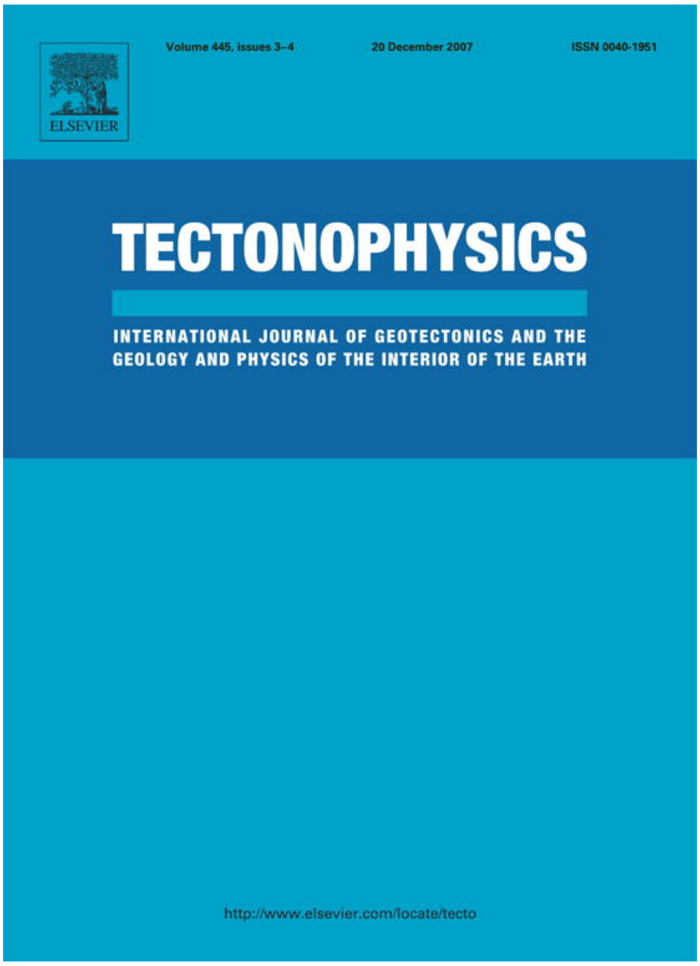
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Temporal variation in the geometry of a strike–slip fault zone: Examples from the Dead Sea Transform

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Abstract

The location of the active fault strands along the Dead Sea Transform fault zone (DST) changed through time. In the western margins of Dead Sea basin, the early activity began a few kilometers west of the preset shores and moved toward the center of the basin in four stages. Similar centerward migration of faulting is apparent in the Hula Valley north of the Sea of Galilee as well as in the Negev and the Sinai Peninsula. In the Arava Valley, seismic surveys reveal a series of buried inactive basins whereas the current active strand is on their eastern margins. In the central Arava the centerward migration of activity was followed by outward migration with Pleistocene faulting along NNE-trending faults nearly 50 km west of the center. Largely the faulting along the DST, which began in the early–middle Miocene over a wide zone of up to 50 km, became localized by the end of the Miocene. The subsidence of fault-controlled basins, which were active in the early stage, stopped at the end of the Miocene. Later during the Plio-Pleistocene new faults were formed in the Negev west of the main transform. They indicate that another cycle has begun with the widening of the fault zone. It is suggested that the localization of faulting goes on as long as there is no change in the stress field. The stresses change because the geometry of the plates must change as they move, and consequently the localization stage ends. The fault zone is rearranged, becomes wide, and a new localization stage begins as slip accumulates. It is hypothesized that alternating periods of widening and narrowing correlate to changes of the plate boundaries, manifest in different Euler poles.

Keywords: Dead Sea Transform fault; Tectonics; Active faulting; Fault zone geometry; Fault segmentation; Middle East

1. Introduction

This work aims to examine whether changes in the geometry of faulting exhibit any pattern. It is done by comparing the changes in at several sections of the Dead Sea Transform fault (Fig. 1). Changes in the geometry of fault zones as slip accumulates through time are inherent to their evolution. The changes reflect mechanical properties of the crust and bear on the nature of the seismic activity. Fault zones may begin as simple systems that become more complex in time or the opposite — begin complex and evolve toward simplicity. A common type of geometrical complexity at the surface is fault segmentation and overlapping segments. Pull-apart basins, which are formed between such overlapping segments, are commonly shorter than the total offsets on strike–slip faults. This can be explained by changes in the fault geometry and formation of such basins after some slip had accumulated. Aydin and Nur (1982) suggest that pull apart structures grow longer and wider in time, preserving a length/width ratio of about 3/1.

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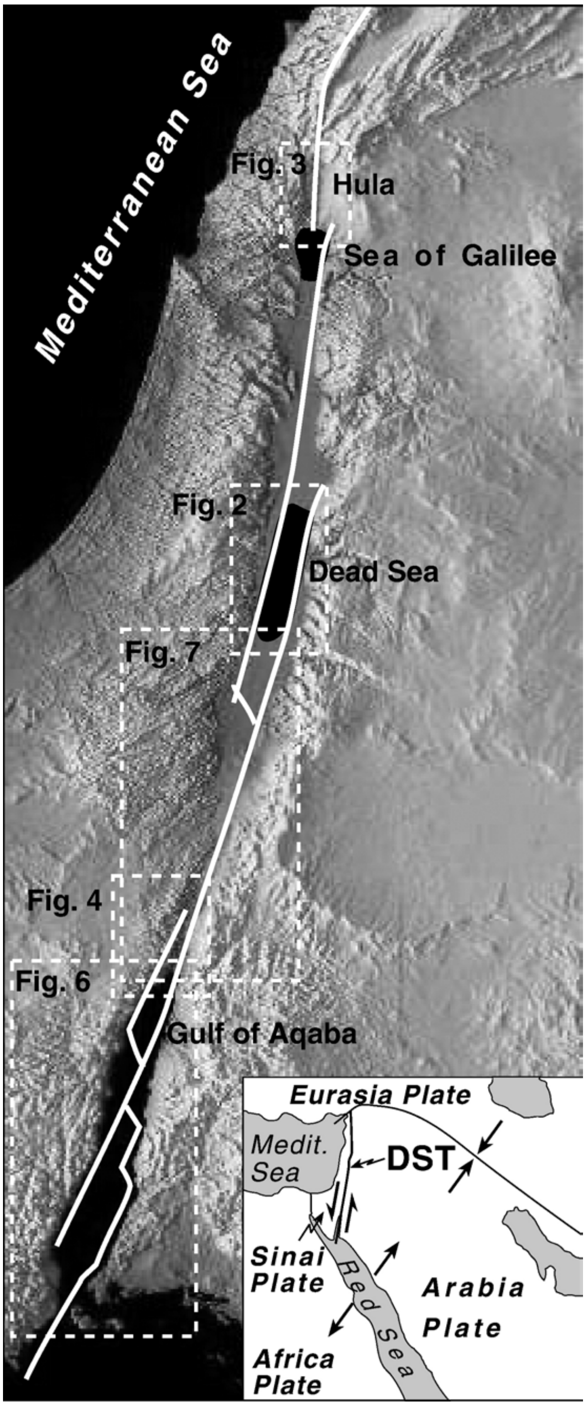

Fig. 1. Location of examples for changes in geometry of the Dead Sea Transform (solid white line) fault zone. Inset: Plate tectonics in the Middle East and the Dead Sea Transform fault (DST). Shaded relief map from Hall (1994).

Reduction of fault trace complexity with increasing cumulative slip was suggested in previous studies (e.g., Stirling et al., 1995). The geometrical evolution was linked to evolution of earthquake distribution models, beginning with Gutenberg–Richter distribution when the fault zone is complex and becoming Characteristic Earthquake distribution when the geometry becomes simple (Wesnousky, 1994). The distribution pattern is therefore important to seismic hazard assessment. Similar evolutionary trends are noticed in theoretical models of damage evolution as well (Lyakhovsky et al., 2001). The detailed fault zone geometry is one of the keys to understanding the physics, mechanics, and kinematics of faulting.

Migration of faulting activity has been documented in many extensional structures. For example, young continental rifts (<10 Ma), such as in southern Kenya, commonly comprise asymmetric rift basins bound by steep border faults that accommodate most of the strain across the rift. The older, more evolved Asal Rift in Djibouti is a much narrower than the transitional rift sector at the northern end of the Main Ethiopian Rift. Strain and magmatism have migrated from the border faults to a narrow zone within the rift valley (Ebinger and Casey, 2001). The West Antarctic Rift system also began as broadly distributed extension throughout much of West Antarctica in the Late Cretaceous but the second stage of extension was focused in a narrow zone primarily in the Victoria Land Basin (Huerta and Harry, 2007). Strike–slip environments also evolve because of plate motions. For example, the location of the shear zone between the Pacific and the North American plates was initially west and later moved to where the San Andreas Fault is now (Garfunkel, 1973). The Altyn Tagh fault system comprises multiple fault strands in a ~100 km wide zone whereas the current active strand is very narrow. The system also evolved by the sequential formation and death of shortlived fault strands (Cowgill et al., 2004a).

This paper presents examples from the Dead Sea Transform (Figs. 2–7). A new study in the Dead Sea basin shows narrowing of the fault zone in time as fault activity migrates toward the center of the basin (Fig. 2). Data from other localities in the Hula Basin (Fig. 3) and the Gulf of Aqaba (Figs. 4, 5 and 6) show similar characteristics, but the Arava Valley at the southern DST exhibits a different behavior, where buried fault-bound sedimentary basins below the Arava Valley (Fig. 7) demonstrate that such basins have limited longevity. A group of faults in the Negev shows migration of Quaternary activity away from the center of the valley (Avni et al., 2000).

1.1. The Dead Sea Transform fault

The Dead Sea Transform (DST) fault accommodates sinistral motion between the Arabia plate and the Sinai subplate (Fig. 1), transferring the opening at the Red Sea

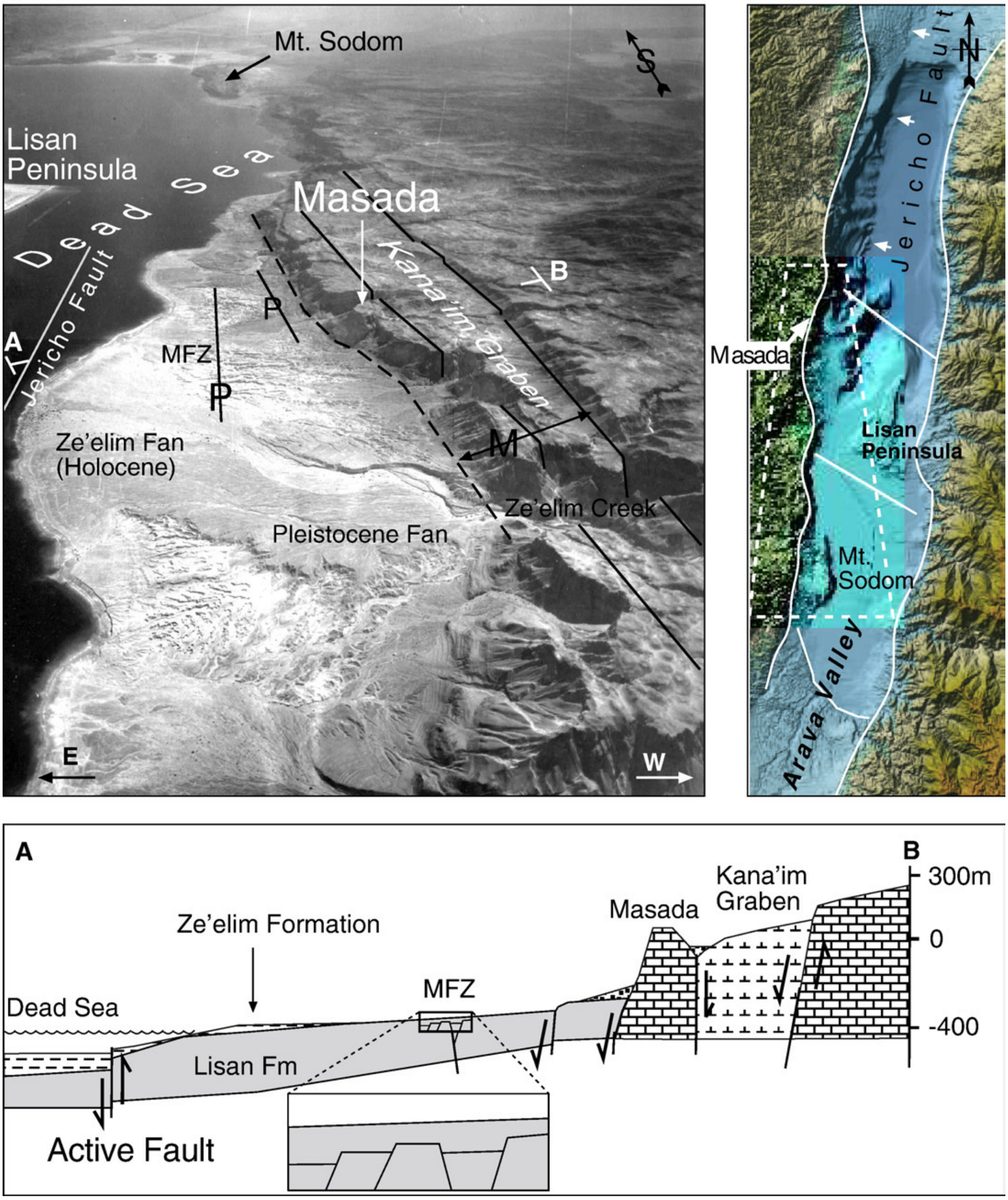

Fig. 2. Top right: The Dead Sea basin, shaded relief map from Plate XI in Hall et al. (2005). The Jericho Fault (white arrows) is the active branch where the steep western slope meets the relatively flat bottom of the lake. The eastern and western margin normal faults, which form formidable cliffs, as well as three transverse NW-striking faults are shown in white. Dashed white line marks the approximate extent of the photo on the left. Left: An oblique airphoto taken in the 1940s showing faults (solid lines) and scarps on the western margin of the Dead Sea basin. The fault at the bottom of the main cliff (dashed) is overlain by the Lisan Formation, which is not faulted there. MFZ is the syn-Lisan Masada Fault Zone (Marco and Agnon, 1995). M marks Miocene faults, P is for Pleistocene–Recent. Bottom: An E–W section showing the early Kana'im Graben west of Masada, Pre-Lisan Formation faults in the graben east of Masada, syn-Lisan faults in the Masada Fault Zone, and a post-Lisan fault offsetting the Dead Sea deposits at the bottom of the lake. No vertical exaggeration.

to the Taurus–Zagros collision zone. No pre-Miocene precursors neither rejuvenation of earlier structure was proved (Garfunkel and Ben-Avraham, 2001). Hence, the general location of the DST was established in the Miocene. The interpretation of left-lateral shear along the DST since the Middle Miocene is based on observations from four independent sources: regional plate tectonics, local geology, seismology, and geodesy. The

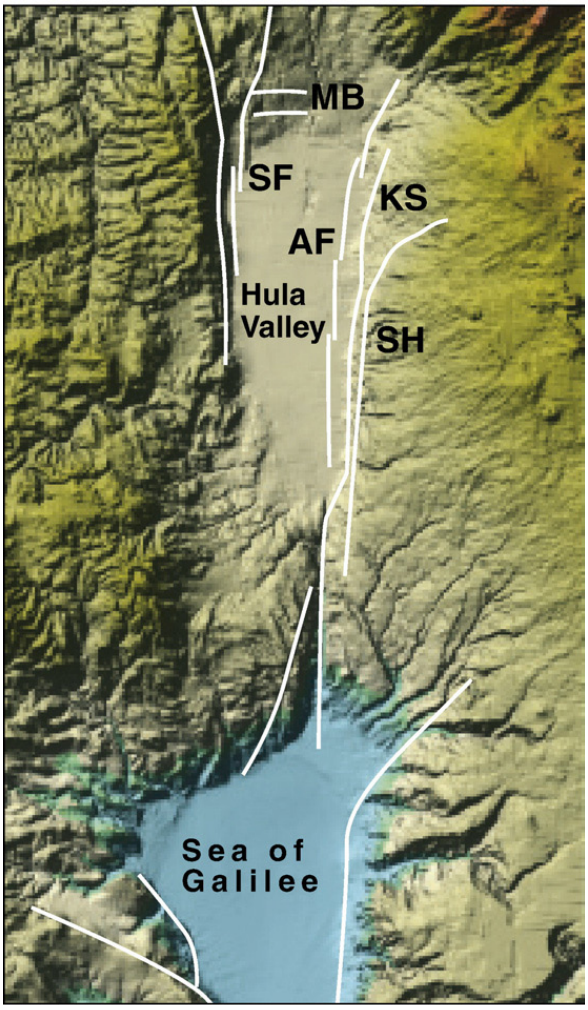

Fig. 3. Faults around the Hula Valley. The age of the northern E–W striking border fault, Maayan Baruch Fault (MB), is constrained by displaced travertine and an unfaulted basalt flow. The basalt, which was K/Ar-dated at 0.88±0.15 ma (Heimann and Ron, 1987), marks the end of activity of the Maayan Baruch Fault. The major activity of the eastern border Shamir Fault (SH), which created a 400-m-high scarp, predated a 2-ma-old basalt flow. Younger activity of the Kefar Szold Fault (KS), west of the Shamir Fault, offset a 0.4 ma basalt flow. Two younger N–S striking faults, Azaz (AF) and Shehumit (SF), displace Late Pleistocene sediments closer to the center of the valley. The Shehumit Fault displaces a Late Pleistocene conglomerate unit. The Azaz Fault displaces by 60 m a travertine unit whose C14 age is 25±0.8 ma (Heimann and Ron, 1987). Younger Early Holocene displacements of the Azaz Fault were dated using OSL (Zilberman et al., 2000). Shaded relief map from Plate XI in Hall et al. (2005).

plate tectonics shows that the opening of the Red Sea, where the Arabian plate is breaking away from Africa, is transferred to the collision with Eurasia via sinistral shear along the DST (Quennell, 1956; Freund, 1965; Garfunkel, 1981; Joffe and Garfunkel, 1987). Sinistral motion explains the systematic offset of numerous pre-Miocene geologic features by a total of ∼105 km (Quennell, 1956; Freund, 1965; Bartov et al., 1980). Historic and prehistoric seismicity were associated with sinistral offsets (Ellenblum et al., 1998; Klinger et al., 2000; Niemi et al., 2001; Meghraoui et al., 2003). Focal mechanisms of moderate-to-large earthquakes show sinistral motion along the DST (Salamon et al., 1996; Baer et al., 1999; Klinger et al., 1999). And finally, geodetic measurements are consistent and confirm the other evidence of 4±1 mm/yr sinistral shear (McClusky et al., 2003; Wdowinski et al., 2004; Reilinger et al., 2006), suggesting stable tectonic regime. The complex geometry of the fault is apparent in pull-apart grabens, which are associated with releasing bends, and pressure ridges that formed where restraining bends occur. Garfunkel (1981) maintains that the pull-apart basins are all shorter than 105 km, the total sinistral offset, because they began to form at a later stage, after some motion had already accrued. This view is supported by seismic surveys that reveal earlier buried basins, which are no longer active (Frieslander, 2000). A different view is presented by Eyal et al. (1981), who do not link the pull aparts lengths with their ages. Various evolutionary schemes of pull-aparts are discussed is several papers (e.g., Mann et al., 1983; Wakabayashi et al., 2004).

Several authors noted explicitly that the detailed shape of the DST had changed through time (Garfunkel, 1981; Heimann and Ron, 1987; ten-Brink and Ben-Avraham, 1989; Rotstein et al., 1992; Heimann and Ron, 1993; ten Brink et al., 1999). The widest zone of about 50 km of distributed faulting is found in the Galilee, where the early-stage (Miocene) faults were associated with formation of basins (Freund et al., 1970; Shaliv, 1991) and with rotation of rigid blocks about sub-vertical axes (Ron et al., 1984), although the linkage to the transform movement is not well established. Subsequent post-Miocene deformation took place mostly in the form of normal faulting on E–W trending faults and the transform movement is currently localized in a very narrow zone. The deformation in the south is characterized by a 20–30-km-wide zone with primarily strike–slip and some normal slip on faults trending sub-parallel to the main transform fault.

In the following sections, I describe several examples of changes in the Dead Sea fault zone geometry.

1.2. The Dead Sea Basin western fault zone

A north–south trending fault zone was found in outcrops of the late Pleistocene Lisan Formation, about 1 km west of the Dead Sea shore near Masada, the 2000-yr-old Jewish rebel stronghold (Fig. 2). Most of the Masada Fault Zone (MFZ) planes strike north, paralleling the main graben faults and morphological trends.

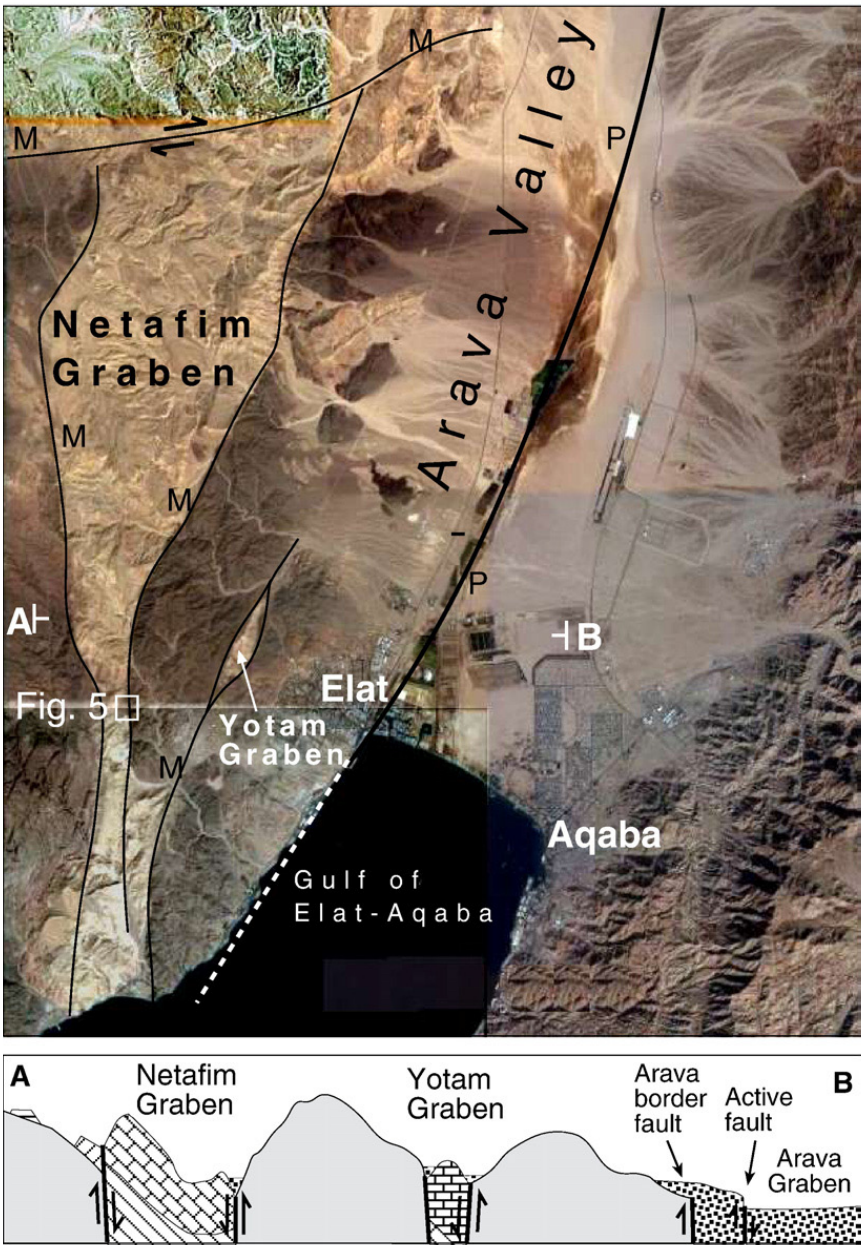

Fig. 4. The main faults in the Elat region show characteristic North-striking graben-and-horst structure that formed since the Miocene. On the Google Earth mosaic Precambrian crystalline basement outcrops appear dark and carbonate sediments, mostly of Cretaceous age are light-beige. Thin lines mark inactive faults, the active Arava fault is solid black. After Garfunkel (1970). The schematic section A–B shows that the oldest faults are not associated with significant topographic differences, elevations of the Netafim Graben and the horsts on both sides are the similar. Younger faults that formed the low Arava Valley are overlain by unfaulted Plio-Pleistocene clastic units and alluvial terraces. The youngest active fault, which offsets Pleistocene alluvial terraces, is in the valley. M—Miocene faults, P—Pleistocene–Recent. After Garfunkel (1978).

The faults are overlain by continuous horizontal layers of the Lisan Formation, indicating that they are syndepositional. The documented slip events on some of these faults provide a paleoseismic record (Marco and Agnon, 1995; Marco et al., 1996; Marco and Agnon, 2005). The MFZ planes dip 50° to 70° eastward as well as westward, with normal displacements up to 2 m. The average strike is 360°. In the absence of suitable markers, it is impossible to document horizontal slip. However, horizontal slip is probably negligible because the prevailing dips are typical of normal faults, and because none of the typical deformations that often accompany strike–slip is present (e.g., push-up swells or rhomb grabens near changes of strike). An upper bound of the maximum E–W extension across the 300-m-wide fault zone is estimated by considering 6 fault planes dipping 50° to 70° with 2 m vertical slip on each. This estimate yields 1.4%–3.3% extension.

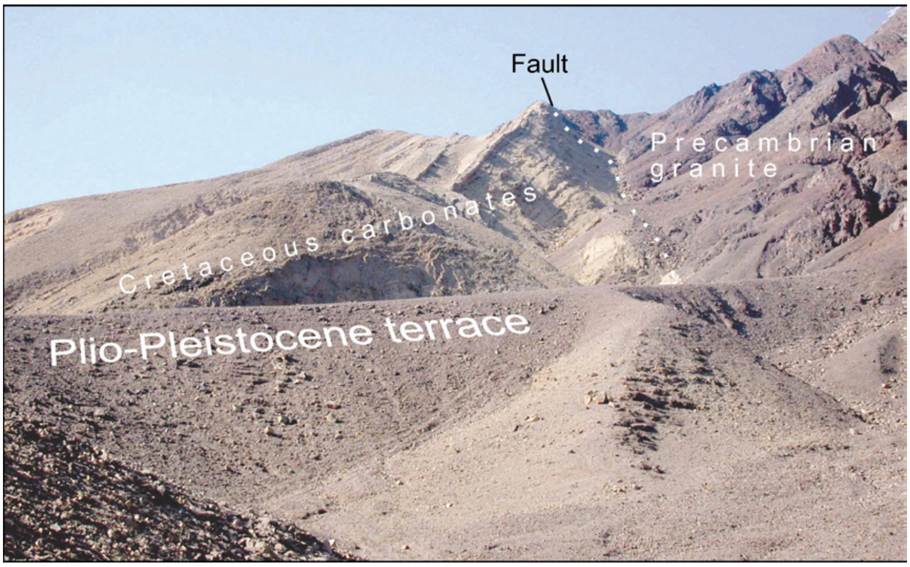

Fig. 5. Photo showing an unfaulted terrace which overlays an early stage, currently inactive fault (location marked on Fig. 4). The terrace contains large diorite cobbles derived from a pluton now separated from this outcrop by a 200-m deep canyon. The age of the terrace is estimated as Plio-Pleistocene (Garfunkel, 1970).

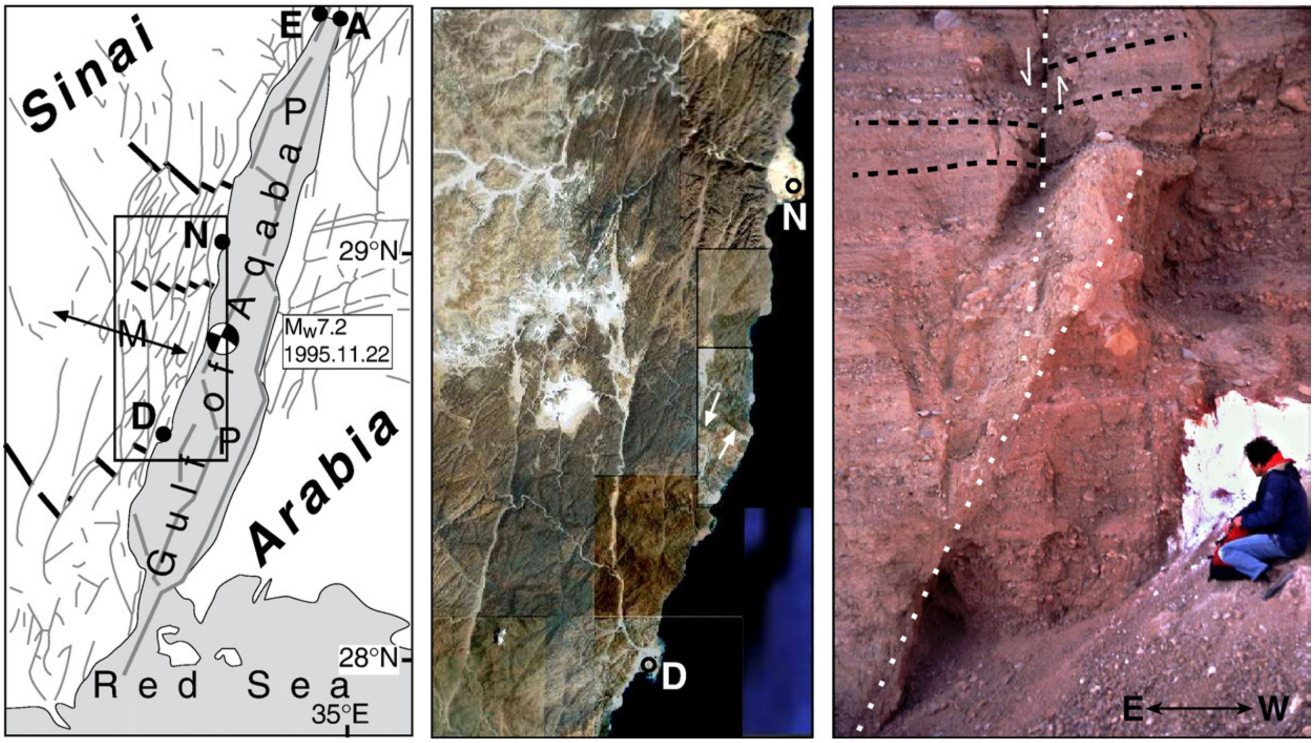

Fig. 6. Left: A ~25-km-wide zone in the eastern Sinai by the Gulf of Aqaba is crossed by several sub-parallel NNE-striking faults (Eyal et al., 1980). Basaltic dikes (solid bold lines) of early Miocene age are displaced left-laterally (Eyal et al., 1981) but the currently active faults (solid grey) are at the center of the gulf (Ben-Avraham, 1985). The focal mechanism and associated deformation of the M_w7.2 earthquake of November 22, 1995 confirm the sinistral nature of the faulting in the gulf (Baer et al., 1999; Klinger et al., 1999; Shamir et al., 2003). Letters mark the locations of the towns of Elat (E), Aqaba (A). M marks Miocene faults, P marks Pleistocene–Recent (Eyal et al., 1980). Rectangle shows location of satellite image. Center: A Google Earth mosaic that shows incision of shore-parallel wadis along fault traces and an example of sinistrally-offset pluton boundary (white arrows). Nuweiba—N, Dahab—D. Right: Faults (white dots) offset Quaternary fan delta alluvium on the shore of the gulf in Nuweiba, east of the inactive faults of eastern Sinai. Examples of displaced horizons are emphasized with black dashed lines. White rock on the right is Early Cretaceous sandstone.

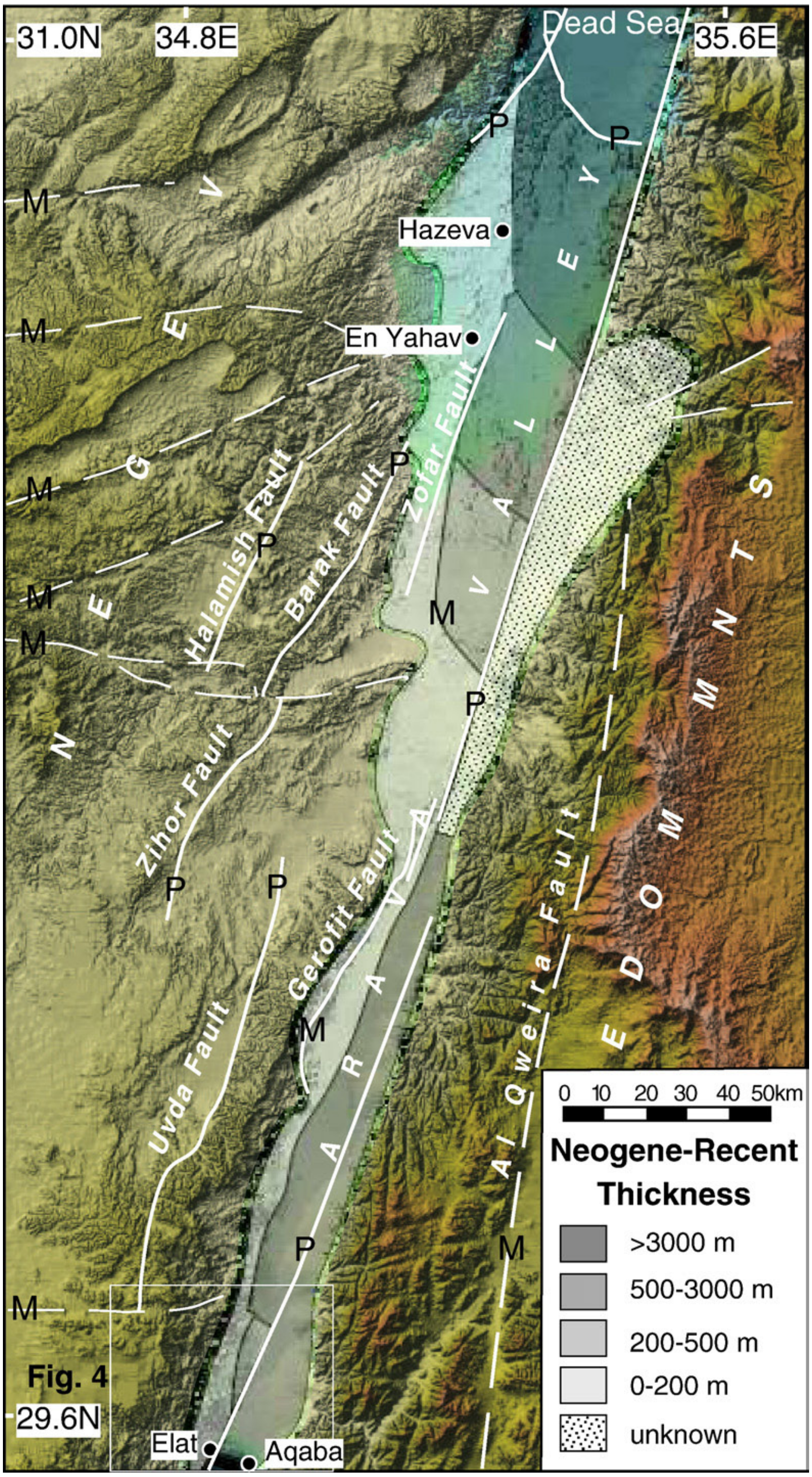

Fig. 7. Subsurface basins in the Arava Valley (after Frieslander, 2000) show different fault patterns than the currently active faults (solid white), which offset Pleistocene–Recent alluvial fans in the valley (after Garfunkel et al., 1981). Other faults, Barak, Halamish, Zihor, and Uvda, which offset Pleistocene units, have been documented west of the Arava (Avni et al., 2000). All these faults have become active after the now buried basins became inactive. Other strike–slip faults like the Al Quweira are also inactive at present. East-to-northeast trending faults (dashed) are part of a dextral shear belt, which is displaced 105 km by the Dead Sea fault system (Quennell, 1956). M marks Miocene faults, P marks Pleistocene–Recent faults. Shaded relief map from Plate XI in Hall et al. (2005).

If the dips of the fault planes do not change at depth, the fault planes that bound ~100-m-wide grabens should intersect at less than 300 m below the surface. It is highly likely that the observed faults merge downward to form a single fault and they are the upper part of a "negative flower structure", accommodating some E–W extension as suggested by gravity models and subsurface data (Kashai and Crocker, 1987). Similar structures are observed in seismic images across the DST north of Lake Kinneret (Rotstein and Bartov, 1989) as well as other strike–slip faults (Sylvester, 1988). The geometry and extent of the fault planes as they appear in high-resolution seismic reflection and ground-penetrating radar indicate that they are tectonic features that accommodate E–W extension (Agnon et al., 2006). However, low heat flow on the order of 38–42 mW/m^2 (Ben-Avraham et al., 1978; Eckstein, 1978) and absence of magmatic activity indicate that the overall extension across the Dead Sea fault zone is small.

An active fault, the Jericho Fault, is recognized in the bathymetry of the Dead Sea (Fig. 2), where the steep slope on the west meets the flat bottom along a straight line about 3 km east of the Masada fault zone (Neev and Hall, 1979; Hall, 1996). Seismic profiles confirm that this break in the slope is a fault (Ben-Avraham et al., 1993). Frieslander and Ben-Avraham (1989) suggest that the continuity of magnetic anomalies over the western part of the Dead Sea, in contrast to their abrupt termination on the eastern margin of the basin, indicates that the faulting in the western part of the basin is mainly normal. Although they suggest that the major strike–slip motion takes place on the fault in the eastern side of the Dead Sea basin, direct submarine observations (Lazar and Ben-Avraham, 2002) reveal a sharp vertical active fault scarp at the northern end of this line, confirming Neev and Hall's (1979) interpretations.

Garfunkel et al. (1981), as well as other authors, noted that the location of the Dead Sea basin margin faults have changed through time. ten-Brink and Ben-Avraham (1989) suggest that the basin has widened westward by the collapse and tilting of margin blocks. They also suggest northward migration of the transverse faults of the basin in the Pleistocene, which results in northward growth of the basin. An E–W seismic reflection image about 10 km north of the Dead Sea basin is interpreted to show early-stage inactive faults that are overlain by graben fill on the west and a single active fault, which offsets the fill, near the center of the basin (Kashai and Crocker, 1987). This structure, together with gravity data, led ten Brink et al. (1999) to suggest that it is the result of continuous changes in relative plate motion. A different evolution is suggested by Shamir et al. (2005) and Shamir (2006), who maintain, based on seismic reflections and earthquake epicenters relocated relative to controlled detonations, that the shear in this section of the DST evolved from an early, probably Miocene–Pliocene, stage of localized strike–slip motion primarily along the Jericho fault to a late stage (Pliocene–Recent) when shear has been distributed over internal fault sets. Some micro earthquakes align with secondary faults, which have either minor or no bathymetric expression, all are located within the deep part of the basin and none is associated with the MFZ neither with the main basin boundary fault. However, the absence of alignment of epicenters with the Jericho Fault and Lazar and Ben Avraham's (2002) submarine observations might indicate that it is currently locked rather than inactive.

The faults of the MFZ do not displace the top of the Lisan Formation (Marco and Agnon, 1995). The uppermost few meters of the Lisan Formation were deposited after faulting stopped and migrated ~3 km eastward. Seismic profiles in the Dead Sea reveal thinning of recent sediments toward fault segments (Ben-Avraham et al., 1993), showing a similar pattern of active syndepositional fault scarps that are larger than the Masada fault scarps. A single small N-striking normal fault is found between the MFZ and the major boundary cliff (Fig. 2). It displaces the top of the Lisan Formation by 2–3 m. The margin faults, which bound the graben and form the 300-m-high escarpment have been dormant in the post Lisan time. This is evident in the unfaulted Lisan beds that fill the creeks and are continuous across the escarpment (Garfunkel et al., 1981). Another graben, called Kana'im, is found west of Masada. This graben appears to be inactive with a very mild topographic expression, shows no signs of recent activity, and therefore predates the Dead Sea graben. An E–W cross section shows the migration of the major activity toward the center of the graben (Fig. 2).

1.3. Hula Valley faults

The tectonic evolution of the Hula pull-apart basin was studied by Heimann and Ron (1987) and Heimann (1990). Subsidence in the Hula Valley occurred by normal faulting around it since the Pliocene (Garfunkel et al., 1981). A group of faults that barely offset the flat Holocene soil of the valley (Fig. 3) appears to be very young, whereas the activity of the border faults, which formed the relief, predated the activity of the faults that displace the valley fill (Heimann and Ron, 1987). The present active zone is narrower than the earlier stage, similar to the Dead Sea Basin.

1.4. Southern Arava (Elat region) faults

The southern Arava (Elat region) is a deep valley, extending northward from the tip of the gulf of Aqaba. Its western bound is a fault-controlled escarpment, comprised of a series of sub-parallel normal faults. North-striking graben-and-horst structure characterizes the Late Cenozoic structure of the region (Fig. 4). Despite vertical offsets of up to 1.5 km, the topographic elevations of the horsts and graben are essentially the same, preserving evidence of a post-faulting erosional surface (peneplain) stage (Garfunkel, 1970). Pleistocene terraces, which overlay the marginal faults (Fig. 5), are not faulted (Eyal, 1973). The peneplains faulted farther to the east by the Arava border fault, which is overlain by Quaternary unfaulted clatic units that fill the main valley. The major activity of the graben-and-horst faults west of the Arava postdated the Middle–Miocene Raham Conglomerate, a coarse conglomerate unit, which in the Elat region is comprised of mainly Eocene limestone and chert pebbles. The Raham conglomerate is a key unit because it was formed together with the onset of faulting and predates the major displacement on the margin faults. The Middle–Miocene age of the Raham Conglomerate was estimated stratigraphically (Garfunkel et al., 1974). Its tectono-stratigraphic position resembles that of the Hazeva Formation in southern Israel (Garfunkel et al., 1974; Calvo et al., 2001) and the Hordos Formation in northern Israel. The Hordos Formation was radiometrically dated using intercalated basaltic flows to Early–Middle–Miocene (Shaliv, 1989).

East of the main Arava border fault escarpment, closer to the center of the valley, a swarm of normal faults offsets the Quaternary alluvial fans (Porat et al., 1996; Amit et al., 1999). The youngest fault is farther east at the center of the valley, where the 1068 AD earthquake rupture was identified (Amit et al., 2002). Hence, this section of the DST also exhibits centerward migration of faulting, much like the Dead Sea and Hula Valleys.

1.5. Southeast Sinai faults

The eastern coast of the Sinai Peninsula at the Gulf of Aqaba is crossed by several sub-parallel NNE-striking faults (Fig. 6). Contacts between various magmatic and metamorphic rock units of the Precambrian crystalline basement provide piercing points for measuring offsets on these faults. NW-striking Miocene basaltic dikes, K–Ar-dated to 22–19 ka, are also offset sinistrally by the same amount. This observation indicates that the whole sinistral movements on these faults postdates the Miocene dikes (Steinitz et al., 1978; Eyal et al., 1981). The Pleistocene alluvium in the same area is not faulted, suggesting that the faults became inactive. The steep bathymetry of the gulf margins, faulted young fan delta deposits on the shore (Fig. 6) and coral reefs (Reches et al., 1987; Shaked et al., 2002, 2004), seismic reflections (Ben-Avraham et al., 1979; Ben-Avraham and Garfunkel, 1979), and epicenter locations (Baer et al., 1999), all indicate that the currently active faulting is at the center of the gulf.

1.6. Central Arava faults

The evolution of the Arava Valley is more complex than the surface exposures reveal. Seismic imaging reveals several deep fault-bounded basins (Fig. 7), most of which are overlain by continuous sediments (Frieslander, 2000). In contrast, the surface mapping appears to show a single continuous sinistral active fault accompanied by secondary normal faults, which offset Late Pleistocene–Recent alluvial units (Zak and Freund, 1966; Garfunkel et al., 1981). Pleistocene–sub Recent activity was also documented farther north on the same fault (Klinger et al., 2000; Niemi et al., 2001). The structure of the Arava section of the DST in the Miocene was somewhat similar to the current Dead Sea. The subsiding basins accumulated sediments until their bounding faults became inactive, perhaps in the Pliocene, resulting with the filling and subsequent burial of the basins. The thickness of the main fill unit, the Miocene Hazeva Formation, is over 2.5 km (Calvo et al., 2001). The overlying Quaternary sediments are faulted in a different manner. The timetable of the changes is not well constrained. The main active fault, which strikes 020°, offsets Quaternary alluvial fans but it is not clear when it became active. The Barak Fault, which strikes 035°, offsets the Arava Formation dated at about 1 ma (Avni et al., 2000). The morphology of the fault scarp suggests minor Holocene activity but a cluster of microearthquakes and InSAR-detected deformation were recently located near it (Finzi, 2005). Subsurface data show that the vertical offset of the Barak Fault is much smaller than that of other faults in the Arava (Frieslander, 2000), indicating either very low slip rate or very short activity time. Similar features are exhibited by the Zihor, Uvda, and Halamish faults (Fig. 7).

Inactive N-striking faults east of the Arava exhibit sinistral displacements of Precambrian igneous units. A sinistral offset of 40 km along the Al Quweira Fault was deduced from the displacement of distinctive andesitic rocks found on both sides of the fault (Barjous and

Mikbel, 1990). Field survey of the surface alluvium and examination of satellite images do not reveal any fresh fault scarps or other indication of recent faulting.

Hence, the Arava fault zone seems to have changed from wide (Miocene) to narrow (Plio-Pleistocene) and then again to wide in the Late Pleistocene.

2. Discussion

Strike–slip fault zones usually include also normal faults, where releasing bends form pull-apart grabens. Most pull-apart grabens in the world are shorter than the total horizontal offset of the main fault. Hence, the changes in the geometry, which brought about the formation of the grabens postdated the initiation of the faulting. Changes in the geometry of plate boundaries is also a direct outcome of the kinematics of plate motions (Garfunkel, 1975). These changes manifest fundamental properties of the faulting process. They can be described by comparing the recent fault patterns to the more ancient ones. The location of recent seismicity along the DST is largely in agreement with the geological mapping of active faults (Hofstetter et al., 2007) but is not readily correlatable with individual strands within the fault zone. Microseismic activity, which is recorded within a wide zone (Fig. 8) either reflects the activity of a wide fault zone with many branches, many of which are not expressed at the surface (e.g., Basson et al., 2002) or a large scatter due to uncertainty in the epicenter locations. I therefore consider active the ones that offset Pleistocene–Recent units.

Observations in the Dead Sea basin show that it is a pull-apart graben, bounded by normal faults, whose activity migrates centerward. Microseismicity and deformed Holocene and sub-recent sediments, including an active fault scarp, are confined to the central part of the basin (Lazar and Ben-Avraham, 2002; Shamir, 2006). In contrast, the main graben boundary normal faults are overlain by un-faulted Late Pleistocene sediments (Garfunkel et al., 1981; Marco and Agnon, 1995). There are two possible scenarios that can explain this structure. In the first, the entire fault zone is active at the same time, i.e., all or most of the faults act simultaneously but with different rates. The different rates of the vertical slip components are expressed in the topography, the higher the rate the larger the topographic expression. In the second scenario, activity is limited to the bounding faults, which becomes inactive as time goes by while new ones form closer to the center of the basin. This scenario portrays initially wide pull-apart grabens. The grabens become narrower and longer as slip proceeds through time. Ultimately, in the final stage, strike–slip movement at the center of the basin tears it apart by the sediments that fill the basin ultimately burying it completely. Buried inactive basins that have been documented by seismic reflections below the Arava Valley resemble Wakabayashi et al.'s (Wakabayashi et al., 2004) model F (their Fig. 8).

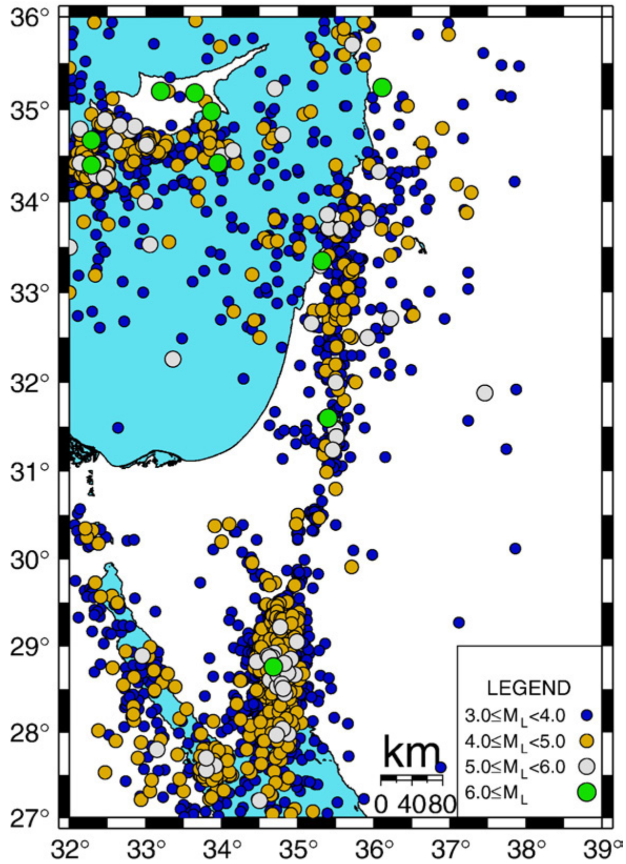

Fig. 8. Epicenters in the Middle East, 1904–2006 [www.gii.co.il].

The second scenario is compatible with the observation of Late Pleistocene alluvial terraces and lake deposits that overlay the faults at the margins but are not faulted, indicating that the faults are currently inactive. However, there is no evidence that the faults at the center were not active together with the marginal faults in earlier stages.

Similar centerward migration of faulting has been documented in the pull-apart grabens along the DST in south Sinai–Gulf of Aqaba (Eyal et al., 1981), in the Elat region (Garfunkel, 1970; Eyal, 1973), and in the central Arava (Frieslander, 2000). The faulting along these segments of the DST began over a wide belt and later became localized. Important exception is the group of faults in the eastern Negev, which offset Pleistocene terraces (Fig. 7). These include the Uvda, Zihor, Halamish, and Barak faults (Avni et al., 2000). This example shows a recent widening of the fault zone, perhaps the onset of another cycle of faulting in a wide zone. The small offset in the subsurface of the Barak Fault (Frieslander, 2000) indicates that it is a new fault and not a reactivated older one. In that case, the localization is only a temporary trend in the evolution of the fault zone and periods of widening may occur intermittently with narrowing. Shamir et al. (2005) suggest that widening occurred also north of the Dead Sea but they base this claim on subsurface data, whereas only one fault appears to offset the Holocene surface (Garfunkel et al., 1981; Reches and Hoexter, 1981; Gardosh et al., 1990).

The width of the initial stages is hard to determine for the whole fault. The Precambrian basement exposures in the Sinai and east of the Gulf of Aqaba–Arava section exhibit about a 25-km-wide belt with abundant fault branches. Other sections where mostly Cretaceous sedimentary units are exposed exhibit narrower zones. However, this difference might be misleading because the sedimentary sequence is capable of dissipating significant amount of displacement. Exposed example for this is observed in sections in the Elat region that show disharmonic structures, where faults in the crystalline basement become folds in the overlying sediments (Garfunkel, 1970).

The precise timing of shifts in activity is poorly constrained. A significant reorganization of the plate motions is manifested in a shift of Sinai–Arabia Euler poles from 32.7°N/19.9°E for the total motion (since early Miocene) to 32.8°N/22.6°E for Pliocene to Recent faults and pre-Pliocene (Joffe and Garfunkel, 1987). Plio-Pleistocene alluvial fans in the Arava are displaced sinistrally 15–30 km (Ginat et al., 1998), whereas the Al Queira fault east of the Arava accumulated 40 km of displacement. These data are compatible with a major shift the location of faulting that occurred approximately at the end of the Miocene or the early Pliocene. The activation of a wider fault zone in the Negev took place in the Pleistocene (Avni et al., 2000).

Processes of plate subduction, collision, and accretion are associated with changes in plate motions and plate shapes. These changes alter the stress field near plate boundaries, which drive changes in the geometry of the plate boundary faults. In the West Antarctic Rift system the transition from broad to focused extension was suggested to be the result of a changes in plate motions and/or thermal regime (Huerta and Harry, 2007). Similar reasoning is also advocated by ten Brink et al. (1999) for explaining the multiple fault structure of the Dead Sea basin. A plausible mechanism for slip-dependent change in the detailed shape of a fault zone is work hardening, i.e., the strength of the crust increases as fault slip grows. The result is localization of faulting in narrow belts until another change occurs in the plate motions and another cycle begins. A 3-D model that fits the spatial and temporal features of several rift systems shows diffuse stretching at the onset of rifting that is followed by local necking (Agnon and Eidelman, 1991). Cowgill et al. (2004b) suggest that changes in the topography due to thrust faulting in restraining bends along strike–slip faults affect the evolution of the fault zone. Although the examples in the present study refer primarily to releasing bends this hypothesis may be examined when data on subsidence rates as well as detailed accurate plate motion records are available. A significant correlation exists between major events at the collision zone of Arabia and Eurasia and the DST. The continental collision and suturing of Arabia with the Turkish terrains in the middle Miocene resulted in the complete annihilation of subduction of the Neo Tethys ocean in eastern Turkey and at the same time the DST was formed (Bozkurt and Mittwede, 2001). The collision between Arabia and Africa has accelerated the convergence of Arabia and Eurasia in the early Pliocene time (Bozkurt and Mittwede, 2001). This apparently led to the development of the North and the East Anatolian faults in the early Pliocene (Bozkurt and Mittwede, 2001; Kocyigit et al., 2001). The northernmost segment of the Dead Sea fault joined the East Anatolian fault in southern Turkey forming the Anatolian–Arabian–African triple junction (Karig and Kozlu, 1990). Between the late Miocene when continent–continent collision began and the Early Pliocene changes occurred in the tectonic regime, basin type and deformation pattern, e.g., from folding and thrusting to strike–slip faulting (Kocyigit et al., 2001). The development of these fault systems provided the mechanism for the tectonic escape of the Anatolian block toward the Aegean arc. Hence, the relative plate motions about the new DST post-Miocene faults were associated with a new Euler pole of rotation (Joffe and Garfunkel, 1987) and also correlate with major rearrangement of the adjacent plate boundaries.

3. Conclusions

The faulting along the DST began in the early–middle Miocene over a wide zone of up to 50 km and later by the end of the Miocene became localized. The subsidence of fault-controlled basins, which were active in the early stage, stopped. New faults were formed in the Negev west of the main transform during the Pleistocene. They indicate that another cycle has begun involving widening of the fault zone.

It is suggested that the localization of faulting takes place as long as there is no change in the stress field. When the geometry of the plates changes due to their movements, the stresses change, the localization stage ends, and the fault zone is rearranged. It becomes wide and a new localization process develops as slip accumulates.

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