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The structure of the Dead Sea basin

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Abstract

The Dead Sea basin is located along the left-lateral transform boundary between the Arabian and Sinai plates. Its structure and history are known from surface geology, drilling, seismic reflection and other geophysical data. The basin comprises a large pull-apart, almost 150 km long and mostly 8–10 km wide, which is flanked by a few kilometres wide zones of normal faulting. The basin formed at about 15 Ma or earlier, close to the beginning of the transform motion, and it reached about half its present length before the end of the Miocene. A strong negative gravity anomaly records a thick sediment basin fill: >5 km under half its length, reaching a maximum of ≥10 km. The fill includes a few km of salt (ca. 6–4 Ma) which forms several diapirs. At any one time large parts of the basin subsided simultaneously, but the site of fastest subsidence seems to have shifted northward. Sedimentation rates reached at least hundreds of metres per million years or more in the Miocene, and ≥1 km/Myr in later periods.

The basin structure is dominated by longitudinal faults: intrabasinal faults which delimit the pull-apart and which are the extensions of the major strike-slip faults north and south of the basin, and normal faults which extend along the basin margins. The latter faults express a small component of extension across the basin, whereas the pull-apart resulted from the much larger lateral motion along the basin. In addition, transverse faults divide the pull-apart into several segments with somewhat different histories. The pull-apart grew by becoming longer parallel to the transform motion. At shallow levels this was probably achieved by normal slip on transverse listric faults while the fill between them was little deformed. The crust under the basin was stretched and thinned during basin lengthening, which caused its subsidence. Basin formation was accompanied by uplifting of its flanks by ≥1 km. Sparse igneous activity occurred along the basin and its flanks. Its presence suggests a thermal anomaly in the lower lithosphere beneath the basin and adjacent parts of the transform.

Keywords: pull apart; transform; Dead Sea

1. Introduction

The Dead Sea, with a water level at about 400 m below sea, is situated within a large pull-apart that formed along the intra-continental Dead Sea transform (sometimes also called rift). Being almost 150 km long, this is one of the largest pull-aparts known on earth. Understanding of its internal structure and history can contribute to the understanding of large pull-aparts in general.

The geology of the Dead Sea basin and its regional setting have been intensively studied for over 150 years and were treated in numerous papers. The purpose of this work is to present an updated summary of the tectonics and geophysical features of this basin. First we briefly review the Dead Sea transform

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in order to outline the regional tectonic framework. Then the structural, geophysical and stratigraphic data bearing on its architecture and history are summarized. Finally we discuss the basin's origin, manner of growth and deep structure, and how they relate to the transform motion.

2. The Dead Sea transform

The Dead Sea transform, some 1000 km long, forms the boundary between the Arabian plate and the Sinai sub-plate which is an appendage of the African plate (Fig. 1; Quennell, 1958, 1959; Freund, 1965; Wilson, 1965; Freund et al., 1970; McKenzie et al., 1970). It joins the divergent plate boundary along the Red Sea with the zone of plate convergence along the Alpine orogenic belt in Turkey. The transform formed as a result of the mid-Cenozoic breakaway of Arabia from Africa, which until then were parts of a continuous continent.

2.1. Regional setting (Fig. 1)

The transform crosses a continental area that was consolidated during the Late Proterozoic Pan-African Orogeny. During most of the Phanerozoic this area behaved as a rather stable platform and it was covered by a few kilometres of mostly marine sediments (Picard, 1943, 1959; Bender, 1968, 1974; Garfunkel, 1988 and references therein). The platformal history was punctuated by a few phases of tectonic and magmatic activity. Most important were rifting events in Permian(?), Triassic and Early Jurassic times which were related to the formation of the Eastern Mediterranean branch of the Neo-Tethys and shaped its passive margins. The crust on both sides of the Dead Sea basin and the nearby parts of the transform is 30–35 km thick, but it thins considerably on approaching the margin of the Eastern Mediterranean basin (Ginzburg and Folkman, 1980; Ginzburg et al., 1981; El-Isa et al., 1987).

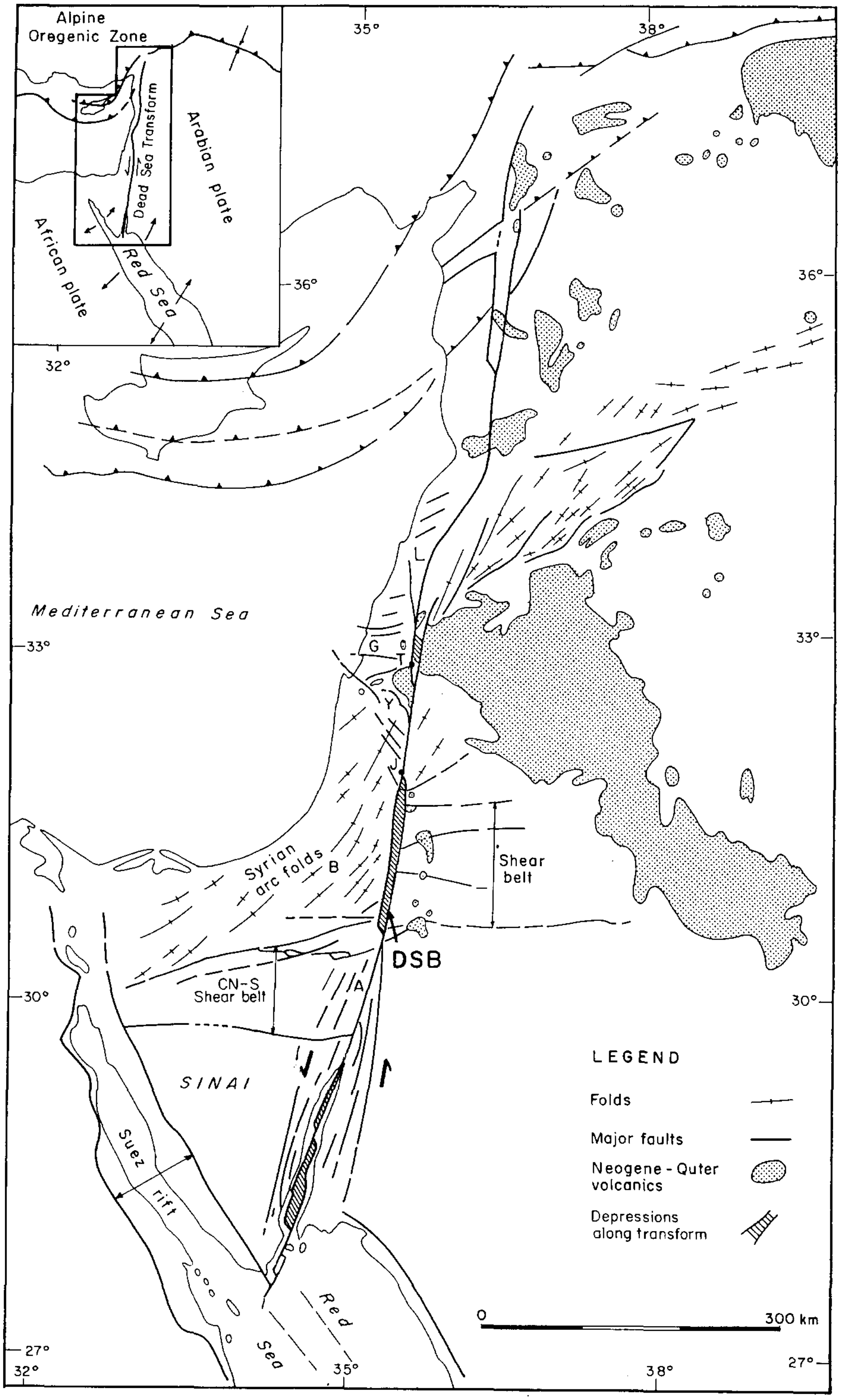
The closure of the neighbouring part of Neo-Tethys began in the Late Cretaceous, when marine sediments were still being deposited over the area, and was echoed by mild compressional deformation (Syrian arc phase). This produced a bundle of NNE–SSW- to ENE–SSW-trending folds, and a group of lineaments trending close to east–west along which right-lateral shearing of up to a few km took place (the central Negev–Sinai shear belt). The deformation continued into the Miocene.

2.2. The displacement and history of the transform

The young continental breakup phase, beginning at about 30–25 Ma, led to the detachment of Arabia from Africa, which became two distinct plates. Their separation created the Red Sea where incipient seafloor spreading takes place (Coleman, 1984). North of the Red Sea the Arabian–African plate motion was mostly taken up by the Dead Sea transform, but a part of the motion was accommodated by opening of the Suez rift (Freund, 1965; McKenzie et al., 1970; Joffe and Garfunkel, 1987 and references therein; LePichon and Gaulier, 1988). The continuing activity of both lines is evidenced by faulting of young sediments and by ongoing seismicity.

The Dead Sea transform trends at a large angle to the Red Sea axis, so it is quite close to the direction of plate motion (Fig. 1). The amount of motion — some 105 km left laterally — is obtained by matching the platformal sedimentary cover and some units of the basement complex across the transform and south of about 34°N (Quennell, 1959; Freund et al., 1970; Bandel and Khouri, 1981). The most accurate value is obtained by matching the lineaments of the central Negev–Sinai shear belt (Fig. 1; Quennell, 1959; Bartov, 1974). An independent estimate of the motion, about 100 km close to north–south, is obtained from the regional plate kinematics based on data from only the Gulf of Aden and the Red Sea. Though the amount and direction of motion obtained in this way are not very accurate, this result proves that the offset inferred from the local geology is consistent with the data on the regional plate kinematics.

Fig. 1. Regional setting of the Dead Sea basin. Inset show the present plate configuration. *DSB* = Dead Sea basin. Other place names: *A* = Arava Valley; *B* = Beer Sheva Valley; *G* = Galilee; *J* = Jericho; *L* = Lebanon; *T* = Tiberias; *Y* = Yizreel (Esdraelon) Valley. *CN–S* = Central Negev–Sinai (shear zone).


Alpine
Oregenic Zone
Arabian plate
Dead Sea Transform
African plate
Red Sea
35°
38°
36°
32°
33°
30°
27°
Mediterranean Sea
G
T
Y
J
L
Syrian arc folds
B
Shear belt
DSB
CN-S
Shear belt
A
SINAI
Suez rift
Red Sea
LEGEND
Folds
Major faults
Neogene - Quter volcanics
Depressions along transform
0
300 km

The history of the transform motion is not well constrained. The youngest rock bodies known to be affected by the entire transform offset are 25–20 Ma old dikes, which indicates that the transform motion began later than at 20 Ma (Eyal et al., 1981; Steinitz et al., 1981). The link between the transform motion and the opening of the Red Sea provides an additional constraint (Joffe and Garfunkel, 1987 and references therein). Magnetic anomalies in the southern Red Sea record an opening of 75 km since 5 Ma, which is only a fraction of the total opening, showing that most of the opening occurred earlier. This implies that the coeval transform motion did not exceed some 40 km. Thus, here too most of the offset must have occurred earlier than 5 Ma. At a constant slip rate the entire offset could have been achieved since 14–12 Ma (Middle Miocene). However, still older igneous activity and local subsidence along the transform — near Tiberias and along the Dead Sea basin — suggest that the transform originated earlier, perhaps at 18 Ma (Early Miocene). In this case the initial transform motion was coeval with right-lateral shearing along the central Negev–Sinai shear belt, which continued into Middle Miocene times. This possibility needs further study.

2.3. Internal structure of the transform

The structure and physiography vary considerably along the transform, because its trace deviates slightly, but by variable amounts, from a small circle about the Euler pole of the Arabian–Sinai plate motion. Therefore, as the transform moves, its two sides converge in some places and diverge in others (Quennell, 1958, 1959; Garfunkel, 1981). In Lebanon the transform trace, which is expressed by a major fault zone that crosses a high terrain, bends to the right, so this is a compressional segment (Figs. 1 and 2). Farther south minor plate separation dominates, so the southern half of the transform is largely leaky. It is almost entirely expressed by a valley whose floor is mostly below sea level — the transform valley. This valley is usually 5–15 km wide and is generally delimited by narrow zones of normal faults with total vertical throws varying from about 1 km to perhaps ≥ 10 km. Though this superficially resembles an extensional rift valley, the transform valley has a markedly different internal structure and is much narrower than extensional rifts.

The internal structure of the transform valley is dominated by left-stepping en-echelon strike-slip

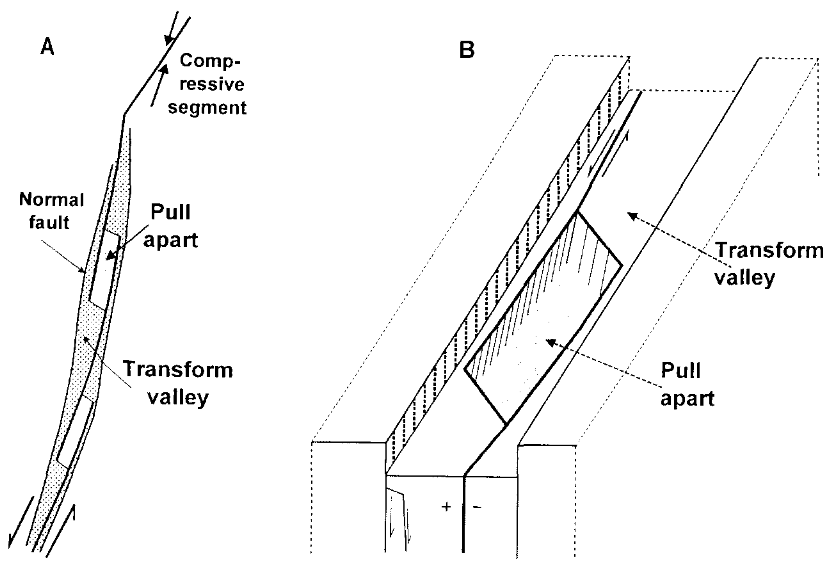


Fig. 2. Sketch showing the basic structural relations along the Dead Sea basin (see text for discussion). (A) The structural pattern in plan view along the part of the transform enclosing the Dead Sea basin. The main strike-slip faults trend at an angle to the transform valley. (B) Schematic block diagram showing the architecture of the Dead Sea basin. As lateral motion continues the edges of the basin move apart and it becomes longer parallel to the transform.

faults (Fig. 2; Quennell, 1958, 1959; Freund, 1965; Garfunkel et al., 1981; Garfunkel, 1981). This fault arrangement produced several pull-aparts (rhomb-grabens) which form deep basins. The largest ones are the Dead Sea basin and the basins beneath the Gulf of Elat (Aqaba) (Quennell, 1958; Ben-Avraham et al., 1979; Garfunkel, 1981; Ben-Avraham, 1985). The pull-aparts are bordered by extensions of the major strike-slip faults, but the lateral motion along these fault segments is variable and in places it is very small (see below). As transform motion continues, the pull-aparts become longer in the direction of motion, i.e., along the transform, and their area increases (Fig. 2B). Thus they are a sort of mini-spreading centres. The pull-aparts alternate with structural saddles along which the valley narrows or becomes indistinct, and there compressional structures are often present.

Because of the en-echelon arrangement of the strike-slip faults, their trends deviate from the overall trend of the transform (Fig. 2). Therefore, motion along these faults leads to some separation between the transform margins, which is augmented by the normal faulting along the transform valley margins. As a result, minor transverse separation (less than 5% of the lateral motion) took place along much the southern half of the transform and it seems to have increased with time (Garfunkel, 1981).

During the development of the transform its margins were deformed to varying degrees. Deformation was strongest in the Galilee and farther north, where it extended to distances of 50–100 km and more from the transform. This was probably related to slip along the curved part of the transform in Lebanon. The flanks of the Dead Sea basin are little deformed, however.

2.4. *Uplift and magmatism*

The formation of the Dead Sea transform, as well as of the other young plate boundaries in the region, was accompanied by regional uplifting and widespread igneous activity. The areas bordering the transform were uplifted by about 1 km and in places considerably more. The shape of the uplifts varies along and across the transform. Prior to the continental breakup, the region crossed by the transform was below sea level until about 40 Ma, and 15–12 Ma (Middle Miocene) it was still low enough for the sea to briefly extend 20–30 km inland of the present coast (now they are up to ca. 500 m above sea level). Thus, the transform flanks were uplifted mostly after 10 Ma ago, i.e., while the transform was active.

Igneous activity accompanying the transform development produced mainly alkali olivine basalts, basanites and nephelinites (Garfunkel, 1989 and references therein). On a regional scale the sites of most voluminous and areally widespread volcanism are not obviously related to the transform, but are situated mostly east of it, up to a distance of a few hundred kilometres. However, some igneous activity was aligned along the northern 2/3 of the transform, as far south as the Dead Sea basin. Most of it is observed on the transform shoulders, but in places magmas preferentially ascended along the transform itself.

3. The Dead Sea basin

The Dead Sea basin, about 150 km long, extends from the southern Jordan valley near Jericho to the central Arava valley (31°50′N to 30°20′N, Figs. 1 and 3). The area east of the basin forms a plateau 1.0–1.4 km above sea level which descends gradually away from the transform. The area on the west forms a broad arch whose crest reaches only 0.6–0.8, rarely 1.0 km, above sea level. The basin is delimited by fault scarps trending close to north–south, but the western fault scarp becomes indistinct along the northern Arava valley, so there the western border of the basin is physiographically indistinct. The northern and southern ends of the Dead Sea basin do not have any topographic expression, but there the transform valley is relatively narrow. Several accentuated slopes cross the basin floor obliquely and divide it into distinct parts.

The northern part of the basin is largely covered by the Dead Sea which is one of the most saline lakes in the world. Its waters contain >30% of dissolved salts, mainly Mg, Na and Ca chlorides, as well as high concentrations of K and Br which are exploited commercially. North and south of the lake the valley floor is 300–400 m below sea level, whereas about half of the lake floor is a flat rectangular area more than 700 m below sea level (water depth >300 m) (Neev and Hall, 1979). These two topographic levels

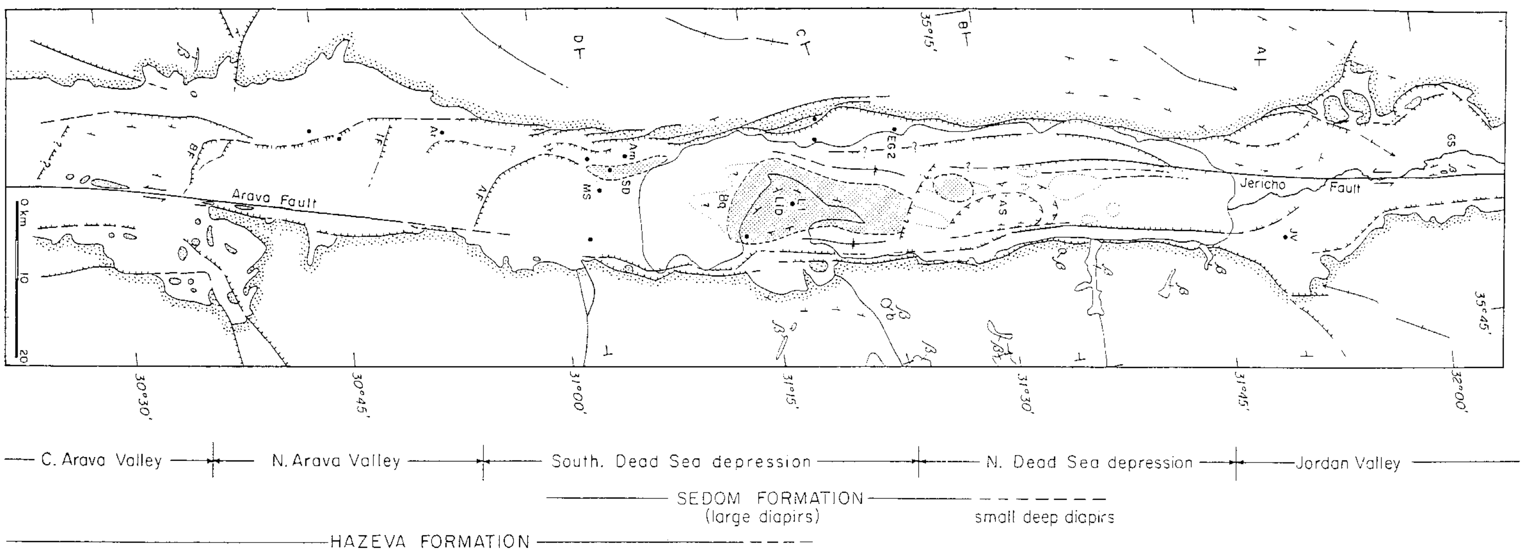


Fig. 3. Map showing the main features of the Dead Sea basin, compiled from different sources (after Garfunkel and Ben-Avraham, 1994). Edge of outcrops of pre-basin rocks is shown by irregular stippling. Diapirs expressed in the topography are shown by stippling; extent of diapirs not affecting the surface are shown by fine points. β = Neogene to Recent basalts; large dots = main boreholes. Abbreviations: *AF* = Amazyahu fault; *Am* = Amiaz well; *Ar* = Arava Well; *AS* = Arnon sink; *Bq* = Boqeq faults; *BF* = Buweirida fault; *EG 2* = En Gedi 2 well; *GS* = Grain Sabt dome; *IF* = Iddan fault; *JV* = Jordan Valley well; *L1* = Lisan 1 well; *LiD* = Lisan diapir; *MS* = Melekh Sedom well; *SD* = Sedom diapir.

are separated by accentuated slopes which extend across the basin. The flat lake floor is flanked on the west and east by 4–5 km and ≤ 2 km wide slopes with average gradients of 7° and 30°, respectively, which separate it from the faults scarps along the basin borders.

South of the lake the valley floor is mostly 400–300 m below sea level, except for two >100 m high hilly areas which formed over the Lisan and Sedom salt diapirs (Fig. 3). A few decades ago the Dead Sea waters extended over this area (which was therefore called the 'southern basin of the Dead Sea') to a distance of more than 20 km south of the Lisan hill which then formed a peninsula. This shallow portion of the lake mostly dried out because of a >10 m drop in the water level and because of human activity. This part of the basin is delimited in the south by the more than 50 m high scarp of the Amatzyahu fault which crosses obliquely the transform valley. Further south the Dead Sea basin extends along the Arava valley whose floor, somewhat dissected by erosion, rises gradually southward from some 300 m below sea level to about 100 m above sea level where the Dead Sea basin tapers out (south of 30°30′N).

4. Structure: geological and geophysical data

Surface geology, boreholes and seismic reflection data allow us to outline the principal structural features of the Dead Sea basin down to depths of several kilometres. This information is supplemented by gravity, magnetic and seismic refraction data which provide important constraints on the deeper structure. The main features are outlined in Figs. 3 and 4, but many details are still incompletely known.

4.1. Overall basin architecture

In upper crustal levels longitudinal strike-slip and normal faults are the most prominent elements controlling the basin structure, but transverse faults that extend across the basin are also important. Evidence for young slip along the major faults, present-day seismic activity, and the physiographic expression of the basin show that it is still tectonically active.

The structure of the Dead Sea basin is dominated by a large pull-apart that formed between the left stepping Jericho and Arava strike-slip faults that extend north and south of the basin, respectively (Figs. 3 and 4; Quennell, 1958, 1959; Neev and Hall, 1979; Garfunkel et al., 1981; Garfunkel, 1981; Kashai and Crocker, 1987; Ten Brink and Ben-Avraham, 1989). The pull-apart forms a deep trough, mostly some 8 km wide, between the extensions of these major strike-slip faults. These structures are embedded in a wider depression which comprises the entire width of the transform valley, 13 to 17 km, and whose marginal zones, on the two sides of the pull-apart, were shaped by normal faulting (Fig. 2). The western marginal zone, 4–5 km wide, extends from near Jericho to the Arava and comprises several blocks down-faulted by a few kilometres. Mesozoic sections drilled in these blocks conform to the pattern of facies changes and erosive unconformities of the adjacent area on the west (which trend obliquely to the transform). This proves that the marginal blocks were not displaced laterally, so that transform motion took place east of them. The marginal zone east of the pull-apart is quite narrow, often less than 2 km wide. Thus in cross section the basin has an overall asymmetric structure, with its deepest portion — comprising the pull-apart — being closer to the eastern border fault (Figs. 3 and 4; cf. Ben-Avraham, 1992). This asymmetry is present all along the basin, and it is expressed in the bathymetry of the Dead Sea.

4.2. Transverse faults

While this basic architecture persists all along the basin, several transverse faults or fault zones extend obliquely across the basin and divide the deep trough into distinct segments with different histories (Fig. 3). In the south the Buweirida fault forms a N-facing scarp which extends across the Arava. South of this fault more than 2 km of tilted Miocene sediments (Hazeva Formation, see below; Bartov, 1994) are exposed over a distance of some 20 km, whereas north of the scarp the basin floor is underlain by younger sediments. Seismic reflection revealed another important transverse fault line, also down-throwing north, near 30°50′N (Iddan fault; Ten Brink and Ben-Avraham, 1989; Frieslander et al., 1994). It is not certain whether it reaches the surface, though a distinct photo-lineament crosses the Arava in this area.

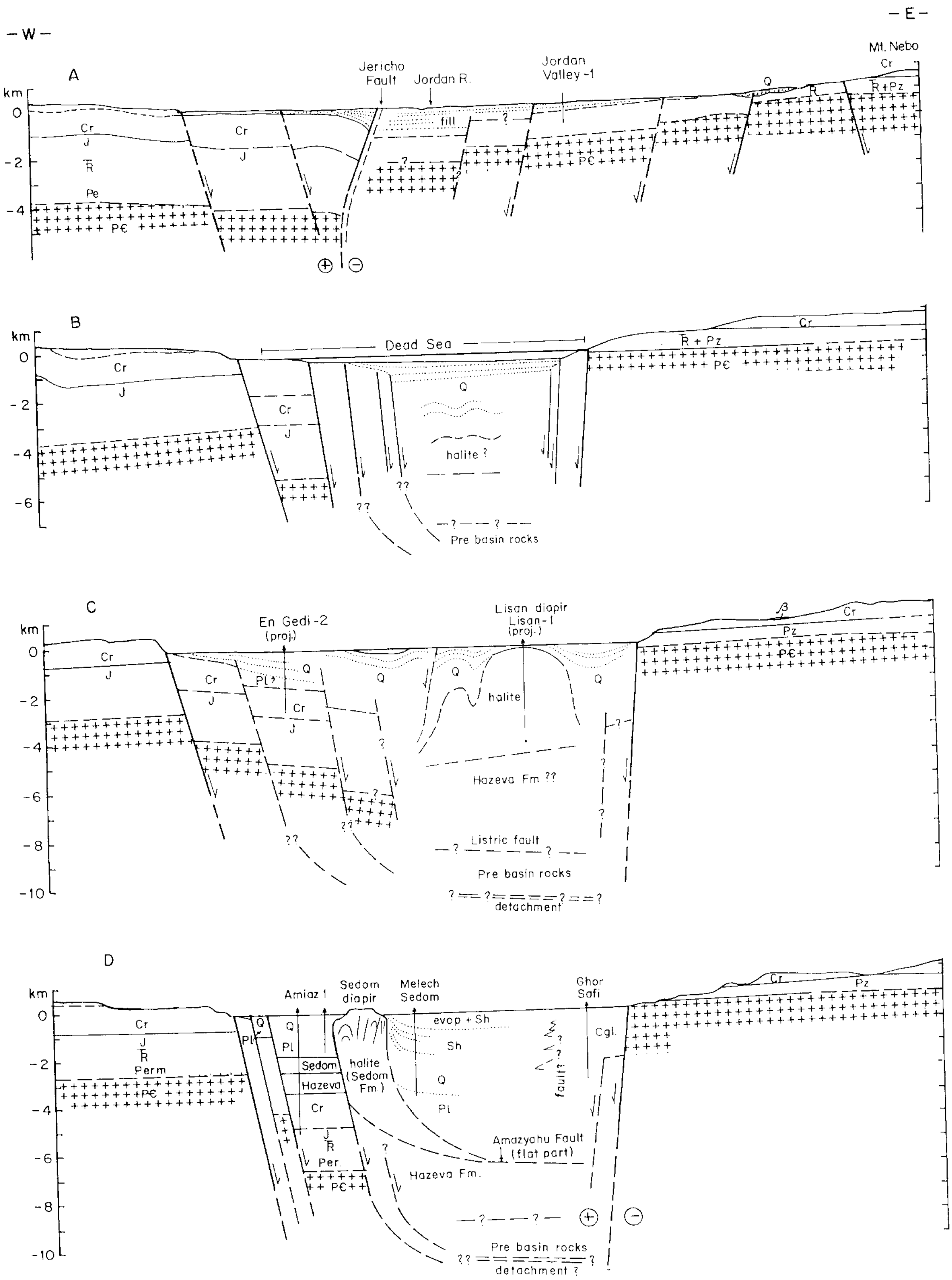


Fig. 4. Sections across the Dead Sea basin (after Garfunkel and Ben-Avraham, 1994). For location of sections, see Fig. 3.

Some 20 km farther north is the Amatzyahu fault (Fig. 3). Seismic reflection data (Kashai and Crocker, 1987; Ten Brink and Ben-Avraham, 1989) show it to be a listric fault: it flattens out at a depth of 5–6 km some 5 km north of its surface trace, and can be identified up to 15 km north of its surface trace. In the north the sediments above the fault surface are delimited by a NW–SW-trending fault or fault zone called the Boqeq fault (Fig. 3; Ben Avraham et al., 1990), but its trace was not well mapped; it probably down-throws to the north (see below). North of this line a thick salt body, whose base is some 5 km deep, forms the Lisan diapir (Bender, 1974; Abu Ajamieh et al., 1989).

The more than 300 m high slopes which delimit on the north and south the flat area at the bottom of the Dead Sea (in the northern part of the basin) are most likely also tectonically controlled, but hitherto faults could not be identified along them. The structure beneath the southern slope is obscured by diapirs (Fig. 3), but a magnetic anomaly (see below) suggests that a nearly WNW–SE-trending fault (or fault zone) offsets the basement beneath this slope. The northern slope is obscured by the sub-aqueous delta of the Jordan River, which according to seismic reflection data consists of an up to 350 m thick sediment wedge overlying flat beds (Neev and Hall, 1979). Perhaps some faults (NNW–SSE? trending) cross the area east or north of the delta, but they are not evident in the topography; alternatively, the slope is formed by a flexure.

4.3. Gravity data

The gravity field over Dead Sea basin was measured both on land and over the lake (Folkman, 1981; Ten Brink et al., 1993; Ginzburg et al., 1994). The basin was found to be marked by a prominent negative anomaly which is delimited by steep gradients along the basin margins (Fig. 5). Over most of the basin area the Bouguer anomaly is more negative than -100 mgal, reaching a minimum of -175 mgal over the Lisan diapir, which is about 300 and 200 mgal lower than the values east and west of the basin, respectively. There the free air anomaly has a minimum of -150 mgal (some 250 and 200 mgal lower than over the areas east and west of the basin, respectively). On the other hand, there is hardly any Bouguer anomaly low over the structural saddles north and south of the basin.

The accentuated negative gravity anomaly outlines the area underlain by a thick fill. It shows that in the north the basin ends quite abruptly near Jericho. In the south the anomaly changes more slowly along the Arava, recording a gradual southward thinning of the basin fill, until the gravity low becomes indistinct. Modelling of the gravity field (Fig. 5; Ten Brink et al., 1990, 1993) shows that while over the saddles north and south of the basin there is hardly any fill, within the transform valley the low-density fill is thicker than 5 km under much of the Dead Sea basin, reaching a maximum of about 10 km under the Lisan diapir. These may be under-estimates, because the fill was assumed to have a uniform and rather low density (2.15 g/cm^3). The gravity data are compatible with a Moho under the basin not deeper than under its flanks, but its exact depth is not well constrained. The gravity data also show the asymmetry of the basin and the continuity of the western zone of marginal blocks.

These results have the important implication that beneath most of the deep trough of the Dead Sea basin the crust is at least 7–10 km thinner than beneath the basin flanks. In contrast, under the saddles north and south of the basin the crust appears to be little modified and there is hardly any depression along the transform.

4.4. Magnetic anomalies

Several studies of the magnetic field over the Dead Sea and surrounding areas were carried out (Neev and Hall, 1979; Folkman, 1981; Hatcher et al., 1981; Frieslander and Ben-Avraham, 1989). These works revealed a major discontinuity in the magnetic anomaly pattern along the transform, and the Dead Sea basin in particular, with the anomaly pattern being displaced by ca. 105 km left-laterally. It is noteworthy that the anomalies west of the basin extend uninterrupted over the western marginal step faults. This corroborates the borehole data which show that the blocks forming this step were not displaced laterally relative to the area west of the basin.

The magnetic anomaly over the deep part of the Dead Sea is smooth. Two-dimensional modelling of

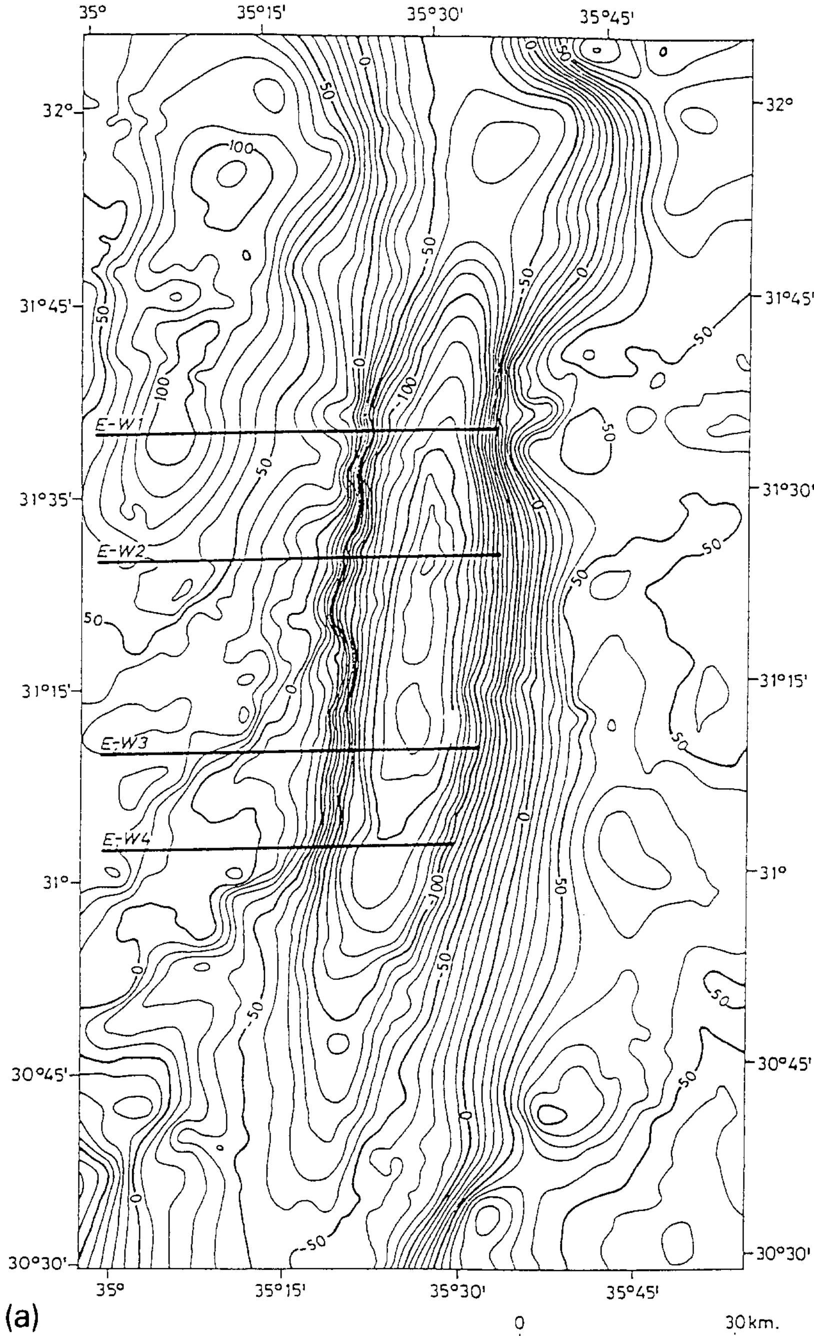


Fig. 5. The gravity anomaly over the Dead Sea basin and its interprettion (after Ten Brink et al., 1993). (A) Map of Bouguer anomaly. (B) Interpreted cross section (locations shown in A).

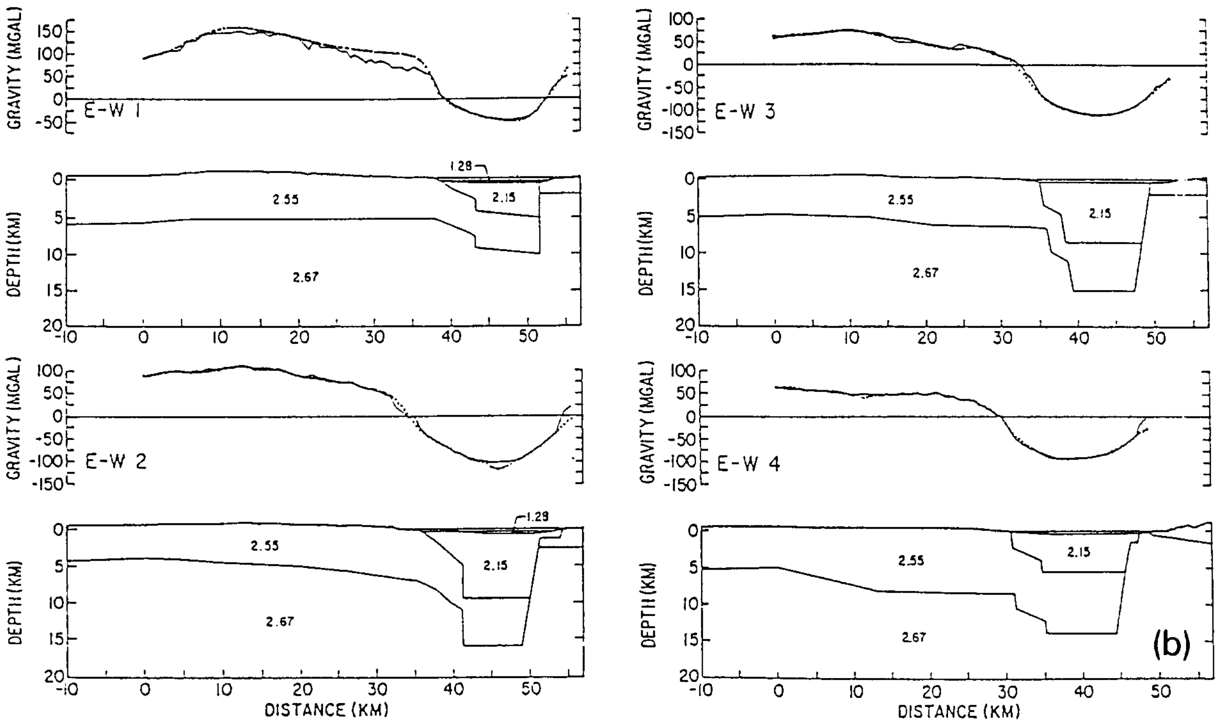


Fig. 5. Continued.

the anomaly across the basin (lat. 31°34′N) shows the magnetic basement to be about 10 km deep, which is not very different from the gravity data. Over the slope at the south of the Dead Sea (about 31°31′N) there is a pronounced WNW–ESE-trending gradient. It probably expresses a major fault in the underlying basement, which forms the boundary between two segments of the basin. South of this gradient the anomalies are irregular. A few very short-wavelength anomalies are present, and they are interpreted as due to shallow igneous bodies (see below).

4.5. Seismic refraction data

A deep seismic refraction profile was obtained in 1977 along the west margin of the transform from the Sea of Galilee to the Gulf of Elat (Aqaba), passing along the west margin of the Dead Sea basin (Ginzburg et al., 1981; Perathoner et al., 1981). As the pre-transform rocks are at shallow depth all along this profile, it imaged the western margin of the basin rather than its deepest part. This profile reveals an about 32-km-thick crust with an about 5-km-thick transition zone at its base in which V_p changes smoothly from 6.72 to 7.9 km/s, which is the sub-Moho velocity.

Seismic refraction profiles west of the basin (Ginzburg et al., 1981; Ginzburg and Folkman, 1980) show that 30–40 km west of the southern part of the Dead Sea the crust is about 33 km thick and that it thickens southward by a several kilometres. However, west of the transform the basal transition zone was not observed, the upper crust is thicker than under the margin of the Dead Sea while the lower crust ($V_p \geq 6.5$ km/s) has the same thickness, and $V_p = 8.0$ km/s below the Moho. A deep seismic reflection profile west of the Dead Sea (Rotstein et al., 1987) imaged a highly reflective lower crust up to about 20 km west of the lake, but farther west it shows a transparent lower crust. These data suggest a change in crustal structure within ca. 20 west of the Dead Sea from about latitude 31°15′N and southward (there are no data from further north). Refraction studies east of transform (El-Isa et al., 1987) revealed an about 32-km-thick crust some 50 km away from the basin, and greater thicknesses farther east; the basal transition zone was identified in these profiles, and $V_p = 8$ km/s below the Moho. This resembles the crust under the western basin margin rather than the crust farther west.

4.6. Heat flow

Heat flow measurements in abandoned wells on the western flank of the basin gave an average value of some 42 mW/m^2 (Eckstein, 1976, 1979). Temperature gradients in wells in the western part of the basin are about 20–23°C/km down to depths of 3–5 km (Kashai and Crocker, 1987), which is compatible with the low heat flow measured west of the basin. Locally, however, where hot waters rise along the marginal faults (sometimes producing hot springs) the heat flow reaches about 90 mW/m^2. Similar and higher values were also obtained near the Zarqa Main hot springs near the east shore of the lake (Abu Ajamieh et al., 1989).

Geothermal measurements in the northern part of the Dead Sea gave an average heat flow of 38 mW/m^2, while lower values were obtained in the southwestern part of the lake (Ben-Avraham et al., 1978). The heat flow in the lake may be higher, however, because much heat can escape through the highly conductive salt of the diapirs rather than through the less conductive overlying sediments. Thermal modelling shows that in this case the heat flow may reach about 60 mW/m^2 (Ben-Avraham, 1996), but this has not been verified.

5. The basin fill and its history

The Dead Sea basin is mostly covered by late Quaternary to Recent sediments, but limited exposures of older beds and borehole data supplemented by seismic reflection data allow to outline the main features of the basin fill, though the distribution of the different units and their lateral changes are still incompletely known. The fill can be divided into three main units (Zak and Freund, 1981). (a) The clastic Hazeva Formation, of Early to Late Miocene age, which consists predominantly of sandstones and conglomerates of fluviatile and lacustrine origin. (b) The evaporitic Sedom Formation, of probably Late Miocene and Pliocene age, which consists predominantly of halite, of lagoonal origin. (c) The post-evaporitic series, of Pliocene to Recent age, which consists largely of coarse to fine clastics of fluviatile and lacustrine origin and of some lacustrine carbonates and evaporites. The latter two units are included in the Dead Sea Group.

5.1. The Hazeva Formation

The Hazeva Formation (Bentor and Vroman, 1957; Garfunkel and Horowitz, 1966; Bender, 1974; Sneh, 1981; Sa'ar, 1985; Zilberman, 1992; Bartov et al., 1993; Garfunkel, 1996) is found both inside the Dead Sea basin and over the nearby areas, but it is much thicker in the basin. Vertebrate fossils and pollen show that the Hazeva Formation is of Early to Late Miocene age (Horowitz, 1987; Goldsmith et al., 1988). Because of its wide distribution this unit provides the best record of the Miocene tectonic history of the Dead Sea basin and of the areas flanking this part of the transform.

The outstanding feature of the Hazeva Formation is that it consists of material of distant derivation. Most abundant are quartzose sandstones which could have been derived only from the early Palaeozoic to Early Cretaceous sandy series and the basement rocks outcropping in southern Sinai and in Saudi Arabia, some 300 km and more away from the Dead Sea. Less distant areas could not supply the sandstones because they are covered by predominantly carbonatic marine series (of mid-Cretaceous to Eocene age). Also present are pebbles which are dispersed in the sandstones or form conglomerate beds. Most pebbles consist of sedimentary rocks, often of types not known in situ; most conspicuous is a special variety of chert of Eocene age (called 'allochthonous' or 'imported'). These exotic pebbles were derived from now largely eroded facies belts that existed inland of the presently preserved sections. Occasionally pebbles derived form the basement are also present.

The distant provenance of the Hazeva clastics and their distribution show that they were transported by a river system that extended a few hundred kilometres inland of the Dead Sea and which drained into the Mediterranean Sea via the Beer-Sheva valley (Garfunkel and Horowitz, 1966). The functioning of such a drainage system indicates the existence of a regional slope that extended across the transform, and that the transform flanks were still not uplifted enough to obstruct sediment transport. The Hazeva sediments were trapped in the Dead Sea basin as well as in much shallower erosive and structural lows in the neighbouring regions, but in places they also extended over the adjacent higher areas. However, the regional seaward slope probably did not allow the accumulation of a widespread thick sedimentary cover.

Within the Dead Sea basin the Hazeva Formation is from about 1 to ≥2.5 km thick (Fig. 4; Kashai and Crocker, 1987; Horowitz, 1987; Bartov et al., 1993; Bartov, 1994). The thick sections occur at least along the southern half of the basin. South of the Buweirida fault a >2-km-thick section is exposed. Farther north the Hazeva Formation is ≥2.5 km and 0.7 km thick in the Arava-1 and Amiaz-1 wells, respectively, located on the western marginal blocks; even thicker sections are expected in the adjacent deep part of the basin. A ≥2-km-thick section was recently drilled in the Sedom Deep-1 well near the margin of the deep trough, just east of the Sedom diapir (Gardosh et al., 1995). Still farther north, seismic data show that the salt body forming the Lisan diapir is about 5 km deep (Abu Ajamieh et al., 1989), whereas he gravity data indicate about 10 km of fill (Ten Brink et al., 1993). It is speculated that the difference indicates that there several kilometres of Hazeva beds underlie the salt.

In contrast with these thick sections, the coeval sections west of the Dead Sea basin are generally only a few hundred metres thick. This is interpreted (Garfunkel, 1996) as indicating that the basin originated already in the Early Miocene (earlier than 15 Ma), and that by some time in the Late Miocene (at 8 Ma?) the basin floor subsided a few kilometres. Sedimentation rates reached several hundred metres per Ma, but the site of greatest subsidence is not known; it could have been east of the Sedom diapir or even farther north. The Miocene depression was probably longer than 50 km (even taking into account the subsequent lengthening of the basin — see below), and was about 7 km wide south of the Buweirida fault, but further north its width is unknown. Because the Dead Sea basin is a pull-apart that was produced by lateral motion, its formation before the Middle Miocene implies that the transform was already active at that time. Thus the Dead Sea basin developed during most or the entire transform history.

During the early history of the basin (until ca. 10 Ma?) sedimentation kept pace with its subsidence, i.e., the basin did not form a depression. This is indicated by the transport of the Hazeva sediments across the basin and into the northern Negev. The supply of these sediments was cut off sometime in the Late Miocene, when uplifting of both transform flanks disrupted the river system that transported the Hazeva sediments. Since then sedimentation in the basin usually did not keep pace with its subsidence and it became a deep topographic trough.

5.2. The Sedom Formation

This evaporitic unit is confined to the Dead Sea basin (Neev and Emery, 1967; Zak, 1967; Bender, 1974; Neev and Hall, 1979; Abu Ajamieh et al., 1989). It is known mainly from the Sedom diapir which exposes a >2-km-thick section, and from the Lisan diapir where a still thicker section was drilled. These sections consist largely of halite (>70%) and of some gypsum and clastics. Minor amounts of potassium salts were drilled in both diapirs, but they seem to be more voluminous in the Lisan diapir. Farther north diapiric structures deform the floor of the Dead Sea, and still farther north halokinetic deformation is the probable cause of undulations of the sediments a few hundred metres below the flat lake floor (Neev and Hall, 1979). Thus, thick evaporites underlie at least a 50-km-long segment of the Dead Sea basin. This area overlaps the northern part of the basin which existed already during deposition of the Hazeva Formation. However, the depression during evaporite deposition may have been longer, as part of the still undated basin fill south of the Amatzyahu fault may be of the same age as the evaporites.

The Sedom diapir is an about 12-km-long salt wall which rose along the fault on the western side of the deep trough (Fig. 3; Zak, 1967). Seismic reflection data (Kashai and Crocker, 1987; Ten Brink and Ben-Avraham, 1989) suggest that the salt rose from beneath the flat part of the Amatzyahu fault. In its original horizontal position this salt body covered more than half the width of the deep trough. On the marginal block west of the Sedom diapir the evaporitic section is much thinner (670 m) than in the diapir, which shows that this block was structurally higher than the adjacent trough already during evaporite deposition. Under the Lisan diapir the evaporites are much more voluminous — this diapir is about 20 km long and dome-shaped, and its base is about 5 km deep (Bender, 1974; Neev and Hall, 1979; Abu Ajamieh et al., 1989). The difference suggests that the Lisan salt body formed in a more subsiding part of the basin than the salt that forms the Sedom diapir. This implies that the Boqeq fault was active during evaporite deposition, down-throwing to the north.

The great volume of halite, the presence of potassium salts, and the variations of the Cl/Br ratio in the halite section of the Sedom diapir indicate that the evaporites precipitated from brines of an ultimately marine origin. The Sedom Formation is probably correlative with a ca. 1-km-thick evaporitic section drilled in Zemah-1 well ca. 15 km south of Tiberias, as well as with marine sediments (locally gypsum bearing) exposed in this region and in the Yizreel (Esdraelon) valley graben (Fig. 1) which branches off the Dead Sea transform (Zak and Bentor, 1972; Marcus and Slager, 1986; Shaliv, 1991). From this correlation it is inferred that a narrow arm of the Mediterranean Sea extended into the transform valley and reached the Dead Sea basin from the north. However, other palaeogeographic settings cannot yet be ruled out.

Volcanics intercalated in the marine beds of the Tiberias region gave K–Ar ages of 6.3–5.7 Ma (Messinian) (Shaliv, 1991), and this is also the age of pollen in the evaporites in the Zemah-1 well (Horowitz, 1987). However, in the Sedom diapir and in nearby wells only Early Pliocene pollen were found (Horowitz, 1987). These data suggest that either evaporites of Miocene age exist in the Dead Sea basin as well (under the Lisan diapir?), or that evaporite deposition in the Tiberias region began earlier than in the Dead Sea basin.

The thick Sedom Formation records important subsidence of a large part of the Dead Sea basin, but its timing is uncertain. Since evaporites can accumulate very fast, the 2–3-km-thick Sedom Formation could have formed in less than 1 Ma if an adequate brine supply were available (Zak, 1967). However, the above age constraints do not exclude deposition over a 2–3 Ma long period. Thus sedimentation rates could have been anywhere in the range of 1–3 km/Myr. If the higher value is correct, a brief period of catastrophic subsidence can be inferred, but a more likely explanation is that the evaporites filled a pre-existing depression which formed by older subsidence that was faster than deposition (when the supply of Hazeva clastics was interrupted?). The lower value can be interpreted as recording syn-depositional basin subsidence, but filling of a pre-existing starved basin cannot be excluded.

5.3. *Post-evaporite series*

After the end of evaporite accumulation, several kilometres of sediments were deposited along a large part of the basin (north of about latitude 30°45′N in the Arava), but their distribution is not known in detail (Neev and Emery, 1967; Zak, 1967; Kashai and Crocker, 1987; Horowitz, 1987; Gardosh et al., 1990). During this period the filling of the Dead Sea basin did not keep pace with its subsidence, so it became a deep depression into which the drainage of a large area was diverted, while there are no indications of any drainage lines leaving the basin. Therefore a land-locked lake must have existed all or most of the time in the topographically lowest part of the basin. The lake-level rose and fell in response to climatic fluctuations which determined the balance between influx of water and water losses by evaporation. This produced alternations of fluviatile and lacustrine beds, sometimes separated by unconformities.

The bulk of the post-evaporite basin fill consists of clastics that were derived from the Hazeva Formation and from the Cretaceous–Eocene carbonatic sequences which cover the areas surrounding the basin, with minor contributions from older sandstones and basement rocks exposed south of the basin. Coarse sediments, often of fluviatile origin, were deposited mainly in the marginal parts of the basin. In the central lake mostly fine clastics accumulated, but some evaporites also formed — varved carbonates, sulphates and rarer halite which was probably derived from dissolution of the Sedom Formation.

The post-evaporitic sequence is best known from the area south of the Lisan diapir, where it contains pollen of Pliocene and Quaternary age (Fig. 4; Kashai and Crocker, 1987; Horowitz, 1987; Ten Brink and Ben-Avraham, 1989). These sediments, up to 3–5 km thick, form a large roll-over structure which developed in response to the syn-depositional activity of the underlying shallow-dipping listric Amatzyahu fault. The average sedimentation rate reached 1–1.5 km/Myr; the subsidence rate was somewhat smaller, close to 1 km/Myr, because the withdrawal of salt forming the Sedom diapir from the deep basin also contributed to its subsidence. South of the Amatzyahu fault a few kilometres of well-bedded sediments, that were imaged on seismic reflection profiles, are probably also younger than the evaporites. Over the marginal blocks west of the Sedom and Lisan diapirs the post-evaporite sections are only 1.5–2 km thick.

The distinct morphologic expression of the depression occupied by the Dead Sea suggests that it was the site of the fastest young subsidence along the basin. Seismic reflection data (Neev and Hall, 1979) show that the flat part of the lake floor is delimited by active longitudinal faults. The fastest subsidence occurs in the southeastern part of the lake where its floor is deepest (−730 m). There the sediments outline a bowl-shaped structure — the Arnon Sink — whose amplitude increases downwards; dips reach 5–6° about 0.5 km below the lake floor. It is not clear whether this is related to faulting at depth or to halokinesis which was active in the nearby area (Fig. 3).

During the last high-stand of the lake, from about 50,000 to 15,000 years ago, water level reached 180 m below sea level, i.e., some 220 m above the present level of the Dead Sea. During this stage the lake, called Lake Lisan, extended from the northern Arava to Tiberias, a distance of about 220 km (Neev and Emery, 1967; Begin et al., 1974). The sediments deposited during this period, called the Lisan Formation, still cover most of the basin floor, though they are often much dissected by erosion. The exposed sections, up to some 40 m thick, consist of alternating varves of fine clastics and of chemically deposited carbonate (aragonite). Near the lake margins coarse clastics, deposited mainly in sub-aqueous alluvial fans, are common.

6. Seismicity and young deformation

6.1. Seismic activity

The ongoing deformation of the Dead Sea basin, and the transform as a whole, is recorded by its seismicity (Arieh, 1967; Wu et al., 1973; Ben-Menahem et al., 1976; Ben-Menahem and Aboodi, 1981; Van Eck and Hofstetter, 1989, 1990; Salamon, 1993; IPRG, 1994). Earthquakes along the Dead Sea transform were known since antiquity. Shocks of M_L = 6.5–7 occur about once in a century. Within the Dead Sea basin about five events in this century had $M_L \geq 5$, the strongest being the $M_L \approx 6.5$ Jericho earthquake of July 1927. This low level of seismicity in the basin and along the transform in general is representative of the last 1000 yr (probably of the last 3000 yr). It can account for less than half the long-term slip rate deduced from the regional plate kinematics (Garfunkel et al., 1981). The reason for the discrepancy is not known. Deciphering the seismicity in the last 50,000 yr as recorded by disturbed layers in the Lisan Formation (Marco et al., 1996) may clarify this question.

The instrumental record reveals diffuse seismic activity along most of the Dead Sea basin. The earthquake foci could not be located accurately enough to be associated with particular faults, but they may cluster along the transverse faults (IPRG, 1994). The 1927 Jericho earthquake was considered to have originated on the Jericho fault north of the Dead Sea basin (e.g., Ben-Menahem et al., 1976). An epicentral location some 60 km farther south was suggested recently (Shapira et al., 1992), though this seems to conflict with the distribution of damage. The focal mechanism of this event indicates left lateral slip along the transform (Ben-Menahem et al., 1976). Other weaker events within the Dead Sea basin had similar focal mechanisms, but normal motions along faults trending nearly north–south were also found (Van Eck and Hofstetter, 1989; Salamon, 1993).

6.2. Recent fault motions

Continuing slip on many faults in the basin is evidenced by their morphologic expression and by the offset and deformation of young sediments. The ongoing activity of the Jericho and Arava faults is well expressed (Garfunkel et al., 1981; Reches and Hoexter, 1981; Gardosh et al., 1990). It is noteworthy that just north and south of the Dead Sea basin, near Jericho and in the central Arava, the young sediments along these faults are somewhat compressed (Garfunkel, 1981; Gardosh et al., 1990; Rotstein et al., 1991). This results from minor local bending to the right of the fault traces, which took place late in their history (Garfunkel, 1996). Within the Dead Sea basin compression has not been documented, however.

The intrabasinal longitudinal faults bordering the deep trough under the Dead Sea are active and displace the lake floor (Neev and Hall, 1979; Ben-Avraham et al., 1993). The great thickness of the young fill south of the Dead Sea shows that this region also continues to subside, and that here too the intrabasinal longitudinal faults moved in geologically young times, but more slowly than farther north. In contrast, the marginal faults which produce prominent scarps along the basin borders did not move much in the last few 10^4 years. This is evidenced by the fact that the Lisan beds often extend continuously from the basin into canyons which were eroded into the marginal scarps, while displacements of the Lisan and younger sediments or of the land surface were documented only along parts of the marginal fault system (Garfunkel et al., 1981).

The ongoing activity of the transverse Amatzyahu and Buweirida faults is recorded by their clear morphologic expression. The activity of other transverse structures is not documented, as noted above, though they are expressed as topographic slopes that extend obliquely across the basin.

In addition to the main faults, minor faults displace young sediments in places (Gardosh et al., 1990; Marco et al., 1996), but the conditions of exposure do no allow to trace them laterally. Therefore, and because of the very young age (a few 10^4 yr old) of the sediments which cover most of the basin floor, the full scope and long-term effects of this diffuse deformation cannot be evaluated. Continuing growth of the main diapirs is expressed by deformation of the Lisan Formation.

7. Igneous activity

The formation of the Dead Sea basin was accompanied by very little igneous activity. Until now igneous rocks were not sampled within the basin. However, short-wavelength magnetic anomalies in the southern part of the Dead Sea (Frieslander and Ben-Avraham, 1989; Ben-Avraham, 1996) show the existence of shallow strongly magnetized bodies, which are interpreted as igneous rocks. Some 40 km north of the Dead Sea a basalt body is exposed along the transform in the Grain Sabt uplift (Fig. 3; Bender, 1968, 1974). These bodies must have been emplaced along the transform, as they are not linked to any igneous activity on the adjacent basin flanks. Other igneous bodies emplaced along the transform are known farther north, indicating that magmas rose along the transform, but only small volumes reached close to the surface (Garfunkel, 1989).

Sporadic igneous activity occurred all along the eastern side of the Dead, Sea basin (Bender, 1968, 1974). K–Ar dating (Barberi et al., 1980; Steinitz and Bartov, 1991) shows that at about 9–6 Ma basalt flows were extruded over the plateau east of the basin, before the incision of the canyons into this area, i.e., they pre-date most uplifting of eastern basin flank. Younger activity, at 3.5–0.5 Ma, produced several cones and minor flows on the plateau and slopes east of the Dead Sea. Basalt flows descended along the Zarqa Main canyon at 3.4 Ma, when it was still incompletely eroded, and also at about 0.5 Ma. The continuation of these volcanics under the Dead Sea is recorded by a strong localized magnetic anomaly (Frieslander and Ben-Avraham, 1989). They are probably cut off by marginal faults, but it is not clear whether their laterally offset extension can be identified.

8. Summary and conclusions

The foregoing data, though still incomplete, provide a general picture of the Dead Sea basin and allow to discuss its formation and growth. Since the basin is located along an intra-continental transform and its internal structure is dominated by a large pull-apart, the basin architecture and development must have been controlled by lateral motion, while transverse separation had only a minor role. To interpret the basin from this point of view, several of its features are most important.

The available data show that the Dead Sea basin formed at about 15 Ma or somewhat earlier, i.e., in the initial stages, or even at the beginning, of the transform activity. Thus, the lateral motion during the basin's history was close to the total transform offset of 105 km, and most probably >80 km.

The basin history, as recorded by its fill, comprises three stages with distinct sedimentary regimes. In the first stage — when the Hazeva Formation formed — deposition more or less kept pace with the basin subsidence, and sediments could be transported across the basin. This regime ended in the Late Miocene (10–8 Ma) when uplifting of the transform flanks obstructed the supply of clastics. By then the basin attained at least half its present length and a depth of a few kilometres. During the next stage, at about 6–4 Ma, the basin was briefly invaded by an arm of the sea, which resulted in deposition of a several kilometres thick evaporitic series — the Sedom Formation — along an at least 50-km-long area. In the third stage, up to 5 km of fluviatile and lacustrine sediments — mainly clastics — were deposited in a more than 100-km-long area. During this stage, and probably already since the end of deposition of the Hazeva Formation, the sediment supply usually did not keep pace with the basin subsidence, so it became a land-locked depression into which the drainage of the surrounding areas was diverted.

This history reflects directly the response of surficial sedimentation processes to the basin formation, but it also records deep seated tectonic processes. The distribution of the basin fill shows that its development was characterized by simultaneous subsidence of large parts of the pull-apart, though the subsidence rates probably varied laterally. Moreover, at any period the subsiding area overlapped to a great extent the area which subsided in previous periods: The Sedom Formation covers a part of the basin of Hazeva times, while the post-evaporite sediments covers areas which subsided earlier (Fig. 3). Thus, the basin growth was more complex than simple successive additions of more northerly portions as suggested by Zak and Freund (1981), though the most strongly subsiding area probably shifted northward. Young subsidence appears to have been fastest in the northern part of the basin, having produced the depression occupied by the Dead Sea. Earlier, during deposition of the Sedom Formation, the greatest subsidence occurred farther south — under the Lisan diapir. On the other hand, the southern extremity of the basin in the central Arava did not subside after deposition of the Hazeva Formation. Subsidence rates reached 0.5–1 km/Myr, but their temporal and lateral variations are not well constrained.

The ongoing deformation, as recorded by young fault motions and seismic activity, occurs over most of the basin area, though with variable intensity: it is stronger in the low area north of the Amatzyahu fault than south of it. This correlates with the distribution of young subsidence, which was also greatest in the northern half of the basin. It is likely that such a relation existed throughout the basin history.

The shallow structure of the basin comprises a central pull-apart and marginal blocks. The pull-apart is almost 150 km long, 7–8 km wide, and is filled by low-density sediments with a maximum thickness of 10 km or more; in its central 70 km the fill is well over 5 km thick. The marginal blocks west of the pull-apart form an about 5-km-wide zone, while in the east they are narrower. Longitudinal faults are the most prominent structures. They include intra-basinal faults which delimit the pull-apart and marginal faults which control the marginal blocks; these fault zones extend north and south of the pull-apart (Figs. 2–4). The longitudinal faults bordering the pull-apart, being continuations of the major strike-slip faults north and south of the basin, accommodated significant lateral motions (mainly on the east, see below), as well as prominent vertical offsets (perhaps reaching 10 km). Another type of intra-basinal structures includes transverse faults or fault zones which cross obliquely the deep part of the basin and divide it into segments with different subsidence histories. Between these faults the basin fill is little deformed, being mildly faulted and flexed. Strong deformation is known only near diapirs formed by the salt of the Sedom Formation.

The deep structure of the Dead Sea basin is little known, but the available geophysical data show that the crust was modified under the basin. Seismic refraction data show a crustal thickness of about 33 km some 30–50 km west and east of the basin as well as under the western basin margin, but under the deep part of the pull-apart the crystalline crust is only 2/3 as thick (and perhaps thinner) according to the gravity data. The heat flow in deep wells in the western part of the basin and in the northern Dead Sea is quite low — about 40 mW/m^2

(≈1 HFU), similar to the average the heat flow measured on the basin flanks. This shows that if a thermal anomaly is associated with basin, it has a limited areal extent. However, the basaltic igneous activity along the eastern basin flank and along the transform, and perhaps in the basin itself, suggest higher than normal temperatures in the lower lithosphere beneath the basin.

These features can be used to clarify how the structure and development of the Dead Sea basin relate to the lateral transform motion by utilizing on the observation that this basin is essentially a pull-apart (rhomb-graben) that formed between two major left-stepping strike-slip faults along a transform plate boundary. This allows to interpret the basin within the framework of the basic structural relations of pull-aparts (Garfunkel, 1981, 1996; Aydin and Nur, 1982; Mann et al., 1983). These relations show that pull-aparts grow by becoming longer parallel to the lateral slip that produces them, while their edges move apart (Figs. 2 and 6). Consequently, their interiors must be dilated or stretched parallel to the lateral motion (Fig. 2B), e.g., as a result of activity of transverse normal faults or of zones of diffuse extension which trend across the pull-aparts.

To apply this notion to the Dead Sea basin, the important observation regarding its shallow structure is that the transverse faults are the only structures which allow basin lengthening, whereas between these faults the fill is little deformed. To explain the basin lengthening in a way compatible with this observation, Arbenz (1984) proposed that the basin grew as a result of activity of several transverse listric faults which continue downward to a mid-crustal detachment zone. In this model the rocks within the growing part of the basin slide above the flat-lying portions of the listric faults without much internal deformation, while new sediments continue to accumulate over them. This allows considerable lengthening of the basin parallel to the transform while leaving the shallow basin fill — especially its upper part — little deformed, though above the curved parts of the faults the basin fill becomes titled and forms roll-over structures.

Such a deformation style is indeed exemplified by the Amatzyahu fault (cf. figures in Kashai and Crocker, 1987; Ten Brink and Ben-Avraham, 1989). Its activity caused a N–S-oriented extension of per-

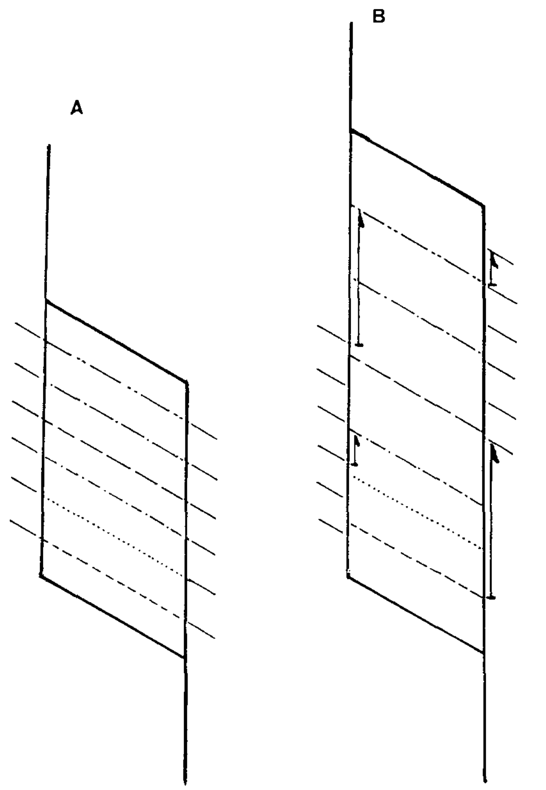


Fig. 6. Effects of lengthening of a pull-apart by stretching of its floor. (A) Markers within the basin at some time. (B) The position of these markers after the some growth of the pull-apart. This results in transfer of the lateral motion across the stretched part of the basin floor.

haps 5 km during deposition of the post-evaporite basin fill. If the other transverse faults are similar, their activity can account for the basin lengthening at the required rate. However, the Amatzyahu fault — the only one whose deep structure has been documented — flattens out within the basin fill, while its relation with deeper structures, especially in the basement, is not known. Thus, it is not obvious that it has the role envisaged in the Arbenz model. This model also predicts that the oldest parts of the basin fill and the pre-transform sediments should have dips reaching 30–40° where they are crossed by the listric faults. Hitherto there is no evidence for such dips, though the available seismic reflection data may not suffice to test this aspect of the model. Thus, while the Arbenz model may well identify the principal style of shallow deformation in the Dead Sea basin, much more information regarding the deep basin structure is needed before this model can be evaluated and applied.

In any case the deep deformation is expected to be complex. During the early basin history the basement was shallow enough to be in the brittle field, so it is expected to have been faulted. However, as the pull-apart deepened, the thermal blanketing effect of its fill may have caused parts of the basement that originally deformed in a brittle manner to deform in a semi-brittle or ductile manner. Thus, the deformation style and its variation with depth may well have changed with time.

Regardless of any particular model, the thinning of the crust beneath the pull-apart can be interpreted as a result of its lengthening and stretching of its floor parallel to the transform, and this is considered to be the cause of the continuing basin subsidence. The subsidence history of the Dead Sea basin indicates, therefore, that at any one time large segments of the pull-apart floor were stretched simultaneously, similar to the situation in the Quaternary.

Another relevant basic feature of pull-aparts it that their growth and internal deformation results in the transfer of lateral slip from one of their sides to the other (Fig. 6). Stretching of any interior portion of a pull-apart necessarily involves partitioning of the lateral motion between the longitudinal faults along the adjacent sides of the pull-apart. Therefore the indications that subsidence of the Dead Sea basin, and by inference also the stretching of its floor, tended to be fastest along its northern part, imply that this was the area where transfer of lateral motion across the basin was most important. Thus, lateral motion must have been important along both sides of the northern part of the basin, whereas on going south lateral motion took place mostly on its eastern side. This could have influenced the basin asymmetry.

Interpretation of the crustal thinning under the pull-apart as a result of stretching during its lengthening calls for a mass balance of the crustal material which relates the amount of stretching to the lateral motion that produced the pull-apart. In our case the lateral motion exceeded 80 km while now the basin reaches a length of 130–150 km (depending on where exactly it ends in the south). Thus, if the present arrangement of the main strike-slip faults persisted during the entire basin history, then its floor would have been stretched by a factor of >2. Unfortunately, it is difficult to apply and evaluate such calculations, because some crucial input data are still uncertain. For example, the present crustal thickness under the basin is not known with sufficient precision to test such a conclusion. Moreover, a change in crustal volume by addition of igneous material cannot be ruled out. This possibility is raised by the igneous activity along the transform and by the evidence that magmas ascended along the transform (Garfunkel, 1989). The presence of a transition zone in which $V_p = 6.7$ to 7.9 km/s at the base of the crust under the basin margin (Perathoner et al., 1981) can also be interpreted as evidence for intrusion or underplating by basic igneous rocks which have the such a seismic velocity.

Perhaps the most crucial problem is that the way in which the shallow deformation extends downward is unknown. Because of the rheological stratification of the crust and lithosphere (Kirby, 1985; Carter and Tsenn, 1987; Cloetingh and Banda, 1992), moderately or shallow dipping detachment zones or structural decollements may exist beneath the transform. This will decouple the shallow structures from the deformation in the lower crust of underlying mantle. In particular the lateral motion may not occur directly below the pull-apart. In such a case the increase in area during the growth of the pull-apart seen on the surface may not be matched at depth by an equal amount of separation between the plates (Garfunkel, 1996).

Because of these problems, quantitative modelling of the crustal extension beneath the Dead Sea transform is premature. These considerations highlight the need for better data concerning the structure of the lower part of the basin fill and the crustal structure.

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References

Abu Ajamieh, M.M., Bender, F.K., Eicher, R.N., El Kaysi, K.K., Nimri, F., Qudah, B.H. and Sheyyab, K.H., 1989. Natural

resources in Jordan: inventory, evaluation, development program. Amman, Jordan, Natural Resources Authority, 224 pp.
Arbenz, J.K., 1984. Oil potential of the Dead Sea area. Seismica Oil Exploration, Tel Aviv, Rep., 84/111, 44 pp.
Arieh, E.J., 1967. Seismicity of Israel and adjacent areas. Geol. Surv. Isr. Bull., 43: 1–14.
Aydin, A. and Nur, A., 1982. Evolution of pull-apart basins and their scale independence. Tectonics, 1: 91–105.
Bandel, K. and Khouri, H., 1981. Lithostratigraphy of the Triassic of Jordan. Facies, 4: 1–23.
Barberi, F., Capaldi, G., Gasperini, P., Martinelli, G., Santacroce, R., Scandone, R., Treuil, M. and Varet, J., 1980. Recent basaltic volcanism of Jordan and its implications for the geodynamic history of the Dead Sea Shear Zone. Atti Accad. Naz. Lincei, 47: 667–683.
Bartov, Y., 1974. A structural and paleogeographic study of the central Sinai faults and domes. Ph.D. Thesis, Hebrew Univ., Jerusalem, Israel, 143 pp. (in Hebrew with English abstract).
Bartov, Y., 1994. The geology of the Arava: explanatory notes to geological map. Geol. Surv. Isr. Rep., GSI/4/94, 16 pp.
Bartov, Y., Frieslander, U. and Rotstein, Y., 1993. New observations on the structure and evolution of the Arava rift valley. Isr. Geol. Soc., Annu. Meet., p. 11 (abstract).
Begin, B.Z., Ehrlich, A. and Nathan, Y., 1974. Lake Lisan — the Pleistocene precursor of the Dead Sea. Geol. Surv. Isr. Bull., 63, 30 pp.
Ben-Avraham, Z., 1985. Structural framework of the Gulf of Elat (Aqaba), northern Red Sea. J. Geophys. Res., 90: 703–726.
Ben-Avraham, Z., 1992. Development of asymmetric basins along continental transform faults. Tectonophysics, 215: 209–220.
Ben-Avraham, Z., 1996. Geophysical framework of the Dead Sea. In: T.N. Niemi, Z. Ben-Avraham and J. Gat (Editors), The Dead Sea. Oxford Univ. Press (in press).
Ben-Avraham, Z., Hanel, R. and Vilinger, H., 1978. Heat flow through the Dead Sea rift. Mar. Geol., 28: 253–269.
Ben-Avraham, Z., Almagor, G. and Garfunkel, Z., 1979. Sediments and structure of the Gulf of Elat. Sediment. Geol., 23: 239–267.
Ben Avraham, Z., Ten Brink, U. and Charrach, J., 1990. Transverse faults at the northern end of the southern basin of the Dead Sea graben. Tectonophysics, 180: 37–47.
Ben-Avraham, Z., Niemi, T.N., Neev, D., Hall, J.K. and Levy, Y., 1993. Distribution of Holocene sediments and neotectonics in the deep north basin of the Dead Sea. Mar. Geol., 113: 219–231.
Bender, F., 1968. Geologie von Jordanien. Borntraeger, Berlin, 230 pp.
Bender, F., 1974. Geology of Jordan. Borntraeger, Berlin, Suppl Ed. of Vol. 7, 196 pp.
Ben-Menahem, A. and Aboodi, E., 1981. Micro- and macroseismicity of the Dead Sea rift and off-coast eastern Mediterranean. Tectonophysics, 80: 199–234.
Ben-Menahem, A., Nur, A. and Vered, M., 1976. Tectonics, seismicity and structure of the Afro-Eurasian junction — the breaking of an incoherent plate. Phys. Earth Planet. Inter., 12: 1–50.
Bentor, Y.K. and Vroman, A., 1957. The geological map of Israel, scale 1:100 000, Sheet 19: Arava Valley, with explanatory notes. Government Printer, Jerusalem, 66 pp.
Carter, N.L. and Tsenn, M.C., 1987. Flow properties of continental lithosphere. Tectonophysics, 136: 27–63.
Cloetingh, S. and Banda, E., 1992. Mechanical structure of Europe's lithosphere. In: E. Blundell, R. Freeman and S. Mueller (Editors), A Continent Revealed: The European Geotraverse. Cambridge Univ. Press, pp. 80–91.
Coleman, R.G., 1984. The Red Sea: A small ocean basin formed by continental extension and sea floor spreading. 27th Int. Geol. Congr., 23: 93–121.
Eckstein, Y., 1976. The measurement and interpretation of terrestrial heat flow in Israel. Geol. Surv. Isr. Rep. Hydro/3/76, 170 pp.
Eckstein, Y., 1979. Review of the heat flow data from the Eastern Mediterranean region. Pure Appl. Geophys., 117: 150–179.
El-Isa, Z., Metchie, J., Prodehl, C., Makris, J. and Rihm, R., 1987. A crustal structure study of Jordan derived from seismic refraction data. Tectonophysics, 138: 235–253.
Eyal, M., Eyal, Y., Bartov, Y. and Steinitz, G., 1981. The tectonic development of the western margin of the Gulf of Elat (Aqaba) rift. Tectonophysics, 80: 39–66.
Folkman, Y., 1981. Structural features in the Dead Sea–Jordan rift zone, interpreted from a combined magnetic-gravity study. Tectonophysics, 80: 135–146.
Freund, R., 1965. A model for the development of Israel and adjacent areas since the Upper Cretaceous times. Geol. Mag., 102: 189–205.
Freund, R., Garfunkel, Z., Zak, I., Goldberg, M., Weissbrod, T. and Derin, B., 1970. The shear along the Dead Sea rift. Philos. Trans. R. Soc. London, Ser. A, 267: 107–130.
Frieslander, U. and Ben-Avraham, Z., 1989. The magnetic field over the Dead Sea and its vicinity. Mar. Pet. Geol., 6: 148–160.
Frieslander, U., Bartov, J., Garfunkel, Z., Rotstein, Y. and Rybakov, M., 1994. The structure of the Arava — Initial results from geophysical data. Isr. Geol. Soc., Annu. Meet., p. 26 (abstract).
Gardosh, M., Reches, Z. and Garfunkel, Z., 1990. Holocene tectonic deformation along the western margins of the Dead Sea. Tectonophysics, 180: 123–137.
Gardosh, M., Shulman, H. and Salhov, S., 1995. Findings from Amiaz East-1 and Sedom Deep-1 wells and their implications for the geology of the southern Dead Sea basin. Isr. Geol. Soc. Annu. Meet., p. 36 (abstract).
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 and E. Tchernov (Editors), The Zoogeography of Israel. Junk, Dordrecht, pp. 7–34.
Garfunkel, Z., 1989., Tectonic setting of Phanerozoic magmatism in Israel. Isr. J. Earth Sci., 38: 51–74.
Garfunkel, Z., 1996. The history and formation of the Dead Sea basin. In: T.N. Niemi, Z. Ben-Avraham and J. Gat (Editors), The Dead Sea. Oxford Univ. press (in press).

Garfunkel, Z. and Ben-Avraham, Z., 1994. Guidebook to the Dead Sea field trip. Tel Aviv Univ., 36 pp.

Garfunkel, Z. and Horowitz, A., 1966. The Upper Tertiary and Quaternary morphology of the Negev, Israel. Isr. J. Earth Sci., 15: 101–117.

Garfunkel, Z., Zak, I. and Freund, R., 1981. Active faulting in the Dead Sea rift. Tectonophysics, 80: 1–26.

Ginzburg, A. and Folkman, Y., 1980. The crustal structure between the Dead Sea rift and the Mediterranean Sea. Earth Planet. Sci. Lett., 51: 181–188.

Ginzburg, A., Makris, J., Fuchs, K. and Prodehl, C., 1981. The structure of the crust and upper mantle in the Dead Sea rift. Tectonophysics, 80: 109–119.

Ginzburg, A., Folkman, Y., Rybakov, M., Rotstein, Y., Assael, R. and Yuval, Z., 1994. Israel–Bouguer gravity map, scale 1:500 000. Inst. Pet. Res. Geophys., Holon, Israel.

Goldsmith, N.F., Hirsch, F., Friedman, G.M., Tchernov, E., Derin, B., Gerry, E., Horowitz, A. and Weinberger, G., 1988. Rotem mammals and Yeroham Crassostreids: Stratigraphy of the Hazeva Formation (Israel) and the paleogeography of Miocene Africa. Newsl. Stratigr., 20: 73–90.

Hatcher, R.D., Zeiz, I., Reagan, R.D. and Abu-Ajameh, M., 1981. Sinistral strike-slip motion on the Dead Sea rift: confirmation from new magnetic data. Geology, 9: 458–462.

Horowitz, A., 1987. Palynological evidence for the age and rate of sedimentation along the Dead Sea Rift, and structural implications. Tectonophysics, 141: 107–115.

IPRG (Institute for Petroleum Research and Geophysics), 1994. Seismicity of Israel and Adjacent Areas, scale 1:750,000.

Joffe, S. and Garfunkel, Z., 1987. The plate kinematics of the circum Red Sea — a re-evaluation. Tectonophysics, 141: 5–22.

Kashai, E.L. and Crocker, P.F., 1987. Structural geometry and evolution of the Dead Sea–Jordan rift system as deduced from new subsurface data. Tectonophysics, 141: 33–60.

Kirby, S.H., 1985. Rock mechanics observations pertinent to the rheology of the continental lithosphere and the localization of strain along shear zones. Tectonophysics, 119: 1–27.

LePichon, X. and Gaulier, J.M., 1988. The rotation of Arabia and the Levant fault system. Tectonophysics, 153: 271–294.

Mann, P., Hempton, M.R., Bradley, D.C. and Burke, K., 1983. Development of pull-apart basins. J. Geol., 91: 529–554.

Marco, S., Stein, M., Agnon, A. and Ron, H., 1996. Long-term earthquake clustering: A 50 000-year paleoseismic record in the Dead Sea graben. J. Geophys. Res., in press.

Marcus, E. and Slager, J., 1986. The sedimentary–magmatic sequences of Zemah-1 well (Dead Sea rift, Israel) and its emplacement in time and space. Isr. J. Earth Sci., 34: 1–10.

McKenzie, D.P., Davies, D. and Molnar, P., 1970. Plate tectonics of the Red Sea and East Africa. Nature, 224: 125–133.

Neev, D. and Emery, K.O., 1967. The Dead Sea: Depositional processes and environments of evaporites. Geol. Surv. Isr. Bull., 41, 147 pp.

Neev, D. and Hall, J.K., 1979. Geophysical investigations in the Dead Sea. Sediment. Geol., 23: 209–238.

Perathoner, B., Fuchs, K., Prodehl, C. and Ginzburg, A., 1981. Seismic investigation of crust–mantle transition in continental rift systems — Jordan–Dead Sea rift and Rhinegraben. Tectonophysics, 80: 121–134.

Picard, L., 1943. Structure and evolution of Palestine with comparative notes on neighboring countries. Bull. Geol. Dep. Hebrew Univ. Jerusalem, 4: 1–187.

Picard, L., 1959. Geology and oil exploration of Israel. Bull. Res. Counc. Isr., 8G: 1–30.

Quennell, A.M., 1958. The structural and geomorphic evolution of the Dead Sea rift. Q. J. Geol. Soc. London, 114: 2–24.

Quennell, A.M., 1959. Tectonics of the Dead Sea rift. 20th Int. Geol. Congr. Assoc. Serv. Geol. Afr., pp. 385–405.

Reches, Z. and Hoexter, D.F., 1981. Holocene seismic and tectonic activity in the Dead Sea area. Tectonophysics, 80: 235–254.

Rotstein, Y., Yuval, Z. and Trachtman, P., 1987. Deep seismic reflection studies in Israel — an update. Geophys. J. R. Astron. Soc., 89: 389–394.

Rotstein, Y., Bartov, Y. and Hofstetter, A., 1991. Active compressional tectonics in the Jericho area, Dead sea rift. Tectonophysics, 198: 239–259.

Sa'ar, H., 1985. The provenance and deposition of sandstones in the sedimentary fill of the Dead Sea rift. M.Sc. Thesis, Hebrew University, Jerusalem, Israel, 92 pp. (in Hebrew).

Salamon, A., 1993. A seismotectonic analysis of earthquakes in Israel and adjacent areas. Ph.D. Thesis, Hebrew Univ., 135 pp. (in Hebrew with English abstract).

Shaliv, G., 1991. Stages in the tectonic and volcanic history of the Neogene basin in the Lower Galilee and the Valleys. Geol. Surv. Isr. Rep., GSI/11/91, 100 pp. (in Hebrew with English abstract).

Shapira, A., Avni, R. and Nur, A., 1992. A new estimate of the epicenter of the Jericho earthquake of 11 July 1927. Isr. J. Earth Sci., 42: 93–96.

Sneh, A., 1981. The Hazeva Formation in the Arava Valley. Isr. J. Earth Sci., 30: 81–91.

Steinitz, G. and Bartov, Y., 1991. The Miocene–Pliocene history of the Dead Sea segment of the Rift in light of K–Ar ages of basalts. Isr. J. Earth Sci., 40: 199–208.

Steinitz, G., Bartov, Y., Eyal, M. and Eyal, Y., 1981. K–Ar age determination of Tertiary magmatism along the western margin of the Gulf of Elat. Geol. Surv. Isr. Curr. Res., 1980: 27–29.

Ten Brink, U.S. and 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. and Ben-Avraham, Z., 1990. Uplift and possible Moho offset across the Dead Sea transform. Tectonophysics, 180: 71–85.

Ten Brink, U.S., Ben-Avraham, Z., Bell, R.E., Hassouneh, M., Coleman, D.F., Andreasen, G., Tibor, G. and Coakley, B., 1993. Structure of the Dead Sea pull-apart basin from gravity analysis. J. Geophys. Res., 98: 21,877–21,894.

Van Eck, T. and Hofstetter, A., 1989. Microearthquake activity in the Dead Sea. Geophys. J. Int., 99: 605–620.

Van Eck, T. and Hofstetter, A., 1990. Fault geometry and spatial clustering of micro-earthquakes along the Dead Sea–Jordan rift fault zones. Tectonophysics, 180: 15–27.

Wilson, J.T., 1965. A new class of faults and their bearing on continental drift. Nature, 207: 343–347.

Wu, F.T., Karcz, I., Arieh, E.J., Kafri, U. and Peled, U., 1973. Microearthquakes along the Dead Sea rift. Geology, 1: 159–161.

Zak, I., 1967. The geology of Mount Sedom. Ph.D. Thesis, Hebrew Univ., Jerusalem, Israel, 208 pp. (in Hebrew with English abstract).

Zak, I. and Bentor, Y.K., 1972. Some new data on the salt deposits of the Dead Sea area, Israel. Proc. Symp. on Geology of Saline Deposits, pp. 137–146.

Zak, I. and Freund, R., 1981. Asymmetry and basin migration in the Dead Sea rift. Tectonophysics, 80: 27–38.

Zilberman, E., 1992. Remnants of Miocene landscape in the central and northern Negev and their paleogeographical implications. Geol. Surv. Isr. Bull., 83, 54 pp.