Active tectonics of the Eastern Mediterranean region: deduced from GPS, neotectonic and seismicity data

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Abstract
This paper reviews the main tectonic features of the Eastern Mediterranean region combining the recent information obtained from GPS measurements, seismicity and neotectonic studies. GPS measurements reveal that the Arabian plate moves northward with respect to Eurasia at a rate of 23 ± 1 mm/yr, 10 mm/yr of this rate is taken up by shortening in the Caucasus. The internal deformation in Eastern Anatolia by conjugate strike-slip faulting and E-W trending thrusts, including the Bitlis frontal thrust, accommodates approximately a 15 mm/yr slip rate. The Northeast Anatolian fault, which extends from the Erzincan basin to Caucasus accommodates about 8 ± 5 mm/yr of left-lateral motion. The neotectonic fault pattern in Eastern Anatolia suggests that the NE Anatolian block moves in an E-ENE direction towards the South Caspian Sea. According to the same data, the Anatolian-Aegean block is undergoing a counter-clockwise rotation. However, from the residuals it appears that this solution can only be taken as a preliminary approximation. The Eulerian rotation pole indicates that slip rate along the North Anatolian fault is about 26 ± 3 mm/yr. This value is 10 mm/yr higher than slip rates obtained from geological data and historical earthquake records and it includes westward drift of the Pontides of a few millimetres/year or more. GPS measurements reveal that the East Anatolian fault accommodates an 11 ± 1 mm/yr relative motion. GPS data suggest that Central Anatolia behaves as a rigid block, but from neotectonic studies, it clearly appears that it is sliced by a number of conjugate strike-slip faults. The Isparta Angle area might be considered a major obstacle for the westward motion of the Anatolian block (Central and Eastern Anatolia). The western flank of this geological structure, the Fethiye-Burdur fault zone appears to be a major boundary with a slip rate of 15-20 mm/yr. The Western Anatolian grabens take up a total of 15 mm/yr NE-SW extension. The fact that motions in Central Anatolia relative to Eurasia, are 15-20 mm/yr while in Western Anatolia and Aegean Sea they are 30-40 mm/yr could suggest that Western Anatolia decouples from Central Anatolia and the Isparta Angle by the Fethiye-Burdur fault zone and Eskişehir fault. It is also hypothesized that the differentiation of tectonic styles and velocities in the Anatolian-Aegean block are related to differences between the slabs lying under the Cyprus and Hellenic arcs.

Key words *Global Positioning System (GPS) – neotectonics – seismicity*

1. Introduction

Recent tectonics of the Eastern Mediterranean region is very complex and has been studied intensely for the last 30 years since initiation of the plate-tectonic concept. The Eastern Mediterranean region, which defines the region lying between the Caspian Sea and the Adriatic Sea through Caucasus, Anatolia, Aegean Sea and Greece, is the one of the world's most seismically active regions (figs. 1-3). Deadly earthquakes occurring in this region during the last 20 years are: 1975 Lice (M_s = 6.6, Turkey, Arpat, 1977a; Eyidoğan, 1980); 1976 Çaldıran (M_s = 7.1, Turkey, Arpat, 1977b); 1983 Horasan-Narman (M_s = 6.8, Turkey, Barka *et al.*, 1983); 1988 Spitak,

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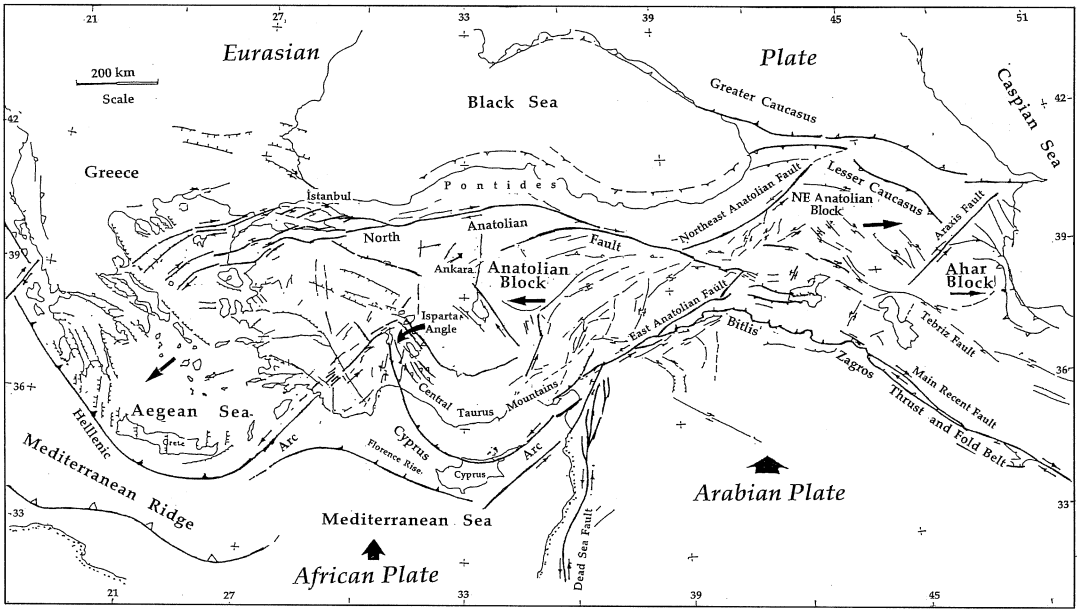

Fig. 1. Distribution of active faults in the Anatolian region. Modified from Şengör *et al.* (1985), Şaroğlu *et al.* (1987), Barka (1992) and Armijo *et al.* (1992).

Armenia (M_s = 6.7, Haessler *et al.*, 1992); 1990 Iran (M_s = 7.7, Fuenzalida, 1995); 1991 Rach (M_s = 7.0, Georgia, Fuenzalida *et al.*, 1997a); 1992 Erzincan (M = 6.8, Turkey, Barka and Eyidoğan, 1993; Fuenzalida *et al.*, 1997b); 1995 Western Macedonia (M_s = 6.6, Greece, Stiros, 1995); 1995 Dinar (M_s = 6.1, Turkey, Eyidoğan and Barka, 1996), (fig. 3).

Global kinematic models (NUVEL-1A; DeMets *et al.*, 1990, 1994) based on the analysis of oceanic spreadings, fault systems, and earthquake slip vectors indicate that the Arabian plate is moving in a north-northwest direction relative to Eurasia at a rate of about 25 mm/yr, averaged over about 3 Ma. This motion, resulting in a continental collision along the Bitlis-Zagros fold and thrust belt, is thought to cause intense seismic activity (fig. 2). The African plate is moving in a northerly direction relative to Eurasia, at a rate of about 10 mm/yr. The differential motion between Africa and Arabia (~ 15 mm/yr) is thought to be taken up predominantly by left-lateral motion along the Dead Sea transform fault (fig. 1), (*e.g.*, Freund *et al.*, 1970). GPS velocities for the two sites located south of the Bitlis suture both indicate NW oriented motion relative to Eurasia (18 ± 5 mm/yr), (Reilinger *et al.*, 1995, 1997) somewhat slower, but not statistically different from NUVEL-1A estimates (24 ± 6 mm/yr), (fig. 4).

The analysis of 1988-1994 GPS measurements in Turkey indicates that the Anatolian block is undergoing counter-clockwise rotation about a pole located north of Sinai (33.4°E, 31.1°N), (fig. 4, Reilinger *et al.* 1995). The Anatolian block escapes from Eastern Anatolia due to Eurasia and Arabian collision and moves onto the African oceanic plate along the Hellenic arc. In this paper, we will review the main tectonic elements of the Eastern Mediterranean region by combining recent information obtained from GPS measurements, seismicity and neotectonic observations.

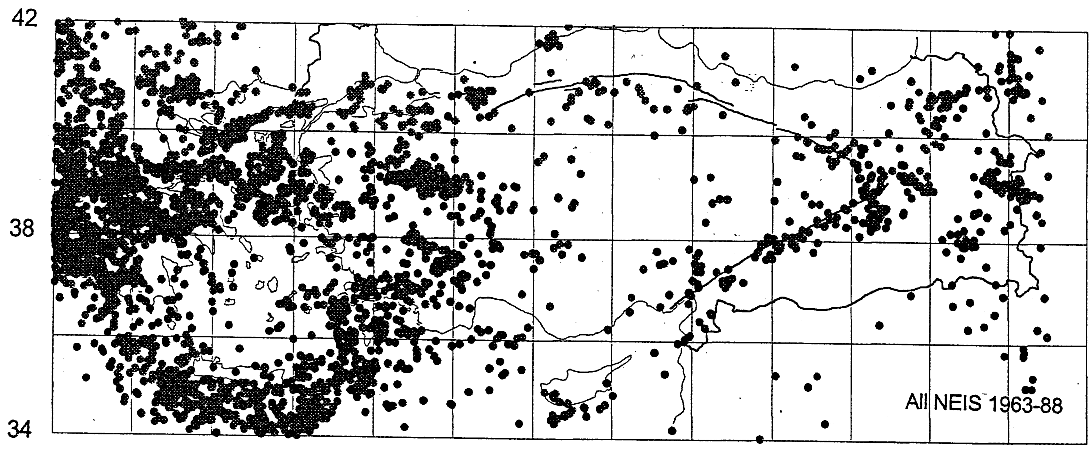

Fig. 2. Seismic activity of the Anatolian region between 1963-1988. Modified from Jackson (1994).

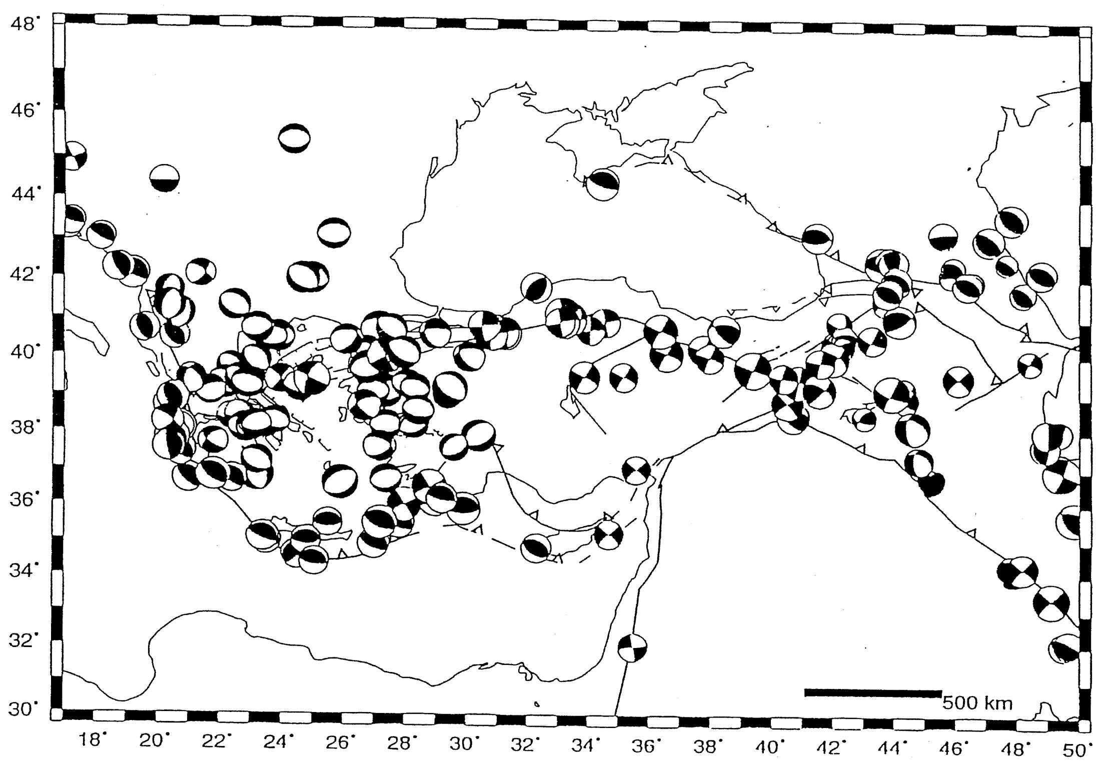

Fig. 3. Distribution of fault plane solutions in the Eastern Mediterranean region, documented from McKenzie (1972, 1978), Jackson and McKenzie (1984). From Reilinger *et al.* (1997).

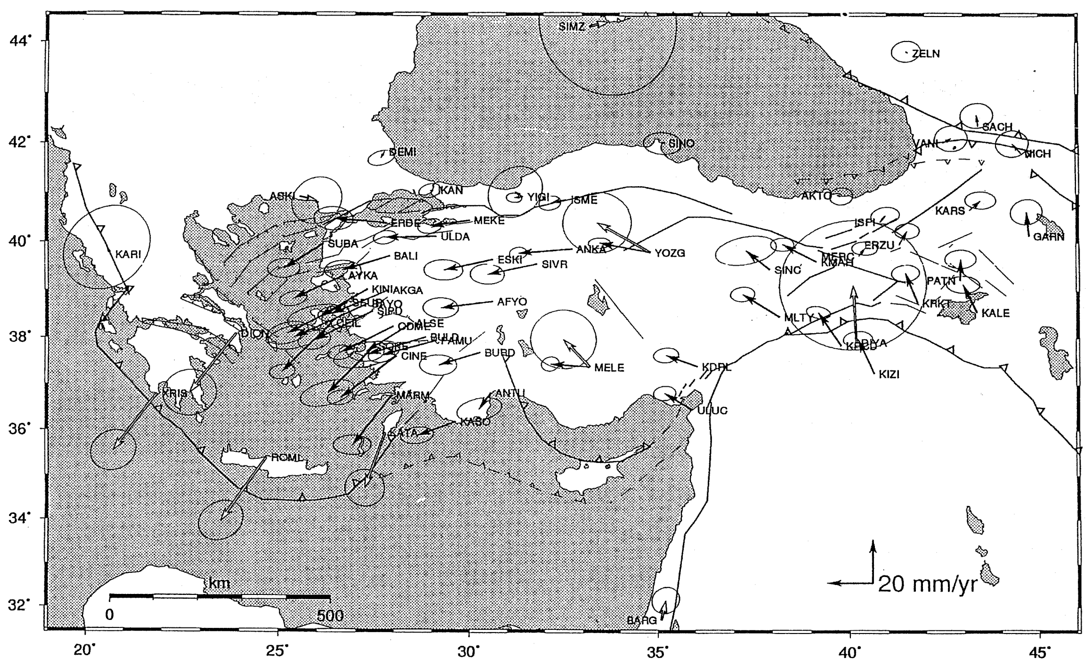

Fig. 4. Distribution of GPS velocity vectors measured between 1988-1994 in the Eastern Mediterranean region (Reilinger *et al.*, 1997). Open arrows are SLR measurements taken from Robbins *et al.* (1995).

2. Eastern Anatolia and Caucasus

Neotectonics in Eastern Anatolia (figs. 5 and 6) involves four types of structures: NE-SW trending sinistral faults (Horasan-Narman fault, Erzurum fault zone, Malazgirt faults), NW-SE trending dextral faults (Çaldıran-Tutak-Karayazı fault, Balıkgölü fault), E-W trending thrusts (Muş-Van thrust) and N-S trending extension cracks and/or normal faults (giving rise to volcanic activity, Nemrut and Süphan volcanoes), (fig. 6), (Şaroğlu and Güner, 1981; Şengör *et al.*, 1985; Dewey *et al.*, 1986; Barka and Kadinsky-Cade, 1988).

The 1975 Lice earthquake, M = 6.6, occurred in the western part of the Bitlis frontal thrust (Eyidoğan, 1980). The 1976 Çaldıran earthquake had a right-lateral mechanism and created a 50 km long surface rupture (Arpat, 1977b), (fig. 7). The 1983 Horasan-Narman earthquake, M = 6.8 took place on the NE-SW trending left-lateral Horasan-Narman fault east of Erzurum (Barka *et al.*, 1983).

Figure 8 shows aftershock distribution and fault plane solution of the main shock of the 1983 earthquake (Eyidoğan, 1990). The 1988 Spitak earthquake occurred south of the Pambak-Sevan fault and exhibited dominant thrust solution even though the Pambak-Sevan fault is dominantly right-lateral strike-slip (Haessler *et al.*, 1992). It is suggested that this earthquake was related to a positive flower structure along the Pambak-Sevan fault due to the compressional bend along the fault. From GPS measurements (fig. 4), we estimate a 7 ± 2 mm/yr right-lateral slip on the Çaldıran-Tutak-Karayazı fault system and about 15 mm/yr N-S shortening is taken up by Eastern Anatolia including Bitlis thrust (Reilinger *et al.*, 1995).

Fault slip rates for the range-bounding faults of the Lesser and Greater Caucasus are difficult to estimate given the wide station spacing. Overall shortening roughly N-S is estimated in 10 ± 2 mm/yr, (Reilinger *et al.*, 1997). There is weak evidence that this shortening is distributed more or less equally between the lesser and Greater Caucasus. The recent 1991 Rach, Georgia, earthquake $M_s = 7.0$, had a thrust solution, consistent with the type of deformation observed in the Caucasus (Fuenzalida *et al.*, 1997a).

3. The Northeast Anatolian fault zone

The Northeast Anatolian fault zone extends from near the city of Erzurum, Turkey northeast to the Caucasus mountains (figs. 1 and 5; Barka and Kadinsky-Cade, 1988). This fault zone may include the Borjomi-Kazbeg fault which crosses the Caucasus mountains as defined by Philip *et al.* (1989). The Northeast Anatolian fault zone consists of several seg-

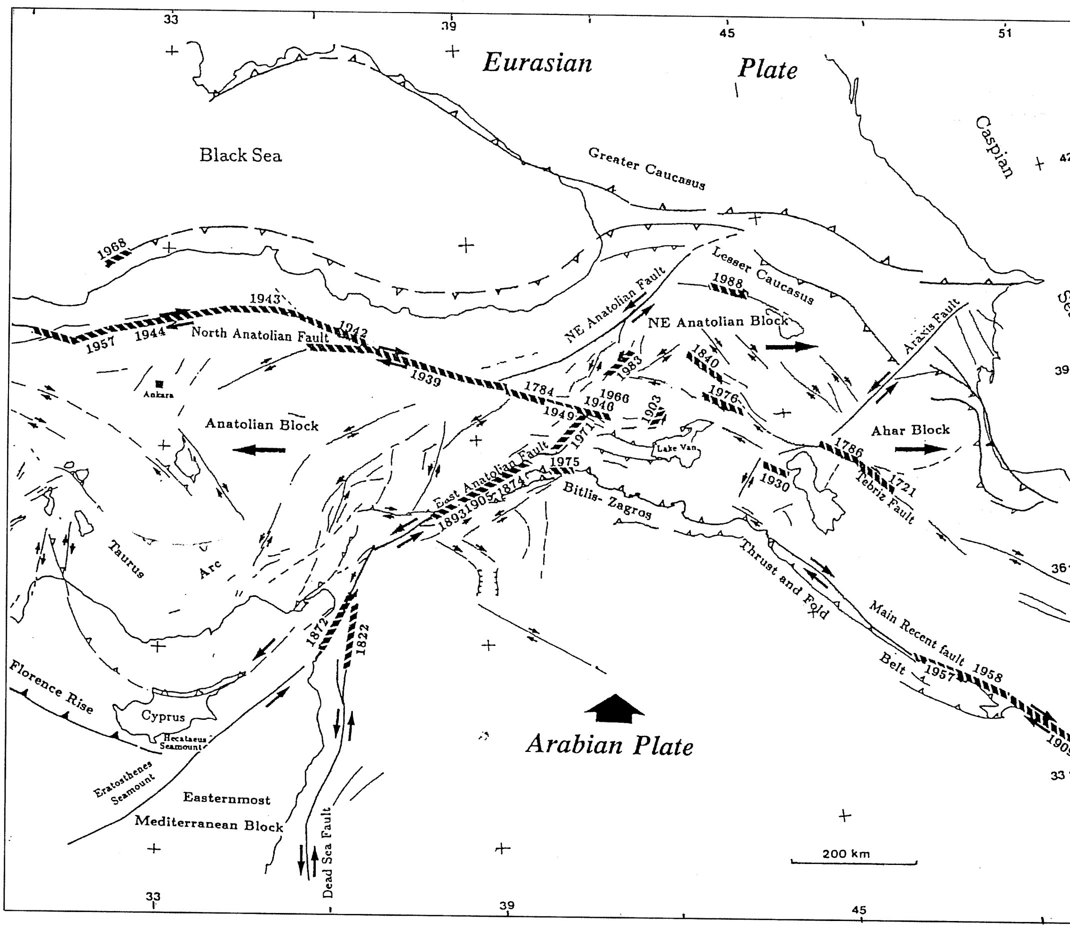

Fig. 5. Distribution and extents of surface rupture of large earthquakes in Eastern Anatolia, Caucasus and Western Iran for the last 250 years documented mostly from Ambraseys (1970, 1975, 1988, 1989), Ambraseys and Finkel (1995), Tchalenko (1977).

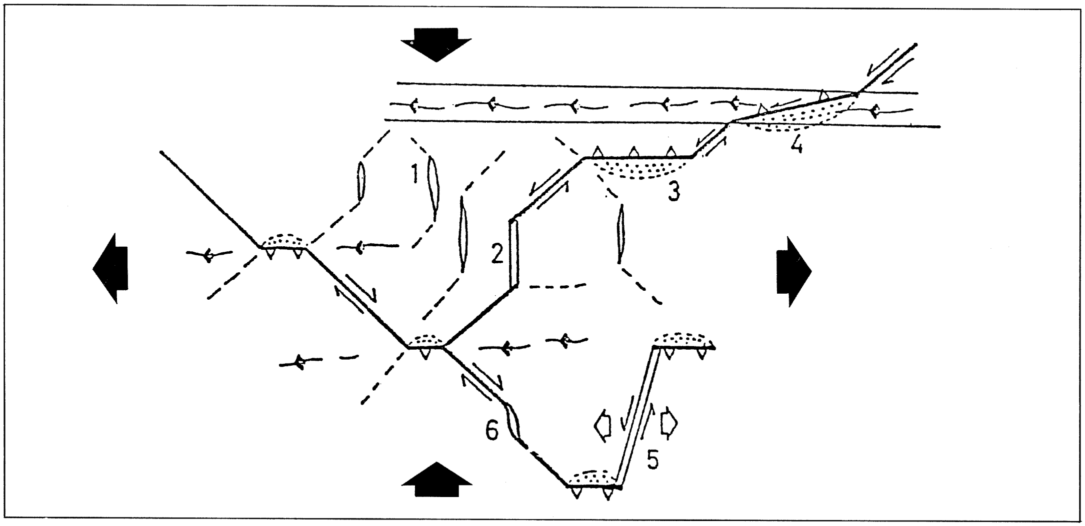

Fig. 6. Deformation pattern and types of basins resulting from the N-S shortening in Eastern Anatolia. From Şengör *et al.* (1985). 1 = Extension cracks; 2 and 6 = pull-apart basins; 3 and 4 = ramp basins; 5 = oblique extension.

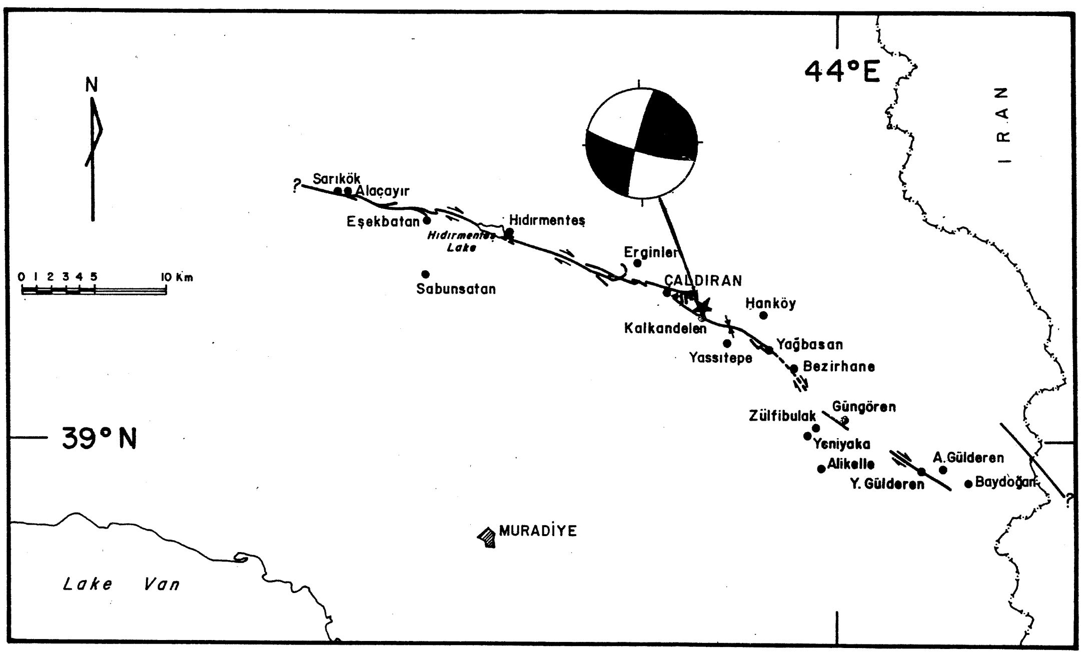

Fig. 7. Rupture zone of the 1976 Çaldıran earthquake ($M = 7.1$), Eastern Anatolia and its fault plane solution (Arpat, 1977b; Şaroğlu *et al.*, 1983; Jackson and McKenzie, 1984). The maximum right-lateral offset was about 3.5 m which took place in the northeastern half of the rupture zone (Şaroğlu *et al.*, 1983).

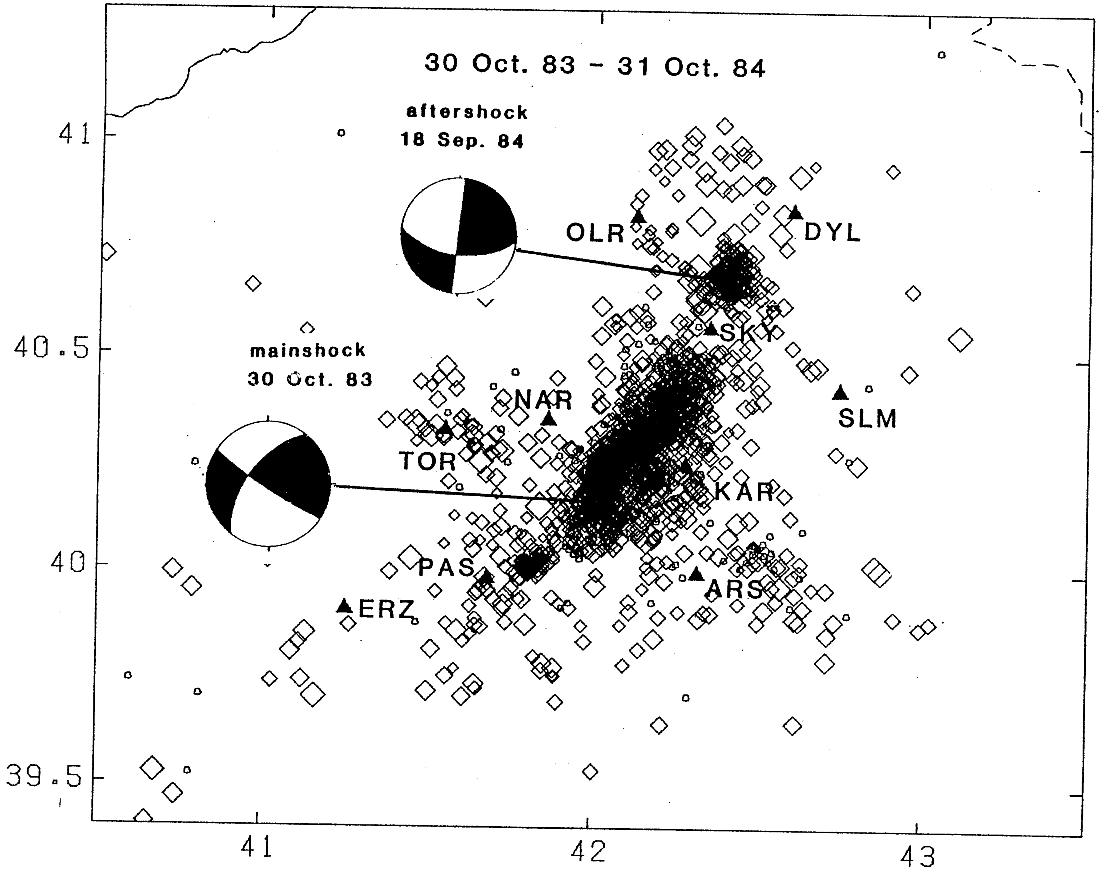

Fig. 8. Aftershock distribution and fault plane solutions of the 1983 Horasan-Narman, M = 6.8, and 1984 Kömürlü earthquakes in Eastern Anatolia (Eyidoğan, 1990).

ments with a total length of approximately 350 km. The fault zone steps and widens to the northeast (in places up to 10 km). Tatar (1978), who first studied this fault zone, suggested that it has an oblique motion, consisting mostly of left-lateral slip with a subordinate thrust component. Earthquake records indicate that it is less active than the segments of the North Anatolian fault. GPS stations in Northeast Turkey provide some bounds on fault slip rates. All of these stations are consistent with left lateral slip. Total motion between stations AKTO and ERZU is 8 ± 5 mm/yr (fig. 4), (Reilinger *et al.*, 1997). Geologic estimates of average fault slip for the segment crossing the Caucasus are given by Philip *et al.* (1989) as 18-25 mm/yr. However, this evidence is disputed by Triep *et al.* (1995) who question even the presence of a left-lateral structure crossing the Caucasus. Westaway (1994) gives an estimate of slip across this zone (termed the Erzurum Tiblisi fault zone) of 0.9 mm/yr from summed seismic moments and of < ~ 6 mm/yr from kinematic considerations.

4. Margins of the Black Sea

The Northern margin of the Black Sea has been identified as active thrusting along the

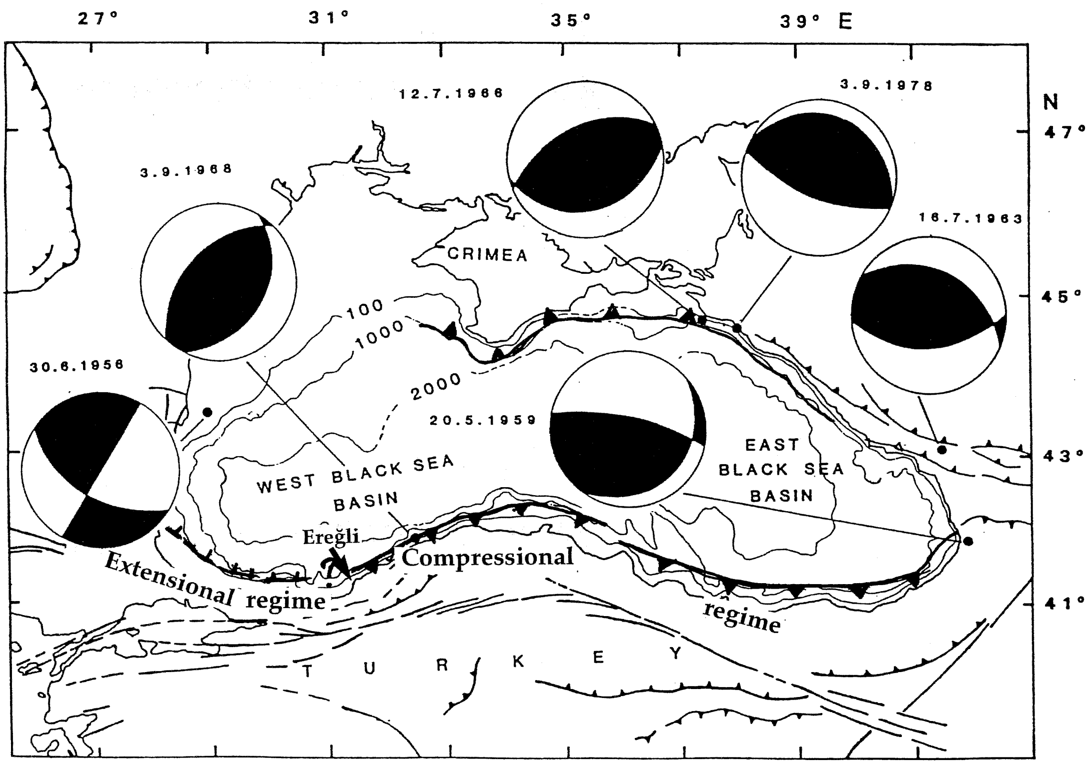

Fig. 9. Tectonics of the Black Sea (modified from Alptekin *et al.*, 1986). Most of the southern and northern margins show shortening, except north of the Marmara Sea region (west of the 31°E) and Bulgarian and Romanian coasts where it is of extensional nature (Barka, 1991).

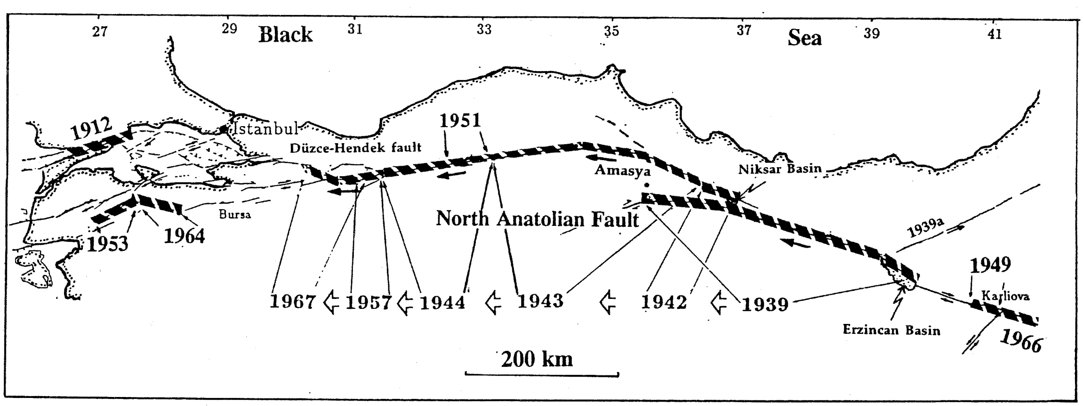

Fig. 10. Distribution of surface ruptures of consecutive large earthquakes between 1939 and 1967 and distribution of the rupture zone along the North Anatolian fault (Barka, 1992).

Greater Caucuses from Georgia to the Crimea (fig. 9), (Jackson and McKenzie, 1984; Alptekin *et al.*, 1986). The southern margin of the Black Sea has not been studied in detail due to the low rate of seismic activity (fig. 2). However, the 1968 Bartın earthquake (M = 6.8) testifies that this margin can produce destructive earthquakes (fig. 10). This margin can be divided into two subsections: the eastern section extending from the Lesser Caucasus to Ereğli, and the western section extending from Ereğli to Bulgaria. The first section shows a compressional nature, while the latter one is under extensional tectonics. Slip rates along these sections are lower than 5 mm/yr.

5. The East Anatolian fault

The left-lateral East Anatolian fault is the southern boundary of the westward escaping Anatolian block (figs. 1 and 6), (Şengör *et al.*, 1985; Dewey *et al.*, 1986; Barka and Kadinsky-Cade, 1988; Şaroğlu *et al.*, 1992; Taymaz *et al.*, 1992; Westaway, 1994). According to Arpat and Şaroğlu (1972) and Seymen and Aydin (1972) the fault zone has 15-27 km left-lateral post-Miocene displacement revealing about 6 mm/yr slip rate. From seismicity and relative plate motions, Taymaz *et al.* (1991) estimated a 25-31 mm/yr slip rate along the East Anatolian fault. The GPS-derived slip rate for the East Anatolian fault (11 ± 2 mm/yr), (Reilinger *et al.*, 1995) is slightly higher, but statistically consistent with geological estimates, as well as with estimates based on historic earthquakes (6-10 mm/yr). In contrast, the GPS slip rate is significantly lower than some estimates derived from plate kinematics reconstructions (~ 30 mm/yr). For this century, 1905 (M = 6.8), 1908 (M = 6.1), 1971 Bingöl earthquake (M = 6.7), and 1986 Sürgü earthquake (M = 6) occurred on the East Anatolian fault (Ambraseys and Finkel, 1987, Taymaz *et al.*, 1992). However, during the nineteenth century, there were large events associated with the East Anatolian fault, for example the 1866 (M = 6.8), 1874 (M = 7.2), 1874 (M = 7.1), 1893 (M = 7.1), rupturing two thirds of the fault zone (Ambraseys, 1989), (fig. 6).

6. The North Anatolian fault

The right-lateral North Anatolian fault is the northern boundary of the anticlockwise rotating Anatolian-Aegean Block and extends over 1500 km, from Karlıova in Eastern Turkey to the Greek mainland (*e.g.* Ketin, 1969; McKenzie, 1972; Şengör, 1979). The GPS data provide slip rates for the North Anatolian fault (26 ± 3 mm/yr). The actual rate might be less than this estimate when we consider the western drift of the Pontides up to 5 mm/yr (Oral *et al.*, 1995). The GPS-derived slip rate for the NAF is significantly higher than long term (offset of geological features) and intermediate term (historical earthquakes) slip rates (5-17 mm/yr), (*e.g.*, Barka, 1992; Westaway, 1994) and significantly lower than seismic and plate closure estimates (~ 38 mm/yr), (*e.g.*, Taymaz *et al.*, 1991). The implications of these discrepancies are difficult to assess given the large uncertainties in the geological and seismic/closure estimates. A total of 23 large earthquakes (M > 6.5) occurred during the twentieth century along the entire length of the fault zone (7 in the North Aegean, 6 in the Marmara region and 10 along the main part). Amongst these earthquakes, six westward migrating large earthquakes (M > 7) occurring between 1939-1967, created a remarkable continuous 900 km long rupture zone (Ketin, 1948), (fig. 10), with maximum offsets of 7.5 m (Koçyiğit, 1989; Barka, 1992, 1996). The 1992 earthquake, M = 6.9, which occurred in the Erzincan basin is the most recent destructive earthquake along the fault. Figure 11 shows fault plane solutions of the main shock and the largest the aftershock, and aftershock distribution (*e.g.*, Barka and Eyidoğan, 1993). Neotectonic structures along the western side of the big bend of the North Anatolian fault indicate thrusting in Neogene sediments suggesting that the rotational motion of the Anatolian block is not perfectly tangential. This imperfect rotation might be also responsible for the thrusting along the western part of the Black Sea coast (the 1968 Bartın earthquake) and the western drift of Pontides depicted from the GPS measurements.

In the Marmara Sea region, the North Ana-

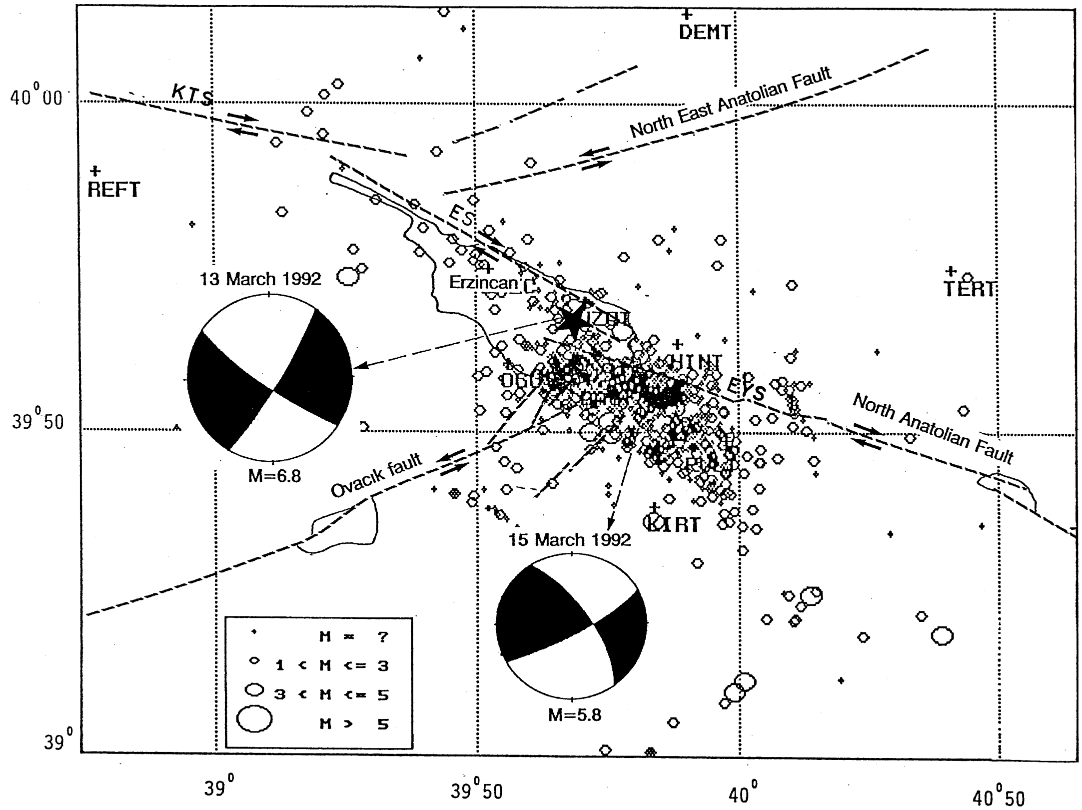

Fig. 11. Aftershock distribution (Ergintav *et al.*, 1992) and fault mechanism of the 13 and 15 (ISC data) March 1992 earthquakes (M = 6.8, M = 5.8 respectively), from Barka and Eyidoğan (1993). The epicenter is located in the eastern half of the Erzincan basin (NEIS). EYS, ES and KTS are segments of the North Anatolian fault. REFT, DEMT, TERT, KIRT, HINT, UZAT and Erzincan are location of seismic stations. The unruptured segment of the North Anatolian fault is EYS, which last ruptured in 1784 (Ambraseys, 1975).

tolian fault zone splays into three strands and forms a diffused boundary between the Anatolian block and the Black Sea (fig. 12). Barka and Kadinsky-Cade (1988) proposed a pull-apart model for the Marmara Sea region to account for the deformation and dynamics of the strands of the North Anatolian fault. This model has been generally accepted with small modifications (*e.g.*, Wong *et al.*, 1995; Ergün and Özel, 1995; Akgün and Ergün, 1995; Koral and Öncel, 1995). Barka (1992) further investigated the extent of the active strands of the North Anatolian fault beyond west of the Marmara Sea toward the North Aegean Sea and suggested that three strands cross the Marmara Sea and North Aegean region with an identical geometric pattern (figs. 12 and 13). The distribution of historical earthquakes in the last 2000 years (Ambraseys and Finkel, 1991) reveals that the northern strand has accommodated more large earthquakes than the other two southern strands. Similar results have been obtained from the GPS measurements which indicate that at least 60% of the motion is taken up by the northern strand (Straub and Kahle, 1995; Straub, 1996).

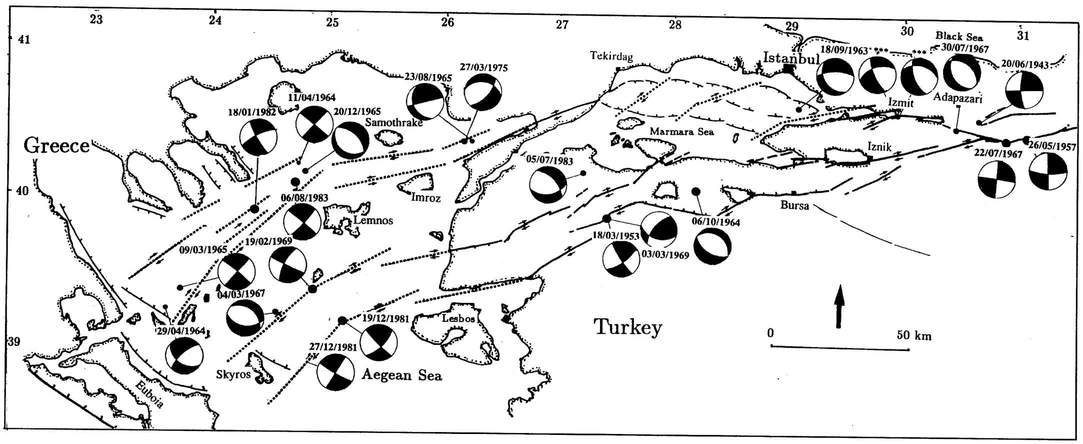

Fig. 12. Fault plane solutions of major earthquakes along the North Anatolian fault in the Marmara and North Aegean regions (compiled from McKenzie, 1972, 1978, Jackson *et al.*, 1982; Crampin and Evans, 1986; International Seismological Center, ISC, 1981, 1982, 1983).

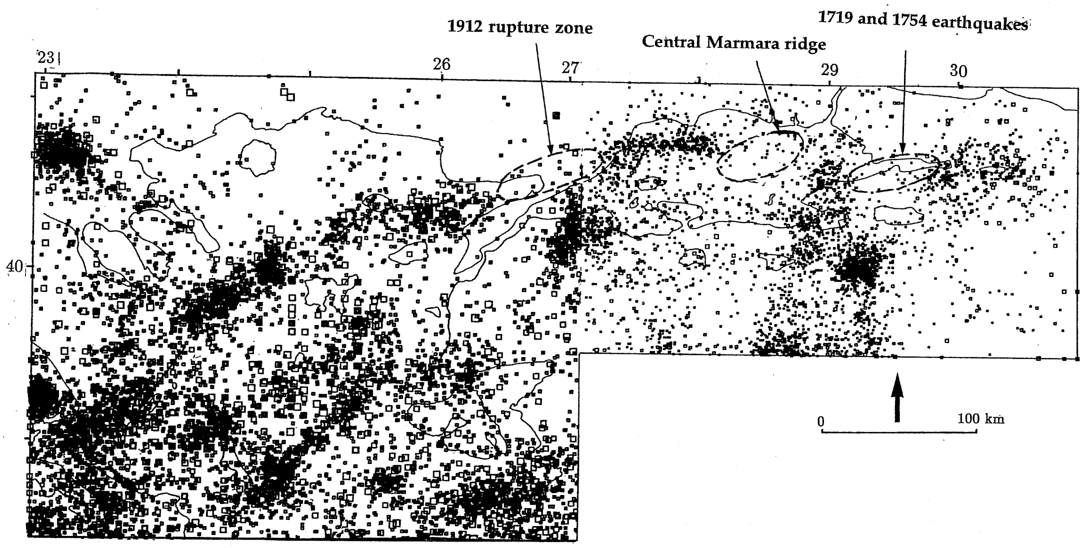

Fig. 13. Distribution of all earthquakes, $M > 2.5$ in the Marmara Sea and North Aegean regions for the period 1964-1984 (ISC data, 1984). The ellipses in the Marmara Sea region indicate the locations of seismically quiet areas which coincide with major strike-slip segments.

7. Central Anatolia

The Tuz gölü fault (Arpat and Şaroğlu, 1975; Aksaray-Şereflikoçhisar fault, Şengör *et al.*, 1985) is one of the more prominent active features in Central Anatolia (figs. 1 and 4). Şaroğlu *et al.* (1987) suggested that in spite of some offset streams at the southern part of the fault indicating right-lateral faulting, structural observations along the fault indicate that this fault is a high angle thrust fault with a right-lateral component. The N-S and NE-SW trending volcanic cones in the vicinity of the Tuz Gölü fault are consistent with both this reverse faulting and right-lateral motion (Pasquare *et al.*, 1988; Emre, 1991). The NNE-SSW trending Ecemiş fault (fig. 4) is a major left-lateral fault in Central Anatolia (Yetiş and Demirkol, 1984). Along the fault, morphological features such as offset streams and ridges and fault line scarps are very clear. The main part of the fault zone extends from Yeşilhisar to the Pozantı area, where the Ankara-Adana highway cut exhibits a considerable thrust component on the fault.

Historical (0-1900 A.D.) and instrumental earthquake records show that seismic activity in Central Anatolia has been low relative to Western Anatolia (Ambraseys, 1970, 1975, 1988; Ambraseys and Finkel, 1987), (figs. 2 and 3). The 1938 Kırşehir earthquake, $M = 6.8$, is the only large earthquake that occurred during this century in this region (Parejas and Pamir, 1939; Ketin, 1969), (fig. 6). The 1717 and 1835 Ecemiş earthquakes occurred near Kayseri (Öztin and Bayülke, 1990) can be listed as other important events in Central Anatolia.

The Central Taurus Mountains (Özgül, 1983) are a major neotectonic feature of South-Central Anatolia which is parallel to the Cyprus arc. During the early-middle Miocene time Central Anatolia was covered by a large lake whilst the Taurus Mountains were under sea level where extensive carbonate deposition occurred (Şaroğlu *et al.*, 1983). Since the mid-late Miocene the Taurus Mountains have been uplifted at least 1000 m relative to the Central Anatolian Plateau. The uplift has been interpreted as a wide anticlinal fold by Şaroğlu *et al.* (1983).

In summary, when the geometry and compressional nature of the Taurus and Pontic arcs, conjugate right-lateral Tuz Gölü and left-lateral Ecemiş faults and N-S and/or NE-SW orientations of volcanic cones are considered together, we can suggest that Central Anatolia is under approximately N-S or NNE-SSW shortening while it is rotating anticlockwise along the North Anatolian fault. This shortening is probably related to the collisional processes along the Cyprus arc between Africa and Anatolia. Although the total amount is not clearly known, it is assumed to be approximately 10 mm/yr.

8. The Isparta Angle

The Isparta Angle (figs. 1 and 14), (Blumenthal, 1963) constitutes the junction between the Cyprus and Hellenic arcs and is a tectonic assemblage which has a complex tectonic history. It has been studied intensely for the last 25 years (*e.g.*, Brunn *et al.*, 1971; Graciansky, 1972; Monod, 1977; Poisson, 1984, 1990; Ricou *et al.*, 1979; Gutnic *et al.*, 1979; Şengör and Yilmaz, 1981; Yilmaz, 1983, 1984; Poisson *et al.*, 1984; Şenel, 1983; Hayward, 1984; Robertson and Woodcock, 1984; Waldron, 1984; Marcoux, 1987; Akay and Uysal, 1988; Robertson, 1990; Kissel *et al.*, 1993; Frizon De Lamotte *et al.*, 1995). This zone consists of several different tectonic entities, such as the Lycian Nappes, the Antalya Nappes, the Beyşehir-Hoyran Nappes and the Alanya Massif which were emplaced between the late Cretaceous and the late Miocene (*e.g.*, Şengör and Yilmaz, 1981). According to paleomagnetic data (Kissel *et al.*, 1993) the angular shape of the Isparta Angle is related to the post Eocene tectonic activity. GPS measurements made in Turkey during the period of 1988-1992 indicate that the center of the Isparta Angle (Antalya site) has slower motion, less than 10 mm/yr, relative to Eurasia (Oral, 1994). By contrast, Central Anatolia (east of the Isparta Angle) moves westward relative to Eurasia at about 15 mm/yr and Western Anatolia (west of the Isparta Angle) moves in a SW direction at ~ 30 mm/yr (fig. 4). Neotectonic structures in and

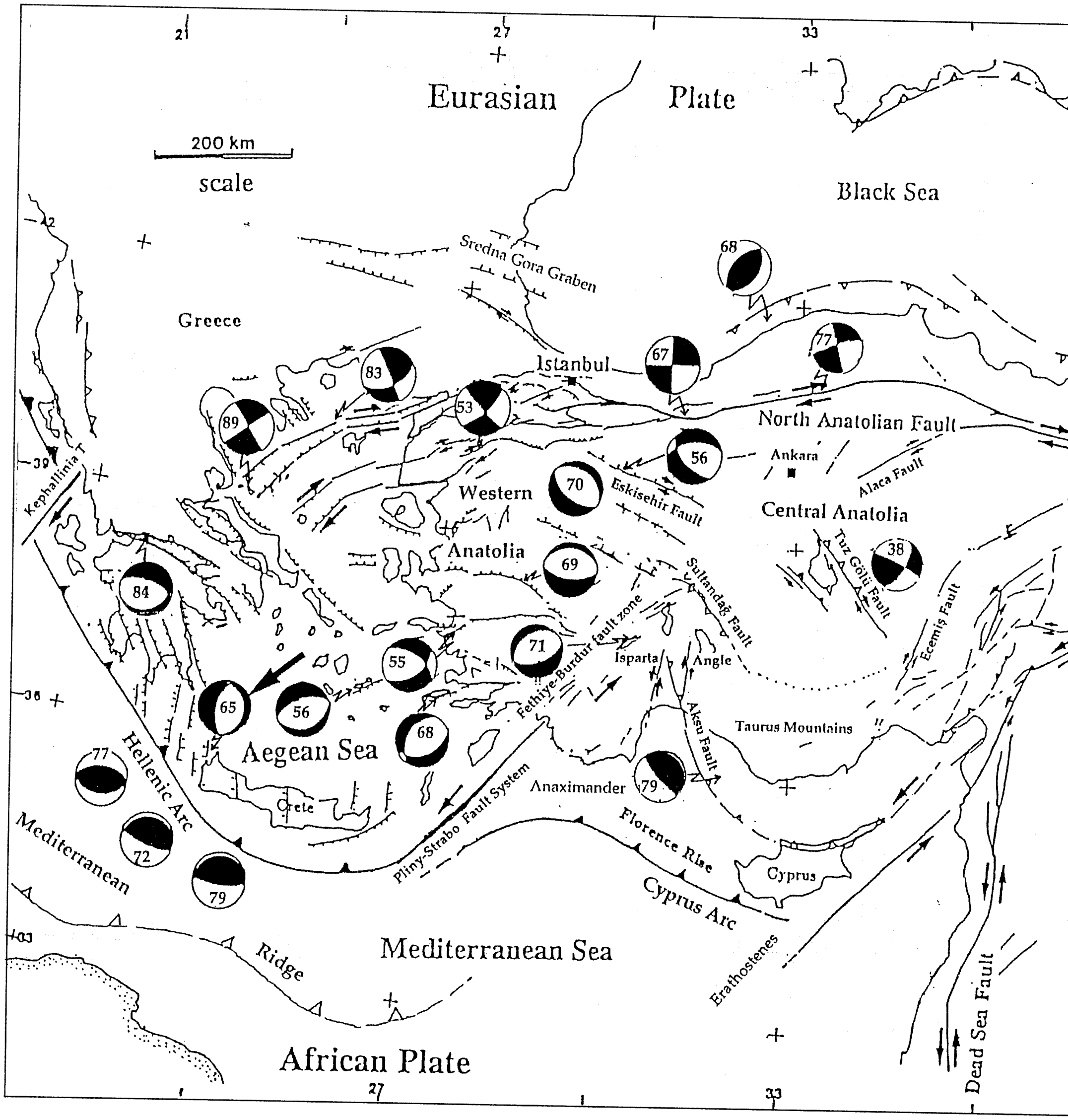

Fig. 14. Neotectonic structures in the Eastern Mediterranean region. Fault plane solutions of major earthquakes, $M > 6.5$, documented from Alptekin *et al.* (1986), Jackson and McKenzie (1984), Taymaz *et al.* (1990, 1991) and Ekström and England (1989) showing that motion along the North Anatolian fault is dominantly dextral and Western Anatolia and the Aegean have dominant extension. Two digit numbers in the fault plane solutions are dates of the earthquakes.

on the margins of the Isparta Angle, where the eastern flank is bounded by the NW-SE trending Sultandağ thrust fault (Boray *et al.*, 1985) and the western side is made up of the transtensional left-lateral NE-SW trending Fethiye-Burdur fault zone (Dumont *et al.*, 1979; Taymaz and Price, 1992; Price and Scott, 1994), are compatible with these GPS results (figs. 14). This tectonic nature of the Isparta Angle also suggests that the eastern flank of the Isparta Angle may slow down the westward motion of Southern Central Anatolia.

9. Cyprus arc

Subduction along the Cyprus arc has been a long discussed controversial subject in the literature (*e.g.*, McKenzie, 1972; Biju-Duval *et al.*, 1978; Büyükaşikoğlu, 1979, 1980; Harsh *et al.*, 1981; Rotstein and Kafka, 1982; Jackson and McKenzie, 1984; Rotstein and Ben-Avraham, 1986; Kempler and Ben-Avraham, 1987; Ben-Avraham *et al.*, 1988). Seismic data indicate subduction beneath the Florence Rise (fig. 14). The occurrence of subduction is not clear south of Cyprus (Büyükaşikoğlu, 1979, 1980; Jackson and McKenzie, 1984). In the latter area, subduction is either complicated or prevented by the sea mounts, such as the Eratosthenes (Kempler and Ben-Avraham, 1987; Ben-Avraham *et al.*, 1988). The Anaximander sea mount is another significant structure between the Hellenic and Cyprus arcs located in the SW corner of the Isparta Angle. Offshore Isparta Angle, seismic reflection studies indicate that the Florence Rise, the Anaximander sea mount and the Antalya basin including structures in the vicinity of the Cyprus arc are all compressional features (Biju-Duval *et al.*, 1978; Kempler and Ben-Avraham, 1987; Poisson, 1990). The subduction along the Cyprus Arc does not create an arc parallel extension on the over-riding Central Anatolia, as in the Western Anatolia/Aegean. This might suggest that the dip angle of the slab is different from the slab being subducted along the Hellenic arc (*e.g.*, Kempler and Ben-Avraham, 1987; Wortel and Spakman, 1992), (fig. 15). This relatively lower angle of subduction is also supported by

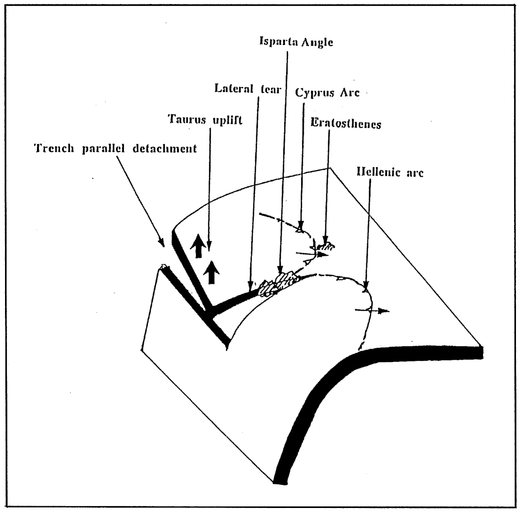

Fig. 15. Eastern Mediterranean slab geometry under the Hellenic and Cyprian arcs. The idea is partly inspired from Büyükaşikoğlu (1979), Wortel and Spakman (1992).

the location of the volcanic activity in Central Anatolia, which occurs at a considerable distance from the trench.

10. Western Anatolia

In Western Anatolia, E-W and WNW-ESE trending grabens and the related normal faults are the dominant neotectonic features (Philippson, 1918; Ketin, 1968; McKenzie, 1978; Dewey and Şengör, 1979; Jackson and Mc Kenzie, 1984; Şengör, 1982, 1987; Şengör *et al.*, 1984). Among these, Gökova, Büyük Menderes, Gediz, Bakırçay and Simav grabens and the Kütahya and Eskişehir faults are the most prominent (figs. 1 and 14). A number of major normal fault events occurred along these faults, for example 1899 Büyük Menderes, 1928 Torbalı, 1955 Balat, 1969 Alaşehir, 1969 Simav, 1970 Gediz and 1995 Dinar earthquakes (*e.g.*, Ambraseys, 1988), (fig. 16). The October 1, 1995, Dinar earthquake, (M = 6), occurred along the NW-SE trending Dinar fault (fig. 16). Study of the surface rupture and

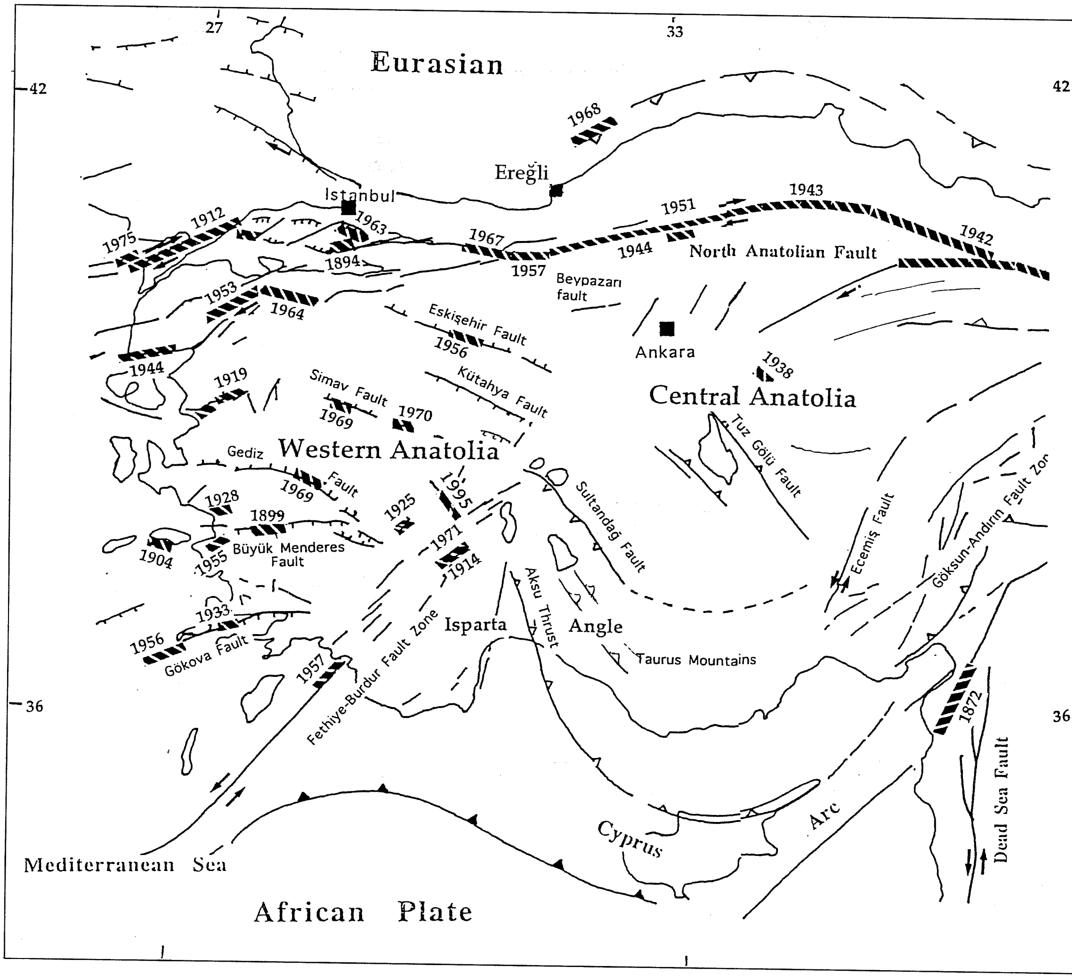

Fig. 16. Distribution of earthquake ruptures in the Anatolian region between 1800-1995. Documented from Ketin (1948), Ambraseys and Finkel (1987), Ambraseys (1988), Westaway (1990).

inversion of broad-band *P* wave seismograms illustrated that the Dinar earthquake had dominantly normal faulting (Eyidoğan and Barka, 1996). The WNW-ESE trending Eskişehir fault, a right-lateral fault with a significant extensional component limits the extension in Western Anatolia (fig. 17). The fault extends between Uludağ and Afyon (Şaroğlu *et al.*, 1987). The 1956 Eskişehir earthquake (Öcal, 1959), $M = 6.5$, occurred along this fault and its mechanism (Jackson and McKenzie, 1984) consisted of right-lateral and extensional components being consistent with the structural nature of the fault (fig. 14). The extension in Western Anatolia steps from the northwest end of the Eskişehir fault to Bulgaria (the Sredna Gora graben, Richter, 1958) across the Marmara Sea where the strands of the North Anatolian fault cross (fig. 14). The extensional regime also influences the North Anatolian fault that splays into three strands and forms a number of pull-apart structures displaying a diffused boundary across the Marmara Sea and North Aegean (Dewey and Şengör, 1979; Şen-

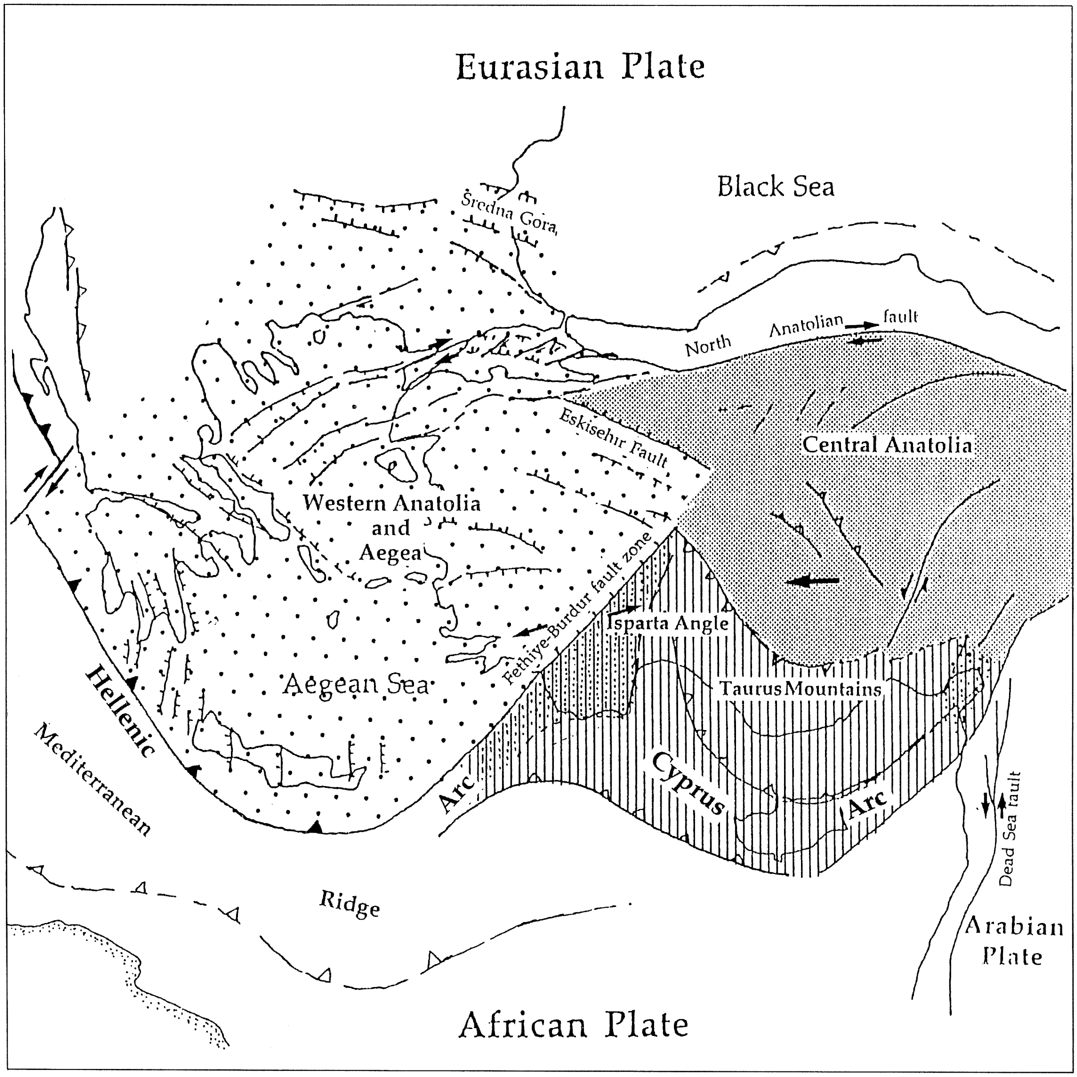

Fig. 17. Neotectonic sub-division of the Anatolian Block. Dotted zones indicate NE-SW extension, shaded area is Central Anatolia where NNE-SSW shortening currently occurs, the hatched area correspond to the complex structures of the Cyprian arc, within this, dotted parts are the extensional areas (belonging to the western flank of the Isparta Angle). Otherwise it represents compressional tectonics.

gör *et al.*, 1985; Barka and Kadinsky-Cade, 1988; Barka, 1992). The spatial distribution extension in Western Anatolia is quite uncertain because of the relatively large uncertainties on velocity estimates and the limited station density. The available data are consistent with about 10 ± 5 mm/yr extension across the Bozdağ horst, Büyük Menderes and Gediz grabens and 5 ± 5 mm/yr extension across the Gulf of Gökova, north of the Marmaris penin-

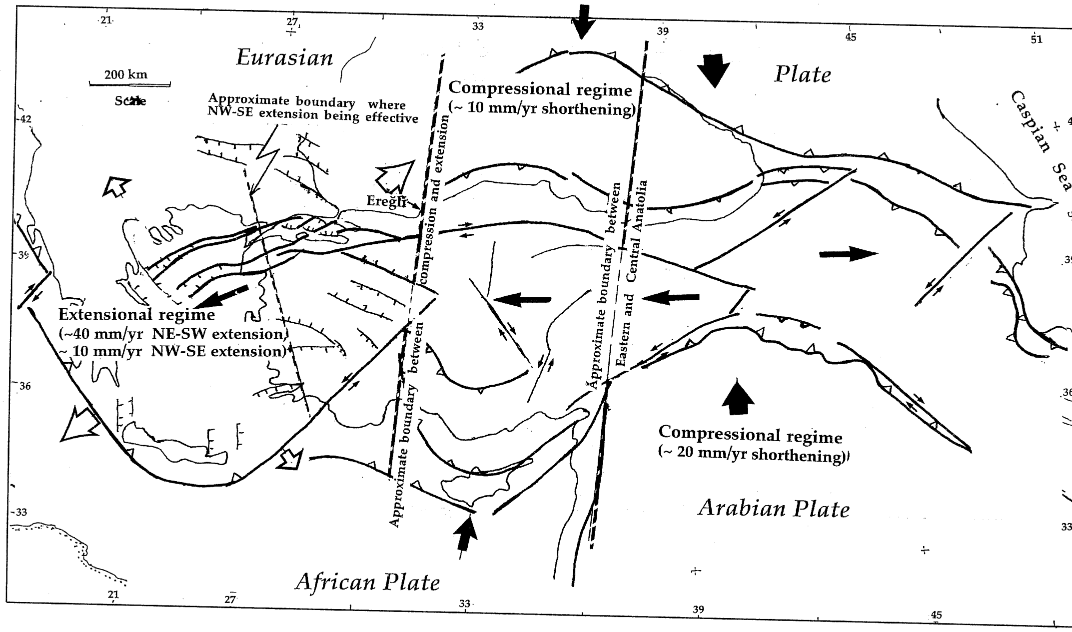

Fig. 18. Simplified tectonic map of the Eastern Mediterranean region showing the main tectonic domains and related extension and compression directions and rates obtained mostly from the GPS measurements (Reilinger *et al.*, 1997). Solid large arrows indicate directions of shortening in Eastern and Central Anatolia, the same as northward motions of Arabia and Africa. Open arrows show directions of extension in Western Anatolia and Aegean regions. Smaller solid arrows indicate side escapes of continental blocks.

sula. The velocities increase from 24 mm/yr in the Southern Marmara Sea to 36 mm/yr towards the western flank of the Hellenic arc, and to 30 mm/yr towards the eastern flank of the Hellenic arc. The GPS vector also has a sudden 10 southerly kink along a line shown in fig. 18. The same line joins the horizontal westward opening «>» shape gulfs along the west coast of Anatolia such as Saros, Edremit, İzmir and Gökova.

11. The Hellenic arc

The differences in tectonic regimes in Western Anatolia/Aegean and Central Anatolia could be related to the kinematics of the subducting slabs under them. For example, it is now well established by seismic tomographic studies (*e.g.*, Maulkamp *et al.*, 1988; Spakman, 1991; Wortel and Spakman, 1992) that the slab being subducted along the Hellenic trench and/or the Mediterranean Ridge Accretionary Complex seems to have a high dip angle and thus, the Hellenic trench has a retreating nature (*e.g.*, Le Pichon and Angelier, 1979, 1981; Toksöz and Kasapoğlu, 1988; Taymaz *et al.*, 1990; Royden, 1993). The Hellenic subduction is believed to be responsible for the extension in Western Anatolia and the Aegean since at least the middle Miocene (*e.g.*, Seyitoğlu and Scott, 1991, 1992; Barka *et al.*, 1994; Armijo *et al.*, 1996). According to paleomagnetic data (Kissel and Laj, 1988; Kissel *et al.*, 1993), the Hellenic arc was a rectilinear feature before the Langien (middle Miocene) and since then it has formed as a southward arc. Le Pichon *et al.* (1995) suggested that the African promontory began to collide with central part of the arc in Pliocene.

12. Discussion

Even though GPS measurements have provided valuable information on the and kinematics of the plate motions, there are a few inconsistencies between velocity vectors obtained from GPS measurements and fault kinematics and rates obtained from active fault studies. These inconsistencies and some interpretations are discussed below.

a) The active fault pattern in Eastern Anatolia suggests that the Northeastern Anatolian block escapes east-northeastward. Philip *et al.* (1989) suggested the motion of Eastern Anatolia is northward and two strike-slip faults the left-lateral Northeast Anatolian fault and right-lateral fault along the Talesh mountains are the main boundaries for this indentation (fig. 19a). Figure 19b shows expected slip lines (fault pattern) from a wedge shape compressed Prandtl cell. This fault pattern was used by Cummings (1976) for the Mojave desert in California. A similar fault pattern exists in Eastern Anatolia which agrees with eastward escape thrusting on oceanic crust of the Southern Caspian Sea (fig. 19c). Although this is confirmed by only two GPS sites, Erzurum and Kars, the Garni site located in Armenia shows a northward motion. However, the fault plane solution of 1978 and 1980 earthquakes along the approximately N-S trending Talesh fault (Berberian, 1981; Jackson and McKenzie, 1984) indicated thrusting which is consistent with eastward motion of the Northeast Anatolian block. Thus, the Talesh mountains are more related to thrusting than right-lateral strike-slip faulting.

b) Eastern Anatolia is separated from Central Anatolia by an approximately N-S trending imaginary line as a continuity of the Dead Sea fault even though the eastern tip of the Anatolian block has a wedge geometry. To the east of this line there is at least 20 mm/yr shortening taken up by the zone extending from the Bitlis thrust to Greater Caucasus. To the west of this line, Central Anatolia, the deformation pattern suggests that there is NNE-SSW shortening while it is rotating anticlockwise along the North Anatolian fault. This is related to the collisional nature of the Cyprus arc. Thus, the 10 mm/yr northward motion of Africa is taken up by a wide region from the Cyprus arc to the Crimean thrust including active faults in Central Anatolia and thrusting along the southern shore of Black Sea (fig. 18).

c) Neotectonic structures and some GPS results show that the eastern flank of the Isparta Angle which is the northwestern continuation of the Cyprus arc into SW Anatolia, has less motion relative to Central Anatolia. This difference is taken up by the Sultandağ mountains and Aksu thrust (fig. 18).

d) The NE-SW extension in Western Anatolia is separated from NNE-SSW compression in Central Anatolia by a N-S line connecting the tip of the Isparta Angle and Ereğli where the Black Sea coast sharply turns to the south. However, the real separation has a wedge geometry defined by the Eskişehir and Fethiye-Burdur faults (fig. 18). At the tip of the wedge, the eastward continuation of the Simav graben defines this wedge better, however, geological observations suggest that to the north of this line there are two other kinematically similar faults, the Kütahya and Eskişehir fault which should be included in Western Anatolia. This is not very clear from the GPS measurements since rates along these faults are not more than a few millimetres/year. However GPS velocities increase in Western Anatolia relative to Central Anatolia.

e) Two extension directions exist in Western Anatolia NE-SW and NW-SE caused by the western and eastern flank of the Hellenic arc respectively (fig. 18). This is due to the fact that the southern tip of the arc had already collided with Africa in the Pliocene, there are only oceanic crusts left south of the eastern and western flanks of the Hellenic arc. From the GPS vectors the western flank causes a higher rate of extension than the eastern flank. However the NW-SE extension became effective to the west of a line connecting the tips of the Gulf along the west coast of Anatolia such as Gökova, Edremit and Saros (personal discussions with S. Müller and F. Oktay, 1996). Results of this second order extension GPS vectors have at least ten sudden directional changes along this line (discussion with Le Pichon, 1995). This also results in occurrence of NW-SE, ENE-WSW and NE-SW trending nor-

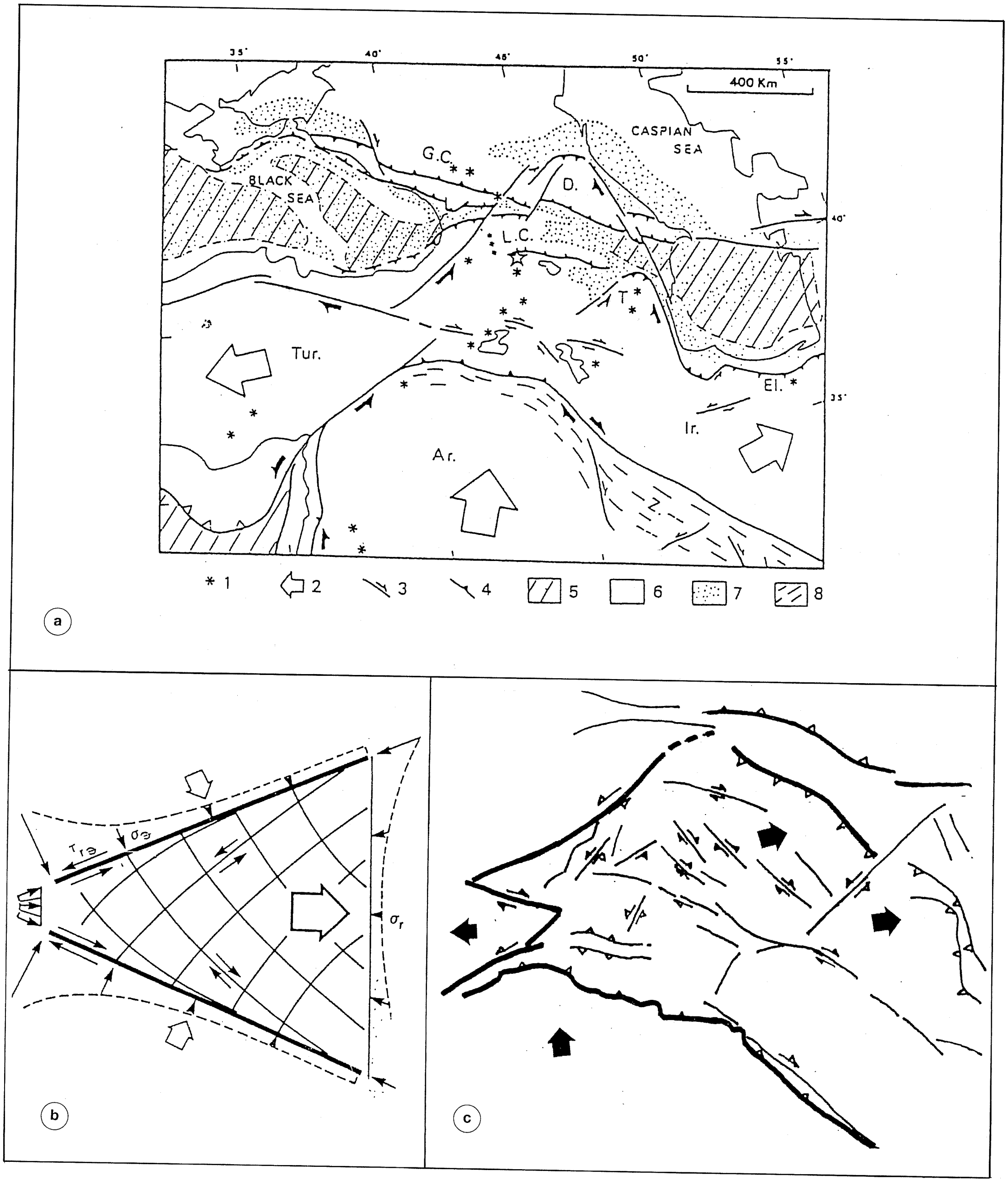

Fig. 19a-c. Neotectonic models for the Eastern Anatolia and Caucasus regions. a) From Philip *et al.* (1989) suggesting that the main motion of the Eastern Anatolia is NNE direction; b) Prandtl's compressed cell showing the fault patterns (Cummings, 1976); c) fault pattern in Eastern Anatolia and surrounding regions which is very similar to the fault pattern obtained from Prandtl's cell indicating that Eastern Anatolia moves westward rather than to the north.

mal and normal-oblique faults in Western Anatolia. Consequently, smaller continental blocks sliced by these faults drift towards the southwest.

f) This interpretation illustrates that continental collision causes sideway escapes of wedge shape continental blocks (Burke and Şengör, 1986). The tectonic features of Western Anatolia show that wedge shape geometry can occur under an extensional regime where internally sliced blocks drift in certain directions.

13. Conclusions

The Eastern Mediterranean region offers ideal examples of continental collision (*e.g.*, in Eastern Anatolia), oceanic subduction (*e.g.*, along the Hellenic arc) and transition between partly collision and partly subduction (*e.g.*, along the Cyprus arc), moderate to extensive internal deformations within the overriding continental block (*e.g.*, Anatolia). In this region, all these processes are interacting together at present. The counter-clockwise rotation of the Anatolian-Aegean block is only a preliminary interpretation. A combined analysis of 1988-1994 GPS measurements and neotectonic observations suggest:

a) NE Anatolia moves east-northeastward.

b) The North Anatolian fault has a slip rate 26 ± 3 mm/yr, but at least 5 mm/yr should be extracted for the western drift of Pontides.

c) The present neotectonic regime in Western Anatolia and the Aegean Sea differs from that in Central Anatolia. In Western Anatolia and the Aegean, approximately E-W and WNW-ESE trending graben and normal faults are dominant structures and this region is extending in a SW-NE direction at a rate of roughly 30-40 mm/yr (Oral *et al.*, 1995; Le Pichon *et al.*, 1995). In contrast, neotectonic structures in Central Anatolia consist of strike-slip and thrusting, indicating that Central Anatolia is undergoing a NNE-SSW shortening (see also Şengör *et al.*, 1985) and is moving in a more westerly direction at a rate of about 15-20 mm/yr (Oral *et al.*, 1995). Furthermore, internal deformation in Central Anatolia appears to be significantly less than in Western Anatolia as testified by seismic activity (*e.g.*, Ambraseys, 1970, 1975). This suggests that Central Anatolia is more rigid even though it accommodates a number of large active faults. Western Anatolia is separated from Central Anatolia by the WNW-ESE trending Eskişehir fault and the NE-SW trending Fethiye-Burdur fault zone (fig. 14). GPS measurements indicate that the interior of the Isparta Angle zone is a relatively stable domain, located between Central and Western Anatolia. However, each of its flanks has been affected by deformation due to the motion of the neighboring tectonic blocks with respect to Eurasia.

d) The segmentation of the subduction in the Eastern Mediterranean into the Hellenic and the Cyprus arcs is believed to be responsible for compartmentalization of tectonic regimes on the overriding Anatolia (*i.e.*, the extension in Western Anatolia and Aegean region and compression in Central Anatolia).

It is hoped that continuing GPS efforts will clarify problems related to the kinematics and dynamics of the Eastern Mediterranean region as the amount of sites and accuracy GPS measurements increase.

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