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The East Anatolian Fault: an oblique collisional belt

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ABSTRACT

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The tectonic setting for the area around the Arabia–Africa–Anatolia triple junction is described from combined Landsat–SPOT satellite image analysis and field observations. Since the Late Miocene the motion along the East Anatolian Fault generated major structures corresponding to shortening with a sinistral-slip component. The 150 km wide area of the triple junction is highly deformed by folding and thrusting, distributed mainly on the border of the Anatolian block. In addition, kinematic reconstructions confirm that the Arabia–Anatolia relative motion is both sinistral and convergent. The sinistral strike-slip faults of the East Anatolian Fault zone are of second-order and local consequences of the N–S Arabia–Anatolia collision.

Introduction

In contrast to oceanic plates, where the deformation is concentrated in narrow zones, convergent continental plates display wide areas of deformation along their boundaries. The various aspects of deformation occurring along the edges of continental and oceanic plates are expressed by the different rheological behaviour of these two domains (McKenzie, 1972). Molnar and Tapponnier (1975) have demonstrated that the continental crust undergoes deformation more easily than oceanic crust.

The deformation observed near the contact of three continental plates, Anatolia, Arabia and Africa, is largely the result of the convergence of the African and Arabian plates with respect to the Anatolian, where the relative motions are convergent and/or lateral. This paper aims to shed light on the deformation which has occurred in the vicinity of the Anatolia–Arabia–Africa triple junction, since the Late Miocene.

The fragmentation of the African craton during the Oligocene gave rise to the opening of the Red Sea and the Gulf of Suez, and later the Gulf of Aqaba, separating Arabia from Africa (McKenzie, 1972; Cochran, 1981; Izzeldin, 1987; Bayer et al., 1988). The northward motion of Arabia with respect to Africa was essentially taken up by the Dead Sea Fault (DSF). The DSF is an approximately 1000 km long, left-lateral strike-slip fault (Fig. 1), extending from the Red Sea in the south up to the East Anatolian Fault (EAF) in the north and passing through the Gulf of Aqaba and the Dead Sea. It forms the northeastern boundary of Arabia (Fig. 1a) and was activated during the Late Miocene, after the opening of the Gulf of Suez and when the motion of Arabia relative to Africa jumped to the Dead Sea Fault (Lyberis, 1988; Steckler et al., 1988). Thus, the individualization of Arabia was accomplished by the formation of the DSF.

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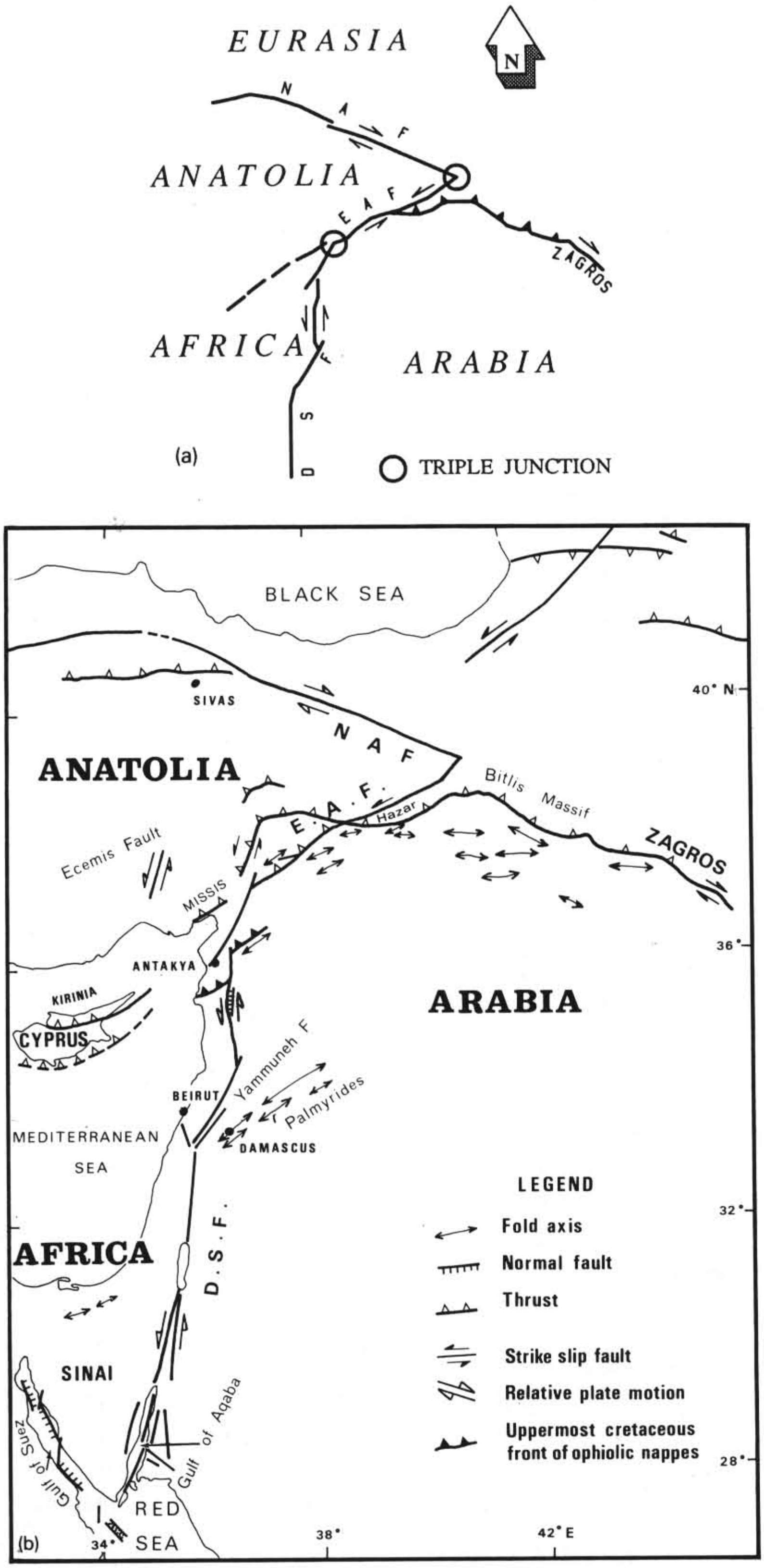

Fig. 1. (a) Sketch of plate boundaries in the eastern Anatolia. (b) General tectonic setting near the Dead Sea Fault (DSF), the East Anatolian Fault (EAF) and the North Anatolian Fault (NAF).

During the Late Oligocene to Middle Miocene, the northern margin of Arabia formed the submerged southern border of the Tethyian Ocean, connecting the Mediterranean to the Indian Ocean. In the Burdigalian, the Miocene transgression reached its culmination. The Arabian shield, the Taurus ranges, eastern Anatolia and parts of central Anatolia were submerged. After the Serravallian, the sea retreated and the northern Arabian shield emerged during the Pliocene–Quaternary interval (Lüttig and Steffens, 1976). Basaltic volcanic activity, which started during the Late Miocene became widespread in Late Pliocene–Pleistocene time (Lüttig and Steffens, 1976).

In the northeast of the Arabian Platform the intercontinental collision along the Bitlis-Zagros suture took place in a 12–14 Ma period (Falcon, 1974). Important volcanic activity, comparable with that of the northern Arabian shield (Giannerini et al., 1988), occurred during this period in central Iran (Berberian and Berberian, 1981).

Following the Eocene Pontide collision, and preceding the Late Miocene, the northern part of the Anatolian block (Black Sea border) formed the southern border of Eurasia. The southern limit of the Anatolian block is defined by a Late Cretaceous mountain range, stretching from the Aegean Sea in the west to the Zagros belt in the east. When the Arabian shield impinged on the southern border of Eurasia in the Late Miocene Anatolia was already a discrete tectonic unit.

The northward motion of Arabia gave rise to the westward escape of the Anatolian block along two intracontinental transforms, the North Anatolian Fault (NAF) in the north, and the East Anatolian Fault in the southeast. Right-lateral slip along the NAF is known to have taken place since the Late Miocene (Sengör, 1979). The EAF extends from the northern end of the DSF to the eastern end of the NAF (Fig. 1). The E–W oriented Bitlis compressional zone, which begins at the EAF, continues through to the Zagros uplands.

The EAF has been identified largely on the basis of seismic observations (McKenzie, 1972, 1978; Jackson and McKenzie, 1984). Earthquakes are associated with ENE–WSW oriented faults exhibiting sinistral strike-slip motion (Arpat and Saroglu, 1971; McKenzie, 1976; Hempton, 1983). It is for this reason that the EAF is thought to contribute as a purely left-lateral strike-slip fault, or a transform, to the taking up of the Anatolian westward extrusion and, therefore, forming a conjugate structure with the NAF. However, further studies (Perinçek, 1979; Perinçek and Ozkaya, 1981; Yazgan et al., 1983) have shown that this region is dominated by considerable thrusting and folding.

Stereoscopic studies of Landsat-MSS images combined with field observations have allowed us to establish a structural scheme of the deformation in the western part of the EAF and its junction with the DSF.

The northern part of the Dead Sea Fault (DSF)

The northern end of the DSF is characterised by narrow, N–S oriented, fault-bounded basins such as the Ghab Basin (Fig. 2). The normal-fault component of these structures is easily recognised on the Landsat-MSS images. Further north, at the latitude of Antakya (N36°15′) the DSF system dies out. The Altinözü massif lies east-southeast of Antakya and is limited, in the east and the west, by N–S to NNE–SSW oriented faults. The Mesozoic bedrock of the massif is overlain by Eocene carbonates followed by Upper Miocene to Pliocene marls and reef limestones (Dubertret, 1962). The massif is highly fractured by numerous small normal and oblique-slip faults. The western flank of the massif is bounded by sinistral-normal faults, such as the Harbiye fault, which displays several hundreds of meters of vertical displacement. To the north the Altınözü massif is bounded by the 30 km wide Amik basin with a Pliocene–Quaternary fill. This basin extends northward through the Karasu valley.

The Amanos massif bounding the Amik basin to the north and the Karasu valley to the west (Figs. 3 and 4) consists essentially of an ophiolite complex which has been shown to be continuous with that of the Baër-Bassit massif (Parrot, 1976) on the African plate to the south (Fig. 2). This massif is a post Miocene, NE–SW trending, broad anticline (Tinkler et al., 1981) (Fig. 5A). The

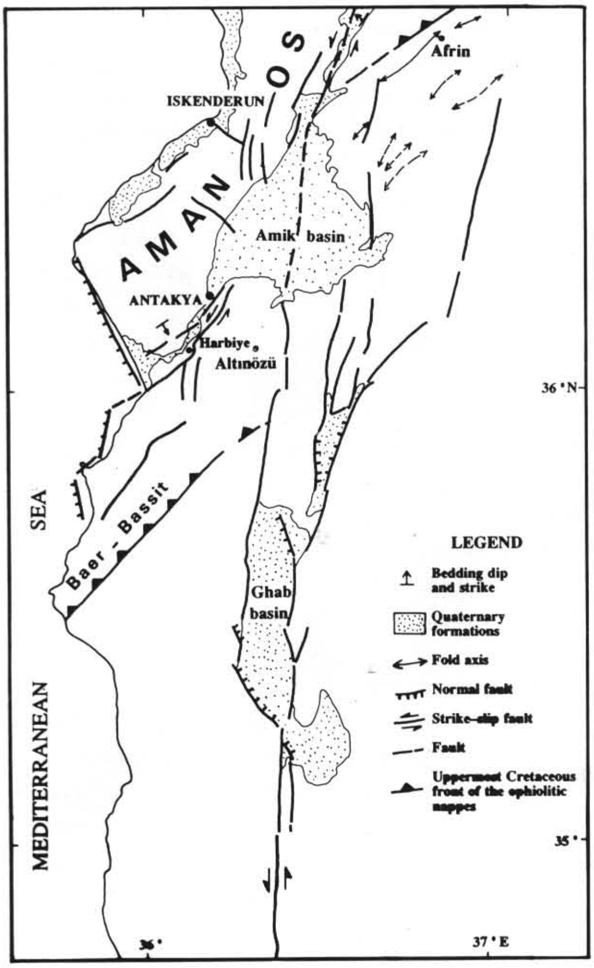

Fig. 2. Structural map of the northern end of the DSF.

boundary between the N25° trending Karasu valley and the Amanos massif is defined by a series of parallel faults (Dubertret, 1962, Hatay and Adana sheets of the Geological map of Turkey 1962; Perinçek et al., 1987) rather than a single fault trace. This fault zone, referred to here as the Amanos Border Fault (ABF), is linked to the left-lateral DSF under the Amik basin.

From the southern margin of the Amik Basin to the Mediterranean Sea, Miocene marine sediments dip to the east-southeast towards the NNE–SSW trending Harbiye-Antakya Fault. Striations on the exposed fault section of this fault show a left-lateral displacement with a minor extensional component. However, north of the Amik basin the Neogene sediments, which have a wedge-shaped geometry thinning eastwards onto the Arabian platform, dip to the west-northwest towards the NNE–SSW trending Amanos massif. The changing dip directions of the Neogene sediments across the Amik basin coincide with the northward projection of the DSF, under the basin fill, to the ABF. Although the DSF is dominated by left-lateral displacement, in excess of 2 km of dip-slip is suggested from the offset of the Eocene sequence in the central section of the ABF.

In the eastern Amanos, NNE–SSW faults, with several meters of sinistral strike-slip, can be observed in the Maastrichtian sediments. The secondary faults cutting the Late Cretaceous and Eocene sediments also display left-lateral displacements. Both the stratigraphic offset of Late Miocene sediments across the Amanos boundary to the south of the Amik Basin and the fault plane striations show NE–SW extension (transtension).

North of the Amik Basin, the kinematic indicators and stratigraphic displacements of the Eocene sediments, along the ABF zone show, in contrast, left-lateral movement associated with significant NW–SE compression, which increases in intensity towards the Maras flysch basin (Figs. 3 and 4). The deformation arising from this continuing compression (Fig. 5A) has much in common with foreland propagating fold and thrust belts.

Thus, the largely transtensional DSF system of the Altinozu massif may be linked to the transpressional ABF by a fault running under the Amik Basin, the presence of which would account for the opposing dip directions of the Neogene sediments from north to south. The intersection of this hidden fault and the ABF is considered to be the location of the Arabia–Anatolia–Africa triple junction.

Structures at the northwestern end of Arabia

The NE–SW Palmyrides fold belt which lies to the north of the Arabian plate, where the DSF meets the southern part of the Bekaa Valley, represents intraplate Miocene compression which reactivated Mesozoic extension faults (Chorowicz et al., 1987). Between the northern margin of the Palmyrides and the Kürtdag massif, the Mesozoic

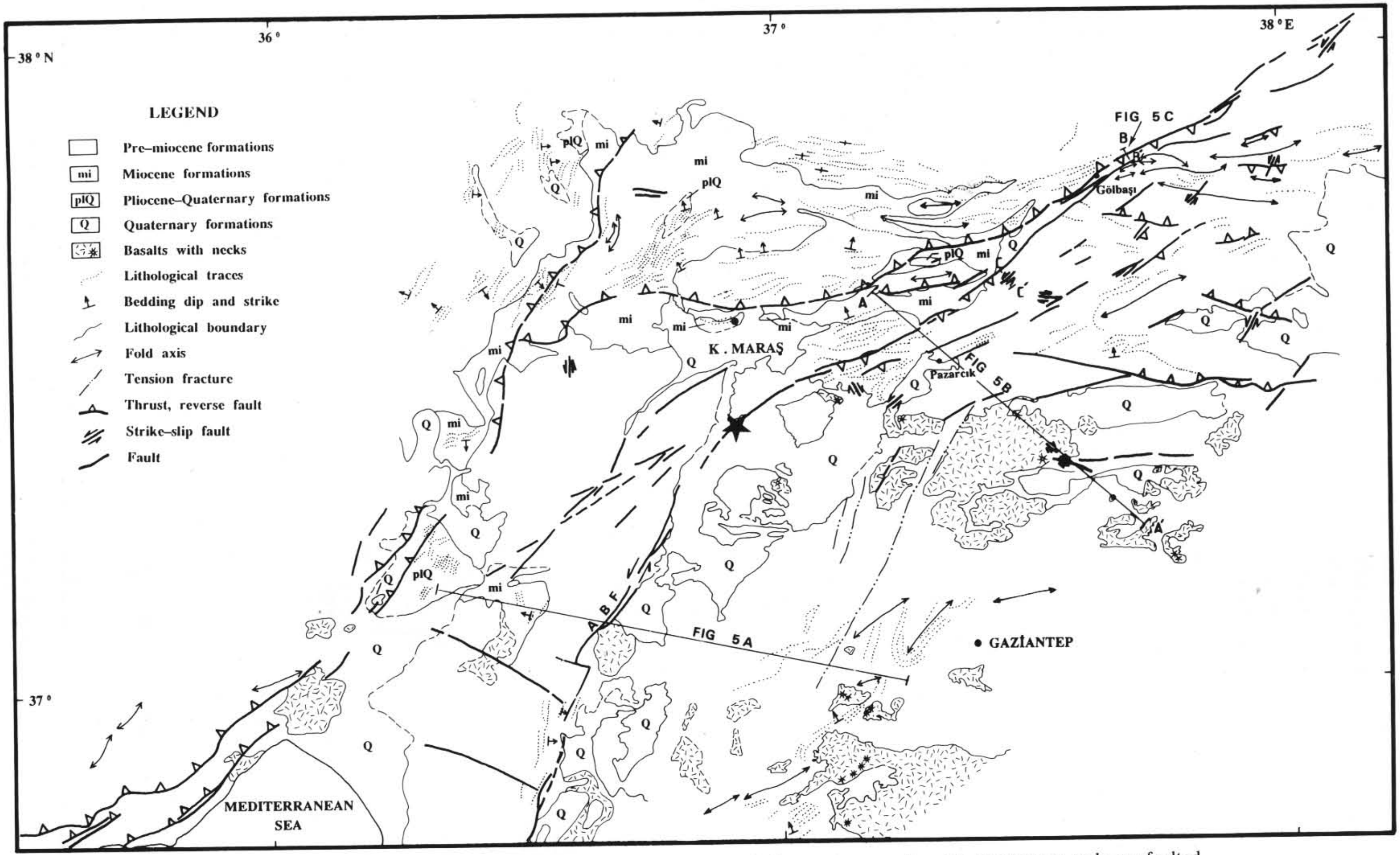

Fig. 3. Structural map of the North Amanos-Maras compressive belt. Star indicates the area where the Quaternary rocks are faulted.

to Cenozoic sediments of the Arabian platform are undeformed (Fig. 4). From the Kurtdag massif northwards, the Middle to Late Miocene and pre-Neogene sedimentary cover is characterised by NE–SW trending folds and thrusts (Ponicarov et al., 1967). The deformed sedimentary cover is referred as the Border Fold System by Braud and Ricou (1975). The folding and thrusting of the Border Fold System is consistent with compression along the Arabia–Anatolia boundary (Yazgan et al., 1983). The Pliocene–Quaternary basaltic lavas (Innocenti et al., 1982) fill topographic lows along the Fold System; this may have been permitted by faulting.

In the vicinity of the Karasu valley and eastwards to the Border Fold System area, numerous

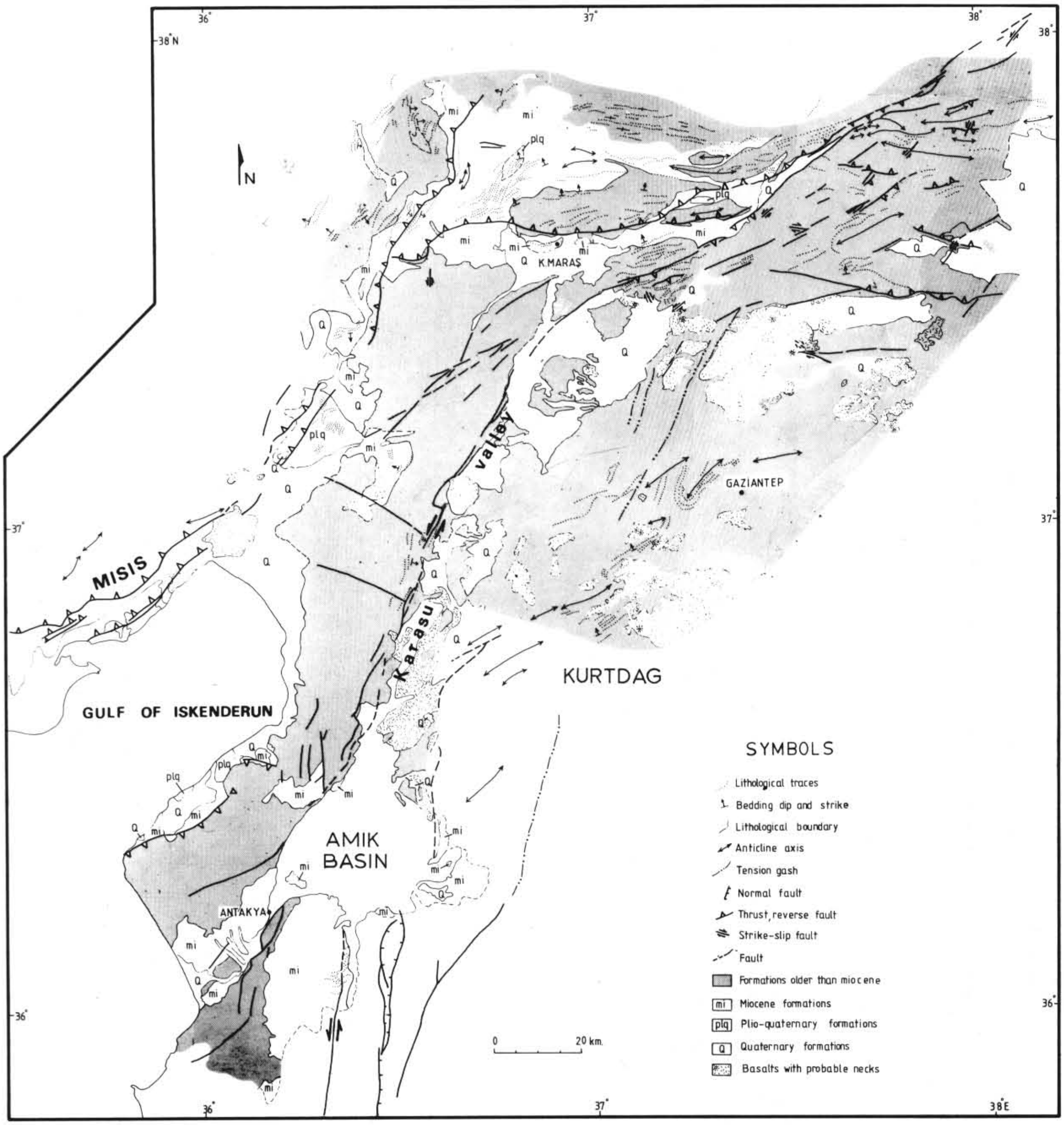

Fig. 4. Structural map of the junction area between Africa, Arabia and Anatolia.

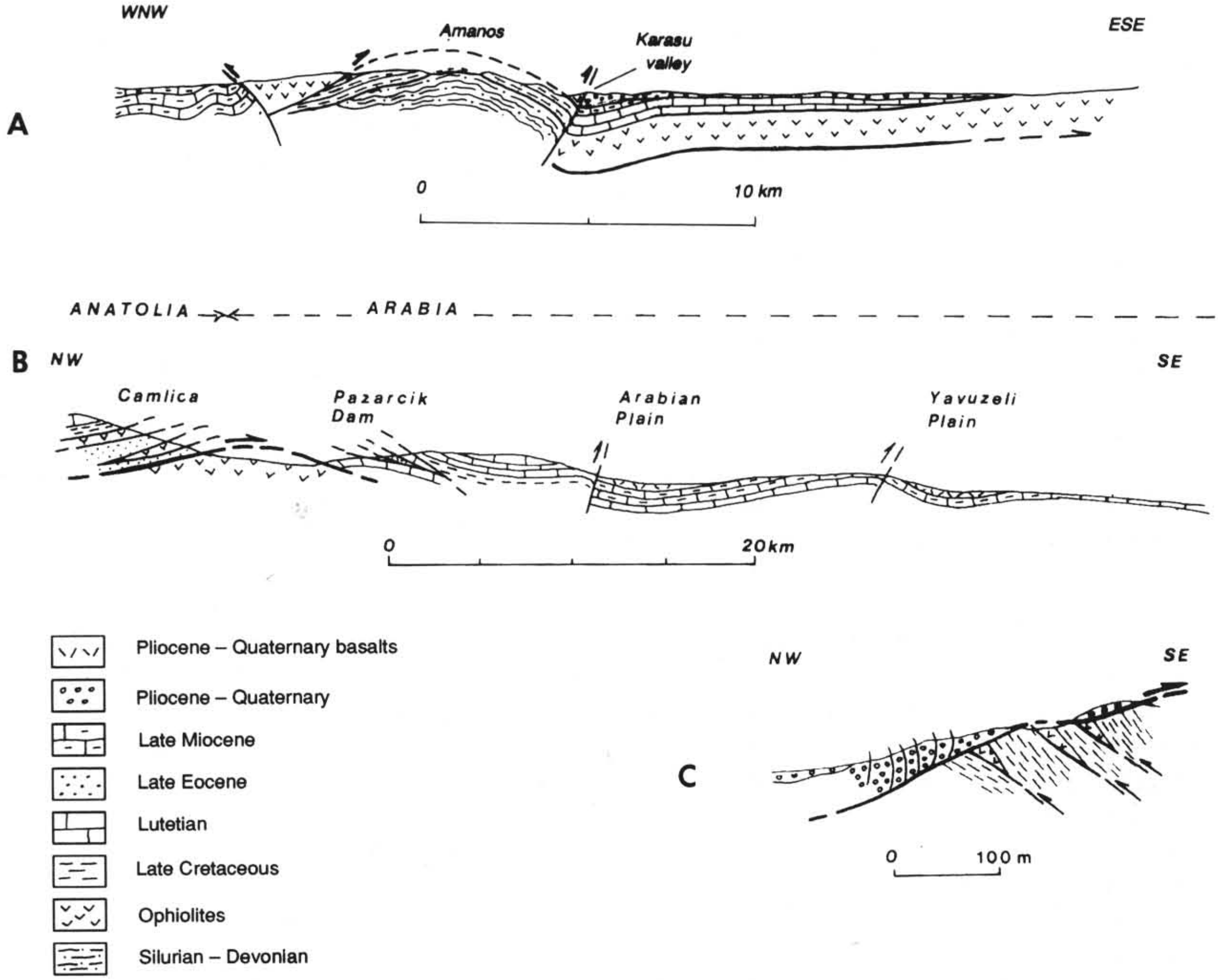

Fig. 5. Schematic cross-sections across the East Anatolian fault zone (for the location, see Fig. 3).

N25° trending, tension gash-like fractures (up to 50 km long), showing left lateral displacement are observed (Fig. 4). Between the Karasu valley and these fractures, the elevation of the Arabian platform pre-Neogene sequences decreases towards the Amanos Border Fault. This dip of the Arabian platform and the tensional fractures may be the results of NNW–SSE compressional flexuring of the crust under the Karasu valley.

Structures near the Arabia–Africa–Anatolia triple junction

The available structural schemes (see Sengör et al., 1985) place the Arabia–Anatolia–Africa junction at the northernmost end of the the Karasu valley, near Maras. The fault (or faulted zone) that separates the Amanos massif from the Karasu valley has more than 2 km of vertical offset. This fault dies out south of Maras, where a NE–SW oriented lineament cuts through the alluvial basin of the Aksu River and reaches the EAF zone. From SPOT satellite images, this feature can be seen to offset, in a sinistral sense (about 320 m), the Holocene fluvial deposits of the Aksu River (Fig. 3).

The Neogene sediments of the northern Amanos massif are strongly deformed (Fig. 3). The sedimentary cover of this area mainly consists of Early to Late Miocene (Tortonian) flysch, unconformably overlain by Pliocene–Quaternary continental conglomerates. The distal flysch series, studied in detail by Gökçen et al. (1986) in the Missis area, are assigned to the Karatas formation. They are deep marine deposits of the northern African margin, deposited before the Arabia–Africa separation. Since the Late Miocene, the flysch-related series have been thrust over the Amanos Massif towards the northern border of the Afro–Arabian shield (Fig. 3).

The flysch series and the Eocene limestones are imbricated by numerous N–S to NE–SW oriented oblique thrust planes dipping steeply to the west. The slickenside pitches are subhorizontal and show a sinistral contractional motion. The northwestern part of the Maras area is dominated by thick Mesozoic limestone formations of the Taurus, which overthrust the flysch series along a major, NE–SW oriented, reverse sinistral fault plane. The N–S shortening affects mostly the Miocene flysch, the unconformably overlying Pliocene–Quaternary sediments are less deformed.

The faulted, 20–30 km wide Maras anticline, which affects the Eocene and Miocene sediments, displays an abrupt structural change on the southern edge of the Anatolian block. To the east, the Maras anticline is connected to the well exposed E–W trending folds involved in the EAF structure. One can see, therefore, that the Maras region, which constitutes the northwestern corner of the Arabian plate (Fig. 3), is affected by thrusts and folds showing a right-angle geometry to the DSF. The area affected by this particularly intense shortening is more than 150 km wide. The sedimentary series involved in the structure belong to the African, Arabian and the Anatolian plates.

On the Mediterranean flank of the Amanos massif, the Mesozoic ophiolites are unconformably overlain by Late Miocene and Pliocene–Quaternary deposits, comprising the Messinian evaporites of the Iskenderun basin. On the southwestern part of the Amanos, the Neogene series are covered by ophiolites along a reverse fault (Fig. 4). In the same area, Messinian marls, interbedded with gypsum-bearing layers, are folded about E–W to ESE–WNW axes. The Missis massif, which runs along the Mediterranean coast, consists of Miocene sediments affected by Tortonian folds and thrusts which converge towards the south-southeast (Kelling et al., 1988).

The compressive structures observed on the northern and western margins of the Amanos massif are coeval (Late Miocene–Present) and may well be related. The Gulf of Iskenderun, which lies between the Amanos and Missis massifs, has the form of a flexural basin developed on the African crust ahead of the Anatolian thrust front. However, the Amanos-Maras compressive belt, which can be traced upto the Missis mountains, may correlate with the Kirinia convergent zone north of Cyprus. If this is so then the Kirinia-Missis-Amanos belt may delineate the African–Anatolian boundary.

Structure of the Arabia–Anatolia boundary

The EAF structure is generally considered to be a continuous major strike-slip fault bounding the Anatolian block and the Arabian plate (McKenzie, 1978; Jackson and McKenzie, 1984). However, east of Maras, in the area of the Arabia–Anatolia contact, compressive structures are found. The Maras anticline is linked eastwards to the Arabian platform by folds and thrusts. Cross-sections across the EAF area show that the dominant structures occurring between Maras and Golbasi are thrusts, involving Mesozoic, Eocene and Neogene formations.

Figure 5B illustrates the structure across the Arabia–Africa boundary, near Pazarcik, comprising the Border Fold system. The Mesozoic ophiolites (serpentinites), Eocene limestones and Miocene–Pliocene sediments are imbricated. In this area and near the presumed location of the EAF, we noted a recent, non-eroded, thrust surface. This thrust surface, dipped gently to the north (about 10°) and was characterized by N00° oriented slickenside lineations, indicating southwards transport of the hanging wall (site 26, Fig. 6).

Further east, the 10 km wide Gölbasi basin is associated with the EAF zone structures, cutting Mesozoic and Neogene sediments. In the southern border of this basin, unconsolidated red continental conglomerates of probable Pliocene–Quaternary age, are deformed. They lie with a

70-90° northward dip upon the Mesozoic formations by means of a basal thrust (Fig. 5C). The conglomerates are affected by several secondary sinistral reverse faults showing a relative southward motion (site 27, Fig. 6). Thus, the Gölbasi basin structure is interpreted as an assymetric flexural basin.

In the steep-sided Göksu valley (site 26, Fig. 6), east of Gölbasi, the Eocene limestones and Miocene sediments are imbricated in E–W pack-

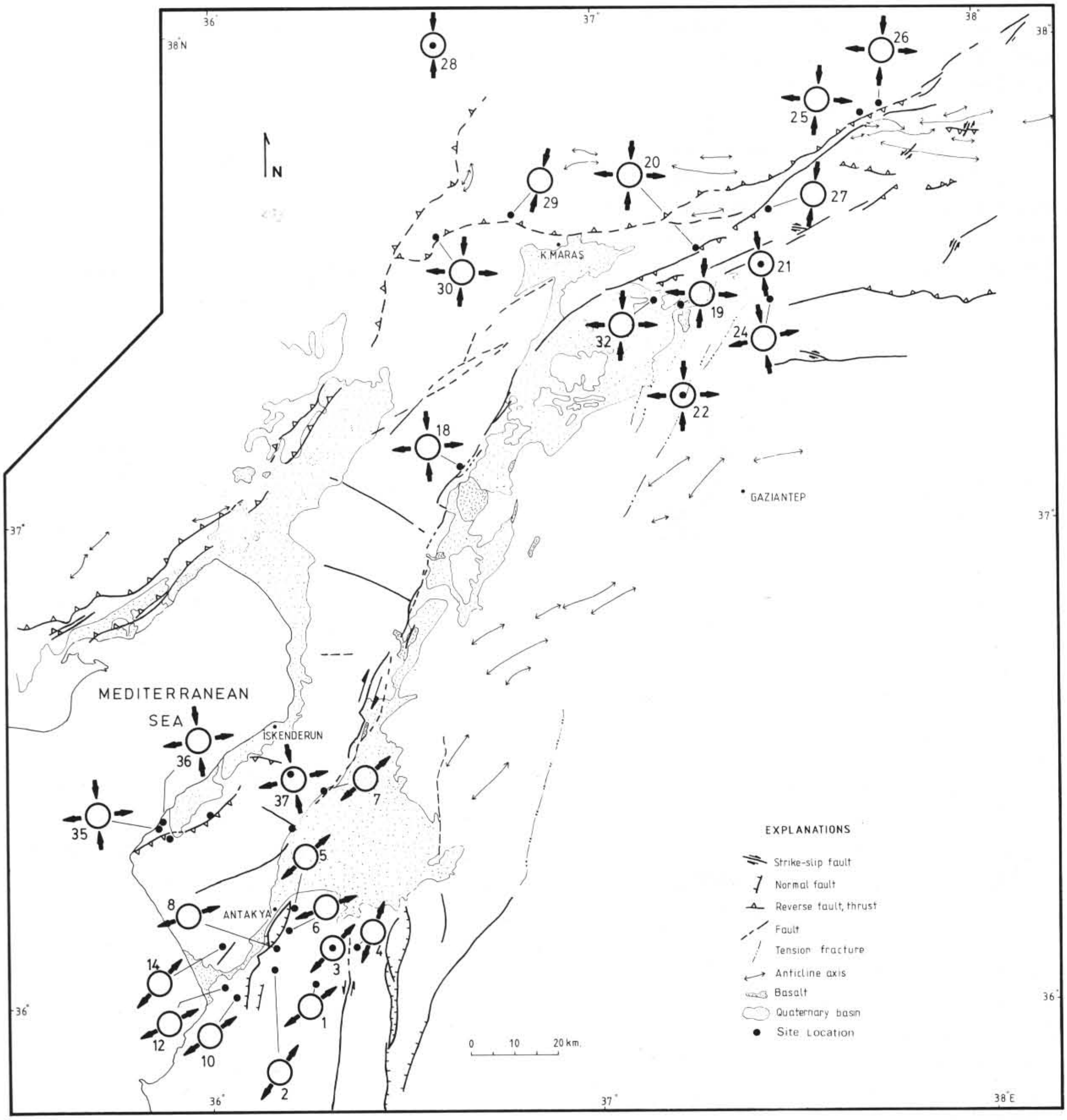

Fig. 6. Fault population mechanisms near the Africa–Arabia–Anatolia triple junction since the Late Miocene, reconstructed after field analysis. Black arrows represent the calculated mechanisms of extension (minimum principal stress axes σ_3 horizontal, centrifugal arrows), compression (maximum principal stress axes σ_1 horizontal, centripetal arrows) and strike-slip (σ_1 and σ_3, horizontals).

ages. Conglomerates of assumed Pliocene age are also affected by reverse faults, implying a N–S shortening. There is no evidence of dip-slip faulting similar to those observed by Hempton (1983) south of Lake Hazar.

The northern edge of the Arabian plate has been, and is still, subject to compressive tectonics. The sedimentary cover displays a large, 5–10 km wide, curved fold structure, located along the EAF (Figs. 4 and 5B). The Eocene and the Miocene limestones form two anticlines, separated by the Adiyaman syncline which is covered by Pliocene–Quaternary basaltic lavas (Innocenti et al., 1982). Further south, the Arabian plate seems to be undeformed. Folds and faulted folds occurred only in the vicinity of the Anatolia–Arabia boundary.

We conclude that deformation along the EAF zone is expressed mainly as a fold and thrust belt. Between Maras and Golbasi, there is no evidence of a major strike-slip faulting along the EAF zone.

The mechanisms of deformation

In the area investigated, detailed studies of the brittle fracturing were carried out in order to establish the stress pattern associated with the Africa–Arabia–Anatolia triple junction since the Late Miocene. Analysis of fault planes and slickenside lineations from field measurements enabled us to reconstruct the orientations (trends and plunges) of the principal stress axes in 37 sites. Folds, tension gashes and stylolites were

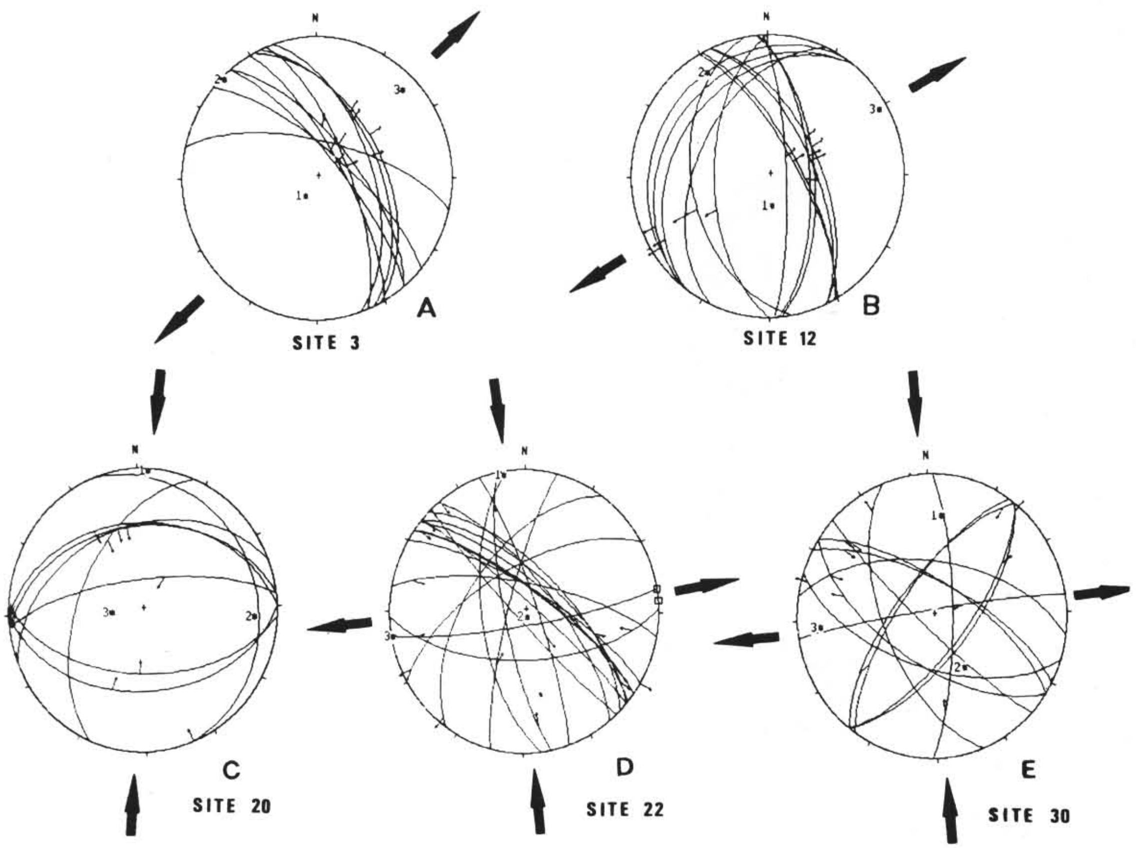

Fig. 7. Examples of fault plane solutions of the Altinozu Massif and the East Anatolian Fault zone. The faults cut the Late Miocene sediments. (A) Site 3. (B) Site 12. (C) Site 20. (D) Site 22. (E) Site 30. Schmidt lower-hemisphere projections. Fault planes are shown as curves, slickenside lineations as dots with centrifugal arrows (normal motion). *1*, *2* and *3* = the principal stress axis: σ_1, maximum principal stress axis; σ_2, intermediate principal stress axis; σ_3, minimum principal stress axis. The open squares correspond to poles of tension gashes. (For location, see Fig. 6).

also taken into account. Each motion recorded by the discontinuities corresponds, in general, to a single tectonic event. With these methods it has been possible to establish the directions of shortening and of extension for each tectonic event (Etchecopar et al., 1981).

The basic assumptions are that, for any location, a tectonic event is characterized by one homogeneous stress tensor and the resulting movement has the same direction and sense as the shear stress. If these assumptions are satisfied, then the deviatoric stress tensor of a tectonic event can be obtained using, for example, the quantitative computer-aided methods proposed by Etchecopar and co-workers (1981).

The results obtained are summarized in Figure 6. The stress tensors calculated from the fault populations in the whole area show a change south and north of latitude 37°N. South of the Amik basin, the stress field calculations gave NE–SW oriented extension (σ_1 = maximum principal stress axis, vertical; σ_2 = intermediate principal stress axis and σ_3 = minimum principal stress axis, horizontal). This direction of extension is compatible with sinistral strike-slip along N–S trending faults related to the DSF system, but with an oblique extensional component implying transtension. The N–S oriented faults in the Late Miocene series of the Altinozu massif are of sinistral extensional motion and the NW–SE faults show normal fault geometry (Figs. 7A and B). North of Antakya, the faults affecting the Late Miocene sediments result from a NE–SW extension (sites 5 and 7, Fig. 6), which is comparable to the one obtained for the Altinozu massif faults.

North of latitude N37°, measurements along the Amanos Border Fault were made in the Cretaceous and Eocene limestones. The tensor solution gives a strike-slip stress patterns (σ_1 and σ_3, horizontals; σ_2, vertical), with a N–S oriented compression and E–W extension (site 18, Fig. 6). Thus, we conclude that the main change of the stress pattern occurs in the central part of the Amanos massif.

The whole fault population measured in the area situated between the northern part of Amanos to the west, and Golbasi to the east, gives a coherent stress pattern. The fault planes are generally NE–SW oriented with a sinistral component and are associated with NW–SE dextral faults (Fig. 7C and D). There are also faults with a pure reverse offset (Fig. 7E). The observed oblique-slip and reverse faults are associated with the faults and thrusts illustrated in Figure 5B and C.

The stress tensor computed for the Amanos-Gölbasi area gives a strike-slip solution with horizontal N–S shortening and horizontal E–W extension (Fig. 6). This stress field also dominates the thrust zone of the northern Amanos, the northeast border of Arabia, as well as the EAF zone. This E–W extension is consistent with the westwards motion of the Anatolian Block (McKenzie, 1978), although the motion along the EAF has a significant N–S shortening component (Figs. 5C and 6).

Kinematic reconstruction

The sinistral displacement along the Dead Sea Fault is estimated to be around 105 km near the Dead Sea (Freund et al., 1970; Garfunkel, 1981), but north of the Palmyrides it diminishes. For this region, the geometry of the northernmost boundary of the ophiolitic nappes may be used as a marker to assess the relative displacement along the DSF. This nappe was emplaced during the Maastrichtian and has since been covered by sediments (Dubertret, 1962; Parrot, 1976). The front of the nappe crops out on both sides of the DSF: in the east, northwest of Afrin; in the west, in Baër-Bassit (Parrot, 1976) (Fig. 2). Al Maleh et al. (in press) have shown from petrological and sedimentological studies that the front of the ophiolite nappes on either side of the DSF can be correlated and they have not been eroded back from their final emplacement positions. The nappe boundary is offset by about 80 km along the DSF. Although the nappe front geometry does not constitute an incontestable criterion for horizontal offset, the displacement must correspond to the upper limit of relative sinistral displacement on the DSF at its northern end, as proposed by Garfunkel (1981).

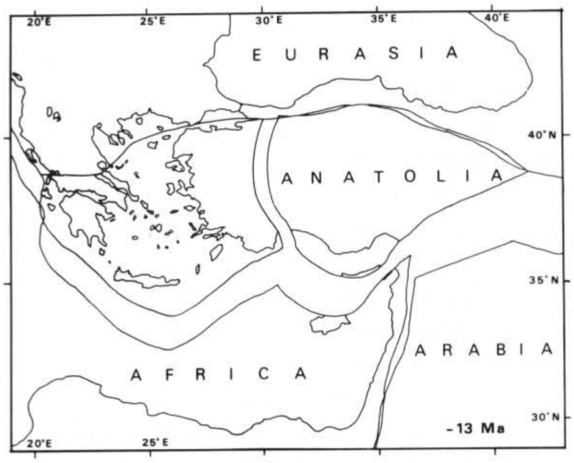

Fig. 8. Rigid block kinematic reconstruction 13 Ma ago, at the beginning of the recent phase of motion, according to the parameters in Table 1. The relative motions are with respect to the Eurasian plate.

At the latitude of Amanos, the rotation about the Eulerian Pole of Le Pichon and Gaulier (1988) gives a N–S direction for the Arabia–Africa relative motion. This corresponds to a N–S displacement of Arabia with respect to the N25° trending Amanos massif, previously assumed to be part of Africa. Therefore, this oblique convergence would imply a significant shortening component along the Amanos–Arabia limit. For an 80 km N–S displacement, this component is 34 km. Using this hypothesis, we interpret the Karasu valley as a flexural bend on the Arabian plate, which is obliquely overthrust by the Amanos massif, in a half-ramp basin structure (Fig. 5A). This interpretation also explains the compressive structures observed on both sides of the Karasu valley. This oblique thrusting starts as far north as N36°30′.

Figure 8 shows a reconstruction of the plates. The finite rotations of the rigid blocks for the last 13 Ma do not take into account, in this first approach, the internal deformation within Anatolia. We used the Eulerian Pole solution of Jackson and McKenzie (1984), for the Anatolian motion along the dextral NAF with respect to Eurasia, with a 80 km of displacement, and that of Chase (1978) for the relative Eurasia-Africa motion. We chose the location of the pole given by Le Pichon and Gaulier (1988) to describe the Arabia-Africa relative motion, which is not significantly different from the location of the pole determined by Garfunkel (1981). We assumes that the total displacement along the northern DSF was 80 km (Table 1).

TABLE 1

Parameters of plate kinematic reconstruction

	Latitude °N	Longitude °E	Angle (°)
Arabia/Eurasia	33.49	8.37	5.62
Arabia/Africa [1]	32.15	22.57	4.08
Africa/Eurasia [2]	29.2	−23.5	1.84
Eurasia/Anatolia [3]	14.6	34.0	−1.2
Arabia/Anatolia	37.61	0.75	4.62

Compiled from:
[1] Le Pichon and Gaulier (1988),
[2] Chase (1978),
[3] pole of rotation from Jackson and McKenzie (1984). The Eurasia–Anatolia rotation angle is calculated for 80 km of dextral motion along the North Anatolian Fault.

For the western part of the EAF (at 38°N, 38°E), we found that the relative motion between Arabia and Anatolia was about 327 km (2.5 cm yr^{-1}), with a N09° orientation (Fig. 9). Geometrical construction yields 193 km (1.5 cm yr^{-1}) for the Arabia–Anatolia convergence rate and 222 km (1.9 cm yr^{-1}) of sinistral displacement along the EAF. The calculated Arabia–Anatolia movement has the same directions as the tension fractures observed north of the DSF northern end.

On the calculated 222 km left-lateral motion along the EAF zone, only 22 km can be demonstrated from the field evidence (e.g., Dewey et al., 1986). On the other hand, recent field observa-

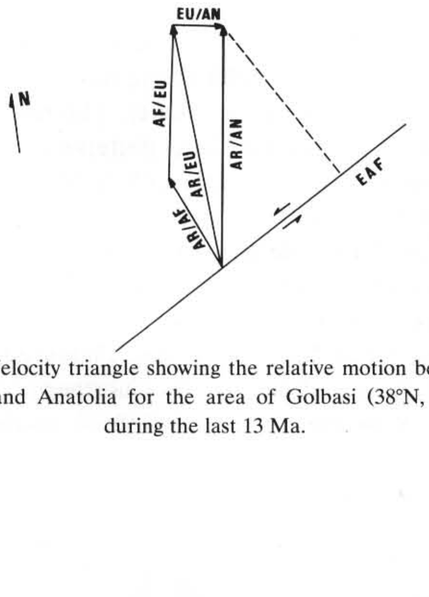

Fig. 9. Velocity triangle showing the relative motion between Arabia and Anatolia for the area of Golbasi (38°N, 38°E), during the last 13 Ma.

tions show that the area between the EAF and the NAF zones is characterised by strong compressional tectonics. In the extreme north an E–W oriented thrust, with a minimum of 10 km horizontal displacement, affects Late Miocene evaporite layers of the Sivas Basin (Fig. 1B). Post Miocene compressive tectonics have also been described by Metin (1984) from the area between Maras and Sivas. In addition, our field observations show that the left-lateral offset along the Eçemis Fault (Fig. 1B) has a significant reverse component. East of the area investigated, Yazgan et al. (1983) and Yazgan and Chessex (1989) show that the imbrication of the Eastern Taurus metamorphic rocks, involving Eocene metamorphic sequences (Maden Formation) and cut by the EAF, took place during the Late Miocene to Pliocene–Quaternary collision event. These metamorphic rocks overthrust the Late Miocene sequences of the Arabian platform.

These observations, combined with the deformation we describe for the area north of Amanos, indicate that extensive contractional deformation (folds and thrusts) is distributed across the whole eastern segment of the Anatolian block. We suggest that the global Anatolia–Arabia relative motion of 327 km in a N09° direction, has been largely accommodated by contraction distributed across the eastern part of the Anatolian block. We conclude that eastern Anatolia cannot be considered as a rigid block and the 22 km of sinistral offset corresponds to strike-slip faulting within a fold and thrust belt but does not constitute a major plate boundary.

Discussion and conclusions

The presence of extensive folding and thrusting within the Anatolian block demonstrates that the Anatolia–Arabia relative motion has an important convergent component and the contact between these two plates cannot be considered as simple transform fault. The deformation related to the Africa—Arabia–Anatolia triple junction is distributed across a large area in the vicinity of Maras (Fig. 4). Arabia is less deformed, and only its borders are implicated in the general structural frame that has evolved over the last 13 Ma. This is in contrast to the southern border of the Anatolian block which underwent shortening over a zone at least 150 km wide. The existing crustal anisotropies may have caused the differences in deformation. The fact that Arabia once belonged to the African craton, which has not undergone any important deformation for the last 500 Ma, may explain the homogeneous inheritance of the Arabian crust in contrast to the Anatolian block, which is a tectonic collage of several crustal terrains welded together during the Late Cretaceous —Eocene time (Sengör and Yilmaz, 1981) and before the Late Miocene Anatolia–Arabia collision. The deformation observed near Maras, especially the corner-like geometry north of Amanos, can be described by a rigid-plastic model where Arabia is an indenter similar to India in its collision with Asia (Molnar and Tapponier, 1975).

The Amanos massif has a particular role in the triple junction mechanism. Its meridional part contributes to the relative Africa–Arabia motion, where Amanos obliquely overthrusts Arabia. Its northern part lies in an area of deformation where the three plates, Anatolia, Africa and Arabia meet. The geometry of the Amanos border fault and the adjacent Karasu valley can be compared to the Bekaa valley segment of the DSF to the south where the horst-anticline feature of Mount Lebanon dominates. The Bekaa valley, lying between the Yammuneh and Serrhaya faults (Hancock and Atiyah, 1979; Walley, 1988) (Fig. 1B), has been affected by shortening (transpression) during the general sinistral event (Fig. 1B). An analogy could be drawn between the Karasu and the Bekaa valleys, which are bounded on their western flanks by the Amanos and Yammuneh faults, respectively.

The change in the tectonic regime takes place in the Amik Basin. The region south of this basin is subject to an NE–SW oriented extension consistent with the DSF sinistral motion. The northern part of the valley is located between the folded sedimentary cover of the Arabian platform and the thrust sheets of the northern Amanos. The northern part of the Amanos Border Fault has a sinistral, reverse-slip motion in its northern part.

The EAF related compression deformation features may continue west of Amanos. The Missis mountains may also be connected to northern Cyprus (Fig. 1B) through the Kirinia thrust (Biju-Duval et al., 1978) resulting from the Africa–Anatolia relative movement which is less than the relative motion between Arabia–Anatolia.

The major tectonic feature of southeastern Turkey is the Bitlis main thrust, which was a result of the collision between Arabia and the Anatolia-Iran border. The westwards extension of this thrust cuts the EAF zone in the Golbasi area (Fig. 1B). The Lice earthquake (6th September, 1975) shows that this thrust is still active (Jackson and McKenzie, 1984).

The EAF starts from Karliova, some 100 km north of the Bitlis-Zagros crush zone. It extends to the Lake Hazar within a sinistral motion (see Hempton, 1983). The western part of the EAF, from Golbasi to Maras, is affected by N–S shortening, oblique to the Arabia–Anatolia boundary. Therefore, it seems obvious that sinistral strike-slip faults are second-order and local consequences of the N–S collisional Arabian–Anatolian thrusting. The reverse sinistral displacement along the Arabia–Anatolia contact may be correlated to the reverse dextral one along the Zagros zone (McKenzie, 1978). Finally, we conclude that the EAF zone can be considered with the Zagros range as part of a "conjugate" fold and thrust belt in an Eurasia—Arabia convergent frame.

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References

Aksay, A., Tekeli, O., Ürgün, B.M. and Isik, A., 1988. Amanos'larin Paleozoyik birimleri ve Mezozoyik platform karbonat istifleri. MTA Gen. Müd., Jeol. Etüd. Dair. Rap., 66 pp.

Al Maleh, K., Delaune-Mayer, M., Mouty, M. and Parrot, J.-F., in press. Relations du fond de la nappe ophiolitique du NW syrien avec son substratum de part et d'autre de la faille du Levant: Baer-Bassit, Kurd Dagh. C.R. Acad. Sci. Paris.

Arpat, E. and Saroglu, F., 1972. The East Anatolian Fault System and thoughts on its development. Bull. Mineral Res. Explor. Inst. Turkey, Ankara, 78: 33–39.

Bayer, H.J., Hötzl, H., Jado, A.R., Rocher, B. and Voggenreiter, W., 1988. Sedimentary and structural evolution of the northwest Arabian Red Sea margin. Tectonophysics, 153: 137–151.

Berberian, F. and Berberian, M., 1981. Tectono-plutonic episodes in Iran. In: H. Gupta and F. Delaney (Editors), Zagros, Indu-Kuch, Himalayan Belt. Am. Geophys. Union, Geodyn. Ser., 3: 5–32.

Bergougnan, H., 1987. Etudes géologiques dans l'est Anatolien. Thèse, 86-33, Univ. Paris 6, 606 pp.

Biju-Duval, B., Lapierre, H. and Letouzey, J., 1976. Is the Troodos Massif (Cyprus) allochtonous? Bull. Soc. Géol. Fr., 18 (5): 1347–1356.

Braud, J. and Ricou, L.-E., 1975. Eléments de continuité entre le Zagros et al Turquie du Sud-Est. Bull Soc. Géol. Fr., 17 (6): 1015–1023.

Chase, C.G., 1978. Plate kinematics: the Americas, East Africa, and the rest of the world. Earth Planet. Sci. Lett., 37: 355–368.

Chorowicz, J., Henry, C. and Lybéris, N., 1987. Tectoniques superposées synsédimentaires des secteurs des golfes de Suez et d'Aqaba. Bull. Soc. Géol. Fr., 8 (3): 223–234.

Cochran, J.R., 1981. The Gulf of Aden: Structure and evolution of a young ocean basin and contintal margin. J. Geophys. Res., 86 (B1): 263–287.

Dewey, J.F., Hempton, M.R., Kidd, W.S.F., Saroglu, F. and Sengor, A.M.C., 1986. Shortening of continental lithosphere: the neotectonics of Eastern Anatolia, a young collision zone. In: M.P. Coward and A. Ries (Editors), Collision Tectonics. Geol. Soc. Spec. Publ., 19: 3–36.

Dubertret, L., 1962. Carte géologique Liban, Syrie et bordure des pays voisins, Scale 1/1.000.000. Inst. Geol. Nat., Paris.

Etchecopar, A., Vasseur, G. and Daignieres, M., 1981. An inverse problem in microtectonics for the determination of stress tensors from fault striation analysis. J. Struct. Geol., 3 (1): 51–65.

Falcon, N.L., 1974. Southern Iran: Zagros mountains. In: A.M. Spencer (Editor), Mesozoic–Cenozoic Orogenic Belts Geol. Soc. London, pp. 199–211.

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, 267: 107–130.

Garfunkel, Z., 1981. Internal structure of the Dead Sea leaky transform (rift) in relation to plate kinematics. Tectonophysics, 80: 81–108.

Giannerini, G., Campredon, R., Féraud, G. and Abou Zakhem, B., 1988. Déformations intraplaques et volcanisme associé: exemple de la bordure NW de la plaque Arabique au Cénozoique. Bull. Soc. Géol. France, 4 (6): 937–947.

Gökçen, S.L. Kelling, G., Gökçen, N. and Floyd, P.A., 1988. Sedimentology of a Late Cenozoic collisional sequence: the Misis Complex, Adana, southern Turkey. Sediment. Geol., 59: 205–223.

Hancock, P.L. and Atiya, M.S., 1979. Tectonic significance of mesofracture systems associated with the Lebanese segment of the Dead Sea transform fault. J. Struct. Geol., 1 (2): 143–153.

Hempton, M.R., 1983. Results of detailed mapping near Lake Hazar (eastern Taurus Mountains). In: O. Tekeli and M.C. Göncüoglu (Editors), Int. Symp. on the Geology of the Taurus Belt.

Innocenti, F., Manetti, P., Mazzuoli, R., Pasquare, G. and Villari, L., 1982. Anatolian and north-western Iran. In: R.S. Thorpe (Editor), Andesites. Wiley, New York, pp. 327–349.

Izzeldin, A.Y., 1987. Seismic, gravity and magnetic surveys in the central part of the Red Sea: their interpretation and implications for the structure and evolution of the Red Sea. Tectonophysics, 143: 269–306.

Jackson, J. and McKenzie, D.P., 1984. Active tectonics of the Alpine-Himalayan Belt between western Turkey and Pakistan. Geophys. J.R. Astron. Soc., 77: 185–264.

Jackson, J. and McKenzie, D.P., 1988. The relationship between plate motions and seismic moment tensors, and the rates of active deformation in the Mediterranean and Middle East. Geophys. J., 93: 45–73.

Kelling, G., Gökçen, S.L., Floyd, P.A. and Gökçen, N., 1987. Neogene tectonics and plate convergence in the eastern Mediterranean: New data from southern Turkey. Geology, 15: 425–429.

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

Lüttig, G. and Steffens, P., 1976. Explanatory notes for the Paleogeographic Atlas of Turkey from the Oligocene to the Pleistocene. Bundes Geowissenschaft. Rohst., Hannover, 64 pp.

Lyberis, N., 1988. Tectonic evolution of the Gulf of Suez and the Gulf of Aqaba. Tectonophysics, 153: 209–220.

McKenzie, D.P., 1972. Active tectonics of the Mediterranean region. Geophys. J.R. Astron. Soc., 30: 109–185.

McKenzie, D.P., 1976. The East Anatolian Fault: a major structure in Eastern Turkey. Earth Planet. Sci. Lett., 29: 189–193.

McKenzie, D.P., 1978, Active tectonics of the Alpine-Himalayan Belt: the Aegean Sea and surrounding regions. Geophys. J.R. Astron. Soc., 55: 217–254.

Metin, S., 1984. The geology of an area amongst the Derbasi (Develi), Armutalan and Gedikli (Saimbeyli) willages in the Eastern Tauride Mountains. Istanbul Univ. Muh. Fak. Yerbimleri Dergisi, C. 4: 1–2, 45–66 (in Turkish).

Molnar, P. and Tapponier, P., 1975. Cenozoic tectonics of Asia: effects of a continental collision. Science, 189: 419–426.

Maden Tetkik ve Arama Enstitüsü, 1962. Geological Map of Turkey, scale 1/5000.000, Hatay and Adana sheets. Maden Tetkik ve Arama Enstitüsü, Ankara.

Parrot, J.-F., 1976. Assemblage ophiolitique du Baër-Bassit et termes effusifs du volcano-sédimentaire. Pétrologie d'un fragment de la croûte océanique téthysienne charriée sur la plate-forme syrienne. Thèse Doc. Etat, Univ. Nancy, France, 333 pp.

Perinçek, D., 1979. Geological investigation of the Celikhan–Sincik–Koçali area (Adiyaman Province). Istanbul Univ. Fen. Fak. Mecm., Seri B, 44: 127–147.

Perinçek, D., and Ozkaya, I., 1981. Tectonic evolution of the Northern Margin of Arabian Plate. Bull. Inst. Earth Sci. Hacettepe Univ., 8: 91–101 (in Turkish).

Perinçek, D., Günay, Y. and Kozlu, H., 1987. Dogu ve güneydogu bölgesindeki yanal atimli faylar ile ilgili yeni gözlemler. 7th Biannual Petrol. Congr. (Turkey) 15 pp.

Philip, H., Cisternas, A., Gvishiani, A. and Gorshkov, A., 1989. The Caucasus: an actual example of the initial stages of continental collision. Tectonophysics, 161: 1–21.

Ponicarov, V.P. (Editor in chief), 1967. The Geological map of Syria. Explanatory notes. Part. I. Stratigraphy, Igneous rocks and Tectonics. V.O. Technoexport, Moscow, 230 pp, with geological maps at scale 1:500,000.

Sengör, A.M.C., 1979. The North Anatolian transform fault: its age, offset and tectonic significance. J. Geol. Soc. London, 136: 269–282.

Sengör, A.M.C., and Yilmaz, Y., 1981., Tethyan evolution of Turkey: A plate tectonic approach. Tectonophysics, 75, 181–224.

Sengör, A.M.C., Görür, N. and Saroglu, F., 1985. Strike-slip faulting and related basin formation in zones of tectonic escape: Turkey as a case study. In: T.R. Biddle and N. Christie-Blick (Editors), Strike-slip Deformation, Basin Formation and Sedimentation. Soc. Econ. Paleontol. Mineral., Spec. Publ., 37: 227–264.

Steckler, M.S., Berthelot, F., Lybéris, N. and LePichon, X., 1988. Subsidence in the Gulf of Suez: impications for rifting and plate kinematics. Tectonophysics, 153: 249–270.

Tinkler, C., Wagner, J.-J., Delaloye, M. and Selçuk, H., 1981. Tectonic history of the Hatay ophiolites (south Turkey) and their relation with the Dead Sea Rift. Tectonophysics, 72: 23–41.

Walley, C., 1988. A braided strike-slip model for the northern continuation of the Dead Sea Fault and its implications for Levantine tectonics. Tectonophysics, 145: 63–72.

Yazgan, E. and Chessex, R., 1989. The Malatya Geotraverse and its bearing on tectonics of Eastern Taurus Belt. Unpublished rep., MTA, 50 pp.

Yazgan, E., Michard, A., Whitechurch, H. and Montigny, R., 1983. Le Taurus (Turquie orientale), élément de la suture sud-téthysienne. Bull. Soc. Géol. France, 7 (25): 59–69.