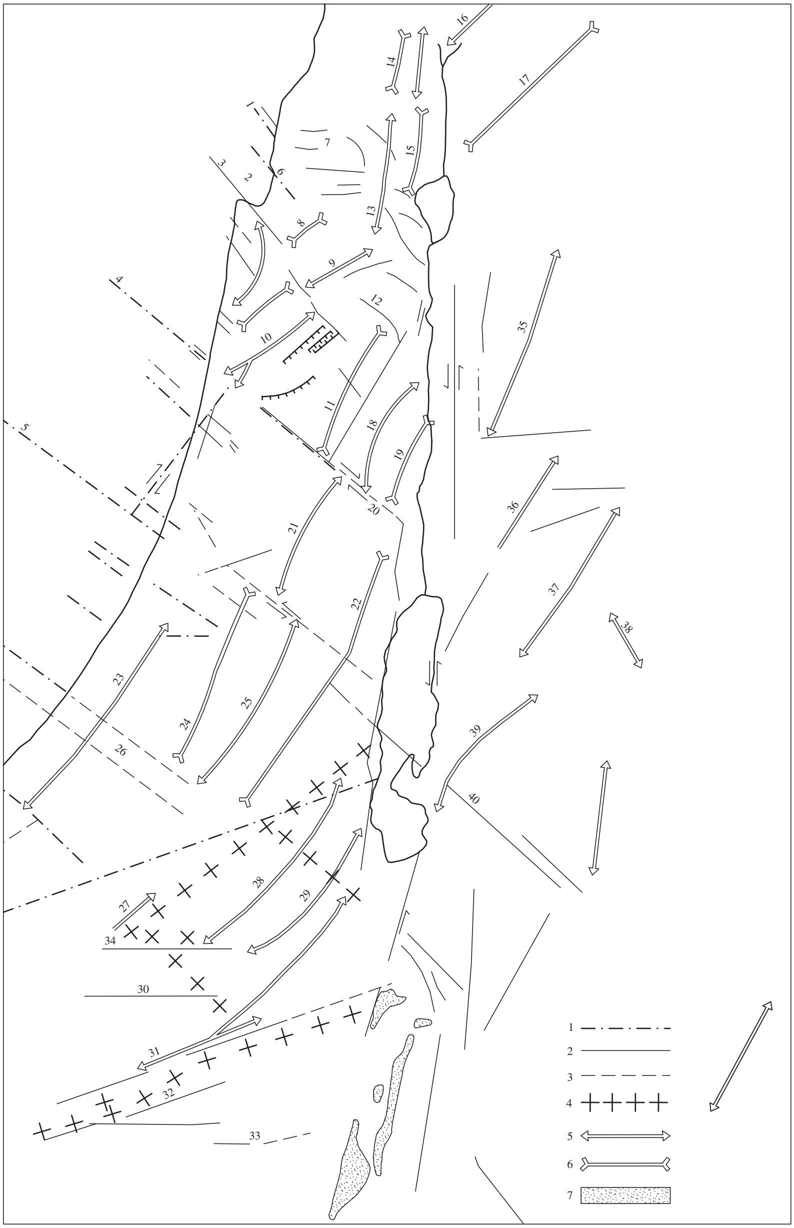


Figure 4.8.1. Folds and morphotectonic lineaments of the southern Levant (modified from Flexer et al. 1984). Legend: (1) inferred fault zones, (2) major fracture lines and lineaments, (3) geophysically based fault zones, probably in the basement, (4) subsurface ancient features, (5) anticline, (6) syncline and (7) Precambrian basement outcrops. Structure names: (1) Akhziv, (2) Qishon Canyon, (3) Carmel, (4) Dor, (5) Palmahim, (6) Akko, (7) Peqi'in, (8) Shefar'am, (9) Nazareth, (10) Umm el Fahm, (11) Nablus, (12) Gilbo'a, (13) Galilee, (14) Alma, (15) Rosh Pinna, (16) Hermon, (17) Golan, (18) Fari'a, (19) Sartaba, (20) Fatza'el, (21) Ramallah, (22) Judean Desert, (23) Helez, (24) Bet Guvrin, (25) Hebron, (26) Afiq Canyon, (27) Qeren-Rogem, (28) Hatira, (29) Hazera, (30) Avedat, (31) Ramon, (32) Arif, (33) Paran, (34) Zin, (35) Ajlun, (36) Suweileh, (37) El Qasr, (38) Es Safra, (39) Karak and (40) Karak-El Fiha.

The folds are well preserved in the Negev and Judean Hills, and, due to Plio-Pleistocene uplift, form broad upwarps of anticlinorial style, culminating at Makhtesh Ramon, Makhtesh Hatira and Judea. To these are attached the synclinoria or downwarps: to the west, the Haluza and Shefela foothills and to the east the Paran-Zin and Judean Desert regions.

Most of the Negev and Judean Desert landscape is characterized by a "Jura"-type fold morphology, with erosion cirques, cluses, ruzes and hogbacks on the frequently steep flanks of the asymmetric anticlines, with mesas and tabular features in the synclines. The fold morphology is less noticeable in western Judea, Samaria and the Galilee, most likely due to the effect of subsequent faulting and the influence of the more humid Mediterranean climate, forming caliche crust and terra rossa soils which cover most of the rocks.

The folds of the central and northern Negev highlands are all strongly asymmetric, with a steep flank to the southeast, a feature generally attributed today to the presence of reverse faults at depth. Other structures, such as the Hebron anticline, are strongly asymmetric to the west. Still other structures are roughly symmetric or of the box-fold type, with steeply dropping flanks on both sides. The folds are often described as concentric or parallel, i.e. not changing the thickness of their beds as a result of the compressive stress (changes in thickness inherited from the sedimentological pattern are common, but irrelevant to this topic). Superimposed younger domal structures are common, mainly in the anticlines of the northern Negev and the Ramon structure (Eran 1982). Folding started in the late Turonian, and the folds received their final shape between the Eocene and Miocene. They became uplifted and, especially to the north, block faulted during the Neogene and still more so during the Pleistocene.

The location of the Syrian Arc fold belt, together with the Zagros–Taurus, Cyprean and Hellenic Arc thrust zone, between Eurasia to the north and Arabia–Africa to the south, has interested geologists for decades. The advent of plate tectonics led to new explanations of the relationship between these two continental masses, with the basic concept that the African–Arabian plate is moving relatively northward and underthrusting the Eurasian plate (Scott 1981). The situation is even more complex, since the Arabian and African plates, now separated, were united ever since the Precambrian and up to Neogene times, forming the Arabo-Nubian Massif, eventually surrounded by the Syrian Arc fold system.

The Arabo-Nubian Massif is a classical cratonic nucleus, surrounded by concentric morpho-tectono-sedimentary belts. Such a picture is shown by many authors (Picard 1939, Henson 1951, Powers et al. 1966). Four such geologic provinces are observed: the Arabo-Nubian Massif; the stable shelf; the unstable shelf including a series of Paleozoic, Mesozoic and Tertiary lows and highs, having a complex inverted tectonic history, whose last tectonic product is the Syrian Arc; and the mobile belt of the Alpine thrust zone, with nappes and ophiolite emplacement (Fig. 1.4).

The thrust zone (the fourth belt) in the Taurus and Cyprean Arcs appears to be seismically active (Rotstein & Ben-Avraham 1986). Southward from the thrust zone the degree of tectonism decreases, through the fairly simple folded zone of the Syrian Arc extending from Sinai through Israel, to Transjordan and Palmyra in southern Syria. A good deal of the folds is buried under the eastern Mediterranean Sea.

Geographically, the Syrian Arc folds belong mainly to the third belt, the unstable marginal shelf of the Arabo-Nubian Massif. This is a foreland area, characterized by consistent shelf-type carbonate sedimentation, exceeding several thousands of meters in thickness, deposited during the Mesozoic and early Tertiary times. The southern part of the belt, toward the stable zone, underwent primarily epeirogenic and only mild horizontal movements. The stronger deformations to the northwest are represented by either local unconformities or facies changes, with anticlinal features and faulted uplifts. Here the dominant tectonic feature is the series of folds with axial planes paralleling the thrust zone and the outline of the Craton, dipping to the north or south, with high angle reverse faults (50–70°), with their steeply dipping southern or northern flanks, respectively.

The structures of the Syrian Arc share many common geometrical characteristics, which is expected from a genetic structural belt:

(1) The structures are trending generally to the northeast, with slight deviations according to the shape of the huge flat "S". In Sinai the trend is east–northeast, in the northern Negev northeast, in Judea and Samaria north–northeast, while the Damascus–Palmyra folds again repeat the northern Negev and Sinai bearings.

(2) The structures are arranged in provinces or clusters, and within each cluster the folds are organized in en-echelon rows. The prominent clusters in Israel are, from the south northward: the central Negev or Ramon area, the northern Negev, the northwestern Negev lowland province (where most of the structures, except for the Qeren-Rogem, are buried in the subsurface), the Judea–Samaria anticlinorial areas (including Mt. Carmel and Umm El Fahm), the coastal plain province (entirely buried, with Sadot and Helez the leading structures), and the Galilee (Fig. 4.8.1).

(3) The Sinai, Negev and Damascus–Palmyra folds are all strongly asymmetric, with the steep flank to the southeast. Other anticlinal structures, such as Hebron and Judea, are also strongly asymmetric, but with the steep flank to the west, and still other structures, such as Fari'a, are roughly symmetric or of the box-fold type with steeply dipping flanks on both sides.

(4) The asymmetry of the folds is attributed to the presence of basement controlled reversed faults (see Chapter 5) at depth (De Sitter 1962, Mimran 1976). The existence of this feature was substantiated by several deep oil wells where strata repetitions were discovered by many seismic profiles revealing the reverse faults at depth and by theoretical geometrical calculations (Reches et al. 1981).

(5) Magnetic and gravimetric anomaly lineaments accompany many of the structures. The magnetic anomalies represent an elongated escarpment (fault or flexure) at the top of the crystalline rock types, with different magnetic properties (Domzalski 1967). The gravimetric anomalies reflect both the basement features

and, mostly, the steep flexures of the top of the thick carbonate sequence of Judea Group.

(6) Geological and geophysical analyses show, with a high degree of confidence, that there is a genetic relationship between magnetic anomalies, the assumed deep seated basement reverse faults which penetrate the sedimentary cover and are expressed in seismic profiles, and the external epidermic folds associated with gravimetric and morphometric anomalies (Fig. 4.8.2).

(7) The ages of the folding, according to various authors, range from the early Turonian, during the Senonian to post-middle Eocene and even Neogene (Picard 1943, p. 34; 1959; Bentor & Vroman 1960; Freund 1961; Braun 1964; Flexer 1971; Eyal & Reches 1983). Apparently the folding took place in several stages. Activity and intensity are different in time and space (Fig. 4.8.3). Evidence for the tectonic activity comes usually from angular unconformities between Turonian, Senonian and Eocene strata, variations in the thickness of Senonian rocks, or diastems. Stratigraphic condensation of the Senonian succession marks the crest of the anticlines, whereas continuous, fully developed thick sections characterize the synclines (Bentor & Vroman 1960, Flexer 1971, Bartov & Steinitz 1977).

Gvirtzman (1979), based on studies carried out in the Shefela coastal plain and Be'er Sheva regions, distinguished two phases of folding. During the first, from the Senonian to the middle Eocene, broad, long wavelength (30–40 km) and long (70–80 km) structures came into being. On these, many narrow (5 km) and

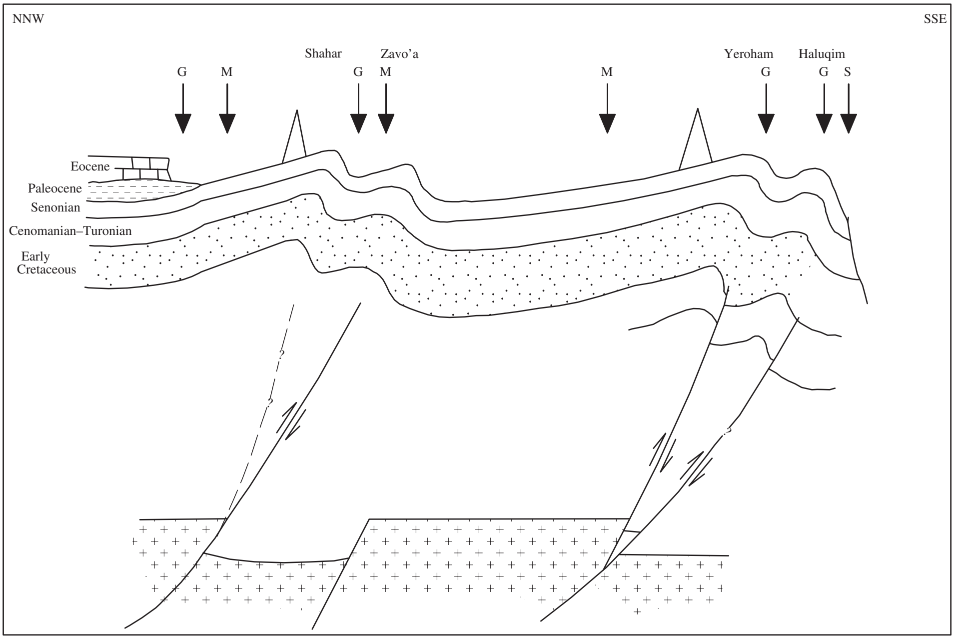


Figure 4.8.2. A typical cross section through Syrian Arc folds, showing the association between structure, gravimetric (G), magnetic (M) and seismic (S) anomalies.

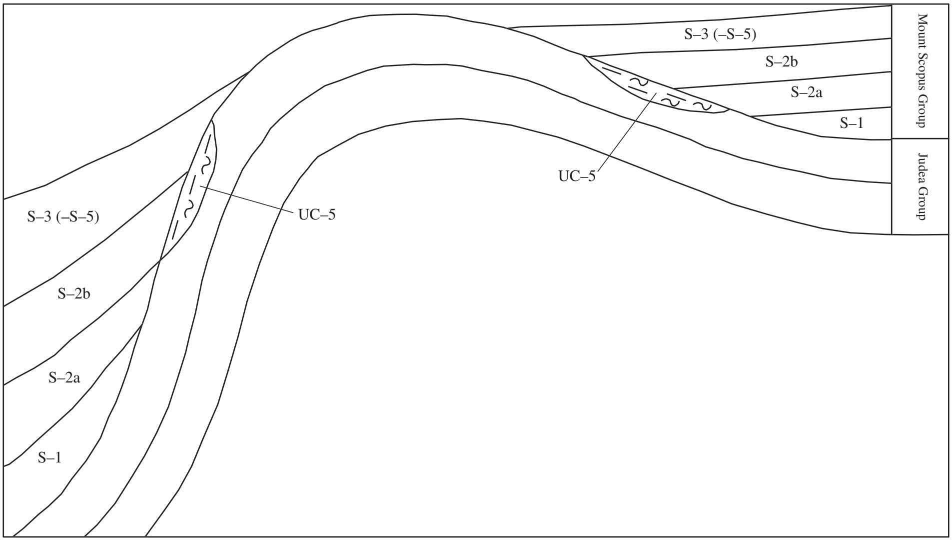


Figure 4.8.3. Schematic cross section through the Judean Hills, showing onlap relations of Senonian and earlier formations, as defined by ostracode biozonation (modified from Flexer et al. 1989): (UC-5) *Neocyprideis vandenboldi*, Turonian, (S-1) *Phyrocythere lata*, late Coniacian, (S-2a) *Cythereis rosenfeldi rosenfeldi*, early Santonian, (S2-b) *Veenia fawwarensis fawwarensis*, middle Santonian, (S3) *Limburgina miarensis*, late Santonian and (S5) Campanian.

shorter (25 km) anticlines and synclines were superimposed during a second phase, lasting from the late Eocene to the early Miocene. Bartov et al. (1980a), in a detailed study of the Areif en Naqa structure, found that asymmetric folding, accompanied by faulting, took place in the early Coniacian, persisting through the late Campanian–Maastrichtian. Post-Eocene folding formed the present-day symmetric component of the structure.

Eyal & Reches (1983) found that a west to west–northwest trending compressional stress field was responsible for the creation, in late Cretaceous–Neogene times, of the folds west of the Jordan Rift Valley, referred to as the "Syrian Arc Stress". These authors distinguished yet another stress field, active in Neogene–Recent times, with dominating horizontal extension, trending east to east–northeast, in all rocks inside the Rift and its neighboring regions.

Mimran (1984), investigating the Fari'a structure in Samaria, concluded that the structural interpretation of folding development in this structure does not conform to the model of folding migration, as suggested by Freund & Zak (1973). He found significant folding in late Turonian times, the late early-Eocene, between the middle Eocene and Pliocene and a fourth, post-Pliocene phase. Honigstein et al. (1988), while analyzing some of the Negev and Judean Desert structures, by studying ostracode biozonation, found that late Turonian–early Santonian times are distinguished by one of the main folding phases of this belt, forming angular unconformities by onlap position of the sediments and local faulting. Recurrent and continuous movements during the Santonian and Campanian are responsible for the thickness variations, caused by condensed sections on high structures, and extended sequences in low areas, as well as by faults (Fig. 4.8.3).

(8) A comparison of the geometric parameters of the northern Negev anticlines, measured in cross sections perpendicular to the fold axes, shows that the amplitude of the folds is directly related to the width of the steep flank (Eran 1982). Obviously, the highest amplitudes and widths in one single anticline are those measured in the section which passes through the crest of the fold. These parameters therefore indicate the extent of development of the anticlines. The comparison has been made on the uppermost Turonian unit.

(9) The cores of the Syrian Arc structures expose, in general, Triassic to Turonian strata, whereas Senonian to Paleocene rocks are associated with the flanks. Younger strata, such as of Eocene and occasionally Senonian age, are restricted to the synclines. It should be noted that the Qeren-Rogem and Giv'at Hayil are the only exposed structures in Israel where Paleocene (Taqiye Formation) or Eocene (Adulam Formation) rocks compose the core of the anticline. Deep subsurface analysis shows (see Paragraph (6), above) the possibility of basement participation controlling the shape, tectonic style and amplitude of the anticlines.

(10) Some calculations of the relationships between the mean throws of the reverse faults in the anticlines and their amplitude, as reflected by the top of Judea Group show a ratio of 1 : 6–1 : 8.

Various authors have speculated on the originating mechanisms of the Syrian Arc fold system. Today, most scholars seem to agree that the Negev structures are the result of reverse faults at depth. Such faults have by now been identified in six deep wells, located on five anticlines of the northern Negev. Freund et al. (1975) assume that every anticline in Israel is underlain by a reverse fault, which originated as a Mesozoic normal fault; such an assumption would explain why the anticlines are asymmetric to one side or the other, or sometimes box-shaped, as well as the varying outlines of several anticlines, interpreted as the surface expression of the zigzag shaped subsurface outline of the Mesozoic normal faults.

The concept of structural inversion of Freund et al. (1975) is commonly accepted, although not every present-day anticline was previously a graben; sometimes the concept fits a province of several associated folds. Druckman (1984), examining the Negev structures, concludes that contrary to Freund et al. (1975), the early Mesozoic and Tertiary structures are not related genetically, nor geometrically. He prefers interference between broad and shallow Mesozoic basins and narrow, high amplitude Tertiary folds, as an explanation for the existing fold pattern.

Neev et al. (1982) relate formation of the Syrian Arc folds to compression, resulting from an oblique collision of "slices" of the crust, due to a systematic northward counterclockwise conveyance movement of the slices between the sinistral (NNE trending) shears. It is agreed, however, that the tensional regime which prevailed in Israel since the Triassic up to early Cretaceous, has been changed during the late Cretaceous to a compressional one; Fig. 4.4.2 shows the change in stress regime in our region.

Arbenz (1984) summarizes that during the formation of a collisional thrust belt, due to the northward migration of the Afro-Arabian Plate into a subduction zone along the Iranian–Turkish microcontinent, a broad belt of the Israeli–Lebanese–Syrian continental foreland underwent mild "thick skin" (i.e. basement included) compressional deformation, resulting in the early state of the Syrian Arc. He states that the geodynamic cause for this foreland or "cratonic type" deformation is not properly understood.

Intensive tectonic activity occurred during the Senonian in Cyprus, Turkey and Syria (Robertson 1978, Biju-Duval et al. 1979). This is expressed by the lower intensity of the Syrian Arc folding (Krenkel 1924) and by fault structures, discerned by lateral thickness variations of the sections, from nil to a few hundred meters (Bentor & Vroman 1960, Freund 1965, Flexer 1971, Bartov & Steinitz 1977). These thickness variations are related to the tectonic evolution of the paleostructures, which started in the Turonian (Braun & Hirsch 1987) and continued during the Senonian (Fig. 4.8.3).

Thickness changes in low amplitude and long wave structures are, in general, difficult to interpret, because of the poorly exposed lateral continuity of the sediments, and because of hiata. In cases where the thickness varies over short distances without changes in the homogenous chalky lithology, ostracode biostratigraphy can be used for the resolution and interpretation of different types of tectonic mechanisms (Honigstein et al. 1988). Some outcrop sections in the northern Negev and Judean Desert were chosen as case studies, because they are well exposed and easily accessible.

The resolution of ostracode biozonation has allowed us to discern, within the late Coniacian–Campanian homogenous chalky units, three basic types of reduction in thickness:

(1) Onlap relationships, in which horizontal strata were deposited progressively against initially inclined surfaces. Here, the tectonic event predated deposition of the beds.

(2) Condensed successions, containing all litho- and biostratigraphic units, against nearby extended sequences. Such cases indicate simultaneous and continuous or repeated tectonic activity during the deposition of the sediments.

(3) Thickness variations, revealed by biozonation, on both sides of a fault, which can determine (Fig. 4.8.3) the age and movement of the fault (Flexer et al. 1989).

Although the case studies described by Honigstein et al. (1988) represent only a relatively small part of the entire tectonic history of the Syrian Arc in Israel, they contribute greatly to knowledge of the different tectonic mechanisms of this folded belt. It is shown (Honigstein et al. 1988) that the Syrian Arc structures exhibit a multiple history of folding, uplift and faulting, resulting in thickness variations due to onlap relationships and stratigraphic condensation.

The main tectonic phase, observed in the case studies, took place during the late Turonian–early Santonian; a minor peak faulting activity occurred during the middle part of the Campanian. The stratigraphic condensation, continuing during the Santonian and Campanian, indicates activity during the entire period.

The results obtained by Honigstein et al. (1988) agree well with those of numerous authors, describing a late Turonian and syn-Senonian activity of the Syrian Arc folding belt (e.g. Bentor & Vroman 1960, Freund 1965, Flexer et al. 1970, Flexer 1971, Bartov et al. 1972, Bartov & Steinitz 1977, Braun & Hirsch 1987). One should remember that the Senonian phase described here represents only one stage in the evolution of the Syrian Arc folding system, which continued into the Neogene (e.g. Picard 1943, p. 34; Eyal & Reches 1983), or even into the Quaternary (but see also Chapters 9 and 10).

In the context of the present review, the age of the latest folding activity related to the formation of the Syrian Arc fold belt is of particular interest. Freund (1965, p. 192) infers a post-Pliocene differential continuation of the upheaval of the Negev highlands, from a detailed study of the Hazeva Formation in the northern Negev.

Evidence for tilting and the timing of the last phase of deformation in the southern Hebron hills is indicated by the structural position of the Ziqlag and Bina formations on the main structure, and indicates they are from Miocene to Recent (Gilat 1980, Gilat & Arkin 1978). This seems to require further confirmation as far as recent movements are concerned. Eran (1982) describes evidence for a post-Pliocene displacement along the Hazera lineament (i.e. the line of inversion along the eastern Hazera anticline). However, this is probably not related to folding, but rather to warping or faulting. Eyal & Reches (1983) state that about half of the Syrian Arc folding developed during the Neogene, or even during the Pleistocene, with amplification of folds occurring primarily by limb rotation.

Mimran (1984) finds a "post-Neogene" age for the latest folding phase in the Fari'a anticline (NE Samaria). Subsurface evidence from electrical resistivity measurements may possibly indicate a Holocene subsidence of the Boqer anticlinal ridge in relation to the Yeroham syncline, but seems to require further study and corroboration by supporting evidence.

In conclusion, it appears that the folding of the Syrian Arc system came to an end during Neogene times (post-Hazeva Formation). Any folding movements posterior to this time were connected either with shearing, faulting, or continuation of the folding of the Syrian Arc fold system, but it will be difficult to resolve this question in an unequivocal manner. This situation does not preclude theoretical reasoning about the stress fields involved in the formation of the present fold pattern, during different periods, and about the question of whether this pattern developed independently from the mechanism that brought about the formation of the rift system (Red Sea, Gulf of Suez and Jordan Valley), or whether it was a reaction to, or part of the stresses and motions that formed the rift system.

4.9 THE TRANSVERSAL FAULT SYSTEM OF ISRAEL

The southern Levant is crossed by a possibly polygenetic system of transversal faults, particularly conspicuous between the Jordan Rift Valley and the Mediterranean Sea, but also known from Transjordan (Fig. 10.2.1). By and large, both the intensity and frequency of the faults increase from the south northward, probably as a result of the decrease in width of the Sinai–Israel sub-plate. However, a conspicuous shear belt, composed of at least six major faults, occurs in the Negev and reflects the basement tectonic control on the relatively thin sedimentary cover. The trends of the fault systems vary slightly in the different geographic provinces. The main trend in the Negev is east–west, whereas in central Israel both east–west and ESE–WNW bearings exist. The Galilee fault system exhibits an east–west and both WNW–ESE and WSW–ENE directions. Timewise, the transversal fault system of Israel represents a multiphase geological evolution.

Surface and subsurface mapping of Israel, together with magnetic and gravimetric studies, portray a variety of lineaments across Israel that strike in NW–SE, WNW–ESE and east–west directions. Some of the features are barely visible faults of presumably a left or right lateral slip component, whereas others are quite well-documented tectonic lineaments, with or without strike-slip movement (Figs 4.8.1, 10.2.1 and 10.2.2). For all practical purposes the division between the fault systems is on the basis of geographic zonation: the Negev and Sinai, central Israel (Judea, Samaria and the coastal plain) and the Galilee. The fault systems in the three separate regions show some difference in characteristics (bearing, extent of basement control, type of fault, sense of motion) which may suggest different generation.

4.9.1 The Negev–central Sinai shear belt

The transversal shear-fault belt of the Negev and central Sinai is a dominant tectonic feature of this region. Elucidation of how recently this system has been active is of prime importance to the assessment of the tectonic stability of the Negev. The following brief review will first summarize some general information regarding the shear-fault belt, and then proceed to deal with the shear lines adjacent to the northern Negev anticlines, the Zin and the Sa'ad-Nafha, and some other unnamed subsurface lineaments.

The central and southern Negev are transversed by six shear zones (Fig. 10.2.1), oriented ENE–WSW to east–west (Bartov 1974): the Thamad–Wadi Sudr line, the Paran–Areif en Naqa–Buruq line, the Arif–Batur line, the Ramon– Minshara line, the Sa'ad-Nafha–Helal line and the Zin line. Excluding the Zin, all these shear zones extend into central Sinai. Each of these exhibits one or more of the following: variations in the direction of stratigraphic throw in various segments of the shear zone; changes of inclinations of the fault plane; offset (usually dextral) of cross-cutting structural elements; oblique to horizontal slickensides on the fault plane; development of small-scale folds (anticlines, domes and half domes, monoclines, V-synclines) and extensional features (rhomb and wedge-shaped grabens), with their axes being subparallel to, or at a low angle with, the trace of the shear zone.

Based on some of these criteria, Bentor & Vroman (1954a) were the first to conclude that the Paran–Areif en Naqa–Buruq, Arif–Batur, Ramon–Minshara and Zin lines have a major strike-slip component. Freund (1965) claimed that their horizontal slip cannot exceed a few kilometers, based on the offset of some early Turonian facial belts. This conclusion was later substantiated by the restoration of offset tectonic elements (e.g. dikes), or by estimation of the amount of slip from the degree of shortening of strata, in compressional structures located at the termination of some of these lines (Bartov 1974, Baer 1981, Zilberman 1983). The maximum horizontal offset is inferred along the Ramon line, not exceeding 4 km. The maximum vertical throw along the Thamad and Paran–Minshara lines is in the order of 1 km (Bartov 1974). A smaller yet significant stratigraphic throw is recorded at the Arif–Batur line, up to 400 m (Baer 1981).

The incongruence between the direction of paleo-stresses of the Syrian Arc system (Senonian to Neogene?) as inferred from the analysis of meso-structural elements, and the direction of macro-structural elements (e.g. the Syrian Arc anticlines) in the central and northern Negev, led Eyal & Reches (1983) to conclude that the orientation, as well as the recurrent nature of movement along the Negev–Sinai shear belt, were induced by the yielding of pre-existing weakness lines in the crystalline basement to WNW oriented compressive stresses. Similar views were quoted by previous authors (see references in Bartov 1974, p. 111).

Conclusive evidence for the earliest movements along some of the transversal shear belt is based on facial changes of Senonian rocks across the Areif en Naqa and Zin lines. The latest documented movement along the shear lines must have postdated the Hazeva Formation, as it was demonstrated that the Hazeva sediments along the Sa'ad-Nafha, Paran, Arif–Batur and Ramon lines are deformed relics, preserved due to downfaulting (Bartov 1974, Zilberman 1983), rather than fillings of pre-existing depocenters, as assumed by Bentor & Vroman (1951). The overlying clastic sediments of the Arava Formation were shown to be undisturbed along these lines, such as the Karkom graben along the Paran line (Bartov 1974, Avni 1998). While this is by no means closed to debate, the top conglomerate of the Hazeva Formation (Ashalon Member) is considered to be of late Miocene to early Pliocene age, whereas the Arava Formation is considered as Pliocene (Garfunkel & Horowitz 1966, and see Chapter 7 of the present book). These age relations prompted Horowitz (1979, p. 54) to include these faults within the Eritrean stage, a view also maintained throughout the present work (see Chapter 9).

Several early Miocene dikes in Sinai, 20–22 Ma old (Steinitz et al. 1978), are offset by some of these faults. The youngest age cluster obtained from fission track dating of epidotes, from several transverse fault systems in the Precambrian outcrops of southern Sinai, is Oligocene to middle Miocene, 32–11 Ma (Bartov & Steinitz 1977, Steinitz et al. 1978). These ages can provide an upper limit to the movement along the Negev–central Sinai shear belt, assuming that the dated systems are coeval to those that transect the sedimentary cover further north, and that the epidote mineralization closely followed the faulting.

Shear zones with marked similarities to those of the Negev–central Sinai are known from Transjordan, east of the Jordan Rift Valley. Quennell (1959) and Bartov (1974) demonstrated that they can be matched with their Negev–Sinai counterparts, by restoring the 105–110 km movement along the Dead Sea Transform (see also Chapter 10). This indicates that the formation of the shear zones preceded the movement along the Dead Sea Transform, and that they do not form a conjugate set to the Jordan Rift Valley system faults (Bartov 1974). This observation, while demonstrating that the formation of these shear zones is pre-early Miocene, does not provide an upper age limit to the movement along the shear lines.

The following are short descriptions of each of the fault lines, from the south northward:

The Thamad fault is one of the longest in the system. It starts in the Arava Valley north of Elat and continues westward to the Gulf of Suez, a traverse of approximately 200 km. The fault system is associated with volcanism and iron mineralization (Bartov 1974). In the fault segment crossing the Negev, the northern side is the downthrown block, with a vertical displacement of 0–1,000 m (Garfunkel 1970). Garfunkel presumes that this is a rejuvenated fault, with a right lateral movement of 2.5 km.

The Paran fault extends along 150 km from Jebel Abu Kandu to the west, through Jebel Harim, Jebel Areif en Naqa, Har Batur, Har Haspas, the Beroqa and Karkom grabens and the anticlinal structures of Rekhes Menuha and Kippat Eshet, meeting the Arava Valley to the east (Fig. 3.1.3). The trace is marked by a continuous fault system, accompanied by many small fold and fault controlled structural depressions, which form a conspicuous morphotectonic belt (Zilberman 1985). Several strike-slip offsets, of 200–600 m, were measured on dikes and half domes which were displaced by the Paran fault in Sinai (Bartov 1974). Tectonic activity, inferred from the fold structure of Areif en Naqa in eastern Sinai, was traced back to the Triassic (Bartov et al. 1980a). Faulting commenced in the Senonian and continued to the middle Eocene, while a second phase lasted during the Neogene and a third one in the Pleistocene (Zilberman 1985).

The Arif–Batur fault extends along some 85 km, from Sinai in the west to the Arava margin in the east. The line is marked by a continuous fault system, accompanied by many small folds and fault controlled depressions, which form a conspicuous morphotectonic belt. The belt system disappears in the easternmost part of the line, from Har Massa to the Arava margin, where its only manifestation is a line of small structures. The trace of the Arif–Batur line is crossed, near the Arava margin, by two southeast trending flexures, which occur on the continuations of the Ramon and the Sa'ad-Nafha lines, but are not disturbed by the Arif–Batur line (Zilberman 1983). The line was defined by tectonic activity which began in the early Santonian and continued until the early Eocene. Tectonic stability characterized the central Negev during most of the early and middle Eocene, and the small structures which were formed along the line were covered by the thick sequence of the Eocene Avedat Group.

The stable tectonic stage came to an end after deposition of the Hazeva Formation, when the entire line was reactivated by an intensive tectonic phase. The present morphotectonic character of the line appeared during the post-Hazeva Formation tectonic event. This event was associated with intense folding activity, followed by the development of an east–west trending fault system, which exhibits a considerable vertical throw and is accompanied by structural patterns characterizing dextral strike-slip fault systems (Zilberman 1983). The intensive structural deformation along the line was followed by rapid erosion, which formed the present drainage system of the area. The eroded materials were deposited in the structural lows, which existed along the line, and along the main channels of the drainage system. This complex sequence is part of an extensive clastic cover, which was deposited south of the Ramon structure after the post-Hazeva Formation tectonic event, and is known as the Arava Formation (Zilberman 1983, Avni 1998).

Several terraces which were preserved along the main wadi channels in the structural lows, reflect a regional deposition and truncation events that postdate the Arava Formation. These can be attributed either to tectonic activity along the base level of the present drainage system, which flows toward the Arava basin, or to climatic events that took place during the Pleistocene and left similar terraces in other parts of the Negev (Zilberman 1983, Avni 1998).

The Ramon fault has a zone of tectonic deformation with a width of 1–3 km and length of some 40 km (Garfunkel 1964). The zone is associated with thrust faults, steep flexures, overturned folds and complex strata undulations. A lateral displacement has not been discerned along the fault, but a thrust movement related to a deep-seated basement fault was suggested (Garfunkel 1964).

The Sa'ad-Nafha fault extends from the western margins of the Arava to Jebel Helal in north central Sinai. This shear belt is characterized by segments of presumably normal faulting with a strike-slip component (Nahal Masor), wedge shaped syncline with several morphotectonic basins containing faulted Abu Treife and Hazeva formations (Mahmal, Nahal Hava and Marzeva), several steep anticlinal domes (Hamran, Qemer, Nafha and Sa'ad), and two half domes bordering upon a graben (Jebel ar Risha). Detailed observations by Zilberman (1983) in the Har Sa'ad and Har Nafha area demonstrate that this line was active after the deposition of the middle Eocene Avedat Group. In the Mahmal depression further east, the top conglomerate of the Hazeva Formation is downfaulted against the Eocene Matred Formation. However, the latest age of activity along this fault cannot be constrained, because of the absence of post-Hazeva sediments. The maximum measured vertical throw is in the order of 100 m. The amount of lateral slip is estimated to exceed 200 m (Zilberman 1983).

The Zin line is a broad faulted zone, containing a conjugate set of normal and shear faults oriented east–west, NW to WNW and NNW to NNE. These faults crosscut and terminate the southern plunges of the Boqer, Haluqim and Hatira anticlines. The Zin line is probably the surface expression of a deep seated basement fault zone, having reverse and right lateral oblique displacement. A right lateral component of displacement has been detected along this line, as evidenced by local slickensides having lateral component of movement; en-echelon, stepped left pattern of surface faults; and structural analysis of horizontal elongation produced by en-echelon normal faulting (G. Shaliv, Water Planning for Israel, Tel Aviv 1988, pers. comm.).

A different model for the Zin line can also be suggested, by which the deep seated basement is normally faulted, still with its downthrown side to the south, which marks the plunges of Boqer and Ketef Shivta anticlines, together with gravity anomalies.

According to Shamir (1983) the Zin line is an east–west tectonic strip, containing a variety of deformation phenomena: faults (east–west to WNW); plunges of monoclines from the northeast (Hatira, Haluqim, Boqer, Shivta), forming flexures parallel to the line (primarily the Havarim–Darokh flexure); asymmetric half domes (Ma'ale Zin and Darokh) leaning against faults, which have a reverse to normal, dextral slip component; a variety of meso-structures, and thinning of the Senonian Menuha Formation chalk toward the faults. The activity along the Havarim–Darokh segment has been independent of other segments of the Zin line, and inherent in the clockwise rotation of the Hatira anticline; it began prior to or during the deposition of the Menuha Formation and succeeded after deposition of the Maastrichtian Ghareb Formation.

Several researchers (Bartov 1974, Zilberman 1983) exclude the Zin fault from the other five central Sinai–Negev shear zones, while others (Shaw 1948, Bentor & Vroman 1951, De Sitter 1962) tend to include it as an integral part of the system. The Zin line differs from other faults mainly in its surface expression, not comprising a single continuous fault line, but a very broad, 10–15 km wide zone, composed of numerous normal faults in segments. The Zin line also does not extend eastward to the Arava Valley as clearly as the other central Sinai–Negev features. Nevertheless, in spite of these differences, the Zin line is a major structural feature that is probably associated (as are the other five) with a basement fault. The Zin line has been active since the Senonian (Bartov et al. 1976, Shamir 1983), and also in Pleistocene times (R. Gerson, Department of Geography, Hebrew University, Jerusalem 1988, pers. comm.).

Faulting along the Zin line is known to have displaced the Quaternary Sede Zin Conglomerate and younger deposits. Alluvial gravels correlative to the Sede Zin Conglomerate along Nahal Besor, which traverses the Haluqim and Boqer anticlines adjacent to the Zin line, are not deformed, indicating that the faulting along the Zin line, during at least the past 0.5 Ma, was not accompanied by deformation on these northeast trending folds. There is no evidence suggesting that a kinematic link exists between the northeast trending folds and the Zin line faults, under the present stress field.

Meso-structure analysis by Shamir (1983) indicates that two separate stress fields have been active during development of the Zin line. The earlier, having a compressional axis direction of 290–310°, is believed to be related to folding of the Syrian Arc structures; the later, having a compressional axis direction of 330–008°, is believed to be related to the Dead Sea stress regime, active since the Miocene or Pliocene. These results are consistent with similar meso-structure studies made by Eyal & Reches (1983) on a more regional scale. The northeastern Negev anticlines, north of the Zin line (the area of Shivta to Haluza) are underlain by several east–west trending magnetic anomalies, which may indicate basement faults (Domzalski 1967). The southern anomaly offsets the Qeren axis and forms the Qeren promontory; another, a middle anomaly forms en-echelon relations of the Qeren and Rogem structures; a third, northern anomaly offsets the Hazerim and Rogem structures. If the magnetic anomalies do indeed reflect basement faults offsetting overlying sedimentary structures, they can be regarded as a northward continuation of the Negev shear belt.

To sum up, tectonic activity along the central Sinai–Negev shear zone in general, and along the Sa'ad-Nafha fault in particular, began in the Senonian. Several Miocene dikes, with absolute ages ranging from 18–22 Ma, have been displaced 0.6–2.5 km by various faults of the system (Steinitz et al. 1978). Zilberman (1983) showed faulting activity along the Sa'ad-Nafha fault in the Mahmal area, which affected the early to middle Miocene Hazeva Formation. Lately, some low magnitude earthquakes have been detected close to the Sa'ad-Nafha fault (A. Shapira, Geophysical Institute of Israel 1984, pers. comm.), which may indicate that it is currently active. The Zin fault was active in the Senonian (Bartov et al. 1976), but Shamir (1983) indicated that its latest age of activity cannot be constrained at this stage. R. Gerson (Department of Geography, Hebrew University, Jerusalem 1988, pers. comm.) had shown that the Zin line was active after the beginning of Matuyama, 780 Ka ago.

4.9.2 Central Israel fault pattern

The fault pattern of central Israel differs from the Negev shear belt by its clockwise-bearing rotation (20–30°) and the unproved genetic relationship to the basement. Three provinces of faulting can be differentiated, but the association between the faults over the provinces is highly debatable.

The eastern Samaria region is distinguished by several NW–SE faults. The major ones, from south northward, are the Samia, Fari'a, Buqei'a and Malih faults (Mimran 1972, Begin 1975a) and the Gilboa fault, the latter comprising a segment of the Carmel–Gilboa system. The Samia fault is eventually a fault strip, about 2 km wide, trending generally NW.

South of coordinate 158, within the Samia fault strip (Begin 1975a), the down-faulted blocks are on the eastern side. Fault planes are almost vertical, sometimes accompanied by fault breccia. Vertical displacements are about 250 m between Jericho and En Samia, decreasing northward. From coordinate 159 northward the displacement increases again, reaching 250 m in the Kariuth–Jalud region. Northwest of Nu'eima, the faults of the Samia strip are discontinuous and mostly parallel to each other, creating horsts (Qubbet e Najame), grabens (En Mu'akar) and step faults, marked in the topography by steep slopes and quite deep valleys. The blocks on both sides of the major faults usually dip toward the fault. Between Jericho and Wadi Abiad, vertical slickensides are found on the steeply northeastward dipping fault plane, indicating a vertical downthrow of the eastern block. A vertical fault plane, striking 310° at coordinates 1852/1483, shows slickensides dipping 45° southeastward, indicating a considerable dextral horizontal component.

Another prominent fault trends east–west along Wadi Batum, with vertical throws of up to 100 m. The Fari'a, Buqei'a and Malih faults form a horst and graben system, begun in post-Jurassic times, still active during the Neogene and possibly Quaternary times (Mimran 1972, 1984). Several east–west structural lines cross western Samaria, forming a horst and graben system associated with iron mineralization and volcanism.

4.9.3 Offshore lineaments

NW–SE trending offshore lineaments have been studied mainly by geophysical methods. Their westward continuation on land is highly debatable. The following is a brief description (after Flexer et al. 1984) of the lineaments, from the north southward (Fig. 4.8.1).

The Carmel–Gilboa sinistral strike-slip (De Sitter 1962, Freund 1965, Domzalski 1967) and its extension into the eastern Mediterranean (Ben-Avraham & Hall 1977, Kafri & Folkman 1981).

The Dor-Fatza'el line, which offsets the Fari'a and Ramallah anticlines and terminates the Nablus syncline quite abruptly. This line also coincides with a hinge, south of which there is a significant thinning of the sedimentary cover overlying the basement (Domzalski 1967, Ginzburg & Folkman 1981). Mart et al. (1978) suggest a dextral offset along this conjectured fault, while Ginzburg & Folkman (1981) imply a sinistral movement.

The Palmahim lineament is an off- and onshore feature. While its submarine extension is considered by some merely as a line of rootless disturbance (Almagor & Garfunkel 1979, Garfunkel et al. 1979), others (Mart et al. 1978, Ginzburg & Folkman 1981) suggest that shear movement occurred along it. In accordance with the second interpretation, the inland extension of the Palmahim line was inferred as offsetting the Hebron and Ramallah anticlines.

The Gaza-Afiq-Be'er Sheva Neogene erosion channel (Neev 1960) is probably superimposed on a fault zone (Gvirtzman 1970). While there is no documented shear movement along its inland portion, Mart et al. (1978) suggest that lateral displacement occurred along the submarine extension of this line. The abrupt termination of some structural axes across the inferred lineament in the Be'er Sheva Valley is worth noting. The El Arish pre-Neogene channel is probably of an origin similar to that of the Afiq-Gaza one. Additional smaller lineaments of similar trends are inferred from morphological, geological, seismic and aeromagnetic data (cf. Domzalski 1967, Klang & Gvirtzman 1983). They are presented with no further elaboration in Fig. 4.8.1.

The structural scheme presented in Fig. 4.8.1, while highly interpretative and undoubtedly still in need of further research, suggests that central and northern Israel is divided into several segments by tectonic lines bearing WNW–ESE (Flexer et al. 1984).

4.9.4 Galilee

The Galilee is the most densely fault-infested zone in Israel, characterized by several sets of faults. A very good agreement exists between the fault pattern and geomorphology, indicating quite recent tectonic activity. Three fault provinces can be discerned in the Galilee: the eastern Lower Galilee with its tilted blocks; the central and western Galilee, characterized by a horst and graben system; and the northwestern Galilee, with an ENE system, continuing to the Lebanon.

The eastern Lower Galilee is distinguished by a conspicuous consistent fault pattern, trending WNW–ESE and forming a system of tilted blocks, with their downthrown side heading to the north. From the south northward the following blocks are observed: Gilboa (geographically belongs to Samaria), Qumi, Kokhav HaYarden, Sarona, Poriyya, Arbel and Migdal. The geomorphological pattern matches the structural habitat. Faulting commenced in late Miocene times (first phase) than in late Pliocene (second phase), displacing the Intermediate Basalt of middle Pliocene age, 3.7–3.1 Ma old (Mor & Steinitz 1985). Structural "reorganization" of the Galilee during the Pleistocene (Picard 1943, p. 40) caused rejuvenation of the faulting (the third phase).

The central and western Galilee are characterized by east–west trending faults, forming a horst and graben system. From the south northward the most prominent structural units are: the Nazareth high, the Tur'an Valley, the Tur'an high, Bet Netofa Valley, Yodfat high, Sakhnin low, Karmon-Hazon high, Bet HaKerem depression and the high block of Bet Gan-Meron. Another horst and graben system also occurs in the western Galilee (Freund 1965), which is distinguished by its "piano keys" type, i.e., the faults are dying out eastward and strengthen westward. The subsurface structure of the western Galilee shows a highly developed east–west fault pattern, forming a horst and graben system, although most of the faults have either sinistral or dextral lateral displacement. The northwestern corner of the Galilee is distinguished by ENE long faults, having mainly right lateral sense of motion (Ron et al. 1984). This group of faults continues toward the Lebanon.

4.9.5 Structural analysis

Structural studies of the Galilee (Ron et al. 1984) test and confirm a kinematic model, showing that rotation of faulted blocks and strike-slip displacement along their boundaries are two contemporaneous aspects of a single deformation process. One of the fundamental predictions of this rigid block rotation model is a quantitative relationship among the faults spacing (width of blocks), fault slip and sense and magnitude of rotations. Blocks would rotate clockwise in a left lateral fault domain, and anticlockwise in a right lateral system. The total strain, therefore, can be calculated from the above structural parameters. Paleomagnetic measurements of rotation, independent of the spacing and slip data, provided a rigorous test for the validity and accuracy of the model.

The deformational history of northern Israel was reconstructed, based on meso- and macro-structures, paleomagnetic data and the geometric model. The Galilee has been intensively deformed by three sets of faults. A right lateral set, trending northeast, and its left lateral conjugate, trending NNW. The maximum principal compressive stress axis trended east–west, in accordance with the sense of shear on the conjugate sets of faults. Westward these faults splay and curve, to trend east–west with dip-slip normal displacements. The conjugate sets are truncated by late Miocene–early Pliocene peneplain and are therefore older. A test of east–west trending normal faults responds to the horsts and grabens in the Galilee. These faults displace the middle Pliocene basalts and the late Neogene peneplain. The faults are therefore younger than the conjugate lateral sets.

The stress field responsible for the normal faulting phase had rejuvenated some of the older lateral faults. Both deformation phases are associated with a continuous north–south extension of the region. The transition from an east–west compressional field to a north–south tensional one may have resulted from the formation of the morphotectonic depression along the nearby segment of the Jordan Rift Valley.

CHAPTER 5

The lithostratigraphy of the Embryonic and Eritrean stages

Lithostratigraphy, as such, should have been an objective description of rock units and their characteristics. This seems like a simple task when dealing with sedimentary formations of marine origin, uniformly covering considerable areas, in which lateral changes of lithology are the exception, rather than the rule. A similar treatment can be applied to large-scale magmatic structures, either plutonic or volcanic, which occupy a substantial volume. The difficulties begin with rock units deposited in small-scale basins, usually continental. There, lateral changes of lithology and thickness are the rule, not the exception, so that a detailed description of a given sequence could become meaningless at the next available outcrop, within a very short distance. In such instances, the term "type section", so popular in lithostratigraphy, seems to lose its primary significance and in my opinion, should only be applied with care.

The major question then remains: what could be used instead? It seems that there is no other way than applying also genetic parameters, such as fluviatile, lacustrine, paludine and so on, implying that lithology changes from the center of the basin or channel outward, while thickness always diminishes to zero, more often than not over a relatively short distance. The last-mentioned fact brings us to the next complication: erosion phases must also be regarded as "lithostratigraphic" units, if a reliable reconstruction of geological history is to be attempted. These are naturally of no lesser importance than rock units for reassembling the geological history, however they usually yield subordinate information, as compared with the hard evidence encompassed within the rocks.

The earliest conspicuous faulting manifested in the Jordan Rift Valley, of the Eritrean phase, is dated to the late Miocene, or the latest part of the middle Miocene (Schulman 1962, Horowitz 1987c, Steinitz & Bartov 1991). This phase was not restricted to the limits of the present-day Rift, nor its preferred north–south bearing. Subsidence, more extensive than of Syrian Arc synclines in neighboring regions, is detected along the Jordan Valley, preceding the Eritrean faulting. This subsidence may have been initiated by faulting of the Red Sea to the south, may also be connected with the last folding stages of the Syrian Arc, but, however, is clearly accentuated along the Rift.

The rock units described here are thus divided into three categories: the earliest, Embryonic stage formations, from the Oligocene through the termination of the middle Miocene, which are thicker within the Rift as compared with adjoining areas due to synclinal subsidence, rather than any conspicuous faulting. The second suite, of Eritrean formations from the late Miocene through the earliest Quaternary, was deposited in taphrogenic basins along but also outside the Jordan Rift. Both the Embryonic and Eritrean suites were deposited in the Jordan Valley while the region comprised a system of intermediate basins, either on the routes of rivers to the sea, or for marine ingressions filling up this series of depressions. These formations were also deposited and widely crop out outside the Jordan Rift limits, while within the Rift they are usually known mostly from boreholes, buried under younger strata, always attaining greater thicknesses than outside the Rift limits. Their nature and extent are controlled both by tectonics and by sea level changes, while climate plays a subordinate role. The stratigraphic terms applied to these formations and erosion phases are, therefore, those defined for the late Cenozoic sea levels changes (Tables 1.4.1 and 7.1, Fig. 9.1.1).

The third division, of Levantine formations spanning approximately the last two million years, refers to sequences accumulated within the Rift in its present-day structural configuration, namely a north–south oriented endoreic system having no connection with the sea. Their characteristics are chiefly affected by structural and hydrological conditions, both of which are frequently changing, with the latter considerably influenced by climate. Their stratigraphy is thus based on climatostratigraphic principles, commonly used for the Quaternary all over the world (Tables 1.4.1 and 11.1, Fig. 6.6.2). Most of these formations bear witness to human settlements, which are described and discussed in Chapters 11 and 12. Separate discussions are dedicated to igneous rocks, most of which are not necessarily confined to the Rift.

A problem frequently faced while considering the stratigraphy of the Jordan Rift Valley is the proliferation of names. This largely originates from two sources: students investigating certain outcrops, applying local formation names; and the political situation which, for many years, did not encourage contacts between Jordanian and Israeli geologists. An effort is made in Chapter 7 to unite terminologies, based as far as possible on priorities, on either side of the river. This problem becomes more acute with the arrangement of formations into groups. It seems that a real chaos exists here, so that probably the best policy would be to refrain from using "groups" at all, unless these are indeed clearly defined.

The present chapter deals primarily with exposed rock units, while only subordinate attention is focused on the subsurface. Although many formations are described here from outcrops which are outside the Rift limits, their equivalents are known from boreholes drilled in the Jordan Rift proper. The rock units are described for each period in ascending order, from north to south. Two exceptions are: when a certain unit is more thoroughly studied, its description precedes others; and some local, restricted or problematic occurrences, although given a formal stratigraphic status, are put at the end of the relevant section. Distribution of most of the rock units described appears on the geological maps, Figs 3.2.1–3.2.4.

Chapter 6, is dedicated to the subsurface continuous sequences from boreholes, where stratigraphy could be resolved mainly by pollen analysis; after clearing this, the exposed sections' unconformities are discussed. An overall discussion of chronostratigraphy, lists of relevant fossils, correlation of outcrops with the subsurface and reconstructions of the environments, are advanced in Chapter 7. Locations mentioned in the present chapter are marked on the relevant paleogeographic maps accompanying Chapter 7. Correlations are summarized in Table 7.1.

5.1 EMBRYONIC STAGE FORMATIONS

The Embryonic stage occupies the period ever since the end of the Eocene, when the Tethys had retreated from the area now occupied by the Jordan Rift Valley (see Section 4.6), until some time in the late middle or early late Miocene, when the Eritrean stage faulting began (Horowitz 1992a, Chapter 9). It is subdivided into two phases, approximately following the main Mediterranean transgressive cycles (Gignoux 1955, p. 554). First is the middle Oligocene cycle, transgressing into an intricate drainage system formed earlier toward retreating seas at the transition from the Eocene, its courses mainly occupying structural lows formed by the Oligocene uplift and folding. The extent of the middle Oligocene transgression is manifested by marine sediments on the eastern shore of Lake Kinneret and, further south, by similar deposits in the southern Dead Sea basin and by filling up of channels connected to the latter area with fluviatile and lacustrine deposits. Regression in the late Oligocene renewed erosion, at approximately the same channels as before.

This cycle was followed, during the early and middle Miocene complex of transgressions, by renewed filling up, partly of these channels but also of others newly formed, mainly with fluviatile and lacustrine sediments. The process ceased rather abruptly some time in the later part of the Miocene by the onset of the Eritrean faulting.

5.1.1 Oligocene

Fossil-bearing marine Oligocene sediments are known with certainty from two localities in the Jordan Rift Valley: the first comprises several outcrops on the eastern escarpment, near the settlement of En Gev on Lake Kinneret's eastern shore (Michelson 1972, Michelson & Lipson-Benitah 1986), near Khirbet Shuna some 10 km south of the lake, and in Wadi Taiyiba, another 10 km southward (Bender 1974a, p. 90). The second occurrence is known only from underground, namely the deepest section, almost 100 m thick, of the Sedom Deep 1 borehole, drilled at the southern end of the Dead Sea (Horowitz 1996a), which is dealt with in Chapter 6. Other outcrops, possibly of Oligocene age (Bentor & Vroman 1951; Horowitz 1979, p. 69; Horowitz 2000a) but with no characteristic fossils, are of the Zefa Formation and Abu Treife Series, outside the Rift but intimately connected with its ancient drainage system. The first is located in the Ef'e syncline, west of the northern Arava (Shahar 1973), the latter on the northern flanks of the Ramon structure. The Wadi Bustan Member (lower part of Dana Formation), or at least its lower part, may also be of Oligocene age (see Section 5.1.2.5).

Other occurrences of fossiliferous marine Oligocene beds are known from the Kingdom of Jordan, quite far eastward, from several boreholes and from outcrops in Wadi Sirhan (Wetzel & Morton 1959, p. 170; Daniel 1963), and from the coastal plain of Israel to the west (Avnimelech 1936, Martinotti 1981).

The sequence near En Gev, described in detail in Michelson (1972, but published formally only in Michelson & Lipson-Benitah 1986), comprises three formations: Fiq, Susita and En Gev Sands, which are known only from limited outcrops east of Lake Kinneret, from the vicinity of En Gev down to about 20 km south of the Lake, where the Taiyiba Beds and the lower part of the Usdom Group (Wetzel & Morton 1959) seem to correlate with the Fiq and Susita formations. The sequence overlies unconformably middle and late Eocene limestones and chalks of the Zor'a Formation ("Sar'a Chalk and Flint Formation" in Transjordan), and is overlain, again unconformably, by the clastic Miocene Herod Formation (upper part of the Usdom Group in Transjordan). Slight unconformities, both erosional and angular, are also quite common within the sequence.

5.1.1.1 *Fiq Formation*

Author: Michelson (1972, formally published in Michelson & Lipson-Benitah 1986).

The type section is in Wadi Fiq (Nahal En Gev) on a hill north of Susita near En Gev, on the eastern shore of Lake Kinneret (Fig. 5.1).

The base of the Fiq Formation overlies a regional unconformity developed on middle Eocene rocks, and is quite significant in the landscape due to a typical light brown-greenish cliff of detrital limestone, standing in contrast to the soft underlying chalk. The upper contact is much less definitive, and the Fiq grades into the overlying Susita Formation rather conformably. There is a typical horizon rich in *Pecten* shells at the transition, which forms the base of the overlying Susita Formation. The unconformity plane at the base of the Fiq is wavy and irregular. Above and near it numerous casts, filling burrows, are common, as well as barite concretions and glauconite grains. The Fiq Formation outcrops are known only from a very limited area in the En Gev–HaOn region, approximately the southern half of (the eastern shore of) Lake Kinneret, somewhat more than 10 km long (but see also discussion regarding the Taiyiba Beds). The thickness at the type section is 45 m, and is usually maintained wherever this unit crops out.

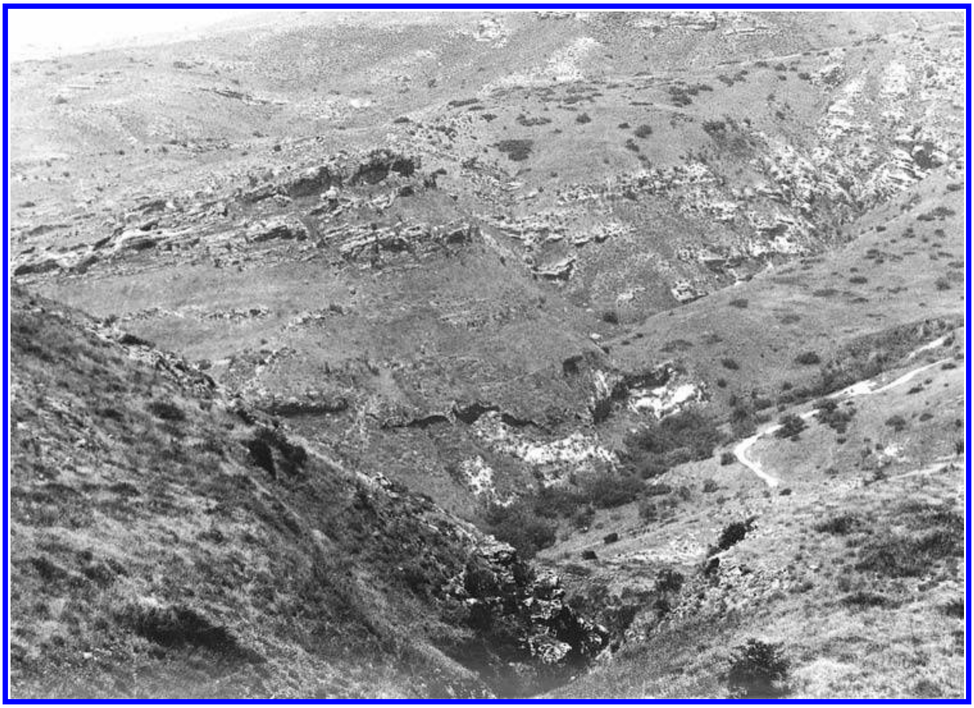


Figure 5.1. The Oligocene sequence east of Lake Kinneret: the Fiq Formation makes a soft landscape, overlying the Eocene Zor'a Formation at the bottom of the wadi. The well-bedded Susita Formation overlies the Fiq, itself overlain in the far distance by En Gev Sands and the Herod Formation.

The Fiq Formation is built chiefly of detrital limestones (pelbiosparite), glauconitic, quite rich in foraminifers, with some chalk, marly limestones and marl. The latter are slightly glauconitic, whitish–yellowish. The lower part of the Fiq is very rich in glauconite, up to 30% of the rock, diminishing upward. Burrows are occasionally also present in the upper part of the sequence, which is more chalky and contains some flint nodules.

The typical fossils of the Fiq Formation are detailed in Section 7.1.1. They comprise fragments of large benthonic foraminifera, common in the pelbiosparite facies, together with planktonic globigerinids which become more frequent toward the middle part, but the benthonic elements again prevail in the upper levels. Some ostracodes and *Lithothamnion* algae are also found, as well as mollusks. The environment of deposition is littoral-neritic. According to Lipson (1971), the lower part of the Fiq, rich in glauconite, is of late Eocene age (Foraminifera Zone P15, possibly P16) overlain unconformably by its upper part, of late early Oligocene age (P19). Reconsidering, Michelson & Lipson-Benitah (1986) assigned the Fiq to a single biozone of early Oligocene age of zones P18–P19.

5.1.1.2 *The Taiyiba Beds*

Authors: Nasr & Medaisco (1947, unpublished, cited in Daniel 1963); amended by Wetzel & Morton (1959).

The type section is in Wadi Taiyiba, some 20 km south of Lake Kinneret, 2.5 km east of the village of Waqqas (Fig. 5.2). The Beds overlie unconformably, over an erosional relief, the Eocene Sar'a Chalk and Flint (Zor'a) Formation, and are overlain, again unconformably, by the Usdom Group (Herod Formation) sands.

It is quite difficult to assess the exact distribution of the Taiyiba Beds. Daniel (1963, p. 377) remarks that "similar beds are not known elsewhere in East Jordan" (Transjordan). However, he suggests correlations of the Taiyiba Beds with other, rare occurrences of possibly Oligocene sediments penetrated by several boreholes, quite far away from the outcrop, and with the Oligocene sequence reported in Wetzel & Morton (1959) from Wadi Sirhan, some 150 km to the southeast. Incidentally, Wetzel & Morton had not correlated these outcrops, but rather thought that the Taiyiba Beds were deposited by a sea connected to the Mediterranean, not the Indian Ocean as in the Wadi Sirhan sequence. The occurrences of *Austrotrillina* in the En Gev suite seem to corroborate Daniel's correlation, which means that the

Figure 5.2. Taiyiba Beds, at the outlet of Wadi Taiyiba south of Lake Kinneret, overlain by the Herod Formation.

middle Oligocene sea, transgressing from the southeast, covered large parts of Transjordan.

The basal 8.5 m comprise glauconitic marls, sandy at the top. These are overlain by 11 m of yellow-greenish marls, bearing limonite nodules, while the uppermost 21 m form cliffs of white to yellowish hard limestone, containing glauconite. Fossils include foraminifera, bivalves such as *Pecten* and oysters, accompanied by echinoids, Bryozoa, charophytes and Cichlid fish teeth (Bender 1974a, p. 90).

5.1.1.3 *Susita Formation*

Author: Golani (1962), "Susita Conglomerate"; amended by Michelson (1972, formally published in Michelson & Lipson-Benitah 1986).

The type section of the Susita Formation is near the archaeological site of Susita, in the same locality as the underlying Fiq Formation (Fig. 5.3).

The Formation overlies the Fiq conformably, commencing with a 3–10 m thick marly horizon particularly rich in *Pecten* shells, always appearing as a guide horizon at the base of this unit. The Susita Formation is unconformably overlain by the En Gev Sands, over a gentle erosional relief, occasionally with slight

Figure 5.3. Close up of the Susita Formation, east of En Gev, dipping westward into Lake Kinneret.

angular contacts. In places where the En Gev Sands are missing, either due to non-deposition or erosion, the Herod Formation overlies the Susita unconformably.

Susita Formation outcrops are known from the same localities as the underlying Fiq. The Formation was subdivided into two members, the lower HaOn and the upper Noqev (also "Nuqeb") Member, named after settlements in the area. A slight angular unconformity of 5–6° separates the two members. The thickness of the HaOn Member is some 100 m, while Noqev attains 45 m in the type sections. The thickness of the Susita decreases southward, near the settlement of HaOn at the southeastern end of Lake Kinneret.

The main rock types of the Susita Formation comprise fossiliferous yellow marls and marly limestones, greenish sandy ("sugary") dolomite containing glauconite and fossils, some lumachel horizons and detritic sandy or bioclastic limestones, some quartzolite, flint and quartzite concretions, some chalky white limestones and yellowish quartz sands. The HaOn Member is characterized by lumachel, dolomites and marls, forming small cliffs in the landscape. The overlying Noqev Member is typified by detritic sandy limestones, with thin layers of quartzolite and sands.

Typical fossils include both benthonic and planktonic foraminifers, bryozoa, corals, echinids, gastropods and bivalves, such as plentiful *Pecten,* oysters and others, and common algae. The fossils, especially the foraminifera (Lipson 1971, Michelson & Lipson-Benitah 1986), indicate an Oligocene age for both the Fiq and Susita formations. The environment of deposition of the Susita sediments is littoral-neritic, but mudcracks noticed in the upper part of Noqev Member indicate a coastal environment which had occasionally dried up.

At Wadi Taiyiba, the lowermost 49 m of the Usdom Group (Wetzel & Morton 1959) are composed of 38 m of white-yellowish sands and fine-grained, hard calcareous sandstones containing oysters and reworked(?) foraminifers. Occasional Eocene limestone pebbles occur at the base, which rests unconformably over the Taiyiba Beds. The upper 11 m comprise hard white and pink sandy limestone, with oysters and ostracodes, overlain unconformably by the Herod Formation.

5.1.1.4 *En Gev Sands*

Author: Golani (1962), "En Gev Sandstone"; amended by Michelson (1972, formally published in Michelson & Lipson-Benitah 1986).

The type section of the En Gev Sands is in the vicinity of the settlement of En Gev, on a hill south of Susita (Fig. 5.4). Golani (1962) included the En Gev Sands in the overlying Herod Formation, for lithological considerations. Michelson (1972), following a much more detailed study, had shown that the two are separate units, considerably differing from one another. The En Gev Sands overlie unconformably the Noqev Member of the Susita Formation. The unit is overlain unconformably by the Herod Formation or the Lower Basalt. Both unconformities are erosional, occasionally slightly angular.

The En Gev Sands crop out in limited areas east of Lake Kinneret, between En Gev and HaOn, with a single small occurrence some 3 km north of En Gev.

Figure 5.4. En Gev Sands, east of En Gev.

The thickness is 80 m at the type section, where the base is not exposed. Michelson estimates the maximum thickness as about 90 m.

The Sands principally comprise yellow quartz grains unconsolidated or slightly cemented by carbonate, with thin laminar chalky or marly limestone horizons, 0.5–1.5 m thick, which may occasionally contain some quartz grains. Most of the sands are fine grained, some 15–25% are medium, sometimes containing carbonate and glauconite particles.

The Sands are sterile of any fossils. Michelson (1972) tends to regard the sand as of fluvio-lacustrine origin. Granulometric analyses of the En Gev Sands led Giveon (1984) to suggest deposition in a continental alluvial fan, with occasional occurrences of shallow seasonal lakes and playas, under-arid or semi-arid climates. This may put the Sands in the lower part of Palynozone Ma, typified by a similar paleoclimate.

5.1.1.5 *Zefa Formation*

Author: Shahar (1973) "Zefa Member"; amended by Horowitz (2000a).

Shahar (1973) defined this unit as the lowermost member of the Hazeva Formation in the central northern Negev. Considering its unique nature, different lithology and paleogeographic significance, Horowitz suggested that the Zefa be regarded as an independent formation. The Formation is named after Nahal (wadi)

Zefa (Hebrew for viper), where the type section is located (Fig. 5.5), a tributary of Nahal Ef'e (Hebrew for another kind of viper), which leads through Nahal Hemar (Hebrew for asphalt) to the southern end of the Dead Sea, a distance of some 20 km.

The Zefa Formation is known only from the Ef'e syncline, where it covers an area of about 2 km^2, attaining up to 18 m in thickness. The Zefa overlies unconformably the Campanian–Maastrichtian Ghareb Formation, and is in turn unconformably overlain by the Shahaq Conglomerate, the lowermost member of the Miocene Hazeva Formation, as defined by Sneh (1967, 1981). Both the underlying and top unconformities are erosional.

The base of the Zefa Formation comprises coarse grit, gravel and breccia, which grade upward to sandy marls, oncolite rich limestones and back to clastics, similar in nature to those at the base. The base unit of coarse grained grit is up to 2.5 m thick, reddish-black due to an abundance of iron oxides. Most of the pebbles are made of flint, with some of limestone and chalk, of Maastrichtian or Eocene age. The pebbles, which are more abundant at the base, are well rounded, varying in size from a few up to 50 cm. This layer is overlain by gray-white fine-grained sand, up to 6 m thick, which occasionally contains flint, limestone or

Figure 5.5. Zefa Formation in Rotem Plain, northern Negev, underlying a ledge of the base conglomerate of the Hazeva Formation, termed the Shahaq Conglomerate or Shu'alim Member.

orthoquartzite pebbles, up to several centimeters across. Cross-bedding and limonite concretions are quite common.

This clastic unit is overlain by slightly sandy variegated marls, up to 7 m thick, rich in limonite, barren of any fossils. In the center of the area occupied by the Zefa Formation a light-gray limestone bed overlies or occasionally entirely replaces the marls. The limestone is slightly friable, containing unidentified plant remains and oncolites, several centimeters in size, oval in shape. The top unit of the Zefa Formation comprises several meters of breccia, with fragments up to 10 cm across, most of them made of flint, some of limestone, embedded in a sandy or rarely carbonate matrix.

Horowitz (1979, p. 69; 2000a), Hirsch & Shahar (1993) and Hirsch (1996) assign an Oligocene age to the Zefa. Calvo et al. (1998) deny the stratigraphic position as defined by Shahar, claiming that the Zefa and its overlying rocks were misidentified, thus suggesting that it constitutes a middle member within the Hazeva Formation, overlying unconformably its Gidron Member and older units. This view is not accepted here, for reasons discussed in detail in Chapter 7.

5.1.1.6 *Abu Treife Series*

Author: Bentor & Vroman (1951).

The sequence is named after Wadi Abu Treife, several kilometers north of the eastern tip of Makhtesh Ramon in the central Negev highlands, where the type section was described (Fig. 5.6). Only three small exposures were defined as belonging to this unit, all in the same vicinity, in Abu Treife, Mahmal and Marzeva, preserved in low structures which accompany the Sa'ad-Nafha fault. The Series always overlies middle Eocene rocks with erosional unconformity, possibly with a slight difference in dip; its upper surface is eroded, or covered unconformably by the base conglomerate of the Hazeva Formation. The maximum thickness is over 150 m.

The lower part of Abu Treife comprises a coarse base conglomerate, consisting of large sub-angular limestone and subordinate chert pebbles derived from the close vicinity. These are cemented by hard, red, sandy–argillaceous limestone. The thickness of the base conglomerate at the Mahmal outcrop approaches 80 m. It is overlain by variegated, partly calcareous sandstones with flint splinters, accompanied by subordinate amounts of green and red shales, gypsiferous marls, hard limestones and dolomites. No fossils are reported.

Bentor & Vroman attributed a late Eocene or Oligocene age to Abu Treife, based on its structural and stratigraphic position and a fluvio-lacustrine environment of deposition, although they noted its resemblance to the Hazeva, in the Abu Treife basin. Both Garfunkel & Horowitz (1966), who visited only the Abu Treife outcrop, and Zilberman (1992) thought that the Series is in fact yet another occurrence of the Hazeva Formation. The discrepancy was cleared up during a recent visit, with A. Sneh and Y. Avni (Geological Survey of Israel), to both the Mahmal

Figure 5.6. Limestones and sands of the Abu Treife Series in the Mahmal basin just north of Makhtesh Ramon.

and Abu Treife outcrops. It seems that the two basins exhibit different rock units. The first indeed contains a suite of beds different from the Hazeva Formation, typified by a dark-red poorly sorted base conglomerate, overlain by alternating yellow sandstone and carbonate horizons. The top is truncated, overlain by a well-bedded, coarse, medium-sorted conglomerate with only a reddish hue. Moving to the Abu Treife basin, this conglomerate appears at the base of the sequence, overlain by the typical rocks of the Hazeva Formation, including the characteristic top conglomerate (see below). Although the Abu Treife basin contains the "wrong" sequence, the name "Abu Treife" is retained here for the one underlying the Hazeva, in honor of Bentor and Vroman.

5.1.2 Early and middle Miocene

Middle Miocene sediments and volcanics are very abundant in the Jordan Rift Valley, extending over wide areas, also beyond its limits, but displaying considerably lesser thicknesses. These are chiefly of fluviatile or fluvio-lacustrine origin, for which the subsiding Rift region acted as a series of intermediate basins, filled up mainly with clastic sediments. The marine regression during the Oligocene–Miocene boundary caused renewed incision of channels leading to the sea, removing much of the Oligocene sediments. Sedimentation and filling up of these channels was initiated by the Miocene rise in sea level, which raised the erosion base level and lowered the energy of rivers. During the peak of the Miocene transgression, as was the situation in the Oligocene, the energy of transporting agents considerably diminished, so that lakes and lagoons were formed in most of the inland basins, constituting the bulk of known sediments.

It is quite difficult to assess whether the sea itself reached the area now occupied by the Jordan Rift Valley. The closest truly marine Miocene sediments, intercalating with the continental strata, are in the northern central Negev, near Yeroham, where oyster banks occur (Goldsmith et al. 1988). Westward, the marine sediments increase their share in the sequence. It is possible that the sea had reached further inland, but that none of its deposits remained due to erosion caused by the regression which followed the peak of transgression. Bender (1974a, p. 88) reports from Dahal, some 20 km south of the Dead Sea in the northern Arava, a sandy marl within the upper part of the Wadi Bustan Member (lower part of Dana Formation, which could be of Oligocene age) containing *Ammonia beccarii*. He concludes that the occurrence of this foraminifer indicates a near-shore marine depositional environment. However, these bugs have been found living today in brackish springwaters near the Dead Sea (Ehrenberg 1849, Almogi-Labin et al. 1991), so their paleogeographic significance is doubtful. Bender further reported littoral sediments south of Gav Ha'Arava watershed, but these had been attributed by Horowitz (1974) to a marine ingression from the south, originating in the proto-Red Sea during the middle Miocene.

The Miocene rocks are known from two basins within the Jordan Valley, one to the north, from Lake Kinneret south to Marma Feiyad and beyond, designated the Herod Formation or Usdom Series (at least in part), while the other covers the southern Dead Sea and goes as far as the boundary between the northern and southern Arava and is named the Hazeva Formation in Israel and Dana Formation (or "Conglomerate") in Transjordan (Ibrahim 1993, Sneh et al. 1998a). The Dana Formation is subdivided into the Wadi Bustan and Dhira Ibn Salih Members (formerly the Lower and Upper Syntectonical Conglomerates or the Edh Dhira Beds). The two domains are separated by a northeast trending structural high, crossing the Rift Valley some 60 km north of the Dead Sea (Bender 1974a, p. 95).

5.1.2.1 *Herod Formation*

Author: Bentor (1946), amended by Schulman & Rosenthal (1968) to include the lower part of the Beida Formation of Rofe & Raffety (1965); by Michelson (1972) to include the sequence east of Lake Kinneret; and by Shaliv (1991) to include a variety of occurrences in the central Jordan and Yizre'el valleys.

Some confusion arises as to the exact name of this formation. It was originally named "Herod", which is the English name for this King. In both Latin and Hebrew

"Hordos" or "Hordus" is used, and this name is also widely applied in the literature. The Herod Formation constitutes the lower part of the "Tiberias Series" (Picard 1943), named after Har Hordus (Mount Herod, also known as Mount Berenice), 1.5 km south of Tiberias.

No "official" type section was ever published for the Herod Formation. The author (Bentor 1946) defined the Formation as Miocene clastic sediments exposed in the vicinity of Tiberias and southward. Reference sections were published by Wetzel & Morton (1959), Schulman (1962), Schulman & Rosenthal (1968), Michelson (1972), Shaliv (1991) and others. A reference section is suggested by Shaliv west of Tiberias hot springs, since it is the thickest known sequence, interbedded with six basalt flows which enabled radiometric dating (Fig. 5.1.1).

The Herod Formation overlies unconformably a variety of older rock units, tilted and eroded prior to its deposition (Schulman 1962, Mimran 1984), but with no sign of faulting. West of Lake Kinneret and the Jordan River the Formation mostly overlies late Cretaceous rocks, except for some localities to the south, where the Eocene Zor'a Formation underlies. Eastward, the Herod rests on late Cretaceous, Eocene and Oligocene formations. The Herod Formation sediments interfinger with flows of the Lower Basalt. The number of these flows and their stratigraphic location within the sediments vary from place to place. The Herod

Figure 5.1.1. Herod Formation, interfingered with flows of the Lower Basalt, just south of Tiberias.

Formation is overlain west of the Jordan River by the Umm Sabune Conglomerate and the Bira and Gesher formations, always truncating the upper parts of the Herod.

The Formation is known from outcrops and boreholes in the central Jordan Valley, from the region south of Marma Feiyad up to approximately midway along Lake Kinneret, on both sides of the lake (Michelson 1972, Shaliv 1991). A sequence of conglomerates at the northwestern edge of Lake Kinneret, more than 100 m thick, was assigned by Saltzman (1964) to the Herod Formation. This sequence also comprises, among the usual late Cretaceous and Eocene components, Jurassic and early Cretaceous pebbles. The latter seem to indicate that the provenance of at least some of the gravel is to the north, in Mount Hermon. Shaliv (1991) regards this outcrop as belonging to the Umm Sabune Conglomerate. A suite of outcrops described in Saltzman (1964), Kafri & Heimann (1994) and Kafri (1997) from various localities in the Galilee is assigned by the authors to the Herod Formation, termed the “Hordos Conglomerate” (see below).

The thickness at the reference section (Fig. 5.1.1) described in Shaliv (1991) is more than 750 m. Since the top is always truncated, the original thickness must have been larger, but it is however impossible to specify numbers. Laterally, the Herod thickness grades to nil, while younger formations overlie rocks older than the Herod.

The base of the Herod sequence, at the reference section south of Tiberias (Shaliv 1991) comprises some 10 m thick base conglomerate (“Lower Conglomerate Member” of Shaliv), built up of local components, usually chalk pebbles, overlying with erosional and slight angular unconformity chalks of the Senonian Mount Scopus Group. The rest of the sequence (the “Silty-Limy” or “Lacustrine” Member of Shaliv) is made of alternations of reddish calcareous sands and silts, with lenses and layers of coarse and fine, very well-cemented conglomerates, and horizons of hard limestones. The fine clastics and limestones are well bedded, usually platy. The pebbles are mostly limestone and chalk, occasionally flint, very poorly sorted, 5–50 cm across.

The sequence is interfingered by six basalt flows, the lowest of which overlies an irregular erosional relief of the underlying sediments. The pebbles above the lower basalt flows are from carbonate rocks of the Judea Group and weathered basalts, poorly sorted, medium rounded, 2–40 cm across. It seems that the roundness and sorting indicate a nearby source for the pebbles, but somewhat more remote than the provenance of the base conglomerate components. Above the third and fourth basalt flows, the lower parts of the conglomerates are made solely of basalt pebbles. Resistant benches, up to 5 m high, occur between the basalt flows, made of very well-cemented coarse conglomerates, with interlayers of carbonate silts bearing small pebbles, red clays and yellowish limestones. The pebbles building the upper benches are mainly derived from Eocene limestones.

South of the reference section the basalt intercalations become considerably thicker, until at the Yavne’el escarpment the Lower Basalt completely replaces the

Herod Formation sediments. In several localities, both outcrops and boreholes, similar sediments of various thicknesses interfinger with dated Lower Basalt flows. All of these are attributed by Shaliv to the Herod Formation.

South of Bet She'an Valley, in the Marma Feiyad–Wadi el Malih region, Rofe & Raffety (1965) described the Beida Formation. Its lower part is correlated with the Herod, which overlies early Cretaceous through Eocene rocks (Schulman & Rosenthal 1968) unconformably. The sequence was subdivided (Shaliv et al. 1988) in the same manner as near Tiberias, with the addition of a capping "Upper Conglomerate" overlying the lower part with angular unconformity. Reconsidering, Shaliv (1991) attributed this unit to the overlying Umm Sabune Conglomerate. The Lower Conglomerate Member, some 20 m in thickness, comprises local components, and is exposed at only a few sites west of Marma Feiyad, overlying unconformably the Judea Group rocks. The second, the "Silty Carbonate Conglomerate Member", is exposed in the entire region, built of variegated thin beds of authigenic dolomite, alternating with marls, sandstones and carbonaceous silts. These are overlain by reddish siltstones and white chalks.

The sequence, some 30 m thick, is truncated by the Umm Sabune Conglomerate. To the east the section becomes thicker, up to 150 m, and includes anhydrite lenses up to 1 m thick. The lower part of the thicker sequence also includes chalk and limestone layers, followed by a variety of siltstones and carbonaceous sandstones alternating with marls and micro-conglomerates (Schulman & Rosenthal 1968). Part of the Herod Formation was penetrated by the Zemah 1 borehole, drilled south of Lake Kinneret. Details of this occurrence are given in Chapter 6.

The outcrops east of Lake Kinneret, described in detail in Michelson (1972), attain a maximum thickness of 242 m. The Formation rests on a basal basalt flow, 10–30 m thick, overlying unconformably the Susita Formation, En Gev Sands and Eocene rocks. In other localities in the vicinity 3–4 basalt flows were observed within the Herod Formation (Michelson 1982). Characteristic lithology comprises frequent alternations of conglomerates, calcareous sandstones, siltstones, marly chalks and marls. It is overlain either by the Gesher Formation or by flows of the Intermediate and Cover basalts.

The outcrops in Wadi Taiyiba (Wetzel & Morton 1959) comprise 103 m, with no volcanics. The basal 14 m overlie unconformably the Oligocene sediments, comprising pink marls with fine gravel. These are topped by 89 m of brick red to pink conglomerates, marly in part, with flint and Eocene limestone pebbles, containing some glauconite and reworked microfossils. A bed of lacustrine sandy limestone occurs in the middle of this sequence. The top is exposed. The sequence is considered part of the Usdom Group (Daniel 1963). Other outcrops occur along the eastern flank of the Jordan Valley, down to the area opposite Marma Feiyad and somewhat further south (Bartov 1994).

No fossils were ever found in outcrops of the Herod Formation, except for an oyster bed almost at its top (Sneh 1993) indicating estuarine conditions. A lacustrine or fluvio-lacustrine environment of deposition is deduced for the sands of the

Herod Formation east of Lake Kinneret, while the silts and limestone horizons of the Herod also seem to indicate a lacustrine environment, a conclusion shared by all investigators of this Formation (Avnimelech 1937, Cailleux 1949, Shaliv 1991). Despite the lack of fossils, almost all authors agree that the Herod Formation is of middle Miocene age, which was strengthened by its radiometric ages, obtained on the interfingering basalt flows (Shaliv 1991).

5.1.2.2 *Hazeva Formation*

Author: Blake (1935), "Kornub and Hosb Sandstone"; amended by Bentor & Vroman (1954b) "Hatseva Formation"; by Harash (1967); by Sneh (1981); and by Calvo et al. (1997).

The type section of the Hazeva Formation is near a settlement of that name in the northern Arava, where it is very well developed (Fig. 5.1.2). Many reference sections are given by various authors, at different localities in both Israel and Transjordan, using synonyms such as "Hasb" or "Hosb" Sandstone (Arabic for Hazeva) or Series, "Kurnub Sands and Sandstones" and others (Bentor 1960, Daniel 1963), or Wadi Bustan Member (lower part of the Dana Formation) in Transjordan (Ibrahim 1993). The "Lower Syntectonical Conglomerate" in south-western Transjordan, and possibly also part of the Dhira Ibn Salih Member (upper

Figure 5.1.2. Mashaq and Gidron members of the Hazeva Formation in Nahal Paran, affected by the Paran fault, topped (upper right hand corner) by the flat-lying Arava Formation.

part of the Dana Formation) or "Upper Syntectonical Conglomerate" (Bender 1974a, p. 92), seem to correlate with the Hazeva Formation as well.

The Hazeva Formation sediments overlie, with angular and erosional unconformities but no faulting, a variety of pre-Miocene rocks. The youngest of these are the Oligocene Zefa Formation in Nahal Ef'e and the Abu Treife Series in Mahmal basin, if the latter age assignment is accepted; if not, the youngest rocks overlain by the Hazeva are of Eocene age. It is overlain by a suite of Pliocene and younger formations, the oldest of which is the Arava Formation (Bentor & Vroman 1957, Sneh 1981), and the Dhira Ibn Salih Member ("Upper Syntectonical Conglomerate") or maybe only its upper parts, on the Transjordanian side of the Arava (Bender 1974a, p. 92; Ibrahim 1993). Slight unconformities are also quite common within the Hazeva, causing appearances of different parts of the Formation in certain localities, due to variable channel fillings resulting from pulses of subsidence (Sneh 1999).

The truncation of the top of the Hazeva is quite substantial in places such as the Arava, where the upper unit is entirely missing from outcrops. The uppermost Ashalon Member of the Hazeva (Harash 1967) is known only from regions far away from the Rift, such as Dimona–Yeroham to the west, which may possibly correlate with part of the Dhira Ibn Salih Member (upper part of the Dana Formation) in Transjordan. A small outcrop on the western Dead Sea shore, near En Boqeq, may represent this Member. However, a very thick sequence topping the Hazeva in some of the boreholes (see Chapter 6) and downfaulted structures is considered by some (Calvo et al. 1998) as the uppermost part of this Formation, while it is referred to here as the separate Hufeira Formation (Section 5.2.1.3), having been deposited in an entirely different morphotectonic domain and drainage system (see Section 7.2.4).

The Hazeva Formation is very abundant in the northern Arava, attaining some 120 m in thickness, extending westward to the central and northern Negev (Bentor & Vroman 1954b, 1957, Harash 1967, Eidelman 1979, Sneh 1981, Eyal 1984, Yechieli 1987 and others). There it is developed mainly in synclines, where thicker sequences had been accumulated, better protected from subsequent erosion. A maximum thickness of some 145 m was penetrated at Mishor Yamin (Weiner et al. 1999), a syncline separating the Hatira and Hazera anticlines. Occasional outcrops are also known from structurally high localities (Bartov 1974, Zilberman 1992). Farther west, the Hazeva interfingers with littoral Miocene sediments of the Ziqlag Formation, grading further westward into the marine Ziqim Formation (Horowitz 1979, p. 70). Southward the sequences are very limited (Ginat 1997, Avni 1998), also extending into the eastern Sinai (Garfunkel & Horowitz 1966). The northernmost outcrops are in the vicinity of the town of Arad, west of the Dead Sea, where oyster and echinid fragments occur within the sequence (Agnon 1983); to the east they are reported from Edh Dhira, just east of the Lisan Peninsula, but further east of the northern Arava outcrops of the Hazeva Formation are quite rare, except for subordinate occurrences (Bender 1974a, p. 92).

Sneh (1981) subdivided the Hazeva Formation in the northern Arava into three members, the lowermost Shahaq Conglomerate, overlain by the Mashaq and Gidron members, for which he presents detailed columnar sections. Harash (1967) subdivided the Hazeva in the Yeroham–Dimona region, 35–40 km west of the southern tip of the Dead Sea, into five members, Shu'alim, Mingar, Yeroham, Aro'er and Ashalon. The lowermost one seems identical to Sneh's Shahaq Conglomerate; the overlying Mingar is correlated with the Mashaq and part of the Gidron; no correlation with the Arava is suggested for the oyster-bearing, littoral Yeroham Member, while the Aro'er Member is correlated by Harash with the upper part of the Gidron. The uppermost Ashalon Member does not occur in the Arava (Sneh 1981). Shahar (1973) defined the Rotem Member in the Ef'e syncline east of Dimona, in which he included the Mingar, Yeroham and Aro'er Members. It is overlain here also by the typical Ashalon conglomerates.

Calvo et al. (1997) subdivided the Hazeva in the Karkom Graben, some 20–40 km west of the central Arava, into a lower part similar to Sneh's three members, and an upper part comprising up to 2,000 m of clastics assigned to the Rotem and Karkom Members, equal to the Hufeira Member in the northern Arava. This subdivision is not acceptable here, being based on Calvo's misidentification of the Zefa and Rotem Members. The Hufeira is considered in the present context an independent formation, having been deposited in a system entirely different than the Hazeva (see Sections 7.1.3.2 and 7.1.4).

The Shahaq Conglomerate is usually several meters thick, attaining some 22 m at its best, in Nahal Shahaq, some 40 km south of the Dead Sea at the western rim of the northern Arava. It is usually exposed in rather restricted outcrops, in the form of a fan-shaped geometry. Shahaq is, however, missing in many locations, or covered where the overlying Mashaq Member extensively overlaps. The pebbles, sub-angular to sub-rounded, 1–40 cm across, badly sorted, mostly comprise limestone and flint derived from the Eocene Avedat, or rarely from the Senonian Mount Scopus Group. In most places the conglomerate is well cemented by microcrystalline calcite, forming small cliffs, while sometimes it can be rather loose. The size and composition of the pebbles vary with distance from the source rocks, which seem to be in the immediate vicinity. Long-distance transport is subordinate, indicating a local origin for the conglomerates. The Conglomerate is usually massive, but crude bedding is occasionally observed. In places the upper part of the Member includes some algal limestones and marls.

The Mashaq Member overlies conformably the Shahaq Conglomerate, but frequently overlaps, resting on older rocks. The contact is usually abrupt, occasionally gradual. It is overlain by the Gidron Member but, when the latter is truncated, by younger units such as the Arava Formation and others. The maximum thickness at the type locality in Nahal Mashaq, close to Nahal Shahaq, where the base is not exposed, is 25.5 m (Yechieli 1987), but the unit varies in thickness over rather short distances.

This Member was subdivided into two units by Bentor & Vroman (1957), which were treated as a single entity by Sneh (1981), because this separation is apparent only in a restricted number of outcrops. The lower part of the Mashaq comprises limestones, gray to light-greenish sandy marls, some clays and gravels, forming prominent cliffs. The upper part consists mainly of gray-pink, hard limestone, with sandstone lenses, which in places appear as small cliffs. Facies changes over very short distances are quite common. Limestones predominate to the west while marls take the upper hand eastward and southward, indicating deepening of the basin (Sneh 1981).

The Gidron Member covers the widest areas in the northern Arava, overlying the Mashaq conformably, occasionally overlapping earlier rocks. The transition from limestones to sandstones and red clays marks the boundary. Its typical red colors make it very prominent in the landscape. It is overlain, unconformably, by the Arava Formation or younger rocks. The thickness in the reference section is 35.5 m, but composite sections are up to 80 m. The unit is considerably thicker in boreholes (Sneh 1981), attaining more than 200 m.

The Gidron Member comprises alternations of red clayey, calcareous mudstones, siltstones and fine to very fine quartz sandstones, unimodal, well-sorted, with subordinate mica, feldspar and occasionally redeposited glauconite grains. The sandstones display horizontal bedding or planar and trough cross beds. The red clays are washed over the outcrops, giving the entire Hazeva Formation a typical reddish hue. Lateral changes in lithology are quite common over short distances, which makes exact correlations difficult.

The Ashalon Member is known only from regions outside the Rift, such as Dimona–Yeroham and Arad to the west. It overlies conformably the Aro'er Member, which is correlative to the upper part of the Gidron in the northern Arava. The top is always truncated by erosion, occasionally covered by a thin veneer of Quaternary sediments. A small outcrop on the western Dead Sea shore, near En Boqeq, may represent this Member (Garfunkel 1997), but alternatively could belong to the younger Hufeira.

The Ashalon Member consists of several meters of sandy conglomerates, most of the pebbles made from well-rounded and sorted flint. This flint is traditionally termed "imported", since no known source could be located in the vicinity. The flint is very rich in foraminifers, which were at first thought to indicate a late Cretaceous age (Reiss, in Garfunkel & Horowitz 1966). The rock is considerably different from late Cretaceous flints of the northern Negev, thus "imported". Later, detailed analyses of the microfauna yielded an early middle Eocene age (Benjamini 1984), and similar flints were found within sequences of this age, the Paran Formation in the central Negev (Horowitz 1974, Benjamini 1984). The roundness and sorting seem to indicate, anyway, a long transport for these pebbles.

A travertine layer occasionally interfingers (Enmar et al. 1998a, Enmar 1999) with the Hazeva Formation, attaining in places up to 15 m in thickness. The unit is mainly known from the northern Arava, where it occasionally occurs as large patches of travertine, usually sandwiched between the base conglomerate (Shahaq Conglomerate) and the overlying sands of the Mashaq. The principal rock type is stromatolitic, coarse crystalline, thickly enveloping organic structures and pebbles with layered crusts. Cylindrical oncoids, formed around roots and stems, are very common. Plant remains are numerous, including a variety of broad leaves and palm hands (Enmar & Heimann 1999). These, together with the wide distribution of the travertines, may indicate a humid climate, corresponding to Palynozone Mb.

No fossils have ever been reported from the Shahaq Conglomerate. Mashaq limestones contain freshwater mollusks, ostracodes, stromatolites, oncolites and algal debris, indicating a lacustrine environment in the Arava region; the Yeroham Member contains oyster beds, indicating a brackish estuary environment. However, none of these fossils are indicative of age. Clays of the Gidron Member occasionally contain plant remains. Vertebrate fossils are quite common in the Mingar Member in the Yeroham–Dimona basin and the Rotem basin; the latter may be somewhat younger since the bearing strata contain fragments of reworked oysters (Tchernov et al. 1987, Goldsmith et al. 1988). Bones are also found in the Arava, in the Gidron Member (Savage & Tchernov 1968). The same outcrops also yielded numerous plant remains, usually silicified wood. Many unidentified silicified wood fragments are found within the otherwise barren sequence of the Ashalon, which could not be used for dating, but their abundance indicates a humid climate, typical for Mb times. Its conformable contact over the upper member of the Hazeva could well justify assigning a late-middle Miocene age.

Tchernov et al. suggest an early Miocene, possibly Burdigalian age for the vertebrates, based on correlations with somewhat similar assemblages, mainly from northern Africa. The taxa indicate that the environment of the bone bearing beds is mainly riverine, partly lacustrine. The vegetation must have been rich, with forested and marshy habitats, but also with open, fairly dry country not too far away. The Ashalon Member was deposited in a river valley, which drew its water from quite far away, possibly Transjordan (Garfunkel & Horowitz 1966) or the central Negev.

5.1.2.3 *Dana Formation*

Author: Wetzel & Morton (1959), "Edh Dhira Beds"; amended by Bender (1963), "Lower" and "Upper Syntectonical Conglomerates"; and by Parker (1970).

The Formation was named primarily after Ghor Edh Dhira, east of the Lisan Peninsula (Figs 5.1.3 and 5.1.4), where the type section was described (as a reference section of the Usdom Series). The sediments overlie unconformably the Eocene Sar'a Chalk and Flint (Zor'a) Formation and are overlain, over an angular and erosion surface, by the late Pleistocene Lisan Formation. The exact extent of the Edh Dhira Beds is not known, since they were correlated with a variety of sequences, at least some of which are certainly of different ages, like the Sedom Formation. The thickness is 191 m at Ghor Edh Dhira, but varies, attaining in places almost 300 m. The Beds were termed the "Dana Formation" by Parker,

Figure 5.1.3. Dhira Ibn Salih Member, upper part of the Dana Formation near Edh Dhira, dipping east toward the Dead Sea. The faulted part is covered by the "Upper Syntectonical Conglomerate".

after Wadi Dana, between Dana and Shubak, occasionally also the "Dana Conglomerate Formation" (Ibrahim 1991, 1993) or the "Dana Conglomerate" (Sneh et al. 1998a).

The basal 9 m chiefly comprise conglomerates of flint and limestone pebbles, in a sandy matrix. These alternate with sandy limestones rich in glauconite, with subordinate pebbles. Above lie 104 m of pink, yellow and gray, finely sandy marly clays, with lenses and massive beds of sandstones and gravel, the latter comprising flint and limestones pebbles. These are overlain by a 12 m thick massive conglomerate bed. The top is made of 66 m of gray, green or pink sandy marls, alternating with lenses of sandstone and flint gravel.

Very few rare and usually inconclusive fossils are reported, and the suggested age is "?Oligocene–Miocene–Pliocene", most probably based on too liberal correlations (Wetzel & Morton 1959, p. 169; Bender 1974a, p. 89).

5.1.2.4 *Wadi Bustan Member (lower part of the Dana Formation)*

Author: Bender (1963, p. 51) "Lower Syntectonical Conglomerate"; amended by Ibrahim (1993).

Figure 5.1.4. Wadi Bustan Member, lower part of the Dana Formation near Edh Dhira, also known as the "Lower Syntectonical Conglomerate".

The Wadi Bustan Member (Fig. 5.1.4) has intimate geographical connections with the Jordan Rift Valley, which led Bender to apply a genetic name for the sequence, probably inappropriate (see Chapter 7) since it does not overlie faulted relief. The type section is near Gharandal, east of the central Arava. The Wadi Bustan Member overlies unconformably a variety of formations of Eocene and older ages, and is overlain by the Dhira Ibn Salih, the upper member of the Dana Formation. The contact between the two units is set by Bender and Ibrahim where a distinct scarp of thick-bedded conglomerates with no marl intercalations is developed.

The Wadi Bustan Member is common along the eastern escarpment of the northern Arava, known mostly from subsequently downfaulted blocks, where the sediments were partially sheltered from erosion. The Member does not occur much further east of the Rift. The thickness at the type section is 90 m, comprising alternating marls, sandy marls, layers and elongated lenses of unsorted conglomerates. The latter contain sub-rounded blocks up to 2 m across, consisting exclusively of transported and redeposited rocks of Cretaceous and Tertiary age. In addition, up to half a meter thick chert beds and fine-crystalline limestones are developed in the section. Another sequence, approximately 300 m thick, is described by Bender (1974a, p. 89) from Shubak, some 20 km north of Petra, on the highlands. It comprises alternating sandy and gypsiferous marls and shales, thin-bedded limestones, fine calcareous sandstones, sandstones and conglomeratic sandstones.

The lowermost beds in Gharandal contain freshwater gastropods and charophytes. The upper beds at Dahal, 25 km south of the Dead Sea, in the Arava, contain some foraminifera. The sequence at Shubak yielded charophytes and Cichlid (freshwater fish) remains. Bender therefore concluded that the Wadi Bustan Member in the Arava had some connections with marine environments, while the outcrops on the highlands should be regarded as of fluvio-lacustrine origin. Bender (1974a, p. 89; 1974b) assigned a "?middle to upper Oligocene, ?Miocene" age to the Lower Syntectonical Conglomerates.

Khalil (1992) reports *Globigerina officinalis subbotina*, *G. senilis* and *G. tripartita* from the Edh Dhira area, all thought to be of Oligocene age. Macumber & Edwards (1997) indicate, from the same region, that a doleritic intrusion penetrates through the Dana Formation, most probably its lower part, in Wadi Karak. Other, seemingly similar intrusions upstream the same wadi (Fig. 5.3.4), yielded ages of 18.3 ± 0.9 and 18.9 ± 0.8 Ma (Barberi et al. 1980). If the fossils are indeed indicative and the intrusions of a similar age, the lower part of the sequence may well be of Oligocene age.

5.1.2.5 *Ghor el Qatar Series*

Author: Ionides & Blake (1940), "Series of Grain Sabt"; amended by Bender (1974a, p. 93).

The unit is named after a region of the Ghor (upper surface of the Jordan Valley) 25–30 km north of the northern end of the Dead Sea, occupied by the diapiric(?) uprise of Grain Sabt, also known as the Zahrat el Qurein dome, where the type section is located (Fig. 5.1.5). The sequence, according to Bender, overlies the Shagur Formation with an angular, erosional unconformity, but this contact is highly questionable (see below). It is overlain, also disconformably, by the Lisan beds and by basalt flows, possibly also intruded by a basalt dike. It is questionably known from several localities along some 50 km north of the Dead Sea, on the eastern side of the southern Jordan Valley. The maximum thickness at Ghor el Qatar is about 350 m (but this may seem so due to possible repetitions in the sequence), diminishing in other localities, possibly due to subsequent erosion and faulting, or to deposition in shallower basins.

The sequence comprises alternating conglomerates, conglomeratic sandstones, sandstones, marls and marly clays. The argillaceous–marly–fine sandy matrix of the coarser clastics is often reddish–brownish, reminiscent of terra rossa colors. Huckriede (in Bender 1974a, p.94) identified *Melanopsis praemorsa*, ostracodes and poorly preserved vertebrate and plant remains in the Ghor el Qatar Series, which indicate a fluvio-lacustrine environment but not an exact age. He assigned

Figure 5.1.5. Ghor el Qatar Series at Zahrat el Qurein, north of the Dead Sea.

an early Pleistocene age, based on the stratigraphic position of the beds as described by Bender.

Wetzel & Morton (1959, p. 170) suggested correlation of the Ghor el Qatar Series with the Edh Dhira and Taiyiba Miocene sequences, which was not accepted by Bender, who regarded the Series as considerably younger stratigraphically. Recent observation by Y. Bartov, Jr. (Department of Geology, Hebrew University of Jerusalem 1999, pers. comm.) indicate an abundance of "imported" flint pebbles in the sequence, typical for the top members of the Hazeva Formation (see Section 5.2.1.2), which clearly supports Wetzel & Morton's opinion. In contrast, the dark-red color of these beds, and particularly the occurrence of basalt pebbles which never appear in the Hazeva Conglomerates (pers. obs.), shed doubt on the above assignment. It is now apparent that the minimum age for the top of the Ghor el Qatar Series deposition is late Miocene, but it seems that its main bulk is of middle Miocene age, or possibly much younger if Bender's view is accepted, as by Horowitz (1974 and further publications).

Several outcrops assigned by various authors to the Ghor el Qatar, such as those near Mashari'a, some 30 km south of Lake Kinneret (Macumber & Edwards 1997), most probably belong to the Herod Formation (pers. obs.). Others, in the southern Jordan Valley, show the distinct lithological characteristics of the Hazeva or Dana formations (Y. Bartov, geological Survey of Israel 2000, pers. comm.).

5.2 ERITREAN-STAGE FORMATIONS

The Eritrean faulting was mainly active throughout the later part of the Miocene and the earliest Pliocene. Its outcome affected the Jordan Rift Valley and surrounding regions until the end of Palynozone QI, earliest Quaternary, by controlling subsidence, hydrography and environments.

5.2.1 Late Miocene

The transition from middle to late Miocene is marked by considerable faulting and trough formation (Horowitz & Horowitz 1990, Calvo et al. 1997). As a result, late Miocene sediments show considerable thicknesses, in the order of hundreds or thousands of meters, in boreholes drilled in the Hula Valley, south of Lake Kinneret and in the southern Dead Sea (Horowitz & Horowitz 1990, Horowitz 1996a), and at Jebel Hufeira, where the sequence is exposed. Outside the Rift such sequences are only questionably known from the Karkom graben in the central Negev (Calvo et al. 1997). In other, then low-lying terrains, sections are limited to 200 m and less. However, most regions surrounding the Jordan Valley were subject to erosion, which removed large quantities of older formations.

5.2.1.1 *Umm Sabune Conglomerate*

Author: Schulman (1962); amended by Shaliv (1991).

The Formation is named after Wadi Umm Sabune (Nahal Tzabun), near Belvoir, a subsidiary of Wadi Bira (Nahal Tavor) just west of the central Jordan Valley, where the type section is located, comprising a sequence of conglomerates composed largely of basalt pebbles (Fig. 5.2.1). Shaliv (1991) expanded Schulman's definition to include various conglomerate beds of local origin, not necessarily dominated by basalt pebbles but nevertheless overlying a faulted relief, thus enlarging considerably the geographic extent of this unit.

At the type locality, the Umm Sabune Conglomerate unconformably overlies the Lower Basalt over an erosional, faulted relief. In other localities this unit unconformably overlies a variety of older rocks. The upper part of the sequence grades into the overlying Bira Formation, at the type section, para- or disconformably (Schulman 1962). Shaliv (1991) indicates a gradual transition, both vertically and horizontally, into the lower beds of the Bira. In other localities Shaliv shows in his sections (p. 40) unconformable contacts.

According to Shaliv, Schulman, and Schulman & Rosenthal (1968), the Umm Sabune Conglomerate covers wide areas, beyond the extent of the Herod Formation, mainly in the central Jordan Valley but extending northward to both sides of Lake Kinneret and possibly to southern Lebanon, southward down to Marma Feiyad and further south, and westward to the Yizre'el Valley. In the central, deeper parts

Figure 5.2.1. Umm Sabune Conglomerate near Menahemya, central Jordan Valley. The middle, platy part of the outcrop is a Late Miocene Basalt flow.

of the Yizre'el Valley, known mainly from boreholes and only a few outcrops, the conglomerates seem to be replaced by brown to yellowish clays ("Clay Series" of Picard 1943) resting on what appears like the Lower Basalt, overlain by Bira Formation marls. Similar beds are described by Braun (1992) near Tiberias, which he termed the "Yachza'el Formation", agreeing to its correlation with the Umm Sabune. Since it does not seem necessary to add yet another term, Braun's is not used here. Sneh (1996) suggests that the lower, mainly limestone and conglomerate member of the Bira correlates with the Umm Sabune (Fig. 5.3.7).

It appears that safe assignment of various gravel units to the Umm Sabune cannot be justified only by lithology. As long as the stratigraphic position is quite clear, such as the "Clay Series", the correlation can be accepted. In other localities, where conglomerates are unconformably overlain by the Bira or younger Formations, correlation may be doubtful, such as in the southern Golan and northwestern corner of Lake Kinneret outcrops, where the conglomerates seem to be intimately connected with the underlying Herod Formation (Saltzman 1964, Michelson 1972). Thus the Umm Sabune can be safely defined only when overlying Miocene sediments with an angular unconformity, and overlain by Pliocene rocks. Such is exactly the case near Marma Feiyad, where the Herod Formation (previously termed the "Beida Formation" in Rofe & Raffety 1965) dips 60°SE, while its uppermost 40 m, assigned by Shaliv (1991) to the Umm Sabune, dip 15°SW and are covered by the Pliocene Gesher Formation (Schulman & Rosenthal 1968).

The thickness of the Umm Sabune Conglomerate is some 200 m at the type area, in the central Jordan Valley (Schulman 1962), where it fills a faulted, deeply incised relief, diminishing over short distances to complete absence. Other outcrops and well sequences described in Shaliv (1991) are usually several meters to several tens of meters thick. The thickest seems to be the "Clay Series" sequence, attaining up to 260 m in one of the boreholes in the Yizre'el Valley.

The lowermost 50 m of the type section (Schulman 1962) comprise very coarse conglomerates, with unsorted basalt pebbles up to 60 cm across, loosely cemented. Above lie some 130 m of coarse to fine rudites, arenites and siltites, with fine quartz grains. Upward, clear bedding is gradually developed. This part is better cemented than the lower, with cement made of weathering products of basalt and some calcite. The uppermost 20 m are very finely bedded or occasionally laminated, made of fine limestone and basalt clastics, with some unidentified plant remains.

No fossils except for the plants are reported from the Umm Sabune, so a direct age assignment is impossible. Most authors preferred an early Pliocene age, based on stratigraphic grounds (Schulman 1962; Horowitz 1979, p. 76; and many others). Shaliv (1991), based on radiometric ages obtained on basalts below, within and above the conglomerates, suggests a late Miocene age, which is accepted here, supported also by morpho-structural considerations (see discussion in Chapter 7).

The Umm Sabune Conglomerate fills a continental relief system, usually river channels, with occasional small ponds or lakes, most likely formed following the regressing late Miocene sea, flooded partly by the Tortonian sea, which deposited the lowermost Bira. The erratic, rather rare outcrops and the problematic correlations make precise paleogeographic reconstructions somewhat questionable. If however the correlation of the Umm Sabune and Hordos Conglomerates is accepted, as proposed by Shaliv (1991), then the system must have drained to the Mediterranean (Kafri 1997).

5.2.1.2 *Hordos Conglomerate*

Author: Kafri & Heimann (1994).

The Hordos Conglomerate occurs in the Galilee, is named after the Herod Formation and is believed by Kafri (1997) to constitute its uppermost part. A type section is not given. The conglomerates had earlier been referred to by many investigators, usually as "Neogene Conglomerates". Shaliv (1991) suggested that they should be considered correlative to the Umm Sabune Conglomerate.

The Hordos Conglomerate overlies rocks of various ages, over an erosional and taphrogenic relief, largely peneplained. The top is usually not covered, eroded, but in places in the Galilee it is overlain by the Intermediate or Cover Basalt (Fig. 5.2.2), or by younger gravels and soils. It grades westward into the

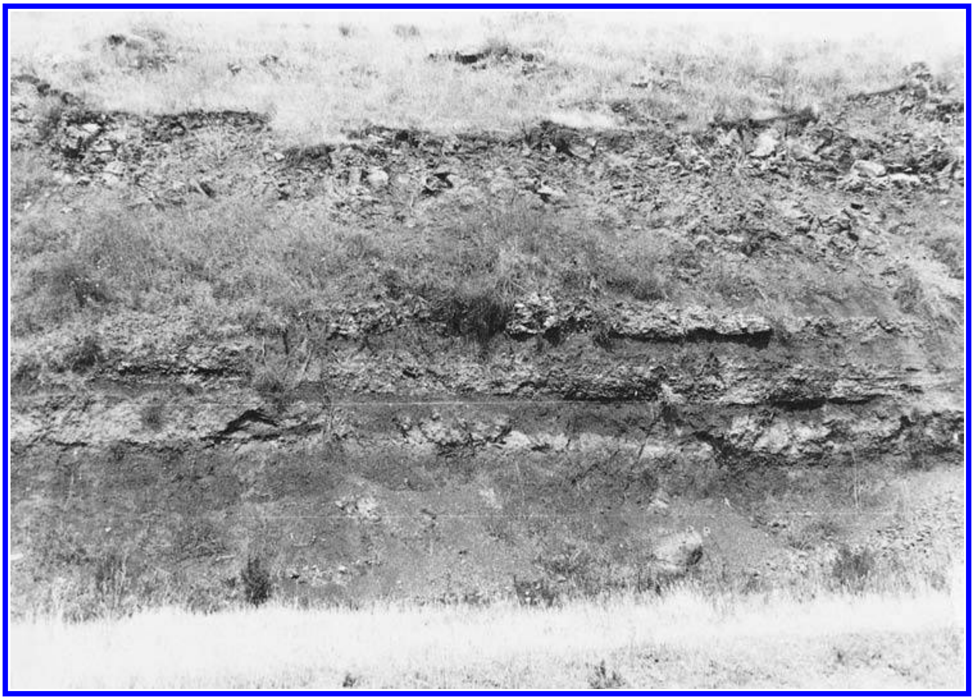


Figure 5.2.2. Hordos (Herod) Conglomerate covered by the (?)Cover Basalt at Livnim, some 5 km west of Lake Kinneret. Photo courtesy of U. Kafri.

late(?) Miocene Bet Nir Conglomerate. The Hordos Conglomerate is known from patchy occurrences in the entire eastern Galilee, more abundant in its southern sector, filling a drainage system leading to the Mediterranean (Kafri 1997). The usual thicknesses close to and inside the Jordan Valley are several tens of meters, up to a maximum of 140–150.

The Conglomerate, generally reddish, comprises alternations of gravels and brown-reddish siltstones and mudstones. The pebbles are polymictic, of different sizes and shapes, derived from the underlying country rocks, late Cretaceous or Eocene limestones, dolomites, chalk and quartzolite. No basalt components are present, most probably since this part of the country is quite far from the Lower Basalt occurrences. The matrix consists of either siltstones or travertine-like brown carbonate cement. No fossils have ever been found within the unit. The coarse gravels were deposited in a high energy stream bed, while the fine mudstones represent low energy slack waters of floodplains. Kafri (1997) reconstructed the drainage direction as leading westward, to the Mediterranean.

The lower age limit of the Hordos Conglomerate can only be assumed indirectly, by the fact that it overlies a faulted peneplain. Such is known for the end of the middle Miocene from the entire southern Levant (Garfunkel & Horowitz 1966, Horowitz 1974). The upper age limit is somewhat better defined by radiometric ages of the overlying basalts, ranging around four million years (Kafri 1997). The suggested correlations with the Bet Nir and Umm Sabune Conglomerates place the sequence in the late Miocene, but its typical red color indicates a humid, warm climate, which is more in accord with Mb times, possibly the later part.

5.2.1.3 *Hufeira Formation*

Author: Calvo et al. (1997), amended here.

The Formation is named after Jebel Hufeira, east of En Yahav in the central northern Arava, some 50 km south of the Dead Sea, where the type section is located (Fig. 5.2.3). Calvo et al. referred to the Hufeira as an upper member of the Hazeva Formation, which it overlies over an angular unconformity. The Hufeira is here regarded as an independent formation, since it was deposited in a completely different tectono-sedimentary system than the Hazeva (see Section 7.2.4, Fig. 7.4), also making its distribution entirely unlike the latter.

According to Calvo et al. (but see reservations in Chapter 7), the Hufeira grades laterally into the Rotem Member as defined in Shahar (1973), and unconformably overlies all previous units of the Hazeva Formation or older rocks. It is conformably overlain by the Karkom Member which, in turn, is overlain by various

Figure 5.2.3. Hufeira Formation at Jebel Hufeira, northern Arava. Photo courtesy of R. Calvo.

rock units, the oldest of which is the Arava Formation. These units are known from the northern Arava and the central Negev. The thickness of the Hufeira is in the order of 2 km; the Karkom, which could be regarded as a member of the Hufeira, may exceed in some places 100 m.

The Formation consists of alternations of coarse sandstones and yellowish shales. Some of the sand layers are rich in well-rounded flint, quartz, some limestone and "imported" flint pebbles (see Chapter 5.1.2.2). The Karkom Member is made of reddish sandstones and clays, rich in somewhat angular Eocene limestone pebbles, with subordinate flint gravels. Occasionally the conglomerates are very rich in "imported" flint pebbles, appearing in a similar manner to those from the Ashalon Member. No fossils are reported, but the stratigraphic and structural position led Calvo et al. (1997, 1998) to assign a late Miocene age.

5.2.1.4 *Dhira Ibn Salih Member (upper part of the Dana Formation)*

Author: Bender (1968b) "Upper Syntectonical Conglomerate"; amended by Ibrahim (1993).

This unit, same as the Wadi Bustan Member, has intimate geographical connections with the Rift Valley which led Bender to apply a generic name for the sequence, probably inappropriate (see Chapter 7). The type section is located 2 km ENE of Gharandal, in the central Arava. The Ibn Salih Member is nowadays referred to as the upper part of the Dana Formation (Fig. 5.1.3).

The Ibn Salih Member overlies the Wadi Bustan without a visible, or only slight disconformity, or unconformably rests on a variety of older rock formations. In places such as near Shubak, 45 km NNW of Ma'an, it overlies the Wadi Bustan with a distinct erosional and slight angular unconformity. In this locality the Ibn Salih Member is overlain by a 10 m thick basalt sheet. The thickness is usually in the order of tens of meters, attaining up to 120 at the western side of Jebel Harun, near Petra. The sequence is known from numerous outcrops in the Arava and eastward, on the Transjordanian Plateau, its northernmost occurrence probably in Edh Dhira, east of the Lisan Peninsula (Bender 1974a, p. 91; and see Section 5.1.2.4).

The Ibn Salih Member commences with a scarp, consisting of 1–8 m thick coarse-conglomerate beds, alternating with brown-ochre and pale-yellow, fine conglomeratic sandstones. The coarse clastic constituents are late Cretaceous and Eocene, sub- to well-rounded rocks, cemented by a calcareous-sandy, hard matrix. No fossils are reported, thus, based solely on similarity in lithology, Bender broadly correlated the sequence with the Hazeva Formation.

5.2.2 Pliocene

The late Miocene Eritrean faulting considerably changed the face of the southern Levant, mainly by connecting the Jordan Valley and the Mediterranean, through the Yizre'el Rift Valley. Hydrographic connections with the Persian Gulf ended, the previous late Miocene troughs had already been filled up with sediments and most of the landscape became peneplained. Consequently, the Pliocene transgression flooded vast regions with shallow seas, lagoons and lakes, fed by a newly developed river system, which left its gravels as testimony (Garfunkel & Horowitz 1966; Horowitz 1973; 1974; 1979, p. 79). This paleogeography is expressed by lateral facies changes, so that lagoonal sediments grade into lacustrine, themselves gradually changing to fluviatile, all indeed synchronous. Formations were named according to the different lithologies, so that time correlative units were assigned different names over relatively short distances. Horowitz (1973) proposed to refer to the entire Pliocene sequence of the northern sector of the Jordan Valley and surrounding regions, including all its lateral facies, as the "Bira Series" (a term first suggested in Picard 1943), after the well-studied Bira Formation of the central Jordan Valley. The use of this term was subsequently neglected, when it was realized that it caused further confusion, between the "Formation" and the "Series". In addition, considerable volcanic activity is known from the region at that time.

To the south, the term "Usdum Series" (or "Usdom Group") was coined by Wyllie (1931), later amended by Wetzel & Morton (1959), Bentor & Vroman (1960) and Zak (1967). The term was used through the years for a variety of formations around the Dead Sea, in the central and southern Jordan Valley and in the Arava. Some of these sequences are referred to since the beginning of geological research in the southern Levant (see Chapter 2) by numerous investigators. Wyllie included in his definition only the outcrops at Mount Sedom; Wetzel & Morton included also the Miocene Hazeva Formation sediments of the eastern side of the Dead Sea; Bentor & Vroman referred only to the Sedom Formation; while Daniel (1963) included in the term all the formations from the Jordan Valley and vicinity from the Oligocene until the present day. It is quite clear, then, that this term has no practical use at present, when further knowledge has permitted splitting the "Series" into better-defined distinct formations.

The Pliocene formations were studied in great detail in the central Jordan Valley by numerous investigators. The Pliocene Mediterranean had reached the area of the Jordan Valley through the Yizre'el Valley, making the central Jordan Valley a meeting point of continental and marine environments of deposition. Thus, formations in this region are described here first, and are taken as reference for discussing contemporaneous rock units elsewhere.

5.2.2.1 *Bira Formation*

Author: Picard (1943, following Blake 1928), "Bira Series"; amended by Schulman (1962), "Bira Marl".

The Bira Formation is named after Wadi Bira (Nahal Tavor), joining the central Jordan Valley, which is the type area (Fig. 5.2.4). Synonyms are "Tonmergel"

Figure 5.2.4. The south facing flank of Wadi Bira (Nahal Tavor), approaching the central Jordan Valley. The white sequence at the base is the Bira Formation, overlain by the Intermediate Basalt, Gesher Formation (white again) and Cover Basalt.

(Picard 1932), "Brackish Water Formation" (Bentor 1946), "Brackish–Lagoonal Series" or "Brackwater Series" (Schulman 1959) and "Poriyya Formation" (Bentor 1960). It is not clear when exactly the term "Bira Marl" was replaced by "Formation", but the latter is now in use and is retained here. A type section is given in Schulman (1962) at Wadi Umm as Sawalil, a small tributary of Wadi Bira, only for the basal member of the Formation, while for the rest only reference sections are available, in many localities in the central Jordan and Yizre'el valleys (Schulman 1959, 1962, Shaliv 1991).

In places where the Umm Sabune occurs, such as Belvoir, the Bira Formation overlies the Conglomerate dis- or paraconformably (Schulman 1962). Otherwise, the Bira overlies discordantly a variety of older rocks, over a structural and erosional relief. The Bira is conformably or paraconformably overlain by the Gesher Formation, and partly by the Intermediate and Cover basalts in places where the Gesher does not occur, such as the Yizre'el Valley, where the entire Pliocene sequence is of lagoonal nature.

Bira Formation sediments, occasionally intercalated by volcanics of the Intermediate Basalt and Fajjas Tuff, are known from the entire central Jordan Valley, extending westward and covering large areas in the Yizre'el Valley and vicinity, interfingering to the west with marine sediments ("Yagur Facies") of Pliocene age. To the north and east, particularly in the Golan Heights, the Bira is replaced by a variety of freshwater and fluviatile sediments, termed the "Gesher Formation" (see below). Southward, evaporites are progressively more common. Thicknesses vary considerably, from several hundred meters in the deeper basins such as the central Jordan and the central parts of the Yizre'el valleys, to complete disappearance at the margins of the depositional basins. Typical thicknesses are between several tens and 200 m.

The lowermost 20 m, where developed, comprise alternations of coarse and fine, unsorted clastics composed of basalt pebbles in a variety of weathering stages. These, especially the fine-grained horizons, bear badly preserved fossils of small mollusks, grading in places to biocalcarenite. Toward the margins of the basin a lumachel is developed, built of hard limestone rich in oyster shells, which is also known from the eastern side of the central Jordan Valley (Wetzel & Morton 1959) and the En Gev region (Michelson 1972). Two horizons of such lumachel are described by Picard (1943) from the western part of the Yizre'el Valley.

The main body of the Bira comprises a complex of clayey, chalky, marly and gypsiferous sediments, especially rich in pyrite and organic matter, both of which are oxidized at the outcrops. Gypsum is quite common, with irregular occurrences. Occasionally, thin beds of hard, limonitic crystalline limestone occur, as do beds of black-brown clay or cryptocrystalline dolomite. These are finely stratified, very soft rocks, white to yellow at the outcrops, which are usually covered by slumps.

The top of the Bira is sandy, limonitic, brownish, indicating according to Schulman a paraconformable transition to the overlying Gesher Formation. In places, such as near the settlement of Gesher, the sands grade into gypsum up to 14 m thick. The lower, main part is a massive, pure gypsum, bearing some authigenic magnetite, overlain by a thin layer of clays rich in gypsum, iron minerals and organic matter. Further up several thin, black gypsum layers occur, grading upward into shales with gypsum, iron and some calcite. Calcite increases within several meters, until the rocks become fine, platy dolomite, forming the base of the Gesher Formation. Raab (1998), based mainly on geochemical grounds, suggested uniting the gypsum and dolomite occurrences at the top of the Bira with the platy "limestone" (f1 of Schulman) basal member of the Gesher Formation, made chiefly of dolomite, referring to this unit as the "Dolomite–Gypsum Member".

Unidentified plant remains and redeposited Cretaceous and Eocene foraminifera are very common in the Bira Formation (Reiss, in Schulman 1962). Foraminifers indicating a general Neogene age and brackish environment are *Ammonia beccarii* and a variety of benthonic miliolids (Weiler 1961). Others indicating lagoonal or shallow marine environments are ostracodes such as *Cyprideis torosa* (Rosenfeld, in Shaliv 1991), clams, of which *Crassostrea* and *Ostrea* (possibly *edulis*) are common, accompanied by *Chama*, *Lucina*, *Cardium* (Bentor 1946), *Lithophaga* and others, and gastropods such as *Cerithium*. Occasionally the environment of deposition becomes less saline, evidenced by occurrences of freshwater mollusks such as *Melanoides tuberculatus*, *Theodoxus jordani* and *Melanopsis praemorsa* (Lewy, in Shaliv 1991). Clupeid fish remains are reported from the gypsum layers (Avnimelech & Steinitz 1951). Unfortunately, none of these fossils are age indicative, since most of them already appeared at the beginning of the Miocene, living until the present day.

The age of the Bira Formation is considered Pliocene by most investigators. Radiometric datings obtained by Shaliv (1991) and by Heimann et al. (1996) on neighboring basalts led these authors to suggest a late Miocene age for the Bira. This problem is discussed in detail in Chapter 7.

5.2.2.2 *Gesher Formation*

Author: Schulman (1962).

The Formation is named after the settlement of Gesher, on the western rim of the central Jordan Valley, 2 km from the two type sections, at Nahal Hemed and Nahal Ghareb. The Gesher, or parts of the Formation, was known by several names: "Pliozäne Süsswasserkalke" (Blanckenhorn 1912); "Plattenkalk" overlain by "Oolithkalk" (Picard 1932); "*Melania* Limestone" or "*Hydrobia fraasi* Beds" (Blake 1935); "Limnic" or "Freshwater Beds" (Picard 1943); "Freshwater Series" (Bentor 1946) and "Degania Formation" (Bentor 1960).

The Gesher overlies the Bira Formation (Fig. 5.2.4), mostly paraconformably, but in several localities the latter grades conformably into the former (Schulman 1962). It seems also that the Bira grades laterally into the Gesher, particularly to the east (Horowitz 1973), overlying unconformably the Herod Formation (Michelson 1972). To the west, in the Yizre'el Valley, the freshwater Gesher disappears and the entire Pliocene sequence was deposited in a lagoon environment, therefore termed "Bira", closer to the sea. Besides the Bira, the Gesher overlies unconformably a variety of older rocks in different locations. The Gesher Formation is overlain unconformably, over an erosional relief, by the Cover Basalt. An intercalation of volcanics, the Intermediate Basalt and Fajjas Tuff, interfingers the Gesher sediments.

The Gesher occurs, at thicknesses of 0–70 m but usually 20–50, all around the central Jordan Valley. It is somewhat thicker southward, in Marma Feiyad (110–120 m), eastward, in the Golan (80 m) and northward, at the northwestern corner of Lake Kinneret, where it attains some 120 m (Saltzman 1964).

The type sections were subdivided by Schulman into three members. The lowermost, "f1" (Picard's "Plattenkalk"), some 22 m thick, overlies conformably the gypsum beds of the underlying Bira Formation. It is characterized by very fine bedding mainly of extremely white limestones. In places white chalk occurs, with some ostracodes possibly indicating a hypersaline environment (Rosenfeld et al. 1981). Fish remains, most probably Cichlidae, are found at the base (Avnimelech & Steinitz 1951), but other than that no fossils are reported. Primary limonite and manganese nodules are quite common; some of the rocks are saline, some are silicified.

The middle member, "f2" ("Main Oolite" of Bentor 1946), is 17.5 m thick, distinctive in the landscape as it forms gray-white, hard cliffs, of medium to coarse bedding. The rocks are mainly detritic limestones, composed chiefly of oolites, accumulated around ostracode shells. Epigenetic silicification and dolomitization are quite common, forming very hard rocks. The abundant ostracodes indicate a freshwater environment of deposition (Rosenfeld et al. 1981), as do the numerous shells of the gastropod *Hydrobia fraasi*. The uppermost 6 m make up the third member, "f3", which is typically variegated, contains abundant clays and is rich in freshwater fossils, organic matter and plant remains. The Gesher Formation does not show frequent facies changes, except when approaching the lake shores, where the sediments may contain some clastic components.

The fossils mentioned above, particularly the ostracodes and *Hydrobia*, although not age indicative (range Neogene–Recent), are very good evidence of the freshwater environment of deposition of the Gesher Formation. Most investigators agree on a Pliocene age for this unit, but when exactly within this stage is a matter of debate (see Chapter 7).

5.2.2.3 *Tel Hai Formation*

Author: Picard (1952), "Lower Neogene Lacustrine Series"; amended by Rosenberg (1960), "Süsswaser Mergel und Kalke"; by Glikson (1966), "Tel Hai Freshwater Limestone", "Tel Hai Formation". Sneh (1996) views both the Tel Hai Formation and the Tanur Conglomerate as a single entity, termed the "Kefar Gil'adi Formation", an opinion not favored here.

The Formation is named after Tel Hai, a settlement at the northwestern corner of the Hula Valley, where the type section is located (Fig. 5.2.5). The lacustrine limestones overlie unconformably middle Eocene rocks, their top is always truncated, and they seem to grade laterally into the Tanur Conglomerate. The Formation as defined is known only from limited occurrences in the type area, and its thickness attains some 400 m.

The main rock types are interbedded detritic and fine grained limestones, accompanied by marl, chalk, calcarenites, shales and some conglomerate horizons. The limestones, usually fine but occasionally medium to coarse grained, are limonitic, hard, thinly bedded, platy or laminated, forming small cliffs. Fossil gastropods are abundant, especially in the limestones and chalk, consisting of freshwater taxa such as *Hydrobia fraasi*, *Chondrina jamina*, *Lymnaea*, *Planorbis* and terrestrial *Helix*. They indicate a lacustrine environment, but are unfortunately not very helpful with age, since all are known from the early Miocene until today.

Figure 5.2.5. Steeply dipping beds of the Tel Hai Formation, near this settlement.

Based mainly on paleogeographic considerations, Horowitz (1973) placed the Tel Hai in the Pliocene. Sneh (1996) suggested a late Miocene age, correlative to the Umm Sabune Conglomerate.

5.2.2.4 *Tanur Conglomerate*

Author: Picard (1952), "Lower Neogene Pudding Conglomerate"; amended by Rosenberg (1960), "Nummuliten-Marmor Konglomerate"; and by Glikson (1966). Sneh (1996) views both the Tel Hai Formation and the Tanur Conglomerate as a single entity, termed the "Kefar Gil'adi Formation", an opinion not favored here.

The Formation is named after the Tanur (oven) waterfall, north of the Hula Valley, where it was misidentified. The type section is several kilometers to the south, near the town of Qiryat Shemona. The Tanur Conglomerate (Fig. 5.2.6) unconformably overlies middle Eocene limestones, but in most of the outcrops contacts with other rock units are along faults. It grades in some places into freshwater limestones of the Tel Hai type. The Tanur Conglomerate, as was originally defined by Glikson, is confined to the northeastern corner of the Hula Valley and somewhat northward. Other conglomerates of similar nature are reported (Horowitz 1973) from both flanks of the Valley, extending southward as far as areas close to Lake Kinneret (see Chapter 7). The thickness at the type section is in the order

Figure 5.2.6. Tanur Conglomerate near Metulla, northern Hula Valley, with its typical red matrix.

of 100 m, but since the base is not exposed and the top truncated, this is only a minimum figure. Thicknesses in other exposures diminish, possibly due to faulting and subsequent erosion. Glikson maintains that since the Tanur and Tel Hai are correlative, the former may attain a thickness of up to 400 m.

The Tanur Conglomerate is composed of sub-angular to well-rounded limestone pebbles and cobbles, well cemented by calcarenite. The rock is stratified, hard, forming a landscape resembling late Cretaceous rocks in the region. The pebbles comprise two main types of light colored limestones, one microcrystalline, homogenous, dense, with calcite veins, the other made largely of *Nummulites*, accompanied by other middle Eocene fossils. The environment of deposition is no doubt fluviatile, but no fossils are known from the matrix. When interfingering with the Tel Hai, the typical freshwater fossils of the latter occur.

A gravel unit defined in Heimann (1985) as "Egel Gravel", 10–40 m thick, consisting of Jurassic, Cretaceous and possibly Pliocene components, in the northern Hula Valley, was considered by Horowitz (1973), Sneh (1996) and U. Kafri (Geological Survey of Israel 1998, pers. comm.) as a correlative of the Tanur Conglomerate, viewing the differences in pebbles lithology as arriving from different sources. Since nothing conclusive in Heimann's descriptions changes this view, it is still held here. Based mainly on paleogeographic considerations, Horowitz

(1973) placed the Tanur in the Pliocene. Sneh (1996) suggested a late Miocene age, correlative with the Umm Sabune Conglomerate.

5.2.2.5 *Zahlé Beds*

North of the Hula Valley in the Beqa'a syncline, Dubertret (1952 and other publications) describes a sequence attaining 800–900 m of lacustrine sediments, similar in every respect to the Tel Hai freshwater limestone. This sequence, termed the "Marne Lacustre de Zahlé", grades laterally and is overlain by a suite of conglomerates 500–600 m thick, named the "Poudingue de Zahlé". It seems quite safe (Horowitz 1973) to correlate the Zahlé Beds with the Tel Hai and Tanur suite, and other occurrences around the Hula and southward. Dubertret (1963) suggested that the Zahlé Beds are correlative with the "Conglomérat du Sahel Sahra", near Damascus. In addition to the abundant freshwater mollusks already described in Blanckenhorn (1897), Kansou (1961) found teeth of *Hipparion* in the lacustrine sediments, leading him to assign a Pontian age to these deposits, and a Pliocene age for the overlying conglomerates. Unfortunately, this tiny horse is not age-conclusive, occurring from the late Miocene until some time in the middle Quaternary (Tchernov 1996).

5.2.2.6 *Sedom Formation*

Author: Wyllie et al. (1922), "Usdum Series" (*s. str.*); amended by Bentor & Vroman (1960), "Sedom Formation"; and by Zak (1967).

The Formation is named after Mount Sedom at the southwestern tip of the Dead Sea ("Jebel Usdum" or "Khashm Usdum" in Arabic), where the legendary Biblical city of Sodom is thought to have been located. The type section is taken at this locality, which is the only outcrop of these rocks (Fig. 5.2.7), which are known from numerous boreholes.

The base of the Sedom Formation is not exposed at the type locality, but is known from several boreholes in the vicinity to overlie, most probably conformably, late Miocene (Palynozone Mc, Fig. 6.1.2) sediments, which seem correlative to the Hufeira Formation. In some boreholes the Formation overlies older rocks. The Sedom is unconformably overlain by various types of caprock, or by the Amora (possibly conformably) and Lisan formations. The outcrops at Mount Sedom cover an elongated north–south oriented strip, some 10 km long and $1\text{–}1\frac{1}{2}$ km wide. The thickness is estimated by Zak (1967) at 1,500–2,000 m. An almost similar occurrence is known from a borehole drilled in the Lisan Peninsula, and several others are suggested to cause and underlie diapirs in the northern Dead Sea basin (Neev & Hall 1979) and in the Jordan Valley, north of the Dead Sea (Belitzky 1996, Belitzky & Mimran 1996). Shulman & Ben-Avraham (1999) consider, based on a geophysical survey, that Mount Sedom and the Lisan Peninsula are a single rocksalt body.

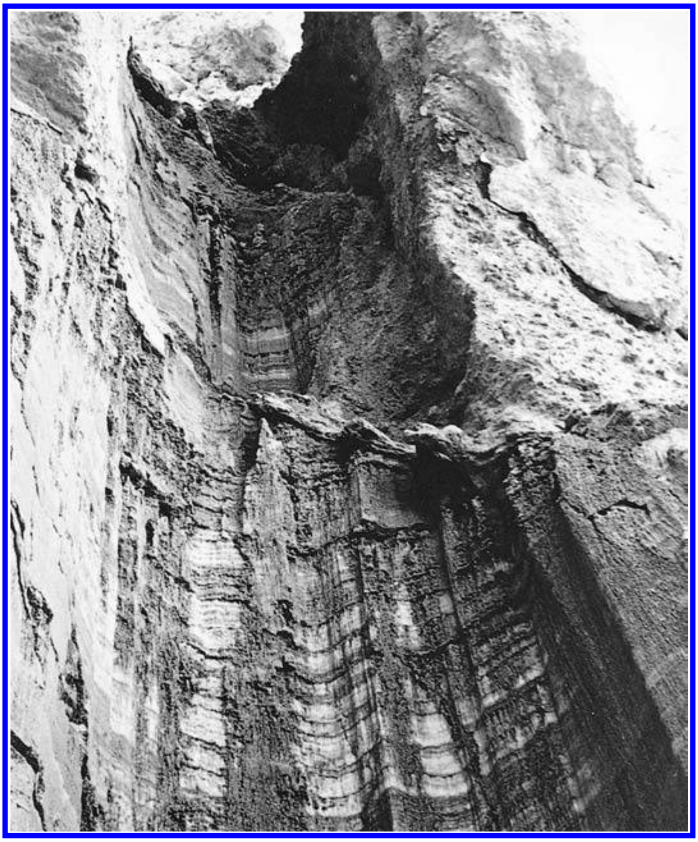


Figure 5.2.7. Rocksalt of the Sedom Formation, exposed in a sinkhole at Mount Sedom.

The Sedom Formation comprises principally massive rocksalt, alternating with thinner shale units. The salt units are mainly halite, accompanied by some anhydrite, dolomite, clay and silt, as thin laminae or interspersed within the salt. The shale units consist of finely laminated dolomite, silt and clay, accompanied by fine to coarse-bedded red or white sandstones and white or gray gypsum and anhydrite. The Formation is subdivided into five members, of which the lower two, the Karbolet (cockscomb) Salt and Shale and the Lot Salt, comprise the major part, some 1,400 m. They are overlain by the Benot (daughters of) Lot Shale, the Me'arat (cave of) Sedom Salt and the Hof (beach) Shale and Salt, attaining together somewhat less than 500 m in thickness.

The Karbolet comprises mainly rocksalt, with intercalations of silt, clay, dolomite, sandstone, anhydrite and gypsum. The overlying Lot also consists chiefly of rocksalt, with subordinate dolomite, clay and anhydrite. The Benot Lot is made of sandstones, anhydrite and shales, partly dolomitic, overlain by the massive, almost pure rocksalt of the Me'arat Sedom. The sequence is topped by the Hof

Shale and Salt Member, which is however always truncated, built of sandstones, clay, salt and anhydrite.

Except for the numerous pollen grains (see Chapter 6) and occasionally large quantities of unidentified plant debris, the Sedom Formation is poor in fossils. Notable are fish fossils of *Mugil priscus* (H. Steinitz in Bentor & Vroman 1960), very common in some horizons of the Formation. Footmarks of birds and a large mammal are reported in Zak (1967), as well as some insect remains, but these were not identified. The Sedom Formation was assigned a variety of ages by different authors, from Cretaceous by Lartet (1869), who mis-correlated the fossil fish with others known at that time from the Lebanon (see Chapter 2), through the Oligocene to Miocene (Bentor 1960), Pliocene by most investigators and Pliocene–Pleistocene(?) by Zak (1967). Based on K–Ar ages of basalts east of the Rift Valley and their estimated rates of erosion, Steinitz & Bartov (1991) suggested an age for the base of the Sedom Formation younger than 7–9 Ma, maybe as young as 6 Ma, while the top was assigned an age of 4–3.4 Ma.

The environment of deposition is accepted by everyone to have been hypersaline lagoonal, with remote but clear connections with the Mediterranean Sea, as evidenced by the geochemistry of the evaporites (Zak 1967, Raab 1998).

Mount Sedom is considered by most investigators to be a diapir, which continues its upward movement even today. In times when this process pushed the salt into the Rift Valley lakes (Raab 1998), its upper part had dissolved and the insoluble residue accumulated as caprock, as was noted already by Anderson (1852). Various such stages, attaining up to 40 m in thickness, are described in Zak (1967) and in Raab (1998), but dating these uplift and dissolution stages is far from certain. Others suggested that Mount Sedom is a tilted block, based, among other considerations, on the fact that several boreholes sunk into the rocksalt encountered underlying Miocene sediments (Horowitz & Horowitz 1990). This view is also supported by geochemical observations (J. Charrach, Israel Potash Co., 1999, pers. comm.). Even if the last opinion is favored, it is clear that some streaming of the rocksalt indeed took place.

5.2.2.7 *Amora Formation*

Author: Wyllie et al. (1922), "Gypseo-Argillaceous Division" or "Gypsum and Shale"; amended by Picard (1950), "Foothill Series"; by Bentor & Vroman (1960), "Foothill Formation"; and by Zak (1967).

The Formation is named after the lost Biblical city of Gomorra (Amora in Hebrew), the type section is in Mount Sedom (Fig. 5.2.8). The Amora Formation overlies, probably unconformably, the Sedom Formation and some of its caprock horizons. It is overlain, over an erosional and structural relief, by the late Pleistocene Lisan Formation and younger sediments and lake terraces. Picard (1950) reported that the Amora is also covered by the Samra Formation, a possibility mentioned also by Zak (1967), probably based on erroneous identification

Figure 5.2.8. Tilted Amora Formation hogbacks, overlying Sedom Formation massive rocksalt (dark layer in middle left), at Mount Sedom.

of the latter. The sequence crops out as a series of patches encircling the foothills of Mount Sedom, forming a typical landscape by the alternation of soft and hard rocks, as described in Picard (1950, p. 6): "Opposing erosion, the gypsum beds form conspicuous cathedral-like peaks". The sequence does not crop out anywhere else, but was penetrated (Fig. 6.1.2) by numerous boreholes (see Chapter 6), where it overlies the Sedom conformably. The thickness, according to Picard (1950) and Bentor & Vroman (1960), is more than 700 m; Zak (1967) placed the lower part of the "Foothill" in the Sedom Formation, leaving for the Amora proper somewhat more than 400 m.

The Amora Formation comprises laminar or shaley chalky marls, anhydrite and gypsum, rocksalt, sandstones and conglomerates of light colors, usually gray–yellowish, olive gray, brownish and whitish. It was subdivided by Zak into five members: the lowermost, "Amc", comprises marl, sand and conglomerates more than 130 m thick, the gravel made of flint, limestone, quartz porphyry and a variety of other igneous rocks. This is overlain by 70 m of the "Ama" Member, mainly of marl and anhydrite with subordinate conglomerates, consisting of the same variety of gravel as the underlying unit. The third member, "As", consists of 6 m of massive rocksalt, on which lies "Am", 130–150 m chiefly of marls, again with subordinate gravel of similar composition to those underlying. The top member, "Ams", is built chiefly of marl, chalk and sand, attaining more than 100 m in thickness.

No fossils are reported from the Amora Formation, except for plant debris and numerous pollen grains (see Chapter 6). The age is believed to be Pliocene (Picard 1950, Horowitz & Horowitz 1990), Miocene–Pliocene (Bentor & Vroman 1960) or Pleistocene (Zak 1967). Based on K–Ar ages of basalts east of the Rift Valley and their estimated rates of erosion, Steinitz & Bartov (1991) suggested an age for the base of the Amora Formation of 4–3.4 Ma, while the top was placed at a little more than two million years, which broadly conforms with the palynostratigraphic assignment. The environment of deposition is considered terrestrial by Bentor & Vroman, lagoonal by Zak (1967) and Horowitz & Zak (1968), with a remote connection with the Mediterranean Sea.

5.2.2.8 *Oolitic Formation*

Author: Picard (1931).

The Formation is named after its distinct petrographic nature. The type section was made near Nabi Musa, on the Jerusalem–Jericho road, some 10 km west of the northern tip of the Dead Sea (Fig. 5.2.9), approximately 100 m below sea level. The Oolitic Formation overlies unconformably late Cretaceous rocks, while the eroded top is exposed. The Formation, as defined, is known only from three small outcrops at its type area, where the thickness is a little more than 6 m.

The Formation is chiefly made of pinkish oolitic limestones, with black manganese coloring of their fine cracks. The single oolites measure $\frac{1}{2}-\frac{3}{4}$ mm, of oval or spheroid shape, strikingly symmetrically developed, mostly with a white outer

Figure 5.2.9. Oolitic Formation, unconformably overlying Senonian beds west of Jericho.

peel and a reddish ferruginous center. The limestones often contain flint pebbles, 2–3 cm across, and some fine-grained flint splinters. No fossils are reported, the presumed environment of deposition is lacustrine, possibly of a limited extent. Picard assumed a Pliocene to early Pleistocene age for the Oolitic Formation, based on his recognition that its sedimentation was confined to the Jordan Rift Valley, postdating initial rifting. He further suggests that it may be correlative with the Bira and Gesher formations of the central Jordan Valley.

5.2.2.9 *Mazzar Formation*

Author: Shahar et al. (1966).

Named after Nahal Mazzar, near Hazera erosion cirque, some 25 km southwest of the Dead Sea, where the type outcrop is described, the Mazzar lies conformably within and interfingers (Fig. 5.2.10) with the Arava Formation (Elron 1980, Eyal 1984). Various outcrops of presumably similar rocks have been assigned to the Mazzar by many authors (for details see Avni 1998) in the entire northern and central Arava. This poses a problem for delineating its distribution, since although the rocks show some similarity, they are not necessarily of the same age (Livnat & Kronfeld 1990), and their fossil assemblages are also different. At the type locality

Figure 5.2.10. Mazzar Formation in contact with Arava Formation conglomerates at Nahal Nemiyya, some 5 km west of the northern Arava.

the microfauna indicates a brackish environment, while many other occurrences are typified by freshwater fauna. This led Avni (1998) to suggest that the entire suite of outcrops was laid down in a single large lake, occupying the entire northern Arava, an opinion not favored here due to the difficulties mentioned above. It is therefore suggested to view the Mazzar in its original context, as a marginal marine intercalation within the Arava Formation. Other lacustrine sediments are treated here separately, particularly when they could be dated (see Section 5.2.2.14).

The thickness at the type area is about 6 m, comprising alternations of well-bedded light gray marls, with pink, reddish, yellow or gray sands cemented by carbonate, both occurring in layers a few centimeters thick (Eyal 1984). Fine ripple marks and load structures, north–south oriented, probably indicating a beach, are present. Numerous unidentified carbonized plant remains are reported. Typical microfossils are the foraminifera *Ammonia beccarii* and *Elphidium* sp., and the ostracode *Cyprideis littoralis*, indicating a Neogene to recent brackish environment, possibly marine. These are similar to the assemblage designated the "Lido facies" of the Bira Formation near Tiberias (Reiss, Gerry, in Eyal 1984) and at Mount Sedom (Zak 1967), or the "Lido Formation" in the southern Dead Sea and northern Arava (Zak & Freund 1981). Similar fossil assemblages are reported from the En Gedi 2 and Arava 1 boreholes, at depths corresponding to Palynozone Pa (Neev & Emery 1967, Bartov, in Elron 1980, Horowitz & Horowitz 1990). Zak (1967), Horowitz (1974), Elron (1980) and Eyal (1984) suggest a broad correlation of the Mazzar Formation with the Pliocene Bira and Gesher formations to the north.

5.2.2.10 *Samra Formation*

Author: Picard (1943), "Samra Series"; amended by Bentor & Vroman (1954b), "Samra Beds"; by Bentor & Vroman (1957), "Samra Stage"; by Bentor & Vroman (1960); and by Bender (1974a, p. 95).

The Formation is named after Khirbet Samra, 6 km NE of Jericho, where its gravels were quarried. The lower contact is not exposed in the type locality, where the unit, gently dipping to the east (Fig. 5.2.11), underlies the late Pleistocene Lisan Formation with a slight angular unconformity. Bentor & Vroman (1957) report the Samra as overlying unconformably the Arava Formation and older rocks, overlain by the Lisan in the same manner described in Picard (1943, p. 146). Picard reports a thickness of 10–12 m from the type locality, while Bentor & Vroman measured up to 25 m in the northern Arava. Picard described the Samra from the southern Jordan Valley, while Bentor (1960) includes occurrences from the northern Arava, extending northward up to Beth She'an, possibly also in the eastern Judean Desert.

Picard (1943) describes the Samra at the type area as comprising some 6–10 m of calcareous sandstones at the lower part, practically without quartz but with fine flint particles, with a thin intercalation of calcareous marl. These are topped by 1–2 m of hard loam, fine flint sand and flint conglomerate, with pebbles several centimeters across. Northward, in the Marma Feiyad area, Picard describes the

Figure 5.2.12. En Feshha Conglomerate at the northwestern end of the Dead Sea.

section is located (Fig. 5.2.13) about 3 km east of the village. The unit unconformably overlies a rugged relief cut into Mesozoic and Cenozoic rocks. The upper contact is problematic, since Bender claims that the Shagur is overlain by the Ghor el Qatar Series, over an angular unconformity. If indeed the latter correlates with the Dana Formation (see Section 5.1.2.6), which is still questionable, such a situation seems impossible. The Shagur was observed by Bender in extensive areas, exclusively on the eastern side of the southern Jordan Valley and the Dead Sea. The maximum thickness maybe more than 100 m, but is highly irregular, due to deposition on previous relief and subsequent faulting and erosion.

The sequence comprises massive thick-bedded conglomerates, containing sub-rounded blocks up to 1 m across, embedded in a hard, calcareous, calcareous sandy or siliceous matrix. The conglomerates alternate with, or grade into hard, thick-bedded travertines and into hard, often siliceous and locally conglomeratic fine sandstones, which bear black weathering colors. Fossils collected near Shagur (Huckriede, in Bender 1974a, p. 92) include *Melanopsis* cf. *praemorsa*, *Trichia*, *Poiretia*, ostracodes and plant remains, among which date palms were identified. These fossils are indicative of a fluvio-lacustrine environment, possibly with some springs, but unfortunately not of age. Bender assigned the Shagur Formation a late Pliocene to earliest Pleistocene age.

Figure 5.2.11. Samra Formation gravel beds, gently dipping east toward the southern Jordan Valley, just north of Jericho.

Samra as a lacustrine limestone with *Hydrobia*, *Bythinia*, *Planorbis* and *Melanoides* mollusk shells. These beds were assigned by Schulman & Rosenthal (1968), followed by others, to the Gesher Formation. Bentor & Vroman (1957) describe the Samra from the northern Arava as consisting of fine-grained, partly calcareous sandstone and silt, oolitic chalk and limestone, finely bedded calcareous shales, green clays, conglomerates and breccias, with flint and limestone pebbles and fragments. They also report fossils similar to those described in Picard, but these are not indicative of a precise age. Both authors suggested a "Lower Pleistocene" age and a fluvio-limnic environment of deposition for the Samra Formation. Picard further indicated that both the Samra and the Oolitic Formation are time equivalent, different facies of the same system. Picard (1943, p. 147) describes a sequence of well-cemented conglomerates in Wadi Malih, west of Marma Feiyad, designated the "Wadi Malih Gompholites", sandwiched within the Samra. Schulman & Rosenthal (1968) maintain that these should be younger. Considering the nature and especially the hard calcareous cementation of the conglomerates, Picard's view is preferred here.

Daniel (1963, p. 370) states that the Samra sequence rests with an angular unconformity on the Pliocene Usdom Series. He included also the late Pleistocene Naharayim Formation gravels (Picard 1965) within the Samra, thus acquiring a thickness of 65 m, which seems unacceptable (Horowitz 1979, p. 144). The tendency of numerous authors to call "Samra" every gravel bed in the Jordan Valley, from the Pliocene almost to the present day, disregarding its stratigraphic position (e.g. Bender 1974a, p. 97; Begin et al. 1974; Avni 1998; and many others), is not favored here (see also Horowitz 1974).

5.2.2.11 *En Feshha Conglomerate*

Author: Roth (1970).

The unit is named after En Feshha, one of the main springs on the northwestern shore of the Dead Sea. No type section is provided. Raz (1983) renamed the Conglomerate "Darga", which is used occasionally, although "En Feshha" has priority. The Conglomerate overlies unconformably various formations of the late Cretaceous Judea Group and is overlain, again disconformably, by the late Pleistocene Lisan Formation. It appears that terraces of earlier Quaternary lakes are cut into the En Feshha, but no dates are provided for these (Mor 1987). The lower and upper contacts are both angular and erosional. The En Feshha is known from intermittent outcrops almost all along both the western and eastern shores of the Dead Sea, except for wadi outlets where it is missing altogether. The thickness can attain up to 12 m, usually quite difficult to tell due to the irregular nature of the Conglomerate and its being considerably faulted (Figs 5.2.12 and 3.1.6).

The En Feshha Conglomerate comprises polymictic gravels, representing almost all previous rock formations occurring in the vicinity. The components appear in different sizes, sphericity, sorting, roundness and cementation, usually coated by a dark to blackish patina. Cross bedding is quite common, the cement comprises usually sand and silt, rich in redeposited foraminifera of Danian and older ages (Reiss, in Raz 1983), occasionally very hard, dolomitic. Autochtonous travertines are sometimes interbedded within the gravel horizons.

No fossils are reported from the sequence. The environment of deposition suggested by Raz (1983) and Mor (1987) is fluviatile or fluvio-lacustrine, over an erosional relief, before the main faulting of the Dead Sea Rift, as seen by its faulting and predating the channeling of the wadis. No age is specified, but Raz suggests, on morpho-structural grounds, correlation with the Amora and Samra formations. The typical hard carbonate cementation maybe attributed to contact with seawater, which penetrated the region during the period of formation of the Sedom and Amora (Zak 1967), which was also responsible for the diagenetic dolomitization of older rocks (Raz 1983).

5.2.2.12 *Shagur Formation*

Author: Bender (1974a, p. 93).

The Formation is named after the village of Ash Shagur in the southern Jordan Valley, 6–7 km NE of the northeastern corner of the Dead Sea, where the type

Figure 5.2.13. Westward dipping Shagur Formation near Kufrein, northeastern Dead Sea corner.

5.2.2.13 *Arava Formation*

Author: Bentor & Vroman (1957), "The Arava Conglomerate"; amended by Garfunkel & Horowitz (1966); by Horowitz (1974); by Sneh (1982); by Eyal (1984); and by Avni (1998).

As seen from the list of authors (which is most probably not complete), the Arava Formation has evoked plenty of interest and debate ever since it was first recognized as an independent lithostratigraphic unit. Before, it was considered part of the Hazeva Formation, or the Dhira Ibn Salih Member (upper part of the Dana Formation). The debate is basically about how extensive is the Arava, both laterally and vertically. The two principal approaches are that either the term should apply to a limited time span, during which the Formation represents a river system leading to the Pliocene Sedom embayment, whose terminal base level is the Mediterranean (Bentor & Vroman 1957; Garfunkel & Horowitz 1966; Horowitz 1974; 1979, p. 76; 1992a, p. 359; Horowitz & Horowitz 1990), or that it should include the entire conglomerate sequence of rivers and wadis leading into the Dead Sea, almost until the present day (Sneh 1982). The first view is preferred here, for it keeps the Arava Formation within well-defined chronostratigraphic and paleogeographic limits.

The Formation was first defined in the northern Arava, with no definitive type section, where it overlies, always unconformably, the Hazeva (Fig. 5.2.14) and older formations. The top is cut and filled up by younger gravel beds, of which the oldest is the HaMeshar Formation (Zehiha Formation of Avni 1998, erroneously). The Arava Formation is known from vast areas in the southern and central Negev, the northern Arava up to the Dead Sea, where its typical pebbles were found within the Amora Formation (Zak 1967). The thickness varies considerably from just a few meters up to several tens in the deeper channels. The conglomerates of the Arava interfinger with the Mazzar Formation in several localities south of the Dead Sea (see Section 5.2.2.9), while in boreholes such as Arava 1 (Horowitz & Horowitz 1990) with sediments bearing spectra of Palynozones Pa and Pb.

The Arava Formation is made of loosely cemented conglomerates in a reddish sandstone matrix, occasionally with sand lenses. The pebbles vary in size, composition and sphericity according to their relative locations in the channels. The sequence is well stratified, commonly with imbrications and cross bedding. Two principal facies are distinguished in the northern Arava (Eyal 1984), the En Yahav facies typified by perfectly rounded pebbles, and the Dohan facies much less so. The first is notable for its magmatic and metamorphic pebbles, besides the local constituents, indicating a distant provenance such as the Edom mountains in

Figure 5.2.14. The Arava Formation makes up the highest surface, overlying the Hazeva Formation in the northern Arava. This surface is cut and filled by several younger lake terraces, in front.

southern Transjordan or the Neshef hills close to Elat. The other facies, of a local nature, comprises less-rounded and sorted pebbles, derived from sources nearby. Another, of only local significance, is the Neqarot facies (Baer 1981), which contains a suite of pebbles originating in the Makhtesh Ramon erosion cirque, indicating its drainage to the Arava at that time. Enmar et al. (1998a) report travertines at the top of the Arava Formation, which they christened the "Arava Travertine".

Avni (1998) subdivided the Arava Formation into two members, the Saggi (Na1) base conglomerate and the Kuntilla (Na2), whose lower part is lacustrine, grading laterally and topped by fluviatile pebble beds. No fossils have ever been recovered from the Arava Formation, which was deposited in a large fluvio-lacustrine system.

5.2.2.14 *Other unnamed, possibly Pliocene, occurrences*

Shahar (1969) described a 5 m sequence of conglomerate and gray chalk near Nabi Musa, overlying late Cretaceous rocks. The conglomerate makes the lowermost 2 m, grading laterally and vertically into the overlying chalk. The base is composed of well-sorted limestone, chert and chalk pebbles in a marly matrix, overlain by poorly sorted chert and limestone components. The chalk contains marine Pliocene ostracodes and foraminifers and, according to Shahar, is reminiscent of both the Bira and Mazzar formations.

Bender (1974a, p. 92) describes an isolated occurrence, some 10 km north of Gharandal in the Arava Valley, of more than 40 m of alternating oolitic limestones, chalky, partly siliceous limestones, fine-crystalline limestones and friable marls. These contain poorly preserved microfossils such as smooth shelled ostracodes and "dwarfed" foraminifera such as ?*Rotalia* and ?*Discorbis*, indicating in general a Neogene age and a shallow estuarine to shallow marine environment of deposition.

5.2.3 Earliest Quaternary

The late Pliocene global sea level drop ended the flooding of the Jordan Valley and its surrounding regions, where lagoons, lakes and rivers deposited the suite of Pliocene formations described above. The base level was lowered with the regression, causing channeling westward toward the retreating sea accompanied by slight structural disturbances and volcanism. The early Pleistocene Calabrian and Sicilian transgressions again raised the sea level, by approximately 120 and 90 m, respectively, compared to the present day. These higher seas could not reach as far as the Jordan Valley region, depositing marine sediments only on the Mediterranean coastal plain (Horowitz 1979, p. 84–115). The higher erosion base levels did however cause silting up of the drainage systems, leaving behind predominantly fluviatile sediments, with several intermediate lakes occupying basins within the slowly subsiding Jordan Rift Valley (Levin & Horowitz 1987).

Outcrops are quite rare in the Rift, due to subsequent subsidence and erosion, but the sequences are well known from numerous boreholes (Horowitz & Horowitz 1990) drilled in the Jordan Valley (Fig. 6.1.2). On the Rift shoulders, the combination of a relatively short duration of deposition in rather flat channels, followed by considerable uplift and erosion, resulted in rare, patchy exposures of the early Pleistocene fluviatile sediments. The usual scarcity of fossils in these deposits (except for pollen grains in boreholes, but outcrops are highly oxidized), does not make their exact assignments any easier. All of these factors make for the prolific nomenclature applied to rocks of this relatively short time span, less than a million years.

5.2.3.1 *Melekh Sedom Sands*

Author: Clark et al. (1984).

"Melekh Sedom Sands" is a term used by oil companies operating in the Jordan Rift Valley to describe a sequence of clastics, attaining a maximum thickness of 920 m penetrated in the Melekh (King of) Sdom (Sedom) 1 borehole, drilled in the southern Dead Sea basin (but this thickness is apparent, as are all numbers concerning this unit, since dips are known to exist but have never been accurately measured). The sequence rests on sediments of Palynozone Pb, of late Pliocene age, and is covered by deposits of Palynozone QII, of later early Pleistocene. The nature of both contacts, known only from drillings, is not clear but seems conformable. The Melekh Sedom Sands are known from numerous boreholes in the Jordan Valley, from the Hula basin where some 400 m were penetrated, down to the southern Dead Sea, of varying thicknesses, but always in the order of several hundred meters (Levin & Horowitz 1987). Southward, in the northern Arava east of the international border, the Melekh Sedom Sands are exposed, overlying the Arava and occasionally the Hazeva formations, attaining some 300 m in the Hazeva 5 borehole (Eyal 1984). These outcrops (Fig. 5.2.15) are marked as "Hazeva" on the geological maps (Fig. 3.2.4), erroneously in my opinion.

Levin (1985) subdivided the sequence at the type borehole into three units. The "Lower Shale Unit" comprises 430 m of carbonaceous shales, partly silty or sandy, occasionally limonitic. The overlying "Sandy Unit" is some 350 m thick, consisting mainly of sand, alternating with sandy or carbonaceous shales in the lower and upper parts, almost pure sand at the middle 180 m. The sand is fine to medium, poorly sorted, almost exclusively rounded to sub-angular quartz grains. This "Sandy Unit" is the only part of the sequence referred to by Levin as "Melekh Sedom Sands". The uppermost 320 m, the "Upper Shale Unit", resemble very much in their composition the lowermost unit. To the north, in the central Jordan Valley and the Hula, the clastics are mainly conglomerates, with interfingering basalt tongues and possibly some paleosols.

The Melekh Sedom Sands are very rich in palynomorphs, discussed in Chapter 6, indicating an early Pleistocene (Palynozone QI) age and a fluvio-lacustrine

Figure 5.2.15. Melekh Sedom Sands at the Transjordanian sector of the northern Arava.

environment of deposition. Weinberger (1991), who made a detailed petrographic study of the Sands, concluded a braided stream depositional environment, while the finer sediments were deposited in a lacustrine domain.

5.2.3.2 *HaMeshar Formation*

Author: Garfunkel & Horowitz (1966); amended by Avni (1998).

The sequence is named after the HaMeshar (flatland) Valley in the central Negev, where it was first recognized and its type section located. It was previously known by various names which were, however, not specific to this unit alone, comprising also other gravel beds, such as the “Quaternary–Neogene Hammada” (Bentor & Vroman 1957, Picard 1959), the “Alluvium and undivided Neogene–Quaternary Series” (Shaw 1948) and so on. The HaMeshar Formation overlies the Pliocene Arava with a slight unconformity and also a variety of older formations, its upper surface usually exposed and eroded by all wadis leading to the Dead Sea (Garfunkel & Horowitz 1966; Horowitz 1974; Horowitz 1979, pp. 115–118). The Formation is known from the entire southern Negev, where it covers considerable areas, usually by a veneer several meters thick. At the type locality, a rather deep channel of the drainage system in which the HaMeshar was deposited, the sequence attains several tens of meters. However, Avni (1998) considers the lower part of this section as the upper part of the Arava Formation, assigning what is left a new name, the Zehiha Formation, based on erroneous correlation with the Nahal Zihor lake sediments (see Section 11.2.5.1).

The HaMeshar Formation (Fig. 5.2.16) consists almost entirely of conglomerates, whose flint, limestone and dolomite pebbles are derived from the close vicinity. Rare magmatic, metamorphic and quartzite pebbles are only found close to outcrops of the Arava Formation, from where they were redeposited. The local character of the HaMeshar is particularly conspicuous when compared with the underlying Arava and Hazeva formations. The size of the constituents varies from sand up to several tens of centimeters, sorting is poor, the pebbles are angular to sub-rounded, and often the angular and sub-angular predominate. The matrix is loose, comprising carbonate rich silts or clays, occasionally of a reddish hue. Sandy horizons within the gravel layers are made of medium to coarse flint and limestone grains, with subordinate quartz, loosely cemented by a carbonate or clayey matrix. Clay and chalk horizons occur in places.

The Formation is made up of two principal sedimentary facies, deposited either in ancient channels or on floodplains. The former are thicker, comprising better-rounded gravel, occasionally bearing fine-grained layers indicating a transition to

Figure 5.2.16. HaMeshar Formation at the confluence of Nahal Hiyyon and the central Arava.

temporary lacustrine conditions (Avni 1998). The floodplain deposits comprise rather thin, extensive sedimentary sheets, typified by angular to sub-angular components. Fossils have never been found, and the early Pleistocene age assigned by Garfunkel & Horowitz is based on the stratigraphic position, since the HaMeshar overlies the Pliocene Arava, while cut by the wadis leading to the Dead Sea, and thus predates the main Levantine rifting.

Garfunkel & Horowitz (1966), followed by Horowitz (1979, p. 117), reconstructed the drainage system and concluded that it led to the Mediterranean, without any notable depression in the Arava at that time. Conversely, Avni (1998) states that the drainage system led to the Dead Sea.

5.2.3.3 *Ar Risha Gravels Formation*

A sequence of some 80 m of poorly cemented fluviatile gravels is described by Ibrahim (1993) from the area of Ar Risha, east of the central Arava (Fig. 5.2.17), termed the Ar Risha Gravels Formation. The gravels cover unconformably late Cretaceous rocks, cropping out in a narrow, 25-km long strip along the eastern Arava, affected by faulting, gently dipping to the west. Their upper contact constitutes the present-day erosion surface. The stratigraphic position is not clear: Ibrahim (1993) suggests they are of early to middle Pleistocene age, while Miocene is suggested by Sneh et al. (1998a).

Figure 5.2.17. Ar Risha Gravel at Jebel Ar Risha, east of the central Arava.

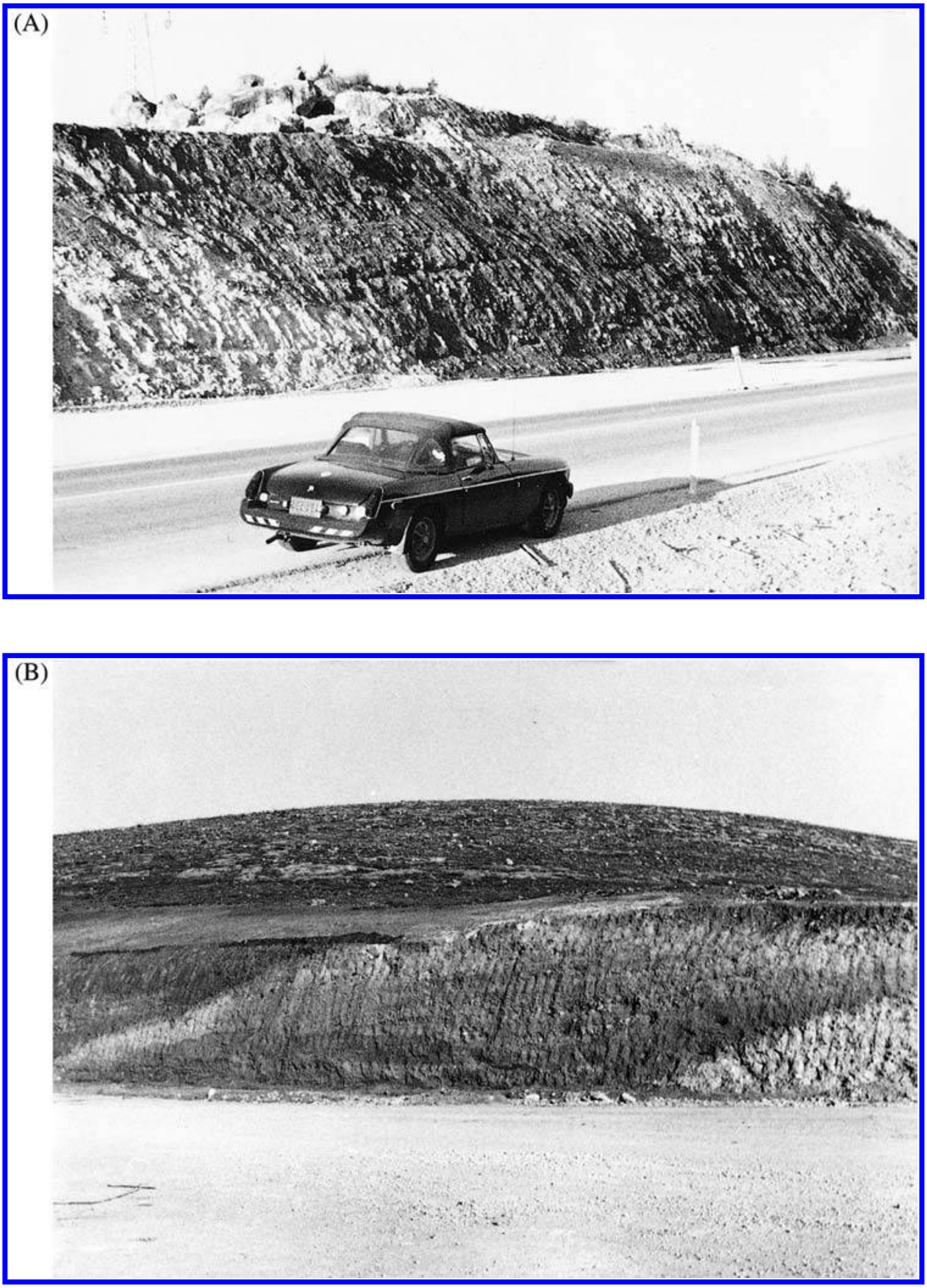


Figure 5.2.18. Channel fill of the Bethlehem Conglomerate, in a roadcut: (A) west of Jerusalem, (B) east of Jerusalem.

5.2.3.4 *Bethlehem Conglomerate*

Author: Gardner & Bate (1937), "Bone Bearing Beds of Bethlehem"; amended by Horowitz (1974).

The unit is named after the city of Bethlehem, south of Jerusalem, where during the 1930s Gardner, Bate and Stekelis had excavated what they described a "solution pothole" (Fig. 12.1), finding numerous bones and questionable artifacts (Stekelis 1940). Solomonica (1948) and Horowitz (1974) recognized that this was not a local unit but rather a regional occurrence left by a fluviatile system. The type section is at Bethlehem, where some 10–12 m of this unit have been excavated.

The fluviatile sediments usually fill erosion channels cut into the late Cretaceous formations of the Judean hills (Fig. 5.2.18), but in places are also incised into early Tertiary rocks, or a late Neogene erosive surface (Horowitz 1979, p. 118). The top is hardly ever covered, except by soil. The Bethlehem Conglomerate is exposed in many localities on the crest of the Judean hills, at elevations of approximately 800 m above sea level, but also on both flanks of the watershed at lower altitudes (Horowitz 1979, p. 119). To the west it is known down to some 200 m above the present-day Mediterranean, where the Conglomerate overlies the Pliocene Pleshet Formation and is termed the Ahuzam Formation (Buchbinder & Sneh 1983–1984). Thicknesses in the centers of channels may attain 15–20 m.

The sequence comprises conglomerates and gravel embedded in red and green clays, talus and flint agglomerates, with components ranging in size from small, well-rounded pebbles to angular blocks over 1 m across. Some gravel horizons are cemented by a hard carbonate matrix. The clayey matrix comprises chiefly kaolinite, with up to 1% titanium, indicating a rather long transport and partly basaltic provenance, most probably east of the Jordan Valley, since no such rocks are known west of the Rift in Judea. Mammal bones unearthed in the Bethlehem "pit" (Hooijer 1958) include *Archidiskodon planifrons*, *Leptobos*, *Giraffa* cf. *camelopardalis*, *Felis*, *Hippopotamus*, *Bos*, *Hipparion*, *Rhinoceros* cf. *etruscus*, *Stegodon* and *Elephas*. Among other species, turtle bones are known from Bethlehem, while crocodile remains were found at Mount Scopus, in northern Jerusalem. These fossils led to the generally accepted view of a Villafranchian age for the Bethlehem Conglomerate. An occurrence of alluvial sediments which looks just like the Bethlehem Conglomerate was termed the Dhuleil Formation (Baubron et al. 1985), cutting into the Transjordanian highlands, sandwiched between basalts 3.5–5 and 2.3–3 Ma old (Fig. 5.2.19).

5.2.3.5 *Mahanayim Marl*

Author: Heimann (1990).

The unit was named after a nearby settlement, and the type section is at the Ayyelet HaShahar quarry (Fig. 5.2.20), about 2 km north of Mahanayim. The

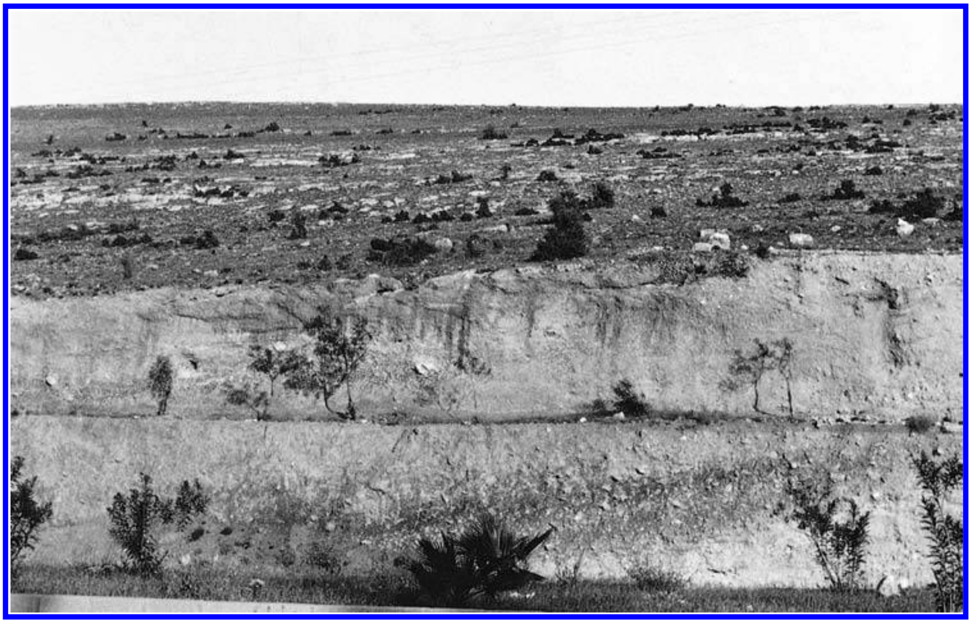


Figure 5.2.19. Dhuleil Formation, some 60 km north of Amman on the main road to Irbid.

Figure 5.2.20. Mahanayim Marl, at the base of Ayyelet HaShahar quarry, southern Hula Valley.

base is not exposed, the bed is overlain, probably conformably, by the Ruman Basalt. The outcrop at the quarry is the only one known from the Hula Valley, only 3 m thick. It comprises gray marl, chalk and clays, rich in *Melanopsis* shells. The overlying basalt yielded K–Ar ages of 2.3 Ma at the base, up to 1.6 Ma at its top.

Based on these ages, Heimann correlated the Mahanayim and the basalt with a sequence drilled 11 km to the north, in the Notera 3 borehole, at depths of 1,000–1,600 m, which overlies Pliocene deposits (see Chapter 6), bearing pollen spectra of Palynozone QI (Horowitz & Horowitz 1985).

5.2.3.6 *Amud Conglomerate*

Author: Kafri & Heimann (1994).

The unit was defined in and is named after Nahal Amud, west of Lake Kinneret. It overlies unconformably late Cretaceous or early Tertiary rocks, filling a channel system which cuts through the late Miocene Herod Conglomerate. The top is exposed, in places turned into hard caliche crust (Nari), or covered by some soil. Affected by subsequent erosion, the Amud Conglomerate occurs in lens-shaped bodies several meters thick, tens of meters long, in a few localities along Nahal Amud (Fig. 5.2.21).

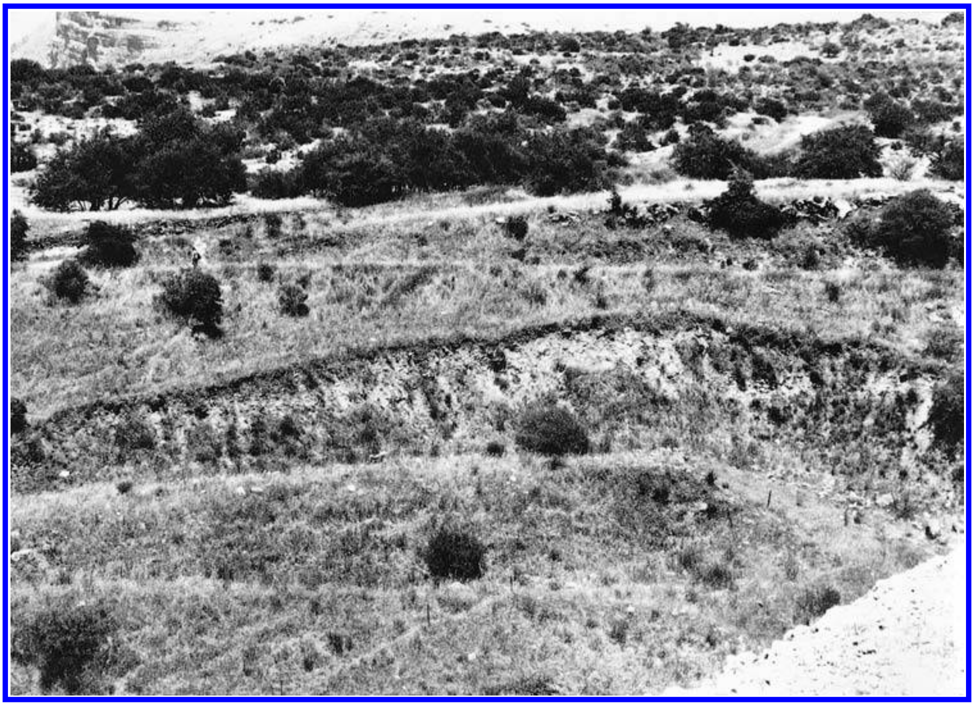


Figure 5.2.21. A high terrace of the Amud Conglomerate in Nahal Amud, some 10 km west of the northwest corner of Lake Kinneret.

The Conglomerate is made of rounded to sub-angular polymictic pebbles, from a few up to 20 cm across, mostly derived from nearby Cenomanian–Turonian carbonate rocks. Weathered basalt pebbles also occur, diminishing in number as one goes WNW, derived from the Amud Basalt, yielding K–Ar ages of 2.7–2.3 Ma. The matrix is a slightly reddish chalk, which forms the caliche crust on the surface.

The Amud Conglomerate fills a drainage system that flowed in a general WNW direction toward the Mediterranean Sea, most probably at the same time or slightly younger than the Amud Basalt. Other conglomerate horizons of similar appearance, located some 15–20 km to the north, in which artifacts maybe embedded, are covered by the Dalton Basalt which yielded an average age of 2.43 Ma (Ronen 1996).

5.3 MAGMATISM

Late Cenozoic volcanism is quite common in the vicinity of the centrl and northern Jordan Rift Valley, active almost continuously since the beginning of the Miocene up to the present day. It seems, however, that this part of the Rift crosses only a corner of the large El Shamah-Jebel Druze (also known as Harrat ash Shaam) volcanic field, where eruptions commenced as early as the Oligocene (Tarawneh et al. 2000), extending over southern Syria, eastern Transjordan and northern Saudi Arabia (Figs 5.3.1 and 5.3.2). Along the rest of the Jordan Valley, to the south, only rare, restricted magmatic occurrences are known, mostly concentrated east of the Rift. Intrusions, usually small scale, are found all over the southern Levant, in most cases tied up with corresponding phases of volcanic activity. These are mentioned in the relevant context.

Petrographically, all Cenozoic volcanics of the Near East (except for rare xenoliths), including lavas, intrusions and pyroclastics, comprise alkali olivine basalts (Garfunkel 1989). This was observed in the Lower Galilee and central Jordan Valley (Bentor 1946, Schulman 1962, Oppenheim 1962); in Transjordan (Bender 1968a, Boom 1968); in southwest Syria (Dubertret 1929, 1966, Wolfart 1966, Ponikarov et al. 1967); and more specifically in the Golan (Michelson 1972, Mor 1973, 1986, Lang et al. 1979, Brenner 1979).

5.3.1 The Red Sea Dike System

The Red Sea Dike System (Fig. 5.3.5) intruded over an area several hundred kilometers wide, on both sides of the Gulf of Suez and Red Sea (Garfunkel 1989). Often these dikes did not penetrate the sedimentary cover, while in some locations, such as near the Gulf of Suez, several sills and flows also belonging to this system are known (Bartov 1994). The magmas have tholeiitic affinities, yielding K–Ar ages in the range of 25–19 million years (Steinitz et al. 1978).

Close to the Jordan Rift Valley only two occurrences of this system are known. A vent system in Nahal Ashosh, some 60 km south of the Dead Sea at the western rim

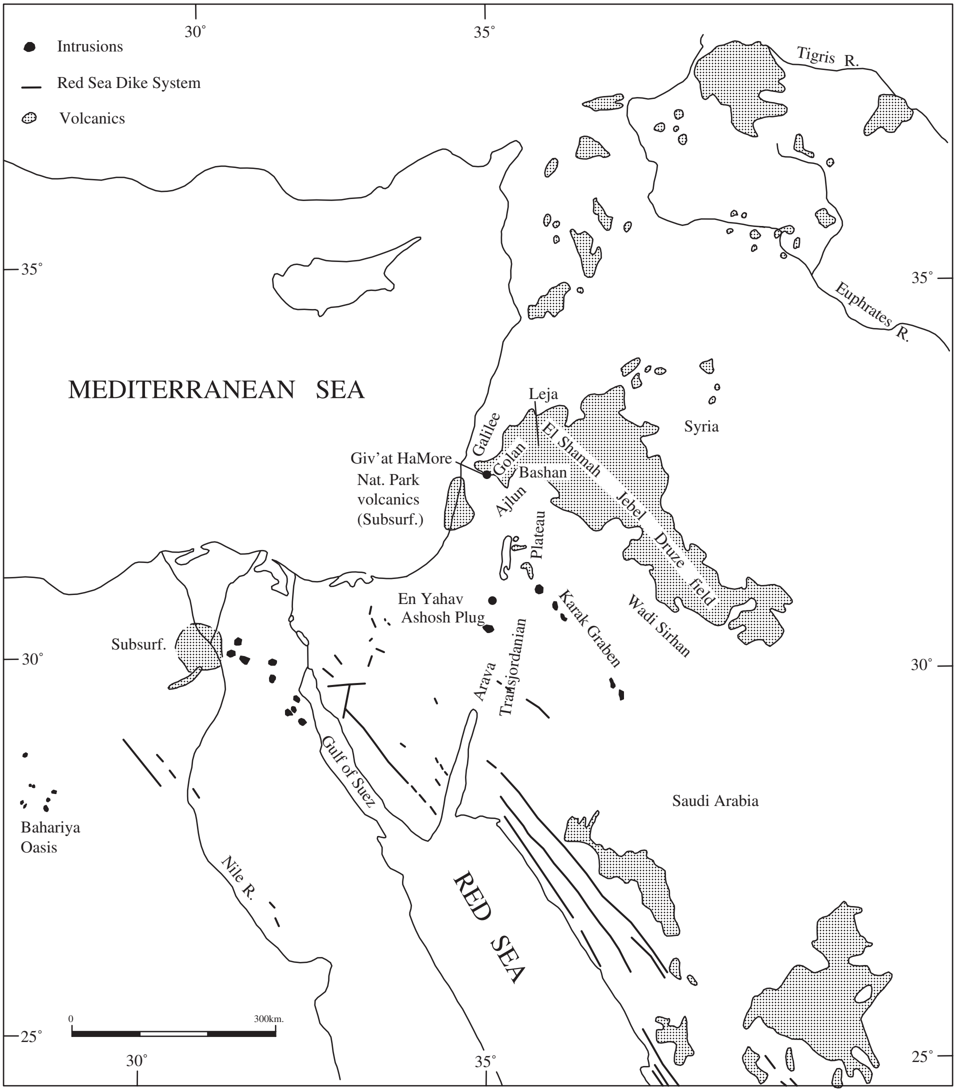


Figure 5.3.1. Distribution of magmatics in the southern Levant, after Garfunkel (1989), Bartov (1994) and Sneh et al. (1998a).

of the Arava (Fig. 5.3.3), comprises a few small plugs of basalt with gabbro xenoliths and pyroclastics, intruding middle Eocene rocks (Levitte 1966), and yielded K–Ar ages of approximately 21 Ma (Steinitz et al. 1978). Several basalt plugs exist in the Karak graben (Fig. 5.3.4) east of the Lisan Peninsula (Bender 1974a, p. 105), one of which intrudes the lower part of the Dana Formation, unfortunately not yet dated (Macumber & Edwards 1997). Others yielded K–Ar ages of 18.3 ± 0.9 and 18.9 ± 0.8 Ma (Barberi et al. 1980).

Figure 5.3.2. A general view of the Harrat ash Shaam volcanic field, northeastern Transjordan.

Figure 5.3.3. The largest plug at Nahal Ashosh, some 5 km west of the northern Arava.

Figure 5.3.4. One of the intrusions in the Karak graben, some 15 km east of the southern Dead Sea.

Figure 5.3.5. A dike of the Red Sea System, intruding Eocene rocks in central Sinai.

5.3.2 Lower Basalt

Author: Blake (1928), "Lower Basalts"; amended by Schulman (1962).

The term was applied by Blake and Schulman to a suite of basalts occurring in the Lower Galilee and the central Jordan Valley (Fig. 5.3.6), underlying or interfingering a variety of Neogene sediments. No type section was proposed, but many reference sections were described and several names suggested for the sequence. Picard (1928) called both the Lower and Cover basalts "Pliozän Basalt"; Bentor (1946), "Miocene Basalt"; Schulman (1959), "Lower Volcanics"; Oppenheim (1959), "Miocene Lavas". Shaliv (1991) and others retained the name "Lower Basalt", which is in wide use today.

The base of the Lower Basalt, rarely exposed, unconformably overlies Cretaceous through Eocene strata west of the Jordan Valley. To the East, near En Gev, it overlies the Oligocene formations (Michelson 1972). The Basalt interfingers with the Herod Formation, and is covered by a variety of younger formations, the oldest of which is the Umm Sabune Conglomerate (Schulman 1962).

The exposed Lower Basalt sequence attains some 450 m in thickness, in Belvoir, but the base there is buried. Drillings in the area proved the occurrence of thicker sections; the thickest sequence penetrated in a borehole, in the center

Figure 5.3.6. The Lower Basalt sequence of flows at Belvoir, facing the central Jordan Valley.

of Yizre'el Valley, is some 650 m, again with the base not reached (Shaliv 1991). These figures can, however, vary considerably over short distances. In places the Basalt occurs as a solid body of flows (Fig. 5.3.6), while to the north, near Tiberias, it appears as several tongues interfingering with the Herod Formation sediments (Fig. 5.1.1). Three or four flows, each 5–15 m thick, are known east of Lake Kinneret, near En Gev, while to the south no occurrences are known on the eastern flank of the Jordan Valley. The Lower Basalt also occurs in the eastern and central parts of the Yizre'el Valley, occasionally displaying considerable thicknesses. Basalts of similar stratigraphic position are also known from eastern Transjordan (Bender 1974a, p. 105) and southern Syria (Ponikarov et al. 1967). The present-day volume of the Lower Basalt west of the Jordan River is estimated at about 200 km^3, the original volume at about 300 (Shaliv 1991).

The petrography of the Lower Basalt was studied, among others, by Bentor (1946), Oppenheim (1959), Schulman (1962), Dicker (1969) and Brenner (1979). The Lower Basalt comprises alkali olivine basalts of various textures. It contains olivine phenocrysts of different sizes, their outer layers usually weathered to idingsite, occasionally the entire crystals, but at least in part maintaining their idiomorphic forms. The basalts also commonly contain phenocrysts of pyroxene, sometimes partly replaced by chlorite, and of plagioclase. The matrix is fine to medium, comprising principally plagioclase needles, surrounding the olivine, pyroxene and ore minerals crystals, in a trachytic texture. In some of the samples flow textures are common, while cracks and vacuoles are usually filled up with calcite crystals, and occasionally by zeolites.

The field appearance of the Lower Basalt is quite typical. In the Belvoir outcrops it comprises several tens of lava flows, 10–50 m thick, each topped by aa clinker. Paleosols between basalt flows are very rare. The alternating flows and aa clinker give the slopes a "bedded" appearance. Further north, in the Yavne'el area, the clinker tops are less abundant, but some paleosols occur. In this locality at least two occurrences of limestone lenses containing *Dreissensia* shells are reported in Schulman (1962, p. 12). Further northward, at the Poriyya escarpment, paleosols are thicker and more common, but the sequence is limited to six Lower Basalt flows which interfinger with the Herod Formation sediments. The Lower Basalt, in outcrops, is typified by very severe weathering, which accentuates the numerous fissures, filled by calcitic clay minerals, an appearance which led to the term "rotten basalt". Pyroclastic rocks are quite rare in the Lower Basalt sequences. Only some thin lenses of tuffs are known from Poriyya (Bentor 1946), another lens in Belvoir (Schulman 1962) and several meters of pyroclastics found in the latter locality by Shaliv (1991).

Schulman (1962) considered that the Lower Basalt flows were fed and erupted through fissures. Dikes which maybe responsible for that are known mainly from the Yizre'el Valley and the Lower Galilee, where they are 100–1,400 m long, appearing in two systems, one which strikes WNW, the other SW. The two systems cross each other, and seem synchronous (Dicker 1969). In Giv'at HaMore, located in the center of the Yizre'el Valley, Oppenheim (1962) and Dicker (1969) found at least four large-scale intrusions, in the form and size of stocks. Relics of volcanoes connected with Lower Basalt effusions are quite rare, but several possible candidates are mentioned in Shaliv (1991) east of Belvoir, in Giv'at HaMore and in Ramot Menashe, further west (Avnimelech 1948).

The rare fossils occurring in beds and lenses within the Lower Basalt are not indicative of age. Before the advent of radiometric dating, all investigators used stratigraphic criteria to date the Lower Basalt, by the synchronous, interfingering Herod Formation. Overlying the Eocene and Oligocene, topped by Pliocene sediments, the most reasonable guess was Miocene for both. Bentor (1957) defined the age of the Lower Basalt as middle Miocene; Schulman (1962), as Miocene; Freund et al. (1965), as middle to late Miocene, based on paleomagnetism. Radiogenic ages for the Lower Basalt were first published by Siedner & Horowitz (1974), followed by Steinitz et al. (1978), Shaliv (1991) and others, all indicating a middle Miocene age (for details see Chapter 7) for the main body of these volcanics, in the range of 17.5–12 Ma. Although Shaliv considers also volcanics younger than 10 Ma as part of the Lower Basalt, it seems that these constitute a separate phase, discussed below.

5.3.3 Late Miocene Volcanics

A few limited occurrences of volcanics (Fig. 5.3.7) with ages clustering around nine to six million years ago, constitute this subordinate group that, in my opinion, should be regarded separately from the Lower Basalt, since the latter was originally defined as interfingering and coeval with the Herod Formation. Shaliv (1991) reports a flow which interfingers with the Umm Sabune Conglomerate at its type section, yielding 8.4 ± 1.2 Ma (Fig. 5.2.1). At the same locality, the Umm Sabune overlies a basalt sheet 10.1 ± 0.3 Ma old; a small intrusion at Giv'at HaMore, in the central part of the Yizre'el Valley, is 9.9 ± 0.3 Ma old; some other volcanic occurrences across the Yizre'el Valley yielded ages within this range, while a somewhat younger one, 8.5 ± 3.7 Ma, was dated east of Lake Kinneret, underlying the "Cover Basalt".

The lower flows of the plateau basalts exposed on the Transjordanian highlands, east of the Dead Sea, also belong to the late Miocene phase, with ages of 9.3 ± 0.2 and 8.8 ± 0.2 Ma (Steinitz & Bartov 1991). Other volcanic rocks of this age group are known from several boreholes in the Jordan Valley (see Section 5.4.4).

5.3.4 Intermediate Basalt

Author: Schulman (1962), "Intermediate Lavas".

The name "Intermediate Lavas" was originally given in Hebrew, but when translated into English became "Intermediate Basalt", by which this unit is known in the literature. The term was applied by Schulman to two basalt flows, intercalating with the Bira Formation sediments in the section at Belvoir, after he realized that

Figure 5.3.7. Late Miocene Basalt (under the electric pole), overlying the lowermost part of the Bira Formation at Marma Feiyad, central Jordan Valley.

his earlier (1959) assignment of these flows to the Lower Basalt was wrong. No "official" type section is given.

No other lower or upper contacts of these flows could be outlined, since by definition they should intercalate with the Bira sediments. However, Shaliv (1991) correlated several occurrences of basalts with this unit, based solely on radiometric ages. Flows of the Intermediate Basalt are known from almost all localities where the Bira is exposed or penetrated by boreholes. The flows, usually between one and four, are never very thick, commonly in the order of 10–15 m or less. Basalt flows interfingering with the Gesher Formation in the Golan Plateau were also assigned to the Intermediate Basalt (Michelson 1972), based on the assumption (accepted by Horowitz 1973) that the lower part of the Gesher in the Golan is correlative to the Bira Formation of the central Jordan Valley (see Chapter 7).

The flows are distinctive, different from the older Lower or younger Cover basalts, by their petrography and appearance. They are very rich in zeolites, occurring both in the matrix and as cavities fillings. Most of the rocks comprise black, highly alkaline olivine melabasalts. The tops of the flows are smooth, glassy, highly oxidized, in contrast with the aa clinker tops of the Lower and Cover basalts, indicating underwater solidification. Some differences are seen in flows of the Intermediate Basalt from one outcrop to another, and also their setting

Figure 5.3.8. The Intermediate Basalt sequence of flows near Yavne'el, west of the central Jordan Valley.

within the Bira is not the same, so Schulman concludes that correlations of particular flows are impossible.

The age of the Intermediate Basalt flows is that of the Bira sediments. Most investigators regarded it as Pliocene, but radiometric datings led Shaliv (1991) to suggest a late Miocene age. A detailed discussion of the ages and relations of the Intermediate and Cover basalts is presented in Section 7.1.5. If the attitude presented there is accepted, most occurrences described below for the Cover Basalt should indeed come under the current heading, of Intermediate Basalt (Fig. 5.3.8).

5.3.5 Fajjas Tuff

Author: Schulman (1962), following Picard (1932), "Basalttuffe des Altdiluvium".

The Formation is named after Wadi Fajjas (sometimes written "Fejjas" or "Fejas"), Nahal Yavne'el in Hebrew, where the type section is located, some 8 km west of the southernmost tip of Lake Kinneret. The Fajjas Tuff interfingers with the Gesher Formation sediments conformably, usually lying on the "Main Oolithic" ("f2") Member, but occasionally overlies the Bira Formation or the Lower Basalt. It is overlain by the Cover Basalt, but the exact nature of contact is not clear.

Only a small number of outcrops of the tuff is known, most of them located in the northern part of the western margins of the central Jordan Valley (Fig. 5.3.9). Several occurrences of pyroclastics within the Gesher sediments, attributed to the Fajjas (Horowitz 1973, 1974), are known north of Lake Kinneret, in the Buteiha and Korazim regions. The best developed, longest sequence crops out where the type section is located, attaining more than 50 m in thickness. The Fajjas consists principally of brownish–grayish pyroclastics, deposited in a lake concurrently with the Gesher Formation, showing graded bedding. The pyroclastics comprise palagonitic and lapilli tuffs, usually considerably weathered, and agglomerates containing volcanic bombs and pieces of the country rocks. Several subordinate basalt flows accompany the tuffs. Some are composed of granular olivine melabasalts, rich in ultramafic, plutonic xenoliths, while others consist of scoria, with volcanic bombs.

A vent connected with the Fajjas Tuff, through which at least some of the pyroclastics were ejected, was found just west of the prehistoric site at Ubeidiya (pers. obs.). The stratigraphic relations of the pyroclastics and flows are not always clear. Some of the latter may actually belong to the Cover Basalt (Schulman 1962). The age of the Fajjas Tuff, besides radiometric datings discussed in Chapter 7, is similar to that of the Gesher Formation, for which most students agree on Pliocene, but disagree on what part of it. Shaliv (1991), based on what he considers the similarity of radiometric ages, suggests the Fajjas is a pyroclastic member of the Cover Basalt.

Figure 5.3.9. Fajjas Tuff, near the southwestern tip of Lake Kinneret.

Conversely, if the attitude presented in Section 7.1.5 is accepted, the Fajjas belongs to the Intermediate Basalt.

5.3.6 Cover Basalt

Author: Schulman (1959), following Blake (1928), "Upper Basalt Flows".

The name, first suggested by Picard (1936), originated from the supposed considerable geographic extent of this unit in the southern Levant, and from its morphotectonic setting, covering wide areas on both sides of the northern part of the Jordan Rift Valley, frequently forming the top flat layer. The Cover Basalt was mentioned by Lartet (1869), and since then has been given a variety of names by numerous investigators, usually implying its plateau nature. The type section is near Khirbet Jabul, where the sequence is best developed west of the Jordan River. Mor (1986) distinguished two members of this Formation in the Golan, the very abundant Cover Basalt, accompanied by subordinate pyroclastics termed the Bnei Yehuda Scoria.

The lower contacts of the Cover Basalt are always unconformable, over an erosional and occasionally taphrogenic relief, overlying a variety of older rocks, of which the youngest is the Gesher Formation (Fig. 5.3.10). Its upper surface is also always erosional, covered by soils and in places by Quaternary deposits.

Figure 5.3.10. Cover Basalt, overlying the Gesher Formation at Poriyya, near Lake Kinneret.

The Cover Basalt is known from large areas in southwestern Syria (Dubertret 1933), Transjordan (Daniel 1963; Bender 1968a; 1974a, p. 105) and northern Israel (Picard 1936, Mor 1986, Kafri 1997). The unit was also penetrated by several boreholes, discussed in Chapter 6. The thickness varies, and is usually several tens of meters, attaining up to 150 in places, as does the number of individual flows in the sequence, up to a maximum of 10 (Schulman 1962).

The mineral composition of the Cover Basalt resembles very much the Lower Basalt, comprising granular or ophitic olivine (or idingsite) basalts, occasionally grading to plagioclase basalts, rarely porphyritic. The usual thickness of individual flows is 15–20 m, occasionally topped by paleosols. The flows are quite similar to one another. The base is glassy, irregular, with a typical "sole" up to 1 m thick, made of massive rock with a small number of spherical vesicles. Above is 1–4 m of thick massive rock, with no vesicles at all but with well-developed sheeting, grading to columnar basalt, which takes up more than half of the flow's thickness. The columns are 20–120 cm across, increasingly vesicular upward. The top is typically horizontally jointed.

The Cover Basalt was most probably fissure erupted, which is evidenced by the vast areas and the lack of conspicuous volcanoes connected with this phase west of the Jordan River (Schulman 1962). Mor (1986) suggested that the western part was indeed formed by fissure eruptions through what are now dikes, while the eruption of the Cover Basalt in the Golan Plateau is connected with several rather flat, large cones. It seems that the cones described by Mor could not be the sole source, and were accompanied also by fissure eruption. Garfunkel (1989) suggested that the Golan is a very large, rather flat shield volcano. Heimann (1990) described three volcanic cones from the Korazim block, connected with lava flows of the Cover Basalt. Intrusive rocks connected with the Cover Basalt were penetrated in the Zemah 1 borehole, just south of Lake Kinneret (Marcus & Slager 1985). The drill went through eight rather thick occurrences of gabbro, from 1,740 down to 3,994 m, the thickest attaining some 200 m. Several thinner horizons were also penetrated. The age given for the gabbro, 3.8 ± 0.4 to 2.88 ± 0.11 Ma, seems to indicate its connection with the Cover Basalt, itself occurring between 486 and 1,184 m at this borehole, with an age of 4.48 ± 0.3 Ma.

The age is problematic: the Cover Basalt should, as defined, be younger than the Gesher Formation, but the age of the latter is not certain; the basalt is hardly ever overlain by any sediments in outcrops, and if it is, unconformably; it is quite difficult to define the Cover Basalt safely in boreholes, where it may occur sandwiched between dated sediments. Radiometric ages (Heimann et al. 1996) are in the range of 5.5 million years for the base, through 3.3 for the top. These may be problematic, since other authors have published different figures (Siedner & Horowitz 1974, Mor 1986). The problem increases due to the similarity of various basalts, making field definitions of what exactly belongs to the "Cover" quite risky.

A detailed discussion of the ages and relations of the Intermediate and Cover basalts is presented in Section 7.1.5. If the attitude presented there is accepted, most occurrences described here for the Cover Basalt should indeed come under the former heading, of Intermediate Basalt.

5.3.7 Dalton Basalt

Author: Mor et al. (1987).

The Dalton Basalt is named after the Dalton Plateau in the eastern Upper Galilee, where its outcrops were described by Schlein (1961) and mapped by Glikson (1964). The flows overlie unconformably a variety of older rocks, but possibly conformably a suite of conglomerates and paleosols (Fig. 5.3.11), in which artifacts

Figure 5.3.11. Dalton Basalt, overlying a red paleosol near Kibbutz Yir'on, west of the Hula Valley. Photo courtesy of A. Ronen.

are reported by Ronen (1996). The top is usually eroded, exposed, and in several localities is covered by various paleosols, some of which contain a variety of Paleolithic sites.

The Dalton Basalt occurs at the eastern Upper Galilee and slopes toward the Hula in patches, with several cones from which some of the alkali olivine basalt flows erupted, yielding ages of 2.7–1.7 Ma, which make them correlative with the Ruman Basalt of the northern Korazim area.

5.3.8 Amud Basalt

Author: Kafri & Heimann (1994).

The term "Amud Basalt" was applied to several rather small outcrops of flows, a dike and a few epiphyses of alkali olivine basalt in Nahal Amud (Fig. 5.3.12), west of Lake Kinneret. These overlie unconformably a variety of late Cretaceous or early Tertiary rocks, the dike cuts through and "bakes" the late Miocene Herod Conglomerate. K–Ar ages for these occurrences are within the range of 2.5–2.2 Ma, thus they seem to be part of the same event as the Dalton Basalt to the north and the Ruman Basalt, northeastward.

Figure 5.3.12. Amud Basalt in Nahal Amud, some 10 km west of the northwest corner of Lake Kinneret.

5.3.9 En Yahav dike

Besides the Ashosh plug (above), this is the only other magmatic occurrence close to the Arava. Volcanics on the Transjordanian Plateau east of the Arava are at least 20 km from the Rift Valley. The dike, accompanied by a caldera with pyroclastics and volcanic bombs, is situated on the western rim of the Arava (Fig. 5.3.13), some 50 km south of the Dead Sea, and is east–west oriented (Bentor & Vroman 1957). The dike intrudes Eocene rocks and seems to be younger than the Hazeva Formation, which was affected by the explosions (Steinitz et al. 1978). The rock was determined as tanbushite (Levitte 1966). Five K–Ar point ages for the En Yahav dike (Steinitz et al. 1978) range between 8.8 ± 1.7 and 3.9 ± 0.5 Ma, while an isochron yielded an age of 2.7 ± 0.6 Ma, a figure preferred by the authors. Recently Steinitz et al. (2000) redated the dike by the $^{40}Ar/^{39}Ar$ method, which yielded ages of 6.36 ± 0.22 and 6.42 ± 0.36 Ma. I leave the reader to decide which age is reliable.

5.4 IN-RIFT MAGMATICS

Almost all investigators agree that the Levantine rifting postdates the Cover Basalt (but see discussion in Section 7.1.5), as originally defined (Schulman 1962,

Figure 5.3.13. En Yahav dike cutting through Eocene rocks near En Yahav, just west of the northern Arava.

followed by many, and see also Chapter 9). Thus, all occurrences of the Cover and older basalts are not treated here, although all but the Red Sea Dikes are also found within the Jordan Rift limits. Of the volcanics mentioned below, discrimination is made between lavas erupted out of the Rift limits but flowed down along river valleys, reaching the depression, and magmatics actually intruding or erupting within the Valley itself. However, most of the in-Rift volcanism is younger, and is thus treated in Chapter 11.

5.4.1 Mechki Basalt

Author: Mor (1986), following Dubertret (1952) and Picard (1963). The name is occasionally spelt "Meshki".

The unit is named after a village in southern Lebanon, the location Mor believes is the source of this basalt. Picard named the southern termination of the flows in the northern Hula Valley "Hasbani Basalt". Mor, following radiometric datings, distinguished between the older Mechki and a younger Hasbani Basalt *s. str.* (see below). No type section is given.

The flows of Mechki Basalt follow the Hasbani River (Nahal Snir) southward (Fig. 5.4.1), for a distance of some 36 km, attaining a maximum width of some 2 km close to the source, unconformably overlying a variety of older rocks. It is overlain, north of the Hula Valley, by the Hasbani Basalt. The rocks comprise alkali olivine basalts, with no pyroclastics, yielding (Mor 1986) radiogenic ages in

Figure 5.4.1. Mechki Basalt, just north of the Hula Valley.

the range of 2.5–3.0 Ma. Dubertret (1952) indicates that the basalt is synchronous with "Neogene" conglomerates (possibly the Tanur Conglomerate) in the Beqa'a region, near Hasbaya, assigning it a Pliocene age.

5.4.2 Ruman Basalt

Author: Heimann et al. (1987).

Ruman Basalt occurs at the central and northern parts of the Korazim block, named after Tel Ruman, a volcanic cone which is thought to be the source of the flows. The Ruman Basalt overlies the Cover Basalt and is overlain by the Gadot-Hazor Formation (Fig. 5.4.2), with both contacts unconformable. Many authors considered this unit as the upper part of the Cover Basalt (Picard 1963, Schulman 1967, Horowitz 1973).

Heimann dated the Ruman at between 2.9 and 1.6 Ma. Several outcrops in the southwest corner of the Hula Valley also belong to this phase, but may have erupted from a different source. Basalts at the Golan escarpment opposite Korazim and the southern Hula, which yielded ages in the range of 2.5–2.0 Ma, probably also belong to the Ruman phase. According to Heimann (1990) the Ruman Basalt extends eastward to the Golan, where it covers considerable areas down to Lake

Figure 5.4.2. Ruman Basalt at Ayyelet HaShahar quarry, southern Hula Valley. Arrow indicates overlying Gadot-Hazor sediments.

Kinneret. However, Mor (1986) placed these Golan basalts in the Ortal Formation, 1.61–0.79 Ma old, which is somewhat younger than the Ruman. The Dalton Basalt occurs in the eastern Upper Galilee and slopes toward the Hula in patches, with several cones from which some of the flows erupted (Mor et al. 1987), yielding ages of 2.7–1.7 Ma, which make them correlative with the Ruman Basalt.

5.4.3 Grain Sabt

Grain Sabt (also known as "Zahrat el Qurein") is a supposedly diapiric dome, situated some 25 km north of the Dead Sea, close to the Jordan River, in the Ghor el Qatar area (occasionally written "Ghawr el Katar"). A small basalt sheet, less than half a square kilometer, covers unconformably an outcrop of Ghor el Qatar clastics (Fig. 5.4.3), and is overlain by Lisan sediments over an erosional relief (Bender 1974a, p. 105, Belitzky & Mimran 1996). The exact relations of the Basalt and Ghor el Qatar Series is not clear, since in another publication in the same year Bender (1974a, p. 95) mentions that the basalt intrudes the sediments. Macumber & Edwards (1997) also maintain that it forms a dike within the deposits. A recent visit to this site showed the Ghor el Qatar Series to be considerably tilted, up to the vertical (Fig. 5.1.5), which may put a question mark

Figure 5.4.3. Grain Sabt Basalt, on top(?) of the Ghor el Qatar Series at Zahrat el Qurein, north of the Dead Sea.

over observations claiming that the basalt is younger than the Ghor el Qatar Series.

5.4.4 Magmatics in boreholes

Magmatic occurrences are known only from boreholes drilled in the northern and central Jordan Valley, and none have been found in any of the numerous drillings in the Dead Sea and Arava regions. Two of the boreholes, Notera 3 in the Hula Valley and Zemah 1 just south of Lake Kinneret, were the subject both of detailed palynostratigraphic studies and radiogenic datings, so that chronostratigraphy based on the two methods could be compared (Horowitz & Horowitz 1985, 1990, Heimann 1990), discussed in detail in Chapter 7.

Magmatics dating 8.82 ± 1.04 Ma from 2,430 m (average of five measurements) occur in Notera 3 (Heimann & Steinitz 1989), and 9.5 Ma from 3,281 m in Zemah 1 (A. Heimann, Geological Survey of Israel 1998, pers. comm.). Shaliv reported three basalt horizons in the Belvoir 1 borehole, spanning the depth interval of 396–774 m, with ages of 8.7–9.7 Ma. These occurrences are attributed by Heimann to the Lower Basalt, while according to Shaliv (1991) the Lower Basalt is older, and these ages should correspond to the flow of the Intermediate Basalt, in the Umm Sabune Conglomerate. In the present context, however, (see Section 5.3.3) all volcanics older than the Bira but younger than the Herod Formation are regarded a separate stratigraphic entity, of Late Miocene Volcanics.

Rocks corresponding in age to the Cover Basalt are reported from several boreholes in Heimann (1990): Notera 3, 1,660–2,425 m; Kinneret 4, 90 m; Ami'ad 1, above 89 m; Rosh Pinna 1, above 195 m; and Zemah 1 (A. Heimann, Geological Survey of Israel 1998, pers. comm.) from 532 down to 2,674 m. These are regarded here as occurrences of the Intermediate Basalt (see discussion in Chapter 7.1.5). The Mechki Basalt is identified in Hula SH2 at a depth of 354 m and possibly at Notera 3 at 1,245–1,580 m. The Ruman Basalt in Hula SHJ1 at 158 m and possibly in Ayyelet HaShahar 2b at 69 m. Equivalents of the Hasbani Basalt (see Chapter 11) are reported from Notera 3 at 550–900 m and Hula 1 at 180 m, which may also correspond to the Nahal Orvim section (Mor 1986) of the Dalwe Basalt, at the western Golan escarpment.

The radiometric ages at Notera 3 raise a possibility that some of the magmatic horizons described in Heimann & Steinitz (1989) and Heimann (1990) as flows are indeed intrusive. The two suites of samples from 550 and 900 m yielded practically a similar age; but the age at 550 m conforms with the palynostratigraphy, while the lower one is much too young, and so should be considered intrusive. A similar situation occurs with the suite of samples from the 1,245–1,580 m interval, where the uppermost could be considered a flow, while the lower ones are most probably intrusive. Most of the magmatics in Zemah 1 are intrusive, being made of gabbro (Marcus & Slager 1985), so their datings only represent minimum ages.

CHAPTER 6

Palynostratigraphy, continuous sequences and unconformities

Palynostratigraphy is probably the most efficient tool for delineating the stages of evolution of the Jordan Rift Valley for several reasons. Variations in the composition of the vegetation through time, caused by evolution, by climatic or geomorphological changes, or a combination of these, are recorded in the pollen spectra in great detail, due to the abundance of these microfossils. Almost every analyzed sample yields at least tens, but usually hundreds of grains, so the statistical foundations are considerably better as compared with larger fossils, which are much more scarce. Pollen assemblages reflect both the immediate environment of deposition and regional conditions, because the grains are brought by wind and water, assembled within any kind of sediment, thus making possible correlations of deposits from basins which differ environmentally, such as in their salinity, depth, etc. Other microfossils, which live in the basin itself (if at all), reflect its peculiar conditions, but in most cases prove useless for long-range stratigraphic correlations.

The main reason for the inadequacy of most organisms to serve as guide fossils is the relatively short duration of the late Cenozoic, a time span in which the pace of evolution did not allow for major changes, a problem particularly when freshwater fossils are concerned. (It is true that larger mammals did evolve tremendously during the late Cenozoic, but they are rarely found as fossils, and certainly not in boreholes nor in continuous sequences.) Thus, with most "practical" fossils their age assignment is usually "Neogene to recent". Plants are no different in this respect, but plant associations are, since their composition depends on environmental conditions. These plant associations typically respond to climate, and their reorganization primarily represents climatic changes which, in turn, constitute a major foundation (although not the sole one) of the late Cenozoic time scale. The study of continuous pollen assemblages through rock sequences therefore results in climatostratigraphy. If this is tied up satisfactorily with other methods of stratigraphy, such as those based on the evolution of foraminifera, oxygen isotopes and radiogenic datings, the ensuing time scale can be used reliably.

In this respect the continental basins are entirely different from marine environments, where conditions are uniform over wide areas, enabling the use of such organisms as foraminifera, radiolaria and others for chronostratigraphic correlations. The Jordan Rift Valley was a continental environment for the major part of its history, but even in times when the sea invaded it the restrictive conditions in the resulting shallow bays, estuaries and lagoons changed very rapidly both spatially and in time.

The combination of continental deposition and considerable subsidence of the Rift's floor, which also enhanced erosion of its synchronously uplifting flanks, created the situation where most of the sequences, particularly the complete, continuous ones, are only preserved below the surface. As there is no use for larger fossils, which are hardly ever recovered from boreholes, it remains for the smaller ones to tell the story. Of these, mollusks, bearing on environment and somewhat on stratigraphy, are occasionally present in a state permitting identifications; diatoms and ostracodes, as well as scarce foraminifera, usually serve only for paleoenvironmental reconstructions, except on rare occasions. Consequently, palynostratigraphy remains the only practical tool for deciphering the stratigraphy, chronology and environments of the Jordan Rift Valley formations on a continuous basis.

Continuity is indeed the greatest merit of palynology in predominantly continental sequences. All other dating possibilities, be they based on fossils other than pollen, on radiogenic methods, or on the evolutionary stages of human artifacts, accurate as they are, only date points of the sections, which may at times be quite far apart. Knowledge of the continuous sequences has another merit, by providing an excellent reference for understanding the incomplete ones, outside the subsiding basins. The missing parts of sections usually represent periods of uplift and erosion, which are as important for deciphering the Rift's history as its subsidence. The last part of this chapter thus deals with the major unconformities, enlightened by the continuous sequences.

6.1 SOURCES OF MATERIAL

Three principal sources of material are available in the Jordan Rift Valley for palynostratigraphic purposes: recent sediments and shallow drillings, studied in great detail, which serve as a foundation for understanding the remote past; deep boreholes, giving the overall, continuous stratigraphic and environmental outline of the history of the region; or outcrops and prehistoric sites, which need to be dated and their environment reconstructed.

The first group was studied for two purposes. Initially for the understanding of recent pollen deposition in the Rift Valley basins, in relation to present-day vegetation and general environmental conditions, which would serve as a reference for deciphering past trends (Rossignol 1969a, Horowitz 1969, Weinstein 1979, Baruch 1993). This stage was followed by detailed analyses of radiocarbon and otherwise dated late Pleistocene through Holocene sequences, obtained from shallow drillings in the Hula, Kinneret and Dead Sea lakes, which delineated continuous environmental changes throughout this time span in great detail (Rossignol 1969b, Horowitz 1971, Weinstein-Evron 1983, 1989, Baruch 1993). Comparison of the results obtained from these studies with sequences from the oceans and other continents (Horowitz 1989c, Fuji & Horowitz 1989) made possible the construction of climatostratigraphic models. These create the basis for correlation of the palynostratigraphy of the Jordan Rift Valley formations (Horowitz & Horowitz 1985, 1990) and prehistoric sites (Horowitz 1989a, 1996b, Weinstein-Evron 1990) with the global late Cenozoic time scale.

The second group of sources for palynostratigraphic studies, which indeed made a fundamental change of approach to the chronostratigraphy of the Jordan Rift, comprises 13 deep boreholes drilled along the Valley, from the Hula down to the northern Arava, in the search for oil and natural gas (Fig. 6.1.1). These boreholes each penetrated sequences in the order of kilometers of thickness, which were sampled at intervals in the order of tens of meters (Horowitz & Horowitz 1990, Horowitz 1996a). Such sampling intervals would, at first glance, seem quite sparse, but considering the rates of accumulation involved, over a relatively short geological time span, is sufficient for obtaining the general picture. The sampling intervals were also affected by practical considerations, or, in other words, there is no way of analyzing some 40 km of borehole sequences every centimeter (as indeed would be much better) in one's lifetime. Distribution of the Neogene palynozones in the analyzed boreholes is shown in Fig. 6.1.2, while the Quaternary ones are further detailed in Fig. 6.1.3.

The samples comprised drill cuttings, not cores. Again, this was a result of being practical, since I had not found any way of convincing an oil company (and rightly, on their behalf) to extract continuous cores several kilometers long. Naturally, the type of samples and their spacing put limitations on the conclusions obtained. However, the recurrence of results along as many as 14 drilled sequences (the other one in the Israeli offshore), indicates their validity in reconstructing the general climatostratigraphic trends of the late Cenozoic in the southern Levant, which seems sound enough.

The chronostratigraphy of the palynozones is based on three sources (for a detailed discussion and terminology see Section 1.4.2). The Quaternary sector was correlated with the oxygen isotope curves obtained from the oceans, themselves dated by magnetic reversals (Horowitz 1989c, and further elaborated in this work). The Neogene palynozones had been identified in a borehole, Bravo 1, drilled off the coast of Israel, where the strata also contained foraminifera, on which the common biostratigraphic zonation is based (Horowitz & Derin 1987). Radiometric datings of volcanics along the drilled sequences (Heimann 1990, Heimann et al. 1996, among others) resulted in points of absolute ages, scattered along the entire section. Occasionally, some discrepancies occur between the stratigraphic and absolute ages, which are discussed in more detail in Chapters 7 and 11.

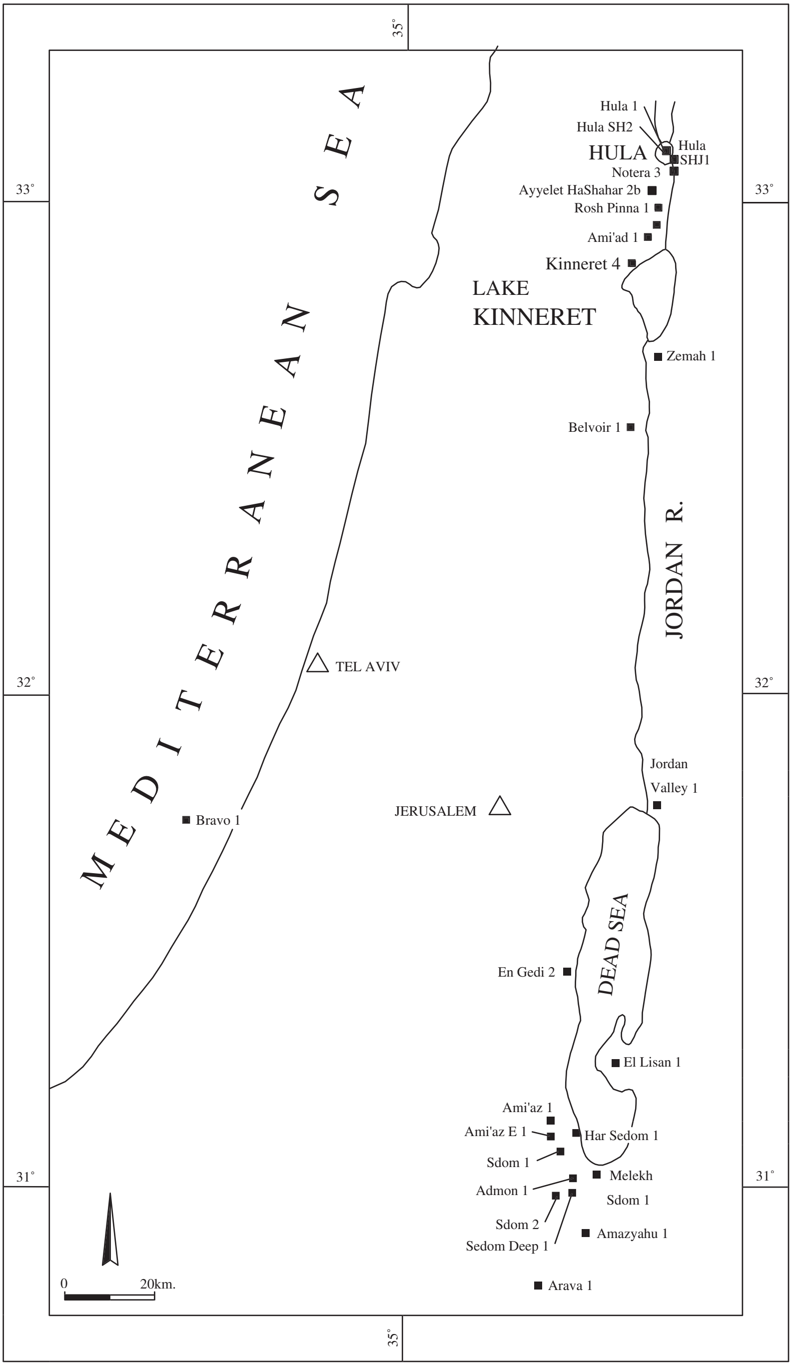


Figure 6.1.1. Location map of boreholes analyzed palynostratigraphically and others.

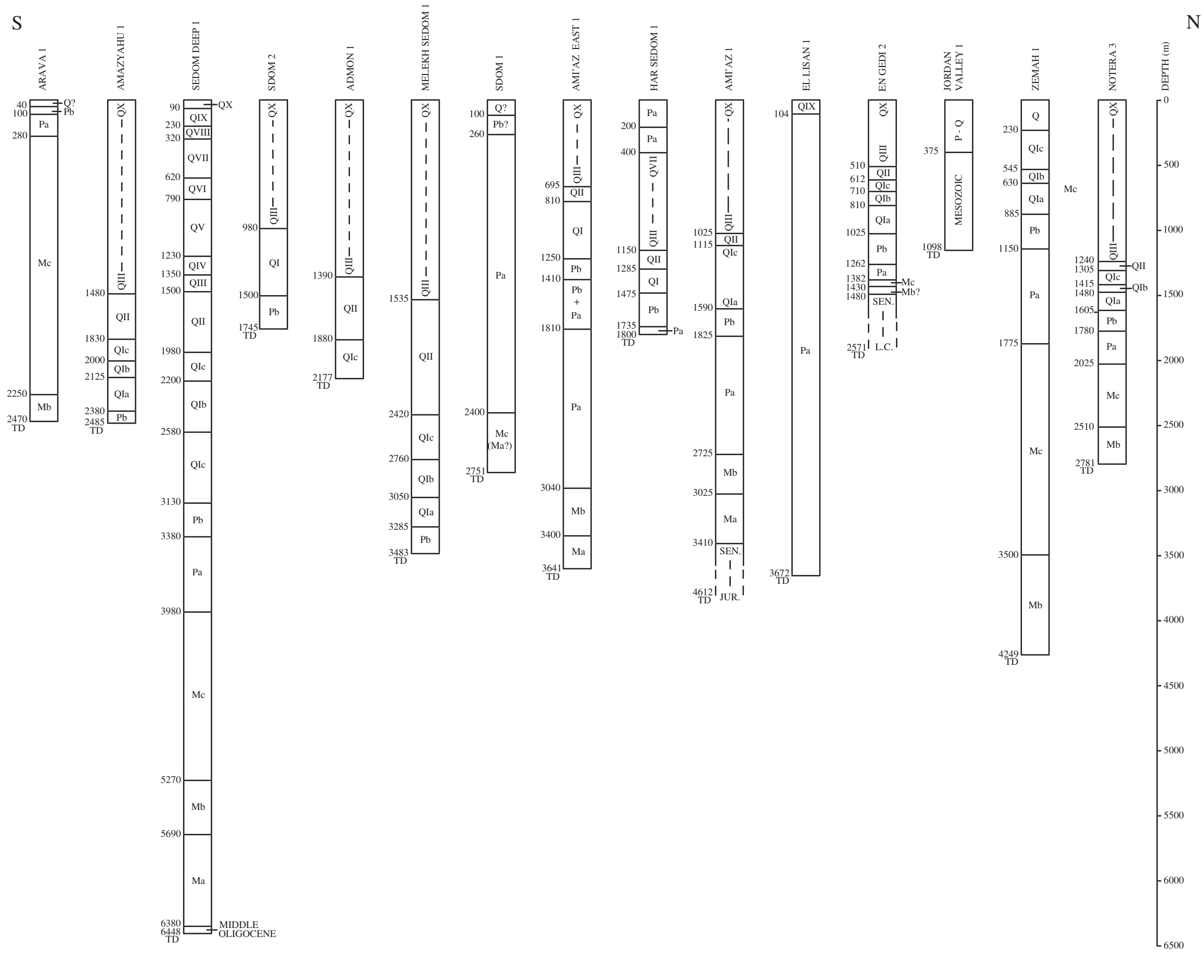


Figure 6.1.2. Distribution of the palynozones in the boreholes analyzed. For details of the Quaternary Palynozones please refer to Fig. 6.1.3.

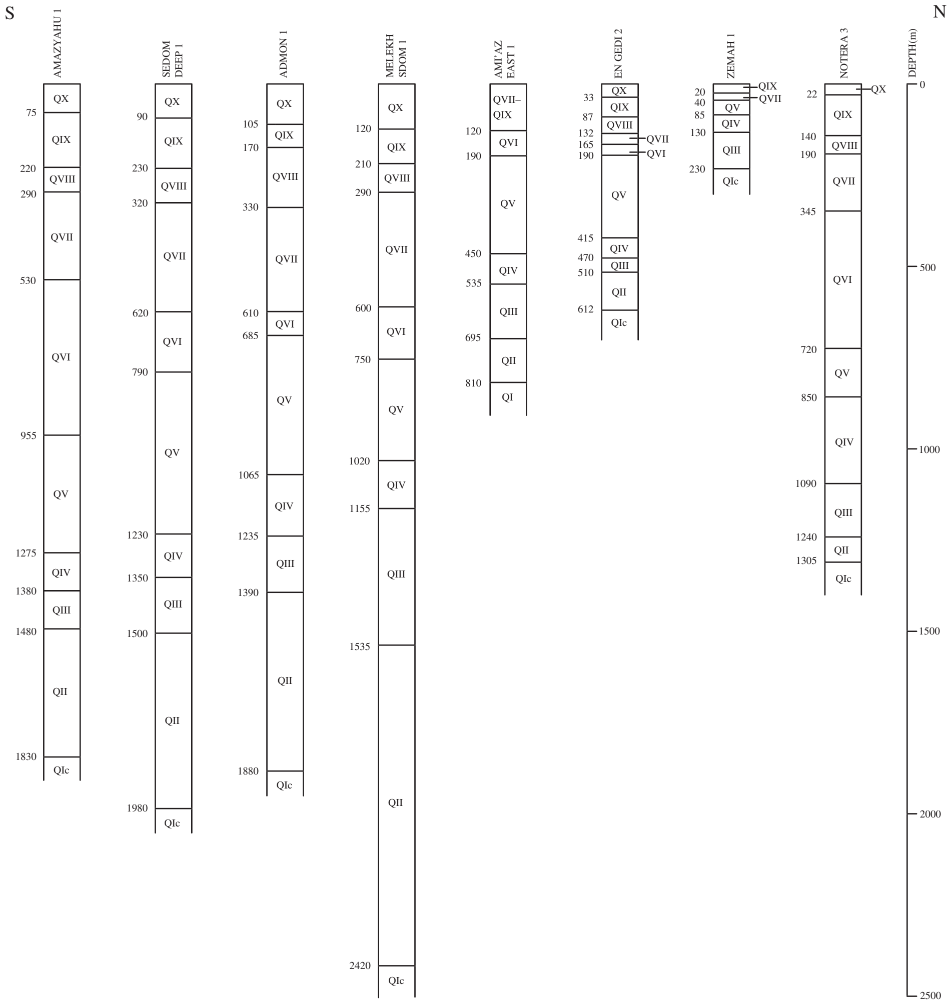


Figure 6.1.3. Distribution of the Quaternary Palynozones QII–QX in the boreholes analyzed.

The third group of sources is quite heterogeneous: the only common thing for the samples is their being collected from outcrops, or occasionally shallow archaeological and other artificial excavations. Compared with the former groups, these samples never make a considerable sequence, not to mention continuous, which naturally affects their stratigraphic significance. Even worse, such samples are often badly oxidized (Horowitz 1992a, p. 204), or quite frequently contaminated with recent pollen (Weinstein-Evron 1986, Kronfeld et al. 1988). Despite these grave limitations, such outcrop samples enabled construction of a time scale for the Jordan Rift Valley quite accurately, even before the deep sequences had been drilled and analyzed (Horowitz 1979, Chapter 6). In the present state of knowledge their contribution has been reduced to solving local problems of stratigraphy and environments. These pollen spectra are discussed in their relevant context, namely Neogene and earliest Quaternary outcrops in Chapter 7, Quaternary and prehistoric sites in Chapter 11.

6.2 PRESENT-DAY POLLEN

The study of present-day pollen is an essential basis for understanding paleoenvironmental conditions. This involves investigations of pollen production, dispersion and transport, deposition and preservation, all summed up in the nature of the pollen spectra obtained from samples, and what they tell about past environments, vegetations and transporting processes. A detailed discussion of the methodologies involved, particularly in arid regions, is advanced in Horowitz (1992a, Chapters 4–8), so to avoid unnecessary duplication and due to lack of space, it will not be repeated here.

Recent pollen deposition in the Jordan Valley was studied particularly in surface samples collected from its three lakes. Airborne pollen distribution was studied across the entire Israeli area (Weinstein 1979). The Hula was investigated by Horowitz (1971) and Weinstein (1982); Lake Kinneret by Horowitz (1969) and Baruch (1983); and the Dead Sea by Rossignol (1969a) and Baruch (1993).

The most surprising outcome of these studies is the general uniformity of the proportions of pollen derived from the regional vegetation in the composition of spectra all along the Rift, be it the Hula lying in a typical Mediterranean domain, or the Dead Sea which is situated in an extremely arid environment. The differences in the sections of the pollen spectra representing local floras of marshes, rivers, wadis, springs or saline playas, are naturally conspicuous. The similarities in the regional pollen spectra result mainly from the nature of the transporting agents. Northern winds are very common along the Jordan Valley, carrying pollen of Mediterranean trees such as oaks and cedars even as far as the Gulf of Aqaba (Horowitz 1966). The water bodies on the way act like pollen traps, thus typical Mediterranean spectra were recovered from recent sediments of the Dead Sea. Other, dry surfaces do not act as traps, so that wadi sediments around the Dead Sea are altogether devoid of Mediterranean pollen (Horowitz 1992a, p. 253), which again reappear in the sediments from the Gulf of Aqaba, further south.

The Jordan River also carries pollen southward, down to the Dead Sea. It seems however that the role of winds is much more significant, as is evident in the similarity of recent pollen spectra from the Dead Sea and the Gulf of Aqaba. Another factor that helps the dominance of Mediterranean pollen in the spectra is the extremely low masking factor of the local desert vegetation to the south, where the poor plant cover and very low quantities of grains produced (Horowitz 1992a, p. 8) contribute very little to the total amount of deposited grains, thus enabling a fair representation of the vegetation to the north.

6.3 BIASES IN THE POLLEN SPECTRA

The term "biases" refers either to over- or under-representation of a certain group (or groups) of pollen in the spectra, or to a sudden change in the percentage of one of the constituents (a single or a group of taxa), contrary to the expected, presumably natural trend. Numerous biases are indeed the basis of palynological discussions in general, such as differential pollen production, transportation, preservation, etc., and their combination makes the pollen spectrum different from its mother vegetation. These, but especially their application in the *Palynology of arid lands*, are discussed in detail in Horowitz (1992a, Chapter 8). Biases are thus divided into two distinct groups: the first includes those peculiar to the region, comprising a continuous, occasionally widespread phenomenon, which hardly changes through time; the second involves biases of a local character, either in space or time, indicating particular, sometimes short-lived aspects. Here, only the biases peculiar to the Jordan Rift Valley are discussed.

The most conspicuous bias was mentioned above, namely the uniformity of pollen spectra comprising regional elements along the entire Jordan Rift Valley, resulting from transportation of the Mediterranean elements, accompanied by the low masking factor of the poor local vegetation to the south. Another regional bias, significant only when recent pollen spectra are dealt with or occasionally concerning archaeological sites, is the occurrence of pollen produced by imported and agricultural plants. These do not yield much information about the natural environments, but are helpful for understanding the transportation paths of the grains, or agricultural and plant husbandry processes. Indeed, when the present is taken as a reference for past environments, the "exotic" pollen are omitted altogether from the calculations.

Other biases found along the Jordan Valley are usually of a more localized nature, but occasionally are very helpful in paleogeographic reconstructions. The over-representation of pollen produced by marsh vegetation in the Hula Valley was used to delineate changes in the marsh–lake proportions during the late Quaternary (Horowitz 1971, 1973). Pollen brought by the Jordan River into Lake Kinneret indicate the river's path within the lake (Horowitz 1969). Occurrences of redeposited Pliocene pollen in sediments of Lake Kinneret, close to saline springs, help in understanding the source of salinity and reconstructing the springs' history (Horowitz 1970).

High percentages of pollen originating in hydrophil vegetation attest to the higher activity of springs. This may follow a general, regional increase in arboreal pollen, in which case the springs are an outcome of increased amounts of rain; but sometimes, when these two curves do not run parallel, the excess of hydrophil pollen may indicate exposure of aquifers by faulting. Springs in arid regions, of higher water output, may also support Mediterranean trees, which could lead to mistaken conclusions on the general climate (Weinstein-Evron 1987), particularly when samples are collected close to the source. Only an analysis of distant contemporaneous sediments could support or refute such interpretations (Horowitz 1987b).

6.4 LATE CENOZOIC POLLEN ASSOCIATIONS AND FLORAS

A basic distinction must be made between the arboreal and non-arboreal groups, which principally emerges from the level of possible identifications (Horowitz 1992a, p. 203). Most pollen grains of non-arboreal plants can be identified in practice merely at the family level, with only a few identifiable by genus, and very rarely by species. Since most of these plant families occur along the entire late Cenozoic, the evolutionary significance of their grains (and consequently their use for stratigraphy) is rather limited. Although the genera and species change in different environments, most of the families show an almost global distribution, especially those producing large amounts of pollen such as Gramineae, Chenopodiaceae, Compositae and so on. Changes of the various non-arboreal pollen shares in the pollen spectra therefore remain applicable only to local paleoenvironmental reconstructions. Contrary to the non-arboreal, arboreal pollen yields more information on the composition of the late Cenozoic flora of the southern Levant, although in some cases even these could not be identified below the family level, for lack of an adequate reference collection or due to less than ideal preservation, particularly in the case of palynomorphs of older periods such as the Oligocene and Miocene.

Analyzed by the major constituents, the late Cenozoic sequences of pollen associations of the southern Levant are divided into four groups, representing floras of different climatic and environmental regimes, broadly corresponding to the Oligocene and Miocene ("M" palynozones); Pliocene ("P"); earliest Quaternary (Palynozones QI and QII); and the rest of this period, from Palynozone QIII until the present day, QX.

The oldest, Oligocene flora, was only encountered in the three deepest samples from the Sedom Deep 1 borehole, where the spectra are dominated by triporate pollen, most probably Juglandaceae of subtropical origin. This trend continues into the Miocene, which is characterized throughout by pollen derived from trees of the families Juglandaceae, Betulaceae and Pinaceae. The genera could not however be safely recognized. The pollen assemblages, as well as fossil plants such as a variety of palms (*Palmoxylon*) and *Leguminoxylon* (Lorch & Fahn 1959, Zohary 1959), recovered from the Hazeva Formation, together with its vertebrates (see Section 7.1.3.2), all indicate a subtropical lowland environment, which persisted in the southern Levant from the Oligocene to the end of the Miocene, fluctuating between wetter and drier phases (Horowitz 1990).

The Pliocene flora is represented by entirely different pollen assemblages. The most conspicuous constituents are members of the *Quercetalia*, accompanied by a conifer, most probably *Picea orientalis*. This constitutes a rather characteristic northern Mediterranean, or Pontian, flora, indicating the onset of this environment and flora in the region (see also Suc 1989). This general environmental trend did not basically change until the present, but went through two rather different floral associations. During Palynozones QI and QII the arboreal pollen flora comprises almost solely *Picea orientalis*, which was substituted by *Quercetalia* during the following palynozones, to become the typical Mediterranean flora of the present. Climatic changes from the Pliocene onward have caused differences in percentages of the various pollen groups consisting the spectra, but the basic composition remained Mediterranean.

It is quite interesting to follow the change in the environmental gradient along the Jordan Valley throughout the late Cenozoic, which is to some extent minimized in the pollen spectra, due to the averaging effect of long-range transporting agents (see above). The Miocene palynozones are richer in arboreal pollen to the south, in samples from the Dead Sea region, as compared with the Hula. This trend changes in the Pliocene, when higher AP values are recorded in samples from the north, a trend which is gradually accentuated until today.

This pattern is also seen in the sediments, expressed particularly by their overall coloration. Red-beds are typical of a warm, humid environment (see a brief discussion in Section 7.3.3). Since the southern Levant was never cold during the Cenozoic, as seen from the rich and varied associations of marine fossils, it is chiefly humidity that determines the formation of red-beds. Data for the Oligocene are scant, but the rocks deposited during this period (if the proposed correlations are accepted) are reddish to the south, yellowish-white to the north. The rare early Miocene rocks are always whitish, representing the overall aridity of that period, as are the late Miocene occurrences. The middle Miocene southern formations, Hazeva and Dana, are typically red, but the northern Herod is much less so. Early Pliocene rocks are whitish all over, most possibly because the climate was quite cool; late Pliocene sediments retained the pale colors to the south, but their northern counterparts are reddish, reflecting the general warming and the higher rainfall to the north. The same trend holds for the entire suite of Quaternary deposits.

6.5 LATE CENOZOIC PALYNOSTRATIGRAPHY OF THE JORDAN VALLEY

The late Cenozoic palynostratigraphy of the Jordan Valley was first proposed by Horowitz & Horowitz (1985) for the Notera 3 borehole, drilled in the center of the Hula basin, penetrating a sedimentary column (with several volcanic intercalations), covering the time span from the present day continuously down to the middle Miocene. This sequence was subdivided into 14 palynozones, two for the Miocene, two for the Pliocene and ten for the Quaternary. An older, late

Oligocene–early Miocene palynozone, Ma, was penetrated by the Ami'az 1 and Bravo 1 boreholes (Horowitz & Derin 1987), the former drilled near the southern end of the Dead Sea, the latter off the Mediterranean coast. Palynozone Ma and earlier middle Oligocene sediments were later encountered in another borehole drilled south of the Dead Sea, Sedom Deep 1, during 1992–1993 (Horowitz 1996a). No formal palynozone status was attached to the middle Oligocene section, since only three samples in a single borehole were analyzed.

The Sedom Deep 1 borehole, being the deepest ever drilled into the Jordan Rift Valley fill, down to 6,448 m below the surface, penetrating the most complete continuous sequence, is used here as a reference section for the late Cenozoic palynostratigraphy. Additional information from the other 12 deep boreholes in the Jordan Valley, as well as from the one in the Mediterranean (Fig. 6.1.1), is also included.

6.5.1 Presentation

Pollen diagrams accompany the descriptions of the palynozones, which are therefore kept short. The Neogene and early Quaternary diagrams are from the most recently drilled Sedom Deep 1 borehole, while most of the Quaternary was sampled and studied in more detail in a nearby well, Admon 1 (Fig. 6.1.1).

Each pollen diagram (Figs 6.5.1–6.5.10) is divided into four columns. The left shows relations within the arboreal pollen group, the total AP taken as 100%. The second column from the left displays relative shares of grains originating from three vegetation groups which comprise the regional flora, excluding pollen derived from hydrophil and halophil vegetation of the Rift Valley local habitats such as lakes, springs, wadis and playas. In the second column, the first group comprises arboreal pollen, the second group represents pollen originating from steppe vegetation, while the third includes grains produced by desert plants. The total number of pollen derived from the regional vegetation is taken as 100%.

The third column presents composition of the local, mostly hydrophil and halophil pollen flora, derived from plants growing in close connection to the locality of deposition, in wadis, playas, lake shores and springs, the total of which is taken as 100%. The fourth column is based on the total number of pollen counted in every sample, taken as 100%, and gives the ratios of local to regional elements, arboreal pollen as part of the entire spectrum, and fern spores. In the two left columns each pollen curve is presented on the vertical axis, so they are overlapping, the third column is cumulative, while the right column is cumulative for the fern spores, local and regional vegetations, and overlapping for the AP curve. This form of presentation was found (Horowitz 1992a, p. 229) to be most suitable for displaying the peculiar characteristics of this region.

The arboreal pollen presented in the left column is divided into four groups, representing different environments. *Quercetalia*, comprising mainly oaks, typical of the Mediterranean garrigues and forests. Conifers, such as pine, cypress, cedar and unidentified Pinaceae (mainly in the Miocene), which seem to grow in drier domains, usually highlands. *Picea orientalis* which, although also a conifer, is known today only from some secluded localities in western Turkey, where it grows in a euxine forest, probably indicative of a rather wet, temperate environment. Betulaceae and Juglandaceae, are of wet subtropical lowlands affinity.

The second column from the left displays the relationships of three pollen groups: arboreal, discussed above, generally representing moister environments; steppe, of the Irano-Turanian domain; and desert, indicative of the Saharan regime. The steppe vegetation is represented in the southern Levant by pollen of *Artemisia* and *Ephedra* as the main constituents, occasionally accompanied by *Scabiosa*, *Echinops*, *Astragalus*, *Centaurea*, Cruciferae, Liliaceae, Papilionaceae and Malvaceae. The most common pollen grains indicating a desert environment in this region (A. Danin, Department of Botany, the Hebrew University of Jerusalem 1982, pers. comm.) originate from Compositae and Umbelliferae, accompanied by occurrences of *Plantago* and *Asphodelus*.

The third column shows the composition of pollen derived from the local, hydrophil and halophil wadi, playa, lake shore and springs plants, which are shown cumulatively. The left curve is for hydrophil, including both trees such as *Tamarix*, which is the principal component, accompanied by *Acacia*, *Moringa*, *Ziziphus*, *Phoenix*, *Populus*, *Acer* and *Salvadora*, and herbaceous plants such as *Typha*, *Rubus*, *Polygonum* and Labiatae. All these grow in intimate connection with perennial water, either springs or high groundwater tables. Some of these plants can withstand salinity, but in general the group indicates freshwater occurrences near the locality of deposition. The following curve is for Gramineae and Cyperaceae, which bear environmental characteristics similar to the former, so the two left curves are mostly regarded together as indicative of freshwater occurrences in the region. The rest of the area on this column represents the proportion of Chenopodiaceae in the local spectra, indicating saline conditions.

The last column to the right is calculated from the total number of counted pollen, and is used primarily to show relations of the local and regional elements within the pollen flora. The arboreal pollen curve given in this column is less conclusive than the curve displaying AP as part of the regional vegetation, and is only given here for comparison with data presented in diagrams from other boreholes (Horowitz & Horowitz 1990). The fern spores shown in this column seem to follow the hydrophil vegetation curve trends, but may eventually indicate redeposition from saline springs, dissolving older evaporites (Horowitz 1970).

6.5.2 The palynozones

6.5.2.1 *Middle Oligocene*

The middle Oligocene sediments, penetrated only by the Sedom Deep 1 borehole in the southern Dead Sea basin, are represented solely by three analyzed samples (Fig. 6.5.1), which yielded pollen spectra rich in triporate palynomorphs,

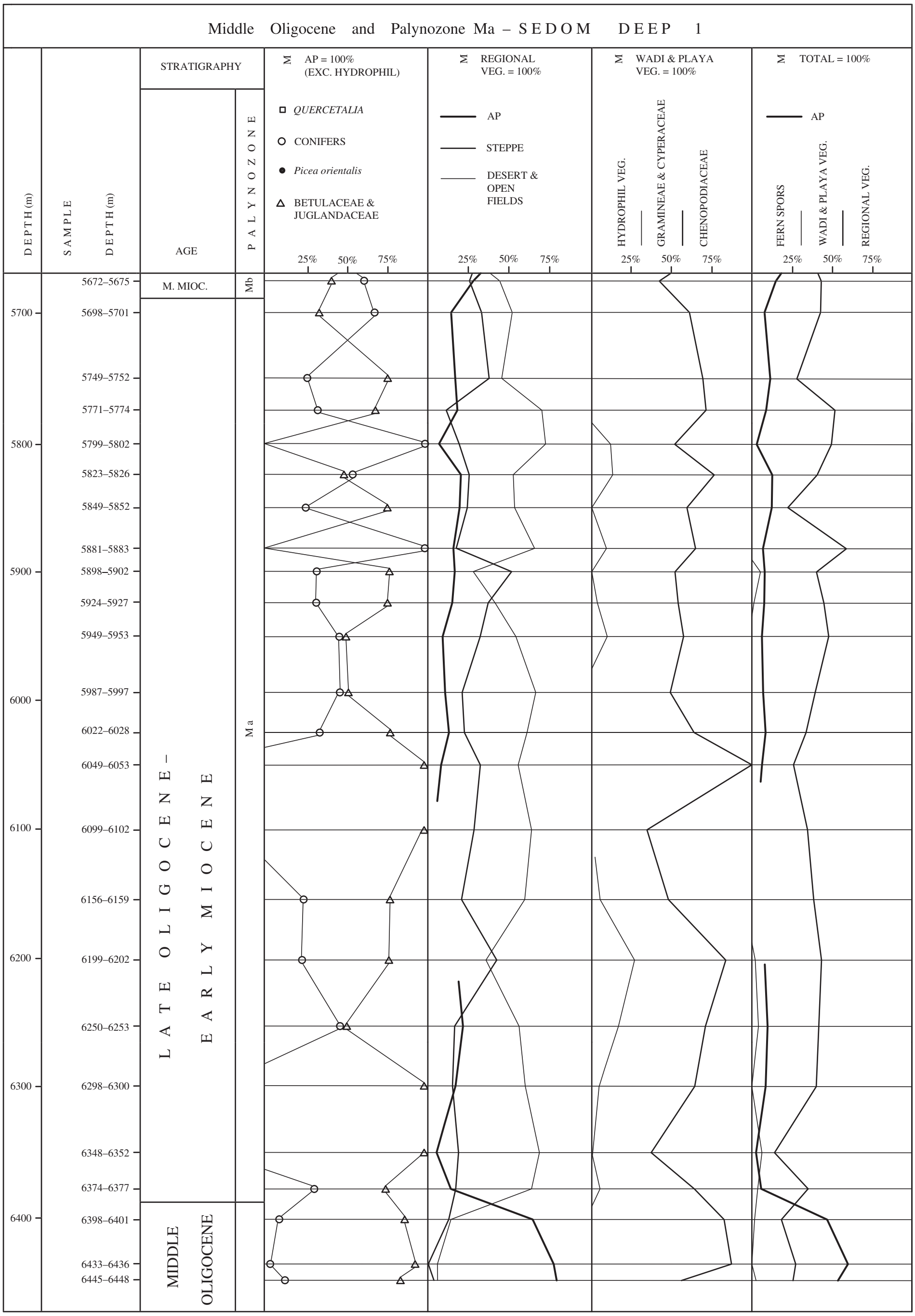


Figure 6.5.1. Pollen diagram of the middle Oligocene and Palynozone Ma, Sedom Deep 1 borehole.

comprising up to 70–80% of the total, typical for this period in the Mediterranean region. The pollen spectra indicate a wet subtropical climate and lowland, possibly marsh forests. Other microfossils, particularly nannoplankton assemblages recovered from this interval, verify the palynological age assignment and indicate a marine environment of deposition (Baker 1994).

6.5.2.2 *Late Oligocene through Miocene*

Sediments of Palynozone Ma (Fig. 6.5.1) were analyzed in three boreholes, Sedom Deep 1 and Ami'az 1 in the southern Dead Sea basin and Bravo 1 in the Mediterranean offshore. They yielded pollen spectra rich in non-arboreal components, of which Compositae are the most common. The arboreal part consists of pollen derived from trees of the families Juglandaceae and Betulaceae, resembling *Juglans*, *Pterocarya*, *Platycarya*, *Alnus*, *Corylus* and *Carpinus*, accompanied by some unidentified Pinaceae. The pollen assemblages indicate a warm, dry, subtropical environment, possibly bearing a savanna vegetation, for late Oligocene through early Miocene times. Palynozone Ma is correlated with the biostratigraphic Foraminifera zones N3 (P22) up to N6, of late Oligocene through the early part of the early Miocene.

Sediments of Palynozone Mb (Fig. 6.5.2), as well as the overlying palynozones, were encountered in most boreholes analyzed. They are characterized by arboreal pollen of types quite similar to those of the underlying Ma, namely Pinaceae, as yet unidentified, Betulaceae and Juglandaceae. Typical arboreal pollen ratios for Mb are around 40–50% of the pollen derived from regional vegetation, while the rest of this group chiefly originates from desert plants, such as Compositae, accompanied by a few pollen of steppe plants, especially in the lower part. Palynozone Mb is correlated with Foraminifera zones N7–N14, of late early through middle Miocene age. The pollen spectra indicate a moist subtropical climate, with a wet, partly marshy savanna vegetation.

Table 6.5.2.1 compares composite pollen spectra for Palynozone Mb from three localities: the Hula Valley to the north, the Mediterranean offshore to the west and the southern Dead Sea basin. The principal difference which emerges is that arboreal pollen are more abundant in the southern Jordan Valley than to the north and west, while an opposing picture is displayed by desert constituents, especially the Compositae. The composition of the AP spectra is also quite different, with the north dominated by conifers, decreasing west and markedly south, opposing the picture displayed by the Betulaceae–Juglandaceae group. The trees thus indicate drier environments to the north. Steppe pollen also support this conclusion, being rare to the north, increasing southward.

Palynozone Mc (Fig. 6.5.3) is typified by its characteristic regional vegetation pollen, which is represented almost exclusively by Compositae, with very low values of arboreal pollen, lower even than those encountered in Ma times. The Palynozone corresponds to Foraminifera zones N15, Nl6 and Nl7, of late Miocene age. The pollen spectra indicate a very dry, desert environment of the

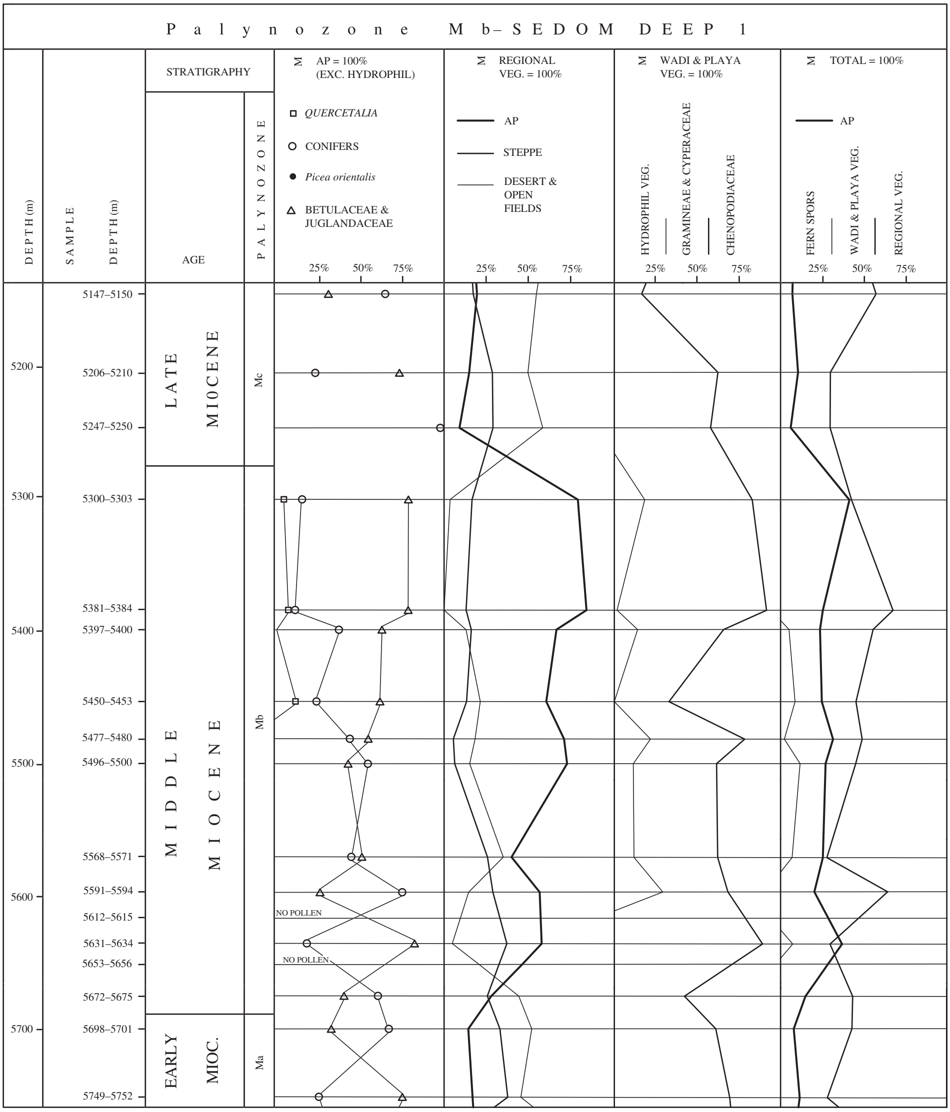


Figure 6.5.2. Pollen diagram of Palynozone Mb, Sedom Deep 1 borehole.

Saharan type. As before, arboreal pollen, although rare, are more abundant to the south.

6.5.2.3 *Pliocene*

The Pliocene sequence was subdivided into two palynozones (Figs 6.5.4 and 6.5.5): Pa, corresponding to Foraminifera zones N18, N19 and the lower part of N20, early through middle Pliocene, and Pb, upper N20 and N21, of late Pliocene age, which bear pollen spectra considerably different from those of

Table 6.5.2.1. Percentages of pollen groups in composite borehole samples of Palynozone Mb, from the northern Jordan Valley (Notera 3), the Mediterranean offshore (Bravo 1) and the southern Dead Sea (Sedom Deep 1).

% of pollen/Borehole	Notera 3	Bravo 1	Sedom Deep 1
Quercetalia out of AP	0	0	3.88
Conifers out of AP	88.24	47.79	32.04
Betulaceae and Juglandaceae out of AP	11.76	52.21	64.08
AP out of regional	49.79	47.88	62.80
Steppe out of regional	0.84	9.32	19.51
Desert out of regional	49.37	42.80	17.68
Hydrophil out of local	0	1.48	11.63
Gramineae and Cyperaceae out of local	32.16	48.15	61.63
Chenopodiaceae out of local	67.84	50.37	26.74
Local out of total	74.40	34.01	49.14
Spores out of total	0.52	6.55	4.00
AP out of total	12.49	28.46	29.43
Total counted	953	397	350
Number of samples	14	4	12

the underlying Miocene sediments, the "M" palynozones. The dominant Pliocene arboreal pollen are derived from two groups of trees, the *Quercetalia* and a conifer, most probably corresponding to *Picea orientalis*. It is quite difficult to be exact about this identification: of all the conifer pollen examined in reference collections, the Pliocene grains looked very much like those produced by *P. orientalis*. It must be stressed, however, that N. Drivaliari (Laboratoire de Palynologie, U.S.T.L., Montpellier 1989, pers. comm.) assigned these grains to *Abies*, or to the type genus *Cathaya*, which may be correct as well. At any rate, the different composition of the arboreal pollen group, as compared with the Miocene, probably represents initiation of the Mediterranean climatic and vegetation domain (Suc 1989, Fauquette et al. 1998). The age of Palynozone Pa is further strengthened by occurrences of Tabianian planktonic foraminifera in the corresponding sequence in the Zemah 1 borehole, 1,750–1,777 m below the surface (Gerry & Derin 1983). For details see Section 7.1.5.

Palynozone Pa is richer in arboreal pollen as compared with the overlying Pb. Shares of up to 40% of the pollen produced by regional vegetation are typical for Pa, while Pb is characterized by figures of 20–30% for that group. There is no essential difference in composition of the non-arboreal pollen group between the Pliocene and the Miocene and indeed the same taxa continue to the present day. Part of it may be a result of the problems arising in exact identification, mentioned above.

During the Pliocene the offshore sections (Horowitz & Derin 1987) show somewhat lower arboreal pollen percentages than those obtained from the

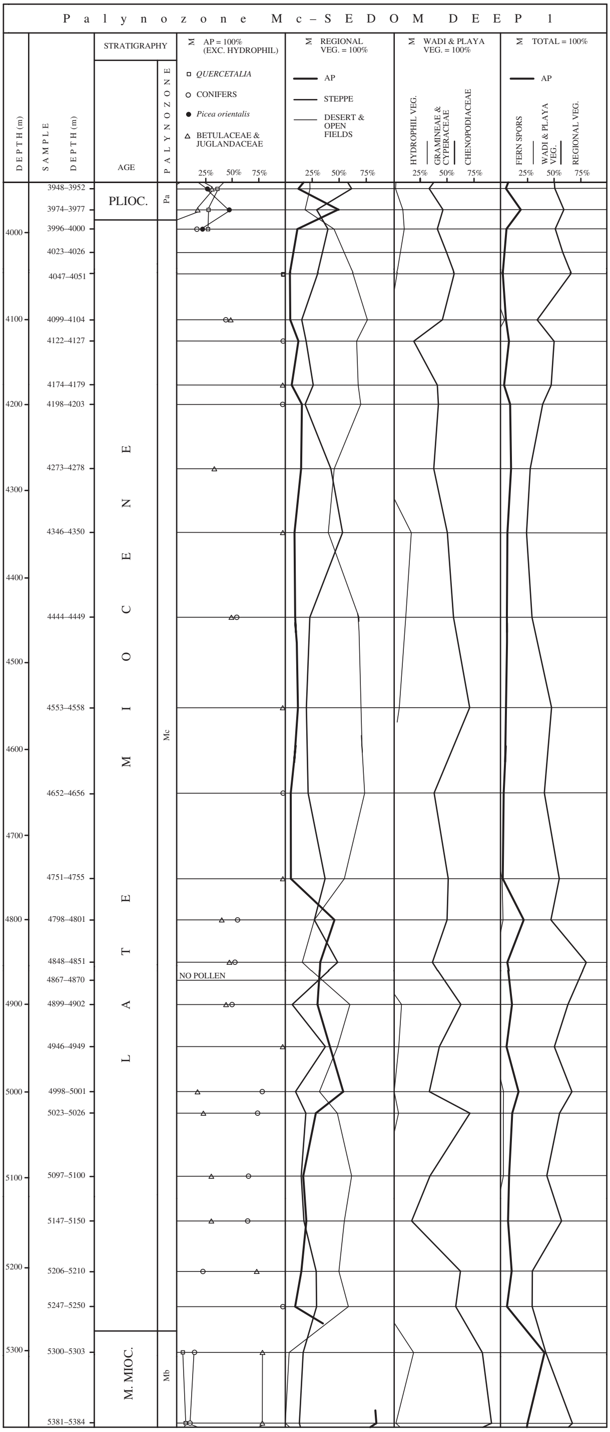


Figure 6.5.3. Pollen diagram of Palynozone Mc, Sedom Deep 1 borehole.

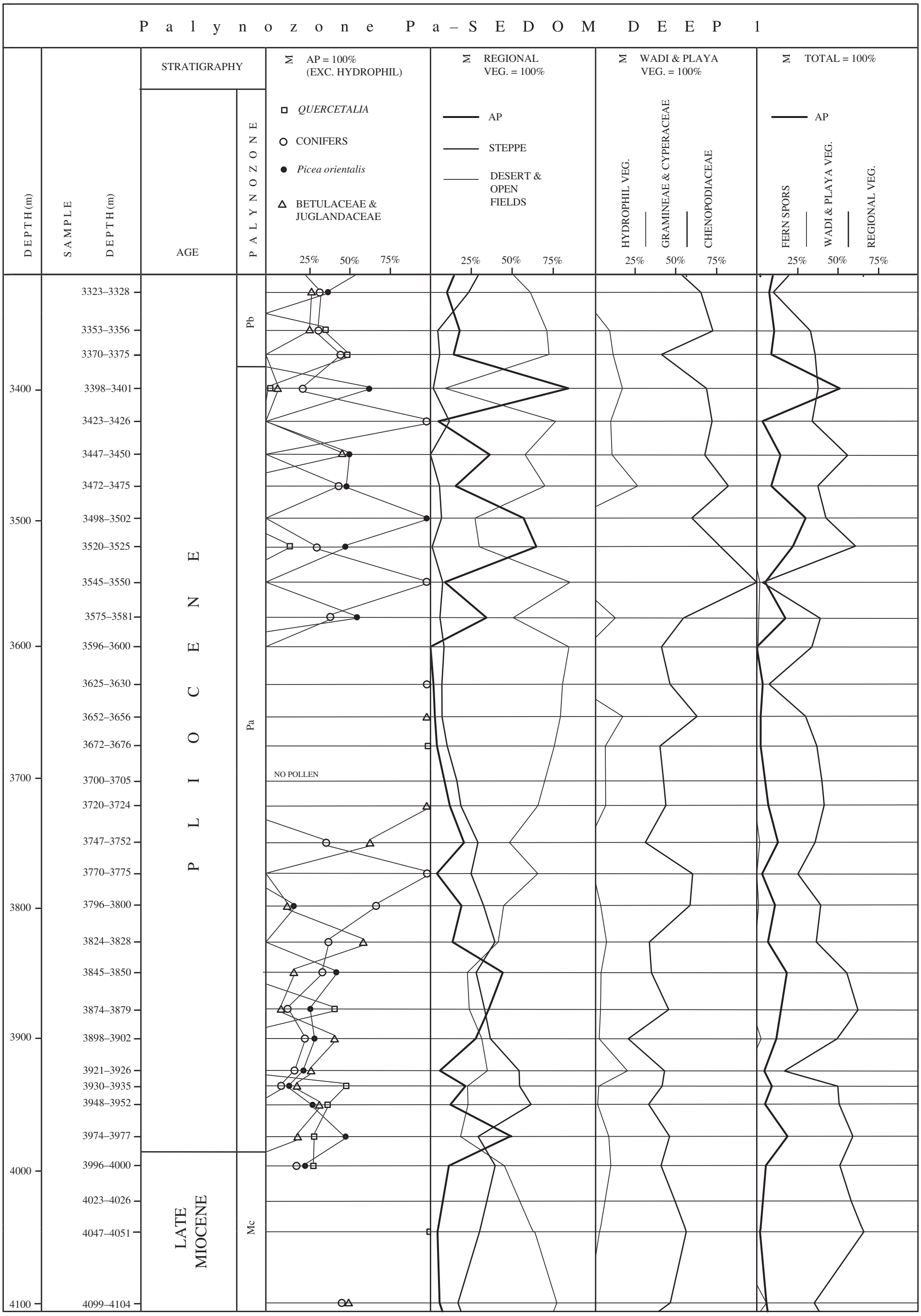


Figure 6.5.4. Pollen diagram of Palynozone Pa, Sedom Deep 1 borehole.

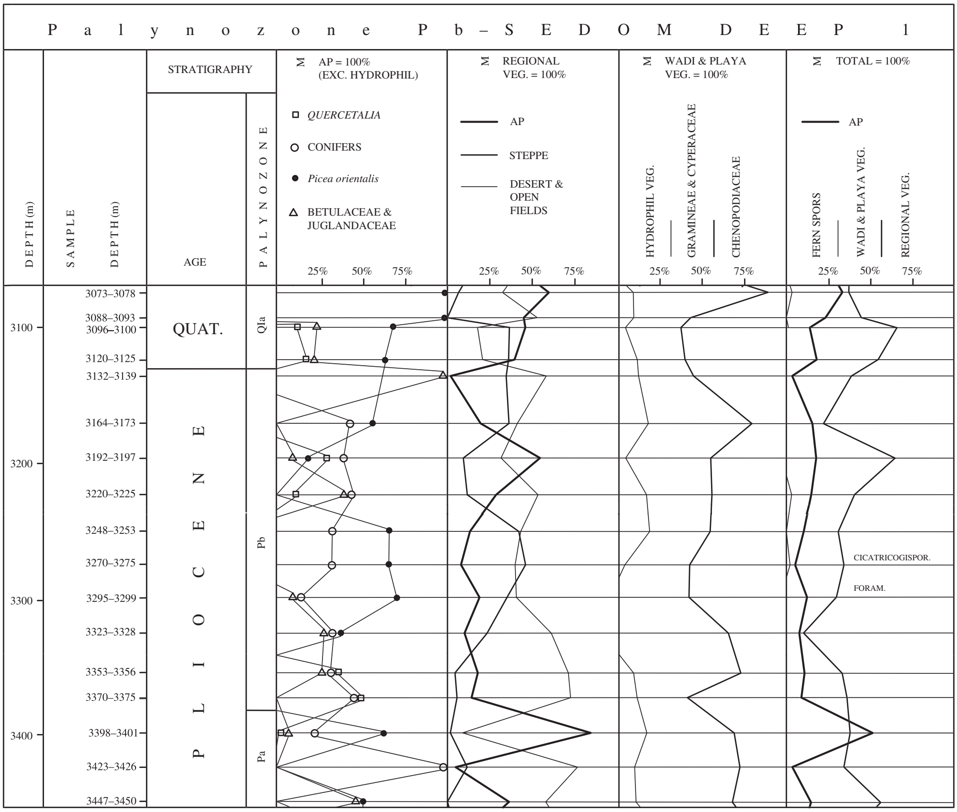


Figure 6.5.5. Pollen diagram of Palynozone Pb, Sedom Deep 1 borehole.

Jordan Rift region, but within these Betulaceae–Juglandaceae grains are more abundant (Table 6.5.2.2). The reason for the slight over-representation of non-arboreal pollen in the offshore sequences is probably connected with the activity of the Nile, which commenced in the early Pliocene. The Nile waters carry and deposit offshore considerable amounts of non-arboreal pollen (Horowitz 1979, p. 188), especially those originating from its prolific bank and delta vegetation, which may have also partly contained Betulaceae–Juglandaceae.

The area of Israel constituted a lowland during the Pliocene (Horowitz 1979, p. 75), which most probably also supported marsh vegetation of types similar to the Nile. To the northeast, where the landscape became higher in elevation and slightly drier, the oaks grew, while conifer forests most probably covered the even higher hills and mountains. The pollen spectra indicate a wet northern Mediterranean, Pontian type climate for Pa, becoming slightly drier, possibly also warmer, in Pb times.

As for Mb, Table 6.5.2.2 compares composite pollen spectra for Palynozone Pa from three localities: the Hula Valley to the north, the Mediterranean offshore to the west and the southern Dead Sea basin. Contrary to the Miocene, the northern

Table 6.5.2.2. Percentages of pollen groups in composite borehole samples of Palynozone Pa, from the northern Jordan Valley (Notera 3), the Mediterranean offshore (Bravo 1) and the southern Dead Sea (Sedom Deep 1).

% of pollen/Borehole	Notera 3	Bravo 1	Sedom Deep 1
Quercetalia out of AP	27.82	14.13	10.37
Conifers out of AP	0	5.43	28.89
Picea orientalis out of AP	72.18	57.61	41.48
Betulaceae and Juglandaceae out of AP	0	22.83	19.26
AP out of regional	50.57	42.99	25.76
Steppe out of regional	1.90	21.96	27.48
Desert out of regional	47.53	35.05	46.76
Hydrophil out of local	1.07	0	8.85
Gramineae and Cyperaceae out of local	58.01	56.41	38.54
Chenopodiaceae out of local	40.93	43.59	52.60
Local out of total	47.15	39.90	41.97
Spores out of total	8.72	5.37	0.77
AP out of total	22.32	23.53	14.75
Total counted	596	391	915
Number of samples	10	4	24

and western spectra are richer in arboreal components, in which *Picea orientalis* prevails but oaks also occur in significant proportions. Conifers other than *Picea* are more abundant to the south, together with pollen derived from steppe plants. All these indicate that the subtropical environmental gradient typical of the Miocene had been completely replaced by a different regime, the Pliocene, typified by rains coming from the north or northwest. Notably, the northwestern Mediterranean pollen reflect similar climatic trends (Fauquette et al. 1998).

6.5.2.4 *Quaternary*

The Quaternary palynozones show alternations of two principal types of pollen spectra. One is rich in arboreal components, mainly dominated by *Picea* at the beginning of the period, which is replaced upward by oak, characterizing Palynozones QI, QIII, QV, QVII and QIX. The other, in which non-arboreal steppe and desert elements prevail, typifies Palynozones QII, QIV, QVI, QVIII and QX. These alternations indicate oscillations of humid and dry Mediterranean climates: rather wet pluvials, during which even the southern part of the country became covered by steppe vegetation, while to the north *Picea* or later oak forests grew, most probably dominated by winter deciduous varieties (winter is stressed here because there are some summer deciduous trees in the southern Levant, such as certain acacias). The pluvial climates, when reaching their maximum effect, are typified by short cold, dry periods. The interpluvials, during which desert characterized the southern part of the Levant while to the north Mediterranean vegetation progressively covered the region, in a manner quite similar to the present-day situation but possibly somewhat drier; the interpluvials occasionally culminated in peaks of very dry climate when almost the entire country was a desert. Naturally, transition climates between these characterize some parts of the sequence. The lowermost Quaternary Palynozones, QI and QII, previously regarded as of "Preglacial Pleistocene" age (Horowitz 1979, p. 6), correspond to Foraminifera Zone N22, while the rest of the overlying sequence ("Glacial Pleistocene" and Holocene) to N23.

Palynozone QI (Figs 6.5.6 and 6.5.7), studied in detail by Levin & Horowitz (1987), is subdivided into three parts. The lower, QIa, is typified by very high proportions of arboreal pollen, sometimes exceeding 50% of the total, dominated, especially in its upper part, by *Picea orientalis*; the middle, QIb, is characterized by lower arboreal pollen shares, as compared with the underlying and overlying sequences, and higher percentages of steppe and desert elements. *Picea orientalis* is usually the common arboreal pollen, although *Quercetalia* are fairly frequent, especially in the middle part; the upper, QIc, resembles QIa by the very high arboreal pollen values, again dominated by *Picea orientalis*. The pollen spectra indicate a very humid, northeastern Mediterranean climate for QIa and QIc, somewhat drier during QIb. Similar trends are also seen in pollen diagrams from the northwestern Mediterranean (Fauquette et al. 1998).

Table 6.5.2.3 compares composite pollen spectra for Palynozone QIa, from three localities, the Hula Valley to the north, the southern Dead Sea basin and the southern Mediterranean offshore. As for the Pliocene, the north is richer in arboreal pollen, with dominating *Quercetalia*. The west and south display a greater influence of steppe vegetation, while the west is again considerably wealthier in Betulaceae–Juglandaceae. It seems that, commencing in the Pliocene, the present day environmental gradient of the southern Levant was already well established. The generally low landscape and the Nile are probably sources for the Betulaceae–Juglandaceae, as seen during the previous Pliocene times.

Palynozone QII (Fig. 6.5.8) is typified by very low arboreal pollen shares, comprising mainly *Picea orientalis* and some oak. Pollen from desert vegetation, dominated by Compositae, increase slightly as compared with the underlying QI, but among the regional vegetation steppe elements predominate. The pollen spectra indicate a dry Levantine Mediterranean climate, probably not very much different from the present day. Table 6.5.2.4 compares composite pollen spectra for Palynozone QII from two localities, the Hula Valley to the north and the southern Dead Sea basin. The offshore sequences for this part of the Quaternary were not studied palynologically in sufficient detail, and are thus not included in comparisons of this and the following palynozones. Clearly, arboreal pollen dominate the north, with increasing shares of oaks, while steppe prevails to the south.

Palynozone QIII (Fig. 6.5.8) displays higher arboreal pollen shares than the preceding QII, dominated by *Picea orientalis*, pine and oak. AP percentages are in the order of the present day or somewhat higher. The pollen spectra indicate a Mediterranean climate, somewhat wetter than at present.

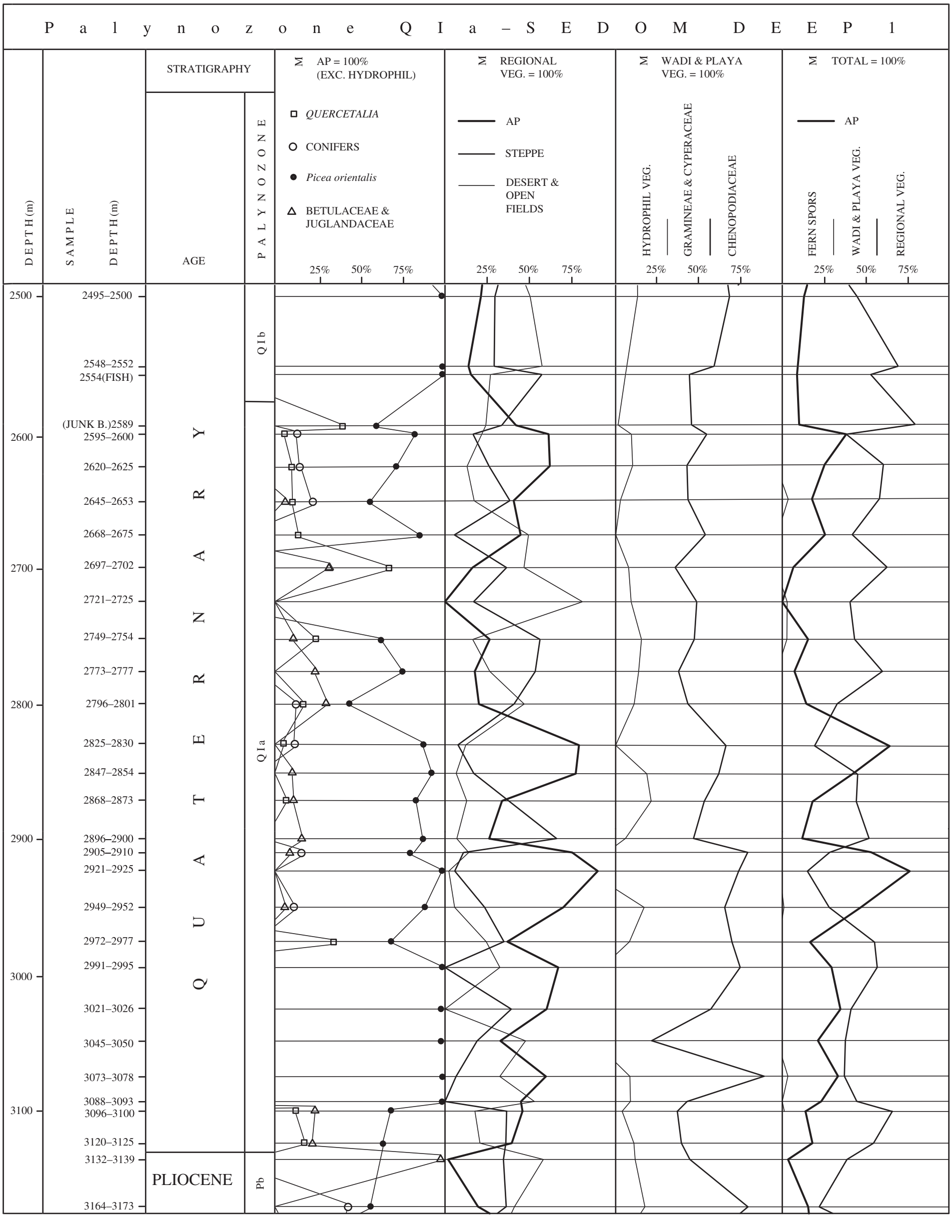


Figure 6.5.6. Pollen diagram of Palynozone QIa, Sedom Deep 1 borehole.

Palynozone QIV (Fig. 6.5.9) is characterized by a drop in arboreal pollen and a rise in desert vegetation pollen. Toward the top the latter decreases to give place to an increase of steppe elements. Oak is the main constituent of the arboreal pollen spectra, which indicate a dry Mediterranean climate.

Palynozone QV (Fig. 6.5.9) displays considerable peaks in the arboreal pollen curve, among the highest for the middle and late Pleistocene. The AP are

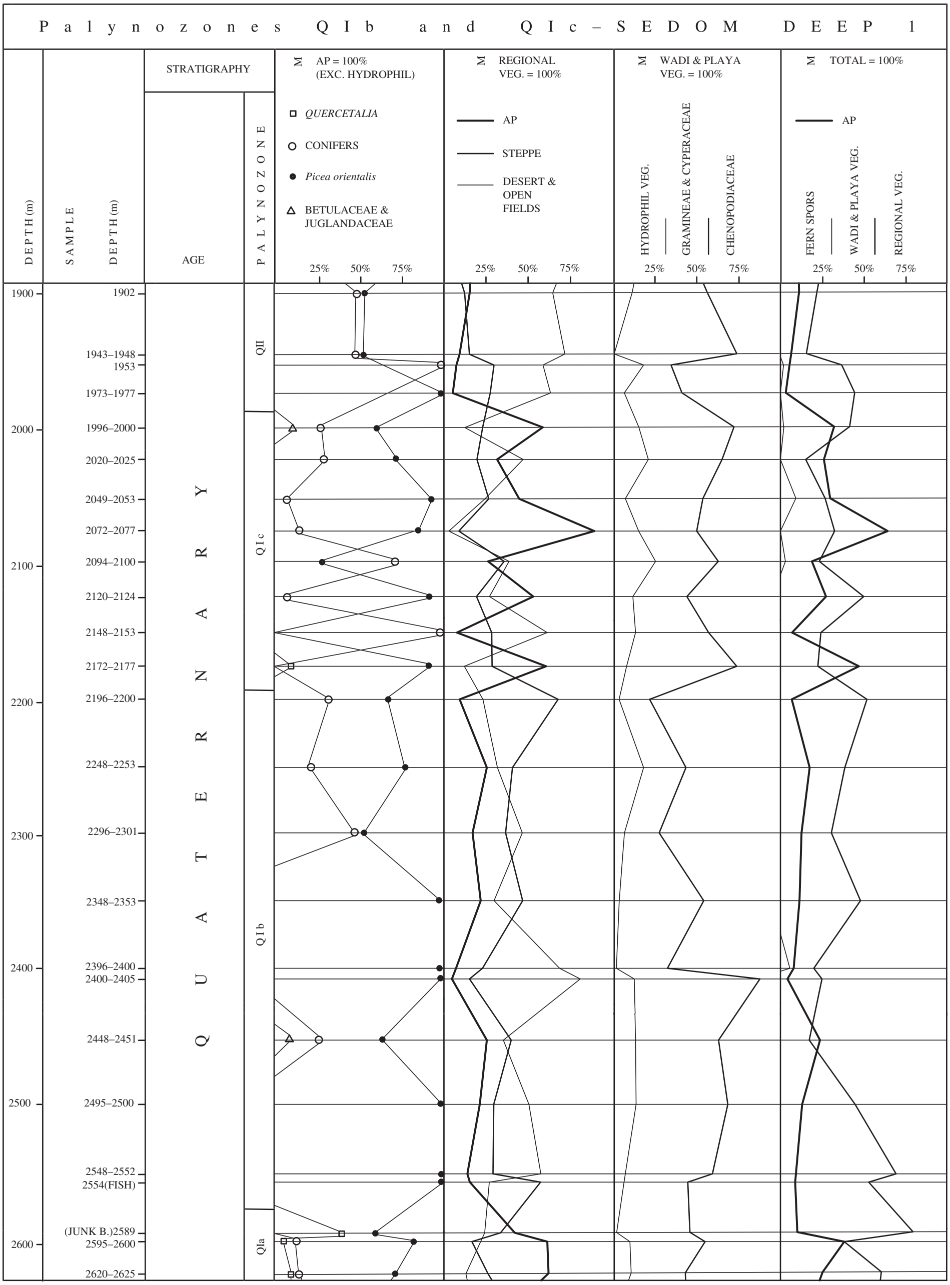


Figure 6.5.7. Pollen diagram of Palynozones QIb and QIc, Sedom Deep 1 borehole.

dominated by oak, especially the winter deciduous varieties, but pine also makes its appearance, most probably *Pinus halepensis*. Some last occurrences of *Picea orientalis* are still recorded in this palynozone. The QV sequence shows rather low values of desert and steppe vegetation pollen. The pollen spectra indicate a

Table 6.5.2.3. Percentages of pollen groups in composite borehole samples of Palynozone QIa, from the northern Jordan Valley (Notera 3), the Mediterranean offshore (Bravo 1) and the southern Dead Sea (Sedom Deep 1).

% of pollen/Borehole	Notera 3	Bravo 1	Sedom Deep 1
Quercetalia out of AP	46.96	25.42	5.88
Conifers out of AP	0	0	5.88
Picea orientalis out of AP	52.17	33.90	82.68
Betulaceae and Juglandaceae out of AP	0.87	40.68	5.56
AP out of regional	78.77	34.91	49.84
Steppe out of regional	0.68	28.99	27.20
Desert out of regional	20.55	36.09	22.96
Hydrophil out of local	15.87	0	9.25
Gramineae and Cyperaceae out of local	66.67	72.73	39.69
Chenopodiaceae out of local	17.46	27.27	51.06
Local out of total	29.86	38.17	45.57
Spores out of total	0.95	8.52	0.53
AP out of total	54.50	18.61	26.87
Total counted	211	317	1139
Number of samples	5	3	25

wet Mediterranean climate, considerably more so than at present, reminiscent of the Pontic environment, most probably with some summer rains. Table 6.5.2.5, which compares composite pollen spectra for Palynozone QV from the Hula Valley and the southern Dead Sea, shows environmental trends similar to those commenced in Palynozone QII times.

Palynozone QVI (Fig. 6.5.9) is characterized by relatively low AP shares along the sequence. The main arboreal pollen is oak, with subordinate *Pinus halepensis*. Desert plant pollen are quite abundant, and steppe vegetation grains show median values. The pollen spectra indicate a considerable deterioration in climate, as compared with QV. Table 6.5.2.6 shows no change in the previous environmental gradient.

Palynozone QVII (Fig. 6.5.10) is typified by an arboreal pollen peak composed of a mixture of oak and pine. The pollen spectra indicate an amelioration of the climate, toward a wet Mediterranean environment, less so than during QV but with the same gradient as before (Table 6.5.2.7).

Palynozone QVIII (Fig. 6.5.10) is characterized by rather low arboreal pollen shares, consisting of oak and some pine. The desert and steppe plant pollen curves show a peak, which decreases toward the upper part. The pollen spectra indicate a dry Levantine environment.

Palynozone QIX (Fig. 6.5.10) is characterized by very high arboreal pollen percentages, consisting mainly of the winter deciduous oak *Quercus ithaburensis*, accompanied by the evergreen *Quercus calliprinos* and some pine grains. Pollen of desert plants take a relatively small share of the spectra, while those of steppe plants have median values. The pollen spectra indicate a wetter climate as compared with the present day, but never again reaching the paradise of QV.

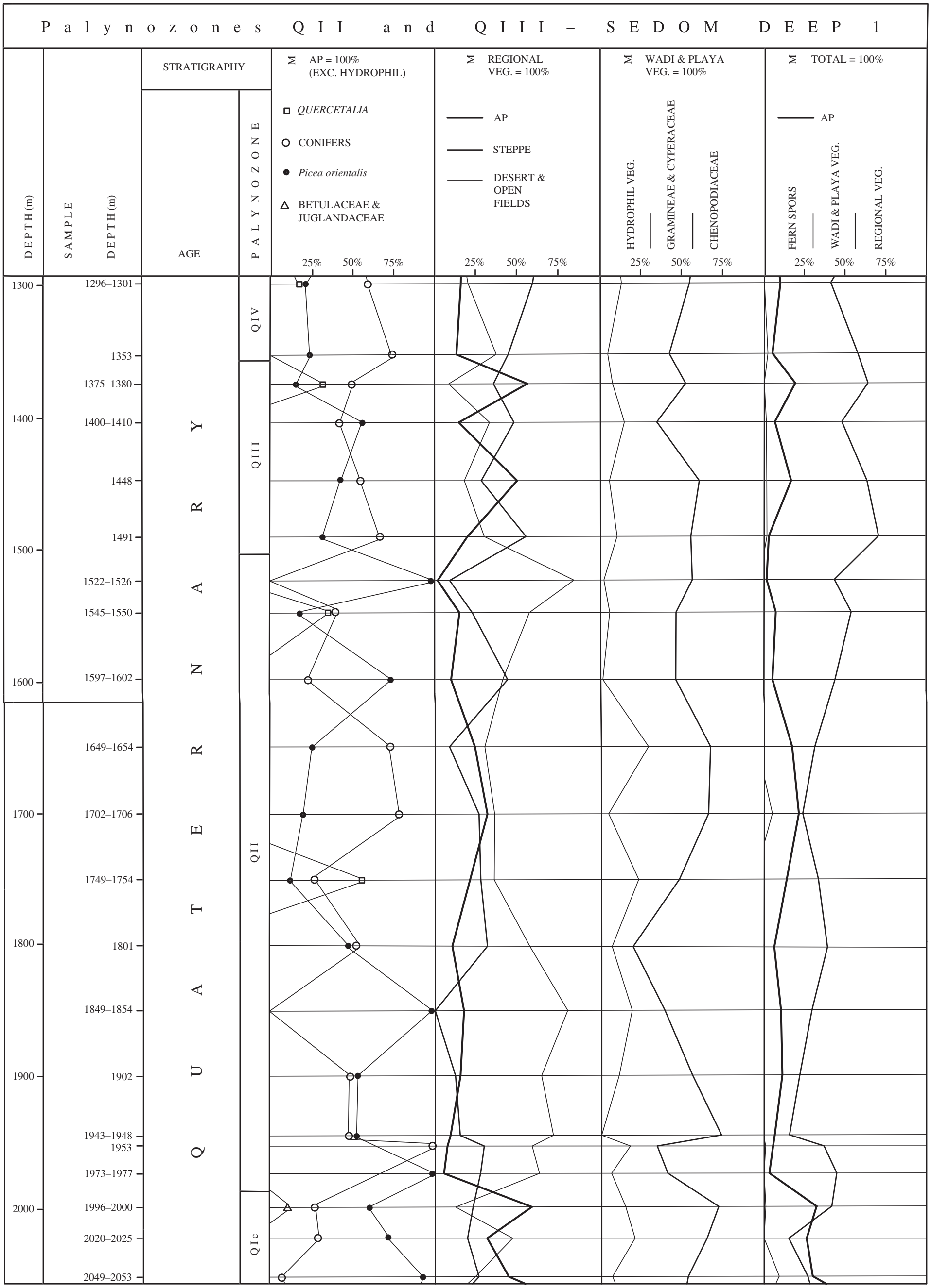


Figure 6.5.8. Pollen diagram of Palynozones QII and QIII, Sedom Deep 1 borehole.

Table 6.5.2.4. Percentages of pollen groups in composite borehole samples of Palynozone QII, from the northern Jordan Valley (Notera 3) and the southern Dead Sea (Amazyahu 1).

% of Pollen/Borehole	Notera 3	Amazyahu 1
Quercetalia out of AP	58.82	8.96
Picea orientalis out of AP	41.18	89.55
Betulaceae and Juglandaceae out of AP	0	1.49
AP out of regional	36.17	6.79
Steppe out of regional	4.26	64.54
Desert out of regional	59.57	28.67
Hydrophil out of local	1.47	13.83
Gramineae and Cyperaceae out of local	79.41	17.13
Chenopodiaceae out of local	19.12	69.04
Local out of total	58.12	42.73
Spores out of total	1.71	3.74
AP out of total	14.53	3.63
Total counted	117	1844
Number of samples	3	19

Palynozone QX (Fig. 6.5.10) shows very low AP and very high proportions of the pollen of desert plants particularly at the bottom. Both curves change direction toward the middle part of QX, which is followed by a drop in AP and an increase of steppe elements to their present-day values. The pollen spectra indicate slight fluctuations of the present-day type of climate and environments, within the north–south oriented environmental gradient (Table 6.5.2.8).

6.6 CLIMATOSTRATIGRAPHY

The entire sequence of late Cenozoic climatic fluctuations which have affected the southern Levant can be subdivided into two styles: long term, in the order of several million years duration, during the Oligocene, Miocene and Pliocene; and phases which persisted only for tens of thousands of years, or maximally 200–300 Ka, typical for the Quaternary. In a most general way, humid climates are characterized by high to medium arboreal pollen shares in the regional components diagrams, while desert environments are indicated by very low percentages, or sometimes the total absence, of AP, accompanied by dominance of pollen grains produced by desert plants, notably Compositae.

6.6.1 Oligocene and Neogene

The climatic development of the region throughout the Oligocene and Neogene (Horowitz 1992a, p. 373, 1996a) can be summarized as follows (Fig. 6.6.1), based

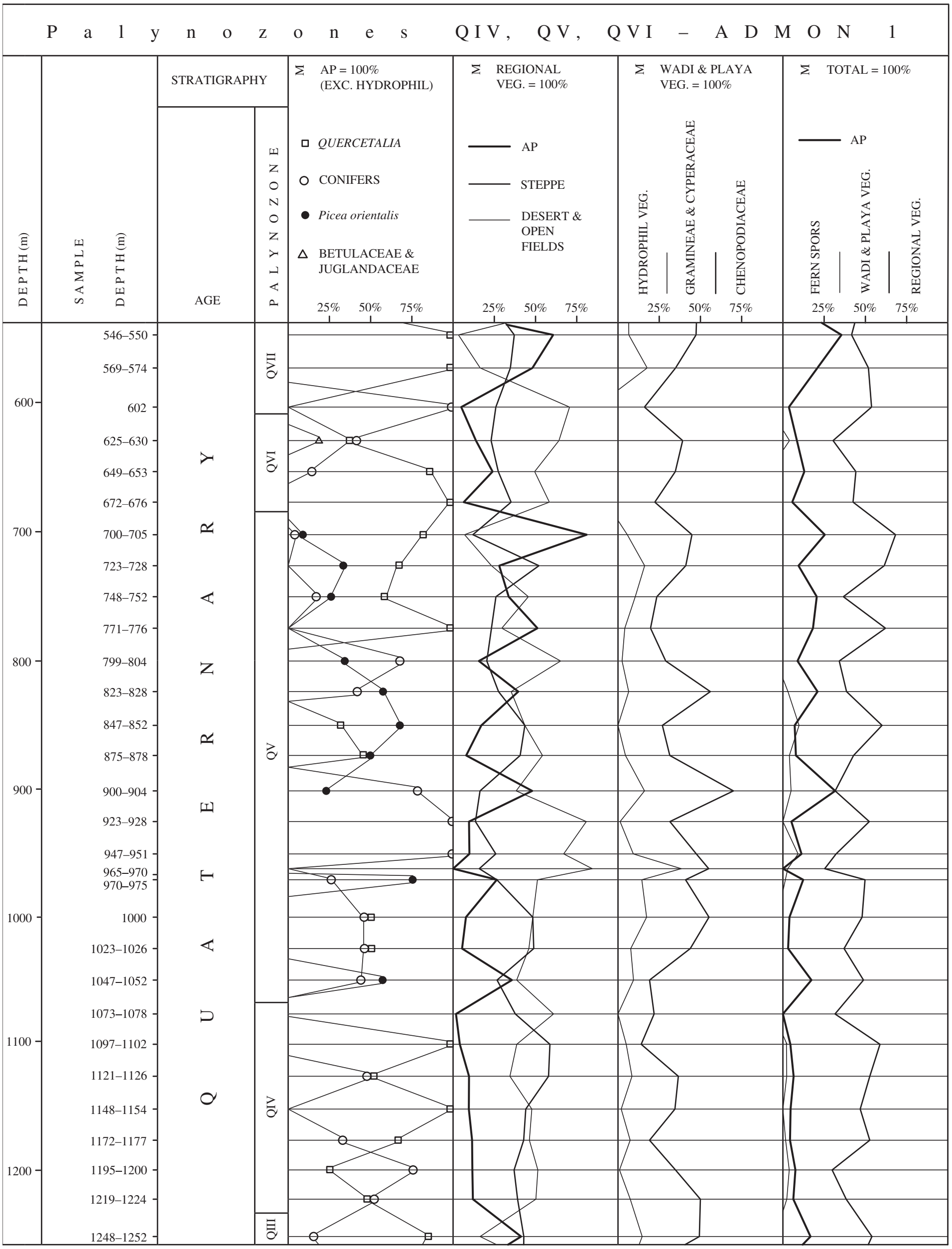


Figure 6.5.9. Pollen diagram of Palynozones QIV, QV and QVI, Admon 1 borehole.

on the pollen diagrams (Figs 6.5.1–6.5.5). The middle Oligocene enjoyed a wet subtropical, maybe even truly tropical, climate. A desert, occasionally oscillating between dry savanna and an extremely arid environment, existed during Palynozone Ma times, latest Oligocene through early Miocene. Throughout this

Table 6.5.2.5. Percentages of pollen groups in composite borehole samples of Palynozone QV, from the northern Jordan Valley (Notera 3) and the southern Dead Sea (Sedom Deep 1).

% of pollen/Borehole	Notera 3	Sedom Deep 1
Quercetalia out of AP	84.81	40.96
Conifers out of AP	14.92	54.22
Picea orientalis out of AP	0	4.82
Betulaceae and Juglandaceae out of AP	0.28	0
AP out of regional	84.78	27.30
Steppe out of regional	1.17	50.00
Desert out of regional	14.05	22.70
Hydrophil out of local	4.01	10.62
Gramineae and Cyperaceae out of local	80.45	22.86
Chenopodiaceae out of local	15.54	66.51
Local out of total	47.96	57.89
Spores out of total	0.72	1.47
AP out of total	43.51	11.10
Total counted	832	1496
Number of samples	12	9

Table 6.5.2.6. Percentages of pollen groups in composite borehole samples of Palynozone QVI, from the northern Jordan Valley (Notera 3) and the southern Dead Sea (Amazyahu 1).

% of pollen/Borehole	Notera 3	Amazyahu 1
Quercetalia out of AP	63.30	80.00
Conifers out of AP	32.70	5.00
Picea orientalis out of AP	0	14.44
Betulaceae and Juglandaceae out of AP	0	0.56
AP out of regional	39.55	18.97
Steppe out of regional	4.48	38.67
Desert out of regional	55.97	42.36
Hydrophil out of local	7.48	13.79
Gramineae and Cyperaceae out of local	76.00	26.10
Chenopodiaceae out of local	16.52	60.11
Local out of total	66.08	66.54
Spores out of total	4.28	1.45
AP out of total	11.73	6.07
Total counted	1358	2965
Number of samples	19	22

time the south witnessed more rainfall than the north, most probably of monsoon origin. Water availability gradually increased in Mb times, latest early Miocene through middle Miocene, when the lowlands were covered with a wet savanna of palms, Juglandaceae and Betulaceae, and probably also other trees which are not represented by fossils, while conifers grew on the drier highlands. The north continued to be drier than the south. This environment changed entirely toward Mc times, late Miocene, when hardly any arboreal pollen at all are recorded.

Table 6.5.2.7. Percentages of pollen groups in composite borehole samples of Palynozone QVII, from the northern Jordan Valley (Notera 3) and the southern Dead Sea (Sedom Deep 1).

% of pollen / Borehole	Notera 3	Sedom Deep 1
Quercetalia out of AP	81.73	97.26
Conifers out of AP	18.27	2.74
AP out of regional	75.06	30.42
Steppe out of regional	7.98	37.92
Desert out of regional	16.96	31.67
Hydrophil out of local	10.29	12.98
Gramineae and Cyperaceae out of local	73.86	24.86
Chenopodiaceae out of local	15.85	62.15
Local out of total	54.70	59.93
Spores out of total	8.30	0.33
AP out of total	27.77	12.09
Total counted	1084	604
Number of samples	11	6

Pollen diagrams are dominated by Compositae, indicating arid to extremely arid environments, probably a Sahara-type desert for that period. The Oligo-Miocene climatic cycles are in the order of six million years duration, with only subordinate fluctuations.

In Pliocene times climatic cycles became shorter, in the order of 1–1.5 million years. The entire region was subject to temperate conditions, with rains arriving from the northern, European direction. Arboreal pollen spectra are dominated by such trees as *P. orientalis* (or *Abies*?) and oaks, while the environmental gradient became drier southward. Thus Pa times, early through middle Pliocene, enjoyed a rather cool, humid, temperate climate, while during Pb, late Pliocene, the temperate environment became drier and possibly somewhat warmer.

6.6.2 Quaternary

In contrast with the frequency of Oligocene and Neogene climatic cycles, Quaternary oscillations are much shorter in duration, seemingly becoming progressively shorter as the present day is approached. The reason for the latter phenomenon may well be artifactual, due to the considerably larger body of data acquired for the later part of the Quaternary, as compared with its earlier periods. Oxygen-isotope curves for the last 3.5 million years (Shackleton & Hall 1984, Shackleton et al. 1990) seem to show equal durations of all oscillations. However, the Brunhes Chron is characterized by oscillations of larger magnitudes, which are more readily distinguished as climatic stages on land. Thus the earlier palynozones encompass numerous oxygen isotope stages, their number gradually diminishing the younger they become (Fig. 6.6.2).

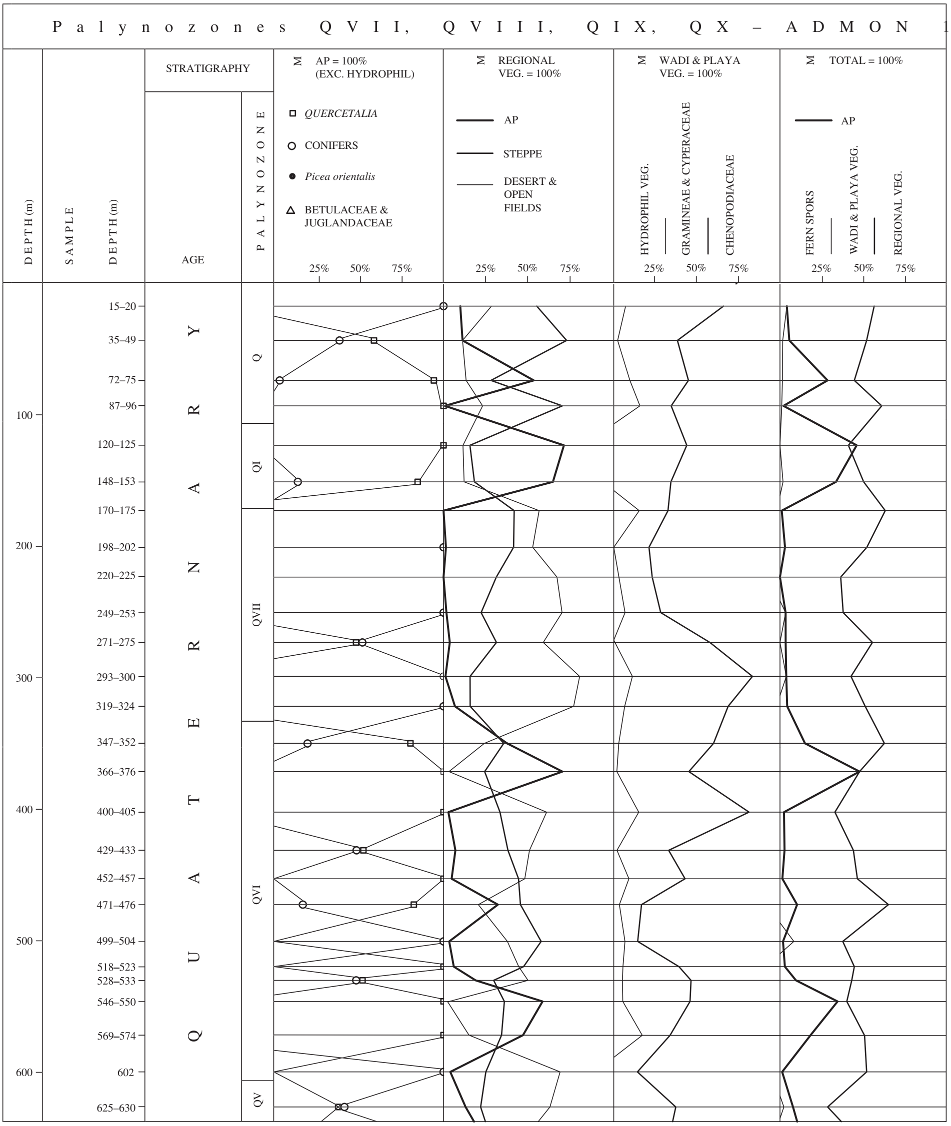


Figure 6.5.10. Pollen diagram of Palynozones QVII, QVIII, QIX and QX, Admon 1 borehole.

Several periods are recorded for the Quaternary, at the peaks of which almost the entire southern Levant had turned to a desert. These interpluvial phases characterize Palynozones QII, QIV, QVI, QVIII and QX, separated by humid pluvial periods during QIII, QV, QVII and QIX. The interpluvials are occasionally interrupted by short phases, somewhat more humid than the arid stages, but not as much as the pluvials. The Quaternary desert phases of the southern Levant seem to correlate quite well with the European interglacials, as recorded in deep sea

Table 6.5.2.8. Percentages of pollen groups in composite borehole samples of Palynozone QX, from the northern Jordan Valley (Notera 3) and the southern Dead Sea (Amazyahu 1).

% of pollen/Borehole	Notera 3	Amazyahu 1
Quercetalia out of AP	94.44	70.21
Conifers out of AP	5.56	29.79
AP out of regional	50.70	19.50
Steppe out of regional	4.23	51.87
Desert out of regional	45.07	28.63
Hydrophil out of local	6.41	12.11
Gramineae and Cyperaceae out of local	76.92	20.79
Chenopodiaceae out of local	16.67	67.11
Local out of total	45.61	60.80
Spores out of total	12.87	0.64
AP out of total	21.05	7.52
Total counted	175	625
Number of samples	2	4

cores (Horowitz & Weinstein-Evron 1986; Cheddadi 1988; Horowitz 1992a, p. 373), in the Alps (Horowitz 1979, p. 173; Horowitz & Horowitz 1985; Horowitz 1992a, p. 383) and in speleothems of Israel (Bar-Matthews et al. 2000).

The information available on Quaternary paleoclimates in the southern Levant is much more extensive than for the Neogene, coming from several disciplines which seem to corroborate each other, such as geology, pedology, geomorphology, paleontology, palynology and isotope analyses (Horowitz 1992a, p. 374; Bar-Matthews et al. 2000). These are summarized in Table 6.6.1. Thus two types of "end member" climates have been discerned (Horowitz 1979, p. 344) for the Quaternary of the southern Levant: interpluvial and pluvial, occasionally interrupted by stadials or interstadials. All comparisons are made with present-day environments, taken as reference. At present it seems more appropriate to discuss the average characteristics of the interpluvials and pluvials, which span most of the period discussed, while regarding the stadials/interstadials as drier or wetter phases, respectively.

Pollen diagrams obtained from Israel for the last 150 Ka, dated by radiocarbon for their upper parts and uranium series for the lower (Horowitz & Weinstein-Evron 1986), have shown quite clearly that pluvial climates of the southern Levant correspond to the European and oceanic glacial phases, while interpluvials correspond to the interglacials, both interrupted by phases of opposing trends of rather short duration. To the north, particularly in the higher elevations of northern Syria, Iran and Turkey, the glacials are recorded in pollen diagrams as dry periods (van Zeist & Bottema 1988), a conclusion drawn from the paucity of arboreal pollen.

However, El Moslimany (1990) indicates that trees growing in the Zagros mountains of Iran are more sensitive to heavy snowfall than to drought, thus

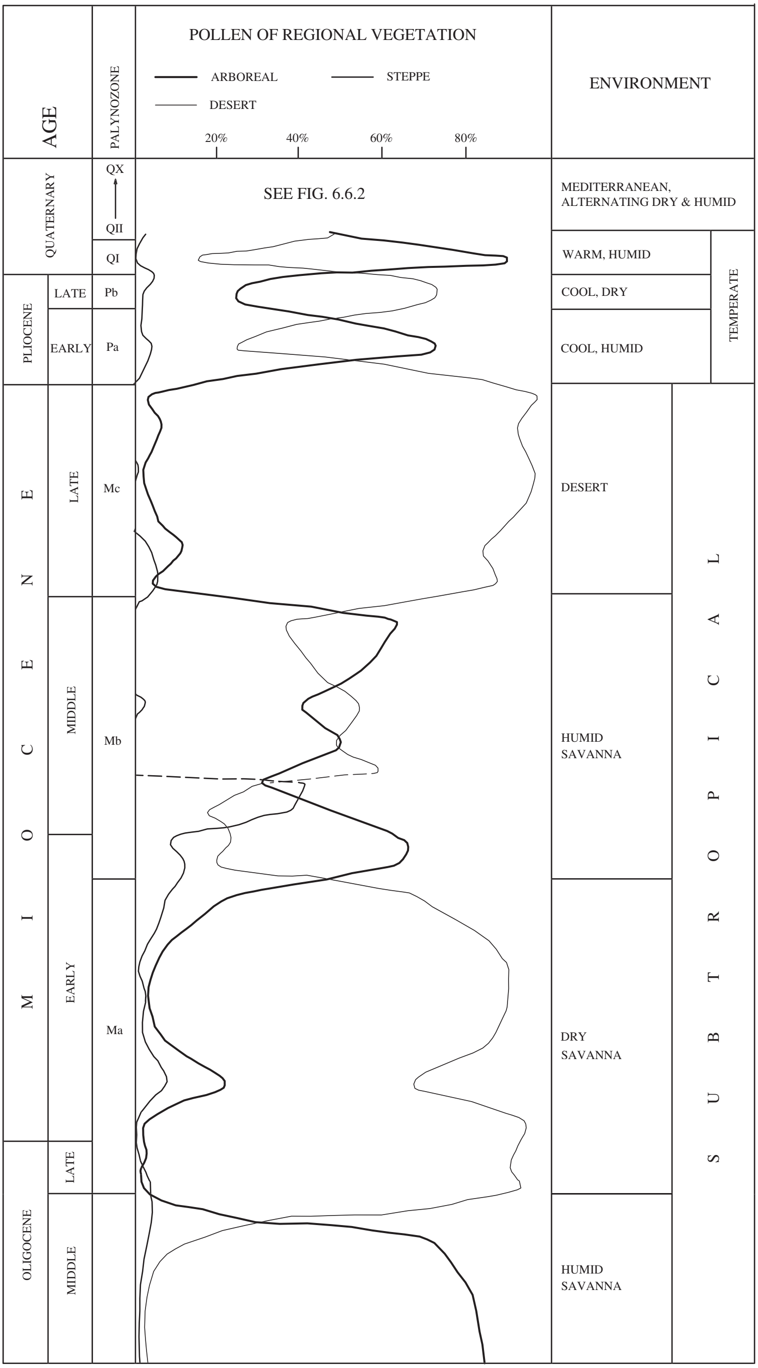


Figure 6.6.1. Oligocene and Neogene palynozones and climates of the southern Levant, based on pollen diagrams of regional vegetation (=100%) from the Bravo 1 (Mediterranean offshore) and Sedom Deep 1 (southern Dead Sea) boreholes, not to scale. Curves are overlapping.

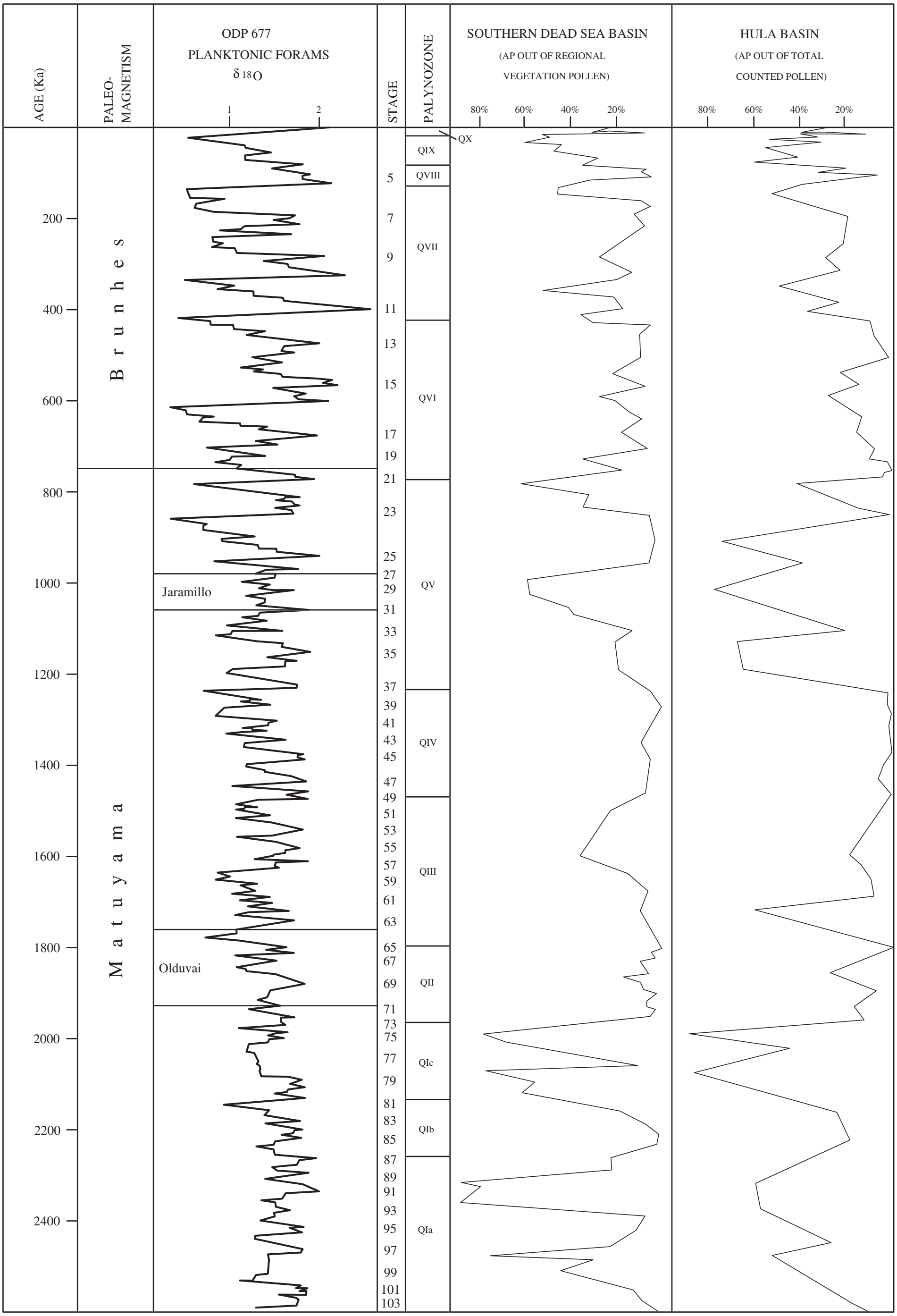


Figure 6.6.2. Quaternary oxygen-isotope curves from the oceans, compared with major trends of pollen diagrams from the Jordan Rift Valley.

Table 6.6.1. Characteristics of pluvial and interpluvial climates of the southern Levant.

	Pluvial	Interpluvial
Vegetation to the north, according to pollen spectra	Deciduous oak forests	Evergreen oaks and Mediterranean maquis
Vegetation to the south	Steppe	Desert
Rift Valley lakes	Extensive	Restricted
Fluviatile sediments	Aggradation	Erosion
Soils	Red, well leached	Gray, calcareous
Average $\delta^{18}O$ of speleothems (PDB)	−3‰ to −4‰	−4‰ to −6‰
Rain regime	Moderate rains all year around	Winter thunderstorms
Origin of rains	Possibly Atlantic Ocean	Mediterranean

concluding that the last glacial was indeed not arid but, on the contrary, was characterized by considerable winter precipitation, which at higher elevations caused heavy snow accumulations that hindered the growth of trees, particularly oaks, in that region. The data from Israel support Van Zeist & Bottema's view, since the maxima of glacials are recorded here as cold dry periods, judging not only from pollen but also from the types of sediments and from shrinkage of the Jordan Rift Valley lakes (Horowitz 1973, Kislev et al. 1992, and see below).

Several researchers claim that glacials correspond to dry climates in the southern Levant (Gat & Magaritz 1980), or that interglacials were humid (Livnat & Kronfeld 1985). Gat & Magaritz observed a dry climate phase in Israel 18 Ka ago, which indeed was also seen in pollen diagrams from the Hula (Horowitz 1971, Weinstein-Evron 1990). This was a very short phase, however, that may reflect the influence of the very cold climate to the north, advancing southward at the peak of the last glacial maximum (LGM). Weinstein (1976) pointed out that this last phase of the Pleistocene was markedly colder than most preceding pluvial periods. It is possible that short, dry phases would occur in the region at the peaks of glacials, but in general periods recorded by low sea levels enjoyed humid, pluvial climates. Since these short dry phases (or also, short wet phases within an interpluvial) are very limited in duration, apparently occurring only in extremes of both pluvial and interpluvial climates, they were considered as interstadials and stadials.

Weinstein (1976), Weinstein-Evron (1983, 1990) and Bar-Matthews et al. (2000) also showed that climatic stages, both pluvials and interpluvials, are frequently interrupted by short phases, which would inflict an almost opposite paleoclimate for a given phase. It is therefore essential that conclusions be drawn from longer continuous sequences, rather than from point evidence, as seems to be the basis for all those studies that appear to contradict the general paleoclimatic trends. It is frequently tempting to study these opposing subordinate oscillations, since they are quite prominent in the landscape.

Bar-Matthews et al. (2000) have indeed analyzed a continuous speleothems sequence, still maintaining that during the pluvial, not its extremes, rain quantity was lower than at present. On the other hand they state that the cave enjoyed a constant water supply over the year. Their conclusions are based on recent processes (Ayalon et al. 1998), when two styles of water infiltration into the cave are observed: fast, typical of the strong winter rains and slow, seepage water that remains in the upper vadose zone for up to several decades. Since the pluvial speleothems only indicate the slow drip, the authors concluded there was less rain. In view of the rich arboreal vegetation and extensive pluvial lakes, and also in line with suggested climatic models, it seems that what is recorded in the cave deposits could simply mean no thunderstorms, but rather gentle rains all through the year. In this case the quantity of water actually reaching the cave could be decreased, since such a rain style is used much more efficiently by the considerably richer vegetation.

Correlation of the Levantine interpluvials with odd numbered oxygen-isotope stages, and pluvials with even numbered, was suggested by Weinstein-Evron (1983, 1990), Horowitz & Weinstein-Evron (1986), Cheddadi (1988) and Frumkin et al. (1997), recently corroborated by Bar-Matthews et al. (2000). These studies dealt mainly with the late Quaternary, the last ca. 150 Ka, while others (Fuji & Horowitz 1989, Horowitz 1989c) extended these correlations through the Brunhes Chron and further, to 3.5 million years ago. It seems that a good correlation exists between the climatic curves for the Jordan Valley, based on relative distributions of pollen components of the regional vegetation, and oxygen-isotope curves obtained for that period from the oceans (Shackleton & Hall 1984, Shackleton et al. 1990), as can be seen in Fig. 6.6.2.

Correlation of the humid pluvials of the southern Levant with the European glacials (and vice versa), is even clearer where the Levantine coastal plain is concerned. Red, well-leached soils and accumulations of river terraces, both typical of humid climates all over the Levant, always occur in tandem with low sea levels, typical of glacial times (Horowitz 1979, p. 346; Besançon & Sanlaville 1984).

6.6.2.1 *Interpluvial climate*

Interpluvial climates, of which the present day is a fair representative, are characterized by pollen spectra which display some 10–20% arboreal components to the north of the country, where Mediterranean environments prevail, diminishing southward, to the Negev, where no AP occur at all, reflecting the desert conditions in that region (Fig. 6.6.3A). The situation, as noted above, is different in the Dead Sea, which although surrounded by a desert, yields pollen spectra typical of the Mediterranean domain, due to long-range transportation by the winds and the Jordan River, aided by the trapping effect of the lake's water surface (see above).

The major contributor to recent arboreal pollen is the evergreen oak, *Quercus calliprinos*, partly replaced by pine on the coastal plain, depending on the prevailing

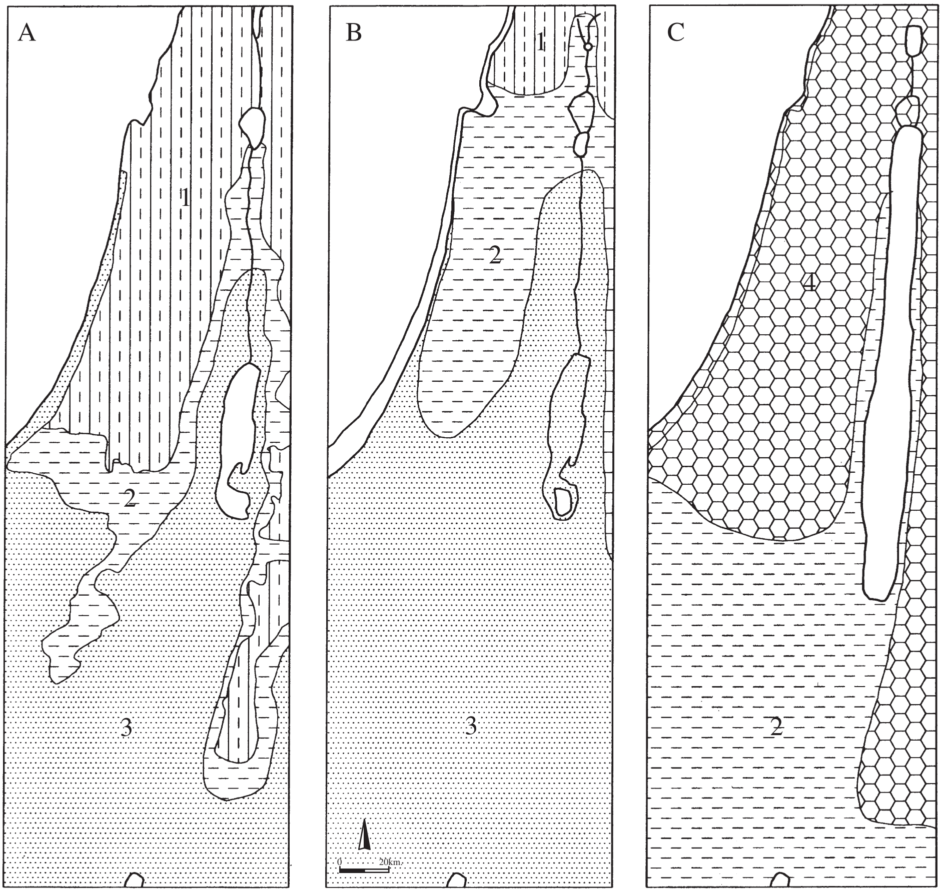


Figure 6.6.3. Vegetation of the southern Levant: (A) present day, also representing average interpluvial conditions. (B) reconstructed extreme interpluvial and (C) reconstructed pluvial; (1) Mediterranean maquis, (2) Irano-Turanian steppe, (3) Saharan desert and (4) Mediterranean forests, mostly deciduous oaks. Thick lines: schematic reconstruction of Mediterranean shorelines and Jordan Valley lakes.

wind directions during flowering seasons (Weinstein 1979). Lakes are present in the Jordan Rift Valley in three localities: the Hula to the north, Lake Kinneret at the northern end of the central Jordan Valley and the Dead Sea to the south, all three rather limited in area (see Chapter 12). Typical soils of the present day are gray, lime-rich, with a rather shallow profile and very little organic matter in the Mediterranean domain, while to the south bare rocks make most of the landscape. Wadis are subject to erosion caused by the frequent winter floods, as indeed is almost the entire southern Levant. Loess is eroded by wind from its former accumulations in southern Israel, and in no place are any significant new deposits being formed. Eolian sediments that do accumulate at present are sand dunes along the coastal plain, penetrating landward where not obstructed, and also in several parts of the Arava. Speleothems of the Soreq Cave (Bar-Matthews et al. 2000) are characterized by intermittent growth of crystals, which contain fine clastics impurities. Average $\delta^{18}O$ values for the speleothems are between $-4‰$ and $-6‰$ (PDB), while wetter phases within the interpluvials are characterized by increasing amounts of isotopically light rains, with average $\delta^{18}O$ values of $-6‰$ to $-8‰$.

Peaks of the interpluvial climates in the southern Levant are characterized by pollen spectra which, compared to the present day, are extremely poor in arboreal constituents, with none to the south and only some 3–5% to the north of the region. The prevailing components of interpluvial assemblages are pollen of steppe plants to the north and desert to the south, reflecting distribution of their parent vegetations (Fig. 6.6.3B). The Jordan Rift Valley lakes were restricted to shallow marshes to the north, while a terminal hypersaline lake existed in the Dead Sea region, both subordinate in area, their sediments known only from boreholes. Hardly any paleosols, wadi deposits or loess are known for the interpluvials, due to erosion that prevailed throughout the entire region. The extreme climate responsible for this suite of phenomena must have had very little rains, the rare ones which did reach the region brought by occasional, possibly strong thunderstorms, hitting only the northern parts of the region.

6.6.2.2 *Pluvial climate*

Pluvial climates are characterized by pollen spectra considerably richer in arboreal components as compared with the present day. Arboreal pollen also occur in sediments from the south of the country, in proportions similar to the present-day northern regions, while to the north the percentage in pluvial times is more than double than at present. The main increase in arboreal pollen is derived from winter deciduous oaks, growing in the much expanded Mediterranean environment (Fig. 6.6.3C).

Mediterranean sea levels were much lower than at present, thus continental sediments are found far west of the recent coastline. Lakes in the Jordan Rift Valley occupied considerably larger areas than today, sometimes covering the entire valley floor. Their sediments are known from many localities all over, both from outcrops and boreholes. Paleosols are quite abundant, comprising mainly red loams, also known from the southern regions of Israel (Magaritz 1986), where hardly any soil is formed today. Gravel and silt were being accumulated in the wadis in considerable quantities (Besançon & Sanlaville 1984, Goldberg 1986), filling and silting up previous channels, which had been cut down by the floods during preceding interpluvials. This process is known from all over the Levant, extending southward to southern Sinai and northward up to Turkey. Loess was accumulated in sequences several tens of meters thick (Ginzbourg 1963), mainly in the northwestern Negev, but extending to the central Negev as well, in thinner sheets as one goes south. The loess was being subject to pedogenic processes, quite extensive in the northern Negev, less so to the south. Speleothems of the

Soreq Cave (Bar-Matthews et al. 2000) are characterized by large crystals, practically devoid of clastic debris, indicating a slow but constant supply of water. Average $\delta^{18}O$ values for the speleothems are between −3‰ and −4‰ (PDB), while the peaks of the pluvials are characterized by decreasing amounts of isotopically heavy rains, with average $\delta^{18}O$ values of −2.5‰ to −3‰. These isotopic compositions indicate that rains at that time originated in a cold sea, possibly the Atlantic Ocean, rather than the Mediterranean during the interpluvials, some arriving in the southern Levant in summer time also. During the short peaks of glacial maxima, such as the LGM some 19–18 Ka ago, the southern Levant became cold and dry.

The pluvial climate must have been considerably more humid than the present day (or other interpluvials), with rains which also reached the south of the country. Rains were mostly gentle and lengthy, probably originating from warm fronts, while thunderstorms seem to have been quite rare. It should be stressed that "humid" does not necessarily entail larger absolute quantities of rain. It is sufficient for plants that the same amount would be equally distributed over the entire year and the rains be gentler (as we do with artificial irrigation), to allow for considerably more prolific growth. The question remains, however, whether lakes of the Jordan Valley could expand to their pluvial sizes without an increase in the absolute rain quantities, a problem pertaining as well to other pluvial phenomena. The increased summer cloudiness and the generally lower temperatures could be the answer, but it is still not certain whether these alone would be sufficient, or some increase in rain quantity would be required.

6.7 PALEOCLIMATIC MODELS

Modeling of the late Cenozoic climates of the Levant should account for both the long-range variations during the Neogene, and the short duration ones typical of the Quaternary. During the Tertiary the global climatic belts had slowly evolved until they became well established by the beginning of the Neogene. Horowitz (1977a) suggested a connection between the development, contrast and magnitude of the climatic belts and geotectonic processes, principally mountain building and uplift or relaxation of latitudinal chains. Increased activity of these processes during the Quaternary caused an accentuation of the effects of the Milankovitch (1941) curve, which resulted in greater contrasts between the different global climates, culminating in the onset of glacials and interglacials.

6.7.1 Neogene climates

While numerous models were suggested for the global Quaternary paleoclimates, especially for the alternating glacials and interglacials, less is known about the Neogene, probably due to the relative poverty of data acquired from continuous continental sequences for that period. Most studies of Neogene paleoclimates concentrated on marine sediments and faunas, while pollen diagrams from uninterrupted continental (or even marine, for that matter) Neogene sequences are only rarely available. It is very difficult and sometimes quite hazardous to assume continental climates exclusively from records which reflect only marine environments, very frequently differing from conditions on land. Just an example is the occurrence of coral reefs, generally regarded as representing tropical to subtropical climates. Tropical, in many instances, is tied up almost automatically with rain forests and the like, but one must remember that corals flourish happily in places where the surrounding continent is a bare, extremely arid desert, for instance the extensive reefs of the Red Sea.

Attempts to acquire data and come to conclusions on Neogene paleoclimates from fossil plants seem to pose yet another problem. Most localities of fossil plants analyzed, if not all, are intimately connected with freshwater sources, namely springs, lakes or marsh deposits. Plants typical of such environments, both arboreal and non-arboreal, are much less sensitive to the general climate as long as water is locally available. Thus when successions of spring deposits or marsh sediments were analyzed, the conclusion was that throughout the Neogene there was hardly any climatic change (Gregor & Velitzelos 1987), or that any changes were subordinate (Velitzelos & Gregor 1987).

Apparently the major difference in the global circulation pattern between the Neogene and Quaternary is the onset, during the latter period, of such critical conditions as would amplify the effects of variations in solar insolence, described by the Milankovitch (1941) curve, toward the formation of glaciers on the one hand and planetary deserts on the other. A possible trigger for such a mechanism could be (Horowitz 1977a) the global uplift of the Alpine orogeny (Guillaume & Guillaume 1982), which barred the way of hot, humid, tropical air beyond certain latitudes and created Hadley cells, typical of today's atmospheric circulation, resulting in considerable accentuation and contrasts in the global climatic belts.

It seems that the causes of the Neogene environmental shifts, of rather long duration, are hardly relevant for the Quaternary, and that the latter climatic oscillations are best explained by amplified effects of the Milankovitch curve. Causes of the Neogene shifts should possibly be sought in the late Tertiary–Quaternary gradual development of global climatic belts, when the Earth changed from a more uniform subtropical environment to the present day configuration of prevalent latitudinal zonation. The development of global climatic belts seems to be intimately connected with uplifts of generally east–west oriented orogenic belts, such as the Alps and Himalayas.

During the Oligocene and Miocene, the Levant was situated within the then widely distributed subtropical monsoon system, with long term alternations of dry and wet, culminating in a Sahara-type desert in the course of the Miocene Ma and Mc palynozones. This is based on the characteristics of Levantine floras (Horowitz 1990), indicating subtropical environments, which are also typical of the northern and northwestern Mediterranean regions in southern Europe at that time (Bessais & Suc 1987, Benda & Meulenkamp 1990).

The middle Miocene amelioration of desert conditions in the Levant, typical of Mb times, was probably caused by global cooling at that time, corresponding to the initial glaciation of Antarctica (Van Zinderen Bakker & Mercer 1986), rather than by migrating climatic belts, as thought previously. It seems highly plausible that the middle Miocene glaciation of Antarctica was caused by some accentuation of the contrast between the climatic belts, resulting from the uplift of the Alps at that time. Such an increased contrast was not sufficient to create glaciation of the North Pole, which is at sea level, but could initiate ice accumulation on Antarctica, due to its considerably higher elevation.

In the course of the Pliocene the global climatic belts regime was further established, the Levant was affected by the southern European temperate domain, which was wetter at the beginning, Pa, becoming dryer toward the late Pliocene, Pb. Pollen analyses of northern Mediterranean regions indicate a pronounced cooling for the Pliocene (Suc 1989). This change is part of a global trend, expressed in a gradual contraction of the tropical belt throughout the Cenozoic (cf. Traverse 1988, p. 291), but was also accentuated by the regional extensional tectonics, opening a path for polar air to the Levant (Mercier et al. 1987).

6.7.2 Quaternary climates

The primary point is, why the great difference in frequency and amplitude between the Quaternary short-term oscillations and the previous, considerably longer and more subtle climatic changes? It is widely accepted that the Quaternary oscillations follow astronomical changes in relations between the Earth and the Sun, first calculated by Milankovitch (1941), followed by many. But these relations existed also before the Quaternary, so an additional mechanism seems to be involved in the considerable amplification of their influence during the last 2.5 million years. Horowitz (1977a) proposed that this supplementary mechanism depends on geotectonic processes, causing considerable uplift of the Alpine and other latitudinal mountain chains, at the beginning of the Quaternary (Mercier et al. 1987). This uplift effectively blocked movements of warm air from the equator to the north, thus considerably strengthening the activity and influence of the resulting Hadley cells. By the natural general symmetry of the atmosphere these effects are also seen in the southern hemisphere, although in a slightly milder way.

Periods when higher amounts of solar radiation were absorbed by the Earth, as is the present day and the interglacials, are typified by increased activity of the tropical monsoon belt, resulting in greater amounts of rains in this region. Concurrently, this activity inflicted higher barometric pressures both north and south of the equator by the descending hot air, which evoked formation of the planetary deserts. The warmer temperate regions at that time caused melting of the polar ice caps. Lower amounts of radiation reversed the process, so that less rains are recorded in the tropics, the planetary deserts enjoyed some rain, while the colder temperate regions caused accumulation of snow and ice at the poles, resulting in a glacial climate.

These processes were further amplified by the effects of albedo, both in the planetary deserts and the polar regions. The former, due to their poorer vegetation in interglacial times, could absorb very little solar radiation, thus amplifying the drying effects of the descending hot air, and by not being able to produce low pressure necessary for bringing rain, they further increased aridity. In the same way, the larger areas covered by snow during glacial times could again absorb very little radiation, which resulted in further cooling. Both these amplification processes could start acting only when certain limiting values were reached, which was enhanced by the greater activity of the Hadley cells, an activity increasing as a function of the uplifting latitudinal mountain chains during the Quaternary.

Notably, earlier ice ages in the history of the Earth are also connected with periods of latitudinal mountain building, such as the late Precambrian, the Silurian and the Permian, when orogenies caused accentuation of the climatic belts, resulting in widespread polar glaciations.

Climatic models for the Quaternary of Israel, based on numerous previous studies and suggestions, are summarized in Horowitz (1992a, p. 373). However additional data, mainly obtained from isotopic measurements, geochemistry and mineralogy, accurately and continuously dated, inspired some modifications of the earlier models. Also, the previous discussions centered on understanding the "end members" of Quaternary climates, which caused some confusion since during most of the time it is the transitions that count, while the extremes only existed for short periods. It now seems more reasonable, particularly when the additional data are at hand, to give more attention to the longer periods of glacial and interglacial climates rather than the short peaks treated earlier. Most of the new information was acquired by a team from the Geological Survey of Israel, headed by M. Bar-Matthews and A. Ayalon, who meticulously studied speleothems from the Soreq Cave in the Judean Hills west of Jerusalem, spanning (for the time being) the last 120 Ka (Bar-Matthews et al. 1997, 1998a,b, Ayalon et al. 1999, Bar-Matthews et al. 2000). Radiometric dating was carried out by A. Kaufman, the Weizman Institute for Sciences, Rehovot.

The point of reference is the present-day conditions, for which the climate can be observed and measured; inferences for the past are based on the proxy data described above, for which the basis is the pollen spectra and diagrams of the various palynozones. Other proxy data, such as isotopic compositions of speleothems, water and fossils, paleosols, and modes of erosion and deposition, wherever available, have also been included.

At present, rain falls in the southern Levant only in wintertime, precipitating annually up to 800 mm in the north of Israel, diminishing gradually to the south and southeast, until in southernmost Israel the total yearly amount does not usually exceed 30–40 mm (Fig. 3.4.1). Consequently, the environment is Mediterranean to the north, while southward the desert gradually takes the upper hand (Fig. 3.6.1). Precipitation originates in the central and eastern Mediterranean from cyclonic thunderstorms during the winter, resulting from a combination of several factors: high barometric pressure over the Sahara keeps any rains from reaching the region in summer; at the same time solar radiation and descending hot air cause a warming of the eastern Mediterranean; the only connection between the Mediterranean and the Atlantic Ocean is the restricted path of the Straits of Gibraltar; the freshwater supply to the Mediterranean by rivers is not sufficient to compensate for evaporation, thus the basin has a negative water balance; as a result the Atlantic Ocean waters flow into the Mediterranean, becoming warmer and saltier as they propagate eastward, due to evaporation, until they become heavy enough to sink to the bottom. Consequently, the Mediterranean is quite well mixed, storing solar energy (and oxygen) in its entire water column.

Solar energy thus stored in the Mediterranean during the long, hot summer, is only released to the atmosphere on contact with cold air masses arriving from the north, which can reach the Mediterranean region only in winter, when the high Saharan barometric pressure decreases and the polar high increases. The combination of cold fronts and stored energy creates cyclonic thunderstorms. The confluence, over southern Europe and the Mediterranean, of northern, polar winds and western winds, the latter also being pushed southward during the European winter, defines the paths of thunderstorms and accounts for the present-day distribution of rains in the Levant (Fig. 6.7.1A).

In addition, the heat budget of the Atlantic Ocean also affects cyclogenesis and precipitation in the eastern Mediterranean. When the energy stored in the Atlantic during summer time is higher than average, due both to increased insolation and Gulf Stream activity, warm air masses blowing westward over Europe in winter neutralize the polar cold fronts already there, blocking their way to the Mediterranean. Consequently the European winter becomes stormy, rainy and snowy, while the Levant suffers from a mild, dry winter. Depending on the amount of energy released by the Atlantic Ocean waters, the dry winter could last for the entire season, or occasionally only its earlier part. This climatic system is considered responsible for rains in interglacial climates.

At the height of the interglacial (Fig. 6.7.1B) the southern Levant becomes practically an integral part of the Saharan high pressure system throughout most of the year. The Mediterranean may reach a state of positive water balance, due to the considerably increased activity of the Nile, probably helped by rivers coming from Europe, which benefit from glacial meltwater. The positive water balance caused stratification and sapropel deposition, preventing solar energy and oxygen from being stored in the Mediterranean during the summer, thus the source for

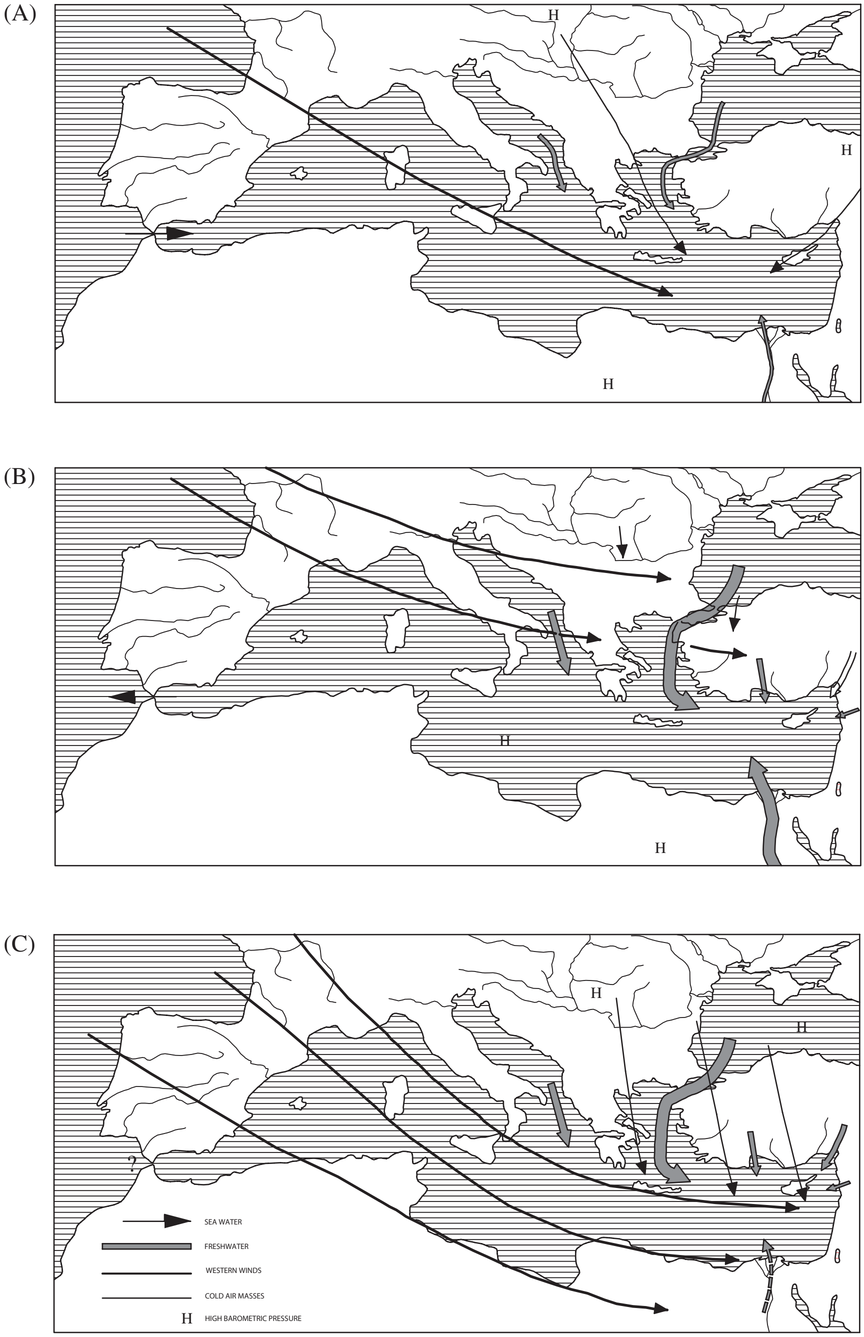


Figure 6.7.1. Climatic models for the Quaternary of the southern Levant: (A) present day, also representing average interpluvial conditions, with only winter thunderstorms, (B) extreme interpluvial conditions, with only few winter thunderstorms, (C) pluvial, with gentle rains all year round. Note direction of seawater flow in Straits of Gibraltar.

winter rains diminished considerably during the interpluvials peaks. It seems that only rare thunderstorms could have reached the Levant under these conditions, most probably from the northwest. The northwestern origin is suggested (Horowitz 1992a, p. 379) because, dry as the region was, the north was wetter than the south (Tables 6.5.2.4–6.5.2.8). The light isotopic composition of rains during times of sapropel deposition could possibly hint at their source in the Atlantic Ocean.

The idea which is occasionally raised, that monsoons reached as far as the southern Levant in interglacial times, has no support in the evidence, since this should have been expressed by a better-developed vegetation to the south, which clearly is not the case during the Pliocene and Quaternary. This model seems to explain the phenomena observed for interpluvial climates in Israel, discussed above, which occur in times of high Mediterranean sea levels, thus justifying the correlation of interpluvials with the interglacials.

In glacial times the situation must have been different. The global climatic belts were located south of their present position (Horowitz 1977a); tropical regions received less insolation (Rossignol-Strick 1983), resulting in decreased monsoon rain quantities and Gulf Stream activity (Beuning 1998, Jolly et al. 1998), also decreasing the efficiency of the Saharan high pressure over the Mediterranean Sea; concurrently, the western winds belt became more active over the Mediterranean, pushed there by the high pressure over Europe, created by the greater extent of glaciers. Consequently, the Mediterranean water became colder, and its balance more positive, due to rains coming in summertime from the Atlantic, decreased evaporation because of increased summer cloudiness, with some increase in water supply from the rivers of southern Europe; while the decrease in the quantity of Nile water, typical of the glacial periods, apparently did not much alter the water balance. All these factors made the Mediterranean only poorly mixed, which prevented the bulk of its water from acting as a solar energy reservoir, thus diminishing considerably the extent of energy available to generate cyclonic thunderstorms.

On the other hand, the lower pressure over the Sahara did not hinder the western winds, which could thus bring rains from the Atlantic Ocean to the Levant (Fig. 6.7.1C) and indeed also to much of the present Sahara (Williams & Faure 1980, Schulz 1987a,b). The rain regime of the Levant in glacial times was therefore characterized by rather gentle rains, some also falling during summer (Horowitz & Gat 1984), spreading southward considerably more than at present, probably (but not necessarily) in somewhat larger annual averages. It is quite evident that the heat budget of the temperate sector of the Atlantic Ocean was, in those times, more significant for the Levantine rains than the depleted energy store of the Mediterranean Sea. Since the Mediterranean was rather cold in these times, it seems that the Atlantic served as the main source of rains during the pluvial periods.

It should be stressed again that the most significant outcome of pluvial climates is the gentle style and more even annual distribution of rains, rather than the absolute quantity. Such a rain regime can explain the pluvial characteristics of the southern Levant: the development of rich flora, its composition and distribution; the depositional and erosional processes; the style of crystallization of speleothems; the extensive pedogenesis resulting in widespread paleosols; and the extension and distribution of lakes and loess deposits. The occurrence of all these pluvial characteristics in the southern Levant in times of low Mediterranean sea levels indicates quite clearly that pluvial climates corresponds to the glacial phases.

6.7.3 Conclusion

It seems that the uplift of the Alpine orogeny was already sufficient to initiate the formation of "embryonic" global climatic belts in middle Oligocene times, and that these were wider and warmer than at present. This placed the Levant within the wet subtropical domain, an extension of the tropical monsoon-controlled belt of the present-day tropical style. Further development of this system created the planetary deserts belt, which is typical for Ma and Mc times in the southern Levant, grading into a temperate climate to the north, getting somewhat cooler as one goes northward.

A global cooling in middle Miocene times was probably sufficient to create an ice cap over Antarctica, due to its higher elevation as compared with the North Pole area, where no ice is known at this time. This global cooling resulted in amelioration of the planetary desert conditions which had affected the Near East, causing the spread of trees through the entire region. The former situation was resumed in Mc times, when extreme desert conditions prevailed in the Near East. Indeed, mangroves are known from southern and southwestern Europe in middle Miocene times (Bessais & Suc 1987).

Further cooling of the Mediterranean region took place in Pliocene times, extending into the earliest Quaternary. Consequently, a temperate climate of northern origin spread southward to the Near East, causing the region to be covered by pontic type forests. Long term oscillations of this kind of climate resulted in Pa being more humid, Pb less so, while QI was even more humid than Pa. The Pliocene cooling could have resulted from a relaxation in the African–Eurasian plates convergence (Fig. 10.4.1), which resulted in the development of extensional structures in the entire Alpine–Mediterranean zone at that time (Mercier et al. 1987, Udintsev et al. 1994, Kempler 1994). The widespread extension resulted in a general lowering of Alpine structures, thus opening the way for colder polar air to move southward and affect the Mediterranean region's climate (for more details see Chapter 10).

During the Quaternary, Israel was mostly under the influence of the Mediterranean regime, characterized by winter rains, which was interrupted by pluvial climates, initiated by the glacials to the north. The onset of this regime since QII times, some two million years ago, can most probably be attributed to an accentuation of the climatic belts' characteristics, resulting from further uplift of the Alpine orogenic belt in that period, which commenced some two million years ago (Mercier et al. 1987, Yilmaz 2001).

6.8 MAJOR UNCONFORMITIES

Unconformities represent time units, just as do rock sequences, but unfortunately bearing no concrete physical evidence. As such they are frequently neglected, because more often than not they are not sufficiently conspicuous. This is even more so with continental sequences, where lithological changes and minor lacunae are quite common, obliterating the major ones due to the similarity in their appearance. Also, sediments deposited in environments such as the fluviatile or eolian display what seem like unconformities, caused by the lateral migration of rivers on a floodplain, or changes in wind directions. Such phenomena could be (and were) mistaken for unconformities, although in fact the environment of deposition existed continuously. A secure way to encounter unconformities in such environments is to first define a complete, continuous sequence in the region, then compare it with the known rock formations, to find out what is missing, if at all. Fortunately, the continuously subsiding basins of the Jordan Rift Valley had accumulated such complete stratigraphic columns ever since the Oligocene, though known only from the deeper boreholes. This has its limitations, but these are surpassed by the abundance and extensive preservation of palynomorphs in the buried sediments.

The next place to look for continuous sections is the deep sea, where they were encountered, in the Mediterranean offshore (except for a part of the Miocene, see Table 1.4.1). Palynostratigraphic correlations of the marine sequences with those obtained from the Jordan Valley, gave certain support for the continuity of the latter. The lithology of the buried formations in the Jordan Valley pretty much resembles rocks cropping out in the vicinity of the Rift, with which they could thus be connected. This provided the foundations for comparative studies, which helped in dating the outcrops, in most cases devoid of or very poor in fossils (pollen grains are usually obliterated by oxidation at the surface), by analogies with the pollen-rich borehole sequences. Similar pollen assemblages in the marine and continental sediments made it possible to correlate them, while microfossils of the former were used for age determinations, based on the standard international geological time scale.

Unconformities may emerge from three principal causes. Global, resulting from large-scale sea-level changes, such as those applied in delimiting geological periods. Regional, ensuing from climate or changing climates, particular to a set of conditions in a certain territory; or local, depending on structural disturbances. Needless to say that, to complicate things, since each of the three causes acts independently, any combinations of two or even all three is possible. Consequently, several of the palynozones described in the present Chapter, are known only from boreholes drilled in the deeper basins of the Jordan Rift Valley and the Mediterranean offshore. Occasionally, but certainly not always, their missing counterparts, the erosive surfaces, could be located and dated on the elevated terrains surrounding the Rift Valley, especially when they are contained between two well-defined sedimentary units. However, in places where successive erosional phases had affected a certain region, leaving very little separating rock formations or none at all, such recognition becomes very difficult, a situation which is unfortunately quite common.

6.8.1 Global sea-level changes

The earliest of these pertaining to the scope of the present work is the major regression separating the Eocene and Oligocene. Indeed, until the end of the former period the entire southern Levant (see Chapter 4) was submerged by the seas, including the area now occupied by the Jordan Valley. The latter region does not seem to bear any special characteristics until the end of the Eocene. The emergence of the southern Levant is a combined result of regression of the Tethys and uplift of the entire region, more accentuated to the south and gently sloping northwestward (Martinotti 1981, Begin & Zilberman 1997). It resulted in erosion of the entire Levant through the formation of channels toward the retreating seas, a process which went on until the beginning of the Miocene, forming at its termination a vast peneplain (Zilberman 1992). The seas retreated in two principal directions, to form the embryonic Mediterranean to the northwest and the Persian Gulf domain (Iraq–Iran basin) to the east and northeast. This is the reason why late Eocene and early Miocene rocks are extremely rare, while early Oligocene sediments are entirely missing from most of the southern Levant (Benjamini 1984, Sneh et al. 1998a). No information from boreholes is known for these periods in the Jordan Valley, since none penetrated that deep.

An interlude in this process occurred during the middle Oligocene transgression, which caused deposition of various rocks filling up the main channels and lowlands. This is indeed the first time when extended subsidence is known for the Rift area, in comparison with its neighboring terrains.

The next global regression, late Oligocene–early Miocene, further accentuated the peneplanation and channeling of the country. This time unit is represented by sediments of Palynozone Ma, fully encountered only in boreholes drilled in the subsiding Dead Sea basin and the Mediterranean offshore. The earlier phase of Palynozone Ma is most probably represented by the En Gev Sands and possibly the uppermost part of the Zefa and Abu Treife, while its later stage may comprise the lowermost sectors of the Hazeva and Herod formations. However, the greater part of Ma duration in most regions of the southern Levant is manifested by erosion, which removed parts or all of the Oligocene sediments, wherever these were laid down at all. Consequently, in most of the region, the peneplanation which started after the regression of the Eocene sea lasted throughout the Oligocene to the early Miocene, so that this combined and rather long unconformity usually separates middle Eocene and middle Miocene rocks. In the elevated regions, which make up the major part of the southern Levant, subsequent deposition was subordinate or non-existent, so that erosion continued to the present day, its results alternating between peneplanation in times of tectonic quiescence, and channeling during periods of structural disturbances.

Miocene sedimentation, chiefly confined to the Mb times of higher seas, filled up the lowlands, occasionally also forming thin veneers of deposits on the highlands, so that toward the late Miocene the entire region became a vast peneplain, evidenced by the widespread rather thin top conglomerates of the Hazeva and Herod formations.

A drop in sea level, which occurred in the middle Miocene, is also apparent in the eastern Mediterranean, resulting in the absence of Foraminifera zones N10, N11 and N12 from the offshore boreholes (Horowitz & Derin 1987), and deposition of evaporites in the Gulf of Suez region (McClay et al. 1998). It seems that this is a global event, possibly induced by the initial major glaciation of Antarctica which occurred at that time (Van Zinderen Bakker & Mercer 1986), as suggested in Horowitz (1992a, p. 383), but alternately could have resulted from local uplift of the southern Levant and the eastern Mediterranean. This regression is very well documented by an unconformity separating the lower and upper Hazeva (Calvo et al. 1997, 1998), laid down in a drainage system leading to the Mediterranean, but much less clear in the Herod Formation to the north. The latter, which was deposited in a suite of water bodies connected to the Persian Gulf, shows several minor unconformities (Michelson 1972), but none comparable to the one in the Hazeva. This fact seems to support a local origin for the middle Miocene regression, thus confined to the eastern Mediterranean, but on the other hand it may have resulted from the considerably longer route of the Herod system. The long distance to the Persian Gulf, and the very gentle relief involved, made erosion quite slow, so that possibly it did not much affect the region as far as the Jordan Valley because of the rather short duration of this regression.

The mid-Miocene regression is accompanied by some decrease in rain, as seen in pollen diagrams from southern Europe (Bessais & Suc 1987) which is a possible result of the cooling, typical of the tropical and subtropical regions up to the present day (Beuning 1998). A decline shown in the arboreal pollen diagrams obtained from the Jordan Rift Valley, in the middle of Palynozone Mb, may correspond to this paleoclimatic trend, indicating that subsidence and accumulation continued in the Rift regardless of the retreating sea. This decline is not seen in the pollen diagram from the Mediterranean, which indicates that the seeming continuity in the AP curve may in fact hide the hiatus.

The next drop in global sea level marks the transition from the Miocene to the Pliocene, represented by the unconformity separating the Hazeva and Herod from the overlying formations (Schulman 1962, Michelson 1972, Shaliv 1991, Ginat 1997, Avni 1998 and many others). Due to the Eritrean faulting (see below), which enabled a rather rapid early Pliocene submersion of the Jordan Valley through the downfaulted Yizre'el Valley, this unconformity is confined mainly to the late Miocene, of a relatively short duration, and so is not represented by a specific Pliocene palynozone. Rather, most of Palynozone Mc is denoted by erosion in much of the region surrounding the Jordan Rift basins, its sediments known only from boreholes or localities subsequently uplifted, such as Jebel Hufeira in the northern Arava, the Karkom basin in the central Negev (Calvo et al. 1997) or Belvoir in the central Jordan Valley (Schulman 1962). Other, rare occurrences in the central Jordan Valley, the Golan Heights or the Hula Valley have been assigned a late Miocene age (Sneh 1996).

This unconformity is very clearly seen anywhere in the southern Levant and again the only localities where sedimentation was continuous are the Mediterranean offshore and the Jordan Rift basins (Horowitz & Derin 1987, Horowitz 1987c). Peneplanation of the elevated regions continued along the same lines observed for the previous stages of low sea levels, somewhat modifying and further flattening the landscape.

The Pliocene constitutes two successive transgressive cycles, the Tabianian (or Zanclean) and the Piacenzian, separated by a regression. Erosion and unconformities connected with this drop in sea level separate the Bira and Gesher formations in the central Jordan Valley (Schulman 1959, Shaliv 1991), the Sedom and Amora in the Dead Sea region (Zak 1967) and the Saggi and Kuntilla members of the Arava Formation to the south (Avni 1998). The last conspicuous global regression is at the transition from the Pliocene to the Quaternary. This again caused erosion, so that Quaternary formations, wherever they overlie Pliocene ones, do so over clear unconformities. As before, peneplanation continued, while continuous sequences are known only from the Jordan Rift basins and the Mediterranean offshore.

6.8.2 Regional climatically induced unconformities

These mainly apply to the Quaternary, a period of conspicuous frequent climate changes which considerably affected sedimentation and erosion in the southern Levant, especially in the continuously subsiding Jordan Rift Valley and its shoulders. The wetter periods of the Quaternary, designated by the odd-numbered palynozones, are characterized by gentle rains all year round, lush vegetation and expansion of the rift valley lakes. These led to accumulation of gravel in rivers leading to the lakes, and to deposition of lacustrine sediments on relatively large areas, including part of the rift shoulders. Naturally, the subsiding rift basins had accumulated greater thicknesses (Horowitz 1989b) of correlative sediments. Both the wadi and lake sediments are thus quite widespread, as is known from numerous localities in and around the Jordan Rift Valley.

The drier, interpluvial Quaternary stages, expressed in the even-numbered palynozones, are typified by winter thunderstorms and rather limited lakes in the Jordan Valley. The style of rains and the poor vegetation allowed for extended erosion, so that no sediments are preserved in the wadis at all, while some or occasionally all preceding pluvial deposits are eroded too. The sediments which had been accumulated during the dry climates in the small lakes were subsequently covered by those deposited by the expanding lakes of the following wetter period, thus hardly ever exposed. The net result of alternations of the wetter and drier periods is that the even-numbered palynozones are known only from the subsurface, where continuous sequences are preserved (Horowitz & Horowitz 1985, 1990). Hardly any exposures of these exist, save for some thin marginal facies on locally elevated blocks, such as at Gesher Benot Ya'aqov (Horowitz 1973).

This picture is naturally different for the last dry period, the Holocene (Palynozone QX), whose sediments did not yet have the chance to be covered. Very restricted occurrences of Holocene sediments are also known from some of the wadis leading to the Jordan Valley (Vita-Finzi 1964), which did not have time enough to be completely eroded.

The sediments of the pluvial formations overlie each other successively, and the separating unconformities are not too clear anyway due to the considerable similarity of these types of sediments. Thus the initial stratigraphic scheme for the Quaternary of the Jordan Valley (Picard 1943, Table facing p. 160) included only formations deposited in a humid climate, which made many authors regard the entire period as "pluvial", subdividing it into "Pluvial A", "Pluvial B" and so on, with no "interpluvials" at all. Unfortunately, this approach is still occasionally used today. The interpluvial formations were first defined from their limited outcrops by Horowitz (1973) for the Hula Valley. Later, when deep drilling and palynostratigraphy were involved, the continuity of the sequence became much clearer.

6.8.3 Local structural unconformities

The three structural stages concerned with formation of the Jordan Rift Valley from the Oligocene onward, also partly affecting other regions of the southern Levant, are the Embryonic, Eritrean and Levantine, which left numerous angular unconformities separating many of the relevant rock formations. These were preceded by the Syrian Arc (see Chapters 4 and 9), which created folds along the southern Levant, controlling sedimentation throughout its long time of activity, from the late Cretaceous through the Eocene, so that corresponding rock units are always thicker in the synclines (see Chapter 4).

The Embryonic stage is apparent in the Jordan Valley by gentle angular unconformities within both Oligocene and Miocene formations, seen more clearly to the north (Michelson 1972), in the Oligocene formations east of Lake Kinneret and the Herod sediments on both sides of the lake. To the south, in the Dead Sea and Arava regions, the unconformities are much less pronounced, but are nevertheless still there (Sneh 1981).

The second stage, Eritrean faulting, has two phases. The first acted during the late Miocene, creating two or three deep troughs along the Jordan Valley and also outside its limits, in addition to several rift valleys of which the most conspicuous are the Yizre'el and the Tripoli. The troughs along the Jordan Valley became local base levels which captured the drainage, so that late and occasionally middle Miocene sediments were eroded and transported into them. The Eritrean faulting thus helped stress the late Miocene erosion initiated by the global regression of this time. Since the erosion following the drop in sea level was long recognized (Garfunkel & Horowitz 1966), it took a while until the role played by the first stage of the Eritrean faulting was acknowledged (Horowitz 1987c). The combined effect of both causes is to make exposures of sediments deposited during Palynozone Mc very rare, as was first recognized and is known chiefly from deep boreholes in the Jordan Rift basins (Horowitz & Horowitz 1990, Horowitz 1996a).

The second Eritrean phase acted in latest Miocene or earliest Pliocene times, accentuating the former processes but also creating new troughs, particularly in the southern Dead Sea area. These further helped erosion of the late and middle Miocene formations, and occasionally even older ones.

The Levantine faulting, which created the north–south oriented Jordan Rift Valley in its present shape, commenced in QII times, some two million years ago, continuing until the present day. This phase has two principal aspects, both affecting the resulting unconformities: namely the subsidence of the Rift on the one hand, accompanied by synchronous uplift of the neighboring regions on the other (Fig. 3.1.1). Both resulted in extensive erosion of the uplands, where hardly any Quaternary deposits are now preserved, while deep in the Jordan Rift basins continuous sedimentation took place as before (Horowitz 1987a). Thus in regions bordering the Jordan Valley unconformities prevail, which in many cases result in complete absence of most sedimentary units.

The Levantine faulting, besides causing continuous subsidence of the Rift's floor, also acted in several "intra-graben" phases, of which the most distinct are two: one following Palynozone QV, during which the Ubeidiya and Mishmar HaYarden formations were deposited to the north, and correlative beds in the Dead Sea region. Distinct angular unconformities between QV and earlier formations and the overlying younger units are apparent along the Jordan Valley (Fig. 11.2.1). The second was active at the end of Palynozone QIX, some 20–15 Ka ago (Fig. 9.6.4), resulting in the formation of the northern Dead Sea basin and Lake Kinneret, as well as considerable deepening of the Hula basin. These caused severe erosion toward the deeper lakes, very prominent to the south where the Lisan sediments are deeply incised, but apparent also along the entire Jordan Valley.

CHAPTER 7

Chronostratigraphy, paleogeography and environments of the Embryonic and Eritrean stages

The three topics discussed in this chapter bring together, when integrated and synthesized, an understanding of the natural landscapes and their evolution. These subjects are mutually interrelated, and also intimately connected with the lithostratigraphy and palynostratigraphy discussed in Chapters 5 and 6.

The main problem faced here was how to arrange their presentation, and in what order. I chose to deal with each separately, in stratigraphic order, so that the line of thought concerning a certain theme is not lost. The other way would be to discuss each time slice in full, which is not inferior to the order preferred here. The discussed chronostratigraphy is determined by the palynozones, which are the only source of continuous information about the Jordan Valley sequences and events. They do not always exactly correlate with the usual stratigraphic terms, but due to lack of other fossils in uninterrupted sections, palynostratigraphy seems the best choice.

This order of discussion involves some unavoidable repetitions, but so would any other, for that matter. The summary of natural landscapes for each palynozone will thus be found under "Natural environments". The present chapter deals with the Embryonic and Eritrean stages of development of the Jordan Rift Valley, namely up to and including Palynozone QI. Correlations are summarized in Table 7.1.

The Levantine stage, spanning the Quaternary Palynozones QII–QX, is coincidentally characterized by two events in the region. The Dead Sea became a terminal basin for a newly formed endoreic drainage system (see Chapter 9); and humans became part of the scene. The latter event justifies, in my opinion, separate discussions of the chronostratigraphy, as well as occasionally a refinement of the ages beyond the palynozone level, in Chapter 11, while the paleogeography, environments and landscapes concerned with human settlement in the southern Levant are dealt with in Chapter 12.

Table 7.1. Correlations of the Embryonic- and Eritrean-stage formations.

Age (Palynozone)	Northern sector	Southern sector
Eritrean-stage formations		
Earliest Quaternary (QI)	Mahanayim Marl Amud Conglomerate Mechki Basalt Ruman Basalt Dalton Basalt Amud Basalt Cover Basalt Bethlehem Conglomerate Dhuleil Formation	Melekh Sedom Sands HaMeshar Formation Ar Risha Gravel Formation Ghor el Qatar Series?
Pliocene (Pa and Pb)	Zahlé Beds Tel Hai Formation Tanur Conglomerate Gesher Formation Bira Formation Intermediate Basalt Fajjas Tuff Jabal Bakiya Formation	Oolitic Formation Samra Formation Shagur Formation En Feshha Conglomerate Amora Formation Sedom Formation Mazzar Formation Arava Formation En Yahav dike Grain Sabt Basalt
Late Miocene (Mc)	Umm Sabune Conglomerate Hordos Conglomerate? Lowest part of Bira Formation Late Miocene Volcanics Uppermost Herod Formation?	Hufeira Formation Uppermost part of Ashalon Member Dhira Ibn Salih Member? (upper part of Dana Formation)
Embryonic-stage formations		
Early and middle Miocene (upper Ma and Mb)	Herod Formation Hordos Conglomerate? Lower Basalt Barada Formation	Hazeva Formation Ghor el Qatar Series? Wadi Bustan Member (lower part of Dana Formation) Ashosh and Karak plugs Red Sea Dike System
Oligocene (lower Ma and middle Oligocene)	En Gev Sands Susita Formation Fiq Formation Taiyiba Beds	Zefa Formation Abu Treife Series Lowest part of Dana Formation? Lowest 400–500 m at Sedom Deep 1 borehole

7.1 CHRONOSTRATIGRAPHY

Chronostratigraphy is obtained from two different dating techniques, the results of which are, unfortunately, too often contradictory. The "classic" relative chronology relies principally on biostratigraphy, which delineates evolutionary trends in fossils or their associations, applying relevant changes for dating the stratigraphic stages. The "modern" approach administers "absolute" dating methods, mostly based on radiogenic techniques, for outlining the timing of geological events, expressed in radiometric "years". Both approaches are widely used in chronostratigraphy, but the trouble begins when the results of either do not agree with those for the other. This situation is especially common in late Cenozoic sequences, which involve geologically short durations, since neither biostratigraphy nor radiogenic dating are sufficiently sensitive or refined, particularly when continental sediments are concerned. This is so for biostratigraphy because evolution in numerous plant and animal groups is occasionally too slow, and this is further complicated by the length of the dispersal times necessary to travel considerable geographic areas over land. Radiogenic dating is inadequate because many of the methods are not yet refined enough, and possibly also not all its limitations are fully understood. Other dating techniques, such as those based on oxygen isotopes or paleomagnetism, are only extensions of these two basic methods.

The main reason for difficulties in the chronostratigraphic assignment of continental formations lies however with the rarity of material suitable for dating by either method. Fossils are frequently scarce in such rocks, except for pollen. However these, being composed of organic matter, easily oxidize and disintegrate in outcrops, even when one digs deep into the exposure. Oxidation particularly affects palynomorphs in warm climates, such as were prevalent in the southern Levant. Thus palynostratigraphy can be used to its full advantage only when borehole material is available, which is rarely the case for financial reasons. Even when drilling samples are ample, as is the case with the Jordan Rift Valley, their correlation with outcrops may be quite problematic, since the latter have hardly any clues that could be followed.

The same applies to radiometric dating methods: it is quite rare to find continuous sequences of datable rocks, either in subsurface sections or in outcrops. Furthermore, it is quite frequently difficult to assess exactly the nature of magmatic rocks in boreholes, particularly the fine-grained varieties, where at least some of the occurrences may consist of intrusions. This is very difficult to tell from the drill cuttings, which are the most common samples available, for practical reasons.

One therefore faces two problems. The first concerns tying up the palynostratigraphy of continental sequences with the international subdivisions of geological time. Considering the above limitations, a possible solution is to correlate pollen spectra obtained from boreholes drilled in continental sediments with those extracted from sequences of marine origin, where in addition a variety of datable fossils are found (Fig. 6.6.1). For the Quaternary part, which is based globally on climatostratigraphy, correlations involve the climatic aspects of the palynozones, compared with those obtained from oxygen isotopes of the oceans (Fig. 6.6.2).

The second problem involves correlating palynostratigraphic units from boreholes with their exposed lithostratigraphic counterparts, as the two occurrences may occasionally comprise different rock types. Three methods are applied here for tackling this problem. The first, which is not too frequently applicable, involves lithological correlation of boreholes and outcrops. The question marks over this method concern the normally poor quality of borehole samples, almost always comprising well cuttings, which would never, for instance, reveal the true nature of a conglomerate horizon, so common in the Jordan Valley sequences. Another drawback concerns the fact that while most boreholes in the Jordan Valley were sunk into deep basins, the outcrops usually represent only marginal facies, where different types of sediments accumulate. Thus the most common method of correlating marine sequences becomes almost useless with the continental rocks of the Jordan Valley, and could only rarely be practically applied.

The second method relies on the effects of climate on sedimentary processes in this part of the world, which give a particular "character" to each rock unit. Dry periods, with rare rains occurring as thunderstorms, are characterized by diminishing areas of lakes in the Jordan Valley. These result in increased erosion on the one hand, and accumulations of coarser gravel on the other, such as we see in palynozones Ma, Mc, or Pb. The climate is probably not the only cause, since all three palynozones also represent times of low sea levels. However, the climate effect is increased in continental environments, typical of the Jordan Valley. Humid climates, typified by higher and gentler precipitation and extended vegetation cover, cause increase in extension of the Rift Valley water bodies, less erosion and accumulation of fine grained clastics or chemical sediments.

Wet subtropical climates, such as during Palynozone Mb, are characterized by the formation of red-beds. Cooler climates, typical of "P" palynozones, involve much less oxidation and grayish–whitish sediments, usually rich in carbonates and well cemented. The climatic trends concluded for the sequence of palynozones obtained for the Jordan Valley, mainly from boreholes, could thus be used to further support or refute proposed correlations.

The third approach involves the application of sequence stratigraphy principles. The outcropping sequences can be correlated with a series of known worldwide or regional events, for which transgressive cycles and global climate changes seem the best reference, due to the fact that these were used primarily to define chronostratigraphic stages (Fig. 9.1.1). Once the sequence stratigraphy of events has been established, several point dates scattered along the sections can be used to assure their chronology. These point dates could either be radiogenic or based on evidence provided by fossils, paleoclimate or lithological correlations.

It would thus be appropriate to outline the basic principles pertaining to definitions of the late Cenozoic geological time units, in their sequence stratigraphic order, before any further step is taken in dating and correlating the various rock units. In the first place, the Oligocene, Miocene and Pliocene were defined as major transgressive cycles (Gignoux 1955, pp. 467, 554). The Quaternary, although not originally defined as a transgressive complex during the first stages of research, seems also to agree with this definition (Horowitz 1992a, p. 344), commencing with the Calabrian sea-level rise, continuing with a general regressive process which is superimposed by eustatic sea-level changes, affected by glaciation and deglaciation cycles. Taking these definitions into account, it would be clear that complete sequences can only be found in the deeper marine domain, while the transgressive cycles on land would usually be diachronous, almost always more complete seaward. Exceptions to this rule are found only in subsiding continental basins, such as the Jordan Rift Valley, where sediments are accumulated regardless of sea-level changes.

Consequently, a distinction is made here between the rock units known from boreholes drilled in the Rift Valley, which represent quite faithfully the entire period, and their counterparts outside the continuously subsiding region, in regions frequently given to uplift. There, sedimentation took place only during times of higher sea levels, while erosion and channeling followed the regressive phases. Naturally, such behavior is not unique to marine sediments, but applies also to continental formations, the deposition of which largely depends on erosive energy, which diminishes in times of high base levels.

The most common continental sequence representing a marine transgressive cycle would therefore commence with deposition of a base conglomerate, which becomes younger landward. As the base level rises the clastics become progressively finer until, at the maximum height of transgression, the hinterland may develop lakes, or occasionally even be invaded by the sea itself. An inverted process follows the regression, when fine clastics overlie the lacustrine, lagoonal or littoral sediments, becoming coarser upward, until the cycle terminates with a top conglomerate. From there on, following further regression, erosion and channeling modify the landscape. This sequence becomes thinner toward the hinterland, until it reaches a place where the base and top conglomerates may practically form a single unit, while further landward only erosion prevails. This ideal picture may, naturally, be affected locally by structural disturbances.

When the base level remains static for a long enough period, provided there is no uplift of the hinterland the continuous erosion would finally result in formation of a rather flat peneplain. Such occurrences are therefore indicative of tectonically quiet periods.

The distinction between base and top conglomerates is usually straightforward: the former is coarse, built of badly sorted, poorly rounded local components, filling channels, rather limited in area, mostly well preserved due to their subsequent cover by finer clastics; the top conglomerate is well sorted and rounded, due to its long-distance transportation. It overlies rather flat surfaces, filled up previously by the transgressive sediments, thus comprising thin, wide gravel bodies, frequently eroded following the subsequent drop in sea level.

Another type of conglomerate that needs consideration is the one termed "syntectonic". These are gravel accumulations which follow faulting and relief accentuation, always of local rather than regional base level changes. Their gravels are poorly sorted, hardly rounded, occasionally breccias, always arriving from nearby. Care should be taken when labeling a conglomerate "syntectonic", as in many instances concerned with the Jordan Valley it was found that these are indeed regional occurrences (e.g. Bender's "Lower" and at least part of the "Upper Syntectonic Conglomerates").

The chronostratigraphy of magmatic formations of the Jordan Rift Valley is based on both their stratigraphic setup and radiometric ages. These magmatics were traditionally defined as being sandwiched between any two sedimentary units. When the latter are dated by the use of fossils or sequence stratigraphy methods, the magmatics are referred to by the geological stages during which they had been formed. Incidentally, many magmatic occurrences, especially in outcrops, do not always have clear contacts with sedimentary formations. These include various isolated exposures, or in boreholes where it is not always clear whether magmatics are flows or intrusions. In these cases, when only radiogenic ages are available, correlations with "traditionally" dated formations is made, but with reservations.

7.1.1 Middle Oligocene

The Oligocene is one of the more interesting, but lesser known, geological periods of the southern Levant. To demonstrate, a computer search at the library of the Geological Survey of Israel in 1999 resulted in only 76 studies concerned with the Oligocene of Israel, most of which are not even fully dedicated to this period. This was the transition interval from the Tethys Ocean realm, which terminated some time at the late Eocene, to the principally continental domain that brought the region to its present-day configuration. Being sandwiched between two large-scale global regressions, over an uplifted region, the Oligocene transgression was subordinate in the southern Levant, leaving only marginal shallow seas deposits, with rarely any continental counterparts, in contrast with the situation in neighboring countries such as Egypt (Said 1962, p. 219). The considerable regression at the end of the Oligocene caused severe erosion in the southern Levant, leaving behind canyons several hundred meters deep (Neev 1960, Martinotti 1981, Buchbinder & Siman-Tov 2000), removing much of the previous sedimentary evidence, which was poorly developed anyway.

Relics of rocks formed during this period are therefore found only in some rare, sheltered localities, such as the southern Israeli coastal plain synclinorium to the west (Avnimelech 1936), or the rift valleys of the Jordan and Wadi Sirhan, where occurrences of fossiliferous marine Oligocene beds are known (Wetzel & Morton 1959, p. 170; Daniel 1963; Michelson 1972; Michelson & Lipson-Benitah 1986; Horowitz 2000a). In addition, subordinate Oligocene beds were penetrated by several boreholes in the Kingdom of Jordan, several tens of kilometers east of the Jordan Valley, and in western Israel, particularly offshore. The one exception in the Mediterranean coastal plain is at the Haruvit group of boreholes (Derin 1976), where an Oligocene sequence of some 600–700 m was penetrated. Conspicuously, however, Oligocene sequences in the Jordan Valley limits are thicker than in neighboring terrains, including synclines of the Syrian Arc. The Oligocene is thus the first period during which accelerated subsidence is seen along the Jordan Valley.

The Oligocene sediments (the closest magmatics are known for that period only from eastern Transjordan) comprise a suite of formations known from the eastern flanks of the Jordan Rift Valley, east and south of Lake Kinneret (the Fiq Formation, Susita Formation, En Gev Sands and Taiyiba Beds), and a sequence of 400–500 m penetrated at the bottom of Sedom Deep 1 borehole, just south of the Dead Sea. Various fossils found in the marine rocks, pollen and nannoplankton recovered from the borehole, do not leave much doubt as to their Oligocene age. Possible Oligocene units are the Zefa Formation in the northern Negev, the Abu Treife Series in the central Negev and the lower part of the Dana Formation east and south of the Dead Sea, along the eastern Arava. The sterile nature of the Zefa and Abu Treife call for some consideration before this age is safely assigned to them. The rare fossils of the lower Dana, although possibly indicating marine connections, are of a wide stratigraphic range, and not significant enough for firm conclusions.

Typical fossils of the Fiq Formation are fragments of large benthonic foraminifera, common in the pelbiosparite facies, such as *Discocyclina*, *Lepidocyclina*, *Nummulites*, and rotaliids, together with *Globigerina* (Lipson 1971, Michelson & Lipson-Benitah 1986). Planktonic globigerinids become more frequent toward the middle part, indicating some deepening of the sea, but upward the benthonic elements again prevail. One of the more interesting components of the faunal assemblage in the Fiq Formation is *Austrotrillina*, a foraminifer rarely found in the Mediterranean but very common in the Indian Ocean domain. Some ostracodes and *Lithothamnion* algae are also found, as well as mollusks. *Cassigerinella chipolensis* and *Globanomalina micra* at the base indicate, according to Lipson, an early Oligocene age. The planktonic globigerinids of the middle part led Lipson to assign this unit to Blow's P18–P19 zones of the early Oligocene, overlying unconformably the lower sector, of late Eocene (P15 and possibly P16). Subsequently, Michelson & Lipson-Benitah (1986) defined the entire Fiq as of early Oligocene age.

Foraminifera reported from the Taiyiba Beds include *Textularia, Uvigerina* and *Operculina*; megafossils are mainly bivalves such as *Pecten*, *Chlamis* (*Aequipecten*) *judaica* and oysters, accompanied by echinoids, Bryozoa, charophytes and Cichlid fish teeth (Bender 1974a, p. 90). The outcrops near Esh Shuna contain *Echinolampas*, *Clypeaster* and *Pholadomya*. The fauna led Wetzel & Morton (1959) to assign the Taiyiba Beds an Oligocene age, accepted also by Bender. These Beds show great lithologic similarity to the Fiq Formation, and both thus should, in my opinion, be treated as a single stratigraphic entity, for which the name Taiyiba Beds has priority.

The Susita Formation overlies the Fiq conformably, commencing with 3–10 m of thick marly horizon, particularly rich in *Pecten* shells, always appearing as a guide horizon at the base of this unit. Its typical fossils include foraminifers such as *Archaias*, *Austrotrillina*, *Operculina*, miliolids and others, indicating an early Oligocene age for the lower part of the HaOn Member (P20), and middle Oligocene to early Miocene for its upper part, with a clear affinity to the Indian Ocean province. An abundant planktonic foraminifer pointing to the same age is *Globigerina ampliapertura* (Lipson 1971). Megafossils include Bryozoa, corals, echinids, and gastropods such as *Turitella, Cerithium*, *Bithium* and *Conus*. Bivalves are plentiful, including *Pecten*, *Cardium*, *Arca*, *Lucina*, oysters and others. *Lithothamnion* and other, as yet unidentified algae are common. No diagnostic fossils are reported from the Noqev Member, so its age is therefore considered "younger than early Miocene" by Michelson & Lipson-Benitah (1986).

The age of the Susita, according to Michelson (1972) and Michelson & Lipson-Benitah (1986) is early Oligocene to early Miocene, or even the early part of the middle Miocene. This assignment resulted from the wide stratigraphic range of the upper HaOn fossils, pushing all the way up to the Miocene. It seems however more conceivable to regard the entire Susita as of early-middle Oligocene age, since it does not seem to represent a long period of deposition, nor does it display the unconformity of the Oligocene–Miocene boundary, so prominent in the coastal plain. This unconformity is most probably located between the overlying En Gev Sands and the Herod Formation. Consequently Horowitz (1979, p. 70) assigned the Noqev Member a middle Miocene age, erroneously correlating it with the lower part of the Herod Formation. It seems however that the fossils, especially the foraminifera mentioned above, justify an Oligocene age for both the Fiq and Susita formations. This suite was deposited during the height of a transgressive cycle, which also corroborates the Oligocene age.

At Wadi Taiyiba, the lowermost 38 m of what is referred to as the "Usdom Group" by Wetzel & Morton (1959) are composed of white–yellowish sands and fine-grained, hard calcareous sandstones containing oysters and reworked(?) foraminifers. Occasional Eocene limestone pebbles occur at the base, which rests unconformably over the Taiyiba Beds. These are overlain by 11 m of hard, white and pink sandy limestone, with oysters and ostracodes, overlain unconformably by the red sandstones of the Herod Formation (which is also included in the "Usdom Group" at this locality). This sequence, when considering its lithology and setting, seems to correlate with the Susita Formation and maybe also to at least part of the En Gev Sands, further north.

The middle part of the Zefa Formation is a slightly friable limestone, containing unidentified plant remains and oncolites. The latter are several centimeters in size, oval in shape, probably originating from freshwater blue algae (Buchbinder, in Shahar 1973). It seems quite clear from the descriptions and columnar sections presented in Shahar (1973) that the Zefa Formation sequence represents a complete transgressive cycle, prior to the Hazeva Formation, with base and top conglomerates, the middle marls and limestones having been deposited in a shallow lake or lagoon. The Eocene pebbles found in the Zefa define its maximum age, so it is sandwiched between the Eocene and the Miocene Hazeva Formation (see below) and hence an Oligocene age is the most plausible assumption.

Admittedly, this is not based on direct evidence, but lacking any indicative fossils it appears the best guess. It seems that the Zefa represents the middle Oligocene, again based on sequence stratigraphy grounds. It is however possible that its deposition started already in the later early, while its upper part may be stretched into the late, Oligocene. Another problem with the dating of the Zefa Formation arises from the fact that, in terms of environment, it was deposited under wet subtropical conditions, similar to the overlying Hazeva, so the sediments of both rock units present very similar appearances. This resemblance convinced some authors (Calvo et al. 1997) to include the Zefa as a middle member of the Hazeva (erroneously, in my opinion). Similar reasoning also made Shahar (1973) consider the Zefa as a lowermost member of the Hazeva, not a separate formation. Horowitz (1979, p. 69; 2000a), Hirsch (1996) and Hirsch & Roded (in preparation) all agree on an Oligocene age for the Zefa.

The setup of the Abu Treife Series recalls that of the Zefa. Both are superficially quite similar in lithology to the Hazeva, but both quite undoubtedly underlie the base conglomerate of the latter, which makes assignment of the Abu Treife to the Oligocene as safe as that of the Zefa. Bentor & Vroman (1951) suggested late Eocene or Oligocene age for the Series, because it overlies middle Eocene rocks unconformably, its exposures are located on a structure much higher than where the Miocene Hazeva Formation is usually found and because of its lithology, different in some details from the latter. As mentioned in Chapter 5.1.1.6, only the outcrop at Mahmal indeed belongs to the Abu Treife, while the one at Abu Treife itself is Hazeva.

The apparent similarities in lithology of the Zefa, Abu Treife and Hazeva (although careful examination readily reveals the differences between the first two and the Hazeva) could be explained, even if the first two are of Oligocene age, while the latter is Miocene. All three were deposited under approximately similar environmental conditions, in fluvio-lacustrine systems affected by transgressive cycles and wet subtropical climates. This, together with lack of diagnostic fossils, makes exact age assignments of the Zefa and Abu Treife so difficult.

Bender (1974a, p. 89, 1974b) assigned his “Lower Syntectonical Conglomerates”, now known as the Wadi Bustan Member, which makes up the lower part of the Dana Formation (or Edh Dhira Beds), a “?middle to upper Oligocene, ?Miocene” age. However, in his opinion Oligocene is the more likely choice. Khalil (1992) reports *Globigerina officinalis subbotina, G. senilis* and *G. tripartita* from the Edh Dhira area, all thought to be of Oligocene age. Macumber & Edwards (1997) indicate, from the same region, that a doleritic intrusion penetrating the lower part of the Dana Formation in Wadi Karak is similar to intrusions upstream, K–Ar dated at approximately 19 Ma (Barberi et al. 1980). If indeed these views are accepted, at least for the lower part of the sequence, than we have in the Arava a similar situation to the Zefa Formation, only the lower Dana shows a stronger evidence of marine origin for some of its layers, by their fossils.

In terms of sequence stratigraphy and partly also of lithology, the Zefa and Abu Treife recall very much the situation at En Gev, of the upper Fiq, Susita and En Gev Sands complex, although the latter is better developed and contains diagnostic marine fossils. A similar situation is also seen at the Wadi Taiyiba suite, as well as for the middle Oligocene and the lower part of Palynozone Ma at Sedom Deep 1, in the southern Dead Sea. A correlation between the five occurrences was thus proposed (Horowitz 2000a), based on these similarities. It is suggested that deposition of all these sequences was controlled by the Oligocene transgression, reaching the Jordan Valley from both east and west. The lower section of the Wadi Bustan Member may well belong to this system.

7.1.2 Palynozone Ma, late Oligocene to early Miocene

A complete sequence of late Oligocene through early Miocene rocks is known in the Jordan Rift Valley only from the Sedom Deep 1 borehole, where sediments of Palynozone Ma seem to have been accumulating continuously ever since the end of the middle Oligocene. Ma was also encountered in the Ami'az 1 borehole, where its late Oligocene base is probably missing, the early Miocene part directly overlying Senonian rocks. Palynozone Ma corresponds to Foraminifera zones N2 (P21) through N7 offshore (Horowitz & Derin 1987), of late Chattian–Burdigalian age (Derin & Reiss 1973).

In all other locations late Oligocene–early Miocene rocks are entirely missing, most probably due to erosion toward the low sea level. The Jordan Rift Valley region, being quite far away from either the Mediterranean to the west or the Persian Gulf to the east, could therefore be expected to accumulate sediments only in times of highest sea levels, that is, the later middle Miocene. This is true for the areas closely neighboring the Valley, where sediments of the Hazeva Formation to the south and the Herod to the north mostly represent this time span. However, within the continuously subsiding parts of the Jordan Valley depression, a complete Oligo-Miocene sequence was encountered in the southern Dead Sea region, although at present it is known only from a single borehole (Horowitz 1996a, 2000a). It is not clear what the situation is in other basins along the Rift, since none of the drillings penetrated any deeper than Palynozone Mb.

The En Gev Sands are sterile of any fossils, so the problem of both their exact age and environment of deposition remains a matter of speculation. Michelson (1972) tends to regard the sand as of either fluviatile or lacustrine origin. It seems, however, that at least the limestone horizons were laid down in a shallow lake or a near-shore lagoon. Granulometric analyses of the En Gev Sands led Giveon (1984) to suggest a continental alluvial fan as the principal environment of deposition, with occasional occurrences of shallow seasonal lakes and playas, under arid or semi-arid climates. These climatic conditions correspond quite well to what is known for Palynozone Ma, whose lower part is of late Oligocene age. Notably, yellow sands of similar appearance make up the upper part of the Abu Treife Series, which underlies the Hazeva base conglomerate.

Traditionally, the En Gev Sands were thought to be a part of the Miocene Herod system (Horowitz 1979, p. 70). The main reason appears to be their lithological nature, as sandstones are a common sediment type in many inland Miocene formations. This fact alone led Golani (1962) to include the En Gev within the Herod, while Michelson (1972) and Michelson & Lipson-Benitah (1986), even though they regarded it as an independent stratigraphic unit, still attributed the En Gev to the Miocene. Another reason is the way the fauna of the underlying Susita Formation was frequently viewed: most (but not all!) of its components, especially the numerous megafossils and benthonic foraminifera, also lived through to the Miocene. Thus it was usually referred to as "Oligo-Miocene" (Lipson 1971, Michelson 1972, Michelson & Lipson-Benitah 1986). Any overlying unit would thus "naturally" be of Miocene age.

Apart from the fact that both the En Gev and Herod formations comprise sandstones, there are significant differences between the two in almost any characteristics of the sands. It seems to me that the En Gev Sands should be regarded as representing the final stage of the Oligocene transgressive cycle, namely the late Oligocene regression, and should thus be attributed this age (Horowitz 2000a).

The lowermost 38 m of what is referred to as the "Usdom Group" by Wetzel & Morton (1959) at Wadi Taiyiba, are composed of white–yellowish sands and fine-grained, hard calcareous sandstones containing oysters and reworked(?) foraminifers. This sequence, considering its lithology and setting, seems to correlate with the Noqev Member of the Susita Formation and at least partly with the En Gev Sands, further north. It also resembles part of the Abu Treife and Zefa.

As it appears, the lower part of the Hazeva Formation (its base, designated the Shahaq Conglomerate), the upper part of the Wadi Bustan Member of the Dana Formation in Transjordan (if its lower sector is indeed of Oligocene age, as suggested in Bender 1974b; Horowitz, 2000a), and the basal sequence of the Herod (its "Lower Conglomerate"), may all have been deposited simultaneously, during the early Miocene. The evidence for that is not entirely well founded, but certainly each of these units is itself diachronous, its seaward extension deposited during longer periods. Seaward, which is west for the Hazeva system, but east–northeast for the Herod, the time spans represented by early Miocene sediments increase, until in the Mediterranean offshore or the Persian Gulf complete Miocene sequences are found (Horowitz 1974).

Magmatics of early Miocene age, termed the Red Sea Dike System, are rare in the vicinity of the Jordan Valley, but quite common in regions bordering the Red Sea and the Gulf of Suez. Only two occurrences of this system are known close to the Jordan Rift. A plug (or vent) in Nahal Ashosh, some 60 km south of the Dead Sea at the western rim of the Arava, comprises a few small basalt intrusions with gabbro xenoliths and pyroclastics penetrating middle Eocene rocks (Levitte 1966), which yielded K–Ar ages of 20.4 ± 0.7 Ma (Steinitz et al. 1978). Several basalt plugs are in the Karak graben east of the Lisan peninsula (Bender 1974a, p. 105). One of these, a tholeiite, yielded a K–Ar age of some 19 Ma (Barberi et al. 1980).

7.1.3 Palynozone Mb, middle Miocene

Palynozone Mb, which was encountered in numerous boreholes drilled along the entire Jordan Valley and the Mediterranean offshore, corresponds to Foraminifera zones N7–N14, of late Burdigalian, Langhian and Serravallian age. Although the middle part (N10–N12) is missing offshore, it seems that it is fully represented in the continuously subsiding Jordan Valley basins. The palynozone is characterized by pollen spectra indicating a wet subtropical climate, an environment in which red-beds are typically deposited, a result of oxidization of iron-bearing minerals under conditions of high temperatures and humidity. Another mineral common to such environments is dolomite, as opposed to calcite typical for cooler surroundings.

This, naturally, is one of the main reasons for assigning both the Herod–Barada and the Hazeva–Dana systems to the Mb time interval, since no other period in the history of the southern Levant provides such suitable environmental conditions for deposition of their typical rocks. Of course, the middle Oligocene also displayed quite similar environments, but this time span is already taken by formations pre-dating those mentioned above. Both systems are overlain by rocks rich in calcite, deposited in a cool domain, typical of the Pliocene Pa and Pb palynozones.

7.1.3.1 *Herod Formation and Lower Basalt*

No identifiable fossils were ever reported from outcrops of the Herod Formation, which seems quite strange, considering that all investigators agree that the environment of deposition is fluvio-lacustrine. The Formation is quite rich in dolomites (Shaliv 1991), particularly in its upper parts, probably post-genetic or diagenetic. This may have to do with dissolution and recrystallization processes which could have wiped out any remains of fossils, whose total absence still remains enigmatic. Some rare mollusk fossils occurring in beds and lenses within the Lower Basalt are not indicative of age.

Despite the lack of fossils, all authors agree that the Herod Formation and the synchronous, interfingering Lower Basalt are of Miocene age. Before the advent of radiometric dating, investigators had applied stratigraphic criteria to date both rock units. As it overlies the Eocene and Oligocene at En Gev, and is topped by Pliocene sediments all over, the most reasonable guess was Miocene. Bentor (1957) defined the age of the Lower Basalt as middle Miocene; Schulman (1962) as Miocene; Freund et al. (1965) as middle to late Miocene, based on preliminary paleomagnetic measurements. Radiogenic ages for the Lower Basalt were first published by Siedner & Horowitz (1974), followed by Steinitz et al. (1978) and Shaliv (1991), confirming earlier assignments, all broadly indicating a middle Miocene age.

In contrast with other fossils, pollen grains are quite abundant in correlative borehole samples assigned to the Herod Formation (see Chapter 6.5) from Zemah 1 and Notera 3, representing Palynozone Mb (Horowitz & Horowitz 1985, Marcus & Slager 1985). Correlation of these subsurface sequences with the Herod is based on the stratigraphic setting underlying the Pliocene sediments (the base of Mb was not reached in either of these boreholes), their lithological characteristics and the radiometric ages of interfingering magmatics. Radiometric dates obtained for the base of Palynozone Mc in the Zemah 1 borehole (see below) make the underlying Mb somewhat older, within the middle Miocene, as suggested in Horowitz & Derin (1987, see also Table 1.4.1).

The assignment of the Herod Formation to the Miocene is further strengthened by radiometric ages obtained from the interfingering basalt tongues. Siedner & Horowitz (1974) measured four samples from the Poriyya escarpment, of flows within the Herod Formation clastics, obtaining ages in the range of 14.6 ± 0.4 to 12.2 ± 0.4 Ma. Steinitz et al. (1978) indicate isochron ages of 14.5 ± 0.3 Ma for the lowermost flow at the same section, while the uppermost was dated at 4.9 ± 1.3 Ma.

Shaliv (1991) made a much more detailed radiometric study of the Lower Basalt in the Yizre'el and central Jordan valleys, as well as the southern Golan east of Lake Kinneret, measuring some 130 samples, mostly from outcrops and some from boreholes. Shaliv's results for all his samples generally indicate older ages than measured before, explained by him as "due to improvements of the analytic system". Measurements from the Poriyya escarpment indicate an interval between 16.8 ± 0.4 Ma for the lowermost flow, which overlies more than 300 m of Herod clastics, to 12.3 ± 1.2 for the uppermost, some 10 m below the truncated top of the sequence. These dates show that the time interval taken by the Herod Formation could have been longer than previously thought, most probably from the middle or late part of the early Miocene, possibly Burdigalian, through to the end of the middle Miocene, Serravallian or early Tortonian. The Lower Basalt at Belvoir, where no sediment intercalations are known, yielded 15.7 ± 0.2 Ma in the lower part of the section (whose base is not exposed), while a flow most probably overlying the truncated uppermost part was dated at 10.1 ± 0.3 Ma, attributed here to the late Miocene volcanic phase. Only a few samples from the southern Golan were dated by Shaliv, yielding ages in the range of 16–8.5 Ma. These ages make the Poriyya, Belvoir and southern Golan sequences broadly correlative.

7.1.3.2 *Main body of the Hazeva and Dana formations*

The first to identify interfingering, in the Be'er Sheva region, of the middle part of the Hazeva Formation with marine middle Miocene sediments were Bentor & Vroman (1957). Based on that, they assigned an early to middle Miocene age to the Formation. They further extended the age into the Pliocene by including also the Arava Formation (Conglomerate) within the Hazeva, though noting the unconformity and discontinuity between the two units. Earlier investigators suggested a variety of late Tertiary ages, such as Pliocene (Blake 1928), upper Pliocene (Blanckenhorn 1931), middle Miocene and possibly Pliocene (Picard 1943) and occasionally also Oligocene to Miocene (Bender 1974a, p. 92, 1974b), for the Transjordanian correlative sequences.

Westward, the Hazeva interfingers with the Ziqlag Formation, which includes the foraminifer *Borelis melo*, some 50 km from the Dead Sea. This microfossil led Derin & Reiss (1973) to attribute the formation to the middle to early-late Miocene, Serravallian–Tortonian. The Ziqlag is the shallow marine reefy equivalent of the deep-water Ziqim Formation, in which the upper part of Palynozone Mb is located; its lower part lies within the upper part of the Bet Guvrin Formation (Horowitz & Derin 1987). In the central northern Negev a littoral-estuarine facies appears, the Yeroham Member of the Hazeva, with its typical oyster banks. Toward the later part of the Miocene, deposition of Ziqlag-type sediments followed the regressing sea, so that in the offshore boreholes they constitute the top of the Ziqim. The Ziqlag, which interfingers with the Hazeva, represents the maximum of the transgression, hence it is, in this region, broadly of middle Miocene age.

No direct radiometric datings of the Hazeva Formation, the Edh Dhira Beds, the Wadi Bustan or the Ghor el Qatar (if indeed the latter belongs to this system, which is still highly speculative) is possible, for lack of interfingering volcanic rocks. Two ages obtained from post-Hazeva events put an upper limit on its time of deposition. Fission-track dating of a metamorphic process (locally known as the "Mottled Zone"), which had affected sediments of the Hazeva Formation west of the Dead Sea (Kolodny et al. 1973), yielded an age of 13.6 ± 2 Ma. The basaltic dike of En Yahav, which penetrates the Hazeva, yielded K–Ar ages of 8.8 ± 1.7 to 3.9 ± 0.5 Ma (Steinitz et al. 1978), or recently 6.4 Ma (Steinitz et al. 2000). These ages are in broad agreement with the stratigraphic ones.

No fossils have ever been reported from the Shahaq Conglomerate. Mashaq limestones and the Ghor el Qatar Series contain freshwater mollusks such as *Melanopsis*, *Melanoides* and *Planorbis*, ostracodes, stromatolites, oncolites and algal debris, indicating a lacustrine environment in the Arava and Dead Sea regions; the Yeroham Member contains oyster beds (*Crassostrea gingensis* according to Tchernov et al. 1987, and possibly some *Ostrea edulis*, according to Harash 1967), indicating a brackish estuary environment. However none of these fossils are indicative of age. Clays of the Gidron Member occasionally contain plant remains, among which J. Lorch (Department of Evolution, Systematics and

Ecology, the Hebrew University of Jerusalem 1998, pers. comm.) identified fruit and large leaves of *Lotus* sp. The Ashalon Member contains silicified wood.

Vertebrate fossils, discovered by Neev (1960), are quite common in the Mingar Member (Shahar's Rotem) in the Yeroham–Dimona basin, some 5–10 m below the oyster beds. Bones collected from the Rotem basin, 10 km eastward may be somewhat younger since the bearing strata contain fragments of reworked oysters (Tchernov et al. 1987, Goldsmith et al. 1988). Bones are also found in the Arava, in the Gidron Member (Savage & Tchernov 1968). The beds yielded 19 mammal taxa, mostly of African affinity, such as the proboscideans *Prodeinotherium* and *Gomphotherium*, the artiodactyl *Canthumeryx syrtensis*, two species of *Dorcatherium* and *Gazella negevensis*, rodents such as *Megapedetes*, the lagomorph *Kenyalagomys*, and the creodont *Anasinopa haasi*. These are accompanied by unidentified carnivores and primates, numerous fish remains, among which only *Lates* and Clariidae were identified. Unidentified amphibians also occur, and reptiles such as the large water turtles Trionychidae and the crocodile *C. pigotti*. The same outcrops also yielded plenty of fossil wood, usually silicified, particularly of a variety of palms (*Palmoxylon*), as well as *Leguminoxylon* (Lorch & Fahn 1959, Zohary 1959).

Tchernov et al. suggest a late early Miocene, possibly Burdigalian, age for the vertebrates, based on correlations with somewhat similar assemblages, mainly from Africa, which conforms with the setting of the bones, underlying the oyster beds. The animal and plant taxa, as well as the sediments, indicate a mainly fluviatile partly lacustrine environment of deposition for the bone-bearing beds. The vegetation must have been rich, with forested and marshy habitats, but also with open, fairly dry country not too far away, correlating very well with the type of environments indicated by Palynozone Mb pollen spectra.

No fossils are reported for the Edh Dhira Beds, and the suggested age is "?Oligocene–Miocene–Pliocene", most probably based on too liberal considerations (Wetzel & Morton 1959, p. 169). Bender (1974a, p. 92) correlated the Edh Dhira section with both the lower and upper members of the Dana Formation. If Bender's assumption, which seems logical, is accepted, the lower part of the sequence conforms with the Hazeva Formation as known from the western Arava, while the upper part could be coeval with the Ashalon, a late middle Miocene Member of the Hazeva. On the other hand, again if Bender is right, the lower sequence of the Edh Dhira may be of Oligocene age, thus probably corresponding to the Zefa Formation and the Abu Treife Series.

The lowermost beds of the Wadi Bustan Member in Gharandal contain internal molds of freshwater gastropods and charophytes. The upper beds at Dahal, 25 km south of the Dead Sea, in the Arava, contain some foraminifera, such as a "forerunner" of *Streblus*, of Oligocene to Miocene age, together with *Ammonia beccarii*. The sequence at Shubak yielded, apart from charophytes, Cichlid (freshwater fish) remains. Bender (op. cit.) therefore concluded that the outcrops of the Wadi Bustan Member in the Arava had some connections with marine environments, while the ones on the highlands, further east, were regarded as of fluvio-lacustrine origin. Bender further correlated the Wadi Bustan with the Taiyiba Beds, thus assigning an Oligocene to Miocene age to the former also.

It seems that the Hazeva Formation, the Edh Dhira Beds, the upper part of the Wadi Bustan Member (excluding the Taiyiba Beds, contrary to Bender's view) and the Herod–Lower Basalt complex, all bear rather similar sequence-stratigraphic characteristics. They overlie a considerable pre-existing erosional relief, developed mainly in synclines of the Syrian Arc and the Jordan Rift, but in places cut into elevated structures of the former system, with no previous faulting. The sedimentation usually begins with local, coarse–base conglomerates, over which finer clastics such as sandstones, siltstones and clays are superimposed. The middle parts of the suites comprise estuarine, lagoonal or freshwater, mostly chemical sediments, which are in turn overlain by fine clastics, grading to coarser toward the top, where top conglomerates are usually developed, seen only where they have escaped subsequent erosion. The upper parts of the suites are always truncated prior to any subsequent deposition, except in the deeper basins of the Jordan Valley where complete, continuous sequences were accumulated and preserved. The truncation indicates a large scale regression and drop in sea level, which terminated the depositional cycles. Sediments of all these units indicate deposition under wet subtropical conditions. It is still quite difficult, in the present state of knowledge, to see the Ghor el Qatar Series as part of this system.

All these units are therefore regarded here as a single event-stratigraphic entity. Since fossils are rare and inconclusive, this view is based on the similarities in lithologic nature and implications of the three units, as well as their sequence-stratigraphic setting and paleoclimatic inferences. All overlie Eocene rocks, over a considerable unconformity. In some rare localities part of this hiatus is to some extent filled up with Oligocene sediments, so that these formations occupy the second transgressive cycle, counting from the Eocene. The Miocene fossils of the Hazeva Formation and its interfingering marine strata to the west, as well as the radiometric ages obtained for the basalts within the Herod, seem to corroborate this age assignment. It is however quite difficult to place the upper limit, especially since the tops of the sequences are so frequently truncated to some degree or another, so that these units may well have been deposited also during the early stages of the late Miocene.

A sequence similar to those described above is known from the Barada River valley (Dubertret 1963, 1966), near Damascus, some 70 km northeast of the Hula Valley, where "red-beds" more than 400 m thick are intercalated by several basalt flows. The setting and lithology seem similar to the Herod Formation, so that the two were correlated (Picard 1943, Horowitz 1974). Indeed, there exists a series of such basins along the SW–NE oriented synclinorium bounding the Hermon–Anti-Lebanon range to the southeast (Picard 1943, p. 79). These grade into marine Miocene sediments of the northeastern Syria–southwestern Iraq basin, deposited by the sea which invaded from the Persian Gulf at that time.

The Ashalon Member, constituting the uppermost, top conglomerate part of the Hazeva Formation in the central and northern Negev, is also most probably of Mb age, at least its later part. This is inferred because in many outcrops the sediments contain numerous remains of fossil wood, indicating a wet climate corresponding to Mb, rather than the next, the extremely dry Mc. It may thus represent the regression following the middle Miocene, predating the Eritrean faulting and the Messinian drop in sea level. The "Mottled Zone" metamorphism mentioned above also affected the Ashalon Member, thus its fission track age of 13.6 ± 2 Ma, if accepted, puts an upper limit on this unit at the late-middle Miocene.

7.1.4 Palynozone Mc, late Miocene

The terminal Miocene regression, which was so pronounced in the region as to cause an almost complete drying out of the Mediterranean, the Red Sea and the Persian Gulf, so that chiefly anhydrite and rocksalt were deposited in these basins at this time (Hsü 1972), was the main reason for ceasing sedimentation and commencing erosion of the previous Hazeva, Herod and other correlative suites. Added to this, the Eritrean faulting commenced its activity during the late Miocene (Horowitz 1979, p. 61; Steinitz & Bartov 1991), affecting some of the area now occupied by the Jordan Valley, but by no means confined only to this region. The faulting is manifested as well in a variety of other regions, from the Red Sea through the Near East, up to Turkey. The combination of these two factors, regression and faulting, resulted in extensive regional erosion, thus leaving behind very little in the way of rock units. Again, the subsiding sectors of the Jordan Rift Valley, in which deposition continued, are the only localities where complete late Miocene sequences were accumulated and preserved. Such sequences are principally known from boreholes (Horowitz & Horowitz 1985, 1990), constituting Palynozone Mc.

Most of the late Miocene rocks outside the Jordan Valley basins comprise coarse clastics, in which no fossils have ever been found. Within the Rift, finer sediments had accumulated in water bodies, where pollen grains are quite abundant, designating the late Miocene sequence as Palynozone Mc. Pollen spectra of Palynozone Mc are characteristic, typified by their regional vegetation components, which are represented almost exclusively by Compositae, with very low values of arboreal pollen, lower even than those encountered in Ma times, indicating a Saharan type desert environment. The pollen spectra served for correlation of these rocks with the marine late Miocene of the offshore subsurface, dated by foraminifera (Horowitz & Derin 1987) to the N15–N17 zones, comprising the Tortonian and Messinian. Palynozone Mc constitutes the upper part of the Ziqim and the entire Mavqi'im Formation, in the Mediterranean offshore.

The late Miocene formations are thus divided into two distinct groups. One comprises a suite of truly syntectonic clastics of the Jordan Rift Valley, filling up the newly formed troughs of the Eritrean faulting system; the other, made up of top conglomerates typical of the surrounding regions, following the regional regression. The two groups are always separated by areas where erosion prevailed. While there is no problem in dating the first group, wherever penetrated by boreholes and palynologically analyzed, the second suite can only be attributed a late Miocene age according to its sequence-stratigraphic setting. Volcanics are missing from most of the exposed late Miocene sequences, except for a single basalt flow within the Umm Sabune Conglomerate (Fig. 5.2.1), dated at 8.4 ± 1.2 Ma, and a flow overlying the "lower Bira" at Marma Feiyad (Fig. 5.3.7), dated at some 6 Ma (Shaliv 1991), and possibly some other rare occurrences.

The first group of truly syntectonic clastics includes all occurrences of Palynozone Mc sediments penetrated in the Hula, the central Jordan Valley and the Dead Sea basins and their outcropping equivalents, the Umm Sabune Conglomerate to the north and the Hufeira Formation to the south, possibly also the outcrop near En Boqeq (which I believe corresponds to the earlier Ashalon) and (doubtfully) the uppermost Ghor el Qatar. The assignment suggested in Calvo et al. (1997, 1998) for the state of the Rotem Member is far from convincing. The main problem is that, if these authors' opinion is accepted, it puts the Rotem in the late Miocene, which is contrary to the age indicated by the vertebrates it yielded. I therefore prefer to retain the previous view of Harash (1967), which is to regard the "Rotem" as part of the "classic" Hazeva Formation, a view also shared by Shahar (1973) and Hirsch & Roded (in preparation). The second group, of top conglomerates, comprises such units as the Hordos Conglomerate (if Shaliv's correlation is adopted) in the Galilee and possibly(?) the uppermost part of the Ashalon Member of the Hazeva Formation in the northern Negev, both occurring outside the Jordan Valley limits.

The upper part of the Herod Formation, at the Poriyya escarpment, is rich in authigenic dolomites, indicating an aridification trend (Shaliv 1991) which most probably corresponds to the lower part of Palynozone Mc, still predating the Eritrean faulting. Similar enrichment in dolomites is seen also in the Hufeira Formation. There, going upward in the section, they display lighter oxygen isotopes composition (Calvo et al. 1996), most probably indicating the warming typical for Mc.

The Mb/Mc transition is dated at 10.4 Ma, while the basalt in the Umm Sabune is dated at 8.4 Ma. If these dates are indeed reliable, they contain the age of faulting within a narrow range, probably some nine million years ago. Heimann & Steinitz (1989) obtained an age of 8.82 ± 1.04 Ma on five samples from 2,427 to 2,431 m in Notera 3, representing the lower part of the late Miocene Palynozone Mc. An age of 9.5 Ma was measured in Zemah 1 at a depth of 3,281 m, also in the lower part of Mc (A. Heimann, Geological Survey of Israel 1998, pers. comm.). The figures from Zemah should however only be considered as minimum ages, since analysis was performed on gabbro, which may be intrusive.

The two phases of the Tortonian transgression may have reached the central Jordan Valley. The earlier one could have deposited the oyster bed occurring in the upper part of the Herod Formation (Sneh 1993), while the later could be connected with sedimentation of the basal part of the Bira Formation. These two rock units, for which no specific names have ever been formally applied on the assumption that they are an integral part of the Bira Formation, are thus indeed correlative to the Umm Sabune Conglomerate. This "lower Bira" is termed in Sneh (1996) "Late Miocene units", while Braun (1992) referred to the sequence as the "Yachza'el Formation".

7.1.5 Palynozones Pa and Pb, Pliocene

Comparisons of the pollen spectra from the Jordan Valley characterizing palynozones Pa and Pb with those obtained from the Mediterranean offshore (Horowitz & Derin 1987), indicate that Pa and Pb span the entire Pliocene, as defined by the microfauna in the Bravo 1 borehole. The Tabianian age for Palynozone Pa is further strengthened by occurrences of planktonic foraminifera in the Zemah 1 borehole sequence, at a depth of 1,750–1,777 m (Gerry & Derin 1983). These include *Globigerinoides trilobus*, *G. obliquus extremus*, *G.* spp., *Orbulina universa*, *O. suturalis*, *Globoquadrina altispira*, *Planulina ariminensis*, *Cancris auriculus*, *Asterigorinata mamilla* and *Neoeponides* sp., indicating that the Tabianian sea had reached as far as the central Jordan Valley at the time of its maximum extent.

Palynozone Pa corresponds to Foraminifera zones N18, N19 and the lower part of N20, early through middle Pliocene, while Pb corresponds to upper N20 and N21, of late Pliocene age (Horowitz & Derin 1987). Pliocene arboreal pollen are dominated by two groups of trees, the *Quercetalia* and a conifer, most probably *Picea orientalis*. The different composition of the arboreal pollen group as compared with the Miocene probably represents the beginning of development of the Mediterranean climate domain (Suc 1989, Fauquette et al. 1998). Palynozone Pa is richer in arboreal pollen as compared with Pb. Shares of up to 40% of the pollen produced by regional vegetation are typical for Pa, while Pb is characterized by figures of 20–30% for that group. There is no essential difference in the composition of non-arboreal pollen groups in the Pliocene and Miocene.

Some ten pollen spectra obtained from outcrops of the Sedom and Amora formations (Horowitz & Zak 1968, Horowitz & Horowitz 1992) are similar to those of Palynozones Pa and Pb, respectively; three pollen assemblages from the Neve Ur 2 borehole, covering the interval of 215–290 m, and two from the Gideon 2 borehole, at 190–235 m (personally analyzed), are identical to those of Palynozone Pa. These intervals, in the central Jordan Valley, are considered by Shaliv (1991) correlative to the type Bira Formation. It would thus be only reasonable to assign all these units a Pliocene age.

Based on the numerous pollen spectra, a northern Mediterranean temperate climate was concluded for the Pliocene, being more humid for Pa and somewhat drier but of a similar nature for Pb. This type of climate is unique for the Pliocene, different from all other periods in the Jordan Valley's history. The environmental gradient completely changed from the Miocene, so that during the Pliocene it was cooler and more humid to the north, becoming warmer and drier southward (Table 6.5.2.2). The lower temperatures are immediately expressed in the sediments, which are much less oxidized, typified by grayish–whitish hues of their abundant carbonates, the considerable amounts of which represent the copious lakes and lagoons and the connection with the sea. Southward, where the climate was warmer, oxidation and reddish colors do appear occasionally, especially in sediments of the Arava Formation.

Similarly to the Miocene, sea level and climate together dictate the general type of sediments. It seems quite clear that the more humid Pa, which was laid down in times of higher sea level as compared with the Pb, is richer in chemical precipitates and fine-grained sediments, while the latter is usually characterized by conglomerates. It seems that although sea level may have had more influence, the effect of the climate cannot be underestimated.

Fossils found in the Bira Formation, summed up in Shaliv (1991), include unidentified plant remains and very common redeposited Cretaceous and Eocene foraminifera (Reiss, in Schulman 1962). Foraminifera indicating a general Neogene to Recent age and brackish environment are *Ammonia beccarii* and a variety of benthonic miliolids (Weiler 1961). Other fossils, denoting lagoonal or shallow marine environments are ostracodes such as *Cyprideis torosa* (Rosenfeld, in Shaliv 1991), clams, of which *Crassostrea* and *Ostrea* (possibly *edulis*) are common, accompanied by *Chama*, *Lucina*, *Cardium*, *Lithophaga* and others, and gastropods such as *Cerithium* (Bentor 1946). Occasionally the environment of deposition becomes less saline, evidenced by occurrences of freshwater mollusks such as *Melanoides tuberculatus, Theodoxus jordani* and *Melanopsis praemorsa* (Lewy, in Shaliv 1991). Clupeid fish remains (herring and sardine family) are reported from the gypsum layers (Avnimelech & Steinitz 1951).

Fossils of the Gesher Formation include abundant ostracodes indicating a freshwater environment (Rosenfeld et al. 1981), as do the numerous shells of the gastropod *Hydrobia fraasi*. Freshwater gastropods are also abundant in the Tel Hai Formation, especially in the limestones and chalks, such as *Hydrobia fraasi*, *Chondrina jamina*, *Lymnaea*, *Planorbis* and terrestrial *Helix*. The Tanur Conglomerate did not yield any fossils. Kansou (1961) reported teeth of *Hipparion* found in the Zahlé Beds lake sediments, which made him assign them a Pontian age, while the Pliocene was suggested for the overlying conglomerates. It should however be noted that *Hipparion* lived in this region at least until the middle Pleistocene (Hooijer 1958, Tchernov 1996). A tooth of the Pliocene proboscidean *Tetralophodon* was found (Horowitz 1974) in the Gesher sediments, just northeast of Lake Kinneret, near Sheikh Ali.

The Sedom Formation is poor in fossils, except for numerous pollen grains typical of Pa (even in outcrops, the dense salt protecting them from oxidation), and occasionally unidentified plant debris. Notable are fossil fish, *Mugil priscus* (Steinitz, in Bentor & Vroman 1960), indicating a connection with the sea.

Footmarks of birds and a large mammal are reported in Zak (1967), as well as some unidentified insect remains. No fossils are reported from the Amora Formation, except for plant debris and pollen grains of Pb affinity.

The Mazzar Formation yielded foraminifera, *Ammonia beccarii* and *Elphidium* sp., and an ostracode, *Cyprideis littoralis*, indicating a Neogene to Recent marine brackish environment, similar to those found in what was designated the "Lido Facies" of the Bira Formation near Tiberias (Reiss, Gerry, in Eyal 1984) and on Mount Sedom (Zak 1967). Similar fossil assemblages are reported from the En Gedi 2 and Arava 1 boreholes, at depths where pollen spectra of Palynozone Pa are encountered (Neev & Emery 1967, Bartov, in Elron 1980, Horowitz & Horowitz 1990). Picard (1943) describes part of the Samra as a lacustrine limestone with *Hydrobia*, *Bythinia*, *Planorbis* and *Melanoides* shells. Fossils collected from the Shagur Formation (Huckriede, in Bender 1974a, p. 92) include *Melanopsis* cf. *praemorsa*, *Trichia*, *Poiretia*, ostracodes and plant remains, among which date palms were identified. An outcrop described by Bender (1974a, p. 92), situated somewhat north of Gharandal, contains poorly preserved smooth-shelled ostracodes and "dwarfed" foraminifera such as ?*Rotalia* and ?*Discorbis*, indicating in general a Neogene age and a shallow estuarine to shallow marine environment of deposition.

Based on indirect evidence of K–Ar ages of basalts east of the Rift Valley and their estimated rates of erosion, Steinitz & Bartov (1991) suggested the base of the Sedom Formation is younger than 7–9 Ma, maybe as young as 6 Ma, while the top was assigned an age of 4–3.4 Ma. The base of the Amora is placed at 4–3.4 Ma and the top at a little more than two million years. The Sedom ages correspond quite well with their assignment to Palynozone Pa (Horowitz & Horowitz 1990), as does the age for the base of the Amora. The top of the Amora seems somewhat too young, which could have resulted from inclusion of younger rocks in this formation, a rather common practice with many investigators (cf. Kashai 1988).

All the rock units discussed under Pa and Pb overly the Miocene Hazeva and Herod, or older formations. The contacts are almost always unconformable, occasionally angular, predominantly faulted and erosional. Exceptions are known only from boreholes drilled into the deeper Jordan Valley basins, from the Hula down to the southern Dead Sea, where the Neogene sequence is continuous, and possibly also in Belvoir, where the Bira (*s. str.*) overlies the late Miocene Umm Sabune Conglomerate and the lowermost part of the "Bira" paraconformably. These rock units are overlain in the deep boreholes by earliest Quaternary sediments of Palynozone QI, while in exposures, where QI rocks are rarely preserved, by younger Quaternary formations. More often than not, the contact between the Pliocene and the overlying units is unconformable, the older ones being faulted and eroded. This setting puts the major bulk of the Bira and its associated units within the Pliocene, as indeed was suggested by almost all investigators.

The lowermost part of the Bira was already deposited in late Mc times. This is based on occurrences of evaporites in the Zemah 1 borehole, at depths of

1,840–2,280 m, at the upper part of Palynozone Mc, which were deposited from sea water (Raab 1998), possibly a shallow transgression during the later part of the Tortonian, just before the Messinian. If the layers interfingering and covering the Umm Sabune Conglomerate at the Belvoir 1 borehole do indeed belong to the Bira Formation, as suggested by Shaliv (1991), then the ages of interfingering volcanics, 7.0 and 7.3 Ma, could date this minor transgression to the Tortonian, as also noted by Sneh (1993). An outcrop of Bira-like sediments also underlies a basalt flow at Marma Feiyad, dated at some 6 Ma (Shaliv 1991). This exposure is regarded by Sneh (1996) as late Miocene "Bira".

Unfortunately, none of the fossils other than pollen recovered from the Bira, Gesher, Tel Hai or the other rock units are indicative of age, since most of them already appeared at the beginning of the Miocene, and live until the present-day. However, E. Tchernov (Department of Evolution, Systematics and Ecology, the Hebrew University of Jerusalem 1999, pers. comm.) maintains that *Lucina* is typical of the Pliocene, not earlier. The *Hipparion* found in the Zahlé Beds occurs from the late Miocene until some time in the Quaternary (Tchernov 1996). It seems that, for the sake of correlations, attention should be drawn to the characteristic environments expressed by both fossils and lithology, which calls for the application of eco-stratigraphic principles. For some reason, neither this assemblage of fossils (or pollen, for that matter, if assemblage is considered) nor the types of rocks are found at any time pre- or postdating the "P" palynozones, whenever all the rock units concerned are securely dated. Added to this, whenever "P" was positively, safely identified, the rocks always indicate the characteristic Pliocene environments of the Jordan Valley and its surroundings.

Another important point is the zoogeographical significance of the fossil mollusks corresponding to the Pliocene palynozones, which are entirely Mediterranean or southern palearctic (E. Tchernov, Department of Evolution, Systematics and Ecology, the Hebrew University of Jerusalem 1999, pers. comm.). This is in accord with the significance of the pollen spectra from Palynozones Pa and Pb, and in great contrast with the characteristic tropical affinity of the Miocene fossils, or entirely Mediterranean nature of the Quaternary assemblages. It seems thus that rocks in which the above suite of fossil mollusks occur cannot be assigned neither a Miocene nor a Quaternary age.

Attention should be drawn to the significance of the two most common fossils in almost any of the rock units discussed here. These are *Ammonia beccarii*, prevalent in the lagoonal chalky–marly facies, and *Hydrobia fraasi*, very abundant in the freshwater limestones and chalks. The stratigraphic range of these two is Neogene through Recent, with *Ammonia* reported also from the Ubeidiya Formation (Almogi-Labin et al. 1995), while both genera are living in the Dead Sea region even today, where spring freshwater mingles with the hypersaline lake's water (Ehrenberg 1849, Almogi-Labin et al. 1991, 1992). Based on these occurrences, the last-mentioned authors questioned the validity of using this foraminifer as an indicator for the Pliocene transgression. However, analyses of entire borehole sequences by Derin and Gerry in Zemah 1 (in Marcus & Slager 1985), or Reiss in En Gedi 2 (in Neev & Emery 1967) never encountered *Ammonia* in any other horizon but those yielding Pa and Pb pollen spectra. *Hydrobia* is hardly referred to in the context of boreholes, for two reasons: most of these had been sunk in the deeper basins, where salinity was probably higher due to the increased influence of penetrating seawater; and mollusk shells are rarely preserved in well-cuttings.

We thus conclude that although far from being strictly time indicators, the Pliocene was a period during which these two bugs had been flourishing, and were considerably more abundant than at any other period in the history of the Jordan Valley, each prolific in its specific environment. It appears that the relevant environments were much more common at that time, so it seems therefore quite safe to use this assemblage of fossils for correlation, in conjunction with all other seemingly inconclusive parameters mentioned above, thus regarding all units mentioned in Chapter 5.2.2 as of Pliocene age. Truly, this only constitutes circumstantial evidence, but plenty of it. Naturally, wherever outcrops are subordinate, it would be quite difficult to discern Pa from Pb. Needless to repeat, whenever pollen spectra are available no doubt as to the Pliocene age exists. The only authors who challenged the Pliocene assignment to the Bira are Shaliv (1991) and Heimann et al. (1996), who attributed the Formation to the late Miocene, based on radiogenic ages obtained for the presumably overlying the Cover Basalt. The discrepancy seems to lie in the stratigraphic definition of the Cover Basalt, which was never satisfactorily presented. The original definition by Schulman (1962) attributed to the Cover Basalt all volcanics overlying the Bira or Gesher formations, assuming that both predate this eruption phase. However, the type section was proposed by Schulman at a locality where the "Cover Basalt" overlies some Bira sediments, with no Gesher at all. Heimann et al. (1996) maintain that (p. 58) "The definition of a type section (*s. str.*) in volcanic sequences is not clear in general, and in this case in particular".

Horowitz (1973, 1974) and Siedner & Horowitz (1974), followed by Horowitz (1979, p. 155) regard only those volcanics strictly overlying the Gesher Formation west of the Jordan Valley as "Cover Basalt", thus attributing it an earliest Quaternary ("Preglacial Pleistocene") age, based on radiogenic dates of ca. two million years for flows unconformably overlying the Gesher. This, naturally, is only a minimum age for the Gesher, which may be considerably older. They further regarded the Fajjas Tuff as volcanics interfingering with the Gesher, while the term Intermediate Basalt was saved for eruptions at the time of the Bira deposition, as primarily suggested for both by Schulman (1962).

Conversely, Heimann et al. named "Intermediate Basalt" those flows interfingering with late Miocene formations, such as the Umm Sabune Conglomerate, suggesting a view of this phase as either the last of the Lower Basalt, or an independent minor one, at times of volcanic "quietness", separating the major eruptions of the Lower and Cover basalts. Heimann et al. also include in their

"Cover Basalt" the Fajjas Tuff, and in fact any volcanics that even partly overlie the Bira sediments, regardless of which part of the Bira sequence it is (as long as the basalts are not of Quaternary age), or where, geographically, the sections are. The last note is important, since Horowitz (1974) viewed the Gesher Formation east of the Jordan Valley as a time equivalent of the Bira to the west, the two differing only somewhat in the degree of salinity of the respective water bodies from which they had been deposited.

It should be noted that if any of these three basalts does not appear in a clear stratigraphic context they are impossible to tell apart. As Heimann et al. (1996, p. 59) put it: "This is especially critical when dealing with late Cenozoic basalts in northern Israel which have similar field appearance and paleoflow directions, similar petrographic features, and a similar major element geochemistry." Unfortunately, most (if not all) of their dated samples are not in a clear stratigraphic context. If one accepts Schulman's and Horowitz's definitions, only flows overlying the Gesher Formation west of the Jordan Valley should have been treated under the term "Cover Basalt"; however no such sample was dated by Shaliv nor by Heimann et al. All dated samples which do have connections with the Bira (or the Gesher on the Golan Heights), seemingly overlie the sediments, sometimes unconformably.

The significance of these field relations may be viewed in different ways. The simplest is advanced in Heimann et al., who maintain that their "Cover Basalt" is entirely younger than the Bira, but contemporaneous with the Gesher. Based on that, they assign the Bira Formation and the Intermediate Basalt a late Miocene age (late Tortonian through Messinian according to Shaliv 1991), contrary to all other investigators. This assignment is problematic for several reasons. The Bira sediments interfinger westward with the "Yagur Facies" containing typical Pliocene (Astian) fossils (Picard & Solomonica 1936); the Bira sediments in the Yizre'el Valley contain two oyster or lumachel banks (Picard 1943), indicating a twofold transgressive cycle typical of the Pliocene; the Bira and Gesher complex is transgressive, while the late Miocene, particularly the Messinian, is a period of extensive Mediterranean regression; pollen spectra favor correlation of the Bira with the marine Pliocene; and other fossils, although each not entirely conclusive, all point to the Pliocene as the most plausible age.

The solution to these discrepancies may lie in a different view, suggested here, but hinted at already by Bentor (1946), Vroman (1958) and in part by Sneh et al. (1998a), who marked in their maps the type section of the "Cover Basalt" as "Lower and Intermediate Basalt". It seems to me that all the samples dated by Heimann et al. (1996) represent volcanic activity at the same time as the Bira deposition, not later than the Gesher, as it would have been if Schulman's definition had been strictly followed. Notably, Siedner & Horowitz (1974), who adopted Schulman's definition precisely, published dates for the Cover Basalt in the range of two million years. The picture is thus similar to what is known for the relations of the Herod sediments and the Lower Basalt, of contemporaneous deposition and volcanism.

Since the center of Pliocene volcanic activity, as noted by Heimann et al., traveled northward with time, it may well be that eruptions had inundated different parts of the Bira basin at disparate localities through time, seemingly overlying the latter Formation. The same holds for dates obtained for the "Cover Basalt" interfingering with the Gesher Formation on the Golan, east of the Jordan Valley. If one accepts the proposition that the Gesher there fully represents a freshwater facies of the lagoonal Bira closer to the sea, located to the west, then it becomes clear that the ages obtained for the "Cover Basalt" by Heimann et al. correspond to the time of deposition of the Bira Formation, not of a younger unit that overlies it. Thus, the time interval allocated for the Bira (and maybe for the lower part of the overlying Gesher west of the Jordan Valley), would be in the range of 5.5–3.3 Ma, which fits nicely with the ages suggested by the palynostratigraphic correlations to Palynozone Pa (5.3–3.5 Ma), and possibly the beginning of Pb (Table 1.4.1).

It therefore emerges that, keeping Schulman's definitions, volcanic activity in the central and northern Jordan Valley can be divided into four distinct phases. The Lower Basalt, during the interval 17.5–12 Ma (Shaliv 1991); the late Miocene volcanism around 9–6 million years ago; the Intermediate Basalt, as recognized here, at 5.5–3.3 Ma (Heimann et al. 1996); and the Cover Basalt, a phase which commenced some 2.5–2.7 Ma ago, practically continuing until the present-day, particularly on the Golan Heights. This continuity is clearly seen there (A. Sneh, Geological Survey of Israel 1999, pers. comm.), although in some places the sequence could be subdivided into distinct "basalts" or "flows" (cf. Mor 1986). It would therefore appear that the Intermediate Basalt is indeed a considerable volcanic phase, active throughout most of the Pliocene, including what was previously termed the "Fajjas Tuff". This phase inundated much larger areas than previously thought, when some of its activity was mistakenly attributed to the Cover Basalt. It is suggested not to use particular names for the rare late Miocene flows, which do not correspond in time to any of the three major phases. At present it seems sufficient to regard them simply as "flows within this or that unit", mainly due to their scarcity and subordinate distribution.

It seems that the Pliocene age for Palynozones Pa and Pb encountered in numerous boreholes (Horowitz & Horowitz 1990), the Bira–Gesher formations, the Sedom–Amora suite and the Intermediate Basalt (as originally defined, but considerably more widely distributed) is quite well established. The problem remains whether satisfactory correlations with this suite could be proposed for other rock units, despite their fragmentary distribution of limited, solitary outcrops, poverty in fossils or lack of diagnostic ones, and oxidized exposures yielding no pollen. In terms of distribution, the two large areas in which Pliocene rocks occur are the central Jordan Valley, whose exposures extend westward to the Mediterranean through the Yizre'el Valley, and the Dead Sea–Arava region. The main body of formations to the north comprises the Bira–Gesher and the Intermediate Basalt (as defined above), while to the south the Sedom–Amora makes up the bulk of the Pliocene sequence.

Horowitz (1973) proposed correlation of the Bira–Gesher suite with the Tel Hai–Tanur formations to the north, extending further to the Zahlé Beds in the Lebanon. The late Miocene age proposed by several authors for the Lebanese and the northern Hula formations (Kansou 1961, Sneh 1996) is not accepted here, except possibly for the lowermost part, as in the central Jordan Valley. The Pliocene age for these is based on their typical fauna, lithology, transgressive nature and paleoclimatic implications: the saline Bira becomes fresher as one goes north, where the freshwater supply is ample, contrary to the environmental gradient for the Miocene. This trend is seen even further in the Beqa'a, where an extensive sequence of freshwater sediments comprises the Zahlé Beds. The Tel Hai–Tanur complex extends somewhat southward, known from boreholes in the Hula Valley and the northern part of the Korazim block, where it interfingers with flows of the Intermediate Basalt (Fleischer 1968, Horowitz & Horowitz 1985, Heimann 1990).

A similar situation occurs on the southern Golan Heights, where lagoonal sediments are not known, the entire Pliocene interval taken by freshwater deposits of the Gesher Formation. In this area the Gesher is also interfingered by basalt flows, whose radiometric dates (Heimann et al. 1996) do not leave any doubt concerning their Pliocene age. Based on these correlations, Horowitz (1973) suggested regarding all the northern units as the "Bira Series", representing the Pliocene transgressive complex, their differences in lithology resulting from the location of any particular place in a system consisting of rivers, lakes and lagoons, connected further west to the Mediterranean. Hypersaline lagoons in which gypsum and rock-salt were deposited during Pa are known southward, from the Zemah 1 borehole (Marcus & Slager 1985) and from the vicinity of the Gesher and Menahemya settlements (Schulman 1959, 1962), most probably extending further south (see below).

There is no direct, physical connection between the Pliocene formations of the central Jordan Valley and those to the north, in the Hula and Beqa'a, nor with those known from the southern Golan, nor with the Dead Sea–Arava region. The various possibilities for such connections are discussed when dealing with paleogeography, in Section 7.2.5.

Discussion of the southern rock units is centered around the Sedom–Amora complex, which is the best studied and dated. Unfortunately no volcanics occur in the southern Pliocene, so that no direct help in dating, as is so useful to the north, could be found in this region. However, the numerous pollen in both outcrops and boreholes seem to leave no doubt as to the Pa nature of the Sedom and Pb of the Amora. Correlation of the Arava Formation with the Sedom–Amora suite, particularly with the younger Amora, was suggested by many (Zak 1967, Horowitz 1979, p. 79, Agnon 1983, Sa'ar 1986, Steinitz & Bartov 1991). The correlation is based on occurrences of magmatic pebbles typical of the Arava conglomerates within the Amora, and on paleogeographic considerations which tied up the Arava both with the Sedom and the Amora (Garfunkel & Horowitz 1966).

The Mazzar Formation was correlated by Zak and by Horowitz with the lower part of Sedom, where fossils of *Mugil* and remains of other marine fish occur. The Mazzar interfingers with the Arava conglomerates south of the Dead Sea (Agnon 1983, Eyal 1984, Avni 1998), which makes it a part of the Pliocene suite, as also appears from its typical fossils. Zak (1967) proposed that the Mazzar represents the maximum level of the Pliocene sea in the region, at the beginning of deposition of the Sedom Formation. This assumption is supported by finds in the Zemah 1 borehole (Gerry & Derin 1983), where planktonic foraminifera indicating a deeper sea occur only at the lower part of Palynozone Pa. On similar grounds, the Oolitic, Shagur and lower part of the Samra are all regarded as correlatives of the maximum Tabianian marine transgression.

Occurrences of Precambrian magmatic pebbles in both the Arava and Amora formations led to their correlation (Zak 1967, and many others). The Sedom Formation, which includes several fine clastics intercalations, does not carry magmatics, hence its correlation with the Arava is occasionally doubted (Agnon 1983). The lack of magmatics could be explained by the distant location of the source, and by the lengthy period needed for the Arava river system to develop, thus supplying the magmatics to the Dead Sea basin only at the last stages of the Pliocene. This may also be supported by the paleoclimate, which was drier during Pb, possibly characterized by floods which would carry the coarse pebbles a greater distance. The larger extent of the Sedom basin (Shulman & Ben-Avraham 1999) could also hinder pebbles from arriving at its central part, where most of the rocksalt was accumulated.

The restricted occurrences of isolated "spots", probably belonging to the Mazzar, in many localities along the Arava (Bender 1974b, Sneh et al. 1998a, Avni 1998), is important mainly for paleogeographic reconstructions. Some confusion arose when almost every limy–marly occurrence was referred to as "Mazzar" (or if more gravelly occasionally as "Samra"), so that only those bearing the characteristic fossils are included here within the Formation. The lacustrine Kuntilla Member (Na2) of the Arava (Avni & Rosenfeld 1996, Avni 1998) directly overlies the base conglomerate of this Formation (Saggi Member, Na1), thus occupying a similar position as the Mazzar in relation to the Pliocene sequence of events, representing the highest sea level at the time. The suggestion of Avni to regard the "Arava Lake" as correlative with the Mazzar, is therefore fully acceptable, but not including all the outcrops suggested by him, some of which proved to be younger (Livnat & Kronfeld 1990).

A crucial problem of correlation exists between the southern and northern ends of the Dead Sea, which is surrounded by steep escarpments both east and west, from which most evidence was eroded. There are however two possible links: the entire northern basin of the Dead Sea is lined, according to geophysical data, by rocksalt correlative to the Sedom or Amora Formation, underlying a thin veneer of late Quaternary sediments (Neev & Hall 1979); the En Feshha Conglomerate is quite abundant on both shores of the Dead Sea, almost along its entire length.

Raz (1983) suggests, on morpho-structural grounds, its correlation with the Amora Formation to the south and the Samra to the north, an idea already raised in Horowitz (1974).

The Samra (as originally defined by Picard 1943, and not as later misused for any gravel bed), Oolitic and Shagur formations, as well as the rocks described by Shahar (1969) from Nabi Musa, all occurring north of the Dead Sea, have in common typical (although not age-conclusive) Pliocene mollusks, and a similar morpho-structural setting. They are all affected by the principal faults of the Levantine system, thus predating the early Quaternary. It seems that all these units are broadly synchronous, between themselves as well as with the Sedom–Amora suite, as proposed in Horowitz (1974) and others, seemingly with no opponents.

An occurrence of lacustrine Pliocene sediments, termed the Jabal Bakiya Formation, is reported by Baubron et al. (1985) from the confluence of Wadi Dhuleil and Wadi Zarqa cutting into the Transjordanian highlands, some 20 km north of Amman, 40–50 km east of the Jordan Valley. This sequence, sandwiched between basalts 7.5–4.5 and 4–5 Ma old, is of early Pliocene age, most probably corresponding to at least part of the Bira–Sedom formations complex. If indeed it was connected with the Jordan Valley, and was not merely a local lake, it testifies to the considerable extension of the early Pliocene inundation.

7.1.6 Palynozone QI, earliest Quaternary

The earliest Quaternary Palynozone QI (Levin & Horowitz 1987) corresponds to the lower part of Foraminifera Zone N22 (Horowitz & Derin 1987, Table 1.4.1). This time span was previously termed "Preglacial Pleistocene" (Horowitz 1979, p. 6), while some authors consider it to represent the latest part of the Pliocene. Correlations of paleoclimatic trends with oceanic oxygen isotope curves (Horowitz 1989c) show that QI corresponds to stages 103–73 (Fig. 6.6.2), the early part of Matuyama. QI is subdivided into QIa, stages 103–87, typified by very high proportions of arboreal pollen, dominated by *Picea orientalis*; QIb, stages 87–81, has lower arboreal pollen values, in which *Quercetalia* are frequent, while the upper, QIc, stages 81–73, resembles QIa by the very high arboreal pollen values, again dominated by *P. orientalis*.

Palynozone QI is very distinctive in respect of its lithology and pollen spectra in the boreholes, particularly in the Dead Sea region, where it was designated the Melekh Sedom Sands, recently identified also from outcrops in the northeastern Arava. Its correlation with outcropping rock units may however turn quite difficult, since most exposures, save for two exceptions, are devoid of pollen, due to subsequent oxidation. Correlation is thus based on sequence stratigraphy: whenever a formation overlies the late Pliocene rocks or erosive surface, and is in turn cut by the entire sequence of Quaternary erosion phases, it is considered as of QI age. Examples are the HaMeshar and Ar Risha Gravels formations, the Bethlehem Conglomerate, the Mahanayim Marl and the Amud Conglomerate.

The Ghor el Qatar Series, previously thought to be a part of this system (Horowitz 1974, 1979, p. 119), is no longer considered as such, and lately a Miocene age is preferred (see Chapter 5.1.2.6). It is however my feeling that the last word has not yet been said, so this Series may find its way back to QI times.

An occurrence of alluvial sediments termed the Dhuleil Formation is reported by Baubron et al. (1985) from the confluence of Wadi Dhuleil and Wadi Zarqa, cutting into the Transjordanian highlands, some 20 km north of Amman, 40–50 km east of the Jordan Valley. This sequence, sandwiched between basalts 3.5–5 and 2.3–3 Ma old, is most probably of QI age. Similar sediments were personally observed some 65 km north of Amman (Fig. 5.2.19) in roadcuts, covered by caliche crust.

Other fossils indicating a similar age would be helpful but except for a single occurrence have not yet been found. Excavations of the Bethlehem Conglomerate in this city yielded mammal bones (Hooijer 1958) of *Archidiskodon planifrons*, *Leptobos*, *Hipparion*, *Giraffa* cf. *camelopardalis*, *Felis*, *Hippopotamus*, *Bos*, *Rhinoceros* cf. *etruscus*, *Stegodon* and *Elephas*, which led to the generally accepted view of a Villafranchian age for this formation. Pollen spectra with high arboreal shares, dominated by *Picea orientalis*, were obtained (Horowitz 1993, pers. obs.) from the Bethlehem outcrop on Mount Scopus, Jerusalem, as well as from outcrops of the HaMeshar Formation in the southern Negev, corroborating the proposed correlations.

The only radiogenic ages for basalts interfingering with QI sediments come from the Notera 3 borehole in the Hula basin (Heimann & Steinitz 1989, Heimann 1990), for the interval 1,245–1,580 m (QI was identified at 1,300–1,600, Horowitz & Horowitz 1985). The basalts yielded an average age of 2.72 ± 0.16 Ma, which is, although barely, within the limits set for QI.

7.2 PALEOGEOGRAPHY

Paleogeographic reconstructions are based primarily on two foundations: chronostratigraphic correlations of the various rock units involved, linked with interpretation of their specific environments of deposition and the particular geographic distribution of each. The limitations set on exact age assignments in primarily continental environments are detailed above, in Section 7.1, while conclusions about past depositional environments seems more straightforward, posing less critical questions. Distribution may turn problematic due to subsequent erosion, coverage or structural disturbances. It is however appropriate, if one variable is not entirely secure, to make reconstructions of only a preliminary nature. Thus care is taken and prudence applied when the paleogeography is outlined, since there still remain plenty of data to be collected or refined. It is, however, my belief that the present body of known information is sufficient for delineating at least the general outlines of the development of the paleogeography of the southern Levant from the Oligocene throughout the late Cenozoic and up to the present day.

The first period of subsidence for which there is positive evidence along the Jordan Valley Rift, is the middle Oligocene, when accumulations of sediments both in the central Jordan Valley and the southern Dead Sea basins exceed those known outside the Rift, by almost an order of magnitude. This calls for some consideration of earlier paleogeography, Eocene through early Oligocene which served as a foundation for events to come. Depositional patterns until the middle Eocene were largely controlled by the distribution of structures connected with the Syrian Arc (Benjamini 1980, 1984, Sneh 1996), with no apparent preference of any lineament hinting at the future Jordan Rift Valley.

An uplift affected the southern Levant some time at the end of the middle Eocene, causing the retreat of the Tethys and channeling of the elevated terrains (Neev 1960, 1979, Martinotti 1981), particularly in the coastal plains of southern Israel and Sinai (Derin 1974). This was somewhat accentuated by a eustatic regression at this time. These channels were filled up during the late Eocene Priabonian transgression, which spread all the way down to the Gulf of Aqaba (Benjamini 1984). Relics of rocks deposited during this stage are known from all over the southern Levant, many of which are apparently concentrated along the Arava and the northern part of the Jordan Valley. They are also known in other localities, and there is no increase in their thickness while approaching the Rift. It thus seems that their present distribution is concentrated in low structures, since these are the places where the rocks were better sheltered from subsequent erosion.

The drop in global sea level during the transition from the Eocene to the Oligocene caused considerable erosion, along the same lines as its predecessor following the middle Eocene. It is also possible that further uplift of the southern Levant aided the process. When the sea rose again, toward the middle Oligocene, it found ways to reach the Jordan Valley, from both the east–northeast, to the central Jordan Valley, and the west (or the south? see below) to its southern sector.

7.2.1 Middle Oligocene

Evidently, due to the paucity of data, it is quite difficult to outline in any detail the middle Oligocene paleogeography of the southern Levant (Fig. 7.1). Following the retreat of the late Eocene sea and differential uplift at that time (and some volcanism, but far from the Jordan Valley), drainage systems began to create channels toward the falling base levels, both the Tethys to the west–northwest and the Indian Ocean to the east–southeast. The paths of these drainage systems were, to a great extent, dictated by the topography formed due to the differential uplift, in conjunction with structures inherited from the folding of the Syrian Arc, which left two main low-lying regions draining the Jordan Valley. To the north, through

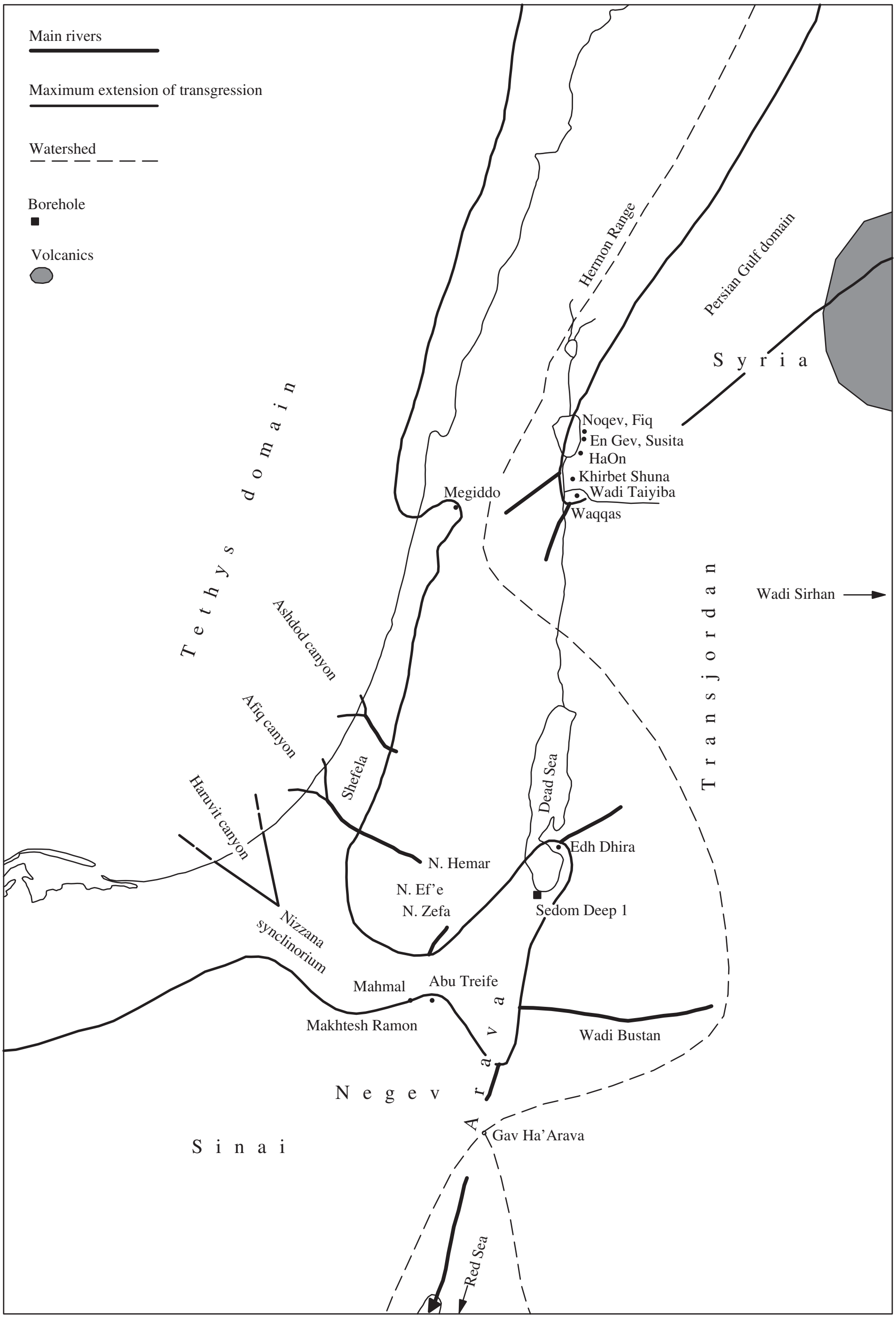


Figure 7.1. Middle Oligocene paleogeography. N.: Nahal (wadi, stream). Some of the data from Buchbinder & Siman-Tov (2000).

the great synclinorium south of the Hermon range, which continues to the Iraq–Iran basin; and to the south, a lowland extending from northern Sinai through the central Negev, joining the southern Dead Sea and Arava.

The ensuing rivers flowed mainly through synclines, but occasionally also cut through slowly rising anticlinal ridges. Elevated structures were eroded, while the seas invaded some of the lower-lying synclines during the middle Oligocene transgression, leaving littoral-neritic, lagoonal or paludine sediments, depending on the morphology, occasionally over a base conglomerate, frequently over basal beds rich in glauconite. The beds deposited during the maximum of the transgression were covered by clastics following the late Oligocene regression, which also caused increased erosion in most localities. It seems that this erosion phase, together with the general shallowness of the invading middle Oligocene seas, which did not favor any considerable accumulation of sediments, are the principal reasons for the poverty of rocks representing this time span in the southern Levant. The erosion, prevailing during the entire Oligocene in large parts of the southern Levant, caused peneplanation of most of the area. Relics of this peneplain were discerned in several localities in the Negev (Garfunkel & Horowitz 1966, Zilberman 1992), and possibly also in the Galilee (Matmon 2000).

Within the area now occupied by the Jordan Rift Valley, at least two localities were subject to subsidence in excess of the neighboring regions, east of Lake Kinneret and somewhat southward, and the southern Dead Sea basin. There is no sign of faulting connected with this subsidence, certainly not predating it, so it seems that ideas claiming that rifting had already begun in late Cretaceous times (Freund et al. 1970), or that it happened throughout the Oligocene (Sneh 1996), do not find any support in middle Oligocene paleogeography. The notable accelerated subsidence in the Jordan Valley area during the middle Oligocene marks the first step of the Embryonic stage of tectonics particular to this region, which was already different from its surroundings.

The types of rocks and the rich, varied benthonic fauna recovered from the Fiq Formation as well as the Taiyiba Beds, indicate a littoral-neritic environment of deposition for the two units, during the early-middle Oligocene, as well as for the Susita Formation (Lipson 1971, Michelson & Lipson-Benitah 1986). Mudcracks noticed in the upper part of the Noqev Member indicate a coastal environment which had occasionally dried up. Whenever some subtle uplift or minor drop in sea level occurred the region emerged, and was immediately subject to erosion.

The lower part of the "Usdom Group" at Wadi Taiyiba seems to have similar characteristics as the Susita. The transgressive nature of these units is clear from the considerable amounts of glauconite grains and limonite nodules they contain, especially in the basal parts of the first two. The overlying En Gev Sands, and possibly also the lowermost sands of the "Usdom Group" at Wadi Taiyiba, were deposited in fan deltas following retreat of the sea during the late Oligocene. Thus an almost complete transgressive cycle of Oligocene age seems to be represented in the area east and southeast of Lake Kinneret.

Slight angular and erosional unconformities exist between the various units, as well as at the base and top of the Oligocene suites. These indicate that the basins of deposition were shaped by slow, continuous subsidence, which did not affect other structures of the Syrian Arc, which was apparently not active during this period. Similar processes also continued, controlling the deposition of the overlying Miocene sediments (Garfunkel & Horowitz 1966, Michelson 1972).

A transgressive cycle is also evident for the Zefa Formation, the Abu Treife Series and the Wadi Bustan Member (if their Oligocene age is accepted), with their middle part deposited in a marshy or lagoonal environment, possibly connected to the sea. However, due to the very limited distribution of these formations and the doubt concerning their age due to lack of any diagnostic fossils, all paleogeographic inferences concerning these units are only of a preliminary nature.

It is quite difficult to assess the exact distribution of the suite exposed east of Lake Kinneret, nor the Taiyiba Beds, mainly because the entire area is covered by younger volcanics. Daniel (1963, p. 377) remarks that "similar beds are not known elsewhere in East Jordan" (Transjordan). However, he suggests correlations of the Taiyiba Beds with other, rare occurrences of possibly Oligocene sediments penetrated by several boreholes, quite far to the east of the outcrops, and with the Oligocene sequence reported in Wetzel & Morton (1959) at Wadi Sirhan, 150 km to the southeast. Incidentally, Wetzel & Morton had not connected the Jordan Valley and Wadi Sirhan outcrops, but rather thought that the Taiyiba Beds were deposited by a sea connected to the Mediterranean, and not the Indian Ocean as in the Wadi Sirhan sequence.

The occurrences of *Austrotrillina* in the En Gev suite seem to corroborate Daniel's correlation, namely that the middle Oligocene sea transgressed from the east–northeast, and covered most of northern and eastern Transjordan, while the south must have been elevated (Bender 1974b). This conclusion, arrived at also by Michelson (1972) and Horowitz (1974), is further strengthened by the total lack of Oligocene sediments from most synclines of the central hilly backbone of Israel. The closest middle Oligocene outcrop is near Megiddo, about 40 km west of the central Jordan Valley, situated on the western flank of the central hilly backbone of Israel, where a sequence of some 50 m is described in Avnimelech (1939), from a large syncline directly connecting this area with the Mediterranean coastal plain. Oligocene rocks occur subordinately in parts of the western Shefela synclinorium (Avnimelech 1936), and are also known from northern Sinai, in a coastal or very shallow marine facies (B. Derin, Consulting and Geological Services Ltd. Ramat Gan 1999, pers. comm.).

Middle Oligocene sediments usually comprise only subordinate sequences in the Israeli Mediterranean offshore (Derin & Reiss 1973, Martinotti 1981, Buchbinder & Siman-Tov 2000). In several boreholes they are missing altogether, where Miocene rocks directly overlie a variety of Eocene units, or occasionally even early Cretaceous formations. The sole exception to this rule is in the northeastern Sinai, where the group of Haruvit boreholes encountered more than 600 m of Oligocene sediments, rich in sands, unconformably overlying late Eocene rocks, both filling up a canyon more than 1 km deep, cut down to the Albian Yakhini Formation (Derin 1976). Two other canyons are known from boreholes northward, at Gaza and Afiq (Buchbinder & Siman-Tov 2000), but these are not as deep as the one at Haruvit.

Furthermore, an indication that the northwestern Negev region was elevated at that time comes from Miocene sediments which fill up a previously cut erosion channel several hundred meters deep (Neev 1960), where no Oligocene was encountered. The cutting of this canyon must thus have predated the early Miocene, most probably due to the regression at the Oligocene–Miocene boundary. Martinotti (1981) also pointed out a similar paleogeographic setting while analyzing middle Oligocene rocks from boreholes and outcrops in southwestern Israel. Conversely, Zilberman (1992) and Buchbinder & Zilberman (1997) propose that the southern Levant was low-lying during the Oligocene, while the canyons were cut only in Miocene times, an idea not in accord with micropaleontological investigations of the canyon fill (Derin 1976, Martinotti 1981), which indicate some are lined with middle Oligocene, others with early Miocene and occasionally even late Eocene sediments. Incidentally, there probably was a subsequent Miocene phase of channeling that affected different regions (see below), which may be the reason for confusion.

A barrier, possibly already created by the Syrian Arc folding, existed during the Oligocene (and most of the Miocene, see below) between the Tethys domain to the west–northwest and the Indian Ocean realm to the east–southeast. This barrier was wider to the south, gradually narrowing northward, but still occurs in northwestern Syria, where late Eocene and Oligocene marine sediments are practically absent (Krasheninnikov 1994).

The middle Oligocene layers penetrated by the Sedom Deep 1 borehole just south of the Dead Sea, comprising gray marls rich in pollen and nannoplankton, have been deposited in a marine or lagoonal environment. It is impossible to say with certainty from which direction the sea invaded at that time, since all we have is a single subsurface locality, with no fossils indicative of provenance. The middle Oligocene marls are overlain at the borehole by red beds of the late Oligocene–early Miocene Palynozone Ma, deposited in a predominantly continental environment. It seems that in this locality too, although the base was not penetrated, the Oligocene sequence represents a transgressive cycle.

To the south, when the occurrences of Sedom Deep 1 in the southern Dead Sea and Wadi Bustan in the Arava are connected with the Zefa and Abu Treife (if indeed all are of Oligocene age, a reservation which must be kept in mind), the natural continuation to the Mediterranean would be through the large Nizzana synclinorium. It seems that the canyon of Haruvit, with its exceptional thickness of sandy Oligocene rocks, justifies the plausibility of this assumption. The possibility raised by Bender (1974b), that the middle Oligocene sea could have reached the Arava and possibly also the Dead Sea from the Red Sea to the south, cannot be dismissed altogether. Marine middle Oligocene (Foraminifera Zone N2, referred to by the authors as "late Oligocene") is known from the northern Red Sea, at Midyan (Dullo et al. 1983, Abou Ouf & Gheith 1998), so the sea could easily have invaded northward. It seems however that Gav Ha'Arava may have already been elevated in Oligocene times, since it is known to have acted as a watershed during the earliest Miocene (Garfunkel et al. 1974). The large amounts of sand in the considerable Oligocene sequence at Haruvit must have arrived there from east of the Arava, as no other source was available at that time west of this region. These seem to support the Mediterranean connection.

7.2.2 Palynozone Ma, late Oligocene to early Miocene

Palynozone Ma spans the late Oligocene and a major part of the early Miocene, a period of global drop in sea level. This regression continued the trends of its predecessor, the late Eocene–early Oligocene, in almost all respects (Fig. 7.2). The main process was erosion toward the retreating seas, causing deepening of channels and peneplanation of the higher terrains. The regressing seas had "pulled up" sediments, which make up the uppermost part of the middle Oligocene formations at the beginning of Ma, and continued to remove them and parts of the underlying rocks when the seas were farthest away. Deposition of base conglomerates began when the sea began to rise again, toward the later part of Ma. The earlier stage of Ma is thus characterized by such formations as the En Gev Sands, the top conglomerate of the Zefa Formation and possibly the upper sandy part of the Abu Treife, while its later part is typified by the base conglomerates of both the Hazeva–Dana and the Herod–Barada. The southern Dead Sea basin continued to subside, accumulating a complete sequence of Ma sediments. Unfortunately no data are available for the central Jordan Valley at that time.

The areas covered by these formations are very limited, and confined to the deeper parts of the drainage systems. The subsequent middle Miocene transgression caused filling of the channels, mainly with Mb sediments, so the best way to reconstruct Ma paleogeography would be to follow up the distribution of Palynozone Mb formations.

In the Lake Kinneret region, the En Gev Sands, and possibly also the lowermost sands of the "Usdom Group" in Wadi Taiyiba, were deposited in fan deltas, following the retreat of the sea in late Oligocene times. The middle Oligocene marls are overlain at the Sedom Deep 1 borehole, in the southern Dead Sea basin, by red-beds of the late Oligocene–early Miocene Palynozone Ma, deposited in a continental environment. The lagoonal–paludine middle part of the Zefa is covered by clastics, most probably also following the late Oligocene regression, at the beginning of Ma, while the Abu Treife also behaves along comparable lines. Further regression caused erosion in most localities, strongest during the Oligocene–Miocene boundary time. It seems, as before, that this erosion phase, together with the general shallowness of the invading middle Oligocene seas, which did not

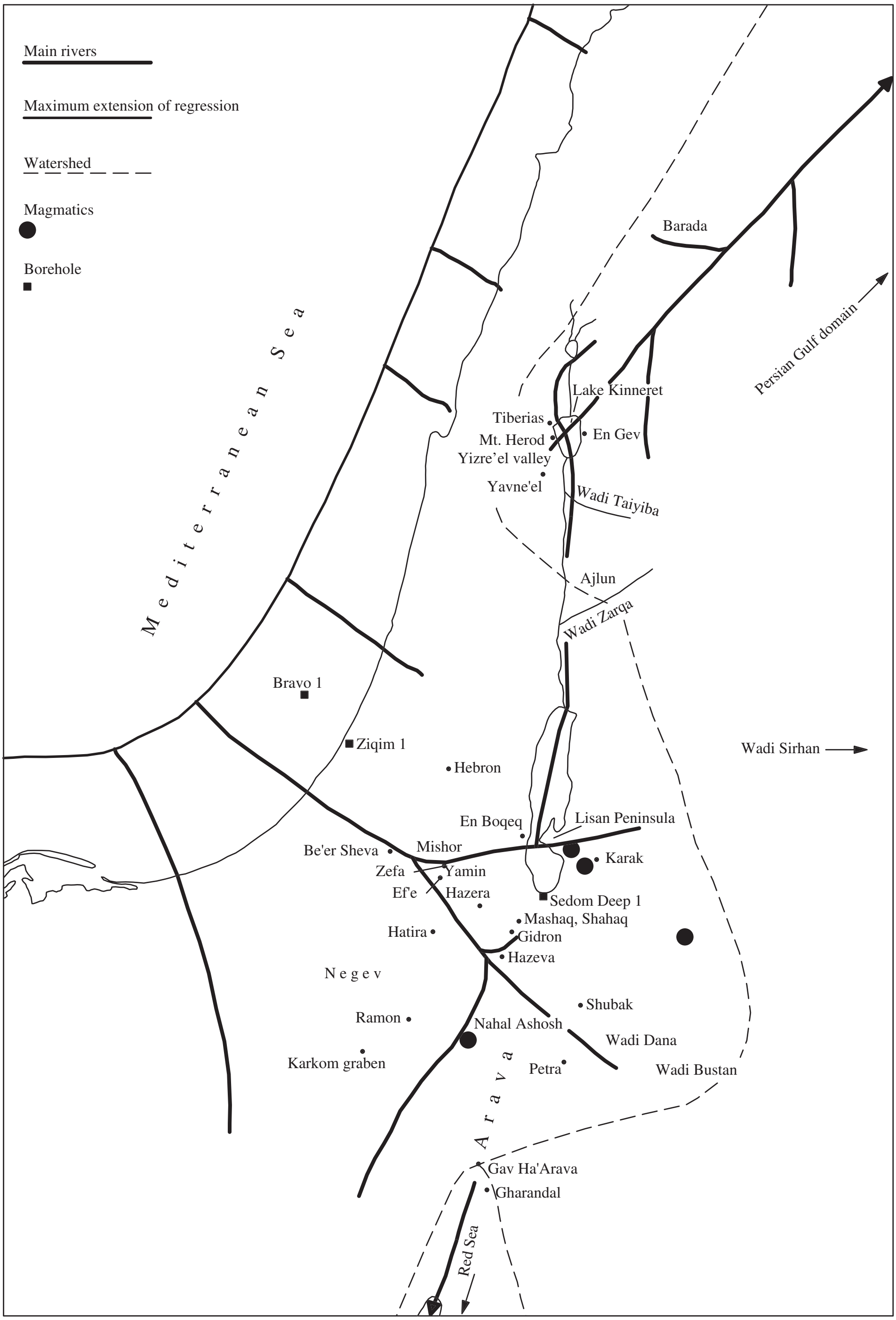


Figure 7.2. Palynozone Ma paleogeography, late Oligocene to early Miocene.

favor any considerable accumulation of sediments, is the principal reason for the poverty of rocks representing the Ma time span in the southern Levant.

While reconstructing the Oligocene paleogeography, both Michelson (1972) and Horowitz (1974; 1979, p. 73) indicated the possible connection of the pre-Herod and Herod drainage systems to the Persian Gulf. Michelson founded his conclusions only on occurrences of the foraminifer *Austrotrillina* in the Oligocene En Gev suite, seeing no fundamental paleogeographic difference between these rocks and the overlying Herod sediments. Horowitz (1974) added paleo-geomorphological criteria, particularly those regarding the early and middle Miocene watershed, to support Michelson's ideas.

The channeling thus went from the Jordan Valley in two directions: the north-eastern sector was drained to the Persian Gulf, continuing the middle Oligocene tradition, while the southern regions were drained to the Mediterranean (Horowitz 1974). In contrast with the location of the middle Oligocene principal river, the late Oligocene system draining the southern Dead Sea and the Arava, as well as large parts of the Negev, had changed its main course. The new path now went through the synclines of the northern Negev and the Be'er Sheva synclinorium, joining the Mediterranean some 40 km to the north of the previous, middle Oligocene outlet. This shift is seen by the depth and extent of the northern channel (Neev 1960), as well as by the extensive sequences of sediments, especially those representing almost the entire Miocene period, in boreholes such as Ziqim 1 and Bravo 1, where they are more than a kilometer thick (Derin & Reiss 1973). The Haruvit boreholes to the southwest, where the middle Oligocene sequence is exceptionally developed, penetrated only subordinate Miocene sections.

The change of paleogeography to the south during Ma times most probably resulted from the beginning of uplift affecting the central Negev, elevating the Ramon block. The northern Negev fold belt was high during the middle Oligocene, as can be seen from the pre-Hazeva breaching of the Hatira erosion cirque, subsequently subsiding, as shown by the silting up of the cirque with Hazeva Formation sediments. This stands in contrast to the Ramon structure, which was breached only later during the Miocene (Baer 1981, Zilberman 1992, Plakht 1999).

A conspicuous watershed existed during the late Oligocene–early Miocene, following the middle Oligocene heritage, separating drainage systems which led to the Mediterranean, which were later filled up by the Hazeva–Dana suite, and those which led to the Persian Gulf, occupied by the Herod–Barada. This watershed ran along the central areas of Israel and Lebanon, down to the southern end of the Hebron anticline. It crossed the present Jordan Valley north of the Dead Sea, most probably in the region now separating the Ajlun and Zarqa anticlines from the Hebron and Hatira elevated structures to the west, heading east of the Valley.

Another watershed is to the south, separating the Hazeva–Dana from the Raham system (Garfunkel et al. 1974), the latter flowing in a general southward direction toward the Red Sea (Horowitz 1979, p. 70). The watershed is situated at Gav Ha'Arava, which serves as such for most of the Jordan Valley's history, save possibly for the Pliocene (see below). The southward channeling certainly began during (or before) Ma times, as these channels are filled up with the Raham Formation, a time equivalent of Palynozone Mb.

Magmatic activity in Ma times was quite restricted, manifested only close to the southern Jordan Valley by several minor intrusions in the Karak graben east of the Dead Sea, and a plug in Nahal Ashosh, some 60 km south of the Dead Sea on the western rim of the Arava.

7.2.3 Palynozone Mb, middle Miocene

Both the Hazeva–Dana and Herod–Barada systems were deposited in a series of channels, in which alternations of clastic fluviatile sediments and fine grained, occasionally chemical, lacustrine and lagoonal deposits are quite frequent (Fig. 7.3). When superficially observing the outcrops of these formations one gets the false impression that their lithology chiefly comprises sandy red-beds (see e.g. names given to the Hazeva by earlier investigators, in Chapter 5.1.2.2). However, a closer look reveals that lake and lagoon limestones and marls invariably make up considerable parts of both the Hazeva and Herod formations. Regarding the Hazeva (Sneh 1981), the Shahaq Member is almost exclusively conglomeratic, of fluviatile origin; the Mashaq contains only 5–10% sand; while the Gidron contains up to 50% sand. The Herod sequence in the Poriyya escarpment does not contain more than 20–25% sand (Shaliv 1991). Both Avnimelech (1937) and Cailleux (1949) concluded a lacustrine or fluvio-lacustrine environment of deposition for the sands of the Herod Formation east of Lake Kinneret. The silts and limestone horizons also seem to indicate a lacustrine environment, a conclusion shared by all investigators of this Formation. Certainly, red-beds make up only part of the sequences. Both the sandy appearance and the red coloration of entire outcrops seem to result from the form weathering, which causes smearing of the exposures by down-washed sands and red clays, giving them their typical looks.

The sequences are better preserved in lower structures, such as synclines or younger grabens, on both sides of the Jordan Valley, where subsequent erosion hit them to a lesser degree. In most cases, the Hazeva and Herod sediments, even if they were deposited on the higher structures as suggested by Zilberman (1992), were subsequently washed away. Typically, in southern Transjordan east of the Arava and the Dead Sea, almost all outcrops of the Dana Formation are restricted to the Jordan Rift, with only subordinate occurrences in smaller rifts on the Transjordanian Plateau (Bender 1974b, Sneh et al. 1998a). A similar situation exists for the Herod–Barada suite in northeastern Israel and southwestern Syria. A conspicuous rift valley in which correlative sediments were preserved lies further east, at Wadi Sirhan in eastern Transjordan.

Notably, when Hazeva sediments do occur on earlier higher structures of the Syrian Arc, such as the southern part of the Hebron anticlinorium or some of the lower northern Negev anticlines, their sequences are very much compressed, with

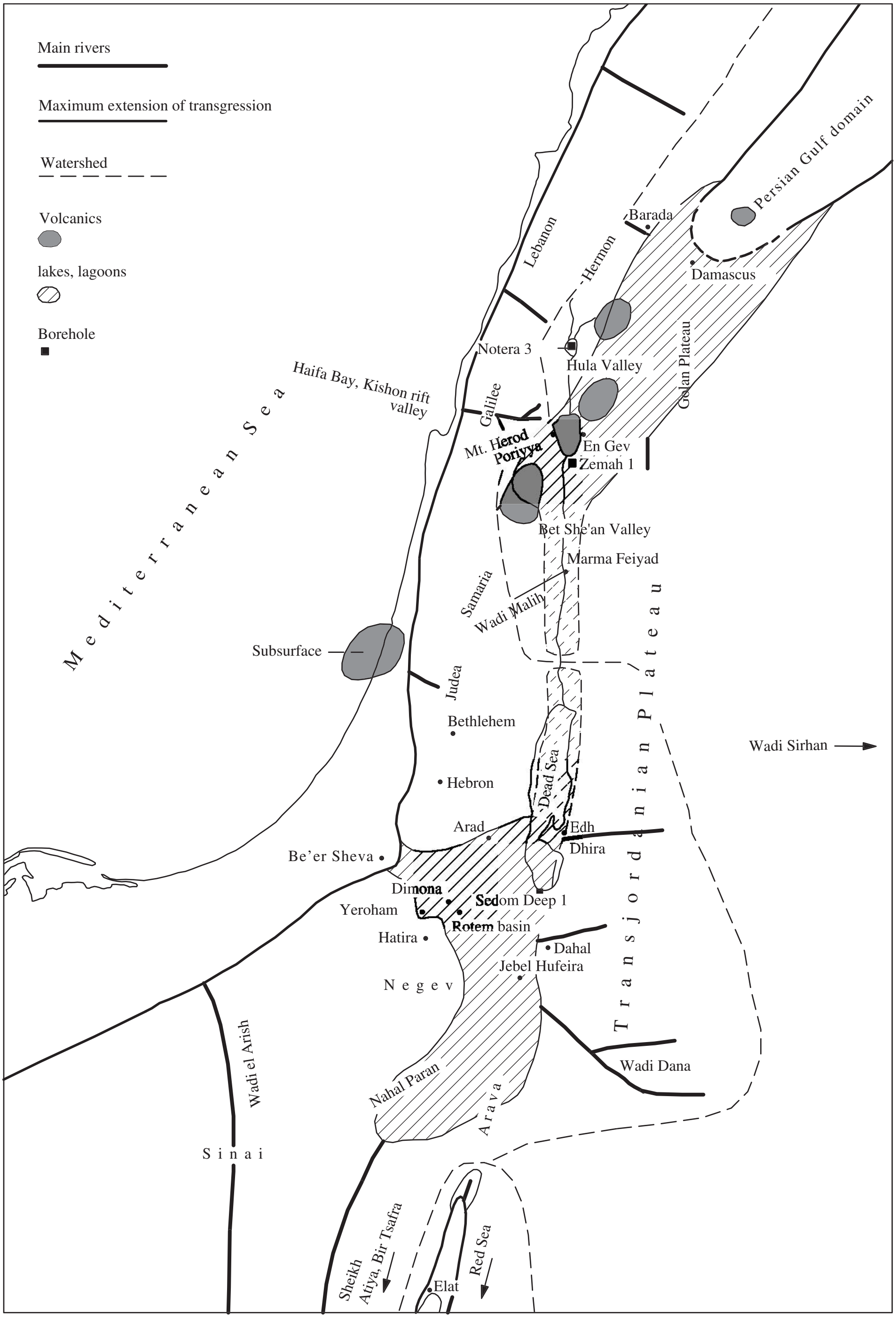


Figure 7.3. Palynozone Mb paleogeography, middle Miocene.

angular and erosional unconformities within the section (Braun 1963, Aharoni & Aizin 1966, Garfunkel & Horowitz 1966, Horowitz 1974). This indicates structural control on the sedimentation, with some uplift during the Miocene. This uplift was only moderate, as can be seen from the composition of pebbles in the Hazeva fluviatile sediments. The base conglomerates are dominated by late Cretaceous pebbles. Further incision exposed the early Cretaceous, which supplied sands. Eocene pebbles prevail in Hazeva top conglomerates, because the middle Miocene peneplain arrived at almost the same altitude as the Oligocene one, thus silting up the older sources of sands, eroding only the then abundant Paran Formation, of early middle Eocene age, which made the uppermost surfaces exposed at that time. Such flattening of the landscape could only have occurred if uplift was minimal.

The sequences preserved on the rims of the Jordan Rift Valley are no different from the ones outside its area, indicating that the Eritrean faulting postdates the deposition, and that the exposures were protected from subsequent erosion only because of their low elevations. On the other hand, Mb sequences in basins along the Jordan Valley are considerably thicker than in the Syrian Arc synclines, indicating accelerated subsidence (Horowitz 1987c), a continuation of the basins' behavior during the Oligocene.

The Hazeva–Dana sediments are widespread in the Arava sector of southern Transjordan and in large parts of the Negev, deposited in fluvio-lacustrine systems which led to the Mediterranean. The directions of flow of the main channels and the distribution of intermediate basins where better developed sequences accumulated, as well as the provenances of the clastics, were discussed in detail in Garfunkel & Horowitz (1966). Some modifications were later suggested (Zilberman 1992, Ginat 1997, Avni 1998, Sneh 1999), but the general picture does not differ much. The southern extension of the Hazeva system goes quite far, some 100 km south of Elat in eastern Sinai (Garfunkel 1970), known from Sheikh Atiya and Bir Tsafra, 30–40 km west of the Gulf of Aqaba. Most of the Hazeva system was drained to the Mediterranean through the northern Negev and Be'er Sheva area (Neev 1960, Garfunkel & Horowitz 1966, Horowitz 1974), following channels cut during the late Oligocene–early Miocene regression and uplift (see above). Northern Sinai, west of the Negev, was drained through Wadi el Arish, while other Miocene drainage systems are known further west in northern Africa.

The northern limit of the Hazeva system in the western highlands is set by the Judea–Hebron anticlinorium, which was subject mainly to erosion at this time. Several questionable, thin gravel occurrences on the anticlinorium, such as near Bethlehem (Horowitz 1974), may have been deposited by contemporaneous, probably subordinate rivers, but no positive proof of age is available for that. Similarly, their limit on the eastern highlands is approximately at the same parallel. Within the Jordan Valley, the extent of the Hazeva or its correlatives quite far northward, to several tens of kilometers north of the Dead Sea, depends only on whether such correlations are accepted, which is still highly conjectural. This is so since almost any reddish clastic unit was called "Hazeva" or "Dana" at one time or another, without much data to substantiate such assignments. Although most of the Hazeva–Herod channels and intermediate basins are better developed in synclines, several of the channels had cut through anticlines, forming superimposed valleys. The most prominent of these cut the northern Negev and northern Sinai Syrian Arc anticlines (Neev 1960, Garfunkel & Horowitz 1966, Horowitz 1974, 1979, p. 73).

The Hazeva–Dana–Ghor el Qatar(?) system, draining central and southern Transjordan, almost the entire Negev and parts of Sinai, supplied sediments in the order of a kilometer and more of thickness to its marine counterparts (Derin & Reiss 1973). In contrast, the central anticlinorial sector of Israel, from the Hebron anticline northward to the Lebanon, was drained to the Mediterranean only by several subordinate rivers during the Miocene, the courses of which were outlined by Gvirtzman (1970), Issar & Kafri (1972), Kafri & Heimann (1994) and Kafri (1997). These were quite short rivers, draining a rather low-lying hinterland, even more so to the north, which is also expressed in the thicknesses of marine sediments, decreasing in that direction to mere tens of meters in the western Galilee coastal plain. The sediments deposited by this river system in the Galilee are termed the Hordos Conglomerate, also subordinate in thickness.

Incidentally, there is no evidence for a marine transgression through the area of the present Haifa Bay or the Yizre'el Valley eastward, nor for any large channel draining these regions westward during early or middle Miocene times. This indicates that in these periods the Yizre'el Valley was still a part of the elevated anticlinorial belt separating the Herod drainage from the Mediterranean. Notably, Pliocene sediments in the western Yizre'el Valley (Gerry 1965) directly overlie early Cretaceous formations in boreholes, with no Miocene beds present at all. A similar situation exists with the Herod–Lower Basalt complex. While to the east both are present and well developed, only basalt is known from most Yizre'el Valley boreholes, overlying various Cretaceous beds, usually the older ones, thinning as one goes west (Shaliv 1991).

The Kishon Rift Valley, which is the western continuation of the Yizre'el Valley approaching Haifa Bay, was thus formed in late Miocene times, as shown by the lack of Miocene sediments, while Pliocene deposits overlie older rocks (Horowitz 1974). In other elevated regions, such as Wadi Malih west of Marma Feiyad, the Herod Formation rests on early Cretaceous or even Jurassic rocks (Schulman & Rosenthal 1968, Mimran et al. 1989). The conspicuous watershed concluded for the Oligocene and early Miocene, dividing the Jordan Valley into two distinct northern and southern provinces, persisted also through the middle Miocene, separating drainage systems which led to the Mediterranean on the one hand, mainly the Hazeva–Dana–Ghor el Qatar(?), and to the Persian Gulf, the Herod-Barada, on the other.

The valleys filled up by Herod sediments and Lower Basalt volcanics occupy a large, northeast plunging synclinorial structure, leading from the eastern Lower

Galilee through the region of Damascus, connecting to the Persian Gulf, partly buried under younger volcanics of the Golan Plateau and El Shamah–Jebel Druze field (Fig. 5.3.1). This synclinorial basin, a part of the Syrian Arc fold belt, already existed in Senonian and Eocene times (Flexer 1968), controlling facies and deposition of marine formations during this period, as well as during the continental and partly marine Oligocene. This Syrian Arc mega-structure was somewhat accentuated by differential uplift and subsidence throughout the Miocene, and possibly also somewhat later. This can be seen from angular unconformities within the Miocene formations, controlling facies and deposition of the Herod River system, as seen in the En Gev and Marma Feiyad regions (Schulman & Rosenthal 1968, Michelson 1972).

To the west and southwest, in the areas of the central Yizre'el Valley and on the eastern flanks of the Samaria anticlinorium, the pre-Miocene erosion had cut down to the early Cretaceous and Jurassic rocks, which served as the main source for the considerable amounts of sand in the Herod Formation. Some sand could also have arrived from the Hermon anticline through the Hula Valley (see below). The Samaria, in contrast to the Yizre'el Valley, remained an elevated structure until the present day. The channels filled up by the Dana–Hazeva–Ghor el Qatar(?) sediments cut deeply into the elevated structures of Transjordan, exposing Paleozoic–early Mesozoic sandstone formations known in general as "Nubian Sandstone", which were the source of sand in these formations. At least one of the more elevated anticlines of the northern Negev, the Hatira, had also been quite deeply cut by pre-Hazeva erosion, exposing early Cretaceous or Jurassic sandstone beds in the central parts of its erosion cirque (Garfunkel & Horowitz 1966, Plakht 1999), which also supplied some of the sand for the Hazeva, although in lesser quantities.

The Herod Formation fills a previously cut channel system, mainly occupying synclines of the Syrian Arc, which led from northeastern Israel and northwestern Transjordan through the Damascus region (Picard 1943, p. 79) to the extended Miocene Persian Gulf neritic basin. In contrast with the Oligocene, however, this basin was connected to the Mediterranean during much of the Miocene period (Krasheninnikov 1994), through northern Syria.

The northernmost occurrence of the Herod known in the Jordan Valley is from the Notera 3 borehole in the Hula Valley, where 270 m of Palynozone Mb were penetrated without reaching its base (Horowitz & Horowitz 1985). A questionable outcrop, assigned by Saltzman (1964) to the Herod but by Shaliv (1991) to the Umm Sabune Conglomerate, at the northwestern corner of Lake Kinneret, contains Jurassic pebbles. These two occurrences, if indeed correlative, may indicate a Herod channel leading southward from the Hermon or Lebanon ranges, which are the only localities where the Jurassic is exposed, to the main En Gev channel.

Magmatic activity during Mb times was considerable, manifested in volcanic extrusions which affected northeastern Israel and southwestern Syria, resulting in substantial volumes of the Lower Basalt (Shaliv 1991). The southernmost occurrence is south of Marma Feiyad(?), and the northernmost in the Jordan Valley is from the Notera 3 borehole, in the Hula basin. The Lower Basalt or its correlatives occur in northeastern Transjordan and southwestern Syria, further north. These volcanics appear either as massive accumulations, or as flows interfingering with the Herod–Barada sediments. Several contemporary intrusions are known from boreholes and outcrops in the central Jordan and Yizre'el valleys, summarized by Shaliv, occurring as dikes, stocks and pipes, as well as volcanoes. No volcanics at all are known for that period from the southern parts of the Jordan Valley or its surroundings.

The Hazeva and Herod suites represent a typical deposition cycle caused by transgression, seen in all localities beside some of those later affected by the Eritrean faulting. The process started with base conglomerates, continuing with gradually finer clastics until when the sea was high enough it caused stagnation of the drainage systems, where lagoonal and freshwater sediments were accumulated, in lower regions on an otherwise well-developed peneplain. Seaward, these are replaced by estuaries depositing rocks such as oyster banks, turning further away to coral and algal reefs, which grade into deeper marine marls, both in the Mediterranean and Persian Gulf domains. The late Serravallian regression, which later continued into the Tortonian except for a minor rise within the latter period, achieving its climax in Messinian times, inverted this process, causing deposition of gradually coarser clastics, terminating with top conglomerates, usually developed as sheetwash gravel bodies over extensive areas. The ages of the top conglomerates vary in different localities, spanning the entire interval mentioned above.

This cycle was interrupted at least once by the regression phase known from the eastern Mediterranean and the Gulf of Suez during zones N10–N12 (Table 1.4.1). This drop in sea level, which was not recognized for the Persian Gulf domain, caused a conspicuous unconformity within the Dana–Hazeva sequences (Calvo et al. 1997, 1998).

The maximum level of the middle Miocene sea is securely known for the Hazeva system to be in the Yeroham–Dimona region, some 30–40 km west of the Dead Sea (Harash 1967), possibly extending to Arad, 15–20 km further east (Agnon 1983); neither marine sediments nor evaporites are known for Palynozone Mb at the Sedom Deep 1 borehole, which penetrated the deepest part of the southern Dead Sea basin. No clear marine intercalation is known for the Herod Formation exposures. However, the geochemistry of evaporites from the Zemah 1 borehole, at a depth of some 3,800 m, 300 m below the top of Palynozone Mb, indicates (Raab 1998) deposition from a mixture of landlocked seawater and groundwater flushing basalts or gabbro, which may hint at some marine connections.

7.2.4 Palynozone Mc, late Miocene

The late Miocene is a period of major changes in paleogeography in the Jordan Valley region, caused by a combination of extensive global decline in sea level and the Eritrean faulting (Fig. 7.4). For the first time in its history, parts of the area now occupied by the Jordan Rift became independent, endoreic drainage base levels drawing channels from all around, although certainly not in its present shape. Two rather extensive troughs are known, one from the central Jordan Valley, extending north over a shallow sill, into the southern part of Lake Kinneret, and possibly even further north (Ben-Avraham et al. 1990b, 1996, Reznikov et al. 1999), and another from the northern Arava and southern Dead Sea region (Horowitz 1987c, Frieslander et al. 1997). The first subsided some 1,700 m during the Mc period at the place where the Zemah 1 borehole was drilled (Marcus & Slager 1985), and the second subsided some 2,000 m, as encountered in the Arava 1 borehole (Horowitz 1987c). Some 500 m of Mc are known from the Hula (Kashai & Goldberg 1984, Horowitz & Horowitz 1985), but most of this sequence predominantly comprises basalt. It is not clear whether this basin was an independent small trough, or only part of a canyon leading south to the central Jordan Valley. The figures denote subsidence, not sediment accumulation, since rocks in both boreholes frequently display dips, so that calculation of the true thicknesses is quite difficult.

The combination of troughs, which commenced subsidence somewhat after the beginning of Palynozone Mc, and regressing seas throughout most of its duration, caused a division of the southern Levant into three distinct drainage domains. One encompasses the troughs, which accumulated sequences of great thicknesses, the second is made up of areas far away from the troughs both to the east and west, in which processes affected by the retreating seas caused deposition of top conglomerates of both the Hazeva and Herod formations, and the third comprised the areas separating these two, regions which were eroded either toward the troughs or seaward. During the advent of Mc, the boundaries between these three domains changed continuously, the erosive zones becoming gradually larger.

It is quite difficult to define exactly the limits of the late Miocene troughs because they are known mainly from boreholes, where considerable Mc sequences have been encountered. Detailed geophysical data are not always available, and at any rate results are occasionally questionable, especially within the narrower parts of the Rift Valley. An example is from Zemah 1, whose drilling was based on high resolution seismic reflection profiles (Fig. 8.4.5) indicating a buried anticline (Rotstein et al. 1992), but which penetrated a great thickness of Neogene–Quaternary fill accumulated in a trough. Magnetic measurements had shown (Ginzburg & Ben-Avraham 1986, and see also Chapter 8.4) that the anticline shape on the seismic reflection profiles resulted from deep seated magmatic intrusions.

Outcrops of considerable volumes of Mc sediments in the Rift Valley are limited to a single occurrence, in Jebel Hufeira almost 50 km south of the Dead Sea, where the thickness conforms to the order penetrated by boreholes. Other outcrops, of limited thickness and still extremely conjectural, may be near En Boqeq on the western shore of the Dead Sea, or the ones attributed by Bender to the Ghor el Qatar Series (see Section 5.1.2.6). Another questionable outcrop, seemingly of

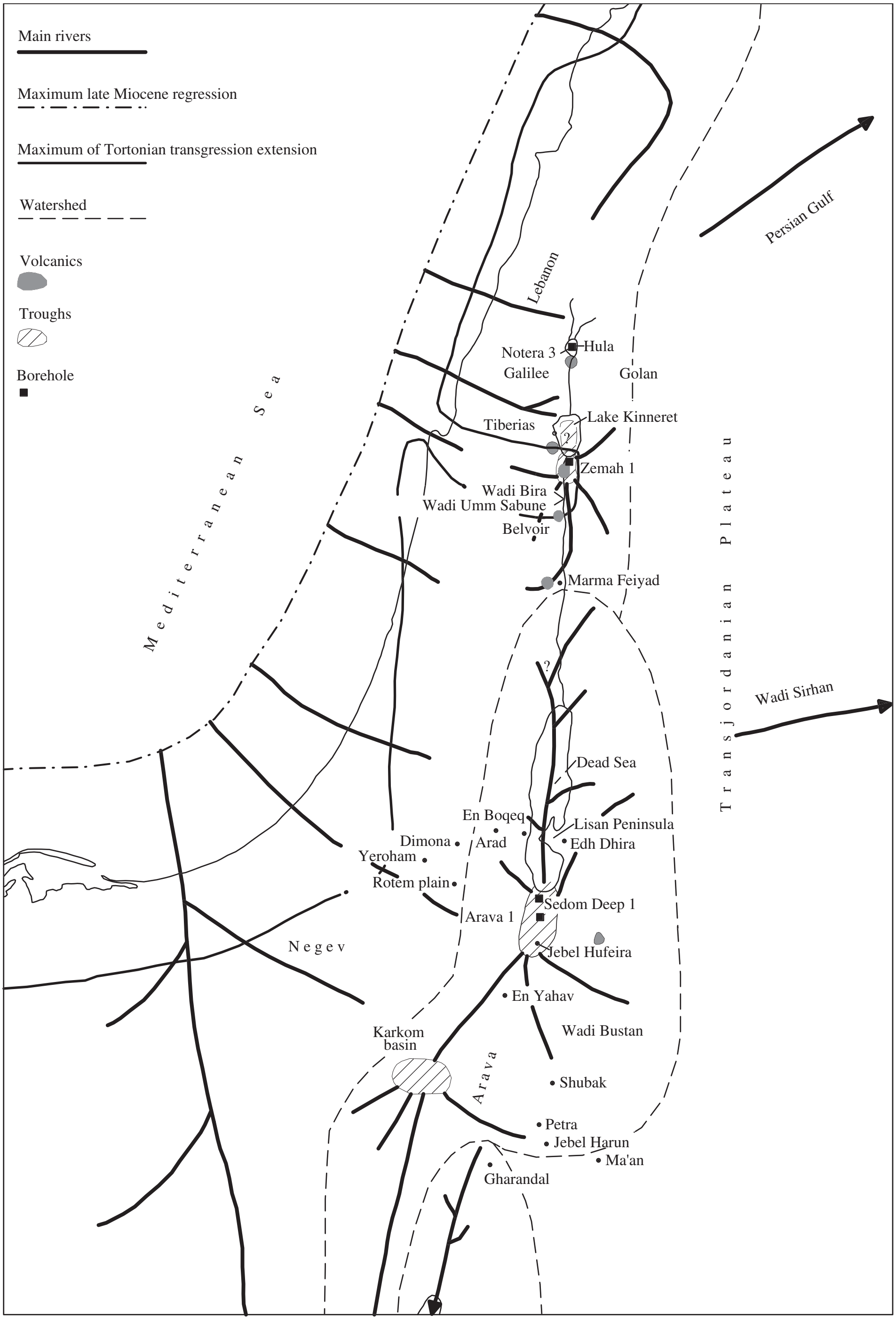


Figure 7.4. Palynozone Mc paleogeography, late Miocene.

a similar nature, is reported from outside the Jordan Valley, in the Karkom basin of the central Negev (Calvo et al. 1998).

In regions where the retreating seas were in control, the late Miocene regression inverted the depositional processes of Mb times. This caused coverage by gradually coarser clastics, terminating with top conglomerates, usually developed as sheetwash gravel bodies over extensive areas. A further drop in sea level caused erosion, hitting the gravel beds first and cutting the lower section later. These regions are quite far away from the Jordan Valley, and thus are not discussed here in further detail.

The areas surrounding the Jordan Valley, in particular those closer to the troughs, had been subject to severe erosion leading toward the latter. It is impossible to delineate exactly any channel systems for this phase, mainly because subsequent faulting and erosion wiped out most of the evidence. It is clear however that considerable volumes of the upper parts of both the Herod and Hazeva formations are missing in areas close to the Rift Valley, to be found as a fill in the troughs. Younger sediments overlying the Herod and Hazeva are always deposited over clear structural and erosive unconformities.

It is not clear at what exact stages the troughs acted as endoreic basins, as intermediate, or how far the Tortonian sea transgressed inland. The Mc sequences penetrated in the boreholes Zemah 1, drilled in the central Jordan Valley (Marcus & Slager 1985), and Sedom Deep 1 at the southern Dead Sea (Baker 1994) both contain rocksalt and evaporites. The geochemistry of the lower Mc evaporites in Zemah, at 2,567 m, indicates (Raab 1998) deposition from seawater trapped in a continental basin, while the upper Mc evaporites, at some 2,000 m, were laid down directly from seawater. This would mean that two late Miocene shallow transgressions known for the Tortonian (Fig. 9.1.1) may have reached the central Jordan Valley, each for a short while.

The earlier one may also have deposited the oyster bed occurring in the upper part of the Herod Formation (Sneh 1993), while the later could be connected with sedimentation of the basal part of the Bira Formation, outside the trough limits. There is no clear indication concerning the direction this sea could have come from, whether from the east, as concluded for Mb, or from the west, as is quite clear for the overlying Pliocene sediments. If indeed this sea deposited the lowermost part of the Bira, it stands to reason that it was connected to the Mediterranean to the west.

Unfortunately no geochemical data are available for the Sedom Deep 1 borehole, the only locality in the southern Dead Sea basin where late Miocene evaporites were encountered. In contrast, Mc at Notera 3, in the Hula basin, does not contain evaporites at all, only subordinate dolomite in very thin horizons, which may support the assumption above, that this sequence filled up a canyon leading to a deeper trough, in the Kinneret and central Jordan Valley area.

Magmatic activity is subdued in Mc times as compared with the previous Mb. Single flows are known to interfinger within the Umm Sabune Conglomerate and its correlatives, and are also noted from both the Notera 3 and Zemah 1 boreholes, in the Hula and central Jordan Valley. As during Mb, no magmatic activity is known south of the central Jordan Valley.

7.2.5 Palynozones Pa and Pb, Pliocene

Three prominent depocenters are known for the Pliocene formations of the Jordan Rift Valley (Figs 7.5.1 and 7.5.2): the Beqa'a–Hula to the north; the central Jordan-Yizre'el valleys; and the Dead Sea, particularly its southern sector. Deposition of the Pliocene sequences is dictated by two cardinal processes inherited from the previous late Miocene paleogeography: peneplanation and channeling of most of the southern Levant, and outcomes of the Eritrean faulting, subsequently flooded by the Pliocene sea. There seems to be general agreement concerning the connections of the central Jordan Valley to the Mediterranean through the Yizre'el Valley, based on the close to perfect continuity of outcrops along this path. Thus the twin transgressions of the Pliocene first deposited the chiefly lagoonal Bira Formation during the Tabianian (or Zanclean, Zanclian) high sea level, which reached the central Jordan Valley, while the Piacenzian, when the sea level did not rise as far as before, caused sedimentation in the lakes of the Gesher. As mentioned above, it is possible that the Tabianian transgression was not the first to lay down Bira-like lagoonal sediments, as the lowermost part of the sequence may have been deposited already in Palynozone Mc times by the rising late Tortonian sea.

No connections with the Indian Ocean, as suggested for the Oligocene and Miocene, occur in the Pliocene (however nothing conclusive can be said about the late Miocene). It seems that some uplift of the Transjordanian Plateau, as part of the Eritrean movements, had completely closed this option. This is seen by occurrences of Bira–Gesher-type sediments some 40 km east of the Jordan Valley, the Jabal Bakiya Formation (Baubron et al. 1985), in the channel of Wadi Zarqa which was earlier cut into the uplifting terrain. The ensuing paleogeography of Pliocene times was of a rather flat country, dotted with lagoons and lakes, fed by wide, meandering rivers.

The two transgressions are separated by a period of lower sea level, which led to the deposition of evaporites in the central Jordan Valley, resulting also in the Gesher Formation overlying the Bira paraconformably, over a gentle erosional relief, in the same vicinity. Marine influence, even during the Tabianian, did not reach too far eastward, so that on the Golan and Transjordanian highlands only freshwater lakes existed, depositing Gesher-type sediments all along the Pliocene. On the other hand westward, in the Yizre'el Valley, closer to the sea, most (if not all) of the sequence is taken up by lagoonal, Bira-type sediments, occasionally grading into truly shallow marine deposits such as oyster beds. At the same time, magmatic activity caused effusions of the Intermediate Basalt, interfingering with the Bira–Gesher suite or, in certain regions, replacing altogether the sedimentary cover with basalt flows.

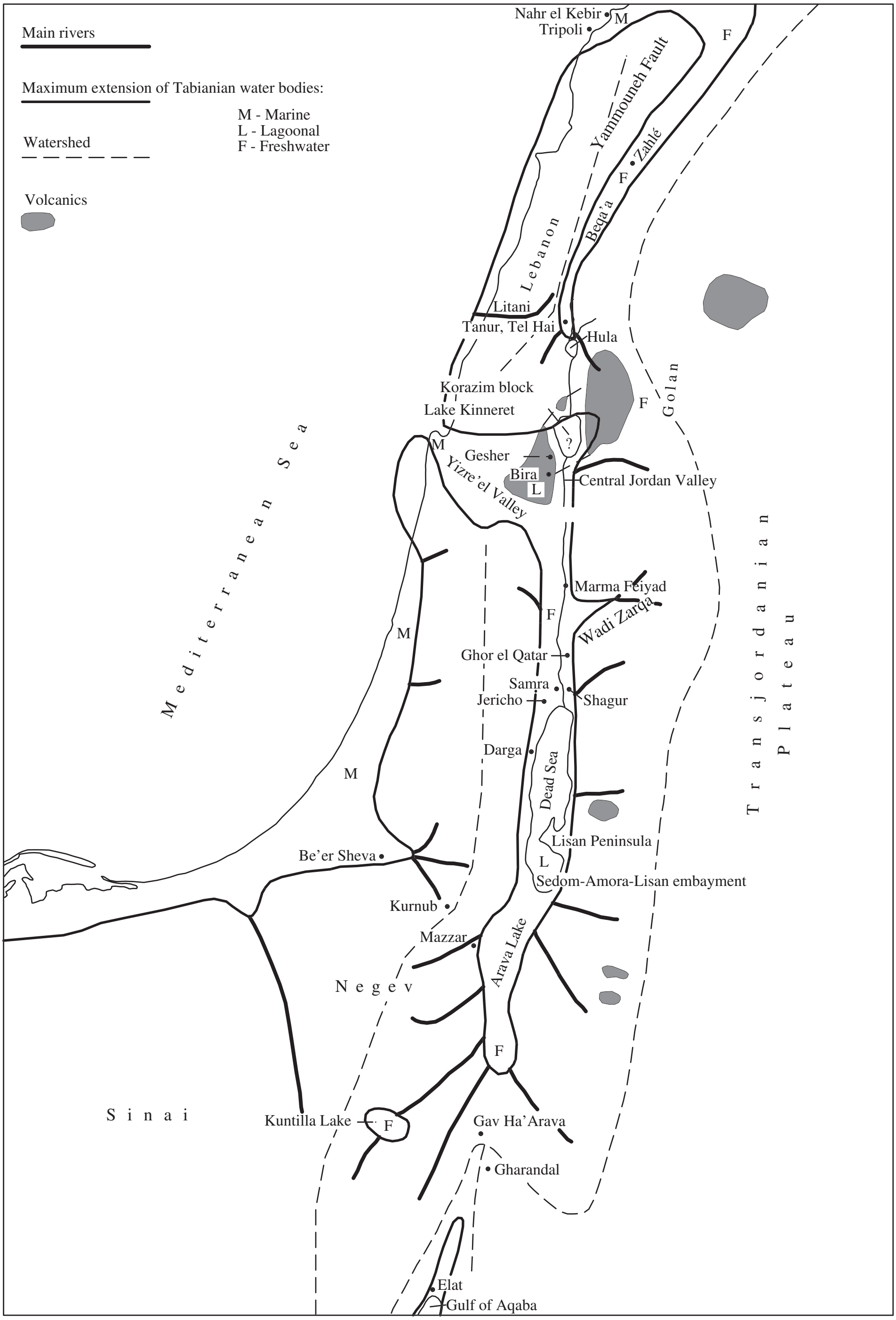


Figure 7.5.1. Palynozone Pa paleogeography, early Pliocene.

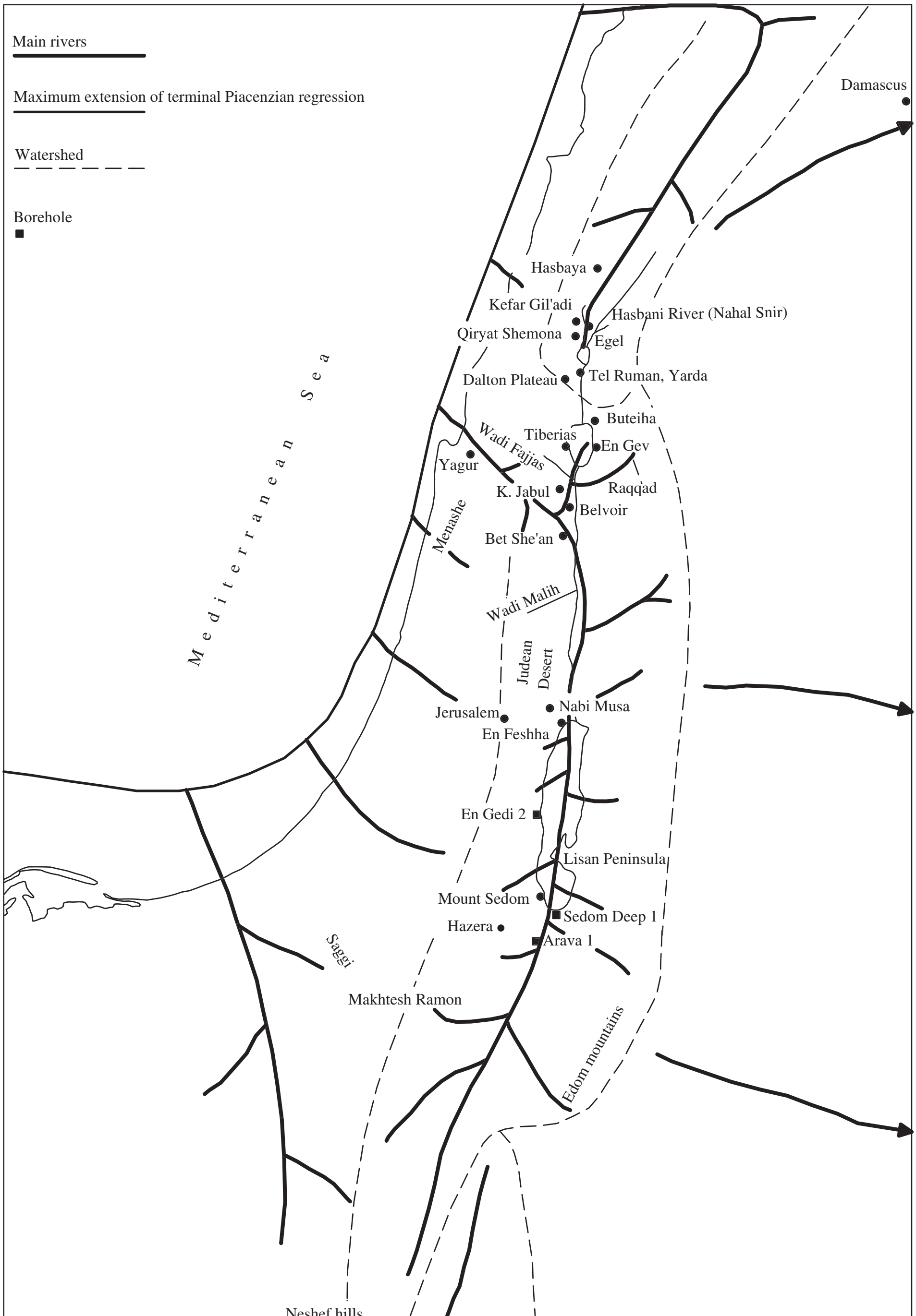


Figure 7.5.2. Palynozone Pb paleogeography, late Pliocene.

The paleogeography of the two interconnected central Jordan–Yizre'el valleys, within the triangle bounded by Marma Feiyad to the south, Lake Kinneret to the north and the Mediterranean Sea westward, including the western sector of the Golan Heights, seems quite clear from the ample available evidence. The connections of this embayment, both to the south and north, are however much more questionable. The generally accepted view (Horowitz 1979, p. 79) called for interconnection of the three Pliocene depocenters, broadly through the area of the present Jordan Valley, with the seemingly sole outlet to the Mediterranean through the Yizre'el Valley.

According to this reconstruction, the Pliocene sea twice invaded the Dead Sea region, depositing the Oolitic, Shagur and lower part of Samra north of the Dead Sea, while the Mazzar and Sedom formations were laid down to the south, as well as the Kuntilla lacustrine member of the Arava Formation, during the earlier Tabianian transgression. Evaporites of marine affinity in the Amora were thought to have been laid down by the Piacenzian sea (Zak 1967). This is questionable, since no trace of this sea is known to the north, from where it should have arrived, thus these evaporites could have been redeposited from the Sedom Formation. A river system leading to the Sedom–Amora embayment from a general southern direction was filled up with Arava Formation clastics in times of higher sea levels. A parallel drainage system was suggested to the north of the central Jordan Valley, where lakes developed in the Hula–Beqa'a region, depositing the Tel Hai and Zahlé limestones, while rivers leading to these were filled up with the Tanur and other correlative conglomerates.

The principal objections raised against the above seemingly elegant reconstruction are concerned with the discontinuity of outcrops in areas separating the central Jordan Valley from its southern and northern time-correlative units. There is no single unquestionable occurrence of Pliocene sediments along more than 40 km between Marma Feiyad and somewhat north of the Dead Sea to the south; neither are there such sediments along almost 40 km from the northern tip of Lake Kinneret northward (Sneh et al. 1998a). A reservation is in place for the region north of Lake Kinneret: the gravel beds lining the hill slopes along the entire distance from the northern Hula Valley down to Lake Kinneret were correlated by Horowitz (1973) with the Tel Hai–Tanur suite. However, due to the lack of fossils or clear contacts, such a correlation remains only a possible conjecture, no more. Besides the geographic discontinuity, the rocks themselves are quite different in each depocenter.

A reconstruction of the northern extension of the Bira–Gesher basin depends on understanding the role played by the Korazim block during the Pliocene. Heimann (1990) states that the Korazim block was elevated throughout Neogene times, and so may have blocked connection of the Hula and Kinneret. This argument should be regarded with caution, since the Korazim is an elevated block even today, when the Jordan River does connect both lakes. Fleischer (1968) clearly showed, based on analysis of numerous drillings across the Korazim block, that hardly any Neogene–Quaternary sediments are preserved there below the Cover or the Ruman basalts, topping the block to the south and north respectively (Heimann & Ron 1993). Saltzman et al. (1981) view the block as an extension of the Syrian Arc folds, not as a faulted structure. Contrary to the latter view, Heimann & Ron (1993) regard the Korazim as an elevated pressure ridge, whose segments rotated $11.4° \pm 4.1°$ counterclockwise at some time after the late Pliocene or early Quaternary.

If indeed the Korazim block acted as a watershed during the Pliocene, another outlet of the Tel Hai–Tanur–Zahlé system to the Mediterranean should be considered. The two plausible possibilities are either the Litani River Valley in southern Lebanon, which presently connects the Beqa'a Valley to the Mediterranean, or Nahr el Kebir (the big river) Valley, north of Tripoli (not to be confused with another river of the same name, near Latakia, further north). The latter (Figs 10.2.2 and 10.2.3) has several peculiar characteristics: it is very wide and shallow, drained by a few relatively short rivers. To the west it is covered by marine Pliocene sediments, while the Beqa'a side (which is not drained by those rivers) is paved by the Zahlé Beds. The area between the Mediterranean coast and the Beqa'a is a relatively low graben, not exceeding several hundred meters in elevation, separating the considerably higher north–south oriented ranges of the Lebanon and the Ansariyeh mountains.

This lowland is entirely topped by "Pliocene–Quaternary" basalts (Dubertret 1962), later designated "Homs Basalt", which also cover, unfaulted, the Yammouneh Fault. Radiometric datings of these basalts yielded ages in the vicinity of five million years (Butler et al. 1997) which, together with the taphrogenic and sedimentological parameters, recall strongly the almost parallel situation of the Yizre'el Valley, whose central sector was also uplifted only after the Bira deposition.

These make the Nahr el Kebir River Valley the best candidate for an outlet of the Hula–Beqa'a system to the Mediterranean Sea, in a manner similar to the outline seen in the Yizre'el Valley, connecting the sea with the central Jordan Valley. The proposed scenario is that the Nahr el Kebir Valley is yet another one of the Eritrean system rifts, which joined the Beqa'a faulted synclinorium westward to the Mediterranean during the Pliocene. Marine sediments deposited by the rising sea to the west grade to lacustrine deposits eastward, the sequence interfingered by volcanics at the same time. The rift was later slightly uplifted and arched, during the Quaternary, again in the same manner observed for the Yizre'el Valley (Horowitz 1979, p. 53). The fault running from Tripoli to the east–northeast, bounding the southern end of the Nahr el Kebir Valley (Dubertret 1962), was assigned to the Eritrean system by Horowitz (op. cit.). Recent studies showed (Butler et al. 1997) that the Yammouneh Fault is also a part of the Eritrean system, inactive during the last five million years.

If this assumption is accepted, the Hula was lying at the upper reaches of the Pliocene Zahlé system, drained northward. Several wide, rather shallow hanging valleys are found around and north of the northern sector of the Hula Valley

(U. Kafri, Geological Survey of Israel 1998, pers. comm.), one of which is partly filled up with the Egel Conglomerate (Heimann 1990), equivalent to the Tanur (see Section 5.2.2.4). No clear indications concerning flow directions could be found in this preliminary state of research, but a general northward direction seems favored.

Another possibility is that the Pliocene Hula was drained to the north, and the Beqa'a to the south, both connecting to the Mediterranean through the Litani River Valley. It seems however from the Litani's narrow, superimposed gorge and its sequence of Quaternary terraces, that this river captured the Beqa'a drainage only some time later, in the earliest Quaternary, following the blockage of El Kebir by the Homs Basalt, followed by uplift of the Lebanese mountains, similar to the order of events in Israel (Horowitz 1979, p. 328). Considering both possibilities, the El Kebir option seems the more promising one. However, until more details are available (which is quite difficult these days, considering the political situation), no definite answer can be suggested for the Pliocene drainage of the Hula; southward, northward and if so by which route. At the moment the El Kebir option seems to me the preferred one.

The other options for the Dead Sea Pliocene basin are that the sea could have penetrated either from the south, connecting the region to the Red Sea via the Arava and the Gulf of Aqaba, or from the west, joining the Mediterranean through the northern Negev and Be'er Sheva region. The first view may be supported by occurrences of several small, but quite distant and discontinuous exposures of Mazzar-type sediments, all along the Arava down to the Gulf of Aqaba (Bartov & Bein 1994). The second is backed up by the existence of marine-littoral Pliocene sediments in the Be'er Sheva basin, reaching at least 20 km east of the city (Goldbery 1980, Buchbinder & Zilberman 1997, Voznesensky & Shimshilashvili 1999), about 40 km from the Dead Sea, and possibly extending eastward down to Kurnub (Bentor 1960), a distance of only 35 km from Sedom. It should be noted that the last mentioned is questionable, and does not appear in later publications (Sneh et al. 1998a). Incidentally, some authors (Elron 1980) even raised the possibility that the Dead Sea region was connected both to the Mediterranean via the Yizre'el Valley, and to the Gulf of Aqaba through the Arava.

It seems easier to refute the Aqaba connection than to support the Yizre'el Valley one, by the obvious path of the Arava Formation river, which flowed northward through the northern Arava, bringing magmatic pebbles from the Elat region to the Dead Sea as first suggested by Garfunkel & Horowitz (1966), with additions and slight modifications by Zilberman (1992), Ginat (1997) and Avni (1998). The southern part of the Arava was drained, as before, southward to the proto-Gulf of Aqaba, depositing the Eilot Conglomerate (Horowitz 1979, p. 76).

The view that the Dead Sea domain was connected to the Mediterranean is also strengthened by occurrences of fossil *Mugil* and clupeid fish in the Sedom Formation, which are typical Atlantic–Mediterranean dwellers. In addition, both the Sedom and Amora sediments yielded sporomorphs that could have been brought to the basin only by the River Nile and the eastern Mediterranean longshore current (Horowitz & Zak 1968, Horowitz 1992a, p. 362). Several endemic animals inhabiting springs around the Dead Sea seem to be descendants of typical Tethyan or circum-Mediterranean creatures, that apparently could reach the region only during the Pliocene (Por 1963). In addition, it seems quite clear that the watershed at Gav Ha'Arava was already established in Oligocene times, continuously acting as such through the Miocene (Garfunkel & Horowitz 1966) and Pliocene (Ginat 1997), and apparently not changing its role since.

These, however, do not settle the Be'er Sheva or Yizre'el Valley dispute, since both ways directly connect with the same "necessary" sea and, admittedly, the two options are quite plausible. The Miocene watershed south of Marma Feiyad had to be breached in order to bring the sea from the north, but contrary to the one at Gav Ha'Arava, this did happen at some time, as is the situation today; so it might well have resulted from the Eritrean faulting, but no definite proof is at hand for that. Added evidence is that the Eritrean phase did indeed affect the Marma Feiyad region, since there the Umm Sabune Conglomerate overlies the Herod Formation, over a faulted terrain (Schulman & Rosenthal 1968, Shaliv 1991). Another clue is advanced by Begin (1975b), who analyzed the flow directions of conglomerates in the upper part of the Samra Formation north of Jericho, concluding they indicate a north–northeastward oriented drainage system.

For the western option it is necessary that the region separating the Dead Sea and Be'er Sheva would have been much lower, as indeed it was during the early and middle Miocene, when it was crossed by rivers coming from Transjordan. Since most of the uplift of the western hilly backbone took place only during the Quaternary (Horowitz 1979, p. 54), postdating the rifting and Pliocene sedimentation in the Yizre'el Valley, this possibility is not as farfetched as may be thought at first glance. It appears that, for the time being, the question should remain open, although the Yizre'el Valley option is at present preferred.

The Mediterranean apparently invaded the southern Dead Sea region only once, its maximum effect being at the beginning of the Tabianian, as is also known from other localities along the Jordan Valley and in the Be'er Sheva basin (Buchbinder & Zilberman 1997). A supposed later transgression, some time in the Piacenzian, is only indirectly recorded in the Amora salt, solely by its geochemical signature (Zak 1967). This view is opposed by Raab (1998), again based on geochemistry, who maintained that the Amora salt was deposited from inland waters showing no connection with the sea. The lack of any other evidence for Piacenzian marine sediments along the entire Jordan Valley, including its deepest basins, supports Raab's view.

The northern Dead Sea basin and the area north of it may have been affected by the Piacenzian transgression. These regions are thought to be underlain by evaporites (Neev & Hall 1979, Belitzky & Mimran 1996), which could have been deposited there during Pb times. It seems that they are not correlative with the main body of Sedom and Lisan evaporites of Pa age, since there is no connection between the two occurrences (Shulman & Ben-Avraham 1999). These questionable evaporites do not however crop out anywhere, nor were they ever penetrated by any drilling, so their supposed existence and age are far from certain. Also, as Raab indicated for the Amora, they could have been deposited from hypersaline inland water bodies.

The time of highest Tabianian sea level is marked in the northern Arava by the lagoonal Mazzar sediments, grading into lake deposits of the Arava Lake (Avni 1998) and other freshwater bodies, which may have extended over wide areas up to eastern Sinai, but probably just for a short while. Notably, the lacustrine sediments of Kuntilla Lake (Avni & Rosenfeld 1996, Avni 1998) contain *Hydrobia*, a mollusk never abundantly found in beds other than Pliocene in the southern Levant, whenever these are safely dated. It is quite difficult, due to the scarcity of outcrops and fossils, to define an exact boundary between lakes and lagoons or the extent of the lagoonal domain. Bender (1974a, p. 92) describes an isolated occurrence, some 10 km north of Gharandal in the Arava Valley, of some 40 m of shallow estuarine to shallow marine sediments, containing "dwarfed" foraminifera such as ?*Rotalia* and ?*Discorbis*. It is not clear whether this occurrence, located very close to Gav Ha'Arava, was deposited by the Mazzar–Sedom embayment, or may represent the northernmost extension of the Gulf of Aqaba at the time.

When the Tabianian sea regressed, the rivers followed its course, depositing the En Feshha Conglomerate and the upper, gravelly part of the Samra. At the same time, it seems that the regions south of the Dead Sea, where the Arava Formation was deposited, were already subject to partial erosion, helped by the drier climate of Pb times. This regression is probably responsible, due to the creation of rivers, for the inclusion of the typical magmatic components within the Amora Formation.

The central Jordan Valley shows a similar sequence of events. The Tabianian sea reached the region, depositing the Bira lagoonal sediments. These are fed from the east by an extensive freshwater supply, causing the eastern (and northern?) regions to deposit the freshwater Gesher sediments on the Golan and Jabal Bakiya on the already channeled Transjordan highlands. The regression caused a westward flow of the freshwater, depositing Gesher sediments over the Bira in the western central Jordan Valley, while rivers deposited gravel to the east. The rivers followed the retreating sea, depositing a top conglomerate, which in most places was removed by a further drop in sea level at the Pliocene–Quaternary boundary.

Again, a similar paleogeographic development is seen in the Beqa'a and Hula valleys, except that the sea did not invade as far as their counterparts to the south, since it appears that these regions never subsided below sea level. In addition, the prevalence of lacustrine sediments to the north is an outcome of the paleoclimate, since during the Pliocene the north was more humid than the south (see Section 6.5.2.3).

7.2.6 Palynozone QI, earliest Quaternary

The QI paleogeography of the Jordan Valley (Fig. 7.6) seems to follow the preceding Pliocene outline, in that the three basins, the Hula, central Jordan Valley

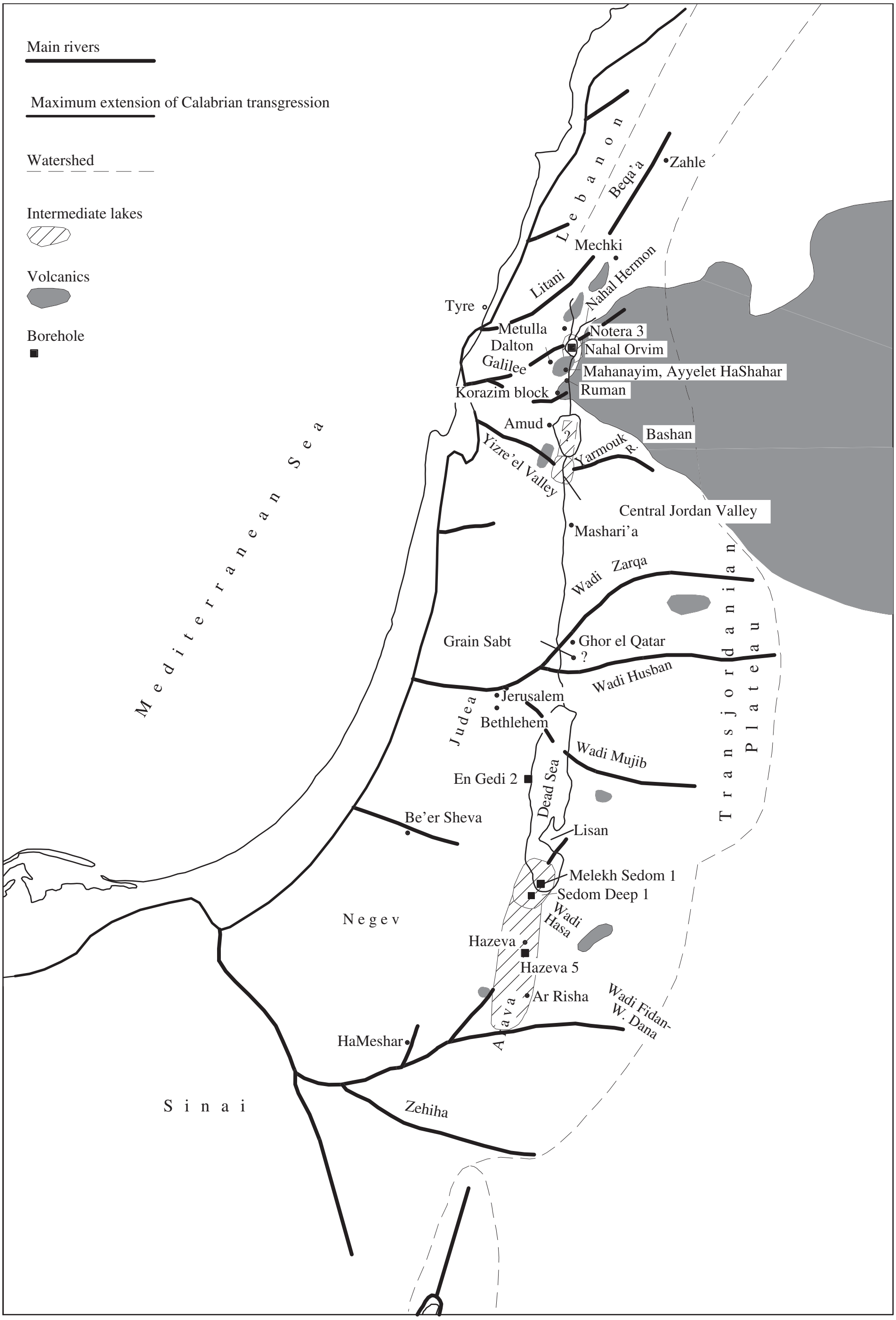


Figure 7.6. Palynozone QI paleogeography, earliest Quaternary.

and southern Dead Sea, accumulated considerable thicknesses of sediments due to continued subsidence. Subsidence in the Hula during QI times, lasting approximately one million years or somewhat less, is some 400 m (Horowitz & Horowitz 1985); it is almost 700 m in the central Jordan Valley (Levin & Horowitz 1987), while the maximum recorded for the southern Dead Sea, in the Sedom Deep 1 borehole (Horowitz 1996a), is 1,150 m. The En Gedi 2 borehole, situated at the southern boundary of the northern Dead Sea basin, penetrated some 400 m of QI sediments. These are subsidence figures, not true thicknesses which are difficult to calculate due to changing dips that cannot be accurately measured in the boreholes.

QI sediments to the north, in the central Jordan Valley and the Hula, are mainly marl and chalk, sometimes organic, with gravel horizons and basalt flows, the latter occupying a greater part of the sequence in the central Jordan Valley as compared with the Hula. To the south, in the Dead Sea region, sandstones prevail (Sa'ar 1986) with no volcanics, hence the name Melekh Sedom Sands. These sediments indicate, as do the pollen spectra (Horowitz 1992a, p. 336), a predominantly fluvio-lacustrine environment, while the lack of any evaporites suggests intermediate, freshwater lakes, not endoreic basins, on the route of rivers leading from Transjordan to the Mediterranean Sea.

A reconstruction of paleogeography outside Jordan Valley boundaries relies chiefly on confidence in the correlations proposed above (see Section 7.1.6), since there is no physical continuity, as only a few outcrops of QI are known from inside the Rift's margins. Three secluded, limited occurrences within the Rift, the HaMeshar Conglomerate and the Ar Risha Gravels Formation of the central Arava, the Melekh Sedom Sands south of the Dead Sea and the Mahanayim Marl in the Hula Valley, believed to belong to the same stage, are of no great help in reconstruction due to their discontinuity. The age of the Ghor el Qatar Series, thought to be part of this system (Horowitz 1979, p. 119), is not certain. West of the Jordan Valley, three formations remain as candidates for representing the drainage of the Rift basins to the Mediterranean, the Amud Conglomerate in the Galilee, the Bethlehem Conglomerate in Judea and the HaMeshar Formation of the southern Negev.

Patches of all three are exposed between the Dead Sea and the Mediterranean, at different elevations on both flanks and on top of the present-day watershed. Their occurrences hint at the existence of river systems crossing the Jordan Rift, flowing in a general westward direction to the Mediterranean, passing over the then rather flat country (Garfunkel & Horowitz 1966; Horowitz 1979, p. 115; 1992a, p. 363; Kafri 1997; Matmon et al. 1999). The flat landscape (Figs 7.7 and 9.6.2) assumption is corroborated by the types of vertebrates found near Bethlehem, now at the peak of the Judean Hills watershed, comprising long-legged animals living in a rather flat savanna landscape.

Of the three systems, reservations were only expressed concerning the one crossing the southern Negev, in which the HaMeshar Formation was deposited.

Figure 7.7. QI flatland, developed on the Judean Hills south of Bethlehem, incised due to subsequent uplift.

Avni (1998) claims that this unit, named by him "Zehiha", overlies the Arava Formation unconformably (as already noted by Garfunkel & Horowitz 1966), but flows into the Arava Valley, contrary to what was suggested by the latter authors and by Horowitz (1975a; 1979, p. 117; 1992a, p. 363). Garfunkel & Horowitz followed the HaMeshar almost continuously from the central Arava to the Mediterranean, through eastern Sinai, crossing over the presently elevated area of the central southern Negev.

Avni based his idea that the drainage of "Zehiha" led to the Dead Sea on the eastward tilting of the Negev following the deposition of the Arava Formation, allegedly proposed by Garfunkel & Horowitz (1966). However, careful examination of the latter publication reveals that the tilting was proposed for post-HaMeshar times, and not as erroneously cited. Thus the data advanced by Avni are not convincing enough to change the former view. In addition, the name "Zehiha" is mistakenly applied to the HaMeshar, since pollen analyses show (Horowitz 1993, unpubl.) that the lacustrine unit originally named Zehiha by Ginat (1997) was most probably deposited during Palynozone QIII, while the HaMeshar corresponds to QI.

The main problem in reconstructing QI paleogeography is to delineate how the Jordan Valley basins connected with the river systems. The only seemingly clear path is the HaMeshar system, connecting the central and possibly northern Arava westward to the sea. To the north, the slowly uplifting Metulla block disconnected the Beqa'a from the Hula, a process that could be helped by effusions of the Mechki Basalt. This had created two paths leading to the Mediterranean, the Litani draining the Beqa'a and a channel connecting the Hula, which most probably crossed westward through southern Lebanon, pouring out to the sea south of Tyre. The post-Pliocene age of the Litani can be concluded from its cutting across the abandoned channels of the Tel Hai–Tanur–Zahlé system (U. Kafri, Geological Survey of Israel 1999, pers. comm.).

The central Jordan Valley, separated from the north by the uplifted Korazim block, seems to have been drained to the Mediterranean through the Lower Galilee, depositing the Amud and other correlative conglomerates (Kafri 1997, Matmon et al. 1999). The Yizre'el Valley appears to have already been uplifting in QI times, since no sign whatsoever indicates that drainage crossed this way.

The southern Dead Sea basin could not have been drained to the Mediterranean through the Be'er Sheva region, since this area was probably elevated already before the Quaternary (Zilberman 1992). The more plausible possibility is that it somehow connected southwestward to the HaMeshar system through the northern Arava where the Melekh Sedom Sands and the Ar Risha Gravels Formation accumulated at that time. It is however also possible that the southern Dead Sea drained to the north, since at the En Gedi 2 borehole QI sediments occur in considerable thickness, somewhere along a possible route connecting to the Bethlehem channel.

The eastward extensions of the QI drainage systems seem clearer, when analyzing the larger rivers crossing the Transjordanian highlands. The locations of these river mouths (Fig. 1.2) in the Jordan Valley are intimately connected with considerable accumulations of QI sediments. The Yarmouk River fed the central Jordan Valley; Wadi Zarqa and possibly Wadi Husban acted as headwaters for the Bethlehem system, while the Mujib and Hasa supplied the southern Dead Sea. Wadi Fidan–Dana and their tributaries were most probably connected with the HaMeshar system. A question still remains to the north, concerning the drainage that led to the Hula basin from the east. The best candidate seems the now deeply incised Nahal Hermon, but there may have been others, now overlain by younger volcanics, of which the more plausible is Nahal Orvim, filled with a succession of basalt flows dating 1.8–1.0 Ma (Horowitz 1979, p. 159).

Subdued volcanic activity is known for QI times, which is certainly a quiet period as compared with the extensive Pliocene eruptions of the Intermediate Basalt. The volcanism is expressed by limited occurrences of the Cover Basalt (as defined in the present text), which includes part of the Bashan Group east of the Hula and Korazim, the Mechki Basalt north and within the Hula, the Dalton west of the Hula, the Ruman in the southern Hula and on the Korazim block and the Amud Basalt west of Lake Kinneret. None of these eruptions covered considerable areas, nor did they extend over long periods of time. Age-equivalent basalts were also penetrated by numerous boreholes in the Hula, Korazim, Kinneret and central Jordan Valley regions, but none were ever of any great extent (Fleischer 1968, Siedner & Horowitz 1974, Marcus & Slager 1985, Mor 1986, Heimann 1990).

7.2.7 Conclusion

Ever since its inception in Oligocene times, the Jordan Valley went through several major stages of paleogeographic evolution. During the first, from the Oligocene to the early part of the late Miocene, the central and northern parts of the region were drained to the Persian Gulf, while its southern sector was connected to the Mediterranean. An interlude during the late Miocene, when several troughs were formed which for part of the time acted as terminal base levels, leads to the Pliocene–earliest Quaternary stage, when the entire Jordan Valley was drained to the Mediterranean. However, during this stage too the Valley was divided, albeit in a different way than before; the Hula was drained northward, through the Beqa'a and the Tripoli graben, and so was never flooded by the Mediterranean, being higher than sea level; the central Jordan Valley, extending down to the Arava, was connected to the Mediterranean via the Yizre'el Valley. Being at sea level or below, it enjoyed flooding by the Pliocene transgressions.

Following a long period of structural quiescence, the Jordan Valley basins were practically filled up with sediments, so that when the sea level dropped considerably at the Pliocene–Quaternary boundary a river system was formed, crossing the Valley on its way from the eastern highlands to the Mediterranean, while subsiding basins within the Valley acted as intermediate lakes, in QI times. This landscape ended with the onset of QII, when extended subsidence of the Levantine Rift, accompanied by uplift of both its shoulders, created the present day endoreic system. Only a little while later the Hula was also connected to this hydrographic net, to assume its present-day role as the principal source of freshwater.

This rather complex history, besides what it left for the geologist, is probably best documented by the rich and varied ichtyologic assemblage of freshwater fish dwelling in the Jordan system. These include almost 30 species, originating from three principal provenances: the palearctic, connected to the Euphrates system; the tropical, with strong affinities to the Nile and other African water bodies; and marine fish which became adapted to freshwater. Some of these, of all three groups, evolved to form a fourth very prominent group of species and even genera endemic to the Jordan system. Krupp (1987) made a thorough study of the history of colonization, concluding that the Euphrates is the source area for the palearctic Jordan fish, which later migrated to the Orontes River. This conclusion, based on zoological grounds, conforms with the inferred paleogeography of the Oligocene and most of the Miocene, when the central and northern Jordan Valley was part of the Persian Gulf domain, closely connected to the Euphrates; during the Pliocene the northern sector was drained to what now constitutes the Orontes system; while the central and southern sectors were connected directly to the

Mediterranean. The fish of marine and tropical origin most probably colonized the Jordan Valley in Pliocene times, when the Nile was close and potent enough to provide a comfortable passageway from Africa, and the Mediterranean invaded the Valley for a considerable length of time. This process could have also extended into QI times. The endoreic Quaternary Jordan Valley was an ideal cradle for endemism.

7.3 NATURAL ENVIRONMENTS

Natural environments are usually characterized by their biota, comprising both animal and plant associations, a direct outcome of the combination of topography and climate. The composition of plant or animal associations also depends on many local factors, such as refuges, migrations, hydrography, salinity, types of soils, etc. Paleogeography, climate and local hydrological conditions define also, to a great extent, the types of sediments left behind, or the style of erosion in a certain period (see Section 7.1). These variables are thus applied to delineate the natural environments of the Jordan Rift Valley and its surrounding regions. The Jordan Valley is environmentally unique, due to its particular morphology, hydrology and climatic regime. Thus a differentiation is made here between regional and local environments. The term "regional" applies to the entire southern Levant, where evidently more than a single natural environment may occur, depending on climatic gradients or topography; the term "local environment" refers to any of the numerous natural habitats particular to the various parts and conditions within the Jordan Rift itself.

Reconstructions of past environments are here based primarily on pollen assemblages, recovered from sediments deposited within the Jordan Rift Valley basins, discussed in Chapter 6. Since pollen grains are brought to the basins both from the surrounding local vegetation growing close by, and from distant regional provenances by winds, rivers and other agents, they can testify for regional and local environments alike (Horowitz 1992a, Chapter 8). A distinction between the two groups of constituents is not always straightforward, with some overlapping always present, so other criteria are also employed. These include, as the particular case may be, any additional data obtained from other groups of fossils, types of rocks, modes of erosion and so on.

7.3.1 Middle Oligocene

Pollen spectra of the middle Oligocene in the Sedom Deep 1 borehole comprise very high percentages of arboreal origin, at 70–90% of the regional components (50–60% of the total). These, mainly Juglandaceae and Betulaceae, indicate a lowland wet subtropical biotope, possibly marsh forest or very dense savanna, for the regional environment. The higher, drier lands were covered by conifers, whose pollen percentages are subordinate, in the range of 10% of the total arboreal count, indicating that such environments were either restricted in area or quite far away from the Jordan Valley. It would be impossible to further elaborate, since only a single locality was analyzed. The littoral-neritic fossils recovered from the middle Oligocene formations near Lake Kinneret seem to support a subtropical nature for the sea at that time, which is indicated also by other Oligocene faunas recovered from the Levantine coastal plain, including coral reefs (Avnimelech 1936).

The middle Oligocene local pollen components in Sedom Deep 1 comprise chiefly Gramineae and Chenopodiaceae, indicating a primarily wet environment, probably with patches of higher salinity, possibly shallow coastal surroundings with nearby marshes. A rather similar local environment, shallow lake or lagoon, can also be concluded from the fossils and mudcracks in the central Jordan Valley and the Arava formations, as well as the oncolites in the middle part of the Zefa.

During the middle Oligocene the southern Levant must have occupied a place in the then widely distributed, wet subtropical climatic belt, most probably affected by monsoons.

7.3.2 Palynozone Ma, late Oligocene to early Miocene

The late Oligocene seems quite different from the middle, which is apparent from the pollen spectra in both the southern Dead Sea and the Mediterranean offshore. Although the types of arboreal components did not change, their overall percentages dropped drastically when Palynozone Ma commenced, to less than 20% of the regional and less than 10% of total counted grains. Among the trees conifers become more common, while Compositae, indicating dry environments, are the most abundant constituents of the regional spectra. These pollen assemblages indicate a warm, dry subtropical environment, possibly bearing a reduced savanna vegetation, for the late Oligocene through early Miocene. The increase in conifer pollen does not necessarily indicate any expansion of these trees; it is sufficient that the areas previously occupied by other, hydrophil trees shrank due to increasing aridity, to reduce their masking factor, which seemingly increases the share of xerophil elements. Sedimentological analyses of the En Gev Sands (Giveon 1984) also indicate deposition in an arid environment.

Local environments for Ma times did not change much from the middle Oligocene, remaining warm and wet, and somewhat saline close to water bodies, as can be seen from the predominant red-beds in Sedom Deep 1 spanning the entire Ma, in the southern Dead Sea region, or the fluvio-deltaic sands at En Gev at the beginning of Ma. The percentages of local pollen out of the total had increased in Ma, but this again may seem so only as a result of a reduction in the coverage of the regional vegetation due to greater aridity. Toward the later part of Ma, river valleys began to be filled up with coarse conglomerates such as the Shahaq Member of the Hazeva, as a combined result of rising sea level and the rain regime, most probably characterized by rare but powerful thunderstorms.

Pollen are not available from the northern sector of the Jordan Valley for Ma, but comparison of the clastics could hint at the environmental gradient at that time. While those penetrated by Sedom Deep 1 are mostly red-beds (see discussion below) formed under warm but humid conditions, the En Gev Sands are typically white, indicating lesser availability of water, and showing that the south was already more humid than the north during Ma times, a situation that persisted until the end of Mc.

Palynozone Ma continues until the transition of Foraminifera Zone N6 to N7, in the early part of the Burdigalian. Thus, in terms of the regional environments, there is no difference between the late Oligocene and the early part of the early Miocene. The southern Levant that had occupied the wet subtropical climatic belt during the middle Oligocene was drying out at the beginning of Palynozone Ma.

7.3.3 Palynozone Mb, middle Miocene

The semi-arid environment typical of Palynozone Ma changed considerably at the beginning of Mb times, in the later part of the Burdigalian. The pollen spectra, rich in arboreal components, mainly Juglandaceae and Betulaceae, indicate a moist subtropical climate, with a wet, partly marshy savanna vegetation for Mb. This is also strongly supported by the vertebrate fauna described above (see Section 7.1.3.2), as well as by the corresponding littoral-neritic fossils, giant oyster beds, etc. As previously, during Ma, the higher elevations and the drier localities were covered by conifers. Since Mb sediments are known from all along the Jordan Valley, as well as offshore, they provide a chance to better depict the environmental gradient along the southern Levant for this time, by their pollen assemblages.

The main trends in Mb pollen spectra are portrayed in Table 6.5.2.1. It is quite evident that arboreal pollen, as part of the total spectra, diminish northward, from almost 30% in the southern Dead Sea and the southern Mediterranean offshore to less than 13% in the Hula. Notably, where the arboreal pollen section of the regional vegetation is concerned the difference is much less dramatic, from somewhat more than 60% in the south to somewhat less than 50% in the Hula. It seems that these figures should be regarded as indicators for the regional gradient, which was wetter to the south and somewhat drier to the north. The composition of arboreal pollen spectra stresses this environmental gradient. While to the south these are dominated by Juglandaceae and Betulaceae, conifers take the upper hand to the north.

Among the non-arboreal pollen, those derived from desert plants are more common to the north, almost 50% of the regional elements, decreasing to less than 20% at Sedom Deep 1. An opposite trend is displayed by pollen of steppe vegetation. The gradient is displayed as well by the sediments, and the lacustrine–lagoonal deposits of the Herod Formation are considerably more dolomitic and less reddish (Shaliv 1991) than those of Hazeva (Sneh 1982), the latter being dominated by limestones.

The greater differences in arboreal pollen, as part of the total, are seen to be the result of morphology. The wider floodplains of the Hazeva–Dana system most probably maintained extensive areas covered by local non-arboreal vegetation, connected with an abundance of water. This is also supported by the relatively high shares of hydrophil plant pollen to the south. Conversely, the prevalence of conifers in the north may have resulted partly from the higher topographies in that region, due to the already elevated Hermon and Lebanon ranges.

The local environments of the Hazeva–Herod systems, as expressed by the high percentages of Gramineae pollen, are mostly fluvio-lacustrine, as is also shown by their typical sediments as well as the occurrences of various fish and crocodiles in the Hazeva sediments. Most of the sediments are either fine clastics or chemical precipitates, representing a combination of high sea level, rather flat landscape and steady, gentle, ample rains.

The typical red colors of the Hazeva–Herod rocks result from a combination of climate and the environment of deposition. However, the types of required climate and environment seem debatable. Authors such as Walker (1967) claim that the red color was formed after deposition, in an arid environment. Glennie (1970) brings forth a variety of opinions about red-beds and their applicability as typical desert sediments. According to Glennie (it seems that opinions have not changed much up to the present), the only subjects on which all geologists agree are that most red-beds sequences are terrestrial in origin, colored by the presence of red-stained clays and a coating of red iron oxide on grains of sand and silt. Other than that, most researchers think that humidity is necessary for the oxidation of iron-rich minerals such as hornblende, augite and pyroxene, in order to form the red coloration.

At present red coloration is typical of warm humid climates such as the tropics and wetter subtropics, with their typical laterites, or the northern Mediterranean region, where such conditions are typical in the summers. Therefore, Horowitz (1992a, p. 136) did not accept the idea that an arid environment is suitable for the formation of red-beds. Rather, these were formed whenever a combination of humid climate and higher temperatures, of a wet subtropical or wet Mediterranean character, took over the desert. When the climate changes to arid the red-beds are preserved, in the desert conditions, from being reduced again. In a temperate climate, or when they are buried in marine environments, red-beds are reduced and lose their typical colors (Glennie 1970).

7.3.4 Palynozone Mc, late Miocene

Pollen spectra characterizing Palynozone Mc are almost devoid of arboreal components, with overwhelming percentages of Compositae, indicating an extremely arid environment, in nature similar to the present Sahara (Horowitz & Horowitz 1985, 1990, Horowitz 1992a, p. 373). The rare arboreal pollen are of the same groups encountered before, in Mb times, possibly indicating that the southern

Levant was still within the influence of the subtropical belt, but its driest, northernmost part. No other fossils are known, except for marine foraminifera indicating a warm sea (Derin & Reiss 1973).

The typical sediments are gravels deposited in the channels leading to the seas and to some extent to the newly formed Eritrean faulting basins, both within the limits of the Jordan Valley (Shaliv 1991) and outside its boundaries (Calvo et al. 1997). The deeper basins, as well as almost the entire Mediterranean at that time, mainly accumulated evaporites (Marcus & Slager 1985, Baker 1994). These intermingle with coarse clastics, probably brought to the deeper localities by the rare but possibly powerful floods, caused by occasional thunderstorms. This type of erosion and deposition could not be hindered by the poor vegetation, also a result of the extremely arid climate.

It seems that the environmental gradient during Mc times retained the previous Mb characteristics, namely drier to the north. This is however very difficult to prove, since the entire southern Levant region was so dry that differences between north and south are hardly discernible. Local environments are typified by salinity, a natural consequence of the rarity of rains, and also the tectonics possibly creating endoreic inland basins.

7.3.5 Palynozones Pa and Pb, Pliocene

A dramatic change of natural environment marks the Mc–Pa transition in the southern Levant, affecting fauna, flora and sediments. In conjunction with the late Miocene Eritrean faulting and the rise in sea level at the beginning of the Tabianian, the Pliocene landscape is entirely different than before. Pliocene pollen spectra considerably differ from the Miocene ones, and are dominated by arboreal pollen of *Quercetalia* and a conifer, *Picea orientalis*, representing a progression from the subtropical monsoon-controlled climate to the onset of a Mediterranean environment (Suc 1989, Fauquette et al. 1998). This change is part of a global trend, expressed in a gradual contraction of the tropical belt throughout the Cenozoic (cf. Traverse 1988, p. 291), but was also accentuated by tectonics in southeastern Europe which opened the way for polar air to reach the Levant.

Palynozone Pa is richer in arboreal pollen as compared with the overlying Pb. Shares of up to 40% of the pollen produced by regional vegetation are typical for Pa, while Pb is characterized by figures of 20–30% for that group. There is no essential difference in the composition of the non-arboreal pollen group between the Pliocene and the Miocene and, indeed, the same taxa continue to the present day. This may partly be a result of problems of identification, mentioned in Chapter 6. Pliocene climates of the northwestern Mediterranean show similar trends (Fauquette et al. 1998).

During the Pliocene the offshore sections (Horowitz & Derin 1987) show somewhat lower arboreal pollen percentages than those obtained from the Jordan Rift region, but are richer in Betulaceae–Juglandaceae grains (Table 6.5.2.2). The reason for the slight over-representation of non-arboreal pollen in the offshore sequences is probably connected with the activity of the Nile, which commenced at the early Pliocene. The Nile waters carry and deposit offshore considerable amounts of non-arboreal pollen (Horowitz 1979, p. 188), originating from its bank and delta vegetation, a habitat which may also have supported the Betulaceae–Juglandaceae. The area of Israel constituted during the Pliocene a lowland (Horowitz 1979, p. 75) in which oaks most probably grew, while to the northeast, where the landscape became higher in elevation, conifer forests covered the hills and mountains. The pollen spectra indicate a wet northern Mediterranean, Pontian type climate for Pa, becoming slightly drier in Pb times.

The typical Pliocene malacofauna and fish (E. Tchernov, Department of Evolution, Systematics and Ecology, the Hebrew University of Jerusalem 1999, pers. comm.), all indicate a circum-Mediterranean to southern Palearctic affinity, completely different from the Miocene subtropical habitats, which agrees with conclusions drawn from the pollen spectra.

Increased salinity is the key to understanding the nature of the sediments accumulated in the Pliocene depositional basins, in spite of the relatively humid regional climate. Salinity originated from three sources: seawater that penetrated deep into the Jordan Valley at the beginning of the period, and continental donations from either entrapped seawater, or groundwater dissolving salts, mainly from basalts (Raab 1998) but probably also from late Miocene evaporites. Salinity is expressed in the types of sediments and their prevalent colors, mostly white–gray chalks, limestones, plenty of dolomites and hard, calcareous–dolomitic cementation of many of the gravel beds. This factor is readily clear where hypersaline basins had accumulated evaporites, but has a considerable effect even in lakes, such as the Bira Lake (Raab 1998) or the Arava Lake (Avni 1998). The effects of salinity are not felt to the north, where the Tel Hai, Tanur and Zahlé Beds display lithological and faunal trends which are typical only of their freshwater southern counterparts, especially the Gesher Formation.

Autochtonous fossils are rare throughout all the Pliocene formations of the Jordan Rift Valley, but when they do occur their typical assemblages almost entirely comprise dwarfed constituents, with very restricted species diversity, a characteristic of ecologically stressed environments (Horowitz 1992a, p. 434). These indications also stand for rocks regarded as representing lacustrine environments. Pollen grains, on the other hand, are numerous, indicating that salinity was restricted to the depositional basins, not to the regional environments outside the water bodies. Among the local components confirming salinity, Chenopodiaceae are very common (Table 6.5.2.2), more so to the south as expected from the newly developed environmental gradient.

Similarly to the early and late Miocene, the lower sea level of Piacenzian times, in conjunction with the drier Pb climate, make gravels and conglomerates much more common than any other type of sediment in the Jordan Valley and surroundings for this period.

7.3.6 Palynozone QI, earliest Quaternary

The spruce and oak forests, so typical of Palynozone QI, their pollen abundant even in sediments from the southern Negev, reflect best of all the environments of this period. An abundance of freshwater, following the retreat of the sea from most of the region, characterizes the environments of deposition. The sediments, many of which appear as typical red-beds, were deposited in wide, meandering rivers with copious floodplains, which crossed several deeper, intermediate basins in the Jordan Valley on their way to the Mediterranean. QIa and QIc are the more humid phases, QIb is somewhat drier.

The north–south environmental gradient, as seen in the pollen spectra of QIa (Table 6.5.2.3), is well developed, but the south is not too poor in arboreal pollen either. Their lesser values to the west are probably a result of the pollen imported by the Nile (seen by the increased proportion of tropical fern spores) and the wide, shallow coastal plain, which contributed large amounts of non-arboreal constituents. Of the trees, oaks prevail to the north and conifers to the south, following the southward decrease in rain, while the coastal lowlands maintained the Betulaceae–Juglandaceae trees. The non-arboreal part shows, as expected, a prevalence of Gramineae–Cyperaceae to the north, replaced to the south by Chenopodiaceae.

Freshwater lakes, probably controlled by somewhat extended subsidence, are known from the Hula, the central Jordan Valley, and the southern Dead Sea. The relatively flat landscape was most probably covered by forests to the north, grading southward into park forests, in which various long legged and characteristic savanna animals roamed. These included (Hooijer 1958) African animals such as three kinds of elephants, giraffes, rhinoceros and hippopotamus, together with palearctic ones, *Leptobos*, *Hipparion* and *Bos*. The rivers were infested with crocodiles.

The strange combination of African and palearctic fauna, living in a northern Mediterranean forest habitat, testifies to the great potential of the southern Levant as a refuge area. Indeed, although a Mediterranean environment has prevailed in this region ever since the Pliocene, both tropical and palearctic animals and plants are still found in the southern Levant even today (see Section 3.6.2.1).

CHAPTER 8

Geophysics

Avihu Ginzburg and Zvi Ben-Avraham

Despite the uniqueness of the Jordan Valley as an outstanding feature of the Earth's crust, geophysical methods were not used in its exploration until some 50 years ago. While geological studies began in the middle of the 19th century, geophysical means which could be easily used in the field were developed essentially after World War II, mainly for mineral and petroleum prospecting. The earliest forerunner of systematic investigations in the Jordan Valley was the bathymetric mapping of the Dead Sea by Lynch (1849). Modern marine geological studies began with a bathymetric study of the southwestern part of the Dead Sea by Neev & Emery (1967).

The earliest references to geophysics were in seismicity studies. Willis (1928) treated historical earthquakes in the Holy Land. The first recorded geophysical measurement in the Jordan Valley were gravity observations conducted using a pendulum by Lejay (1938), in which the gravity low associated with the Rift was first glimpsed. Sporadic resistivity soundings mainly for water exploration were conducted but those, by their very nature, provided only shallow information.

The Weizmann Institute of Science initiated a systematic gravity survey in 1951. The Geophysical Institute of Israel continued the project in the 1960s. This project was partly supported by oil companies, which began petroleum exploration in the early 1950s. The Natural Resources Authority of Jordan undertook gravity investigations in Transjordan.

Airborne magnetic coverage of Israel was carried out in 1967 and 1968, and a magnetic map of the country is available (Folkman & Yuval 1976, Rybakov et al. 1997). The magnetic survey of Transjordan was conducted in 1979 (Hatcher et al. 1981).

The first seismological observatories were established in Israel as a part of the WWSN network in 1954. A modern seismological network has been in operation since 1982 (Shapira 1982). Seismological stations have been in operation in Transjordan since 1986. Seismic reflection surveys were conducted for petroleum exploration, mainly in the southern Dead Sea basin and the northern Arava. Some reflection work was carried out in the central and northern parts of the Jordan Valley.

Long-range refraction profiles were shot in the Jordan Valley, Arava and Gulf of Aqaba as a part of a crustal study of Israel (Ginzburg et al. 1979a). A similar program was conducted in the area east of the Jordan Valley (El-Isa et al. 1987a). Thus the deep structure of the crust in the Jordan Valley and its flanks is known. High-resolution seismic profiling of the Dead Sea was begun by Neev & Hall (1976) and yielded information about the shallow structure of the lake deposits. Very meager heat flow is available in the Jordan Valley (Ben-Avraham et al. 1978). The water-covered portions of the Jordan Valley, the Dead Sea and Lake Kinneret were studied by a variety of marine geophysical methods (Ben-Avraham et al. 1996, Ben-Avraham 1997). These included seismic reflection, seismic refraction, bathymetry, magnetics, gravity, heat flow and electrical conductivity. Important data were made available for the southern part of the Dead Sea through the extensive seismic profiling obtained by oil companies in the course of petroleum exploration.

In the following paragraphs we shall review the available geophysical data and interpretations, and synthesize the relevant parts into a structural model of the Valley.

8.1 REGIONAL BACKGROUND

The gravity map of the Levant (Fig. 8.1.1A) gives a good overview of the regional crustal structure. Transjordan is characterized by broad negative Bouguer anomalies with amplitudes varying between −10 mgal to the west and −65 mgal to the east, typical of a continental crust. Similar anomalies are observed over Sinai and the Negev, while over central and northern Israel the gravity anomalies gradually become more positive northward and westward, indicating a change in the crustal structure in these directions. The dominant regional trend of the gravity anomalies is the westward change, in which gravity values increase dramatically, reaching +100 mgal in the Levantine basin. This increase indicates a major change in the regional crustal structure, and can be ascribed to a thinning to the west, as first suggested by Folkman (1976). The crust thins underneath the continental shelf, while an oceanic crust covered by a thick succession of post-Jurassic sediments was encountered underneath the central part of the Levantine basin (Makris et al. 1983).

On this regional gradient is superimposed a north–south trending alignment of narrow gravity minima of considerable amplitude, which extend from the southern tip of the Sinai Peninsula to the northern Lebanon. This elongated series of gravity anomalies includes the Jordan Valley. Within it are a number of tectonically controlled depressions, including the Dead Sea and Lake Kinneret. These depressions are considered by most geoscientists studying this area to be pull-apart basins, formed by the left lateral strike-slip movement along a transform fault, commonly referred to as the "Dead Sea Transform". The transform fault separates the

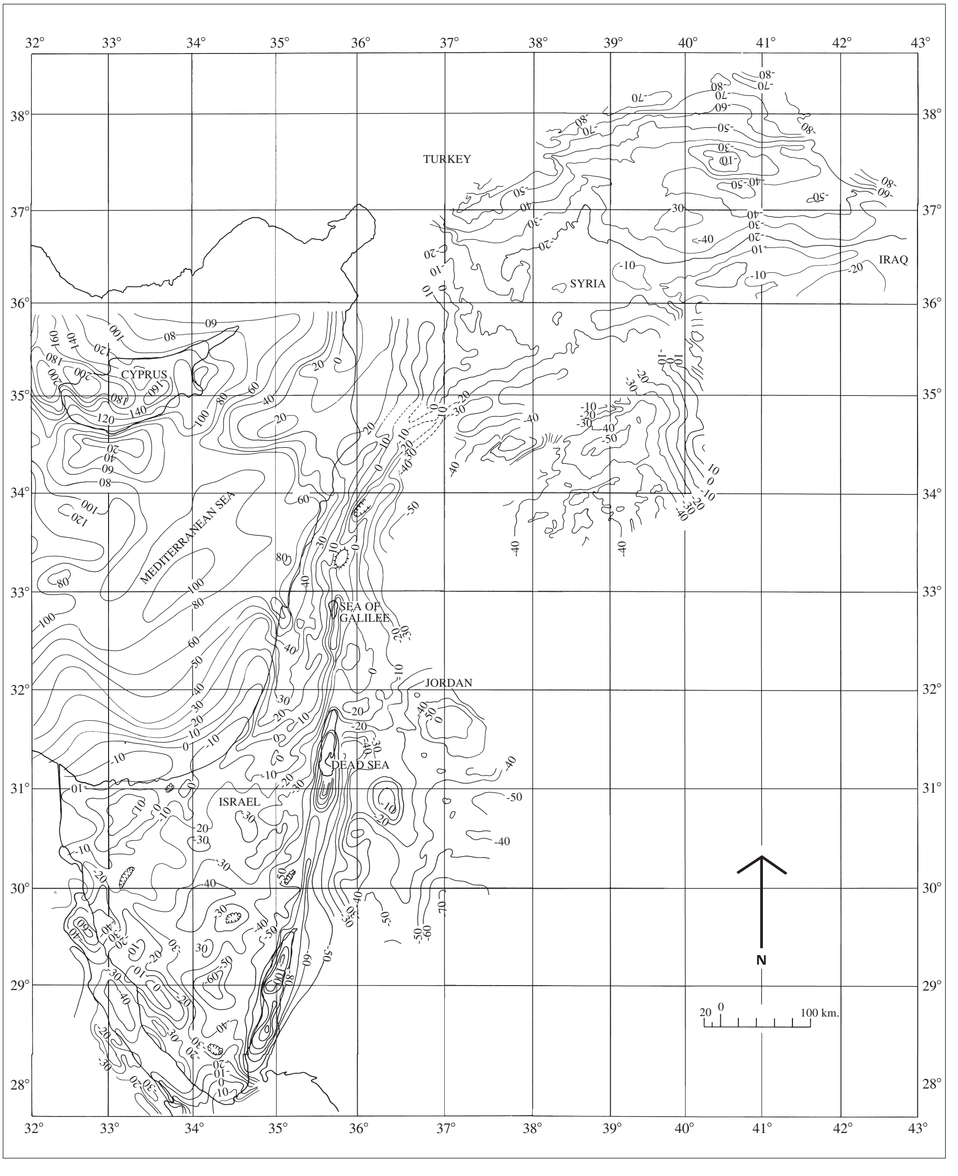


Figure 8.1.1A. Compilation map of Bouguer anomalies in the eastern Mediterranean and Levant. Note the elongated north–south trending gravity low extending along the Dead Sea Rift from the Red Sea to Syria (after Ben-Avraham & Ginzburg 1990).

Arabian and African plates with an estimated total lateral movement of 105 km (Quennell 1959, Freund et al. 1970).

Long-range refraction studies indicate the presence of a continental crust under southern Israel, reaching a thickness of about 35 km (Ginzburg et al. 1979a). The thickness of the crust decreases gradually northward, and under central and northern

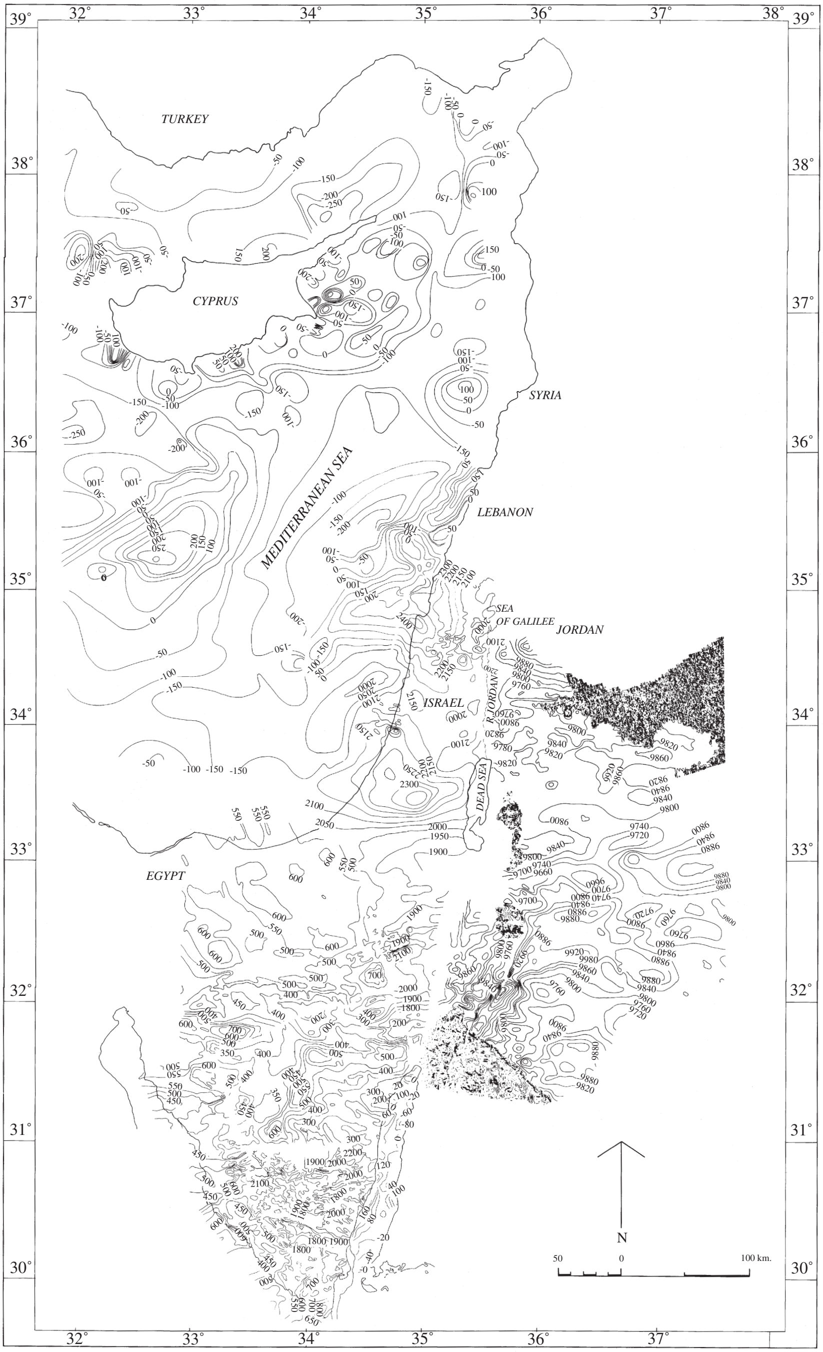


Figure 8.1.1B. A compilation map of total field magnetic anomalies over the eastern Mediterranean and Levant (after Ben-Avraham & Ginzburg 1990).

Israel it is only some 25 km thick, while still maintaining its continental characteristics. Westward a crustal thinning typical of passive margins is observed. At a distance of about 130 km west of the Israel shoreline, in the Levantine basin, an oceanic two-layered crust overlain by a 14 km column of Triassic to Recent sediments has been interpreted (Makris et al. 1983, Ginzburg & Ben-Avraham 1987). A similar refraction study carried out in Transjordan (El-Isa et al. 1987a) shows that the crust underneath Transjordan is continental, with a maximum thickness of 40 km to the east, gradually thinning westward to the region east of the Jordan Rift Valley. The velocity structure of the crust in Transjordan is different from that in Israel, thus suggesting that the crust on either side of the Jordan Rift maybe different. The crustal thinning mentioned may either be the easternmost indication of the thinning observed in Israel, or an indication of a local thinning associated with the Jordan Rift, also observed by Ginzburg et al. (1979a).

The magnetic field over the Levant (Fig. 8.1.1B) exhibits short-wavelength anomalies produced by the exposed Precambrian igneous rocks of the Arabian Massif in Sinai, southern Transjordan and the extreme south of Israel. Broad anomalies originating from the sediment-covered basement feature over the rest of the area.

Geological and geophysical evidence suggests that the Levant is composed of several crustal domains (Ben-Avraham & Ginzburg 1990). The deep crust of the Negev in southern Israel is probably the continental crust of the Arabo-Nubian Massif. North of it are located the Judea–Samaria domain in central Israel and the Galilee–Lebanon domain (Fig. 8.1.2). The boundaries between the domains, the Hebron line and the Carmel line, extend from the continental margin to the Jordan Rift Valley.

8.2 THE DEAD SEA–ARAVA DEPRESSION

Examination of the compiled and generalized Bouguer gravity map of the Levant (Fig. 8.1.1A) shows a prominent elongated series of interconnected gravity minima following the trend of the Jordan Rift. The anomaly extends from the Gulf of Aqaba to northern Syria, and is very prominent along the Jordan Valley.

In a recent cooperative Israeli–Transjordanian–US project, supported by US Aid following the Israeli–Transjordanian peace treaty, the gravity networks of Israel and Transjordan were joined. Thus a detailed Bouguer gravity map covering the entire Jordan Valley and its flanks is available, as shown in Fig. 8.2.1. The gravity minima follow the topography of the rift and comprise prominent elongated north–south trending lows. The maximum width of the negative anomalies is some 25 km, but narrows in places to as little as two. The flanks are characterized by steep gradients, steeper on the eastern flank from the Dead Sea southward, and on the western flank from the Dead Sea northward. Within the regional gravity

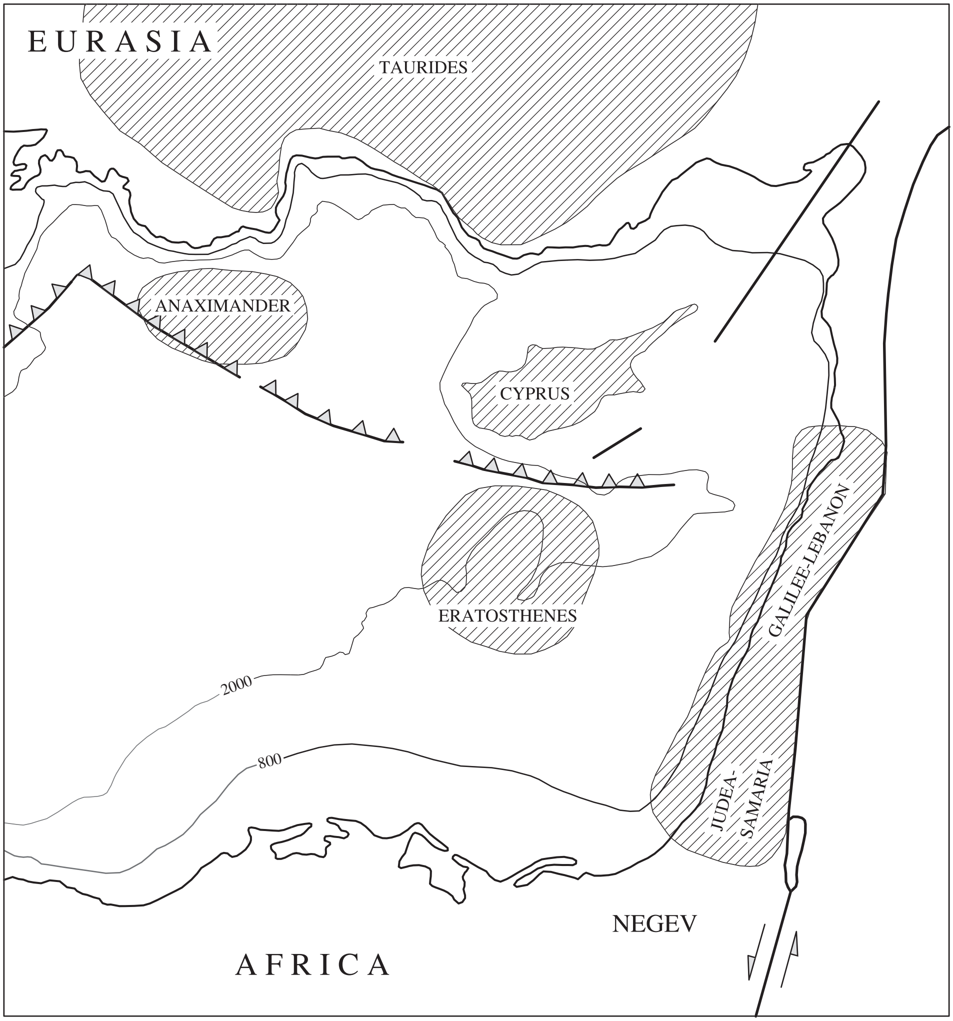


Figure 8.1.2. A map showing the schematic distribution of different crustal domains in the eastern Mediterranean, Levant and adjacent areas. The domains, which have a different crustal structure from the adjacent areas, are represented by the hatched areas (after Ben-Avraham & Ginzburg 1990).

low, four distinct separate basins can be easily distinguished. They are, from south to north: the southern Dead Sea–Arava, the northern Dead Sea, Lake Kinneret–Bet She'an and the Hula basins. The Dead Sea–Arava depression is by far the most prominent of these and a large amount of geophysical data has been collected in it over the years.

The largest and most prominent gravity anomaly in the Jordan Valley is that associated with the Dead Sea. This anomaly extends from the Elat–Dead Sea watershed in Gav Ha'Arava to somewhat north of the Dead Sea (Fig. 8.2.1).

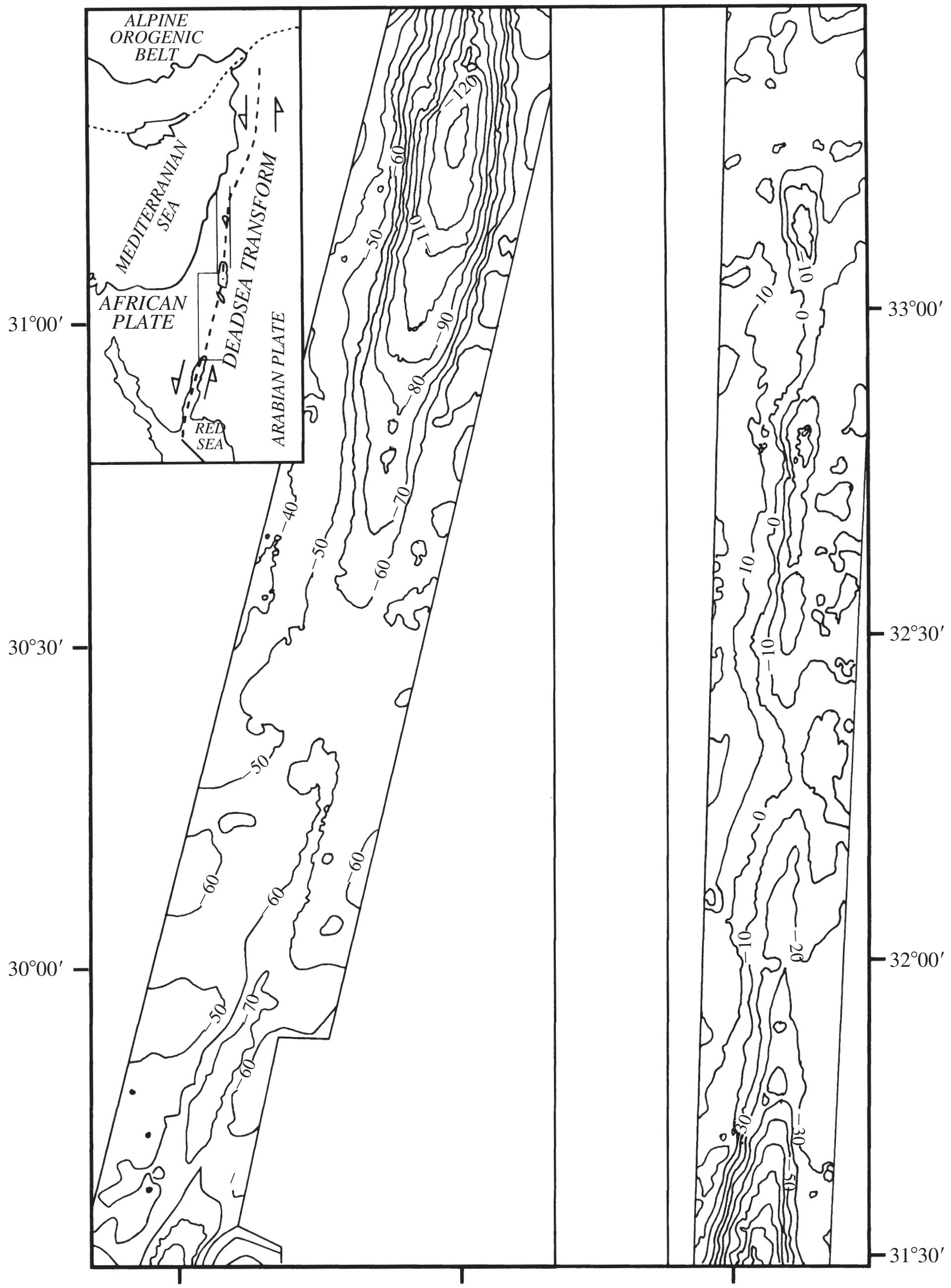


Figure 8.2.1. Gravity maps of the Dead Sea Rift from Elat to northern Israel. Note the main Bouguer anomalies of the Jordan Valley: the Dead Sea, Lake Kinneret and the Hula basin (after ten Brink et al. 1999).

8.2.1 Topography and bathymetry

The Dead Sea depression is the most prominent feature in the Jordan Valley. It is the lowest depression on the Earth's crust, the current water level being 409 m below sea level. It is an abrupt asymmetrical discontinuity. Escarpments on the western flank mark a drop of 300–500 m from the plateau. On the east the plateaus are higher, 900–1,000 m above sea level. The asymmetry is clearly displayed in Fig. 1.3A, which is based on the Hall (1996, 1997) digital terrain map (DTM). The northern end of the lake bounds the depression to the north. The watershed between the southward and northward drainage in the Arava marks the southern end of the Dead Sea–Arava depression and of the Jordan Valley.

Lynch (1849) first surveyed the bathymetry of the lake; it was his mapping that showed the existence of a shallow basin south of the Lisan Peninsula, and a deep basin to the north. The first modern mapping of the lake was by Neev & Emery (1967), who surveyed its southwestern portion. In 1967 a survey was conducted which comprised echo sounding, magnetic surveying and single channel seismic reflection profiling (Hall & Neev 1975). The southern basin, which is very shallow and slopes gently to the north, is now covered by the evaporating pans of the Israeli and Transjordanian potash works. The area between the Lisan Peninsula and the western shore of the lake is now, due to the recent decline in the lake's water level, a mud flat. Most of the floor of the northern basin is flat. The basin, which is the main water body, reflects the asymmetry shown by the topography. The lake floor slopes from the shoreline to the bottom on the eastern shore at 30°, while the slope on the western shore is gentler at about 7°. The northern end of the basin is marked by the gently sloping Jordan delta.

8.2.2 Seismic refraction

A long-range refraction experiment was conducted in southern Israel in 1977 (Ginzburg et al. 1979a,b, Perathoner et al. 1981), providing information about the deep structure of the crust along the Jordan Rift and the adjacent area to the west. Some years later a seismic refraction experiment was carried out east of the Rift (El-Isa et al. 1987a,b). The refraction profile along the Rift was about 600 km long, extending from Lake Kinneret to the southern end of the Gulf of Aqaba. The initial evaluation was based on first arrivals (Ginzburg et al. 1979a); late arrival information was subsequently used to obtain the detailed velocity structure of the crust underneath the Rift and the adjacent areas (Ginzburg et al. 1979b).

Information in the southern portion of the Dead Sea section was obtained from a reversed refraction profile, while in the northern part only an unreversed profile was completed. Energy propagation from the Dead Sea shots was excellent. The structural models computed from the seismograms were plotted on a longitudinal section (Fig. 8.2.2), which also shows the relative Bouguer anomaly. The 2.5 km/sec layer represents the Rift fill in this section. The 6.15 km/sec layer represents the

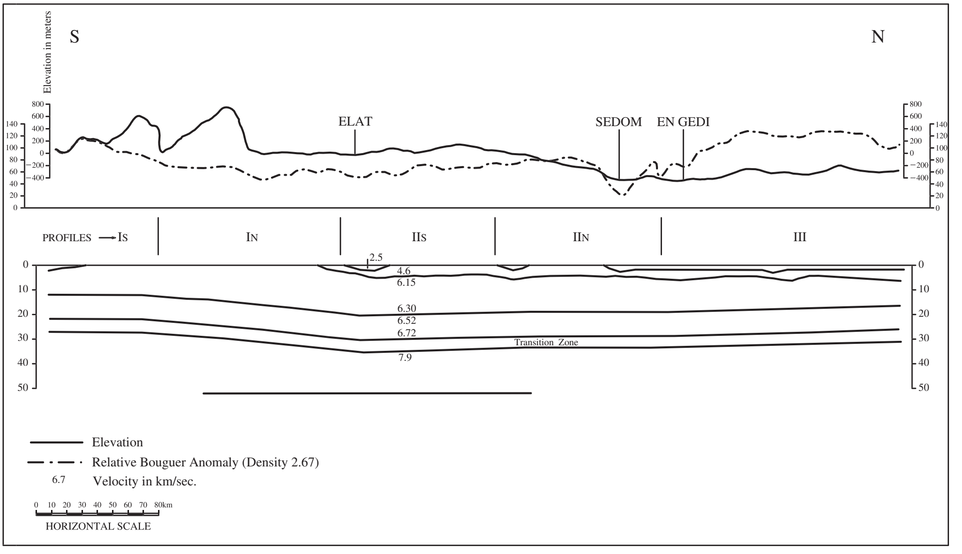


Figure 8.2.2. A structural section along the Dead Sea Rift and the Gulf of Elat based on seismic refraction data. Note the thickening of the crust toward the Red Sea (after Ginzburg et al. 1979b).

upper crust, 6.52–6.72 km/sec indicates the lower crust, while the last layer is a transition zone, in which the velocity increases to the top of the upper mantle, represented here by the 7.9 km/sec velocity. The results indicate that the Jordan Valley and its immediate surrounding are underlain by a thinner than usual crust, about 30 km, compared with 40 km in the Negev and southern Israel. The Rift is also characterized by the presence of a transition layer at the base of the crust. This transition zone was not recorded beneath Israel and northern Sinai, where a sharp velocity contrast marks the base of the crust (Moho).

The results obtained in Transjordan (El-Isa et al. 1987a,b) indicate a continental crust with a different velocity structure from that observed beneath southern Israel. An east–west cross section from Transjordan across the Rift Valley (Fig. 8.2.3) to Israel shows that the thickness of the crust decreases gradually in Transjordan, from 40 km in the east to just over 30 east of the Dead Sea. A transition zone, thicker than that observed under the Rift, is present at the crust–upper mantle boundary underneath Transjordan. Given the lack of a measured east–west crustal section across the Jordan Valley, any attempt at explaining the variations is at best tenuous. However, the modeling of a regional gravity profile across the Rift, just north of the Dead Sea (ten Brink et al. 1990) also suggests that two different crustal blocks exist on either side of the depression.

The thickness of the fill and the depth to basement in both basins of the lake were not resolved, because the seismic recording stations of the 1978 experiment were deployed only along the western margins of the lake. To obtain this

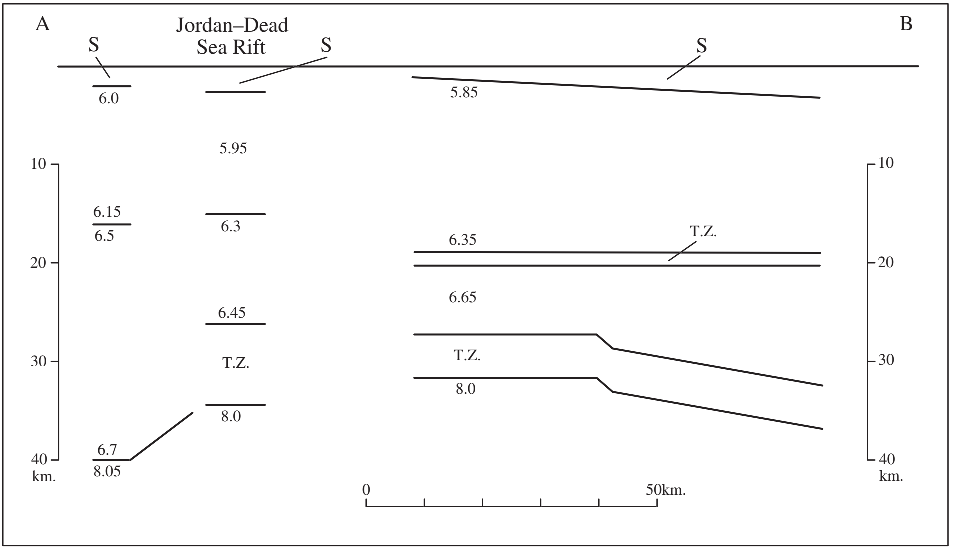


Figure 8.2.3. An east–west section across the Dead Sea and adjacent area based on refraction measurements. The numbers shown on the section indicate the P-wave velocities of the different velocity layers. T.Z. indicates the velocity transition zone at the base of the lower crust. The difference in crustal structure on either side of the valley is evident (after El-Isa et al. 1987a).

information a seismic refraction experiment along the axis of the lake was conducted in 1996 (Ginzburg & Ben-Avraham 1997). Ocean Bottom Seismometers were deployed in the northern basin and land stations north and south of it. Travel-time interpretation of this profile (Fig. 8.2.4) shows clearly the difference between the northern and southern basin. In the northern basin the depth to basement varies from 4 km to the north to about 7 near the Lisan Peninsula. The young (Pleistocene) sedimentary cover with a velocity of 2.0 km/sec varies from 2 km to the north to 4 km to the south. The 3.0 km/sec layer represents the pre-Rift sedimentary section, which overlies the 6.0 km/sec basement. In the southern basin the depth to basement varies rapidly from 7 km to the north to some 14 to the south. As can be seen in Fig. 8.2.4, the 2.0–2.8 km/sec layer includes two separate occurrences of a 4.2 km/sec, typical of evaporite layers. The southern occurrence represents the Lisan diapir, which was penetrated by drilling. The northern high-velocity layer within the low-velocity sediments is probably another massive salt layer, possibly a diapir.

8.2.3 Structural interpretation of the gravity data

Gravity data were collected in the Dead Sea by a marine gravity survey of the northern basin of the lake. These were integrated with land data from Israel and Transjordan, to give Free-Air and Bouguer anomaly maps of the Dead Sea

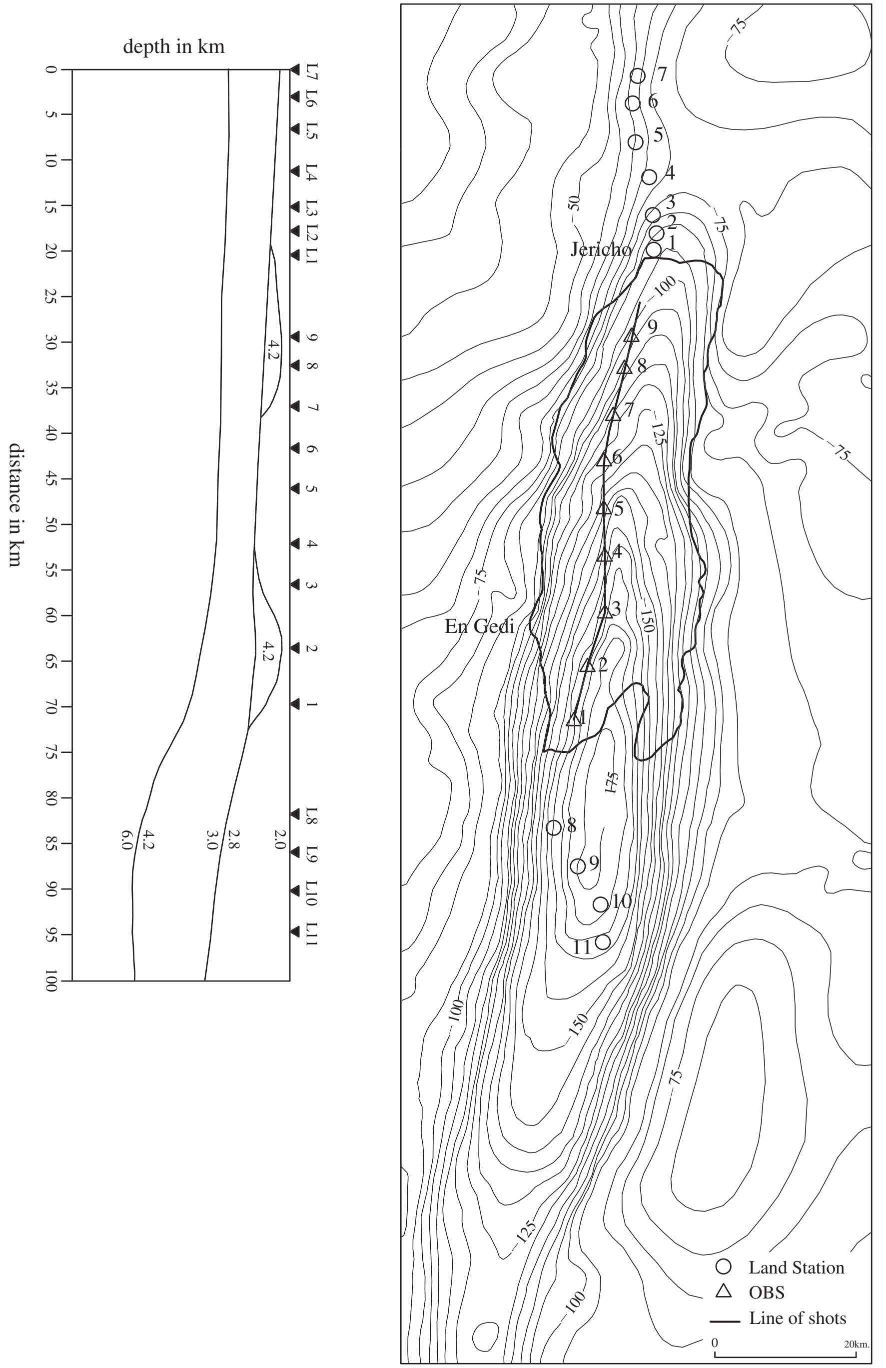


Figure 8.2.4. A north–south seismic refraction section approximately along the axis of the Dead Sea. Note the thickening of the sedimentary section in the south (after Ginzburg & Ben-Avraham 1996).

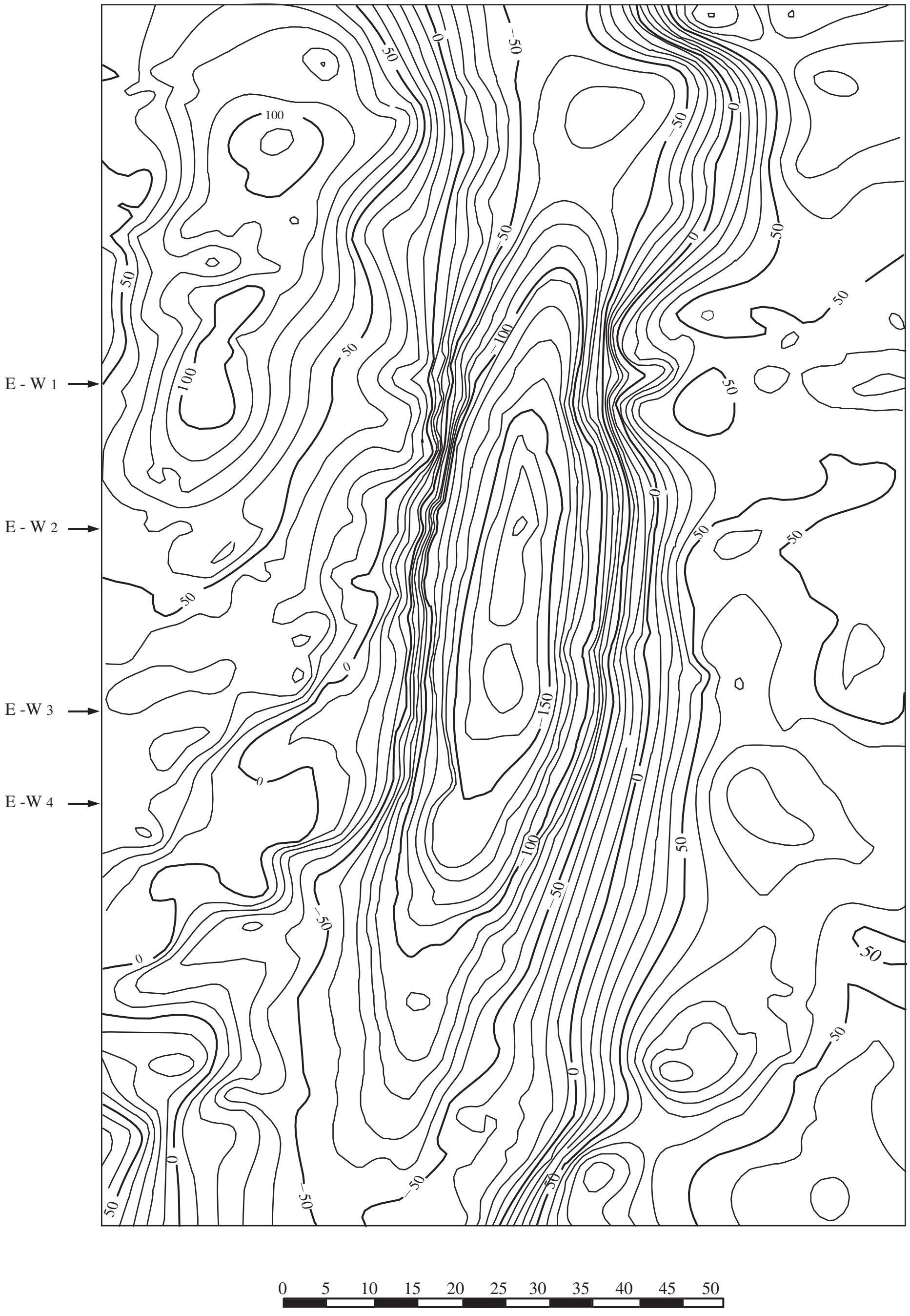


Figure 8.2.5A. Free-Air anomaly map of the Dead Sea.

depression and its flanks (ten Brink et al. 1993), shown in Fig. 8.2.5A. The Bouguer anomaly map was reduced using variable densities, thus eliminating the effects of the rugged topography.

The steep gradients seen on the Bouguer anomaly map, roughly following the −100 mgal contour (Fig. 8.2.5B), define the extent of the Dead Sea–Arava depression. The lowest anomaly on the map shown here (after ten Brink et al. 1993) is −175 mgal, located south of the Lisan Peninsula. The depression so defined ends some 5 km north of the lake. The northern end is marked by a constant gravity gradient to the south, which is probably produced by a slope rather than a normal fault. To the south the depression is divided into at least two parts, the deeper part extending to about 30°55'N, where a change in gradient marks the thinning of the depression fill. The southern end of the depression is at 30°40'N. The depression is about 130 km long and 7–18 km wide.

In order to better delineate possible faulting, the first horizontal derivative of the gravity map was used. This marks the short-wavelength anomalies associated with changes in gradient, namely with faults. The maxima on the first derivative map (Fig. 8.2.5C) shows locations of the north–south trending eastern and western boundary faults. Note that the trends of the maxima do not meet at either end, separated by a zone of minima, probably marking the deepest part of the depression. From the skewness of the first derivative anomalies, ten Brink et al. (1993) deduce that the western boundary fault is mostly normal, while the eastern is vertical or slightly reversed. Based on gravity modeling across the Jordan Valley, ten Brink et al. (1993) conclude that no major intrusion of the upper mantle into the lower crust under the Rift can be supported by the gravity data. They also conclude that the Dead Sea depression is only partially compensated isostatically.

Four two-dimensional east–west models are presented (Fig. 8.2.6A), whose locations are shown in Fig. 8.2.5A. The densities assumed are $1.28 \times 10^3\,\mathrm{kg/m^3}$ for water, $2.15 \times 10^3\,\mathrm{kg/m^3}$ for rift fill, $2.55 \times 10^3\,\mathrm{kg/m^3}$ for pre-Rift sediments and $2.67 \times 10^3\,\mathrm{kg/m^3}$ for basement rocks. As can be readily seen, the models indicate the existence of step faults on the west and of a neat vertical single fault on the east. The calculated thickness of the fill varies from about 5 km in the northern basin to a maximum of 9 north of the Lisan Peninsula. The thickness of the fill is reduced southward to about 4 km on profile EW-4, to the south. The density model is summarized in the north–south profile (Fig. 8.2.6B).

8.2.4 Magnetic studies

The Earth's magnetic field is a useful tool for the study of the basement configuration. Several magnetic surveys were carried out in the Dead Sea and their results shed light on the manner of faulting along the margins of the Dead Sea, its sub-bottom structure and the occurrences of basaltic flows in the basin.

A magnetic survey was conducted over part of the Dead Sea in 1974 (Neev & Hall 1979). In 1983 a detailed marine magnetic survey was carried out in the

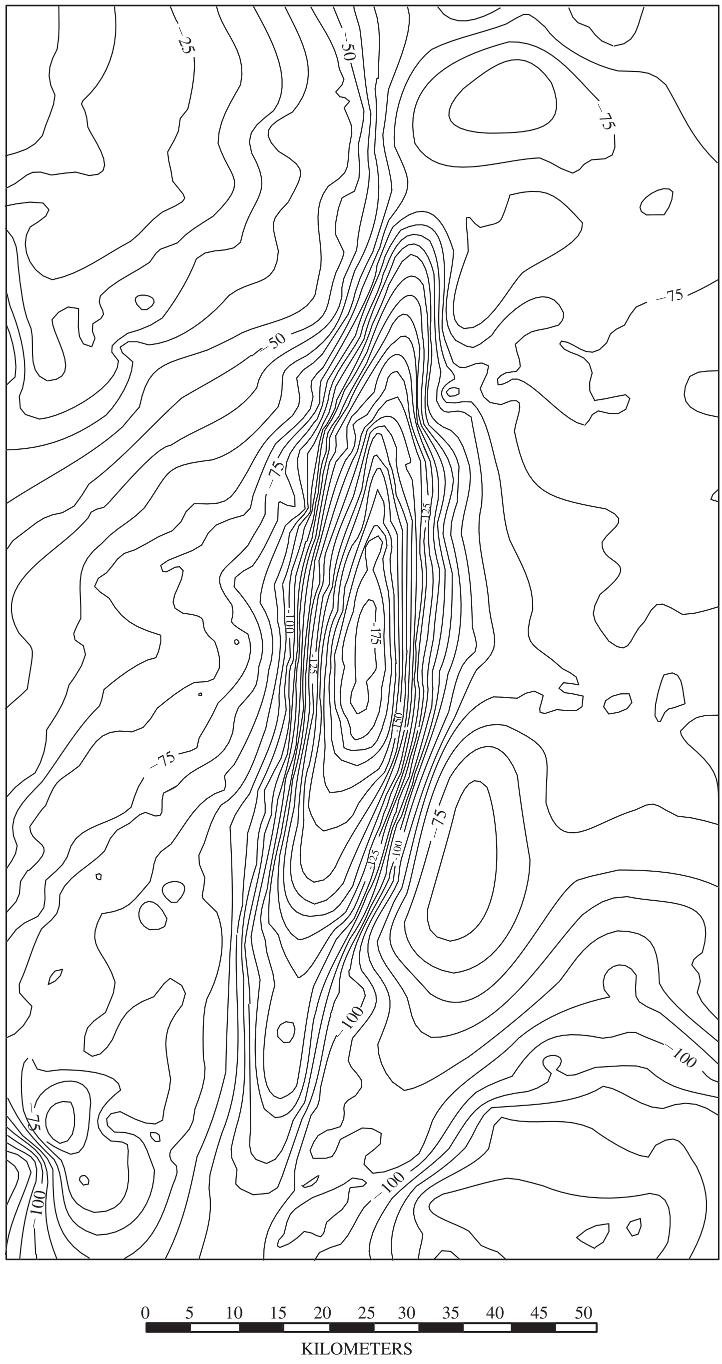


Figure 8.2.5B. Bouguer anomaly map of the Dead Sea.

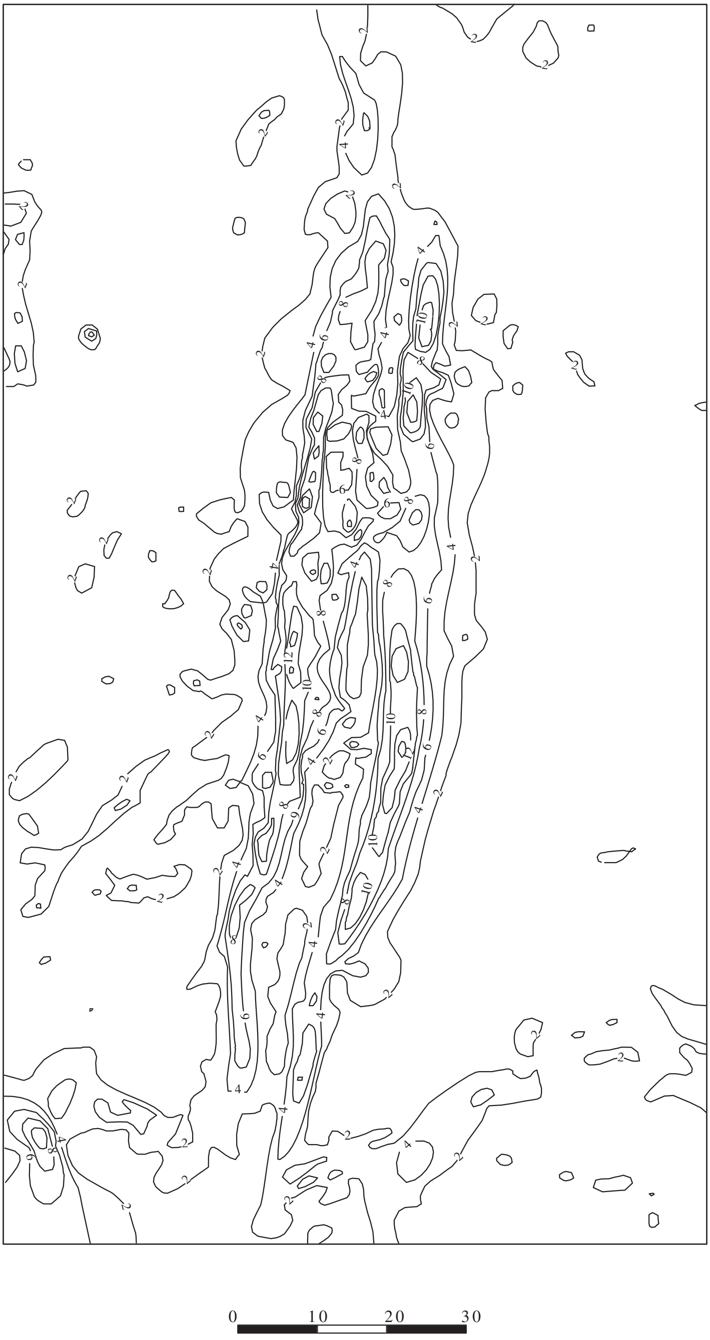


Figure 8.2.5C. Map of the horizontal first derivative of gravity of the Dead Sea. Note the elongated maxima marking the location of maximum gradients (after ten Brink et al. 1993).

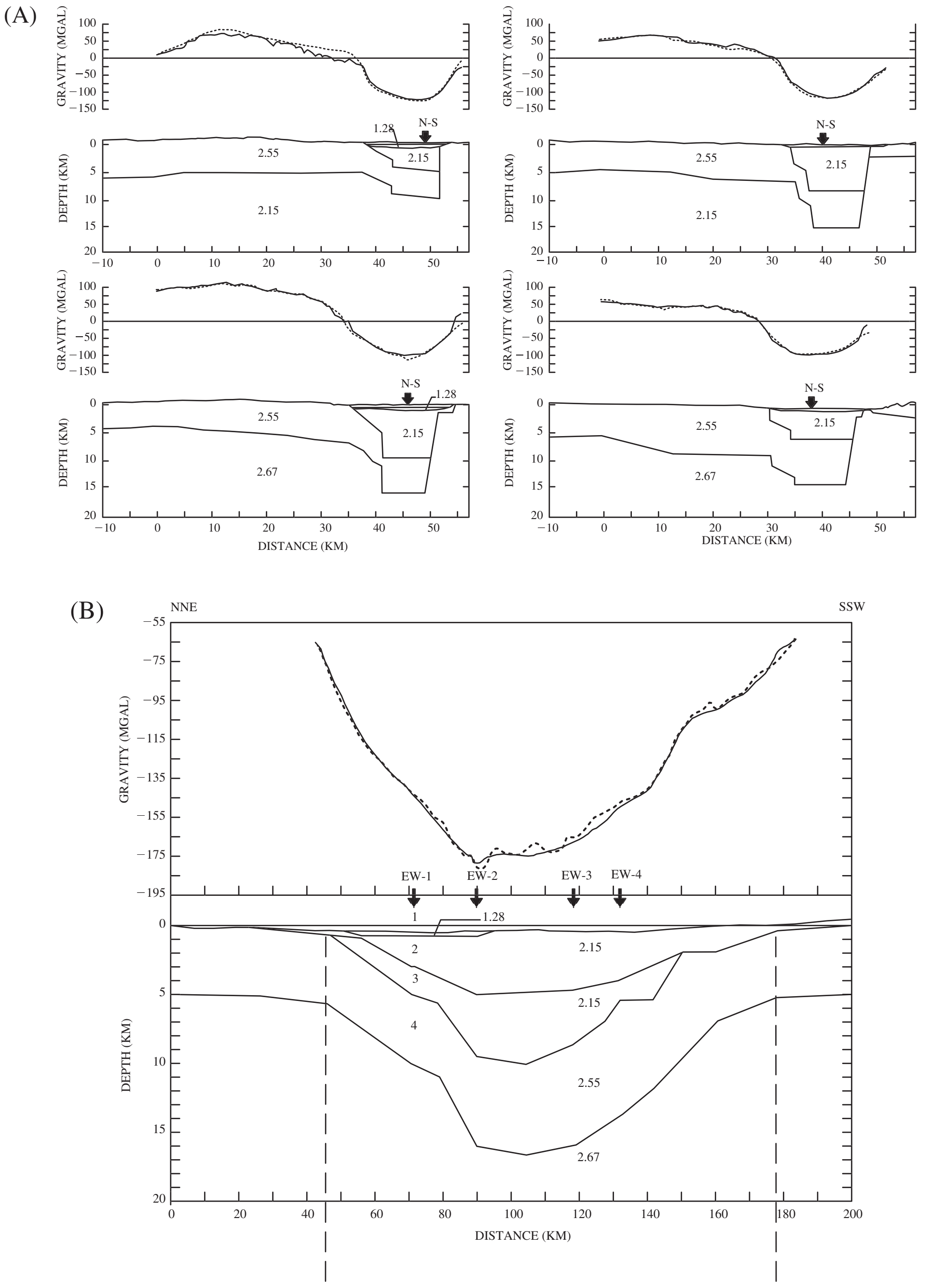


Figure 8.2.6. (A) Four east–west Free-Air gravity profiles (dotted lines) compared with calculated gravity (solid lines); density in 1×10^3 kg/m^3; (B) North–south gravity profile along the axis of the Dead Sea. Calculated and observed gravity values as in A. Arrows indicate the intersections with the east–west profiles.

northern part of the basin (Frieslander & Ben-Avraham 1989). This survey comprised 900 km of track lines, with east–west lines spaced 1 km apart and north–south lines 1.5 km apart. An aeromagnetic survey with a closely spaced grid was conducted in the area west of the Lisan Peninsula, between the northern and southern parts of the basin (Ram 1989), comprising some 1,000 km of profiles. Another aeromagnetic survey was carried out in 1991 by the Israel National Oil Co. Earlier aeromagnetic measurements were made over the land areas west (Folkman 1976, 1981) and east (Hatcher et al. 1981) of the Dead Sea basin. To date no integrated magnetic coverage of the Dead Sea and the adjacent area is available. The maps presented here are merely a compilation.

The magnetic anomaly map of the northern Dead Sea basin is smooth, with a few isolated anomalies (Frieslander & Ben-Avraham 1989). It can be divided into two distinct parts at 31°31'N (Fig. 8.2.7). North of this latitude, the field is smooth and trends north–south and northwest at the northernmost part of the lake. South of latitude 31°31'N, the magnetic field is very different from the field to the north. Most of the smooth north–south contours to the north change their trend to northwest, west–east and northeast, intersecting the shoreline at a high angle. In the central part of the basin the magnetic contours have a similar trend (Ram 1989). Isolated short-wavelength anomalies, some of high amplitude, were mapped mostly along the basin margins. These anomalies indicate the presence of small basaltic bodies buried at shallow depths. Several of these anomalies also exist west of the Lisan Peninsula and in the northern part of the southern basin (Ram 1989). The most pronounced local anomaly (253 nT) is the positive anomaly on the eastern margin of the northern part of the basin, some 2 km south of Wadi Zarqa Ma'in, where young basaltic flows exist. An aeromagnetic survey of this area (Kovach et al. 1990) indicates that the basalt flows onshore are associated with magnetic anomalies of about 3,500 nT. The anomaly at sea is probably associated with the submarine continuation of these flows. An examination of the compiled magnetic map (Fig. 8.2.8) shows that the anomalies continue uninterrupted from the land area to the east into the basin (Frieslander & Ben-Avraham 1989, Ram 1989). This continuity and the gradient into the basin suggest that the faulting along the western margin has been mostly normal. Across the eastern margin it can be seen easily that despite the gap in the data the magnetic contours are discontinuous. Further, the trend and pattern of the anomalies and their wavelength are completely different from those in the basin and land areas to the west. This suggests that major lithological changes occur across the eastern basin margin, attributed to a strike-slip movement, which brought into contact different types of crust (Hatcher et al. 1981). The magnetic data suggest that the southern margin of the northern part of the basin is bounded by a transverse fault. There is no evidence for transverse faulting on the northern margin of the basin. Based on modeling of the magnetic data, Frieslander & Ben-Avraham (1989) estimated that the thickness of the total sedimentary section, including rift fill in the northern basin of the Dead Sea, might reach 10.5 km. From their model (Fig. 8.2.9) it appears that

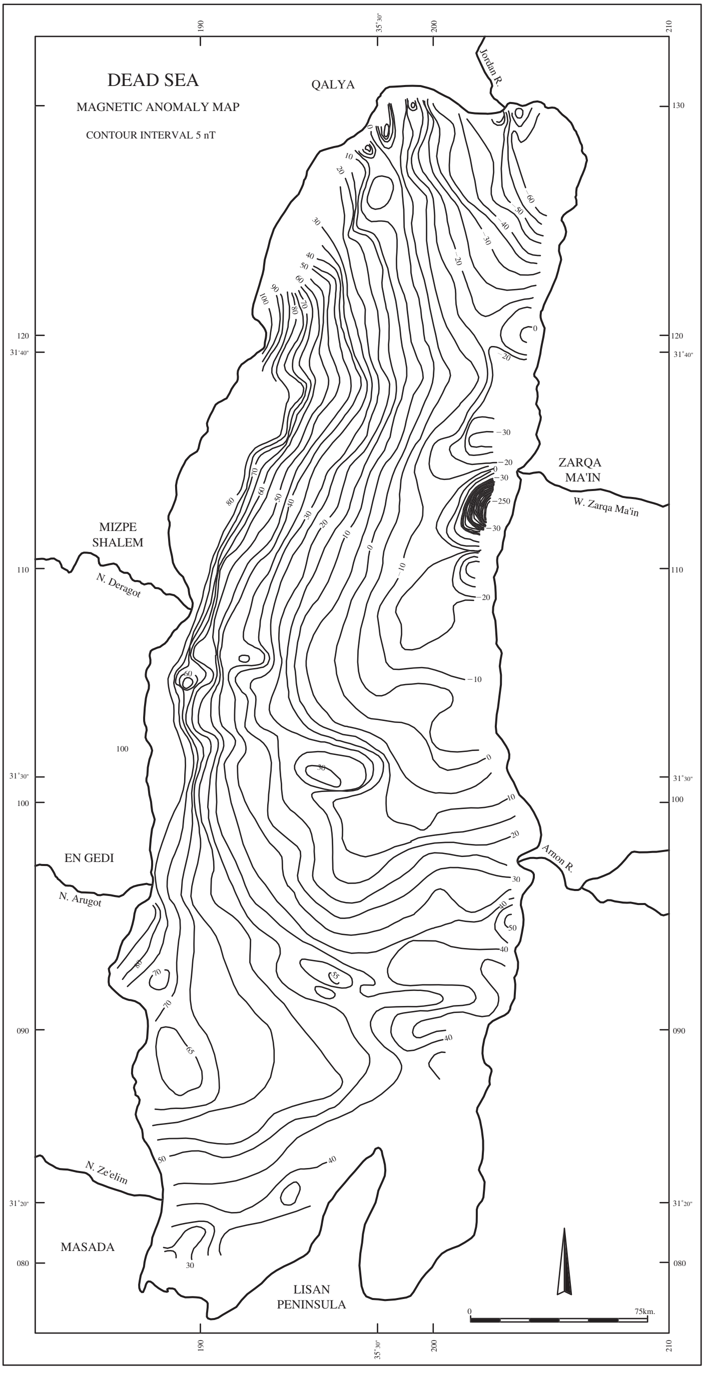


Figure 8.2.7. Magnetic map of the northern basin of the Dead Sea, contour interval 5 nT (after Frieslander & Ben-Avraham 1989).

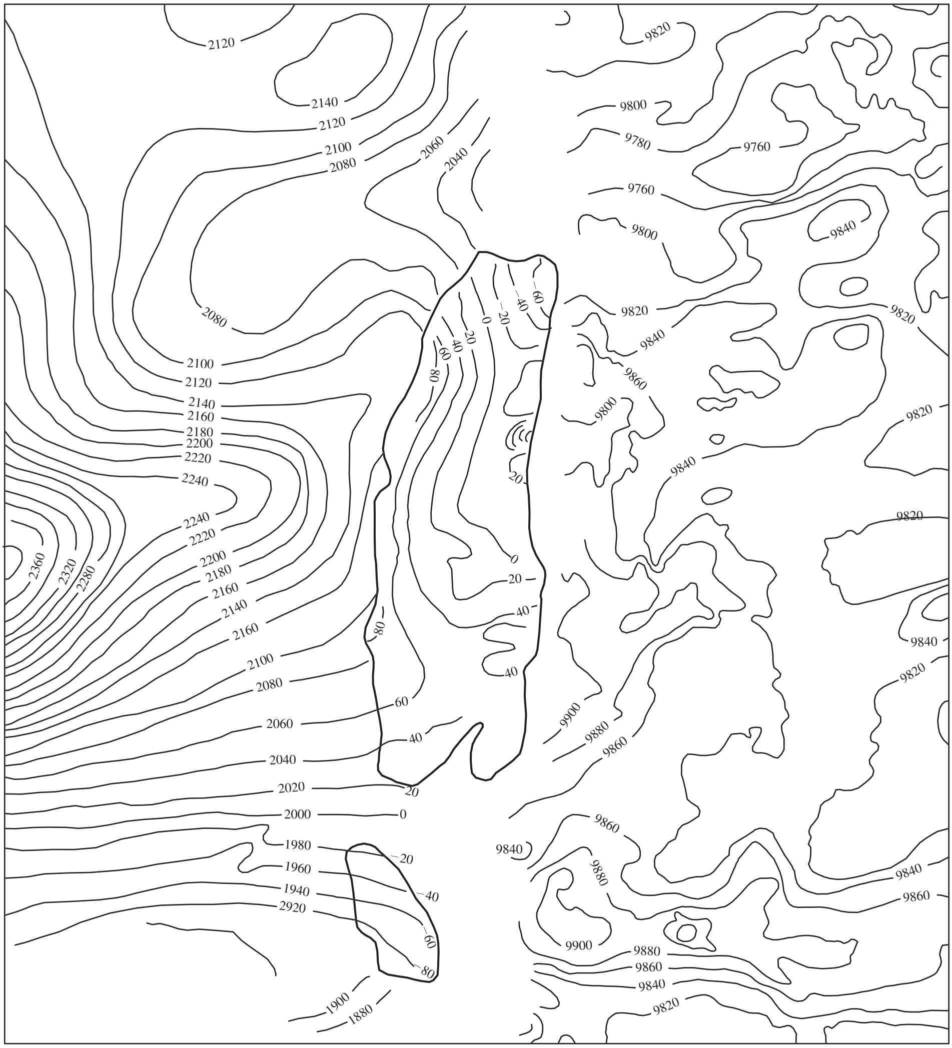


Figure 8.2.8. Magnetic total field anomaly map of the Dead Sea and adjacent areas, based on a compilation of airborne magnetic data from Israel and Jordan. Note the different patterns of the anomalies on either side of the rift (after Ben-Avraham 1997).

the basement underlying the northern part of the basin probably has a similar composition to that in the west, but different from that to the east, thus supporting the suggestion of two different crustal blocks existing across the eastern margin.

8.2.5 Seismic reflection measurements in the northern Dead Sea basin

Seismic reflection coverage in the northern Dead Sea basin is limited to sparker surveys, which were conducted in the lake since 1974. Multichannel seismic

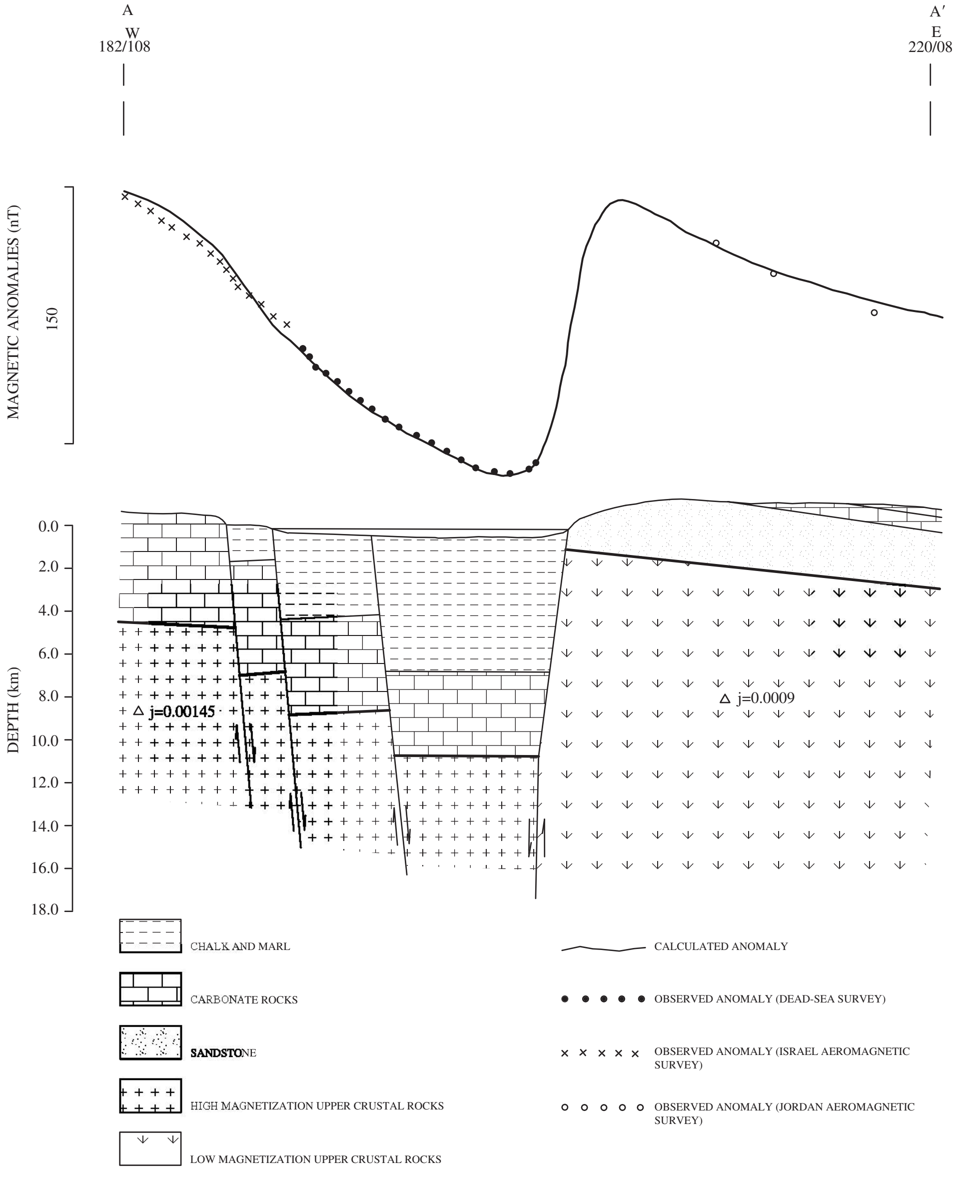


Figure 8.2.9. Calculated magnetic model across the Dead Sea (after Frieslander & Ben-Avraham 1989).

reflection coverage, which was obtained in the course of oil exploration, is limited to a small area near En Gedi.

Neev & Hall (1979) conducted the first reflection survey in the Dead Sea. Penetration of two-thirds of a second two-way time was achieved in this sparker survey, which comprised continuous single-channel recordings. Neev & Hall

(1979) mapped four prominent reflectors over the area. They ascribed the deepest reflector to the Pliocene Sedom Formation, which is composed mainly of salt. Two of the other reflectors were within the Lisan Formation, while the shallowest one presumably represents the beginning of the present lake, at 16–18 Ka ago.

Based on their interpretation of the seismic data, Neev & Hall (1979) suggested that intensive deposition took place in Lake Lisan (the predecessor of the present lake). The deposition was accompanied by intensive erosion of the basin flanks. Due to the shrinkage and eventual complete desiccation of Lake Lisan, salt was precipitated in the deepest part of the basin. This was followed by additional deposition of sediments, resulting from the continuing erosion in the same area. During the following period the Holocene sediments were deposited. Neev & Hall (1979) identified several diapiric structures. It was assumed that the movement of the salt of the Sedom Formation formed these structures, which are mainly concentrated along the western margin of the lake. Of these structures, Neev & Hall (1979) named two large diapirs, which have penetrated the lake bottom, the En Gedi diapir and the Jordan River delta diapir.

The sparker data provided important information about active faulting in the northern basin. Neev & Hall (1979) mapped a major north–south trending linear border fault. In addition they interpreted secondary faults, which branch off to the north–northeast. A set of step faults was mapped along the western margin. Neev & Hall also suggested that two west to northwest trending faults extend north and south of the Lisan Peninsula. Ben-Avraham et al. (1993) and Niemi & Ben-Avraham (1997) concluded that the easternmost longitudinal fault of the western margin, which they refer to as the "western intrabasinal fault", is composed of three segments with slightly different trends (Fig. 8.2.10). Kashai & Croker (1987) report a multichannel reflection profile, some 10 km north of the Dead Sea, which shows a flower structure indicating strike-slip faulting. Rotstein et al. (1991), in their study of the same area, also noted compressional features displayed by this fault. Ben-Avraham (1997) considers this fault to be the onshore extension of the western intrabasinal fault. No fault bounding the basin was found at its northern end.

A large amount of high-resolution 3.5 kHz data were collected during geophysical surveys in 1983 and 1984 (Ben-Avraham et al. 1985, Hall & Ben-Avraham 1985) when some 1,500 km of profiles were recorded. In 1993, 200 km of additional high-resolution profiles were acquired. Ben-Avraham et al. (1993) and Niemi & Ben-Avraham (1997) reviewed the 3.5 kHz data set. The penetration of the high-resolution sparker surveys is limited to the uppermost 30 msec two-way time below the lake bottom. However, these profiles yielded much useful information about active tectonism in the lake. In their analysis of the 3.5 kHz data Ben-Avraham et al. (1993) concluded that the shallow uppermost sediments which were deposited in the northern part of the basin are continuous and flat lying. They correlated the four prominent reflectors they have mapped, with alternations of marl and rocksalt. This correlation was verified in the course of a coring program, recently undertaken in the northern Dead Sea basin (Ben-Avraham et al.

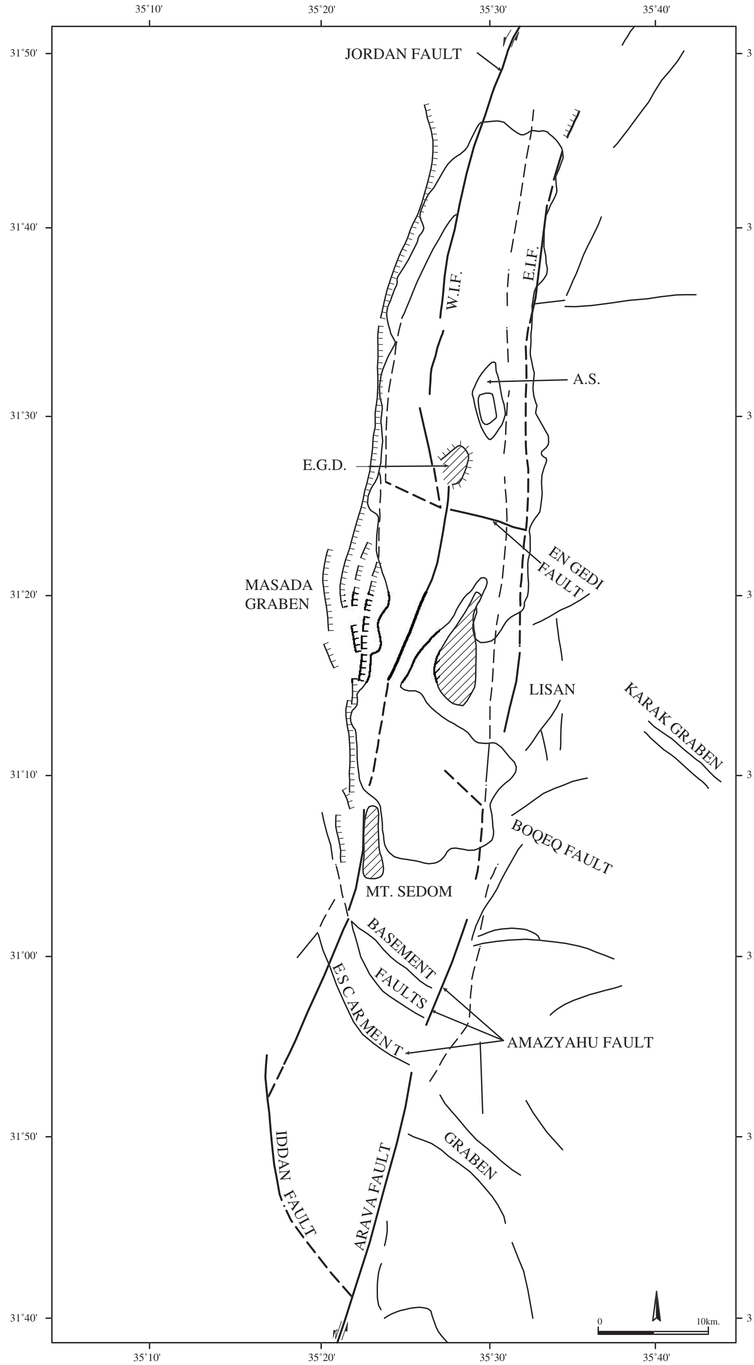


Figure 8.2.10. Structural interpretation map of the Dead Sea. Solid lines mark faults with extrapolations shown by dashed lines. The heavy lines mark the main boundary faults of the basin. Hatched areas show the locations of the Mount Sedom, Lisan and En Gedi (E.G.D.) diapirs. W.I.F – western intrabasinal fault; E.I.F. – eastern intrabasinal fault; A.S. – Arnon sink.

1999). In their survey Ben-Avraham et al. (1993) re-mapped the Arnon sink, first reported by Neev & Hall (1979), which is located in the north central part of the basin. The sink shows the asymmetry of the present subsidence. The maximum subsidence occurs above the sink, which is a rhomb-shaped depression located closer to the eastern shore of the lake, in proximity to the Lisan and En Gedi diapirs (Fig. 8.2.10). Ben-Avraham et al. (1993) suggested that the trend of the depocenter indicates tectonic control of the subsidence and deposition. Further, the proximity to the diapirs may point to subsidence controlled by salt movement.

8.2.6 Reflection data in the southern Dead Sea basin

Numerous multichannel seismic surveys were conducted in the southern sub-aerial and land parts of the southern Dead Sea basin during oil exploration activities. A number of deep holes were drilled in the area, thus affording correlation of the reflection data with the geology. As is well known, reflection seismology is by far the best tool for investigating the subsurface, as it has the required resolution and yields a graphical output close in appearance to a geological cross section. Over the years several articles based on several generations of reflection surveys have been published. Studies based on seismic reflection interpretation include Arbenz (1984), Manspeizer (1985), Kashai & Croker (1987), Ben-Avraham & ten Brink (1989), Ben-Avraham (1997) and Gardosh et al. (1997).

The quality of the reflection data is such that good information is obtained from the Rift fill and, in places, its base. The information from deeper below the surface is partial and discontinuous and our knowledge of the pre-Rift structure is incomplete. The seismic sections give a good idea of the structure and type of faulting along the western margin of the southern basin. The western boundary fault is illustrated in Fig. 8.2.11. The step faults are clearly seen, as well as the base of the fill. The Sedom Deep 1 well reached a depth of 6,445 m (about 3.9 msec TWT) and bottomed in middle Oligocene sediments, still part of the fill. The prominent reflections under the well site represent the salt sequence. A number of salt pillows were mapped in the area (Gardosh et al. 1997). On the eastern side of the basin, multichannel reflection data obtained for oil exploration show that the boundary fault is a steeply dipping normal fault, similar to the faults observed on the western side. According to ten Brink & Ben-Avraham (1989) the deformation in the southern basin is along longitudinal and transverse faults, while the sediments between these faults are largely undisturbed.

The transverse faults in this area listed from north to south, are: the En Gedi fault (in the northern part of the basin), the Boqeq fault, the Amazyahu fault and the Iddan fault (Fig. 8.2.10). Of these, the most prominent is the Amazyahu fault, which has a surface expression in an escarpment (Fig. 3.1.7), as well as a prominent expression on the seismic cross section. It has been interpreted by Kashai & Croker (1987), ten Brink & Ben Avraham (1989) and Gardosh et al. (1997) as a normal, down to the north listric fault, which flattens at depth

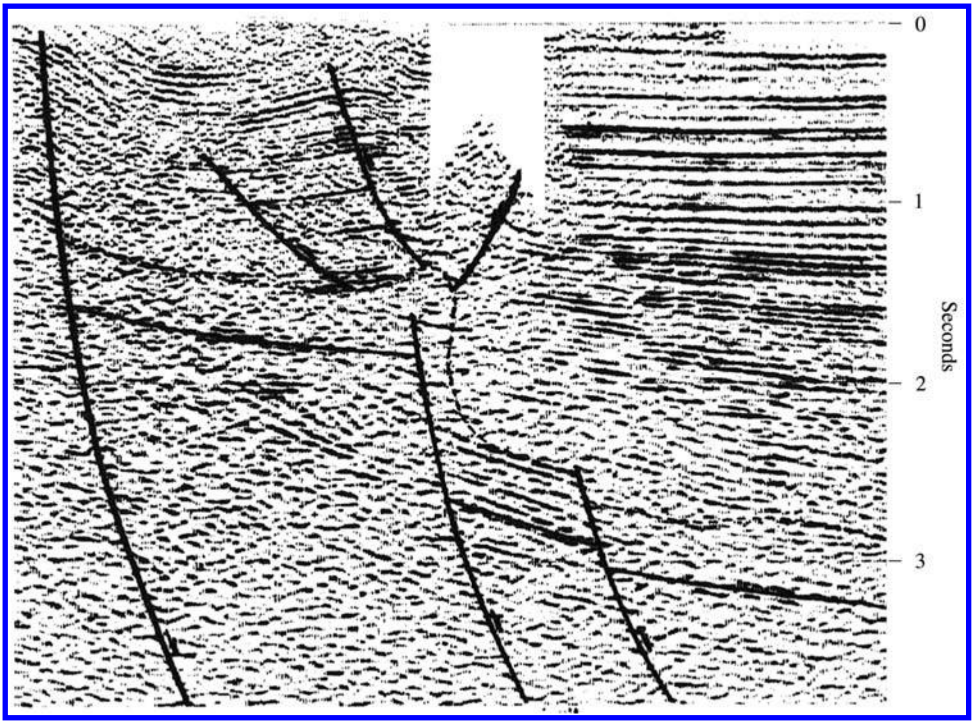


Figure 8.2.11. Interpretation of a seismic reflection-time section obtained south of Mount Sedom. The down-to-the-east step faults of the western flank of the Dead Sea are clearly seen. The dashed line shows the top of a salt layer encountered in a nearby well (after Gardosh et al. 1997).

against the salt layer, with some salt emplaced underneath it. Gardosh et al. (1997) also interpret a transverse ridge in the pre-Rift section on the upthrown block. Two interpretations of the Amazyahu fault are shown in Figs 8.2.12 and 8.2.13. The interpretation of the fault as listric was the basis of a theory by Arbenz (1984), according to which the strike-slip was formed by north–south tension which caused slices limited by listric faults to slide down. In this connection it should be noted that the only major listric fault which was interpreted by oil company geophysicists in the Dead Sea was the Amazyahu fault. It should also be mentioned that an alternative interpretation of the Amazyahu fault is possible. A set of down to the north normal step faults, covered by a salt intrusion, can be seen on some of the old migrated time sections. Modern pre-stack depth imaging shows clearly that this is the more likely structure of this fault (Fig. 8.2.13).

The Boqeq fault (Ben-Avraham et al. 1990a) is a down to the south normal fault, separating the southern part of the basin from its northern part. The Iddan fault, marking the southern end of the depression, is a down to the north normal fault, with a possible salt body against it. The sparker data from the northern basin and multichannel data from the southern basin indicate that the Dead Sea is located in a true two-sided rift. The seismic data indicate that the deposition of the rift fill was rapid, took place during the Pleistocene and progressed together with the faulting (see the thickening of the sediments against the Amazyahu fault in Fig. 8.2.13).

South of the Iddan fault the only seismic data available are a series of east–west high-resolution reflection lines. Varying styles of faulting along the western margin of the Arava are reported. At least one of these lines shows a flower-structure type of fault pattern, a feature associated with strike-slip faulting (U. Frieslander, Geophysical Institute of Israel, Holon 1999, pers. comm.).

8.2.7 Seismicity

The Dead Sea is the most seismically active basin in the Jordan Valley. Documented evidence of strong earthquake occurrences in the area date back to Biblical times. Catalogues of earthquakes which occurred in the past 4,000 years have been published in numerous papers, including Amiran (1951), Arieh (1967), Ben-Menahem et al. (1976), Ben-Menahem (1979), Shapira (1979), Ambraseys & Melville (1988), Turcotte & Arieh (1988), Ben-Menahem (1991) and Ben-Menahem & Aboodi (1981). The epicenters of historic earthquakes cannot be located accurately on the basis of macro-seismic observations. However several strong and particularly destructive earthquakes were located at or near the Dead Sea. These include the earthquakes of 1060, 1293, 1458, May 23 1834 (intensity X on the modified Mercali scale and magnitude 5.8) and July 11 1927 (intensity IX, magnitude 6.2). The recalculated location of the 1927 earthquake is in the Dead Sea (Shapira et al. 1993). Niemi & Ben-Avraham (1994) show, by their analysis of the 3.5 KHz and sparker seismic reflection sections, sediment slumping in the northern Dead Sea basin, suggesting that the 1927 earthquake triggered the slumping.

During the period 1903–1981, seismological information was obtained from macro-seismological observations and instrumental data, from observatories at Ksarah (Lebanon), Helwan (Egypt), Jerusalem and Elat, as well as occasional recordings of strong earthquakes in Europe. Since 1982 the Israel Seismic Network has been in operation (Shapira 1982). The activation of the Transjordanian stations in 1986 has improved the accuracy of epicenter location.

Shapira et al. (1986) have demonstrated that despite the large concentration of seismic stations in the Dead Sea area, focal depth determinations are not sufficiently accurate. This inaccuracy leads to errors in epicenter location, which in turn precludes the association of a particular event to a known fault. However, despite the poor constraints on focal depths, van Eck & Hofstetter (1990) concluded that in the Dead Sea depression earthquakes originate at depths not exceeding 12–15 km. This suggests that the deformation in this depression takes place within the brittle upper crust. Composite fault plane solutions of earthquakes in and around the Dead Sea indicate a left-lateral strike-slip mechanism in the majority of the cases examined (van Eck & Hofstetter 1989, Shapira 1997). Individual solutions, however, show both normal and reverse components of slip (Fig. 8.2.14). Two

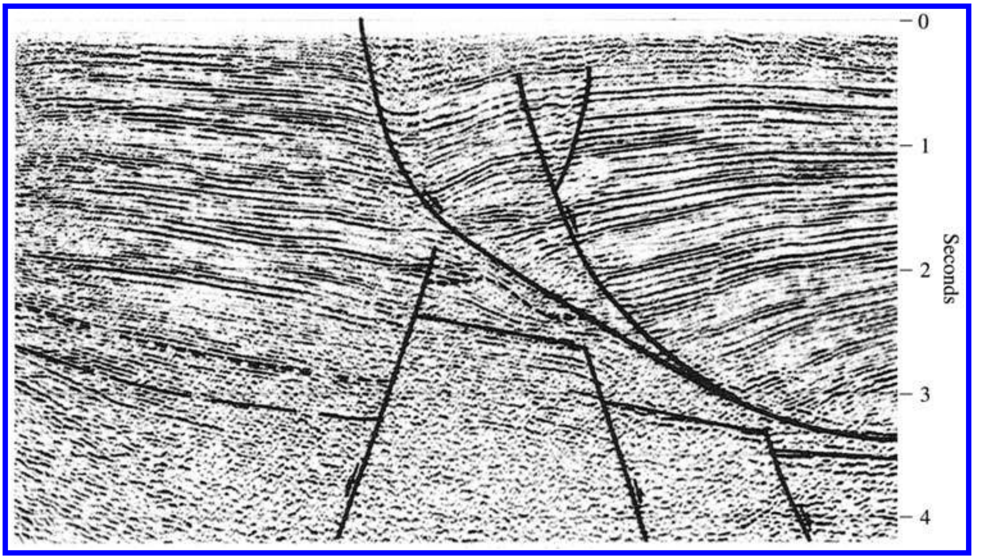


Figure 8.2.12. A north–south seismic reflection-time section crossing the Amazyahu fault south of the Dead Sea. The Amazyahu fault is interpreted as a listric fault (after Gardosh et al. 1997).

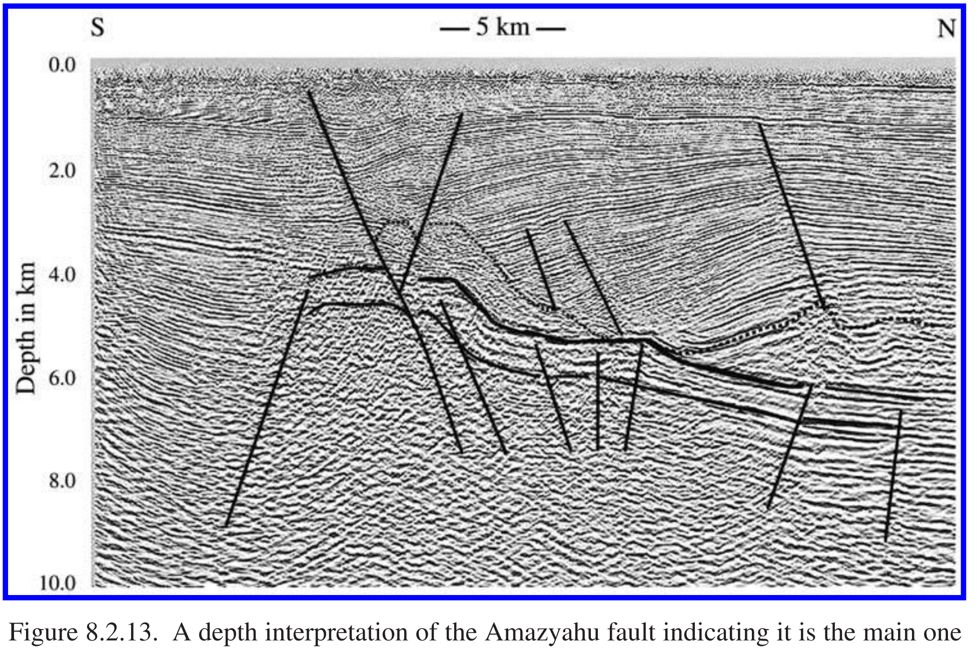


Figure 8.2.13. A depth interpretation of the Amazyahu fault indicating it is the main one of a down-to-the-south series of step faults.

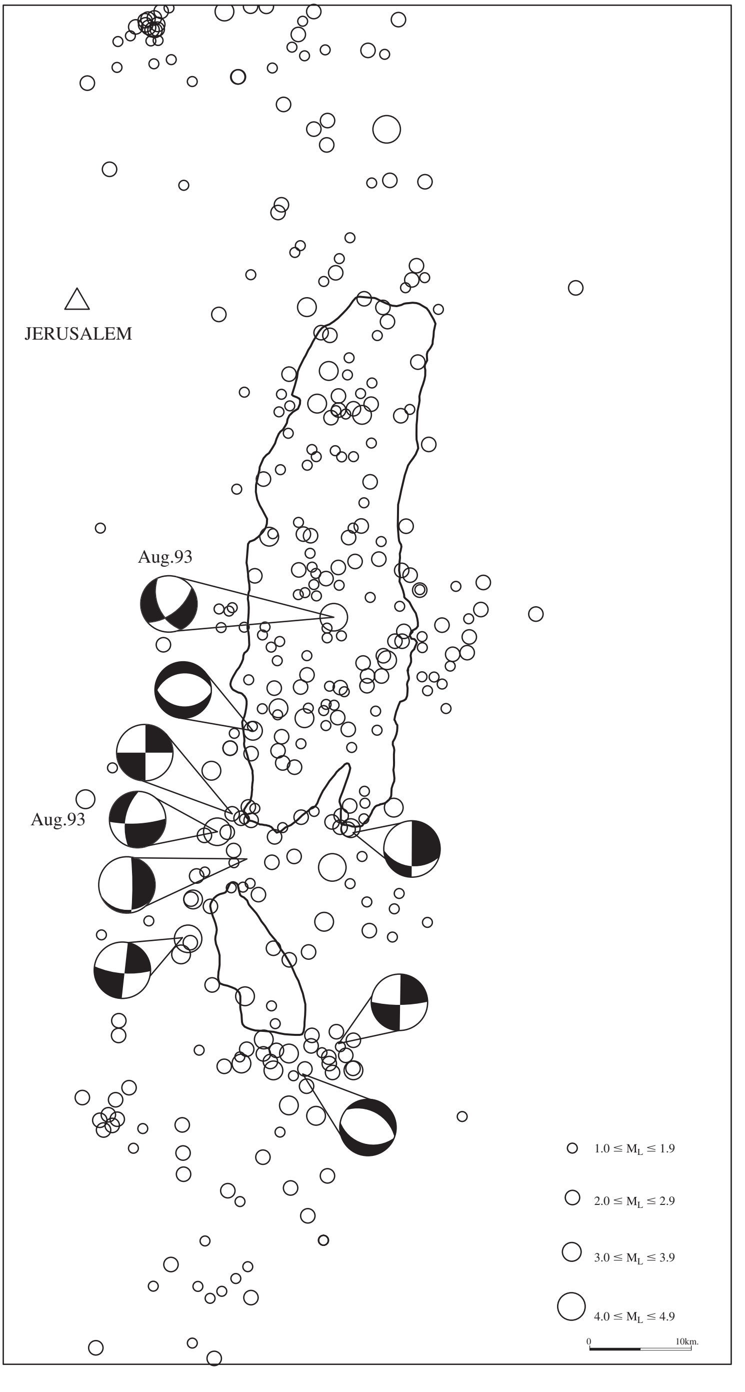


Figure 8.2.14. Location map of earthquake epicenters in the Dead Sea, with selected fault-plane solutions (after Shapira 1997).

earthquakes which occurred on August 3 1993, with magnitudes of 4.1, show two types of slip. The focal plane solution of the northern earthquake shows a northwest– southeast direction of slip with a left lateral strike-slip and a normal component. In the southern earthquake a classical solution of strike-slip with a north–south strike was obtained.

An examination of the aerial distribution of epicenter locations indicates possible clustering in a number of distinct areas, around the fault which separates the Lisan Peninsula from the northern basin. Other areas are around the Arnon sink, where active subsidence takes place, and south of the Jordan delta.

8.2.8 Heat flow in the Dead Sea basin

Geothermal measurements were made in the northern Dead Sea basin in 1975 (Ben-Avraham et al. 1978). Onland measurements of temperature gradients were made in wells west of the Dead Sea (Eckstein & Simmons 1978) and east of it (Galanis et al. 1986).

Nineteen measurements were taken in the lake. At the time of measurement, before the 1979 overturn of the meromictic regime, water temperatures were found to be very stable, agreeing within less than 0.01°C at locations as far as 40 km apart. Ben-Avraham et al. (1978) found an adiabatic temperature distribution from the lake bottom (335 m) to a depth of 185 m. Lake bottom water temperature variations were small. Neev & Emery (1967) reported nearly uniform temperatures, both with depth and season. This unique situation was ideal for obtaining reliable heat-flow measurements in the northern Dead Sea basin at depths exceeding 100 m.

The mean value of the corrected heat-flow data for the northern Dead Sea basin is 38 mW/m^2. This value is comparable with figures obtained in wells west of the lake (Eckstein 1975, Eckstein & Simmons 1978) and with heat-flow values (Galanis et al. 1986) east of the lake. A numerical model of heat-flow distribution from the Dead Sea and surroundings (Mass 1978), indicates that the actual heat flow in this area is probably 63 mW/m^2. Most of the heat escapes through the diapirs, composed of conductive salt, which are covered by an insulating layer.

8.2.9 Conclusion

Based on the interpretation of the geophysical measurements in the Dead Sea basin presented above, one can arrive at a composite structural model. Several models have been proposed, such as those of Neev & Hall (1979), Kashai & Croker (1987), ten Brink & Ben-Avraham (1989) and Ben-Avraham et al. (1993). The model presented here is that of Ben-Avraham (1997). The Dead Sea basin is a depression bounded by faults on the east, west and south (Fig. 8.2.10). Deformation in the basin takes place mainly along the longitudinal and transverse faults, while the sediments in the intervening segments are relatively unaffected. From seismological and heat-flow data it can be concluded that deformation takes place in the upper crust at a depth of 12–15 km. No evidence of substantial invasion of upper mantle material into the lower crust underneath the basin has been indicated.

The longitudinal faults show strike-slip movement occurring on some segments, and normal on others. It appears that the main strike-slip motion in the northern basin takes place at present on the western intrabasinal fault (the Jordan fault). In the southern part the main motion takes place along the eastern border (Arava) fault (ten Brink & Ben-Avraham 1989). Earthquake solutions (van Eck & Hofstetter 1989) suggest that some overlap between the two faults may occur near the Lisan Peninsula, where the basin is deepest.

The Dead Sea basin is asymmetrical to the east, the asymmetry being expressed in the topography, bathymetry and magnetic field, in which the anomalies west of the depression continue smoothly over the lake to its eastern margin. The continuity of the magnetic anomaly across the western margin indicates that the lateral slip there is minimal. The western margin is characterized by a set of normal step faults, extending into the lake. The presence of tilted blocks (Kashai & Croker 1987, ten Brink & Ben-Avraham 1989, ten Brink et al. 1993) indicates that normal faulting was dominant along the entire western margin of the Dead Sea basin. From seismic reflection data and drill holes it is evident that the rapid subsidence and accumulation of over 6 km of sediments took place during the late Cenozoic (see Chapter 6). According to Garfunkel (1981) and Joffe & Garfunkel (1987) the rapid subsidence is associated with the slight change in the relative motion of Africa and Arabia that occurred at the beginning of the Pliocene. This change caused extension and consequently subsidence in the Dead Sea. It is possible that a shear component was introduced at the time, which created the western intrabasinal fault as a strike-slip fault.

The depression is divided by transverse faults into a number of distinct segments, in which different styles of faulting can be seen. The nature of the transverse faults is unclear. This lack of clarity is illustrated by the examination of two major transverse faults, the Amazyahu and Boqeq faults. The most spectacular transverse fault is the Amazyahu, which appeared on early record sections as a huge listric fault. It has been considered as such in all the subsequent interpretations which have been published. A careful examination of some of the time migrated sections which are available and depth imaging, currently carried out by the authors, indicates that the Amazyahu fault may be in fact a zone of normal step faults. It was suggested by Ben-Avraham et al. (1990a) that the Boqeq fault, which separates the Lisan Peninsula from the southern basin, was a normal fault during Miocene–early Pleistocene times, but changed its faulting motion to strike-slip during the late Pleistocene. The examination of the depth imaging obtained by pre-stack depth migration of seismic sections suggests that the Boqeq fault is predominantly a normal fault.

8.3 THE DEAD SEA–BET SHE'AN SEGMENT

The geophysical data available for this part of the Jordan Valley are rather limited, since it did not attract much interest for oil exploration. The available data thus consist of regional gravity and magnetic coverage and some multichannel seismic sections in the Jericho area and south of Bet She'an, most of which are not in the public domain.

8.3.1 Topography

This part of the Jordan Valley lies between the Dead Sea depression to the south and Lake Kinneret to the north. Topographically, the asymmetry between the Israeli and Transjordanian sides is maintained, the eastern area being higher (Fig. 1.3A). At its southern end, in the vicinity of Jericho, the Valley is about 18 km wide. The width decreases gradually northward, reaching only 8 km some 15 km south of Bet She'an. From that point northward the valley broadens and joins the Yizre'el Valley to the west and the elongated Lake Kinneret–Bet She'an depression. The slopes of the valley sides are variable. From Jericho northward, over a distance of 45 km, the slopes on either side of the valley are much gentler than along the Dead Sea. Some steepening of the slopes is noted to the west, caused by a monoclinal fold dipping into the valley. Picard (1987) considered the entire Jordan Rift to comprise subgrabens, separated by pivotal uplifted area such as this part of the valley. Picard commented that the subgrabens are asymmetrical, with the sense of asymmetry changing from east to west and vice versa along the Rift.

8.3.2 Crustal structure

As has been mentioned before, evidence from long-range refraction experiments in Israel and Transjordan (Ginzburg et al. 1979a,b, El-Isa et al. 1987a) indicates the existence of a different crust on either side of the valley. The differences are both in velocity distribution and depth. The crust is about 25 km thick under the Jordan Valley, while a thickness of up to 40 km east of the valley is reported by El-Isa et al. (1987a). Ginzburg & Folkman (1980) report a thinner crust, 25 km and less, under central and northern Israel. Ben-Avraham & Ginzburg (1990) suggested that the thinner crust under central Israel, with further thinning under the Galilee, are terrains which were accreted in Paleozoic times.

A multichannel seismic reflection line shot in the course of oil exploration some 10 km north of the Dead Sea, just south of Jericho, is of particular interest. Kashai & Croker (1987) published this line. Their interpretation shows the western boundary fault and the block faulted Jurassic sequence stepping down to the east, overlain by the Cretaceous section. A thin layer of fill, not exceeding 600 m in thickness, unconformably rests on the overlying Tertiary rocks. At the eastern part of this profile the entire sequence is affected by a set of faults. This is a flower structure, which indicates strike-slip faulting. This fault is thought to be the extension of the western intrabasinal fault (Rotstein et al. 1991, Ben-Avraham et al. 1993).

8.3.3 Structural interpretation of the gravity data

Gravity anomalies are the main source of structural interpretation in this segment of the Jordan Valley. The integrated Bouguer gravity anomaly map (Fig. 8.2.1) published by ten Brink et al. (1999) shows a series of elongated low amplitude local gravity minima at the center of the valley, aligned north–south. The gravity gradients bounding the valley anomaly are in most cases not very steep, but well defined. The gradients to the east are mostly fairly diffused. This has partly to do with the regional gravity trend, caused by the deepening of the Moho from west to east (Fig. 8.3.2). South of Bet She'an a trend of gravity lows is disrupted by a very distinct saddle, located in the narrowest part of the valley some 20 km south of Bet She'an, which marks the southern border of the Lake Kinneret–Bet She'an depression.

A regional east–west running gravity profile through Azraq–Amman–Tel Aviv was analyzed by ten Brink et al. (1990). As can be seen the model (Fig. 8.3.1), which is a compilation of refraction data with velocities converted to densities, includes a major step in Moho depth underneath the Jordan Valley. The calculated and observed gravity anomalies (Bouguer for the land areas and Free Air for the Mediterranean and areas below sea level) show a good fit for this model, which includes a thin layer of fill and does not require a rift. ten Brink et al. (1990) concluded that there is no upper mantle material intrusion under the Dead Sea strike-slip, and that the valley uplifted margins are not isostatically compensated. They also remarked that the change in crustal thickness may support the accepted view of a 105 km left lateral offset along the strike-slip, since a right lateral shift of 105 km would juxtapose similar crustal thickness across the strike-slip.

Three gravity profiles were computed across the Jordan Valley, at the locations shown in Fig. 8.3.2. It can be seen on the model calculated for profile A-A that the fill is only 4 km thick. This is a minimum value since the fill includes an unknown thickness of basalt, which would reduce the amplitude of the anomaly. This means that it would take a thicker layer of fill to satisfy the observed anomaly than the thickness shown on the model. Profile B-B shows that the fill here is not thick and that the flanks of the valley are not faulted. Profile C-C shows a monoclinal fold dipping into the valley on the western margin and probably just dipping layers on the east as well. The magnetic map compiled by Folkman & Yuval (1976) is important for the study of the subsurface structure in the northern part of this segment. The short-wavelength anomalies evident on this map are indicative of the distribution of basalt in the valley and its margins.

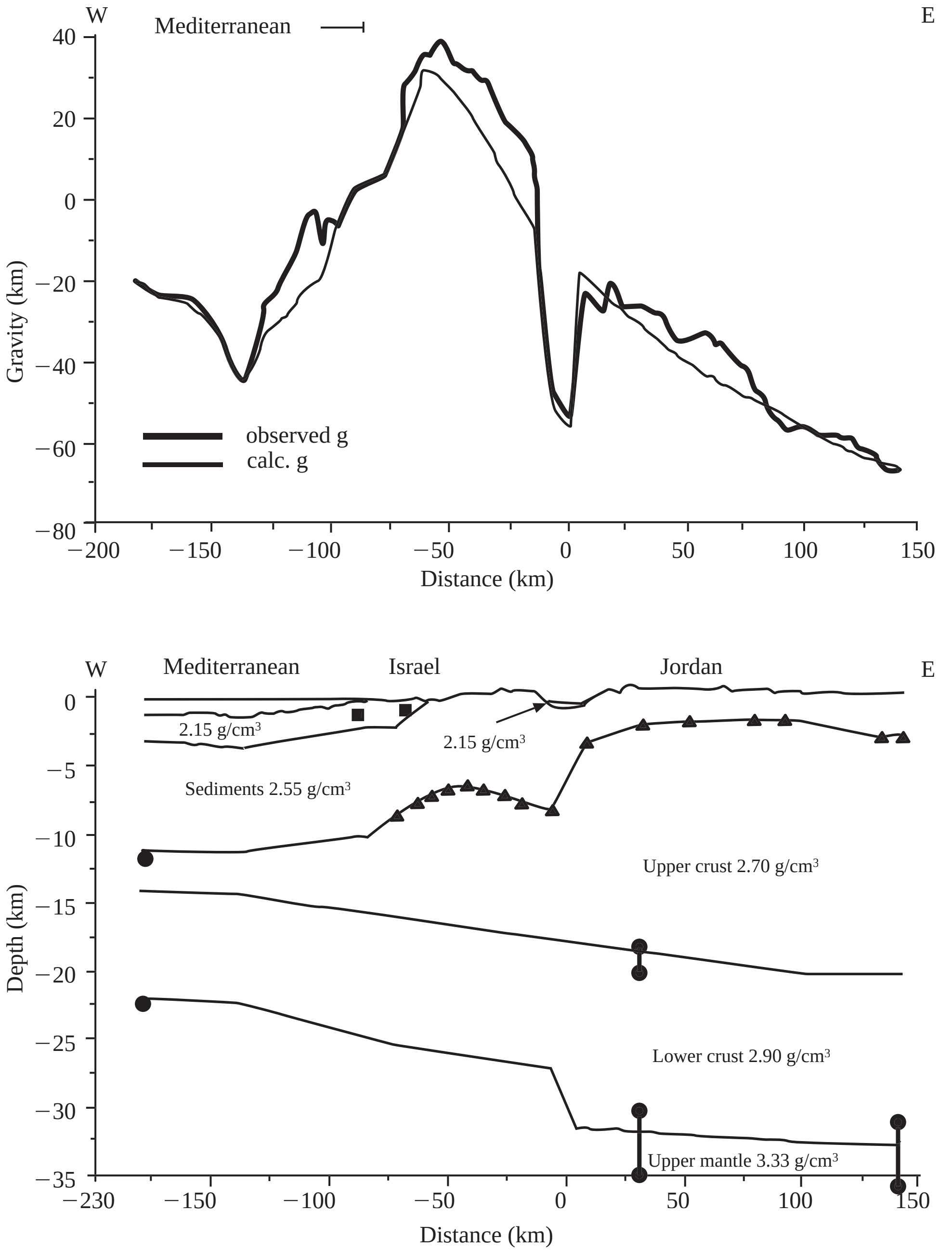


Figure 8.3.1. An east–west gravity model across Transjordan and Israel calculated using structure based on refraction data. Note the difference in crustal structure between Transjordan and Israel, the step in Moho depth and the absence of a rift (after ten Brink et al. 1990).

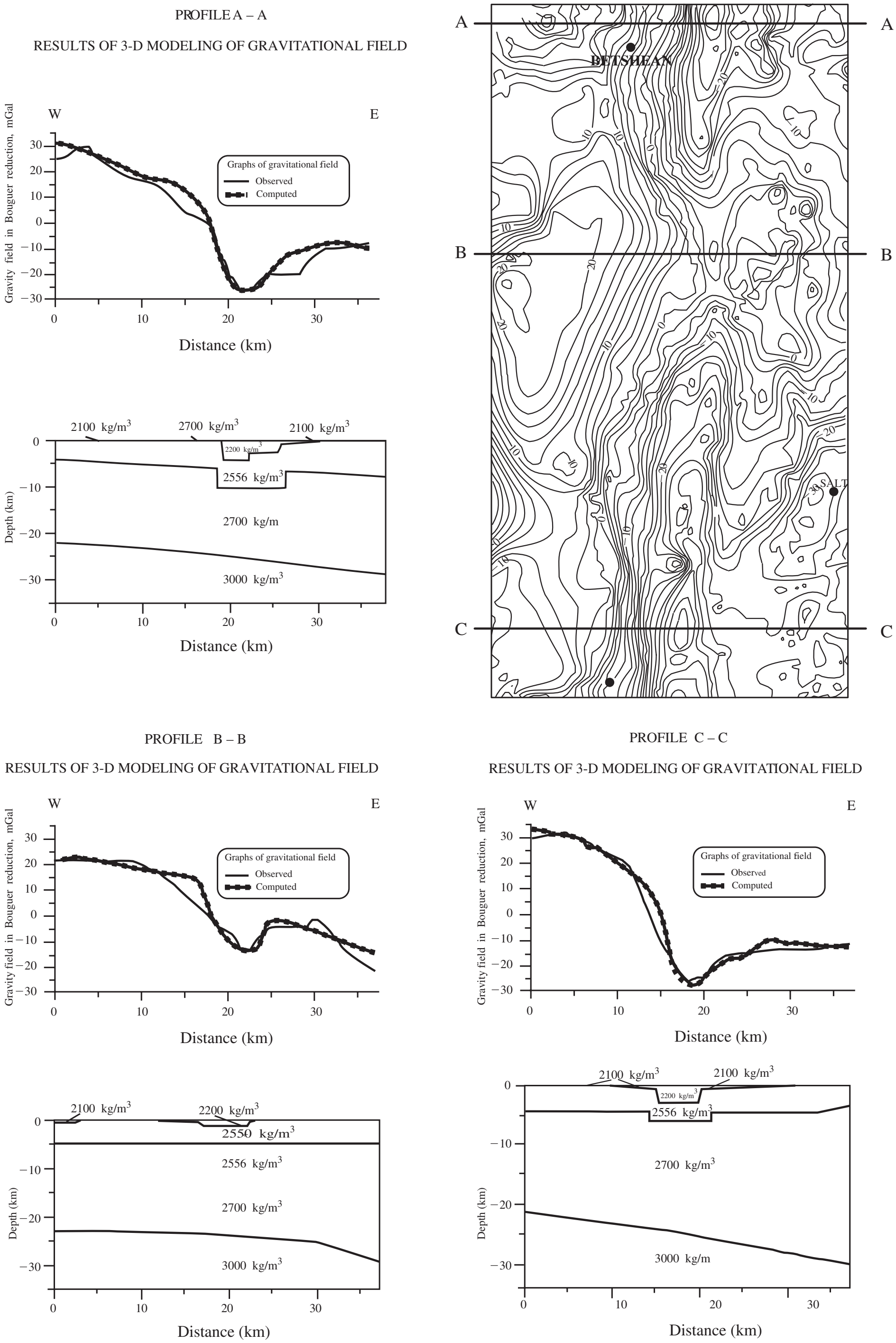


Figure 8.3.2. Three east–west gravity profiles calculated across the Jordan Valley. Location of the lines is shown on the Bouguer anomaly map. Note the absence of a rift on profile B-B.

8.3.4 Seismicity

The distribution of earthquake epicenter locations for this segment of the Jordan Valley is interesting. The epicenter location map (Fig. 8.3.3) shows clearly that only a small part of the seismic activity in this segment takes place in the valley. The main part of the seismic activity actually occurs along the faults, which branch off to the west.

van Eck & Hofstetter (1989) analyzed earthquakes which occurred from 1982 to 1989, as shown in Fig. 8.3.3. The focal plane solutions for selected earthquakes show a good deal of change. The solution for 3b points to a left lateral strike-slip motion with a reverse component, while 3c shows dip-slip motion with a strong dip to the east. As has been noted by Shapira (1997), it is difficult to correlate a given earthquake with a known fault. The results of the analysis do point however to the complexity of the tectonic activity in this area and to the fact that no single faulting mechanism is responsible for the formation of the valley. The quiet zones indicate that there may not be a continuity of active faults in this segment.

8.3.5 Conclusion

The area between the Dead Sea and Bet She'an depressions is essentially slightly uplifted, bounded partly by longitudinal faults and partly by monoclinal folds. Gravity modeling based on refraction results points to the existence of two different types of continental crust along the strike-slip. The sense of motion along the longitudinal faults varies from left lateral strike-slip near Jericho to mainly normal further north. The strike of the strike-slip fault is apparently NNE–SSW, crossing the valley from west to east. The thickness of the fill in this part of the Jordan Valley does not exceed 4 km and reaches a minimum just south of the Bet She'an depression.

8.4 THE BET SHE'AN–LAKE KINNERET DEPRESSION

Bet She'an–Lake Kinneret is the second-largest depression in the Jordan Valley. It extends from Bet She'an in the south to Lake Kinneret in the north, all of it still below sea level. Lake Kinneret is located in the northern part of the depression. The structure of the Kinneret basin appears to be more complex than that of others within the Jordan Rift Valley, such as the Dead Sea or the Hula basins. The complexity arises from the intersection of the north–south main rift fault system with the NW–SE trending faults on the western side of the lake. Structural and geophysical interpretation is difficult in this area, because of extensive occurrences of

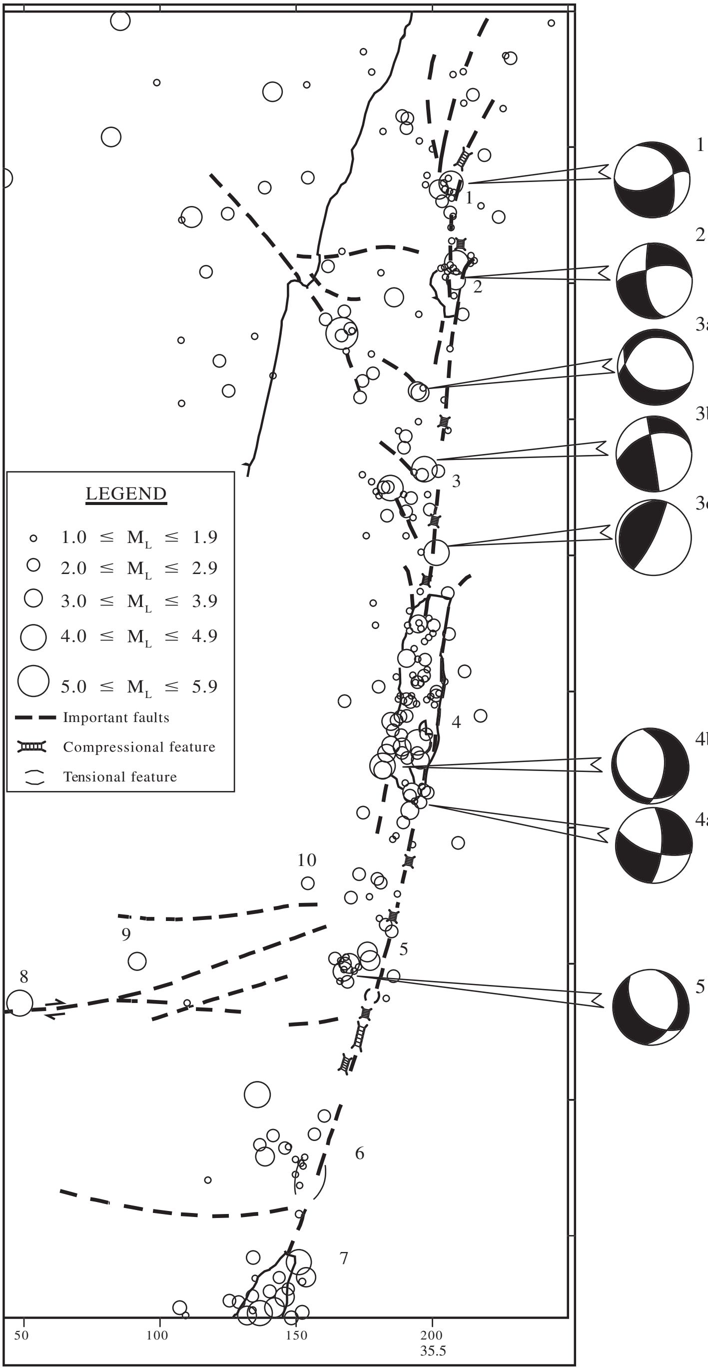


Figure 8.3.3. Location of epicenters and some fault-plane solutions in the Jordan Valley (after van Eck & Hofstetter 1990).

Plio-Pleistocene basalt flows and intrusions, which cover the area. Here we shall have to rely very heavily on gravity and magnetic data due to the lack of extensive seismic coverage.

8.4.1 Topography and bathymetry

The southern end of this depression is marked in the topography by a considerable widening of the valley, to a width of about 18 km. A north–south slope that curves westward, toward the Yizre'el Valley at its northern end, marks the western margin. North of Bet She'an the valley narrows again and is bounded to the west by a steep gradient, which continues about two-thirds of the way on the western shore of the Sea of Galilee. The western margin of the valley is marked by a north–south trending escarpment, which, in its northern part, parallels the eastern shore of the lake.

A detailed bathymetric map of the lake was compiled by TAHAL (1969) from surveys conducted in 1961, 1964 and 1968. The surveys comprised north–south profiles spaced 100 m apart. Later detailed local surveys revealed features which were not mapped previously. The need to map fine details, which could indicate subsurface structural features, has led to the new detailed bathymetric mapping of the lake (Ben-Avraham et al. 1990b). The new map is based on a very close grid of traverses and accurate navigation and shows much detail which was not noticed previously.

The main features of interest noted on the bathymetric map are as follows. The deepest part of the lake, deeper than 250 m below sea level, is located in the north-eastern part, where the lake is widest. The deep part of the lake forms an elongated depression trending NNE–SSW. Note that the contours in this depression are more disturbed than elsewhere in the lake. The deepest point is at 256 m below sea level. Steep slopes mark the margins of the lake. The eastern margin trends almost north–south, and the slope has a relief of about 20 m to the south and 40 m to the north. A very conspicuous feature is a scarp at elevations of −223 to −231 along coordinates 238–239N, in the southern part of the lake, which steepens from east to west. Results of seismic reflection profiles indicate that the scarp is the surface expression of a fault (Ben-Avraham et al. 1990b). The scarp separates the shallower southernmost part of the lake from its main part. The floor of this shallow terrace is disturbed, which suggests the possibility of an active structural element in this area. The western margin is broad and regular, with a north–south steepening of the gradient noted in its northern part. An examination of the 3.5 kHz sections obtained in this area shows that this feature is a local crack, which separates two parts of a slope with equal inclinations.

The bathymetry indicates active faults along the margins. Along the western margin, fault segments follow the shape of the shoreline (Fig. 8.4.1). The fault along the eastern margin runs almost straight, north–south, not following the undulations of the shoreline.

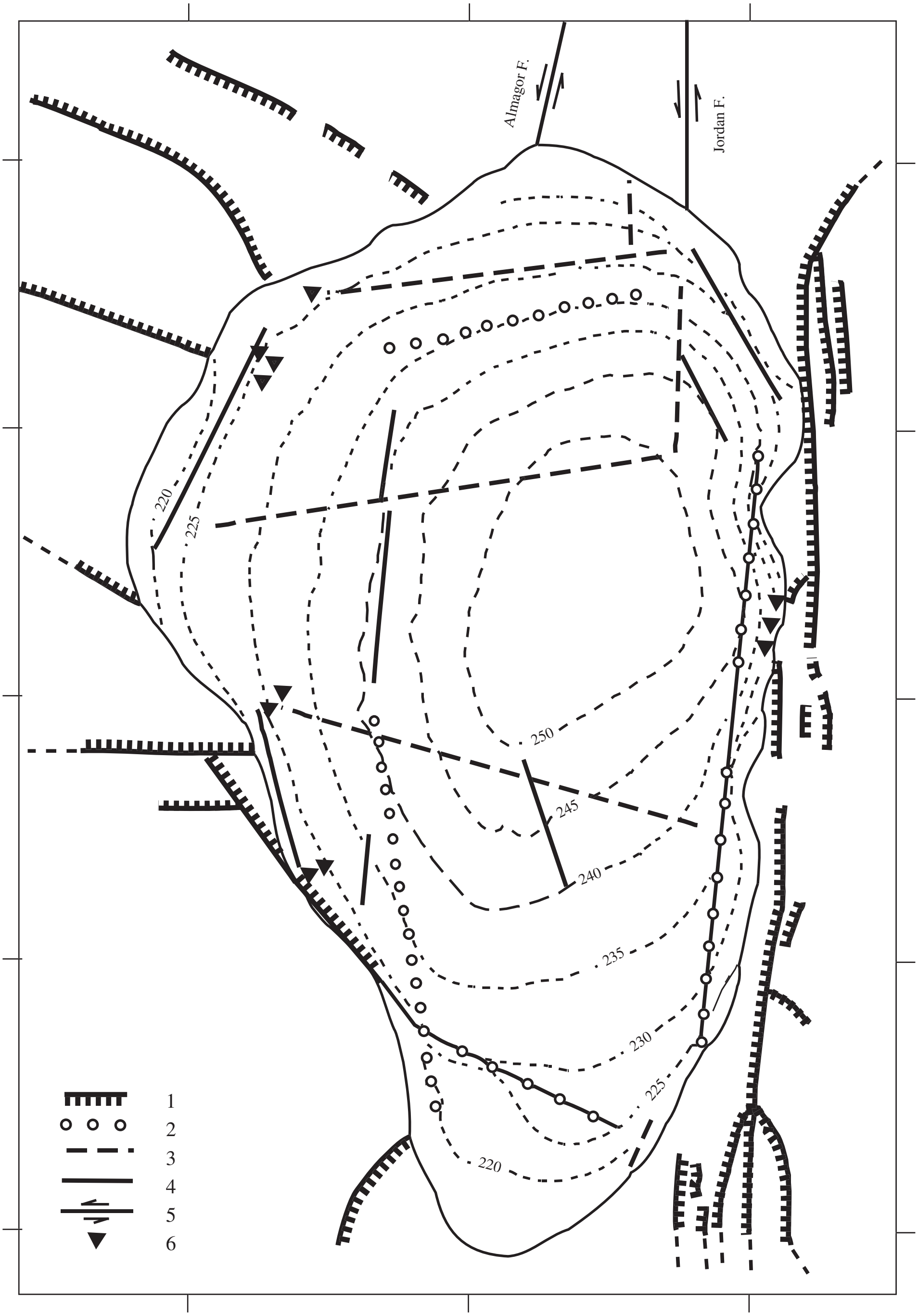


Figure 8.4.1. Structural elements map of Lake Kinneret based on bathymetry: (1) faults on land, (2) faults based on gravity data, (3) faults based on magnetic data, (4) faults based on bathymetry, (5) strike-slip faults on land and (6) hot springs (after Ben-Avraham et al. 1990).

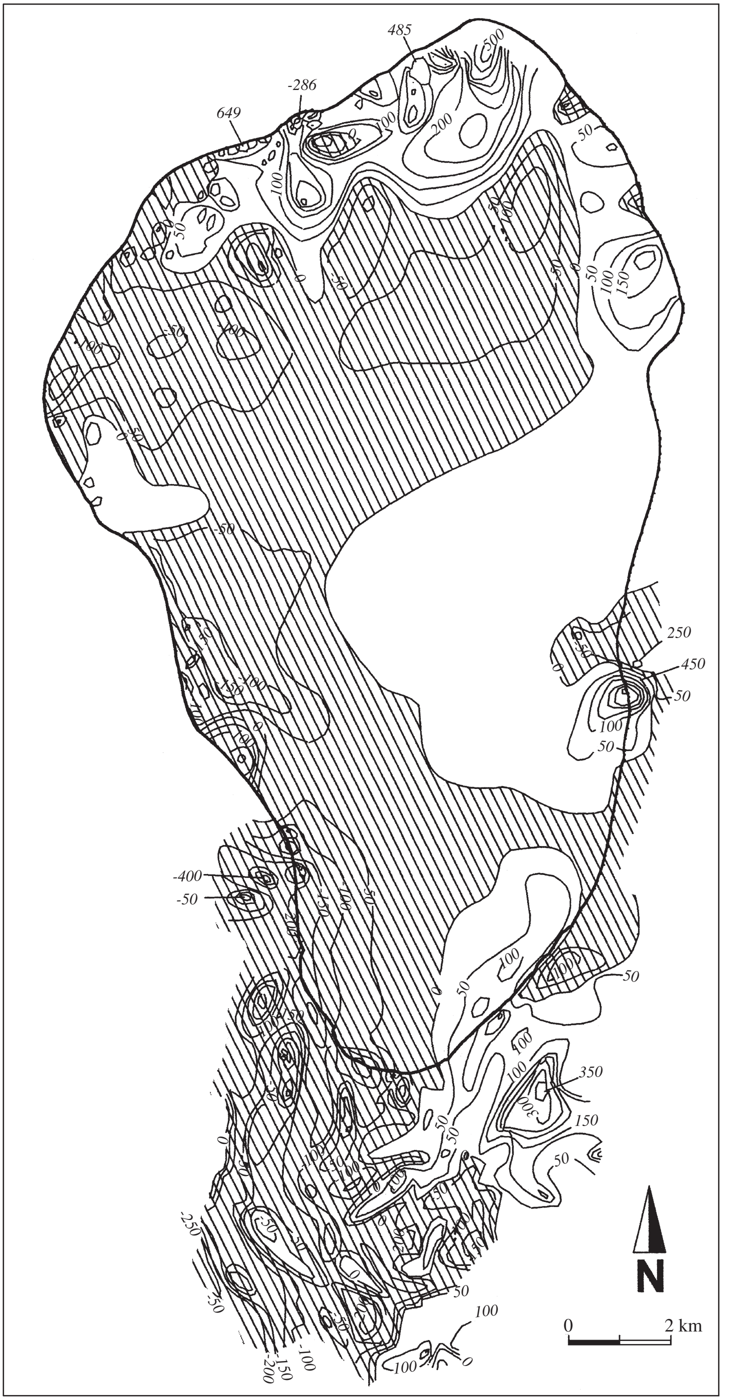


Figure 8.4.2A. Magnetic anomaly map of Lake Kinneret.

8.4.2 Structure derived from gravity and magnetic measurements

Regionally, the aeromagnetic map of Israel (Folkman & Yuval 1976) shows the flanks of the depression and the effects of the near-surface basalt occurrences. Since the map does not provide information regarding the floor of the depression, detailed magnetic studies were conducted in Lake Kinneret and in the land area south of it (Ben-Avraham et al. 1980, Ginzburg & Ben-Avraham 1986). The magnetic anomaly map (Fig. 8.4.2A) shows that the lake and the area south of it are divided into two distinct parts: a magnetically quiet zone in the central part of the lake and a magnetically disturbed zone along the lake margins which extends south of the lake. The amplitude of these anomalies is 50–300 nT, with widths varying from 200 m to 1 km. The magnetic anomaly zone on the western margin of the lake continues to about 3 km south of the lake, where its trend changes to NW–SE. The magnetic anomaly zone along the eastern margin of the lake runs through the site of the Zemah 1 well, meeting the western anomaly some 5 km south of the lake.

Magmatic rocks are prevalent in this area and some of the anomalies mapped coincide with known bodies. Thus, it may be assumed that the causative bodies underlying the magnetic anomalies are basic igneous rocks. The susceptibility value used in the interpretation was 0.025 S.I., based on measured values (Domzalski 1967). Since remanent magnetization is weak in this area (Nur & Helseley 1971, Ron et al. 1984), it was not considered in the evaluation. Two types of causative bodies were interpreted: shallow bodies causing narrow anomalies with steep (300–400 nT/km) gradients, and elongated anomalies with moderate gradients caused by deep seated bodies with a large vertical extent. The interpreted structural elements derived from the magnetic map are shown in Fig. 8.4.2B. A model was calculated for the Zemah 1 well, where thick basalt was encountered (Marcus & Slager 1985), which comprises a tabular body simulating the Pliocene basalt flow found in the well and the underlying gabbro-evaporite sequence, which extends to a considerable depth (Ginzburg & Ben-Avraham 1986). The anomaly trend along the western margin of the lake is due to a complex causative body with a limited lateral extent and extending to a great depth. This trend of elongated magnetic anomalies along the western margin, with their NW–SE trending extension to the south, marks the extent of the Kinneret basin (Fig. 8.4.2).

Detailed gravity measurements in Lake Kinneret were made in 1988 (Ben-Avraham et al. 1996), and incorporated into the compilation gravity map of the Jordan Rift (ten Brink et al. 1999). The Free-Air anomaly map expresses to a large extent the topography of the area. It does show however that in the lake, where the topographic effects are minimal, the gravity minimum is located closer to the eastern margin of the lake and south of the bathymetric low (Fig. 8.4.3). The Free-Air minimum is separated from the western margin of the lake by a steep gradient, trending north–south. The Bouguer anomaly map (Fig. 8.4.3), in which the gravitational effects of the topography have been removed, shows the regional

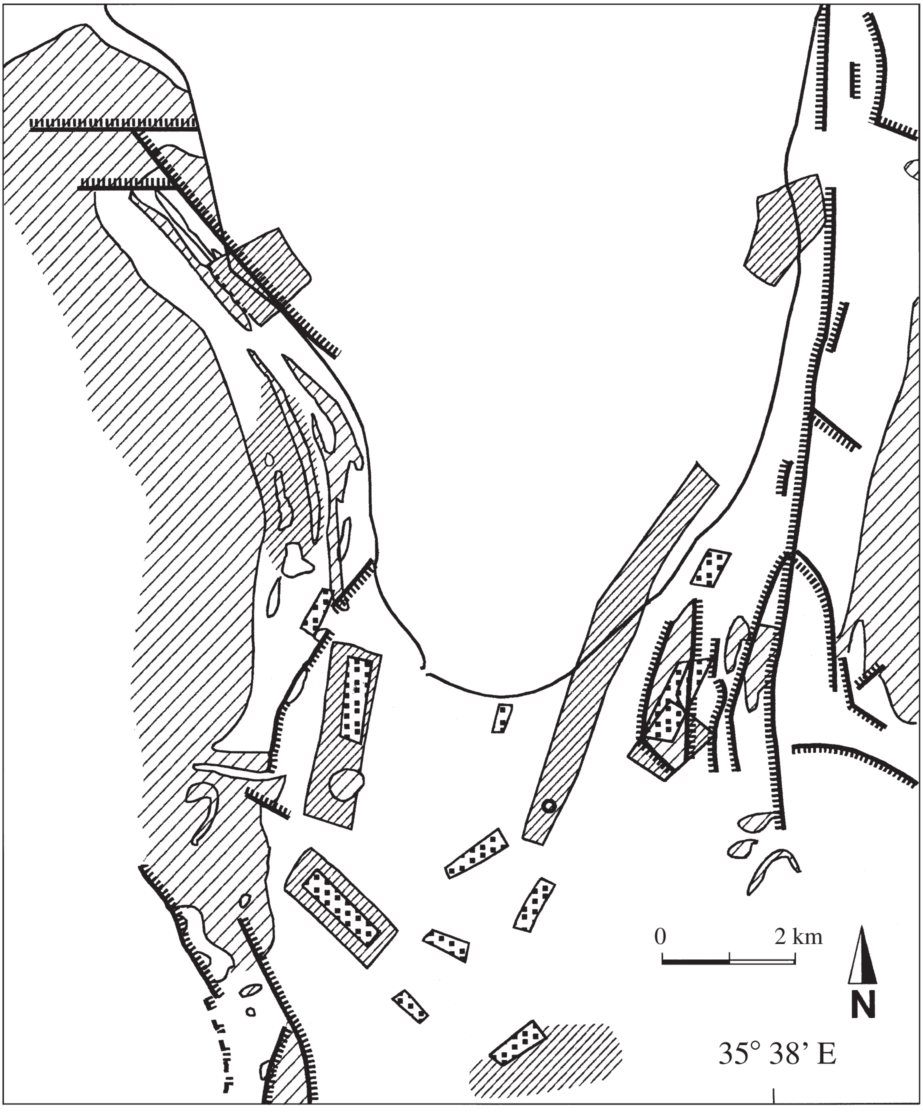


Figure 8.4.2B. Structural interpretation of the magnetic anomalies in Lake Kinneret. Widely spaced hatched areas indicate basalt outcrops, rectangles with dots indicate the locations of shallow magnetic bodies and closely hatched rectangles show the location of deep magnetic bodies (after Ginzburg & Ben-Avraham 1986).

gradient, in which gravity values decrease from +40 mgal to the west to 0 mgal east of the lake. This gradient is the gravity expression of the east–west shoaling of the Moho. It tends to maximize gradients to the east and minimize to the west. The Lake Kinneret Bouguer minimum of −32 mgal is located in the southern part of the lake, close to the eastern margin. It is separated from both margins by gravity

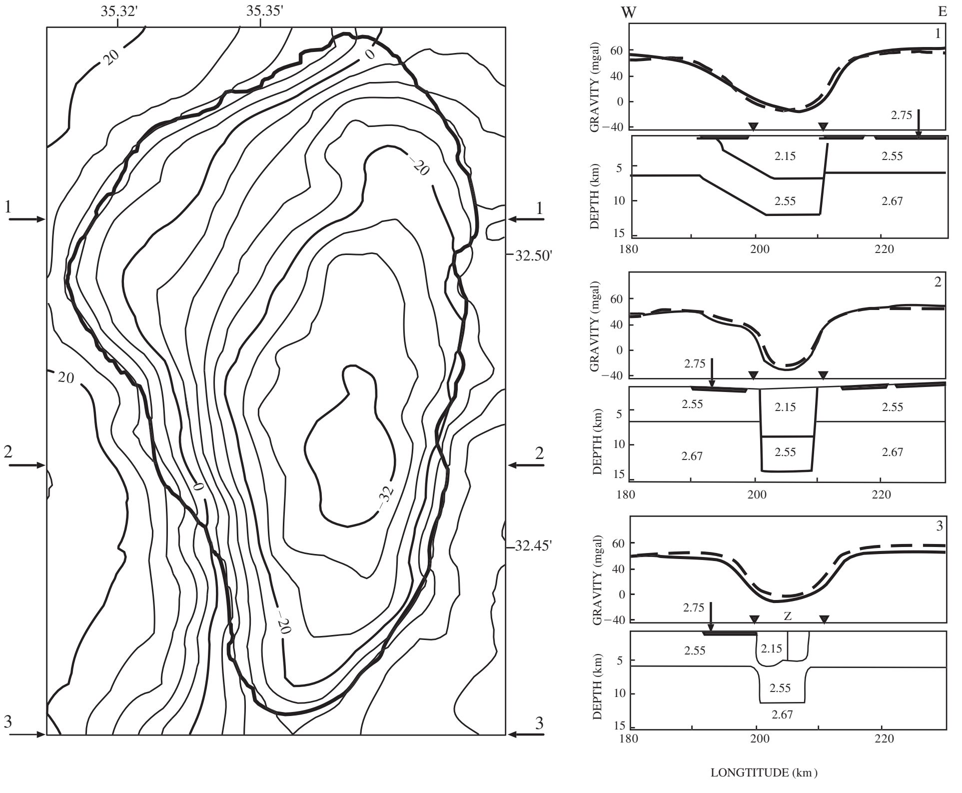


Figure 8.4.3. Bouguer anomaly map of Lake Kinneret and three calculated east–west gravity models. Locations of the profiles are shown on the map (after Ben-Avraham et al. 1996).

gradients, which parallel the margins from the southern end to latitude 32°48'N. The map of the first horizontal derivative of the gravity field (Fig. 8.4.4) accentuates the short-wavelength anomalies. This map, together with the Bouguer anomaly map, shows that the lake is divided into two sub-basins. The southern sub-basin occupies the narrower part of the lake and includes the gravity minimum. It extends southward to include Kinarot Valley. The second sub-basin occupies the broader northern part of the lake and extends to the land areas of Buteiha to the northeast and Ginnosar to the west. The Kinneret–Bet She'an depression is bounded on the west by an elongated trend of first derivative anomalies, which trend north–south to the lake and then follow its western shoreline to its widest point. To the east an elongated first derivative anomaly follows the shoreline to the Buteiha Valley. These elongated anomaly trends indicate the location of faults with a considerable normal component, which delimit the depression on either side.

Ben-Avraham et al. (1996) calculated three two-dimensional gravity models across Lake Kinneret, based on all the geological and geophysical data available, and their locations are marked on the Bouguer map (Fig. 8.4.3). The densities used

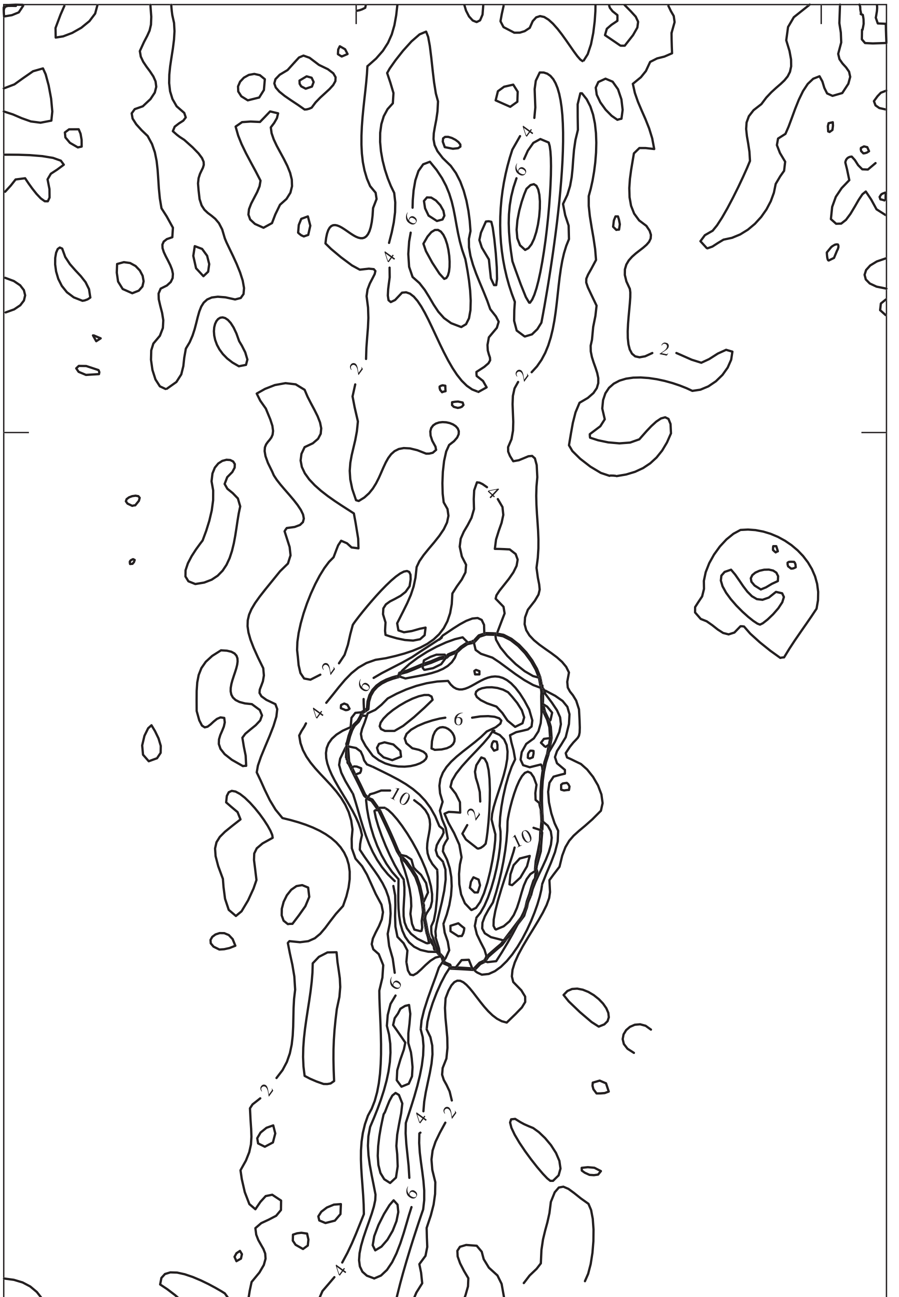


Figure 8.4.4. Map of the first horizontal derivative of gravity in the northern Jordan Valley and adjacent areas. The elongated closed maxima indicate the location and trend of the main faults (after Ben-Avraham et al. 1996).

for calculation of the gravitational effect are shown in the diagram. The difference between the northern and southern sub-basins is evident. The northern sub-basin has less sediment fill, bounded by a fault on its eastern margin, whereas to the west faulting plays a minor role. In the center a full rift is interpreted, with some 8 km of fill. The third profile, located south of the lake, through the Zemah 1 well, shows a full rift structure and some 5 km of fill. One should bear in mind the effect of unknown thickness of basalt flows and intrusions, particularly south of the lake. Since the high density basalt within the fill would increase the average density, the calculated depth should be considered a minimum figure, and the actual thickness of the fill in the southern basin of the Kinneret–Bet She'an depression could be considerably higher. The structure as inferred from the gravity data is shown in Fig. 8.4.3.

8.4.3 Seismic investigations

A number of seismic refraction and reflection experiments have been conducted in the lake and the onshore part of the depression over the years. A seismic refraction experiment in which two east–west profiles were shot was conducted in Lake Kinneret (Ben-Avraham et al. 1981). In the northern profile a 3.6 km/sec refractor underlying a 2.5 km/sec layer was detected at a depth of 420–520 m. In the southern profile a different velocity sequence was found. Here a 2.1 km/sec layer overlies a 2.5 km/sec layer at a depth of 620 m. The difference in the velocity structure suggests different stratigraphic sequences in the north and south of the lake, which could be separated by faults. The increased depth to the 2.5 km/sec refractor indicates that the southern part of the lake has a deeper fill. The same seismic experiment included a number of shallow penetration profiles. Several energy sources such as a sparker, a boomer and an air-gun system were used. All yielded low penetration records, which gave information on the uppermost sediments. Zones of deformation were observed along the margins of the lake, but some evidence for deformation in the center was also obtained.

Seismic profiles of 3.5 kHz were measured over the entire lake, with variable spacing (Ben-Avraham et al. 1986). In general, no penetration was achieved in most parts of the lake because of the high gas content in the uppermost sediments. However, in some locations good penetration was achieved. One such location was over a terrace in the southern part of the lake, where the sedimentary structure shows up very clearly. Another area where good penetration was achieved is near the hot springs, in the northwestern part of the lake, where folded structures are present. The deformation includes some active faulting.

A seismic reflection profile shot in the course of oil exploration in the Zemah area (Kahsai & Croker 1987) is shown in Fig. 8.4.5. The main features interpreted are a faulted anticline under a thin cover of young sediments, with scattered partial reflections below the mapped horizon. On the western side of the section a flower-structure fault zone is shown, indicative of a strike-slip movement. In a

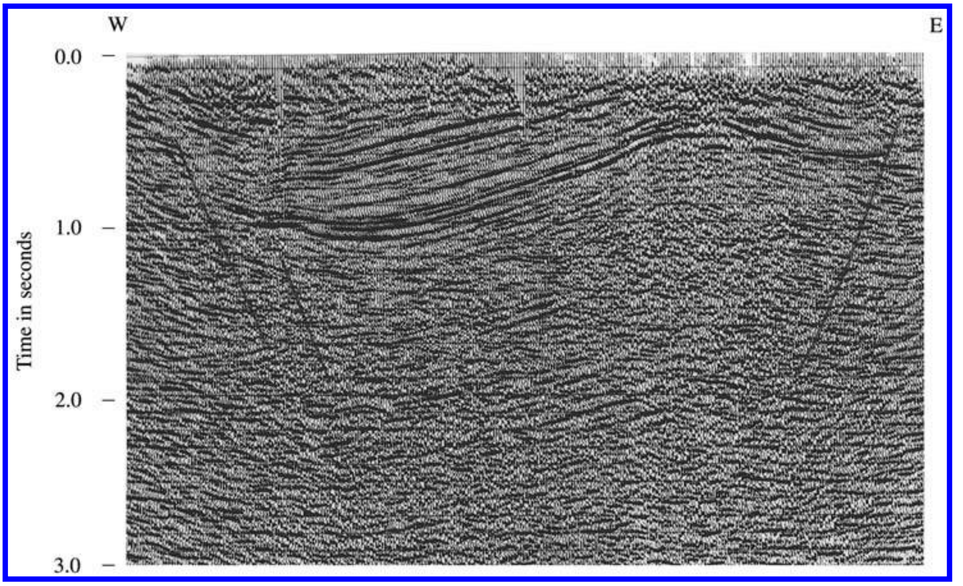


Figure 8.4.5. East–west seismic reflection section shot south of Lake Kinneret. Note the prominent apparent Zemah anticline and the boundary faults on the west flank of the valley.

later paper, Rotstein et al. (1992) showed a high-resolution version of the same section. Their interpretation is of two anticlines (Zemah and Ubeidiya) separated by a central syncline and flanked by two main fault zones, east and west. The quality of both sections published, below the base of the upper fill, leaves a lot of latitude to the interpreter, but two zones of disruption are evident. The Zemah anticline may well be a tilted fault block comprising basalt and evaporites, as revealed by the Zemah 1 well (Marcus & Slager 1985). The evidence for the main faults is there, but the determination of the type and sense of motion is open to question.

8.4.4 Seismicity

The southern part of the Lake Kinneret–Bet She'an depression, between the lake and Bet She'an, is not very active seismically (Fig. 8.3.3). van Eck & Hofstetter (1989) ascribed most of the epicenters which were located in this area to the fault delimiting the eastern margin of the Lake Kinneret depression, itself seismically active, with most of the epicenters concentrated along the eastern margin of the lake and at its northern part. A focal-plane solution obtained for an earthquake cluster in the north of the lake indicates a north–south left lateral strike-slip motion. This motion could be associated with the fault which was mapped by Rotstein & Bartov (1989) by seismic reflection profiling in the land area just north of Lake Kinneret.

8.4.5 Heat flow in Lake Kinneret

Heat flow was measured in Lake Kinneret in 1975 (Ben-Avraham et al. 1978). It is unusual to measure a temperature gradient in shallow water bodies, because of the thermal instability of the bottom water layer. In the case of Lake Kinneret the saving factor was the availability of a temperature record of the entire lake for many years. The heat-flow measurements were recorded at 10 permanent stations over a period of 6 years prior to the study. Temperatures are more stable in the central part of the lake. The corrected temperature gradient varies between 0.91 and 1.09×10^{-3}°C/cm. The mean value of the corrected heat flow for Lake Kinneret is 75 mW/m^2, which is significantly higher than values in the Dead Sea.

8.4.6 Conclusion

The structural interpretation of the geophysical data in Lake Kinneret–Bet She'an depression relies heavily on the compiled gravity data, but is partially supported by seismic and seismicity information. The structural picture that emerges is shown in Fig. 8.4.6. Lake Kinneret is divided structurally into two parts (Ben-Avraham et al. 1996), the northern part being wider and bathymetrically deeper than the southern. It is bounded to the east by a north–south trending fault, which is seismically active. The western margin is shaped by the NW–SE trending branching faults of the eastern Galilee. The depth of fill in the northern sub-basin may reach 5–6 km. The southern sub-basin, which extends southward into the Kinarot Valley and further south to Bet She'an, is narrower than the northern sub-basin. It is a full rift bounded by faults on either margin, from Lake Kinneret to about 12 km north of Bet She'an. There is some evidence indicating a strike-slip component of motion on both faults. This sub-basin is deeper than the northern, with depth of fill reaching a maximum under the southern part of the lake, considerably further south of the bathymetric deep which probably marks the location of present subsidence. The depth calculated here is about 8 km, this being a minimum value because of the effect of the unknown thickness of basalts on the computation. The depth of fill decreases gradually southward. The resemblance between the Dead Sea and Lake Kinneret basins is noteworthy. In both depressions the northern part of the lake is bathymetrically deeper, while the southern is structurally deeper, containing a thicker section of Plio-Pleistocene fill.

8.5 THE HULA DEPRESSION

The Hula basin occupies the northernmost part of the Jordan Valley. Topographically it is separated from Lake Kinneret by the Korazim block to the south, the latter bounded by a fairly gentle gradient leading to the Golan Heights

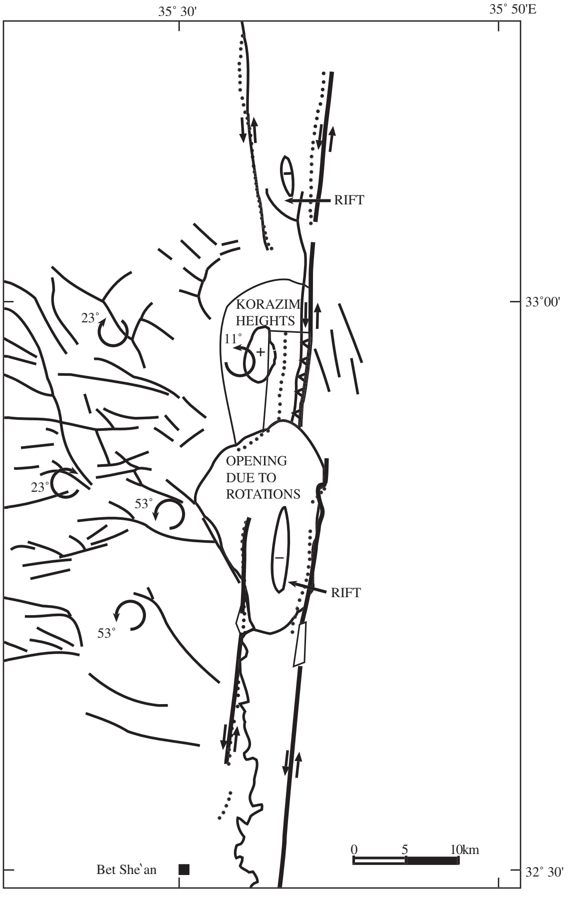


Figure 8.4.6. Structural elements of the northern Jordan Valley. Lines with arrows mark interpreted strike-slip faults. Hatched areas mark location of faults from seismic data and dotted lines mark location of maximum gravity gradients. Circular arrows show block rotation in Galilee (after Ben-Avraham et al. 1996).

to the east, with the Jordan River gorge separating the two. Further north the basin is bordered to the west by a steep escarpment, while a gentler gradient leads to the Golan Heights to the east.

8.5.1 Gravity and magnetic interpretation

The Bouguer anomaly map shows the limits of the Hula basin very clearly (Fig. 8.4.4). It is a closed gravity low bounded by steep gravity gradients on the east, south and west, with a gentler gradient marking its northern limit. The closed elongated highs on the map of the first horizontal derivative of the gravity field (Fig. 8.4.4) point to the existence of faults on the margins of the basin. The fault on the western margin trends NNW–SSE while the fault on the east trends north–south. The negative Bouguer anomaly associated with the Hula basin is about 25 mgal. Using density values derived from an exploration well drilled in the basin, Klang (1984) has recalculated the depth of fill, which consists of alternating basalt and low density material such as peat, lignite, marl and sandy layers. His results indicate a depth of about 5 km for the high-density sediment base of the fill (Fig. 8.5.1).

The magnetic map of northern Israel comprises two distinct parts. West of 35°15'E the magnetic field is smooth with a gradual increase westward. The eastern

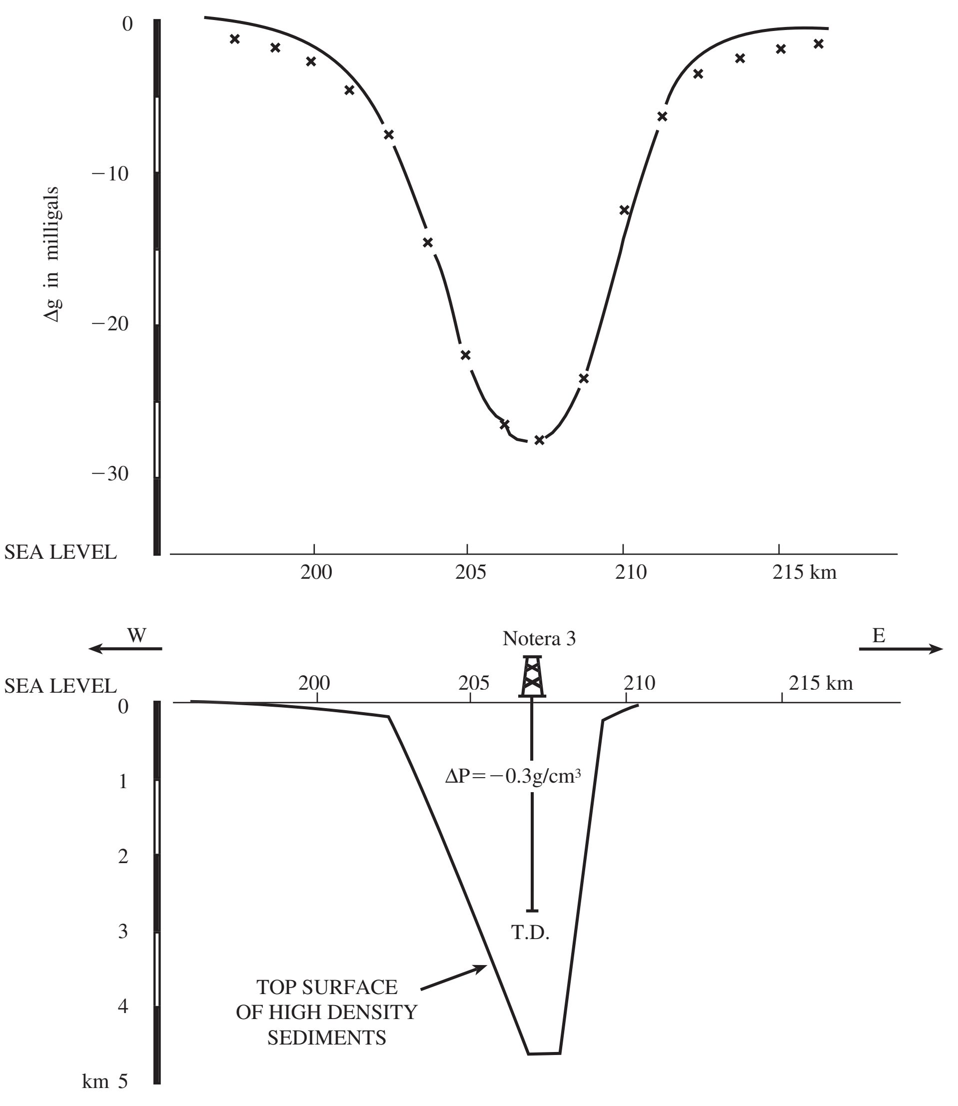


Figure 8.5.1. A two-dimensional gravity model calculated across the Hula basin (after Klang 1984). Upper sector: solid line is the observed Bouger anomaly after reduction of the assumed regional background; ××× represents the calculated anomaly of the illustrated model. Lower sector: solid line represents the top surface of high density sediments.

part is characterized by short-wavelength anomalies superimposed on the regional gradient. The anomalies over the Golan are extremely short wavelength and are aligned SW–NE. The short-wavelength magnetic anomalies are obviously associated with the widespread volcanic occurrences in the eastern Galilee, the Hula basin and the Golan Heights. The gravity and magnetic fields over this area resemble those of a regional component with minor superimposed effects. Since the magnetic field is involved, upper mantle effects are excluded. Hence, the combined interpretation of both fields can give the crustal structure of this area. Folkman (1981) suggested the following interpretation: after filtering the short-wavelength magnetic anomalies along an east–west profile across the Galilee he obtained a smooth curve, showing a westward increasing magnetic gradient. Since no westward shoaling of the magnetic basement is known, the increase in magnetic values can only be explained by a westward increase in the magnetic susceptibility of the crust, that is, an increase in its mafic content. The model shown in Fig. 8.5.2A demonstrates this change, which indicates a different crustal type on either side of the Jordan Valley. This interpretation is well supported by the gravity data. It has been established that the westward crustal thinning in northern Israel takes place from the continental shelf westward. Therefore the thickness of the crust under northern Israel is fairly uniform. Hence, the westward rise in the Bouguer anomaly values can only be explained by an increase in the density of the crust, as is shown in Fig. 8.5.2B. The gravity model is almost identical with the magnetic, thus confirming the existence of two different crusts east and west of the Jordan Valley.

8.5.2 Seismic reflection and seismicity

A seismic reflection study (Rotstein & Bartov 1989) shows the geometry of the layers across the Jordan fault, between Lake Kinneret and the Hula. The seismic data (Fig. 8.5.3) show that just north of Lake Kinneret, the down dipping Golan block to the east dips about 70° to the west, separated from the Korazim block by a wide fracture zone. The fault shows clear overthrusting from the west, which indicates compression caused by the convergence of the two blocks due to strike-slip motion. While the eastern block shows continuous coherent reflectors, the reflections across the Korazim block to the west of the main fault are badly disrupted, with near-horizontal reflectors at shallow depth and indications of doming, accompanied by small-scale faulting deeper down. Rotstein & Bartov (1989) attribute the push up on the Korazim block to the compression across the Jordan fault. The positions of the fault in the Golan block, which are hidden by basalt, can be estimated from the location of the Jordan fault at depth and on the surface. These positions appear to trend to the NNE, from the gorge of the Jordan River near Lake Kinneret. Further north, the fracture zone is narrower and located further east. The interpretation is that the western boundary fault of Lake Kinneret continues obliquely to the north to become the eastern boundary fault of the Hula

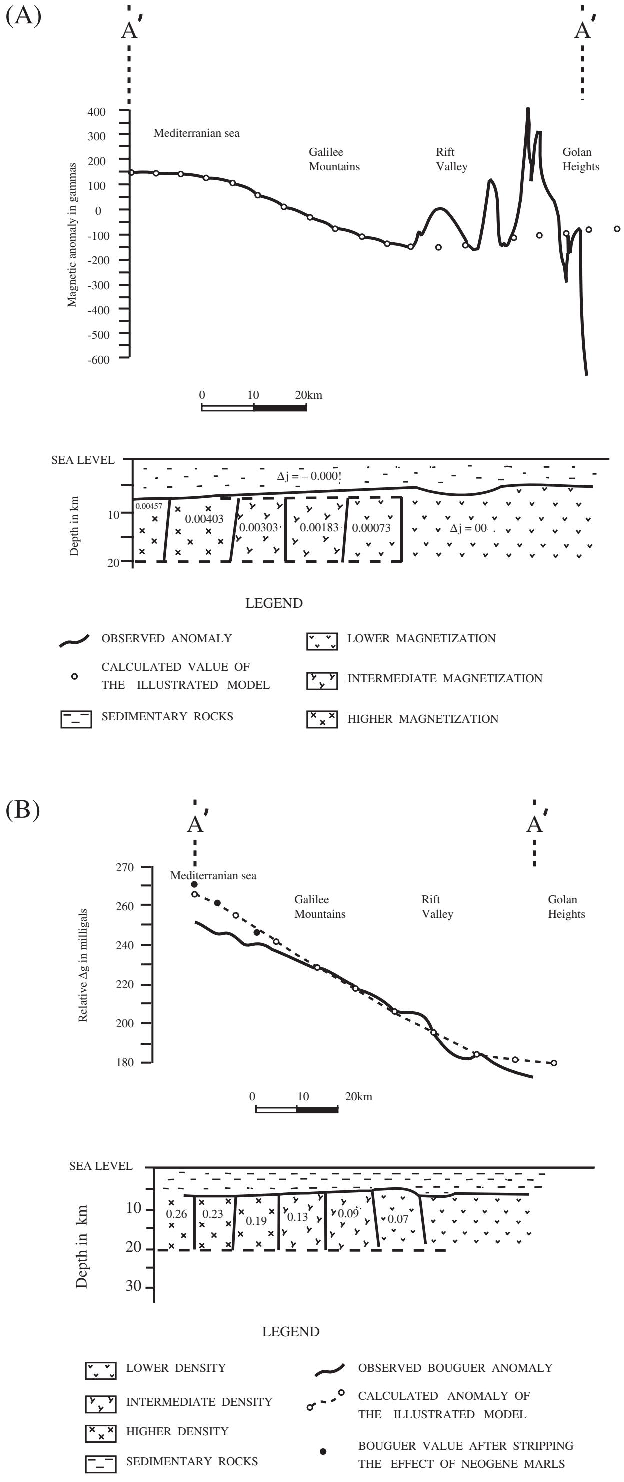


Figure 8.5.2. Gravity and magnetic models across the Galilee. (A) Magnetic model. (B) Gravity model (after Folkman 1976).

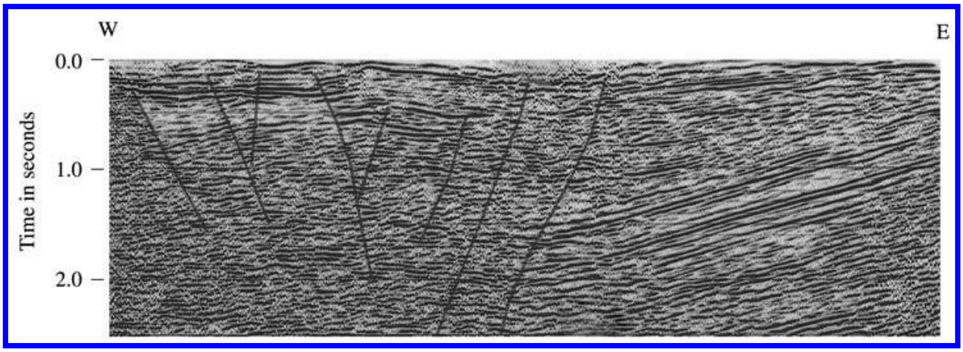


Figure 8.5.3. Seismic reflection section shot across the Valley just north of Lake Kinneret. Note the prominent strongly west-dipping event to the east and the overthrust block to the west (after Rotstein & Bartov 1989).

depression. The western boundary fault of Lake Kinneret veers to the north-east and is evident at the eastern side of the seismic reflection section shown in Fig. 8.5.3.

Earthquake epicenter locations between the Hula and Lake Kinneret are aligned along the Jordan fault, but the most active part is apparently the western boundary fault of the Hula basin. Focal plane solutions for these earthquakes (van Eck & Hofstetter 1989) are that of a north–south strike-slip motion, with a vertical component. The Hula basin is thus as close to a classical pull-apart basin as one can observe in the Jordan Valley, with evidence coming from all geophysical data available.

8.5.3 Conclusion

The northernmost part of the Jordan Valley, north of Lake Kinneret, shows the features one would expect in a pull-apart basin, albeit small. North of Lake Kinneret we have good evidence of a NNE–SSW striking left lateral strike-slip fault with a reverse faulting component. This fault causes compression and an uplift of the Korazim block. The fault forms the eastern margin of the Hula depression to the north. Strike-slip motion is apparently transferred to the western margin in the small two-sided Hula rift basin, north of the Korazim block. This basin is rather shallow, with only 5 km of fill.

8.6 OVERALL STRUCTURE OF THE JORDAN VALLEY

Thus far we have described and discussed the geophysical results obtained in the Jordan Valley within the individual depressions which compose the valley. We now treat the Jordan Valley as a unit and examine the implications of the geophysical results for the internal structure of the valley and the mode of its formation.

The Jordan Valley is only a part of a major continental transform, the Dead Sea strike-slip, extending from the Red Sea to southern Turkey. This part of the transform has attracted the attention of investigators over the past 150 years. The Jordan Valley has been referred to as a geosuture, a rift valley (and even a ramp valley) and the mode of its formation has been the source of many discussions. However, in the past three decades the view that the Jordan Rift is in fact a transform fault separating two plates, with pull-apart basins developed by its left lateral motion, has gained wide acceptance. The early investigators had to rely on surface geology and occasional drill holes and of course much intuition. At present we have at our disposal an accumulation of geophysical data (albeit insufficient in some places) which gives us an insight into the concealed internal structure of the crust and the Valley. In this section we shall review the structure of the Jordan Valley and discuss the tectonic implications of the geophysical results.

The regional structure of the crust under the Jordan Valley and the adjacent areas is known from refraction studies and from the interpretation of gravity and magnetic data. From refraction data it appears that a slightly thinned crust underlies the southern part of the Jordan Valley. The velocity structure of the crust is different on either side of the valley, indicating the presence of two different crustal types adjacent to each other along the Rift. This difference becomes more prominent toward the north. Between Jericho and Bet She'an, a large difference in crustal thickness is evident, the crust west of the Jordan Valley being considerably thinner. Still further north, in the Galilee, no refraction data are available, but gravity and magnetic modeling point to the presence of denser, more basic crust under the Galilee. The geophysical evidence for the juxtapositioning of two different crustal types along the Jordan strike-slip is convincing.

The topographic division into three main depressions, the Dead Sea–Arava, the Lake Kinneret–Bet She'an and the Hula, is accompanied by matching gravity anomalies which follow the same trend. The Bouguer anomalies do not reflect topography but rather the subsurface structure. The depressions are typified by considerable gravity minima, the most striking of which is that of the Dead Sea. The Dead Sea gravity anomaly marks the extent of the depression. The steep gravity gradients along the margins mark the location of the main boundary faults, while the southern end of the gravity low marks the southern margin of the depression. The deepest part of the anomaly, located south of the Lisan Peninsula, marks the zone of maximum deposition of Plio-Pleistocene sediments, which is corroborated by seismic refraction. From the refraction data it is clear that the Dead Sea depression is divided into two basins, the northern being shallower than the southern. The depth to basement to the north is 6–7 km, with a relatively thin fill. The southern basin, now subaerial, is structurally the deepest part of the depression. Refraction data show that the fill may reach a thickness of 10 km south of the Lisan Peninsula, while the depth to basement is some 14 km.

South of the Dead Sea, in the northern Arava, the gravity data indicate further subdivisions.

The analysis of seismic reflection data shows that the eastern and western margins of the Dead Sea comprise of a number of step faults with considerable vertical displacements, which have depressed the basement to its present depth. Diapiric structures arising from the evaporitic layers were mapped, the main ones being the Mount Sedom, En Gedi and Lisan diapirs. The sub-basins are separated by a number of transverse faults. Of these, the Boqeq fault located south of the Lisan Peninsula bounds the deep southern basin to the north. To the south, the main transverse fault is the Amazyahu fault, which comprises a number of normal step faults. These fault blocks are associated with salt, which was pushed up from the deep part of the basin against them, and was covered by the south dipping sediments, which were deposited toward the growth faults. The Amazyahu fault was previously interpreted as a listric normal fault, flattening against a salt layer. South of this fault the sediments are semi-horizontal, thinning to the south, while the next sub-basin is limited to the south by the normal Iddan fault.

In the northern basin, longitudinal faults were located by single channel seismic surveys. The easternmost boundary fault on the western margin is segmented, and a fault with indications of strike-slip motion mapped by reflection profiling near Jericho may comprise its northern extension.

The Dead Sea is extremely active seismically. It appears that most of the deformation which occurs now, takes place along the longitudinal and transverse faults only. At present the seismic activity in the northern basin is concentrated along the western margin (the Jordan fault), while in the southern basin the seismic activity is located along the eastern margin (the Arava fault). Both strike-slip and normal motions were found along these faults. Seismic and heat-flow studies indicate that the deformation is active within the sediments and the brittle upper crust, which holds true for the entire Jordan Valley.

The segment of the Jordan Valley between the Dead Sea and the Bet She'an depression is bounded essentially by monoclines on either side, with occasional longitudinal faults. The young fill is fairly thin, reaching a minimum thickness around 32°15'N where the valley is narrowest. Earthquake activity is sparse in the valley itself and is concentrated at zones of interaction with faults, which trend NW–SE into Samaria and toward Mount Carmel. Here again, both strike-slip and vertical motion have been interpreted, with variable strikes.

Between the Dead Sea and the southern extension of Lake Kinneret, no two-sided rift segments can be observed in the subsurface. The existence of a longitudinal fault in the center of the valley cannot be resolved with the data at hand. The addition of a number of east–west reflection lines across the valley could resolve this issue. At present, the possible lineup of sparsely located earthquake epicenters, along a roughly north–south trend, is the only possible hint for the existence of such a fault.

The next depression north, the Kinneret–Bet She'an, is the second largest in the valley. Like the Dead Sea it is a composite depression, divided into two parts.

The northern part, although wider and bathymetrically deeper, is shallower, with a thinner fill of 5–6 km. It is bounded to the east by a north–south fault, while to the west it is limited by NW–SE trending faults. The southern basin is narrower and includes the southern part of the lake, the Kinarot Valley and the Bet She'an Valley. It is bounded by faults on either side. The thickness of the rift fill reaches a maximum of 8 km in the southern part of the lake, gradually decreasing to the south.

North of Lake Kinneret a clear indication of compression and left-lateral motion along a fault between the Golan and Korazim blocks has been mapped seismically. This fault apparently trends NNE from the gorge of the Jordan River near Lake Kinneret, forming the eastern margin of the Hula depression, which is an almost classical pull-apart basin. It is bounded to the east and west by faults. The western margin fault is seismically active, with a north–south strike-slip motion and a normal component. The depth of fill of this depression is estimated at 5–6 km.

The results of the geophysical investigations of the Jordan Valley, although sadly lacking in some parts, give a good general structural picture. The valley is a tectonically-controlled depression, which separates two blocks with a different crustal structure, thickness and composition. The three basins located along the valley, namely the Dead Sea–Arava, the Lake Kinneret–Bet She'an and the Hula, are seismically active with both strike-slip and normal motions. Geophysical evidence for the existence of strike-slip faulting was found in parts of the Jordan Valley, with gaps such as the region between Jericho and Bet She'an, which for the time being must be filled via geological considerations and conjecture. In all, despite the gaps in our data, and while other solutions may be possible, the geophysical results can in our view be best explained by a left-lateral transform fault separating the Arabian plate from the Sinai sub-plate.

CHAPTER 9

Structural history

The structure of the Jordan Rift Valley attracted attention for a long time (see Section 2.1), as expressed in the wealth of studies dealing with this topic (see bibliographies of Arad et al. 1997, 1998, Inbar et al. 1989, Arad & Bartov 1994, Qummou et al. 1997). Its complex and elusive characteristics, many of which are buried under the younger sediments, made understanding and mapping the structures very difficult, resulting in the rather sad fact that there are rarely two geologists who agree on any single issue, nor two maps that look exactly alike. One of the main drawbacks which hindered reconstruction of the Rift's history is the uncertainty in chronology of the rock units, due to lack of suitable indicative fossils. This problem has, at least to a certain degree, been overcome during recent years by the development and application of palynostratigraphic and radiometric methods, which have considerably clarified the chronology.

Palynostratigraphic studies were carried out by Horowitz and friends at the Palynological Laboratory of the Institute of Archaeology, Tel Aviv University. They concentrated mainly on analyzing 13 continuous sequences penetrated by drillings along the Rift, in the Hula, the central Jordan Valley and Dead Sea basins, the data only recently having become available. To these are added pollen analyses of numerous samples collected from exposures of various formations, either natural or excavated for archaeological purposes.

Radiometric datings, based on the K–Ar and $^{40}Ar/^{39}Ar$ total gas methods, were performed on almost all of the copious magmatic rocks occurring in both outcrops and boreholes in the central and northern Jordan Valley. These were first carried out by Horowitz et al. (1973) and Siedner & Horowitz (1974), later continuing in much more detail at the Geochronological Laboratory of the Geological Survey of Israel. The results appeared in numerous publications by Heimann, Steinitz, Shaliv, Mor and their associates, as detailed in the reference list at the end of this book. Unfortunately these methods could not be applied to the southern part of the Jordan Rift Valley, due to the lack of magmatic activity in this region. Uranium series and radiocarbon (Kaufman 1971; Vogel, in Horowitz 1979, p. 151; Schwarcz et al. 1979; 1980; Livnat & Kronfeld 1985; 1990; Clark et al. 1997), as well as other dating methods, are particularly concentrated around the Dead Sea and the Arava (Schramm et al. 1997, Enzel 1997, Amit et al. 1997, Marco 1997), but were also used elsewhere (Kronfeld et al. 1988), and were occasionally applied to time structural disturbances.

Although less age-indicative, geophysical investigations along the Jordan Valley (see Chapter 8) have contributed considerably to understanding its structure. Invaluable sources are chiefly the studies made by the Department of Geophysics and Planetary Sciences at Tel Aviv University, at the Geophysical Institute of Israel, and by the oil companies, most of the latter released only in recent years. A combination of the results obtained from geophysical, radiogenic and palynostratigraphic techniques is the basis for the structural history outlined here.

The present chapter deals with dating events involved in the evolution of the Jordan Rift Valley. Some matters of dispute are discussed here, chronological as well as structural and tectonic; but the main issue of present-day disagreement, whether the Rift was formed by vertical or lateral movements, and its geotectonic implications and significance, is left for Chapter 10.

Downfaulting and basin formation usually require compensation by uplift somewhere not too far away from the subsiding region. The contrast between downfaulted and uplifted neighboring areas is further emphasized by isostasy, balancing the load of sediments eroded from high localities, and accumulated in the lower ones. It is however much easier to date the stages in evolution of basins by their sediments, than to time the history of eroded highlands. An artificial problem in studying the structure and evolution of the Jordan Rift Valley results from "allocation" of areas of interest, causing a situation in which most studies are concerned with either the Rift itself, or the bordering highlands. Only very few investigators integrated subsidence and uplift to clarify tectonic processes, as Picard had already done in 1943, following Wellings (1938) and other researchers of this time.

The inner sector of the morphologically elongated Jordan Rift Valley displays three types of structures, active throughout its entire history: deep basins, in which subsidence and accumulation of sediments make for considerable sequences; these are separated by areas in which subsidence is moderate or even completely absent; and by several distinct, uplifted structures. The type of activity changes from place to place with time; the only area which is known to consistently subside is the southern Dead Sea basin (Horowitz & Horowitz 1990), which is also the locality where most information is available, from numerous deep oil drillings. However, even the boundaries of this basin are changing over time, with differential movements of its constituting blocks.

Various authors suggested five cumulative stages in the tectonic development of the Jordan Valley. The relevance of the two earlier stages to the history of the depression is rather disputable, while the others are much clearer, though even their mechanism is far from agreed by all investigators. Initially, there was supposedly delineation of the area occupied by the present Rift, which may have already begun in late Precambrian times; this was followed by folding of the Syrian Arc during the late Mesozoic–Cenozoic. Three successive phases, the Embryonic Oligocene through early late Miocene "synclinal" subsidence, the late Miocene–early Pliocene Eritrean faulting and the Quaternary Levantine faulting, then accomplished the process of the Jordan Rift formation.

9.1 TECTONICS OR EROSION BASE LEVEL CHANGES?

There are two principal periods concerning the Jordan Rift Valley, in terms of its drainage system erosion base level; up to the end of Palynozone QI, when the region was for most of the time drained to the sea, while from QII times until the present day an endoreic, internal drainage system was formed, whose terminal base level is the Dead Sea. During the first period, global sea-level changes controlled deposition and erosion, on top of which the effects of structural disturbances are superimposed. The second phase is principally affected by climate changes, also of a global nature; humid periods are characterized by high lake levels, slowing down erosion and promoting sedimentation in both the Rift Valley lakes and in wadis leading to them, while dry climates show opposite trends. As before, tectonic movements cause the superimposed effects. It is therefore our intention here to connect phenomena observed in the Jordan Rift, first with global base level data, be they transgressions, regressions, small-scale sea-level oscillations or changing climates, as the case may be in the area and period under discussion. Only if these cannot give satisfactory explanations, or if global data contradict the observations in the Jordan Valley, will tectonics be called on for help.

Although the above may sound logical, the common tendency and practice in local geology is to apply "tectonic" causes to numerous events. For example (with no references, for obvious reasons), many conglomerate beds were thought to indicate taphrogenic processes; some do, no doubt, but most do not, being a result of fluviatile activity under eustatically changing erosion base levels. Whenever marine sediments were recorded inland, "subsidence" was invoked, while in most cases these only manifested a global rise in sea level. Conversely, declines in sea level are too frequently regarded as "tectonic uplift", and the list is still long. Consequently, if one adds together only a small part of the "tectonics" suggested to have affected the Jordan Rift Valley, it should have trembled dramatically every day, in yo-yo fashion. As this was certainly not the case, we should exercise some prudence before "tectonizing" almost everything.

Starting from the Oligocene, until the termination of Palynozone QI (Fig. 9.1.1) several principal periods of complex sea-level rises and falls are known (Vail et al. 1977). The Oligocene transgressive cycle, its duration almost 15 million years, is characterized by low sea levels both at its beginning and end, separated by the

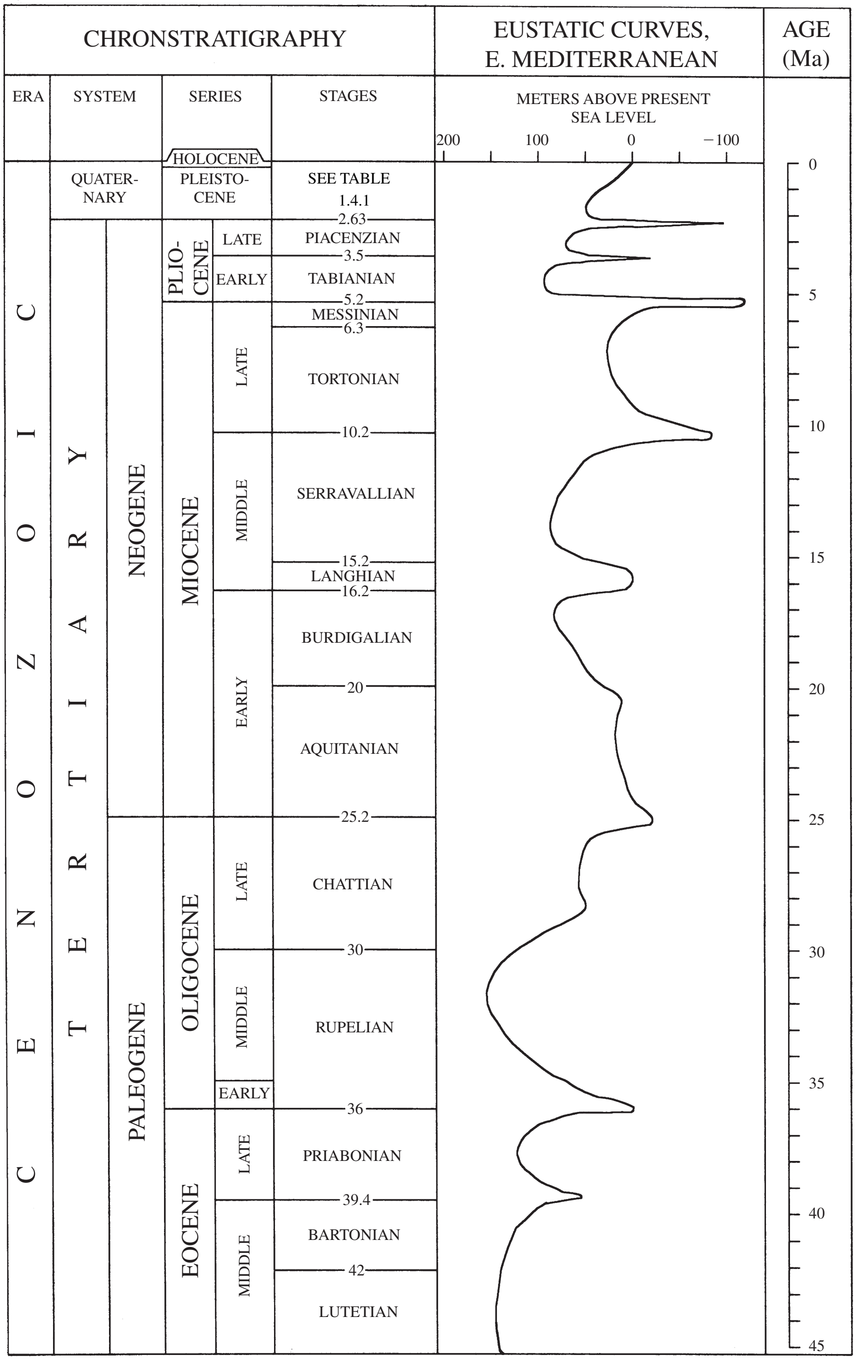


Figure 9.1.1. Eustatic sea levels of the eastern Mediterranean. Modified from Vail et al. (1977) according to local data.

high seas of its middle part. The transition from the Rupelian to the Chattian is marked by a considerable drop in sea level. This regression was (Vail et al. 1977, p. 85; Martinotti 1981, p. 31) "The greatest fall of sea level since Triassic times … (which had occurred) … at the beginning of the late middle Oligocene". This time interval witnessed the post-Eocene erosion of the southern Levant, followed by deposition of the middle Oligocene sequences of En Gev and the southern Dead Sea. The subsequent regression, which continued into the early Miocene, had almost five million years to remove much of the middle Oligocene sediments, occasionally also affecting older formations.

The one thing this cycle cannot explain is the excessive thickness of Oligocene sediments in the Jordan Rift Valley, as compared with the rest of the southern Levant, which calls for some differential subsidence. This subsidence is probably the reason why the sequences were accumulated, and were subsequently preserved from erosion by being sheltered in the low lying regions. The uninterrupted subsidence is concluded from late Oligocene and early Miocene (Ma) rocks found in the southern Dead Sea basin, comprising a continuous section. Although not perfectly continuous, the En Gev–Wadi Taiyiba sequence is also much better developed in comparison to other Oligocene occurrences outside the Jordan Valley.

The Miocene, again a complex transgressive cycle, encompasses three main stages of rise in sea levels (Vail et al. 1977, Haq et al. 1988, Buchbinder & Zilberman 1997, B. Derin, Consulting and Geological Services Ltd. Ramat Gan 1997, pers. comm.). The earliest rise occurred in the early middle Miocene, late Burdigalian, N7–N9, followed by a regression during N10–N12 and another transgression during the Serravallian, N13–N14. The third rise in sea level, less extensive than the middle Miocene ones, was during the Tortonian, N15–N16.

The twin middle-Miocene events explain the deposition of the Hazeva Formation outside the Rift limits, with all its characteristics (Calvo et al. 1997). The picture is not as clear with the Herod, in which a middle unconformity is not apparent (Shaliv 1991). This could be explained by the location of Herod exposures, which are within or very close to the central Jordan Valley subsiding basin. It may well be that accelerated subsidence caused accumulation of a continuous sequence. Similarly, the Hazeva sequences inside the southern Dead Sea basin are continuous (Horowitz & Horowitz 1990), which calls for subsidence as a probable explanation. Another possibility may result from the different histories of the Mediterranean and Persian Gulf domains during the Miocene. It may well be that the regression observed for N10–N12 in the former, did not affect the latter at all. Since regions crucial for demonstrating this or another hypothesis, such as the Barada sequence, the Herod's correlative outside the Rift, cannot at present be accessed by us, this point may need future clarification.

The subordinate Tortonian ingression is known only in the central Jordan Valley (Sneh 1993, Raab 1998), while to the south it reached as far east as Be'er Sheva, filling up a canyon (Buchbinder & Zilberman 1997). At approximately the same time, deep basins were formed within and outside the Jordan Valley limits, which certainly call for a tectonic cause, the Eritrean faulting (see below). This phase was responsible for barring the direct westward connection of the Dead Sea to the Mediterranean, but also for forming a new connection of the Jordan Valley with this sea via the Yizre'el Rift. Another new connection, to the north, linked the Hula and Beqa'a with the Mediterranean through the Tripoli graben, following disconnection of the Oligo-Miocene system from the Persian Gulf domain.

The Pliocene sea could subsequently access the Jordan Valley while it was rising in Tabianian (Palynozone Pa) times, depositing the Bira and Sedom formations and all their associates. The drop in sea level at the Pliocene–Quaternary boundary ceased this deposition, while causing the formation of channel systems toward the retreating Mediterranean. The subsequent rise of the Calabrian–Sicilian sea clogged these channels with predominantly fluviatile sediments during Palynozone QI times.

Next came the formation of the Jordan Rift in its present-day shape of an inland system, in Palynozone QII times, a stage which certainly requires tectonics. This Levantine faulting, which is still active, controlled the locations of the lakes and rivers of the Jordan system, their extent and characteristics resulting from a complicated interplay of changing Quaternary climates and structural disturbances (Horowitz 1988, 1989b).

9.2 INITIAL STRUCTURE? THE "GEOSUTURE" CONCEPT

Ever since Lartet (1865) noted that different rock formations are exposed east and west of the Jordan Rift Valley, various explanations have been offered for this phenomenon. Bender (1974a, pp. 121, 124–125) proposes the existence of a longitudinal zone of crustal weakness ("hinge line", "geosuture"), of approximately similar direction and location as the present day Jordan Rift Valley, but predating the rifting by a very long time. Zilberfarb (1987) recognized the southward continuation of this belt into the area occupied by the Gulf of Aqaba, where metamorphism and mineralization are considerably more extensive than outside its limits. Subsequently, geophysical investigations indicated that the underlying crust is indeed significantly different east and west of the Jordan Valley along at least some 300 km (see Chapter 8), thus justifying Bender's claim (Fig. 9.2.1), which rests on the following criteria:

(1) The nature of Precambrian structural elements, especially the dikes and joints systems in southern Transjordan. Numerous dikes, particularly those made of granites–aplites, are directed parallel to the southern Arava Rift.

(2) The number of Precambrian dikes increases to up to 30% of the total volume of basement rocks, when approaching the Rift zone from the east at any locality in Transjordan. Bender maintains that these intrusions must have taken place in a

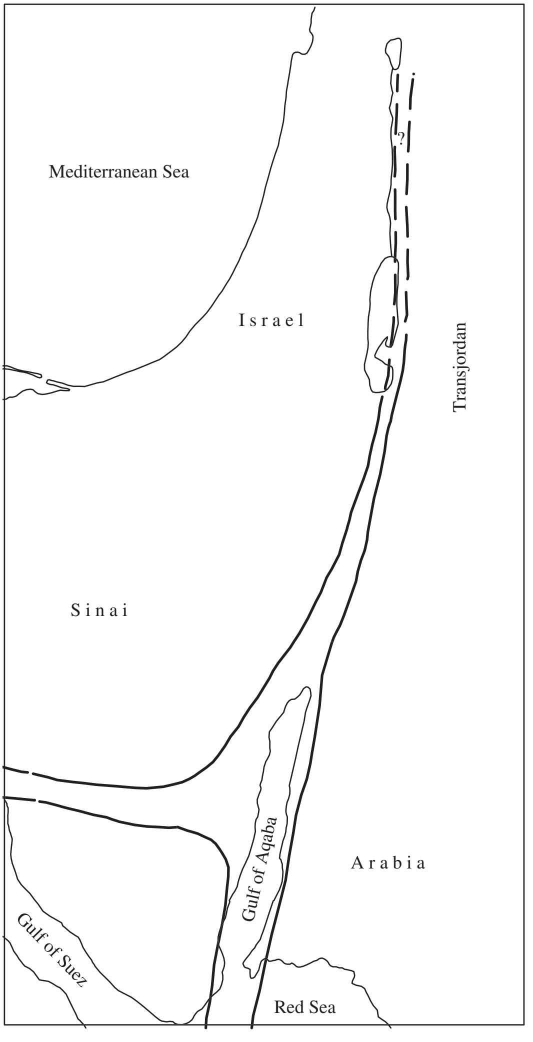


Figure 9.2.1. Outline of the late Precambrian "geosuture".

zone of structural weakness, not just elsewhere within the consolidated, rigid basement rock complex. The north–south bearing and wider distribution of dikes also extend to the Gulf of Aqaba (Zilberfarb 1978).

(3) The upper surface of the Precambrian basement complex comprises a well-developed vast peneplain at some distance to the east of the Rift. When approaching the Rift area, however, the peneplain assumes a distinct erosional relief, filled up with coarse conglomerates of early Cambrian age, in the area occupied by the present Rift zone.

(4) The occurrences of gneisses, slates, graywackes and conglomerates of late Proterozoic age, which form a "roof rocks complex" in graben structures within and in the close vicinity of the geosuture, all the way from the Dead Sea to the

Gulf of Aqaba, in contrast to the absence of such rocks away from the Dead Sea–Arava Rift on the eastern Transjordanian Plateau.

(5) The latest Proterozoic to early Cambrian was characterized by extensive quartz–porphyry volcanism, again limited to areas of structural weakness, close to the present Rift Valley.

(6) The nature of Cambrian facies boundaries and distribution of mineralization zones along the Arava, particularly of copper and manganese, also of Cambrian age, are features that could have been affected by the existence of a major paleogeographic change in the region of the present Rift, or resulting from hydrothermal activity along a structurally active zone.

(7) The rapid changes in thicknesses and facies of Triassic, Jurassic and Cretaceous sedimentary formations, when one crosses the present Rift zone, are taken by Bender as indicators of the existence of some kind of hinge zone along this region. Incidentally, these same facies changes were considered by others (see Section 10.3) as indicators of a late Mesozoic–Cenozoic left-lateral faulting along the Jordan Rift. In order to clarify which explanation is more plausible, Horowitz (1979, p. 58) and Mart (1991) showed the distribution of isopachs and lithofacies of several sedimentary groups, from Paleozoic through Senonian, west of the Jordan Rift (see also Figs. in Chapter 4). The assumption being that if there was a hinge line it would affect these formations to the west, regardless of the later faulting to the east. Indeed it was seen that such was the case, and most isopachs and lithofacies lines already ran sub-parallel to the Jordan Rift during the Paleozoic and Mesozoic, long before the Rift was formed.

Bender (1974a, p. 125) thus summarizes the geosuture concept: "The taphrogenic structural movements which initiated the formation of the Graben occurred along the pre-existing zone of weakness, and apparently started in the (?)*Upper Eocene–Oligocene* (first continental influence on the depositional environment of marine upper Eocene sediments: fine-conglomerates, ostracode-layer at the eastern side of the Southgraben)." The "Southgraben" is defined by Bender as the sector of the Rift extending from the Dead Sea to the Gulf of Aqaba.

The question, however, still remains: was this geosuture, if its existence is accepted, indeed the forerunner of the Jordan Rift, controlling its tectonic development, location and direction, or does it only explain the differences in rock formations on both its flanks? In order to answer this question, it would be useful to delineate the extensions of the above described criteria, beyond the limits of the "Southgraben". It is impossible to do so north of the Dead Sea, where the Precambrian and Paleozoic rocks are covered by younger ones, except by geophysics; however, most of these phenomena are also observed west of the southern Arava and in Sinai (Picard 1943, Bentor 1961, Bartov 1994), as well as along the Gulf of Aqaba.

Deep channeling of the late Precambrian peneplain, filled by Cambrian coarse clastics, is very distinct to the west of the southern Arava, at Timna (Karcz & Key 1966, Segev 1984). Gneisses and slates are present west of the southern Arava, near Elat (Bentor 1961), extending southward down to southern Sinai. These rocks are also exposed over wide areas in Saudi Arabia (Bartov 1994), making a belt several tens of kilometers wide, which runs parallel to the Red Sea. It seems that the extensive distribution of such rocks precludes them from being indicators of a narrow geosuture.

Conglomerates of late Proterozoic age are known from the southern Dead Sea all the way to southern Sinai (Bartov 1994). When approaching the northern tip of the Gulf of Aqaba, however they leave the Rift Valley and turn southwestward, so that they are missing at the eastern side of the Gulf. A similar pattern is also seen for the distribution of quartz–porphyry (Segev 1984) and copper–manganese mineralizations (Bentor 1952). The Volcano-Conglomeratic Complex is also found in Saudi Arabia and the Eastern Desert of Egypt, several tens of kilometers from the Red Sea. Graywackes are known from the subsurface in Makhtesh Ramon, 25–30 km west of the Rift Valley (Weissbrod 1981).

It seems quite clear that, if the criteria set by Bender and Zilberfarb for delineating a late Precambrian geosuture are valid (which personally I do not doubt), its coincidence with the Arava and Gulf of Aqaba is striking. However, this belt bifurcates at the northern end of the Gulf of Aqaba (Fig. 9.2.1), with an offshoot which turns southwestward into Sinai, in a manner broadly parallel to the Syrian Arc folds. The geosuture may have controlled the shape of this fold belt, as suggested long ago by Picard (1939, p. 54) and others. The geosuture could have controlled the formation and direction of the southern sector of the Jordan Rift Valley and the Gulf of Aqaba, but certainly not the northern extension of the rift system. It seems that the Arava–Sinai sector should be termed "hinge line", while the southward extension is indeed a geosuture.

A note concerning the age of rifting should be inserted here. Bender (1974a, p. 125) maintains that isopachs of late Cretaceous sedimentary formations indicate "an area of subsidence within the present Rift zone". No such claims, to the best of my knowledge, were made by any other investigator. The general approach, even of authors assuming an early formation of the Rift such as Freund et al. (1970) or Quennell (1959, 1996), is that this process postdates the Turonian.

9.3 SYRIAN ARC

The Syrian Arc is discussed in detail in Section 4.8, which ends with the statement: "This situation does not preclude theoretical reasoning about the stress fields, involved in the formation of the present fold pattern, during different periods, and about the question of whether this pattern developed independently from the mechanism that brought about the formation of the rift system (Red Sea, Gulf of Suez and Jordan Valley), or whether it was a reaction to, or part of the stresses and motions that formed the rift system." Numerous researchers have answered this question positively (myself, for one, in several publications); others went even further to regard both the Syrian Arc folding and Jordan Rift formation as the outcome of a single cause (Freund et al. 1970, and many of their followers).

This was concluded for several reasons. Some of the synclines crossing the Jordan Valley are accentuated by younger movements; structures both east and west of the Rift suffered considerable uplift in late Cenozoic times, concurrently with subsidence of the Rift's floor. Angular unconformities are observed on anticlinal flanks for Oligocene and Miocene sediments intimately connected with the Rift Valley, such as the En Gev suite, and the Herod and Hazeva formations; these formations are considerably thicker in synclines of the Syrian Arc, and are sometimes known only from such structures. The paths of major Oligocene and Miocene drainage systems principally follow the structural lows of the Syrian Arc, especially the larger synclinoria. On the other hand, some of the higher structures are cut by antecedent Oligocene and Miocene rivers, which seems to indicate concurrent folding.

It seems to me now that I was mistaken, and that the Syrian Arc folding should be regarded as an independent tectonic phase which commenced some time in the late Cretaceous, ending its activity as such toward the end of the Eocene. It is of course true that its structures, both anticlines and synclines, were accentuated at later dates, but these should be connected with differential uplifts caused by different tectonic regimes. The reasoning behind this suggestion is as follows.

The Jordan Rift cuts through various structures of the Syrian Arc, the cutting angles varying at different locations (Fig. 3.1.1), so that the two systems do not represent a common stress bearing or tectonic lineament. This is seen clearly from the southern end of the Dead Sea northward, and is even more apparent north of the Hula Valley (Dubertret 1962, Quennell 1996). The shape of the Syrian Arc fold belt is controlled by an ancient hinge line (Shahar 1994, and see above) delineating the boundaries of the African–Arabian craton crystalline basement (Rybakov et al. 1999), while the Rift partly follows this feature, which is clearly seen south of the Dead Sea, but continues southward, in the Gulf of Aqaba, along the geosuture, while the hinge line turns west of the Rift (Bartov 1994). The north–south direction of the Rift is very prominent when Pliocene and Quaternary features are concerned, less so with Miocene and Oligocene ones. However, the principal depocenters where subsidence attains greater figures are all aligned longitudinally, and also for the latter periods.

It remains then to explain the uplifts which affected the Syrian Arc structures since the Oligocene, possibly until the present day. The striking but frequently overlooked characteristic of these uplifts is that most, if not all, affected previous anticlines and synclines alike. Such is the case with the gradual Miocene uplift of the northern Negev (Zilberman 1992), which finally barred the way of the Hazeva river system, or the northeastern highlands, which ended the Herod–Barada drainage to the Persian Gulf. The Miocene slow, continuous uplift caused gradual breaching of anticlines by rivers, but this drainage pattern ended when the uplift culminated beyond a certain level. A similar process is seen during the Quaternary from QII times onward, when most older structures were elevated, regardless of whether they were anticlines, synclines or even previously formed Eritrean rift valleys (Horowitz 1979, p. 54).

The combination of different structural bearings and uplifting affecting also low structures, means we must see the Syrian Arc and the Jordan Rift Valley as separate tectonic provinces. It is however quite understandable why combining the two was so easy. One readily observes uplifting of anticlines, while the behavior of synclines, mostly buried under younger sediments, is considerably more obscure. It is therefore only natural to draw the activity of the conspicuous and well-known Syrian Arc folding up to the present day. It should however be clear that subsidence along the Jordan Valley must have been compensated by uplift affecting both flanks of the Rift, most probably since Oligocene times.

The one feature which seems crucial for separating the structures formed by the Syrian Arc folding from later movements is the facies and thickness changes of Senonian formations. These show two typical facies, one for rocks deposited on the anticlines, the other for those laid down in the lower structures (Bentor & Vroman 1991, and see Section 4.5 and Fig. 4.8.3). Thus when a synclinal Senonian occurs on top of elevated structures, such as the Karak anticline or the Amman region, uplift is undoubtedly younger than the Syrian Arc. Unfortunately, Senonian rocks are rarely preserved on many of the high structures, in which case distinction is very difficult.

It seems that reconsideration should be given to the ideas of Picard (1943), summarized by Bentor (1960, p. 9): "The major phase of compressional folding ended in late Eocene and was followed by a period of tensional stress."

9.3.1 Late Eocene

The late Eocene is a transition time from the tectonic domain of the Syrian Arc, to the Embryonic stage of the Jordan Rift Valley formation. Deposition in early to middle Eocene times clearly follows structures delineated by the folding, where considerably thicker sequences were deposited in synclines, as compared with the thinner sedimentation, occasionally entirely missing, on the higher Syrian Arc anticlines (Bender 1974a,b, Benjamini 1984, Sneh 1988, Sneh et al. 1998a). Considerable uplift is recorded in the entire southern Levant for the transition from the middle to the late Eocene, the latter rocks always laid down on a clear unconformity. However, the pattern of late Eocene outcrops does not depend, in any way, on a lineament that could be attributed to the Jordan Rift Valley (C. Benjamini, Department of Geology and Mineralogy, Ben Gurion University of the Negev 1999, pers. comm.).

The middle to late Eocene uplift is very well documented in the southern coastal plain of the southern Levant, where a considerable channel was cut at this time, some 700 m deep into the elevated terrain, penetrated by the Haruvit boreholes (Derin 1976) located close to the coastline between Rafah and El Arish. This uplift is indeed manifested in the entire southern Levant, at least down to the Gulf of Aqaba, where the late Eocene transgression deposited rocks which overlie early to middle Eocene and older formations over a pronounced unconformity (Sakal et al. 1966; Bender 1974a, p. 88; Benjamini 1984).

The subsequent global regression, at the boundary of the Eocene and Oligocene periods, caused erosion of most of the late Eocene formations, so that the exposures in southern Israel and the Arava, the areas near Lake Kinneret, and possibly also other localities not yet studied in sufficient detail, consist only of early late Eocene rocks, Foraminifera Zone P15 (Lipson 1971, Benjamini 1980). Younger Eocene rocks, of zones P16–P17, are known only from the subsurface of the coastal plain, such as the Haruvit boreholes (Derin 1976), where they line the bottoms of the erosion channels, consisting of highly terrigenous sediments. These overlie several tens of meters of brown, red and black shales, of uncertain age, which may represent the erosion between P15 high sea level and the younger, latest Eocene rise which inundated only the lower reaches of the Haruvit canyon.

It appears from the above situation that a significant drainage system was developed toward the retreating Tethys, possibly also the Indian Ocean or the Red Sea (if it existed then), following the uplift and regression. The crucial question is, however, whether the late Eocene rocks of the Arava and Gulf of Aqaba were deposited by a sea that arrived from the south, or by the Tethys. Until this problem is resolved, it is quite difficult to assess whether the late-middle Eocene uplift should be regarded as the final "kick" of the Syrian Arc or is in fact the initial precursor of tectonic movements connected with the Jordan Rift Valley, which could be the case should the sea have come from the south. The last possibility, if verified, could attest to the existence of a "proto-Red Sea" in late Eocene times. It seems appropriate, at the present state of knowledge, to refer to this uplift as a transition phase, a term which would be justified no matter what future research reveals.

9.4 EMBRYONIC STAGE

The Embryonic stage of the Jordan Rift Valley is defined as the period during which subsidence along the present-day sector of the Rift discussed here (from Gav Ha'Arava northward to the Hula) is accelerated, in comparison with neighboring regions, but with no signs of faulting, not even detected by geophysics (Bruner et al. 2000). This definition naturally does not preclude faulting outside the region specified, particularly to the south, as discussed below. The subsidence is superimposed on previous Syrian Arc structures, causing the formation of folded basins along the Jordan Rift Valley. In contrast, other synclines of the Syrian Arc do not show any conspicuous subsidence during this stage.

Subsidence along the Rift affected both low and high structures of the Syrian Arc, such as the synclinal En Gev region east of Lake Kinneret and the previously elevated southern Dead Sea region. The Embryonic subsidence was accompanied by differential uplifts on both sides of the Rift (Garfunkel & Horowitz 1966; Bender 1974a, p. 102; Zilberman 1992). Since the subsidence is seemingly synclinal and the uplift is more apparent in the anticlines, many investigators consider these movements as continuation phases of the Syrian Arc folding (see also above). It is quite difficult indeed to decide, since both opinions may be correct: the possibility that further, somewhat differential Syrian Arc activity accentuated subsidence in two locations along the Rift, at Lake Kinneret and in the southern Dead Sea region, cannot be entirely ruled out.

These subsidence phases are seen here to represent the Embryonic stage of the Jordan Rift formation, because I do not see any reason to regard their locations as merely accidental. Ever since the Oligocene, but in particular during the early and middle Miocene, accelerated subsidence clearly affected the Jordan Valley lineament in a considerably more pronounced way than the neighboring areas east and west of the Valley, whose dominant tendency during much of these periods is uplifting, which had affected both synclines and anticlines of the Syrian Arc.

Most of the data presented here are based on pollen analyses of numerous boreholes drilled into the Jordan Rift Valley fill, many of them in recent years. The locations of these boreholes are given in Fig. 6.1.1, and the occurrences of the Neogene palynozones within the drillings are shown in Fig. 6.1.2. Rates of subsidence during the Neogene palynozones are calculated in Fig. 9.4.1.

Oligocene and early Miocene subsidence of the Jordan Rift Valley lineament is known only in the southern Dead Sea and central Jordan Valley basins, separated by a conspicuous watershed dividing the Persian Gulf and Mediterranean systems. No information is available for the Hula basin, where rocks of this age are neither exposed nor penetrated by drillings. The deepest borehole in the Hula, Notera 3 (Horowitz & Horowitz 1985) reached to Palynozone Mb, but not its base. It seems however that the Hula was separated to the north by a watershed comprising the Lebanon–Hermon high structures. This is concluded from occurrences of Jurassic pebbles originating in these highlands, which are found to the south near Lake Kinneret, in Herod sediments of middle Miocene age. Another, much clearer watershed is at Gav Ha'Arava, separating the Mediterranean and Red Sea domains (Garfunkel & Horowitz 1966).

9.4.1 Oligocene

The most intriguing period in the evolution of the Jordan Rift Valley is its inception during the Oligocene, a rather elusive chapter in the geological history of the southern Levant. This is the time of last emergence from under the waves of the Tethys Ocean, which had submerged the region until nearly the end of the

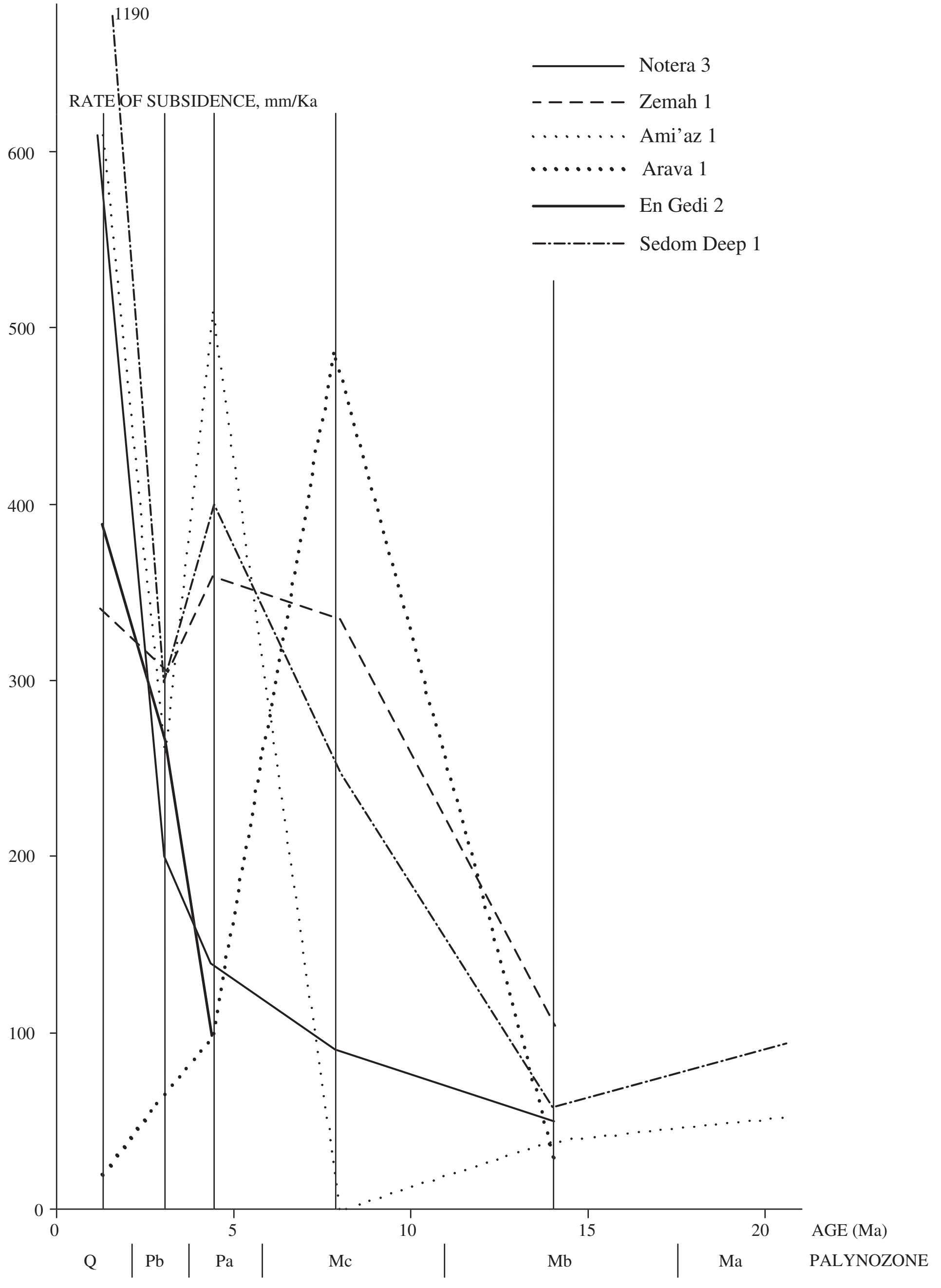


Figure 9.4.1. Late Cenozoic rates of subsidence of different basins along the Jordan Rift Valley, according to apparent thicknesses of palynozones from the specified boreholes.

Eocene period (Sakal et al. 1966; Bender 1974a, p. 88; Benjamini 1984), almost continuously since the early Mesozoic (see Chapter 4). The Oligocene uplift of the southern Levant is only the tip of the iceberg, a process which affected the entire Arabo-Nubian Massif at that time (Picard 1939, Omar & Steckler 1995, Steininger & Rögl 1996). The effects of this uplift were felt far beyond the limits of the region discussed, manifested in a total lack of marine Oligocene sediments at least as far as northwestern Syria (Krasheninnikov 1994), and their partial absence in the Mediterranean offshore (Derin & Reiss 1973, Martinotti 1981).

The uplift began, naturally, with large-scale, regional erosion which not only hindered any considerable accumulation of Oligocene deposits, but also wiped out large bodies of previously laid rocks over a period which lasted at least 15 million years. This is evident by the scarcity of late Eocene rocks and the close to complete absence of latest Eocene and Oligocene rocks in the entire region (C. Benjamini, Department of Geology and Mineralogy, Ben Gurion University of the Negev 1999, pers. comm.). The prevailing erosion caused the development of a vast peneplain at this time (Zilberman 1992), which is still preserved at numerous locations, such as the central Negev highlands. Thus, since very little is left of the Oligocene sediments, conclusions must be drawn from the substrate over which Miocene and younger rocks have been laid down unconformably, which was formed mainly during middle to late Eocene through Oligocene times. The main difficulty involved is the length of this hiatus, which in many cases is quite hard to define with any accuracy.

Positive evidence for the Oligocene uplift of the Arabo-Nubian Massif, accompanied by rift formation taphrogenic processes, comes from the southern Arava, just south of Gav Ha'Arava and extending southward along the Gulf of Aqaba, where several isolated outcrops of the Miocene Raham Conglomerate occur (Garfunkel et al. 1974; Bender 1974a, p. 88). Such initial rifting processes are also known from the Gulf of Suez and the Red Sea, dating to the early Oligocene (Robson 1971, Omar & Steckler 1995, Purser & Bosence 1998).

The Raham Conglomerate comprises up to 300 m of mainly coarse clastics, with several finer, clayey, silty or sandy horizons, in which a shallow marine intercalation was detected in a few outcrops, its maximum northern extension reaching almost to Gav Ha'Arava, at Jebel el Khureij, 65–70 km north of Aqaba. The Conglomerate rests, over angular and erosional unconformities, on top of a variety of older rocks, the youngest of which is of Eocene age. The coarse components comprise pebbles originating in almost the entire geological sequence, from the Precambrian onward, but is dominated by Eocene constituents.

Garfunkel et al. correlated the Raham Conglomerate with the Hazeva Formation and the "Lower Syntectonical Conglomerates" of the Dana, attributing it a different name since it was deposited in a drainage system leading to the Red Sea, rather than the Mediterranean base level of the latter two. They maintain that (p. 62) "The occurrence of clasts from different parts of the section requires that they were all exposed side by side, which could result only from tectonic uplifting and subsequent stripping of the overlying sediments". This calls for a structural relief of at least 1,200 m in the Taba region, south of Elat, where Precambrian and Eocene pebbles occur together within the Raham Conglomerate. Garfunkel et al. suggest that the relief could even attain up to 2 km. They also conclude that (p. 62) "The angular unconformities at the base of the Raham Conglomerate suggest still older (than Miocene) tectonism, but since the various outcrops are probably not strictly contemporaneous, an older and distinct phase cannot be demonstrated".

Despite the above reservation, it seems clear that the Raham Conglomerate was deposited in a northward extension of the Red Sea Rift during the Miocene, an extension which must have been formed prior to that period. The extensive body of Eocene marine limestone pebbles comprising the Raham, does not leave any other possibility but the Oligocene as the time of faulting and beginning of erosion. The composition of the Raham Conglomerate pebbles indicates that during the Oligocene the regions bounding the Gulf of Aqaba were uplifted and eroded.

The area now occupied by the Gulf of Aqaba, its Oligo-Miocene predecessor conveniently termed the "ancestral" or "proto-" Gulf of Elat (Aqaba) by Garfunkel et al. (1974), including its northern extension up to Gav Ha'Arava, served as an erosion channel leading southward to the Red Sea. No definitely Oligocene rocks are reported from the Gulf of Aqaba itself, which indicates that during most of this period the downfaulted "proto-bay" was eroded. Shallow marine Oligocene sediments are known from Midyan, at the junction of the Gulf and the Red Sea (Dullo et al. 1983) on the Saudi Arabian coast. Even if the Oligocene sea had invaded the proto-Gulf, every clue was removed by subsequent erosion (or not identified as such yet).

Bender (1974a, p. 88; 1974b) presents a somewhat different view, by regarding the Lower Syntectonical Conglomerates as of Oligocene age, possibly extending to the early Miocene (although, for lack of diagnostic fossils, he puts a question mark on this assignment). If this view is accepted, then the Lower Syntectonical Conglomerates could be broadly correlative to other, possibly Oligocene sediments in the Negev, such as the Zefa Formation and the Abu Treife Series. At any rate, based on this assignment Bender also proposes that faulting and rift formation already began in Oligocene times.

The northernmost area where pre-Raham faulting was reported is the outcrop at the Nahal Raham graben, somewhat more than 20 km north of the city of Elat. The Oligocene faulting apparently did not extend further north, certainly not into the sector of the Jordan Valley discussed in this book, since in no place Miocene sediments are known to overlie faults, not even of subordinate magnitudes. The Oligocene tectonic lineament did however affect the regions to the north, which subsided in reaction to faulting to the south, continuing its bearing northward in a pattern of folding, or possibly sagging over faults in depth. The Oligocene folding is quite clear from its outcrops east of Lake Kinneret, seen by the angular unconformities in the En Gev section (Michelson 1972).

The Oligocene uplift and faulting to the south, which is known down to the Gulf of Aden (Nichols & Watchorn 1998), continued northward in a pattern of gentle, large-scale undulations, which defined the ensuing hydrographic network of the southern Levant at that time. The watershed at Gav Ha'Arava (Fig. 7.1), which separated the central Negev lowlands from the Red Sea system, still maintains its role today in a manner similar to Oligocene times. Other morphotectonic features were later on faulted and breached, such as the watershed at Marma Feiyad, then separating the lowlands formed during the Oligocene. These lowlands were the now elevated central Negev, connecting the southern Dead Sea and northern Arava to the Mediterranean via the Zefa–Abu Treife channel and the Ramat Matred–Nizzana synclinorium; and the central Jordan Valley–Damascus synclinorium, drained to the Persian Gulf.

It is quite interesting to note that the Ramon anticlinorium, which is now the highest structure in the central Negev, was considerably lower in relation to the northern Negev during the Oligocene and probably much of the Miocene. This is evident from the meticulous studies of Plakht (1996, 1999), who mapped terraces in both the Ramon and Hatira erosion cirques, indicating that while the latter was already deeply cut before the Miocene, the former bears only Quaternary terraces. This also emerges while analyzing pebbles from the younger formations in the region (Garfunkel & Horowitz 1966, Baer 1981, Zilberman 1992); the Miocene Hazeva Formation sediments contain components derived from the Hatira, while those originating in Ramon are known only from the Pliocene Arava Formation.

It seems that subsidence continued into the late Oligocene in the Jordan Valley, as evidenced from the lower part of Palynozone Ma, well developed in the Sedom Deep 1 borehole, at the subsiding southern Dead Sea basin. It is only natural to assume that while subsidence continued in the lower structures of the Jordan Rift, uplift of the neighboring regions also occurred, causing further erosion. Although the late Oligocene tectonic activity of the Jordan Valley region went along the same lines as before, further uplift of the central Negev (and possibly also southern and central Transjordan) caused a conspicuous change in hydrography. The uplift propagated northward, so that in the late Oligocene the Ramon structure became higher than the northern Negev region. This process made the northern Negev a main path for the subsequent Hazeva river system connecting Transjordan to the Mediterranean.

This change is clearly seen from the distribution of the sediments of the Hazeva Formation (Garfunkel & Horowitz 1966, Zilberman 1992), silting up the Hatira erosion cirque and occurring on top of many of the northern Negev higher structures. The late Oligocene northward transition of the principal drainage systems is also clear from an analysis of their outpourings into the Mediterranean. A thick sequence of middle Oligocene sediments, up to 700 m is located in Haruvit, where Miocene sediments are subordinate (Derin 1976). In contrast, considerable Miocene sections more than 900 m thick are developed to the north, in the Gaza region, overlying only subordinate Oligocene sequences, in the order of less than 100 m (Derin & Reiss 1973).

The Oligocene tectonic activity thus has four principal aspects: regional uplift, more accentuated to the east and south, which affected areas on both sides of the Jordan Valley; faulting of the southern parts of the Rift Valley, but only those connected with the Red Sea system; large-scale, shallow undulations which determined both regional and local hydrography; and extended subsidence at least in two localities along the present-day Rift, the southern Dead Sea and Lake Kinneret regions.

The exact age of the Oligocene uplift is hard to determine, due to the paucity of sediments of this age in the critical regions. The task is somewhat easier where marine sediments occur, from boreholes drilled in the southern coastal plain of Israel. This region is quite far away from the Jordan Rift Valley, so that the tectonic processes may have affected it somewhat later or in a slightly different way than those known from the Rift. Martinotti (1981) studied in detail the Oligocene of the coastal plain, which in most localities does not exceed several tens of meters in thickness (except for Haruvit), while in others it is altogether missing. He states (p. 31): "An uplift characterized the area which today comprises the continental shelf in the west, the coastal plain, and the foothills belt in the east, as documented by the above unconformity." (Sediments of the early-middle Oligocene *Globigerina ampliapertura* Zone unconformably overlie the lower part of the *Globigerinatheka semiinvoluta* Zone, of earliest-late Eocene age.) Martinotti thus concludes that the uplift "occurred no later than the earliest middle Oligocene … Was intense and took place within a short time, as attested to by the subsequent deep erosive channeling in Ashdod 1" (borehole in the coastal plain, some 55 km due west of Jerusalem, where an unusually thick sequence, 282 m of *Globigerina ampliapertura* Zone were penetrated). The history of Oligocene channeling in this region is also discussed by Neev (1960, 1979) and by Gvirtzman & Buchbinder (1978), who arrived at similar conclusions, but without the accuracy given by the detailed microfaunal studies.

It is not easy to assess the magnitudes of the Oligocene uplift, due both to its differential nature and the paucity of sediments of this period. Notably, all Oligocene occurrences, in outcrops as well as subsurface, both those that are safely attributed this age and the questionable ones, contain abundant quartz sands. The youngest sandstones which could have served as provenance in the southern Levant are of early Cretaceous age. These are covered by a variety of carbonates extending from the late Cretaceous through the late Eocene, attaining an order of 1 km in thickness (see Chapter 4). These should have been stripped off by the combination of uplift and erosion, at least in some localities to expose the underlying sandstones by the middle Oligocene, which gives an idea of the uplift involved. Similar orders of magnitude are also seen from the depths of pre-middle Oligocene buried canyons of the coastal plain. In one of these, penetrated by the Gaza 1 borehole, an early Cretaceous boulder was encountered within the middle Oligocene sediments (Martinotti 1981).

The amounts of Oligocene uplift can be indirectly, but less accurately, determined also from the composition of Miocene formations filling up the canyons cut previously, such as the Raham Conglomerate containing Precambrian through Eocene pebbles mentioned above. The inaccuracy may result from continuation of the uplift processes into the Miocene. The composition of early and middle Miocene clastics is discussed below, in Section 9.4.2.

Neev (1960, p. 18; 1979, p. 132) proposed that "the great regional uplift and the deep erosion which accompanied it" were followed by "gradual and regional subsidence". It seems that this "subsidence" was indeed the middle Oligocene global transgression, which submerged the canyons cut previously, and did not have anything to do with local tectonism. This is concluded from the continuation of erosion and channeling during the late Oligocene–early Miocene, which caused further deepening of the canyons (Martinotti 1981, Buchbinder & Zilberman 1997), and occasionally also "cleaned" them of their Oligocene sediments.

Subsidence of several hundred meters can be seen east of Lake Kinneret from the considerable accumulation of Oligocene sediments in this locality and the angular unconformities within the sequence. When exact figures are concerned, subsidence of the southern Dead Sea basin is more problematic. Some 100 m of middle Oligocene, and particularly several hundred meters of late Oligocene rocks were encountered in the Sedom Deep 1 borehole. The true thickness cannot be calculated because the base was not penetrated and dips are not known. At any rate, this gives an idea of the order of subsidence in this region during that time.

Estimates of rates of subsidence for the Oligocene sediments in the southern Dead Sea basin are in the order of 50–100 mm per 1,000 years (Horowitz 1987c, 1996a). These calculations took into account the entire sequence of Palynozone Ma, since it is quite impossible to differentiate its late Oligocene from the earliest Miocene part. The Oligocene sections near Lake Kinneret are of the same order of thickness as those from the Dead Sea, so similar rates of subsidence are also concluded for this region. It should be noted that although the Oligocene sequences in both localities are hundreds of meters thick, the actual rates of accumulation are quite low, since the time involved is relatively long, almost 15 million years for the middle Oligocene through Palynozone Ma interval. This provides some scale for the tempo of tectonic activity during the Embryonic stage of the Jordan Rift Valley, which was slow but continuous. The figures advanced above may in fact be somewhat larger, before compaction of the sediments took place, but probably remain in the same order.

9.4.2 Early and middle Miocene

The early and middle Miocene saw a continuation of these trends, except for a considerable change in the hydrography of the Negev. The central Negev channel became practically inactive, and the Hazeva system drained to the Mediterranean through the Dimona–Be'er Sheva area, some 30–40 km to the north of the former Oligocene outlet, as discussed above.

Except for the Jordan Valley, considerable subsidence is also known for this time in the coastal plain and offshore, as evidenced by extensive Miocene sequences in these regions (Gvirtzman 1970, Derin & Reiss 1973), in contrast with the underlying subordinate Oligocene ones. Only the Haruvit canyon, incidentally, displays an inverted situation. This indicates a northward shift of the drainage system, which commenced already in Oligocene times (Neev 1960, 1979, Gvirtzman & Buchbinder 1978, Martinotti 1981), but reached the Jordan Valley only in the early Miocene. This shift must have resulted mainly from tectonics, causing uplift of the central Negev in relation to the northern.

The central–northern sector of the Jordan Valley, northward from the watershed somewhat south of Marma Feiyad up to the Hula region, seems to have continued to act in the Oligocene tradition. Here again, the poverty of Oligocene sediments can give only a general idea of the drainage system at that time, but the early to middle Miocene path is much clearer, flowing from west of the Jordan Valley in a general northeastward direction toward the Persian Gulf basin. As opposed to the situation in the central and northern Negev, there is no evidence in the northern province of structural changes to justify any new drainage pattern, only a continuation of the previous, Oligocene one.

Besides the paucity of Oligocene sediments, another factor that at least partially masks the real picture to the north is the Neogene volcanic activity. This caused extensive coverage by basalt sheets over vast areas, which possibly blurs the true paleogeographic pattern of the Oligocene, and also partly the Miocene. As a result, only the general nature of the tectonics for this region could be determined.

Components of both the Hazeva (Sa'ar 1986) and Herod (Michelson 1972, Shaliv 1991) formations, particularly the predominant sandstones, indicate further uplift of their provenance regions, where early Cretaceous and possibly older rocks must have been exposed. East of the Rift to the south, these supplied the Hazeva–Dana–Ghor el Qatar(?) system, while west of the Valley to the north, such formations attended the Herod. Some Jurassic pebbles found in the Herod at the northwestern corner of Lake Kinneret, indicate that the Hermon or Lebanon range was already considerably uplifted and stripped during the Miocene, and possibly earlier.

The early to middle Miocene continued uplift is clearly demonstrated by angular unconformities observed in both the Herod and Hazeva sequences. However, the most conspicuous unconformity in the Hazeva is erosional, resulting from tectonic disturbances of the eastern Mediterranean basin or from the global cooling. The Persian Gulf domain, or at least its western reaches, was only slightly affected by this middle Miocene phase, or else the central Jordan Valley was too far away from the affected ocean, which appears to explain why such an unconformity was never detected within the Herod Formation clastics.

Extended subsidence continued in the Jordan Valley basins, both the southern Dead Sea and the central to northern sector in the vicinity of Lake Kinneret, throughout the early and middle Miocene. The subsidence, possibly in pulses

(Sneh 1999), caused the formation of the intermediate basins of the Hazeva and Herod systems, respectively, in which considerable sequences of the relevant fluvio-lacustrine sediments had been accumulating, usually an order of magnitude thicker than their counterparts outside the Jordan Rift Valley limits. Questionably, the southern Jordan Valley also exhibits accelerated subsidence, where an intermediate basin in which the Ghor el Qatar sequence (if its correlation with the Hazeva is proved) was accumulated. It is however too soon to say, in the present state of knowledge, whether this was a continuation of the southern Dead Sea or an independent basin. Conserving the Oligocene tradition, no faulting is known for this period from the area north of the watershed at Gav Ha'Arava.

The rates of subsidence for Palynozone Ma are discussed above for the Oligocene, since its lower part was deposited at that time. These did not change much throughout Palynozone Mb, when the slow but steady accentuation of the basins along the present-day Rift Valley continued, in the same embryonic style as before. Contrary to the Oligocene, which was never encountered, Palynozone Mb (and the overlying ones) is also known from the Hula and the central Jordan Valley basins. Accumulation rates could therefore be calculated for all four basins (if indeed the Ghor el Qatar belongs to this system) of the Dead Sea Rift (Horowitz 1987c, 1996a), and are in the same order as during the Oligocene, 30–120 mm per 1,000 years. Notably, the lower values are from the Hula, the southern Jordan Valley(?) and the southern Dead Sea basins, 30–70 mm/Ka, while a higher number was obtained for the central Jordan Valley, at the Zemah 1 borehole, of 120 mm/Ka. The thicknesses of the exposed Herod sequences within the Rift are also somewhat greater than those of the Hazeva, not taking into consideration the Hufeira Formation which was deposited during Palynozone Mc.

The subsurface thicknesses of Palynozone Mb can only be defined in the southern Dead Sea basin, where its base was penetrated, as opposed to the situation in the Notera 3 and Zemah 1 boreholes, where drilling stopped within Mb. Even so, the sequence at Zemah 1 is the longest encountered for Mb, some 750 m. This figure is close to almost similar thicknesses measured for the exposed outcrops of its correlative Herod Formation, which indicates the considerable magnitude of the subsiding basin. Palynozone Mb thicknesses in drillings sunk in the southern Dead Sea basin do not exceed 300–400 m (Fig. 6.1.2). They are considerably thicker than the tens of meters of the Hazeva Formation, occasionally up to 150, in the Arava and northern Negev, indicating a rather restricted basin in which subsidence was somewhat accelerated.

9.5 ERITREAN STAGE

The Eritrean stage is generally characterized by tensional structures which are occasionally of considerable magnitude causing subsidence in various parts of the entire Levant, and only moderate uplift in others. The term "Eritrean lineament" was coined by Shalem (1956), to describe the fault system whose predominant bearing is N30W, parallel to the Red Sea. This general direction is prevalent from Bab el Mandeb in the south, northward to the southern Levant, extending over belts hundreds of kilometers wide on both sides of the Red Sea Rift. Horowitz (1979, p. 61) recognized the continuation of this fault system into the central and northern Levant, where synchronous activity affected large areas. The N30W trend veers gradually as one goes from the southern Levant northward, until it reaches a N45E direction in the Bay of Alexandretta. The central Levant sector, particularly the Lebanon, is characterized by directions closer to east–west in the major faults (Fig. 10.2.2).

In addition, the system is accompanied by roughly north–south oriented faults, of which the most prominent are the Yammouneh and others, in the Lebanon and further north. Such north–south directed faults must also have affected the Jordan Valley, but seem of a subordinate nature in that region, mainly because the subsequent Levantine stage is mostly characterized by a similar bearing. Most of the north–south oriented faults of the Jordan Valley are thus either Levantine or Eritrean, the latter subsequently rejuvenated, and the two are quite difficult to tell apart. Only in certain rare cases, such as the western bounding fault of the Korazim block, is the Eritrean age clear. This fault, as with the Yammouneh, is overlain by unfaulted basalt some 5 Ma old.

An array of east–west trending wrench faults (Fig 10.2.1), which particularly affected Sinai and Transjordan, was also included by Horowitz in the Levantine sector of the Eritrean stage. The activity along these "Transversal Faults" had already begun before the Miocene, so this array is dealt with in detail in Section 4.9. However all these faults were considerably accentuated during the Eritrean stage.

The inclusion of all the above fault systems under the "Eritrean" heading is based on their contemporaneous activity. The age of initial activity of the Eritrean stage is at the beginning of the late Miocene, as can be seen in numerous localities where older rocks are affected. These faults are covered by undisturbed late Miocene rocks in some localities, but such exposures are not too common; much more frequently Pliocene formations overly the Eritrean structures, so that for a long time it was generally thought that the correct age was the Miocene–Pliocene transition (cf. Horowitz 1979, p. 61). Other suggestions were proposed for the age of Eritrean activity, such as middle Miocene (Zak & Freund 1981, Hirsch & Shahar 1993), or late Miocene (Zilberman 1992), but these were all based on intelligent assumptions (sometimes correct), rather than on detailed chronostratigraphy.

More accurate datings emerged only from recent discoveries in both boreholes (Horowitz & Horowitz 1990) and outcrops (Shaliv 1991, Calvo et al. 1997, 1998), which made the exact age clearer. Excessive thicknesses of Palynozone Mc in boreholes and its equivalents in outcrops, particularly the Umm Sabune to the north and the Hufeira to the south, both covering previous faults, place the age of initial Eritrean activity at the beginning of the late Miocene. Radiogenic constraints on the age of faulting come from the central Jordan Valley, where a basalt sheet 10.1 ± 0.3 Ma old is faulted, while another, 8.4 ± 1.2 Ma old covers the fault (Shaliv 1991).

Much of the data presented here come from the palynostratigraphy of boreholes penetrating the Jordan Rift Valley fill, whose locations are given in Fig. 6.1.1, while occurrences of the Neogene palynozones in the drillings are shown in Fig. 6.1.2. Rates of subsidence during the Neogene are displayed in Fig. 9.4.1.

Two main phases of accelerated tectonic activity, first recognized by Zak & Freund (1981), each followed by a period of relative quietness, characterize the Eritrean stage. The earlier one occurred at the beginning of the late Miocene, the second at the Miocene–Pliocene transition, each of which, in its turn, abruptly changed the previous landscape. Incidentally, past inability to differentiate between the two distinct phases of activity created much debate on the timing of the Eritrean stage.

Later movements along Eritrean lines, usually of subordinate magnitudes outside the Rift limits, affected younger rocks such as the Pliocene Arava Formation (Avni 1998) to the south, or the Bira–Gesher complex of the central Jordan Valley (Shaliv 1991). It is however quite difficult to define exact ages for faulting of Pliocene rocks outside the deeper basins of the Valley since in most localities they are not covered by younger ones, and so further faulting may have resulted from the consecutive Levantine stage, which is an altogether different system (see below) but is north–south oriented just like numerous Eritrean faults. Continuous sequences of Pliocene–Quaternary sediments are known only from boreholes, where actual faults are very difficult or entirely impossible to trace, which does not make their history any clearer. Conclusions from drillings are thus based on abrupt changes in the rates of subsidence of relevant rock units, dated by both palynostratigraphy and radiogenic methods. At any rate, rejuvenation of Eritrean faults due to later activities, such as the Levantine faulting and uplift, can be seen up to the present day (Horowitz 1979, p. 41; Matmon 2000).

Eritrean differential movements affected the Jordan Rift Valley in two ways, by faulted structures formed within its proper area, usually subsiding sectors; and by a combination of subsidence within the Valley and slight uplift outside the Rift margins, which determined erosion, hydrography and the nature of its connections with the Mediterranean, which further superimposed the effects of changes in global sea level. The combination of all these factors is also accompanied by changing climates, resulting in the intricate pattern of the region during the later part of the Neogene.

9.5.1 Late Miocene

Conspicuous late Miocene (Palynozone Mc) subsidence is known from the Jordan Rift Valley from two locations, the northern Arava and the central Jordan Valley, affecting hydrography in different ways. Another trough of a similar nature is suggested at Nahal Paran (Calvo et al. 1998), in the southern–central Negev. Rates of subsidence calculated for the southern sector of the Rift in the late Miocene are in the order of 250 mm per 1,000 years at Sedom Deep 1 and up to 500 mm/Ka at the Arava 1 borehole (Horowitz 1987c, 1996a). The extensive subsidence created a trough into which large volumes of the previous Hazeva Formation were eroded, thus cutting the drainage connection of Transjordan to the Mediterranean. At the same time both regions east and west of the Dead Sea were moderately uplifted (Begin & Zilberman 1997), which only accentuated disruption of the previous Hazeva–Dana river system. This uplift most probably later prevented a direct westward connection of the Rift area to the Mediterranean even in Pliocene times, with its higher transgressive sea levels.

The central sector of the Jordan Valley was also cut off from the sea by a combination of trough formation and uplift of the Rift's shoulders. However, isolation from the Persian Gulf domain was compensated for by a newly formed path of the central Jordan Valley to the Mediterranean, through the concurrently subsiding newly formed graben in the Yizre'el Valley. This lowland permitted the invasion, however limited, by the late Miocene Mediterranean into the central Jordan Valley, which was not an endoreic system at least during part of Mc times. The central Jordan Valley trough is a double structure, as seen from geophysics (Reznikov et al. 1999), comprising two deeper sectors separated by a shallower sill, the northern one within the limits of the present Lake Kinneret (see Section 8.4).

It is quite difficult to reconstruct tectonic activity (if at all) of the Hula at that time, for lack of data. The only clear occurrence of Palynozone Mc rocks in this area is the prevalently volcanic sequence almost 500 m thick, penetrated by the Notera 3 borehole (Horowitz & Horowitz 1985), which displays a rate of accumulation in the order of 100 mm/Ka (Horowitz 1987c). On the other hand, the Hermon–Lebanon highlands were already considerably elevated at that time, as seen from Jurassic pebbles in the Herod Formation at the northwestern corner of Lake Kinneret. The Hula Mc sequence may have been deposited either in a subsiding intermediate basin or in a canyon, leading in a general southward direction, to the trough at the central Jordan Valley.

It is impossible to delineate with any accuracy the faults which were active in or along the Jordan Valley, constituting the initial phase of the Levantine stage. In the rare cases that these could be observed, such as in Belvoir (Schulman 1962, Shaliv 1991), where the Umm Sabune Conglomerate overlies faulted Lower Basalt, the fault trends conform with the general N30W bearing of the Eritrean system. It seems quite logical to assume that the deeper troughs in the Jordan Valley are also bounded by similarly trending faults. However the combination of the troughs' limited geographical extent and considerable depth would probably require some north–south oriented faulting, to allow for the extensive subsidence.

The principal uplifts that can be discerned in the first phase of the Eritrean stage are all except one located outside the Jordan Valley limits, but can only be delineated in a most general way. The two most significant for changing the landscape are the Transjordanian and southern Syrian highlands, whose uplift was sufficient to end the drainage connection of the central Jordan Valley to the Persian Gulf system; and the entire Negev, which became high enough to form a watershed between the Mediterranean and the Jordan Valley. The Korazim block, separating the Hula from the rest of the Jordan Valley to the south, may have been elevated during this phase, causing a dramatic change in the paleogeography of that region, which thus commenced its northward draining.

The eastern highlands and the Negev only suffered moderate uplifts during the late Miocene, as shown by Baubron et al. (1985) and Steinitz & Bartov (1991) for Transjordan and by Garfunkel & Horowitz (1966), Zilberman (1992) and Avni (1998) for the Negev. Steinitz & Bartov had shown that extensive plateau basalts dated at 9.3 ± 0.2 through 5.1 ± 0.2 Ma cover the Transjordanian highlands 5–30 km east of the Dead Sea, overlying a vast peneplain, whose principal erosion phase commenced only after cooling of these volcanics. This situation is further stressed by basalt outpours dated to an interval of 3.7 ± 0.4 to 0.66 ± 0.15 Ma, which flowed in channels cut into the previous plateau, getting progressively younger as they fill up a steeper and deeper relief. Further north and somewhat further away from the Rift, at the headwaters of Wadi Zarqa, Baubron et al. indicated that basalts dating from around five million years ago fill up the river's channel, implying that uplift preceded the Pliocene, most probably in late Miocene times.

The moderate uplift of the Negev is indicated by the pattern of the developing late Miocene drainage system flowing into the Rift, where the Arava Formation was consequently deposited during the Pliocene. This system comprised wide, meandering rivers accompanied by vast floodplains and was devoid of deep canyons, a system that could only form on a relatively low relief. The only exception is the Ramon and northern Negev, which show a somewhat more conspicuous uplift at that time, although not a considerable one, possibly due to activity along the Sinai–Negev transversal faults. This is clear from the pattern of the Arava Formation drainage (Avni 1998), which recognizes the Ramon elevated structure and flows to the northern Arava on its southern flanks, approximately following the present-day Nahal Neqarot course, which became a canyon only later (see below). The northern Negev uplift had cut the earlier connection of the Dead Sea region westward to the Mediterranean (Zilberman 1992).

Hardly any direct data are available for the Judea–Samaria highlands, where younger uplift phases caused erosion which wiped out practically all evidence. Indirectly, however, the late Miocene occurrences of the Shefela and coastal plain lowlands to the west (Buchbinder & Zilberman 1997) seem to support the idea of a generally low-lying landscape also for the central parts of the country, west of the Jordan Valley. Late Miocene sediments are generally found filling and silting up previous, occasionally rather deep channels, further occurring as a thin veneer covering considerable areas of the western lowlands. A similar situation is also observed for the Galilee highlands, of a rather low relief in which shallow, meandering rivers flowed (Kafri 1997).

9.5.2 Miocene–Pliocene transition

This phase of Eritrean activity was recognized as the principal one for a long time, due to its conspicuous effects on paleogeography. These are further emphasized by the rapid rise in sea level at the beginning of the Pliocene, the Tabianian (or Zanclean) transgression, which flooded large portions of the Jordan Rift Valley through the deepening graben of the Yizre'el Valley, reaching as far south as the northern Arava, where it deposited the Mazzar Formation at its peak.

A parallel situation occurred to the north, where the Hula was connected to the Mediterranean through the subsiding Beqa'a and the faulted Bay of Tripoli, where another graben system was formed at that time (Fig. 10.2.3). The subsidence of the Beqa'a occurred both as an accentuation of the synclinorium separating the Lebanon and Hermon mountains, and downfaulting, at least along the Yammouneh Fault, which was never active since (Butler et al. 1997). These northern regions were somewhat more elevated than the Jordan Valley, a situation persisting until the present day, so that the rise in sea level caused their transformation into vast lakes, fed by the abundant rains of the humid Palynozone Pa times. During the following, drier Pb, in conjunction with the lower sea level at that time and the development of drainage toward the Mediterranean, these northern Hula–Beqa'a lakes gave place to rivers.

The paucity of late Miocene rocks along the Jordan Rift Valley makes the assessment of exact age of this phase quite difficult. In most cases early Pliocene formations lie on top of middle Miocene ones, over a faulted and eroded relief, in which case a distinction between the two phases of the Eritrean stage is impossible.

Although tectonic activity of the Miocene–Pliocene transition went largely along the same lines as its predecessor in the late Miocene phase, some details are altogether different. Probably the most interesting feature is the "northward migration" of the southern Dead Sea basin, from its center of subsidence in Mc times, the northern Arava, to the new depocenter of Mount Sedom–Lisan Peninsula. This phenomenon was already noted by Zak & Freund (1981), who took it as evidence for strike-slip movement along the Jordan Rift Valley (see Chapter 10). The exact age of this phase is apparent from several drillings and exposures in the southern Dead Sea region, palynologically analyzed.

The late Miocene Palynozone Mc at the Arava 1 borehole and the Hufeira Member outcrops in the northern Arava are very thick, in the order of 2 km; however, Pliocene sediments do not exceed several tens of meters in outcrops south of the Dead Sea, the maximum figure being the 240 m penetrated by Arava 1. Only several kilometers to the north, in the area presently occupied by the southern

Dead Sea, the situation is inverted; Palynozone Mc is entirely missing at the Ami'az 1 and Ami'az East 1 boreholes (Horowitz & Horowitz 1990), where Pa, which directly overlies Mb, attains up to 1,200 m in thickness. This figure increases eastward, with the correlative exposed part of the Sedom Formation estimated by Zak (1967) at 1,500–2,000 m, while the El Lisan 1 borehole encountered more than 3,500 m of the Sedom evaporites (Bender 1974a, pp. 91, 139).

A parallel setup, of a twin trough, also occurs in the central Jordan Valley. It is quite tempting to suggest a similar "migration" there, but no evidence yet supports this idea, since the northern sector of the twin trough was never drilled. On the other hand, while Palynozone Mc in the Zemah 1 borehole is of the same order of thickness as at Arava 1, Pa and Pb are also considerably thicker in this same location, almost 900 m, so the parallelism of north and south is not obvious.

The situation in the Hula and Beqa'a valleys is not entirely clear, because no distinction can be made there between the two Eritrean phases for lack of conclusive evidence. Only a single deep borehole was drilled in this part of the Jordan Valley, while the few rare fossils occurring in outcrops cannot distinguish the late Miocene from the Pliocene. The Beqa'a and Hula subsided, a process that most probably began during the late Miocene, separating the Hermon and Lebanon anticlinal ranges, connecting both basins with the Mediterranean via the graben north of Tripoli. The Beqa'a became a deeper faulted syncline, the Hula a rift, bounded by faults both to the east and west (see Section 8.5).

The late Miocene age of faulting of the Tripoli graben and the Yammouneh Fault suggests a striking similarity to the Yizre'el Valley. The latter was certainly faulted at the beginning of the late Miocene, as shown from the radiometric ages of basalts interfingering with the Umm Sabune Conglomerate (Shaliv 1991), which postdates the initial faulting. In addition, late Miocene marine sediments are known from the Zemah 1 borehole, drilled in the central Jordan Valley (Gerry & Derin 1983, Raab 1998). The sedimentary sequence (Dubertret 1966) and ages of the non-faulted Homs Basalt covering the Yammouneh Fault, 5.7 ± 0.5 and 5.2 ± 0.2 Ma (Butler et al. 1997), in the Tripoli graben are similar to those reported from the Yizre'el Valley.

The most pronounced uplift of the second phase of the Eritrean stage occurred along the eastern flank of the Jordan Rift Valley, as shown (see above) by the ages of basalts filling up the deep channels leading westward (Steinitz & Bartov 1991). West of the Rift, the Pliocene sediments form extensive bodies in which the thicknesses are quite uniform, in the order of 200 m, indicating a low-lying relief. This is further supported by the absence of canyons west of the valley. Though they are prominent in Transjordan, they are not known for this period in Israel, where the deposits usually contain very few conglomerate horizons. Gravel beds are more common at the tops of Pliocene sequences, following the regressing Piacenzian sea.

The pattern and nature of Pliocene rocks along the entire coastal plain of Israel, which do not differ much from what is seen along the Jordan Rift Valley, seem to support the idea of a relatively flat landscape (Buchbinder & Zilberman 1997). Further west, however, which is outside the scope of this book but worth mentioning, this second Eritrean phase is manifested by considerable deepening of the Mediterranean, followed by its flooding during the earliest Pliocene, which terminated the late Miocene (Messinian) desiccation and deposition of evaporites. This subsidence also initiated the activity of the Nile River (Nachmias 1969, Gvirtzman 1970) at that time.

9.5.3 Pliocene

The Pliocene is generally a tectonically quiet period in the history of the southern Levant (Zak & Freund 1981), except for continuous, most probably taphrogenic, subsidence in the southern Dead Sea and the central Jordan Valley basins. Slight subsidence is also evident for the northern Arava and the Hula (Fig. 9.4.1). Outside the limits of the Jordan Valley to the south, the southernmost Arava and the Gulf of Aqaba continued their subsidence along the bordering faults (Garfunkel 1970), but no figures are available. To the north, continuous deepening of the Beqa'a syncline caused the accumulation of up to 700 m of the Zahlé Beds, at a maximum calculated rate of some 260 mm/Ka.

Along the Jordan Valley, the maximum subsidence is recorded for the southern Dead Sea basin, with figures in excess of 500 mm/Ka for Palynozone Pa times, decreasing to 300 during Pb. The central Jordan Valley basin subsided at some 350 mm/Ka during Pa, decreasing to 300 at Pb; the Hula basin shows very low values for Pliocene subsidence, less than 150 mm/Ka for Pa, increasing to 200 during Pb. The higher rates for the Beqa'a (above) were most probably the cause of the northward drainage of the Hula at that time.

In contrast with the situation in the south, the northern Dead Sea basin shows somewhat accelerated subsidence in Pb times, as seen from the En Gedi 2 borehole, drilled in the southwestern corner of the northern basin. Rates for Pa are in the order of 100 mm/Ka, but increase to almost 300 for Pb. This seems to be the first time when subsidence hits this region (unfortunately no detailed palynozonation is available for the Jordan Valley 1 borehole, drilled just to the north of the Dead Sea). Pb subsidence is most probably better expressed in the deeper parts of the northern Dead Sea basin, but these were never drilled. It is however quite plausible to assume that evaporites in this basin (Neev & Hall 1979) accumulated following its deepening during Pb times, which seems to indicate a further northward migration of subsidence, in the manner shown by Zak & Freund (1981) for the northern Arava and southern Dead Sea regions. This stage of subsidence, accompanied by deposition of evaporites, may have extended further northward to the Jordan Valley, if the assumptions of Belitzky (1996) and Belitzky & Mimran (1996) prove true. These authors suggested that several uplifted areas in the southern and central Jordan Valley could result from the diapiric effects of underlying evaporites (see also Fig. 8.2.10).

When details are compared, the difference between the southern and northern Dead Sea basins becomes apparent. The late Miocene subsidence of the northern Arava practically stopped at the beginning of the Pliocene, when subsidence of the southern Dead Sea basin was initiated, at the Miocene–Pliocene transition. Pb subsidence of the southern Dead Sea basin still shows somewhat higher rates than those calculated for the En Gedi 2 borehole to the north; if indeed, as seems logical, the central part of the northern basin subsided more than its marginal southwestern corner (where En Gedi 2 is located), the subsidence ratio between the two basins may be in favor of the northern during Pb times, but however the southern also continued subsiding. At any rate, no sign of a considerable trough is available for the late Pliocene northern Dead Sea basin, supported by the observation that the southern was never disconnected from its northward drainage system (Zak 1967).

The Yizre'el Valley graben (and possibly also its twin structure, the Bay of Tripoli), through which the Jordan Valley was connected to the Mediterranean during the Pliocene, shows very little subsidence during the Pliocene, being filled up by only some 200 m of sediments during the entire period. This figure compares with the late Miocene, when the Umm Sabune Conglomerate and its equivalents (Shaliv 1991) attain as much as 260 m, indicating moderate subsidence even before the Pliocene. Uplift seems to have been minimal during the Pliocene, both east and west of the Jordan Valley. At the headwaters of Wadi Zarqa (Baubron et al. 1985) sediments of the Jabal Bakiya Formation, which very much resembles the Gesher, are sandwiched between early and late Pliocene basalts, with no considerable elevation differences between the three units.

9.6 LEVANTINE STAGE

The Levantine stage, a term proposed by Horowitz (1979, p. 53), comprises downfaulting along the Jordan Rift Valley and uplift of its flanks, both roughly north–south oriented. The combination of extended subsidence along the Jordan Valley, particularly of the southern Dead Sea basin, and substantial uplift of its eastern and western shoulders, resulted in considerably accentuated morphology. The formation of an endoreic drainage system, whose terminal base level lies within the southern Dead Sea, seems an inevitable outcome of the Levantine combined subsidence and uplift stage, with both processes being of similar importance. The Levantine faults and uplifts are superimposed on all types of former structures, downfaulted and elevated alike, representing a new tectonic trend in the development of the Jordan Rift Valley.

In terms of morphology and development of drainage systems, the Levantine stage is divided into two distinct phases. Its inception was during Palynozone QI times (Horowitz 1992a, p. 331), when the Jordan Rift Valley began subsiding almost all along its length, at rates an order of magnitude higher than before (note the different scales of Figs 9.4.1 and 9.6.1). Uplift of the shoulders was minimal at this time, which permitted direct westward drainage of the eastern highlands to the Mediterranean over the flat landscape remaining from the Pliocene period. The subsiding Rift basins acted as the intermediate lakes of westward-leading rivers throughout QI (Levin & Horowitz 1987; Horowitz 1992a, p. 363; Matmon et al. 1999).

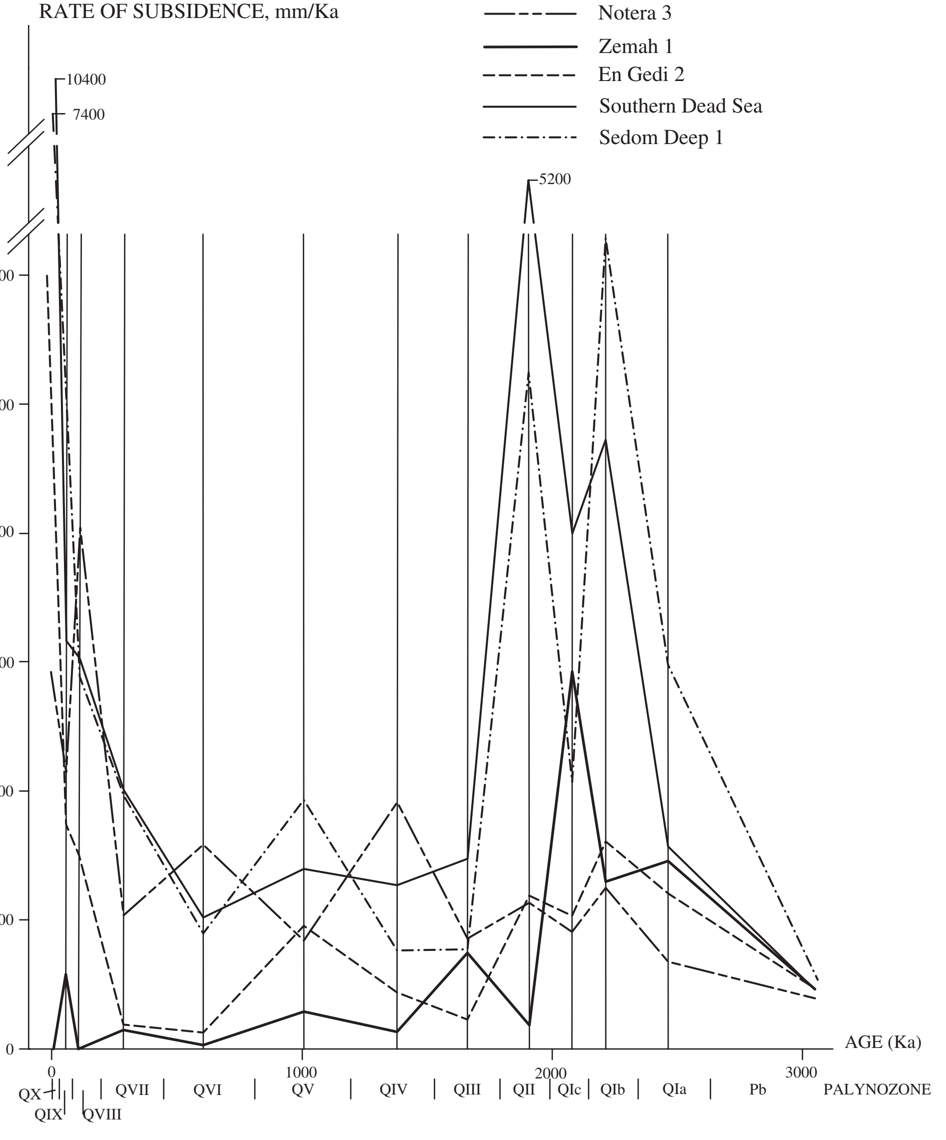


Figure 9.6.1. Quaternary rates of subsidence of different basins along the Jordan Rift Valley, according to thicknesses of palynozones from the specified boreholes.

The second morphological phase commenced during Palynozone QII, when further even more rapid subsidence of the southern Dead Sea basin, accompanied by discernible uplift of the eastern and western shoulders, made the downfaulted region a terminal base level, a process that completely and dramatically changed the landscape, so that a new watershed began forming to the west of the Rift (Horowitz 1992a, p. 366; Matmon et al. 1999). Major sectors of river systems, formerly flowing to the Mediterranean, were redirected eastward, a process continuing until the present day.

The western highlands uplift tectonics are much more pronounced along the central and northern sectors of Israel and the Lebanon, extending northward the formerly uplifted highlands of the central and northern Negev. The subsidence along the Jordan Rift Valley decreases northward, so that the morphological contrast is gradually fading in this direction. This is clearly seen by the extent of drainage capturing (Fig. 1.2). The southern Negev is entirely captured, being drained to the Dead Sea; drainage of the central parts of the western highlands is, broadly considering, equally divided between the Mediterranean and the Jordan Valley; a large part of the upper Galilee is drained to the Mediterranean. Further north, the Litani River still keeps the QI tradition up to the present day. This river drains the Beqa'a sector of the Rift to the Mediterranean, a situation shared by various rivers up to southern Turkey, their channels superimposed on the slowly uplifting western Rift shoulders. The trend is apparent from the location of the western highlands watershed, becoming closer to the Jordan Valley as one goes northward (Fig. 1.2).

The situation to the east is different. The main rivers led to the Jordan Valley in Pliocene times, so that the extended subsidence of the Rift only caused further cutting and deepening of their courses (Michelson 1973, Steinitz & Bartov 1991), a process helped by continuous uplift of the eastern shoulders. Further east, the uplift takes the upper hand in defining hydrography, by creating an extended north–south oriented structurally low belt (Horowitz 1979, p. 152), in which internal drainage developed, forming such inland basins as El Jafr and El Qeisiya in Transjordan, or El Hijane in southwestern Syria (Fig. 1.2).

Figure 9.6.1 summarizes the Quaternary rates of subsidence and accumulation along the Jordan Rift Valley, separately for each basin. Two curves are presented for the southern Dead Sea basin: one averages results obtained from palynostratigraphic studies of several boreholes (Horowitz 1989b; 1992a, p. 331; Horowitz & Horowitz 1990) drilled in the region, while the other is based specifically on the newly analyzed deepest borehole in the basin, Sedom Deep 1 (Horowitz 1996a). The northern Dead Sea basin is represented by the En Gedi 2 borehole, drilled at its southwestern end; the central Jordan Valley by Zemah 1, which penetrated into the deepest part of this basin, as did Notera 3, which represents the Hula. The distribution and thicknesses of Quaternary palynozones in all the boreholes, analyzed palynologically, are presented in Fig. 6.1.3.

It is quite straightforward to calculate rates of subsidence when detailed palynostratigraphic information is available from numerous boreholes (Horowitz

1987c; 1989b; 1992a, pp. 327–334). The question is how to define, with any accuracy, rates of uplift in primarily eroded highlands. To gain some knowledge, Quaternary terraces of the Judean hills were examined. These terraces were formed (Horowitz 1974, Goldberg 1986) by the alternating humid and dry climates, combined with uplift of the hilly regions, and are known all over the Levant (Besançon & Sanlaville 1984). The humid climates caused silting up and gentle erosion (see Section 6.6.2), while the thunderstorms and floods typical of the dry periods resulted in deep incision of the wadis. Consequently nick points were formed, each broadly representing a pair of palynozones, a humid followed by a dry one (Fig. 3.5.2).

Generally, four such main terraces and nick points are known (Horowitz 1979, p. 173; Besançon & Sanlaville 1984) throughout the southern Levant, cutting into the peneplain formed in QI times (Figs 7.7 and 9.6.2). The lowermost two, which are the youngest, contain frequent occurrences of artifacts, and so could easily be correlated with the Palynozones QIX and QVII deposits in the Jordan Valley. The two older, upper ones, west of the Jordan Valley, were assigned to Palynozones QV and QIII, based on sequence stratigraphy principles (Horowitz 1988, 1989a). East of the Rift, at least in one locality upstream the Zarqa River, the third terrace contains artifacts and vertebrate fossils correlative with Palynozone QV (Parenti et al. 1997). The elevations of both the QI peneplain and the four nick points are compared with the present-day wadi floor near Bethlehem, where Neuville (1951) carried out a detailed survey of prehistoric sites. The ages assigned to these (Table 1.4.1) are median ages of the corresponding palynozones, and rates of uplift were calculated (Table 9.6.1, Fig. 9.6.3).

The rates of uplift obtained are only for one locality, but the uniformity of appearances of terraces throughout the southern Levant seems to lend the numbers

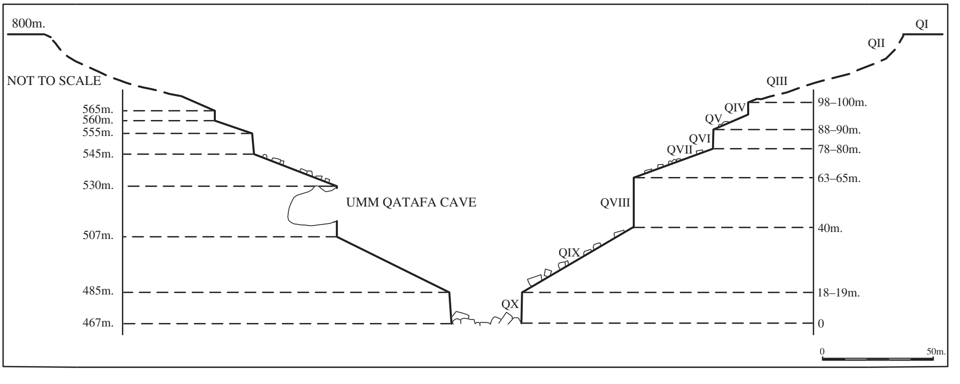


Figure 9.6.2. Cross section of Wadi Hareitun, where the Umm Qatafa Cave is located, some 20 km west of the Dead Sea, modified from Neuville (1951). The succession of terraces is correlated here with the Quaternary palynozones, and their elevation above the wadi floor is used for calculation of rates of uplift, presented in Fig. 9.6.3.

Table 9.6.1. Quaternary rates of uplift of the Judean anticlinorium, from measurements taken near Bethlehem by Neuville (1951). The nick points of the benches and terraces are correlated partly by prehistoric implements, partly by sequence stratigraphy with the even-numbered, wetter palynozones. Note that "younger" refers only to even-numbered palynozones. For further explanation see text.

Palynozone	Elevation above present-day sea level (m)	Age (middle of palynozone)	Age interval to younger palynozone	Elevation above younger palynozone (m)	Average rate of uplift (mm/Ka)
QI	800	2.31 Ma	560 Ka	235	420
QIII	565	1.75 Ma	730 Ka	15	21
QV	550	1.02 Ma	740 Ka	20	27
QVII	530	280 Ka	240 Ka	45	188
QIX	485	40 Ka	40 Ka	18	450
Present day	467	–	–	–	–
Interval QI to present-day	–	2.31 Ma	2.31 Ma	333	144

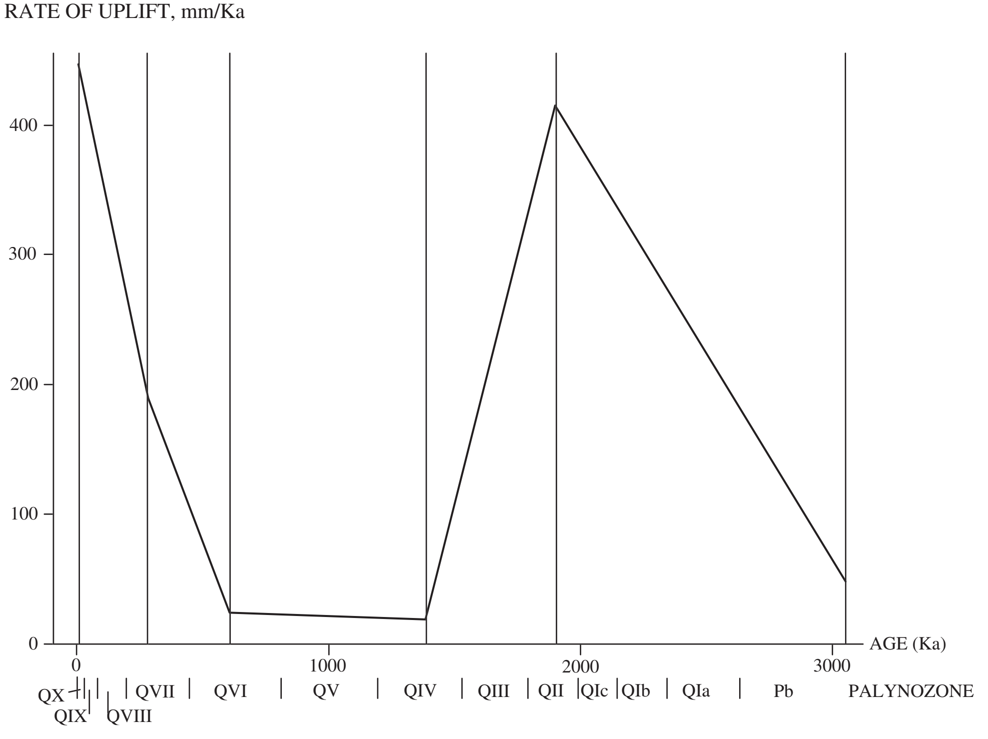


Figure 9.6.3. Rates of Quaternary uplift at Wadi Hareitun, some 20 km west of the Dead Sea, calculated from elevation of terraces correlated with the palynozone. A cross section of the wadi is shown in Fig. 9.6.2. The rate for the Pliocene is adopted from Begin & Zilberman (1997).

more than a local significance. Also, the calculations only represent minimum figures when compared with the Rift Valley, where counterparts of the terraces are located well below the elevation of the present-day wadi floor, taken as a datum. As an example, the QI peneplain is only 333 m above this datum, but its correlatives are buried well under the Dead Sea (Figs 6.1.3 and 9.6.1), so that the actual uplift is much more extensive.

The figures in Table 9.6.1 thus give an idea of uplift timing, and the relations between various time intervals. A look at these figures immediately reveals the correlation between the subsidence of the Rift and uplift of the hills, which seem to go broadly hand in hand. Whenever accelerated subsidence is recorded for the Dead Sea basin, an extended rate of uplift of its flank is apparent, indicating that the two opposing trends are indeed caused by a single tectonic process.

As seen from Figs 9.6.1 and 9.6.3, and Table 9.6.1, the Quaternary history of subsidence and uplift can be subdivided into three main stages, with QI–QII and QVIII–QX showing extensive activity, while the interval QIII–QVII is characterized by moderate disturbances. QI and QII are discussed separately, since their landscapes are so different. Details of the subsidence and uplift along the Quaternary are given below for each of the stages.

9.6.1 Palynozone QI

The relative regional tectonic quietness of the Pliocene, particularly the almost unnoticeable uplift, extended also into the earliest Quaternary, except for the considerably accelerated subsidence of the Jordan Rift basins. However, paleogeographic changes had occurred, especially outside the Rift margins, mainly resulting from the retreat of the Pliocene sea, accompanied by the further development of drainage systems toward the Mediterranean (Fig. 10.4.2) and slight uplifts of several rather small but critical blocks. To the south, the slight uplift of the Ramon and northern Negev structures (Zilberman 1992, Avni 1998) is quite obvious from the inversion of the drainage direction south of the Ramon, from the northwestward course of the Arava Formation into the southward-flowing HaMeshar Formation channel (erroneously termed "Zehiha" by Avni).

The developing westward drainage, leading to the Mediterranean, had reached the southern Negev and central Arava in these times (Garfunkel & Horowitz 1966), capturing much of the areas' previous channels. This ceased the activity of a major part of the Arava Formation system, previously flowing toward the Dead Sea, which now drained at this stage directly to the Mediterranean through the Kuntilla plains, Wadi Quraiya and Wadi el Arish, depositing the HaMeshar Formation on its way, from the central Arava onward. Another such drainage system developed during the earliest Quaternary north of the northern Negev elevated terrain, crossing Judea from the area north of Amman where the Dhuleil Formation was deposited, via the southern Jordan Valley through to the Mediterranean, depositing the Bethlehem Conglomerate on its way (Horowitz 1979, p. 118).

No earliest Quaternary rocks, in fact no Quaternary rocks at all, are known from the central sector of the Yizre'el Valley graben. Boreholes sunk in this area reveal that the present-day valley floor consists mostly of middle or late Miocene and Pliocene sediments and volcanics (Shaliv 1991), indicating that the region has been under erosion for a long time. Relics of lagoonal Pliocene formations testify to a connection of the central Jordan Valley with the Mediterranean at that time (Sneh 1993, and others), which leaves the earliest Quaternary the best candidate for the beginning of uplift. Further north, in the Galilee, another parallel drainage system developed at this time (Kafri 1997, Matmon et al. 1999).

The existence of these three major river systems, with their shallow channels and wide floodplains, indicates that the earliest Quaternary, save for the two slightly elevated Ramon–northern Negev and Yizre'el blocks, was a period of no considerable uplift, at least west of the Jordan Rift Valley. The situation to the east is far from clear, for lack of data and uncertainty of dating. This region was certainly higher, as evidenced from the directions of flow of the main river systems draining the Transjordanian Plateau to the Mediterranean (Zak & Freund 1981). Whether the higher elevation of the eastern highlands is a Pliocene heritage, or was also slightly accentuated during the earliest Quaternary, is impossible to tell with any certainty.

Further north, in southern Lebanon, another such drainage system developed, most probably at the same time, connecting the Beqa'a to the Mediterranean via the Litani River, and is still active to the present day. This system remained active, in contrast with the other three to the south, because the Beqa'a had not subsided much since the end of the Neogene, as can be seen by its extensive coverage by the Pliocene Zahlé Beds, themselves hardly overlain by any younger rocks. This process is helped by the earliest Quaternary uplift of the Metulla block, separating the Hula and the Beqa'a. This age is assigned to the uplift since the elevated block is covered only by faulted Pliocene rocks and erosional features, and not by any younger ones (Heimann 1990, U. Kafri, Geological Survey of Israel 1998, pers. comm.). Also, there is an intimate connection, and great similarity in rocks, between the Hula and the Beqa'a during the Pliocene, which does not extend into the Quaternary (Horowitz 1973).

Within the Jordan Rift Valley, the Pliocene troughs continued sinking during QI, but at a considerably quicker pace as compared with former times. The southern Dead Sea basin had accumulated several hundred meters, and up to more than 1 km, of the Melekh Sedom Sands, during Palynozone QI times at rates of accumulation occasionally exceeding 3 m per 1,000 years. The northern Dead Sea, the central Jordan Valley and the Hula show somewhat more modest figures, in the range of 500–700 mm/Ka (Horowitz 1989b). However, as can be seen from the sediments (Sa'ar 1986) and the pollen assemblages (Horowitz & Horowitz 1990; Horowitz 1992a, pp. 335–336), neither of these served as a terminal base level in QI times. Rather, they acted as intermediate basins for the principal river systems connecting Transjordan to the Mediterranean.

A more detailed observation of the numbers calculated for rates of accumulation in the various boreholes drilled in the Rift and palynologically analyzed (Fig. 9.6.1), reveals some differences in the behavior and timing of the four principal basins along the Jordan Valley. Both the Hula and the northern Dead Sea show relatively (compared with the others) moderate subsidence during the entire period of Palynozone QI, in the order of 600–700 mm/Ka. But the southern Dead Sea basin subsided at rates of up to 3,000 mm/Ka, with maximum values attained during QIb, while the highest figure for the central Jordan Valley, some 2,350 mm/Ka, was reached only later, during Palynozone QIc times.

Uplift of the hilly flanks cannot be accurately calculated for Palynozone QI, since its rate (Table 9.6.1), 420 mm/Ka, is for the interval up to QIII times, and this could well represent QII activity. It seems that this may be the case when considering the paleogeography of QI times, of a rather flat peneplain crossed by several large, meandering shallow rivers, with extensive floodplains. QII erosion cut deeply into this peneplain (Fig. 9.6.2). On the other hand, one must be careful with this conclusion because, although correlation of the peneplain with Palynozone QI seems quite safe, it may well represent only the earlier QIa part of the palynozone, while in QIb and QIc we may envisage at least some uplift. This uplift, if any, had not yet affected the hydrographic network, since it is clear that the southern Dead Sea became an endoreic basin only in QII times (Horowitz 1992a, p. 336).

9.6.2 Palynozone QII

This is the crucial time, when the southern Dead Sea basin was established as a terminal base level, a situation persisting to the present day. Rates of subsidence of this region calculated for Palynozone QII times attain (Horowitz 1989b) up to 5.2 m per thousand years, accumulating more than 1 km of corresponding sediments (Fig. 9.6.1), a figure only ever surpassed by QIX–QX activity (see below). In contrast, the northern Dead Sea, the central Jordan Valley and the Hula suffer only subordinate subsidence during QII. The first and last basins do not accumulate more than 100 m of sediments, and similar figures are also encountered in boreholes drilled close to the southern Dead Sea basin but outside its limits, such as Ami'az East 1. No QII rocks are known at all from the central Jordan Valley, neither exposed nor recognized in drillings (Figs 6.1.3 and 9.6.1).

At the same time, considerable uplift is recorded for the hilly region west of the Dead Sea (Table 9.6.1), with a calculated minimum rate of some 420 mm/Ka. This uplift most probably affected the entire western belt (Fig. 3.1.1), as can be seen by the eastward drainage direction of the entire area of Israel at that time (Horowitz 1992a, p. 366). Further, the extended incision of wadis to the east of the Jordan Valley (Steinitz & Bartov 1991), although not as accurately dated as the western sector, hints at a broadly synchronous uplift of the eastern belt.

One of the problems concerning the drainage system that began to develop during QII, which finally evolved to form the present-day system, is the question of when was the Hula connected to the Dead Sea following its separation from the Beqa'a by the Metulla block, uplifted in QI. Tables 6.5.2.4 and 6.5.2.8 compare pollen spectra obtained from the Hula and the southern Dead Sea, of Palynozone QII with those of QX. The pollen spectra from the two palynozones indicate approximately similar natural environments, so that any differences could be blamed on transportation. Most of the constituents show similar characteristics in both tables, except for the percentages of arboreal pollen. When their share in the regional vegetation pollen is compared, the difference between north and south is much greater in QII than in QX times.

Detailed calculations indicate that arboreal pollen percentages are 5.3 times higher in the Hula, for Palynozone QII; 2.45 for QIII; 2.6 for QIV; 3.1 for QV; 2.1 for QVI; 2.5 for QVII and 2.6 for QX. The higher ratio for QII as compared to the overlying ones, and the persistence of ratios ever since QIII until the present day, indicates that transportation of arboreal pollen, produced mainly in the northern parts of the country, was not as effective in QII times as it was later on. In other words, this gives rise to the speculation that the Hula was possibly not connected to the Dead Sea system at the immediate time of its formation, but that this happened only later, during QIII.

Sediments of QII are hardly known from the Hula, but the overlying QIII Gadot Formation comprises mostly sterile chalk (Horowitz 1973), which may indicate that the basin was not properly drained at the time. The few mollusks recovered from the Gadot are dwarf forms, usually found only near the shore of the lake at that period. Colonization of the Hula by a rich and varied fauna only came later, as known from the prolific Mishmar HaYarden Formation, corresponding to Palynozone QV (Tchernov 1973). The delay in freshening the Hula, compared with the time of changing pollen transportation, is not surprising, since it takes time for a river to accomplish the former process, while the latter is immediate.

9.6.3 Palynozones QIII–QVII

This is a relatively quiet period in the tectonic evolution of the Levantine Rift Valley (Fig. 9.6.1), characterized by continuous subsidence of the Hula and southern Dead Sea basins accompanied by slow but steady uplift of its shoulders (Table 9.6.1). The central Jordan Valley behaved differently during this interval, showing slight subsidence in QIII times, even less in QV, when the Erk el Ahmar and Ubeidiya formations were laid down, respectively (Horowitz 1988). Other palynozones are hardly represented in this region, where the entire sequence, from the top of Palynozone QI until the present day, takes up only the uppermost 230 m in the Zemah 1 borehole. The average rates of subsidence for the Hula and southern Dead Sea throughout this period are in the order of 500–1,000 mm/Ka, an order of magnitude less than for QII and QI in the latter locality.

Although structural evolution is generally continuous and usually steady during this period, a conspicuous faulting phase is known to postdate QV times, affecting the entire Jordan Valley, termed by Picard (1943) "Intragraben tectonics". The faulting is especially manifested in the Hula and central Jordan valleys, where the Erk el Ahmar and Ubeidiya formations and their correlatives are considerably tilted. It also affected the Dead Sea region, as seen by a sudden increase in spring discharge, recorded in the pollen spectra at the top of Palynozone QV (Horowitz 1992a, p. 338), most probably resulting from exposure of aquifers by the faulting. Interestingly, this faulting phase, however conspicuous in exposures, did not have any detectable effects on the subsidence of the valley floor (Horowitz 1989b), nor on the rate of uplift of the western hills.

Another such faulting phase, less noticeable in outcrops, affected the Jordan Rift Valley some time during the middle part of Palynozone QVII. Its results are seen south of the Hula Valley, in the Gesher Benot Ya'aqov area and especially at its prehistoric sites, where strata of the Ayyelet HaShahar and Benot Ya'aqov formations, corresponding to Palynozones QVI and QVII, are considerably tilted (Horowitz 1973, Goren-Inbar 1995). As for the previous structural disturbance, pollen spectra indicate increased outflow from springs in the Arava (Horowitz 1987b; 1992a, p. 337). In contrast with the former, this faulting phase was accompanied by accelerated subsidence and uplift in QVIII times.

9.6.4 Palynozones QVIII–QX

The period covered by the last three Quaternary palynozones, approximately 130,000 years, sees a rejuvenation of intense tectonic activity both in the Jordan Valley, which goes through extended subsidence (Fig. 9.6.1), and the Rift's shoulders, whose uplift continues at a considerably more expeditious pace (Table 9.6.1, Fig. 9.6.3). Again, as during the QIII–QVII interval, the one exception is the central Jordan Valley, which subsided very little during QIX, but shows no such activity at all for QVIII or QX (Horowitz 1989b).

Extended subsidence is recorded for QVIII in the southern Dead Sea basin, an increase which began during Palynozone QVII, attaining rates in the order of 1,500 mm/Ka. The northern Dead Sea basin shows, for the first time in the Rift's history, an increase in subsidence to an average rate of 700 mm/Ka, as compared with 100–150 before. This stage is the precursor of the present-day deep trough, which developed very rapidly in this region. However the most active basin during QVIII is the Hula, with rates of subsidence in the order of 2,000 mm/Ka, the highest ever recorded for this area.

This was also a time of pronounced uplift, seen clearly west of the Hula Valley, apparently also affecting its eastern flank. On the western highlands, a swarm of north–south oriented clastic dikes (Horowitz 1975b) cut through Eocene limestones in Nahal Yir'on, some 10 km due west of Lake Hula. The filling comprises, among all kind of fallen debris, some Acheulian artifacts, of a type comparable with those occurring at Benot Ya'aqov Formation, corresponding to Palynozone QVII (Horowitz 1992a, p. 409). These tensional features are clearly connected with the uplift of the eastern upper Galilee, as seen from their mechanics and bearing (Arkin 1989). The artifacts convey a maximum age for the uplift, which thus postdated QVII.

To the east, Heimann (1990) reports faults younger than 200 Ka affecting the western slopes of the Golan Heights, which could belong to the same phase, that must have taken place at approximately 100 Ka ago. Extended uplift is also recorded for the Judean Hills at that time (Table 9.6.1), reaching the highest rate ever, 450 mm/Ka which, as discussed above, should be considered a minimum figure.

The structural trends observed for Palynozone QVIII in the two basins of the Dead Sea continue into QIX, while the central Jordan Valley shows some subordinate subsidence for QIX. The Hula, on the other hand, slows its pace, and the subsidence rate decreases to half the value it was during QVIII. Uplift continued both east and west of the entire Jordan Valley, as can be seen from the elevations of terraces in wadis cutting the highlands. Terraces formed during Palynozone QVII, containing Acheulian artifacts, are at least several tens of meters higher than those corresponding to QIX (Table 9.6.1), the latter bearing Middle and Upper Paleolithic artifacts (Bender 1974a, p. 100; Horowitz 1979, pp. 120–121).

The generally accelerated movements of QVIII and QIX reached a peak toward the end of the latter palynozone, when a distinct faulting phase (Fig. 9.6.4), active at approximately 18–14 Ka ago (Neev & Emery 1967; Horowitz 1979, p. 54), completely changed the landscape of the entire Jordan Rift Valley. This activity has continued throughout Palynozone QX to the present-day, with another accelerated faulting phase at approximately 5 Ka (Horowitz 1973, 1989b, Horowitz & Horowitz 1990) also affecting the entire length of the Valley.

These disturbances are expressed all along the Jordan Valley. Two new troughs have been formed, in the Dead Sea and north of the central Jordan Valley, now housing the northern Dead Sea deep basin, whose floor reaches some 750 m below sea level, and Lake Kinneret, down to 250 m below sea level. The rates of subsidence involved in this stage are the highest ever recorded (Fig. 9.6.1): up to 10,400 mm/Ka for the southern Dead Sea, some 3,000 mm/Ka for the southern corner of the northern Dead Sea, and approximately 2,500 mm/Ka for the central part of Lake Kinneret. The center of the northern Dead Sea basin subsided at least 350 m during the last 16 Ka, at the unbelievable rate of some 22 m/Ka. This process ended, rather abruptly, the existence of Lake Lisan (Begin et al. 1974), and was typical of QIX times along almost all the Jordan Valley.

The shorelines of Lake Lisan are known at an average elevation of 180 m below sea level, all along the Jordan Valley. The lake itself, as seen from the sediments, was rather shallow (Begin et al. 1974). To the north, in the central Jordan Valley, the elevation of Lisan Formation sediments is only slightly lower than the above figure, while around the Dead Sea these are found down to at least 150 m lower than the shoreline, a position they acquired due to the terminal QIX faulting and subsidence. Around the Dead Sea, this phase is conspicuous in almost all the wadis leading to the new, much deeper basin, where impressive waterfalls were

Figure 9.6.4. Faulted Lisan Formation at Wadi Mallaha, some 25 km north of the Dead Sea. This fault is the extension of the one bounding the western side of the deep northern basin of the Dead Sea, clearly seen at the upper part of Fig. 3.6. Photo courtesy of S. Belitzky.

formed. Such waterfalls are not known in the much more stable central Jordan Valley at that time. The new Lake Kinneret was very quickly filled up with water, so the net erosion base level changed only slightly since Lake Lisan times, and is now only some 30 m below the latter.

The Hula shows some increase in rate of subsidence in this phase, up to 1,500 mm/Ka. An interesting question concerning the Hula basin is when it had acquired its present-day morphology of a rather deep valley, flanked by steep escarpments, a problem seen when regarding their hydrography (Fig. 3.1). Both east and west of the Hula Valley, there are hardly any large rivers or wadis leading to the lake. The two exceptions are Nahal Hermon, leading westward to the north-eastern corner of the Hula Valley, and Nahal Dishon, draining the Naftali mountains to its southwestern part. These two rivers seem to be a heritage of Pliocene times, since sediments of this age line their courses (Heimann 1990).

The watershed to the west is very close to the escarpment, approaching the Hula Valley in some places to a distance of only a few hundred meters. The few wadis, both east and west of the Valley, are very steep and short, while to the north several waterfalls are prominent. The profile of the Jordan River sector connecting the Hula to Lake Kinneret is convex (Belitzky, in prep.), indicating its immaturity. All these features, in contrast with the much better-developed river systems flowing to the central Jordan Valley or the Dead Sea, seem to indicate that the bordering escarpments of the Hula Valley, as well as its present floor elevation, are very young, and have hardly had time to develop proper drainage. It should also be noted that the Dan Travertine and the Ashmura Formation, both of QIX age, are considerably faulted (Horowitz 1973, Heimann 1985, 1990).

Heimann & Sass (1989) maintain that deposition of the Dan Travertine, whose top is some 20 Ka old, stopped due to an abrupt change of relief, from flat to steep, so that the water ran much more rapidly in the deeper channels, which did not allow enough time for proper deposition of carbonates. It thus seems only logical to connect the formation of these escarpments with the faulting and extended subsidence at the end of QIX, also helped by uplift of the shoulders. Uplift of the Rift's flanks continues to the present day, as can be seen from terraces equivalent in age to Palynozone QIX, which are up to several tens of meters high above the wadi floors (Horowitz 1979, p. 121; and Table 9.6.1).

9.7 CONCLUSION

Three types of processes are woven together continuously throughout the history and development of the Jordan Rift Valley: subsidence, which evidently affected mainly the areas occupied by the Rift itself, but during certain stages exceeds its limits; uplift, expressed outside the Rift on both flanks, but known also from the behavior of certain blocks within its boundaries; and volcanism, primarily active beyond the Rift's margins, but known also from its subsiding northern basins. The time relations between these three agents are intricate, mainly because in most cases the limits of dating error, both radiometric and stratigraphic, are too large to define exactly whether any two events are indeed precisely synchronous, or one predates the other. Added to this is the difficulty of exactly defining the ages of uplifting, since the ensuing erosion features hardly ever contain any solid datable material.

The most important questions thus concerned with the tectonic evolution of the Jordan Rift Valley are: is the subsidence a result of uplift, or vice versa; or are the two a single process? Is there an intimate, genetic connection between volcanism, subsidence and uplift; which activity predates the other, or are they synchronous? Or maybe magmatics just occur in the region, but are not connected directly with the rifting?

9.7.1 Subsidence

Subsidence appears to have acted along the Rift in two styles, by deepening of the synclinal axes or by the formation of taphrogenic troughs, or occasionally by a combination of both. The two processes change location and amplitude in time, and usually the pace of faulting exceeds synclinal subsidence by an order of magnitude (Figs 9.4.1 and 9.6.1).

The northern Arava and southern Dead Sea, as well as the central Jordan Valley, went through synclinal deepening during the Oligocene, which lasted until the beginning of the late Miocene, when troughs were formed in the former and latter regions. The area north of the Dead Sea underwent similar subsidence during the early and middle Miocene, but no data are available for its behavior in Oligocene times. Following the late Miocene trough formation, the northern Arava became quiet again, continuing its slow synclinal subsidence to the present-day. The central Jordan Valley trough continued its accelerated subsidence until the end of Palynozone QI, when this region calmed down. At the end of Palynozone QIX a trough was formed north of the central Jordan Valley, housing Lake Kinneret in its present shape, actively subsiding until the present-day, in a region that up to that stage suffered only mild synclinal activity. Unfortunately no chronological data are available for the part of Lake Kinneret that forms the northern extension of the central Jordan Valley trough (see Section 8.4), which may indeed be older.

The southern Dead Sea basin became a trough in Pliocene times, continuing this style of subsidence to the present-day. Its neighbor, the northern Dead Sea basin, usually subsided at a very slow pace, with only two exceptions: a somewhat accelerated activity of uncertain nature during the late Pliocene and earliest Quaternary, while at the end of Palynozone QIX a very deep trough was formed, in which the northern sector of the Dead Sea is now located. The region separating the Dead Sea from the central Jordan Valley was elevated, showing no subsidence until the late Miocene when it commenced its slow synclinal deepening, persisting in this way to the present.

North of Lake Kinneret the Korazim block was high probably almost throughout the entire history of the Rift, as compared with its bordering basins, the Hula and central Jordan Valley. It probably slowly subsided, in a synclinal way, following the general activity of the Rift. The Korazim was breached by channels at least twice in its history; during the Miocene, when it was occupied by a river joining the Herod system to the south, and from the end of Palynozone III times, when the Hula is drained southward by the Jordan River.

The Hula and its northward extension, the Beqa'a, started subsiding both synclinally and taphrogenically only at the beginning of the Pliocene (Heimann 2000). No evidence for any activity in this region is available before that time, since no Herod Formation equivalents are present at the Beqa'a. This process terminated abruptly in QI times, when the Hula was disconnected from the Beqa'a by the uplifting Metulla block. At this time the Hula became a taphrogenic trough which, for a while until captured by the Jordan River, acted as an endoreic basin. The closure of the Hula may have been helped by basalt effusions, of the Mechki to the north and Ruman to the south (Heimann et al. 1987).

The Hula Valley attained its considerable relief only in Palynozones QVIII–QIX, due to the combined subsidence of the valley floor and uplift of the bordering highlands.

It is tempting to regard the subsidence along the Rift as gradually propagating from the south northward, as suggested by Zak & Freund (1981), but several facts seem to spoil this picture. Oligocene and Miocene subsidence occurred simultaneously both in the southern Dead Sea and the central Jordan Valley; the deepening of the Hula in QVIII preceded the formation of Lake Kinneret to the south, and even further south came the creation of the deep trough of the northern Dead Sea basin at the end of QIX. However, this last phase also considerably deepened the Hula.

9.7.2 Uplift

The southern Levant, as part of a wider terrain, was subject to various phases of subsidence and uplift throughout its history (see Chapter 4). The last regional uplift which is not yet connected with the formation of the Jordan Rift Valley, occurred some time during the late middle Eocene. The following phase, during the Oligocene, is broadly tied up with subsidence along the lineament of the future rift, but the exact timing of the two is doubtful, due to poverty of data. The elevated regions during the Oligocene include the southern half of the Transjordanian highlands, the southern Negev, and central and northern Israel. Southern Transjordan was connected to northern Israel by an elevated area, which formed a regional watershed north of the present Dead Sea. The remaining lowlands, the central Negev to the south of the watershed and the Damascus–Golan synclinorium to the northeast, served as the main waterways connected to the Mediterranean and the Persian Gulf domains, respectively. The Oligocene uplifts, although noticeable, are only minor in rate and amplitude, which helped peneplanation of the entire region over the long period involved in this process.

This trend, of long wavelength and low amplitude, persisted along the same lines during the early and middle Miocene with one difference, namely that the central Negev was uplifted at this time, pushing northward to the northern Negev the connection of the southern Dead Sea–northern Arava with the Mediterranean. The frequent occurrences of Dana Formation sediments east of these regions and their participation in subsequent uplift (Fig. 5.1.3) hint that the Miocene uplift of Transjordan was minimal. Throughout the early and middle Miocene the region north of the Dead Sea continued its role as a regional watershed.

This style changed during the late Miocene, when uplift of the northern sector of the eastern highlands disconnected the central Jordan Valley from the Persian Gulf, the watershed north of the Dead Sea was breached for the first time, while uplift of the Negev propagated further northward, to its northern sector, to block the previous waterway to the Mediterranean. This stage involved more extensive uplifts than before, to account for such drastic changes of hydrography.

The Pliocene and earliest Quaternary show only very slight uplifts, a quiet period which permitted, toward its end, the development of Palynozone QI river systems, crossing the filled up Rift's area from Transjordan to the Mediterranean. The two main considerable uplift phases of the Quaternary, during QII and QVIII–QX, are known both east and west of the Jordan Valley, separated by a period of only mild uplift. These Quaternary uplifts are manifested by their effects on Pliocene and QI rocks (Figs 5.2.13, 5.2.18 and 5.2.19), shaping the region to its present-day characteristics, and again dramatically changing the hydrography. Indeed, when examining the extent of uplift, it appears to be the chief factor in determining the landscape (Sneh 1996), considerably outweighing the contribution of faulting on both sides of the Jordan Valley.

The timing and extent of Quaternary uplift correlates very well with the subsidence of the Rift's floor (compare Figs 9.6.1 and 9.6.3). The extent and broad timing of other uplift phases seem also to correspond to subsidence within the Rift, in the sense that both processes display parallel trends of amplitude. The Oligocene through the beginning of late Miocene, the Pliocene through QI (except for the southern Dead Sea basin), and the interval QIII–QVII, are periods characterized by slow subsidence and uplift, while both are accelerated during the late Miocene and the two Quaternary phases. This correlation, although in need of more refinement of data for better support, hints that the subsidence and uplift are intimately connected processes, involved together in the rifting evolution.

9.7.3 Volcanism

Garfunkel (1989) considers that the Cenozoic magmatism of the Arabo-African continent, like other intraplate magmatism, is caused by hot spots, which comprise a system of plumes or upwellings in the mantle. He noted however that a systematic age progression of magmatic activity is not evident in this region. Garfunkel therefore suggests that the Rift Valley and its surroundings are underlain by hotter than normal mantle, which generated volcanic activity from time to time. Since heat flow data generally show low values for the Jordan Rift Valley (Ben-Avraham et al. 1978, Eckstein 1979, Feinstein 1987), Garfunkel maintains that the anomalously hot mantle is quite deep. Further, since the main body of magmatics, known as the "El Shamah–Jebel Druze field" extends far away from the Jordan Valley, the connection between the two could be accidental. This is also the view of Kashai & Croker (1987), who maintain (p. 55): "Despite the spatial association, the origin of this volcanism is probably not connected with the evolution of the Dead Sea–Jordan rift system." Bender (1974b) regarded the Cenozoic volcanics as (p. 6) "cratogenic magmatic activity of basaltic character". Stein & Hofmann (1992) suggest that the Arabian lithosphere could be underlined by a fossil plume head.

A variety of approaches can be adopted to view the possible connections of volcanic activity, uplift and subsidence processes in and around the Jordan Rift Valley. Geographically (Fig. 5.3.1) most of the volcanics are located quite far away from the Jordan Valley, their bulk concentrated at considerable distances east of the Rift (Garfunkel 1989). The general bearing of the mid-Cenozoic and younger eruption centers follows the Eritrean lineament, namely SE–NW directed. This is clearly seen in the distribution patterns of the Red Sea Dike system, dated at 25–19 Ma, followed by the younger Lower Basalt (17.5–12 Ma), late Miocene basalts (ca. 9 Ma), and Intermediate and Cover basalts (5.5 Ma until the present day). The Jordan Rift touches only the northwestern corner of the large El Shamah–Jebel Druze volcanic field, where volcanism commenced in Oligocene times (Tarawneh et al. 2000) and, except for minor in-rift intrusions and effusions, is free from any conspicuous activity that could be directly related to the rifting. It seems therefore that, when locations are considered, there is no particular reason to connect the tectonic and volcanic phenomena known for the southern Levant.

In terms of timing however, the gross correlation between uplift, volcanism and rifting in the southern Levant is striking. These processes began some time during the Oligocene, continuing until the present day, thus tempting numerous investigators to regard them as different facets of a single phenomenon. A closer look at the details, however, reveals several discrepancies, which raise some question marks over this seemingly fine correlation. It is not clear whether subsidence of the Rift already began in early Oligocene, but it was certainly active during the middle part of that period, when the earliest volcanism is known.

Some later activity is known from Sinai, some 25 Ma ago, at the beginning of the late Oligocene, already at the onset of Palynozone Ma. The next volcanic phase, the Lower Basalt, broadly corresponds to Palynozone Mb times. The subordinate late Miocene activity does not seem to have much significance. The one that follows, pouring the Intermediate Basalt, was active mainly in Pa times. Continuing volcanism, of lesser amplitude, is seen throughout Palynozone QI. This commenced the last phase, comprising the Cover Basalt and the various locally named effusions, active until almost the present day. These all belong to the same cycle, which is more active in the period from QIII through QVII (Heimann 1990). A suggestion could be made that volcanism accompanies the quieter periods of the Jordan Valley Rift's activity, typified by slow, synclinal type subsidence and subordinate uplift. The periods characterized by distinct faulting, and accelerated subsidence and uplift, are accompanied by only subdued volcanic activity.

If this proves true, there seems to be some genetic relationship between processes pertaining to the rifting and volcanism. However, this needs, in my view, further and more accurate proof. Another interesting point, when discussing the distribution of volcanics, is that their considerable volumes are usually found in the lower structures, such as the Yizre'el Valley, the Golan synclinorium or the Tripoli and Wadi Sirhan grabens. I cannot see any explanation for this phenomenon. On the other hand, it should be noted that volcanic activity is notably more pronounced east of the Jordan Valley, a region that also suffered considerably greater uplift, as compared with the area to the west. This connection could support the idea that uplift resulted from stronger upper mantle activity.

Last comes the question of the various seemingly well-defined magmatic phases. When regarding the volcanic occurrences in their stratigraphic context, namely wherever they interfinger with distinct sedimentary units, there seems no problem in stratigraphic recognition. These volcanics were defined (Bentor 1946, Schulman 1959 and many others) as time correlative with the sedimentary formations. However, when volcanics are radiometrically dated (see Chapters 5, 7 and 11 for details), the boundaries between the various phases become questionable, since it seems that somewhere in the southern Levant there always was some volcanic activity, also in those seemingly quiet times. It is thus not such a preposterous idea to regard the volcanism in the vicinity of the Jordan Valley as a continuous process, lasting from the Oligocene until the present day, accompanying rifting and uplift in a general way all along this period.

At present I do not think we can provide definite answers to the questions raised above. The close synchronization of rifting and uplift phases seems quite well established, but it is not entirely clear whether volcanism is also intimately connected to the two processes at all in terms of particular timing, and if it is, exactly in what way. These reservations were already raised by Schulman (1959, 1962), but unfortunately were not answered satisfactorily since (Hirsch & Roded, in prep.). Tectonic modeling proposed for the entire Red Sea province (see Section 10.5), based on the integration of rifting, uplift and volcanism, could possibly better explain their spatial and temporal relationships.

CHAPTER 10

Tectonics: models and debates

The debate about the tectonics of the Jordan Rift Valley principally concerns its setup within the framework of regional plate kinematics, namely the relative movements of the bordering Sinai–Israel sub-plate to the west and the Arabian plate to the east: had the two parted at all? And if so, are they moving along an east–west trajectory, or is a sinistral offset exceeding 100 km involved, with the north–south line of detachment a subsequent outcome? This debate emerges initially from a rather surprising situation: indeed it all stands on determining the elusive structural framework, which first has to be resolved before any suggestions regarding models for its formation are proposed. The strange fact is that, after almost 150 years of research, during which information was and still is being acquired, in thousands of publications and plenty of unpublished, but available information such as oil companies' reports, no definite conclusion is at hand as to whether this north–south Levantine lineament is indeed determined only by faulting, and if so, is it bounded by normal faults or by a large-scale strike-slip system.

Based on either point of view (or belief?), there are three schools of thought concerning the tectonic origin of the Jordan Valley. One, shared by the greater number of geologists and geophysicists, regards this Rift as having been formed by a major sinistral strike-slip movement along a boundary between the Sinai and the Arabian plates (Garfunkel & Ben-Avraham 2000, and see Chapters 4, 8 and Appendix); the second, much smaller group of scientists, agrees with the idea of a plate boundary, but does not see any conspicuous lateral dislocation between the two (Mart 1991); the third, with only a few followers, maintains that since no continuous faulting was ever claimed for this lineament, it does not constitute a true plate boundary. Naturally, if faulting is interrupted there is no place for a lateral offset of the magnitude suggested (Horowitz 2000b, 2001).

Consequently, perceptions differ as to the tectonic framework of the Jordan Rift Valley, and its role in the regional plate-kinematic pattern. Basically, the first group views the valley as a "leaky transform" (Garfunkel 1981), while the second regards it as a direct continuation of the Red Sea spreading ridge and rift system (Horowitz 1979, p. 56), opening due to an east–west oriented regional extension, thus forming an incipient ocean (Mart 1991). Attempts have also been made for a "compromise" between the two groups, essentially suggesting oblique separation along the Jordan Rift Valley (e.g. see Vroman 1973, Mart 1991). The third approach agrees with the second in admitting only an east–west separation, but as this process is in its embryonic stage, when the plates are not (yet?) fully separated, to regard this feature as an incipient ocean is somewhat premature. At any rate, either theory or model connects the Jordan Rift Valley to the Red Sea, in some way or another, so a brief discussion of views on the origin of the latter is outlined below (see Section 10.1).

Needless to say, numerous other theories (see Chapter 2) were previously suggested, trying to explain the tectonics and origin of the Jordan Valley, but these three are the only ones holding until the present day, which naturally does not say that either is necessarily correct. As will be quite clear from the following, I am not in favor of the strike-slip transform model, which made me ask my friend and colleague Zvi Garfunkel to present it as an appendix at the end of this book. The differences between the second and third views are not so profound, so no particular discussion of the "incipient ocean" idea is presented. I do hope that in this way readers can judge for themselves what explanation is preferable, if any. In the same line of approach, the views and theories expressed in Chapters 4 and 8 of the present book, by Akiva Flexer, Avihu Ginzburg and Zvi Ben-Avraham, who also hold to the "strike-slip", are those of the authors.

As for myself, it all started almost 40 years ago, in 1961, when I was a third-year geology student. A meeting of the Association for the Advancement of Science in Israel was held in Jerusalem, at which one of the main topics was the structure and tectonics of the Jordan Rift Valley. The discussion involved a panel comprising the elder professors of the Department of Geology at the Hebrew University, and a young doctoral student, Raphael Freund. He was commissioned to present an old theory by Dubertret (1932), based initially on observations by Lartet (1869, footnote in p. 15), suggesting a left-lateral movement of some 160 km along the Jordan Rift. Somewhat before the meeting this theory was revived and refined by Quennell (1959), which renewed public interest. Freund had quite a challenge to convince all the others, who profoundly believed in the then popular "Rhinegraben" tensional theory, whose chief proponent was Picard (1943 and other publications), but he succeeded against all odds.

The success had twin fathers, professional and personal. These were the days of the advent of the Continental-Drift theory, which was acknowledged by numerous scholars as an elegant way of unifying the global tectonic processes. The suggested model, of lateral movement along a plate boundary, thus appealed to most scientists present at this discussion, since it viewed the Jordan Rift as an integral part of a worldwide system. The personal angle was not missing either, considering that it featured a young, good-looking, fluently speaking student suggesting a (seemingly) novel idea and model, against a bunch of oldtimers. Admittedly, not all were convinced (I was, for some time at least), but the number of "non-believers" quickly dwindled during the coming years. Notably, most of the latter were geologists who have personally been engaged with fieldwork in the Jordan Valley. I myself began to realize the obvious discrepancies several years later, when I was busy with my dissertation, in the northern Jordan Valley.

Until 1979, although I made several attempts, I could not even publish my "decadent" views, since no creditable journal would print such a sacrilege. But the principal problem involved with the "novel" trend goes far beyond my personal difficulties. It lies in the fact that most scientists living outside this region do not even realize that there may be a problem here, so that for instance almost all reconstructions of the Red Sea plate kinematics regard the lateral displacement of its northern extension as a proven fact that does not require any further consideration (see e.g. numerous papers in Purser & Bosence 1998).

10.1 THE RED SEA SPREADING SYSTEM

Regardless of which model is adopted for the tectonic development of the Jordan Valley, it seems there is no dispute that the Rift is a continuation of the Red Sea spreading system (Fig. 1.4), which makes it appropriate to give a short account of the evolution of this feature. Further, much more detailed information can be found in Coleman (1993) and in Purser & Bosence (1998) among other publications.

Several of the authors whose papers appear in the book edited by Purser & Bosence, as well as of numerous other publications, refer to the history of the Red Sea in three stages, denoting pre-, syn- and post-rift processes. It seems that some clarification of this terminology is in order, so that comparisons with the Jordan Rift can be made on similar grounds. "Pre-rift" is a term applied, in connection with the Red Sea, to those rocks which were deposited over a non-faulted substrate, not showing particular relations with rifting, but subsequently faulted. A distinction is however made in the present book between two groups of "pre-rift" rock formations of the Jordan Valley, based on their relations with the rift (see Chapters 5 and 9). One is an older group, which is typified by similar thicknesses within and outside the rift margins, thus comprising the real pre-rift domain. This is followed by younger sequences, characterized by increased thicknesses within the rift area, which denote subsidence of a synclinal or synclinorial type, but which are not deposited over a faulted terrain. I think that the synclinal subsidence is intimately connected with the rifting to come, and should therefore be, and is in this book, referred to as the "Embryonic stage" of the rifting process.

It seems there is no dispute concerning the application of the term "syn-rift", which is self-evident, when referring to those formations usually limited to the subsiding rift area, deposited over a clearly faulted substructure. "Post-rift" either refers to rocks laid down following oceanization, which is the stage when magmatics begin to be intruded into the rift's axial zone as a series of dikes, or to rocks horizontally overlying previously faulted formations. The syn-rift stage of the Red

Sea, according to those who favor also using the term "post-rift", spans the Oligocene and Miocene (or at least its early and middle parts), periods during which the province was extended and faulted, before the process of its oceanization was begun (Chazot et al. 1998).

The validity of the term "post-rift" could easily be disputed (and frequently is), since rifting and faulting continues in the core of the Red Sea almost to the present day (Horowitz 1967, Carbone et al. 1998). Numerous authors, such as Coleman (1993, p. 27) or McClay et al. (1998) ignore this term altogether, using the "syn-rift" heading for all sediments accumulated within the Red Sea rift ever since faulting began. This logical approach is also adopted here. It is anyway of no consequence for the Jordan Rift Valley, where neither phenomenon related to "post-rift" is relevant.

10.1.1 Embryonic stage?

It appears that the earliest possible indications of an embryonic stage (in the sense applied to the Jordan Rift, see above) in the Red Sea area could date to the late Cretaceous. A Maastrichtian–Paleocene arm of the Tethys Ocean invaded the area now occupied at least by the northern half of the Red Sea, down to Jiddah (Jeddah) and somewhat north of Port Sudan (Coleman 1993, p. 26). Rocks were deposited in a variety of environments, fluviatile, lagoonal and marine, all very shallow, over a practically flat relief. A comparable sequence in almost every aspect is also known from western Yemen (Al-Subbary et al. 1998), whose age may extend up to the early Oligocene. These sequences usually attain only several tens of meters in thickness, up to 400 in a place or two in western Yemen.

The earliest magmatic activity possibly connected with the "embryonic" stage of the Red Sea is known from Ethiopia, where the Ashangi Formation basaltic eruptions are considered to be of middle(?) Eocene age (Coleman 1993, p. 79). Coleman regards the evidence for contemporaneous volcanism in Yemen as "unconvincing". He also states that "Assuming these ages to be correct, these middle Eocene mafic magmas could signal the earliest phases of extension. However, there is no structural evidence of regional faulting at this time". Al-Subbary et al. (1998) conclude (p. 134): "The evidence from the sedimentology and stratigraphic relationships in the Medj-Zir Formation shows that during the Paleocene to Oligocene period prior to volcanism and rifting in the southern Red Sea the region was tectonically stable. There is no evidence for doming, peneplanation or significant subsidence in the pre-rift stratigraphy." The question which deserves more attention than is usually given to it, is how "significant" should subsidence be to justify regarding it as an "embryonic" stage in a rift's development.

There seems to be general agreement (Coleman 1993, and most of the papers in Purser & Bosence 1998) that the main phase of volcanism of the Red Sea commenced some 30 Ma ago, in the Oligocene. Uplift and rifting, or at least faulting, began somewhat earlier, since the middle (and possibly early) Oligocene sediments overlie a faulted substrate (Omar & Steckler 1995). The question which remains to be answered is, were the Maastrichtian–Eocene sequences deposited in an already slightly downwarping basin, which could be regarded as an embryonic stage, or they are truly pre-rift sediments, possibly preserved better in the Red Sea region due to its subsequent lower elevations, as compared with the uplifted margins.

The main rationale for defining an embryonic stage in the Jordan Rift was the excessive thicknesses of Oligocene through middle Miocene sequences, in comparison with the neighboring regions. This is certainly not the case with the Red Sea but, on the other hand, all we know there is the marginal facies of the Maastrichtian through Eocene formations, which may or may not thicken toward the central parts. In the Gulf of Suez, which in any case is only a subordinate branch of the main Red Sea rift, these sequences do not seem to thicken toward the center of the basin. The question remains, then, is the "pre-rift" Maastrichtian through Eocene sequence thickening toward the center of the Red Sea, in which case these periods verily represent an "embryonic" stage, or is the thickness maintained approximately constant, indicating a true "pre-rift" domain. This important question, in the present state of knowledge, remains open (and, at any rate, is quite far from the scope of this book).

It is my own notion that there could be a slight subsidence of the Red Sea region, in its wider geographical sense, remotely connected with the Syrian Arc folding of the Levant. No doubt these two processes occurred simultaneously, during the same time span. It is then quite plausible to assume that the northward movement of the Afro-Arabian plate toward Eurasia caused some subsidence along the future lineament of the Red Sea. The idea of a Precambrian geosuture along this line, which dictated the location of the lines of weakness, was proposed by several investigators (Dubertret 1970; Coleman 1993, p. 85 and references therein; Rihm & Henke 1998, among others).

10.1.2 Syn-rift stage

Coleman (1993, p. 27) subdivided the syn-rift sedimentary sequence of the Red Sea into four categories: initial rift continental sediments, syn-rift marine sediments, evaporiates, and post-evaporite marine sediments. The first three are also considered by Purser & Bosence (1998, p. 5) as "syn-rift" formations, while the latter one already makes the "post-rift" stage.

The "initial rift continental sediments" span the Oligocene through early Miocene period, occurring all along the Red Sea, from the Gulf of Suez to the north down to the Gulf of Aden. This group comprises a variety of formations, predominantly continental, usually deposited in fluviatile or lacustrine environments, over angular unconformities, postdating the initial faulting phase of the Red Sea. The thicknesses are, in most cases, in the order of only several tens of meters, occasionally up to a few hundreds, but these figures are always obtained for sequences located quite far away from the center of the rift. In comparison, sequences attaining up to 2,500–3,000 m of late Eocene through early Miocene rocks of the Dogali Formation, which also include volcanics, have been penetrated by boreholes drilled off the shores of Ethiopia and southern Saudi Arabia (Savoyat & Balcha 1989).

Thin intercalations of shallow marine sediments are known from almost all occurrences of the initial rift continental sediments. Fossils indicate two transgressive cycles, the earlier usually defined as "late Oligocene", occasionally labeled "Chattian", followed by an early Miocene, the latter less common. The "late Oligocene" needs clarification: whenever foraminifera were present, it was defined as Zone 2 of Blow (1969), which is recognized here (see Table 1.4.1 and accompanying discussion) as middle Oligocene. Thus it seems that the middle Oligocene Tethys Ocean transgressed along the Red Sea region all the way down almost to its southernmost tip (Coleman 1993, p. 30). On the other side, the Indian Ocean waters covered the Gulf of Aden at the same time (Fantozzi & Sgavetti 1998), but the two oceans apparently did not meet.

Based on the sedimentological and magmatic occurrences, Coleman (1993, p. 32) concludes: "The presence of the continental to lacustrine alluvial sediments associated with volcanic eruptions along the Red Sea margins indicates that the initial Red Sea Basin was not part of a large Nubian–Arabian dome, but consisted of a continental rift valley or a series of graben depressions receiving continental clastics and bimodal volcanic flows with deposition within fault bounded lake basins". Evidently, the added contribution of the Tethys should be considered.

The "initial rift" was a period of extensive magmatic activity, particularly at the southern end of the Red Sea, where the large volcanic fields of Ethiopia and Yemen had been extruded. Notably, however, Oligocene volcanics were lately reported also from eastern Transjordan (Tarawneh et al. 2000); it is my hunch that they are much more common, but in most localities are covered by younger effusions. Chazot et al. (1998) refer to the Oligocene volcanic and intrusive rocks of Yemen, formed during the interval 31–26 Ma, as "pre-rift" activity, stating that (p. 51) "it is apparent that eruption of several thousand meters of the Yemen large igneous province from 31 to 26 Ma predated break up and erosion of the Yemen margin by ca. 5 Ma (i.e. pre-rift volcanism)". This is in clear contrast to the conclusion of Omar & Steckler (1995), who indicated that uplift began some 34 Ma ago, and to the occurrences of Oligocene sediments over faulted relief within the Red Sea, discussed above.

The term "syn-rift marine sediments" refers to considerable accumulations of deep marine deposits in the central parts of the Red Sea, attaining up to a few kilometers in thickness, grading landward at both flanks through littoral into mainly fluviatile formations. Their age is generally referred to as early through middle Miocene. Typically these rocks, known all over the Red Sea from the Gulf of Suez down to Ethiopia and western Yemen, contain fauna indicating the above time span and Tethyan affinities, developed over a strongly faulted substrate. Middle Miocene oyster banks are also known from the Gulf of Aqaba, all along its length.

Correlative sediments of the Gulf of Aden contain fossils of the Indian Ocean province, again not contacting the Tethyan.

The syn-rift marine sediments are intercalated all along the Red Sea, but especially on its eastern rim, by magmatic rocks, with particularly large piles of lavas in Yemen. These volcanic formations, mainly consisting of alkali olivine basalts, but accompanied also by numerous intrusive bodies, are known as far as several hundred kilometers east of the Red Sea, extending from Yemen north to Transjordan and Syria. Rare occurrences, usually of small-scale intrusions, are also known to the west, extending up to several tens of kilometers west of the Gulf of Suez. Most of the intrusive bodies comprise rather long dikes, termed the "Red Sea Dike system" (see Section 5.3.1).

The suite of deep marine sediments is overlain, again all along the Red Sea, by considerable volumes of evaporites, occasionally attaining up to 3–4 km in thickness. The evaporites are interfingered in their lower sectors by deep marine deposits, indicating the interplay of the middle Miocene high global sea level and the tectonic activity of the Red Sea system. Evaporite deposition commenced some time in the middle Miocene, usually explained as due to occasional disruptions of the Red Sea–Mediterranean connection, which was always quite narrow, and thus very sensitive to even minor structural disturbances. This was most probably helped by the middle Miocene drop in the Mediterranean Sea level, as seen from offshore boreholes of the southern Levant (see Table 1.4.1). Evaporites deposition continued intermittently up to the late Miocene, when the sharp Messinian drop in Mediterranean sea level caused desiccation also within the Red Sea domain. Volcanic activity, accompanied by extension and enlargement of the Red Sea basin, continued throughout this period, but uplift was minimal in the adjoining regions (Coleman 1993, p. 35). It seems that in this period, some 10 Ma ago, the initial oceanization of the Red Sea began, by intrusion of basalts into its axial zone.

Evaporites deposition stopped quite abruptly at the Miocene–Pliocene boundary, following a considerable faulting phase (Mart & Ross 1987). The faulted and truncated evaporites are unconformably overlain by nanno oozes, including silty clays, chalk and dolomites, all of which contain pelagic fossils with Indian Ocean affinities (Dubertret 1970). This unconformity, marking the second stage of sea floor spreading in the axial trough of the Red Sea, is present throughout the main trough, but does not extend into the axial zone. The pelagic sediments are replaced by carbonate platform reefy shallow marine rocks at the margins of the basin, grading to fluviatile further landward. This style of deposition continues, with minor changes related to oscillating climates and sea levels, until the present-day, accompanied by volcanism, particularly on the eastern flank of the Red Sea but also in Afar, the emplacement of oceanic crust within the axial zone, and considerable uplift of the neighboring regions.

Concluding, Coleman (1993, p. 39) raises the following points:

(1) There is no evidence for pre-Cenozoic sedimentation trends following the Red Sea lineament.

(2) Intermontane continental red-beds and fluvial (and marine!) deposits follow Red Sea tensional faulting in the Oligocene. The presence of these negates the idea of doming prior to rifting, as also indicated in Purser & Bosence (1998, p. 5).

(3) Extension and widespread early Miocene volcanism both define the shape of the Red Sea basin.

(4) Evaporite sedimentation is correlative to the Messinian and confirms the closed basin nature of early extension in the Red Sea (but as it seems, closure already began in the middle Miocene, see above and Bosworth et al. 1998).

(5) A major tectonic event dates to the Miocene–Pliocene boundary, which denotes the transition to open ocean deposits, connected to the Indian Ocean instead of the Mediterranean or Tethyan domain, as during the Oligocene and Miocene.

(6) Coarse continental derived clastics mark the nearly synchronous uplift of the Arabian, Yemen, Somalian and Ethiopian flanks.

(7) Sedimentation rates in the Red Sea follow these uplifts, and are high to the south, decreasing northward.

10.1.3 Conclusion: structural evolution

The earliest phase ("pulse") of uplift along the entire length of the Red Sea is dated (Omar & Steckler 1995) to approximately 34 Ma ago, some time at the beginning of the Oligocene. The uplift must have been followed by the formation of some kind of a neighboring depression, possibly (but not necessarily) rifting, since the dating is based on apatite fission tracks, denoting erosion. This conclusion is also supported by finds of middle Oligocene marine sediments overlying a faulted relief in several localities along the Red Sea (see above). The amount of extension is quite limited at this stage. The second uplift "pulse" began in the early Miocene, some 25–21 Ma ago, marking the start of the main phase of extension, which is manifested by considerable accumulations of marine sediments of that age along the Red Sea. At the same time swarms of tholeiitic dikes were intruded into the country rocks, particularly east of the of the Red Sea but some also to the west, and are known from Yemen up to east of the Dead Sea (Coleman 1993, p. 127; and references therein).

Omar & Steckler (1995, p. 1341) conclude: "These data support a rigid plate model for continental extension. These results also indicate that the initiation of rift flank uplift, and therefore rifting, and volcanism occurred nearly simultaneously. This conflicts with classical models of active and passive extension that predict sequential development of these features." Notably, the Oligocene uplift and volcanism of the southern Levant support the above model of a rigid plate.

The third pulse, expressed by both extension and uplift (Coleman 1993, p. 128; and references therein), is dated at around 12–10 Ma ago, again affecting the entire length of the Red Sea (McClay et al. 1998), and may have been accompanied by the initiation of the axial formation of oceanic crust, at approximately the same time (Rihm & Henke 1998). The fourth pulse of extension is characterized by rapid dike injection into the southern and central sectors of the axial trough of the Red Sea (Coleman 1993, p. 129, and references therein; Purser & Bosence 1998, p. 4), causing sea-floor spreading. This process, which commenced some five million years ago, marks the transition of the Red Sea to its oceanic, post-rift stage (for those using this term).

10.2 THE JORDAN RIFT VALLEY: A CONTINUATION OF THE RED SEA SPREADING SYSTEM

The Levant is squeezed between two large continental masses, Africa–Arabia to the south and Eurasia to the north, whose relative movements dictate its tectonics, at least ever since the Tethys Ocean area was being reduced, some time in the Cretaceous, by the two approaching continents. The northward movement of Africa–Arabia can be grossly divided into two main stages, the earlier, when Arabia was still an integral part of Africa; and the later, when these two plates began to separate along the Red Sea Rift system. The lines along which separation was active are most probably dictated by zones of weakness in the Arabo-Nubian crystalline massif, generally referred to as "geosutures", which could have been formed by incomplete continental or island-arc accretion at the end of the Precambrian, or possibly other causes. The breakup of the Arabo-African continent into the main African and Arabian plates was accompanied by formation of several smaller ones, among which the more conspicuous are Sinai to the north and Somalia to the south (Dubertret 1970).

The African, Sinai and Arabian plates are still moving northward, the latter two approximately at a similar pace; ever since the early Oligocene the three have also been drifting away from each other; this combined movement is responsible for the structural disturbances and reorganization of the southern Levant (among other regions), expressed also by the formation of the Jordan Rift Valley.

The continuing northward movement of the African–Arabian plates caused a gradual shrinking of the Miocene Tethys, to become the Mediterranean. This is expressed by hiata in the sedimentary sequence of the southeastern Mediterranean, where Foraminifera Zones N10–N12 are missing. To the north this uplift was compensated by large-scale subsidence, reopening the seaway between the northern Mediterranean and the Indian Ocean through the Persian Gulf, which was blocked during the Oligocene. Following this event, the Red Sea became partly isolated from its "parental Mediterranean, by a topographic barrier situated north of the Gulf of Suez" (Purser & Bosence 1998, p. 6), thus depositing considerable sequences of evaporites even before the Messinian salinity crisis. These are very abundant in the Gulf of Suez and along almost the entire Red Sea, and are also known from the Gulf of Aqaba (Mart & Rabinowitz 1986). The further northward drift of the African–Arabian pair, during the late Miocene, completely isolated the Mediterranean from both the Atlantic and the Indian oceans, causing its almost total drying up.

The later Miocene and the beginning of the Pliocene saw an easing of the previously northward-oriented compression caused by the drift of Africa, Sinai and Arabia. This created large-scale extensional structures (and volcanism) all along the Alpine belt from western Europe through southeast Asia (Gignoux 1955, p. 577), including the Middle East, so extensive that one could suggest not only an easing of compression, but rather a slight southward retreat of Africa and Arabia in relation to Eurasia. This extension phase also reopened the Straits of Gibraltar, through which the Atlantic Ocean again filled the Mediterranean, itself becoming considerably deeper by the easing of compression and subsidence of extension structures. The barrier north of the Gulf of Suez remained elevated, but on the other hand the slight southward drift of the Somalian plate, resulting from extension in the Gulf of Aden, opened the entire Red Sea domain to the Indian Ocean, a situation persisting to the present day (Dubertret 1970, and many others).

The timing and locations of these changes in the Mediterranean, which are not correlative to any of the rifting and associated phenomena of the Red Sea system (except for the extension mentioned in the Gulf of Aden), seem to indicate their association with the mutual drifting of the African–Arabian continent, rather than a result solely of Red Sea kinematics connected with differential drift of the two land masses.

Tectonic modeling for the Jordan Rift Valley should answer several questions pertaining to its shape, structure and temporal evolution, detailed in Chapter 9. The initial, Embryonic-stage activity along the Jordan Valley commenced some time in the early Oligocene, comprising accelerated subsidence along the would be Rift region, but not accompanied by any noticeable faulting observed neither on the surface nor in the subsurface (see Chapter 8). This phase of slow synclinal subsidence, with rates of less than 100 mm/Ka, was accompanied by a wide, long wave undulation of large areas in the southern Levant, a process which lasted approximately 25–26 million years, until the beginning of the Eritrean stage. A somewhat accelerated uplift, within the period of the Embryonic stage, affected mainly the southern parts of Israel and Jordan, causing changes of relief and hydrography during the Oligocene–Miocene transition (Fig. 10.4.2).

The succeeding Eritrean stage lasted from the late Miocene until the end of Palynozone QI, an interval of some 6–7 million years, but tectonics were active mainly at the beginning and middle of that period, in two distinct waves. The Eritrean stage is characterized by extensive tensional structures and volcanism, with only subordinate uplift. This stage affected the entire Levant and beyond, including the area subsequently modified by the Levantine stage, in parts of which subsidence was more pronounced, up to rates of 500 mm/Ka. The Eritrean was followed, after a period of a few million years of tectonic rest, by the Levantine stage, which shaped the region into its present form, including the north–south depression and the longitudinal elevated terrains on both its flanks. This most conspicuous phase, which has gone on for the last two million years, is characterized by the subsidence of several basins along the north–south oriented Rift, with rates occasionally exceeding 10,000 mm/Ka, but only subordinate volcanism.

10.2.1 Previous structures

Structures predating the Jordan Rift Valley which could have had some effect on its tectonics must also be taken into consideration. These include mainly the late Precambrian "geosuture" (Section 9.2, Fig. 9.2.1) and the prominent late Cretaceous through Eocene Syrian Arc folding (Sections 4.8 and 9.3, Fig. 4.8.1). The Transversal fault system's (Section 4.9, Figs 10.2.1 and 10.2.2) principal activity postdates the early Oligocene, although at least one of its faults shows subordinate activity already in Senonian times. This system is regarded here as an integral part of the extension involved with the Jordan Rift formation. Other earlier Mesozoic or Paleozoic tectonic phases detailed in Chapter 4 do not seem to have intimate connections with the later Cenozoic processes.

The concept of a geosuture (also referred to as "zone of weakness" or "hinge line") pattern that controls the major lineaments of the Syrian–African Rift system is quite popular among numerous scientists (Dubertret 1970; Bender 1974a, p. 124; Rihm & Henke 1998, and references therein). It seems that the term "geosuture" was used for two different aspects of the Precambrian shield. One pertains to its external boundaries, better described as the "hinge line", while the other refers to true zones of weakness within the continental plate, which should thus be seen as being in accord. The occurrence of large-scale late Precambrian volcanics and conglomerates is considered one of the most conspicuous indications for the existence of a geosuture. These rocks are indeed abundant both along the Red Sea, where they denote a zone of weakness, and along at least the northern edge of the Arabian–African continent, delimiting its plate boundary in the manner of a wide hinge line (or rather hinge zone).

The latter occurrence is known from east of the Dead Sea, striking south along the Arava, but veering west somewhat north of Elat toward southern Sinai and Egypt (Bartov 1994), thus characterizing the northern edge of the Arabo-Nubian late Precambrian continent in the southern Levant. The structure of the Precambrian basement in Israel also shows lineaments parallel to this hinge zone (Rybakov et al. 1999). The Precambrian Volcano-Conglomeratic Complex, besides turning westward into Sinai, also continues southward, to join similar occurrences along the Red Sea, in the manner of a true zone of weakness, which acted as a "stress guide" for the initial opening of the rift system (Rihm & Henke 1998, and references therein).

The shape of this continent edge, or the hinge line, has dictated the tectonic and sedimentary lineaments of the Levant ever since the Cambrian, continuously until the present day (Figs 4.3.1, 4.4.1, 4.5.2–4.5.4, 4.7.1 and 4.8.1), as has been noted

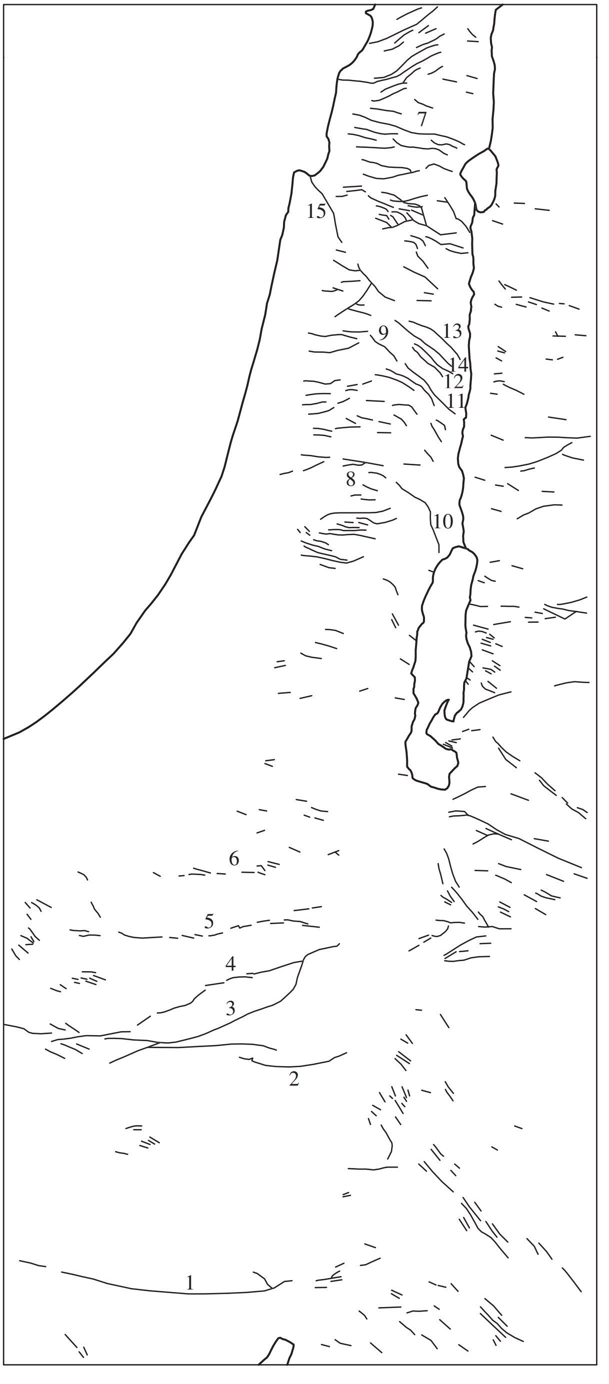


Figure 10.2.1. The transversal and Eritrean fault systems on both sides of the Jordan Rift Valley, from various geological maps: (1) Thamad– Wadi Sudr line, (2) Paran–Areif en Naqa–Buruq line, (3) Arif–Batur line, (4) Ramon– Minshara line, (5) Sa'ad-Nafha–Helal line, (6) Zin line, (7) Galilee, (8) Judea, (9) Samaria, (10) Samia fault, (11) Fari'a fault, (12) Buqei'a fault, (13) Gilboa fault, (14) Malih fault and (15) Carmel fault.

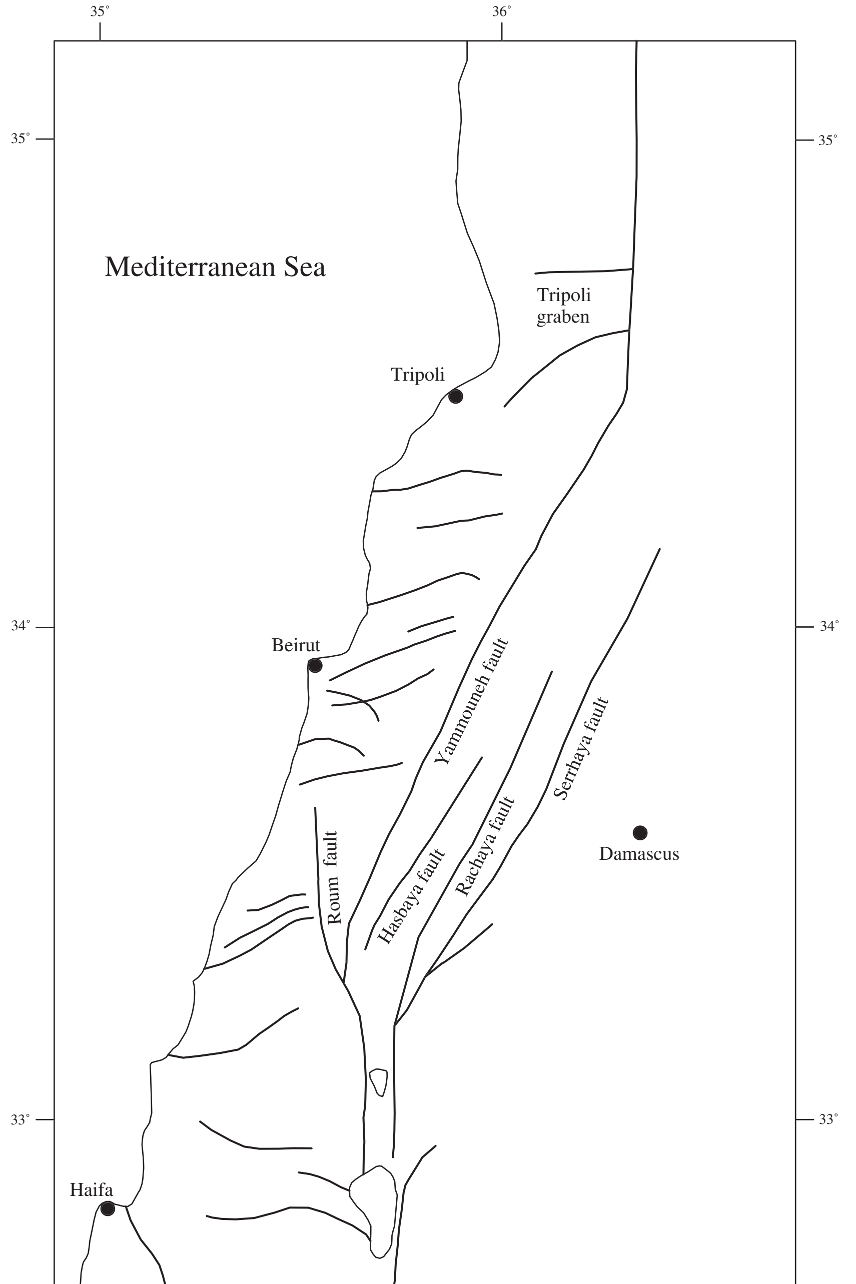


Figure 10.2.2. Continuation of the transversal and Eritrean fault systems north of the Jordan Rift Valley, mainly after Dubertret (1962, 1970).

by numerous authors (among whom are Dubertret 1970; Bender 1974a, p. 129; Horowitz 1979, p. 58; Mart 1991). Based on geophysical studies of the basement in Israel, this is summarized in Rybakov et al. (1999, p. 101): "Specifically, we find that the deep structures in the area have the same orientation as the later Syrian Arc structures, suggesting that the younger features are controlled by an older structural framework." Rybakov et al. (2000, p. 30) further elaborate: "The Bouger and Freeair gravity and seismic refraction data suggest a low-density continental crust east of the DST (Dead Sea Transform) and a dense suboceanic crust to the west of the DST. We assume that this difference existed before the DST motion was initiated. The development of the DST used the boundaries of these different blocks as a weakness zone". Indeed, the same orientation is still maintained in the southern Levant coast line (although to the north it has deviated, ever since the Oligocene, due to uplift connected with the rifting).

It is quite clear from the above that the conspicuous "S" shape of the Syrian Arc fold belt was inherited from the hinge line, not from approaching the Jordan Rift Valley, which did not exist at the time of folding. This fold belt, which overlies deep seated basement reversed faults (Fig. 4.8.2), was formed as a result of the pressure generated by the African–Arabian northward drift, during the late Cretaceous through Eocene. It is regarded here as a compressional tectonic domain, not involving any magmatic extrusions, predating and different from the extension which created rifting, large-scale volcanism and all other associated phenomena. The difference in tectonic style can also be seen in the Arabo-Nubian Massif, whose uplift began only in the early Oligocene. The large wide, shallow bay of the Tethys, which transgressed down to Jidda during Maastrichtian through Eocene times, is regarded as an outcome of the rise in global sea level at that time, which also flooded the entire southern Levant (see Section 4.6).

The accentuation of several Syrian Arc structures is however apparent also in younger periods, resulting from a combination of two processes. The northward movement of Africa and Arabia continues until the present day; and accentuation is apparent where the rifting involved uplift, so that earlier structures such as the Ramon, Ajlun and northern Negev anticlinoria (Fig. 4.8.1) became further elevated, causing the false impression that the Syrian Arc folding continued at least until the Quaternary (Shahar 1994, and to be honest, I myself fell into this "trap", and wrote so in some of my earlier publications).

10.2.2 Embryonic stage

Initial uplift, rifting and volcanism all began along the Red Sea system at approximately the same time, namely the early Oligocene. It was generally thought that Oligocene volcanism was restricted to the southern end of the Red Sea at that time, arriving at its northern reaches only later, at the beginning of the Miocene, but recent datings have revealed Oligocene volcanics also from the El Shamah–Jebel Druze field, in eastern Transjordan (Tarawneh et al. 2000). The Jordan Rift Valley is no different to the Red Sea in this respect, only expressing it in a more humble way. Uplift, roughly following the shape of the Precambrian hinge line, is apparent for the early Oligocene when paleogeography is concerned (see Section 7.2.1, Fig. 10.4.2), but faulting did not reach the surface until much later. The question of whether faulting already affected the subsurface during the Embryonic stage remains a puzzle at present. It is impossible to be sure whether the Kinneret and southern Dead Sea regions, where Oligocene through middle Miocene subsidence was recorded (but only a few hundred meters for the entire Embryonic stage), were at that time underlain by basement faults, forming slowly subsiding sag basins. The northernmost area hit by faulting reaching the surface, during the entire Embryonic stage, is the Gulf of Aqaba all along its present-day extension.

The extended subsidence of the Red Sea and Gulf of Suez, at the very beginning of the Miocene, is expressed by further uplift along the hinge line, which caused the change in the Oligocene paleogeography of the southern Levant into the Miocene landscape (see Sections 7.2.2 and 7.2.3, Fig. 10.4.2). This uplift is apparent all along the Red Sea (Omar & Steckler 1995). The actual rates and style of subsidence along the Jordan Rift Valley did not change much during this transition. Oligocene volcanism is known from eastern Transjordan, but magmatism in the closer area of the Jordan Valley region is known only some 20–19 Ma ago, in the form of minor intrusions at the Karak region east of the Dead Sea and at Ashosh, west of the northern Arava. It is not clear whether these reached the surface at that time. Considerable volcanism covered some areas of the southern Levant during the middle Miocene.

It is quite clear then that the Jordan Rift Valley was an integral part of the Red Sea system ever since its inception in the early Oligocene, during the entire Embryonic stage, as both regions had been developing along similar lines, at approximately the same times. It seems of no great consequence in which manner exactly the complex process of extension along the Red Sea was manifested in the southern Levant. It is quite clear however that uplift, volcanism and subsidence seem to be broadly synchronous in the two regions, supporting the "rigid plates" model proposed by Omar & Steckler (1995).

Previous ideas (Horowitz 1979, p. 51; Mart 1991) suggesting that the Gulf of Suez was the primary continuation of the Red Sea spreading center, while the northward bearing of the Gulf of Aqaba only developed in the Quaternary due to a change in the direction of spreading, thus have no support in the more recently acquired data. Rather, it appears that both gulfs were formed simultaneously, the Suez being a relatively shallow offshoot of the system, while the north–south oriented deeper feature of the Gulf of Aqaba, extending to the Jordan Valley, could have been the main tectonic lineament throughout the entire period of activity (Fig. 10.4.1).

A particular role is played by the Eritrean transversal fault system of the southern Levant (Figs 10.2.1 and 10.2.2), which became activated some time during the Embryonic stage. The system cuts the Red Sea Dike system (Bartov 1974), but could also have been active before, since some of the dikes intrude, or are directed by parts of these faults. It seems that the difference in width of the Red Sea and its northern extension, through the Gulf of Aqaba, the Jordan Valley and possibly further north in the Lebanon and Syria, is compensated on these faults. Notably, they are dextral strike-slip faults in the Negev and Sinai, while in Saudi Arabia their movement, which never exceeds a few kilometers, is left-lateral (Bartov 1994). Unfortunately no clear indication of the sense of movement is available from Transjordan. The east–west direction of all these faults, which is seen in Egypt, Sinai, Israel, Transjordan and Saudi Arabia (and possibly also Lebanon), as well as the main bearing of the central African fault zone, together with their trends of lateral movements, indicate that the regional extension is latitudinally oriented, while the entire spreading system is centered approximately along the 35°E meridian (Fig. 10.2.6).

The deviation of the Red Sea from the meridional direction most probably results from the locations and bearing of much older late Precambrian geosutures within the Afro-Arabian continent. As Rihm & Henke (1998, p. 45) summarize: "In the case of the Red Sea, the line of initial rifting, as activated in late (should rather be "early", A.H.) Oligocene, was located along a zone of structural weakness created during the Pan-African (late Precambrian) period of island-arc accretion … which defined the first-order orientation of the rift. As secondary mechanism, another pre-existing fault system, the – roughly east–west oriented – Central African Fault Zone was utilized for adjusting the first-order orientation to the regional plate kinematic pattern."

10.2.3 Eritrean stage

The Eritrean stage is typified in the Jordan Valley, and indeed in the entire southern Levant including the eastern Mediterranean (Udintsev et al. 1994, Kempler 1994, Yilmaz 2001), by two consecutive stronger normal faulting phases, active approximately along the same lines. A parallel activity is also recorded in the Red Sea by two stages of oceanization, apparent from injection of mid-ocean ridge basalts into its axial zone, estimated as having occurred some ten and five million years ago, respectively (Rihm & Henke 1998). The exact timing of these events in the Red Sea, the Jordan Valley and the eastern Mediterranean still needs refinement, but they seem broadly coeval.

Three principal bearings characterize the Eritrean stage, north–south and NNW–SSE normal faults (although the latter change direction approaching Eurasia), and east–west oriented wrench faults (Figs 1.1, 4.8.1, 10.2.1 and 10.2.3). The more interesting factor during this stage is the minimal uplift involved in the structural changes which took place throughout the late Miocene and Pliocene. This pattern, which extends over regions far away from the Red Sea (Bartov 1994), gives the impression that the main force behind this faulting phase should be sought somewhere outside of this system. It appears that the "backward" drift

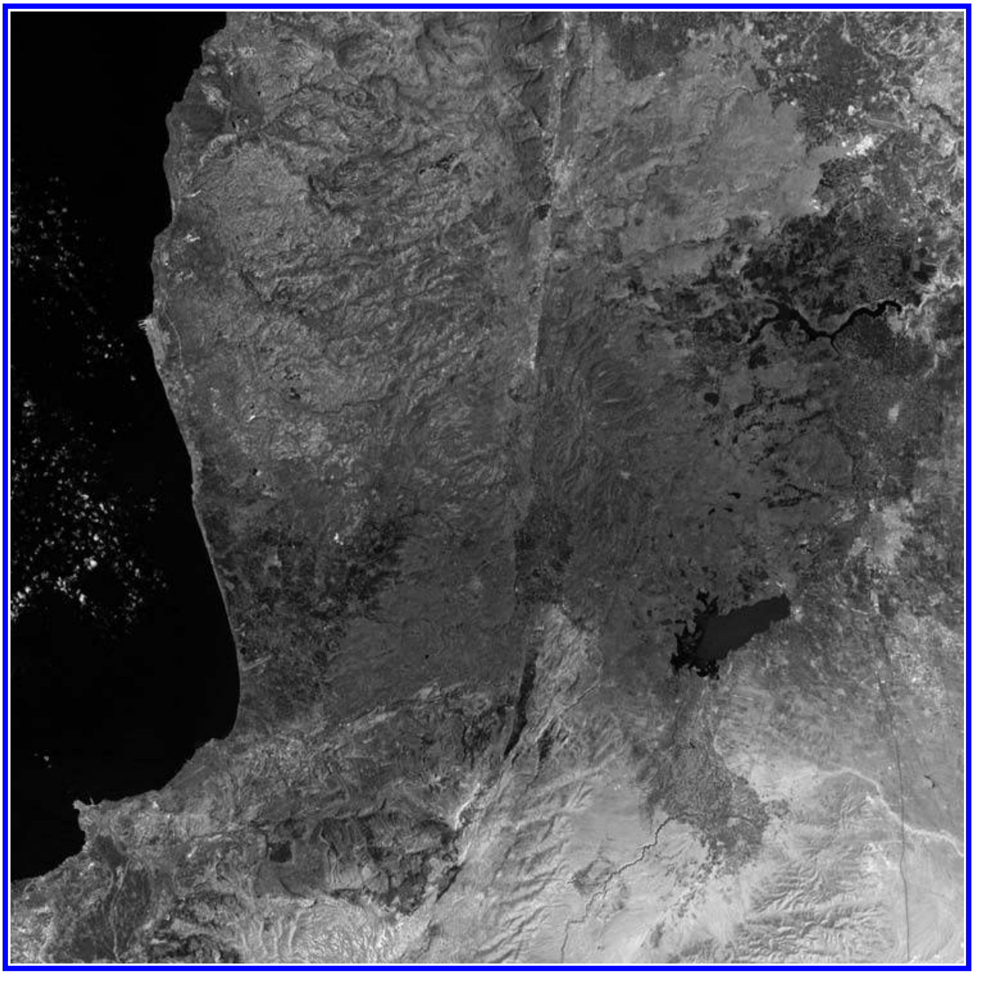


Figure 10.2.3. Satellite image of northern Lebanon. The city of Tripoli is at the southern tip of the bay, itself a result of Eritrean faulting which created the Tripoli graben, extending landward and clearly seen on the image. The Yammouneh fault runs along the image, approximately at the center, overlain by unfaulted Homs Basalt (coordinates 3800–3900N; 200–300E; UTM Grid; 100 × 100 km). Taken from Hall (2000) and based upon an EarthSat GeoCover™ mosaic with LANDSAT 28.5 m imagery. By permission of the Geological Survey of Israel.

of Afro-Arabia, recorded at approximately the same times, could be a good cause for this new tensional regime, combined with some further opening of the Red Sea, most probably along its axial zone. Attempts at explaining the synchronous tensional features of the eastern Mediterranean as of "transtensional" origin (Kempler 1994) seem unjustified, since in such a case compression must occur somewhere, for which there is no convincing evidence anywhere around the region. The widespread volcanism of this stage, again also known from distant regions, is further testimony to a large-scale tensional field, particularly active during the early Pliocene.

It is proposed to regard the Eritrean-stage activity along the Red Sea system, including the Jordan Rift Valley, as an outcome of a much larger phenomenon, encompassing the entire south Eurasian–Mediterranean–African–Arabian domain. A variety of rift valleys or grabens, from very small to country size, were formed in numerous localities in these regions trending in different directions, of which the principal ones are north–south and NNW–SSE. Most of these extensional structures are bounded by normal faults, with their pattern compensated by a suite of mostly east–west oriented wrench faults. The sense of movement of these depends on the relation of the plates where they are located with the general trend of movement. It thus seems (Fig. 10.4.1) that the principal plates of Arabia and Africa were separated along a north–south oriented zone, at least several hundred kilometers wide and quite complicated, now known as the Syrian–African Rift Valley system. The African–Arabian east–west oriented separation was accompanied by some relative southward drift of these plates, which drew away from Eurasia at the same time.

The initial phase, some time during the late Miocene, is the first time when faulting is apparent within the limits of the Jordan Rift Valley, but also in neighboring and even distant regions, such as the Wadi Sirhan graben formed in eastern Transjordan, or similar structures of the eastern Mediterranean. These were accompanied by the injection of oceanic crust into some parts of the Red Sea, as well as the continuation of the oceanization of the Gulf of Aden, processes which caused a slight clockwise rotation of Africa, enough for it to collide with Europe at Gibraltar, which started the drying up of the late Miocene Mediterranean, or the Messinian salinity crisis. The Red Sea system was still connected to the Mediterranean at that time, and was thus also hit by the salinity crisis.

The second, considerably stronger phase of the Eritrean stage marks the Miocene–Pliocene transition. The further combined movement of Africa to the south and west made two major oceanographic changes. On the one hand, there was the flooding of the Red Sea by Indian Ocean waters, all the way north to the gulfs of Suez and Aqaba, by opening the straits at Bab el Mandeb; and on the other, the creation of a new Mediterranean. A barrier was still uplifting north of the Gulf of Suez, slightly but enough to end the connection of the Red Sea with the former Miocene relic of the Tethys; the Gibraltar Straits opened the way for Atlantic Ocean water to the Mediterranean, itself becoming considerably deeper. This extension phase, characterizing the entire Pliocene, is also accompanied by widespread volcanism but very little uplift. Deepening of the Yizre'el Rift Valley made the Jordan Valley an integral part of the Mediterranean, while the parallel but shallower Nahr el Kebir graben in the Lebanon connected the Beqa'a to the sea.

Toward the end of the Eritrean stage tectonic activity subsided, and when the global sea level dropped in the latest Pliocene, the drainage pattern typical for Palynozone QI times, of rivers crossing the southern Levant westward draining the eastern highlands to the Mediterranean, evolved (see Section 7.2.6, Fig. 10.4.2). This relatively quiet period, lasting no more than a million years, may represent the "reconsideration" stage, during which the direction of drift of Africa–Arabia again changed, to resume its pre-Pliocene route to the north (Fig. 10.4.1). Consequently, the eastern Mediterranean also resumed its compression regime during the last 2 Ma (Mercier et al. 1987, Yilmaz 2001). Unfortunately not enough detailed information is available for the Red Sea.

10.2.4 Levantine stage

The Levantine stage, which shaped the southern Levant in its present form, is hardly expressed in the Red Sea system as an independent phase, while the Gulf of Suez seems to have been a rather quiet region since the beginning of the Pliocene (McClay et al. 1998). This results possibly from the strengthening of the principal activity along the Gulf of Aqaba and the Jordan Valley at that time (Mart 1982, 1991), accompanied by weakening of the "Eritrean" bearing. It should however be realized that the Levantine stage, although responsible for a considerable change in the geography and hydrography of the southern Levant, does not necessarily involve a major regional tectonic disturbance. Practically, an extension of 1 or 2 km across the Jordan Valley would do the entire job, since its bordering faults are quite steep (Kashai & Croker 1987), usually in the order of 70–80° (Fig. 3.1.6). This stage is therefore regarded here as a mere subordinate northward expression of activity of the Red Sea spreading system.

Volcanism throughout the Levantine stage is apparent, but by no means really widespread. So is uplift, which accompanies and goes in step with subsidence. When volumes are considered, all phenomena concerned are indeed quite limited, which may explain why this stage is not so pronounced when the entire Red Sea system is viewed, but nonetheless, due to local conditions, it is so significant for the southern Levant.

The northward drift of the Sinai–Arabian plates is apparent from the location and shape of the Hellenic–Persian convergence zone (Le Pichon & Angelier 1979). The fact that this zone does not extend further westward along the Mediterranean indicates that, at least at present, there is no significant northward drift of the entire African plate in relation to Europe. Measurements of the present-day drift of the Sinai plate indicate (Bechor & Wdowinski 1999) a NNW direction, with a rate of shortening of several millimeters per year in relation to Europe, independent of the motion of the African plate. Analyses of earthquakes in this region suggest a similar trend (Salamon et al. 1996). Detailed analysis of tensional features along the western flanks of the Jordan Rift Valley (Arkin 1989) hints that a similar regime was maintained throughout the Quaternary. Connected with these stress relief features is a series of east–west oriented small-scale dextral strike-slip faults (Gilat 1992, 1994), which compensated for the direction of extension along the Rift (Fig. 3.1.4). Such east–west trending small-scale strike-slip faults also occur east of the Arava, but the sense of their movement is not specified (Bender 1974b, p. 34).

To obtain an east–west extension along the Jordan Valley, it is mandatory that its eastern bordering block would drift in a symmetrical way to the western, namely in a general NNE direction. On a continental scale, such movement is apparent from the shape of the eastern sector of the convergence zone along Iran, and also from the distribution of large-scale earthquakes (Fig. 10.2.4) in this region (Degg & Doornkamp 1989, Ambraseys et al. 1994, p. 3). On the micro-scale, even the joints on the western flanks, surprisingly, indicate this sense of motion as well (Arkin 1989). Analysis of meso-structures connected with the Rift indicates (Eyal & Reches 1983, p. 167) "dominating horizontal extension trending E to ENE, in all rocks inside the Rift or proximal thereto". Sagy & Reches (2000) analyzed fault patterns, joint systems and slip partitioning along the western margins of the Dead Sea Rift, concluding (p. 34): "The above structural features are compatible with E–W extension; however, none of them fits the N–S left-lateral shear of

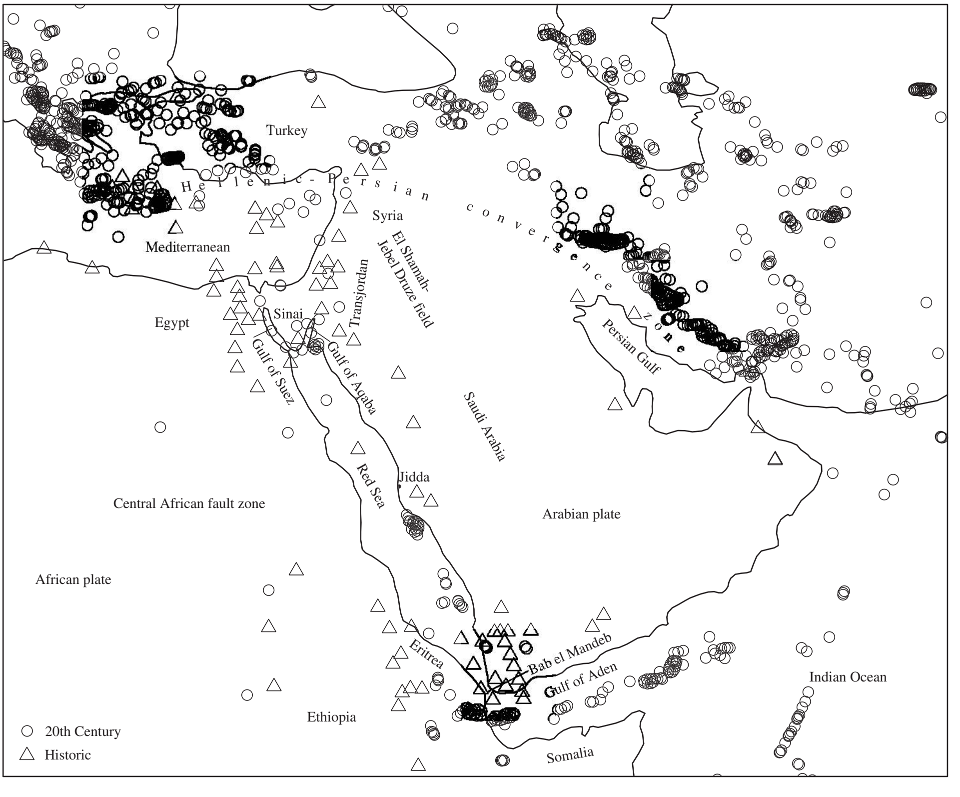


Figure 10.2.4. Earthquakes of the eastern Mediterranean, Persian Gulf and Red Sea regions, magnitudes greater than 5. Compiled from data by Degg & Doornkamp (1989), Ambraseys et al. (1994), and Salamon et al. (1996).

the Dead Sea Transform." GPS monitoring of a station located on the Golan, several kilometers east of Lake Kinneret, indicates that the eastern block moved northward 2.1 ± 1.1 mm/year with respect to Tel Aviv during 1996–1997; at the same time, a station in Elat, west of the Rift, also moved northward with respect to Tel Aviv, at a rate of 2.3 ± 0.9 mm/year (Pe'eri et al. 1999, Wdowinski et al. 2000).

The Levantine-stage motion of the Arabian–Sinai plates, which are not indeed entirely separated north of the southern Arava up to Turkey (Horowitz 2000b, 2001), comprises northward drifting at approximately the same pace, pushed this way principally by the combined opening Gulf of Aden and the mid-ocean rift encircling Africa. In addition, at the same time the two are separated by the east–west opening of the Red Sea and Gulf of Aqaba, a process which propagates further north up to southern Turkey, but the opening gradually diminishes northward. It appears that the northward drift is considerably faster than the latitudinal extension, which is expressed in the bearing and amplitude of the net amount of motion relative to Eurasia. Indeed, compared with the Levantine Rift Valley, the Hellenic–Persian convergence zone is a much more prominent and active feature of the Quaternary tectonics of the Near East.

A set of structures characteristic of the Levantine stage comprises crescentic faults (Picard 1931, p. 99; Kashai & Croker 1987; Horowitz 2000b, 2001), or rather fault zones, which occur in three opposing pairs along the sector of the Jordan Rift Valley discussed here (Fig. 10.2.5). These are, from the south northward, the pair of faults emerging from the northern tips of the Dead Sea (Bender 1974a, Begin 1975a); the Sheikh Ali fault northeast of Lake Kinneret (Michelson 1972, Mor 1986, Rotstein & Bartov 1989, Sneh et al. 1998b) and the complicated fault zone west of this lake (Saltzman 1964); and the pair of faults coming out from the northern Hula Valley (Figs 10.2.2 and 10.2.5), Rachaya heading northeast and Roum veering northwest (Dubertret 1962, Heimann 1985, 1990). In addition another pair borders the center of the Dead Sea, but these are not as clear and could be older, but certainly reactivated, since they offset some wadis leading to the Dead Sea. They comprise the fault heading northeastward from the Lisan Peninsula reaching almost to Amman (Bender 1974a, structural map) and its counterpart fault zone west of Mezada (Gilat 1992, 1994).

The crescentic faults share some common characteristics: their southern sector is always north–south oriented, bending outside the Rift approximately at the end of a deep basin; the fault planes are steeply dipping, usually between 70° and vertical; they occasionally show lateral offsets, in the order of several hundred meters, sinistral east of the Valley, dextral to the west, wherever these could be ascertained; they all affect Pliocene or younger rocks, proving their Levantine affinity; they are quite long, in the order of tens of kilometers. It seems that the inland bending sectors were all active only during the Levantine stage (except possibly the pair connected with the southern Dead Sea basin); the north–south parts, although certainly active at that time, could have been an inheritance of the previous, Eritrean stage, reactivated by the Levantine. This assumption however needs further support. The

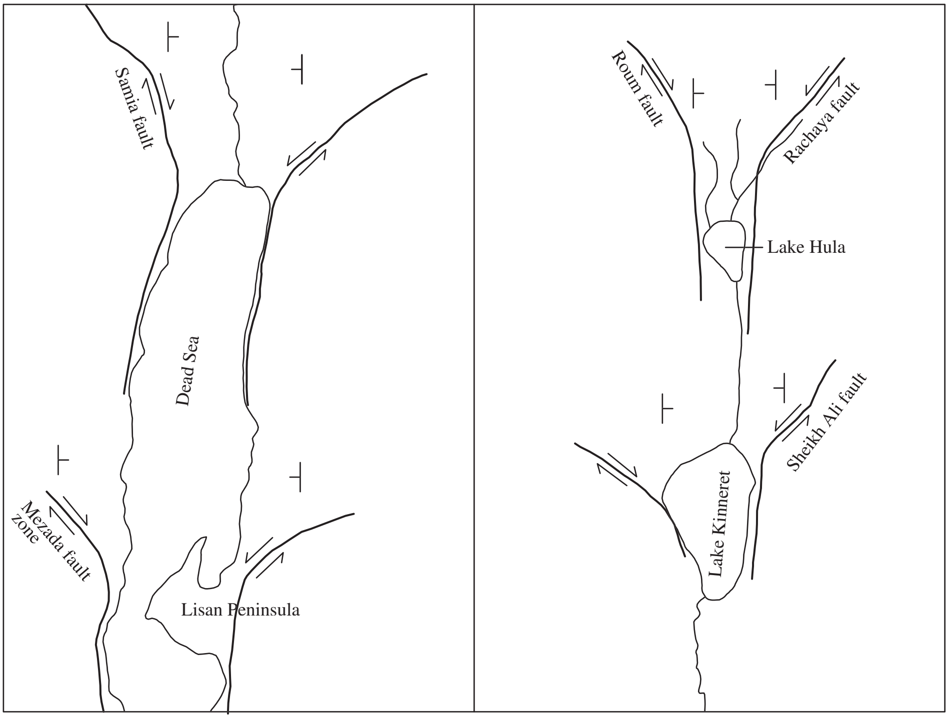


Figure 10.2.5. Crescentic faults bordering the Dead Sea (left), the central Jordan Valley and the Hula (right).

crescentic faults thus bound the four Jordan Valley Levantine basins: the southern Dead Sea, the northern Dead Sea, the central Jordan Valley and the Hula.

Since lateral displacement is never considerable, it is occasionally difficult to observe, while at times it is impossible even to securely define the sense of movement. An example is the Roum fault (or rather fault zone, since this is not a single one), branching off the northwestern end of the Hula into the Lebanon, claimed by Butler et al. (1997) to be dextral, by Heimann (1990) to be sinistral, while Ron et al. (1997) admit that it is impossible to tell, but they prefer the sinistral offset, most probably because of the long-accepted tradition. A look at the detailed map by Heimann (1985) reveals that this fault veers to the northeast several times along its roughly northerly direction within the Hula Valley. These northeast trending sectors are normal faults, which indeed precludes the sinistral possibility, leaving only the dextral option for the Roum fault. Similarly, prompted possibly by the same considerations, Gilat (1992, 1994) mapped sinistral faults in the Mezada area. However he also states that there are northwest trending sets, as well as east–west oriented dextral fault sets.

Some of these crescentic structures do not leave much doubt as to their sense of lateral slip, being dextral west of the Jordan Valley, and sinistral to the east. The

Samia fault-strip (Fig. 10.2.5) which branches off the northwestern tip of the Dead Sea (for description see Chapter 4.9.2), was studied in detail by Begin (1975a), who showed clearly that it has a slight dextral lateral movement component, with very steep fault planes. A sinistral movement that had offset Jurassic beds by some 1,000 m (Sneh et al. 1998a) was measured on the Rachaya fault (Heimann 1990), branching off northeastward from the northeastern end of the Hula Valley. The slight lateral component could also explain the steepness of most north–south trending bordering Levantine faults of the Jordan Rift. This lateral dextral component, although practically minimal, can be seen in the slickensides of the western bordering faults (Gilat 1992, 1994), with their slightly southward oriented (330°) dips. Other dextral movements of similar bearings are described in Arkin et al. (1999b) for the central Jordan Valley, and by Ron (1983) for the Galilee. Both Rotstein & Bartov (1989) and Hurwitz et al. (1999) consider the Sheikh Ali to be a sinistral fault, based on seismic reflection profiles.

The NNW movement of the western Rift shoulders and the NNE movement of the eastern, accompanied by uplift of both, initiated the formation of and movement along the crescentic faults on both sides of the Jordan Valley depression. The "wedges" thus formed within the Rift (Fig. 10.2.5) were pushed slightly northward, creating saddles to the north and basins to the south. The northward "push" of both rims against this wedge also caused an east–west oriented extension just south of the wedge, however limited, that resulted in the formation of rather deep, narrow basins, rapidly subsiding under the load of accumulating sediments. The considerable depth of these basins results mainly from the steepness of their north–south border faults. This extension, wherever it reaches a certain magnitude (possibly also maturity), is accommodated by east–west oriented dextral strike-slip faults west of the Rift (Fig. 3.1.4A), very clear in the southern Dead Sea basin (Gilat 1992, 1994), with offsets in the order of tens to a few hundred meters. It seems that a mirror view of these must have formed east of the valley, where Bender (1974b, p. 34) described east–west trending small scale strike-slip faults (Fig. 3.1.4B), but their sense of movement is not specified.

The Dead Sea reveals a repetition of this process at least twice. At the earlier stage of the Levantine activity the deepest was the southern basin, bounded by the northern Lisan and the Mezada crescentic fault zones, keeping the northern Dead Sea elevated. Subsequently, with the northward advent of activity, the northern Dead Sea crescentic faults defined a new wedge to the north, its northward push causing the formation and subsidence of the northern Dead Sea basin at the end of QIX times, together with elevating the southern Jordan Valley. A similar twin process may also have affected the central Jordan Valley, keeping the area of the present Lake Kinneret elevated for most of the Quaternary, subsequently causing the formation and subsidence of Lake Kinneret at the same time as the northern Dead Sea, while the wedge to the north keeps the Korazim block elevated during the entire Levantine stage. The Hula does not show such a repetition, the wedge to its north elevating the Metulla block.

In the Dead Sea and the Hula, it seems that the southward continuation of the wedges constitutes the substrate of at least the northern part of the adjoining basin by down-bending, being "dragged behind" the slowly moving wedge, so that no faults bound the northern limits of these basins. The picture is inverted in Lake Kinneret, where the northern boundary with the Korazim block is faulted (Horowitz 1973), while the basin's floor is slowly rising to the south, and does not show any conspicuous faulting.

The northward propagation probably dictates also the size of the basins, so that the Dead Sea is longest and widest, followed by the medium-sized central Jordan Valley and the small Hula. It should be noted that this growth came with time, since it is quite evident that the considerable thicknesses of Palynozone QII, denoting the initial subsidence attributed to the Levantine stage, are known only from a limited area in the southern Dead Sea basin. This mechanism can also explain why older, Eritrean basins ceased to subside, as they became parts of newly formed saddles, and so took on the apparent northward migration of the basins.

One of the notable characteristics of the Levantine stage is that hardly any rejuvenation of its previous Eritrean-stage lineament is seen, except possibly for slight movements on the mostly north–south bearing faults, joints and other micro-structures. However, the fact that the westward extensions of the Levantine crescentic faults share a similar bearing to the Eritrean, gave rise to earlier suggestions that the former is an inheritance of the latter. It is quite clear from all the studies cited above that, wherever this could be ascertained, Pliocene beds are only affected by the crescentic faults after having already been laid down in place. This is true for both the north–south and the bending sectors of these faults.

This testifies to the entirely different tectonic regime of the last two million years, attributed here to the change in drift direction of all the plates involved, south of the northern Mediterranean. This may also include Eurasia, if indeed its stationary position in relation to Africa results from a northward drift of the two plates at the same pace, as is occasionally suggested (e.g. Press & Siever 1998, p. 531). The almost steady state of Africa and Eurasia is deduced principally from the fact that there is very little deformation of the sediments deposited in the Pliocene basins of southern Europe (Mercier et al. 1987).

10.2.5 Uplift, rifting and magmatism

Generally speaking, regarding the Red Sea system, there seems to be no debate concerning the gross temporal connections between uplift, rifting and magmatic events, particularly volcanism, as well as their mutual relation to the large-scale regional extension (see discussions in Coleman 1993, Omar & Steckler 1995, and in numerous papers in Purser & Bosence 1998). But when looking more closely at details of timing (Chazot et al. 1998; see Section 9.7.3), it seems that most (but certainly not all) magmatic phases were considerably more active during times of "rest" in the subsidence of the rift or uplift of its shoulders. This apparent discrepancy puzzled many, who tried to resolve it by connecting magmatism with uplift of the rift shoulders, rather than subsidence. However, when the dating of uplift became more accurate, such synchronous relations were found to be equivocal in numerous cases.

For the Jordan Rift Valley and vicinity, the details of temporal correlation are disturbing. The Oligocene, for which considerable uplift is recorded, shows magmatic activity only in rather distant regions, the El Shamah–Jebel Druze field; on the other hand, the middle Miocene, when uplift was moderate, is characterized by extensive magmatism of the Lower Basalt and its correlatives, abundant both east and west of the Rift. Another period with hardly any uplift at all, the Tabianian, again shows substantial volcanism of the Intermediate Basalt. Next, the Cover Basalt predates both the considerable subsidence of the Jordan Rift Valley and the elevation of the western and eastern highlands. The period of considerable uplift of the eastern and western highlands, during the Levantine stage, is typified only by subordinate magmatic activity.

Another notable factor in the relations between rifting and magmatism along the entire Red Sea system is their spatial distribution. Most of the magmatism is outside the rift's margins, occasionally so far away that doubts were expressed whether the connection between the two is indeed genetic, or only circumstantial. Those magmatic occurrences which do have a temporal correlation with the gross timing of the Red Sea system, such as volcanic fields and dike swarms on the Arabian plate, are almost exclusively limited to only one side of the rift. This situation changes to the south, in the Yemen–Horn of Africa region, where most of the activity swings and concentrates on the other, western side of the rift (Fig. 10.2.6). Another interesting point is that both at the narrower northern and southern ends of the Red Sea system, in the El Shamah–Jebel Druze field to the north and the Ethiopian–Eritrean to the south volcanic activity is more extensive than in other regions bordering wider sectors of the rift. Similarly, the uplifted flanks are not equally distributed along the Red Sea, the differences being seen both in the extent of the areas occupied by elevated regions, or their elevations.

A tectonic model is thus required that could answer at least these major discrepancies. Evidently, this is possible only if extension, rifting, uplift and magmatism are regarded as a multifaceted outcome of a single cause, namely the regional extension, and not as separate entities.

When observing the distribution of magmatics and uplifted regions along the Red Sea system (Fig. 10.2.6) it becomes apparent that they are spaced along a several hundred kilometers wide belt, oriented roughly north to south, all the way northward from the East African rift system through Ethiopia–Yemen to southern Turkey. The Afar triple junction with its plume, as well as its northern parallel, the El Shamah–Jebel Druze field, where magmatism began during the Oligocene or possibly even before, seem to be the centers of activity, while a continuation is seen both to the south and north, where younger magmatics predominate in terms of

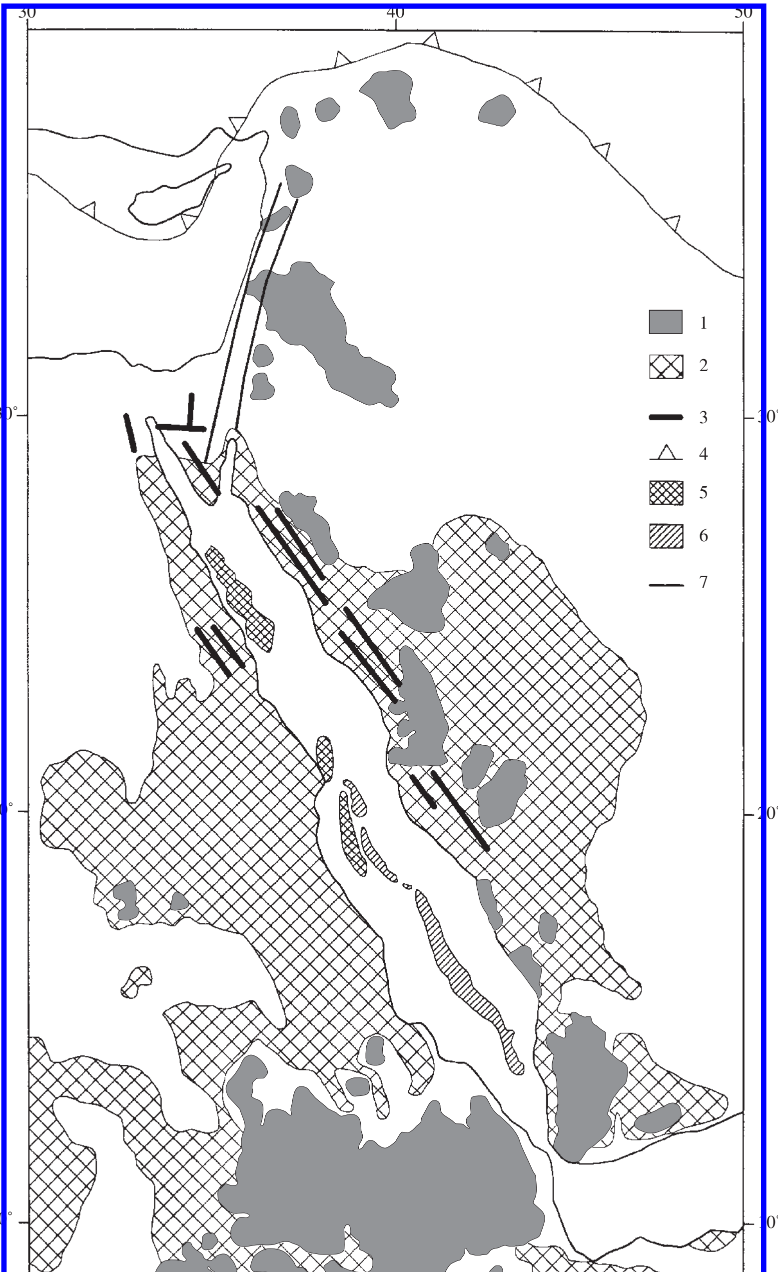


Figure 10.2.6. Extension of volcanics and uplifted areas along the Red Sea Rift: (1) volcanics, (2) uplifted regions with exposed Precambrian rocks, (3) Red Sea Dike system, (4) continental convergence zone, (5) oceanic crust of the Red Sea, exposed in pull-apart basins, (6) oceanic crust of the Red Sea, intruded into its axial zone and (7) uplift along flanks of the Jordan Valley.

volume (Chazot et al. 1998). Magmatic activity continued, however intermittently, until the present day all along this north–south bearing wide belt, where maximum uplift is apparent. Incidentally, Vroman (1981) noted the north–south bias of other major rift valleys around the world.

The sector of this belt north of Afar is cut obliquely by the Red Sea rift, at an angle of some 30°. It seems that the uplift commenced almost at the same time all along the Red Sea (Omar & Steckler 1995), including the southern Levant (see Section 7.2.1), some 34 Ma ago. Notably, the Red Sea itself, with its deviation from the meridional direction, occupies only 28% of the entire length of the Syrian–African rift valley, a fact that is commonly neglected, as more often than not the Red Sea is considered the principal structure of this system.

The question is, if the zone of tension is indeed north–south oriented, could it open an oblique rift at such an angle? Theoretically, the answer is positive, as shown in experiments by Mart & Dauteuil (2000), who observed development of oblique rifting in analog models. They have further shown that oblique rifting results in structures very similar to the Red Sea, with deeper axial basins separated by shallower threshold zones, both subsiding at different rates and located within the composite rift. Indeed, observing the Red Sea one can see that many of its internal deeper structures show an almost meridional bearing, at an angle with the general trend of the mega-structure (Fig. 1.4).

If the above suggestion is accepted, it implies that the African and Arabian plates are separated along a complex line, which is neither precisely parallel to, nor strictly overlying, the active belt of the uppermost mantle. This active belt could consist either of a line of plumes, or may even be continuous: there seems no way to tell at the present state of knowledge. Each process, of rifting, magmatism and uplift, either alone or in any combination of two or even all three in a given period, is thus regarded here as a consequence of upper-mantle activity. Since the general trend of the active belt is meridional, the overall extension is, throughout the Rift's history, east–west oriented. Other directions which occasionally developed, such as during the Eritrean faulting stage (when north–south bearing major faults are very common as well), do not alter this general trend.

The Jordan Rift Valley began its tectonic activity during the early Oligocene, as did the Red Sea. Uplift, faulting and volcanism came in intermittent phases, while subsidence along the depression was continuous, even though varying in pace through time. It is therefore suggested that while the former processes result from plate kinematics, subsidence is principally an outcome of the sediment load. Plate kinematics along the Red Sea–Jordan Valley is not uniform throughout history, and this is expressed by profound differences in the relations between uplift and rifting in periods of distinct tectonic styles. When plume activity is prevalent they go hand in hand. When it is pure extension caused by continental drift, whose driving mechanism is outside or far away from the rift, there is hardly any uplift, only rifting and volcanism. The first process characterizes the Embryonic and Levantine stages, the second is typical of the Eritrean.

Particular attention is paid here to the bearing and sense of movements on the transversal faults, Eritrean and Levantine alike. Both are principally east–west oriented, displaying lateral motions of up to a few kilometers for the Eritrean, less for the Levantine, dextral west of the rift valley, sinistral to the east. These seem to be the best and most reliable indicators for the direction of forces and sense of opening involved with the rifting. These "secondary" fault systems occur all along the Red Sea–Jordan Valley lineament, allowing for the regional latitudinal extension. They therefore fit the definition of transcurrent (rather than "transform") faults (Jackson 1997, p. 673) at this stage, but could also be the embryonic precursors of real transform faults, if and when the Syrian–African rift system further develops to become an ocean. It is quite clear from the above that the use of "transform" to describe the Jordan Rift Valley is wrong. Parts of it are true rift valleys, with branching, perpendicular transcurrent faults, very reminiscent of the relations of mid-ocean rifts and their transform faults, characteristics which prompted a few scientists to regard the Jordan Valley as a prospective incipient ocean (Mart 1991). The only opposition to this idea is set by those sectors of the Jordan Valley which are not faulted at all, being synclines.

The early Oligocene and early Miocene uplift phases of the Red Sea (Omar & Steckler 1995) are accompanied by rifting. These uplift waves also affected the southern Levant, but with no observed faulting, only accelerated subsidence. Both uplift stages are recorded as compression phases in the development of the Hellenic arc (Le Pichon & Angelier 1979), which directly connects them to continental drift processes, in addition to extended upper mantle activity along the Red Sea system at those times. The early Oligocene stage was followed, mainly in the southern sector of the Red Sea, by extensive magmatic activity which began only about three million years later, when rifting was subdued. To the north, the Oligocene volcanism of eastern Transjordan also lagged somewhat behind uplifting. Similarly, the early Miocene uplift, rifting and compression along the Hellenic arc (Mercier et al. 1987) was accompanied by magmatism in the southern Levant, again lagging behind, until it culminated in middle Miocene times in the central Jordan Valley and its surroundings.

The Eritrean-stage phases are different, characterized by rifting with hardly any uplift. Their temporal connection with extension in the Hellenic arc signifies a different tectonic regime, where only subordinate upper mantle activity is recorded for the Red Sea rims. But again, only with the subsequent period of tectonic quietness both in the Jordan Valley and Hellenic arc (Mercier et al. 1987), during the Pliocene and earliest Quaternary, did considerable magmatic activity taken place. The last uplift, in the Levantine stage, is hardly recorded in the Red Sea, but appears in the Hellenic arc as a renewal of compression. It is accompanied, to date, only by minor magmatism.

To conclude, it appears that periods of stronger upper mantle activity, along (but outside) the plates separation region, result in uplift and rifting. Times of "rest", caused by extension derived from large-scale continental drift which eases the regional compression, are characterized by extended magmatism. Some of the basins formed during the rifting stages are then filled continuously with sediments, their load causing further subsidence. The latter process is not endless, however, since other basins had reached an equilibrium after a period of subsidence, such as the late Miocene trough of the northern Arava which had not accumulated sediments since the second phase of the Eritrean faulting, which opened another basin to the north.

10.3 THE ELUSIVE STRIKE-SLIP FAULT

The "leaky transform" with its proposed lateral movement of more than 100 km is indeed an elegant solution, backed by an updated theory, which buries old ideas for the sake of novelty (if one can label "novel" a 70-year-old concept). It is disputed by hardly anyone since it seems to provide almost all necessary answers, so why am I not happy? The answer to that is quite straightforward: I have never seen the strike-slip fault on which this considerable lateral movement should have taken place, and neither has anybody else! Truly, this fault is marked in a general way in numerous papers, but when it comes to detailed geological mapping, nobody had the guts to draw the real line, since it was never located in the field. A partial excuse may be that it is covered by recent alluvium, but this does not hold in places where the valley floor is made of older rocks, such as all the in-Rift elevated structures.

An exception is the southern sheet of the 1 : 250,000 geological map of Israel (Bentor et al. 1965), where a strike-slip fault is marked along the eastern border of the Arava, an area inaccessible to the authors at that time. A detailed mapping of this area by a team that actually walked there (Bender 1974b), does not show any such fault. Bender (1974a, p. 116) makes a rather cynical remark: "A view of the existing literature on the evolution of the Wadi Arabah–Jordan Graben (Dead Sea Rift) leads to the conclusion that among the great number of authors dealing with this subject, only few have actually been there to study it." Bender (1974b, p. 34) clearly states, concerning the eastern Arava faults: "Direct evidence of strike-slip movements has not been observed along any of the above described fault systems." Neither fault is marked (Figs 3.2.1–3.2.4) on the newly published geological maps of Israel (Sneh et al. 1998a, Bartov et al. 2000a), while the area is now accessible, although the senior author do accept the idea of lateral movement, however one with reservations (Sneh 1996). Evidently, if the idea of Sneh (1996) that lateral offset predated the rifting, is accepted, it is only natural that the fault would not be visible, as it would be covered by younger sediments. Only, in this case one wonders, what then is the connection between such a fault and formation of the Jordan Valley?

The cardinal question is then, could a fault (or fault zone) on which more than 100 km of lateral movement is postulated be discontinuous or pass unnoticed?

The answer here is only a matter of faith: I cannot see it otherwise since, as mentioned above, most scholars do not even hesitate in consenting. This wide consensus disturbed me very much, particularly after realizing not only that the bordering faults are typically normal ones, with only a very slight lateral-slip component (Fig. 3.1.6), but that in fact there are areas where one can cross the Jordan Valley without stepping on any fault at all, and where the substrate is not just young alluvium, but Miocene rocks in the area south of Marma Feiyad (between coordinates 190 and 180), Campanian in the central Arava (between coordinates 960 and 950), both regions that acted as watersheds for long periods, so that no in-Rift sediments ever accumulated there to mask this simple picture. Ten Brink et al. (1999), who summarized the anatomy of the Dead Sea Transform, based on geophysics, clearly show on their map that both areas are not faulted, and that the region is characterized by relatively short and discontinuous fault segments (which does not make them have second thoughts on the "transform" idea).

My own answers are different. I am convinced, by simple geometry, that if you move two blocks laterally more than a hundred kilometers, they should be separated by a distinct, clear, and, most importantly, continuous fault (or fault zone, but still continuous). Also, no such large-scale fault in the world is unseen, so that I do not find any excuse for this one to act in such a shy manner.

If indeed, as several students claim, the strike-slip fault is entirely buried, it should have been seen clearly at least in both watersheds, and certainly by geophysicists. Chapter 8 of this book, by Ginzburg & Ben-Avraham, summarizes the last topic, so several citations are in order here. Concerning the Dead Sea–Bet She'an segment, the authors maintain: "Profile B-B (Fig. 8.3.2) shows that the fill here is not thick and that the flanks of the Valley are not faulted. Profile C-C shows a monoclinal fold dipping into the Valley on the western margin and probably just dipping layers on the east as well" (Section 8.3.3) And: "no single faulting mechanism is responsible for the formation of the Valley. The quiet zones indicate that there may not be a continuity of active faults in this segment" (Section 8.3.4). And further, describing the fault north of the Dead Sea, "from left lateral strike-slip near Jericho to mainly normal further north" (Section 8.3.5).

Regarding the Bet She'an–Lake Kinneret depression, Ginzburg & Ben-Avraham maintain: "These elongated anomaly trends indicate the location of faults with considerable normal component, which delimit the depression on either side" (Section 8.4.2). It is now quite interesting, after these statements, to quote the conclusion (Section 8.6) which, surprisingly (or maybe it is time for me to stop raising my brows), is the following: "Geophysical evidence for the existence of strike-slip faulting was found in parts of the Jordan Valley with gaps, such as the region between Jericho and Bet She'an, which for the time being has to be filled on geological consideration and conjecture. In all, despite the gaps in our data, while other solutions may be possible, the geophysical results can in our view be best explained by a left lateral transform fault separating the Arabian plate from the Sinai sub-plate."

Another example can be advanced from earthquakes, which are usually considered reliable representatives of tectonic processes in depth (Figs 3.3.1 and 8.3.3). To cite Avi Shapira, Head of the Geophysical Institute of Israel, Seismology Division (1997, no page numbers) about the Dead Sea: "The pull-apart model, assumed to characterize this area, cannot as yet be verified by the seismological observations, that is, the main faults bordering the lake are less active as compared to those within the lake and so far show a dominant normal faulting mechanism rather than the expected strike-slip." And further north: "The Jordan Valley north of Damiya up to the Sea of Galilee is practically a-seismic. This part of the DST (Dead Sea Transform) is well instrumented and it is quite unlikely that creep will go unnoticed in terms of micro-seismicity." As before, although there is no sign of strike-slip in the Dead Sea, nor movement at all to the north, the term "transform" is liberally applied.

Another example from Shamir (1997, no page numbers): "A major characteristic which seems to dominate the seismic activity along the DST during this period is the spatially non uniform distribution of earthquakes. Activity is localized at specific segments of the DST, specifically where major step-wise discontinuities in the fault structure produce broad pull-apart basins. These are the Gulf of Elat (Aqaba) and the Dead Sea, Kinneret and Hula basins. The DST segments in between these basins, namely the Arava Valley and the lower (central and southern) Jordan Valley, are characterized by a nearly complete lack of seismic activity, including microseismic, during the current (20th) century." Which does not preclude the expected conclusion: "An ad-hoc conclusion is therefore that the basin is responding to strains induced by major left-lateral past earthquakes along the DST faults."

Another problem with earthquakes concerns their focal depths in the Dead Sea area, reported by Rabinowitz & Mart (2000) from depths down to 32 km, while Aldersons et al. (2000) give a maximum figure of 28 km. The last authors (p. 2) also remind the reader that "it is usually assumed that earthquakes in continental strike-slip zones do not occur below 20 kilometers depth".

Such an approach, which is the common one in our places, is in my view only appropriate for religious purposes. No wonder that whenever colleagues are faced with my views, they are considered sacrilege, not even worth discussion. But back to science; I think it was Albert Einstein who once said: "Science is truth, don't confuse me with the facts." Only he said it jokingly.

So much for the elusive fault, so elusive that students were sent with shovels to dig the alluvium in places, in a desperate attempt to find it (Reches & Hoexter 1981, Marco et al. 2000). They surely did find some evidence for lateral offsets, but only in the order of a few meters. Sophisticated attempts to trace the fault, applying high-resolution seismic survey techniques (Shamir et al. 2000), or even a (yellow, for sure) submarine (Lazar et al. 2000), also failed.

The next problem concerns timing. When had the lateral movement (or movements) along the "transform" taken place? Although almost all believe such a movement had indeed taken place, there are hardly two geologists who would agree on the same timing. Here is a real pitfall: if you cannot see the fault, how can you date its activity? Common sense would call for lateral displacement to have taken place in tandem with subsidence along the Jordan Rift Valley, accompanied by uplift and volcanism of its rims, namely from some time during the Oligocene until the present day. I am not going to cite all examples, just a few. Freund (1965 and other publications) and Quennell (1959, 1996) considered that activity began in the late Cretaceous, continuing to the present day in two stages, some 60 km of displacement up to the Miocene and another 45 following the Pliocene; Bartov et al. (1980b) argued that the entire slip took place from the middle Miocene onward; while Sneh (1996) maintains that the 105 km lateral dislocation must have commenced not before the end of the Eocene, but at least most of it took place before the late Miocene or even earlier.

It is then quite interesting to note that Freund, Quennell and Sneh postulated that a major part of the lateral movement (if not almost all, according to Sneh) took place before the Jordan Rift had actually commenced its subsidence. This was also indicated by Butler et al. (1997), who found out that the Homs Basalt, dated at approximately 5 Ma, overlies unfaulted the Yammouneh fault (Fig. 10.2.3) in the Lebanon. This fault, according to almost all investigators, is the "natural" continuation of the Jordan Rift Valley and allegedly fully participated in its displacement. If all these claims prove true, we have the strange situation of a major regional offset which had practically no effect on the rifting, whose last Pliocene and Quaternary phases are the most conspicuous. The situation described above for the eastern Arava (see citation of Bender's) seems to be repeated here: the Lebanon is inaccessible to Israeli geologists, but Dubertret (1970) who actually (1962) mapped this area (and was the first to suggest the lateral offset in 1932), rejected his own idea after becoming familiar with details of Lebanese geology in the field. As for the Syrian part, it is enough to cite Kazmin (2000, p. 83): "Total sinistral displacement along the Syrian segment of LFZ (Levantine Fault Zone) hardly exceeds ten kilometers."

In the elusiveness of the real fault itself, its existence had been concluded merely from circumstantial evidence, which comes from three principal sources: geological and structural elements predating the rifting, allegedly offset by the fault zone; structural elements within the Rift that appear similar to those usually occurring along strike-slip faults; and the plate-kinematic model seemingly necessary for explaining the formation of the entire Red Sea system. The first and last are detailed by Garfunkel (see Appendix), while the other has been taken care of by Ginzburg & Ben-Avraham, in Chapter 8. In addition, the last years saw the advent of a variety of measurements of recent and sub-recent lateral offsets, whose results are extrapolated over time to arrive at the magic 100+ km figure.

10.3.1 Offsets of previous elements

The claim founded on the seemingly offset facies belts, rock formations or structural elements dating from the late Precambrian until the Eocene, across the

Jordan Valley, is based on a quite simple rationale, namely that equivalents of these phenomena west of the Rift occur some 100 km to the north, when crossing the valley to the east. To test the hypothesis, one must first define the primary bearings of the phenomena. Only if these are approximately east–west oriented does the fact that they are offset indeed require lateral displacement. A simple test (Dubertret 1970; Horowitz 1979, p. 58; Mart 1991) shows that most of the facies belts, in the entire period from the Precambrian to the end of the Cretaceous, are parallel to the hinge line (geosuture) of the Afro-Arabian continent west of the Rift, as are indeed almost all structures of the Syrian Arc and the earlier Mesozoic (Figs 4.3.1, 4.4.1, 4.5.2–4.5.4, 4.7.1 and 4.8.1). It is then concluded that these are meaningless, while indications of lateral movement are sought. Furthermore, the fact that these are found some 100 km to the north on the eastern side, where indeed they should be, means that no lateral-slip took place.

The alleged offset of younger elements, from the Miocene onward, has three facets: while Freund (1965 and other publication) and Quennell (1959, 1996) allocated some 40–45 km for this time span, Bartov et al. (1980b) maintained that the entire 105 km slip took place at this period. Conversely Sneh (1996) proposed that from the Miocene onward no lateral movement occurred at all, since similar rock units occur on both sides of the Rift, which is also backed by the observations of Michelson et al. (1987), Shulman & Ben-Avraham (1999) and many others.

Only a single structural element, the Negev–Sinai transversal shear zone (Fig. 10.2.1), is indeed east–west oriented. Bartov et al. (1980b) suggested correlation of the set of faults from Sinai and the Negev with a possibly parallel one from Transjordan, occurring some 100 km to the north. These faults offset the Hazeva Formation sediments, so a post-middle Miocene age of their lateral displacement along the Jordan Rift was proposed. The main problem here is that such a correlation may merely be the outcome of wishful thinking. It stands in contrast to ages suggested by others (Sneh 1996), but the more significant objection is that, from an analysis of the geological map (Sneh et al. 1998a, Figs 3.2.1–3.2.4 above), it is clear that the possibility exists to correlate the Negev faults with Transjordanian counterparts which are a direct continuation, not offset (Mart 1991). In addition Arkin et al. (1999a) have shown that some of these faults emerge outside the limits of the Arava, on both sides. It remains a matter of taste, but the alleged offset of the faults seems, in the best of cases, extremely questionable evidence for strike-slip along the Jordan Rift Valley.

Another roughly east–west oriented feature are rivers that crossed the Jordan Valley during the middle Miocene and earliest Quaternary (Figs 7.3 and 7.6). Freund et al. (1968) had shown that if the eastern side of the Rift had been moved 43 km southward, a match of Miocene channels could be made. This figure came to justify their claim for a 60-km slip before the Miocene (which again contradicts others cited above), but the fact is that they mixed up Miocene and earliest Quaternary rivers (Horowitz 1974), and at any rate these rivers can meet their counterparts also without lateral management, as shown by Mart (1991). This claim seems to me entirely meaningless.

An argument was occasionally raised concerning the crustal structure, which is different across the Jordan Rift Valley (see Chapter 8, Figs 8.1.2 and 8.2.3), being thicker on its eastern side. However this difference extends for at least some 300 km, a figure nobody (yet?) dared to suggest for lateral displacement along this lineament.

10.3.2 Structural considerations

Concerning features typically accompanying strike-slip faults, one would expect synchronous formation of pull-apart basins on the one hand and compressional structures on the other. Both the Hula and the Dead Sea (Fig. 3.1.1) superficially look like pull-apart basins, but indeed they are not, since both lack a northern limiting fault and the typical dimensions (see below). The central Jordan Valley does not even look like one, and although it is probably bounded by a northern fault, it lacks a southern one. The northern Arava basins (Fig 3.1.1B) appear like a series of block-faults (Frieslander et al. 1997, ten Brink et al. 1999). The crust is somewhat thinner under the Dead Sea (Fig. 8.2.3), but is there all along, including the entire sequence of pre-Rift formations, as well as under all other basins of the Jordan Valley (Kashai & Croker 1987, and see Chapter 8). This underlying crust is continental, with no sign of any oceanic characteristics, not even increased heat flow. Concerning heat flow, the words of Feinstein (1987, p. 135), who analyzed the maturity of organic matter in several boreholes in the vicinity of the Dead Sea, are instructive: "Cessation of coalification despite increasing burial depth indicates that the post-Miocene thermal gradient in the Ami'az block could not have exceeded 20–23°C/km. This finding is consistent with present day heat flow and other geophysical information. Tectonic models for the evolution of the Dead Sea Graben requiring a high-thermal regime are inconsistent with the present data. This raises questions as to the mechanism involved in the formation of rhomb-shaped grabens in general, on the one hand, and for the hypothesis of the "leaky" nature of the DST north of the Gulf of Elat, on the other."

In this respect, Kashai & Croker (1987, p. 57), who based their conclusions principally on subsurface data along the entire Jordan Rift Valley, summarize: "Neither are the troughs along the rift system typical, leaky, pull-apart rhomb grabens as defined by numerous workers. Heat-flow, gravimetric and magnetic measurements as well as seismic reflection and some drilling results indicate that the graben fill is not directly deposited on new basalt crust or crystalline basement, but on top of a sedimentary column comparable to the adjacent outcrops on the graben rim. Troughs of the rift are not closed on all four sides by large faults … They are very deep and very narrow, having a depth/width ratio of almost 1 : 1 at certain localities (southern Dead Sea, south of Kinneret and probably central Hula) and a width/length ratio of 1 : 6 (Dead Sea), 1 : 8 (Bet She'an–Kinneret) and 1 : 4 (Hula). These ratios do not seem to fit well into the frame of 1 : 3 ratio for pull-aparts of Aydin & Nur (1982), perhaps because these grabens are not true pull-aparts."

The length of pull-apart basins along a strike-slip fault must reflect, at least closely, the amount of lateral slip. Except for the Dead Sea, none of the other basins along the Jordan Rift, including those comprising the Gulf of Aqaba, even approaches a figure nearing 100 km. But the Dead Sea, seemingly long enough, is not a single basin but two: the southern, subsiding ever since the Oligocene, and the northern, which only became active at the end of the Pleistocene (Figs 9.4.1 and 9.6.1). If indeed one would insist on regarding the northern basin as representing a lateral slip along the Rift, the rate of such slip should be some 50 km during the last 15,000–20,000 years, a figure that seems entirely out of proportion.

It therefore remains a matter of taste (or rather of personal preference) whether to regard the Jordan Valley basins as pull-aparts, or rhomb-shaped grabens. As for synchronous formation, a glance at Figs 9.4.1 and 9.6.1 would immediately reveal that this certainly is not the case.

But the real difficulties are met with compressional structures, which are expected both within the Rift and outside its limits. The only structures within the Rift for which a claim for being a compressional feature was made are the Korazim block (Heimann & Ron 1993) and the area just south of Lake Kinneret (Rotstein et al. 1992). For the first region, Heimann & Ron delineate a series of reverse faults to account for compression. Detailed studies of this area (Fleischer 1968, Horowitz 1973, Belitzky 1987, Harash & Bar 1988) marked these faults as normal; moreover, even Heimann himself (1990) viewed them as normal, so why the unnecessary and sudden change of opinion? Another claim by Heimann & Ron (1993) is that the Korazim block was rotated counterclockwise due to sinistral movement along the Almagor fault, bounding the block to the east. However the amount of lateral offset needed for this rotation is calculated by the authors to be in the order of only 100–150 m since the Pliocene. A somewhat similar situation is in the area south of Lake Kinneret where Rotstein et al. (1992) found "anticlines" on a high-resolution seismic reflection profile (Fig. 8.4.5). These were proven by Ginzburg & Ben-Avraham (1986) to comprise buried magmatic bodies which, if at all, would denote extension. One of these magmatic bodies was penetrated by the Zemah 1 borehole (Marcus & Slager 1985), drilled at the crest of one of these "anticlines", penetrating a trough filling (Fig. 6.1.2). The other intrusion crops out as a volcanic neck near the site of Ubeidiya (Horowitz 1974), used by me for a long time to instruct students what an intrusion looks like. In contrast, Arkin (1989) reports large-scale tensional features along the entire length of the Jordan Rift Valley.

Outside the Rift, one would expect to find considerable compressional structures particularly on the eastern flank, which allegedly moved some 105 km or so northward, or on the western flank, if it indeed slipped southward. But all compressional structures on both sides of the Jordan Valley are considerably older than the Rift itself. Furthermore, during the period of rifting the eastern side (and occasionally also the western) was given to considerable volcanism, which could only denote extension. The fact that the Syrian Arc structures are older than the Rift can easily be seen by the sand contents of the Embryonic stage formations, attesting that the neighboring high structures were already stripped down to their early Cretaceous or even Jurassic formations, the only ones able to supply sand in any quantity. It is also quite clear that the elevated Hermon already considerably affected the facies belts in Eocene times (Sneh 1988).

Bender (1974b, p. 34) summarizes the structural pattern of the eastern side of the Arava: "In the area shown on the three maps, neither overturned strata nor thrust faults are exposed." Thus Quennell's (1996 and other publications) claim, that the Hermon–Anti-Lebanon range was uplifted only during the late Pliocene by underthrusting of the eastern block, has no support when timing is examined.

Another structural problem emerges from the bearing of the Yammouneh and other faults (except for the Roum fault) north of the Hula Valley (Figs 10.2.2 and 10.2.3), which should have acted as a continuation of the Jordan Rift Valley. They all veer to the northeast, so that if indeed the eastern block had moved northward, they would have displayed enormous compressional features, which they simply do not (Dubertret 1962), being connected also with Pliocene basalt effusions. The Roum fault, incidentally, which is suggested as a possible extension of the Rift by Butler et al. (1997), is reported by these authors to have a slight component, a few kilometers only, of lateral slip, but dextral, which is surely the wrong direction. Even according to Quennell (1996) the Yammouneh fault only shows a lateral slip of some 10 km (which according to Butler et al. predates 5 Ma), indicating that this sector cannot be regarded as a continuation of the "transform". Kashai & Croker (1987, p. 56) concluded that "No large scale strike-slip movements are documented along the broad and long synclinal Beqa'a valley … Normal faulting is the dominant structural style".

10.3.3 Plate kinematics

The question then remains whether a lateral-slip along the Jordan Valley is indispensable for explaining the opening of the Red Sea. In terms of time, if indeed this is necessary, the "DST" should have acted as an accommodating structure in tandem with the Red Sea opening, namely from the Oligocene until the present day. Two points seem crucial in this respect: could the Red Sea open by an east–west oriented extension? And if it opened by extension perpendicular to its axis, N60E–S60W, what indeed is its net opening, certainly considerably less than its total width?

The theoretical answer to the first question is positive, as already discussed in Section 10.2.5, and see Fig. 10.3.1. If indeed this is the case, there is certainly no need for lateral movement along the Jordan Valley. The second possibility deserves more consideration, mainly since it is believed by so many scientists.

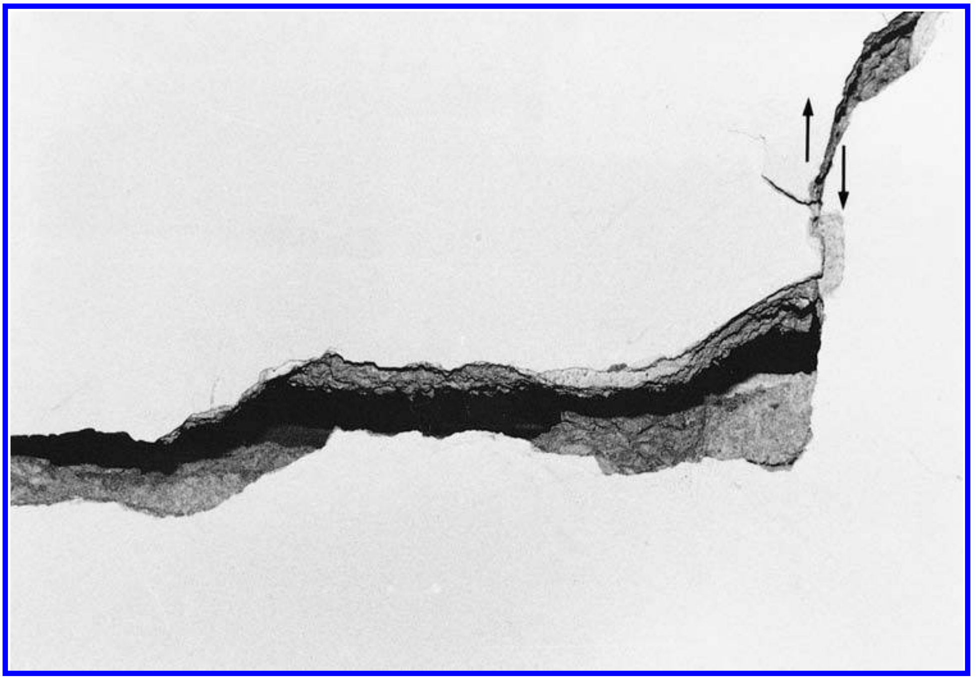


Figure 10.3.1. To demonstrate the possibility of diagonal rifting I have photographed a crack in the wall of my study. The foundations of the old house are slowly subsiding, so that the two parts of the wall are separated along a vertical axis, forming horizontal "rifting". But the horizontal sectors are occasionally offset although the underlying bricks are rectangular, forming diagonal "rifts" in the plaster. The lateral offset involved forms a "strike-slip" fault at the right end of the photo, which runs parallel to the sense of the acting force.

The maximum width of the Red Sea is some 400 km, between Eritrea and Yemen, diminishing to less than 200 at its northern tip. Most of the sea bottom is made of sediment covered Precambrian blocks, usually bounded by very steep faults, which made some people propose that the entire Red Sea opened by a strike-slip mechanism (Rihm & Henke 1998). It is quite clear that extension is maximal in the central Red Sea, diminishing both southward and northward (Coleman 1993, p. 130). Figures for the central sector vary between different authors, but are within the range of several tens up to 150 km. It seems that extension in the northern Red Sea is considerably less, something which is apparent also from the fact that hardly any oceanic crust, if at all, exists in this region (Coleman 1993, p. 124).

It is quite difficult to assess exact numbers for the net extension of the northern tip of the Red Sea. The predominantly steep faults suggest that not much is needed in order to achieve a rift in this place, so that a very liberal estimate would be 30–40 km of extension. Some of this certainly went to the opening of the Suez Rift, itself some 80 km wide; the remaining 20–30 km, if indeed the extension was directed perpendicular to the bearing of the Red Sea, could have created a sinistral slip directed toward the Gulf of Aqaba. A simple calculation shows that such lateral displacement cannot exceed half the amount of extension at a 30° angle, namely 10–15 km. This amount of lateral slip could easily have been accommodated by a diversity of minor strike-slip structures known in eastern Sinai (Bartov & Steinitz 1978).

The figure given in Joffe & Garfunkel (1987) for the opening of the northern Red Sea is considerably higher than suggested above, some 135–140 km. Of this total, 25–35 are taken by the Suez Rift, but from here on the calculation becomes strange, since the 30° angle between the Red Sea and the Gulf of Aqaba is entirely ignored, and the remaining 100 km are fully diverted northward. Since these figures are widely cited, it would be instructive to describe this circular proof for the lateral sinistral offset along the Jordan Valley in Joffe & Garfunkel's words (1987, p. 17): "The main uncertainty in the total motion model arises from insufficient knowledge of the overall opening across the various rifts. The only accurate measurement available is the 105 km motion between the Arabian and Sinai plates along the Dead Sea transform. Together with the 25–35 km opening in the southern Suez rift this gives the motion in the northern Red Sea. These determinations are critical, since they will be used as a basis for estimation of the movements on the other plate boundaries." This seems an evident case of the Red and the Dead (seas) washing each other's hands ("head" would be more appropriate for the rhyme).

10.3.4 Measurements of recent and sub-recent lateral offsets

Such measurements concentrate around two main techniques: GPS stations that gauge the actual present day movement of a certain spot in relation to another (Bechor & Wdowinski 1999, Pe'eri et al. 1999, Wdowinski et al. 2000); and measuring the actual slip of a determined datable phenomenon, usually either geological or archaeological (Marco et al. 2000). Earlier attempts, when dating was far from certain or otherwise controversial, are not discussed here. In spite of numerous reservations, for the sake of discussion both measurements of slip and dates are here referred to as entirely reliable. The average numbers acquired for recent and sub-recent slips are in the order of one to a few millimeters per year.

The problem lies with extrapolation of these figures, in two principal aspects: when did the measured slip commence, and was it indeed linear through time? The measured movements are part of the Levantine stage, which is tectonically different and considerably more active than its preceding Eritrean and Embryonic stages (compare Figs 9.4.1 and 9.6.1), so that it seems highly debatable whether any figure can safely be regarded as significant for processes older than two million years. During this last period the subsidence along the Jordan Valley is far from linear (Fig. 9.6.1), with higher rates observed for the latest periods. Truly, it is not mandatory that rates of subsidence and lateral offset should go hand in hand, but it only makes sense to assume that such would roughly be the situation. One may thus measure today rates of slip considerably larger than was typical for the entire Levantine stage period. These two reservations are more than enough to nullify any extrapolation, before even remembering that there is no general agreement as to the age of the "transform".

Just a single example with numbers: both GPS and archaeologically based measurements (Pe'eri et al. 1999, Marco et al. 2000) yielded roughly similar figures for the sinistral movement along the Jordan fault between Lake Kinneret and the Hula, of some 2–2.5 mm/year. If one extrapolates this figure on the entire 2 Ma span of the Levantine stage, the total slip should have amounted to a maximum of 4–5 km. But the continuation of the Jordan fault is the Rachaya crescentic fault (Figs 10.2.2 and 10.2.5), where a maximum lateral offset of 1 km was measured on Jurassic rocks (Heimann 1990). It seems therefore that such measurements, important as they are for understanding recent crustal movements, could not safely be stretched back over any considerable period of time.

Furthermore, if this rate of slip is applied to account for the more than 100 km proposed offset of the "transform", some 40–50 million years are necessary (a figure which was indeed acknowledged by Freund and Quennell). When the notably slower rates of subsidence of the Eritrean and Embryonic stages are taken into account, this figure must be considerably increased, which is entirely incompatible with the age of the Jordan Rift Valley. The idea of lateral movement, or part of it, predating rifting does not make sense when an explanation of the rifting processes is required. Again, surprisingly(?), the authors who published the results of the measurements claim that they are in accord with the suggested transform model.

10.3.5 Conclusion

As is the case with circumstantial evidence, one has to weigh both sides and decide how significant are the claims, and whether any specific one could entirely nullify the other approach.

We are thus faced with several obscure, rather disturbing points concerning the alleged sinistral offset of 100+ km. Is it possible that a fault of such magnitude is impossible or very difficult to observe and, anyway, where is it located? Then there is its clear discontinuity; its problematic time of activity and the effect of this activity on the actual rifting, if at all. On top of that, there are at least two sectors of the Jordan Rift where one can cross the valley without stepping on any fault, not even buried. No adequate compressional features exist where they would be expected, regionally on the entire eastern block, while a private case concerns the direction of the Yammouneh fault, requiring extreme compression which is not there, since the fault is normal. It is quite instructive to note that other strike-slip faults in the region, although with a lateral-slip of only a few kilometers or even less, such as the Negev–Sinai shear belt (Bartov 1974), do not display the above difficulties and are also clearly visible.

The facies belts and structural elements allegedly offset by the strike-slip simply do not require its "help" to be found where they are; the in-Rift structures are, in the best of cases, inconclusive; opening of the Red Sea could easily (and better, in my opinion) be explained without extensive lateral offset along the Jordan Valley; the structure of the crust across and within the Rift does not necessitate any intervention of slip. Besides all others, the single fact that the Jordan Valley can be crossed without stepping on any fault is sufficient to invalidate such an extensive lateral offset model. The above does not altogether preclude strike-slip faulting along the Jordan Valley, as discussed in Sections 10.2.4 and 10.5; it is the magnitude of slip that makes the difference. Wherever offset can be measured it never exceeds several hundred meters, up to a maximum of 1 km, known from both the eastern and western bordering and crescentic faults.

10.4 TECTONICS OF THE JORDAN RIFT VALLEY

The structural evolution of the Jordan Rift Valley has gone through three distinct phases, the Embryonic, Eritrean and Levantine, each with its own characteristics indicating different tectonic domains. It is thus impossible, in my view, to explain its complex history by a single process, a target aimed at by almost all investigators. Similarly, it is impossible to delineate the tectonics of the depression without regard to its accompanying structural elements, particularly uplift, volcanism, crescentic and branching faults, most of which are located outside the limits of the Rift itself. The significance of the crescentic and transversal faults, as well as the uplift phases, for understanding the movements of the Rift's margins was underestimated for a long time, deserving prompt attention. The Jordan Rift Valley is treated here as an integral part of the widely distributed regional processes, resulting from the combined activity of a series of north–south oriented plumes (or, less likely, a continuous line of upwelling upper mantle), and the kinematic relations of the African–Arabian and Eurasian continents, from the Oligocene onward.

The activity of the plumes commenced simultaneously, or almost, in early middle Oligocene times, giving rise to volcanism and uplift along the Syrian–African Rift system, at least from Syria down to Ethiopia, during the Embryonic stage of the Jordan Valley, which persisted until the beginning of the late Miocene. The uplift caused the initial opening of the Red Sea up to and including the gulfs of Aqaba and Suez. The northern continuation of the Jordan Valley (Fig. 10.4.1) suffered only rather mild synclinal subsidence at that time, being squeezed between its slowly uplifting eastern and western flanks. It is hard to say whether the synclines were formed due to compression which developed at that time between the two uplifting regions, or they are sag basins, overlying normally faulted basement. It seems however that the first possibility should be preferred, since geophysical studies do not indicate such faults in depth (see Chapter 8), but do indicate compression along the valley (Fig. 8.3.3, and see also ten Brink et al. 1999).

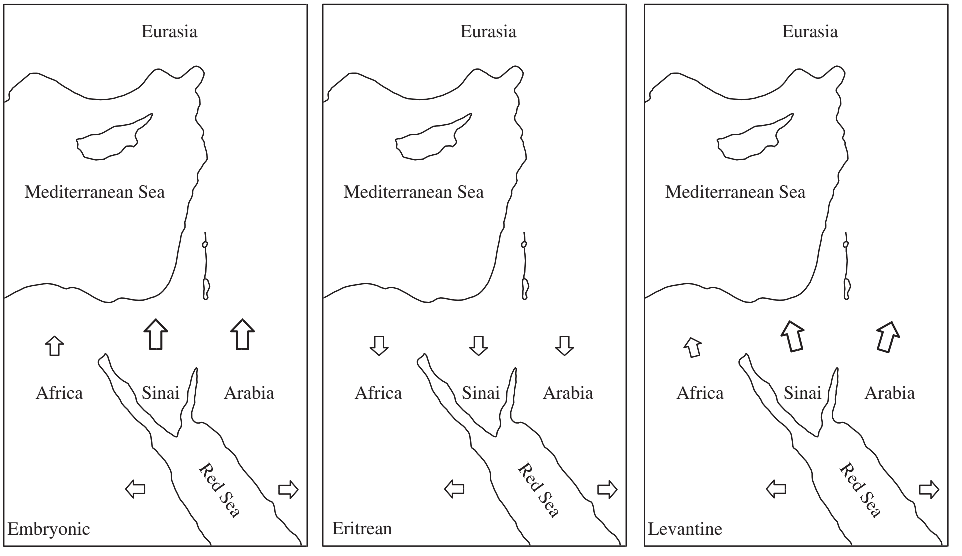


Figure 10.4.1. Principal plate movements during the Embryonic, Eritrean and Levantine stages, in relation to theoretically stationary Eurasia.

It is interesting to note that at this stage faulting affected only the rigid Arabo-Nubian Massif, along its entire length (Fig. 10.2.6). The particular behavior of the Jordan Valley may have resulted from locally developed slight compression, or from the differences in crustal structure on both flanks, which indeed controlled geological processes in the region since the late Precambrian (Figs 4.3.1, 4.4.1, 4.5.2–4.5.4, 4.7.1 and 4.8.1). At any rate, there is practically no extension involved in the Embryonic stage, nor any significant opening of the depression, nor considerable displacements of its shoulders, in any possible direction. Indeed, one is under the impression that this could be a phase of local compression, resulting in the formation of synclinal basins and anticlinal watersheds both within and outside the Jordan Valley. The locations of these structures along the forming depression were most probably controlled by the hinge line (Fig. 9.2.1).

A northward propagation of the uplift of the Sinai sub-plate caused a change in the paleogeography of the region west of the Jordan Valley, from the Oligocene to the early Miocene (Fig. 10.4.2). Although this phase of uplift is also evident to the east, in southern Transjordan, it inflicted no major changes in the drainage pattern of this region at that time. This propagation may have resulted from further uplift, in tandem with the considerable deepening of the Gulf of Suez and the northern Red Sea at that time, causing the accumulation of very thick sequences of early Miocene sediments in these basins. Unfortunately no data are available for the Gulf of Aqaba at that time, but it would be plausible to assume it behaved in a similar way. This uplift was also accompanied by some magmatic intrusion and volcanism in Sinai, practically the only occurrence of such Rift connected activity in this region.

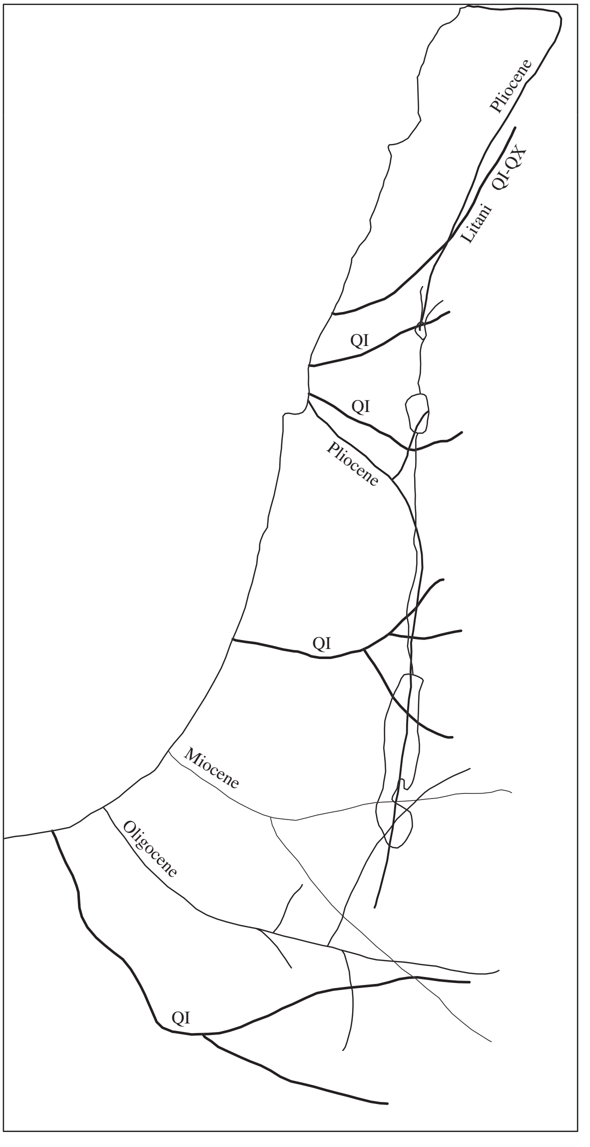


Figure 10.4.2. Locations of drainage systems connecting the Jordan Valley to the Mediterranean, as a function of uplift of the western rims, from the Oligocene onward. The northward advent of uplift affected the Oligocene, Miocene and Pliocene connections, following the same trend. The extension and quiet period following the Eritrean stage caused a return of seaward connections to the south, while the resumed Levantine activity left only the Litani and other northern rivers connecting the Beqa'a to the sea, while the Jordan Valley itself became endoreic.

The Eritrean stage displays an entirely different tectonic regime, dominated by the formation of extension structures (Fig. 10.4.1), with only subordinate uplift. This style is not particular to the Red Sea nor the Jordan Valley, characterizing the entire area affected by the collision of the Afro-Arabian and Eurasian plates, centered in the Mediterranean, including the plates' borders. Late Miocene–Pliocene extension structures are known from southern Eurasia, the Mediterranean and Afro-Arabia, along the entire zone of their contacts. Consequently this wide-scale extension overrides and dominates processes particular to the Red Sea and Jordan Valley systems. The two phases of extended activity of the Eritrean stage, during early late Miocene and earliest Pliocene times, are very conspicuous along and outside both the above systems. The first phase was most probably heralded somewhat earlier, by the extensive middle Miocene volcanism from Ethiopia throughout the Arabian plate to northern Syria, which however was not accompanied by any considerable uplift.

This is the time of maximum opening of both the Red Sea and the Jordan Rift, ensuing from extension on a continental scale, while the Afro-Arabian and Eurasian plates practically moved away from each another. It is thus quite instructive to observe the effects on the Red Sea. Two clusters of dates are available for oceanic crust formation in this region (Fig. 10.2.6), one around 10 Ma, the other at approximately 5 Ma ago. Notably, hardly any dates are available in between these clusters, nor from the younger to the present day. Also, the lack of linear magnetic anomalies in both oceanic crust phases is notable. Rihm & Henke (1998) explain the latter phenomenon by the slow spreading, not likely to produce linear magnetic anomalies, which however does not explain the clustering of dates. Conversely, it may well be that oceanic crust was only injected into the axial zone of the Red Sea during the periods of increased extension, which coincide too well with the clusters of dates, explaining also the lack of magnetic anomalies. Thus the easing of compression resulted in the formation of oceanic crust during both of these short, distinct, non-continuous phases of the Eritrean stage.

The typical tectonic style of the Red Sea system, of uplift on both sides of the simultaneously subsiding rift, did not cease during the Eritrean stage, but its effects are considerably masked by the continental scale extension. The uplift of the Sinai sub-plate continued propagating to the north, blocking the Dead Sea from draining directly westward to the Mediterranean (Fig. 10.4.2). In the same manner, the northward advancing uplift of the Transjordanian highlands had cut the connection of the central and northern Jordan Valley to the Persian Gulf domain, and also deepened the rivers leading to the Jordan Valley. These uplifts were accompanied by faulting along the Jordan Valley, predominantly north–south oriented, which, together with the faults caused by the continental extension inside and outside the depression, resulted in the intricate fault pattern observed all over the Levant. The net tension causing the north–south opening of the Jordan Valley (and further north) was east–west directed, as is clear from the bearings and senses of movement of the transversal fault system active at that time, extending from the Eastern Desert of Egypt to Sinai and Transjordan, and known all the way to northern Lebanon (Figs 10.2.1 and 10.2.2).

The previous synclinal style typical of the Embryonic Jordan Valley also wandered northward, and now forms the Beqa'a. In contrast with the Jordan Valley, the Beqa'a was affected by roughly north–south oriented faulting as well at this time. It is indeed tempting to think that the synclinal subsidence predated the faulting in the Beqa'a, but this is hard to prove, although the faults do affect the Zahlé Beds deposited in the syncline. The northward propagation of the main activity along the Red Sea system actually left behind the Gulf of Suez, which hardly ever subsided since the end of the Pliocene, and represents now a slowly dying branch of this system. It is however plausible to assume that northward propagation occurred also in the Suez branch, inflicting structural disturbances in the eastern Mediterranean. This is however far from the present concerns.

The quiet period following the second Eritrean phase saw large-scale effusions of volcanics in the central Jordan Valley and regions to the west, north and east, in much the same style as during the preceding quiet period, the middle Miocene. This volcanic phase could again be regarded as the precursor of the soon to arrive Levantine faulting stage. But, as with the previous volcanic phase, such an assumption may prove to only be wishful thinking.

The Levantine stage is the time of resumed compression between the Afro-Arabian and Eurasian plates (Fig. 10.4.1), accompanied by the continuing uplift-subsidence activity of the Red Sea system, resulting in the formation of a truly impressive depression in the Jordan Valley, considerably accentuated by the uplift on both flanks. But this last fact should not deceive: the actual volume involved in its formation is indeed not very large. The extensive subsidence affected only limited basins, not the entire Jordan Valley; while the uplift of both its shoulders touched only rather narrow belts. In addition, the opening needed to form the Jordan Valley is indeed very limited, since its bounding faults are close to vertical, so considerable depths in the basins can be achieved without much extension. The fact that there are still intact synclines along the Jordan Valley also indicates that overall east–west extension is minimal.

Two principal forces acted during the last two million years to produce the Jordan Rift Valley in its present-day configuration, the northward relative movement of the Afro-Arabian continent and the north–south oriented upwelling of the upper mantle. The uplift kept on propagating northward along the Jordan Valley, as it did during the entire history of this part of the Rift. The western branch had cut the last, Pliocene Mediterranean connection in Israel through the Yizre'el Valley (Fig. 10.4.2), redirecting the drainage toward the newly formed Jordan Valley basins. The amount of uplift along this branch diminishes to the north and, although affecting the Lebanon too, was not sufficient to cut the connections of the Beqa'a to the sea. To the east, this phase of uplift caused considerable deepening of rivers which already led to the Rift, while further east a series of internal basins was formed along the eastern flanks of the uplifting zone.

The northward movements of both flanks of the Rift created pairs of crescentic faults, which delineated wedges to the north. These, while being pushed northward by the flanks, inflicted the synchronous development of saddles to the north and basins to the south within the Jordan Valley. In addition to the uplift, the tangential pressure along the elevated saddles further enhanced their synclinal nature. Notably, the eastern half of this process was described in Kashai & Croker (1987), who unfortunately did not describe the mirror view to the west. Although these authors recognized the crescentic as partial strike-slip faults, they carefully refrained from commitments regarding the quantity of slips.

It is quite clear from the above that the Jordan Rift Valley is a small-scale copy of the Red Sea. Except for the injection of axial dikes, it bears all the typical characteristics of a mid-ocean ridge, namely a narrow, elongated depression bounded by uplifting belts affected by volcanism, its opening accommodated by perpendicular strike-slip faults. In this respect, it seems fully justified to regard the Jordan Rift Valley as an embryonic ocean. If its development continues along the same lines, it may even reach the stage of dike injection and true oceanization, a stage that took the Red Sea quite a while to achieve. However one should not forget the fate of its neglected brother structure, the Gulf of Suez, a doom that could hit the Jordan Rift as well, if northward propagation continues and leaves it behind (or altogether terminated for some reason). Another point that needs consideration is whether the Jordan Valley marks a plate boundary. If faulting is not complete, it may be premature to regard its flanks as different plates. But again, this remains a matter of opinion.

Some last words about the Red Sea. Although very impressive, the actual activity of the Red Sea spreading system is in fact subordinate, while its major extension phases were caused by the larger-scale continental processes originating from the Afro-Arabian–Eurasian relationships, particularly during late Miocene–Pliocene times. This subdued activity is the main reason why the Red Sea never developed into a real ocean, although its life extends over almost half the period it took the Atlantic to form. It is indeed an embryonic ocean, but if it goes on as it has it will never grow up to be a genuine one. It is true that uplift and volcanism began simultaneously along the entire Red Sea rift, but there is a distinct northward propagation of accentuation of these over time, at least as seen in the southern Levant, accompanied by similar trends in faulting.

CHAPTER 11

Stratigraphy of the artifact-bearing sequences

The earliest clear indications of hominids in the Jordan Valley come from the Erk el Ahmar Formation (Horowitz 1979, p. 296; Braun et al. 1991), exposed south of Lake Kinneret (Fig. 11.2.1), which produced (Fig. 12.1.2.1) chopping tools and flakes (a detailed study of this formation is being carried out by E. Tchernov, Department of Evolution, Systematics and Ecology, the Hebrew University of Jerusalem 2000, pers. comm.). Huckriede (1966) reported "pre-Abbevillian pebble tools of Olduvan and Münzenberger type" artifacts from Abu Habil (Fig. 11.2.3), but subsequent surveys failed to locate any further such implements (Macumber & Edwards 1997). Also, the age of the Abu Habil beds is not yet safely established.

Outside the Jordan Valley, a possible occurrence of such artifacts, not yet proven beyond doubt, is from the correlative Zehiha Formation in the southern Negev (Fig. 11.5.2), deposited in a drainage system leading to the Dead Sea (Ginat 1997). Another chopping tool was found in a correlative layer, the Hasi Member of the Gaza Formation (Lamdan et al. 1977), on the southern coastal plain of Israel. These layers, as well as the Erk el Ahmar, are all correlated with Palynozone QIII (Horowitz 1996b).

Questionable findings of artifacts, still very doubtful, in formations predating the Erk el Ahmar and its correlatives, are reported from two locations, both quite intimately connected with the Jordan Rift Valley, in sediments of Palynozone QI age. The Bethlehem Conglomerate (Figs 5.2.18 and 12.1), exposed within the city limits (Gardner & Bate 1937), yielded a great number of fractured flints thought by Stekelis (1940) to be artifacts, while others (Clark 1961) consider them to be eoliths, naturally broken stones. Stekelis (1992) hesitantly accepted that the flints may have been naturally broken, but he points out that the wealth of fractured long bones necessarily indicates human activity, to extract the marrow.

The other occurrence is from the eastern Upper Galilee, where Ronen et al. (1980) reported a suite of artifacts found (Ronen 1996, p. 99) "apparently below a basalt layer near Kibbutz Yir'on". The basalt, defined and dated by Mor et al. (1987) as Dalton Basalt (Fig. 5.3.11), yielded ages averaging 2.43 Ma for flows close to the find. Here the flints indeed look like man-made implements (Fig. 12.2), but serious doubt still exists whether the basalt indeed covers the gravels where these artifacts were found. If indeed these gravels are found to underlie the basalt, they constitute a part of the Amud Conglomerate.

Since the two last-mentioned occurrences are still open to serious question, the present chapter deals with the stratigraphy and chronology of the rock units starting from Palynozone QII, which incidentally also characterizes the time when the Jordan Valley became an endoreic drainage system, with the onset of the Levantine faulting and uplift stage. Evidently, not each and every one of the formations described below contains artifacts or sites. It seems however appropriate to discuss the entire continuous sequence, even though this is encountered only from the subsurface of the Hula and Dead Sea basins. In this way, the evolution of the natural environments and habitats is clearer, within the framework provided which I hope can be applied to any future finds.

11.1 EROSION, DEPOSITION AND THE LOCATIONS OF SITES

Occurrences of prehistoric sites in the Jordan Rift Valley and neighboring regions depend entirely on geological and geomorphological processes, a combined outcome of tectonics and climate. It is quite clear from descriptions of the rock units, below, that the drier periods are hardly ever represented by exposed sediments (Besançon & Sanlaville 1984, Besançon & Hours 1985, Horowitz 1996b, Parenti et al. 1997, Macumber & Edwards 1997). This results from a combination of erosion, characterizing these periods in the highlands, streams and wadis, and from shrinkage of the Jordan Valley lakes during dry climates. The subsequent wetter phases cause expansion of the lakes, thus burying the former layers deposited by the restricted water bodies. Archaeological sites which, thanks to their young age, were affected by erosion for a considerably shorter duration, and had no chance to be covered by expanding lakes, are very abundant in the later part of QIX and QX sediments.

The only locality in which interpluvial sediments were systematically accumulated is the wide Israeli coastal plain, where sand dunes represent the dry periods (Horowitz 1979, p. 109). Sites are rare in such deposits laid down in a hostile environment, except where marshes occupied some of the dune area such as Evron (see below). Such sites are more common to the north, in the Lebanese and Syrian coastal plains, which always enjoyed a more humid climate than the south (Horowitz 1979, p. 297) and had hardly any sand to form dunes. An exception in the Jordan Valley is Gesher Benot Ya'aqov, a site located in sediments of Palynozone QVI, accumulated and later exposed due to unique local conditions (Belitzky, in prep.). Open-air Paleolithic sites are thus known from pluvial sediments both in the Jordan Valley and the coastal plain; while interpluvial ones have only been found in connection with the latter environment, or in caves (see below).

The Acheulian subdivisions are quite long, due to the rather slow development of tool-making techniques at that time, so that both coastal and inland industries could be (and were) referred to by similar cultural terms. The slight differences between the artifact assemblages led to their classification under "coastal" and "inland" facies (Bar-Yosef 1994), instead of treating the two "facies" as possibly being produced in two different periods.

The only exception, where habitation is concerned, seems to be in caves and rockshelters, where sediments of various types accumulated since their openings were exposed. Even in caves, the sequences are not always continuous, and are occasionally affected by erosion (Weinstein-Evron 1983, Tsatskin et al. 1995). The exposure of cave openings largely depends on rates of uplift and incision, while preservation of the caves depends on the time and tempo of erosion. Thus, the initial Quaternary uplift, during QII times (Fig. 9.6.2), may have exposed caves which could eventually serve for human habitation, but none of these persisted due to time and erosion. Relics of such caves, one of which (Fig. 11.7.1) yielded bones of extinct animals such as *Ursus speleus* (Tchernov, in Horowitz 1979, p. 163), are known from numerous localities in the higher elevations of the western highlands (unfortunately no information is available for the eastern), but no artifacts have ever been reported from them. This may be an outcome of the sad fact that nobody, to the best of my knowledge, has ever seriously surveyed these relics for artifacts.

The second accelerated uplift phase of the highlands commenced some time during QVI–QVII times, exposing several caves facing wadis leading to the Jordan Valley, of which the best known are Umm Qatafa, Zuttiyeh and other caves of Nahal Amud (Figs 11.7.2 and 12.1.6.3), as well as Tabun Cave on Mount Carmel facing the Mediterranean. Consequently, artifact-bearing sediments of the dry Palynozone QVIII are known from these localities (Horowitz 1996b).

This situation explains why the great majority of prehistoric sites are found only in formations deposited during the humid phases, represented by Palynozones QIII, QV, QVII and QIX, both within the Jordan Valley and in the surrounding higher terrains. This leaves us with a crucial question: was habitation of the region continuous ever since at least QIII times (and possibly QI)? But we cannot locate sites since the layers are not there, or as suggested by many the area was populated only during the humid phases, when climate was favorable, with ample food and water. The last idea could possibly be negated by the fact that habitation of cave sites took place even during the dry QVIII times and the later part of QIX and throughout QX, when even open-air sites were preserved. On top of that, when geological conditions were favorable, an interpluvial site such as Gesher Benot Ya'aqov was found in the Jordan Valley. The answer, needless to say, is elusive, since these occurrences could be the exception to the rule. Thus the two options are advanced here as possibilities.

The settlement situation is different north of the Jordan Valley, where interpluvial open sites of Palynozones QIV, QVI and QVIII are known from Syria and the Lebanon. A single QVI open site, Evron, is located in the northwestern Galilee (Horowitz 1996b). Considering the entire southern Levant as a single geographical entity, humans have inhabited the region continuously ever since QIII times, and possibly even before.

11.2 LITHOSTRATIGRAPHY OF THE LEVANTINE STAGE FORMATIONS

In contrast with the previously described Embryonic and Eritrean formations, which are actually better known from regions bordering the Jordan Rift Valley, the Levantine stage rocks have been deposited only within the limits of the depression, following the formation of the Rift as an endoreic system some time during the early Quaternary. Distinct basins were formed along the Rift, in which extensive subsidence caused accumulations of kilometers of lacustrine and fluviatile sediments (Horowitz 1987c, 1989b, 1996a, Horowitz & Horowitz 1985, 1990). Within the area discussed here, three such basins are prominent: the Hula to the north, the central Jordan Valley and the southern Dead Sea. The first two, in spite of the considerable subsidence, suffered also some differential uplifts which resulted in partial exposures of lateral facies of the entire sequences (Horowitz 1979, pp. 129–151); the Dead Sea basin, being the terminal base level of this system, shows extended subsidence and very little uplifting compared to the others, thus its in-Rift formations during the Levantine stage are known almost solely from the subsurface. The three basins were penetrated by deep boreholes (Fig. 6.1.1) which were studied palynologically, which helped assign ages to the various rock units. The northern ones were also dated radigenically. The procedure and results are discussed in detail in Chapter 6.

As a general rule, the wetter periods of the Quaternary caused spreading of lakes over considerable areas of the Jordan Rift Valley, while drier times are characterized by their shrinkage (Horowitz 1992a, p. 366). Consequently most formations deposited during the latter periods are known only from the subsurface, while wetter times are represented also in outcrops, depending also on the structural pattern. The Levantine stage formations were grouped into three suites: the Hula Group, the Jordan Group of the central Jordan Valley and the Upper Dead Sea Group (Horowitz 1979, p. 173). The last term is used for rock units deposited in the Dead Sea region since it became a terminal basin; the term “Dead Sea Group” (Zak 1967) also includes the underlying Sedom and Amora formations, laid down when the area was still connected to the Mediterranean Sea.

Deposits intimately connected with the Jordan Rift Valley and its drainage system, such as those accumulated in the wadis, are also discussed here. Following the extensive subsidence, the Rift’s flanks were almost continuously subject to erosion, accentuated by the synchronous uplift of the highlands, the Transjordanian Plateau to the east and the central Israeli–Palestinian hilly backbone west of the

Jordan Rift Valley. Wadis and rivers leading to the depression from both sides were shaped by a combination of tectonics and changing climates, which is now shown by series of terraces. Occasionally, due to partial structural disturbances, wadis leading to the Dead Sea basin became lakes, usually of short duration (Bentor 1946, Ginat 1997).

Abundant springs, mainly formed by the exposure of aquifers due to faulting, always lined the Jordan Valley, depositing travertines which are quite prominent in places (Horowitz 1979, p. 161; Heimann 1985; Kronfeld et al. 1988; Enmar et al. 1998a,b; Enmar 1999). The eastern flanks of the Rift enjoy more rain than the western, which are in the rain shadow zone. This results in a greater number of perennial rivers and springs, depositing larger volumes of gravels and travertines in comparison with the western side of the Jordan Valley, and this was the case most probably also during the entire Quaternary (Macumber & Edwards 1997).

Finally, some in-Rift volcanism is also known, and is discussed together with other magmatic occurrences (see Section 5.4). The stratigraphy of the in-Rift volcanics is based mainly on the numerous pollen assemblages of the sedimentary beds below and above, discussed in detail in Chapter 6, but is greatly helped by radiometric datings. The distribution of most of the rock units described appears on the geological maps, Figs 3.2.1–3.2.4.

11.2.1 Hula Group

The term was applied by Horowitz (1973) and Horowitz & Horowitz (1985) to a suite of formations confined to the Hula basin, the marginal facies of most of which crop out at the southern end of the valley, in the vicinity of Benot Ya'aqov bridge, due to locally uplifted small blocks (Belitzky 1987). However, the bulk of the Hula Group is buried within the basin itself, and is known from several boreholes such as Notera 3, Emeq Hula 1 and others. Picard (1952) referred to the upper part of the Group, penetrated by boreholes down to 120 m below surface, as the "Hula Series". The Group includes only those formations laid down in the basin since the Dead Sea became an endoreic system in QII times, approximately two million years ago (Horowitz & Horowitz 1985, 1990, Kafri & Heimann 1994).

Most of the Hula Group comprises lacustrine sediments, grading laterally on the valley's rims into fluvio-lacustrine, fluviatile, spring deposits and soils. Under certain conditions, mostly climatic, the Hula Lake shrank to give place to marshes, in which highly organic sediments and often peat were accumulated, especially in the upper part of the sequence. Subordinate volcanics occasionally intercalate within the sedimentary column.

It seems (Horowitz 1973) that most of the sediments were accumulated in rather shallow lakes, the rates of subsidence of the substrate approximately coinciding with the rates of deposition. The Hula basin always supported intermediate lakes, except for some short periods in which the sediments are barren of autochtonous fossils (such as the Gadot and the lower part of the Ashmura formations), possibly indicating temporarily increased salinity. Chemical and isotopic analyses indicate that for most of the Quaternary the Hula was a fresh-water lake (Rosenthal et al. 1989).

Pioneering studies of the Hula Group, both from outcrops and boreholes, were carried out by Picard (1952, 1963, 1965) followed by detailed investigations of the upper part of the sequence, for economic reasons connected with the abundant peat and natural gas (Bein 1967 and others, usually unpublished). Detailed palynostratigraphic studies throughout the last 25 years (see Chapter 6), followed by radiogenic datings (Heimann 1990), enabled clarification of the stratigraphy and chronology of the basin's fill. The various formations of the Hula Group are named after localities in the Hula Valley. The entire Hula Group sequence was penetrated in the Notera 3 borehole, drilled at the center of the basin, where all the formations overlie each other conformably. The rich malacofauna was studied by Tchernov (1973, 1979) and Schütt & Ortal (1993) among others, and the diatoms by Ehrlich (1973). The lists, identifications and conclusions advanced below or in Chapter 12 are based on their studies.

11.2.1.1 *Gonen Formation*

Author: Horowitz & Horowitz (1985).

The type section is at the Notera 3 borehole, from 1,300 up to 1,235 m. The base rests on Melekh Sedom Sands (Palynozone QI) and the Ruman Basalt; the nature of contact is not clear, but seems conformable. The distribution is not known, since the Formation was encountered only in a single borehole. An erosion surface over which the Gadot and Hazor formations were laid down, known from the southern Hula Valley, seems a time equivalent of the Gonen. Lithology comprises sandy, somewhat organic marls deposited in a shallow lake or marshes, in which several basalt flows interfinger. Abundant pollen grains assign the Gonen Formation to Palynozone QII.

11.2.1.2 *Gadot and Hazor formations*

Author: Picard (1952), "Gadot Chalk" and "Hazor Gravel"; amended by Horowitz (1973).

The type sections are located near the settlement of Gadot and at Tel Hazor (Fig. 11.1.1), respectively. At the outcrops both formations, which interfinger with each other, overlie (Fig. 5.4.2) the Ruman Basalt (Heimann 1990), previously considered part of the Cover Basalt (Horowitz 1979, p. 133), over an erosional and taphrogenic unconformity. The two formations (and also the younger Mishmar HaYarden) are unconformably overlain by the Yarda Basalt, the Gadot also by a lateral equivalent of the Notera Formation and the Mishmar HaYarden Formation

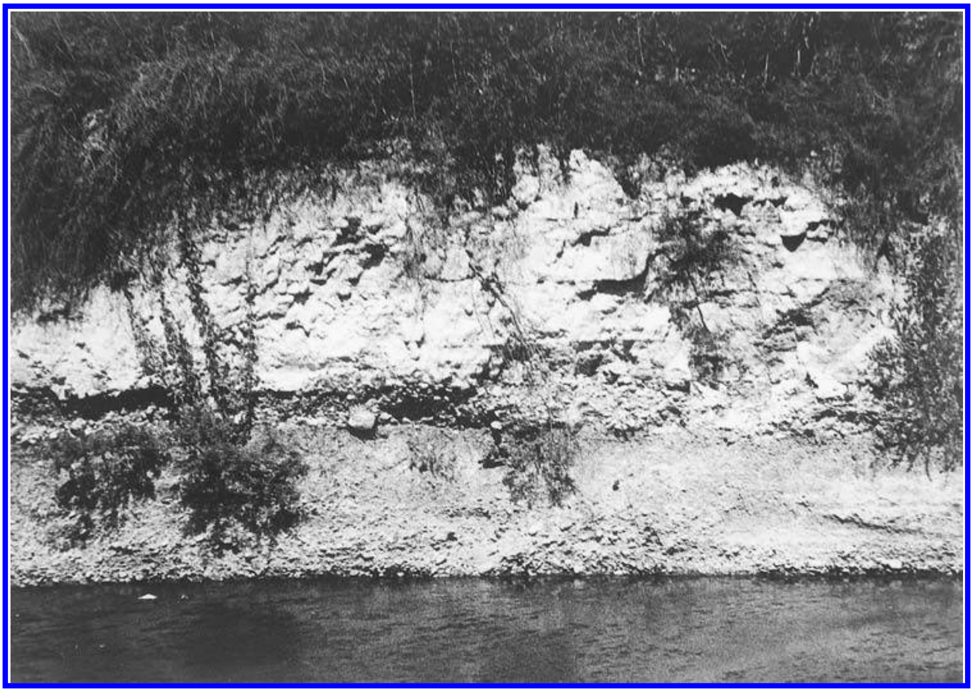


Figure 11.1.1. Lacustrine Gadot and fluviatile Hazor formations, interfingering near Kibbutz Ayyelet HaShahar, southwestern corner of Hula Valley.

at the Benot Ya'aqov bridge outcrops. The Yarda Basalt fills ravines deeply cut into the Gadot, which led Picard (1963) and Schulman (1967) to consider the volcanics older than the sediments. In the Notera 3 borehole the Gadot overlies conformably the Gonen Formation, and is similarly overlain by the Notera Formation. The Gadot and Hazor formations occur in the southern part of the Hula Valley, north of the Korazim Block, covering an area of several square kilometers. The thickness at outcrops is in the order of 25–30 m, attaining up to 100 in some boreholes (Heimann 1990) and 145 at Notera 3 (Horowitz & Horowitz 1985).

The Gadot is made up of white to buff, massive rather dense lacustrine chalk, in which only rare, usually dwarfed forms of *Melanopsis* shells were found. It is almost always covered by a very hard caliche crust, giving it a typical appearance in the landscape. Westward, the chalk grades laterally into the gravel of the Hazor. The gravels are coarse, polymictic, poorly sorted, rounded to sub-angular, in the range of a few centimeters to 20–30 cm across, mostly consisting of Eocene rocks, some late Cretaceous and some weathered basalt components. The Gadot was deposited in a lake extending somewhat south of the present-day Hula Valley, its northward extension unknown. Rivers leading to the lake deposited the Hazor, which extends to the eastward lower reaches of the eastern Galilee hills (Heimann

1990), covered by caliche and soil, where it overlies late Cretaceous through Eocene rocks and also the Dalton Basalt.

The barren, pale, chemically precipitated chalk of the Gadot may indicate a rather saline lake, which did not favor life on the bottom. The lake was shallow, as can be seen from the interfingering gravel horizons, which hardly display any depositional dips. Based on pollen assemblages, the Gadot and the Hazor were assigned to Palynozone QIII. Radiometric datings of the underlying Ruman and the overlying Yarda basalts led Heimann to assign an interval of 1.6–0.9 Ma for the two. This is also supported by an age of 1.81 Ma, obtained by Heimann for a basalt pebble within the Hazor Gravel. Since the Yarda Basalt also overlies the Mishmar HaYarden, the upper age limit is too young. Heimann views the Egel and possibly Si'on gravels as correlative, at least partly, with the Hazor Formation.

11.2.1.3 *Notera Formation*

Author: Horowitz & Horowitz (1985).

The type section is at the Notera 3 borehole, from a depth of 1,090 up to 850 m. The Notera at this borehole is sandwiched conformably between the underlying Gadot and the overlying Mishmar HaYarden Formation. The exact distribution of the Notera is not known, since it was encountered only in a single borehole. The sequence is mainly made of chalk, somewhat organic in the lower part, sandy in the middle; the top is interfingered by several basalt flows. The only fossils are abundant pollen grains, indicating Palynozone QIV. The sediments testify to a shallow lacustrine environment, occasionally grading to paludine.

At the Benot Ya'aqov bridge outcrop, a 2 m thick, grayish-brown paleosol separates the Gadot and Mishmar HaYarden formations. The color and high carbonate contents, as opposed to the more common red paleosols, indicates its formation under relatively dry conditions, supporting its correlation with the Notera Formation.

11.2.1.4 *Mishmar HaYarden Formation*

Author: Picard (1963), "Tilted Villafranchian Beds (*Melanopsis* Stage)"; amended by Horowitz (1973).

The type section was taken at the Benot Ya'aqov bridge outcrop, which is the only one safely assigned to this Formation (Fig. 11.1.2). It is however well developed in the subsurface, and is known from several boreholes. At the outcrop, the Mishmar HaYarden overlies the Gadot Formation unconformably, separated from the latter by a paleosol, which is a lateral equivalent of the Notera; the Formation is overlain, most probably unconformably, by the Yarda Basalt. The thickness at Benot Ya'aqov is only a few meters, comprising white to buff lacustrine chalk and marl, very rich in mollusks and vertebrate bones. The top layer is made of 30–40 cm of peat. In the Notera 3 borehole the sequence reaches 133 m of mainly lacustrine

Figure 11.1.2. Steeply eastward dipping Mishmar HaYarden Formation near the bridge at Benot Ya'aqov, crossing the Jordan River south of the Hula.

marls, somewhat organic, occasionally turning to limestone layers. The upper part contains several paleosol horizons, derived from weathered basalts.

Among the vertebrates, hippopotamus bones are the most common. Mollusks (Tchernov 1973, 1987) include *Theodoxus jordani, Valvata saulcyi, Bulimus hawaderiana, Melanopsis praemorsa, Lymnaea lagotis, L. palustris, Planorbis planorbis, Gyraulus piscinarum, Anisus spirorbis, Segmentina nitidae, Ancylus fluviatilis, Succinea pfeifferi, Pisidium casertanum* and *Unio terminalis.*

Several artifacts, mostly large flakes, were found in the Mishmar HaYarden Formation, and have a general resemblance to assemblages from Ubeidiya (Horowitz et al. 1973; Bar-Yosef in Horowitz 1979, p. 134). Horowitz & Horowitz (1985) attributed the Formation to Palynozone QV. Heimann (1990) considered the Mishmar HaYarden as an upper member of the Gadot Formation; this view is not acceptable, since at the Notera 3 borehole the two are separated by 240 m of the Notera Formation.

11.2.1.5 *Ayyelet HaShahar Formation*

Author: Horowitz (1973).

The type section is at the Emeq Hula 1 borehole, reaching from its total depth of 455 m up to 340. The Ayyelet HaShahar was penetrated in the Notera 3 borehole from 345 down to 717 m, which represents the complete section, sandwiched conformably between the Mishmar HaYarden below and the Benot Ya'aqov Formation above. The lithology is mainly marls, occasionally organic in the lower part of the sequence, known only from Notera 3, while the upper part is highly organic in both sections, with several distinct peat layers. The sequence is interfingered by several basalt tongues. The only outcrop of this Formation is known from the Gesher Benot Ya'aqov area south of the bridge (Fig. 11.1.3), where sediments of the lower sector of Ayyelet HaShahar contain the Middle Acheulian site of Gesher Benot Ya'aqov, rich in artifacts, bones and wood remains (Goren-Inbar 1995, 1998, Werker & Goren-Inbar, in press).

The abundant pollen indicate Palynozone QVI for the Ayyelet HaShahar, both in boreholes and the site. Mollusks remains recovered from the drill cuttings include *Melanopsis, Melanoides* and *Theodoxus*, but *Viviparus* occurs in great numbers at the outcrop, where a variety of vertebrate remains were found, including an elephant skull. A detailed study of melanopsids from beds of the Gesher Benot Ya'aqov site (Heller & Sivan, in prep.) revealed a population of some seven species, with a resemblance to the Ubeidiya assemblage. Besides the layers containing the site, an unconformity, both erosional and structural, which separates the Mishmar HaYarden and Benot Ya'aqov formations at the Benot Ya'aqov bridge outcrop, is a time equivalent of the Ayyelet HaShahar Formation.

Figure 11.1.3. Ayyelet HaShahar Formation, at the Benot Ya'aqov Acheulian site, south of the bridge. Photo courtesy of N. Goren-Inbar.

11.2.1.6 *Benot Ya'aqov Formation*

Author: Stekelis et al. (1937), "Benat Yaqub Gravel"; Picard (1952), "Lower Lacustrine"; Picard (1963), "Jordan Terrace"; amended by Horowitz (1973).

The type section is nearby to the north of the Benot Ya'aqov bridge. Incidentally, the name "Benot Ya'aqov" in Hebrew, or Banat (also Benat) Yaqub in Arabic, is a translation of "Filles de Saint Jacques", an order whose nuns acquired the concession for passage at this locality in the 19th century. In this locality (Fig. 11.1.4) the Formation overlies, over a base conglomerate, a faulted and eroded relief of the Mishmar HaYarden Formation and the Yarda Basalt. The top is truncated by a base conglomerate of the overlying Ashmura Formation. The outcrop faces the Jordan River gorge, along a few hundred meters from south of the former Hula Lake to somewhat north of the bridge at Benot Ya'aqov. The amended type section is at the Emeq Hula 1 borehole, where the Formation is fully developed, from 340 up to 155 m. The Formation is known from many drillings in the Hula Valley, where it is conformably situated between the Ayyelet HaShahar below and the Hulata Formation above. A single, very limited outcrop south of Lake Kinneret (Braun & Tchernov 1991) has a similar lithological and faunistic appearance as the Benot Ya'aqov Formation, bearing plenty of *Viviparus apameae* mollusk

Figure 11.1.4. Benot Ya'aqov Formation just north of the bridge, showing the numerous *Melanopsis* and *Viviparus* (bigger) shells.

shells. It is however questionable whether this outcrop could be regarded as part of the Formation, as it is embedded within the Naharayim Formation conglomerates and loams.

Benot Ya'aqov comprises mainly shallow lacustrine marls, organic to some degree in the lower and upper parts, much more so in the middle sector, and is very rich in mollusks shells both at the outcrops and the boreholes. The lacustrine sediments grade southward, westward and eastward, toward the Valley's margins, into fluviatile gravels, and northward into extensive sheets of the Kefar Yuval Travertine. The most abundant mollusk is *Viviparus apameae galileae*, typical of this Formation but also occurring in the underlying Ayyelet HaShahar (Schütt & Ortal 1993). Other mollusks typical of Benot Ya'aqov are *Semisalsa longiscata, Melanopsis corrugata* and *M. blanckenhorni*. Others, not restricted to this Formation, include numerous *Valvata saulcyi*, *Bythinella* sp., *Bulimus hawaderiana*, *Melanopsis praemorsa* and *Pisidium casertanum.* Less frequent are *Theodoxus jordani*, *Lymnaea lagotis*, *Gyraulus piscinarum* and *Unio terminalis*.

The outcrop at Benot Ya'aqov yielded abundant vertebrate fossils described in detail in Hooijer (1959, 1960), including *Equus* cf. *caballus*, *Elephas trogontherii*, *Dicerorhinus merckii*, *Sus* cf. *scrofa*, *Hippopotamus amphibius*, *Dama* cf. *mesopotamica*, *Cervus* cf. *elaphus*, cf. *Bison priscus* and *Stegodon mediterraneus*. The outcrop is thought to represent the lower part of the Formation, and excavations in this locality yielded rich flint and basalt artifact assemblages of African Middle Paleolithic affinity (Stekelis 1960, Gilead 1970), as well as a human femur (Geraads & Tchernov 1983). This site is referred to here as "Jisr Banat Yaqub" (see remark in Section 11.3.5).

Pollen are especially abundant both at the outcrop and boreholes, assigning the Benot Ya'aqov Formation to Palynozone QVII. Diatoms are very copious, with dominant planktonic species such as *Cyclotella kutzingiana, Melosira ambigua, Stephanodiscus astraea* and *Fragilaria* spp. The sediments and diatoms indicate that the Benot Ya'aqov was deposited in a fairly deep alkaline lake, extending beyond the present-day Hula Valley limits. The middle part was laid down in a smaller, rather shallow lake, or in marshes. The middle and upper parts are known only from boreholes, the former due to shrinkage of the lake, the latter because of subsequent erosion during the dry QVIII.

11.2.1.7 *Hulata Formation*

Author: Picard (1952), "Main Peat"; amended by Horowitz (1973).

The Hulata Formation was defined in the "No. Zero First Test" borehole, at the northern border of the former Hula Lake. The amended section is from the Emeq Hula 1 borehole, where the Hulata is sandwiched conformably between the Benot Ya'aqov and the overlying Ashmura Formation. The Formation was penetrated by numerous drillings all around the Hula Valley, but is not known from outcrops save for some lateral gravel horizons (Heimann 1985).

The sequence at the "No. Zero" borehole comprises 54 m, of which the lower 20 and upper 19 consist of massive peat. The middle part, 15 m, is mainly made of peat with several limnic chalk intercalations. The thickest sequence, which is somewhat more marly, of 75 m, was encountered in the Emeq Hula 1 borehole. Pollen assemblages assign the Hulata Formation to Palynozone QVIII. Diatoms include high percentages of epiphytic species such as *Epithemia* which together with abundant Chrysostomaceae and phytoliths indicate a very shallow, alkaline, eutrophic water environment of deposition.

11.2.1.8 *Ashmura Formation*

Author: Picard (1952), "Upper Lacustrine"; amended by Horowitz (1973).

The amended type section of the Ashmura comprises the uppermost 80 m of the Emeq Hula 1 borehole, where it overlies conformably the Hulata Formation peat. In localities where the Mallaha Formation is developed (see below), it overlies the Ashmura conformably. The Ashmura Formation is known from boreholes all around the Hula Valley, and from outcrops along the Jordan River to the south (Fig. 11.1.5) down to the Benot Ya'aqov bridge. It was deposited on the northern edge of the Korazim Block until about 4,500 years ago, when this area was

Figure 11.1.5. Chalk of the Ashmura Formation east of the Jordan River, just south of Lake Hula.

tectonically uplifted (Carmi, in Horowitz 1979, p. 138). Northward, the Ashmura grades into gravel and paleosols, interfingering with extensive travertines (Heimann 1985).

The upper part of the Ashmura grades laterally into the Mallaha Formation peat, both of which have been deposited concurrently in the Hula Valley until the present-day, the Ashmura in the lake and the Mallaha in the marshes. In places where the Hulata Formation is absent, the Ashmura overlies unconformably older rocks, occasionally over a base conglomerate. The Ashmura is found all over the Hula Valley, deposited in a lake that occupied the region until recently, when it was artificially drained. During the Holocene the lake gradually shrank, giving place to marshes to the north, where peat of the Mallaha Formation was deposited. The type section comprises white to gray lacustrine chalk, with several horizons rich in organic material, but is rather poor in mollusk remains except for the uppermost few meters, which yielded a rich assemblage comprising *Theodoxus jordani*, *Valvata saulcyi*, *Bythinella* sp., *Bulimus hawaderiana*, *Melanopsis praemorsa*, *Melanoides tuberculata*, *Lymnaea lagotis*, *Gyraulus piscinarum*, *Ancylus fluviatilis*, *Succinea pfeifferi*, *Pisidium casertanum*, *Sphaerium* sp. and the clams *Corbicula fluminalis* and *Unio terminalis.*

Diatoms are quite abundant, especially planktonic forms such as *Cyclotella* and *Melosira*. Benthonic diatoms are only accessories, while epiphytic are rare, indicating a fairly deep, alkaline, freshwater lake, similar to the lake in which the Benot Ya'aqov Formation was deposited. The highly organic beds of the Ashmura are typified by diatoms comparable to those of the underlying Hulata Formation. The uppermost part is characterized by an abundance of epiphytic diatoms, indicating shallowing of the lake. Pollen assemblages of the Ashmura assign the major part of the sequence to Palynozone QIX, while the uppermost part is QX. The outcrops at Benot Ya'aqov yielded some artifacts, ranging from Mousterian at the lower part through to Chalcolithic and Early Bronze at the top (Bar-Yosef, in Horowitz 1979, p. 139).

11.2.1.9 *Mallaha Formation*

Author: Picard (1952), "Upper Peat"; amended by Horowitz (1971).

The type section is from the UP 15 borehole, drilled in the center of the marsh area north of the Hula Lake, comprising 12.7 m of peat, conformably but diachronously overlying and grading laterally to the Ashmura (see above). The peat is abundant in the northern part of the Hula Valley (Fig. 11.1.6), occupied by marshes until they were artificially drained recently. Numerous pollen grains indicate the upper part of Palynozone QX, but other fossils are quite rare in the Mallaha. Diatoms are characterized by a fairly poor epiphytic flora, dominated by *Fragilaria* and some acidophilous forms such as *Eunotia*. These are accompanied by abundant sponge spicules, phytoliths and Chrysostomaceae cysts, indicating a somewhat more acid environment in comparison with the underlying Ashmura, of a very shallow,

Figure 11.1.6. Peat of the Mallaha Formation (dark layer in the trench), north of Lake Hula.

eutrophic water body mixed with some acid waters. Eastward, westward and northward the Mallaha grades into soils, some gravel and subordinate spring deposits. Radiocarbon dates indicate the Mallaha deposition commenced some 4,500 years ago at its deepest occurrence (Horowitz 1971). The base is, however, diachronous, transgressive over the Ashmura, and progressively advancing southward.

11.2.1.10 *Marginal gravel beds*

Mor (1986) described a sequence of some 80+ m of gravels chiefly made of Jurassic components, termed the “Si’on Gravel”, occurring in the northern Hula Valley and on the slopes of Mount Hermon, where Jurassic rocks crop out nearby. The Si’on is considered by Heimann (1985) to be the northern correlative of the Hazor Formation. Later, however, Heimann (1990) viewed the Si’on Gravel as a marginal equivalent of most of the lacustrine Quaternary formations of the Hula basin (Fig. 11.1.7). He also adds another term, “Dishon Gravel”, to describe what look like similar gravel occurrences on the eastern slopes of the Galilee hills. An allegedly equivalent unit, made of basalt pebbles, occurs southward along the Golan escarpment down to Lake Kinneret. These are termed by Mor (1986) “Buteiha Gravel”, and are considered similar to the Si’on except for the lithology, which indicates a different source at the same time and under similar conditions.

Figure 11.1.7. Marginal gravel beds of the Ashmura Formation, south of Lake Hula.

Somewhat more problematic is a gravel unit defined in Heimann (1985) as the "Egel Gravel", some 10–40 m thick, consisting of Jurassic, Cretaceous and possibly Pliocene components, cemented by a reddish, hard calcareous matrix, at the northern end of the Hula Valley. On the one hand, Heimann (1985) suggests a correlation either with the Hazor Formation or with the Si'on Gravel. In 1990 Heimann regarded the two as separate entities. Horowitz (1973) Sneh (1996) and U. Kafri (Geological Survey of Israel 1998, pers. comm.) considered these gravel beds correlative with the Pliocene Tanur Conglomerate, viewing the variations in pebble lithology as indicating their arrival from different sources. Since nothing conclusive changes this view, it is still held here.

11.2.2 Jordan Group

Horowitz (1974) applied the term to embody a suite of formations cropping out in the central Jordan Valley, from the time the Rift became an endoreic system until the present day. The central Jordan Valley basin extends from Lake Kinneret southward, most probably down to Marma Feiyad, or somewhat north of this locality. The Tabgha Formation, deposited in Lake Kinneret during the last 18,000 years or so, is also included in this Group. Outcrops of the Jordan Group have been studied in great detail, chiefly due to the abundance of artifacts and sites, especially in its earlier parts, bearing witness to the antiquity of human presence in this region. Subsurface information is much more scanty, since only a single borehole, the Zemah 1 just south of Lake Kinneret, was sunk deep enough.

Tectonically the central Jordan Valley basin acted in a completely different way as compared with the Hula and the Dead Sea, its major phase of subsidence being restricted to the Neogene and earliest Quaternary with more than 4 km of rocks being accumulated in those times (Marcus & Slager 1985; Horowitz & Horowitz 1990; Horowitz 1992a, p. 331). This leaves the Jordan Group subordinate in thickness, involving only the uppermost 250 m of the Zemah 1 borehole (Fig. 6.1.3), a thickness quite similar to the exposed sequence (Schulman 1959). The Group comprises mainly lacustrine formations laid down in wetter climates, the dry periods being represented only by erosion, save for an early one (QIV) known from the Zemah 1 borehole, while Holocene sediments are at the surface, and are not yet completely eroded or covered. During Jordan Group times, the central Jordan Valley was always an intermediate basin on the way to the Dead Sea.

The Jordan Group overlies unconformably, at the outcrops, the Cover Basalt and older rocks (Schulman 1959, 1962). The youngest age obtained for underlying volcanics is in the range of 2.0–1.7 Ma (Siedner & Horowitz 1974). Recent definitions tend to exclude these rocks from the Cover Basalt proper (Heimann et al. 1996), but nonetheless the ages give a maximum limit for the start of the Jordan Group. At Zemah 1, the Jordan Group formations unconformably overlie QI Melekh Sedom Sands, with Palynozone QII missing (Horowitz & Horowitz 1990).

11.2.2.1 *Erk el Ahmar Formation*

Author: Horowitz (1974).

The name is derived from a north-facing cliff (the red cliff) of the Jordan gorge, close to the road from Gesher to Menahemya (Fig. 11.2.1) some 10 km south of Lake Kinneret, which for a long time was the only recognized outcrop of this Formation (Horowitz 1979, p. 140). This outcrop is known for quite a long time, and was discussed by, among others, Blanckenhorn (1914), Blanckenhorn & Oppenheim (1927), Picard (1932, 1943), Schulman (1959) and Picard & Baida (1966a). All these investigators attributed the outcrop to the Ubeidiya Formation, so much so that the only photograph of the "Ubeidiya" in Picard & Baida (1966a) is the type locality of Erk el Ahmar.

The first to note the difference between the two formations, suggesting that Erk el Ahmar may be older, was Baida in 1964 (in Horowitz 1979, p. 140). This was verified by Horowitz (1974), based on field relations and palynostratigraphic considerations, and by Tchernov (1975), who indicated the differences in malacological assemblages of the two. E. Tchernov and his team (Department of Evolution, Systematics and Ecology, the Hebrew University of Jerusalem 1998, pers. comm.), who made a detailed survey of the Formation, have discovered further outcrops

Figure 11.2.1. Tilted beds of the Erk el Ahmar Formation, at its type locality south of Lake Kinneret, overlain unconformably by horizontally lying Lisan Formation sediments.

of the Erk el Ahmar to the south, west of the Jordan River down to Nahal Tavor, covering an area of several square kilometers (Fig. 12.1.2.1). It was suggested (Horowitz 1979, p. 141) that the lower part of a sequence described in Bender (1974a, p. 94) from Abu Habil east of the Jordan River, some 25 km south of Nahal Tavor, may represent the southward extension of the Erk el Ahmar. This correlation is however doubted by Macumber & Edwards (1997) and by Copeland (1998), but see the discussion of the Abu Habil, below.

Close to the type section, the Erk el Ahmar sediments overlie the Cover Basalt, on a faulted, erosional relief (Horowitz 1979, p. 140). In the subsurface it overlies the Melekh Sedom Sands unconformably in the Zemah 1 borehole, with Palynozone QII missing (Horowitz & Horowitz 1990). The Erk el Ahmar is almost everywhere truncated by the late Pleistocene Lisan Formation, over an angular unconformity. In some localities older formations, such as the Naharayim and the Yarmouk Basalt, overlie unconformably the Erk el Ahmar (Tchernov, in Schütt & Ortal 1993). The Yarmouk Basalt is K–Ar dated at 0.8 Ma (Mor & Steinitz 1985). At the Zemah 1 borehole the Erk el Ahmar is overlain by sediments of Palynozone QIV, but it is difficult to assess the nature of the contact, which seems conformable. The thickness both at the type section and the subsurface is 80–90 m, comprising clays and marls, occasionally varved, with rare occurrences of silts and sands, some gravel horizons of basalt, flint and limestone components up to 20–25 cm across, with several thin lacustrine limestones.

The Erk el Ahmar sediments yielded extensive mollusk assemblages (Tchernov 1975, Schütt & Ortal 1993), some vertebrate bones such as a giraffe or camel, and an impressive elephant skull and tusks (Fig. 12.1.2.1A), as yet unidentified. In addition, hominid artifacts such as chopping tools (Fig. 12.1.2.1B) and flakes were unearthed from the upper part of the sequence (Horowitz 1979, p. 296; Braun et al. 1991; E. Tchernov 1998, pers. comm.). Of the rich malacofauna recovered from the Erk el Ahmar, the following are unique to this Formation: *Theodoxus jordani pliocostulus*, *Viviparus unicolor*, *Bythinia multicostata*, *Limnidia hebraica*, *Staja lycica*, *Falsipyrgula barroisi*, *Ginaia picardi*, *Chilopyrgula ahmara*, *Melanopsis aaronsohni*, *M. praecursor*, *Melanoides dadiana*, *M. jordanica*, *Potomida subrectangularis*, and *Dreissena israelica.*

Paleomagnetic studies of the sequence (Verosub & Tchernov 1991, Braun et al. 1991, Ron & Levy 2001) revealed a reverse–normal–reverse alternation, the upper part of which is attributed in both studies either to the beginning of the Olduvai or the Upper Reunion Subchrone, which assigns the Erk el Ahmar an age of approximately two million years. Based on the rich pollen flora, Horowitz (1992a, p. 410) correlated the Formation with Palynozone QIII. The Formation was laid down primarily in a shallow freshwater lake, with rivers supplying the gravel. Chemical and isotopic analyses indicate that it was essentially a freshwater lake (Rosenthal et al. 1989), with salinity in the range of 300–400 mg/l.

11.2.2.2 *Palynozone QIV*

Author: Horowitz & Horowitz (1985).

Some 30 m of sediments attributed to this Palynozone (Horowitz & Horowitz 1990) separate the Erk el Ahmar and Ubeidiya formations in the Zemah 1 borehole. The lower and upper contacts seem to be conformable, but are hard to assess with any certainty. The palynozone is discussed in detail in Chapter 6.

11.2.2.3 *Ubeidiya Formation*

Author: Blanckenhorn (1897), "*Melanopsis* (or *Unio-Melanopsis*) Stufe"; Picard (1932), "Levantinische Stufe", amended by Picard & Baida (1966a,b).

The Formation is named after an abandoned settlement, which also gave its name to the prehistoric site, some 3 km south of Lake Kinneret on the western bank of the Jordan River. The type section is located at the excavations of the site (Figs 11.2.2 and 12.1.4.1). The base is not exposed, but to the west the Ubeidiya overlies the Cover Basalt and Neogene rocks, which are faulted and eroded. The top is always truncated, and is either overlain with angular unconformity by the Naharayim and Lisan formations or else is exposed. Some 30–40 m of the Ubeidiya Formation are known from the Zemah 1 borehole, overlying Palynozone

Figure 11.2.2. Tilted Ubeidiya Formation beds, at its type locality some 3 km south of Lake Kinneret.

QIV sediments with what seems to be a conformable contact. The top is most probably truncated, and is covered by the Naharayim Formation sediments (Palynozone QVII).

The Ubeidiya Formation is typically known from the region south of Lake Kinneret, covering an area of a few square kilometers. Rosenthal (1980, p. 46) suggested that several occurrences of conglomerates and marls, containing *Melanopsis*, known from outcrops and boreholes in the Bet She'an region, should also be regarded as the Ubeidiya Formation. The type section attains 190 m, but this represents only part of the original sequence, considering the nature of the lower and upper contacts. It was studied in trenches dug in the course of excavations. The Formation was subdivided by Picard & Baida (1966b) into four members. The lowermost, the "Clay, Silt and Limestone Member (Li)" is 52 m thick, comprising lacustrine clay and silt varves, hard oolitic limestone beds, soft chalk and marls and silty, occasionally sandy layers. Except for the varves, all horizons are rich in freshwater mollusks and vertebrates, primarily fish. Veins of secondary idiomorphic gypsum are abundant within the varves. Some well-preserved plant fossils were found in the uppermost varves (Lorch 1966), such as *Rhus tripartita*, *Pistacia lentiscus*, *Myriophyllum* and *Ranunculus*. *Melanopsis* shells are very abundant,but this Member is devoid of artifacts and terrestrial vertebrates. Several horizons were deposited in slightly saline environments, yielding *Ammonia beccarii* fossils (Almogi-Labin et al. 1995).

The overlying "Intraconglomerate and Clay Member (Fi)" is 22–30 m thick, displaying different facies on each flank of a miniature anticline, exposed at the site during excavations. The western flank consists entirely of clays and conglomerates, the latter comprising sub-angular components, and the matrix varies between clay, chalk, marl, silt and basaltic sand. Some beds yielded artifacts, among which choppers, flakes and cores are common (Stekelis 1966b). The layers contain numerous freshwater gastropods and ostracodes, together with well-preserved vertebrates bones. The clay beds of this Member are sometimes reddish, unfossiliferous, most probably paleosols. On the eastern flank of the minor fold the conglomerate horizons are thinner, made of smaller components, and often very sandy or silty. Clays are predominant, in which limestone and chalk beds intercalate, and very rich in mollusks fossils. Picard & Baida (1966b) concluded that to the west it was deposited mainly in a river, grading to swamp and lake environment eastward, over a rather short distance (see also Bar-Yosef & Tchernov 1972).

This Member was the first to be excavated (Stekelis et al. 1960) on the assumption that it was the origin of *Homo erectus* remains (Tobias 1966) found scattered in a vineyard at this locality. Stekelis (1966b) classified the artifacts from the lower part of this Member as the "Israel Variant of Olduvan II Culture, Phase I", while those from the upper part are designated as "Phase II". Bar-Yosef (1994) referred to the lithic assemblages as Lower Acheulian. The fauna was determined by Haas (1966) and Tchernov (1987 and other publications), comprising rich assemblages of vertebrates and mollusks. Abundant mammals include *Crocidura russula*, *Canis arnensis*, *Praemegaceros verticornis*, *Hippopotamus behemoth*, *Equus tabeti* and numerous rodents. Mollusks typical of the Ubeidiya are (Schütt & Ortal 1993) *Theodoxus jordani unicarinatus*, *T. j. bicarinatus*, *T. j. tricarinatus*, *Bithynia phialensis syriaca*, *Melanopsis obediensis*, *M. turriformis*, *Potomila littoralis ubeidiyensis* and *Sinanodonta tchernovi*, among many other taxa.

The third, the "Main Silt Member (Lu)", reaches a thickness of 56 m, consisting of lacustrine, occasionally well-bedded white-grayish-yellow silts, with some silty marl and clay layers, very rich in ostracodes. The lower part is mainly well-bedded limnic clay, chalk and chalky marls, in which freshwater ostracodes, gastropods and fish remains are quite abundant. Occasional increased salinity is indicated by rare finds of the foraminifer *Ammonia beccarii* (Almogi-Labin et al. 1995). This is overlain by the uppermost "Upper Conglomerate Member (Fu)" made of 16 m of conglomerates, which range in size from very fine micro-conglomerates to big boulders, mainly of basalt, some flint and a few limestone constituents. The base displays fine clastics, interspersed in a 2 m thick marly chalk layer. The clastic layers reveal repeated cycles of graded, well-rounded and sorted pebbles, occasionally interrupted by less-rounded, unsorted horizons. The uppermost layer contains big, fairly rounded basalt boulders up to 80 cm across and small flat pebbles up to 10 cm in diameter. Neither fossils nor artifacts are found in this Member.

Early investigators suggested a late Pliocene or early Pleistocene age for the Ubeidiya Formation (including Erk el Ahmar). During the 1960s the term "Villafranchian" was applied (Stekelis et al. 1960, followed by others), based on faunal comparisons with European sites. Bar-Yosef & Tchernov (1972) compared the lithic assemblages and suggested a Mindel age, which was followed by Horowitz (1974, 1979, p. 173, 1988). Tchernov (1987), again based on faunal considerations but having collected numerous additional fossils, suggested a Biharian age, approximately corresponding to 1.4 million years. K–Ar ages (Horowitz et al. 1973, Siedner & Horowitz 1974, Mor & Steinitz 1985) placed the Ubeidiya between the 1.7 Ma of the uppermost flows of basalts then considered Cover Basalt, and the Yarmouk Basalt, in the range of 0.8–0.6 Ma. Paleomagnetically (Opdyke et al. 1985, Verosub & Tchernov 1991, Braun 1992) the entire sequence is reversed, as expected for the Matuyama.

The Ubeidiya Formation was deposited in a wide shallow lake which occupied the central Jordan Valley, changing its size, depth and shorelines several times during accumulation of the sequence. Rivers flowing to the lake brought gravel, common in the Formation especially at times of lake shrinkage. Chemical and isotopic analyses indicate a medium salinity, 1,000–1,300 mg/l, for the Ubeidiya Lake (Rosenthal et al. 1989), which most probably received some of its waters from saline springs.

11.2.2.4 *Abu Habil Formation*

Author: Bender (1974a).

An outcrop comprising up to 30 m of hard, conglomeratic, partly pisolitic limestones, was found near the village of Abu Habil east of the Jordan River (Fig. 11.2.3) some 36 km south of Lake Kinneret (Bender 1974a, p. 100), overlying unconformably, over an angular and erosional relief, reddish sandstones and conglomerates. These most probably are part of the Herod Formation, although they are considered both by Bender and Macumber & Edwards (1997) to represent the Ghor el Qatar Series, a possibility which should not be discarded. Huckriede (1966) maintains that the Abu Habil predates the Yarmouk Basalt. The Abu Habil was correlated by Bender and by Tchernov (1987) with the Ubeidiya Formation, while Horowitz (1979, p. 144) thought that at least its lower part may correlate with the underlying Erk el Ahmar, while only the upper may correspond to Ubeidiya. All these age assignments are, however, doubted by Macumber & Edwards (1997) and by Copeland (1998), who maintain that the Abu Habil is considerably younger.

Artifacts were reported from the upper part of the Formation by Huckriede (1966), who defined them as pre-Abbevillian pebble tools of Olduvan and Münzenberger type, which have the same nature as those recovered at Ubeidiya. Huckriede also found a few *Melanopsis praemorsa* shells, together with some bones, possibly of antelopes. These finds are questioned by Macumber & Edwards

Figure 11.2.3. Tilted beds of the Abu Habil Formation, at the village some 40 km south of Lake Kinneret.

(1997) and by Copeland (1998), who could not locate any but Middle and Upper Acheulian artifacts in these layers. It seems however that the latter authors have not searched exactly the same outcrop originally described by Bender and Huckriede, which comprised steeply dipping beds. The newly investigated occurrences show only mild tilting, and so may be younger, possibly belonging to the Tabaqat Fahl Formation, as indeed suggested by Macumber & Edwards. The similarity in lithology is not surprising, since both formations were deposited in almost the same location, under very much the same conditions.

An assemblage of melanopsids from the Abu Habil was recently studied in detail (J. Heller, Department of Evolution, Systematics and Ecology, the Hebrew University of Jerusalem 1999, pers. comm.), comprising *Melanopsis buccinoidea*, *M. jordanoscalaris* and *M. costata*. Heller maintains that they most resemble the assemblages from the Erk el Ahmar Formation, supporting my earlier assignment.

11.2.2.5 *Tabaqat Fahl Formation*

Author: Macumber (1992).

The Formation is named after the Tabaqat Fahl flat hill where the type section is described, bordering the Jordan Valley to the east (Fig. 11.2.4), some 28 km

Figure 11.2.4. Tabaqat Fahl Formation at its type locality.

south of Lake Kinneret opposite the Yizre'el Valley. The Formation overlies unconformably dipping and eroded Cretaceous and early Tertiary rocks on the hills bordering the Rift, while closer to the valley it rests on steeply tilted reddish sandstones and conglomerates, identified here as the Miocene Herod Formation, although Macumber relates them to the Ghor el Qatar Series, perhaps correctly. The upper surface of the Tabaqat Fahl is cut and filled up with the fluviatile Wadi al Hammeh Conglomerate, correlated by Macumber with the Lisan Formation of Palynozone QIX age.

It is quite difficult to assess the exact distribution of Tabaqat Fahl sediments. Other occurrences, which appear in a similar sequence stratigraphic order, are known along the Jordan Valley on both sides, but their exact assignments were never studied in enough detail. The thickness of the Tabaqat Fahl at the Rift Valley rim is approximately 120 m, decreasing eastward inside the wadis. It consists of two distinct parts, the lower comprising dense conglomerates and hard white limestones, occasionally nodular, while the upper, sometimes up to 100 m thick, consists of travertines (referred to as "tufas" by Macumber, but since the *Glossary of Geology* defines the "hard, dense variety" as "travertine", this term seems much more appropriate in this case).

Common fossils are *Phragmites* reeds and *Melanopsis praemorsa* shells, which indicate a spring environment. Middle Acheulian artifacts and sites are

Figure 11.2.5. Gravel beds of the Naharayim Formation, at its type locality south of Lake Kinneret.

embedded in the lower part of the Tabaqat Fahl, while the upper is typified by Upper Acheulian ones. These, together with the sequence stratigraphic position, place the Formation in Palynozone QVII, so it could be regarded as a lateral variant of the Naharayim Formation.

11.2.2.6 *Naharayim Formation*

Author: Nötling (1887), "Alt-Alluviale Jarmuq-Schotter"; amended by Picard (1932), "Naharaim-Schotterstufe"; and by Picard (1965).

The Formation is named after the confluence of the Yarmouk and Jordan rivers south of Lake Kinneret ("Naharayim" means two rivers), where the type locality was defined. The older definitions refer only to the gravel beds (Fig. 11.2.5) sandwiched conformably between thick brown paleosol horizons, termed by Picard (1932) "Basislehme A" and "Hangendlehme B". These were subsequently included by Picard (1965) as parts of the Naharayim Formation. The Naharayim rests unconformably on the Yarmouk Basalt and older formations both in outcrops and boreholes. It is overlain, again unconformably, by the Raqqad Basalt and the Lisan Formation. The Naharayim *s. str.* is known only from the central Jordan Valley. However, stratigraphically similar occurrences of gravel horizons were recognized along the Jordan Valley down to the Dead Sea (Bender 1974a, p. 97), and possibly also further south. Bender (1974a, p. 94) includes within the Naharayim the poorly consolidated gravels with red argillaceous matrix and with Lower Paleolithic artifacts of the Kufrinja–Yabis area, east of the Jordan River, 30–50 km south of Lake Kinneret (Fig. 11.2.6).

The Naharayim sequence, at the type locality, comprises up to 20–30 m of mainly medium-sized basalt and limestone gravels, intermingled with red, occasionally sandy loams, most probably floodplain paleosols and some thin sandy horizons. In a single locality some 10 km south of Lake Kinneret, a thin bed very rich in *Viviparus apameae* shells (Braun & Tchernov 1991, Braun 1992), that could have been laid down in a small shallow lake or pond, was found at the base of the Naharayim, overlying the Yarmouk Basalt; besides, only rare melanopsids occur. The thickness decreases gradually in all directions from the type area to a few meters or less.

Rare artifacts of Upper Acheulian affinity are reported from the Naharayim and its correlatives (Huckriede 1966). It seems that the Naharayim was deposited in a large delta and floodplain complex of the Yarmouk and Jordan rivers, in times of increased water supply. Radiogenic ages for the Naharayim are between 600 and 70 Ka, of the underlying Yarmouk and overlying Raqqad basalts (Siedner & Horowitz 1974), respectively.

Figure 11.2.6. Kufrinja Gravel, at the confluence of Wadi Kufrinja and the Jordan Valley.

11.2.2.7 *Lisan Formation*

Author: Lartet (1869, but described already by former travelers, see Chapter 2), "Dépôt de la Liçan", "Marnes de la Liçan", "Anciens Dépôt de la Mer Morte"; amended by Bentor & Vroman (1954b); by Langozky (1960); by Begin et al. (1974); by Horowitz (1974); and by Eyal (1984).

The Formation was named after the Lisan (Arabic for "tongue") Peninsula, on the eastern side of the Dead Sea. Covering wide areas from Lake Kinneret through to the northern Arava, it is by far the best known and intensively studied formation of the Jordan Valley (265 papers, directly related, in the Geological Survey of Israel library, 1998). As such, it acquired a multitude of synonyms: "Ancient Deposits of Salt Sea" (Hull 1886), "Lisan Mergel" (Blanckenhorn 1914), "Lisan Stage" (Picard 1937), "Lisan Marl" (Picard 1938), "Lisan Series", "Lisan Deposits" (Picard 1943), and so on. There is no "official" type section of the Lisan Formation, most probably a result of the proliferation in studies. Reference sections are however numerous, notably by Zak (1967), Picard & Baida (1966b), Neev & Emery (1967), Begin et al. (1974) and others.

The Lisan overlies a multitude of older formations, but rests conformably only on sediments of Palynozone QVIII, a contact known only from boreholes in the Dead Sea area (see Chapter 6). The youngest formation overlain by the Lisan in the central Jordan Valley, possibly quite conformably, is the Raqqad Basalt (Figs 11.2.7 and 11.8.4). The thickness is some 60–80 m in the Dead Sea region, somewhat diminishing northward, unless one adopts Bentor & Vroman's (1954b, 1957) attitude and includes the entire Upper Dead Sea Group (of which some 600 m were known at the time from drillings), within the Lisan. Lake Lisan extended approximately from the middle of the present Lake Kinneret, south through the northern Arava, down to some 35 km south of the Dead Sea.

Following Langozky (1960), Begin et al. (1974) and others, Horowitz (1979, p. 148) suggested a threefold subdivision of the Lisan Formation into members (Fig. 11.2.8). The lower, termed Hamarmar after Langozky (1960) and the upper, termed Fatza'el (now transliterated "Peza'el") by Horowitz (1974), are clastic. The middle, Ami'az Member (Langozky 1960), predominantly comprising alternating varves of aragonite and clay or silt of lacustrine origin, was subdivided by Begin et al. into the "Laminated" and the "White Cliff" members, given the status of beds by Horowitz. The southernmost extension of the Lisan Formation in the Arava, particularly its upper parts, is very sandy (Fig. 11.2.9). This facies was termed the "Iddan Formation" by Eyal (1984), a name which is partly retained here but with the status of member, as a southern equivalent of the Fatza'el (or Damiya) Member of the central and southern Jordan Valley.

The reference section near Deir Shaman (Begin et al. 1974), 15 km north of the Dead Sea at the middle of the basin, seems to be one of the better studied representatives of the Lisan Formation, reaching a thickness of 58+ m, with the base of the sequence not exposed. The Hamarmar Member comprises 17 m

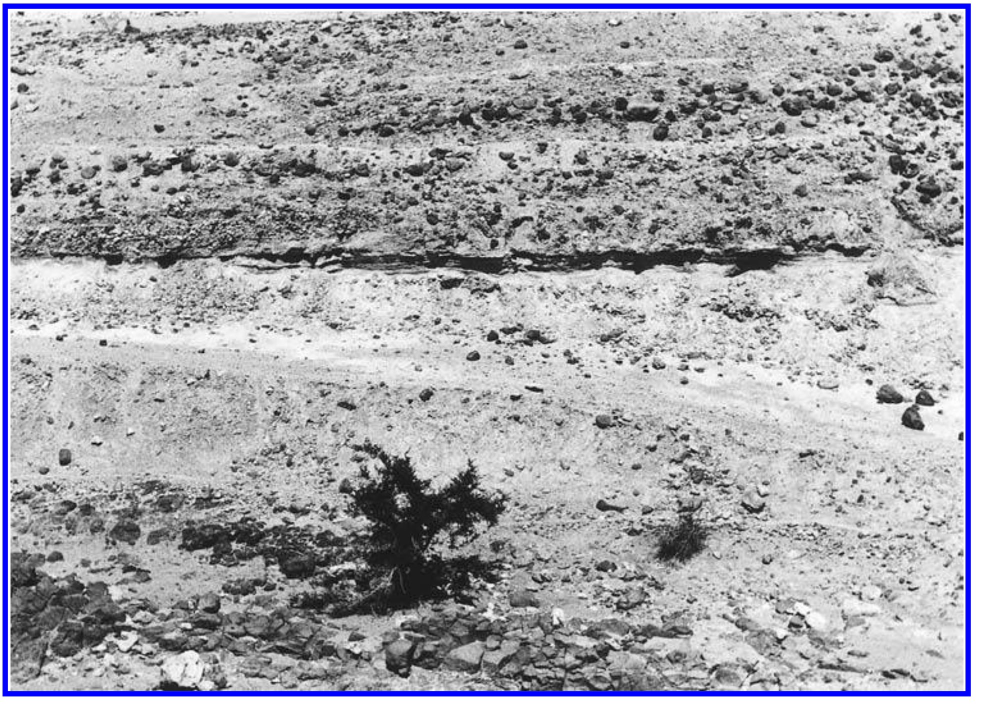


Figure 11.2.7. Lisan Formation, overlying a basalt flow correlative of the Raqqad, at the northeastern corner of the Dead Sea.

predominantly of sands and sandy silts, which grade outward at the basin's margins to gravel. The unit is diachronous, and occasionally the entire Lisan sequence becomes clastic toward the shores of the lake. These gravels acquired several names, such as "Sandy Member" (Bentor & Vroman 1954b) or "Hamarmar" (Langozky 1960) in the southern Dead Sea region, "Samachschotter" (Picard 1932) in the central Jordan Valley, or was (erroneously) thought to represent the Samra Formation (Begin et al. 1974; Bender 1974a, p. 95).

The conformably overlying Ami'az Member is made of 25 m of the Laminated Bed, overlain by 6–7 m of the White Cliff Bed. The former comprises finely laminated, whitish soft varves of alternating aragonite and clays, occasionally silty or sandy, with several subordinate sand or gypsum layers. Sulfur concretions are quite common in this unit. The White Cliff stands erosion somewhat better, with its lower part made of varves capped by a rather hard, 1–2 m thick gypsum layer. The contact between the two beds maybe somewhat unconformable. Close to the lake's shores, especially where wadis are present, the sequence becomes more detritic. This is particularly apparent in the northern Arava, where a vast area was drained into the Lisan Lake through a rather narrow outlet (Begin et al. 1980, Sneh 1982). Close to the shores, in numerous localities algal tufas were formed, probably where some freshwater influx entered the otherwise saline

Figure 11.2.8. Marginal facies of the Lisan Formation in the central Jordan Valley, showing the conglomeratic Hamarmar Member (H) at the base, overlain by the aragonitic Ami'az (A) and the gravely Fatza'el (F).

Figure 11.2.9. Sandy–gravely Iddan Member overlies the Ami'az at the northeastern Arava, some 20 km south of the Dead Sea.

lake. These were described by many, and studied in detail by Buchbinder et al. (1974).

The Fatza'el Member overlies the White Cliff Bed conformably at Deir Shaman but unconformably in many other localities. It is chiefly clastic, consisting of fine clay, silt and sand in the center of the basin, grading to conglomerates intermingled with red clayey loams at the basin's margins. This unit covers areas much more restricted than the underlying Ami'az (which, incidentally, is still regarded by many as the only "true" Lisan), and is found mainly in the central and southern Jordan Valley. The Fatza'el was referred to as the "Hasa Formation" by Vita-Finzi (1966), the "Unnamed Clastic Unit" in Begin et al. (1974), or the "Damiya Formation" by Abed (1985), among others.

The Iddan Member is named after Nahal Iddan, in the northern Arava, a region where its outcrops and the type section are located. It overlies the Hazeva Formation over an erosive surface, its thickness being in the order of tens of meters. The type section is 23 m thick. Eyal subdivided the Iddan into two members, a lower, mainly marly, 8 m thick, and an upper, chiefly sandy, which reaches 15 m. The lower member comprises alternations of white and gray marls, calcareous sandstone and sandy marl, quite loose, topped by 3 m of somewhat harder well-stratified white marl, and is actually not much different than the Ami'az. The Sandy Member commences with a hard layer of cross-bedded calcareous sandstone, while the rest is made of gray soft sand and grit, with several marl and gravel horizons. The latter may occasionally contain magmatic components derived from Transjordan (Bentor & Vroman 1957). No fossils were ever found. Only this "Sandy Member" is here regarded as the Iddan Member.

The Iddan was previously considered an early stage of the Lisan Formation (Bentor & Vroman 1957). Based on its different and distinct sandy lithology, Eyal defined it as an independent unit, probably considerably older than the Lisan. Horowitz & Horowitz (1990) erroneously suggested, based on the sandy characteristics, a possible correlation of the Iddan and Melekh Sedom Sands. Eyal apparently was led by similar considerations when recognizing 302 m of his Iddan in the Hazeva 5 borehole, near the settlement of Hazeva, a sequence which at least its major part indeed constitutes the Melekh Sedom Sands.

Fossils are quite rare in the Lisan Formation, except for diatoms, indicative of salinity and environment, and numerous pollen grains. Begin et al. (1974) distinguished three euryhaline diatom zones within the Ami'az Member, while the Fatza'el is characterized by freshwater flora. Freshwater diatoms are also more common in the entire sequence, both in the northern and southern reaches of the lake (Begin et al. 1980). Pollen, which assign the Lisan to Palynozone QIX, are discussed in detail in Chapter 6. Freshwater mollusks and ostracodes are reported by Bender (1974a, p. 95) from the lower part of the Lisan Formation, which seems, according to his lithological description, to correspond to the Hamarmar, south of the Damiya bridge. Various authors reported rare occurrences of mollusks, particularly *Melanopsis*, from the northern exposures of the Lisan, in the central Jordan Valley. Rosendahl et al. (1997) report a find of wild boar jaws (*Sus* sp.) from the Lisan beds just north of the Dead Sea.

It seems that the lower Hamarmar Member represents the phase of expansion in the initial stages of Lake Lisan, while the upper, the Fatza'el, represents the final shrinkage of the lake, due to structural disturbances creating the deep northern Dead Sea basin and Lake Kinneret some 18,000–16,000 years ago, followed by a drying up of the regional climate (Horowitz 1979, p. 151). The middle, more conspicuous Ami'az Member, was laid down in a rather wide saline lake, saltier in its central parts, becoming fresher both north and southward (Begin et al. 1980).

Numerous geochemical and isotopic studies were carried out, particularly on the Ami'az Member sediments (for example, Stein et al. 1997, among many others), which resulted in detailed information concerning changing salinities, the provenance of water, etc. Radiogenic datings, especially radiocarbon and uranium series based, are again mainly concentrated on the Ami'az Member, whose chemical deposits are suitable for these methods. These studies commenced in the 1960s, when Neev & Emery (1967) gave a span of 70–18 Ka for the sequence. With only minor refinements, almost all investigators arrived at similar figures (Kaufman 1971; Vogel, in Horowitz 1979, p. 151; Schramm et al. 1997; Macumber & Edwards 1997; among others). Thus, the Hamarmar and Fatza'el members are somewhat older and younger, respectively, with the termination of the Lisan attributed to the transition from the Pleistocene to the Holocene, some 11 Ka ago.

No artifacts were found within the lacustrine sediments of the Lisan Formation. However marginal, interfingering fluviatile gravels yielded a sequence from the Middle Paleolithic, found more or less at the base of the section, up to the Epipaleolithic at the top, Fatza'el Member, which is quite rich in sites. Neolithic artifacts and sites are found on the surface of the Fatza'el at many localities (Bar-Yosef et al. 1974, Bar-Yosef 1987, Horowitz 1988, 1989a, Macumber & Edwards 1997).

11.2.2.8 *Tabgha Formation*

Author: Horowitz (1971).

Tabgha Formation was defined in a borehole drilled underwater near a locality of that name, at the northwestern corner of Lake Kinneret. The Formation encompasses the sedimentary sequence laid down in Lake Kinneret since it acquired its present shape some 18 Ka ago (Fig. 11.2.10). As such, its lower part corresponds to the Fatza'el Member of the Lisan Formation. The Tabgha overlies conformably the Ami'az Member in the southern parts of Lake Kinneret, where the latter was deposited. Elsewhere it rests unconformably over older rocks. The Formation is known from numerous shallow boreholes in Lake Kinneret and also from several outcrops around the lake, particularly in the area where parts of the sequence were deposited in times of greater extension of the lake, under more humid climates.

The type section comprises 22 m of loose dark-brown to black clays, with subordinate sand size grains of quartz, chert, basalt, limestone and dolomite. The unit

Figure 11.2.10. Tabgha Formation on the shore of Lake Kinneret.

thickens, up to 30 m, at the lake's center (Ben-Avraham in Horowitz 1979, p. 145), but thins toward the margins, grading into gravel beds and soils. Bones of *Hippopotamus, Bos* and other vertebrates were recovered from exposures at the Buteiha; prehistoric sites were discovered around Lake Kinneret in littoral facies of the unit; and the rather rich mollusk assemblages, found mainly in the littoral facies, are those living today in the lake. The more common are *Melanopsis praemorsa, Theodoxus jordani, Bulimus hawaderiana, Melanoides tuberculata, Unio terminalis, U. semirugatus* and *Corbicula fluminalis* (Tchernov 1973). Pollen assemblages assign the lower sector of the Tabgha to the upper part of Palynozone QIX, while the rest is Palynozone QX of Holocene age.

11.2.2.9 *Rehov Formation*

Author: Neev (1967), "Bet She'an Lake Sediments".

The term "Rehov Formation" is suggested here to avoid confusion with the better-known Bet She'an Travertine, Rehov being a settlement on top of these deposits (Fig. 11.2.11), south of the city of Bet She'an. Shaliv et al. (1991) view the unit as an upper "Marl Member" of the Bet She'an Travertine. The differences in lithology, environment of deposition and the hiatus between the two units, seem to justify the separate reference.

Figure 11.2.11. Rehov Formation in Bet She'an Valley.

A suite of shallow lake sediments is described by Neev from the area in and around Bet She'an, covering some 100 km^2 to the south and east of the city. These rest over a paleosol, which in turn covers the Bet She'an Formation travertines. The top is exposed, or covered by a thin soil. The sequence comprises up to 30 m of white-brownish, occasionally black clays and marls, with very abundant pisolites and *Melanopsis* shells, which yielded radiocarbon ages in the range of 5,000 years. The sediments are rich in Chalcolithic potsherds and flint implements, while Early Bronze settlements are located on the surface.

The Bet She'an Lake existed during the humid Atlantic period, at the mid-Holocene. It came to an abrupt end due to a structural disturbance some 4,500 years ago, which downfaulted its eastern side, forming today a conspicuous scarp (Horowitz 1979, p. 146). Koucky & Smith (1986) proposed that this lake existed also east of the Jordan River, a view negated by later datings (Macumber 1992).

11.2.3 Upper Dead Sea Group

This term was suggested in Neev & Emery (1967) for the sedimentary suite deposited in the Dead Sea basin, since it became a terminal inland base level. The Group encompasses in places thousands of meters of sediments, accumulated due to the considerable subsidence of this region (Horowitz 1989b, 1996a, 1997).

Most of the sequence is known only from boreholes drilled in the deeper parts of the basin (Fig. 6.1.3); the margins, being continuously subject to erosion, display only very poor relics of correlative rocks. The only exception is the late Pleistocene Lisan Formation, discussed above, which has not yet entirely disappeared due to erosion, given the short time since its deposition. The drilled sequences were subject to detailed palynostratigraphic studies, and a thorough discussion of the Upper Dead Sea Group is given in Chapter 6.

The Lisan is conformably overlain in the Dead Sea itself by a suite of sediments deposited since the lake was formed some 15–18 Ka ago. This suite was termed the "Unnamed post Lisan sediments" by Neev & Emery (1967) and by Horowitz (1979, p. 151), stratigraphically equal to the Tabgha Formation of Lake Kinneret. Short, humid climate phases since the termination of Lake Lisan caused expansions of the Dead Sea, depositing lacustrine sediments, separated by rocksalt crystallized under drier conditions. The layers laid down during the middle Holocene Atlantic stage and somewhat later, radiocarbon dated, are very abundant everywhere (Gardosh 1987, Yechieli et al. 1993), and were informally termed by Yechieli the Ze'elim Formation (Fig. 11.3.1). The extremely dry phase terminating QIX times, corresponding to oxygen isotope Stage 1, caused drying up of the

Figure 11.3.1. Ze'elim Formation (the lower terrace) lake sediments fill erosion channels in the Lisan Formation (white cliffs above the terrace), at the outlet of Nahal Darga to the Dead Sea.

extended Dead Sea, as seen by the deposition of a rocksalt layer underlying the Ze'elim Formation some 13–11 Ka ago (Y. Yechieli, Geological Survey of Israel 2000, pers. comm.).

Another stratigraphically equivalent occurrence is south of the Dead Sea, always east of Nahal Ha'Arava which drains the northern Arava to the Dead Sea, where widely distributed sand dunes (Fig. 11.3.2) cover the Lisan Formation (Sneh et al. 1998a). The sand originates from several sources: the numerous Paleozoic–Mesozoic exposures of Nubian Sandstone east of the Arava on the western flanks of the Transjordanian Plateau, the Hazeva Formation, Melekh Sedom Sands and the Iddan Member in the Arava, redistributed by the winds until the present day.

Limited exposures of lake sediments preceding the Lisan, termed the Sayif Formation (previously spellt "Seif") (Sneh 1982), uranium series dated to an interval spanning 275–150 Ka ago (Livnat & Kronfeld 1985), are known from the northern Arava, down to the area just north of the watershed at Gav Ha'Arava (Livnat & Kronfeld 1990). The sequence, several meters thick, mainly comprises lacustrine limestones and marls, truncating unconformably a variety of older formations and overlain disconformably by the Lisan Lake sediments. In places they grade laterally to spring deposits, the Sayif Travertine, or to gravels of the Upper Dome Country Terraces. The age corresponds, according to pollen (Horowitz 1987b, Weinstein-Evron 1987) to Palynozone QVII.

Figure 11.3.2. Sand dunes of the northeastern Arava.

11.2.4 Lakeshore and wadi terraces

Quaternary climatic changes, of alternating dry and wet periods, occasionally helped by structural disturbances (Horowitz 1987a, 1992a, Chapter 9, 1997), caused expansions and contractions of lakes occupying the Jordan Rift Valley basins, changing also deposition and erosion styles in the wadis and rivers leading to those depressions. Several of the expanded lakes filled the depressions up to the bordering flanks, cutting terraces and leaving sedimentary and morphological evidence of their presence. However, the extensive subsidence of the basins' floors, causing the removal of considerable amounts of rocks by erosion, wiped out most of the evidence, except for the youngest, Lake Lisan, whose terraces (Fig. 11.4.1) are particularly conspicuous around the Dead Sea (Vita-Finzi 1964; Bowman 1971; Bender 1974a, p. 95).

Thus, lakeshore sediments and terraces are preserved only in areas where subsidence and erosion are minimal, such as the region of Marma Feiyad separating the central and southern Jordan Valley (Fig. 3.5.3), or the northern Arava, two regions where Miocene sediments make up the valley floors. The relic terraces of the former region (Bender 1974a, p. 94; Horowitz 1979, p. 148) had never been dated, since no clear succession exists in any one locality, and neither fossils nor artifacts have ever been found in their very limited exposures. The northern Arava,

Figure 11.4.1. Terraces left by the retreating Lisan Lake, facing the Dead Sea northeast of the Lisan Peninsula.

which is considerably larger and suffered hardly any subsidence since the end of the Miocene (Horowitz 1992a, p. 328), presents a fine sequence of fluvio-lacustrine sediments and terraces (Figs 11.4.3 and 11.4.4), testifying to the existence of a series of ancient expanded lakes.

The terraces of the northern Arava are mainly cut into the Pliocene Arava Formation (Sneh 1982, Avni 1998). However, contrary to what Sneh and Avni proposed, they are considered here as a separate stratigraphic entity (see also Horowitz 1979, p. 147) and not part of the Arava Formation as suggested by Sneh, or the "Arava Lake" proposed by Avni, for which he joined together all occurrences to form a single large lake, regardless of their different ages. These terraces, built mostly of gravel, occasionally grading to lacustrine limestones or marls, form a succession of four levels (names in parentheses are those given in Horowitz 1979, p. 147). The highest, apparently most ancient, is termed by Sneh the "Shezaf Hills Conglomerate" (4th Lakeshore Terrace). It is cut by the "Upper Dome Country Terraces" (3rd Lakeshore Terrace), which are cut in turn by the "Lower Dome Country Terraces" (Seif Lakeshore Terrace). These are subsequently incised by terraces of Lake Lisan (Lisan Lakeshore Terraces). The two highest terraces are chiefly of fluviatile origin, with no traces left of their corresponding lake deposits, but their connection with expanded lakes could be seen nearby in sequences penetrated by boreholes, (Horowitz & Horowitz 1990). The two lower terraces are intimately connected with lacustrine chalks. The upper of these relates to the Sayif

Figure 11.4.2. "Uppermost Terrace" at the headwaters of Wadi Zarqa, near Sukhne village.

Figure 11.4.3. The Shezaf Hills Conglomerate forms the highest surface on which the water tank is located, to the right (bold arrow). It is cut and filled up here by the Lower Dome Country Terrace, on which a solitary *Acacia* grows, to the left (fine arrow). Near the settlement of Hazeva.

Figure 11.4.4. Upper Dome Country Terraces at Nahal Gidron just west of the northern Arava, cut and filled up by the Lower Dome Country Terraces.

(or "Seif") Formation (Horowitz 1979, p. 147; Sneh 1982; Avni 1998), while the youngest connects with the Lisan (Sneh 1982). This set of four terraces has been correlated (Horowitz 1979, p.173) with the four known expansion phases of lakes in the Jordan Valley since it became an internal drainage system, and thus broadly corresponds to Palynozones QIII, QV, QVII and QIX.

Wadi terraces suffer a similar fate as lakeshore relics. They are hardly preserved close to the basins, where erosion is prominent, again except for the youngest ones, connected with the Lisan or even younger (Picard 1932; Vita-Finzi 1964; 1966; Horowitz 1979, p. 125; Sneh 1982; and others). The better exposures are therefore far away from the base level, upstream along the longer, older wadis. These localities, such as Nahal Zin at its head in the central Negev, present another difficulty, namely that the wadi length is a function of age, and so usually only the later stages could be traced there (Goldberg 1986, Plakht 1996, Horowitz 1996b). Consequently, only very few places where the entire sequence is preserved are known west of the Jordan, such as Wadi Hareitun southeast of Bethlehem (Neuville 1951; Horowitz 1979, p. 121, Fig. 9.6.2), or several of the longer wadis leading to the Dead Sea through the northern Arava, like Nahal Hiyyon (Ginat 1997). To the east, sequences of terraces occur in most larger (and older) wadis in Transjordan. There, unfortunately only the suite at the headwaters of Wadi Zarqa (Figs 11.4.2, 11.4.5 and 11.4.6) was studied in detail (Baubron et al. 1985, Besançon & Hours 1985).

Figure 11.4.5. Dauqara Formation at the headwaters of Wadi Zarqa, underlying the village of that name.

Figure 11.4.6. The Bire Formation (B), partly covered by a younger basalt flow, forms the high terrace under the building. It is cut and filled up by the lower terraces of the Khirbet Samra (K) and Sukhne (S) formations, seen in the foreground. Near the village of Bire, at the headwaters of Wadi Zarqa.

The climatic mechanism of forming wadi terraces and benches in the Near East was summarized in Horowitz (1996b). Wetter climates, typified by gentle winter rains, lack of thunderstorms and some summer rains, cause the accumulation of alluvium and colluvium, helped by the rich vegetation, resulting in the formation of moderate slopes and the silting up of previous channels; while drier climates, characterized by poor plant cover and winter thunderstorms, are responsible for erosion and down cutting, forming steep slopes and deep ravines. Thus, alternations of wet and dry climates, typical of the Quaternary in this part of the world, are recorded in wadis as sets of terraces. Usually the lower, younger terraces are made of gravel and colluvium, while the upper, older ones, went through extended erosion which removed the sediments, leaving a bench cut into the bedrock as evidence (Horowitz 1979, p. 121). Occasionally, however, sediments are found also in connection with upper terraces, as at Wadi Zarqa (see below).

Where fully developed, the sequence comprises four principal stages of such terraces or benches, formed during the Levantine stage (Fig. 9.6.2). Being of climatic origin, this set of four terraces is known from the entire Levant (Besançon & Sanlaville 1984). It is usually quite difficult to correlate sets of terraces, since local structural disturbances may add or subtract one or more. One of the more instructive localities, where a complete set of such terraces with their sediments was preserved and studied, is at the confluence of Wadi Dhuleil and Wadi Zarqa on the Transjordanian highlands some 20 km north of Amman, 40–50 km east of the Jordan Valley.

This sequence, studied by Baubron et al. (1985) and Besançon & Hours (1985), comprises four formations: the older is the "Uppermost Terrace", apparently not given a specific name since no artifacts were found within its alluvial sediments. It is cut and filled by the Dauqara Formation, yielding Developed Olduvan artifacts (Parenti et al. 1997). The Dauqara is cut and filled by the Bire Formation (not to be confused with "Bira"), with Middle and Upper Acheulian implements, which is again cut and filled by the Khirbet Samra Formation (not to be confused with "Samra"), containing Middle Paleolithic flints. This one is also cut and filled, by the Sukhne Formation, on which opinions seem to vary. While Baubron et al. consider it of anthropogenic origin with Bronze Age remains, Besançon & Hours maintain that it contains Kebaran cultures, topped by Pre-Pottery Neolithic structures.

Terraces in Nahal Hiyyon (Ginat 1997) were designated as stages Q1 through Q4 in the development of the wadi, Q1 being the oldest, overlying the Zehiha Formation. Another suite of terraces is reported in Avni (1998) for the area of the northern Arava, also designated as Q1 through Q3. However, it seems that Avni's Q1 is younger than Ginat's, since it cuts much deeper into the Zehiha Formation. Rich assemblages of Upper Acheulian artifacts are found on the surface of the Q2 Terrace in Nahal Hiyyon (Ginat 1997). It is impossible, in the present state of knowledge, to correlate the terraces of Avni and Ginat with those of Sneh, the latter being quite well connected with the Dead Sea sequence.

To the north such complete sequences are quite rare, but occasionally at least parts of them could be observed and dated, such as in Nahal Dishon leading to the Hula (Horowitz 1975b), or Nahal Amud draining to Lake Kinneret (Kafri & Heimann 1994). Typically, the lowermost, youngest two terraces are quite rich in artifacts all along the Rift Valley, the older yielding Upper Acheulian assemblages, the younger Middle Paleolithic through Epipaleolithic (Horowitz 1996b).

11.2.5 Lakes of the Dead Sea system outside the Rift

Wadis and rivers leading to the Dead Sea occasionally but rarely developed small lakes, as a combined result of a temporary blockage of their channel and a wetter climate. One such lake, which deposited some 2 m of sediments of presumably "Pleistocene" age, is reported by Bentor (1946) from a well dug near Yavne'el several kilometers west of the central Jordan Valley. This occurrence was unfortunately, despite my efforts, never found or observed again, so no further details are known.

11.2.5.1 *Zehiha Formation*

Author: Ginat et al. (1996, detailed in Ginat 1997).

The best-studied "external" lake sediments are from the southern Negev, exposed at Nahal Zihor (Ginat et al. 1996, Ginat 1997), a wadi drained to the Dead Sea through the northern Arava, a distance of some 90–100 km. The Formation is named after the nearby Zehiha (extremely arid) hills. The Formation rests unconformably over the Arava and older formations, but in places over a caliche layer comprising the base of what is termed by Ginat the "Red Unit" (Fig. 11.5.1), or over the lower part of this Red Unit. The upper part of the Red Unit is the lateral, fluviatile facies of the lake sediments. The Q1 Terrace is developed on a gravel bed at the top of this unit.

The thickness of the lacustrine sections varies up to some 20 m, covering an area of approximately 20 km^2. This is the only well-defined locality of the Zehiha (Fig. 11.5.2), but other occurrences, somewhat more detritic and possibly correlative, are reported closer to the Arava, notably at Biq'at Demama (silent valley) some 10 km north of Gav Ha'Arava, where a section of 45 m is described in Ginat (1997). Another occurrence of possibly correlative fluvio-lacustrine sediments is from Nahal Hiyyon, at its confluence with the Arava (Ginat & Zilberman 1996).

The type section in Nahal Zihor comprises 10 m of alternating white and green limestones, the latter detritic, with several thin highly organic black clay beds. The

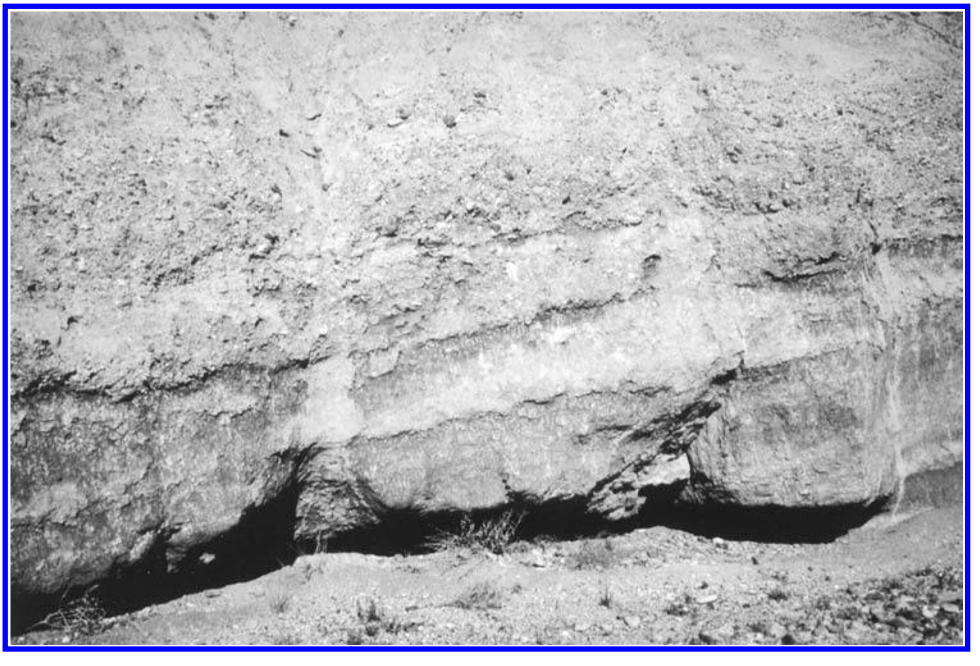


Figure 11.5.1. The "Red Unit" paleosol in Nahal Zihor, southern Negev, the only outcrop of sediments apparently representing Palynozone QII. Photo courtesy of H. Ginat.

Figure 11.5.2. Zehiha Formation lake sediments in Nahal Zihor, southern Negev. Photo courtesy of H. Ginat.

sequence is made of three such cycles, separated by minor unconformities. It grades laterally into a fluviatile “Red Unit”, made principally of sands and gravel, itself grading laterally into a suite of paleosols.

The Zehiha Formation is very rich in freshwater fossils, such as mollusks, ostracodes, fish and charophytes. Typical freshwater ostracodes are *Candona angulata, Cypridopsis* sp. and *Cypris pubera* (Ginat et al. 1996). These ostracodes are also found in the Erk el Ahmar and Ubeidiya formations (Rosenfeld et al. 1981). Of the variety of mollusks found in the Zehiha Formation, *Melanoides dadiana* and *M. jordanica* (Tchernov, in Ginat 1997) are characteristic of the Erk el Ahmar Formation in the central Jordan Valley (Schütt & Ortal 1993), while the rest occur throughout the entire Quaternary sequence. Pollen grains have also been recovered, indicating Palynozone QIII for the lake sediments, QII for the Red Unit where it underlies the lake sediments (Horowitz, in Ginat 1997). Avni (1998) attempted to define a gravely sequence further north as a correlative of the Zehiha. However, this bed constitutes part of the HaMeshar Formation of QI Palynozone age (Garfunkel & Horowitz 1966), which is older than Ginat’s Zehiha lake sediments (Horowitz 1992a, p. 363). Large, crude, Lower Paleolithic artifacts and flakes (Ginat & Saragusti 1996) were found in intimate connection with the Zehiha Formation lakeshore sediments, but it is doubtful if this is their point of origin (I. Saragusti, Department of Prehistory, the Hebrew University of Jerusalem 1998, pers. comm.).

11.2.6 Travertines

Travertines are hardly deposited at present by the numerous springs of the Jordan Valley, except for some very restricted occurrences (Fig. 11.6.1) mostly in the Dead Sea and northern Arava region (Buchbinder et al. 1974, Eidelman 1979, Khoury et al. 1984, Livnat & Kronfeld 1985, Enmar et al. 1998a,b, Enmar & Heimann 1999). This is attributed by Heimann & Sass (1989), who studied fossil travertines in the Hula Valley, to differences in the paleogeographic setting as compared with past times, when large veneers of travertines were formed. Heimann & Sass view water flow in the past as having been mostly sluggish, forming a widespread sheet of shallow ponds, in which the combined lush vegetation and relatively long residence time of the water in the area, led to increased efficiency of calcium carbonate precipitation. Today, in contrast, water flows rapidly in gorges, precipitating only a small fraction of the dissolved carbonate load. The authors blame the paleogeographic changes on late Quaternary rejuvenation of faulting activity. Such rejuvenation is not confined only to the Hula (see Chapter 9), but occurred all along the Jordan Rift Valley (Horowitz 1989b), explaining the paucity of recent travertine deposits.

Several isolated travertine occurrences, such as the exposure in Nahal Zihor (Ginat 1997) and others in the northern Arava or on the eastern side of the Jordan

Figure 11.6.1. Travertine formed recently by the hot springs at Hamamat Ma'in, east of the Dead Sea.

Valley, could not be accurately dated either because they do not interfinger with any other datable beds, do not occur in any sequence, or are beyond uranium series dating limits. Others are stratigraphically better connected, as is discussed below from the north southward. A distinction is made here between two types of occurrences, those actually exposed within the limits of the Jordan Valley and others which are widespread in numerous wadis leading to the Rift. The first group is usually quite well correlated stratigraphically by interfingering with other formations, but the other could only be radiometrically dated, or else by artifact contents, with all the problems involved with such "chronostratigraphic" assignment. Another group of unidentified travertines is very common east of the Jordan River, lining almost the entire stretch from Lake Kinneret down to the northern Arava (Fig. 11.6.2), for which no stratigraphic assignments are available since, save for one or two exceptions, they were never studied in detail.

11.2.6.1 *Kefar Yuval Travertine*

Author: Picard (1963).

This is named after a settlement in the northern Hula Valley, where extensive outcrops are present (Fig. 11.6.3). The Travertine is unconformably sandwiched

Figure 11.6.2. One of the numerous yet unaccounted for travertines lining the eastern flank of the Jordan Valley. Near Damiya.

Figure 11.6.3. Kefar Yuval Travertine, at the northern end of the Hula Valley.

between volcanic flows, overlying the Hasbani (as redefined in Heimann 1985) and Dalwe basalts, dated at around one million years (Heimann & Sass 1989), underlying the Ma'yan Barukh Basalt, previously termed "Hasbani" (Horowitz 1979, p. 157), dated at 73 ± 14 Ka (Siedner & Horowitz 1974), or some red, clayey paleosols. In places where the younger Dan Travertine directly overlies the Kefar Yuval, the two are occasionally quite difficult to differentiate. Heimann & Sass maintain that Kefar Yuval is also overlain by both Dalwe and Hasbani basalts, which seems questionable, considering the numerous Upper Acheulian artifacts (Bar-Yosef 1994) found in intimate connection with the Travertine. Laterally, the spring deposits grade to clastics, designated by Heimann (1985) the "Si'on Gravel".

The Kefar Yuval covers considerable areas, approximately 18 km^2, in the northern Hula Valley, and its thickness varies from 2 up to 30 m. The carbonate rocks are porous to very porous and soft, occasionally hard, massive, with rare traces of bedding, in most localities very rich in well preserved plant molds. Stromatolite-like structures are quite common in places. Fossil plants recovered from the layers (Danin & Lev-Ari, in Heimann & Sass 1989) include leaves of the trees *Quercus ithaburensis*, *Platanus orientalis*, *Celtis australis*, *Salix acmophylla*, *Ficus carica* and *Populus*, all of which grow even today in the vicinity. Water plants are very abundant, mainly reed and cattails, *Arundo donax* and *Phragmites communis*. Mollusk shells, usually *Melanopsis*, are very rare, maybe due to dissolution of their aragonitic shells (Heimann & Sass 1989). Several thousand Upper Acheulian artifacts were recovered from the paleosol cover of the Kefar Yuval Travertine (Stekelis & Gilead 1966). Some of these are coated with travertine (Schwarcz et al. 1980), which indicates an intimate connection of habitation and springs.

Attempts at uranium series dating of the Kefar Yuval Travertine, including also the artifact coating, indicated ages in excess of 350 Ka, which seems too old, resulting most probably from contamination (Schwarcz et al. 1980). Heimann & Sass assumed that the travertine began its accumulation about one million years ago and, together with the overlying Dan Travertine, continued until some 25 Ka ago, their deposition ceasing due to structural disturbances that changed the paleogeography. Horowitz (1979, p. 165) gave a much narrower interval for the Kefar Yuval, broadly correlating it with the Benot Ya'aqov Formation, of Palynozone QVII.

11.2.6.2 *Dan Travertine*

Author: Picard (1963), "Younger Travertines"; amended by Horowitz (1973).

This is named after a settlement in the northern Hula Valley, where extensive outcrops are exposed. The Dan overlies the Kefar Yuval Travertine, but the nature of the contact is not clear. It seems unconformable, but in several localities the latter grades upward into the former and the two are quite difficult to tell apart (Heimann & Sass 1989). In places the two are separated (Fig. 11.6.4) by a flow of

Figure 11.6.4. Dan Travertine, overlying the Ma'yan Barukh Basalt, at the northern end of the Hula Valley.

the Ma'yan Barukh Basalt (formerly termed "Hasbani"), with K–Ar ages in the order of 70 Ka (Siedner & Horowitz 1974). The thickness varies from a few meters up to 20, with the lithology very similar to the Kefar Yuval.

Plant fossils are numerous, mollusks are rare and again of the same types found in the underlying unit. Scarce artifacts of Middle through Upper Paleolithic cultures are occasionally found (Horowitz 1973), which prompted its correlation with the lower part of the Ashmura Formation, of Palynozone QIX times. Radiocarbon dates of the upper layers of the Dan Travertine (Kaufman, in Heimann 1985) are 33,100 ± 1,600 and 25,300 ± 800 years, confirming the correlation suggested above, as does the date for the underlying basalt.

11.2.6.3 *Bet She'an Formation*

Author: Picard (1929), "Kalksinter von Beisan"; amended by Shaliv et al. (1991), who adopted the term "Bet She'an Travertine" suggested in Horowitz (1979, p. 166), but included in it also the Rehov Formation which, for reasons explained above (see Section 11.2.2.7), is not accepted here.

The name, after the city built on this Formation (Fig. 11.6.5), was given by Picard to a suite of spring deposits exposed in and around Bet She'an. The base of the travertines is nowhere exposed, except where they unconformably overlie

Figure 11.6.5. Bet She'an Formation, near this city.

considerably older rocks. It is overlain by the lake sediments of the Rehov Formation (Horowitz 1979, p. 166), and occasionally the two are separated by a paleosol horizon. The travertines cover an area of some 100–150 km^2, comprising a sequence of up to about 60 m (Shaliv et al. 1991), with the uppermost 18 m exposed due to subsequent faulting.

It is made of yellowish-brown friable carbonates, massive with no discernible bedding, bearing numerous plant molds, algal stromatolites and mollusks, such as *Melanopsis* and *Melanoides tuberculata*. The upper 16 m were dated by both radiocarbon and uranium series (Kronfeld et al. 1988), indicating a continuous deposition between 41 and 22 Ka ago, broadly correlative to the Ami'az Member of the Lisan Formation, with which the Bet She'an most likely interfingers eastward. If the sequence indeed comprises the 60 m reported in Shaliv et al. (1991), then the base could well predate the Lisan Lake.

11.2.6.4 *Sayif Travertine*

Author: Sneh (1982).

Named after a wadi in the northern Arava (previously transliterated "Seif"), where the type exposures were designated (Fig. 11.6.6), some 30–35 km south of the

Figure 11.6.6. Sayif Travertine at Nahal Sayif, northern Arava, overlying the Hazeva Formation.

Dead Sea. For some reason the Sayif is termed the "Arava Travertine" by Enmar et al. (1998a), a change of name for which I cannot find any justification, since it seems that it overlies the Arava Formation rather than interfingers with it. The Travertine is unconformably sandwiched between a suite of older formations and the Lisan Formation sediments. It grades laterally in the general direction of the Dead Sea into the Sayif Formation lake deposits (Sneh 1982, Livnat & Kronfeld 1985, 1990), while in the opposite direction to the Upper Dome Country Terraces.

Small outcrops of the Sayif are known from the entire northern Arava, south to the watershed at Gav Ha'Arava (Livnat & Kronfeld 1990). The sequence never exceeds several meters, the rocks are mostly travertine, with some pebbly limestone and gravel, while locally a conglomerate occurs at the base. Occasionally fossil *Melanopsis* and *Melanoides* shells are present, together with numerous unidentified plant molds. Uranium series dates (Livnat & Kronfeld 1985, 1990) are within oxygen isotope Stage 7, while pollen assemblages testify to Palynozone QVII. Livnat & Kronfeld maintain that the travertines represent a more humid climate for Stage 7, which is also supported by pollen analyses by Weinstein-Evron (1987), who assigns the spectra to a humid phase within Stage 7. Horowitz (1987b) indicated that the Sayif Travertine was formed in a dry climate during QVII times, most probably due to exposure of the supporting aquifer by faulting, and the pollen reported by Weinstein-Evron (collected from the travertine itself) represent the local environment, not the regional. Horowitz's claim is based on the extremely high peak in hydrophil plant pollen in the neighboring Amazyahu 1 borehole, during the drier middle part of Palynozone QVII (Horowitz 1992a, p. 337).

11.2.6.5 *Mo'a Travertine*

Author: Enmar (1999).

The sequence is named after ruins in Nahal Omer, some 65 km south of the Dead Sea. It was previously described in Sneh (1982), who did not name it. Morphologically, the Mo'a cuts and fills channels in the underlying (Fig. 11.6.7) and morphologically higher Sayif Travertine and lake deposits, and is in turn cut and filled by sub-recent wadis and travertines. The unit is found in small, isolated outcrops all along the northern Arava, never exceeding a few meters in thickness. It merges laterally to lacustrine sediments of the Lisan Formation.

The usual plant and root molds and rare mollusks occur also in this Travertine, but are useless as age indicators. These include mainly reeds and cattails, and leaves of poplars, willows and maple. Uranium series datings (Livnat & Kronfeld 1985) are within the 60–120 Ka range. Pollen assemblages (Weinstein-Evron 1987, Horowitz 1987b) assign the Travertine to Palynozone QIX. Mousterian artifacts (Bar-Yosef, in Livnat & Kronfeld 1985) are reported from the top of the Mo'a Travertine.

Figure 11.6.7. Mo'a Travertine on the hilltop to the left. Sub-recent travertines line the wadi bed. Near En Yahav, northern Arava.

11.2.6.6 *Sub-recent travertines*

These occur in several localities along the Dead Sea and the northern Arava (Enmar et al. 1998a,b, Enmar 1999), but no other information besides the fact that they are younger than the Mo'a (Fig. 11.6.7), possibly of Holocene age, is available. The thickness is very limited, while composition of the sub-recent travertines testifies to evaporitic sedimentation conditions, indicating a decrease in the water output of the springs, probably corresponding to climatic deterioration during the Holocene.

11.2.6.7 *Travertines in wadis leading to the Jordan Valley*

There are numerous such occurrences on both sides of the Jordan Valley, more abundant to the east, most of which formed by relatively small springs, and have never been studied in much detail. The exceptions are those which contain artifacts that aroused the attention of prehistorians, so that part of them are dated, usually by uranium series.

One of the better known, but as yet undated, is the suite of travertines comprising the upper part of the Tabaqat Fahl Formation (Macumber 1992), where Upper Acheulian sites are embedded (Macumber & Edwards 1997), some 20–25 km south of Lake Kinneret on the eastern flank of the Rift near Pella. This travertine ("tufa" or "tuffaceous limestone" by the authors) forms the upper part of the Formation, while in the lower one Middle Acheulian sites are found. The Travertine underlies the Lisan Formation sediments, and thus in both stratigraphic position and artifact contents resembles very much the Kefar Yuval and Sayif Travertines. The Tabaqat Fahl spring deposits approach the Rift Valley, where they turn into lacustrine sediments, most probably correlative to Palynozone QVII.

Other travertines can be observed along major parts of the eastern Rift flanks, and are particularly common along the Dead Sea and the southern and central Jordan Valley segments (Fig. 11.6.2). They occur at various topographic levels, indicating several phases of stronger spring activity during the Quaternary. However, most of these occurrences have never been subject to detailed studies, let alone datings. On most of the geological maps they are not even marked as spring deposits. At any rate, just as today, the eastern flanks enjoyed more rain-water most of the time, as shown by the abundance and considerable extension of travertines.

Several stream-dissected fossil spring deposits are exposed along the walls of Wadi Fasayil (Fatza'el) some 28 km north of the Dead Sea, west of the River Jordan, interfingering with terrace gravels containing Middle Paleolithic (Levalloiso-Mousterian) artifacts. Two uranium series dates obtained from these travertines (Schwarcz et al. 1980) seem questionable: 136 ± 9 and 63 ± 4 Ka. The first was on a dense, impermeable calcitic travertine which should be reliable, but is considered too old for the artifacts; the second, which is in agreement with estimates for this culture, was obtained on a highly porous powdery sample, seemingly unreliable. No solution is proposed by the authors, but it seems that if the ages suggested by Valladas et al. (1988) for the Mousterian are accepted, the older age may be within the limits.

Schwarcz et al. (1979) uranium series dated several travertine occurrences upstream in Nahal Zin and its tributary Nahal Aqev, located some 50 km west of the Arava. An older assemblage of samples yielded ages in the range of 150–300 Ka, which is correlated by Livnat & Kronfeld (1985) with ages obtained for the Sayif spring and lake sediments in the northern Arava. Younger travertine occurrences, in which Mousterian artifacts are embedded, yielded dates around 50–40 Ka ago. These dates are somewhat younger than those obtained by Livnat & Kronfeld for the Mo'a Travertine, but are comparable to the lower part of the Lisan Formation. A sub-recent travertine in the same locality was dated at 11.8 ± 0.9 Ka.

11.2.7 Caves

Numerous caves occur in wadis leading to the Jordan Valley, and almost all contain various sediments and some accumulations of travertine. Most of these are small and were never inhabited, but several larger ones were used for human

Figure 11.7.1. Old, possibly early or middle Pleistocene cave sediments at the top of the Galilee hills, containing fossil bear bones. The cave itself was eroded due to the conspicuous Quaternary uplift.

shelter over long periods of time. The most famous are in Nahal Amud, a few kilometers west of Lake Kinneret, and in Wadi Hareitun near Bethlehem, less than 20 km west of the Dead Sea, both uranium series dated by Schwarcz et al. (1980). In addition some remains of older, now ruined caves, are preserved at higher elevations on the surrounding hills (Fig. 11.7.1). Factors controlling the exposure of and deposition in these caves are discussed above, in Section 11.1.

Dates obtained from the Zuttiyeh Cave in Nahal Amud (Fig. 11.7.2) suggest that the Acheulo-Yabrudian industry, accumulated there from approximately 150 Ka ago, was replaced by Mousterian at about 95 Ka. The "Ante-Neanderthal" skull fragments (see Chapter 2) were thought at first to predate 164 Ka, but are now considered a real Neanderthal, tied to the Acheulo-Yabrudian layers (Hovers et al. 1994). A single date from Umm Qatafa Cave (Fig. 12.1.6.3), Bed E1, yielded an age of 115 ± 18 Ka for the contained artifact assemblage of ovate and lanceolate hand axes, with a few Levallois-like flakes and other flake tools of Middle Paleolithic aspect. Both dates place the beds at Palynozone QVIII, which is represented hardly anywhere in and around the Jordan Valley besides these caves, and in boreholes.

Figure 11.7.2. Zuttiyeh Cave at Nahal Amud, west of Lake Kinneret.

11.2.8 Volcanics

11.2.8.1 *Bashan Group*

Author: Mor (1973); amended by Mor (1986).

This is named after the biblical appellation for southwestern Syria, a region largely covered by various volcanic formations of the Bashan Group. This term was coined by Mor to include the entire suite of basalts, scorias and tuffs, from the point they overly unconformably Pliocene sediments, until the present day (Fig. 11.8.1). The Group includes five formations, each denoting a distinct period of volcanic activity. From bottom to top these are the Cover Basalt (see above, Fig. 5.3.10), the Mechki Basalt (Fig. 5.4.1), the Ortal Formation (Mor 1986), the Golan Formation (Mor 1973, amended 1986) and the youngest, the Leja Formation (Mor 1986). Each of these formations comprises several members, representing closely related eruptions. The Cover Basalt and members of the other formations or their time correlative volcanics, except for the Leja, occur both on the Golan Plateau and in the Jordan Rift Valley.

The volcanics of the different eruptions are determined chiefly by their chronological (radiogenic ages) and morphological criteria. On the eastern flanks of the Jordan Valley, but west of the main Rift fault, the morphological characteristics are somewhat obscure, so Mor refers to the sequence as the "Undivided Bashan

Figure 11.8.1. Bashan Group: Ortal Formation flows at Nahal Orvim, just east of the Hula Valley, overlying Eocene rocks.

Group" at the Rift margins. The main body of volcanics younger than the Cover Basalt occurs on the Golan and in southwestern Syria, but some flows also reached the Jordan Valley, interfingering with lake and river sediments, or filling up river valleys. These are discussed in detail below, together with the rest of the formations limited to the Rift Valley (see Section 11.2.8).

The Bashan Group thickness, close to the northern Jordan Valley, reaches some 50–60 m east of Lake Kinneret, thickening considerably northward, assumed by Mor (1986) to comprise more than 1 km at the central Golan, gradually thinning again further north, and completely disappearing when approaching the elevated Mount Hermon structure. It seems possible however that the lower, deeper parts of the thicker sequence may indeed be older, such as the Lower Basalt, which is known from several localities east of Lake Kinneret (Michelson 1972). It seems that the Bashan volcanics filled up the large synclinorium separating the Ajlun and Hermon anticlinoria. Since the base of the Bashan is usually not exposed, it is impossible to calculate volumes of the lavas and pyroclastics. Petrographically, the Bashan volcanics (except for rare xenoliths) as well as all others in the region, both lavas and pyroclastics, are made of alkali olivine rocks.

Radiometrically, the base of the Cover Basalt in the Golan is given in Mor (1986) as 3.7 ± 0.23 Ma (but see discussion in Section 7.1.5). Volcanic activity in this region has continued ever since, almost continuously. One of the last eruptions of the Leja Formation was radiocarbon dated at 4,075 ± 160 years, burying a herd of goats, sheep and cattle (Dubertret 1955a). A more recent account comes from monks, who claimed to have witnessed basaltic eruptions in this area during the 17th century (Gushchenko 1979).

The Cover Basalt is known from many localities in the Jordan Valley. Younger members of the Bashan Group, which erupted on the Golan, occasionally reached the northern and central Jordan Valley, and are known from outcrops and boreholes (Heimann 1990). Examples are flows of the Yarmouk and Raqqad basalts (see below) in the central Jordan Valley; the Yarmouk corresponding time equivalents, the Yarda Basalt on the Korazim block and the Anafa flow in the northeastern Hula Valley, where the Raqqad correlative Sa'ar Basalt (Heimann 1985) occurs. Between these two phases, Heimann (1990) indicates almost continuous occurrences of Golan Formation flows in the northeastern Hula Valley, most of them confined to the Golan slopes, not extending far into the depression.

11.2.8.2 *Hasbani Basalt*

Author: Picard (1963); amended by Heimann (1985); and by Mor (1986).

The unit is named after the Hasbani River (Nahal Snir), in whose gorge is the best-exposed type section (Fig. 11.8.2). Picard termed "Hasbani" most of the volcanic

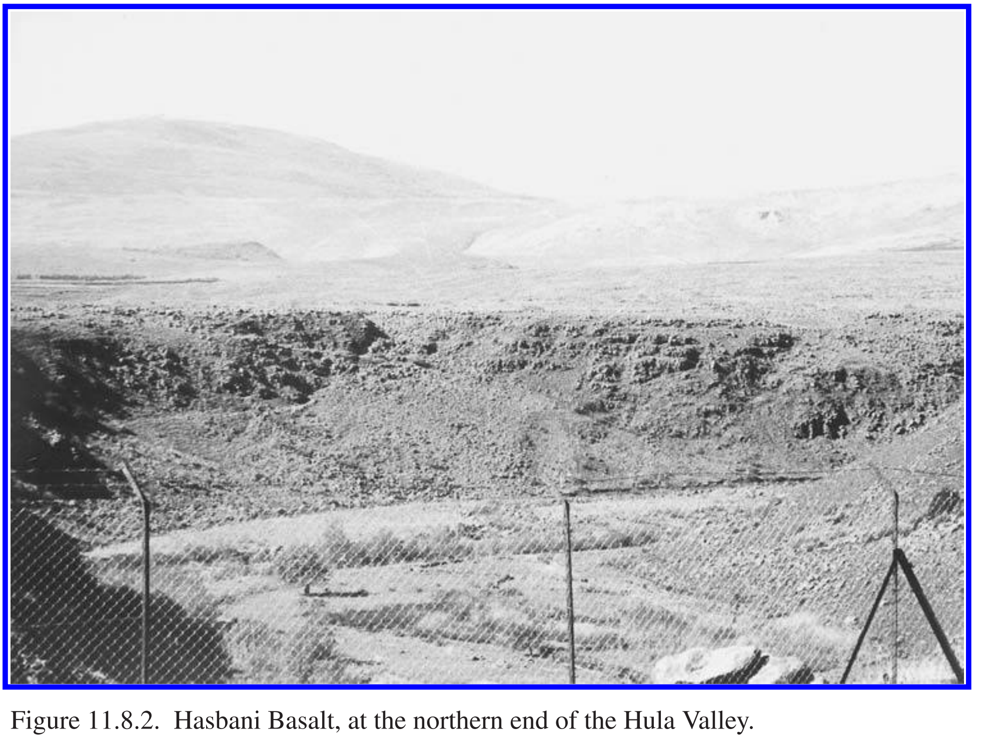


Figure 11.8.2. Hasbani Basalt, at the northern end of the Hula Valley.

occurrences in the northern Hula Valley and southern Lebanon. Heimann (1985) discriminated within Picard's definition an older phase for which he kept the name "Hasbani" and a younger one, which he termed the Ma'yan Barukh Basalt (below). Mor, following radiometric datings, termed the oldest flows the "Mechki Basalt" (see Section 5.4.1), but retained Picard's term for the middle flows. It seems that all three units are there and should accordingly be treated as separate volcano-stratigraphic units.

The Hasbani Basalt overlies unconformably a variety of older formations and is overlain, possibly also unconformably, by the Kefar Yuval Travertine and younger sediments and volcanics. Basalts in several boreholes drilled in the Hula Valley yielded ages similar to those of the Hasbani, in the range of 1.35–0.83 Ma (Mor 1986), and are therefore correlated with this unit (Heimann 1990). The type section for the Hasbani Basalt incorporates some 30 m comprising several flows of alkali olivine basalt, which cover an area of some 450 km^2 north of the Hula and in the Beqa'a. Several basalt flows which occur at the northwestern Golan escarpment and the northeastern Hula Valley, are termed by Heimann (1990) "Golan margins basalts", and yielded ages similar to the Hasbani in the range of 1.0–1.2 Ma. These outcrops should, in my opinion, be considered as part of the Hasbani Basalt, and not as a separate entity.

Equivalents of the Hasbani Basalt are reported from boreholes, Notera 3 at 550–900 m and Hula 1 at 180 m, which may also correspond to the Nahal Orvim section (Mor 1986) of the Dalwe Basalt on the western Golan escarpment.

11.2.8.3 *Yarda Basalt*

Author: Picard (1963); amended by Horowitz (1973).

The unit is named after Khirbet Yarda just south of the Hula Valley, where it was first described by Picard, who attributed it to the Cover Basalt. Schulman (1967) referred to the Yarda as the "Young basalts of Gadot". Siedner & Horowitz (1974), before realizing that the two belong to the same unit, termed the Yarda outcrop (Fig. 11.8.3) at Benot Ya'aqov bridge (after Picard 1963) the "Mishmar HaYarden Basalt", which was later corrected (Horowitz 1979, p. 155).

The Yarda comprises one or two alkali olivine basalt flows, overlying unconformably the Mishmar HaYarden and Gadot formations over an erosional relief. The top is usually exposed and eroded, but in several localities is covered by Holocene sediments of the Ashmura Formation. The flows do not attain more than several meters in thickness and are known from restricted areas in the northern sector of the Korazim block. The Yarda displays normal magnetic polarity and yielded a K–Ar age of 0.64 ± 0.12 Ma (Siedner & Horowitz 1974). Heimann (1990) calculated ages of 0.8 ± 0.13 and 0.9 ± 0.15 Ma for the Yarda, which seem somewhat incompatible with its normal magnetic polarity.

Figure 11.8.3. Tilted Yarda Basalt at Benot Ya'aqov just north of the bridge, overlying a peat layer of the Mishmar HaYarden Formation.

11.2.8.4 *Yarmouk Basalt*

Author: Nötling (1886), "Ez Zeijatin Lava"; amended by Picard (1932), "Yarmouk Basalt" (which also included the younger Raqqad Basalt), and by Michelson (1973).

The unit is named after the Yarmouk River gorge, where it appears as terraces lining both sides, up to 200 m above the present-day talweg (Fig. 11.8.4). The Basalt overlies gravel within the Yarmouk's course and the Erk el Ahmar and Ubeidiya formations when arriving at the central Jordan Valley, where it is overlain by the Naharayim Gravel and the Lisan Formation. Both lower and upper contacts are unconformable. The Basalt occurs in Naharayim, where the Yarmouk pours into the Jordan River, which gave it the unofficial name of "Naharayim Basalt". The thickness is some 25 m, comprising up to four flows.

Radiometric datings by Siedner & Horowitz (1974) and Siedner (in Horowitz 1979, p. 169) gave ages of 0.56–0.68 Ma, while Mor (1986) dated both at 0.79 ± 0.17 Ma, proving that the Yarmouk and Naharayim basalts are a single unit. Heimann (1990) dated ten samples collected from four flows, and arrived at an average age of 0.6 ± 0.15 Ma. The Yarmouk Basalt is known only from the gorge and its outlet to the Jordan Valley, but its source is unknown.

Figure 11.8.4. Yarmouk Basalt, forming the high terrace to the right (Y), at the outlet of the Yarmouk River to the central Jordan Valley. Further incision deepened the gorge, which was subsequently filled up with the lower-lying Raqqad Basalt (R), to the left.

11.2.8.5 *Raqqad Basalt*

Author: Nötling (1886), "Rukkad Lava"; amended by Michelson (1973).

The Raqqad Basalt is named after one of the main tributaries of the Yarmouk River. The type locality is in Wadi Raqqad, where several flows attaining 20–40 m of alkali olivine massive basalt filled the pre-existing valley, and now form terraces some 30–60 m above the present-day talweg. While most investigators saw the Basalt as a separate, distinct volcanic phase, Picard (1932) regarded both the Raqqad and Yarmouk as a single unit. The lava flowed through the Raqqad and Yarmouk gorges (Fig. 11.8.4) to the central Jordan Valley, where it separates the Naharayim Gravel from the younger Lisan Formation (Horowitz 1979, p. 158). The source is not known, but Mor (1989) suggested Tel Hader on the slopes of Mount Hermon as a possibility.

Based on stratigraphic considerations Horowitz suggested an age around 70 Ka for the Raqqad phase, which is quite widespread on the eastern slopes of the Jordan Valley. Mor (1986) gave the radiogenic age of Raqqad as 0.3 ± 0.11 Ma, while Mor (1989) measured ages around 0.4 Ma on a flow at the outlet of the

Yarmouk River, ages which may seem too old in view of the stratigraphic position of the Raqqad Basalt. The question is whether the Raqqad was erupting only during the time span separating the Naharayim and Lisan formations, as it seems from most outcrops, corresponding to Palynozone QVIII, in which case the radiogenic ages are too old, or if it replaces parts of the Naharayim, of QVII age, in which case its ages are within reason. I am inclined to accept the latter view.

Basalt flows which are similar in stratigraphic position to the Raqqad are described in Bender (1974a, p. 109) from the eastern slopes of the Jordan Rift down to the Dead Sea (Fig. 11.2.7), extruded from volcanic cones on the Transjordanian Plateau. These flowed through pre-existing valleys and when arriving at the Jordan Valley overlaid gravel beds in which Upper Acheulian artifacts were found (Huckriede 1966). The flows are in turn usually overlain by the Lisan Formation. Following Picard, Bender also regarded the Yarmouk and Raqqad basalts as a single unit, so that only those flows covering Acheulian bearing gravels are correlated with the Raqqad (Horowitz 1979, p. 158). These include the large volcanoes south and east of Rashadiya, Jebel el Hikkir, Jebel el Rudeisiya and Jebel el Qirana, towering almost 1,500 m above sea level. Craters, calderas and flows are sometimes very well preserved, such as in Tel Burma and Jebel Uneiza.

11.2.8.6 *Ma'yan Barukh Basalt*

Author: Heimann (1985).

This consists of the uppermost flow or two in the northern Hula Valley, termed the Ma'yan Barukh Basalt after a settlement in that region, where they crop out in a limited area (Fig. 11.6.4). These flows are stratigraphically in a similar position to the Raqqad Basalt, sandwiched between the Kefar Yuval Travertine bearing Acheulian artifacts, and the Dan Travertine equivalent in age to the Lisan Formation. The upper flows yielded (Siedner & Horowitz 1974) an age of 73 ± 14 Ka. These flows were regarded at the time as the Hasbani Basalt, which they directly overlie where the travertine was not deposited.

11.3 CHRONOSTRATIGRAPHY, CORRELATIONS

The extent and environments of deposition of the Levantine formations, which span approximately the last two million years, are chiefly affected by structural and hydrological conditions, both of which frequently change, with the latter considerably influenced by climate. Their stratigraphy is thus based on climatostratigraphic principles commonly used for the Quaternary all over the world. The Quaternary sequence of the Jordan Valley was subdivided into ten palynozones, each characterized by its particular pollen assemblages (Horowitz & Horowitz

1985, and see Chapter 6). The climatic trends represented by the spectra were correlated with those obtained from the oxygen isotopes curves of the oceans, thus corresponding to the global stratigraphic scheme for the Quaternary period (Horowitz & Derin 1987, Horowitz 1989c, Table 1.4.1 and Fig. 6.6.2).

Palynozone QI, of earliest Quaternary age, for which only hints but no definite indications of humans are available from the southern Levant, is the last period still affected by the previous Eritrean stage tectonics, and so is discussed in Chapter 7. Although no hints of humans are known for QII, it is included here for geological reasons, being the first to be controlled by the Levantine stage.

The earliest Quaternary (previously termed "Preglacial Pleistocene", Horowitz 1979, p. 6; considered by some the latest part of the Pliocene) Palynozone QII corresponds to the uppermost part of Foraminifera Zone N22, oxygen isotopic stages 73–63, approximately spanning the Olduvai, 1.99–1.82 Ma ago. All overlying Palynozones, QIII–QX, are correlative to Zone 23. Palynozone QIII corresponds to stages 65–49, 1.82–1.49 Ma ago; QIV to 49–38, 1.49–1.25 Ma; QV to 37–22, 1.25 Ma–790 Ka; QVI to 21–12, 790–425 Ka; QVII to 11–6, 425–130 Ka; QVIII to 5, 130–70 Ka; QIX to 4–2, 70–11 Ka; while QX relates to Stage 1, up to the present day. QIII–QV were deposited during the upper part of Matuyama, while the overlying palynozones were laid down throughout the Brunhes Chron.

Most outcrops of the Jordan Valley formations were subject to pollen analyses and these, together with comparisons of sequence stratigraphy and climatic characteritics, serve for connecting the exposures to the palynozones determined in boreholes, shown in Table 11.1. Numerous other fossils, particularly mollusks and vertebrates, were studied from prehistoric and archaeological excavations, some of which are of great help in the chronostratigraphy and reconstruction of past climates, landscapes and environments. These are detailed above, included with the descriptions of the Levantine formations. Occasionally, but rarely, rock units are correlated by their characteristic artifacts. Such correlations, whenever suggested, take into account the problems involved in applying this method.

11.3.1 Palynozone QII

Palynozone QII was defined as the Gonen Formation in the subsurface of the Hula basin. Sediments of this palynozone were never encountered in the central Jordan Valley, but make up considerable sequences in drillings sunk in the southern Dead Sea basin (Fig. 6.1.3). QII is considerably poorer in arboreal pollen compared to the underlying QI, typified by low shares of *Picea orientalis*, with some oaks. No volcanics are known to interfinger with QII. Spectra poor in AP, but dominated by *P. orientalis*, were obtained from a sample of red-beds (Fig. 11.5.1), the Red Unit of Ginat (1997) separating the HaMeshar from the overlying Zehiha Formation lake sediments (as designated by Ginat 1997, which is not in accord with what Avni 1998 called the "Zehiha", not accepted here). If a single pollen spectrum is accepted as significant enough, this is probably the only known outcrop of QII.

Table 11.1. Correlations of the exposed Levantine stage formations.

Age (Palynozone)	Hula Group	Jordan Group	Upper Dead Sea Group	Wadi terraces	Sites and cultures
QX	Mallaha Formation; Uppermost Ashmura Formation	Tabgha Formation; Rehov Formation	Ze'elim Formation; Sub-recent Travertines	"Lower Terrace"; Sukhne Formation	Neolithic through present day
QIX	Ashmura Formation; Dan Travertine	Tabgha Formation; Lisan Formation; Bet She'an Formation; "Knob Limestone" Travertines	Damiya Formation; Fatza'el Member; Iddan Member; Lisan Formation; Fasayil Travertine; Mo'a Travertine	Wadi Hammeh Conglomerate; Nahshon Conglomerate; Khirbet Samra Formation	Epipaleolithic; Upper Paleolithic; Middle Paleolithic
QVIII	Hulata Formation; Ma'yan Barukh Basalt	Raqqad Basalt		Lower beds of Zuttiyeh Cave	Middle Paleolithic; Acheulo-Yabrudian
QVII	Benot Ya'aqov Formation; Kefar Yuval Travertine	Naharayim Formation; Tabaqat Fahl Formation; Kufrinja Gravel	Sayif Travertine; Lower Dome Country Terraces	Baq'a Conglomerate; Bire Formation	Jisr Banat Yaqub; Umm Qatafa Cave; Middle and Upper Acheulian
QVI	Ayyelet HaShahar Formation; Yarda Basalt	Yarmouk Basalt			Gesher Benot Ya'aqov
QV	Mishmar HaYarden Formation	Ubeidiya Formation; upper Abu Habil Formation	Upper Dome Country Terraces	Dauqara Formation	Dauqara; Ubeidiya
QIV	Notera Formation; Hasbani Basalt				
QIII	Gadot and Hazor formations	Erk el Ahmar Formation; Lower Abu Habil Formation	Shezaf Hills Conglomerate; Zehiha Formation	"Uppermost Terrace"	Erk el Ahmar; lower Abu Habil
QII	Gonen Formation		"Red Unit"?		

Erosive features of QII times are very abundant on both flanks of the Jordan Valley, comprising the initial cutting phase of the previous QI peneplain (Fig. 9.6.2). They are much more difficult to identify within the valley, since in no place are QIII sediments known to overly QI rocks in outcrops. An erosion surface overlapped by the Gadot and Hazor formations in the southern Hula Valley seems correlative to the Gonen, but may well be older. The same consideration holds for the erosive surface underlying the highest Shezaf Hills Conglomerate terraces in the northern Arava (Sneh 1982), which cuts into the Pliocene Arava Formation.

11.3.2 Palynozone QIII

Palynozone QIII, typified by an abundance of arboreal pollen dominated by oaks, was designated primarily from the Gadot–Hazor formations of the Hula basin (Horowitz & Horowitz 1985). The palynozone was encountered in the three basins along the Rift (Fig. 6.1.3) and correlates both by its pollen spectra and sequence stratigraphy (Horowitz 1992a, p. 410) with the Erk el Ahmar Formation of the central Jordan Valley. Sequence stratigraphic considerations tie QIII also with the Shezaf Hills Conglomerate (Sneh 1982) of the northern Arava, and with the highest gentle slope of the highlands, where no sediments were left in almost all localities due to a subsequent long period of erosion (Fig. 9.6.2). The only place where such sediments were securely identified is at the confluence of Wadi Dhuleil and Wadi Zarqa (Baubron et al. 1985), termed the "Uppermost Terrace" (Fig. 11.4.2). Such terraces seem however to exist in numerous large wadis in Transjordan (personal observation).

Paleomagnetic studies of the Erk el Ahmar Formation sediments (Verosub & Tchernov 1991, Braun et al. 1991, Ron & Levy 2001) are in accord with its age as attributed by correlation with ocean oxygen-isotopes sequences (Fig. 6.6.2). Horowitz (1979, p. 141) suggested that the lower part of a sequence described in Bender (1974a, p. 94) at Abu Habil, east of the River Jordan some 25 km south of Nahal Tavor, may represent the southward extension of the Erk el Ahmar, a correlation doubted by Macumber & Edwards (1997) and by Copeland (1998), both considering the Abu Habil to be younger. The newly studied melanopsids from the Abu Habil (J. Heller, Department of Evolution, Systematics and Ecology, the Hebrew University of Jerusalem 1999, pers. comm.), which best resemble the assemblages from the Erk el Ahmar Formation, seem to further support my earlier assignment, which was only strengthened while recently visiting the site. Fossil mollusks recovered from the Zehiha Formation (for details see Section 11.2.5.1) as well as pollen grains, are typical also of the assemblages recovered from Erk el Ahmar, which safely assign the Zehiha to Palynozone QIII.

11.3.3 Palynozone QIV

QIV, defined in the subsurface of the Hula basin as the Notera Formation, bears pollen spectra indicating a poor vegetation and dry climate. It is encountered only in boreholes, also in the central Jordan Valley and the Dead Sea (Fig. 6.1.3). Its lateral exposed equivalent is an erosion surface separating the Gadot–Hazor from the overlying Mishmar HaYarden Formation in the Hula Valley, on which occasionally rare, reddish-gray paleosols were preserved. This surface is not clear in the central Jordan Valley, since in no place is a contact seen between the Erk el Ahmar and Ubeidiya formations (which is probably the reason why, before their fossils were studied in detail, the two had been considered a single unit). This erosion surface is again seen in the northern Arava, underlying the Upper Dome Country Terraces (Sneh 1982). Steep slopes, cutting into the gentle ones of QIII, characterize QIV in the highlands (Fig. 9.6.2), with no sediments.

Radiogenic datings of basalts interfingering with QIV sediments in the Notera 3 borehole, Hula basin (Heimann & Steinitz 1989, Heimann 1990), at a depth of 900 m, yielded an average age of 1.36 ± 0.22 Ma, which is in very good agreement with the ages obtained by correlation with the ocean sequence. Heimann correlated these volcanics with the Hasbani Basalt.

11.3.4 Palynozone QV

Palynozone QV was defined in the Hula basin, designated by pollen assemblages typical of the Mishmar HaYarden Formation, extremely rich in oak grains, mainly the winter deciduous varieties. Similar spectra had been extracted from sediments of the Ubeidiya Formation (Horowitz 1979, p. 144) and from every borehole analyzed in the Jordan Valley (Horowitz & Horowitz 1990, Horowitz 1996a), indicating that all are broadly contemporary. An outcrop at Abu Habil was correlated by Bender (1974a, p. 100) and Tchernov (1987) with the Ubeidiya, while Horowitz (1979, p. 144) thought that only the upper part corresponds to this formation. This correlation is doubted by Macumber & Edwards (1997) and by Copeland (1998), but see Section 11.3.2.

The Upper Dome Country Terraces (Sneh 1982) of the northern Arava occur at the same sequence stratigraphic position as Palynozone QV, and are therefore considered synchronous. A similar reasoning was applied (Horowitz 1979, p. 121) to connect the third gentle slope above the bottom of wadis in the western highlands (Fig. 9.6.2) with this palynozone. This slope is barren of any positively identifiable sediments west of the Jordan River, while to the east, far enough from the Rift Valley to withstand erosion, they occur as high terraces in most of the larger wadis. The better studied of these was defined by Besançon & Hours (1985) as the Dauqara Formation, in the upper Zarqa Valley some 40 km east of the Jordan Valley. This formation yielded numerous flakes and other artifacts, considered by Parenti et al. (1997) to represent a somewhat more evolved stage of the Ubeidiya Developed Olduvan. In addition, bones of *Mammuthus meridionalis, Equus* cf. *tabeti*, both known from Ubeidiya, and *Bos primigenius*, considered somewhat younger, were found in conjunction with the artifacts.

It seems reasonable to assume that both the sites of Ubeidiya and Dauqara are of Palynozone QV times. Apparently the site of Ubeidiya is embedded in sediments of the early part of this palynozone, and its later parts eroded during the next dry period, QVI, being close to the subsiding region of the Rift. The Dauqara on the other hand was incised quite deeply in the subsequent interpluvial, but its upper part apparently suffered less from the erosion.

Radiometric ages of the magnetically normal Yarda and Yarmouk basalts, overlying QV at the outcrops, cluster around 600–800 Ka (Siedner & Horowitz 1974, Mor 1986, Heimann 1990), which does agree with the suggested correlation. The age of 1.36 ± 0.22 Ma obtained for basalts within the underlying Palynozone QIV in the Hula, contain the age of QV to the interval also suggested by correlations with the deep sea cores, which thus cannot place the site of Ubeidiya at 1.4 Ma, as suggested by Tchernov (1987) based on correlations of rodent faunas with southern Europe. Paleomagnetic studies of the site of Ubeidiya show all the analyzed samples to be of reversed polarity (Opdyke et al. 1985, Braun 1992), which does not contradict the age proposed by correlation of Palynozone QV with the oxygen isotopes curves (Fig. 6.6.2).

11.3.5 Palynozone QVI

Again of a dry nature, Palynozone QVI is hardly represented by exposed sediments. It was encountered in the Hula and Dead Sea subsurface, but in the central Jordan Valley even such an occurrence is missing (Fig. 6.1.3). It was designated as the Ayyelet HaShahar Formation in the center of the Hula basin, appearing on the margins as an erosive surface underlying the Benot Ya'aqov Formation. An exception to the rule is the Middle Acheulian site of Gesher Benot Ya'aqov (Goren-Inbar 1995, 1998), embedded in sediments most probably deposited during the early part of QVI. This is concluded from pollen spectra obtained from the site (B.H. Conway, Geological Survey of Israel 1998, pers. comm., and pers. obs.), which are poor in arboreal components, very similar to those of the Ayyelet HaShahar Formation in the Notera 3 borehole. The sediments at the site are very rich in sponge spicules, indicating a marsh environment typical of the dry periods. Such spicules occur in the Hula only in respective peaty sediments (Ehrlich 1973). Paleomagnetic analysis of the site's sequence (Verosub et al. 1998) shows the lower part to be of reversed polarity, with the upper sector normal, indicating the Matuyama–Brunhes transition correlative with the beginning of Palynozone QVI (Fig. 6.6.2). The *Melanopsis* assemblage (Heller & Sivan in preparation), which shows an affinity with those from Ubeidiya, also strengthens the assignment of the site to the lower part of Ayyelet HaShahar, following the QV tradition.

This stratigraphic assignment of the Benot Ya'aqov site could cause some confusion. It is not, as claimed by Goren-Inbar (1998), embedded in sediments of the younger formation bearing the same name, of Palynozone QVII age, which yielded entirely different pollen spectra. Furthermore, a site which was excavated long ago in the Benot Ya'aqov Formation (Stekelis et al. 1937, Stekelis 1960), which was originally designated by its Arabic name Jisr Banat Yaqub, is also presently known by the Hebrew name, Gesher Benot Ya'aqov. It seems thus that we have two sites, of different ages (although bearing quite similar industries), now referred to by the same name. The solution preferred here is to retain the Arabic name which has priority for Stekelis' site, north of the bridge, and use the Hebrew for the newly excavated site south of the bridge at Gesher Benot Ya'aqov.

Palynozone QVI is represented in the wadis by a steep slope (Fig. 9.6.2), correlated by sequence stratigraphy (Horowitz 1979, p. 121), devoid of sediments. The erosive surface underlying the Lower Dome Country Terraces (Sneh 1982) in the northern Arava was formed during this period, concluded on similar grounds (Horowitz 1979, p. 147).

The magnetically normal Yarda Basalt indicates weak volcanic activity at this time within the Rift, in the Hula basin. It is not clear whether the correlative Yarmouk Basalt also has an eruption center within the Rift, it seems rather that it is connected to the much more extensive activity on the Golan (Bender 1974a, p. 105; Steinitz & Bartov 1991). Heimann (1990) reports from the Notera 3 borehole in the Hula basin an age of 1.13 ± 0.31 Ma at depth of 550 m for Palynozone QVI, which disagrees with the correlation proposed here, being too old. A possible reason for the discrepancy in this dating may lie in the extremely weathered state indicated for some of the basalts in this borehole (Kashai & Goldberg 1984).

11.3.6 Palynozone QVII

Defined in the Hula as the Benot Ya'aqov Formation, Palynozone QVII is characterized by pollen spectra extremely rich in arboreal components, particularly deciduous oaks, indicating a wet Mediterranean pluvial environment. QVII sediments are widespread in the Jordan Valley both as exposures and in the subsurface (Fig. 6.1.3). To the north of the Hula Valley the lake sediments of Benot Ya'aqov interfinger with extensive spring deposits of the Kefar Yuval Travertine. The correlative Naharayim Formation (Horowitz 1974, Braun & Tchernov 1991) was defined in the central Jordan Valley, extending down to the Dead Sea, possibly even further south, particularly east of the Jordan River, and also includes the Kufrinja Gravel (Bender 1974a, p. 97) and Tabaqat Fahl Formation (Macumber 1992). Some extrusions of the Raqqad Basalt may have taken place already during QVII. The palynozone is represented in the northern Arava by the Lower Dome Country Terraces (Horowitz 1979, p. 147, Sneh 1982).

QVII is represented in the wadis by a prominent gentle slope, second from the bottom (Fig. 9.6.2). In contrast with the older slopes west of the Jordan Valley, the short duration from the shaping of this one was not sufficient for erosion to remove all of the sediments accumulated in the wadis during the humid QVII. Thus this slope is covered in numerous localities west of the Rift by gravel beds designated the Baq'a Conglomerate (Horowitz 1979, p. 122), which yielded artifacts and numerous sites, particularly of Upper Acheulian cultures (Fig. 3.5.2). In the northern highlands, what remained from the earliest Quaternary QI peneplain but also younger rocks, are usually partly covered by red, occasionally thick paleosols of this age (Ronen 1979), in which plenty of Acheulian artifacts were found. These are especially common on both sides of the Hula Valley, to the west in the eastern Upper Galilee in the Yir'on–Bar'am area (Ohel 1986, 1991), and east on the Golan Plateau (Goren-Inbar 1985) at Berekhat Ram.

A similar situation is seen in the Transjordanian highlands where QVII wadi sediments, yielding artifacts of the same nature as those west of the Jordan Valley, are designated as the Bire Formation (Besançon & Hours 1985, Parenti et al. 1997) in the upper Zarqa Valley, but are widely distributed also in other wadis.

This is also the first period from which caves bearing sediments are preserved. The Umm Qatafa Cave (Neuville 1951) in Wadi Hareitun, leading to the Dead Sea, presents an entire sequence of QVII deposits bearing Middle and Upper Acheulian artifacts (Horowitz 1996b). It is highly probable that the human remains found in Zuttiyeh Cave below the Acheulo-Yabrudian layers could also be dated to Palynozone QVII.

11.3.7 Palynozone QVIII

Sediments of QVIII, deposited under a dry climate, have a very restricted distribution, and are known only from the subsurface of the Hula, the Hulata Formation, and Dead Sea basins (Fig. 6.1.3). An exception is the lower beds of the Zuttiyeh Cave in Nahal Amud, leading to Lake Kinneret, where an almost complete sequence of QVIII was preserved, bearing artifacts of the Acheulo-Yabrudian culture, replaced by Middle Paleolithic toward the upper part of the palynozone.

In the rest of the region erosion prevailed, seen as surfaces underlying the Ashmura and Lisan formations all along the Jordan Valley, or a steep cutting (Fig. 9.6.2) into the gentle QVII slope in the wadis (Macumber & Edwards 1997). Volcanic activity in QVIII times, possibly already starting in QVII, is expressed by flows of the Raqqad Basalt, which occasionally arrived at the Jordan Valley, typically located between the Naharayim and Lisan formations. These are known only in the region east of the Jordan River. Correlative volcanics, the Ma'yan Barukh Basalt, occur in the northern part of the Hula Valley, sandwiched between the Kefar Yuval and Dan travertines.

11.3.8 Palynozone QIX

Being the last wet period, sediments of this palynozone are widespread all along the Jordan Valley from Lake Kinneret down to the Arava, comprising the Lisan Formation and its terraces (Sneh 1982). Its northern correlative, the Ashmura Formation in the Hula, is less conspicuous in the landscape, being overlain by younger sediments and soils although it covers most of the basin's floor (Horowitz 1973). Large parts of the northern Hula Valley are lined with the correlative Dan Travertine. In the central Jordan Valley, travertines of the Bet She'an Formation are intercalated with the Lisan lake sediments west of the Rift, while to the east the "Knob Limestone" Travertines (Macumber & Edwards 1997) make up the upper part of the correlative Wadi Hammeh Conglomerate.

North of the Dead Sea the Fasayil Travertine most probably correlates with the lower part of the Lisan; further south in the northern Arava, a similarly lying unit is the Mo'a Travertine. Comparable occurrences are also known upstream in Nahal Zin and other wadis on both sides of the Jordan Valley. Deposits bearing Palynozone QIX pollen spectra are also encountered in the subsurface almost everywhere in the Jordan Valley (Fig. 6.1.3).

The lowest considerable gentle slope of the wadis (Fig. 9.6.2) is almost everywhere developed in conjunction with the Nahshon Conglomerate west of the Jordan Valley (Horowitz 1979, p. 122). Comparative gravel layers east of the Rift are defined as the Wadi Hammeh Conglomerate (Macumber & Edwards 1997), the Khirbet Samra Formation (Besançon & Hours 1985), and occasionally other names. These wadi sediments interfinger directly with the Lisan in numerous localities on both flanks of the Rift Valley, thus leaving no doubt as to their synchronous nature. The lower part of these gravel beds yielded many finds and sites of the Middle Paleolithic Mousterian culture, while their upper part is strewn with Upper Paleolithic and Epipaleolithic artifacts (Bar-Yosef 1987, Macumber & Edwards 1997). Correlative beds and sites are also known from caves and rockshelters on both sides of the Jordan River, as well as from abundant surface red paleosols, which remained from QIX times in numerous localities on the highlands (terra rossa) and the Levantine coastal plain (locally called "hamra").

The major faulting phase which predated the end of Palynozone QIX times (see Chapter 9), completely changed the face of the Jordan Valley. The Hula subsided considerably in relation to its bordering highlands, which is the main reason why the Ashmura Formation is usually covered, in contrast with the Lisan; Lake Kinneret was formed, in which the Tabgha Formation has accumulated ever since; and the northern basin subsided in the Dead Sea region, causing erosion of the Lisan Lake sediments, and its coverage by the fluviatile Fatza'el Member (Hasa or Damiya Formation in Transjordan) north of the Dead Sea, and the Iddan Member to the south. Except for the Iddan, which was never surveyed, all other units are very rich in prehistoric remains of Epipaleolithic affinities. No volcanics are known for QIX, but its plentiful uranium series and radiocarbon datings confirm its age.

11.3.9 Palynozone QX

The interpluvial Palynozone QX was defined by pollen spectra of the Mallaha and the uppermost sector of the Ashmura formations, in the Hula Valley, known both from outcrops and boreholes. Correlative sediments constitute the upper part of the Tabgha Formation in Lake Kinneret and the uppermost sector of the Upper Dead Sea Group to the south (Horowitz 1979, p. 173), including the Ze'elim Formation. Correlations are based both on pollen assemblages and radiocarbon datings. The middle part of QX is considerably more humid than the earlier and later, resulting in the expansion of lakes in all basins of the Jordan Valley, as well as by the formation of at least one new lake near Bet She'an, where the Rehov Formation was laid down, very rich in Chalcolithic remains and covered by Early Bronze potsherds.

Areas outside the relatively limited Rift Valley lakes were severely eroded during the early and late drier stages, but were silted up while the middle wetter phase persisted. This created a typical "Lower Terrace", rich in later prehistoric and archaeological remains. This subordinate terrace, which is not counted when the more prominent pluvial ones are referred to, is very common in most wadis leading to the Jordan Valley (Picard 1932, Vita-Finzi 1966, Horowitz 1979, p. 125). The terrace was termed the Sukhne Formation in the upper Zarqa Valley (Besançon & Hours 1985), or "Fill 4" in Wadi Hasa (Vita-Finzi 1966). The "sub-recent travertines" of the northern Arava could well be formed by an extended spring activity at this wetter time.

11.4 CONCLUSION: CHRONOLOGY OF CULTURES IN THE JORDAN VALLEY

The chronology presented here is based entirely on geological grounds, and does not take into consideration the cultural affinities of artifact assemblages or sites as chronostratigraphic markers. The use of artifacts as "age indicators" is a very common practice in Quaternary geology, in most cases for lack of other age-conclusive fossils or radiogenic datings. This has frequently created confusion; as an example, the cultural term "Lower Paleolithic" encompasses almost the entire Quaternary. It is thus our aim to base the chronology of cultures and sites of the Jordan Valley on palynostratigraphic grounds, which seem more objective and offer a much more refined timetable. This pertains to most of the Quaternary except for the last 80 Ka or so, when cultural evolution was much more rapid and accurately dated, enabling the use of artifacts as reliable "fossils". Most of the sites have been subject to pollen analysis at some time or another (Horowitz 1992a, Section 10.3.1), and thus could be correlated with the continuous pollen diagrams obtained from boreholes in the Jordan Rift Valley (see Chapter 6), themselves dated by a variety of methods (see Section 11.3).

If indeed the fractured flints or bones of Bethlehem prove to have been man-made, or the assemblage (too small to be assigned a definite cultural stage) found near Yir'on is shown unquestionably to underlie the Dalton Basalt (see the beginning of this Chapter), then Palynozone QI is the first period in which humans occupied the southern Levant. Until such proofs are accepted beyond doubt, the chopping tools and flakes recovered from the Erk el Ahmar Formation and its correlatives should be regarded as indicators of the first arrival of *Homo erectus* on the scene in the southern Levant. Again, the assemblage is too small to be assigned a specific culture; Horowitz (1996b), following Hours (1975), referred to it as "Para-Acheulian". This suite of artifacts is no doubt of QIII times, extending from 1.82 to 1.49 Ma ago.

The oldest site in the southern Levant designated as Lower Acheulian, contrary to common knowledge, is not Ubeidiya but Borj Qinnarit, on the Lebanese coast south of Sidon, embedded in sediments correlated with Palynozone QIV (Horowitz 1992a, p. 411). Hours et al. (1973) designated the few core choppers and flakes as Para-Acheulian, but later (Hours 1981) as Early Lower Paleolithic or (Besançon et al. 1988) Early Acheulian. Stratigraphically, it lies midway between the artifact occurrences of Erk el Ahmar and those from Ubeidiya, during the time span between 1.49 and 1.25 Ma. No exposures of QIV, and so no evidence of humans, are known from the Jordan Valley or its neighboring areas. The Lower Acheulian of Ubeidiya (Bar-Yosef 1994, Goren-Inbar 1995) is embedded in the lower part of Palynozone QV sediments, covering the period of 1.25 Ma up to 790 Ka (Horowitz 1996b). The site itself is most probably no younger than one million years. This is possibly also the case with the artifact finds at Abu Habil (Huckriede 1966), although this is doubted by Macumber & Edwards (1997) and by Copeland (1998). The later part of QV, somewhat younger than the site of Ubeidiya, yielded a variety of artifacts from the Dauqara Formation east of the Rift Valley (Parenti et al. 1997).

QVI sediments, spanning the 790–425 Ka interval, are again not exposed in the Jordan Valley except for a small outcrop south of the bridge at Gesher Benot Ya'aqov, where the Middle Acheulian site bearing this name is embedded (Goren-Inbar 1995, 1998, and see remark in Chapter 11.3.5). Other Middle Acheulian industries of that age are known only further north, at the site of Latamne (Clark 1967, 1968), and west, at Evron (Ronen 1979, 1991, Ron et al. 2001).

Middle Acheulian sites of the Jordan Valley and surroundings, such as Jisr Banat Yaqub (Stekelis et al. 1937, Stekelis 1960, and see remark in Section 11.3.5) or the Umm Qatafa beds G and F (Neuville 1951) are found in sediments of the early part of Palynozone QVII (Horowitz 1979, p. 338), oxygen isotope Stage 10, around 400 Ka ago (Fig. 6.6.2). Numerous Middle and Upper Acheulian artifacts are reported from the Naharayim Formation, also of Palynozone QVII age (Huckriede 1966; Bender 1974a, p. 101; Bar-Yosef 1987), but were never studied in detail. Middle Acheulian sites are embedded within the lower part of the Tabaqat Fahl Formation (Macumber & Edwards 1997) just east of the Jordan Rift, opposite the Yizre'el Valley.

The later parts of Palynozone QVII yielded rich and varied assemblages of Upper Acheulian artifacts, especially elaborated handaxes (Bar-Yosef 1994), from several sites. Of these, two stratigraphic levels were identified. The middle QVII, corresponding to oxygen isotope Stage 8, some 270–240 Ka ago, encompasses the site of Ma'yan Barukh (Stekelis & Gilead 1966) and most probably also Berekhat Ram (Goren-Inbar 1985) and Umm Qatafa Layer E (Neuville 1951, Horowitz 1996b). The second, last phase of Palynozone QVII, oxygen isotope Stage 6, is characterized by a wealth of surface finds all around the southern Levant, including the present desert areas (Bar-Yosef 1994). Excavated sites within the Jordan Valley reaches are those from the upper part of the Tabaqat Fahl Formation, east of the central Jordan Valley (Macumber & Edwards 1997), and the Umm Qatafa Cave Layer D in the Judean Desert. It seems that most finds in the Naharayim Formation can be attributed to this stage.

The last culture of the Acheulian, occasionally referred to as a transition into, or the early part of, the Middle Paleolithic (Bar-Yosef 1987), was first termed the "Acheulo-Yabrudian", and renamed the "Mugharan Tradition" by Jelinek (1982). It was found only in a few cave or rockshelter sediments, such as the Zuttiyeh Cave in Nahal Amud, the only one in the Jordan Valley (Gisis & Bar-Yosef 1974). Other well-known sites are the Yabrud rockshelter north of Damascus, the Adloun caves in southwestern Lebanon and the Tabun Cave on Mount Carmel. The pollen diagram from the latter (Horowitz 1992a, p. 321) indicates that the Mugharan culture of Bed E could only be attributed to Palynozone QVIII, isotope Stage 5, 130–70 Ka ago (Horowitz 1996b). No other rocks formed during this dry period are exposed in the Jordan Valley and surroundings.

Contrary to the above age assignment, considerably older dates were obtained by a variety of isotopic methods for the Mousterian and Mugharan cultures. Grün et al. (1991) obtained ESR ages for the Mousterian in Tabun spanning the 102–186 Ka interval; Mercier et al. (1995) published a series of TL dates derived from burnt flints from the same cave, reporting ages for beds C and D (Mousterian) in the range of 171 ± 17 down to 263 ± 27 Ka, while Bed E (Acheulo-Yabrudian) is 270 ± 22 to 331 ± 30 Ka old. Clark et al. (1997) dated the Middle Paleolithic occupation at Ain Difla, upstream in Wadi Hasa southeast of the Dead Sea, by a variety of methods to the 90–180 Ka interval.

These dates stand in contrast with the palynostratigraphy, as well as with numerous other datings of Palynozone QIX and QVIII, including rocks in which Mousterian artifacts are embedded, always yielding ages younger than 70–80 Ka (Horowitz 1979, p. 151; Schwarcz et al. 1979; Schramm et al. 1997; Macumber & Edwards 1997; and many others). The 10 Ka span between 70 and 80 results from inaccuracies and different opinions as to the exact age of oxygen isotope Stage 4. The TL ages are also considerably older than others obtained by various methods from Tabun (Porat et al. 1999), although even the latter are slightly older than suggested here.

As mentioned above, all known occurrences of Mousterian sites are from sediments deposited during the first humid phase of Palynozone QIX, except for the one from Zuttiyeh and possibly that at Tabun. Numerous finds of artifacts line the Jordan Valley (Bar-Yosef 1987, Macumber & Edwards 1997) embedded within the Nahshon, Wadi Hammeh and other correlative conglomerates, which occasionally interfinger with the well-dated Lisan Formation lake sediments. Mousterian sites are common in many wadis leading to the Jordan Valley on both sides (Marks 1976, 1977, Macumber 1992), some of which were pollen analyzed, proving their QIX affinity (Horowitz 1976).

A single date which somewhat precedes the 70–80 Ka figure is from the Zuttiyeh Cave, by Schwarcz et al. (1980), who found that the Acheulo-Yabrudian industry accumulated there from approximately 150 Ka ago was replaced by Mousterian at about 95 Ka. This possibility cannot be ruled out, since no outcrops of Palynozone QVIII, except for those from caves, are known from the Jordan Valley, so it is plausible that Mousterian industry could somewhat predate the onset of Lisan deposition. Weinstein-Evron (1983, 1988), when comparing the Mousterian chronostratigraphy of sites with the detailed palynostratigraphy of the relevant layers in the Hula Basin, also arrived at the conclusion that this culture must have commenced in the Levant about 100 Ka ago. Besides the caves, the Mousterian occupation of Naame, on the Lebanese coastal plain, is embedded within sediments correlative with Palynozone QVIII, termed the Naamean Transgression (Besançon 1981), which supports the conclusion advanced above. For the time being, no explanation is proposed for the considerable discrepancy between the various dating methods (see also Section 1.4.2).

The Mousterian is almost everywhere followed by the Upper Paleolithic suite of cultures, in sediments of oxygen isotope Stage 2, the next humid phase of QIX (Horowitz 1976) which occurred approximately 32–22 Ka ago (Horowitz 1971). But the early (up to about 40 Ka) and middle QIX humid phases are separated by a dry interval which, as usual, is not represented by exposed sediments except in caves. Thus the transition of Middle to Upper Paleolithic is found only at the Amud Cave, west of Lake Kinneret, which is the sole known occurrence in the region (Hovers et al. 1994).

The Upper Paleolithic extended up to approximately 20–19 Ka ago, and is represented by numerous sites all along the Jordan Valley (Bar-Yosef 1987, Macumber & Edwards 1997). Being within the radiocarbon dating range, Upper Paleolithic sites (and, naturally, younger ones) are easy to date. Such sites are also known from the early part of the drier phase of QIX, following Stage 2, itself extending up to some 16–14 Ka ago (Horowitz 1971, 1992a, p. 418, Weinstein-Evron 1990, 1993). Since faulting ended the existence of Lake Lisan some 18 Ka ago, no younger layers overlie the corresponding Last Glacial Maximum dry phase sediments.

The retreat of Lake Lisan gave room for the development of favorable environments in most of the Jordan Valley, which has indeed been inhabited ever since by numerous settlers of varying cultures. These are very often radiocarbon or otherwise dated (Bar-Yosef 1987, Macumber 1992), and are found within the upper QIX and through the QX Palynozone (Horowitz 1992a, Chapter 10.3.2). Thus, the Epipaleolithic, comprising the Kebaran, Geometric Kebaran and Natufian, extends approximately from 20–19 to 10.3 Ka; the Neolithic up to some 7,000 years ago, followed by the Chalcolithic and the Historic era.

CHAPTER 12

Paleoecology of Man

The present chapter, like the previous, discusses the period represented by Palynozones QII through QX, spanning approximately the last two million years, typified by tectonic activity of the Levantine stage and the appearance of humans in the southern Levant. On a global scale, this period is characterized by the Quaternary climatic changes, which are represented in the southern Levant by alternations of wetter and drier periods, as compared with the present-day (dry) conditions (see Section 6.6). To the best of our present knowledge, these three crucial phenomena in the history of the Jordan Rift Valley do not necessarily coincide, since climatic changes are already prevalent in QI times while no hominid traces are known for QII. The reason for that mishap could be geological, namely that Palynozone QII is represented only (if at all!) by a single, questionable outcrop, from which merely one sample yielded a rather poor pollen spectrum. On the other hand, as already discussed in Chapter 11, questionable finds of which one cannot deny the possibility they may at some time be verified (Figs 12.1 and 12.2) indicate the probability that early humans already inhabited the region in QI times (Stekelis 1940, Ronen 1996). Palynozone QI rocks are discussed in detail in Chapter 5, while its age, paleogeography and environments are covered in Chapter 7.

What we are quite sure of is that core choppers and flakes are found at several localities in sediments deposited during Palynozones QIII and QIV (Hours 1975; Lamdan et al. 1977; Horowitz 1979, p. 296; Braun et al. 1991; Horowitz 1996b; E. Tchernov, Department of Evolution, Systematics and Ecology, the Hebrew University of Jerusalem 1998, pers. comm., Ron & Levy 2001). Unfortunately, these finds are not always given appropriate attention, and are entirely disregarded in reviews such as that by Bar-Yosef (1987, 1994), who personally identified most of the finds, as well as by others (Goren-Inbar 1995). These authors regard Ubeidiya as "the most ancient site in the Levant", which may be true if an actual large site is considered, but misses the point that humans inhabited the region for several hundred thousand years before, even if not occupying "real" sites (or such sites have not yet been discovered).

Figure 12.1. The QI site at Bethlehem, during excavations. Photo by M. Stekelis, ca. 1935.

12.1 PALEOGEOGRAPHY AND ENVIRONMENTS

The most dramatic change in the paleogeography of the Jordan Valley at the transition from Palynozone QI to QII is the transformation of the southern Dead Sea into an endoreic basin, following the onset of the north–south oriented Levantine faulting phase (Horowitz 1989b, and see Chapter 9). Ever since, it has functioned as the terminal base level of the newly formed drainage system, which began capturing the previous ones, particularly to the west of the Rift. Concurrently, the Dead Sea became increasingly more and more saturated with various salts, until it reached its present state, and is now the most saline water body on Earth. Besides its being a terminal lake, this process was accelerated by the uplift and dissolution of rock-salt bodies, such as Mount Sedom and the one underlying the Lisan Peninsula.

It is in the southern Dead Sea region where the process of developing an internal drainage system began. QII subsidence is most spectacular in this area, attaining a maximum figure of almost 900 m at the Melekh Sedom 1 borehole, and several hundred meters in others (Figs 6.1.1, 6.1.3 and 9.6.2). There is hardly any subsidence at this time in the central Jordan Valley, while the Hula basin shows a figure of only 65 m for QII times (Horowitz & Horowitz 1985, Horowitz 1989b). In addition to the subsidence, considerable synchronous uplift of the Rift shoulders (Fig. 9.6.2) participated in turning the drainage system endoreic. A detailed discussion of subsidence and uplift is advanced in Chapter 9.

Figure 12.2. Possible QI finds at Yir'on, Upper Galilee. (A) The gravel horizon in which the artifacts were found, presumably underlying the Dalton Basalt (Fig. 5.3.11). (B) A core chopper, viewed from above. Photos courtesy of A. Ronen.

The deeper basins of the Jordan Valley were occupied by extensive lakes during periods of humid climates, typical of the even-numbered palynozones, which shrank in dry periods into marshes or playas, or occasionally dried out altogether. A continuous sequence of such pluvial lakes is known from both the Hula and southern Dead Sea, while the central Jordan Valley did not develop a lake in QVII times, probably due to earlier faulting (Horowitz 1987a). A lake or a series of lakes also developed in the southern Negev, probably in QIII, due to structural partial blockage of the drainage, depositing the Zehiha Formation (Ginat et al. 1996, Ginat & Zilberman 1996, Ginat 1997). Other temporary lakes developed on the Transjordanian Plateau, following its uplift which formed a series of local endoreic depressions (Fig. 1.2), of which the best known is El Jafr, populated by early humans (Huckriede & Wiesemann 1968; Horowitz 1979, p. 153).

Rivers were developing into the new base level to the west, north and south, all along the Jordan Valley, nibbling at the former ones, with some of these newcomers attaining considerable drainage basins (Fig. 1.2). The drainage capturing processes were expedited by uplift of both the eastern and western Rift shoulders (Fig. 3.1.1), which initiated the creation of a novel pattern of watersheds. West of the Jordan Valley, a new watershed divided the westward drainage to the Mediterranean from that eastward to the Dead Sea; on the eastern highlands former rivers continued their westward routes, but were stopped by the Jordan Rift, not going any further west as before, but deepening notably. This is quite obvious from a look at the drainage map (Fig. 1.2), which clearly shows that rivers draining Transjordan to the Rift are considerably longer and better developed when compared with their western younger counterparts. The eastern uplift also created a series of playas fed by local internal drainage systems, developed east of the watershed (Fig. 1.2).

The drainage system to the south of the Dead Sea, which is developed over a vast area of predominantly desert environment, inherited the main channels used by the Pliocene Arava Formation rivers, which also led to the southern Dead Sea basin. The HaMeshar system, which was only active for a short duration, less than a million years of QI, apparently had no enough time to completely destroy its predecessor, the Arava system. This seems to explain why the southern drainage is so well developed, despite the extremely arid climate (although pluvials were somewhat more humid). The watershed at Gav Ha'Arava seems to have remained unchanged ever since the Oligocene, separating the Gulf of Aqaba and the Dead Sea systems.

Volcanism played a subordinate role in Quaternary times, being mainly limited to the northern part of the Jordan Valley. The center of volcanic activity at that time seems to have moved northeastward, to the Golan Heights (Mor 1986) and further east, with only slight echoes reaching the Jordan Valley. These are expressed by several minor basalt flows, arriving in the Hula from the east, and possibly some localized extrusions in the southern and northern ends of this basin. Such flows have also rarely arrived in the central Jordan Valley and the Dead Sea.

Reconstructions of the past environments are based principally on their respective pollen assemblages more than on any other group of fossils, aided by the types of sediments. The main reason is that palynological information for the Jordan Valley is ample and continuous, as opposed to the fragmentary nature of all other data. Other fossils, as well as sediments, reflect only the indigenous habitats where they live or are laid down, while pollen assemblages faithfully represent both local and regional environments. In this way, changes in the local conditions can be attributed either to general causes, such as climatic changes, or to autochtonous ones, for instance structural disturbances or volcanism. Needless to say, pollen grains are the only available fossils for those interpluvial sequences known only from boreholes which, in terms of time, comprise a substantial part of the Quaternary in the Jordan Rift Valley. On top of that, fossils such as vertebrates, mollusks or plants have been studied chiefly from archaeological excavations, where taphonomy plays an important role, so that human tastes, or scavenging, hunting, gathering and agricultural techniques may have considerably affected the spectra of animals or plants found in the site.

The pollen spectra for Palynozones QII through QX indicate a Mediterranean environment, alternating between wetter pluvials and drier interpluvials, with stadials and interstadials of median characteristics. QII still shows some occurrences of Pliocene–earliest Quaternary elements such as *Picea orientalis*, but these diminish quite quickly upward along the sequence, to be replaced entirely by *Quercetalia*, the typical Mediterranean floral components. The environmental gradient remained as during the Pliocene, more humid to the north (Tables 6.5.2.4–6.5.2.8). These climates, and their possible causes, are detailed in Chapter 6.

12.1.1 Palynozone QII

The initial outcome of the Levantine tectonic stage was expressed by the formation of two endoreic basins (Fig. 12.1.1), one in the Hula depositing the Gonen Formation, the other in the southern Dead Sea region, both occupied by restricted lakes, possibly saline to some degree, the southern probably more so (see Section 9.6.2). The hilly country on both sides of the Rift was uplifted and considerably incised; for example in Wadi Hareitun leading from Bethlehem to the Dead Sea, where uplift was measured and dated QII incision cut at least 150 m deep into the QI erosion surface (Fig. 9.6.2). This figure, recorded where the Umm Qatafa Cave is located, is only a minimum one, increasing considerably when the wadi approaches the Dead Sea, a distance of some 20 km.

The uppermost flows of the Ruman Basalt may have erupted during the early stages of Palynozone QII south of the Hula on the Korazim block, blocking the basin from draining southward. It may well be that the channels developing toward the Dead Sea had not yet a chance to reach the Hula in QII times. It is quite difficult to say what was happening during this time in the central Jordan Valley, since all we have is a single deep borehole where QII was not encountered (Fig. 6.1.3). It seems plausible that this area, together with most of the Jordan Valley, underwent erosion toward the rapidly subsiding southern Dead Sea basin.

The vegetation in QII times, as seen from the pollen spectra (no other fossils are known for this interval), resembled the present day in two respects. Trees were more abundant to the north, grading to steppe southward, expressing the typical Mediterranean environmental gradient of the southern Levant; and although *Picea orientalis*, the typical QI tree, still occurs, its place was gradually being taken by

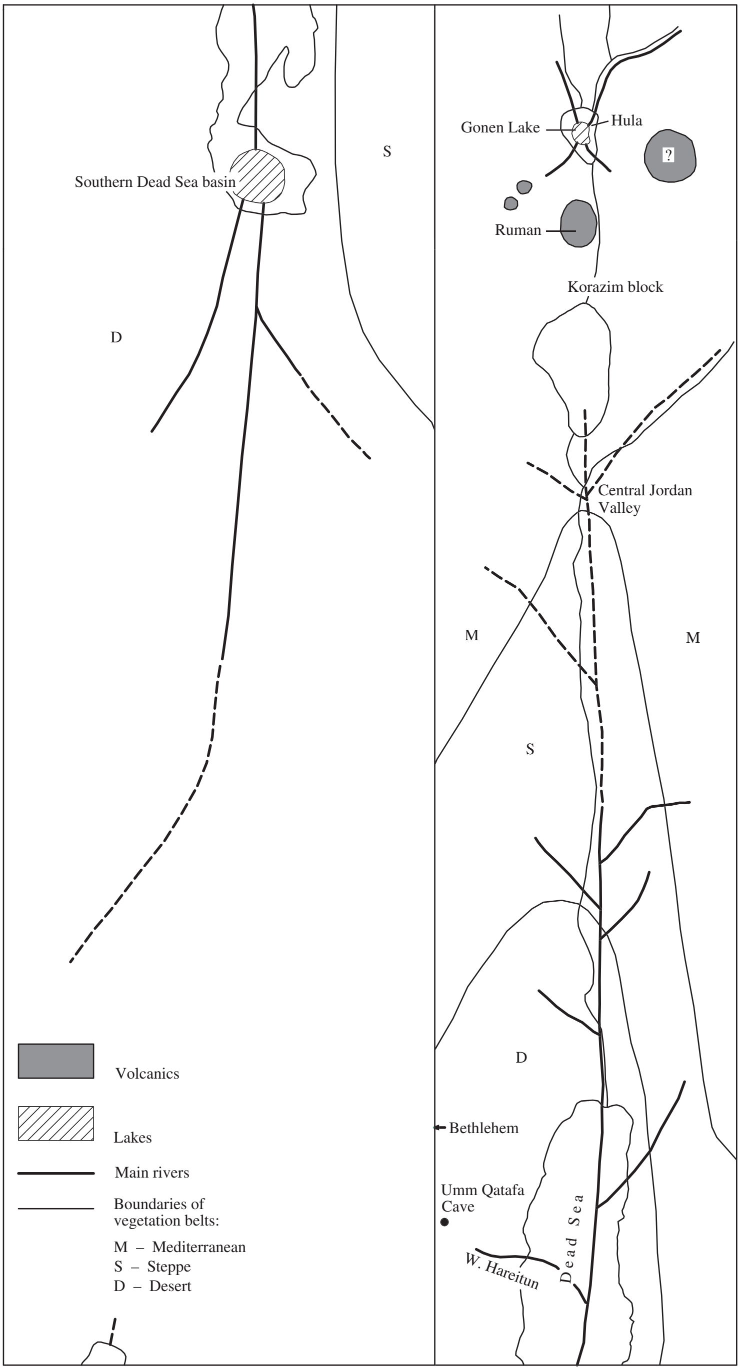


Figure 12.1.1. Paleogeography and paleoenvironments, Palynozone QII.

Quercetalia. Desert vegetation was less developed than at present, which may have resulted either from the environmental gradient, which may not yet have been sufficiently developed, or from the fact that this type of vegetation was quite rare in QI times, and subsequently took some time to fully evolve, when its appropriate habitat became available once more. The climate thus deduced for Palynozone QII is of a dry interpluvial Levantine Mediterranean character.

12.1.2 Palynozone QIII

Lake sediments of Palynozone QIII are well known from three locations in the Jordan Valley (Fig. 12.1.2), one in the Hula comprising the Gadot–Hazor fluvio-lacustrine suite, the other in the central Jordan Valley depositing the Erk el Ahmar Formation and possibly the lower Abu Habil, of a similar nature, while the third flooded the northern Arava, leaving as evidence the Shezaf Hills Conglomerate lakeshore terraces. They also occur in all boreholes drilled in the Jordan Valley (Fig. 6.1.3). Outside the Rift, the Zehiha and possibly other small lakes existed in the southern Negev, as intermediate water bodies on the route of the northward drainage system leading to the Dead Sea. A gentle slope of the wadi shoulders in the hilly country cut into the previous and steeper QII slope. This slope was most probably covered by colluvium and gravel, if we judge from similar younger slopes, but almost none of these sediments has remained due to subsequent erosion. The only exception seems to be the "Uppermost Terrace" at the headwaters of Wadi Zarqa, where alluvial sediments, most probably deposited during QIII, are preserved (Besançon & Hours 1985).

It seems quite clear that there was no hydrological connection between the Gadot and Erk el Ahmar lakes. The Gadot deposited an almost sterile white chalk and dolomites indicating sedimentation in a somewhat saline terminal basin, a heritage of QII times (see Section 9.6.2), rich in Chenopodiaceae pollen, while Erk el Ahmar was basically an intermediate, freshwater lake, as expressed in its wealth and variety of mollusks. Rare occurrences of gypsum may hint that at certain times even this lake became restricted, at least somewhat brackish, but probably only for short durations, possibly due to the existence of saline springs. A paleomorphological study (Belitzky, in prep.) also indicates no southward drainage of the Gadot Lake.

The more interesting question, to which a definite answer cannot as yet be provided, is whether the Erk el Ahmar Lake and the Shezaf Hills of the northern Arava comprised a single lacustrine entity, or if the two were connected by a river. Several occurrences of lakeshore terraces quite high above the present valley floor are known along the central and southern Jordan Valley, down to the Dead Sea (Horowitz 1979, p. 148). In the southern Dead Sea basin QIII sediments occur in sequences of several hundred meters. It seems logical to correlate the Zarqa uppermost terraces with Palynozone QIII, although no positive evidence was found for such a thought, based on sequence stratigraphic considerations. A thin

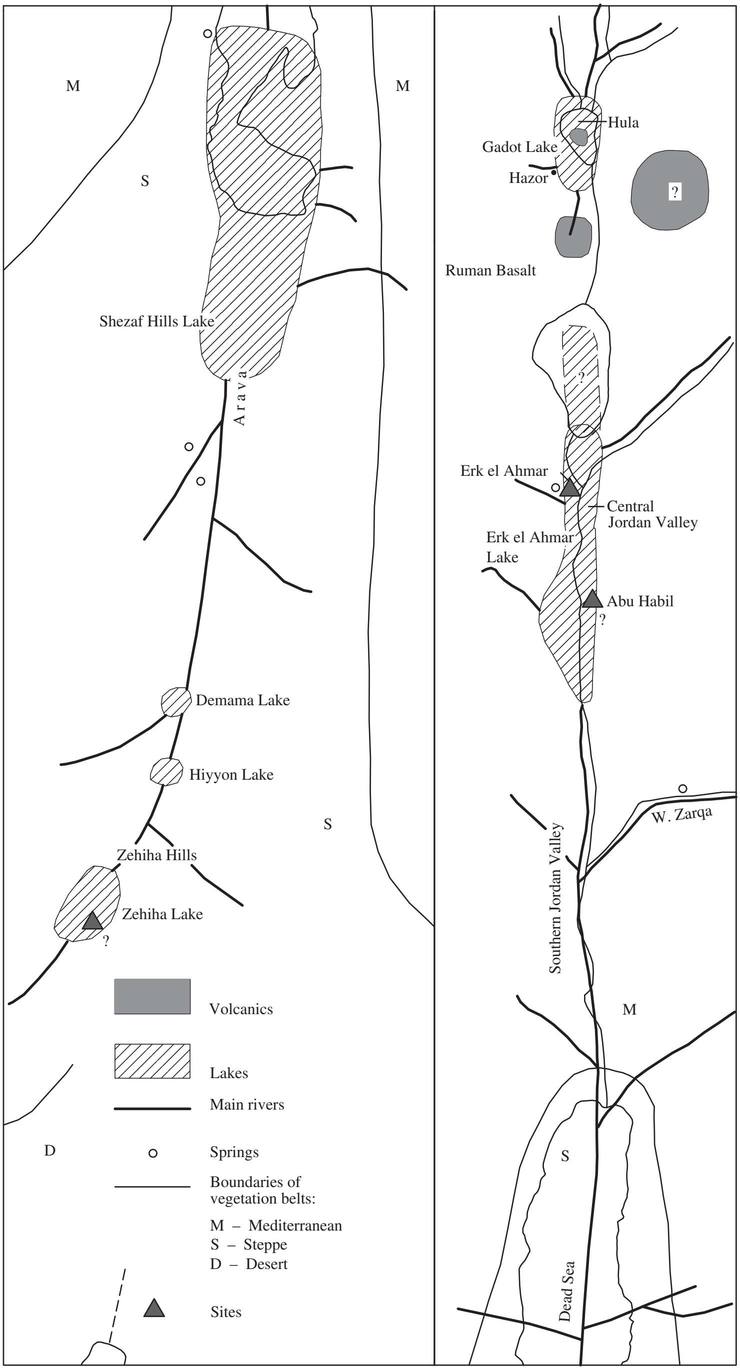


Figure 12.1.2. Paleogeography and paleoenvironments, Palynozone QIII.

basalt flow recorded from the subsurface of the Hula near the base of QIII, together with the uppermost Ruman Basalt extrusion, testify to subordinate volcanic activity at this time, which was unknown further south.

The pollen spectra, rich in arboreal components, the affluent fossils recovered from the Erk el Ahmar sediments, the extension of lakes along and outside the Jordan Valley, the style of gentle erosion in the highlands, all indicate a humid, pluvial climate for Palynozone QIII, of a Mediterranean character, wetter than at present, dominated by *Picea orientalis*, pine and oak forests to the north, grading to steppe southward. This seems like an ideal environment for early humans, whose artifacts are found in the Erk el Ahmar Formation (Figs 12.1.2.1A and B), the earliest undoubted evidence in the Jordan Rift Valley.

Outside the Jordan Valley limits, almost a 100 km south of the southern Dead Sea basin, a tributary leading to it was partially blocked by structural disturbance to form the Zehiha Lake (or series of lakes) where Nahal Zihor runs today and in its vicinity (Fig. 11.5.2). There are possible indications that humans inhabited the shores of this lake in QIII times, but it is not yet satisfactorily proven.

12.1.3 Palynozone QIV

Sediments of this period are known from the subsurface of all three basins of the Jordan Rift Valley, deposited in shallow lakes which occasionally turned into

Figure 12.1.2.1A. Excavation at Erk el Ahmar. The site, with an elephant tusk in situ. E. Tchernov poses for size. Photo courtesy N. Goren-Inbar.

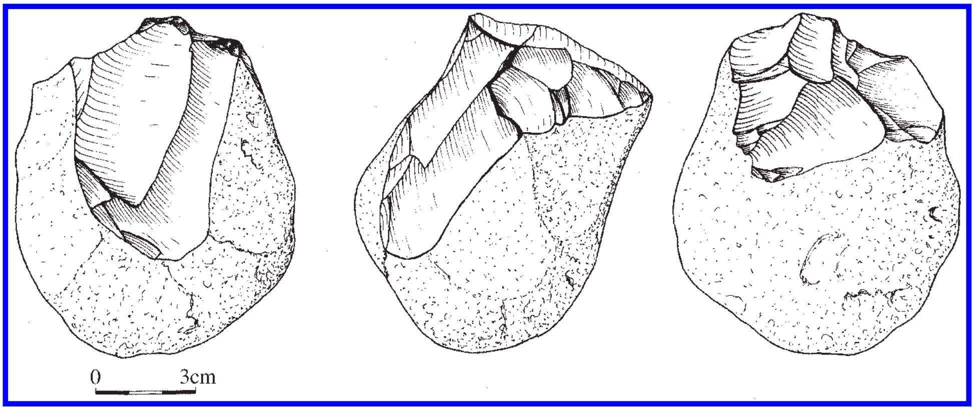


Figure 12.1.2.1B. A core chopper found by E. Tchernov, drawn by R. Penchas.

marshes (Fig. 12.1.3). Both the Hula, where the Notera Formation was deposited, and the central Jordan Valley housed freshwater bodies, while the Dead Sea was saline, as can be seen from the pollen spectra. Considering the restricted sizes of these lakes, it seems improbable that they were joined other than by a river. This is the first period in which, hydrographically, the Hula became part of the Dead Sea drainage system, by breaching the Korazim block, as seen from the pollen assemblages discussed in Section 9.6.2. From this stage on, the Dead Sea is the sole terminal base level of the Jordan Valley endoreic drainage system.

Besides these three small lakes or marshes, the rest of the region was subject to erosion, including the continuously uplifting highlands on both sides of the Rift, but at a much slower pace than before (Fig. 9.6.2). The interpluvial floods cut deeply into the previous gentle slopes of QIII times, removing much of their accumulated sediments. Several basalt flows at the top of the Notera Formation indicate some volcanic activity in the Hula or the Golan Heights which did not extend southward.

Pollen assemblages (no other fossils are known for QIV) are poor in arboreal components, comprising mainly oak, pine and rare cedars in the Hula, which indicates an interpluvial climate, also characterized by the size of the lakes and the style of erosion. A group of plants which is highly represented by its pollen is the Compositae, indicating a desert environment, particularly to the south. The renewed appearance of large numbers of Compositae (the last palynozone typified by this association was the late Miocene Mc), testifies to the development of desert in the southern parts of the Levant, following a long period of milder climate. As noted above when discussing QII pollen spectra, it may well be that a desert environment already prevailed then, but it could take its typical vegetation some time to develop, thus showing fully only in QIV times. No exposures and so no artifacts are known for QIV in the Jordan Valley, nor in its close surroundings.

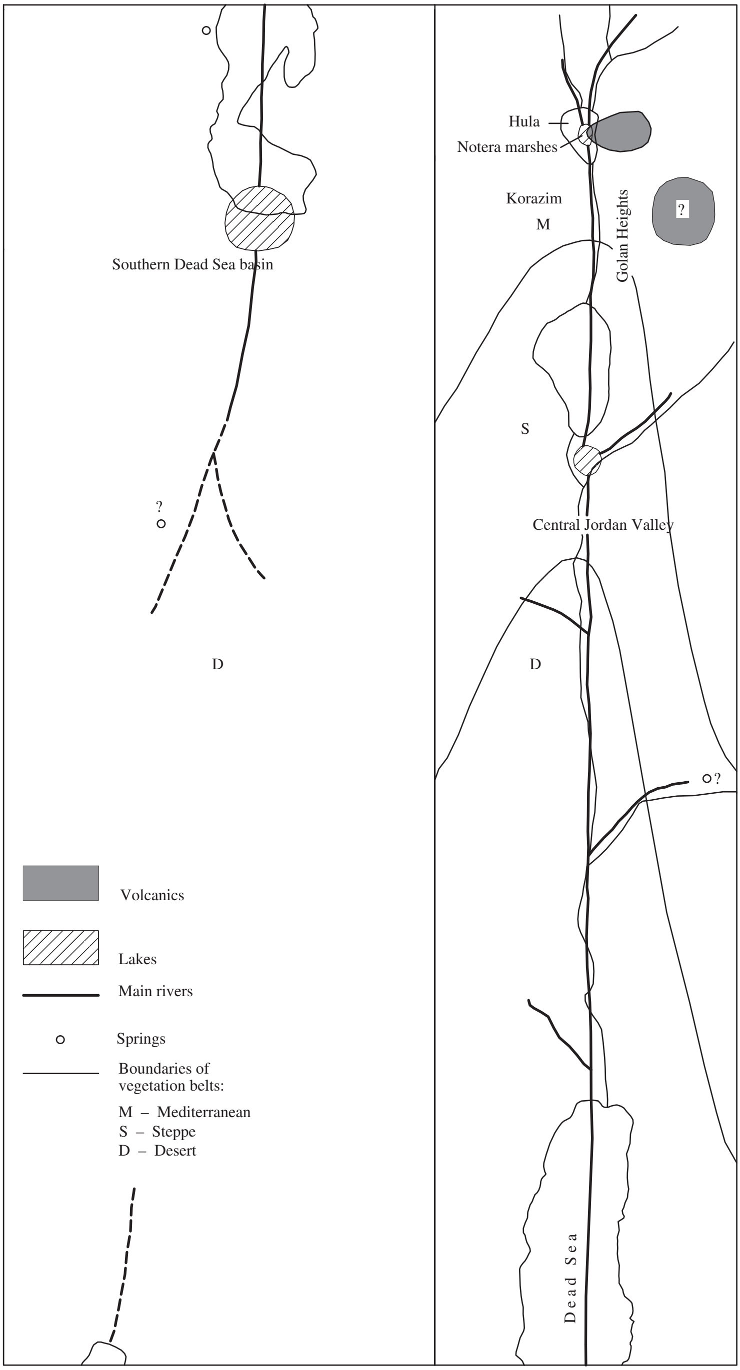


Figure 12.1.3. Paleogeography and paleoenvironments, Palynozone QIV.

12.1.4 Palynozone QV

Palynozone QV, whose sediments are known from several exposures and all boreholes drilled in the Jordan Valley, reflects a return of pluvial conditions, and thus very much resembles QIII (Fig. 12.1.4). The Hula again became flooded to the rims by a lake, depositing the Mishmar HaYarden Formation, only now the lake is connected by a river to the south, where the central Jordan Valley was occupied by the Ubeidiya–Abu Habil lake. Differences in composition of the malacofauna suggest that the Hula and central Jordan Valley were never united to house a single lake, but were always separated by the higher Korazim block, crossed by a river. An extensive lake also occupied the Dead Sea and northern Arava regions, its shores reflected by the Upper Dome Country Terraces. As in QIII, it is not clear whether the central Jordan Valley and the Dead Sea–Arava regions were covered by a single lake or two connected by a river, since no datings are available for the high lakeshore terraces in that area in the southern Jordan Valley. Very restricted volcanism is known for Palynozone QV times, in the form of the Hasbani Basalt of the northern Hula Valley and some extrusions on the Golan Heights.

The highlands continued their slow pace of uplift, being gently eroded under the pluvial conditions, forming moderate slopes at the wadi shoulders. As with QIII, from comparisons with younger pluvial periods it seems that these slopes must have been covered with gravel and colluvium, but almost none of those escaped subsequent erosion. The only exception, as for QIII, seems the Dauqara Formation at the headwaters of Wadi Zarqa, where alluvial sediments, most probably deposited during QV, are preserved (Baubron et al. 1985, Besançon & Hours 1985, Parenti et al. 1997). One could also speculate that a lake still existed at that time where Nahal Zihor is located today, in the southern Negev, but this is not satisfactorily proven.

Among all the Quaternary palynozones discussed in this chapter, QV is the richest in arboreal pollen, dominated by winter deciduous oaks. The northern sector of the Jordan Valley and its neighboring highlands seem to have been quite heavily forested, grading to steppe southward. No significant part was played by desert vegetation at that time. Apart from the pollen, other fossils indicate a substantial biomass; together with the extensive lakes and the gentle erosion, all these point to a pluvial climate, the wettest in the history of the Levantine Rift Valley. No wonder, under such environmental conditions, that people settled in the Jordan Valley and elsewhere in the southern Levant. Artifacts are known from both the Hula and the central Jordan Valley, the famous site of Ubeidiya located in the latter (Fig. 12.1.4.1) with its wealth of cultures and a few *Homo erectus* bones and teeth. The sites of Dauqara, in Wadi Zarqa, were probably also settled during QV, but during its later part, so it is somewhat younger than Ubeidiya.

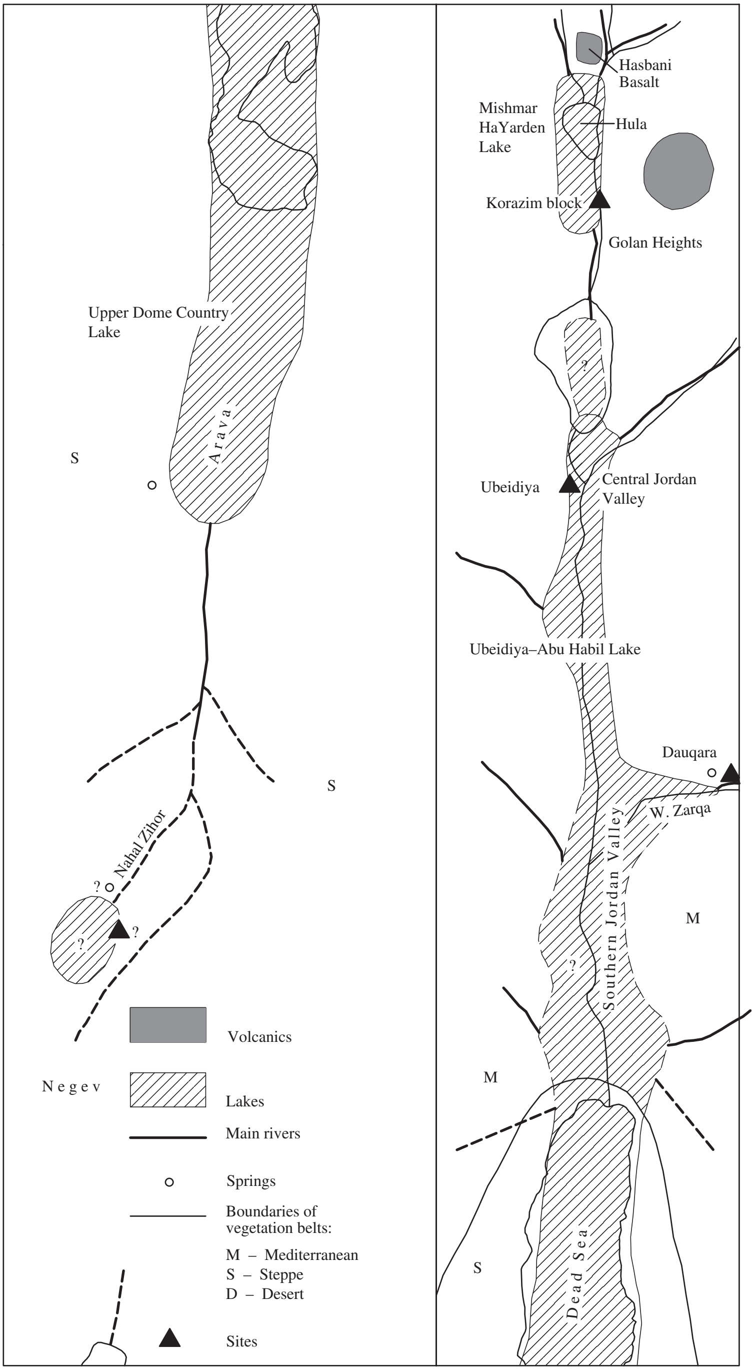


Figure 12.1.4. Paleogeography and paleoenvironments, Palynozone QV.

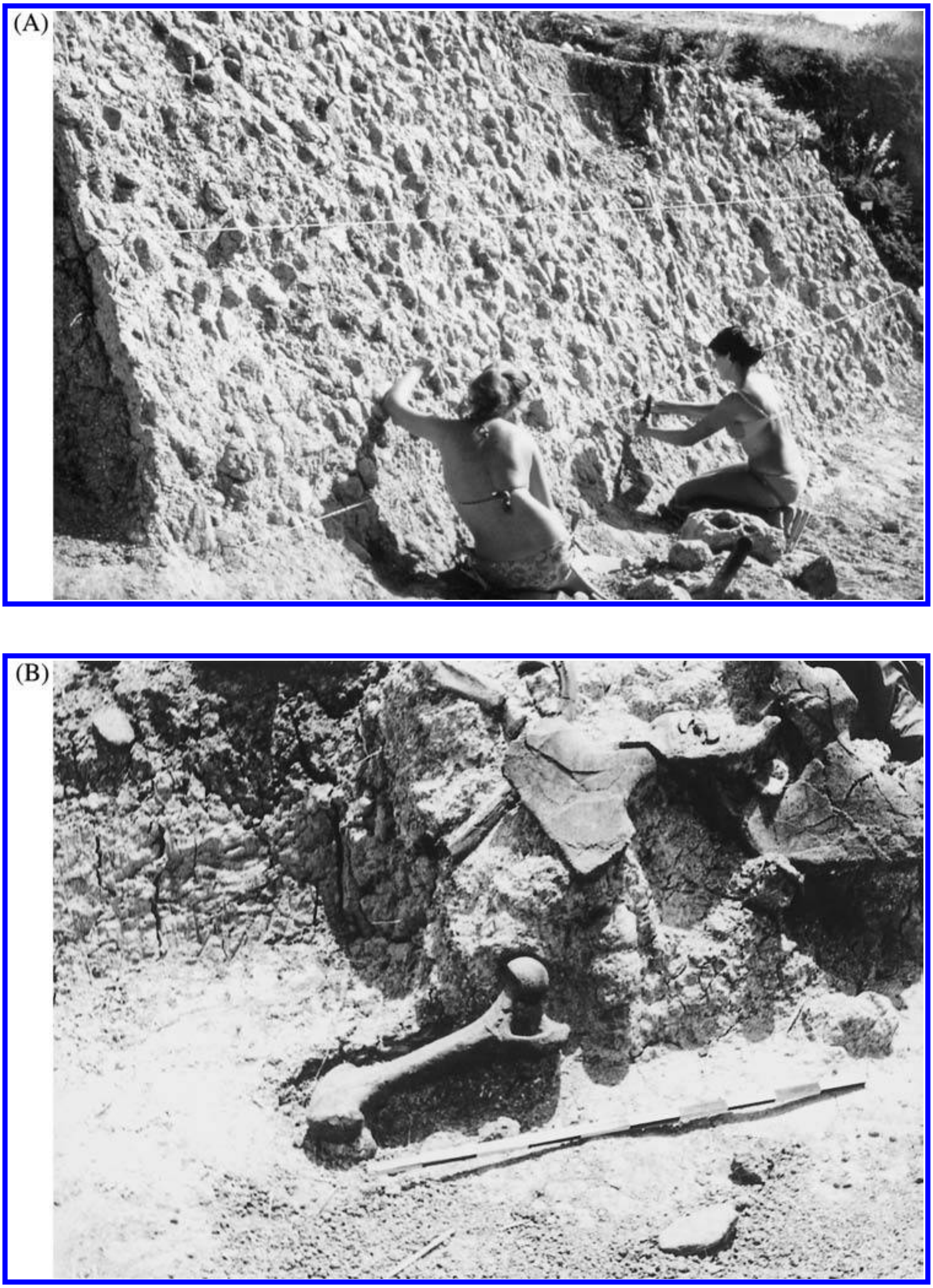


Figure 12.1.4.1. The site of Ubeidiya. (A) Excavating the living floor. Photo courtesy N. Goren-Inbar. (B) Hippopotamus bones in situ.

12.1.5 Palynozone QVI

Lakes again shrank considerably, due to a renewed onset of dry interpluvial conditions (Fig. 12.1.5). The Hula was occupied by a small shallow lake, occasionally turning into a peat bog, depositing the Ayyelet HaShahar Formation; no sediments at all are known from the central Jordan Valley for QVI times, while the southern Dead Sea basin was covered by a rather limited, saline to hypersaline lake or by playas. According to the pollen spectra, it seems that the Hula was connected to the Dead Sea by a river which laid down very little sediments on its way, if at all. A possible southward extension of the Ayyelet HaShahar lake in the Hula basin, or a local pond, deposited the sediments in which the Gesher Benot Ya'aqov site is located (Fig. 12.1.5.1), south of the bridge at Benot Ya'aqov. It is not to be confused with Jisr Banat Yaqub, a younger site located north of the bridge, within sediments of the Benot Ya'aqov Formation of QVII age (Horowitz 1973). Unfortunately the numerous bones and wood remains recovered from the Gesher Benot Ya'aqov site (Goren-Inbar 1995, 1998, Werker & Goren-Inbar, in press) cannot serve as indicators for the regional environment, being dependent on the local wet conditions.

Volcanic activity shows some increase in amplitude in the earlier stages of QVI, forming the Yarda Basalt in the Hula and northern Korazim regions, the Yarmouk Basalt on the Golan, which also reached the central Jordan Valley, and correlative extrusions along the eastern flanks of the Rift down to the Dead Sea, as well as on the Transjordanian highlands. A conspicuous faulting phase followed the volcanism, considerably contorting previous formations in both the Hula and central Jordan Valley. Thanks to these disturbances, the older formations and sites became exposed in these two regions. It seems that the faulting also hit the southern areas of the Jordan Valley, but the evidence comes mainly from boreholes indicating increased spring activity (Horowitz 1992a, p. 338), and it was much less spectacular on the surface.

The highlands on both sides of the Rift continue their slow but steady uplift in QVI times. The gentle slopes formed during the humid QV were incised quite steeply by the prevailing erosion, which most probably removed much of the previously deposited sediments. Pollen spectra display low AP shares, with oak and subordinate *Pinus halepensis*, quite abundant desert plants and median values of steppe vegetation, typical of an interpluvial, dry climate.

12.1.6 Palynozone QVII

Sediments laid down during Palynozone QVII are very abundant along the entire Jordan Valley (Fig. 12.1.6), comprising the Benot Ya'aqov Formation in the Hula, the Naharayim in the central Jordan Valley and its equivalents, particularly the Tabaqat Fahl Formation and the Kufrinja Gravel east of the Jordan River, down to the Dead Sea where they grade in the northern Arava into the Lower Dome Country Terraces. This complex is lined to the north and south by extensive spring deposits, the Kefar Yuval and Sayif Travertines. These sediments indicate that the

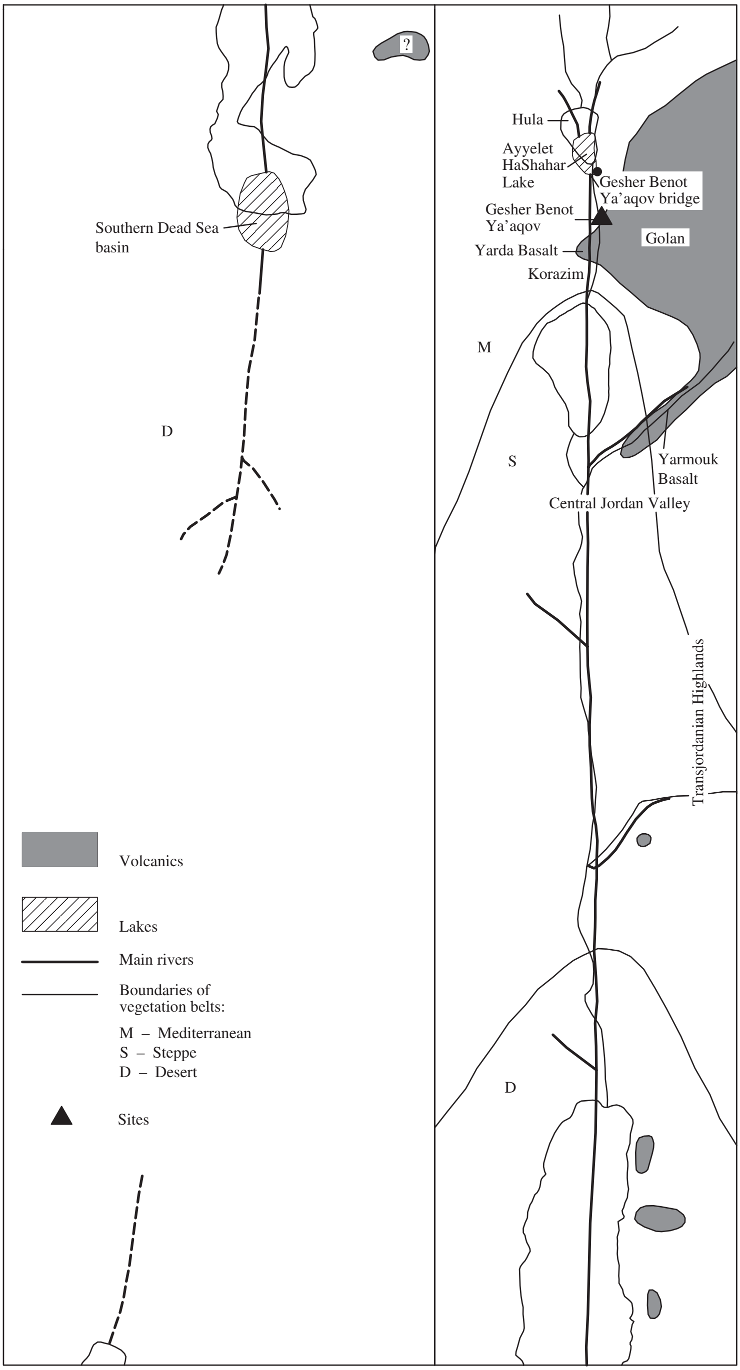


Figure 12.1.5. Paleogeography and paleoenvironments, Palynozone QVI.

Figure 12.1.5.1. Excavations at the Gesher Benot Ya'aqov site. Photo courtesy N. Goren-Inbar.

Jordan Valley was covered at that time by two extensive lakes, one covering the entire Hula Valley and the northern part of the Korazim block, the other centered in the Dead Sea area, extending north and south of the present lake.

The central Jordan Valley was occupied by a large delta with wide floodplains of the Yarmouk and Jordan rivers depositing gravel and thick soils, occasionally turning into a temporary shallow lake. It is quite difficult to delineate the boundary between this deltaic complex and the lake to the south, for lack of adequate exposures. QVII paleogeography inherited the consequences of the extensive faulting observed in QVI times, accompanied by very limited subsidence of the central Jordan Valley but increased rates in the Dead Sea (Fig. 9.6.1). These created a new landscape, with the Dead Sea occupied by a lake, while the central Jordan Valley was drained by a river system leading southward, fed also by a river draining the lake in the Hula area.

No volcanic rocks are known to interfinger with sediments of Palynozone QVII, which are covered by the Raqqad Basalt in numerous localities. If the radiometric ages obtained by Mor (1986, 1989) for the Raqqad, 0.3–0.4 Ma, are correct, its extrusion already began in QVII times. This is doubted since in many places the volcanics cover sediments bearing Upper Acheulian artifacts, which are typical of the upper part of QVII. However, since none of the dated samples was collected from such stratigraphic context, it may well be that some lava flows already

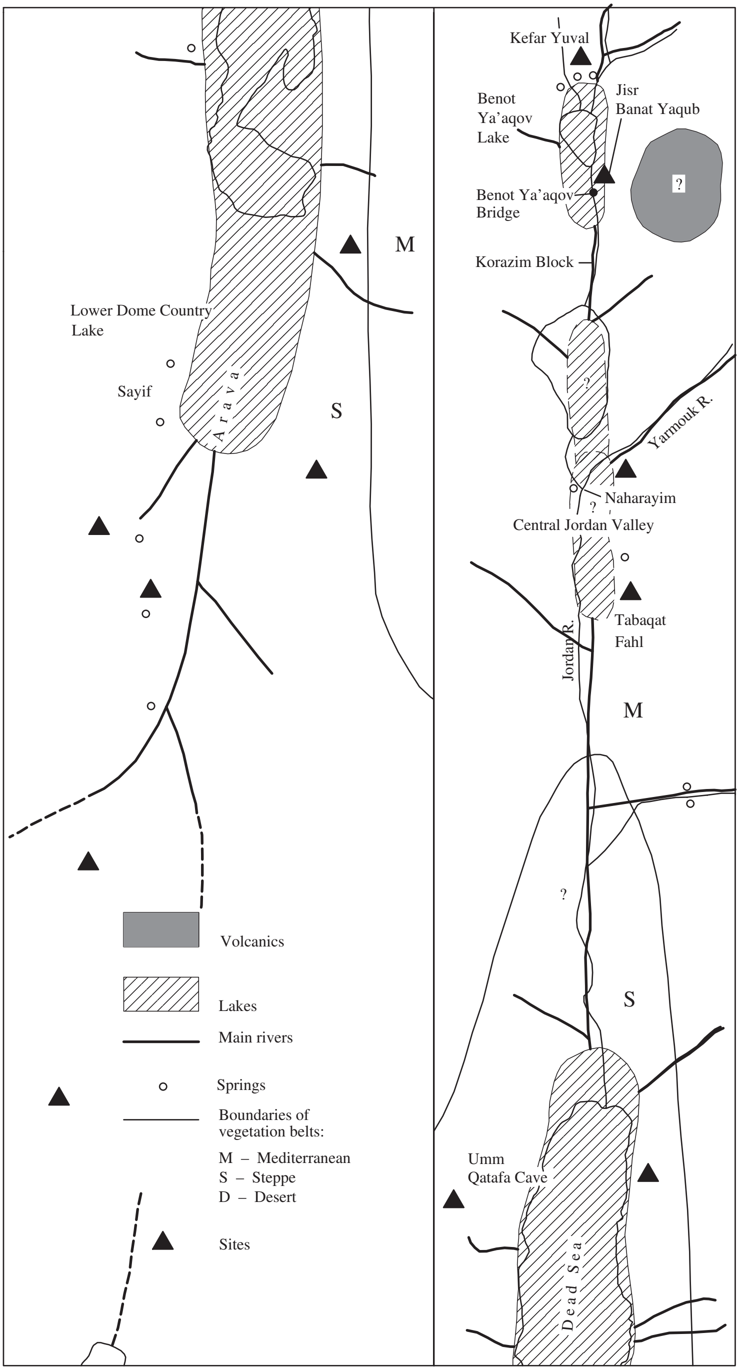


Figure 12.1.6. Paleogeography and paleoenvironments, Palynozone QVII.

extruded during QVII on the Golan Heights had indeed reached the central Jordan Valley.

Erosion on the uplifting, vegetation-covered highlands was mild, resulting in the formation of gentle slopes on which colluvium and gravel accumulated. These are termed the Baq'a Conglomerate west of the Jordan Valley, and the Bire Formation to the east. In contrast with former pluvial periods, due to the shorter time to the present, numerous exposures are still existing, in some of which Middle and Upper Acheulian artifacts and sites are located.

The pollen spectra of QVII are very rich in arboreal components, dominated by oaks to the north and a mixture of oak and pine to the south, indicating well-developed vegetation and pluvial conditions. Notable among the numerous vertebrate fossils are elephants, found in conjunction with almost all Middle and Upper Acheulian sites (Ronen 1975). These are accompanied by a variety of long-legged animals such as deer and horses, indicating a relatively flat landscape and rich pasture before the coming accelerated uplift phase (Fig. 9.6.2).

Like the entire southern Levant, the Jordan Valley was densely populated by humans during QVII times, usually in open-air sites such as Jisr Banat Yaqub south of the Hula (Fig. 12.1.6.1), Ma'yan Barukh to the north or the Tabaqat Fahl sites

Figure 12.1.6.1. The Jisr Banat Yaqub site during excavations in the 1930s. Photo courtesy N. Goren-Inbar.

Figure 12.1.6.2. The Tabaqat Fahl, an area of numerous sites.

Figure 12.1.6.3. Umm Qatafa Cave. Photo courtesy A. Ronen.

east of the central Jordan Valley (Fig. 12.1.6.2), but also in the Umm Qatafa Cave (Fig. 12.1.6.3). Analyzing the settlement pattern, Ronen (1975) suggests that this could be partly a result of the numerous exposures of relevant beds in the region, but he mentions that Upper Acheulian finds outnumber younger cultures located in similarly distributed outcrops, until the Epipaleolithic.

12.1.7 Palynozone QVIII

Palynozone QVIII, like previous dry periods, is characterized by considerable shrinking of the water bodies occupying the Jordan Rift Valley (Fig. 12.1.7). The climate effect was further amplified by accelerated subsidence of the Hula and southern Dead Sea basins, the only localities where sediments of this time were encountered, but solely in the subsurface. The extensive Benot Ya'aqov Lake which previously covered the Hula Valley turned into marshes at the transition from QVII to QVIII, depositing the peat and highly organic sediments of the Hulata Formation. It seems that the Hula was connected by a river to the southern Dead Sea basin, where correlative sediments of a small hypersaline lake were accumulated. The rest of the region was subject to severe erosion.

Volcanic activity is known for this period all along the eastern flanks of the Jordan Valley, from the Golan down to the Dead Sea, forming the Raqqad Basalt and its correlatives. Some activity is also recorded at the northern end of the Hula basin, where the Ma'yan Barukh Basalt is sandwiched between the Kefar Yuval and Dan Travertines. The source of this basalt could be within the Rift, but its extension was very limited, since it was not encountered in boreholes drilled at the center of the basin.

The highlands lining the Jordan Valley began a new phase of accelerated uplift which together with the rare thunderstorms causing floods and the rather poor vegetation resulted in steep incisions into the previous gentle QVII slopes. This process was not all bad, since it opened several new caves where people could dwell, the only localities where sediments accumulated during QVIII outside the two restricted water bodies. Of these, the Umm Qatafa and Zuttiyeh caves are the best known for their artifacts, the latter also for its skull fragments. These are the only sites of QVIII age within or close to the Jordan Valley. Pollen spectra are poor in arboreal components, indicating subordinate Mediterranean maquis to the north, grading to steppe and Saharan desert southward, subject to interpluvial conditions.

12.1.8 Palynozone QIX

This is probably the best-documented period in the history of the Jordan Valley, with hundreds of studies carried out on sediments of the Lisan Formation ever since it was designated by Lartet (1865). Lake Lisan covered the entire area from the present-day elevation of 180 m below sea level at its maximum extension, from

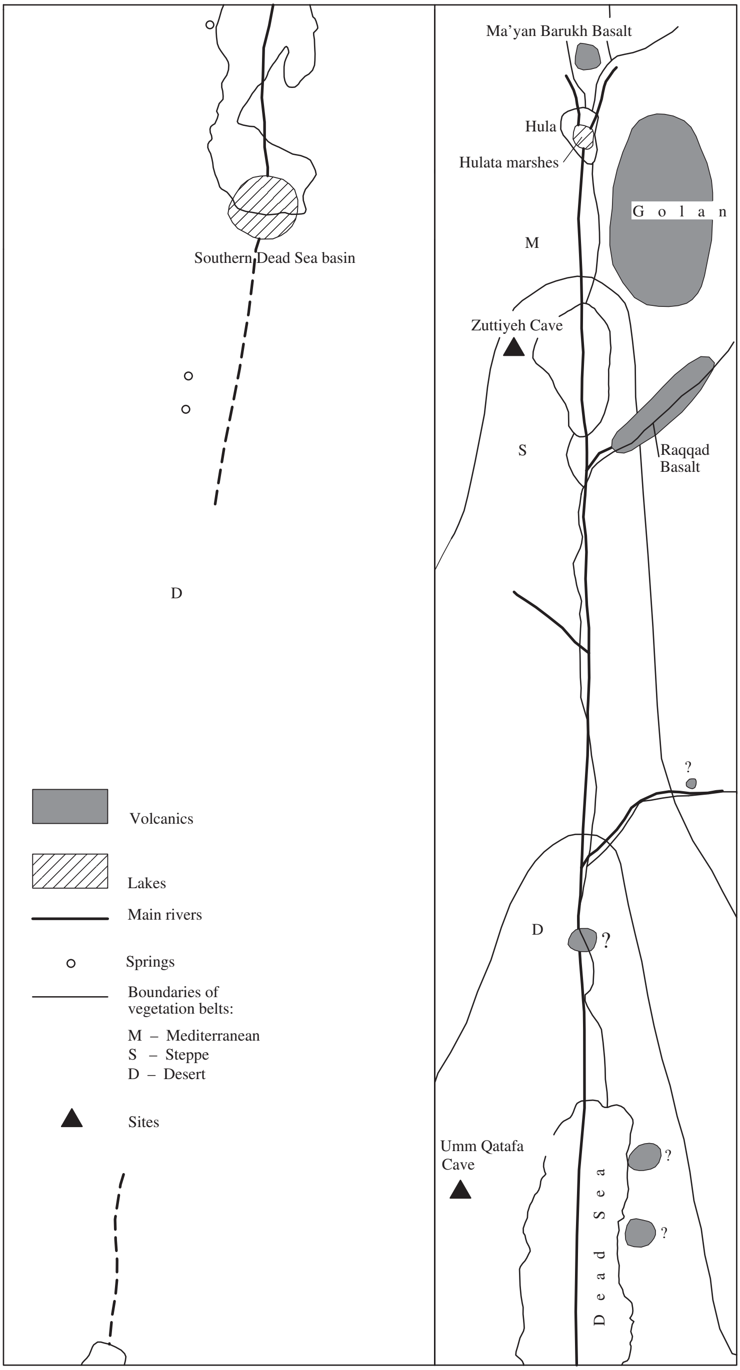


Figure 12.1.7. Paleogeography and paleoenvironments, Palynozone QVIII.

the middle of Lake Kinneret down to the northern Arava (Fig. 12.1.8.1). At the same time the Hula basin and the northern edge of the Korazim block were also entirely flooded by a lake, depositing the Ashmura Formation. The two lakes were connected by a river, their shorelines oscillating according to climate (Horowitz 1979, p. 130; Bartov et al. 2000b). This paleogeography changed dramatically some 18–16 Ka ago (Fig. 12.1.8.2), when a major faulting phase affected the Rift, creating two new troughs, the northern Dead Sea and Lake Kinneret, while at the same time considerably deepening the Hula Valley. As a result both previous lakes shrank, to give place to approximately the present-day landscape.

Palynozone QIX is characterized by three humid pluvial phases, correlated with the European glacial maxima, separated by drier interstadials. The first pluvial phase saw the advent of Lake Lisan over most of the area, depositing the Hamarmar Member some 70–50 Ka ago; during the second, 32–22 Ka ago, the lake acquired its maximum extension, laying down the Ami'az Member. The last humid phase of QIX, 16–11.5 Ka ago, following the major faulting phase, was neither long nor potent enough to fill up the previous Lisan basin again, causing only a slight extension of the new lakes formed in the northern Dead Sea and Lake Kinneret basins. Rivers leading to the somewhat extended northern Dead Sea deposited the Fatza'el Member (the Damiya or Hasa Formation in Jordan) during late QIX times, while the southern Dead Sea was fed by a river system depositing the Iddan Member. The extremely dry phase terminating QIX times, corresponding to oxygen isotope Stage 1, caused the partial drying up of the extended Dead Sea, as seen by the deposition of the rocksalt layer at the base of the Ze'elim Formation, some 13–11 Ka ago.

In the Hula Valley, the Ashmura Lake went through similar phases of climatically controlled expansion and contraction. In contrast with the Lisan, the Ashmura is hardly exposed due to the extended subsidence of the Hula basin in late QIX times, continuing also into QX. This caused coverage of the Ashmura, chiefly by thick soils and younger alluvium, which precludes any discovery of possible sites. The Tabgha Formation is being deposited in Lake Kinneret since its creation, and this lake, too, oscillates according to climate.

Besides rivers, the Ashmura and Lisan lakes were fed also by numerous springs, some depositing travertines. Travertine deposition practically stopped in most springs of the Jordan Valley in late QIX, following the faulting and creation of the steep topography (Heimann & Sass 1989), although spring outflow had not ceased and some are active until the present. The Dan Travertine in the northern Hula Valley is synchronous with the Ashmura Formation, while several travertine bodies are connected with Lake Lisan, which was also lined by numerous small freshwater seepages causing the deposition of algal tufa all along its coasts. To the south in the northern Arava, Lake Lisan was fed partly by springs depositing the Mo'a Travertine, at least during the earlier stage of the palynozone. A considerable supply of freshwater formed travertines of the Bet She'an Formation at the junction of the Jordan and Yizre'el rifts; the Fasayil Travertine occurs in one of the wadis leading to Lake Lisan from the west, and similar occurrences are also known

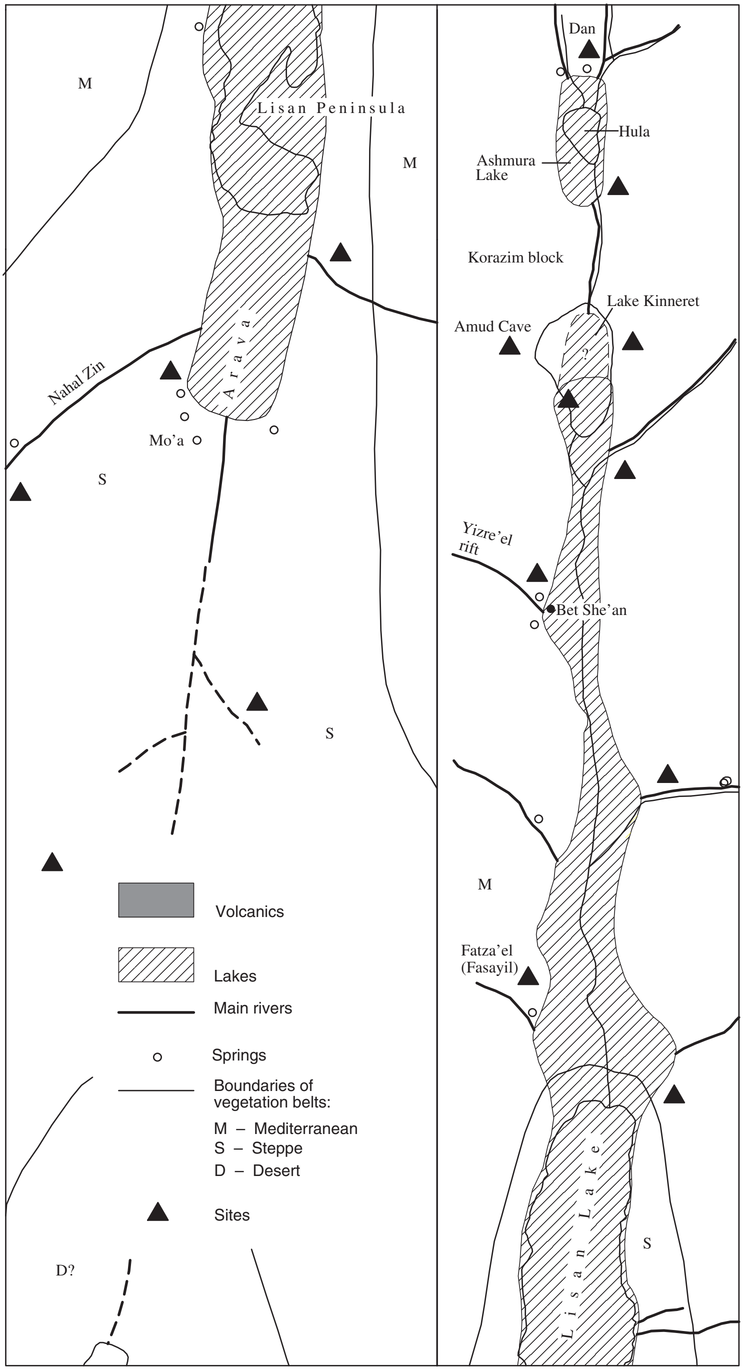


Figure 12.1.8.1. Paleogeography and paleoenvironments, Palynozone QIX, early and middle humid phases.

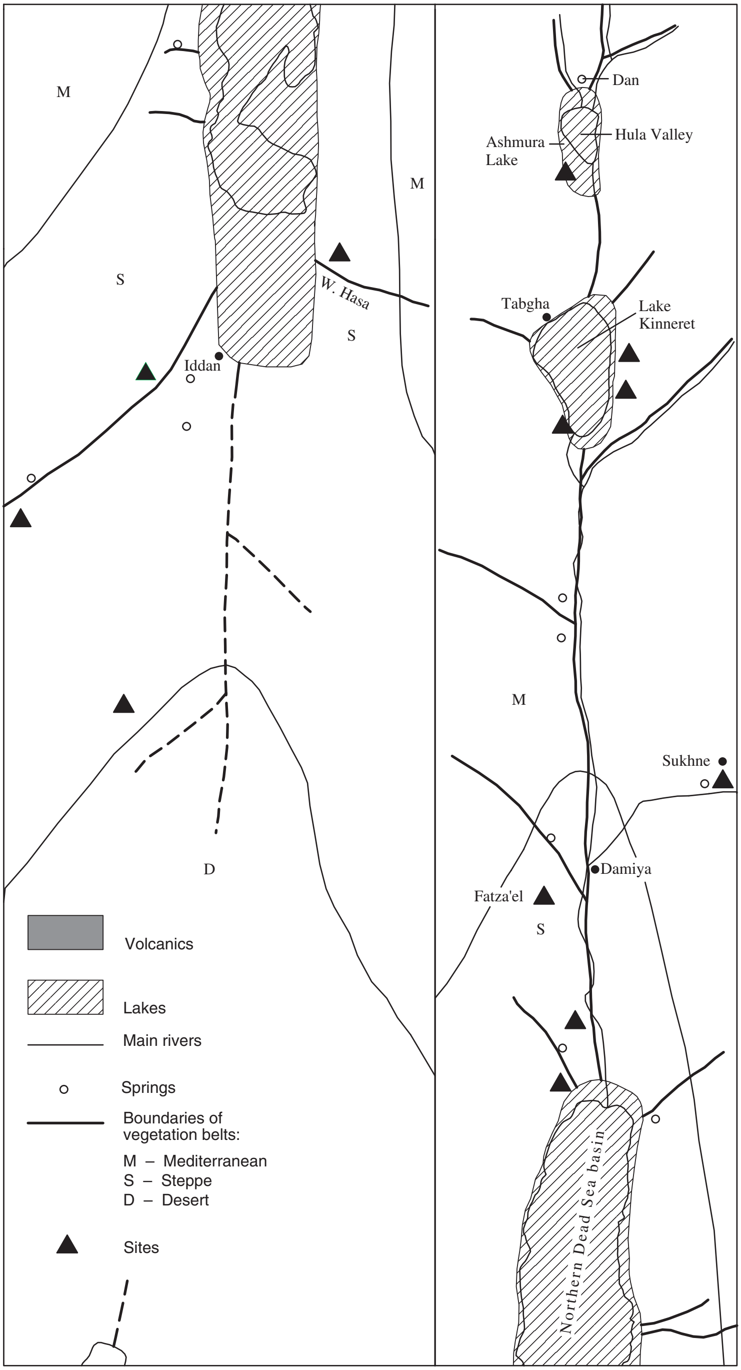


Figure 12.1.8.2. Paleogeography and paleoenvironments, Palynozone QIX, late humid phase.

from other wadis, west but particularly east of the Rift. The latter area is heavily strewn with travertines, which unfortunately have never been studied in great detail. Springs are also known from upstream reaches of some of the main wadis leading to the lake, in which travertines were accumulating, such as those from Nahal Zin. Almost all these travertines yielded Middle through Epipaleolithic artifacts and sites.

The accelerated subsidence of the Jordan Rift Valley's old and new basins was accompanied by considerable uplift of the Rift's shoulders during QIX. However, the pluvial climate and the prosperous vegetation confined incision of the wadis and rivers into the formation of gentle shoulders. Colluvium and soils silted up the channels to form the Nahshon Conglomerate west of the Jordan, and the Khirbet Samra Formation eastward, in which artifacts and sites ranging from Middle to Epipaleolithic are located. Epipaleolithic artifacts and sites are also found in connection with extended lake sediments of QIX latest humid phase. No volcanic activity is known for Palynozone QIX times.

Pollen spectra of the three wetter phases are characterized by the dominance of winter deciduous oaks, indicating some amounts of summer rains; during the drier stadials, evergreen oaks and some pines replace the deciduous trees. Man inhabited the entire southern Levant during QIX, but sites are dramatically more abundant during the wetter phases, occurring also in the present desert areas of the region (Marks 1976, 1977). Sediments of the first humid phase yielded Middle Paleolithic, Mousterian occupations; the second is characterized by Upper Paleolithic settlements, while the third is dominated by Epipaleolithic, particularly the Geometric Kebaran. For the earlier drier interstadial only a single site is known from the Jordan Valley, in sediments of the Amud Cave. This is not surprising, since deposits of this phase are encountered only in boreholes, as is the usual situation with dry climates. We are much luckier from the second interstadial on since no lakes extended to cover the relevant formations, so a complete, very detailed cultural sequence starting from the Upper Paleolithic is known from the southern Levant in general, and the Jordan Valley in particular.

12.1.9 Palynozone QX

The paleogeography at the beginning of Palynozone QX resembles very much the present day, even though the climate was drier. The middle part of this palynozone is characterized by wetter conditions with some summer rains, which caused expansions of lakes in the Hula, Kinneret and Dead Sea basins (Fig. 12.1.9). These expanded lakes suffered the same fate as Lake Lisan, their areas diminishing by faulting and deepening of the basins when the climate was still humid, some 4,500–5,000 years ago. They never attained their former size again, since the climate became drier toward the present-day. In the same manner as before, even the expanded lake sediments of the Hula are encountered mostly in the subsurface due to extended subsidence. The expanding Lake Kinneret deposited the Tabgha Formation also in the Buteiha and Ginnosar valleys; a new lake developed in the

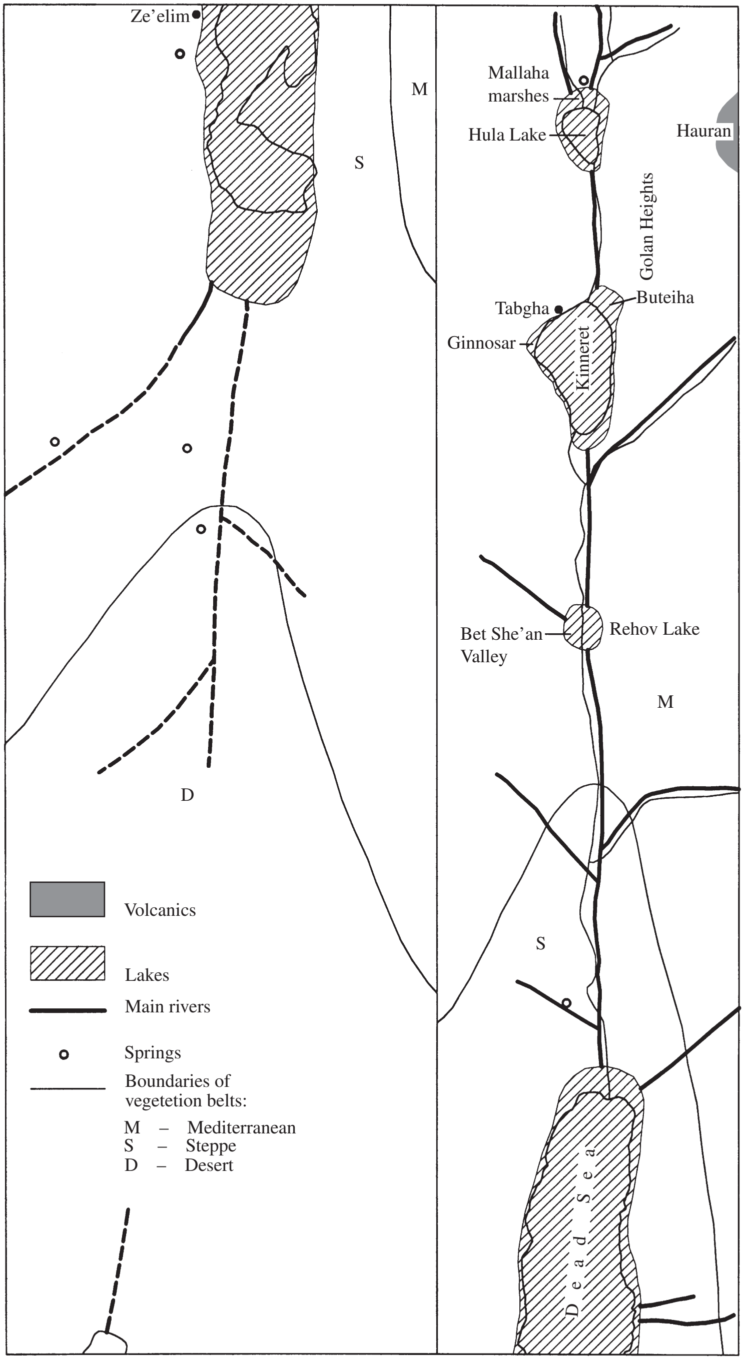


Figure 12.1.9. Paleogeography and paleoenvironments, Palynozone QX, middle.

Bet She'an Valley, in which the Rehov Formation was laid down; while the larger Dead Sea caused accumulation of the Ze'elim Formation in numerous localities landward along its beaches.

Subordinate volcanic activity is known only from the Hauran, quite far away from the Hula, east of the Golan Heights. The highlands on both sides of the Jordan Valley continued their accelerated uplift, being cut rather deeply and steeply by the wadis as a result of the thunderstorms and floods typical of the dry climates. The wetter middle part is however expressed in an accumulation of colluvium, forming the "Lower Terrace". Pollen spectra, basically similar to the present day with minor oscillations, indicate relatively subordinate climatic changes, expressed in changing shares of arboreal pollen along the sequence.

Minor as they were, the climatic oscillations considerably affected the settlement pattern in the southern Levant, as discussed in detail in Horowitz (1992a, Section 10.3.2), since economy in those days was mainly dependent on agriculture. These minor changes are, however, felt much less in the Jordan Valley, where a constant supply of freshwater was always available throughout QX times, so that habitation never ceased close to lakes such as the Hula and Kinneret or to perennial rivers or springs.

12.2 SPRINGS

Springs were always an important source of pure drinking water for animals and humans, supporting life and lush vegetation, and their importance was particularly stressed during drier periods. Indeed, almost all spring deposits in the Jordan Valley contain artifacts, attesting to the intimate connection of human habitation, scavenging, hunting and gathering food with these water sources. It seems necessary, therefore, to discuss occurrences and history of springs in the Jordan Valley and close surroundings.

The Jordan Valley itself, being considerably lower than the surrounding highlands, lies within a partial rain shadow (Fig. 3.4.1). Most of the springs are thus fed by rainwater falling on the western and eastern hilly or mountainous regions, where they fill up aquifer beds. These approach the Rift in two ways, either dipping under the younger fill or faulted against the subsiding blocks. In either case the beds are usually tilted riftward, due to the uplift of both its shoulders, which provides for a constant supply of groundwater. These emerge as springs on the Jordan Valley rims along its entire length, in two ways: small springs of relatively limited discharge, where the aquifer is exposed by erosion; and large ones with considerable water outpours, when the corresponding beds are faulted against the Rift. A minority of springs is within the Rift Valley itself, in localities of high groundwater, usually supplied by the upland aquifers as well, dipping underneath the younger in-Rift sediments. This type of spring is more common along internal faults of the Jordan Valley, particularly the active ones (Golani 1962, Horowitz 1970, Gvirtzman 1997). Springs are, and have been through the entire Levantine stage, considerably more abundant along the eastern Rift flanks, for two reasons: the larger amounts of rainfall on the eastern highlands as compared with the regions bordering the Jordan Valley to the west (Fig. 3.4.1); and the catchment area to the east is considerably larger than to the west.

The most common aquifers consist of limestones and dolomites of the Judea and Avedat groups and various basalts to the north, while sandstones prevail to the south. Consequently, almost all springwater in the region is extremely rich in carbonates. These may deposit travertines or tufas, depending on the residence time of the water at any one locality (Heimann & Sass 1989), which in most cases is a direct outcome of the topography. Thus, occurrences of travertines attest positively to the existence of springs, but the absence of these rocks does not necessarily preclude this possibility. In both cases, however, springs are characterized by their typical vegetation, so that pollen analyses of continuous borehole sequences can be used as a more reliable source of information for spring activity around and within the Jordan Valley (for details see Horowitz 1992a, p. 338).

Reconstruction of spring activity at En Gedi on the western shore of the Dead Sea (op. cit.) shows two styles of increased discharge. Normally, throughout the entire Quaternary freshwater output depends on the climate; it is higher during the pluvials, lower in interpluvials, which seems only natural. However, peaks of spring vegetation pollen also occur during the drier QII, QIV and QVI. These were most probably caused by accentuated faulting which opened up the aquifers during phases which are also known and dated in outcrops (see Section 9.6). It is, however, occasionally difficult to asses faulting phases during the humid pluvials, with their extended spring outputs. Notably, this method was also applied to older strata, indicating quite clearly the Eritrean faulting of the southern Dead Sea basin (Horowitz 1992a, p. 339). As it seems from the pollen diagrams, springs always existed in the Jordan Valley throughout the entire Quaternary, their output varying according to climatic and structural conditions.

Instances of travertine exposure along the Jordan Valley depend on two factors, whether they were deposited in the first place, as discussed above, and if so, were they at all preserved. Being mostly confined to wadis, the fate of travertines follows very much the terraces (see Section 11.1). Thus travertines older than QVII are not securely known, although they may exist east of the Jordan River or the Dead Sea. Nor is there evidence of those travertines formed during the interpluvials, except for a rare occurrence in the northern Arava where the rather flat landscape helped preserve some. Most known spring deposits are restricted to the interval starting with the pronounced faulting of QVI which exposed aquifers, up to the quite recent faulting at the end of QIX which, although further exposing the aquifers, emphasized topography to such an extent that deposition of these rocks ceased almost completely.

An occurrence particular to the Jordan Rift is the warm saline springs common along the Valley from Lake Kinneret southward to the northern Arava. Their appearance is always connected with faults, and more often than not with intersections of these. These springs are fed by meteoric waters which flow down to a depth of 1–2 km below the Rift's floor through the dipping aquifers, where they encounter and dissolve buried salt bodies, usually of Pliocene age, while at the same time being heated by the considerable depth. The hydrostatic pressure within the aquifers raises these waters to the surface through any of the numerous faults of the Rift, usually at the margins but occasionally also at the center of the Valley (Horowitz 1970). Such springs are not known north of Lake Kinneret, which is the northern limit of deposition of evaporites during the Neogene.

Not too many travertines were left by saline springs but some, such as the youngest in the Arava, show clear signs of being deposited in brackish waters. The effects of salinity on the Dead Sea lakes is minimal, since these were in any case highly saline throughout the entire period from QII times on. The principal effect is thus felt in the central Jordan Valley lakes, whose increased salinity affected numerous life forms, so that, as only one example clearly shows, their malacofauna is pronouncedly different from the predominantly freshwater Hula (Rosenthal et al. 1989, Schütt & Ortal 1993) for most of the Quaternary. Brackish foraminifera such as *Ammonia* inhabited the Ubeidiya Lake at some stages (Almogi-Labin et al. 1995). The activity of saline springs in the Levantine stage can be followed by the pollen spectra, which contain redeposited Pliocene sporomorphs (Horowitz & Horowitz 1990), typically carried by the spring water (Horowitz 1970).

This increased salinity of the central Jordan Valley lakes does not seem to have affected Man directly, since it never attained values that hindered life in any way. The only use I can think of that humans could make of the saline springs, is bathing in the warm water.

12.3 SETTLEMENT PATTERN

Any discussion of the settlement pattern of humans in the Jordan Valley and surroundings suffers from two major limitations. The first applies to most of the Quaternary, save for its youngest part, and concerns the almost total lack of sites during the dry interpluvials. The second concerns environmental conditions, which are almost ideal along the entire Jordan Valley, including its drier southern parts. Such ideal conditions seem to have existed throughout the entire time span, some two million years, of human presence in the southern Levant, regardless of climatic or structural changes. It thus seems immaterial to define any specific settlement pattern for a given period, when indeed almost any locality within the valley will do. This paradise is a direct result of the wealth and ample availability of the freshwater supply, persisting throughout the drier periods, and the convenient temperatures all through the year. Needless to say, the pluvial climates provided even better living conditions, the somewhat lower temperatures usually being a blessing in this region.

These two main factors, ample water and convenient temperatures, affected wildlife, vegetation and humans in a favorable way, providing for a rich, varied food supply, which could be obtained with no excessive effort. But there are naturally also drawbacks connected with this ideal environment, particularly parasites and diseases, which flourish under the same conditions so suitable for the fauna and flora. Typical parasites very common in the Jordan Valley until quite recently, include those causing bilharzia, leichmaniasis and malaria, while other dangers may include venomous snakes, scorpions, leeches and so on.

Since the natural environments do not seem to be a limiting factor on settlements patterns, it appears that we have to look at other reasons to explain the pattern of location of sites, as indicated by finds of human relics. Among these, sedimentological and erosional factors may be the crucial ones pertaining to prehistoric sites, at least for the period from QII up to and including QVIII. A detailed discussion of these processes is given in Section 11.1. In contrast with prehistoric, the locations of archaeological sites are affected by different criteria, mainly cultural and economic, dictating their settlement pattern. Wars, principally, seem to control their destruction, aided to a lesser extent by natural forces such as earthquakes and erosion.

Stratigraphically, prehistoric sites of the Jordan Valley are located in several principal domains. They are either embedded within discrete rock formations, which includes both open air and cave sites, or comprise dwellings found on the surface as fully developed sites or only as scattered finds. Such scattered finds are also known from certain well-defined deposits, but hardly ever from caves. The rock formations are those common for the Jordan Valley, with sites known from lake and fluvio-lacustrine sediments, river and wadi colluvium and gravels, and travertines. Common to all these sites is the intimate proximity to water, either lakes, streams or springs. This is not surprising, considering the wealth of water, but could also be a result of the availability of sediments.

The oldest sites and artifact find spots are located in fluvio-lacustrine sediments, for purely geological reasons. These involve the wide distribution of pluvial lakes combined with the younger tectonic activity, particularly at the end of QIX and middle of QX, that exposed these formations in areas such as south of the Hula or the central Jordan Valley. The relatively short time since the exposure did not let erosion wipe these sites out. Such are Erk el Ahmar of QIII, Ubeidiya of QV and Jisr Banat Yaqub of QVII palynozones. The only exception seems the site of Gesher Benot Ya'aqov, embedded in fluvio-lacustrine sediments dated to the beginning of the interpluvial QVI times. The site, incidentally, was not exposed naturally but was discovered during bulldozing activities to widen the River Jordan channel. The continuation of these works demolished this important site in a most vandalistic way.

The last site was only recently discovered, while the pattern of all others gave the impression, held for a long time, that people inhabited the Jordan Valley only at times of the more humid pluvials. Interpluvial sites are known to the north, from the Lebanon and Syria, so the common theory was that during the dry periods people were seeking a more habitable environment, pushing northward to the wetter lands. The cardinal question thus remains open: is the theory of wandering in the footsteps of favorable environments correct, or is the settlement pattern apparent, merely resulting from the fact that no interpluvial sediments are exposed? If the second approach is correct, people could settle the Jordan Valley also during the dry periods, but since their sites were close to the lakes they were subsequently buried under the next phase of lake expansion. This assumption is strongly supported by the fact that artifacts and sites are indeed found whenever interpluvial sediments are rarely located, such as at Gesher Benot Ya'aqov, in some caves or in sediments following the last pluvial period.

The same situation holds for sites located in connection with rivers, wadis and springs, whose sediments exclusively represent the pluvial periods. With these there is a further limitation, namely that older sediments are almost never preserved due to subsequent erosion, which was quite strong on the higher terrains. Thus no such sediments are known at all for the pluvial Palynozones QIII and QV west of the Jordan, while to the east possibly (but not entirely proven) the "Uppermost Terrace" at the headwaters of Wadi Zarqa may represent QIII. Similarly, only a single occurrence could safely be attributed to the later part of QV in the same locality, where the site of Dauqara is embedded. Younger QVII and QIX river, wadi and spring deposits are much more abundant, encompassing numerous prehistoric sites and find spots. However, these too represent only pluvial periods, so that again no interpluvial sites connected with rivers, wadis, or springs, are known around the Jordan Valley. It is of course possible that if these regions were settled at all, any remains were subsequently washed away by erosion, together with the sediments.

The situation with caves is entirely different, since these locations escaped erosion of interpluvial sediments, fully or in part, and thus contain sufficient evidence for settling by humans. Yet again older caves hardly withstood erosion, so that positive evidence only comes from Palynozone QVIII onward. Besides Gesher Benot Ya'aqov this is the only proof that people did indeed settle the region also during the dry interpluvials, except of course for the Holocene, for which ample evidence is at hand. It is still debatable whether caves were used as shelter for people forsaking the hot and dry Jordan Valley, thus serving as a sole habitat, or, as seems to me more reasonable, humans still lived along the streams as during the pluvials, only that no physical evidence for that remained due to erosion.

Prehistoric sites embedded in paleosols are quite common in both the eastern and western highlands bordering the Jordan Rift Valley. Almost without exception, these are red-brown, deep-profiled, well-washed soils, a type which was formed only during the pluvials. No soils older than Palynozone QVII are known for sure, nor are interpluvial soils recognized, except for a very small outcrop only a few meters long of a thin gray paleosol horizon at the bridge of Benot Ya'aqov.

No artifacts were found at this spot, which is too restricted in area to be of any significance when negative evidence is concerned. As in all other environments, paleosols provide a similar type of information, namely that humans inhabited the Jordan Valley surroundings only during the wetter periods. An exception to that rule is Palynozone QX, during which these regions were settled, but one must take into consideration the agriculture of this period, which could have made them more attractive.

It must be stressed that even during the driest periods of the Quaternary, as seen by the pollen spectra (see Chapter 6), water and vegetation were sufficient to support both animals and humans. The idea of deserting the Jordan Valley and its neighboring regions during the interpluvials in search of better habitats seems far-fetched to me. The lack of sites is thus attributed here to the corresponding absence of sediments, accompanied by the severe erosion characteristic of the dry climates.

The basic settlement patterns of historic periods are quite similar to the prehistoric ones, namely attachment of most habitations to water sources. In addition, some other considerations apply as well. First comes the increasing need for arable and pasture lands, which became more and more pronounced following advances in agriculture and animal husbandry, together with an increase in the numbers of inhabitants. Second came the need for exploitation of raw materials, such as copper, salt, sulfur and asphalt, common in the southern, drier parts around the Dead Sea and at the margins of the Arava. This made people settle close to the quarries or gathering areas, occasionally necessitating the import of food or even water. The third aspect is commerce, which prompted two main types of settlements. Large urban habitations which served as commercial centers, usually erected close to water sources, with at least limited areas suitable for agriculture; and small guard points built at crucial locations on the main commerce roads, which in many cases depended on imported supplies, occasionally even water.

12.4 SUBSISTENCE

The Jordan Valley is a land of plenty, even its southern, drier areas. Freshwater is always there, catering for a variety of plants and animals. The two principal biotopes, the Rift lowlands and the marginal elevated terrains, are always represented in the faunal lists and the pollen spectra obtained from sites within the valley. There is however a problem with documentation of the vegetal parts of human diet in sites, due to the fact that these are much less preserved in the region. Also, some of the plant remains may represent firewood, building or matting materials, and not food. The pollen assemblages could thus give an idea about the composition of vegetation, but cannot provide any information about which plants were actually used for food. Fossils found in excavations suggest that throughout most of the prehistoric times, humans mainly consumed meat from a variety of vertebrates and possibly some mollusks.

Here, naturally, we come to the gray area of assumptions: is it possible that meat was humans' main staple food? The answer could be positive, but there are doubts since we are dealing with negative evidence. Looking at a variety of peoples living in the last centuries in areas rich in both game and vegetation, such as the American Indians, it is clear that wild plants, in addition to those intentionally grown, constitute a considerable part of their diet (Niethammer 1974 and references therein). The question still remains whether an analogy is valid, which is not necessarily the case.

An example of what we may have missed in terms of vegetal food, due to the bad preservation conditions of sites, comes from Ohalo II, an Epipaleolithic settlement dated to 19,000 years BP, located at the southern end of Lake Kinneret. The place was settled during the short dry phase corresponding to the Last Glacial Maximum, when Lake Lisan shrank. Following this brief interlude the area was again submerged, so that excellent preservation conditions were created. The excavations revealed (Kislev et al. 1992) a wealth of vegetal remains that could have served for food (and most probably did), such as 11 species of grains, nine species of edible wild fruits, and eight species of other edible wild plants, in addition to numerous bones.

Besides the poverty in vegetal relics in most other sites of the Jordan Valley and surroundings, another problem still remains in trying to assess the diets of prehistoric (and sometimes even archaeological) cultures. Putting it simply, were the fossils actually found in sites used for food? An extreme example of this complexity comes from the famous Neolithic site of Eynan (Perrot 1960) in the Hula Valley, where a joint burial of a man and a dog was discovered. People born into what is termed "Western Civilization" maintain that there could be no doubt the dog was kept as a pet; but this is not the case in other parts of the globe, where dogs are reared for food and even regarded a delicacy. We know of numerous cultures that bury the dead with food supplies for the great journey ahead; could this not be the case at Eynan? I know it sounds preposterous, but still, despite our education and tradition, this possibility would not seem farfetched to someone from southeast Asia.

The same problem applies to the small rodents such as mice and rats whose bones and teeth are so common in most prehistoric and archaeological sites in the southern Levant (and elsewhere, for that matter). Again, we usually regard these animals as pests, but this time one does not have to wander far. I wonder how many admirers of the French cuisine know the recipe for "Grilled rat Bordeaux style" (Entrecôte à la bordelaise), the Canadian "Baked Muskrat", the American "Cajun Muskrat", the ancient Roman "Stuffed Dormice", or the Mexican "Roasted Fieldmice" (Raton de campo asado), detailed in Schwabe (1994, pp. 204–208). In fact, leafing through Schwabe's highly instructive book, one quite soon realizes that almost every animal is considered edible even now, in some place around the globe.

In terms of food acquisition, for which matter the Jordan Valley shows no particular characteristics over most other regions worldwide, the history of mankind could be divided into "passive" and "active" stages. The passive stage involves, in terms of time, almost the entire history of mankind until and including the Middle Acheulian. During this period people must have been primarily sedentary, as can be seen from the size of large, discrete sites such as Ubeidiya, the Dauqara complex, Gesher Benot Ya'aqov, Jisr Banat Yaqub, Quneitra, the Dishon–Yir'on complex and Ma'yan Barukh, while occasional find spots are quite rare. Bones recovered from these sites comprise the entire animal spectrum, from the tiniest mice and birds up to hippopotami and elephants.

It seems, however, that there is no definite proof for the widely accepted assumption that humans were mainly scavengers, feeding on previously hunted or otherwise dead meat. Judging from artifacts, it appears to the modern eye that Olduvan, Lower and Middle Acheulian ones were generally used for butchering, rather than hunting. The only exception may possibly be the large spheroids common at Ubeidiya (Bar-Yosef & Goren-Inbar 1993), a tool the likes of which are presently used for hunting by some African tribes. I am afraid that again, our concepts may not necessarily be in line with those of early humans, who could have used tools in a different way than we tend to think.

A staple always consumed by people in the Jordan Valley is fish, which indicates at least some activity by humans. Of these, the most common in all excavations are bones of the local catfish, *Clarias*, which is very easily caught by hand. Catfish live in shallow, usually muddy environments, and are neither too fast nor too bright. The question is whether with such characteristics its catching could be defined as "active" or falls under "gathering", but this seems to be only semantics. It is quite clear that no artifacts, no specific tools, had to be developed for this action. The story may, however, be different with faster, open-water fish, whose bones have also been recovered from excavations, and which would need some tool to catch.

It is only reasonable to assume, but with merely a single definite positive proof from a younger culture, that vegetal material was also consumed by inhabitants of the Jordan Valley. In this respect, it is notable that hardly any roots or tubers of considerable nutritional value are present in the Jordan Valley region. On the other hand, pulses are plentiful, as are cereals and various fruits (see Section 3.6).

The second stage in the food supply seems to comprise active hunting, possibly with the aid of stone tools, starting some time in the Upper Acheulian. These tools, or some of them, may have been connected to wooden shafts, an assumption which has no support in the archaeological finds of the southern Levant. This lack of evidence may result either from the general bad state of preservation of vegetal material, or may be because wooden handles were indeed not in use. Wooden shafts are known from some rare prehistoric sites of different cultures, particularly in places of extremely arid climate such as the Atacama Desert in northern Chile, where they are well preserved. Fine examples are displayed in the small museum in San Pedro de Atacama.

The idea of active hunting comes from two phenomena, the increasing slenderness of stone tools which makes them lighter to carry and more appropriate for hunting, rather than mere butchering, and the extensive distribution of find spots not connected with any particular site. Upper Acheulian and Middle Paleolithic artifacts are indeed found everywhere, suggesting that people went on hunting trips, losing or leaving their tools behind in all places, probably after the hunting and initial butchering was carried out. It seems that on their way back home they did not carry any surplus weight, particularly tools that could be easily manufactured once they were back at camp.

If this scenario is true, it may have necessitated an additional instrument, namely the invention of a skin bottle or something similar to carry water along. This does not seem so crucial in the Jordan Valley, particularly in its northern reaches, but may have become indispensable in the more arid southern regions. Indeed, the wide distribution of Upper Acheulian find spots in the Arava, Negev and Sinai dry lands, not connected and sometimes quite far away from dwelling sites or water sources, seems to lead to the conclusion that they did use some water-carrying device.

The active hunting stage, most probably accompanied also by gathering of vegetal food, lasted at least 100–150 Ka, during which the implements gradually improved, their size diminishing while the shapes became more and more specific and efficient. It appears that the number of hunted as against scavenged game increases as the prehistoric cultures progressively advance with time, since the bones of smaller animals such as gazelles, ibex and fallow deer seem to be more common in the younger sites (Tchernov 1980). This is far from being a definite proof, because it could merely represent a natural evolutionary trend in the composition of the local fauna. Another factor should also be remembered, that in the course of many earlier excavations, particularly before World War II, no systematic search was conducted for the smaller bones. Archaeologists were much happier to find elephants rather than mice. Another problem concerned with the earlier excavations is that no counts of the bones have been published, so that it becomes quite impossible to discuss diets.

The next development involved plant and animal domestication, husbandry and agriculture. Again, even though the termination of the Pleistocene saw the climate drying out in the entire southern Levant (see Chapter 6), there seems to be no shortage of appropriate environments or of food and water for these activities along the Jordan Valley, even in its southern reaches. Continuous settlement of this region is recorded for the last part of Palynozone QIX and the entire Palynozone QX, a period typified by the advancement of agriculture (Horowitz 1992a, Section 10.3.2, Stern 1992, Gebel et al. 1997, Henry 1998).

To sum up, it appears that the Jordan Valley afforded suitable biotopes, environments and sources of subsistence for humans, providing for continuous settlement throughout the Quaternary. The gaps observed in the chronostratigraphy of prehistoric and some historic cultures are here blamed on geological processes, rather than on human behavior or preferences.

12.5 CONCLUSION: THE JORDAN RIFT VALLEY, A PROPAGATION HUB FOR EARLY MAN?

The primary question that has to be answered, before even touching upon the role played by the Jordan Valley in the evolution and dispersal of people, concerns the source area of this species. Did *Homo erectus* indeed originate in East Africa, as is the common belief (Bar-Yosef 1994 and references therein), or somewhere else, as is hinted at by the consistent accumulation of amounts of evidence? The main difficulty seems to emerge from two disturbing sets of data. The first concerns the dates of appearance of *Homo erectus* in various localities around southeastern and western Asia, as well as southern Europe, all clustering around 1.8 million years, which is also the time this species appears on the East African scene. The second pertains to the vast areas separating the latter region from these, areas in which no clues for *Homo erectus* have hitherto been found.

These two problematic points can be (and are) easily explained by proponents of the East African cradle theory. The fact that Man appears in all these places at approximately the same time (with the emphasis on "approximately"), could result from rather fast migration, during a time span too short to be detected by the accuracy of the available dating techniques. It is quite clear that if this process of dispersal and colonization took even 100 Ka, an adequate interval by all means, radiogenic ages are not sufficiently refined to tell the ages of sites apart. Both the Nile Valley and the Red Sea coastal plain, the most plausible routes connecting East Africa and the Levant, were prone to alternating erosion and deposition phases during the Quaternary, with the first usually having the upper hand under the prevailing dry environments, so that sites which could have been on a migration route were wiped out. Another cause of the lack of information may lie with the sad fact that these regions were never properly surveyed, for political and logistic reasons.

It appears that, in the present state of knowledge, the East African provenance of *Homo erectus* can be considered highly plausible, despite the open questions. As for the trek northward, both the Nile and the Red Sea routes seem convincing. However the first could have the advantage of a constant water supply, while sources along the Red Sea were not plentiful, and were located quite far away from one another. Either of these routes would bring humans eventually to the southern Levant or, if one turns left at the coast of the African continent, to North Africa. It only makes sense to assume that both directions were in use. Incidentally, if they did not swim (see below), both the Nile and the Red Sea routes would bring people to the same starting point for the southern Levant, namely northwestern Sinai.

The one thing which is always stressed in this context is the necessity for a "land bridge" to make human dispersal possible. This was for a long time the rationale behind the idea of the southern Levant, and particularly the Jordan Valley, as the only route to settle the world after getting away from Africa. The idea of being Europe and Asia settled by crossing the Gibraltar, Sicily or Bab el Mandeb straits was occasionally raised. The question that has troubled me for a long time, for which I have no concrete answer, is whether that "land bridge" was indeed a necessary requirement. One thing is quite clear in this respect, namely the occurrence of *Homo erectus* on Java some 1.8 Ma ago (Swisher et al. 1994), which would most probably have necessitated crossing the sea.

Truly, we cannot suggest a way by which humans could have crossed the Mediterranean or the Red Sea. But on the other hand, it is quite clear that less intelligent animals (and plants) do it all the time to settle islands. This relates not only to those which can fly or swim, but seems a general phenomenon. It is quite astonishing to observe the tempo by which newly formed islands, or those completely destroyed by devastating volcanism, such as Krakatoa, are settled by all kinds of plants and animals. This is not the place to discuss the process of island settling in any more detail, but it seems worth mentioning.

Back to the southern Levant, where presumably early Man arrived after the long trek along the Nile Valley (or the Red Sea coast?) approximately 1.8 million years ago. Again, there appears still another reservation here, because the entire picture would change if the chipped flints and bones at Bethlehem prove to relate to human activity, or if the age of the Yir'on artifacts is confirmed (see introductory remarks to Chapter 11). These two questionable findings, if verified, are located in sediments of Palynozone QI, undoubtedly older than 1.8 Ma. Even so, these enigmatic occurrences do not necessarily represent *Homo erectus* activity, but could as well be produced by an earlier type of hominid, which would have arrived from somewhere anyway. Or could they have originated in this part of the world, as already suggested (Repenning & Fejfar 1982)?

The basic idea implies (Bar-Yosef 1994, Fig. 1) that when they reached the Mediterranean, at least part of the population turned right, toward the Levant, where they settled. This however was not for long, since people continued to wander northward, splitting again into two streams, to settle Asia and southern Europe. Bar-Yosef further suggests that this first "sortie" from Africa was followed by several others, in approximately the same style. These ideas are based mainly on the rather lengthy interruptions between the ages of the Acheulian sites, particularly those from the Jordan Valley, and their "African" affinity. Time intervals between the better known settlement periods are in the order of hundreds of thousands of years.

While this fine theory was developed, several facts were neglected, ignored or otherwise underestimated (see Section 11.4). The site of Ubeidiya, disregarding whether its age is 1.4 Ma (Tchernov 1987) or somewhat younger as suggested here, is still heralded by Bar-Yosef and others as the oldest in the southern Levant.

As already shown by many, artifacts considerably older than Ubeidiya are embedded in sediments of Palynozone QIII in at least two localities in Israel, and stone tools of QIV age are found in the Lebanon. So if there indeed was a first "sortie" of *Homo erectus* from Africa it had to predate 1.8 Ma.

The second point regards the temporal distribution of Acheulian settlements. If the Jordan Valley alone is discussed, the time intervals between the sites are indeed considerably long; but when the entire southern Levant is looked at, the actual intervals are much shorter. In fact, as discussed in Section 11.1, the site distribution pattern closely follows exposures of relevant rock formations. It is quite obvious that interruptions of the sequence of settlement ever since Palynozone QII relate to periods of erosion. This brings second thoughts as to the necessity of more than a single wave of humans marching out of Africa to explain the settlement pattern of the southern Levant. Clearly, the geological reasoning advanced above in no way precludes the possibility of several "sorties". It is quite possible both that Ubeidiya being regarded as the earliest occupation of the Jordan Valley, and the extremely important role given to this region in the history of Man, result from good public relations, which may have inspired and made widespread some misleading ideas.

Was then the Jordan Valley the sole route by which humans could have wandered from Africa to conquer the world? Perhaps a book dedicated to this region should answer this positively, an idea that I personally have nourished for a long time, but the answer is most probably negative, for several reasons concerned with the stratigraphy and geology of the southern Levant. There are numerous Acheulian sites and find spots along the coastal plain, covering the same time span as those in the Jordan Valley, and the latter is not alone in offering suitable environments for habitation, especially during the pluvials. The question remains as to the highlands, in which sediments older than QVII are hardly ever found. However, it is instructive that whenever appropriate deposits are found, they do contain artifacts and sites. In the only case where QV sediments were located, the Dauqara Formation on the Transjordanian Highlands, it also yielded a rich assemblage indicating that this region was settled by humans. The one thing to remember is that the prevalent uplift on both sides of the Jordan Valley is quite young, so that elevation did not hinder people from settling there during most of the Quaternary.

Subsistence was widely available on the hills, with probably less pests and more agreeable weather; springs are plentiful even today, in a dry period, while shelter was available in caves and rockshelters. It is my belief that the limited number of sites known from the now elevated highlands results from the subsequent uplift, followed by intensive erosion which wiped out most evidence.

Similarly, the present-day desert areas of Sinai, southern Israel and the Kingdom of Jordan could also have been easily traversed or even settled by humans. Again, wherever the appropriate sediments are present in these largely eroded regions, artifacts and sites are found. However, as in the hilly areas, only pluvial deposits are preserved. The question remains whether humans could live in these dry lands during the interpluvials, for which there is no positive evidence. In terms of environment and subsistence the answer is positive, since it is the carrying capacity of the region that counts, rather than the climate. In this respect, it appears from comparisons with the present day when these deserts are inhabited by Bedouins, that the limited number of individuals making up early "societies" could easily thrive there.

It remains now to consider the paleoenvironmental conditions involved with the "out of Africa" wave, or waves (nicely termed "sorties" for some reason). The principal question is whether a deteriorating environment, accompanied by poorer subsistence and increased competition for the remaining impoverished food supplies, was the pushing force behind the venture into new territories (Bar-Yosef 1994). This problem is addressed here exclusively from the environmental point of view.

The onset of the Quaternary, some 2.6 Ma ago, saw a drying up of the East African habitats, due to a combination of aridification caused by the global initiation of the Ice Age, accompanied by regional tectonic uplift (Partridge 1992). Ever since, equatorial Africa went through alternations of wetter periods corresponding to the interglacials, and drier during the glacials (Horowitz 1977a). This, seemingly, makes for a perfect story. When the climate became drier in East Africa, humans wandered north (probably south, too) in search of more agreeable environments, finally arrived at the Jordan Rift Valley which in glacial times was a veritable paradise, and settled there, which is exactly what one finds: most sites are indeed embedded in pluvial deposits.

This ideal scenario, although admittedly perfectly possible, poses several questions. First is, why did *Homo erectus* not follow the logical way taken by most animals (and plants), which simply wandered in the footsteps of the changing environments. When the climate became drier in East Africa, wetter biotopes could easily be found quite close, to the west, in the former tropical rain-forest area, which became a lush savanna in glacial times (Van Zinderen Bakker & Mercer 1986). Instead, humans ventured to cross new frontiers by trekking northward (most probably not the only direction, but this is the one discussed here). This again raises a question, which pertains to human behavior in general. On the way north was the Nile Valley, with enough resources to supply quite a large population, so why go on? The same question naturally holds true for the Jordan Valley. It does not seem probable that the population increased so much as to create a "boom" that may have necessitated looking for additional territory.

It appears, from discussion of the environments and possibilities, that humans behaved in this respect differently from the typical animal, which would usually only wander back and forth following a suitable biotope, rarely crossing new frontiers. It seems that people wandered in spite of having sufficient food and water supplies and suitable environments all along the way. It appears that it was not entirely environmental stress or forcing that put humans on the move; rather, as seen from geology, the striving for new horizons resulted from curiosity, which is one of the fundamental and, it seems, one of the most ancient characteristics of the genus *Homo*.

APPENDIX

The nature and history of motion along the Dead Sea Transform (Rift)

Zvi Garfunkel

The name "Dead Sea Rift" has been used traditionally in the geological literature to designate the approximately 1,000 km line of crustal-scale fracturing that branches off the northern Red Sea and extends northward to the edge of the Taurus Mountains (Fig. A.1). Much of this zone of fracturing, especially its southern half, is marked by morphologically prominent fault-controlled depressions with a thick Neogene–Quaternary fill. The clear physiographical expression of faults, displacement of very young sediments, and the historical seismic activity along this line prove that it is an active tectonic feature.

The Dead Sea Rift already attracted the attention of geologists more than 150 years ago and its overall outline was recognized in early regional geological studies (Lartet 1869, Hull 1886, Blanckenhorn 1914). Suess (1909) and Gregory (1921) recognized that the Dead Sea Rift was the northern branch of the continental-scale rift system that extended via the Red Sea to East Africa. These early works led mainly to the view that the Dead Sea Rift was an extensional feature dominated by normal faulting. However, some early workers also suggested that important lateral motion along the Rift took place. This view became widely accepted when the role of large strike-slip faults was recognized, and especially following the advent of plate tectonic theory. Consequently the term "Dead Sea Transform" rather than "Rift" is now often used.

The purpose of the present account is to summarize the evidence supporting the latter interpretation and to show how it accounts for the various features of the Dead Sea Transform and for its regional relations.

A.1 GENERAL FEATURES

The Dead Sea Transform (Rift) cuts across a continental area with a normal crust, 30–40 km thick, that was shaped during the late Proterozoic Pan African orogeny. The stabilization of the area was marked by the development of an extensive continental-scale flat erosion surface (peneplain). Afterwards the region behaved

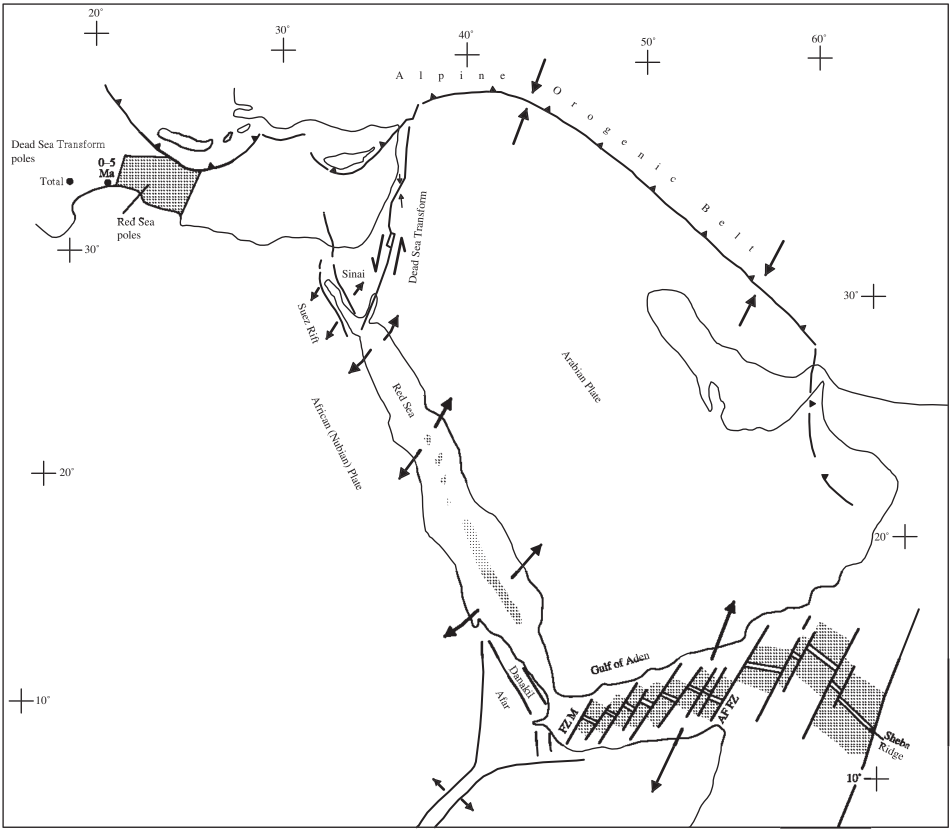


Figure A.1. Plate-tectonic framework of the Red Sea area. Stippling in Red Sea and Gulf of Aden: areas of young sea floor with linear, high-amplitude magnetic anomalies correletable to 10 Ma in Gulf of Aden, 2–5 Ma in southern Red Sea, 2 Ma or less in central Red Sea (shown schematically). Also shown are preferred ranges of locations of Red Sea and Gulf of Aden opening poles (stippled) as well as previous estimates. Poles for the Dead Sea Transform are accurate to within 0.5°. Based on Joffe & Garfunkel (1987).

as a rather stable platform during most of the Phanerozoic and was covered by an extensive veneer of continental and shallow water marine sediments of Cambrian to Paleogene age. The sediments that were deposited in the last major transgressive cycle of mid-Cretaceous to late Eocene age, consisting mainly of carbonates, still cover most of the areas flanking the Rift. However, along its southernmost part recent erosion has exposed older parts of the sediment cover as well as the crystalline basement.

The platformal history was punctuated by a few episodes of volcanic and tectonic activity (Garfunkel 1988, 1998). Most important were: (a) Permian(?), Triassic and early Jurassic rifting phases that were related to the shaping of the eastern Mediterranean passive continental margin, (b) late Cretaceous to mid-Cenozoic mild contractional and shearing deformation that produced a system of structures known as the Syrian Arc. These comprise two domains (Fig. 4.8.1). First, a domain of NE–SW to NNE–SSW trending asymmetric folds that formed over deep reverse faults; most, if not all, of these faults originated as normal faults during the early Mesozoic rifting phases and were later reactivated in the new stress field. Second, a more southern domain that comprises several approximately east–west trending lineaments defined by aligned faults, flexures and folds along which some right lateral shearing also took place – the central Negev–Sinai shear belt (Bartov 1974). Some of these lineaments in the Negev were active in the late Cretaceous whereas others were more active in the Cenozoic, especially in Miocene times (Fig. 10.2.1).

The Dead Sea Rift forms a conspicuous geological discontinuity that is unique in this region, across which rock units and the structure change abruptly. This is interpreted to have resulted from substantial lateral motion (see below) along the Rift that juxtaposed areas that originally were about 100 km apart. The structure along the Dead Sea Rift (Transform) varies considerably, which is expressed in marked along-strike changes in the morphology. South of Lebanon an almost continuous valley is developed along the Transform, but its width is quite variable (<5 to almost 25 km) and its floor is undulating ($+300$ m to $-1{,}600$ m). Much of this valley expresses a graben-like feature that is delimited on both sides by scarps controlled by normal faults. The topographic lows along the valley are the surface expression of spectacular basins that are filled with 5–10 km, locally more, of Neogene–Quaternary sediments and some volcanics. These alternate with topographic-structural saddles where the fill is much thinner or absent and locally some compressive structures are present. In Lebanon the northward continuation of the Transform crosses a major structural saddle that comprises the Lebanon and Hermon–Anti-Lebanon mountains and the intervening synclinal Beqa'a valley. Still further north, topographic lows comprising graben-like and synclinal valleys outline the Rift as far as the Taurus Mountains.

As the Dead Sea Rift developed, its flanks were deformed to different extents by several groups of marginal structures. In the south, along the Gulf of Aqaba (Elat) and the southern Arava, up to 20 km wide zones of marginal faults trending sub-parallel to the Rift are developed. Another zone of considerable marginal deformation is situated west of the Rift north of about 32°N, where the marginal deformation extends to the Mediterranean. Still further north, in Lebanon, marginal faulting and folding affected a wide zone. In this region several major faults splay off from the main fault zone (Fig. 10.2.2), the main ones being the Roum fault in southern Lebanon and the Serrhaya and other fractures splaying into Mount Hermon. The formation of the Dead Sea Rift was accompanied by widespread basaltic volcanic activity in the flanking regions, mainly to the east (Fig. 5.3.1). However, along the Rift itself the igneous activity was not particularly intense, and it is known only along the northern two-thirds of its trace. Also important on a regional scale was considerable uplifting that dominates the present-day topography. Uplifting was most accentuated, but variable, to distances of several tens of kilometers away from the Rift (Fig. 3.1.1). There the elevations often reach 0.8–1.2 km above sea level, but locally elevations reach 2.5–3 km, for example in Lebanon and Mount Hermon and on the two sides of the Gulf of Aqaba.

A.2 DEVELOPMENT OF INTERPRETATIONS

Until the 1960s most workers interpreted the Dead Sea Rift as an extensional feature dominated by normal faulting that resulted from stretching perpendicular to its strike. This was based on the presence of a graben-like valley along most of its southern half, in particular along the better-studied Jordan Valley and Dead Sea segments. The resemblance to well-known extensional rifts, such as the Rhine Graben and the East African rifts, strongly suggested a similar origin. The young uplifting and volcanism that accompanied the development of the Dead Sea Rift also resembled phenomena thought to characterize extensional continental rifts, as envisaged by Cloos (1939). However, the discontinuity of the graben-like valley and its absence in Lebanon led Dubertret (1947) to stress the role of vertical motions and the dissimilarity to extensional rifts, though he did not dispute the occurrence of extension. Bentor & Vroman (1960), recognizing the need to explain the abrupt change of stratigraphic sections across the Rift, presented a variant of the extensional interpretation. They proposed that the Rift formed by relatively recent reactivation of an old fracture that extended along the eastern side of the young depression, and that repeated activity of this fracture during the Phanerozoic allowed the two sides of the Rift to have different subsidence histories.

The possibility of lateral motion along the Dead Sea Rift was raised long ago (Lartet 1869), but Dubertret (1932) was probably the first to present concrete arguments favoring this interpretation. Accepting Wegener's (1922) interpretation that the Red Sea formed by the separation of Arabia from Africa, he argued that this motion required a left lateral offset perhaps exceeding 160 km along the Dead Sea Rift (later Dubertret, e.g. 1947, 1970, abandoned this idea). From another point of view, Wellings (1938) stressed the role of lateral motion in order to explain the difference between the stratigraphic sections across the Rift. The great step forward was taken by Quennell (1959). Based on better knowledge of the local geology, he showed that a left lateral offset of 107 km could explain the abrupt structural and stratigraphic change across the Rift. He also introduced the novel idea that lateral motion could explain the structural peculiarities of the Rift, such as the variable width and discontinuity of the Rift Valley, the presence of accentuated depressions (e.g. the Dead Sea basin), and the presence of local compression (especially in Lebanon). He also identified strike-slip faults along the floor of the Rift valley. Freund (1965) further elaborated these ideas and again stressed the relation between the lateral motion along the Dead Sea Transform and the opening of the Red Sea. At the same time Wilson (1965) also recognized the role of the Dead Sea Transform in accommodating the opening of the Red Sea, as it connects the zone of plate separation along the latter with the zone of plate convergence along the Taurus mountains.

These interpretations were tested and elaborated in numerous subsequent works and provide a framework for explaining many features of the Dead Sea Rift. They can also be integrated with, and used to refine, the regional plate kinematics. Therefore it is now widely accepted that left lateral motion was the dominant factor that shaped the Dead Sea Rift, so this tectonic line is more correctly described as a "Transform" rather than a "Rift". The former term will therefore be used from here on. In what follows these advances will be reviewed in some detail, outlining the successes of this model as well as some still unresolved issues.

A.3 INDICATORS OF THE TOTAL LATERAL MOTION

The arguments for considerable lateral motion along the Dead Sea Transform (Rift) are of critical importance for understanding its nature, so they are considered first. The structure of the Transform, which is quite unlike that of extensional rifts but typical of strike-slip fault zones, and the ongoing faulting along the Transform will be discussed later.

A.3.1 Matching of rock units and structures across the Transform

In principle, in order to demonstrate lateral motion along a fault line it is essential to identify offset rock bodies and structures across the fault line, as this is the necessary outcome of lateral slip. Thus the fault line must appear as a geological discontinuity and all the known features predating the lateral motion should be offset by the same amount. Identification of such a situation, therefore, is generally taken as evidence for the occurrence of lateral motion and this is in fact the only reliable basis for inferring the magnitude of such a motion. Other interpretations of such situations necessarily involve complicated and lengthy histories of the fault lines, and thus require a special proof. When other lines of evidence also favor strike-slip motion, its occurrence is taken as the most likely interpretation.

As stressed by Wellings (1938), Quennell (1959) and Freund (1965), the Dead Sea Transform is an outstanding geological discontinuity, as the rock units and the structures facing each other across much of its length are different. They argued that this was accounted for by lateral motion, as the geological features known to them on its two sides could be matched. Many years of continuing studies have indeed shown that this is the case, indicating a total left slip amounting to about 105 km.

A.3.2 Matching of the Phanerozoic sedimentary cover

The lithological and thickness variations of the Cambrian to Paleogene sediments, especially the Mesozoic units, often follow SW–NE or SSW–NNE trends over considerable distances on the flanks of the Dead Sea Transform (Picard 1959, Bender 1968a, Druckman 1974a, Derin 1974, Goldberg & Friedman 1974, Andrews 1992). As these trends form a substantial angle with the Dead Sea Transform trend, they allow us to distinguish features of the sediment cover that can be used as markers whose matching across the Transform allows us to determine the occurrence and magnitude of lateral motion. Many studies showed that units located some 105 km across the Transform are similar, whereas coeval units facing each other across the Transform are substantially different (Freund et al. 1970, Druckman 1974a, Bandel 1981, Bandel & Khouri 1981, Segev 1984). Another indication is provided by matching of the lines of truncation of various units by the early Cretaceous erosion surface, which also indicate a lateral transform offset of about 100 km (Druckman 1974a). In fact, continuing stratigraphic research allowed repeated testing of the interpretation, showing that when new data regarding various exposed or subsurface units became available they always confirmed the pattern expected to result from lateral offset. The data available at present allow the matching of all known features of the Phanerozoic sedimentary cover in exposures and in the subsurface, up to the middle Cretaceous, in the region south of Lebanon. Matching of late Cretaceous–Paleogene units is problematic, as their lateral variations were largely controlled by the 10–30 km wide syn-sedimentary Syrian Arc structures, but their distribution is also compatible with a 105 km left lateral offset (e.g. the Eocene sequence: Sneh 1988). The many data available from locations close to the Dead Sea Transform also confirm the abruptness of the change across the Transform, showing that it occurs over distances of less than 10 km up to 20–30 km – essentially the width of the "Rift Valley". If the contrasts across the Dead Sea Transform depict the original sediment distribution, it would imply a sharp bend in facies and thickness trends along the Transform during the entire Phanerozoic, and very sharp gradients, e.g. across the Dead Sea basin. In this case it would be an extraordinary coincidence that a shift by 100–110 km could remove the anomaly for all stratigraphic units in the area where enough information exists.

Matching of stratigraphic units across the northern part of the Transform was not tested because lack of data, especially from the subsurface, does not allow sufficiently detailed regional stratigraphic syntheses. On the other hand, neither has the continuity of stratigraphic units across the major faults in this region ever been demonstrated. On the contrary, a detailed study of the mid-Cretaceous section in Lebanon revealed that it changes abruptly across the main Transform fault (Saint-Marc 1974), though the possibility of lateral motion was not examined. Further north the Transform appears to offset the front of late Cretaceous ophiolitic nappes by 70–80 km (Freund et al. 1970, Al-Maleh et al. 1992). These pieces of evidence are compatible with considerable lateral offset, but clearly more information is needed to establish the offset along the major faults crossing this region.

A.3.3 Red Sea Dike System

The areas flanking the southern part of the Dead Sea Transform are crossed by a conspicuous set of NW–SE trending basaltic dikes (Fig. 5.3.1). Close to the Transform itself they are affected by the marginal lateral faults on the two sides of the Gulf of Elat–Aqaba. These are the northern members of a dike system that extends along the flanks of the Red Sea and the Suez Rift (Blank 1977, Eyal et al. 1981, Coleman et al. 1983, Brown et al. 1989). Their K–Ar ages cover the range of 19–25 Ma (Steinitz et al. 1978, 1981, Brown et al. 1989), but $^{40}Ar/^{39}Ar$ ages of dikes in Arabia cluster in a narrower range of 21–24 Ma (Sebai et al. 1991). This "Red Sea Dike System" formed just prior to the main phase of subsidence and incursion of the sea along the Red Sea–Suez Rift line.

Eyal et al. (1981) inferred that the dike segments outside the zones of marginal faulting along the southern part of the Transform were affected by the total Transform offset. In particular, a dike that extends through Karak, SE of the Dead Sea, can be matched with a plug and minor dike (Ashosh plug) at 30°30′N, just west of the Transform (Eyal et al. 1981). These bodies yielded K–Ar ages of 19 and 20.7 Ma respectively, which are indistinguishable within the error bars of the results (Barberi et al. 1980, Steinitz et al. 1978). In this area the Transform flanks are hardly deformed, so marginal structures do not complicate the matching. These are the youngest rock bodies known to be affected by the entire Transform offset.

Another important observation is that the marginal faults in SE Sinai, on the western side of the Gulf of Elat–Aqaba, offset these dikes left-laterally. The offsets of individual faults range from less than one to some 10 km, so the aggregate offset may amount to some 20 km (Eyal et al. 1981). Moreover, in many if not all cases the basement rocks are offset by the same amount, demonstrating that the faulting postdated the dike emplacement. The marginal fault system along the eastern side of the Transform in Jordan and NW Arabia is interpreted as of the same type, though this area was little studied.

A.3.4 Matching of the basement

The Late Proterozoic Pan-African basement is known from exposures along the southern part of the Transform and from a few boreholes further north (Bender 1968c, Eyal et al. 1980, Weissbrod 1981, Bartov 1990, Andrews 1991). Similarly to the platform cover, quite different basement rocks face each other across the Transform, but the great variety of rock types and complex outcrop pattern and insufficiently detailed mapping east of the Transform preclude detailed comparisons. The marginal left-lateral faults along this portion of the Transform further complicate matters. Clearly, future mapping of the basement will provide additional insights into, and tests of, the estimates of lateral offset along the Dead Sea Transform.

Nevertheless, present knowledge allows matching of a few prominent features, compatible with the amount of offset deduced above. Thus, an east–west strip of acid volcanics north of Elat can be matched with similar volcanics exposed between Wadi Dana and the Petra area, and which in both areas form a relief that was covered by similar Cambrian sediments (Segev 1984). A granitic complex in SE Sinai between 29°15′N and 28°45′N and further south matches the granitic complex near and north of Aqaba. On the other hand the latter complex differs markedly from the rock complex near Elat just across the Transform, where metamorphic rocks are abundant. However, these metamorphics resemble exposed ones east of the Transform, around 30°12′N (Jarrar 1985).

The lateral offset also affects a series of immature sediments and some volcanics (about 0.5 to >2 km thick) that formed at the end of the Pan-African orogeny. They occur mainly in the subsurface north of the basement exposures and are known as the Zenifim and Saramuj formations in Israel and in Jordan, respectively (Weissbrod 1981, Andrews 1991). West of the Transform these series are well developed (>640 m thick) less than 60 km north of Elat, but on its eastern side they occur only farther north, compatible with a left lateral offset.

Another relevant observation is the similarity between the aeromagnetic anomaly patterns in Jordan between 30°N and 31°30′N and in the southern Negev west of the Transform, which is compatible with the lateral offset deduced above (Hatcher et al. 1981). The source of these anomalies is in the basement, because in this region the sedimentary cover hardly contains any igneous rocks or iron rich beds. This observation should, therefore, be interpreted as evidence for the lateral offset of lithologic zones in the basement.

A.3.5 The Central Negev–Sinai shear belt

Matching of the features listed above indicates a total offset of about 100 km, but this figure involves an uncertainty of up to some 20 km. This is because facies and thickness variations do not define sharp boundaries, and sometimes it is necessary to extrapolate various units that trend at moderate angles to the Transform, which necessarily involves some uncertainty. A more accurate estimate of the offset can be based, as noted by Quennell (1959), on matching the prominent lineaments of the central Negev–Sinai shear belt that trend almost perpendicular to the Transform (Fig. 10.2.1). West of the Transform these lineaments cross the central Negev and continue westward across Sinai, a distance of some 200 km (Bentor & Vroman 1954a, Shata 1959, Bartov 1990). On reaching the Transform they do not continue eastward along strike, but they can be matched with lineaments that occur further north and extend a few hundred kilometers eastward (Quennell 1959, Bender 1968c). Matching of these lineaments (Fig. 10.2.1) constrains the lateral offset on the Dead Sea Transform to about 105 km, with an uncertainty of less than 5 km (Quennell's estimate was 107 km).

The lateral motion also affected the domain of folds of the Syrian Arc north of the shear belt. Individual folds cannot be matched with certainty across the Transform because they do not persist along strike, except for the Hermon and Fari'a folds that may have originally formed a single anticlinal structure. The reconstruction broadly matches not only the Syrian Arc folding domain, but also the infrastructure of the early Mesozoic rift structures. In particular, a major NNE–SSW trending fault that forms the eastern boundary of a deep Triassic graben that extends east of Jerusalem and further north can be matched with the coeval southern boundary of the Palmyra depression in central Syria (Garfunkel 1998). This is again compatible with the proposed lateral motion along the Dead Sea Transform. Furthermore, Walley (1998) correlated the structural zones of Lebanon west of the Transform with various zones of the Palmyra fold belt, which implies an offset of about 100 km across the Transform in Lebanon.

A.3.6 The regional plate kinematics

Historically the kinematic relationship between the opening of the Red Sea and the Dead Sea Transform provided the first strong argument favoring a large left-lateral offset along the latter (Dubertret 1932). This argument was again presented by Freund (1965), Wilson (1965) and Gass & Gibson (1969). Now it is generally accepted that the Dead Sea Transform forms a part of the western boundary of the Arabian plate, separating it from the Sinai sub-plate – an appendage of the African plate from which it is separated by the Gulf of Suez Rift (Fig. A.1). North of the Red Sea its opening is partitioned between the Dead Sea Transform, which took up most of the Arabian–African plate separation, and the Suez Rift, whose opening took up only a fraction of this plate motion. This view explains the abrupt change in morphology between the Red Sea and its two northern branches, and also explains the great difference in internal structure between them. Moreover, this view leads to a quantitative analysis that allows estimation of the motion on the Dead Sea Transform completely, independently of the geology of its flanks, using merely its orientation relative to the direction of opening across the Red Sea.

The first quantitative estimate of the Arabian–African plate motion (McKenzie et al. 1970) was based on a reconstruction of the Red Sea by matching its coastlines. Improved reconstructions take into account not only the similar shapes of the Red Sea coasts but also the stretching of its margins and the presence of the continental Danakil block to the south (Fig. A.1, Freund 1970). The record of sea-floor spreading in the Gulf of Aden provides additional strong constraints. Taken together these considerations show that the Euler pole of the Arabian–African plate motion is located at about 32–33°N, 22–25°E (Joffe & Garfunkel 1987, Le Pichon & Gaulier 1988), that is, in or near NE Libya (south of the pole of McKenzie et al. 1970). The exact amount of opening of the Red Sea is somewhat uncertain, but minimum figures can be obtained from data on the extent of oceanic-like crust under the basin and on the crustal structure of is margins. Thus, at 17°N the plate separation amounted to 280–320 km, nearly perpendicular to the axis of the Red Sea, while at the northern extremity of the Red Sea the plate separation amounted to 130–150 km in a direction in the range of 20–25° east of north. Further north this motion is partitioned between the Dead Sea Transform and the Gulf of Suez Rift, the major structures in this region. However, the geometry of faulting and the presence of a Precambrian (Pan-African) basement underneath the entire width of the latter indicate an opening of only 20–30 km in a NE–SW direction (Garfunkel & Bartov 1977, Moretti & Colletta 1987). As this is a small fraction of the entire motion, it follows that most of it is taken up by the Dead Sea Transform, i.e. that about 100 km of left-lateral slip took place subparallel to the Transform trend. This is in excellent agreement with the estimate above based on the detailed geology of the flanks of the Dead Sea Transform. Thus the quantitative plate-kinematic analysis shows that the lateral motion along the Dead Sea Transform is a necessary geometric consequence of the opening of the Red Sea, and one is not possible without the other. Any interpretation of the Dead Sea Transform must, therefore, define its plate-kinematic role in a quantitative way.

The conclusion from the plate-kinematic analysis is in line with the overall arcuate shape in a map view of the southern half of the Dead Sea Transform. This led Quennell (1959) to describe the motion along the Transform as a rotation about a "pole of veering" located at 33°N, 24°E, while Freund (1970) estimated the pole of opening of the Red Sea and of the motion on the Dead Sea Transform to be at 32°N, 22°E. These excellent estimates can be refined by taking into account the details of the structure along the Transform (Garfunkel 1981), as will be discussed below. Thus the data from the Dead Sea Transform can be integrated with the data from the Red Sea and Gulf of Aden in order to refine the regional plate-kinematic picture (Joffe & Garfunkel 1987, Le Pichon & Gaulier 1988).

The regional plate kinematics requires that the motion between Arabia and the Sinai sub-plate should occur as far north as their junction with the Alpine zone of plate convergence. This means that the interior of Syria and the extra-alpine part of southeastern Turkey moved some 100 km northward relative to the Levant basin. Thus from the plate-kinematic point of view the doubts about the continuation of the lateral Transform motion in Lebanon and farther north are untenable, though the question as to how this motion was accommodated still needs clarification (see below).

A.3.7 Summary

Two independent approaches allow us to establish the occurrence of lateral motion along the Dead Sea Transform and to evaluate its amount. First, matching of geological features across the Transform. Second, kinematic analysis of the Arabian–African plate separation and its partitioning north of the Red Sea. Both lines of evidence are in excellent agreement, indicating a left lateral offset of some 100 km, the most accurate estimate of about 105 km being based on the matching of structural lineaments.

A.4 HISTORY OF THE DEAD SEA TRANSFORM

While the magnitude of the lateral offset along Dead Sea Transform is quite well constrained, the slip history is more difficult to establish in the absence of reliable syn-offset markers. A general picture can be obtained, however, by combining the information regarding the geology along the Transform with the regional plate-kinematic history.

A.4.1 Initiation of the Dead Sea Transform: local constrains

The geological relations of the Red Sea Dike System, mentioned above, bear directly on the time of initiation of the lateral motion (Eyal et al. 1981). They are the youngest known features to be affected by the entire Transform offset, and in SE Sinai (and probably also in Midyan) they are affected by the total displacement of marginal faults. The dikes next to the Transform yield K–Ar ages of 19–22 Ma, i.e. very early Miocene (Aquitanian). However, this overlaps with the paleontologically established age of the marine beds in the Suez Rift, which clearly postdate the igneous activity along the NW continuation of this dike system (Garfunkel & Bartov 1977, Richardson & Arthur 1988, Ouda & Masoud 1993). These relations raise the possibility that these K–Ar ages are somewhat too young (Garfunkel 1989), similar to what was found by Sebai et al. (1991) in Arabia.

Additional information is provided by the presence of Miocene sediments and volcanics in some basins along the Transform, as the formation of these basins can be associated with its activity. Oligocene beds occur also in a few places, but their interpretation seems less obvious, so they will be discussed separately below.

A basin about 35 km long on the western Transform flank that extends from near Tiberias and southward is filled to the north by about 750 m of sediments (Herod Formation) with several intercalated basalt flows, whereas to the south its entire fill consists of volcanics. The oldest flow, about 350 m above the base of the sediments, gave a K–Ar age about 17 Ma and the oldest volcanics to the south gave an age about 16 Ma, while the youngest volcanics are 10–12 Ma old (Shaliv 1991). The structurally low position of these sediments and volcanics and the large proportion of locally-derived coarse conglomerates among the former indicate that they fill a tectonically controlled depression which is best interpreted as a subsidiary structure along the Transform. Nearby, the Zemah 1 well, along the Transform valley just south of Lake Kinneret, crossed more than 500 m of middle Miocene sediments without reaching their base (Marcus & Slager 1985, Horowitz

1987c), indicating that at that time considerable subsidence occurred also along the main Transform trace.

Within the Dead Sea basin thick continental clastics, predominantly quartzose sandstones, were encountered in some boreholes near its western margin. In the Sedom Deep 1 borehole an approximately 1 km thick section was dated palynologically as having an early and middle Miocene age (Baker 1994, Horowitz 1996a). Closer to the basin margin a section about 700 m thick in Ami'az 1 gave an early and middle Miocene age, while further south a section about 500 m thick (base not reached) in Arava 1 gave a middle Miocene age (Horowitz 1987c). These clastic sections are attributed to the Hazeva Formation which is also widely exposed on the Transform flank west of the Dead Sea basin, where it was dated to the same time interval (Goldsmith et al. 1988). The sections in the basin are, however, considerably – up to four times – thicker than the correlative sections on the Transform flank. This is taken as an indication that the Dead Sea basin was already outlined when these sediments were deposited.

Further south, in the northern Arava valley, the Dead Sea basin is bordered on the west by an eastward dipping flexure. In many places the flexure is overlain by the Hazeva Formation, whose basal beds usually consist of locally derived conglomerates to the west, but eastward they grade into thicker fine lacustrine beds, which indicates deposition along the margins of a depression (Bentor & Vroman 1957, Eidelman 1979, Sneh 1981). Though contemporaneous depressions occupied by lakes formed in synclines of the Syrian Arc on the western Transform flank, the shape of the depression along the Arava is discordant with respect to the trend of the Syrian Arc folds and lineaments. This strongly suggests that it was a newly formed structure along the Transform. These basal beds of the Hazeva Formation were not dated, but their position below the fossiliferous beds indicates an early Miocene or even older age.

Taken together these data are interpreted as evidence that tectonically controlled basins, most likely produced by faulting, formed along the Transform at least 18–17 Ma ago or shortly thereafter, that is, in early Miocene times. Strictly speaking, this does not directly date the beginning of lateral motion. However, since the major depressions along the Transform, the Dead Sea basin in particular, formed as a direct consequence of lateral motion (see below), their formation can be interpreted as recording the occurrence of such horizontal slip. An alternative explanation would be the formation of a broad sag, in places filled with 2 km or more of sediments, which was obliterated, except for its parts that were downfaulted along the Transform. In this author's opinion this does not fit the record of the Hazeva Formation in the Negev (Garfunkel 1997) and is also difficult to justify geodynamically.

These indications need to be related to the evidence of Miocene activity of the central Negev–Sinai Shear Belt. At least some lineaments disturb the Hazeva Formation, while in Sinai dikes related to the Red Sea Dike System are offset right laterally by up to a couple of kilometers (Bartov 1974). As the lineaments of the shear belt are affected by the entire Transform offset, the question arises as to whether the lateral Transform motion postdated the last activity of these lineaments (Bartov 1974). This would require that the lateral Transform motion began not much earlier than 10 Ma ago. However, the late Cretaceous activity of some lineaments (Garfunkel 1993) shows that the Miocene movement actually expressed the rejuvenation of structures that are considerably older than the Transform. This raises the possibility that the youngest phases of their activity occurred after initiation of the Transform (Garfunkel 1997). That such a situation is possible is proven by the record of faulting in California, that indicates coeval activity of two conjugate systems of lateral faults (Jennings 1973). In this interpretation the great thickness of the Hazeva sediments in the Karkom graben along the Paran lineament is thought to have resulted mainly from syn-Hazeva subsidence of this depression (Garfunkel 1997, Calvo et al. 1998). Still, the relation between the Miocene activity of the central Negev–Sinai shear belt and the Transform is incompletely documented and requires further study.

Another problem is raised by the presence of a few occurrences of marine beds of Oligocene age along the Transform, which contrasts with the very limited distribution of sediments of that age on the Transform flanks. This raises the question as to whether the faulting along the Transform already began before the Miocene. One such occurrence is in the southern Golan, just east of Lake Kinneret, where about 150 m of early Oligocene to early Miocene beds are present (Michelson & Lipson-Benitah 1986). Another occurrence is at the base of the Sedom Deep 1 borehole in the Dead Sea basin. According to Horowitz (1996a) the lowest approximately 70 m (base not reached) are marine beds of middle Oligocene age, while overlying rocks are of undifferentiated latest Oligocene–early Miocene age; thus in part they may be pre-Miocene in age. Baker (1994) attributed this interval to the continental Hazeva Formation, and thought the lowest to be a distinct marine unit of middle Oligocene age. The third occurrence is in Midyan, east of the Gulf of Aqaba, where about 50 m of fossiliferous marine beds of late Oligocene beds overlie more than 100 m of coarse clastics, presumably also of Oligocene age (Dullo et al. 1983, Bayer et al. 1988).

These occurrences may be interpreted in either of two ways: they may be relicts of more extensive sediment bodies that were protected from erosion in structural lows along the Rift, or they may indicate the beginning of subsidence, and thus probably also faulting, along the Rift. In the southern Golan the Oligocene beds were downflexed towards the Rift and thus protected from erosion before the deposition of middle Miocene sediments (Michelson 1972). Hence, they may well be a remnant of a more extensive sediment sheet. Indeed beds of broadly the same age are widespread further east in Syria (Ponikarov et al. 1969). The significance of the occurrence in the Dead Sea basin is uncertain. On the one hand, the evidence showing that in much of the Negev further west, extensive pre-Hazeva Formation erosion removed a few hundred meters of late Eocene beds (Avni 1991) raises the possibility that even younger beds were widespread and could have been preserved only in the basin. However, deposition in depressions along the Transform cannot be excluded. Neither is it clear how they relate to the above mentioned basal Hazeva beds. Moreover, the shallow nature of the Oligocene beds along the Dead Sea Transform do not indicate exceptional subsidence which needs to be explained by Rift-related faulting, as do the younger sediments. Thus, the occurrence of Oligocene faulting or localized subsidence along the Transform remains an unproven possibility.

The Oligocene beds in Midyan should predate much of, or even the entire, lateral Transform offset (if it began that early), so that originally they formed within the Red Sea basin and therefore will be discussed below in relation to this domain.

A.4.2 Age relative to other rifts in the region

Given the plate-kinematic role of the Dead Sea Transform, its initiation should also be related to the history of the other young rifts in the region. The essential point is that the motion along the Dead Sea Transform is linked to the history of opening of the Red Sea. Any significant opening of the latter could not have occurred without slip on the Dead Sea Transform (as only a minor part of this opening was transferred to the Suez Rift) and vice versa, so their histories must be compatible. Thus the available field and borehole data from the Red Sea provide considerable insight into the origin of the Dead Sea Transform. In the northern Red Sea significant subsidence, and thus plate separation, began only in the early Miocene 22–20 Ma ago (i.e. postdating the Red Sea Dike swarm), but local subsidence and sedimentation (mainly continental) occurred already in the late Oligocene (Bayer et al. 1988, Montenat et al. 1988, Mitchell et al. 1992). The Suez Rift had a similar early history (Garfunkel & Bartov 1977, Moretti & Colletta 1987, Richardon & Arthur 1988, Ouda & Masoud 1993). However, further south along the eastern margin of the Red Sea, between 22° and 17°N, much more significant rifting and subsidence took place before and possibly also during the emplacement of the Red Sea Dike System (Schmidt et al. 1982, Brown et al. 1989, Coleman 1993).

Still further south, marine late Oligocene beds were recorded in the southernmost Red Sea, while in the eastern Gulf of Aden a thick middle Oligocene rift fill was drilled (Hughes et al. 1991). A thick rift fill of approximately that age was also drilled on the margin of the Red Sea in Ethiopia (Savoyat & Balcha 1989). Thus the early stages of rifting become progressively more pronounced on going southward, implying increasing subsidence in this direction. This is to be expected, because rigid plate kinematics requires that the separation must increase at increasing distances form the corresponding Euler pole. It is inferred, therefore, that Arabia already began to break away from Africa along the Red Sea–Suez trend and the Gulf of Aden in Oligocene times, so it was under way some 30 Ma ago. Omar & Steckler (1995) suggested a somewhat older age for the beginning of rifting based on fission track ages from the rift flanks. However, it remains to be shown that the early erosion implied by the fission track dates that these authors obtained is indeed compatible with the geological record along the Red Sea. However, the beginning of significant subsidence in the northern Red Sea and the Suez Rift only in the very early Miocene (22–20 Ma ago) indicates that until then plate separation was sluggish.

These data can be compared with the indications that the lateral motion along the Dead Sea Transform postdates the Red Sea Dike System and that basins began to form along it close to 18–17 Ma ago. Thus it can be concluded that the Transform motion postdates the beginning of the Arabian and African plate separation. However, if faulting along the Transform began in the Oligocene, then the implication is that fracturing along all the rifts in the region began more or less at the same time. In this case it is possible that the late Oligocene sea reached Midyan through a depression along the embryonic Dead Sea Transform. At first sight this is an attractive model, providing an alternative to a seaway through the Suez Rift, which cannot be accepted in view of the absence of marine Oligocene beds from all boreholes in its central and southern parts. Such a long narrow seaway along the Dead Sea Transform is not simply reconciled with the open marine conditions recorded by the Oligocene fauna in Midyan (Dullo et al. 1983), which are better explained by a widespread short-lived marine ingression. Alternatively, this occurrence could record an ancestral Red Sea, but this is not easy to reconcile with the absence of marine beds in drill holes in the northern Red Sea (Mitchell et al. 1992). Thus, the available data do not lead to a simple self-consistent picture of embryonic faulting of Oligocene age along the Dead Sea Transform, so this remains an unproved possibility.

Additional insights into the age relations between the different rifts can be gained by an examination of their later history, to be discussed below.

A.4.3 Progress of the motion

Quantitative estimates of the progress of Arabian–African plate separation provide additional insights into the history of the Dead Sea Transform. The most important constraint is furnished by the magnetic anomaly sequence in the southern Red Sea that records about 75 km of seafloor spreading in the last 4.7 Ma (Roeser 1975). Using the known plate kinematics, this constrains the slip along the Dead Sea Transform to 30–35 km (Joffe & Garfunkel 1987, but see below). Thus at least two-thirds of the Transform motion took place before the end of the Miocene. Moreover, the magnetic anomalies in the Gulf of Aden show that this rate was more or less constant since at least 11 Ma (end of middle Miocene) (Cochran 1981). At this rate the entire Transform motion could have been achieved in 14–16 Ma. In view of the evidence for older activity along the Transform, this indicates that in its early history the lateral motion was slow and it accelerated at some later stage.

To evaluate the significance of this inference it should be compared with changes in the histories of opening of the Red Sea and Suez Rift. The above constraint on the young rate of Arabian–African plate separation shows that the oceanic crust in the southern Red Sea, about 170–180 km wide in that area (Izzeldin 1987), formed during the last 11–12 Ma, that is when the plate separation was constant. At that rate the entire plate separation in this area (300–320 km at about 17°N) could be achieved in 19–20 Ma. This indicates that in the early stages the opening rate must have been considerably slower, given that here rifting began 25–28 Ma ago or earlier. The onset of pronounced subsidence in the early Miocene probably indicates the acceleration of the opening rate, but there are no indications as to whether this was abrupt or gradual.

In contrast, the opening of the Suez Rift slowed down in the middle Miocene. Moretti & Colletta (1987) and Richardson & Arthur (1988) inferred from the record of several boreholes that since the beginning of the middle Miocene (16–15 Ma ago) most of the subsidence can be accounted for by thermal relaxation following the earlier stretching of the rift floor. This implies that there was almost no additional separation across the Rift. A more complete record (Ouda & Masoud 1993) favors, however, a less drastic slowing of the opening and that opening continued in the Quaternary, which agrees better with the recent faulting and seismicity of the Suez Rift (Garfunkel & Bartov 1977, Salamon et al. 1996).

The changes in the rates of plate motions can be interpreted as follows. After a period of initial rifting in the (mainly late) Oligocene, the Arabian–African plate separation accelerated at the beginning of the Miocene. At some stage a large part of the plate motion was transmitted to the Dead Sea Transform, while the opening of the Suez Rift slowed down. The redistribution of the plate motions may have occurred over a period of several million years in the middle Miocene. Le Pichon & Gaulier (1988) suggested an abrupt reorganization about 12 Ma ago, but this probably better approximates the time when the recent state of plate motions was attained.

In the present state of knowledge this interpretation seems to fit the data better than an age of some 10 Ma or less for the beginning of lateral motion along the Dead Sea Transform, and thus also for most of the Red Sea opening. The implication is that the early stages of the Transform motion were coeval with the younger phases of activity of the Central Negev–Sinai Shear Belt.

Unfortunately, the record along the Transform itself adds little information to the lines of evidence discussed above regarding young offsets, because Neogene rocks cannot be matched reliably across the Transform. Nevertheless a few occurrences deserve some discussion. One is the occurrence of middle Miocene sediments with a few basalt flows dated at 15–16 Ma, about 12, and 8 ± 3.7 Ma east of Lake Kinneret (Michelson 1972, Shaliv 1991). They resemble the broadly coeval Herod Formation west of the lake, as well as beds attributed to this unit but located 40–50 km further south on the western side of the Transform. Thus it appears that similar sediments formed at several locations along the Transform, so particular localities cannot be matched with any certainty. Moreover, the absence to the east of thick middle Miocene volcanics similar to those developed southwest of Lake Kinneret casts further doubt on the matching of the Neogene series across the lake.

Another significant occurrence is the approximately 5 Ma Homs volcanic field (Fig. 10.2.3) that extends across the Transform in northern Lebanon and adjacent Syria (Fig. 1 in Mouti et al. 1992). The trace of the Transform across the field has a prominent morphological expression, including a young sediment-filled depression that interrupts the continuity of the volcanic field, which expresses post-volcanism faulting (Dubertret 1955b, Ponikarov et al. 1969). The parts of the field across the Transform appear mis-aligned, but their edges cannot be reliably matched, as they must have been considerably eroded in the approximately 5 Ma since the extrusion of the volcanics. Thus an earlier attempt to match the edges of this field (Garfunkel 1981) can no longer be maintained. It is unlikely, however, that the lateral offset of the Homs field exceeded some 25 km if its two parts were originally continuous, because then they would have been almost completely separated, but the offset could have been smaller. If valid, this argument implies that here the offset was somewhat smaller than farther south, where it was estimated to be 30–35 km based on plate-kinematic considerations. The significance of this will be discussed below.

A.4.4 Summary

The history of motion along the Dead Sea Transform can be constrained by combining local geological data with the regional plate kinematics. The age of the fill in basins (Lake Kinneret, Dead Sea) along the Transform shows that it was already active 18–17 Ma ago, whereas the relations between the Transform and the Red Sea Dike System strongly suggest that the lateral motion is younger than about 20 Ma. This implies that the Dead Sea Transform formed sometime after the beginning of rifting along the Red Sea–Suez Rift trend in the late Oligocene (30–25 Ma ago). Initiation of faulting along the Transform in the Oligocene cannot be ruled out, but its effects (e.g. subsidence, separation of the Transform flanks) were probably very limited, reflecting the very slow Arabia–Africa plate motion in that period. The regional plate kinematics indicate that most of the lateral motion along the Transform occurred since sometime in the (early?) Miocene, when Arabian and African plate separation accelerated, which was probably linked with the slowing of opening of the Suez Rift. More than two-thirds of the lateral Transform offset took place before the Pliocene. The relations between the Dead Sea Transform and the central Negev–Sinai shear belt raises the possibility that the Transform motion and most of the Red Sea opening postdate approximately 10 Ma, but this does not appear to tie in well with the regional picture. Contemporaneous right lateral motion along the shear belt and early left-lateral transform motion seem to be more likely, but the ages of the different structures still need to be better constrained.

A.5 STRUCTURE OF THE TRANSFORM

The structure along the Transform is controlled by normal and strike-slip faults that produce a series of lows, saddles, and some compressive features. Quennell (1959) showed that the variability of these structures is well explained as a result of lateral motion along a fault line that has minor irregularities in map view. Where the fault trace bends to the right its two sides converge during left lateral motion, and there compressive structures will form, whereas where the fault trace bends or steps to the left, its two sides separate and there extensional pull-apart basins (rhomb-grabens) will form. Such relations were later recognized as characteristic, actually diagnostic, of strike-slip faults in general (e.g. Crowell 1974, Mann et al. 1983, Christie-Blick & Biddle 1985).

A.5.1 Southern half of the Transform

The part of the Transform south of Lebanon is marked mostly by a prominent morphological valley, the Transform valley, whereas in Lebanon and further north such a continuous low is absent. Because of this difference these two parts of the Transform will be discussed separately. Along the southern part two groups of longitudinal faults are most important (Garfunkel 1981). First, there are normal faults which define the entire eastern boundary of the Transform valley and much of its western boundary. They usually produce systems of narrow structural steps across which the total vertical offsets can reach a few kilometers. Second, embedded within the structural low between the normal faults is a system of left stepping en-echelon left lateral faults whose strikes deviate slightly clockwise relative to the overall trend of the Transform. These faults define a series of large pull-aparts and intervening structural saddles. Fault motion is essentially left lateral along the structural saddles, whereas along the pull-aparts the lateral motion is combined with vertical offsets reaching more than 10 km. Characteristic of this second fault system is the systematic formation of pull-aparts (rhomb-grabens) where they are left stepping, and of compressive structures where they bend, even by only a few degrees, to the right (Garfunkel 1981). As noted above, such structural relations are typical, and actually diagnostic of left lateral faults.

The main basins along the southern part of the Transform are the composite Gulf of Aqaba basin, the Dead Sea basin (including the northern Arava), the Lake Kinneret–Kinarot basin, and the Hula basin. They are delimited by the structural saddles of the Straits of Tiran, central Arava, southern Jordan Valley, and Korazim. Several studies discuss in detail the internal structure and history of the basins (Ben-Avraham et al. 1979, Garfunkel 1981, 1997, Ben-Avraham 1985, Kashai & Croker 1987, ten Brink & Ben-Avraham 1989, ten Brink et al. 1993, Frieslander et al. 1994, Garfunkel & Ben-Avraham 1996, Ben-Avraham et al. 1996). The available data, mainly drilling, seismic reflection and gravity, reveal that the pull-apart basins are filled with low-density sediments up to 5–10 km thick and some volcanics. The gravity data also strongly suggest that the Moho depth beneath the basins is close to normal, which implies that the crystalline crust under the thick fill of the basins is considerably thinner than under their flanks. On the other hand, the saddles between the basins have a much thinner fill, and along them the Transform valley is much narrower than along the basins, and locally hardly developed.

Within the scope of the present account only a few major points relevant to the Transform motion can be considered. An important point is that pull-apart basins (rhomb-grabens) grow by becoming longer parallel to the lateral motion that produces them while their width hardly changes, so their area increases as the lateral motion proceeds. Their internal architecture should include, therefore, structures that accommodate lengthening parallel to the major longitudinal faults. Indeed, transverse faults, some of them proven to be listric and thus efficient in accommodating extension, have been identified in many instances. However, there is still insufficient data about their deep structure to allow quantitative interpretations (as is the case with all pull-aparts) and to delineate the structure at the extremities of the depressions. Clearly, pull-aparts are generally more complicated structures than the schematic models of rhombic grabens that are often shown in the literature to illustrate the overall relations (e.g. Garfunkel 1981, 1997), and much more needs to be known about their deep structure. There are also no data regarding the mechanisms of the thinning of the underlying crust, which is indicated by gravity, though it probably expresses extension along the basins as they grow longer. These problems, originating from insufficient data, should not obscure the fact that the basins along the Dead Sea Transform, as all pull-aparts elsewhere, display the same basic structural setting, namely they are developed between en-echelon lateral faults that step in the same sense as the fault slip. Neither should it be overlooked that the small width of the depression along the Dead Sea Transform and the development of a string of basins and narrow saddles along the Transform structure are quite unlike the structures found along any extensional Rift.

Since the pull-aparts grow by becoming longer parallel to the bounding strike-slip faults, their length must be equal to or, most probably, greater than the lateral slip during their formation. This relation has important implications for the history of the basins along the Dead Sea Transform. Only the Dead Sea basin is longer than the entire Transform motion, so it could date from the beginning of the lateral motion. On the other hand, the individual basins that comprise the composite Gulf of Aqaba basin and the basins to the north are considerably shorter than the total Transform offset. This shows that only a part of the lateral Transform motion contributed to their formation. Along the Gulf of Aqaba the marginal strike-slip faults took up a portion of the Transform offset, leaving less to be accommodated along the pull-aparts, but even so the basins do not appear long enough to accommodate the entire remaining offset. Such an explanation cannot be invoked for the northern basins, so it must be inferred that they formed either after a part of the Transform offset had already taken place, or that they stopped growing for a while. In all cases, it follows that the major strike-slip faults along the Transform which control the active pull-apart basins originated sometime after the beginning of the lateral motion, that is, that the major faults were rearranged during the Transform development. Indeed, the history of the Hula basin agrees well with this inference (Heimann & Ron 1993).

The combined effect of motion on the two groups of faults is to produce some divergence between the plates bordering the Transform. This results not only from the obvious contribution of the normal faults, but the lateral motion also contributes to the divergence. First, because the strikes of the major lateral faults deviate from the overall Transform trend, and second, because pull-apart formation involves opening new surface area, as noted above. The amount of divergence varies along the Transform, but is estimated to be less than 10% of the lateral offset (Garfunkel 1981). The southern half of the Transform is, thus, predominantly transtensional. It was also called leaky, in analogy with oceanic transforms where some new area forms. It is this component of divergence that produces the graben-like low along the Transform that makes it superficially resemble extensional rifts, but the structural style along the Transform is quite unlike extensional rifts and proves a different origin.

These features can be directly related to the regional plate kinematics. Quennell (1959) used the overall arcuate shape of the southern half of the Transform to infer a pole of rotation for the relative Arabia–Sinai plate motion (though these terms were not yet introduced). The more detailed understanding of the structure along the Transform allows this picture to be refined (Garfunkel 1981). The occurrence of some divergence along the Transform shows that the direction of relative plate motion deviates by several degrees from the Transform trend, even if its trace can be approximated by a small circle. The analysis shows that an Euler pole at 32.8°N, 22.6°E agrees well with the strikes of the individual faults along the margins of the pull-aparts and along the intervening saddles (Fig. A.1). A rotation rate of 0.283°/Ma during the last 5 Ma (i.e. about 6 mm/year along the Dead Sea) is compatible with all the constraints mentioned above. However, this pole cannot describe the entire Transform offset, as it would lead to plate divergence that exceeds the width of the Transform valley. This indicates that the plate motion changed, such that the component of divergence increased as the Transform motion progressed. The overall motion is better inferred from the overall Transform shape and can be described as a rotation of 4.097° about a pole at 32°42′N, 19°48′E. The change in motion may be related to the structural changes in the internal Transform structure inferred above.

The presence of compressive structures along the Transform is important for its interpretation, because their formation is well explained as a consequence of the lateral motion. Such features are found where the main longitudinal faults bend to the right both on the structural saddles and within the structurally low areas. Along the Arava structural saddle, several pressure ridges are developed (Garfunkel et al. 1981), probably resulting from recent changes in the position of the active fractures along the main strike-slip fault zone. Such changes are indicated by the presence of short and therefore young pull-aparts, which requires that the bounding fractures were activated quite recently (some 100 Ka ago). In the Jericho region north of the Dead Sea compression is evidenced by small thrusts in the valley fill (Gardosh et al. 1990), while near 32°00 N the fill of the Transform valley forms a domal structure about 1 km wide and several kilometers long (Bender 1968a,c). Seismic reflection studies show that in this area the main strike-slip fault zone has a reverse component (Rotstein et al. 1991). The contractional deformation is thought to have resulted from a change in the fault configuration late in the Transform history after a considerable fill had accumulated within the Transform valley. If such deformation prevailed in this area during the entire Transform history it would have inhibited the formation of a Transform valley. A phase of contractional deformation of the fill of the Transform valley was also inferred from seismic reflection data south of Lake Kinneret that show a prominent anticlinal structure (Fig. 8.4.5) at the site of the Zemah 1 well (Rotstein & Bartov 1992). Another upwarp was imaged further west, and there exposures of early Pleistocene beds are strongly contorted, tilted and folded at the Ubeidiya site. Compression is also inferred in the Korazim saddle north of Lake Kinneret, where the main fault zone has a reverse character (Rotstein & Bartov 1989), and intense fracturing and rotations of blocks on vertical axes took place (Heimann & Ron 1993).

A.5.2 Northern part of the Transform

The overall trend of the Transform in Lebanon and further north deviates to the right relative to the arcuate trend of the southern part (Fig. A.1). Therefore any plausible plate-kinematic model in which the Dead Sea Transform takes up most of the plate motion involved in the opening of the Red Sea predicts transpression along this part of the Transform, that is, its flanks converge and are compressed to some extent. The amount of convergence will vary, however, depending on the changes in strike of the Transform trace. This explains the structure and morphology (particularly the limited development of deep basins) along the northern part of the Transform as well as the considerable deformation of its flanks, which are quite different from the situation farther south.

In a more detailed picture, transpression is expected to be very pronounced along the Lebanon segment where the main Transform fault, here called the Yammouneh fault (Fig. 10.2.2), bends 30° and more to the right relative to the southern part of the Transform (Dubertret 1955b, Quennell 1959). Indeed considerable contractional deformation characterizes this segment of the Transform. Most obvious is the formation and accentuation of major flexures along the flanks of the Lebanon and Hermon–Anti Lebanon ranges and within them, as well as along the intervening Beqa'a syncline, which also contributed to the great uplifting of this area. Though these structures formed in part before the Transform motion, they were also deformed during the motion. This is evidenced by the deposition of thick Miocene–Pliocene conglomerates in the Beqa'a Valley and by their subsequent flexing (Dubertret 1955b, 1966), which record uplifting and erosion of the Lebanon range and flexing of its flanks. The recent accentuation of the Palmyra folds also expresses compression in this region, though much of the deformation of these folds predates the Transform (Walley 1998). Additional deformation resulted from the right lateral motion of a set of NNE–SSW tending faults that cross Lebanon north of about 33°45′N. Such faulting is expected to lead to rotation of the fault blocks about vertical axes, which would lead to some east–west shortening (Freund et al. 1970). Indeed, paleomagnetic studies revealed that such rotations, amounting to tens of degrees and leading to east–west shortening and north–south extension, are widespread in the Galilee and in Lebanon (Ron et al. 1984, Ron 1987). Furthermore, some reverse motion on the Yammouneh fault may also occur, similar to the situation where transpression takes place further south (e.g. Rotstein & Bartov 1989, 1992).

Further north the Transform trace bends considerably to the left, striking close to north–south, so here there is hardly any convergence of the Transform flanks, which allows the development of the Ghab pull-apart basin. Still further north the Transform bends again to the right along the Kara-Su valley that seems to be a synclinal valley and maybe delimited by reverse faults, but the details of the structure were not documented. West of it the Amanus range rises to an elevation of 1.7–2.2 km. The Transform flanks are considerably deformed in these regions, but the available data do not allow us to evaluate the relations between their structure and the lateral motion.

These considerations show that while the main features of the northern part of the Transform can be related to its lateral offset, there are still many unresolved questions, mainly regarding the quantitative interpretation of the structure.

An important problem is the role of the major faults that splay off the main Transform trace, especially the splays north of the Hula basin. One of them is the Serrhaya fault that extends along the Hermon–Anti-Lebanon range and on which Walley (1998) postulated a lateral offset of about 50 km. However, significant (>10 km?) lateral offset along this fault is difficult to reconcile with its termination in an area where very little deformation was documented.

The Roum fault in southern Lebanon is another prominent splay. The left-lateral offset of major streams crossing its southern part proves its recent motion, as does the regional seismicity (Garfunkel et al. 1981, Butler et al. 1997). However, the stream offsets decrease and disappear northward and the fault becomes indistinct south of Beirut, where it was interpreted to terminate (Dubertret 1955b). Therefore a large lateral offset all along this fault and its continuation to the Mediterranean coast (as suggested by Butler et al. 1997) appear very problematic.

These considerations favor the view that the splay faults did not take up much of the lateral Transform motion, implying that most of this motion was transmitted northward along the Yammouneh fault and the Ghab basin. This is corroborated by the matching of the Palmyra structures with those of Lebanon (Walley 1998) and by the approximately 80 km offset of the Cretaceous nappe front in northern Syria

(Al-Maleh et al. 1992). The difference between the latter figure and the offset further south resulted from the north–south extension west of the Transform in northern Israel and Lebanon, and possibly it also reflects the diversion of some lateral motion to splay faults.

The most outstanding question regarding the northern part of the Transform is whether the known deformation of its flanks can accommodate the shortening that is predicted by plate kinematics. At present a detailed examination of this problem is possible only along the Lebanon segment, because only there is considerable structural information available. Here the predicted contraction is more than 35 km, reaching a maximum of up to about 50 km, but the deformation of the folds and major flexures during the Transform history can hardly account for even half the expected shortening. Block rotations on vertical axes provide an additional efficient deformation mechanism, but their quantitative role is difficult to evaluate in the absence of sufficient data. North–south extension west of the Transform (>10 km?) and the activity of the Roum fault also contributed to the required deformation. However, if the latter effects are significant, then the question arises as to the nature of the junction between the deformed area west of the Transform and the Levant basin, where the post-Miocene, and probably older, deformation was very limited. Important insights regarding these questions could be gained from knowledge of the exact offset of the Homs volcanic field. As noted above, its shape suggests northward reduction of the offset along the main Transform trace, which in turn can be related to the deformation of the Transform flanks or to diversion of the motion to splay faults. Clearly much more detailed information is needed regarding the structure in the northern part of the Transform.

A.5.3 Summary

The structural organization along the Transform, namely the dependence of the development of basins, saddles, and contractional deformation on the sense of discontinuous stepping and sinuosities of the major longitudinal faults is typical and diagnostic of left-lateral fault zones. Most of the lateral motion was transmitted along the main Transform trace, but some lateral motion was diverted to fractures along the Transform flanks (the marginal faults crossing the flanks of the Gulf of Aqaba, splay faults branching from the northern part of the Transform). In particular, a small part of the motion was absorbed by north–south extension west of the Transform in northern Israel and Lebanon, and perhaps farther north.

Along most of the southern half of the Transform its flanks diverged to some extent, the divergence increasing in the later stages of its history. In contrast, the northern part of the Transform is essentially transpressional. This explains the limited development of pull-apart basins, the accentuated uplifting of its flanks and especially their widespread contractional deformation.

However, insufficient knowledge of the structure in the subsurface hinders quantitative interpretation of the structure. In particular, the fault pattern within the major pull-aparts and the mode of their growth, and the way in which the expected shortening along the northern part of the Transform is accommodated, still need clarification. This should not obscure the clear overall structural relations, and the absence of structures typical of extensional rifts.

A.6 RECORD OF RECENT MOTION

The character of the very young and ongoing activity along the Dead Sea Transform is revealed by the nature of the active faults, by the seismic activity and focal plane mechanisms in particular, and by geodetic measurements. This information provides additional insights regarding the nature of the tectonic activity of the Transform, even though they concern only a very small fraction of its history.

Most important in the present context is the presence of active strike-slip faults along much of the onland part of the Transform valley. They were first recognized by Quennell (1959) and by Zak & Freund (1966), who identified several small pull-aparts along one of the segments. Additional features leading to their interpretation as strike-slip faults were summarized by Garfunkel et al. (1981). It is these faults that continue along the borders of the main pull-apart basins, such as the Dead Sea basin. These faults displace mostly very young sediments, and therefore only a short period of their activity is recorded. Therefore, and because of their lateral motion, they often have a rather subtle expression. Nevertheless, they have been trenched and studied in several locations: north of the Dead Sea, north of Elat and in the Arava (Reches & Hoexter 1981, Gardosh et al. 1990, Amit et al. 1999, Zhang et al. 1999). In all cases very young deposits are considerably deformed, which attests to the importance of these fault lines.

The Dead Sea Transform is well known to be seismically active. In the present context the important observation is that the two strongest instrumentally recorded earthquakes of July 11th, 1927 and November 22nd, 1995, with magnitudes well above 6, had focal mechanisms indicative of left lateral motion parallel to the Transform (Ben Menahem et al. 1976, Shamir et al. 1996). The magnitudes of these earthquakes show that they originated by slip on fractures that were several tens of kilometers long. The focal mechanisms of weaker earthquakes are variable, but a significant proportion shows left-lateral slip along the Transform (van Eck & Hofstetter 1989, 1990). Geodetic measurements also demonstrate the dominance of ongoing left-lateral slip along the Dead Sea Transform (Pe'eri et al. 1999).

A.7 CONCLUDING REMARKS

The foregoing account aimed to present the main observations and considerations regarding motion along the Dead Sea Transform – its nature, magnitude, history and relations with the structure along the Transform and its flanks. Several basic points emerge. First, two independent lines of evidence, namely matching the Transform flanks and analysis of the regional plate kinematics, lead to the same conclusion, that a left-lateral offset of close to 100 km took place along the Transform. Crucial to the latter analysis is the estimate of the amounts of opening of the Red Sea and Gulf of Aden, which are based on their crustal structure. Second, the formation of extensional as well as contractional structures along the Transform can be related to variations in the strike of the major lateral faults, typical of left-lateral motion. Moreover, the relations between the major structural elements – deep depressions and structural saddles and uplifts – along the Transform are as expected along a somewhat irregular left-lateral fault line. On the other hand, the discontinuity of the geological features across the Transform and its structural peculiarities are very different from extensional rifts, and cannot be readily explained as recording merely extension across the Transform. Combined, these points provide the framework in which to interpret the Dead Sea Transform, and actually justify its designation as "Transform" rather than "Rift", though respect for an old tradition can warrant the use of the latter term.

This framework allows us to interpret the features of the Transform by combining the local geological data with plate-kinematic considerations based on data from other plate boundaries in the region. This greatly helps to overcome the incompleteness of the local data, and also offers many opportunities to test the interpretations. Use of this broad approach leads to a self-consistent interpretation of all known features and allows us to explain them as resulting from the lateral offset. In particular it allows us to interpret the structural variability within the simple framework of motion along a strike-slip fault zone with small irregularities in plan view.

While the analysis is quite successful in qualitatively explaining the known structures, and relating them to the lateral motion is the only systematic model offered hitherto, quantitative understanding of the structure encounters significant difficulties. Two main questions stand out. First, how is the predicted shortening of the Transform flanks bordering its transpressive segments, for example in Lebanon, accommodated? Second, how is the growth of the pull-apart basins by becoming longer expressed in the structure of their fill? All presently identified structures are compatible with the expected deformation and contribute to it, but it is not clear that they really account for the required amount of deformation. In evaluating the significance of these difficulties it should be stressed that important, sometimes crucial, information is still missing, such as the structure of the edges of the pull-apart basins, the role of block rotations and lateral motions along the transpressive segments, and more. Therefore, future work is needed to test whether these difficulties are real or just reflect the incompleteness of the data. It should not be overlooked, however, that similar quantitative questions have not been resolved anywhere else, and therefore they should not overshadow the other contributions to understanding the Dead Sea Transform. Neither should the great difference between the Dead Sea Transform and extensional rifts be overlooked.

References

Abed, A.M. 1985. Geology of the Damya formation. *Dirasat* 12(2): 99–108.

Abou Ouf, M.A. & Gheith, A.M. 1998. Sedimentary evolution of early rift troughs on the central Red Sea margin, Jeddah, Saudi Arabia. In B.H. Purser & D.W.J. Bosence (eds), *Sedimentation and tectonics in rift basins: Red Sea–Gulf of Aden*: 135–145. London: Chapman & Hall.

Abraham, H. 1960. *Asphalts and allied substances* (6th edn): 370 pp. Princeton, NJ: van Nostrand.

Agnon, A. 1983. An attempted revision of the Neogene stratigraphy in the Dead Sea Valley. Nazerat: Abst., Israel Geol. Soc. Ann. Mtg., 1.

Aharoni, E. 1964. Litho-electric correlation of the Kurnub Group (Lower Cretaceous) in the northern Negev. *Israel J. Earth Sci.* 13: 63–81.

Aharoni, E. & Aizin, B. 1966. Yatir Conglomerate – continental Neogene in the Kuseife area of northern Negev, Israel. *Israel J. Earth Sci.* 15: 118–124.

Albright, W.F. 1924. The archaeological results of an expedition to Moab and the Dead Sea. *Bull. Am. Sch. Oriental Res.* 14: 2–12.

Albright, W.F. 1960. *The archaeology of Palestine* (5th edn): 271 pp. London: Pelican (Penguin) Books.

Aldersons, E., Ben-Avraham, Z., Hofstetter, A., Kissling, E. & Al-Yasjeen, T. 2000. Focal depths of earthquakes and crustal rheology in the Dead Sea basin. Ma'alot: Abst., Israel Geol. Soc. Ann. Mtg., 2.

Al Laboun, A.A. 1986. Stratigraphy and hydrocarbon potential of the Paleozoic succession in both Tabuk and Widyan basins, Saudi Arabia. In M.T. Halbouty (ed.), Future petroleum provinces of the world. *AAPG Mem.* 40: 373–397.

Almagor, G. & Garfunkel, Z. 1979. Submarine slumping in continental margin of Israel and northern Sinai. *AAPG Bull*. 63: 324–340.

Al-Maleh, K., Delaune, M., Mouty, M. & Parrot J.F. 1992. Relations du front de la nappe ophiolitique du nord-ouest syrien avec son substratum de la part et d'autre de la faille du levant: Baer-Bassit, Kurd-Dagh. *C.R. Acad. Sci. Paris* 314: 1195–1202.

Almogi-Labin, A., Perelis-Grossowicz, L. & Raab, M. 1991. *Ammonia beccarii tepida* – a doubtful indicator of a Pliocene marine ingression into the southern Jordan Valley, Israel. *Israel J. Earth Sci.* 40(1–4): 255–256.

Almogi-Labin, A., Perelis-Grossowicz, L. & Raab, M. 1992. Living *Ammonia* from a hypersaline inland pool, Dead Sea area, Israel. *J. Foram. Res.* 22(3): 257–266.

Almogi-Labin, A., Siman-Tov, R., Rosenfeld, A. & Debard, E. 1995. Occurrence and distribution of the foraminifer *Ammonia beccarii tepida* (Cushman) in water bodies, recent and Quaternary, of the Dead Sea Rift, Israel. *Mar. Micropaleont.* 26: 153–159.

Alsharhan, A.S. & Nairn, A.E.M. 1997. *Sedimentary basins and petroleum geology of the Middle East*: 843 pp. Amsterdam: Elsevier.

Al-Subbary, A.K., Nichols, G.J., Bosence, D.W.J. & Al-Kadasi, M. 1998. Pre-rift doming, peneplanation or subsidence in the southern Red Sea? Evidence from the Medj-zir Formation (Taliwah Group) of western Yemen. In B.H. Purser & D.W.J. Bosence (eds), *Sedimentation and tectonics in rift basins: Red Sea–Gulf of Aden*: 119–134. London: Chapman & Hall.

Ambraseys, N.N. & Melville, C.P. 1988. An analysis of eastern Mediterranean earthquake of 20 May 1202. In K. Lee (ed.), *Historical seismograms and earthquakes of the world*: 181–200. New York: Academic Press.

Ambraseys, N.N., Melville, C.P. & Adams, R.D. 1994. *The seismicity of Egypt, Arabia and the Red Sea*: 181 pp. Cambridge: Cambridge Univ. Press.

Amiran, D.H.K. 1951. A revised earthquake catalogue of Palestine. *Israel Explor. J.* 1: 223–246.

Amiran, D.H.K., Arieh, E. & Turcotte, T. 1994. Earthquakes in Israel and adjacent areas: macroseismic observations since 100 B.C.E. *Israel Explor. J.* 44: 260–305.

Amit, O. & Bein, A. 1982. Organic matter in Senonian phosphorites from Israel – origin and diagenesis. *Chem. Geol.* 37: 277–287.

Amit, R., Zilberman, E., Porat, N. & Enzel, Y. 1997. Geomorphic and paleoseismic evidence for large historical earthquake in the Avrona playa, Dead Sea Rift. Kefar Gil'adi: Abst., Israel Geol. Soc. Ann. Mtg., 3.

Amit, R., Zilberman, E., Porat, N. & Enzel, Y. 1999. Relief inversion in the Avrona playa as evidence of large-magnitude historical earthquakes, southern Arava Valley, Dead Sea Rift. *Quat. Res*. 52: 76–91.

Anderson, H.J. 1852. Geological reconnaissance of part of the Holy Land. In W.F. Lynch (ed.), *Official report of the United States' expedition to explore the Dead Sea and the river Jordan*: 75–206. Baltimore: Murphy.

Andrews, I.J. 1991. Paleozoic lithostratigraphy in the subsurface of Jordan. Natural Resources Authority: *Subsurface Geol. Bull*. 2: 1–30.

Andrews, I.J. 1992. Permian, Triassic and Jurassic lithostratigraphy in the subsurface of Jordan. *Subsurface Geol. Bull.* 4: 1–60.

Aoust, V. d', 1834. Nouvelle note relative à l'origine des bitumens. *Bull. Soc. Géol. France (Ser. 4)* IV: 372.

Arad, V. & Bartov, Y. 1994. Arava region, Israel and Jordan, bibliography of geological research. Jerusalem: Geol. Surv. Israel Rep. GSI/15/94, 86 pp.

Arad, V., Beyth, M., Bartov, Y. & Gavrieli, I. 1998. The Dead Sea and its surroundings, geological and limnological research bibliography (2nd edn). Jerusalem: Geol. Surv. Israel Rep. GSI/2/98, 277 pp.

Arad, V., Bartov, Y. & Horowitz, A. 1997. The Lower Jordan Valley (Lake Kinneret–Dead Sea). Geology and water resources bibliography. Jerusalem: Geol. Surv. Israel Rep. GSI/8/97, 121 pp.

Arbenz, J.K. 1984. Oil potential of the Dead Sea area. Tel Aviv: Sismica Oil Exploration Ltd., Rep. 84/111, 44 pp.

Argyriadis, I., de Graciansky, P.C., Marcoux J. & Ricou, L.E. 1980. The opening of the Mesozoic Tethys between Eurasia and Arabia–Africa. *Mém. B.R.G.M*. 115: 199–214.

Arieh, E. 1967. Seismicity of Israel and adjacent areas. *Bull. Geol. Surv. Israel* 43: 1–14.

Arieh, E. 1985. A note on the seismicity of Israel (1900–1982). *Bull. Seis. Soc. Am.* 75: 881–887.

Arkin, Y. 1989. Large-scale tensional features along the Dead Sea–Jordan Rift Valley. *Tectonophysics* 165: 143–154.

Arkin, Y., Nathan, Y. & Starinsky, A. 1972. Paleocene–early Eocene environments of deposition in the northern Negev (southern Israel). *Bull. Geol. Surv. Israel* 56: 1–17.
Arkin, Y., Ichoku, C. & Karnieli, A. 1999a. Fault traces in the arid Arava Valley floor, Israel, revealed by RADARSAT surface roughness classification. *Canadian J. Remote Sensing* 25(3): 302–311.
Arkin, Y., Karnieli, A. & Ichoku, C. 1999b. Potential use of coherence and interferometric images in geologic interpretation: an example from central Jordan Valley. En Boqeq: Abst., Israel Geol. Soc. Ann. Mtg., 5.
Ashbel, D., 1961. Pleistocene climate in view of old Dead Sea shorelines. Jerusalem: Symp. on the Pleistocene in Israel, Hebrew Univ. Jerusalem, 38–39 (in Hebrew).
Avni, Y., 1991. The geology, paleogeography and landscape evolution in the central Negev highlands and the western Ramon structure. Jerusalem: Geol. Surv. Israel Rep. GSI/6/91, 180 pp. (in Hebrew, English Abst.).
Avni, Y. 1998. Paleogeography and tectonics of the central Negev and the Dead Sea Rift western margin during the late Neogene and Quaternary. Jerusalem: Geol. Surv. Israel Rep. GSI/24/98, 231 pp. (in Hebrew, English Abst.).
Avni, Y. & Rosenfeld, A. 1996. The Pliocene–Pleistocene Kuntilla Lake (southern Negev) – paleoecology and facies relations. *Geol. Surv. Israel, Curr. Res.* 10: 51–54.
Avnimelech, M. 1936. Étude géologique dans la région de la Shephelah en Palestine. Travaux du Laboratoire Géologique de l'Université de Grenoble, 20: 1–144.
Avnimelech, M. 1937. Sur un grès lacustre à l'est du lac de Tibériade et sur ses relations avec la structure du fossé du Jourdain. *C.R. Soc. Géol. France*: 213–214.
Avnimelech, M. 1938. The use of mollusks and of their shells by prehistoric man in Palestine. *HaTeva Veha'Aretz* 5(2): 57–61 (in Hebrew).
Avnimelech, M. 1939. On the geology and morphology of the Megiddo area. *J. Palestine Oriental Soc.* 19: 18–37.
Avnimelech, M. 1948. *The secrets of our rocks, chapters in the geological history of Palestine*. Merhavia (Israel): Sifriat Poalim, 324 pp. (in Hebrew).
Avnimelech, M. 1956. On the history of Lake Huleh. *Bull. Israel Explor. Soc.* 20(3–4): 117–120 (in Hebrew).
Avnimelech, M.A. 1965, 1969. *Bibliography of Levant geology*: Vol. 1: 208 pp.; Vol. 2: 195 pp. Jerusalem: Israel Program for Scientific Translations.
Avnimelech, M. & Steinitz, H. 1951. On fish remains in the continental Neogene of the Jordan Valley (Israel). *Eclogae Geol. Helv.* 44(2): 443–447.
Ayalon, A., Bar-Matthews, M. & Sass, E. 1998. Rainfall–recharge relationships within a karstic terrain in the eastern Mediterranean semi-arid region, Israel: $\delta^{18}O$ and δD characteristics. *J. Hydrol.* 207: 18–31.
Ayalon, A., Bar-Matthews, M. & Kaufman, A. 1999. Petrography, strontium and uranium concentrations, and strontium and uranium isotope ratios in speleothems as paleoclimatic proxies: Soreq Cave, Israel. *The Holocene* 9(6): 715–722.
Aydin, A. & Nur, A. 1982. Evolution of pull-apart basins and their scale independence. *Tectonics* 1: 91–105.
Azzaroli, A. 1968. On the evolution of the Gulf of Aden. Prague: Intl. Geol. Cong. 23, Sec. 1: 125–134.
Baer, G. 1981. The geology of the Arif-Batur lineament, Ma'ale HaMeshar area. MSc thesis, Dept. Geol., Hebrew Univ. Jerusalem, 56 pp. (in Hebrew, English summary, unpubl.).
Ball, M.W. & Ball, D. 1953. Oil prospects of Israel. *AAPG Bull.* 37(1): 1–113.
Baker, S. (ed.) 1994. Sedom Deep 1, geological completion report. Tel Aviv: Oil Exploration (Investments) Ltd., 99 pp.

Bandel, K. 1981. New stratigraphical and structural evidence for lateral dislocation in the Jordan Rift Valley connected with a description of a Jurassic rock column in Jordan. *N. Jb. Geol. Paläont. Abh. V.* 161: 271–308.
Bandel, K. & Khoury, H. 1981. Lithostratigraphy of the Triassic in Jordan. *Facies* 4: 1–26.
Barberi, F., Capaldi, G., Gasparini, P., Marinelli, G., Santacroce, R., Scandone, R., Treuil, M. & Varet, J. 1980. Recent basaltic volcanism of Jordan and its implication for the geodynamic history of the Dead Sea shear zone. *Atti Convegni Lincei* 47: 617–628.
Bar-Matthews, M., Ayalon, A. & Kaufman, A. 1997. Late Quaternary paleoclimate in the eastern Mediterranean region from stable isotope analysis of speleothems at Soreq Cave, Israel. *Quat. Res.* 47: 155–168.
Bar-Matthews, M., Ayalon, A. & Kaufman, A. 1998a. Middle to late Holocene (6,500 year period) in the eastern Mediterranean region from stable isotope composition of speleothems at Soreq Cave, Israel. In A.S. Issar & N. Brown (eds), *Water, environment and society in times of climatic change*: 203–214. The Netherlands: Kluwer.
Bar-Matthews, M., Ayalon, A. & Kaufman, A. 1998b. Paleoclimate evolution in the eastern Mediterranean region during the last 58,000 years as derived from stable isotopes of speleothems (Soreq Cave, Israel). IAEA-SM-349/17, 673–682.
Bar-Matthews, M., Ayalon, A. & Kaufman, A. 2000. Timing and hydrological conditions of sapropel events in the eastern Mediterranean as evident from speleothems, Soreq Cave, Israel. *Chem. Geol.* 169: 145–156.
Bartov, Y. 1974. A structural and paleogeographical study of the central Sinai faults and domes. PhD dissertation, Dept. Geol., Hebrew Univ. Jerusalem, 143 pp. (in Hebrew, English summary, unpubl.).
Bartov, Y. 1990. Geological photomap of Israel and adjacent countries, scale 1 : 750,000. Jerusalem: Geol. Surv. Israel.
Bartov, Y. 1993. Geology of Israel. *Encyclopaedia Hebraica* (2nd edn) 6(1): 21–72. Jerusalem: Encyclopaedia Publ. Co. (in Hebrew).
Bartov, Y. 1994. Geological photomap of Israel and adjacent areas, 1 : 750,000. Jerusalem: Geol. Surv. Israel.
Bartov, Y. & Bein, A. 1994. The geology and hydrogeology of the central Arava, between Yotvata and the Paran. Jerusalem: Geol. Surv. Israel Rep. GSI/5/94, 20 pp. (in Hebrew).
Bartov, Y. & Steinitz, G. 1977. The Judea and Mount Scopus groups in the Negev and Sinai with trend surface analysis of the thickness data. *Israel J. Earth. Sci.*, 26: 119–148.
Bartov, Y. & Steinitz, G. 1978. Horizontal displacements along faults, eastern Sinai. Jerusalem: Geol. Surv. Israel Rep. MM/2/78, 7 pp. (in Hebrew).
Bartov, Y., Eyal, Y., Garfunkel, Z. & Steinitz, G. 1972. Late Cretaceous and Tertiary stratigraphy and paleogeography of southern Israel. *Israel J. Earth Sci*. 21: 69–97.
Bartov, Y., Mimran, Y. & Karcz, I. 1976. Lineaments in the coastal plain of Israel. Jerusalem: Geol. Surv. Israel Rep. MM/4/76, 14 pp.
Bartov, Y., Lewy, Z., Steinitz, G. & Zak, I. 1980a. Mesozoic and Tertiary stratigraphy, paleogeography and structural history of the Gebel Areif en Naqa area, eastern Sinai. *Israel J. Earth Sci.* 29: 114–129.
Bartov, Y., Steinitz, G., Eyal, M. & Eyal, Y. 1980b. Sinistral movement along the Gulf of Aqaba – its age and relation to the opening of the Red Sea. *Nature* 285: 220–222.
Bartov, Y., Sneh, A., Fleischer, L., Arad, V. & Rosensaft, M. 2000a. Map of potentially active faults in Israel. Jerusalem: Geol. Surv. Israel.
Bartov, Y., Agnon, A., Enzel, Y., Reches, Z. & Stein, M. 2000b. Reconstruction of Lake Lisan levels by sequence stratigraphy of shore and lacustrine sediments at Massada. Ma'alot: Abst., Israel Geol. Soc. Ann. Mtg., 16.

Baruch, U. 1983. The palynology of a late Holocene core from Lake Kinneret. MSc thesis, Dept. Prehist., Hebrew Univ. Jerusalem, 33 pp. (unpubl.).

Baruch, U. 1993. The palynology of late Quaternary sediments of the Dead Sea. PhD dissertation, Dept. Prehist., Hebrew Univ. Jerusalem, 258 pp. (in Hebrew, English summary, unpubl.).

Bar-Yosef, O. 1975. Archaeological occurrences in the middle Pleistocene of Israel. In K.W. Butzer & G.L. Isaac (eds), *After the Australopithecines*: 571–604. The Hague: Mouton.

Bar-Yosef, O. 1980. Prehistory of the Levant. *Ann. Rev. Anthropol.* 9: 101–133.

Bar-Yosef, O. 1987. Prehistory of the Jordan Rift. *Israel J. Earth Sci.* 36: 107–120.

Bar-Yosef, O. 1994. The Lower Paleolithic of the Near East. *J. World Prehist.* 8(3): 211–265.

Bar-Yosef, O. 1998. Jordan prehistory: a view from the west. In D.O. Henry (ed.), *The prehistoric archaeology of Jordan*: 162–178. BAR Intl. Ser. 705.

Bar-Yosef, O. *Prehistory of the Holy Land* (in prep.).

Bar-Yosef, O. & Goren-Inbar, N. 1993. *The Lithic Assemblages of Ubeidiya: a Lower Paleolithic site in the Jordan Valley*: 266 pp. Jerusalem: Hebrew Univ.

Bar-Yosef, O. & Tchernov, E. 1972. *On the palaeo-ecological history of the site of Ubeidiya*: 35 pp. Jerusalem: Israel Acad. Sci. Hum.

Bar-Yosef, O., Goldberg, P. & Leveson, T. 1974. Kebaran and Natufian sites in Wadi Fatza'el, Jordan Valley, Israel. *Paléorient* 2(2): 415–428.

Baubron, J.C., Besançon, J., Copeland, L., Hours, F., Macaire, J.J. & Sanlaville, P. 1985. Évolution de la moyen vallée du Zarqa (Jordanie) au Néogène et au Quaternaire. *Rév. Géol. Dyn. Géog. Phys*. 26(5): 273–283.

Bayer, J. 1918. Die Jugendlichkeit der ältesten Kultur Palästinas. *Mitt. Anthrop. Ges. Wien* 49: 15–23.

Bayer, J. 1922. Bemerkungen zu M. Blanckenhorn: Die Steinzeit Palästina-Syriens und Nordafrikas. *Mitt. Anthrop. Ges. Wien* 52(5–6): 279.

Bayer, H.J., Hötzl, H., Jado, A.R., Ruscher, B. & Voggenreiter, W. 1988. Sedimentary and structural evolution of the northwest Arabian Red Sea margin. *Tectonophysics* 153: 137–151.

Bechor, N. & Wdowinski, S. 1999. Current motion of the Sinai subplate from GPS observations. En Boqeq: Abst., Israel Geol. Soc. Ann. Mtg., 14.

Begin, Z.B. 1975a. The geology of the Jericho sheet (Geological map series 1 : 50,000). *Bull. Geol. Surv. Israel* 67: 1–35.

Begin, Z.B. 1975b. Paleocurrents in the Plio-Pleistocene Samra Formation (Jericho region, Israel) and their tectonic implication. *Sed. Geol.* 14: 191–218.

Begin, Z.B. & Zilberman, E. 1997. Main stages and rate of relief development in Israel. Jerusalem: Geol. Surv. Israel Rep. GSI/24/97, 63 pp. (in Hebrew, English summary).

Begin, Z.B., Ehrlich, A. & Nathan, Y. 1974. Lake Lisan, the Pleistocene precursor of the Dead Sea. *Bull. Geol. Surv. Israel* 63: 1–30.

Begin, Z.B., Nathan, Y. & Ehrlich, A. 1980. Stratigraphy and facies distribution in the Lisan Formation, new evidence from the area south of the Dead Sea, Israel. *Israel J. Earth Sci.* 29(3): 182–189.

Bein, A. 1967. Upper peat – Hula Valley; summary report. Jerusalem: Geol. Surv. Israel Rep., 6 pp. (in Hebrew).

Bein, A. 1974. Reef complexes in the Judea Group (Mt. Carmel and the Coastal Plain). PhD dissertation, Dept. Geol., Hebrew Univ. Jerusalem, 154 pp. (in Hebrew, English summary, unpubl.).

Bein, A. 1976. Rudist fringing reefs of Cretaceous shallow carbonate platform of Israel. *AAPG Bull*. 60: 258–272.

Bein, A. & Amit, O. 1982. Depositional environments of the Senonian chert, phosphorite and oil share sequence in Israel as deduced from their organic matter composition. *Sedimentology* 29: 81–90.

Bein, A. & Gvirtzman, G. 1977. A Mesozoic fossil edge of the Arabian Plate along the Levant coastline and its bearing on the evolution of the eastern Mediterranean. In B. Biju-Duval & L. Montadert (eds), Split (Yugoslavia): Proc. Intl. Symp. struct. hist. medit. basins: 95–110.

Bein A. & Horowitz, A. 1986. Papyrus – a historic newcomer to the Hula Valley, Israel? *Rev. Palaeobot. Palynol.* 47: 89–95.

Bein, A. & Weiler, Y. 1976. The Cretaceous Talme Yafe Formation: A contour current shaped sedimentary prism of calcareous detritus at the continental margin of the Arabian Craton. *Sedimentology* 23: 511–532.

Belitzky, S. 1987. Tectonics of the Korazim saddle. MSc thesis, Dept. Geol., Hebrew Univ. Jerusalem, 94 pp. (in Hebrew, English Abst., unpubl.).

Belitzky, S. 1996. Tectonic geomorphology of the lower Jordan Valley – an active continental rift. PhD dissertation, Dept. Geol., Hebrew Univ. Jerusalem, 98 pp. (in Hebrew, English summary, unpubl.).

Belitzky, S. Paleodrainage and morphotectonic evolution at the Gesher Benot Ya'aqov area, northern Dead Sea Rift, Israel (in prep.).

Belitzky, S. & Mimran, Y. 1996. Active salt diapirism at the Zahrat El-Qurein dome, lower Jordan Valley, Jordan. *Israel J. Earth Sci.* 45(1): 11–18.

Ben-Avraham, Z. 1985. Structural framework of the Gulf of Elat (Aqaba), northern Red Sea. *J. Geophys. Res.* 90: 703–726.

Ben-Avraham, Z. 1997. Geophysical framework of the Dead Sea. In T.M. Niemi, Z. Ben-Avraham & J.R. Gat (eds), *The Dead Sea, the lake and its setting*: 22–35. Oxford: Oxford Univ. Press.

Ben-Avraham, Z. & Ginzburg, A. 1990. Displaced terrains and crustal evolution of the Levant and eastern Mediterranean. *Tectonics* 9: 613–622.

Ben-Avraham, Z. & Hall, J.K. 1977. Geophysical survey of Mount Carmel structure and its extension into the Eastern Mediterranean. *J. Geophys. Res*. 82: 793–802.

Ben-Avraham, Z. & ten Brink, U.S. 1989. Transverse faults and segmentation of basins within the Dead Sea rift. *J. African Earth Sci*. 8: 603–636.

Ben-Avraham, Z., Hänel, R. & Villiger, H. 1978. Heat flow through the Dead Sea Rift. *Mar. Geol.* 28: 253–269.

Ben-Avraham, Z., Almagor, G. & Garfunkel, Z. 1979. Sediments and structure of the Gulf of Elat. *Sed. Geol.* 23: 239–267.

Ben-Avraham, Z., Shoham, Y., Klein, E., Michelson, H. & Serruya, C. 1980. Magnetic survey of Lake Kinneret – central Jordan Valley, Israel. *Mar. Geophys. Res.* 4: 257–276.

Ben-Avraham, Z., Ginzburg, A. & Yuval, Z. 1981. Seismic reflection and refraction investigations of Lake Kinneret – Central Jordan Valley, Israel. *Tectonophysics* 80: 165–181.

Ben-Avraham, Z., Levy, Y., Hall, J.K. & Neev, D. 1985. The structure and composition of the uppermost sedimentary layers in the Dead Sea. Tel Aviv Univ. Rep. TAU 1/85, 1–26.

Ben-Avraham, Z., Shaliv, G. & Nur, A. 1986. Acoustic reflectivity and shallow sedimentary structure in the Sea of Galilee, Jordan Valley. *Mar. Geol.* 70: 175–189.

Ben-Avraham, Z., ten Brink, U.S. & Charrach, J. 1990a. Transverse faults at the northern end of the southern basin of the Dead Sea Graben. *Tectonophysics* 180: 37–47.

Ben-Avraham, Z., Amit, G., Golan, A. & Begin, Z.B. 1990b. The bathymetry of Lake Kinneret and its structural significance. *Israel J. Earth Sci.* 39: 77–84.

Ben-Avraham, Z., Niemi, T.M., Neev, D., Hall, J.K. & Levy, Y. 1993. Distribution of Holocene sediments and neotectonics in the deep north basin of the Dead Sea. *Mar. Geol.* 113: 219–231.

Ben-Avraham, Z., ten Brink, U.S., Bell, R. & Reznikov, M. 1996. Gravity field over the Sea of Galilee: evidence for a composite basin along a transform fault. *J. Geophys. Res.* 101(B1): 533–544.

Ben-Avraham, Z., Niemi, T.M., Heim, C., Negendank, J. & Nur, A. 1999. Holocene stratigraphy of the Dead Sea: correlation of high resolution seismic profiles to sediment cores. *J. Geophys. Res.* 104: 17617–17625.

Benda, L. & Meulenkamp, J.E. 1990. Biostratigraphic correlations in the eastern Mediterranean Neogene. *Newsl. Stratigr*. 23(1): 1–10.

Bender, F. 1963. Zur Geologie der Kupfererzvorkommen am ostrand des Wadi Arabah, Jordanien. Hannover: Bundes. F. Boden., Rep., 96 pp.

Bender, F. 1968a. *Geologie von Jordanien*: 230 pp. Berlin: Bebrüder Bornträger.

Bender, F. 1968b. Über das Alter und die Entstehungsgeschichte des Jordangrabens am beispiel seines Südabschnittes (Wadi Arabah, Jordanien). *Geol. Jb.* 86: 177–196.

Bender, F. (ed.) 1968c. The geological map of Jordan, scale 1 : 250,000, 5 sheets. Hannover: Geol. Sur. Fed. Repub. Germany.

Bender, F. 1974a. *Geology of Jordan.* Berlin: Gebrüder Bornträger, 196 pp. (English translation of above, with minor revisions).

Bender, F. 1974b. Explanatory notes on the geological map of the Wadi Araba, Jordan. *Geol. Jb.* B10: 3–62, 3 maps.

Benjamini, C. 1979. The geological history of central and southern Israel during the Eocene. PhD dissertation, Dept. Geol., Hebrew Univ. Jerusalem (in Hebrew, English summary 41 pp., unpubl.).

Benjamini, C. 1980. Stratigraphy and foraminifera of the Qezi'ot and Har Akrav formations (latest middle to late Eocene) of the western Negev, Israel. *Israel J. Earth Sci.* 29: 227–244.

Benjamini, C. 1984. Stratigraphy of the Eocene of the Arava Valley (eastern and southern Negev, southern Israel). *Israel J. Earth Sci.* 33: 167–177.

Ben-Menahem, A. 1979. Earthquake catalog for the Middle East (92 BC–1980 AD). *Bolletino Geofisica Teoretica e Applicata* 21: 245–313.

Ben-Menahem, A. 1991. Four thousand years of seismicity along the Dead Sea rift. *J. Geophys. Res.* 96: 20195–20216.

Ben-Menahem, A. & Aboodi, E. 1981. Micro- and macroseismicity of the Dead Sea rift and off-coast eastern Mediterranean. *Tectonophysics* 80: 199–233.

Ben-Menahem, A., Nur, A. & Vered, M. 1976. Tectonics, seismicity and structure of the Afro-Eurasian junction – The breaking of an incoherent plate. *Phys. Earth Planet. Interiors* 12: 1–50.

Bentor, Y.K. 1946. The hydrological conditions of the area Kinneret–Tiberias–Yavne'el–Hadata–Kefar Kama–Lubiya. Jerusalem: Palestine Water Co., 76 pp.

Bentor, Y.K. 1952. Report on the copper and manganese deposits in Wadi Menayeh. Jerusalem: Geol. Surv. Israel Rep., 34 pp.

Bentor, Y.K. 1957. Cenozoic volcanism of Israel. Mexico: 20th Intl. Geol. Cong., Sec. 1: 121–136.

Bentor, Y.K. (ed.) 1960. *Lexique stratigraphique international III: Asie; Fasc. 10c2, Israel*: 151 pp. Gap (France): Louis-Jean.

Bentor, Y.K. 1961. Petrographical outline of the Precambrian in Israel. *Bull. Res. Counc. Israel* 10G: 17–64.
Bentor, Y.K. 1985. The crustal evolution of the Arabo-Nubian massif with special reference to the Sinai Peninsula. *Precambrian Res*. 28: 1–74.
Bentor, Y.K. & Vroman, A.J. 1951. The geological map of the Negev 1 : 100,000, sheet 18: Ovdat. Tel Aviv: Israel Army Sci. Div. (in Hebrew).
Bentor, Y.K. & Vroman, A.J. 1954a. A structural contour map of Israel (1 : 250,000), with remarks on its dynamic interpretation. *Bull. Res. Coun. Israel* 4(2): 125–135.
Bentor, Y.K. & Vroman, A.J. 1954b. The geological map of the Negev 1 : 100,000, sheet 16: Mount Sedom, with explanatory notes. Tel Aviv: Israel Army Sci. Div., 94 pp. (in Hebrew).
Bentor, Y.K. & Vroman, A.J. 1955. The geological map of the Negev, Sheet 24, Eilath, 1 : 100,000. Jerusalem: Geol. Surv. Israel.
Bentor, Y.K. & Vroman, A.J. 1957. The geological map of Israel 1 : 100,000, sheet 19: Arava Valley, with explanatory notes. Jerusalem: Geol. Surv. Israel, 66 pp., maps.
Bentor, Y.K. & Vroman, A.J. 1960. The geological map of Israel 1 : 100,000, sheet 16: Mount Sedom, with explanatory notes. Jerusalem: Geol. Surv. Israel, 117 pp., maps. (2nd revised English edn of 1954).
Bentor, Y.K. & Vroman, A.J. 1963. The geological map of Israel 1 : 100,000, sheet 17: Nitsana. Jerusalem: Geol. Surv. Israel.
Bentor, Y.K., Vroman, A.J. & Zak, I. 1965. Geological map of Israel, southern sheet, 1 : 250,000. Tel Aviv: Surv. Israel.
Ben-Tuvia, A. & Golani, D. 1993. Fish of Israel. *Encyclopaedia Hebraica* (2nd edn) 6(1): 254–277. Jerusalem: Encyclopaedia Publ. Co. (in Hebrew).
Bergy, P.A. 1932. Le paléolitique ancien stratifié à Ras Beyrouth. *Mélanges de l'Univ. Saint Joseph, Beyrouth*: XVI.
Bernoulli, D. & Lemoine, M. 1980. Birth and early evolution of the Tethys: the overall situation. *Mém. B.R.G.M*. 115: 168–179.
Bertou, J. de, 1838. Voyage de l'extrémité sud de la Mer Morte à la pointe norde du golfe Elaouitique. *Bull. Soc. Géog. France* 10: 18–32.
Bertou, J. de, 1839. Itinéraire de la Mer Morte à Akaba par les wadys el-Ghor, el-Araba et el-Akaba et retour à Hebron par Pétra. *Bull. Soc. Géog. France* 11: 274–331.
Besançon, J. 1981. Stratigraphie et chronologie du Quaternaire continental du Proche Orient. In P. Sanlaville & J. Cauvin (eds), *Préhistoire du Levant*: 33–53. Lyon: Maison de l'Orient.
Besançon, J. & Hours, F. 1985. Prehistory and geomorphology in northern Jordan: a preliminary outline. In A. Hadidi (ed.), *Studies in the history and archaeology of Jordan* II: 59–66. Amman: Dept. Antiquities.
Besançon, J. & Sanlaville, P. 1984. Terrasses fluviatile au Proche-Orient. *Bull. Assoc. Fr. Étude Quaternaire*: 186–191.
Besançon, J., Copeland, L. & Sanlaville, P. 1988. Réflexions sur les prospections géopréhistoriques au Proche Orient. *Paléorient* 14(2): 31–39.
Bessais, E. & Suc, J.-P. 1987. Climate control over the Neogene geographic and chronologic variations in the west Mediterranean vegetations. Montpellier: Interim Colloquium RCMNS, 14.
Beuning, K. 1998. Late-Glacial vegetation near Lake Edward, east Africa. Puerto Vallarta (México): Abst. 15th Biennial Mtg., 89.
Beydoun, Z.R. 1977. Petroleum prospects of Lebanon: re-evaluation. *AAPG Bull*. 61: 43–64.
Beydoun, Z.R. 1988. *The Middle East: regional geology and petroleum resources*: 292 pp. Beaconsfield: Scientific Press Ltd.

Beyth, M. & Segev, A. 1983. Lower Cretaceous basaltic plug in the Timna Valley. *Israel J. Earth Sci*. 32: 165–166.

Bielski, M. 1982. On the problem of the Lower Cambrian age in the Arabian Nubian Massif: In G.S. Odin (ed.) *Numerical dating in stratigraphy*: 943–947. Chichester: John Wiley & Sons.

Biju-Duval, B., Letouzey, J. & Montadert, L. 1979. Variety of marginal and deep basins in the Mediterranean. *AAPG Mem.* 29: 293–317.

Billings, M.P. 1954. *Structural geology:* 514 pp. Englewood Cliffs, NJ: Prentice-Hall.

Binford, S.R. 1966. Me'arat Shovakh (Mugharet esh-Shubbabiq). *Israel Expl. J*. 16: 18–32, 96–103.

Blake, G.S. 1928. *Geology and water resources of Palestine and Transjordan*: 51 pp. Jerusalem: Govt. Palestine.

Blake, G.S. 1930. *The mineral resources of Palestine and Transjordan. II: Palestine*: 41 pp. Jerusalem: Govt. Palestine.

Blake, G.S. 1935. *The stratigraphy of Palestine and its building stones*: 138 pp. Jerusalem: Govt. Palestine Print. Stationar. Off.

Blake, G.S. 1937. Mineral deposits of Palestine and Transjordan. *Palestine News* 3: 137–141.

Blake, G.S. & Goldschmidt, M.J. 1947. *Geology and water resources of Palestine:* 413 pp. Jerusalem: Govt. Palestine.

Blanckenhorn, M. 1897. Zur Kenntnis der Süsswasserablagerungen und Mollusken Syriens. *Paläontographica* 1(44): 71–144.

Blanckenhorn, M. 1910. Das Klima der Quartärperiode in Syrien-Palästina und Ägypten. Stockholm: Intl. Geol. Cong. I: 140–143.

Blanckenhorn, M. 1912. *Naturwissenschafliche studien am Toten Meer und im Jordantal*: 478 pp. Berlin: Friedländer & Sohn.

Blanckenhorn, M. 1913–1915. *Beiträge zur Paläontologie und Geologie von Palästina und Syrien*: 3 Vols. Stuttgart: Schweizbart.

Blanckenhorn, M. 1914. Syrien, Arabien und Mesopotamien. *Handb. Region. Geologie, Heidelberg* 5(4): 1–159.

Blanckenhorn, M. 1920/1921. Die Steinzeit Palästina-Syriens und Nordafrikas. *Das Land der Bibel* 3(5): 1–48; 3(6): 1–26; 4(1): 1–45.

Blanckenhorn, M. 1931. Geologie Palästinas nach heutiger Auffassung. *Ztschr. Dtsch. Paläst. Ver.* 54: 1–50.

Blanckenhorn, M. & Oppenheim, P. 1927. Neue Beiträge zur Kenntnis des Neogens in Syrien und Palästina. *Geol. Paläont. Abh., Neue Folge* 15(4): 321–358.

Blank, H.R. 1977. Aeromagnetic and geologic study of Tertiary dikes and related structures on the Arabian margin of the Red Sea. *Saudi Arabia Dir. Miner. Res. Bull*. 22: G1–G18.

Bloch, M.R. & Picard, L. 1970. The Dead Sea – a sinkhole. *Z. Deutsch. Geol. Ges., Sonderh. Hydrogeol. Hydrogeochem.*: 119–128.

Block, J.W. 1975. Sodom and Gomorrah: a volcanic disaster. *J. Geol. Education* 23(3): 74–77.

Blow, W.H. 1969. Late middle Eocene to Recent planktonic foraminiferal biostratigraphy. In P. Bronnimann & H.H. Renz (eds), *Proc. 1st Intl. Conf. Planktonic Microfossils, Geneva 1967*: 199–475. Leiden: Brill.

Bonen, D. 1980. The Mesozoic basalts of Israel. PhD dissertation, Dept. Geol., Hebrew Univ. Jerusalem, 156 + 59 pp. (in Hebrew, English summary, unpubl.).

Bonney, T.G. 1904. The Kishon and Jordan valleys. *Geol. Mag.* 1: 575–582.

Boom, G. van den, 1968. Zur Petrogenese der Plateaubasalte Nordostjordaniens. *Geol. Jahrb.* 85: 489–496.

Bosworth, W., Crevello, P., Winn, R.D. Jr. & Steinmetz, J. 1998. Structure, sedimentation and basin dynamics during rifting of the Gulf of Suez and north-western Red Sea. In B.H. Purser & D.W.J. Bosence (eds), *Sedimentation and tectonics in rift basins: Red Sea–Gulf of Aden*, 77–96. London: Chapman & Hall.

Botta, P.E. 1833. Observations sur le Liban et l'Anti-Liban. *Mém. Soc. Géol. France* 1(1): 135–160.

Bowman, D. 1971. Geomorphology of the shore terraces of the late Pleistocene Lisan Lake (Israel). *Palaeogeogr. Palaeoclimatol. Palaeoecol*. 9(3): 183–209.

Braun, D. 1992. The geology of the Afiqim region. MSc thesis, Dept. Geol., Hebrew Univ. Jerusalem, 93 pp. (in Hebrew, unpubl.).

Braun, D. & Tchernov, E. 1991. *Viviparus* beds from central Jordan Valley – the southern correlative of Gesher Benot-Ya'akov Formation. Akko: Abst., Israel Geol. Soc. Ann. Mtg., 18.

Braun, D., Ron, H. & Marco, S. 1991. Magnetostratigraphy of the hominid tool bearing Erk el-Ahmar Formation in the northern Dead Sea Rift. *Israel J. Earth Sci.* 40(1–4): 191–197.

Braun, M. 1963. Geology and petrography of Imara and Sherif anticlines. *Israel J. Earth Sci.* 12(2): 78–79.

Braun, M. 1964. Geology of the Imara–Sherif region. MSc thesis, Dept. Geol., Hebrew Univ. Jerusalem, 65 pp. (in Hebrew, unpubl.).

Braun, M. 1967. Type sections of Avedat Group Eocene formations in the Negev (Southern Israel). Jerusalem: Geol. Surv. Israel Stratig. Sec. 4, 14 pp.

Braun, M. & Hirsch, F. 1987. Controls on Albian–Cenomanian carbonate platform sedimentation in the Middle Eastern Region: Kesalon Event, a Middle Cretaceous sea level change in Israel and its correlation with global sea level changes. *AAPG Bull.* 71(5): 533–534.

Brawer, A.J. 1957. The landscape of Israel. *Encyclopaedia Hebraica* 6: 40–105. Jerusalem: Encyclopaedia Publ. Co. (in Hebrew).

Brenner, I.B. 1979. The geochemical relations and evolution of the Tertiary–Quaternary volcanic rocks in northern Israel. Jerusalem: Geol. Surv. Israel Rep. GD/4/79, 202 pp.

Brotzen, F. 1928. Neue steinzeitliche Funde vom Karmel-gebirge in Palästina. *Berliner Museum, Ber. Preuss. Kunst-samml. Berlin* 48: 119–122.

Brown, G.F., Schmidt, D.L. & Huffman Jr., A.C. 1989. Geology of Arabian Peninsula: Shield area of western Saudi Arabia. US Geol. Surv. Prof. Paper 560-A, 188 pp.

Bruner, I., Gardosh, M. & Rosenthal, E. 2000. The results of a new seismic survey in the Bet-She'an Valley and their implications for the development of water resources. Ma'alot: Abst., Israel Geol. Soc. Ann. Mtg., 23.

Buch, L. von 1841. Letter. In E. Robinson & E. Smith 1841. *Palästina und die südlich angrenzenden länder; Tagebuch einer reise im jahre 1838 in bezug auf die Biblische Geographie Unternommen.* Vol. 2: 673. Halle: Weisenhaus.

Buchbinder, B. 1969. Geological map of Hashefela region, Israel, with explanatory notes. Jerusalem: Geol. Surv. Israel Rep. OD/1/68, 13 pp.

Buchbinder, B. & Siman-Tov, R. 2000. Oligocene clastics in the Ashdod area: mass flow and turbidite deposits in a submarine upper-fan setting. Ma'alot: Abst., Israel Geol. Soc. Ann. Mtg., 25.

Buchbinder, B. & Sneh, A. 1983–1984. Marine sandstones and terrestrial conglomerates and mudstones of Neogene–Pleistocene age in the Modi'im area: a re-evaluation. *Geol. Surv. Israel, Curr. Res.* 1983–1984: 65–69.

Buchbinder, B. & Zilberman, E. 1997. Sequence stratigraphy of Miocene–Pliocene carbonate-siliciclastic shelf deposits in the eastern Mediterranean margin (Israel): effects of eustasy and tectonics. *Sed. Geol.* 112: 7–32.

Buchbinder, B., Begin, Z.B. & Friedman, G.M. 1974. Pleistocene algal tufa of Lake Lisan, Dead Sea area, Israel. *Israel J. Earth Sci.* 23(4): 131–138.

Buchbinder, B., Magaritz, M. & Buchbinder, L.G. 1983. Turonian to Neogene paleokarst in Israel. *Paleogeog. Paleoclimatol. Paleoecol.* 43: 329–350.

Burckhardt, J.L. 1822. *Travels in Syria and the Holy Land:* 668 pp. London: Murray.

Büsching, A.F. 1766. *Beschreibung des Todten Meers in Palästina:* 55 pp. Hamburg: Böhm.

Butler, R.W.H., Spencer, S. & Griffiths, H.M. 1997. Transcurrent fault activity on the Dead Sea Transform in Lebanon and its implications for plate tectonics and seismic hazard. *J. Geol. Soc. London* 154: 757–760.

Buzy, D. 1928. Une industrie Mésolithique de Palestine. *Rev. Bibl.* 37: 558–578.

Cailleux, A. 1949. Morphoscopie de quelques sable de Palestine. *Bull. Inst. Egypte* 31 (1948/1949): 177–180.

Callier, C. 1839. Mémoire sur le depression de la Mer Morte et de la vallée du Jourdain. *Nouv. Annal. Voyages* 21: 5–38.

Calvo, R., Bartov, J., Kolodny, Y. & Ayalon, A. 1996. Petrology and geochemistry of lacustrine limestones in the Hazeva Formation: implications for the development of sedimentary basins and morphotectonics of the Rift Valley. *Geol. Surv. Israel, Curr. Res.* 10: 47–50.

Calvo, R., Frieslander, U. & Bartov, Y. 1997. A new stratigraphic division of the Hazeva formation in the Negev and the Arava. Kefar Gil'adi: Abst., Israel Geol. Soc. Ann. Mtg., 13.

Calvo, R., Bartov, Y., Avni, Y., Garfunkel, Z. & Frieslander, U. 1998. Mitzpe Ramon: Israel Geol. Soc. Ann. Mtg. Excursion guidebook, 47–62 (in Hebrew).

Carbone, F., Matteucci, R. & Angelucci, A. 1998. Present-day sedimentation on the carbonate platform of the Dahlak Islands, Eritrea. In B.H. Purser & D.W.J. Bosence (eds), *Sedimentation and tectonics in rift basins: Red Sea–Gulf of Aden*: 523–536. London: Chapman & Hall.

Chazot, G., Menzies, M.A. & Baker, J. 1998. Pre-, syn- and post-rift volcanism on the southwestern margin of the Arabian plate. In B.H. Purser & D.W.J. Bosence (eds), *Sedimentation and tectonics in rift basins: Red Sea–Gulf of Aden*: 50–55. London: Chapman & Hall.

Cheddadi, R. 1988. Palèoclimats au nord de la Mediterranèe oriental depuis 250,000 ans. Analyse pollinique et stratigraphie isotopique de quatre carottes marines. PhD dissertation, Univ. des Sciences et Techniques du Languedoc, Montpellier, 116 pp. (unpubl.).

Christie-Blick, N. & Biddle, K.T. 1985. Deformation and basin formations along strike-slip faults. In K.T. Biddle & N. Christie-Blick (eds), *Strike-slip deformation, basin formation, and sedimentation*: 1–34. Soc. of Econ. Paleontol. Mineral., Special Publication 37.

Clapp, F.G. 1936. Geology and bitumens of the Dead Sea area, Palestine and Transjordan. *AAPG Bull.* 20(7): 881–909.

Clark, J.D. 1961. Fractured chert specimens from the Lower Pleistocene Bethlehem levels. *Bull. Brit. Mus.* 5: 73–99.

Clark, J.D. 1967. The Middle Acheulian occupation site at Latamne, northern Syria. *Quaternaria* 9: 1–68.

Clark, J.D. 1968. The Middle Acheulian occupation site at Latamne, northern Syria; further excavations. *Quaternaria* 10: 1–76.

Clark, S.A., Langozky, Y. & Speight, A.G. 1984. "Har Sedom 1" – recommendation for drilling. Tel Aviv: Sismica Oil Exploration, Rep. 84/113, 13 pp.
Clark, G.A., Schuldenrein, J., Donaldson, M.L., Schwarcz, H.P., Rink, W.J. & Fish, S.K. 1997. Chronostratigraphic context of Middle Paleolithic horizons at the 'Ain Difla rockshelter (WHS 634), west–central Jordan. In H.G.K. Gebel, Z. Kafafi & G.O. Rollefson (eds), *The prehistory of Jordan, II. Perspectives from 1997*: 77–100. Berlin: ex Oriente.
Claude, G. 1929. Une mine d'or dans la Mer Morte. *Rev. Mens. Jerusalem* 24: 468.
Cloos, H. 1939. Hebung, Spaltung, Vulkanismus. *Geol. Rund.* 30: 420–444.
Cochran, J.R. 1981. The Gulf of Aden: structure and evolution of a young ocean basin and continental margin. *J. Geophys. Res.* 86: 263–288.
Cohen, Z. 1971. The geology of the Lower Cretaceous of the Helez field, Israel. PhD dissertation, Dept. Geol., Hebrew Univ. Jerusalem, 98 pp. (in Hebrew, English summary, unpubl.).
Cohen, Z. 1976. Early Cretaceous buried canyon – influence on accumulation of hydrocarbons in Helez oil field, Israel. *AAPG Bull.* 60: 108–114.
Cohen, Z. 1988. Hydrocarbon potential of Israel. Highlights of Basin Analysis. Tel Aviv: OEIL Exploration Dept. (unpubl.).
Cohen, Z., Flexer, A. & Kapstan, V. 1988. The Pleshet Basin, a newly discovered link in the peripheral chain of basins of the Arabian craton. *J. Petrol. Geol.* 11(4): 403–414.
Cohen, Z., Kapstan, V. & Flexer, A. 1990. The tectonic mosaic of the southern Levant: implications for hydrocarbon prospects. *J. Petrol. Geol.* 13(4): 437–462.
Coleman, L. 1867. The great crevasse of the Jordan and of the Red Sea. *Bibliotheca Sacra* 24: 248–262.
Coleman, R.G. 1993. *Geologic evolution of the Red Sea*: 186 pp. New York: Oxford Univ. Press.
Coleman, R.G., Gregory, R.T. & Brown, G.F. 1983. Cenozoic volcanic rocks of Saudi Arabia. US Geol. Surv. Open File Rep. 03–93: 82 pp.
Copeland, L. 1998. The Lower Paleolithic of Jordan. In D.O. Henry (ed.), *The prehistoric archaeology of Jordan*: 5–22. BAR Intl. Ser. 705.
Cox, L.R. 1924. A Triassic fauna from the Jordan Valley. *Ann. Mag. Nat. Hist.* Ser. 9, 14(79): 52–96.
Cox, L.R. 1932. Further notes on the Transjordan Trias. *Ann. Mag. Nat. Hist.* Ser. 10, 10: 93–113.
Crowell, J.C. 1974. Origin of late Cenozoic basins in southern California. In W.R. Dickinson (ed.), *Tectonics and Sedimentation*: 190–204. Soc. Econ. Paleontol. Mineral., Special Publication 22.
Dan, J., Gerson, R., Koyumdjisky, H. & Yaalon, D.H. 1981. *Aridic soils of Israel*: 353 pp. Bet Dagan (Israel): The Volcani Center.
Daniel, E.J. 1963. *Lexique Stratigraphique International*, *Vol. 10 c1: Jordanie*: 293–400. Paris: CNRS.
Danin, A. 1988. Flora and vegetation of Israel and adjacent areas. In Y. Yom-Tov & E. Tchernov (eds), *The zoogeography of Israel*: 129–158. Dordrecht: Junk.
Darlymple, G.B. 1972. Potassium–argon dating of geomagnetic reversals and North American glaciations. In W.W. Bishop & J.A. Miller (eds), *Calibration of Hominoid evolution*: 107–134. Edinburgh: Scottish Academic Press.
Degg, M.R. & Doornkamp, J.C. 1989. *Earthquake hazard atlas*. London: Reinsurance Offices Association, Sec. C: 1–24, D: 1–16.
Delachanal, B. 1883. Sur la composition de l'asphalte ou bitume de Judée. *C.R. Acad. Sci. Paris* 97: 491–494.

Derin, B. 1974. The Jurassic of central and northern Israel. PhD dissertation, Dept. Geol., Hebrew Univ. Jerusalem, 152 pp. (in Hebrew, English summary, unpubl.).
Derin, B. 1976. Haruvit 1 and Haruvit 3. Tel Aviv: Israel Institute of Petroleum and Energy, Rep. 9/10, 15 pp. (unpubl.).
Derin, B. & Gerry, E. 1981. Late Permian–Triassic stratigraphy in Israel and its significance to oil exploration. Jerusalem: Abst. Symposium on Oil Exploration in Israel: 9–11.
Derin, B. & Reiss, Z. 1973. Revision of marine Neogene stratigraphy in Israel. *Israel J. Earth Sci*. 22: 199–210.
De Sitter, L.U. 1962. Structural development of the Arabian shield in Palestine. *Geologie en Minjbouw* 41: 116–124.
Dewey, J.F., Pitman, W.C., Ryan, W.B.F. & Bonin, J. 1973. Plate tectonics and the evolution of the Alpine System. *Geol. Soc. Am. Bull*. 84: 3137–3180.
Dicker, T. 1969. Geology of central Yizre'el Valley. *Israel J. Earth Sci.* 18: 39–66.
Diener, C. 1885. Die struktur des Jordanquellgebietes. *Sitzungsber. Akad. Kaiser. Wiss. Wien* 92(1): 633–642.
Domzalski, W. 1967. Aeromagnetic survey of Israel: Interpretation. Holon (Israel): Institute for Petroleum Research and Geophysics (now Geophysical Institute of Israel) Rep. SMA/482/67, 62 pp.
Druckman, Y. 1974a. The stratigraphy of the Triassic sequence in southern Israel. *Bull. Geol. Surv. Israel* 64: 1–92.
Druckman, Y. 1974b. Triassic paleogeography of southern Israel and the Sinai Peninsula. *Erdwiss Lomm. Osterr. Akad. Wiss*. 2: 79–86.
Druckman, Y. 1977. Differential subsidence during the deposition of the lower Jurassic Ardon Formation in western Jordan, southern Israel and northern Sinai. *Israel J. Earth Sci*. 26: 45–54.
Druckman, Y. 1984. Evidence for early–middle Triassic faulting and possible rifting from the Helez Deep borehole in the Coastal Plain of Israel. In J.E. Dixon & A.H.F. Robertson (eds), *The geological evolution of the eastern Mediterranean*: 203–212. London: Geol. Soc.
Druckman, Y. & Kashai, E. 1981. The Helez Deep and Devorah 2A boreholes and their implication to oil prospects in Pre-Jurassic strata in Israel. Jerusalem: Geol. Surv. Israel Rep. OD/1/81, 23 pp.
Druckman, Y. Hirsch, F. & Weissbrod, T. 1982. The Triassic of the southern margin of the Tethys in the Levant and its correlation across the Jordan Rift Valley. *Geol. Rundschau* 71: 919–936.
Dubertret, L. 1929. Étude des regions volcaniques du Hauran, du Djebel Druze et du Diret et Touloul. *Rév. Géog. Phys. Géol. Dyn.* II: 275.
Dubertret, L. 1932. Les formes structurales de la Syrie et de la Palestine; leur origine. *C.R. Acad. Sci. Paris* 195: 66–68.
Dubertret, L. 1933. Les grandes nappes basaltiques syriénnes; age et relations avec la tectonique. *C.R. Soc. Géol. France* 12: 178–180.
Dubertret, L. 1937. Le Crètacé 1. Le Massif Alaouite ou Djebel Ansariya. *Notes et Mém. Moyen Orient* II: 9–42.
Dubertret, L. 1947. Problémes de la géologie du Levant. *Bull. Soc. Géol. France* 5(17): 3–31.
Dubertret, L. 1952. Carte géologique au 1 : 50,000, feuille de Mardjayoun. Beirut: Min. des Travaux Publiques.
Dubertret, L. 1954. Basaltes et roches vertes du Liban, de Syrie et du Hatay. Algier: C.R. Cong. Geol. Intl. 19: 29–36.
Dubertret, L. 1955a. The ossiferous seams of Khirbet el Umbachi and of Hebariye (Safa). *Ann. Archaeol. Syria* IV, V: 59–76.

Dubertret, L. 1955b. Carte géologique du Liban 1 : 200,000 (with explanatory notes). Beirut: Ministry of Public Works.

Dubertret, L. 1962. Carte géologique du Liban, Syrie et bordure des pays voisins, au 1 : 1,000,000. Paris: Museum d'Histoire Naturelle.

Dubertret, L. 1963. Liban et Syrie. *Lexique Stratigraphique International*, *Vol. 10 c1*: 7–154. Paris: CNRS.

Dubertret, L. 1966. Liban, Syrie et bordure des pays voisins. *Notes et Mém. Moyen Orient* VIII: 250–358.

Dubertret, L. 1970. Review of structural geology of the Red Sea and surrounding areas. *R. Soc. London, Philos. Trans., Ser. A* 267: 9–20.

Dullo, W.C., Hötzl, H. & Jado, A.R. 1983. New stratigraphical results from the Tertiary sequence of the Midyan area, NW Saudi Arabia. *Newsl. Strat.* 12: 75–83.

Eckstein, Y. 1975. The measurements and interpretation of terrestrial heat flow in Israel. PhD dissertation, Dept. Geol., Hebrew Univ. Jerusalem, 171 pp. (in Hebrew, English summary, unpubl.).

Eckstein, Y. 1979. Review of heat flow data from the eastern Mediterranean region. *Pure Appl. Geophys.* 117: 150–179.

Eckstein, Y. & Simmons, G. 1978. Measurements and interpretation of terrestrial heat flow in Israel: *Geothermics* 6: 117–142.

Ehlers, J. 1996. *Quaternary and glacial geology*: 578 pp. Chichester: Wiley.

Ehrenberg, C.G. 1849. Über mikroskopische untersuchungen des Jordan wassers und bodens des Toten Meeres. *Monatsber. K. Preuss. Akad. Wiss. Berlin* 33: 187–193.

Ehrlich, A. 1973. Quaternary diatoms of the Hula basin (northern Israel). *Bull. Geol. Surv. Israel* 58: 1–40.

Eicher, D.H. 1968. *Geologic time*: 150 pp. Englewood Cliffs, NJ: Prentice-Hall.

Eidelman, A. 1979. The geology of the Arava rift margin in the En Yahav region. Jerusalem: Geol. Surv. Israel Rep. MM/10/79, 62 pp. (in Hebrew, English Abst.).

Eig, A. 1931–1932. Les éléments et les groupes phytogéographiques auxiliaires dans la flore palestienne. *Spec. Nov. Reg. Veget. Beih*. 63(1): 1–201; (2): 1–120.

Elazari-Volcani, B. 1936. Life in the Dead Sea. *Nature* 138: 467.

El-Isa, Z.H., Mechie, J. & Prodehl, C. 1987a. Shear velocity structure of Jordan from explosion seismic data. *R. Astronom. Soc. Geophys. J.* 90: 265–281.

El-Isa, Z., Mechie, J., Prodehl, C., Makris, J. & Rihm, R. 1987b. A crustal structure study of Jordan derived from seismic refraction data. *Tectonophysics* 138: 235–253.

El Moslimany, A.P. 1990. Ecological significance of common non-arboreal pollen: examples from dry lands of the Middle East. *Rev. Palaeobotan. Palynol*. 64: 343–350.

Elron, E. 1980. The geology of the Nahal Zin area (northern Hazeva sheet). Jerusalem: Geol. Surv. Israel Rep. MM/6/79, 54 pp. (in Hebrew, English Abst.).

Enmar, L. 1999. The travertines in the northern and central Arava: stratigraphy, petrography and geochemistry. Jerusalem: Geol. Surv. Israel Rep. GSI/1/99, 111 pp. (in Hebrew, English Abst.).

Enmar, L. & Heimann, A. 1999. Travertines of the northern and central Arava: paleogeographic, paleohydrologic and tectonic significance. Dead Sea: Israel Geol. Soc. Ann. Mtg. Excursion guidebook, 1–20 (in Hebrew).

Enmar, L., Heimann, A., Starinsky, A. & Katz, A. 1998a. The Arava travertines: lithology and stratigraphy. *Geol. Surv. Israel, Curr. Res.* 11: 17–22.

Enmar, L., Katz, A., Starinsky, A. & Heimann, A. 1998b. The geochemistry of the travertines and the paleohydrology of springs in the Arava Valley, southern Israel. Mitzpe Ramon: Abst., Israel Geol. Soc. Ann. Mtg., 24.

Enzel, Y. 1997. Estimating the ages of fault scarps along the Dead Sea Transform. *Terra Nostra* 4/97 (no page numbers).

Eran, G. 1982. The geometry of monoclines in the Negev. MSc thesis, Dept. Geol., Hebrew Univ. Jerusalem, 89 pp. (in Hebrew, English summary, unpubl.).

Eshet, Y. 1983. A palynostratigraphic study of the Permo-Triassic succession in the Makhtesh Qatan-2 well, Negev, Israel. *Geol. Surv. Israel, Curr. Res.* 1982: 15–20.

Eyal, A. 1984. The geology of the northern Arava Valley and its western margins in the En Yahav–Hazeva region., Jerusalem: Geol. Surv. Israel Rep. GSI/10/91, 63 pp. (in Hebrew, English Abst.).

Eyal, Y. & Reches, Z. 1983. Tectonic analysis of the Dead Sea Rift region since the late Cretaceous based on mesostructures. *Tectonics* 2: 167–185.

Eyal, M., Bartov, Y., Shimron, A.E. & Bentor, Y.K. 1980. Sinai – geological map 1 : 500,000. Tel Aviv: Surv. Israel.

Eyal, M., Eyal, Y., Bartov, Y. & Steinitz, G. 1981. The tectonic development of the western margin of the Gulf of Elat (Aqaba) rift. *Tectonophysics* 80: 39–66.

Eyal, Y., Eyal, M. & Kröner, A. 1991. Geochronology of the Elat terrain, metamorphic basement, and its implication for crustal evolution of the NE part of the Arabian–Nubian shield. *Israel J. Earth Sci*. 40: 5–16.

Fantozzi, P.L. & Sgavetti, M. 1998. Tectonic and sedimentary evolution of the eastern Gulf of Aden continental margins: new structural and stratigraphic data from Somalia and Yemen. In B.H. Purser & D.W.J. Bosence (eds), *Sedimentation and tectonics in rift basins: Red Sea–Gulf of Aden*: 56–76. London: Chapman & Hall.

Fauquette, S., Guiot, J. & Suc, J.-P. 1998. A method for climatic reconstruction of the Mediterranean Pliocene using pollen data. *Palaeogeogr. Palaeoclimatol. Palaeoecol*. 144: 183–201.

Feinstein, S. 1987. Constraints on the thermal history of the Dead-Sea Graben as revealed by coal ranks in deep boreholes. *Tectonophysics* 141: 135–150.

Feldman, L., Reich, B. & Ben-Sasson, C. 1997. Seismological bulletin of Israel 1900–1995. Holon (Israel): Geophys. Inst. Israel, diskette.

Field, H. 1953. Jebel Anazeh (Enaze or Anaiza) in Jordan, Iraq and Saudi Arabia. *Antiquity* 26: 212–213.

Field, H. 1956. *Bibliography of southwestern Asia*: 230 pp. Miami: Univ. Miami Press.

Fleischer, E. 1968. The subsurface geology of the Hula Valley and the Korazim area. Jerusalem: Geol. Surv. Israel Rep., 15 pp. (in Hebrew, unpubl.).

Flexer, A. 1968. Stratigraphy and facies development of Mount Scopus Group (Senonian–Paleocene) in Israel and adjacent countries. *Israel J. Earth Sci.* 17: 85–114.

Flexer, A. 1971. Late cretaceous paleogeography of northern Israel and its significance for the Levant geology. *Palaeogeog. Palaeoclimatol. Palaeocol.* 10: 293–316.

Flexer, A. & Honigstein, A. 1984. The Senonian succession in Israel – lithostratigraphy, biostratigraphy and sea level changes. *Cretaceous Res.* 5: 303–312.

Flexer, A., Freund, R., Reiss, Z. & Buchbinder, B. 1970. Santonian paleostructure of the Galilee. *Israel J. Earth Sci.* 19: 141–146.

Flexer, A., Dimant, A., Polishook, B. & Livnat, A. 1984. Relation of joints in the Avedat Group (Eocene) to the tectonic pattern of Israel. *Israel J. Earth Sci*. 33: 12–25.

Flexer, A., Gilat, A., Hirsch, F., Honigstein, A., Rosenfeld, A. & Rueffer, T. 1989. Late Cretaceous evolution of the Judean Mountains as indicated by ostracodes. *Terra Nova* 1: 349–358.

Flexer, A., Yellin-Dror, A., Kronfeld, J., Rosenthal, E., Ben-Avraham, Z., Artsztein, P.P. & Davidson, L. 2000. A Neogene salt body as the primary source of salinity in Lake Kinneret. *Arch. Hydrobiol*. 55: 1–17.

Fohs, F.J. 1927. Geology and the petroleum and natural gas possibilities of Palestine and Sinaitic peninsula. *AAPG Bull.* 11(2): 135–149.
Folkman, Y. 1976. Magnetic and gravity investigations of the crustal structure in Israel. PhD dissertation, Dept. Geophys. Planet. Sci., Tel Aviv University, 203 pp. (in Hebrew, English summary, unpubl.).
Folkman, Y. 1981. Structural features in the Dead-Sea–Jordan rift zone, interpreted from a combined magnetic-gravity study anomaly. *Tectonophysics* 80: 135–146.
Folkman, Y. & Yuval, Z. 1976. Aeromagnetic map of Israel, 1 : 250,000, 3 sheets. Holon (Israel): Ins. Petrol. Res. Geophys.
Forbes, J. 1861/1862. On the climate of Palestine in modern compared with ancient times. *Edinburgh New Philos. J.*: 169–179.
Forbes, R.J. 1964. *Studies in ancient technology*, Vol. 1: 199 pp. Leiden: Brill.
Fraas, O.F. 1867. *Das Tote Meer*: 22 pp. Stuttgart: Steinkopf.
Fraas, O.F. 1880. The sulphur of the valley of the Jordan. *Palestine Explor. Fund Quart. St.*: 246–248.
Freund, R. 1961. Distribution of lower Turonian ammonites in Israel and the neighboring countries. *Bull. Res. Coun. Israel* 10G: 79–100.
Freund, R. 1965. A model of the structural development of Israel and adjacent areas since Upper Cretaceous times. *Geol. Mag*. 102: 189–205.
Freund, R. 1970. Plate tectonics of the Red Sea and East Africa. *Nature* 228: 453.
Freund, R. & Zak, I. 1973. The migration of folding and faulting along geological structures. *Teva Ve'Aretz* 151: 264–270 (in Hebrew).
Freund, R., Oppenheim, M. & Schulman, N. 1965. Direction of magnetization of some basalts in the Jordan Valley and lower Galilee. *Israel J. Earth Sci.* 14: 37–44.
Freund, R., Zak, I. & Garfunkel, Z. 1968. Age and rate of the sinistral movement along the Dead Sea Rift. *Nature* 220: 253–255.
Freund, R., Garfunkel, Z., Zak, I., Goldberg, M., Weissbrod, T. & Derin, B. 1970. The shear along the Dead Sea Rift. *R. Soc. London Phil. Trans*. A267: 107–130.
Freund, R., Goldberg, M., Weissbrod, T., Druckman, Y. & Derin, B. 1975. The Triassic–Jurassic structure of Israel and its relation to the origin of the Eastern Mediterranean. *Bull. Geol. Surv. Israel* 65: 1–26.
Frieslander, U. & Ben-Avraham, Z. 1989. Magnetic field over the Dead Sea and vicinity. *Mar. Petrol. Geol.* 6: 148–160.
Frieslander, U., Bartov, Y., Garfunkel, Z., Rotstein, Y. & Rybakov, M. 1994. The structure of the Arava – Initial results from geophysical data. Nof Ginosar: Abst., Israel Geol. Soc. Ann. Mtg., 26.
Frieslander, U., Bartov, Y. & Garfunkel, Z. 1997. The structure of the Arava – new results from geological and geophysical studies. *Terra Nostra* 4/97 (no page numbers).
Frumkin, A., Ford, D.C. & Schwarcz, H. 1997. Paleoclimate of Jerusalem in the last glacial cycle from stable isotopes of speleothems. Kefar Gil'adi: Abst., Israel Geol. Soc. Ann. Mtg., 24.
Fuji, N. & Horowitz, A. 1989. Brunhes Epoch paleoclimates of Japan and Israel. *Palaeogeogr. Palaeoclimatol. Palaeoecol*. 72: 79–88.
Gaillardot, C. 1849. Études géologiques et topographiques sur la Syrie. *Annal. Soc. Emulation Vosges*: 6.
Galanis, S.P., Sass, J., Munroe, R.J. & Abu-Ajamieh, M. 1986. Heat flow at Zerqa Ma'in and Zar and a geothermal reconnaissance of Jordan. U.S. Geol. Surv. Open File Rep. 86–631; 1–110.
Gardner, E.W. & Bate, D.M.A. 1937. The bone-bearing beds of Bethlehem: their fauna and industry. *Nature* 140: 431–433.

Gardner, L.S. 1956. *Oil and gas possibilities in the Hashemite Kingdom of Jordan*: 116 pp. Washington: U.S. Geol. Surv.

Gardosh, M. 1987. Stratigraphy and tectonic activity of the late Quaternary in the Dead Sea basin. MSc thesis, Dept. Geol., Hebrew Univ. Jerusalem, 83 pp. (in Hebrew, English Abst., unpubl.).

Gardosh, M., Reches, Z. & Garfunkel, Z. 1990. Holocene tectonic deformation along the western margins of the Dead Sea. *Tectonophysics* 180: 123–137.

Gardosh, M., Kashai, E., Salhov, S., Shulman, H. & Tannenbaum, E. 1997. Hydrocarbon exploration in the southern Dead Sea area. In T.M. Niemi, Z. Ben-Avraham & J.R. Gat (eds), *The Dead Sea, the lake and its setting*: 57–72. Oxford: Oxford Univ. Press.

Garfunkel, Z. 1964. Tectonic problems along the Ramon lineament. MSc thesis, Dept. Geol., Hebrew Univ. Jerusalem, 68 pp. (in Hebrew, unpubl.).

Garfunkel, Z. 1970. Tectonics of the western margins of the southern Arava. PhD dissertation, Dept. Geol., Hebrew Univ. Jerusalem, 204 pp. (in Hebrew, English summary, unpubl.).

Garfunkel, Z. 1980. Contribution to the geology of the Precambrian of the Elat area. *Israel J. Earth Sci.* 29: 25–40.

Garfunkel, Z. 1981. Internal structure of the Dead Sea leaky transform (rift) in relation to plate kinematics. *Tectonophysics* 80: 81–108.

Garfunkel, Z. 1988. The pre-Quaternary geology of Israel. In Y. Yom-Tov & E. Tchernov (eds), *The zoogeography of Israel*: 7–34. Dordrecht: Junk.

Garfunkel, Z. 1989. Tectonic setting of Phanerozoic magmatism in Israel. *Israel J. Earth Sci.* 38(3–4): 51–74.

Garfunkel, Z. 1993. The Arod Pass area: Structural–stratigraphic relations and their implications for the history of the Ramon lineament. *Israel J. Earth Sci.* 42: 165–176.

Garfunkel, Z. 1997. The history and formation of the Dead Sea basin. In T.M. Niemi, Z. Ben-Avraham & J.R. Gat (eds), *The Dead Sea, the lake and its setting*: 36–56. Oxford: Oxford Univ. Press.

Garfunkel, Z. 1998. Constraints on the origin and history of the Eastern Mediterranean basin. *Tectonophysics* 298: 5–35.

Garfunkel, Z. & Bartov, Y. 1977. The tectonics of the Suez Rift: *Bull. Geol. Surv. Israel* 71: 1–44.

Garfunkel, Z. & Ben-Avraham, Z. 1996. The structure of the Dead Sea basin. *Tectonophysics* 266: 155–176.

Garfunkel, Z. & Ben-Avraham, Z. 2000. The architecture and history of the Dead Sea Transform. Dead Sea: Abst., The first Stephan Mueller Conf. Eur. Geophys. Soc., 27.

Garfunkel, Z. & Derin, B. 1984. Permian–early Mesozoic tectonism and continental margin formation in Israel and its implications for the history of the Eastern Mediterranean. In J.E. Dixon & A.H.F. Robertson (eds), *The geological evolution of the eastern Mediterranean*: 187–201. London: Geol. Soc.

Garfunkel, Z. & Horowitz, A. 1966. The upper Tertiary and Quaternary morphology of the Negev, Israel. *Israel J. Earth Sci.* 15: 101–117.

Garfunkel, Z., Bartov, J., Eyal, Y. & Steinitz, G. 1974. Raham Conglomerate – new evidence for Neogene tectonism in the southern part of the Dead Sea Rift. *Geol. Mag.* 111(1): 55–64.

Garfunkel, Z., Arad, A. & Almagor, G. 1979. The Palmahim disturbance and its regional setting. *Bull. Geol. Surv. Israel* 72: 1–56.

Garfunkel, Z., Zak, I. & Freund, R. 1981. Active faulting in the Dead Sea rift. *Tectonophysics* 80: 1–26.

Garrod, D.A.E. & Bate, D.M.A. 1937. *The stone age of Mount Carmel*: 240 pp. Oxford: Oxford Univ. Press.

Gass, I.G. & Gibson, I.L. 1969. Structural evolution of the rift zones in the Middle East. *Nature* 221: 926–930.

Gat, J.R. & Dansgaard, W. 1972. Stable isotope survey of freshwater occurrences in Israel and the Jordan Rift Valley. *J. Hydrol.* 16: 177–211.

Gat, J.R. & Galai, A. 1982. Groundwaters of the Arava Valley: an isotopic study of their origin and interrelationship. *Israel J. Earth Sci.* 31: 25–38.

Gat, J.R. & Magaritz, M. 1980. Climatic variations in the eastern Mediterranean Sea area. *Naturwissenschaft* 67: 80–87.

Gay-Lussac, J.L. 1819. Analyse de l'eau de la Mer Morte. *Ann. Chim. Phys.* 11: 195–197.

Gebel, H.G.K., Kafafi, Z. & Rollefson, G.O. (eds) 1997. *The prehistory of Jordan, II. Perspectives from 1997*: 661 pp. Berlin: ex Oriente.

Gelberman, E. & Kemmis, J.L. 1987. Triassic to Late Cretaceous tectonic alignments in the southern Coastal Plain and the northwestern Negev. Mitzpe Ramon: Abst., Israel Geol. Soc. Ann. Mtg., 39–40 (in Hebrew).

Geraads, D. & Tchernov, E. 1983. Fémurs humains du Pléistocène moyen de Gesher Benot Ya'acov (Israël). *L'Anthropologie* 87(1): 138–141.

Gerry, E. 1965. Report on Gideon 1 and Gideon 2 wells. Tel Aviv: Israel Oil Co. Micropaleontol. Lab. Rep. OS/3/1965, 5 pp.

Gerry, E. & Derin, B. 1983. Micropaleontological report of Zemah 1. Tel Aviv: Israel Institute of Petroleum and Energy, Rep. 5/83, 3 pp. (unpubl.).

GIF (German–Israeli Foundation) 1997. Proceedings of the 13th GIF Symp. on "The Dead Sea Rift as a unique global site". *Terra Nostra* 4/97 (no page numbers).

Gignoux, M. 1955. *Stratigraphic geology*: 682 pp. San Francisco: Freeman.

Gilat, A. 1980. Geology of the southern Hebron Mountains. *Geol. Surv. Israel, Curr. Res.* 1980: 33–35.

Gilat, A. 1992. Oversized U-shaped canyon development at the edge of rotating blocks due to wrench faulting on margins of the Dead Sea graben. *Terra Nova* 3: 638–647.

Gilat, A. 1994. Tectonics and associated mineralization activity, southern Judea, Israel. Jerusalem: Geol. Surv. Israel Rep. GSI/19/94, 280 pp. (in Hebrew, English Abst.).

Gilat, A. & Arkin, Y. 1978. Geotechnical map of the Eshtamoa Sheet, 1 : 50,000. Geological data for regional planning and the preservation of the environment. Jerusalem: Geol. Surv. Israel Rep. MM/6/78, 17 pp.

Gilat, A., Shirav, M., Bogoch, R., Halicz, L., Avner, U. & Nahlieli, D. 1993. Significance of gold exploitation in the early Islamic period, Israel. *J. Archaeol. Sci.* 20: 429–437.

Gilead, D. 1970. Handaxe industries in Israel and the Near East. *World Archaeol.* 2(1): 1–11.

Ginat, H. 1997. Paleogeography and the landscape evolution of the Nahal Hiyyon and Nahal Zihor basins. Jerusalem: Geol. Surv. Israel Rep. GSI/19/97, 206 pp. (in Hebrew, English Abst.).

Ginat, H. & Saragusti, I. 1996. Lower Paleolithic finds from the region of Nahal Zihor, southern Negev. Elat: Abst., Israel Geol. Soc. Ann. Mtg., 30.

Ginat, H. & Zilberman, E. 1996. Fluvio-lacustrine sediments in Nahal Hiyyon – evidence for a Pleistocene lake near the Arava Valley water divide. *Geol. Surv. Israel, Curr. Res.* 10: 62–66.

Ginat, H., Zilberman, E. & Rosenfeld, A. 1996. New finding of early Pleistocene lake deposits in Nahal Zihor, southern Negev. *Geol. Surv. Israel, Curr. Res.* 10: 55–61.

Ginzbourg, D. 1963. Petrography of the loess in the Beer Sheva basin. MSc thesis, Dept. Geol., Hebrew Univ. Jerusalem, 126 pp. (in Hebrew, unpubl.).

Ginzbourg, D. 1982. Geological maps of the Holy Land - Medeba map to satellite images. Budapest: Abst., 10th INHIGEO Symp., 147–157.
Ginzburg, A. & Ben-Avraham, Z. 1986. Structure of the Sea of Galilee graben, Israel, from magnetic measurements. *Tectonophysics* 126: 153–164.
Ginzburg, A. & Ben-Avraham, Z. 1987. The deep structure of the central and southern Levant continental margin. *Annales Tectonicæ* 1: 105–115.
Ginzburg, A. & Ben-Avraham, Z. 1997. The seismic refraction study of the north basin of the Dead Sea, Israel: *Geophys. Res. Lett.* 24: 2063–2066.
Ginzburg, A. & Folkman, Y. 1980. The crustal structure between the Dead Sea Rift and the Mediterranean Sea. *Earth Planet. Sci. Lett.* 51: 181–188.
Ginzburg, A. & Folkman, Y. 1981. Geophysical investigation of crystalline basement between Dead Sea Rift and the Mediterranean Sea. *AAPG Bull*. 65(3): 490–500.
Ginzburg, A., Cohen, S.S., Hay-Roe, H. & Rosenzweig, A. 1975. Geology of the Mediterranean shelf of Israel. *AAPG Bull*. 59: 2142–2160.
Ginzburg, A., Makris, J., Fuchs, K., Prodehl, C., Kaminski, W. & Amitai, U. 1979a. A seismic study of the crust and upper mantle of the Jordan–Dead Sea rift and their transition toward the Mediterranean Sea. *J. Geophys. Res*. 84: 1569–1582.
Ginzburg, A., Makris, J., Fuchs, K., Perathoner, B. & Prodehl, C. 1979b. Detailed structure of the crust and upper mantle along the Jordan–Dead Sea rift. *J. Geophys. Res.* 84: 5605–5612.
Gisis, I. & Bar-Yosef, O. 1974. New excavations in Zuttiyeh Cave, wadi Amud, Israel. *Paléorient* 2: 175–180.
Giveon, M. 1984. Paleogeographic studies: Hordos Formation and En Gev Sands in eastern Lake Kinneret. MSc thesis, Dept. Geol., Hebrew Univ. Jerusalem, 86 pp. (in Hebrew, unpubl.).
Glennie, K.W. 1970. *Desert sedimentary environments*: 222 pp. Amsterdam: Elsevier.
Glennie, K.W., Boef, M.G.A., Hughes-Clarke, M.W., Hoody-Stuart, M., Polaar, W.F.H. & Reinhardt, B.M. 1974. Geology of the Oman Mountains, 1. *Verh. K. Ned. Geol. Mijnbouwk Genoot* 31: 239–241.
Glikson, Y. 1964. The geology of the Nahal Dishon region. MSc thesis, Dept. Geol., Hebrew Univ. Jerusalem, 52 pp. (in Hebrew, unpubl.).
Glikson, Y. 1966. The lacustrine Neogene in the Kefar Gil'adi area, northern Jordan Valley. *Israel J. Earth Sci.* 15(3): 85–100.
Glueck, N. 1940. *The other side of the Jordan*: 207 pp. New Haven: Am. School Oriental Res.
Gmelin, G.G. 1827. Resultate der analyse des wassers von Totem Meer. *Poggendrof's Annal. Phys. Chem.* 85: 9–10.
Golani, U. 1962. The geology of the Lake Tiberias region and the geohydrology of the saline springs. Tel Aviv: TAHAL (Water Planning for Israel), Geotech. Dept. Rep. 19, 54 pp.
Goldberg, M. & Friedman, G.M. 1974. Paleoenvironments and paleogeographic evolution of the Jurassic system in southern Israel. *Bull. Geol. Surv. Israel* 61: 1–44.
Goldberg, P. 1986. Late Quaternary environmental history of the southern Levant. *Geoarchaeology* 1(3): 225–244.
Goldbery, R. 1980. Use of grain-size frequency data to interpret the depositional environment of the Pliocene Pleshet Formation, Be'er Sheva, Israel. *J. Sediment. Petrol.* 50: 843–856.
Goldreich, Y. 1993. The climate of Israel. *Encyclopaedia Hebraica* (2nd edn) 6(1): 203–222. Jerusalem: Encyclopaedia Publ. Co. (in Hebrew).
Goldsmith, N.F., Hirsch, F., Friedman, G.M., Tchernov, E., Derin, B., Gerry, E., Horowitz, A. & Weinberger, G. 1988. Rotem mammals and Yeroham crassostreids: stratigraphy of

Gvirtzman, H. 1997. The saline-hot springs in the vicinity of the Sea of Galilee: hydrogeological modeling. *Terra Nostra* 4/97 (no page numbers).
Haas, G. 1966. *On the vertebrate fauna of the lower Pleistocene site Ubeidiya*: 68 pp. Jerusalem: Israel Acad. Sci. Hum.
Hall, J.K. 1996. Digital topography and bathymetry of the area of the Dead Sea depression. *Tectonophysics* 266: 177–185.
Hall, J.K. 1997. Topography and bathymetry of the Dead Sea depression. In T.M. Niemi, Z. Ben-Avraham & J.R. Gat (eds), *The Dead Sea, the lake and its setting*: 11–21. Oxford: Oxford Univ. Press.
Hall, J.K. 2000. Orthographic satellite imagery coverage, Israel and adjacent areas. Sheet 16: Northern Levant satellite map. Jerusalem: Geol. Surv. Israel Rep. GSI/19/2000.
Hall, J.K. & Ben-Avraham, Z. 1985. The Dead Sea shallow seismic survey of 5–9 June, 1982. *Geol. Surv. Israel, Curr. Res.* 1985: 79–81.
Hall, J.K. & Neev, D. 1975. The Dead Sea geophysical survey, 19 August 1974 – Final Report No. 1: Jerusalem: Geol. Surv. Israel, 27 pp.
Hallam, A. 1981. A revised sea level curve for the Early Jurassic. *J. Geol. Soc.* 138: 735–743.
Haq, B.U., Hardenbol, J. & Vail, P.R. 1988. Mesozoic and Cenozoic chronostratigraphy and cycles of sea level change. *Soc. Econ. Paleontol. Mineral. Sp. Publ.* 42: 71–108.
Harash, A. 1967. The geology of Yeroham–Dimona plain. MSc thesis, Dept. Geol., Hebrew Univ. Jerusalem, 55 pp. (in Hebrew, unpubl.).
Harash, A. & Bar, Y. 1988. Faults, landslides and seismic hazards along the Jordan River gorge, northern Israel. *Eng. Geol.* 25: 1–15.
Harding, T.P. 1984. Graben hydrocarbon occurrences and structural style. *AAPG Bull*. 68: 333–362.
Hatcher, R.D., Zietz, I., Regan, R.D. & Abu-Ajamieh, M. 1981. Sinistral strike-slip motion on the Dead Sea rift: Confirmation from new magnetic data. *Geology* 9: 458–462.
Heimann, A. 1985. The geology of the northern Hula Valley, with emphasis on ancient spring deposits (travertines). MSc thesis, Dept. Geol., Hebrew Univ. Jerusalem, 157 pp. (in Hebrew, unpubl.).
Heimann, A. 1990. The development of the Dead Sea Rift and its margins in northern Israel during the Pliocene and the Pleistocene. Jerusalem: Geol. Surv. Israel Rep. GSI/28/90, 83 pp. (in Hebrew, English summary).
Heimann, A. 2000. Stages in the development of the northern Dead Sea Transform. Dead Sea: Abst., The first Stephan Mueller Conf. Eur. Geophys. Soc., 32.
Heimann, A. & Ron, H. 1993. Geometric changes of plate boundaries along part of the northern Dead Sea Transform: geochronologic and paleomagnetic evidence. *Tectonics* 12(2): 477–491.
Heimann, A. & Sass, E. 1989. Travertines in the northern Hula Valley, Israel. *Sedimentology* 36: 95–108.
Heimann, A. & Steinitz, G. 1989. $^{40}Ar/^{39}Ar$ total gas ages of basalts from Notera #3 well, Hula Valley, Dead Sea Rift: stratigraphic and tectonic implications. *Israel J. Earth Sci.* 38: 173–184.
Heimann, A., Mor, D. & Steinitz, G. 1987. K–Ar dating of the Ramat Korazim basalts, Dead Sea Rift. Mitzpe Ramon: Abst., Israel Geol. Soc. Ann. Mtg., 54.
Heimann, A., Steinitz, G., Mor, D. & Shaliv, G. 1996. The Cover basalt formation, its age and its regional and tectonic setting: implications from K–Ar and $^{40}Ar/^{39}Ar$ geochronology. *Israel J. Earth Sci.* 45(2): 55–72.
Heller, J. & Sivan, N. *Melanopsis* from the mid-Pleistocene site of Gesher Benot Ya'aqov (in prep.).

Henry, D.O. 1986. The prehistory and paleoenvironments of Jordan: an overview. *Paléorient* 12(2): 5–26.

Henry, D.O. (ed.) 1998. *The prehistoric archaeology of Jordan*: 207 pp. BAR Intl. Ser. 705.

Henson, F.R.S. 1951. Observations on the geology and petroleum occurrences of the Middle East. The Hague: Proc. 3rd World Petrol. Cong., 119–140.

Hermbstädt, S.F. 1822. Chemische zergliederung des wassers aus dem Toten Meere. *Schweigger's J. Phys. Chem.* 34: 153–195.

Himus, G.W. 1954. *A dictionary of geology*: 153 pp. London: Penguin.

Hirsch, F. 1980. Jurassic bivalves and gastropods from northern Sinai and southern Israel. *Israel J. Earth Sci*. 28: 128–163.

Hirsch, F. 1984. The Arabian subplate during the Mesozoic. In J.E. Dixon & A.H.F. Robertson (eds), *The geological evolution of the eastern Mediterranean*: 217–224. London: Geol. Soc.

Hirsch, F. 1996. Geological map of Israel 1 : 50,000, sheet 19-IV: HaMakhtesh HaQatan. Jerusalem: Geol. Surv. Israel.

Hirsch, F. & Roded, R. *Lexique Stratigraphique, Israel* (2nd edn) (in prep.).

Hirsch, F. & Shahar, Y. 1993. Senonian tectonics and Neogene stratigraphy of the Zohar-Ef'e region. Arad: Israel Geol. Soc. Ann. Mtg. Excursion guidebook, 133–151 (in Hebrew).

Hitchcock, E. 1843. Notes on the geology of several parts of western Asia, founded chiefly on specimens and descriptions from American missionaries. *Trans. Am. Geol. Assoc.* 1840/1842: 348–421.

Holland, F.W. 1868. On the peninsula of Sinai. *J. R. Geog. Soc. London* 38: 237–257.

Honigstein, A., Flexer, A. & Rosenfeld, A. 1988. Tectonic activity in the Syrian Arc in Israel, as indicated by Senonian ostracodes. *N. Jb. Geol. Paläont. Mh*. 168: 173–183.

Hooijer, D.A. 1958. An early Pleistocene mammalian fauna from Bethlehem. *Bull. Brit. Mus. (Nat. Hist. Geol.)* 3(8): 267–292.

Hooijer, D.A. 1959. Fossil mammals from Jisr Banat Yaqub, south of Lake Huleh, Israel. *Bull. Res. Coun. Israel* 8G: 177–199.

Hooijer, D.A. 1960. A *Stegodon* from Israel. *Bull. Res. Coun. Israel* 9G(2–3): 104–108.

Hooker, J.B. 1862. On the cedars of Lebanon, Taurus, Algeria and India. *Natur. Hist. Rev.* 11.

Horowitz, A. 1966. Tropical and northern pollen and spores in Recent sediments from the Bay of Elat (Aqaba). *Israel J. Earth Sci*. 15: 125–130.

Horowitz, A. 1967. The geology of Museri Island (Dahlak Archipelago, southern Red Sea). *Israel J. Earth Sci*. 16: 74–83.

Horowitz, A. 1969. Recent pollen sedimentation in Lake Kinneret, Israel. *Pollen et Spores* 11: 353–384.

Horowitz, A. 1970. Palynological tracing of saline water sources in the Lake Kinneret region, Israel. *J. Hydrol.* 8: 22–29.

Horowitz, A. 1971. Climatic and vegetational developments in northeastern Israel during upper Pleistocene–Holocene times. *Pollen et Spores* 13(3): 255–278.

Horowitz, A. 1973. Development of the Hula Basin, Israel. *Israel J. Earth Sci*. 22: 107–139.

Horowitz, A. 1974. *The late Cenozoic stratigraphy and paleogeography of Israel*: 184 pp. Inst. Archaeol., Tel Aviv Univ.

Horowitz, A. 1975a. Cenozoic erosive surfaces of central Sinai. *Israel J. Earth Sci*. 24: 53–55.

Horowitz, A. 1975b. Late Pleistocene clastic dykes and rates of erosion in the Galilee, Israel. *Tel Aviv* 2(3): 93–96.
Horowitz, A. 1976. Late Quaternary paleoenvironments of prehistoric settlements. In A.E. Marks (ed.), *Prehistory and paleoenvironments in the central Negev, Israel*: Vol. 1: 57–68. Dallas: SMU Press.
Horowitz, A. 1977a. Climatic changes: atmospheric reaction to astronomic and geotectonic processes. In S. Horie (ed.), *Paleolimnology of Lake Biwa and the Japanese Pleistocene*: Vol. 5: 302–312. Kyoto Univ. Press.
Horowitz, A. 1977b. The Quaternary stratigraphy and paleogeography of Israel. *Paléorient* 3: 47–100.
Horowitz, A. 1979. *The Quaternary of Israel*: 394 pp. New York: Academic Press.
Horowitz, A. 1982. The Olduvai Event and the beginning of the Quaternary. Moscow: Abst., XI Intl. Quaternary Assn., 105 pp.
Horowitz, A. 1987a. Subsurface palynostratigraphy and paleoclimates of the Quaternary Jordan Rift Valley fill, Israel. *Israel J. Earth Sci*. 36: 31–44.
Horowitz, A. 1987b. Travertines of the arid regions, oxygen isotope stages, and late Quaternary climates of Israel. *Quat. Res.* 27: 103–105.
Horowitz, A. 1987c. Palynological evidence for the age and rate of sedimentation along the Dead Sea Rift, and structural implications. *Tectonophysics* 141: 107–115.
Horowitz, A. 1988. Synthése de la stratigraphie et des paléoclimats quaternaires du rift de la Mer Morte, Israël. *L'Anthropologie* 92(3): 753–767.
Horowitz, A. 1989a. Prehistoric cultures of Israel: correlation with the oxygen isotope scale. In: O. Bar Yosef & B. Vandermeersch (eds), *Investigations in South Levantine prehistory*: 5–17. Oxford: BAR Intl. Ser., 497.
Horowitz, A. 1989b. Palynological evidence for the Quaternary rates of accumulation along the Dead Sea Rift, and structural implications. *Tectonophysics* 164: 63–71.
Horowitz, A. 1989c. Continuous pollen diagrams for the last 3.5 M.Y. from Israel: vegetation, climate and correlation with the oxygen isotope record. *Palaeogeogr. Palaeoclimatol. Palaeoecol*. 72: 63–78.
Horowitz, A. 1990. Neogene palynostratigraphy, paleoflora and paleoclimatic reconstruction in Israel. *Paléobiologie Continentale* 17: 245–257.
Horowitz, A. 1991. A geological definition for the Quaternary. Beijing: Abst., XIII Intl. Quaternary Assn., 138.
Horowitz, A. 1992a. *Palynology of arid lands*: 546 pp. Amsterdam: Elsevier.
Horowitz, A. 1992b. AP factors and late prehistoric settlement patterns of the Levant. Aix-en-Provence: Abst., 8th Intl. Palynol. Cong., 67.
Horowitz, A. 1995. Quaternary sedimentation, erosion and human sites location in Israel. Jerusalem: Abst., 2nd Intl. Symp. Geol. East. Medit. Region, 9.
Horowitz, A. 1996a. Palynostratigraphy of Sedom Deep 1 borehole and the late Cenozoic climate of Israel. Houston: Abst., IX Intl. Palynol. Cong., 67.
Horowitz, A. 1996b. Review of Lower Paleolithic site locations in Israel, possibly controlled by deposition and erosion processes. *Israel J. Earth Sci*. 45(3): 137–145.
Horowitz, A. 1997. Palynostratigraphy and subsidence of the Dead Sea Rift fill. *Terra Nostra* 4/97 (no page numbers).
Horowitz, A. 2000a. The middle Oligocene of the Dead Sea and vicinity. *Geol. Surv. Israel, Curr. Res.* 12: 197–205.
Horowitz, A. 2000b. Crescentic faults, an alternative to the transform theory of Dead Sea Rift tectonics? Dead Sea: Abst., The first Stephan Mueller Conf. Eur. Geophys. Soc., 35.
Horowitz, A. 2001. The lateral movement along the Dead Sea Rift: fact or legend? Isparta: Abst., 3rd Intl. Symp. Geol. East. Medit. Region, 78.

Horowitz, A. & Assaf, G. 1981. Configuration of interpluvial and pluvial climates in the Levant. *Beihefte Tübinger Atlas des Vorderen Orients* 8(A): 110–120.

Horowitz, A. & Derin, B. 1987. Palynological correlation of continental and marine Neogene sequences in Israel. *Ann. Inst. Geol. Publ. Hung*. LXX: 503–507.

Horowitz, A. & Gat, J.R. 1984. Floral and isotopic indications for possible summer rains in Israel during wetter climates. *Pollen et Spores* XXVI(1): 61–68.

Horowitz, A. & Horowitz, M. 1985. Subsurface late Cenozoic palynostratigraphy of the Hula Basin, Israel. *Pollen et Spores* 27(3–4): 365–390.

Horowitz, A. & Horowitz, M. 1990. Neogene–Recent paleogeography of the southern Dead Sea. Tel Aviv: Israel National Oil Co. Rep., 153 pp. (unpubl.).

Horowitz, A. & Horowitz, M. 1992. Pollen analyses from outcrops in Mount Sedom. Tel Aviv: Israel National Oil Co. Rep., 6 pp. (unpubl.).

Horowitz, A. & Weinstein-Evron, M. 1986. The late Pleistocene climate of Israel. *Bull. Assoc. Fr. Étude Quaternaire* 25–26: 84–90.

Horowitz, A. & Zak, I. 1968. Preliminary palynological analysis of an evaporitic sequence from Mount Sedom, Israel. *Rev. Palaeobot. Palynol*. 7: 25–30.

Horowitz, A., Siedner, G. & Bar-Yosef, O. 1973. Radiometric dating of the Ubeidiya Formation, Jordan Valley, Israel. *Nature* 242: 186–187.

Hours, F. 1975. The Lower Paleolithic of Lebanon and Syria. In E. Wendorf & A.E. Marks (eds), *Problems in prehistory: North Africa and the Levant*: 249–271. Dallas: SMU Press.

Hours, F. 1981. Le Paléolithique inférieur de la Syrie et du Liban. Le point de la question en 1980. In P. Sanlaville & J. Cauvin (eds), *Préhistoire du Levant*: 165–184. Lyon: Maison de l'Orient.

Hours, F., Copeland, L. & Aurenche, O. 1973. Les industries paléolithiques du Proche-Orient, essai de correlation. *L'Anthropologie* 77: 229–280; 437–496.

Hovers, E., Goren, Y., Rabinovich, R. & Rak, Y. 1994. The Neanderthal site of Amud Cave, Israel. Nof Ginosar: Abst., Israel Geol. Soc. Ann. Mtg., 43.

Hsü, K.J. 1972. Origin of saline giants: a critical review after the discovery of the Mediterranean evaporite. *Earth Sci. Rev*. 8: 371–396.

Huckriede, R. 1966. Das Quartär des arabischen Jordan-Tales und Beobachtungen über "Pebble Culture" und "Prä-Aurignac". *Eiszeitalter u. Gegenw.* 17: 211–212.

Huckriede, R. & Wiesemann, G. 1968. Der jungpleistozäne Pluvial See von el Jafr und weitere Daten zum Quartär Jordaniens. *Geol. Palaeont.* 2: 73–95.

Hudleston, W.H. 1885. *The geology of Palestine*: 76 pp. London: Stanford.

Hughes, G.W., Varol, O. & Beydoun, Z.R. 1991. Evidence for middle Oligocene rifting of the Gulf of Aden and for late Oligocene rifting of the southern Red Sea. *Mar. Petr. Geol.* 8: 354–358.

Hull, E. 1886. *The survey of Western Palestine, memoir on the geology and geography of Arabia Petraea, Palestine and adjoining districts, with special reference to the mode of formation of the Jordan–Arabah Depression and the Dead Sea*. London: Palestine Exploration Fund, 145 pp.

Hume, W.F. 1901. The rift valleys of eastern Sinai. *Geol. Mag.* 8: 198–200.

Hurwitz, S., Matmon, A., Ron, H. & Heimann, A. 1999. Deformation along the margins of the Dead Sea Transform: the Yehudiyya Block, Golan Heights. *Israel J. Earth Sci.* 48: 257–264.

Husseini, M.I. 1988. The Arabian Infracambrian extensional system. *Tectonophysics* 148: 93–103.

Ibrahim, K.M.K. 1991. The geology of the Wadi Rahma, map sheet No. 3049-IV. Amman: Natural Resources Authority, Geol. Bull. 15: 122 pp., map 1 : 50,000.

Ibrahim, K.M.K. 1993. The geology of the Wadi Gharandal area, map sheet No. 3050-II. Amman: Natural Resources Authority, Geol. Bull. 24: 77 pp., map 1 : 50,000.

Inbar, M., Bartov, Y. & Arad, V. 1989. Hula Valley and Korazim region, bibliography of geological research. Jerusalem: Geol. Surv. Israel Rep. GSI/40/89, 66 pp.

Ionides, M.G. 1951. The Jordan Valley. *J. R. Cent. Asian Soc. London* 38: 217–225.

Ionides, M.G. & Blake, G.S. 1940. *Report on the water resources of Transjordan and their development*: 372 pp. London: Crown Agents.

Irby, C.L. & Mangles, J. 1868. *Travels in Egypt and Nubia, Syria, and the Holy land; including a journey round the Dead Sea, and through the country east of the Jordan*: 150 pp. London: J. Murray.

Issar, A. & Kafri, U. 1972. Neogene and Pleistocene geology of the western Galilee coastal plain. *Bull. Geol. Surv. Israel* 53: 1–13.

Izzeldin, A.Y. 1987. Seismic, gravity and magnetic surveys in the central part of the Red Sea: their interpretation and implications for the structure and evolution of the Red Sea. *Tectonophysics* 143: 269–306.

Jackson, J.A. (ed.) 1997. *Glossary of geology*: 769 pp. Alexandria, VA: Am. Geol. Ins.

Jarrar, G.H. 1985. Late Proterozoic crustal evolution of the Arabo-Nubian Shield in the Wadi Araba area, southwestern Jordan. *Geol. Jahrb.* 61: 1–87.

Jarrige, J.J., Ott d'Estevou, P., Burollet, P.F., Thiriet, J.P., Icart, J.C., Ricbert, J.P., Sehans, P., Montenat, C. & Prat, P. 1986. Inherited discontinuities and Neogene structure: the Gulf of Suez and the northwestern edge of the Red Sea. *R. Soc. Lond., Philos. Trans.* A317(1539): 129–139.

Jelinek, A. 1982. The Tabun Cave and Paleolithic Man in the Levant. *Science* 216: 1369–1375.

Jennings, C.W. (compiler) 1973. State of California preliminary fault and geological map, 1 : 750,000. Sacramento: Calif. Div. Mines Geol.

Joffe, S. & Garfunkel, Z. 1987. The kinematics of the circum Red Sea – a re-evaluation. *Tectonophysics* 141: 5–22.

Joinville, J.S. de, ca. 1255 (Publ. 1868). *Histoire de Saint Louis*: XLIII+ 410 pp. Paris: Soc. de l'Histoire de France.

Jolly, D., Harrison, S.P., Damnati, B. & Bonnefille, R. 1998. Simulated climate and biomes of Africa during the late Quaternary: comparison with pollen and lake status data. *Quat. Sci. Rev.* 17(6–7): 629–657.

Kafri, U. 1997. Neogene to Quaternary drainage systems and their relationship to young tectonics: Lower Galilee, Israel. Jerusalem: Geol. Surv. Israel Rep. GSI/1/97, 50 pp.

Kafri, U. & Heimann, A. 1994. Reversal of the paleodrainage system in the Sea of Galilee area as an indicator of the formation timing of the Dead Sea Rift valley base level in northern Israel. *Palaeogeogr. Palaeoclimatol. Palaeoecol*. 109: 101–109.

Kafri, U. & Folkman, Y. 1981. Multiphase reverse vertical tectonic displacement across major faults in northern Israel. *Earth Planet. Sci. Lett*. 53: 343–348.

Kansou, M. 1961. Découverte de vertébrés pontiens au Liban, dans la plaine de la Békaa. *Bull. Scient. (Yugoslavia)* 6(3): 65.

Karcz, I. & Key, C.A. 1966. Note on the pre-Paleozoic morphology of the basement in the Timna area (southern Israel). *Israel J. Earth Sci.* 15: 47–56.

Karcz, I., Weiler, Y. & Key, C.A. 1971. Lithology and environment of deposition of the "Amudei Shelomo Sandstone" ("N.S.P.") in Nahal Shani, Elat. *Israel J. Earth Sci*. 20: 119–124.

Karge, P. 1917. *Repha'im. Die vorgeschichtliche Kultur Palästinas und Phöniciens:* 755 pp. Berlin: Paderborn.

Kashai, E.L. 1988. A review of the relations between the tectonics, sedimentation and petroleum occurrences of the Dead Sea–Jordan rift system. In W. Manspeizer (ed.), *Triassic–Jurassic rifting*: 883–909. Amsterdam: Elsevier.

Kashai, E.L. & Croker, P.F. 1987. Structural geometry and evolution of the Dead Sea–Jordan rift system as deduced from new subsurface data. *Tectonophysics* 141: 33–60.

Kashai, E.L. & Goldberg, Y. 1984. Notera 3, geological completion report. Tel Aviv: Oil Exploration (Investments) Ltd., Rep. 84/20, 76 pp.

Kaufman, A. 1971. U-series dating of Dead Sea basin carbonates. *Geochim. Cosmochim. Acta* 35(12): 1269–1281.

Kaufman, A., Carmi, I. & Schlessinger, R. 1973. Rehovot radiocarbon measurements II. *Radiocarbon* 15: 179–184.

Kazmin, V.G. 2000. The Syrian segment of the Levantine fault zone. Dead Sea: Abst., The first Stephan Mueller Conf. Eur. Geophys. Soc., 83.

Kempler, D. 1994. An outline of northeastern Mediterranean tectonics in view of cruise 5 of the Akademik Nikolaj Strakhov. In V.A. Krasheninnikov & J.K. Hall (eds), *Geological structure of the north-eastern Mediterranean*: 277–294. Jerusalem: Histor. Prod. Hall.

Kenyon, K.M. 1960. Excavations of Jericho, 1957–1958. *Palestine Expl. Quart.* 92: 1–21.

Khalil, B. 1992. The geology of the ar-Rabba area, map sheet No. 3152-IV. Amman: Natural Res. Auth., Geol. Bull. 22: 105 pp., map 1 : 50,000.

Khoury, H., Salameh, E. & Udluft, P. 1984. On the Zerka Ma'in (Therma Kallirrhoes) travertine, Dead Sea. *N. Jb. Geol. Paläont.* 8: 472–484.

Kirkbride, D.V.M. 1966. Five seasons at the Pre-Pottery Neolithic village of Beidha in Jordan. *Palestine Expl. Quart.* 98: 8–72.

Kislev, M.E., Nadel, D. & Carmi, I. 1992. Epipaleolithic (19,000 BP) cereal and fruit diet at Ohalo II, Sea of Galilee, Israel. *Rev. Palaeobot. Palynol.* 73: 161–166.

Klang, A. 1984. A new interpretation of the gravimetric anomaly of the Hula basin. Holon (Israel): Inst. Petrol. Res. Geophys. Rep. X1/776/81, 3 pp.

Klang, A. & Gvirtzman, G. 1983. Preliminary structural analysis of the Sharon area and its continental shelf. Tel Aviv: Oil Exploration (Investments) Ltd. Rep. 13/30, 47 pp. (unpubl.).

Klaus, W. 1965. *Pterocarya*-pollen im Quartär-Sälz aus Israel. *Pollen et Spores* 7: 529–531.

Klein, C. 1986. Fluctuations of the level of the Dead Sea and climatic fluctuations in Erez–Israel during historical times. PhD dissertation, Dept. Geog., Hebrew Univ. Jerusalem, 207 pp. (in Hebrew, English summary, unpubl.).

Kohn, B.P. & Eyal, M. 1981. History of uplift of the crystalline basement of Sinai and its relations to opening of the Red Sea as revealed by fission track dating of apatites. *Earth Planet. Sci. Lett.* 52: 129–141.

Kolodny, Y. 1980. Carbon isotopes and depositional environment of a high productivity sedimentary sequence – the case of the Mishash–Ghareb formations, Israel. *Israel J. Earth Sci*. 29: 147–156.

Kolodny, Y., Schulman, N. & Gross, S. 1973. Hazeva Formation sediments affected by the Mottled Zone Event. *Israel J. Earth Sci.* 22(3): 185–194.

Kolodny, Y., Taraboulos, A. & Frieslander, U. 1980. Participation of freshwater in chert diagenesis: evidence from oxygen isotopes and boron track mapping. *Sedimentology* 27: 305–316.

Koucky, F.L. & Smith, R.H. 1986. Lake Beisan and the prehistoric settlement of the northern Jordan Valley. *Paléorient* 12(2): 27–36.

Kovach, R.L., Andreasen, G.E., Gettings, M.E. & El-Kayse, K. 1990. Geophysical investigations in Jordan. *Tectonophysics* 180: 49–60.

Krasheninnikov, V.A. 1994. Stratigraphy of the Maastrichtian and Cenozoic deposits of the coastal part of north-western Syria (neoautochthon of the Bassit ophiolite massif). In V.A. Krasheninnikov & J.K. Hall (eds), *Geological structure of the north-eastern Mediterranean*: 265–276. Jerusalem: Histor. Prod. Hall.

Krenkel, E. 1924. Der Syrische Bogen. *Zentralbl. Miner*. 9: 274–281; 10: 310–313.

Kröner, A., Eyal, M. & Eyal, Y. 1990. Early Pan-African evolution of the basement around Elat, Israel and the Sinai Peninsula revealed by single-zircon evaporation dating and implication for crustal accretion rates. *Geology* 18: 545–548.

Kronfeld, J., Vogel, J.C., Rosenthal, E. & Weinstein-Evron, M. 1988. Age and paleoclimatic implications of the Bet She'an travertines. *Quat. Res.* 30: 298–303.

Krupp, F. 1987. Freshwater ichtyogeography of the Levant. *Beihefte Tübinger Atlas des Vorderen Orients* 28(A): 229–237.

Laborde, L. de, 1836. *Journey through Arabia Petraea to Mount Sinai, and the excavated city of Petra, the Edom of the prophecies*: 331 pp. London: Murray.

Lambert, C. 1928. Notes made during excavations. Jerusalem: Archives of the Israel Antiquities Authority, File 193.

Lamdan, M., Ziffer, D., Huster, Y. & Ronen, A. 1977. Prehistoric archaeology survey of Nahal Shiqma. Local Council Sha'ar HaNegev, 79 pp. (in Hebrew).

Lang, B., Shirav, M. & Bogoch, R. 1979. Volcanological aspects of the Har Peres composite volcano, Golan Plateau. *Israel J. Earth Sci.* 28: 27–32.

Langozky, Y. 1960. The petrography and geochemistry of the Lisan Formation. MSc thesis, Dept. Geol., Hebrew Univ. Jerusalem, 54 pp. (in Hebrew, unpubl.).

Lapierre, H. & Rocci, G. 1976. Le volcanisme alcalin du sud-ouest de Chypre et le probléme de l'ouverture des regions tethysiénnes au Trias. *Tectonophysics* 30: 299–313.

Lartet, L. 1865. Note sur la formation du bassin de la Mer Morte ou lac Asphaltite et sur les changements survenus dans le niveau de ce lac. *Bull. Soc. Géol. France (Sér. 2)* 22: 420–463.

Lartet, L. 1866. Sur les gîtes bitumineux de la Judée et de la Coelé-Syrie et sur le mode d'arrivée au milieu des eaux de la Mer Morte. *Bull. Soc. Géol. France (Sér. 2)* 24: 12–31.

Lartet, L. 1869. *Essai sur la géologie de la Palestine et des contrées avoisinantes, telles que l'Egypte et l'Arabie, comprenant les observations recueillies dans le cour de l'Expédition du Duc de Luynes à la Mer Morte*: 292 pp. Paris: Masson.

Lartet, L. 1877. *Exploration géologique de la Mer Morte, de la Palestine et de l'Idumée:* 326 pp. Paris: Bertrand.

Lazar, M., Ben-Avraham, Z. & Blumberg, D.G. 2000. Recent tectonic activity in the northern Dead Sea basin. Dead Sea: Abst., The first Stephan Mueller Conf. Eur. Geophys. Soc., 42.

Lees, G.M. 1931. Salt – some depositional and deformational problems. *J. Inst. Petrol. Tech.* 17(91): 259–280.

Lejay, P. 1938. Exploration gravimétrique des états du Levant sous mandat français: Comité national français de Géodésie et Géophysique: 1–54.

Le Pichon, X. & Angelier, J. 1979. The Hellenic arc and trench system: a key to the neotectonic evolution of the eastern Mediterranean area. *Tectonophysics* 60: 1–42.

Le Pichon, X. & Gaulier, J.M. 1988. The rotation of Arabia and the Levant fault system. *Tectonophysics* 153: 271–294.

Letronne, A.J. 1839. Sur la séparation primitive des bassins de la Mer Morte et de la Mer Rouge. *Nouv. Annal. Voyages Paris* 23: 257–317.

Levin, N. 1985. Palyno- and Litho-stratigraphy of the lower Pleistocene sequence (Palynozone QI) in the "Melekh Sedom-1" borehole. MSc thesis, Dept. Geol., Hebrew Univ. Jerusalem, 82 pp. (in Hebrew, English Abst., unpubl.).
Levin, N. & Horowitz, A. 1987. Palynostratigraphy of the early Pleistocene QI Palynozone. *Israel J. Earth Sci*. 36(1–2): 45–58.
Levitte, D. 1966. Magmatic phenomena in the Arava. MSc thesis, Dept. Geol., Hebrew Univ. Jerusalem, 50 pp. (in Hebrew, unpubl.).
Lewy, Z. 1975. The geological history of southern Israel during the Coniacian. *Israel J. Earth Sci.* 24: 19–43.
Lewy, Z. 1981. A late Bathonian–late Callovian unconformity in the Middle East. *Newsl. Stratigr*. 10: 27–33.
Lewy, Z. 1983. Upper Callovian ammonites and middle Jurassic geological history of the Middle East. *Bull. Geol. Surv. Israel* 76: 1–56.
Lewy, Z. & Raab, M. 1976. Mid-Cretaceous stratigraphy of the Middle East: Nice: *Ann. Mus. Hist. Nat.* 4: 32.1–32.20.
Lipson, S. 1971. Micropaleontological study of some sections in the Golan Heights. Tel Aviv: Israel Inst. of Petrol. Rep. 7/71, 7 pp.
Livnat, A. & Kronfeld, J. 1985. Paleoclimatic implications of U-series dates for lake sediments and travertines in the Arava rift valley, Israel. *Quat. Res*. 24: 164–172.
Livnat, A. & Kronfeld, J. 1990. Pleistocene lakes in the Arava Rift of Israel (Sayif Formation): facies and paleoenvironmental setting. *J. African Earth Sci.* 10(3): 409–420.
Loewengart, S. 1928. Zur Geologie der Küstenebene Palästinas. *Cbl. Für Mineral.* 9B: 498–519.
Lorch, J. 1959. A preliminary report on the pollen flora of Lake Huleh, Israel. *Bull. Res. Coun. Israel* 7D(2): 85–89.
Lorch, J. 1966. A Pleistocene florule from the central Jordan Valley. *Israel J. Botany* 15: 31–34.
Lorch, J. & Fahn, A. 1959. A new species of *Leguminoxylon* from the Mio-Pliocene of Israel. *Bull. Res. Coun. Israel* 7D(2): 65–70.
Lortet, L.C. 1880. Dragages profonds exécuté dans le lac de Tibériade (Syrie), en mai 1880. *C.R. Acad. Sci. Paris* 91: 500–502.
Lovelock, P.E.R. 1984. A review of the tectonics of the northern Middle East region. *Geol. Mag*. 121: 577–587.
Lowell, J.D. & Genike, G.J. 1972. Seafloor spreading and structural evolution of the southern Red Sea. *AAPG Bull*. 56: 247–259.
Luz, B. 1970. Biostratigraphy and quantitative significance of microfauna assemblages in the upper Cretaceous of the Judean Desert. MSc thesis, Dept. Geol., Hebrew Univ. Jerusalem, 52 pp. (in Hebrew, unpubl.).
Lynch, W.F., 1849. *Narrative of the United States' expedition to the River Jordan and the Dead Sea*: 508 pp. Philadelphia & London: Bentley.
Maccurdy, G.G. 1936. Prehistoric Man in Palestine. *Proc. Am. Philos. Soc. Philadelphia* 76(4): 523–541.
Macquer, P.J., Lavoisier, A.L. & Sage, B.G. 1778 (Publ. 1781). Analysis of water from the Asphaltite Lake. *Mem. Acad. R. Sci. Paris*: 69.
Macumber, P.G. 1992. The geological setting of Palaeolithic sites at Tabaqat Fahl, Jordan. *Paléorient* 18(2): 31–44.
Macumber, P.G. & Edwards, P.C. 1997. Preliminary results from the Acheulean site of Mashari'a 1, and a new stratigraphic framework for the Lower Palaeolithic of the East Jordan Valley. In H.G.K. Gebel, Z. Kafafi & G.O. Rollefson (eds). *The prehistory of Jordan, II. Perspectives from 1997*: 23–43. Berlin: ex Oriente.

Magaritz, M. 1986. Environmental changes recorded in the upper Pleistocene along the desert boundary, southern Israel. *Palaeogeogr. Palaeoclimatol. Palaeoecol*. 53: 213–229.

Makris, J., Ben-Avraham, Z., Behle, A., Ginzburg, A., Biese, P., Steinmetz, L., Whitmarsh, R.B. & Eleftheriou, S. 1983. Seismic refraction profiles between Cyprus and Israel and their interpretation. *Geophys. J. R. Astronomical Soc*. 75: 575–591.

Mallon, A. 1932. The five cities of the plain. *Palest. Explor. Fund Quart. St.*: 52–56.

Mann, P., Hempton, M. R., Bradley, D.C. & Burke, K. 1983. Development of pull-apart basins. *J. Geol.* 91: 529–554.

Manspeizer, W. 1985. The rift: Impact of climate and tectonism on Pleistocene and Holocene sedimentation. In K.T. Biddle & N. Christie-Blick (eds), *Strike-slip deformation, basin formation and sedimentation*: 143–158. Soc. of Econ. Paleontol. Mineral., Special Publication 37.

Marcet, A. 1807. An analysis of the water of the Dead Sea and the River Jordan. *Roy. Soc. Lond. Philos. Trans.* 1: 296–314.

Marco, S. 1997. A 70 Kyr paleoseismic record of the Dead Sea Graben – geophysical and geological studies. *Terra Nostra* 4/97 (no page numbers).

Marco, S., Rockwell, T., Heimann, A., Agnon, A. & Ellenblum, R. 2000. Historical earthquake deformations revealed by 3D trenching on Dead Sea Transform. Tokyo: Proc. Hokudan Intl. Symp. Sch. Active Faulting, 261–263.

Marcoux, J. 1978. A scenario for the birth of a new oceanic realm: The Alpine Neotethys. Jerusalem: Abst. 10th Intl. Cong. Sedimentol., 419–420.

Marcus, E. & Slager, J. 1985. The sedimentary-magmatic sequence of the Zemah 1 well (Dead Sea Rift, Israel) and its emplacement in time and space. *Israel J. Earth Sci.* 34(1): 1–10.

Marks, A.E. (ed.) 1976. *Prehistory and paleoenvironments in the central Negev, Israel*: Vol. 1: 383 pp. Dallas: SMU Press.

Marks, A.E. (ed.) 1977. *Prehistory and paleoenvironments in the central Negev, Israel*: Vol. 2: 356 pp. Dallas: SMU Press.

Mart, Y. 1982. Incipient spreading center in the Gulf of Elat, northern Red Sea. *Earth Planet. Sci. Lett.* 60: 117–126.

Mart, Y. 1991. The Dead Sea Rift: from continental rift to incipient ocean. *Tectonophysics* 197: 155–179.

Mart, Y. & Dauteuil, O. 2000. Analogue experiments of propagation of oblique rifts. *Tectonophysics* 316: 121–132.

Mart, Y. & Horowitz, A. 1981. The tectonics of the Timna region in southern Israel and the evolution of the Dead Sea Rift. *Tectonophysics* 79: 165–199.

Mart, Y. & Rabinowitz, P.D. 1986. The northern Red Sea and the Dead Sea Rift. *Tectonophysics* 124: 85–113.

Mart, Y. & Ross, D.A. 1987. Post-Miocene rifting and diapirism in the northern Red Sea. *Mar. Geol.* 74: 173–190.

Mart, Y., Eisin, B. & Folkman, Y. 1978. The Palmahim structure: a model of continuous tectonic activity since the upper Miocene in the southeastern Mediterranean off Israel. *Earth Planet. Sci. Lett*. 39(3): 328–334.

Martinotti, G. 1981. An Oligocene unconformity and its interregional interest. *Geol. Surv. Israel, Curr. Res.* 1981: 30–35.

Mass, D. 1978. A new numerical method for the interpretation of heat flow data with application to the Dead Sea region. MSc thesis, Weizmann Institute of Science, Rehovot, 124 pp. (in Hebrew, English summary, unpubl.).

Masterman, E.W.G. 1902. Observations of the Dead Sea levels. *Palest. Explor. Fund, Quart. St.*: 155–160; 297–299; 406–407.

Matmon, A. 2000. The Lower Galilee – a continental rift!? Ma'alot: Abst., Israel Geol. Soc. Ann. Mtg., 87.
Matmon, A., Enzel, Y., Zilberman, E. & Heimann, A. 1999. Late Pliocene and Pleistocene reversal of drainage systems in northern Israel: tectonic implications. *Geomorphology* 28: 43–59.
McClay, K.R., Nichols, G.J., Khalil, S.M., Darwish, M. & Bosworth, W. 1998. Extensional tectonics and sedimentation, eastern Gulf of Suez, Egypt. In B.H. Purser & D.W.J. Bosence (eds), *Sedimentation and tectonics in rift basins: Red Sea – Gulf of Aden*: 223–238. London: Chapman & Hall.
McKenzie, D.P., Davies, D. & Molnar, P. 1970. Plate tectonics of the Red Sea and East Africa. *Nature* 224: 125–133.
Mercier, J.L., Sorel, D. & Simeakis, K. 1987. Changes in the state of stress in the overriding plate of a subduction zone: the Aegean arc from the Pliocene to the present. *Annales Tectonicæ* I(1): 20–39.
Mercier, N., Valladas, H., Valladas, G., Reyss, J.L., Jelinek, A., Meignen, L. & Joron, J.L. 1995. TL dates of burnt flints from Jelinek's excavations at Tabun and their implications. *J. Arch. Sci.* 22: 495–509.
Michelson, H. 1972. The hydrogeology of the southern Golan Heights. Tel Aviv: Tahal (Water Planning for Israel) Rep. HR/72/037, 89 pp. (in Hebrew, columnar sections in English).
Michelson, H. 1973. "Yarmouk Basalt" and "Raqqad Basalt" – two volcanic phases which flowed through pre-existing gorges. *Israel J. Earth Sci.* 22(1): 51–58.
Michelson, H. 1982. Geological survey of the Golan Heights. Tel Aviv: TAHAL (Water Planning for Israel) Rep., 35 pp.
Michelson, H. & Lipson-Benitah, S. 1986. The litho- and biostratigraphy of the southern Golan Heights. *Israel J. Earth Sci.* 35(3–4): 221–240.
Michelson, H., Flexer, A. & Erez, Z. 1987. A comparison of the eastern and western sides of the Sea of Galilee and its implication on the tectonics of the northern Jordan Rift Valley. *Tectonophysics* 141(1–3): 125–134.
Milankovitch, M. 1941. *Kanon der Erdbestrahlung und seine Anwendung auf das Eiszeitenproblem*: 633 pp. Belgrade: Königlich Serbische Akademie Sp. Publ. 133.
Mimran, Y. 1972. The Tayasir Volcanics, a lower Cretaceous formation in the Shomeron, central Israel. *Bull. Geol. Surv. Israel* 52: 1–9.
Mimran, Y. 1976. Deep-seated faulting and the structures of the northern Negev (Israel). An analytical and experimental study. *Israel J. Earth Sci.* 25: 111–126.
Mimran, Y. 1984. The geology of Wadi Malih. MSc thesis, Dept. Geol., Hebrew Univ. Jerusalem, 68 pp. (in Hebrew, unpubl.).
Mimran, Y., Shaliv, G., Hatzor, Y. & Lang, B. 1989. The margins of the Rift Valley, south of Bet She'an and the eastern Gilbo'a. Ramot: Israel Geol. Soc. Ann. Mtg. Excursion guidebook, 47–65 (in Hebrew).
Mitchell, D.J.W., Allen, R.B., Salama, W. & Abouzakm, A. 1992. Tectonostratigraphic framework and hydrocarbon potential of the Red Sea. *J. Petr. Geol.* 15: 187–210.
Molyneux, A. 1848. Expedition to the Jordan and the Dead Sea. *J. R. Geogr. Soc. London* 18(2): 104–130.
Montague, E.P. 1849. *Narrative of the late expedition to the Dead Sea, from a diary by one of the party*: 366 pp. Philadelphia: Carey & Hart.
Montenat, C., Ott d'Estevou, P., Purser, B., Burollet, P.F., Jarrige, J.J., Orszag-Sperber, F., Philobbos, E., Plaizat, J.C., Prat, P., Richert, J.P., Roussel, N. & Thiriet, J.P. 1988. Tectonic and sedimentary evolution of the gulf of Suez and the northwestern Red Sea. *Tectonophysics* 153: 161–177.

Moore, G.H. & Beke, W.G. 1837. On the Dead Sea and some positions in Syria. *J. R. Geog. Soc.* 7: 456.
Mor, D. 1973. The volcanism of the central Golan Plateau. MSc thesis, Dept. Geol., Hebrew Univ. Jerusalem, 179 pp. (in Hebrew, unpubl.).
Mor, D. 1986. The volcanism of the Golan Heights. Jerusalem: Geol. Surv. Israel Rep. GSI/5/86, 159 pp. (in Hebrew, English summary).
Mor, D. 1989. Volcanic phenomena in the southern Golan Plateau. Ramot: Israel Geol. Soc. Ann. Mtg. Excursion guidebook, 120–125 (in Hebrew).
Mor, D. & Steinitz, G. 1985. The history of the Yarmouk River based on K–Ar dating and its implication on the development of the Jordan Rift. Jerusalem: Geol. Surv. Israel Rep. GSI/40/85, 18 pp.
Mor, D., Levitte, D., Steinitz, G. & Lang, B. 1987. Volcanic history of the Dalton Heights (upper Galilee), according to K–Ar ages. Mitzpe Ramon: Abst., Israel Geol. Soc. Ann. Mtg., 50–52.
Mor, U. 1987. The geology of the Judean Desert in the Nahal Darga region. Jerusalem: Geol. Surv. Israel Rep. GSI/21/87, 112 pp. (in Hebrew, English summary).
Moretti, I. & Colletta, B. 1987. Spatial and temporal evolution of the Suez rift subsidence. *J. Geodyn*. 7: 151–168.
Mouti, M., Delaloye, M., Fontignie, D., Piskin, O. & Wagner, J.J. 1992. The volcanic activity in Syria and Lebanon between Jurassic and actual. *Schweitzerische Mineralogische und Petrographische Mitteilungen* 72: 91–105.
Mülinen, Graf von, 1908. Beiträge zur Kenntnis des Karmels. *Ztschr. Dtsch. Paläst. Ver.* 31: 1–190.
Murris, R.J. 1980. Middle East: stratigraphic evolution and oil habitat. *AAPG Bull.* 64: 597–618.
Nathan, Y. & Shahar, Y. 1977. Continental phosphate mineral assemblages in Senonian unconformities, southern Israel. Zikhron Ya'aqov: Abst., Israel Geol. Soc. Ann. Mtg., 28–29.
Nathan, Y., Shiloni, Y., Roded, R., Gal, I. & Deutsch, Y. 1979. The geochemistry of the northern and central Negev phosphorites. *Bull. Geol. Surv. Israel* 73: 1–41.
Nativ, R. 1997. Saline springs near/in Lake Kinneret, their origin and the mechanisms controlling their discharge. *Terra Nostra* 4/97 (no page numbers).
Nachmias, J. 1969. Source rocks of the Saqiye Group sediments in the coastal plain of Israel – a heavy mineral study. *Israel J. Earth Sci.* 18: 1–16.
Neev, D. 1960. A pre-Neogene erosion channel in the southern coastal plain of Israel. *Bull. Geol. Surv. Israel* 25: 1–21.
Neev, D. 1967. The late Holocene fluviatile history of the Kinneret (Tiberias) and Bet She'an basins. *Israel J. Earth Sci.* 16(1): 50.
Neev, D. 1979. Deep-water gypsum deposits as indicated by the Neogene geological history of the central coastal plain of Israel. *Sediment. Geol.* 23: 127–136.
Neev, D. & Emery, K.O. 1967. The Dead Sea. *Bull. Geol. Surv. Israel* 41: 1–147.
Neev, D. & Emery, K.O. 1995. *The destruction of Sodom, Gomorrah and Jericho: geological, climatological and archaeological background*: 175 pp. New York: Oxford Univ. Press.
Neev, D. & Hall, J.K. 1976. The Dead Sea geophysical survey, 19 July–1 August 1974 – Final Report No. 2. Jerusalem: Geol. Surv. Israel Mar. Geol. Div. Rep. 6/76, 21 pp.
Neev, D. & Hall, J.K. 1979. Geophysical investigations in the Dead Sea. *Sediment. Geol.* 23(1/4): 209–238.
Neev, D., Greenfield, L.L. & Hall, J.K., 1982. Slice tectonics in the Levantine Basin. Jerusalem: Geol. Surv. Israel Rep. MGG/2/82, 29 pp.

Neuville, R. 1951. Le Paléolithique et le Mésolithique du Désert de Judée. *Mém. Arch. Inst. Paléont. Humaine Paris* 24: 1–270.
Nevo, E. 1963. The Jurassic strata of Makhtesh Ramon. Mitzpe Ramon: Abst., Israel Geol. Soc. Ann. Mtg., 11–12 (in Hebrew).
Nichols, G. & Watchorn, F. 1998. Climatic and geomorphic controls on rift sedimentation: Oligo-Miocene syn-rift facies in the Gulf of Aden, Yemen. *Mar. Petrol. Geol.* 15(6): 505–518.
Niemi, T.M. & Ben-Avraham, Z. 1994. Evidence for Jericho earthquakes from slumped sediments of the Jordan River delta in the Dead Sea. *Geology* 22: 395–398.
Niemi, T.M. & Ben-Avraham, Z. 1997. Active tectonics in the Dead Sea. In T.M. Niemi, Z. Ben-Avraham & J.R. Gat (eds), *The Dead Sea, the lake and its setting*: 73–81. Oxford: Oxford Univ. Press.
Niethammer, C. 1974. *American Indian food and lore*: 191 pp. New York: Collier Books.
Nissenbaum, A. 1970. Chemical analyses of Dead Sea and Jordan River water, 1778–1830. *Israel J. Chem.* 8: 281–287.
Nissenbaum, A. 1977. The physicochemical basis for legends of the Dead Sea. In D.C. Greer (ed.), *Desertic terminal lakes*: 159–168. Salt Lake City: Univ. Utah.
Nissenbaum, A. 1978. Dead Sea asphalts – historical aspects. *AAPG Bull.* 62(5): 837–844.
Nötling, F. 1886. Das Tote Meer und der untergang von Sodom und Gomorraha. *Deutsch. Montagsblatt Berlin* 27: 31 & 33.
Nötling, F. 1887. Geologische skizze der umgebung von El-Hammi. *Ztschr. Dtsch. Paläst. Ver.* 10: 59–88.
Nur, A. & Helseley, C.E. 1971. Paleomagnetism of Tertiary and Recent lavas of Israel. *Earth Planet. Sci. Lett.* 10: 376–379.
Ohel, M.Y. 1986. *The Acheulean of the Yir'on Plateau, Israel*: 1–192. Oxford: BAR Intl. Ser. 307.
Ohel, M.Y. 1991.Prehistoric survey of the Bar'am Plateau. *Palestine Explor. Quart.* Jan–Jun: 32–47.
Oliphant, L. 1886. Excavations on Carmel. *Palestine Explor. Quart.* 19: 7–15.
Omar, G.I. & Steckler, M.S. 1995. Fission track evidence on the initial rifting of the Red Sea: two pulses, no propagation. *Science* 270: 1341–1344.
Opdyke, N.D., Lindsay, E. & Kukla, G. 1985. Evidence for earlier date of Ubeidiya, Israel hominid site. *Nature* 304: 375.
Oppenheim, M.J. 1959. Volcanological phenomena in the southeastern Lower Galilee. PhD dissertation, Dept. Geol., Hebrew Univ. Jerusalem, 177 pp. (in Hebrew, English summary, unpubl.).
Oppenheim, M.J. 1962. The geology of the southeastern Galilee lava fields. *Bull. Res. Counc. Israel* 11G(2): 58–84.
Ouda, K. & Masoud, M. 1993. Sedimentation history and geological evolution of the Gulf of Suez during the late Oligocene–Miocene. *Geol. Soc. Egypt, Spec. Publ.* 1: 47–88.
Palmer, E.H. 1871. *The desert of the Exodus: Journeys on foot in the wilderness of the forty year's wanderings, undertaken in connection with the Ordnance Survey of Sinai and the Palestine Exploration Fund*: 576 pp. Cambridge: Belland.
Parenti, F., Al-Shiyab, A.H., Santucci, E., Kafafi, Z. & Palumbo, G. 1997. Early Acheulean stone tools and fossil faunas from the Dauqara Formation. In H.G.K. Gebel, Z. Kafafi & G.O. Rollefson (eds), *The prehistory of Jordan, II. Perspectives from 1997*: 7–22. Berlin: ex Oriente.
Parker, D.H. 1970. *The hydrogeology of the Mesozoic–Cainozoic aquifers of the western highlands and plateau of east Jordan*: 286 pp., maps. Rome: FAO.

Parnes, A. 1971. Later lower Cambrian trilobites from the Timna area and Har Amram (southern Negev, Israel). *Israel J. Earth Sci.* 20: 179–205.

Parnes, A. 1975. Middle Triassic ammonite biostratigraphy in Israel. *Bull. Geol. Surv. Israel* 66: 1–25.

Parnes, A. 1980. Lower Jurassic (Liassic) invertebrates from Makhtesh Ramon, Negev, Southern Israel. *Israel J. Earth Sci.* 29: 107–113.

Parnes, A. 1981. Biostratigraphy of the Mahmal Formation (middle and upper Bajocian) in Makhtesh Ramon, Negev, southern Israel. *Bull. Geol. Surv. Israel* 74: 1–55.

Partridge, T.C. 1992. The palaeoecological background to the transition from *Australopithecus* to *Homo* in southern Africa. Jerusalem: Abst., 3rd Intl. Cong. Human Paleontol., 91.

Pe'eri, S., Wdowinski, S., Shtibelman, A., Bechor, N. & Bock, Y. 1999. Current deformation across the Dead Sea fault as observed from 18 months of continuous GPS monitoring. En Boqeq: Abst., Israel Geol. Soc. Ann. Mtg., 65.

Perathoner, B., Fuchs, K., Prodehl, C. & Ginzburg, A. 1981. Seismic investigation of crust-mantle transition in continental rift systems – Jordan–Dead Sea rift and Rhinegraben. *Tectonophysics* 80: 121–133.

Perrot, J. 1960. Excavations at Eynan (Ein Mallaha); preliminary report on the 1959 season. *Israel Expl. J.* 10(1): 14–22.

Petrbok, J. 1925. La stratigraphie et la paléontologie de la terrace pluviale (Mindélienne) de Jourdain prés de Dagania Alef. *Bull. Intern. Acad. Sci. Bohême Prague* 26: 87–100.

Petrbok, J. 1926a. Stratigraphy and fauna of the Neolithic strata at Kinneret in Palestine. *Bull. Intern. Acad. Sci. Bohême Prague* 27: 448–455.

Petrbok, J. 1926b. On the stratigraphy of the Palestine Paleolithic. *Bull. Intern. Acad. Sci. Bohême Prague* 27: 239–256.

Petrie, W.M.F. & Currely, C.T. 1906. *Researches in Sinai*: 280 pp. London: Murray.

Picard, L. 1928. Über die verbreitung des Pliozäns im nördlichen Palästina. *Centralblatt f. Miner.* 5: 326–335.

Picard, L. 1929. Zur Geologie der Besan-Ebene. *Ztschr. Dtsch. Paläst. Ver.* 52: 24–90.

Picard, L. 1931. *Geological researches in the Judean Desert*: 108 pp. Jerusalem: Goldberg Press.

Picard, L. 1932. Zur Geologie des mittleren Jordantales. *Ztschr. Dtsch. Paläst. Ver.* 55: 169–236.

Picard, L. 1936. Conditions of underground water in the western Emeq. *Bull. Geol. Dept. Hebrew Univ. Jerusalem* 1: 1–24.

Picard, L. 1937. Inferences on the problem of the Pleistocene climate of Palestine and Syria. *Proc. Prehist. Soc. Cambridge* 5: 58–70.

Picard, L. 1938. Synopsis of stratigraphic terms in Palestinian geology. *Bull. Geol. Dept. Hebrew Univ. Jerusalem* 2(2): 1–24.

Picard, L. 1939. Outline of the tectonics of the Earth, with special emphasis upon Africa. *Bull. Geol. Dept. Hebrew Univ. Jerusalem* 2(3–4): 1–66.

Picard, L. 1943. Structure and evolution of Palestine, with comparative notes on neighboring countries. *Bull. Geol. Dept. Hebrew. Univ. Jerusalem* 4(2–4): 134 pp.

Picard, L. 1947. Review of petroleum geology in 1946 "Palestine": search and conceptions. *Quart. Colorado Sch. Mines* 42(3): 122–123.

Picard, L. 1950. Geological report of Jordan Exploration Company concession, Dead Sea, Israel. In: A. Franklin, Oil prospects of the Jordan Exploration Company Ltd. Concession at the south end of the Dead Sea, Israel. Houston, Texas: Jordan Exploration Company Ltd., 48 pp., unpubl.

Picard, L. 1952. The Pleistocene peat of Lake Hula. *Bull. Res. Coun. Israel* II(2): 147–156.
Picard, L. 1954. History of mineral research in Israel. *Israel Economic Forum* 9(3): 10–38, 146–150.
Picard, L. 1959. Geology and oil exploration of Israel. *Bull. Res. Coun. Israel* 8G(1): 1–30.
Picard, L. 1963. The Quaternary in the northern Jordan Valley. *Proc. Israel Acad. Sci. Hum.* 1(4): 1–34.
Picard, L. 1965. The geological evolution of the Quaternary in the central-northern Jordan Graben, Israel. *Geol. Soc. Am., Sp. Pap.* 84: 337–366.
Picard, L. 1987. The Elat (Aqaba)-Dead Sea Jordan subgraben system. *Tectonophysics* 141: 23–32.
Picard, L. & Baida, U. 1966a. Stratigraphic position of the Ubeidiya formation. *Proc. Israel Acad. Sci. Hum.* 4: 1–8.
Picard, L. & Baida, U. 1966b. *Geological report on the lower Pleistocene deposits of the Ubeidiya excavations*: 39 pp. Jerusalem: Israel Acad. Sci. Hum.
Picard, L. & Flexer, A. 1974. Studies in the stratigraphy of Israel: the Triassic. Tel Aviv: Israel Inst. Petrol. and Energy, 62 pp.
Picard, L. & Hirsch, F. 1987. *The Jurassic stratigraphy in Israel and the adjacent countries*: 107 pp. Jerusalem: Israel Acad. Sci. Hum.
Picard, L. & Solomonica, P. 1936. On the geology of the Gaza-Be'er Sheva district. *J. Palestine Oriental Soc.* 16: 180–223.
Plakht, J. 1996. Mapping of Quaternary units in Makhtesh Ramon, central Negev. *Israel J. Earth Sci.* 45(4): 217–222.
Plakht, J. 1999. Quaternary maps of Makhtesh Hatira and Makhtesh Hazera. En Boqeq: Abst., Israel Geol. Soc. Ann. Mtg., 70.
Pococke, R. 1754. *Beschreibung des Morgenlandes und einiger anderer Länder.* Vol. II: *Palästina, Syrien, Mesopotamien, Cypern*: 374 pp. Erlangen: Verlag der Stifthauses.
Ponikarov, V.P., Kazmin, V.G., Mikhailov, I.A., Razvalyayev, A.V., Krasheninnikov, V.A., Kozlov, V.V., Sulidi-Kondratev, E.D., Mikhailov, K.Y., Kulakov, V.V., Fardzhev, V.A. & Mirzayev, K. 1967. *The geology of Syria; explanatory notes on the geological map of Syria 1 : 500,000*: 269 pp. Moscow: Technoexport (in Russian).
Ponikarov, V.P., Kazmin, V.G., Kozlov, V.V., Krasheninnikov, V.A., Mikhailov, I.A., Razvlyaev, A.V., Soulidi-Kondratyev, E.D., Uflyand, S.A.K. & Faradzhev, V.A., 1969. *Geology and mineral resources of foreign countries: Syria*: 216 pp. Moscow: Technoexport (in Russian).
Por, D. 1963. The relict aquatic fauna of the Jordan Rift Valley. *Israel J. Zool.* 12: 47–58.
Porat, N., Zhou, L.P., Chazan, M., Noy, T. & Horwitz, L.K. 1999. Dating the Lower Paleolithic open-air site of Holon, Israel, by luminescence and ESR techniques. *Quat. Res.* 51: 328–341.
Powers, R.W., Ramirez, L.F., Redmond, C.P. & Elberg Jr. E.J., 1966. Geology of the Arabian Peninsula: sedimentary geology of Saudi Arabia. U.S. Geol. Surv. Prof. Paper, 560-D, 147 pp.
Press, F. & Siever, R. 1998. *Understanding Earth*: 682 pp. New York: Freeman.
Purser, B.H. & Bosence, D.W.J. (eds) 1998. *Sedimentation and tectonics in rift basins: Red Sea–Gulf of Aden*: 663 pp. London: Chapman & Hall.
Quennell, A.M. 1959. The structural and geomorphic evolution of the Dead Sea Rift. *Quart. J. Geol. Soc. London* 114: 1–24; 432–433.
Quennell, A.M. 1996. The western Arabia rift system. In J.E. Dixon & A.H.F. Robertson (eds), *The geological evolution of the eastern Mediterranean* (2nd printing of 1984 edn): 775–788. London: Geol. Soc.

Qummou, S., Abu Jassar, W., Arad, V &. Bartov, Y. 1997. *Bibliography of Middle East geology, part I: Jordan*: 315 pp. Amman: Nat. Res. Authority; Jerusalem: Geol. Surv. Israel.

Raab, M. 1998. The origin of the evaporites in the Jordan–Dead Sea Valley in view of the evolution of brines and evaporites during seawater evaporation. Jerusalem: Geol. Surv. Israel Rep. GSI/1/98, 134 pp.

Rabinowitz, N. & Mart, Y. 2000. Seismic tomography of the Dead Sea region: thinned crust, anomalous velocities and possible magmatic diapirism. In B. Vendeville, Y. Mart & J.L. Vigneresse (eds), *From the Arctic to the Mediterranean: salt, shale and igneous diapirs in and around Europe*: 79–92. London: Geol. Soc. Sp. Publ. 174.

Ram, E. 1989. The magnetic field over the Dead Sea-Lisan area. MSc thesis, Tel Aviv University, 55 pp. (in Hebrew, English summary, unpubl.).

Range, P. 1921. Die Entwicklungsmöglichkeiten des Bergbaues in Syrien und Palästina. *Ztschr. f. Pract. Geol.* 20(8): 113–116.

Ravikovitch, S. 1969. *Manual and map of soils of Israel*. Jerusalem: Magnes Press, 96 pp., 1 : 250,000 map (in Hebrew, map also in English).

Raz, E. 1983. The geology of the Judean Desert, En Gedi area. Jerusalem: Geol. Surv. Israel Rep. S/3/83, 110 pp. (in Hebrew).

Reches, Z. & Hoexter, D. 1981. Holocene seismic and tectonic activity in the Dead Sea area. *Tectonophysics* 80: 235–254.

Reches, Z., Hoexter, D.F. & Hirsch, F. 1981. The structure of a monocline in the Syrian Arc system, Middle East. Surface and subsurface analysis. *J. Petrol. Geol*. 4: 413–425.

Reclus, É. 1872. *The Earth*: 573 pp. New York: Harper.

Reiss, Z. 1962. Stratigraphy of phosphate deposits in Israel. *Bull. Geol. Surv. Israel* 34: 1–23.

Renouard, G. 1955. Oil prospects of Lebanon. *AAPG Bull.* 39: 2125–2169.

Repenning, C.A. & Fejfar, O. 1982. Evidence for earlier date of Ubeidiya, Israel, hominid site. *Nature* 299: 344–347.

Reznikov, M., Ben Gai, Y. & Ben Avraham, Z. 1999. Combined gravity and seismic interpretation in Lake Kinneret. En Boqeq: Abst., Israel Geol. Soc. Ann. Mtg., 72.

Rialle, G. de, 1868. L'Antiliban. *Bull. Soc. Géog. Paris (Sér 5)*: 225–265.

Richardson, M. & Arthur, M.A. 1988. The Gulf of Suez–Northern Red Sea Neogene Rift: A quantitative basin analysis. *Mar. Petr. Geol*. 5: 247–270.

Ricou, L.E. 1971. Le croissant ophiolitique peri-arabe. Une ceinture de nappes mises en place au Crétace supèrieur. *Rev. Géog. Phys. Géol. Dyn.* 13: 327–350.

Rihm, R. & Henke, C.H. 1998. Geophysical studies on early tectonic controls on Red Sea rifting, opening and segmentation. In B.H. Purser & D.W.J. Bosence (eds), *Sedimentation and tectonics in rift basins: Red Sea–Gulf of Aden*: 29–49. London: Chapman & Hall.

Ritter, C. 1850. *Der Jordan und die Beschiffung des Todten Meeres; ein Vortrag*: 38 pp. Berlin: Reimer.

Robertson, A.H.F. 1978. The sedimentary evolution of the Trodos Ocean. An excursion guide to the sedimentary geology of southern Cyprus. Jerusalem: Guidebook 10th Intl. Cong. Sedimentol. III: 22–75.

Robertson, A.H.F. & Woodcock, N.H. 1982. Sedimentary history of the south-western segment of the Mesozoic–Tertiary Antalya continental margin, south-west Turkey. *Eclogae Geol. Helv*. 75(3): 517–562.

Robinson, E. & Smith, E. 1841. *Palästina und die südlich angrenzenden länder; Tagebuch einer reise im jahre 1838 in bezug auf die Biblische Geographie Unternommen*: Vol. 1: LXXXIV+442 pp., Vol. 2: 756 pp., Vol. 3: 1116 pp. Halle: Weisenhaus.

Robson, A.R. 1971. The structure of the Gulf of Suez (Clysmic) Rift, with special reference to the eastern side. *Q. J. Geol. Soc.* 127: 247–276.
Roeser, H.A. 1975. A detailed magnetic survey of the southern Red Sea. *Geol. Jahrb.* D13: 131–153.
Rofe & Raffety Consulting Engineers Ltd. 1965. Nablus district water resources survey, geological and hydrological report. Amman: Kingdom of Jordan Central Water Authority, 120 pp.
Rohrlich, V., Metzer, A. & Zohar, E. 1980. Potential iron ores in the Lower Cretaceous of Israel and their origin. *Israel J. Earth Sci*. 29: 73–80.
Ron, H. 1983. Paleomagnetic investigation of the fault structure of northern Israel. Nazerat: Abst., Israel Geol. Soc. Ann. Mtg., 67–69.
Ron, H. 1987. Deformation along Yammouneh, the restraining bend of the Dead Sea transform: Paleomagnetic data and kinematic implications. *Tectonics* 6: 653–666.
Ron, H. & Levy, S. 2001. When did hominids first leave Africa? New high resolution paleomagnetic evidence from the Erk-el-Ahmar Formation, Israel. Elat: Abst., Israel Geol. Soc. Ann. Mtg., 96.
Ron, H., Freund, R., Garfunkel, Z. & Nur, A. 1984. Block rotation by strike-slip faulting: Structural and paleomagnetic evidence. *J. Geophys. Res.* 89: 6256–6270.
Ron, H., Shamir, G. & Eyal, Y. 1997. Deformation of the Misgav Am–Margaliyyot block between the Roum and Margaliyyot faults. Kefar Gil'adi: Israel Geol. Soc. Ann. Mtg. Excursion guidebook, 33–45 (in Hebrew).
Ron, H., Porat, N. & Ronen, A. 2001. Magnetostratigraphy and age dating of the soil sequence at the prehistoric Evron Quarry: evidence for one Ma human colonization. Elat: Abst., Israel Geol. Soc. Ann. Mtg., 97.
Ronen, A. 1975. The Paleolithic archaeology and chronology of Israel. In E. Wendorf & A.E. Marks (eds), *Problems in prehistory: North Africa and the Levant*: 229–248. Dallas: SMU Press.
Ronen, A. 1979. Paleolithic industries. In A. Horowitz, *The Quaternary of Israel*: 296–307. New York: Academic Press.
Ronen, A. 1991. The Lower Paleolithic site Evron–Quarry in western Galilee, Israel. *Sonderveröffentlichungen Geologisches Institut der Universität Köln* 82: 187–212.
Ronen, A. 1996. The earliest hominids in Israel. In K.W. Beinhauer, R. Kraatz & A. Wagner (eds), *Homo erectus heidelbergensis von Mauer*: 99–102. Heidelberg: Jan Thorbecke Sigmaringen.
Ronen, A., Inbar, M., Klein, M. & Brunnacker, K. 1980. Artifact-bearing gravels beneath the Yir'on Basalt. *Israel J. Earth Sci.* 29: 221–226.
Rosenberg, E. 1960. Geologische Untersuchungen in den Naftalibergen. PhD dissertation, Inst. Geol., Zürich Univ., 107 pp. (unpubl.).
Rosendahl, W., Schmitz, M. & Hiyari, M.A. 1997. A first note on new localities of Pleistocene mammals in Jordan. Preliminary results of field research in 1995. In H.G.K. Gebel, Z. Kafafi & G.O. Rollefson (eds), *The prehistory of Jordan, II. Perspectives from 1997*: 493–496. Berlin: ex Oriente.
Rosenfeld, A., Segev, A. & Halbersberg, E. 1981. Ostracode species and paleosalinities of the Pliocene Bira and Gesher formations (northwestern Jordan Valley). *Israel J. Earth Sci.* 30(4): 113–119.
Rosenthal, E. 1980. Hydrogeology and hydrogeochemistry of the Bet She'an, Harod and Mehola valleys. PhD dissertation, Dept. Geol., Hebrew Univ. Jerusalem, 244 pp. (in Hebrew, English summary, unpubl.).

Rosenthal, Y., Katz, A. & Tchernov, E. 1989. The reconstruction of Quaternary freshwater lakes from the chemical and isotopic composition of gastropod shells: the Dead Sea Rift, Israel. *Palaeogeogr. Palaeoclimatol. Palaeoecol*. 74: 241–253.
Rossignol, M. 1969a. Sèdimentation palynologique récente dans la Mer Morte. *Pollen et Spores* 11(1): 17–38.
Rossignol, M. 1969b. Une sequence climatique du Pléistocène dans la region de la Mer Morte (Israël). *Pollen et Spores* 11(3): 603–614.
Rossignol-Strick, M. 1983. African monsoons, an immediate climate response to orbital insolation. *Nature* 304: 46–49.
Roth, I. 1970. Wadi el Qilt (Sheet 12-I) – geological map 1 : 50,000. Jerusalem: Geol. Surv. Israel.
Roth, J.R. 1858. Reise von Jerusalem und dem Toten Meeres durch die Araba bis zum Roten Meere. *Peterm. Geogr. Mitt.* 1–5: 267–272.
Rotstein, Y. & Arieh, E. 1986. Tectonic implications of recent microearthquake data from Israel and adjacent areas. *Earth Planet. Sci. Lett.* 78: 237–244.
Rotstein, Y. & Bartov, Y. 1989. Seismic reflection across a continental transform: An example from a convergent segment of the Dead Sea Rift. *J. Geophys. Res.* 94: 2902–2912.
Rotstein, Y. & Bartov, Y. 1992. Evidence for local shifting of the main fault and changes in the structural setting, Kinarot basin, Dead Sea transform. *Geology* 20: 251–254.
Rotstein, Y. & Ben-Avraham, Z. 1986. Active tectonics in the eastern Mediterranean. *Israel J. Earth Sci*. 35(1): 23–39.
Rotstein, Y., Bartov, Y. & Hofstetter, A. 1991. Active compressional tectonics in the Jericho area, Dead Sea Rift. *Tectonophysics* 198: 2–4.
Rotstein, Y. Bartov, Y. & Frieslander, U. 1992. Evidence for local shifting of the main fault and changes in the structural setting, Kinarot basin, Dead Sea transform. *Geology* 20: 251–254.
Russegger, J. 1841. Über die depression des Toten Meeres und des ganzen Jordantales. *Poggendrof's Annal.* 53: 179–194.
Russeger, J. 1847. *Reisen in Unter-Ägypten, auf der Halbinsel des Sinai und im gelobten Lände.* 292 pp. Stuttgart: Schweizerbart.
Rybakov, M., Goldshmidt, V. & Rotstein, Y. 1997. A new compilation of gravity and magnetic data from the Levant and their preliminary interpretation. *Geophys. Res. Lett.* 24: 33–36.
Rybakov, M., Goldshmidt, V., Fleischer, L. & Rotstein, Y. 1999. The crystalline basement in central Israel, derived from gravity and magnetic data. *Israel J. Earth Sci.* 48: 101–111.
Rybakov, M., Goldshmidt, V. & Rotstein, Y. 2000. A new look at the Dead Sea Transform shoulders. Dead Sea: Abst., The first Stephan Mueller Conf. Eur. Geophys. Soc., p. 30.
Sa'ar, H. 1986. Origin and sedimentology of the sandstones in graben fill formations of the Dead Sea Rift Valley. Jerusalem: Geol. Surv. Israel Rep. GSI/3/86, 81 pp. (in Hebrew, English summary).
Sachsse, R. 1897. Beitrag zur chemischen kenntnis der mineralien, gesteine und gewächse Palästinas. *Ztschr. Dtsch. Paläst. Ver.* 20: 1–33.
Sagy, A. & Reches, Z. 2000. Fault patterns, joint systems and slip partitioning along the western margins of the Dead Sea Rift. Dead Sea: Abst., The first Stephan Mueller Conf. Eur. Geophys. Soc., 34.
Said, R. 1962. *The geology of Egypt*: 377 pp. Amsterdam: Elsevier.
Saint-Marc, P. 1974. Étude stratigraphique et micropaléontologique de l'albien, du Cenomanien and du Turonien du Liban. *Notes et Mém. Moyen Orient* XIII: 1–342.

Sakal, E., Raab, M. & Reiss, Z. 1966. Upper Eocene in the southern Negev. *Israel J. Earth Sci.* 15: 42–45.
Salamon, A., Hofstetter, A., Garfunkel, Z. & Ron, H. 1996. Seismicity of the eastern Mediterranean region: perspective from the Sinai subplate. *Tectonophysics* 263: 293–305.
Salem, R. 1976. Evolution of Eocene–Miocene sedimentation patterns in parts of northern Egypt. *AAPG Bull.* 60: 34–64.
Saltzman, U. 1964. The geology of the Tabgha–Hukok–Migdal area. MSc thesis, Dept. Geol., Hebrew Univ. Jerusalem, 55 pp. (in Hebrew, unpubl.).
Saltzman, U., Schulman, N. & Flexer, A. 1981. Remarks on the structure of the Rosh Pinna block. Katzrin: Abst., Israel Geol. Soc. Ann. Mtg., 51–52.
Sass, E. & Bein, A. 1982. The Cretaceous carbonate platform in Israel. *Cretaceous Res.* 3: 135–144.
Sass, E. & Katz, A. 1982. The origin of platform dolomites: new evidence. *Am. J. Sci*. 282: 1184–1213.
Sass, E. & Oppenheim, M.J. 1965. The petrology of some Cenomanian sediments from the Judean Hills, Israel and the paleo-environmental break of the Motsa Marl. *Israel J. Earth Sci.* 14: 91–122.
Savage, R.J.G. & Tchernov, E. 1968. Miocene mammals of Israel. *Proc. Geol. Soc. Lond.* 1648: 98–101.
Savoyat, E. & Balcha, T. 1989. Petroleum exploration in the Ethiopian Red Sea. *J. Petrol. Geol.* 12: 187–204.
Saulcy, F. de, 1854. *Narrative of a journey round the Dead Sea and in the Bible lands in 1850 and 1851*: Vol. 1: 560 pp.; Vol. 2: 604 pp. London: Bentley.
Schlein, N. 1961. The geology of the Alma region. MSc thesis, Dept. Geol., Hebrew Univ. Jerusalem, 45 pp. (in Hebrew, unpubl.).
Schmidt, D.L., Hadley, D.G. & Brown, G.F. 1982. Middle tertiary continental rift and evolution of the Red Sea in southwestern Saudi Arabia. US Geol. Surv. Rep. USGS-OF-03-6, 56 pp.
Schramm, A., Stein, M. & Goldstein, S.L. 1997. ^{14}C time-scale calibration and relation to global paleoclimate: evidence from U-series and ^{14}C dating of Lake Lisan (paleo-Dead Sea sediments. *Terra Nostra* 4/97 (no page numbers).
Schrötter, H. 1924. Esquisse de géographie et de balnéologie de la Mer Morte, avec renseignements sur la flore du littoral de cette mer. *La Géographie* 42: 162–210.
Schubert, G.H. von. 1840. Über die organischen findlinge des Totem Meeres. *Geol. Anzeig. München* 11: 356–364.
Schubert, G.H. von, & Roth, J.R. 1838/1839. *Reisen in dem Morgenland in den Jahren 1836 und 1837*: Vol. 1: 532 pp.; Vol. 2: 592 pp.; Vol. 3: 596 pp. Erlangen: Palin & Enke.
Schuldenrein, J. & Goldberg, P. 1981. Late Quaternary paleoenvironments and prehistoric site distributions in the lower Jordan Valley: a preliminary report. *Paléorient* 7(1): 57–71.
Schulman, N. 1959. The geology of the central Jordan Valley. *Bull. Res. Coun. Israel* 8G(2–3): 63–90.
Schulman, N. 1962. The geology of the central Jordan Valley. PhD dissertation, Dept. Geol., Hebrew Univ. Jerusalem, 103 pp. (in Hebrew, English summary, unpubl.).
Schulman, N. 1967. Remarks on the Quaternary in the northern Jordan Valley. *Israel J. Earth Sci.* 16: 104–106.
Schulman, N. & Rosenthal, E. 1968. Neogene and Quaternary of the Marma Feiyad area, south of Bet She'an. *Israel J. Earth Sci.* 17: 54–62.

Schulz, E. 1987a. Holocene palaeoenvironments in the central and southern Sahara. In G. Matheis & H. Schandelmeier (eds), *Current research in African Earth sciences* 439–442. Rotterdam: A.A. Balkema.
Schulz, E. 1987b. Die Holozäne vegetation der zentralen Sahara (N-Mali, N-Niger, SW-Lybien). *Palaeoecol. Africa* 18: 143–161.
Schütt, H. & Ortal, R. 1993. A preliminary correlation between the Plio-Pleistocene malacofaunas of the Jordan Valley (Israel) and the Orontes Valley (Syria). *Zool. Middle East* 8: 69–111.
Schwabe, C.W. 1994. *Unmentionable cuisine*: 476 pp. Charlottesville: Univ. Press Virginia.
Schwarcz, H.P., Blackwell, B., Goldberg, P. & Marks, A.E. 1979. Uranium series dating of travertine from archaeological sites, Nahal Zin, Israel. *Nature* 277(5697): 558–560.
Schwarcz, H.P., Goldberg, P. & Blackwell, B. 1980. Uranium series dating of archaeological sites in Israel. *Israel J. Earth Sci.* 29: 157–165.
Scott, B. 1981. The Eurasian–Arabian and African continental margin from Iran to Greece. *J. Geol. Soc. London* 138: 719–733.
Sebai, A., Zumbo, V., Feraud, G., Bertrand, H., Hussain, A.G., Giannerini, G. & Camperdon, R. 1991. $^{40}Ar/^{39}Ar$ dating of alkaline and tholeiitic magmatism of Saudi Arabia related to the early Red Sea rifting. *Earth Planet. Sci. Let.* 104: 473–487.
Seetzen, U.J. 1810. Une courte description des pays autour du lac de Tibériade, du Jourdain et de la Mer Morte. *Annal. De Voyages Paris* 7: 137–191.
Segev, A. 1984. Lithostratigraphy and paleogeography of the marine Cambrian sequence in southern Israel and southwestern Jordan. *Israel J. Earth Sci.* 33: 26–33.
Segev, A., Steinitz, G., Platzner, I. & Starinsky, A. 1985. An Upper Devonian thermal event in southern Israel based on Rb–Sr and K–Ar ages of illites $<2\ \mu m$ from Cambrian rocks. Yotvata: Abst., Israel Geol. Soc. Ann. Mtg., 87–88 (in Hebrew).
Seismic Geocode Ltd. 1984. Report on the interpretation of a non-exclusive seismic survey in the Eastern Mediterranean – Offshore Israel. Swanley: Kent BR8 8JR, UK, Horizon House, Azalea Drive.
Şengör, A.M.C. & Yilmaz, Y. 1981. Tethyan evolution of Turkey. A plate tectonic approach. *Tectonophysics* 75: 181–241.
Shackleton, N.J. & Hall, M.A. 1984. Oxygen and carbon isotope stratigraphy of Deep Sea Drilling Project Hole 552A: Plio-Pleistocene glacial history. In D.G. Roberts & D. Schnitker (eds), *Initial reports of the Deep Sea Drilling Project*, Vol. 81: 599–609. Washington: US Govt. Print. Off.
Shackleton, N.J., Berger, A. & Peltier, W.A. 1990. An alternative astronomical calibration of the lower Pleistocene timescale based on ODP Site 677. *Trans. R. Soc. Edinburgh: Earth Sci.* 81: 251–261.
Shahar, J. 1994. The Syrian Arc system: an overview. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 112: 125–142.
Shahar, Y. 1969. Marine Pliocene (?) strata in the Nebi Mussa region. *Israel J. Earth Sci.* 18: 169–170 (Abst.).
Shahar, Y. 1973. The Hazeva Formation in the Oron-Ef'e syncline, Israel. *Israel J. Earth Sci.* 22: 31–49.
Shahar, Y., Reiss, Z. & Gerry, E. 1966. A new outcrop of marine Neogene in the Negev. *Israel J. Earth Sci.* 15(2): 82–84.
Shalem, N. 1937. Nota preliminare su alcuni nuovi giacimenti preistorici in Palestina. *Arch. Anthrop. Ethnol. Firenze* 67: 123–139.
Shalem, N. 1950. Attributed climatic changes in the Levant. *C.R. Inter. Geog. Cong. Lisbonne* 2: 593–649.

Shalem, N. 1956. Seismicity and Erythrean disturbances in the Levant. *Publ. BCIS, Trans. Sci. Ser. A* 19: 267–295.

Shaliv, G. 1991. Stages in the tectonic and volcanic history of the Neogene basin in the Lower Galilee and the valleys. Jerusalem: Geol. Surv. Israel Rep. GSI/11/91, 94 pp. (in Hebrew, English summary).

Shaliv, G., Hatzor, Y. & Mimran, Y. 1988. Sedimentary, structural and morphological development of NE Samaria, Bet She'an Valley and the Gilbo'a margins in the Neogene–Quaternary. En Boqeq: Abst., Israel Geol. Soc. Ann. Mtg., 102–103.

Shaliv, G., Mimran, Y. & Hatzor, Y. 1991. The sedimentary and structural history of the Bet She'an area and its regional implications. *Israel J. Earth Sci.* 40(1–4): 161–179.

Shamir, G. 1983. Structural problems along the Zin Line. MSc thesis, Dept. Geol., Hebrew Univ. Jerusalem, 89 pp. (in Hebrew, unpubl.).

Shamir, G. 1997. Control of fault structure on the seismo-mechanics of the Dead Sea transform. *Terra Nostra* 4/97 (no page numbers).

Shamir, G. & Feldman, L. 1997. Distribution of current microseismic activity in the Hula basin. Kefar Gil'adi: Abst., Israel Geol. Soc. Ann. Mtg., 103.

Shamir, G. and 13 collab. 1996. The Nov. 22, 1995, Nuweiba earthquake. Elat: Abst., Israel Geol. Soc. Ann. Mtg., 84.

Shamir, G., Bruner, I., Eyal, Y. & Marco, S. 2000. Location of the active strand of the Dead Sea Fault at the northern Dead Sea. Ma'alot: Abst., Israel Geol. Soc. Ann. Mtg., 111.

Shapira, A. 1979. Re-determined magnitude of earthquakes in the Afro-Eurasian junction. *Israel J. Earth Sci.* 29: 107–109.

Shapira, A. 1982. Detectability of regional networks: analysis of the Israel seismic networks. *Israel J. Earth Sci*. 41: 21–25.

Shapira, A. 1997. Seismicity of the Dead Sea basin. In T.M. Niemi, Z. Ben-Avraham & J.R. Gat (eds), *The Dead Sea, the lake and its setting*: 82–86. Oxford: Oxford University Press.

Shapira, A., Manor, O., Oman, S. & Sheshinsky, R. 1986. The empirical relation of earthquake epicenters to mapped faults in Israel. *Israel J. Earth Sci.* 35: 149–157.

Shapira, A., Avni, R. & Nur, A. 1993. A new estimate for the epicenter of the Jericho earthquake of July 11, 1927. *Israel J. Earth Sci.* 42: 93–96.

Sharief, F.A. & Moshrif, M.A. 1989. Chronostratigraphy and hydrocarbon prospects of the Sakaka Formation, Northern Arabia. *J. Petrol. Geol*. 12: 85–102.

Shata, A. 1959. Structural development of the Sinai peninsula (Egypt). Mexico: Proc. 20th Intl. Geol. Cong.: 225–250.

Shaw, S.H. 1948. The geology and mineral resources of Palestine. *Bull. Imper. Inst*. 46(1): 87–103.

Shulman, H. & Ben-Avraham, Z. 1999. The subsurface distribution of the Lisan and Mt. Sedom diapirs. En Boqeq: Abst., Israel Geol. Soc. Ann. Mtg., 85.

Siedner, G. & Horowitz, A. 1974. Radiometric ages of late Cainozoic basalts from northern Israel: chronostratigraphic implications. *Nature* 250: 23–26.

Smyth, W.W. 1868. Physical structure of Palestine. *Quart. J. Geol. Soc.* 24: LI–LVIII.

Sneh, A. 1967. The Hazeva Formation. Jerusalem: Geol. Surv. Israel Rep. OD/6/67: 7–10.

Sneh, A. 1981. The Hazeva Formation in the northern Arava, Israel. *Israel J. Earth Sci.* 30(2–3): 81–92.

Sneh, A. 1982. Quaternary of the northwestern Arava, Israel. *Israel J. Earth Sci.* 31: 9–16.

Sneh, A. 1988. Regional lithostratigraphy of the Eocene Avedat Group, Israel. Jerusalem: Geol. Surv. Israel Rep. GSI/26/88: 1–19.

Sneh, A. 1993. Stratigraphic position of marine Pliocene deposits in the Lower Galilee and the Yizre'el Valley. *Geol. Surv. Israel, Curr. Res.* 8: 74–75.

Sneh, A. 1996. The Dead Sea Rift: lateral displacement and downfaulting phases. *Tectonophysics* 263: 277–292.

Sneh, A. 1999. Cross-rift drainage system successions in the Hazeva Formation (Miocene), Dead Sea Rift area, Israel. *Israel J. Earth Sci.* 48: 247–256.

Sneh, A., Bartov, Y., Weissbrod, T. & Rosensaft, M. 1998a. The geological map of Israel, 1 : 200,000. Jerusalem: Israel Geol. Surv. 4 sheets.

Sneh, A., Lang, B. & Halicz, L. 1998b. The Sheikh Ali fault, NE Lake Kinneret, implications of new volcano-stratigraphic findings. *Geol. Surv. Israel, Curr. Res.* 11: 42–44.

Solomonica, P. 1948. Some stratigraphical inferences. *J. Palestine Orient. Soc.* 21: 97.

Starinsky, A., Bielski, M., Bonen, D. & Steinitz, G. 1980. Rb–Sr whole rock age of syenitic intrusions (Shen Ramon and Gavnunim) in the Ramon area, southern Israel. *Israel J. Earth Sci.* 29: 177–181.

Stein, M. & Hofmann, A.W. 1992. Fossil plume head beneath the Arabian lithosphere? *Earth Planet. Sci. Lett.* 114: 193–209.

Stein, M., Starinsky, A., Katz, A., Goldstein, S.L., Machlus, M. & Schramm, A. 1997. Sr isotopic, chemical and sedimentological evidence for the evolution of Lake Lisan and the Dead Sea. *Terra Nostra* 4/97 (no page numbers).

Steininger, F.F. & Rögl, F. 1996. Paleogeography and palynspastic reconstruction of the Neogene of the Mediterranean and Paratethys. In J.E. Dixon & A.H.F. Robertson (eds), *The geological evolution of the eastern Mediterranean* (2nd printing of 1984 edn): 659–668. London: Geol. Soc.

Steinitz, G. 1976. Paleogeography of the Menuha and Mishash formations in the eastern Ramon area. *Israel J. Earth Sci.* 25: 70–75.

Steinitz, G. 1977. Evaporite-chert associations in Senonian bedded cherts, Israel. *Israel J. Earth Sci.* 26: 55–63.

Steinitz, G. & Bartov, Y., 1991. The Miocene–Pleistocene history of the Dead Sea segment of the rift in light of the K–Ar ages of basalts. *Israel J. Earth Sci.* 40: 199–208.

Steinitz, G., Bartov, Y. & Hunziker, J.C. 1978. K–Ar ages of some Miocene–Pliocene basalts in Israel: their significance to the tectonics of the Rift Valley. *Geol. Mag.* 115(5): 329–340.

Steinitz, G., Bartov, Y., Eyal, M. & Eyal, Y. 1981. K–Ar age determination of Tertiary magmatism along the western margin of the Gulf of Elat. *Geol. Surv. Israel, Curr. Res.* 1980: 27–29.

Steinitz, G., Gvirtzman, G. & Lang, B. 1983. Evolution of K–Ar ages of the Asher Volcanics. *Geol. Surv. Israel, Curr. Res.* 1982: 34–38.

Steinitz, G., Calvo, R. & Bartov, Y. 2000. New $^{40}Ar/^{39}Ar$ age for the En Yahav dike and its implications for the age of the Hazeva Formation. *Geol. Surv. Israel, Curr. Res.* 12: 206–208.

Stekelis, M. 1935. Les monuments mégalithiques de Palestine. *Arch. Inst. Paléont. Humaine Paris* 15: 1–92.

Stekelis, M. 1940. Report on the excavations at Bethlehem. Jerusalem: Dept. Archaeol., Hebrew Univ., 12 pp. (unpubl.).

Stekelis, M. 1960. The Paleolithic deposits of Jisr Banat Yaqub. *Bull. Res. Coun. Israel* 9G(2–3): 61–87.

Stekelis, M. 1960/1961. La necrópolis megalítica de Ala-Safat, Transjordania. *Ampurias* XXII–XXIV: 49–128.

Stekelis, M. 1966a. *The Yarmoukian culture*: 65 pp., 76 Pl. Sha'ar Hagolan (Israel): Museum of Prehistory (in Hebrew).

Stekelis, M. 1966b. *Archaeological excavations at Ubeidiya*, 1960–1963*:* 32 pp., 37 Pl. Jerusalem: Israel Acad. Sci. Hum.

Stekelis, M. 1992. Bethlehem. In E. Stern (ed.), *The new Encyclopaedia of archaeological excavations in the Holy Land* (2nd edn): Vol. 1: 186. Jerusalem: Israel Explor. Soc. (in Hebrew).
Stekelis, M. & Bar-Yosef, O. 1965. Un habitat de Paléolitique supérieur à Ein Guev, (Israël) – note préliminaire. *L'Anthropologie* LXIX: 176–183.
Stekelis, M. & Gilead, D. 1966. Ma'ayan Baruch: an early Stone Age site at the upper Galilee. *MiTekufat Ha'Even* 8: 1–24 (in Hebrew).
Stekelis, M., Picard, L. & Bate, D.M.A. 1937. Jisr Banat-Yaqub. *Quart. Dept. Antiquities Palestine* 6(3–4): 214–215.
Stekelis, M., Picard, L., Schulman, N. & Haas, G. 1960. Villafranchian deposits near Ubeidiya in the central Jordan Valley (preliminary report). *Bull. Res. Coun. Israel* 9G(4): 175–183.
Stern, E. (ed.) 1992. *The new Encyclopaedia of archaeological excavations in the Holy Land* (2nd edn): 4 Vols, 1631 pp. Jerusalem: Israel Explor. Soc. (in Hebrew).
Suc, J.P. 1989. Distribution latitudinale et étagement des associations végétales au Cénozoique supérieur dans l'aire ouest-méditerranéenne. *Bull. Soc. Géol. France* 8V(3): 541–550.
Suc, J.P., Bertini, A., Leroy, S.A.G. & Suballyova, D. 1997. Towards the lowering of the Pliocene/Pleistocene boundary to the Gauss–Matuyama reversal. *Quat. Inter.* 40: 37–42.
Suess, E. 1892–1909. *Das Antlitz der Erde*: Vol. I, 1892: 778 pp., Vol. II, 1888: 703 pp., Vol. III, 1901: 508 pp., Vol. IV, 1909: 789 pp. Vienna: Tempsky.
Suzuki, H. & Takai, F. 1970. *The Amud Man and his cave site*: 439 pp. Tokyo: Univ. of Tokyo Press.
Swisher, C.C., Curtis, G.H., Jacob, T., Getty, A.G., Suprijo, A. & Widiamsmoro, 1994. Age of the earliest known hominids in Java, Indonesia. *Science* 263: 1118–1121.
Symonds, 1841. *Survey of Tiberias Sea and Jordan River.* London: British Admiralty.
TAHAL, 1969. Bathymetric map of Lake Kinneret, scale 1 : 50000, contour interval 1 m. Tel Aviv: TAHAL, Water Planning for Israel.
Tarawneh, K., Ilani, S., Rabba, I., Harlavan, Y., Peltz, S., Ibrahim, K., Weinberger, R. & Steinitz, G. 2000. Dating of the Harrat ash Shaam basalts, northeast Jordan (Phase 1). Jerusalem: Geol. Surv. Israel Rep. GSI/2/2000, 59 pp.
Tchernov, E. 1973. *On the Pleistocene mollusks of the Jordan Valley*: 46 pp. Jerusalem: Israel Acad. Sci. Hum.
Tchernov, E. 1975. *The early Pleistocene mollusks of Erk el Ahmar*: 36 pp. Jerusalem: Israel Acad. Sci. Hum.
Tchernov, E. 1979. Quaternary fauna. In A. Horowitz, *The Quaternary of Israel*: 260–292. New York: Academic Press.
Tchernov, E. 1980. The biostratigraphy of the Middle East. *Préhistoire du Levant* 598: 67–97.
Tchernov, E. 1987. The age of the Ubeidiya formation, an early Pleistocene hominid site in the Jordan Valley, Israel. *Israel J. Earth Sci.* 36(1–2): 3–30.
Tchernov, E. 1988. The paleobiogeographical history of the southern Levant. In Y. Yom-Tov & E. Tchernov (eds), *The zoogeography of Israel*: 159–250. Dordrecht: Junk.
Tchernov, E. 1996. Rodent faunas, chronostratigraphy and paleobiogeography of the southern Levant during the Quaternary. *Acta Zool. Cracov* 39(1): 513–530.
Tchernov, E., Ginsburg, L., Tassy, P. & Goldsmith, N.F. 1987. Miocene mammals of the Negev (Israel). *J. Vertebrate Paleontol.* 7(3): 284–310.
ten Brink, U.S. & Ben-Avraham, Z. 1989. The anatomy of a pull-apart basin: Seismic reflection observations of the Dead Sea basin. *Tectonics* 8: 333–350.

ten Brink, U.S., Schoenberg, N., Kovach, R.L. & Ben-Avraham, Z. 1990. Uplift and possible Moho offset across the Dead Sea transform. *Tectonophysics* 180: 71–86.

ten Brink, U.S., Ben-Avraham, Z., Bell, R.E., Hassouneh, M., Coleman, D.F., Andreasen, G., Tibor, G. & Coakley, B. 1993. Structure of the Dead Sea pull apart basin from gravity analyses. *J. Geophys. Res.* 98: 21, 877–21, 894.

ten Brink, U.S., Rybakov, M., Al-Zoubi, A., Hassouneh, M., Frieslander, U., Batanyneh, A., Goldschmidt, V., Daoud, M., Rotstein, Y. & Hall, J.K. 1999. The anatomy of the Dead Sea Transform: Does it reflect continuous changes in plate motion? *Geology* 27: 887–890.

Tobias, P.V. 1966. *A member of the genus Homo from Ubeidiya:* 12 pp., 3 Pl. Jerusalem: Israel Acad. Sci. Hum.

Traverse, A. 1988. *Paleopalynology*: 600 pp. Boston: Unwin Hyman.

Tristram, H.B. 1864. On the sulfur and bitumen deposits at the south-west corner of the Dead Sea. *Rep. Brit. Assoc.* 34: 73.

Tristram, H.B. 1865a. *The land of Israel*: 651 pp. London: Soc. Promot. Christian Knowl.

Tristram, H.B. 1865b. On the geology and physical features of the valley of the Jordan, the Dead Sea, and the adjacent districts. *Geol. Mag.* 2: 254–255.

Tristram, H.B. 1873. *The Land of Moab: travels and discoveries on the east side of the Dead Sea and the Jordan*: 416 pp. New York: Harper.

Tristram, H.B. 1884. *The fauna and flora of Palestine*: 455 pp. London: Palestine Explor. Fund.

Tsatskin, A., Weinstein-Evron, M. & Ronen, A. 1995. Weathering and pedogenesis of wind-blown sediments in the Mount Carmel caves, Israel. *Quat. Proc.* 4: 83–93.

Turcotte, T. & Arieh, E. 1988. Catalog of earthquakes in and around Israel. Tel Aviv: Israel Electric Corporation Ltd., Prelim. Safety Rep., Appendix 2.5A, 1–18, 6 tables.

Turville-Petre, F. 1927. Prehistoric Galilee. *Antiquity* 1: 299–310.

Udintsev, G.B., Zverev, A.S., Odinokov, Yu. Yu., Efremov, V.N., Eleftheriou, S. & Hall, J.K. 1994. Structure of the post-Messinian sedimentary cover in the eastern part of the Mediterranean Sea. In V.A. Krasheninnikov & J.K. Hall (eds), *Geological structure of the north-eastern Mediterranean*: 33–74. Jerusalem: Histor. Prod. Hall.

Vail, P.R. & Todd, R.G. 1981. Cause of northern North Sea Jurassic unconformities. *AAPG Bull.* 65: 1003.

Vail, P.R., Mitchum Jr., R.M., Todd, R.G., Widmier, J.M., Thompson, S., Sangree, J.B., Bubb, J.N. & Hatlelid, W.G. 1977. Seismic stratigraphy and global changes of sea level. *AAPG Mem.* 26: 49–212.

Valeton, I., Stutze, B. & Goldberg, R. 1983. Geochemical and mineralogical investigation of the lower Jurassic flint-clay bearing Mishor and Ardon formations, Makhtesh Ramon, Israel. *Sed. Geol.* 35: 105–152.

Valladas, H., Reyss, J.L., Joron, G., Valladas, G., Bar-Yosef, O. & Vandermeersch, B. 1988. Thermoluminescence dates for the Mousterian Proto-Cro-Magnons from Qafze Cave (Israel). *Nature* 331: 614–616.

van Eck, T. & Hofstetter, R. 1989. Microearthquake activity in the Dead Sea region. *Geophys. J. Int.* 99: 605–620.

van Eck, T. & Hofstetter, R. 1990. Fault geometry and spatial clustering of micro earthquakes along the Dead Sea–Jordan rift fault zone. *Tectonophysics* 180: 15–27.

van Zeist, W. & Bottema, S. 1988. Late Quaternary vegetational and climatic history of southwest Asia. *Proc. Indian Nat. Sci. Acad.* 54A(3): 461–480.

Van Zinderen Bakker, E.M. & Mercer, J.H. 1986. Major late Cainozoic climatic events and palaeoenvironmental changes in Africa viewed in a worldwide context. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 56: 217–235.

Vaumas, È. de. 1949. Sur la structure de la Galilée libanaise et de la dépression du Houlé. *C.R. Acad. Sci. Paris* 229(19): 943–946.
Vaumas, È. de. 1965. Phénomènes cryogéniques en Israël. *Rev. Géog. Phys. Géol. Dyn.* (2), VII(4): 295–309.
Velitzelos, E. & Gregor, H.J. 1987. Preliminary correlation of Oligocene to Pleistocene phyto-stratigraphic units of the Mediterranean and the Paratethys area. *Ann. Inst. Geol. Publ. Hung*. LXX: 79–83.
Verosub, K. & Tchernov, E. 1991. Resultats préliminaires de l'étude magneto-stratigraphique d'une sequence sédimentaire a industrie humaine en Israël. In E. Bonifay & B. Vandermeersch (eds), *Les premiers Européens*: 237–242. Paris: Comité des Travaux Historiques et Scientifiques.
Verosub, K., Goren-Inbar, N., Feibel, C.S. & Saragusti, I. 1998. Location of the Matuyama/Brunhes boundary in the Gesher Benot Ya'aqov archaeological site, Israel. Seattle: Abst. Paleoanthrpol. Soc. Mtg., A22.
Vignes, L. de, 1864. Éxtraits des notes d'un voyage d'exploration à la Mer Morte, dans le Wady Arabah, sur la rive gauche du Jourdain et dans le désert de Palmyre. *Nouv. Annal. Voyages* 4: 370–372.
Vita-Finzi, C. 1964. Observations on the late Quaternary of Jordan. *Palestine Expl. Quart.* 36: 19–33.
Vita-Finzi, C. 1966. The Hasa Formation: an alluvial deposition in Jordan. *Man* 1(3): 384–390.
Volney, F.C. de. 1825. Voyages en Égypte et en Syrie pendant les années 1783, 1784 & 1785. In *État Physique de la Syrie*, Vol. I, Pt. 2: 274–310. Paris: Parmentier.
Voznesensky, V. & Shimshilashvili, E. 1999. Pliocene Pleshet Formation in the District Ramot, Be'er Sheva. En Boqeq: Abst., Israel Geol. Soc. Ann. Mtg., 91.
Vroman, A.J. 1958. The geological map of Israel, 1 : 50,000, Ser. 1: Galilee. Jerusalem: Geol. Surv. Israel, 6 maps.
Vroman, A.J. 1973. Is a compromise between the theories of tension and of shear for the origin of the Jordan–Dead Sea trend possible? *Israel J. Earth Sci.* 22: 141–156.
Vroman, A.J. 1981. Recent discussion about the north–south bias of the major rift valleys. *Tectonophysics* 80: 271–283.
Wade, A. 1921. Oil prospects in Palestine. *Oil News* (June 11): 584–586.
Waisel, Y. (ed.) 1984. *Vegetation of Israel*: 262 pp. Tel Aviv: Ministry of Defense/The Publishing House, Society for Protection of Nature (in Hebrew).
Walcott, S. 1868. The site of Sodom. *Bibliotheca Sacra* 25: 112–151.
Walker, T.R. 1967. Formation of red beds in modern and ancient deserts. *Geol. Soc. Am. Bull*. 78: 353–368.
Walley, C.D. 1998. Some outstanding issues in the geology of Lebanon and their importance in the tectonic evolution of the Levantine region. *Tectonophysics* 298: 37–62.
Wdowinski, S., Bechor, N. & Na'aman, S. 2000. Current plate motion and crustal deformation across the Dead Sea fault. Dead Sea: Abst., The first Stephan Mueller Conf. Eur. Geophys. Soc., 45.
Weber, H. 1963. Ergebnisse erdölgeologischer Aufschlussarbeiten der DEA in Nordost Syrien. 1. Schichtfolge, Fazies und Tectonik in der haute Djesireh. *Erdöl u. Kohle* 16: 669–682.
Wegener, A. 1922. Die Entstehung der Kontinente und Ozeane. (3rd edn). Braunschweig: Vieweg u. Sohn.
Weiler, Y. 1961. The geology of Nazareth and Tavor mountains. MSc thesis, Dept. Geol., Hebrew Univ. Jerusalem, 33 pp. (in Hebrew, unpubl.).

Weiler, Y. & Sass, E. 1972. Karstic sandstone bodies in the Turonian limestones of Judea, Israel. *Sedimentology* 7: 137–152.
Weinberger, G. 1991. A sedimentological study of the Plio-Pleistocene QI Palynozone of Amiaz-1, Sedom-2, Melekh-Sedom-1 and Amatzia-1 boreholes. Tel Aviv: Rep. Submitted to the Israel National Oil Co., 49 pp. (unpubl.).
Weiner, D., Bruner, I., Goldman, M., Calvo, R. & Bartov, Y. 1999. Subsurface imaging and interpretation of the Mishor Yamin area using multi-geophysical methods. Dead Sea: Abst., Israel Geol. Soc. Ann. Mtg., 94.
Weinstein, M. 1976. The late Pleistocene vegetation and climate of the northern Golan. *Pollen et Spores* XVII(4): 553–562.
Weinstein, M. 1979. Airborne pollen. In Horowitz, A, *The Quaternary of Israel*: 180–185. New York: Academic Press.
Weinstein, M. 1982. The paleoecology of the early Würm in the Hula basin. PhD dissertation, Dept. Archaeol., Tel Aviv Univ., 105 pp. (in Hebrew, English summary, unpubl.).
Weinstein-Evron, M. 1983. The paleoecology of the early Würm in the Hula basin, Israel. *Paléorient* 9(1): 5–19.
Weinstein-Evron, M. 1986. Pollen spectra from the Acheulean site of Mitzpeh Yiron, Israel: a cautionary tale. *Pollen et Spores* XXVIII(2): 157–166.
Weinstein-Evron, M. 1987. Palynology of Pleistocene travertines from the Arava Valley, Israel. *Quat. Res*. 27: 82–88.
Weinstein-Evron, M. 1988. A Middle Palaeolithic sequence from the Hula Valley, Israel. *L'Homme de Néanderthal* 2: 207–222.
Weinstein-Evron, M. 1989. Palynological history of the last Pleniglacial in the Levant. *ERAUL* 42: 9–25.
Weinstein-Evron, M. 1990. Palynological history of the last pleniglacial in the Levant. In J.K. Kozlowski (ed.), *Les industries à pointes foliacées en Europe*: 9–25. Liège: *ERAUL* 42.
Weinstein-Evron, M. 1993. Paleoecological reconstruction of the Upper Paleolithic in the Levant. In J. Pavúk (ed.), *Actes du XIIe Congrès International des Sciences Préhistoriques et Protohistoriques:* 259–270. Bratislava.
Weinstein-Evron, M. 1998. *Early Natufian el-Wad revisited*: 255 pp. Liège: *ERAUL* 77.
Weissbrod, T. 1969a. The Paleozoic of Israel and adjacent countries. Part I – The Paleozoic stratigraphy of southern Israel. *Bull. Geol. Surv. Israel* 47: 1–35.
Weissbrod, T. 1969b. The Paleozoic of Israel and adjacent countries. Part II – The Paleozoic outcrops in southwestern Sinai and their correlation with those of southern Israel. *Bull. Geol. Surv. Israel* 48: 1–32.
Weissbrod, T. 1976. The Permian in the Near East: In H. Falke (ed.), *The Continental Permian in Central, West and South Europe*: 200–214. Dordrecht: Reidel.
Weissbrod, T. 1981. The Paleozoic of Israel and adjacent countries (lithostratigraphic study). Jerusalem: Geol. Surv. Israel Rep. MP/600/81, 276 pp. (in Hebrew, English summary).
Weissbrod, T. & Sneh, A. 1983. Marine intercalations within the Netafim Formation in Southeastern Sinai. Nazerat: Abst., Israel Geol. Soc. Ann. Mtg., 84.
Wellings, F.E. 1938. In B. Willis (ed.), Wellings' observations on the Dead Sea Structure. *Geol. Soc. Am. Bull.* 49: 659–668.
Werker, E. & Goren-Inbar, M. in press. Reconstruction of the woody vegetation at the Acheulian site of Gesher Benot Ya'aqov, Dead Sea Rift, Israel. In B. Purdy (ed.), *The environmental and cultural heritage of wetlands*. Gainesville: Univ. of Florida Press.
West, R.G. 1969. *Pleistocene geology and biology*: 377 pp. London: Longmans.

Wetzel, R. & Morton, D.M. 1959. Contribution a la géologie de la Tranjordanie. *Notes et Mém. Moyen Orient* VII: 95–191.

Wetzel, R., Morton, D.M. & Nasr, S.N. 1947. *Stratigraphic sections of Jordan Valley and Dead Sea, Pt. 1*: 37 pp. Amman: Min. Nat. Econ.

Wiegand, T. 1920. *Sinai*: Vol. 1, 145 pp. Berlin & Leipzig: Wissenschaftl. Veröffentl. Dtsh. Türk. Denkmalschutz-Kommando.

Wilson, J.T. 1965. A new class of faults and their bearing on continental drift. *Nature* 207: 343–347.

Williams, M.A.J. & Faure, H. (eds) 1980. *The Sahara and the Nile*: 451 pp. Rotterdam: A.A. Balkema.

Willis, B. 1928. Dead Sea problem: rift valley or ramp valley? *Bull. Geol. Soc. Am.* 39: 490–542.

Willis, B. & Picard, L. 1932. The Jordan Valley and Judean highlands. *Geol. Mag.* 69: 97–107.

Wolfart, R. 1966. Zur Geologie und Hydrogeologie von Syrien, unter besonderer Berücksichtigung der süd- und nordwestlichen Landestiele. *Beih. Geol. Jb.* 68: 1–129.

Wolfart, R. 1967. *Geologie von Syrien und dem Lebanon*: 326 pp. Berlin: Gebruder Bornträger.

Woolley, C.L. & Lawrence, T.E. 1914/1915. The wilderness of Zinn. *Ann. Palest. Explor. Fund 3*, I–XIII: 1–154.

Wyllie, B.K.N. 1931. The geology of Jebel Usdum, Dead Sea. *Geol. Mag.* 68: 366–372.

Wyllie, B.K.N., Campbell, K.A. & Lees, G.M. 1922. Preliminary notes on Jebel Usdum, Dead Sea area, Palestine. Istanbul: Turkish Petroleum Co., Geol. Rep., 16 pp.

Ya'alon, D.H. 1991. The earliest geological map of the Near East – 1751. *Israel J. Earth Sci.* 40: 251–254.

Yechieli, Y. 1987. The geology of the northern Arava rift and the Mahmal anticline, Hazeva area. Jerusalem: Geol. Surv. Israel Rep. GSI/30/87, 98 pp. (in Hebrew, English summary).

Yechieli, Y., Magaritz, M., Levy, Y., Weber, U., Kafri, U., Woelfli, W. & Bonani, G. 1993. Late Quaternary geological history of the Dead Sea area, Israel. *Quat. Res.* 39: 59–67.

Yilmaz, Y. 2001. Geological and tectonic development of the western Anatolia. Isparta: Abst., 3rd Intl. Symp. Geol. East. Medit. Region, 8.

Yom-Tov, Y. & Mendelssohn, H. 1988. Changes in the distribution and abundance of vertebrates in Israel during the 20th century. In Y. Yom-Tov & E. Tchernov (eds), *The zoogeography of Israel*: 515–548. Dordrecht: Junk.

Yom-Tov, Y & Tchernov, E. 1988. *The zoogeography of Israel:* 600 pp. Dordrecht: Junk.

Zak, I. 1967. The geology of Mount Sedom. PhD dissertation, Dept. Geol., Hebrew Univ. Jerusalem, 207 pp. (in Hebrew, English summary, unpubl.).

Zak, I. & Freund, R. 1966. Recent strike-slip movements along the Dead Sea Rift. *Israel J. Earth Sci*. 15: 13–37.

Zak, I. & Freund, R. 1981. Asymmetry and basin migration in the Dead Sea Rift. *Tectonophysics* 80: 27–38.

Zhang, H., Niemi, T.M., Atallah, M. & Harrison, J.B. 1999. Slip rate of the northern Wadi Araba fault, Dead Sea transform, Jordan, during the last 12,000 years. GSA Abst. Programs A-114.

Zilberfarb, A. 1987. Evidence for the Precambrian existence of the Gulf of Eilat fault zone. Zikhron Ya'aqov: Abst., Israel Geol. Soc. Ann. Mtg., 44–46.

Zilberman, E. 1983. The geology of the central Sinai-Negev Shear Zone, central Negev. Part B: The Arif-Batur lineament. Jerusalem: Geol. Surv. Israel Rep. EG/6/83, 58 pp.

the Hazeva Formation (Israel) and the paleogeography of Miocene Africa. *Newsl. Stratigr*. 20(2): 73–90.

Goren, H. 1999. *Go view the Land (German study of Palestine in the nineteenth century)*: 362 pp. Jerusalem: Yad Izhak Ben-Zvi (in Hebrew).

Goren, H. *"Zieht aus und erforscht das Land": Die deutsche palästina-Forschung in 19. Jahrhundert.* Tel Aviv: Tel Aviv Univ., Inst. German History (in press).

Goren-Inbar, N. 1985. The lithic assemblage of the Berekhat Ram Acheulian site, Golan Heights. *Paléorient* 11(1): 7–28.

Goren-Inbar, N. 1995. The Lower Paleolithic of Israel. In T.E. Levy (ed.), *The Archaeology of Society in the Holy Land*: 93–109. London: Leicester Univ. Press.

Goren-Inbar, N. 1998. Gesher Benot Ya'aqov: the Acheulian cultural sequence. Seattle: Abst. Paleoanthrpol. Soc. Mtg., A8.

Gregor, H.J. & Velitzelos, E. 1987. Evolution of Neogene Mediterranean vegetation and the question of a dry upper Miocene period (Salinity Crisis). *Ann. Inst. Geol. Publ. Hung*. LXX: 489–496.

Gregory, J.W. 1921. *The rift valleys and geology of East Africa:* 479 pp. London: Seeley, Service & Co.

Grossman, P. 1983. Notes on the structural evolution and depositional basins of the lower Jurassic sedimentation – Israel. Nazerat: Abst., Israel Geol. Soc. Ann. Mtg., 29.

Grossman, R., 1984. Regional density analysis of Israel. Holon: Inst. Petrol. Res. Geophys. Rep. 84/44, 3 pp., 1 map, 37 Figs.

Grün, R., Stringer, C. & Schwarcz, H.P. 1991. ESR dating of teeth from Garrod's Tabun Cave collection. *J. Human Evolution* 20: 231–248.

Guettard, J.E. 1751. Mèmoir et carte mineralogique sur la nature et la situation des terreins qui transversent la France et l'Angleterre. *Mém. Acad. R. Sci.* 1746: 363–392.

Guillaume, A. & Guillaume, S. 1982. L'èrosion dans les Alpes au Plio-Quaternaire et au Miocéne. *Eclogae Geol. Helv*. 75/2: 247–268.

Günther, A. 1864. Report on a collection of reptiles and fishes from Palestine. *Proc. Zool. Soc. London* 11: 488–493.

Gushchenko, I.I. 1979. Volcanic eruptions of the world. Catalogue. Izverzheniya Vulkanov Mira. Katalog. Moscow: Izd. Nauka, 474 pp. (in Russian).

Guy, P.L.O. 1924. Prehistoric and other remains in the Huleh basin. *Bull. Brit. Sch. Archaeol. Jerusalem* 4–6: 74–77.

Gvirtzman, G. 1970. The Saqiye group in the coastal plain and HaShefela regions, Israel. PhD dissertation, Dept. Geol., Hebrew Univ. Jerusalem, 170 pp. (in Hebrew, English summary, unpubl.).

Gvirtzman, G. 1979. The geology and geomorphology of Biq'at Be'er Sheva. In Y. Grados, E. Stern & G. Rivlin (eds), *The Be'er Sheva Book*: 333–354. Be'er Sheva: Ben Gurion Institute (in Hebrew).

Gvirtzman, G. 1981. The findings in the Ga'ash-2 well and their implication for oil exploration in NW Israel. Jerusalem: Abst. Symp. Oil Explor. Israel: 29–31.

Gvirtzman, G. & Buchbinder, B. 1978. The late Tertiary of the coastal plain and continental shelf of Israel and its bearing on the history of the eastern Mediterranean. *Init. Rep. Deep Sea Drill Proj., Leg* 42: 1195–1222.

Gvirtzman, G. & Steinitz, G. 1983. The Asher Volcanics – an early Jurassic event in northern *Israel. Geol. Surv. Israel, Curr. Res.* 1982: 28–33.

Gvirtzman, G. & Weissbrod, T. 1984. The Hercynian geanticline of Helez and the Late Paleozoic history of the Levant. In J.E. Dixon & A.H.F. Robertson (eds), *The geological evolution of the eastern Mediterranean*: 177–186. London: Geol. Soc.

Zilberman, E. 1985. The geology of the central Sinai-Negev Shear Zone, central Negev. Part C: The Paran lineament. Jerusalem: Geol. Surv. Israel Rep. GSI/38/85, 53 pp.

Zilberman, E. 1992. Remnants of Miocene landscape in the central and northern Negev and their paleogeographic implications. *Bull. Geol. Surv. Israel* 83: 1–54.

Zohary, M. 1959. *Geobotany*: 589 pp. Merhavia (Israel): Sifriat Poalim (in Hebrew).

Zohary, M. 1973. *Geobotanical foundations of the Middle East*: 2 Vols, 739 pp. Stuttgart: Gustav Fischer.

Zumoffen, G. 1897. L'âge de la pierre en Phénicie. *L'Anthropologie* 8: 272–283; 426–438.