A REVIEW OF THE TECTONICS OF THE AEGEAN REGION

Tolga KOMUT

Boğaziçi University

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TOLGA KOMUT

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ABSTRACT

The Aegean region is a remarkably deforming part of the Alpine-Himalayan orogenic belt, which has the highest seismic activity in Europe. An extensional deformation regime has led to subsidence of the continental crust over all the Aegean region behind the south Aegean consumption boundary. The region that is mainly under pure shear stress is an internally deforming part of the counterclockwise rotating (relative to Eurasia) Anatolian plate. To understand the tectonics and dynamic evolution of the region many geoscientists have collected numerous data and interpreted them.

In the present study, all available data have been compiled into a unique database. For this purpose, as a first step, raw and processed information have been compiled and classified with their sources, date information and criticism of their accuracy and limitations. This step allows the comparison of similar types of study and data in order to understand basic problems more clearly. As the second step, all interpretations of the studies are discussed and critically reviewed by considering associated data and methodology on a regional scale. The goal of the work is to determine the problems with highest priority, find the gaps in the database.

ÖZET

Ege bölgesi geniş bir levha çarpışma zonu olan Alp-Himalaya dağ kuşağının oldukça farkedilir şekilde deformasyona uğrayan bir kısmıdır. Güney Ege yitim zonunun arkasında bulunan ve tüm bölgeyi etkisi altında bulunduran aktif genişleme rejimi bu bölgedeki kıtasal yapıdaki kabuğun çökmesine sebep olmuştur. Bu bölge tümü ile, hareketi saat istikametinin tersi yönünde olan Anadolu levhasının batı bölümü olarak düşünülmekle beraber bölgenin tümü ile basit makaslama kuvvetleri etkisi altında olması bir yaygın (diffuse) deformasyonu göstermektedir. Karmaşık tektonik yapısı gereği bilim adamlarının dikkatlerini üzerine çekmekte ve bölgenin kinematiği, buna sebep olan dinamik mekanizma/lar ve dinamik evrimin açığa kavuşturulması için bir çok araştırmalar yapılmaktadır.

Bu tarama/değerlendirme çalışmasının amacı, öncelikle bölge ile ilgili yapılan çalışmaları levha tektoniği kuramının çıkışından bu yana geçen süreç içerisinde derlemek, böylece bunları belirlenmiş başlıklar altında tek bir bakış açısı altına getirmektir. Bu amaç doğrultusunda, önce ham ve işlenmiş veriler, kaynakları ve zamanları belirtilerek bir araya getirilmiştir. Bu veriler ve ilgili çalışmaların sonuçları güvenilirlikleri, amaca uygunlukları ve yeterlilikleri açısından karşılaştırılmıştır ve tartışılmıştır. Bu yaklaşım aynı tür çalışmaların birbirleri ile kolayca karşılaştırılmasını sağlamakta, temel sorunların ve tartışmaların neler olduğunu ortaya daha açık bir şekilde koymaktadır. Çalışmanın ikinci aşamasında, çalışmalarda yer alan tüm görüşler bölgesel olarak, ham veri yeterliliği kıstası ve kullanılan metod gözden kaçırılmadan, ele alınmış ve eleştirisel bir yaklaşım ile tartışılmıştır. Sonuçta bir durum değerlendirmesi yapmak ve öncelikler belirlemek olanağı elde edilebilecektir.

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1. INTRODUCTION

The deformation pattern in the Mediterranean region that constructs a low elevated part of the Alpine Himalayan belt is rather complex, as usually occurs in continental collision zones. Five arcuate tectonic belts are situated in the Mediterranean region (in Alpine Himalayan belt). Four of them (Gibraltar, Tuscanian, Calabrian, South Aegean) have island arc system characteristics, with marginal seas. The fifth (Cyprus) arc is a relatively narrow thrust zone that is not clearly evident. Much of the structure and seismicity of the Alpine Himalayan collision zone (southern Europe, North Africa, the Middle East and Asia as far east as the Himalayas and China) can be attributed to the closure of the Neo-Tethyan Ocean.

The higher spreading rate in the South Atlantic (40 mm/yr) as compared to that in the North Atlantic (25 mm/yr) has been causing a gradual counterclockwise rotation of the African plate. This rotation results in a N-NW directed push against Eurasia. In general view, the Aegean region is situated in a part of the convergent boundary between the African plate which has rotated counter-clockwise with respect to Eurasian plate during last 92 My (*Mueller & Kahle, 1993*) and Eurasian plate. Between them, a roughly N-S directed lithospheric shortening rate is increasing from west to east (*Mueller et al., 1997*). The African plate (Tethys ocean) is being consumed by subduction towards the north. On the other hand, the Mediterranean lithosphere does not left east around Hatay (e.g., *Şengör, 1982*) where a thrusting and thickening process is taking place between the Arabian and Eurasian plates.

The crust of the Alpine Himalayan belt is constructed of approximately parallel trending units whose configuration developed in paleotectonic times. To both sides of the Aegean Sea, the units that are massifs and suture zones (ophiolite and blue schists) are linked to each other. As the Tethys closed, pieces of continent (terrenes) were swept together (accreted) to form a tampon mass between African and Eurasian plates. Evidence for this accretion may be found in the Turkish and Greek ophiolite suture zones. In the Aegean area, these units and ophiolitic suture zones are cut by prominent post orogenic active extensional structures. The Eocene blueschists and eclogites in the Cyclades in the centre of the Aegean Sea indicate that during the Mid-Eocene the region was characterised by a thick continental crust (> 70 km) because of a compressional regime. The timing and cause of transition from compressional to extensional regime in the Aegean region and the tectonic setting of the Oligo-Miocene calc-alkaline magmatism in NW Turkey are controversial. Estimates of the age of transition from compressional to extensional regime range from Late Oligocene to Late Miocene. The underlying cause of the extensional regime is also debated. One possibility is the gravitational collapse of the major topography created as a result of the mid-Eocene crustal thickening. A second is the rollback of the South Aegean Subduction Zone, creating a back-arc-type extension. The third are the forces of the boundaries created by the collision of the African-Arabian plate with the Eurasian plate. The Aegean region is bounded to the North by the stable continental Eurasian plate, to the west by the Adriatic region, to the east by the central Anatolian, and to the south probably by the oceanic material beneath the Mediterranean Sea (Northern edge of the African plate). The region is situated north of the South Aegean Arc (SAA). Northward motion of the Arabian plate relative to the Eurasian plate and a free retreating subduction boundary along the South Aegean Trench causes lateral motion of the Anatolian mass (*Le Pichon et al., 1995; Oral et al., 1995; Reilinger et al., 1997*). This motion causes right lateral motion along the North Anatolian Fault Zone (NAFZ) and left lateral motion along the East Anatolian Fault Zone (EAFZ) (*McKenzie, 1972; Oral et al., 1995*). The horizontal motions suggest that the westward movement of Anatolian plate carries Anatolia through an extensional zone where the continent is stretched and its thickness halved (e.g., *McKenzie, 1978*). The Aegean area includes the Aegean sea, mainland Greece, Bulgaria southern Macedonia and western Turkey. The area deforms as a part of Anatolian mass and is called as Aegean region. The existence of a calc-alkaline inner volcanic arc, the spatial distribution of earthquakes and detailed tomographic studies indicate the existence of a northward - dipping subducted slab beneath this region. The region is an area of present day N-S extension above the South Aegean Subduction Zone.

Elevation in the Aegean region increases from the deepest basins in NW-SE direction Aegean axial trough to the SW and NE; in basins with a ENE-WSW strike in the North Aegean to the N; Cretan basin to the North Cyclades Plateau and to the South Aegean Island Arc in E-W elongation.

The region SW of the Aegean axial trough is characterised by NW-SE trending grabens that result from the SW stretching of the region. To the NE of this line cross grabens trend NE-SW approximately parallel to the SW, and they have a strike-slip component with considerable dip slip component. These trends become progressively more E-W and NW-SE to the further

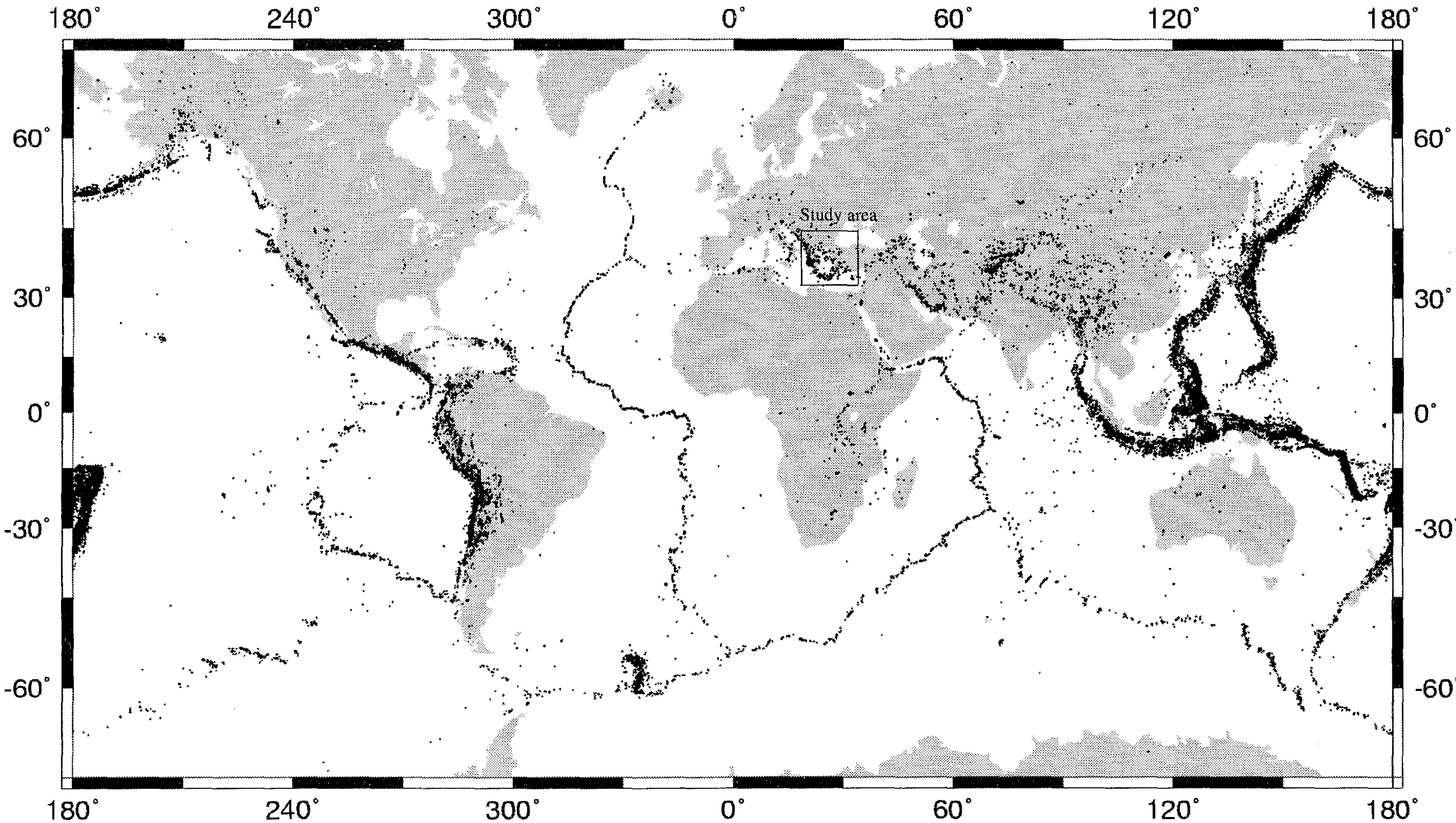


Figure. 1.1. The plate boundaries between the plates outlined by the recent seismicity. Events are plotted with magnitude >5.5. Source: USGS catalogue.

east in western Turkey. The Cretan Basin that is the most aseismic area in the region possibly had been stretched parallel to the SAA. However, possible collision of the SAA with the African promontory stopped further stretching of the Cretan basin. After that event, the Aegean region underwent a new stage of the extensional regime. The Aegean Sea floor (about 350 km mean depth) is seen as a high plateau between the deeper Black Sea floor (mean depth 1300 m) and the Mediterranean Sea floor (Mean depth 1500 m) (*Küleli et al., 1993*). The region is characterised by high heat flow, which is related to thinned and deformed (stretched) continental crust. This thinning is now continuing. For this reason the area is the most seismically active and rapidly internally deforming area of the entire Alpine-Himalayan belt and the continents in the world today (*McKenzie, 1972; Mercier et al., 1977; Ekström & England, 1989; Jackson et al., 1992*). It is called the Aegean extensional region. The location of the region and main names are shown in Figure 1.1 and Figure 2.2, respectively.

The basic plate tectonic hypothesis is that the Earth's uppermost layers are divided into rigid plates, with narrow deforming borders could not rigorously apply to continental crust. For this reason the region continues to serve as a testing ground for new methods and theories as a natural laboratory (*Jackson, 1994*). Between the African and Eurasian plates beneath the Aegean region with continental crust, the boundary cannot be regarded as a single narrow deforming border. The region must be considered as an extensional part of a diffuse (wide) boundary (broad transition zone) between corresponding large plates. Probably brittle crust is being deformed diffusely by micro blocks on continuously deforming ductile material in the region (see *Thatcher, 1995*).

There is a lot of unknown geological detail in the complex Aegean region that needs to be derived. Data are often hidden away due to nationalism and protectionism. For this reason, the production of maps is extremely slow, and available maps very often lack some information. Because of the absence of data from each branch of geoscience and hard problems that are not seen in any other region of the Earth, experts in different fields should join forces to compile a list of major pieces of evidence and of the related geodynamic implications. There is always the danger that structural mapping, data processing, instrumentation or one branch becomes a goal. Furthermore there is a notable lack of dialogue between the different teams (see also *Schmid et al., 1996*). Dialogue and cooperation across borders and across scientific disciplines

Figure.1.2. Location map of Aegean and surrounding area.

are the main requirements for the promotion of future research on the Aegean region or anywhere. Many authors made studies to integrate the geological, geophysical and geodetic data with the kinematic and dynamic problems in the Aegean area. However only limited parts of known information are generally used in geodynamic reconstructions, because it is hard to search the whole literature for a scientist who studying a single branch of geoscience. This is partly because it is hard to reach geological and geophysical wide information. To handle large, multidisciplinary data sets with varying quality and resolution, *Şeber et al. (1997)* have recently adopted a Geographic Information System (GIS). It is an approach for the construction of a multipurpose database to look at the geodynamic problems in a comprehensive and unconventional way. The resulting maps are available at their web site at http://atlas.geo.cornell.edu/. Many authors published review papers that are based on The Aegean region or some parts of the region or that include the region. Some of these consider all the geophysical and geological methods or some selected methods. Some of these are listed in next paragraph.

Previously *Lort (1971)* gave a very brief description of surveys in the eastern Mediterranean. *Horvath & Beckermen (1982)* critically review the geological environment, structure and evolution of Mediterranean back-arc extensional basins as well as the Aegean region. *Jackson & McKenzie (1988)* also gave information for many extensional basins in the Mediterranean region as a general review. *Mueller & Kahle (1993)* made a good general review that was based on geological and geophysical data in the Mediterranean region. *Mercier et al. (1989)* reviewed the structural data in the Aegean region. *Jackson (1993)* made a critical review on deformation of the Aegean region and its surroundings. *Babbucci et al. (1997)* made a general geological review based on available observations in the various branches considering the Alpine Himalayan belt. *Mueller et al. (1997)* gave a general review which considered available geodetic, geophysical and geological data in the Anatolian-Aegean region. *Canıtez & Toksöz (1982)* made a review based on crust and upper mantle structure in Turkey. *Barka (1997)* gave a good review of existing studies on tectonics of the Marmara region. *Berckhemer (1977)* has reviewed the existing results on the South Aegean Subduction Zone. *Jongsma (1975)*, *Udias & Buforn, (1991)*, *Mercier (1981)* and *Morelli (1985)* compared the Aegean subduction system's properties with the other arcs in the world. *Arpat (1976)* made a brief critical review of geophysical and geological data in the Aegean region. *Küleli et al. (1993)* provided a review on existing syntheses based on the Aegean region. *Jackson (1994)* has presented an exhaustive review of the active tectonics of the Aegean region. The review of *Schmid et al. (1996)* is concerned with the active tectonics of the region. Recently *Dadaşbilge (1997)* reviewed the geology and geophysical state of the Aegean region.

This work is a critical and detailed review of the all available geoscientific data and literature in the Aegean Region. Our main goal is to provide up-to-date information that is easy to reach about geoscientific data, dynamic and kinematic hypothesis, which has been studied intensely since initiation of the plate-tectonic concept. The challenge is to determine the problems with highest priority, reveal the gaps in the database and selection of appropriate data for further studies. We try to draw up an inventory of what is known at present based on numerous observational data, trying to link them in an understandable way. To be able to access the large amount of information rapidly all available literature has been compiled into a unique database as a reference library of the study (see Appendix 2.). This database is accessible through a computer and includes a large amount of literature and data information. This study is not perfectly complete but provides the basis of the working goal proposed by this study. The GMT (Generic Mapping Tools) algorithm (*Wessel & Smith, 1995*) was used to prepare many of the maps. The Microsoft Office '97 (Professional) software was used for preparation of some tables and text and especially to make the "comment index" for all scanned texts that come from various parts of the numerous papers. The IRISpen (optical character recognition) system was used to edit texts from hard copy papers.

2. DATA: Review

During the last 25 or 30 years, many geological, geophysical and geodetic data have been collected in the Aegean area, providing important information about the recent tectonic and geodynamic evolution of that region. In this section available raw and processed information and data have been compiled and classified with their sources, date, other information and criticism of their accuracy and limitations. This step allows the comparison of similar types of study and data, to help understand basic problems more clearly. Most of the listed data are taken from literature that is reviewed in this study and some others are taken from references therein when we could not reach the original source.

2.1. GEOLOGICAL MAPS

There are many maps which have been produced by various mainland geological studies. However, in our study some of the well known main geological map series listed. Previously a simplified geological map of western Turkey (1: 500,000) was redrawn from the map at half a million scale in three sheets (İstanbul, İzmir, Denizli) by the Institute of Mineral Research and Exploration, Ankara, Turkey (MTA 1961, 1964). In 1989 the Geological Map of Turkey (1:12,000,000) was compiled. The 1:500,000 geological map of Turkey now has been digitising. There is a geological map of Greece on a scale of 1:500,000 published by Geology and Subsurface Research, Athens (1954). It is known that there is a geological map of Greece on a scale of 1:50,000 but we could not obtain this information. A compilation of the geology of the Mediterranean area on a scale of 1:2,500,000 was drawn by *Biju-Duval et al. (1974).*

2.2. MAGMATICS

2.2.1. Data

An active volcanic arc called the South Aegean Volcanic Arc (SAVA), studied by *Nicholls (1971), Ninkovich and Heezen (1965), Caputo et al. (1970), Ninkovich and Hays (1972), Fytikas et al. (1976), Bellon et al. (1979), Gülen (1990), Altherr et al. (1988). Fytikas et al. (1984).* They discussed the Quaternary evolution of the volcanism in the whole region based on their analysis with some others from previous studies. *Gülen (1990)* studied a substantial number of high-precision isotope data (Table. 2.1) that have been obtained on the magmatic rocks of the region since the 1977 Colloquium in İzmir by various studies. *Ertürk et al. (1990)* also reviewed the tectonic evolution of the Aegean region by considering previous studies. There are many other local studies.

Turkey:

In central Anatolia and western Turkey *Innocetti et al. (1977)* studied the distribution of the rare earth elements in different members of each magmatic series by using stratigraphic and geochronological data. *Ercan (1979)* extensively studied the volcanics of western Turkey, Thrace and Cretaceous volcanites of the coasts of Black Sea. *Ercan (1982)* combined his data with the previous studies (e.g., *Ercan, 1979, 1980; Ercan et al., 1977, 1979).* In addition he reviewed the results of a collateralize project of MTA; Geology Department, İstanbul University; Earth Science Faculty, Ege University; Earth Science Faculty that called "West Anatolia Tertiary Magmatizm and Stratigraphy". He also reviewed some other previous results of these studies. After that, additional radiometric ages were obtained from the northwestern Anatolian volcanics by *Ercan et al. (1985). Yılmaz (1989)* made a review of magmatics in western Turkey and has synthesized his data and compared ideas with the observations of others. *Francalanci et al. (1990)* added new chemical and isotopic data on alkaline magmatism of the Aegean Sea, Western Turkish peninsula. Geochemical results from the Eocene-Miocene volcanics are interpreted by *Ertürk et al. (1990)* in western Turkey (see also *Ercan, 1980; Savaşçın, 1982*).

Table. 2.1. Isotopic ages of Volcanics and granitoids in the Aegean region (after *Gülen, 1990*).

Location	Symbol	Age (M.y.)	References
VOLCANICS:			
Santorini	Va	0.6-1.8	11
Milos	Vb	0.1-3.5	12,15
Antiparos	Vc	4.0-5.4	16
Nisyros	Vd	0.2	9,12
Methana	Ve	0.3-0.9	12
Kos	Vf	1.0-10.0	2,5
Aegina	Vg	2.1-4.4	18,20
Bodrum	Vh	7.8-10.6	21
Patmos	Vi	3.5-7.2	12,21,22
Samos	Vj	7.8-9.3	21
Euboea	Vk	13.2-14.8	2,12,13
Kalogeri	Vl	6.0	15
Söke	Vm	7.0	8
Chios-Çeşme	Vn	14.3-17.0	3
Psara-Urla	Vo	15.0-17.7	7
Skiros	Vp	15.0	13
Karaburun, İzmir	Vq	19.2-21.3	7
Psathoura	Vr	0.7	14
Agios Evstratios	Vs	18.0-23.2	13
Lesbos-Dikili	Vt	15.5-18.0	7,19
Kula	Vu	1.1	7,8
Bergama-Ayvalık	Vv	17.3-23.6	4,7,17
Limnos	Vw	17.8-21.6	13
Bigadiç-Kepsut	Vx	19.6-21.7	8
Alexandroupolis	Vy	30.0-33.1	14
Xanthi-Souflion	Vz	23.6-30.6	14
GRANITOIDS:			
Serifos	G1	8.0-9.5	1
Naxos	G2	11-13	1
Kos	G3	9-12	1
Laurion	G4	15	1
Mikonos	G5	9-11	1
Tinos-Bodrum	G6	13-15	1
Menderes	G7	16.4-18.1	10
Eğrigöz	G8	20-24	6
Tuzla	G9	28	6,12
Çataltepe	G10	20-23	6
Uludağ	G11	25-27	6

(1) Altherr, R. et al. 1988; (2) Bellon, H. et al. 1979a; (3) Bellon, H. et al. 1979b; (4) Benda, L. et al. 1974; (5) Besang, C. et al. 1977; (6) Bingöl, E. et al. 1982; (7) Borsi, S. et al. 1972; (8) Ercan, T. et al. 1985; (9) Di Paola, G. M. 1974; (10) Dora, O. O. et al. 1987; (11) Ferrara, G. et al. 1980; (12) Fytikas, M. et al. 1976; (13) Fytikas, M. et al. 1979; (14) Fytikas, M. et al. 1984; (15) Fytikas, M. et al. 1986; (16) Innocenti, F. et al. 1982; (17) Krushensky, R. D. 1976; (18) Müller, P. et al. 1979; (19) Pe-Piper, G. 1980; (20) Pe-Piper, G. et al. 1983; (21) Robert, U. & Cantagrel, J. M. 1977; (22) Wyers, G. P. 1987.

2.2.2. General Interpretation

The trace element isotope analyses suggest that the calc-alkaline volcanics that are more acidic are the products of an interaction between mantle and crustal reservoirs. Whereas the alkaline volcanics that are more basic come from the mantle with negligible continental crustal (i.e. acidic material) contamination (*Savaşçın, 1990*). On the other hand, the basaltic suite rocks occur as locally developed patches and preferentially follow some distinct zones of fracture or major fault directions especially in the grabens of western Turkey (*Yılmaz, 1989*). These main magmatic rock families were observed in the Aegean region in various ages.

To better understand the recent tectonic nature, the relationship with the geodynamic setting should be recognized. For this reason it is important to find the source characteristics of the different types of magmas in the region. Some relations were sought by researchers between distance from the arc, age and chemical composition of rocks. This relation defines the nature of the subduction zones and their behavior in time. In the Aegean region, subduction related back-arc calk-alkaline volcanism is obvious. It is generally accepted that the volcanic activity along a convergent margin develops only where the depth of the subducted lithosphere reaches suitable values, between approximately 100 and 200 km (*Gill, 1981*, see also *Fytikas et al., 1984*). *Yılmaz (1981)* defined a Middle-Late Cretaceous magmatic arc, related to the Northern branch of the Neo-Tethys. *Borsi et al. (1972)* firstly recognized that the magmatism exhibits a progressive shift in time toward the south in the Aegean region (see also *Fytikas et al., 1979; 1984; Savaşçın & Güleç, 1990; Papadopoulos, 1997*) (Fig. 2.1). This finding suggests that the subduction zones could be shifting towards the south somehow. Accordingly, three volcanic phases are defined by *Boccaletti et al. (1974)*. These are early Miocene (23-14 Ma) in the Central-North Aegean, Late Miocene (14-7 Ma) in the south central Aegean and Plio-Quaternary (4 Ma-present) in the South Aegean Island Arc (see also *Jongsma, 1975; Papadopoulos, 1997*). According to the extensive recent study of *Gülen (1990)* in the Aegean region, a progressive younging trend of calk-alkaline volcanics and granitoids is also observed from NNE towards the south-southwest to the active subduction zone (Fig. 2.2). He distinguishes four volcanic zones in north-south direction. To the North of the active volcanic chain in Aegean SAVA a paleo-volcanic zone is situated along the southern periphery of the Attic Cycladic/Menderes metamorphic belt (*Gülen, 1990*). This zone is referred to as the "inner arc" by *Innocenti et al. (1979)*. Further North an older (Oligocene/Middle Miocene) volcanic zone (*Gülen, 1990*) covers an extensive area in the Northern Aegean region where the Attic/Cycladic/Menderes metamorphic belt forms its southern boundary. The previous observations could be explained in this way. *Fytikas et al. (1984)* also pointed out that Oligocene volcanics occur only in the northernmost part of the belts (Thrace) whereas Miocene products are confined to the southern sector of the belt (Fig. 2.1). The oldest and more northerly position Macedonian volcanics (Alexandroupolis, Xanthi, Souflion) exist according to the data (Figs. 2.1, 2.2) (*Fytikas et al., 1984, Gülen, 1990*). On the contrary there seems to be a reverse trend for the alkaline volcanics based on a

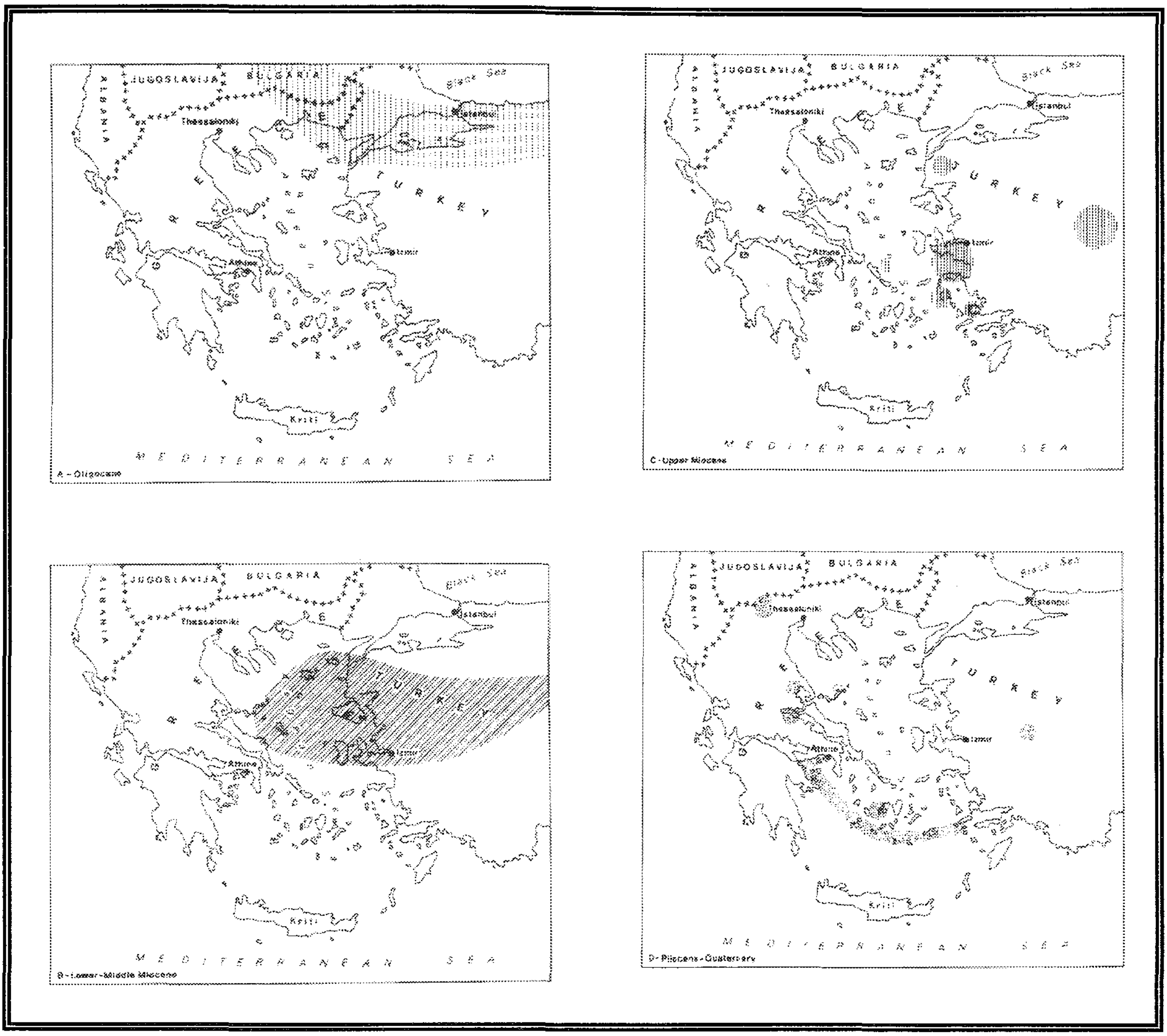


Figure. 2.1. Relative distribution of the volcanic activity in the Aegean area during Oligocene (A), Lower-Middle Miocene (B), Upper Miocene (C) and Pliocene-Quaternary (D).

limited data set relative to more acidic volcanics (calk-alkaline) (Fig. 2.2) (*Gülen, 1990*). These types of volcanics get younger away from the arc to the North. The alkaline more basic volcanics appear to be related not the subducted slab but with the extensional (W-E) tectonic features in the Aegean area like the other extensional back-arc regions (*Dewey & Şengör, 1979; Morrison, 1980; Savaşçın, 1982, 1990*). According to this theory, this negative trend may imply that the extension in the Aegean region might have started first in the south, and then propagated Northward in time (*Gülen, 1990*). This idea well fits with the thin viscous sheet model of *Sonder and England (1989)* to investigate thermal dependent rheology for back arc continental areas. According to the model, extension progresses towards the north.

But *Savaşçın (1990)* reported E-W trend of alkali basaltic magmatism becomes younger towards the west in western Turkey along N-S extension related grabens. But it should not be forgotten that the data are limited (*Gülen, 1990*).

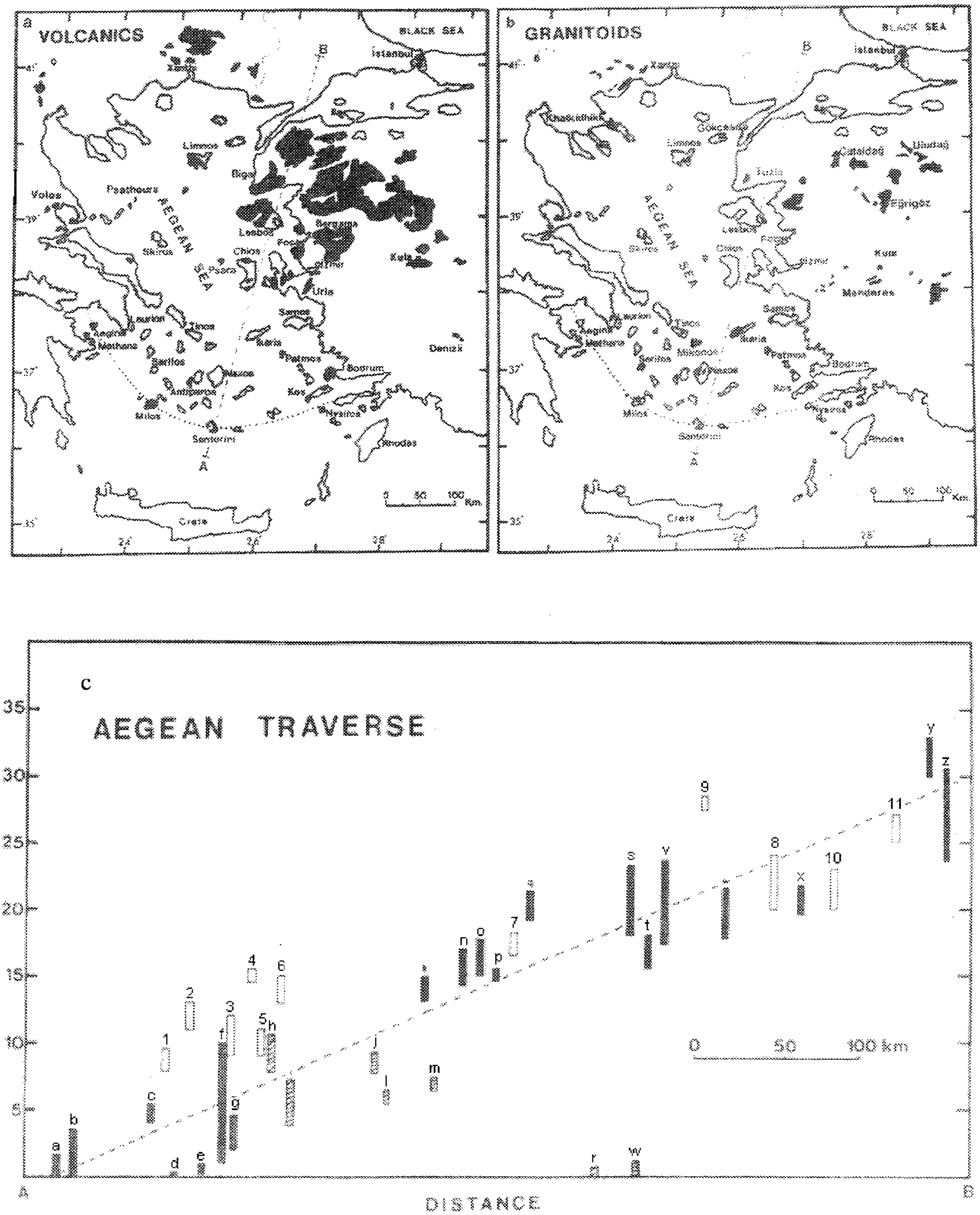


Figure. 2.2. Maps showing (A) the distribution of the Tertiary to Quaternary volcanics, and (B) the Tertiary granitoids in the Aegean region. The active arc is shown by dotted curves. (C) Age-distance correlation plot for the Aegean volcanics and granitoids. The locations are projected onto the Aegean Transverse (line AB in (A) and (B)), to represent relative distances that are orthogonal to the Aegean trench. Symbols denote the localities that are listed in Table. 2.1. Alkaline volcanics are indicated by hatched patterns (after *Gülen, 1990*)

Studies of the composition of the lavas by *Nicholls (1971)* and *Ninkovich & Hays (1972)* show that there is a systematic increase in K_2O and Rb versus SiO_2 with distance from the South Aegean Trench (see *Jongsma, 1975*). Accordingly, *Kuno (1959)* argued that as the Benioff zone depths increase the magma becomes more alkaline. Hence, away from the arc towards the inside of the arc, the K_2O/SiO_2 ratio and also the Al_2O_3/SiO_2 ratio increase (see *Kuno, 1960; 1966; Kushiro & Kuno, 1963; Diskinson & Hatrenton, 1967; Miyashiro, 1972*). *McKenzie & Yılmaz (1991)* also reported an increasing TiO_2/K_2O ratio and TiO_2 percentage away from the arc in the North (Fig. 2.3). However, the study of *McKenzie & Yılmaz (1991)* shows that in western Turkey no trend is observed in neither E-W nor S-N direction in K_2O

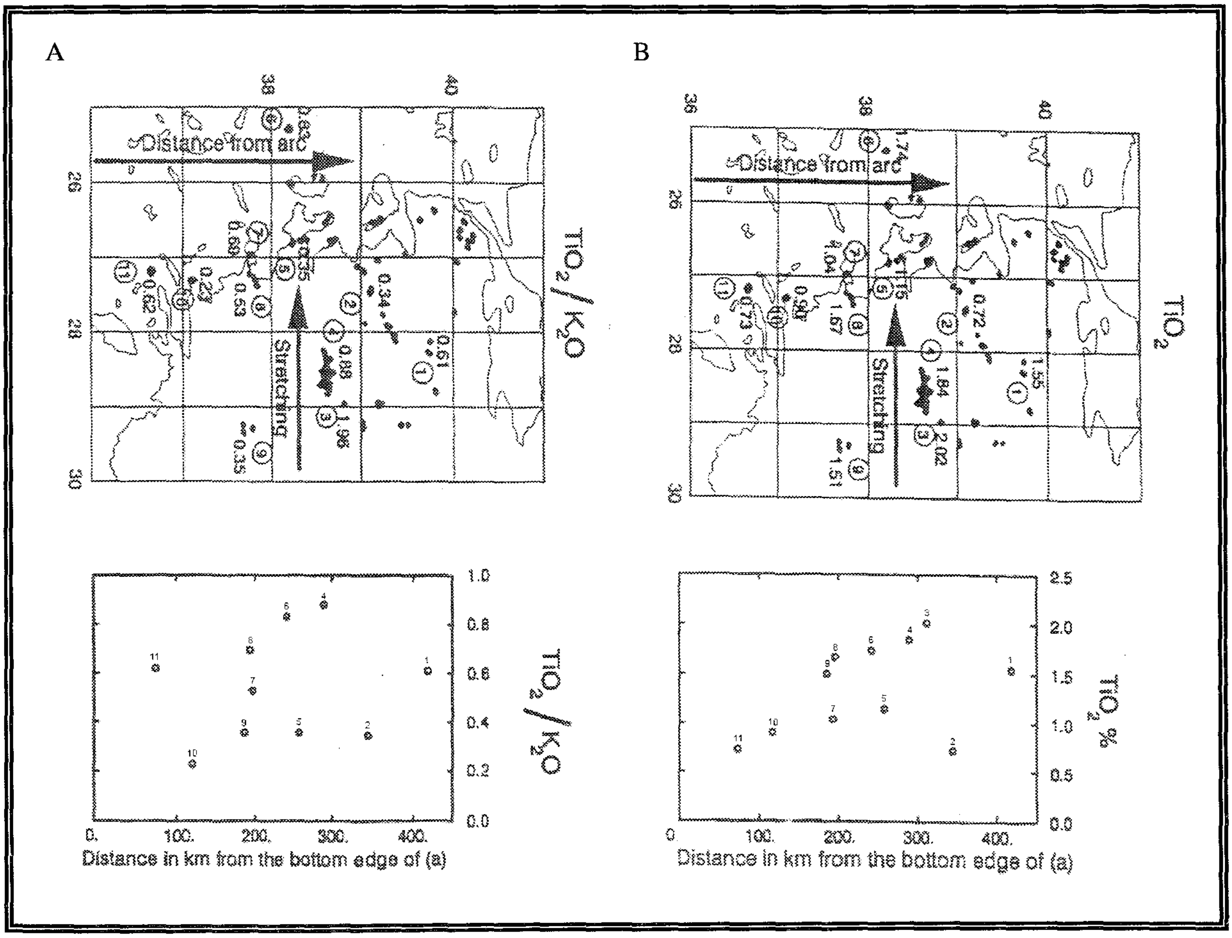


Figure. 2.3. (A) Relations of means of TiO_2/K_2O, (B) mean concentrations of TiO_2 (after *McKenzie and Yılmaz, 1991*)

percentage. *Fytikas et al. (1984)* argue that the volcanism has migrated in time towards a more southerly position from a change in the chemical character of the products erupted, which tend to become richer in K_2O towards the south.

According to the *Yılmaz (1990)*, the calc-alkaline lavas (Late Oligocene to Early Miocene) of western Turkey were erupted at a time when the compressional regime led to crustal thickening and younger east-west extension normal to the compression direction (Table. 2.2). They also suggested that alkaline lavas erupted at the time when N-S extension started. There is an alternative idea that the changes from calk alkaline (Late Oligocene to Early Miocene) to alkaline are interpreted as indicating an increasing asthenosphere contribution resulting from a thinned lithosphere (*Seyitoğlu & Scott, 1992a,b*). This incompatibility of generation of calk alkaline volcanics leads to incompatibility of extension starting time in the Aegean region.

Studies based on detailed geological mapping, establishment of the local volcanic stratigraphy leading to the delineation of volcanic episodes and volcanic/sedimentary rock relationships and the analysis of volcanism by establishing the petrological character of individual volcanic episodes are surprisingly few (see *Yılmaz, 1989*). *Yılmaz (1989)* pointed and that most of the young volcanic studies are geochemical and based on analyses of limited sample populations collected from a small region. For this reason scientists could not find an exact and unique solution and there are a lot of controversial alternative ideas and incompatibilities.

Active Volcanic Arc

In the South Aegean region, an important feature is the chain of volcanic centers stretching from near Corinth to the Bodrum peninsula, which is called SAVA (*Paraskevopoulos, 1956, Nicholls, 1971*) (Fig. 2.2). Active volcanoes in the zone include Santorini (36.5°N, 25.6°E) and Nisyros (36.7°N, 27.1°E) and fumaroles are found on Methana (37,4°N, 23,5°E), Susaki (37.9°N, 23,3°E), Milos (36.8°N, 24.5°E) and Kos (*Ninkovich & Heezen, 1965*). This well-known arc is considered to be the surface expression of still-active subduction of the African plate beneath the Aegean plate (*Caputo et al., 1970; Ninkovich & Hays, 1972; Fytikas et al., 1976*). The obtained K/Ar ages of these volcanics range from 3.5 My to present (*Bellon et al., 1979; Gülen, 1990*) (Table. 2.1). The products of the South Aegean Arc (SAA) form a typical calc-alkaline association, which displays a continuous evolution from basalts to rhyolites. But basaltic lavas are very rare (*Nicholls, 1971*). Granitoids show a systematic compositional variation in the E-W direction (*Altherr et al., 1988*), The eastern granitoids (Samos, Kos, and Bodrum) are monzonites; the central ones are granites (Tinos, Mykonos, Delos, Naxos, Keros, and Ikaria); and the western ones are granodiorites (Laurium, Serifiros) (*Gülen, 1990*)

Table. 2.2. Ages of volcanics and granitoids observed by various geoscientists in the Aegean region. AI- acid intrusion; A- alkali; KA- kalk alkali; G- granitoid; GI- granit intrusion; GE- granit emplacement; GD- gabroic dykes; B- basic; LGD- lekogranit; AV- active volcanism. Extensional phase indicaed by light gray, compression phase indicated by dark gray.

References and Areas beneath volcanic rocks observed	OLIGOCENE		NEOGENE				
			MIOCENE			PLIOCENE	
	Lower	Upper	Lover	Middle	Upper		
Attic-Cycladic Massif	**Attic-Cycladic Massif**						
Fytikas et al., 1984				[AI	AI		
Wijbrans and McDougall, 1988				GE			
Menderes Massif	**Menderes Massif**						
Seyitoğlu et al., 1992			LGD	LGD	BA	BA	BA
Dora et al., 1990					[B	]	
Yılmaz, 1989	[	KA G	KA G	]			
Seyitoğlu et al., 1992	[	KA G	KA G	]			
Benda et al., 1974			[GE				
Becker,platen et al., 1977			[GE				
Yılmaz, 1989		]					
Ertürk et al., 1990						A	
Gülen, 1990				A			
Yılmaz, 1989					B A	[HV	
Savaşçın et al., 1990; Ercan et al., 1985; Bellon et al., 1979				[A	[B A	BA	BA
Yılmaz, 1989; Savaşçın et al., 1990	GI	GI		[A GD			
Savaşçın et al., 1990			]				
Attic-Cycladic-Menderes Massif	**Attic-Cycladic-Menderes Massif**						
Gülen, 1990			GE[	**GE**	GE		
Borsi et al., 1972; Fytikas et al., 1984		KA	]				
Active Volcanic Arc	**Active Volcanic Arc**						
Fytikas, 1984						[AV	AV
Ercan, 1982							AV
Gülen, 1990							AV
Inner Volcanic Arc	**Inner Volcanic Arc**						
Innocenti et al., 1979; Gülen, 1990	[G	G]			KA	KA	
Kalkidiki Greece	**Kalkidiki Greece**						
Gülen, 1990	[G	G]					

(Fig. 2.4). These results could be interpreted as an extension of crustal contamination, which increases gradually from Nysiros (east) towards Aegina (west) along the active volcanic arc (Fig. 2.4) (see *Gülen, 1990*).

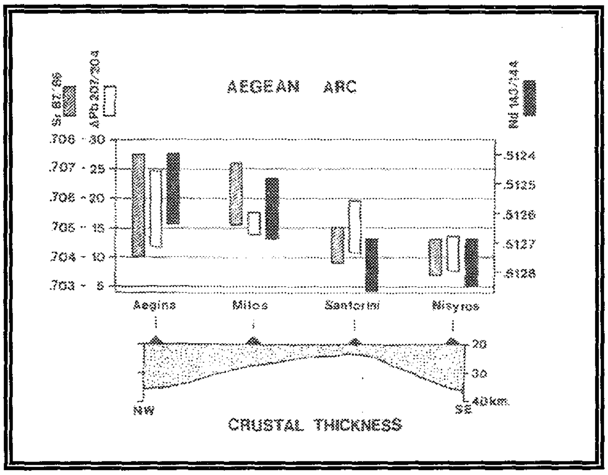


Figure. 2.4. Measured range in Sr 87/86, Nd 143/144, and ΔPb 207/204 for four Aegean active volcanic arc islands plotted against their positions along the arc; also shown is the crustal thickness variation along the arc. The horizontal distance is without scale. ΔPb 207/204 notation is after Hart (1984). The crustal thickness data are from Makris (1978) (after *Gülen, 1990*).

2.3. METAMORPHISM

2.3.1. Data

There are many more or less detailed studies considering metamorphism in the Aegean (e.g., *Hamilton & Strickland, 1840; Erentöz & Ternek, 1968; Phillipson, 1911; Egeran & Yener, 1944; Trikkalinos, 1947; Önay, 1949; Flügel & Metz, 1954; Kaaden & Metz, 1978; Dixon & Ridley, 1987; Schuiling, 1962; Brinkman, 1966; Ketin, 1966; Başarır, 1970, 1975; Dora, 1972, 1975, 1977, 1981; Altınlı, 1973; Dürr, 1975; Altherr et al., 1976; Ben et al., 1976; Dürr et al., 1978; Şengör & Yılmaz, 1981; Öztürk & Koçyiğit, 1983; Lister et al., 1984; Ridley, 1984a,b; Maluski et al., 1987; Avigad & Garfunkel, 1989, 1991; Buick & Holland, 1989, 1991; Candan & Kun, 1989; Kaya et al., 1989; Lister & Davis, 1989; Okay, 1989; İztan & Yazman, 1990; Dora et al., 1990; Okrusch & Bröcker, 1990; Kessel, 1990; Westaway, 1990; Buick, 1991; Seyitoğlu & Scott, 1991, 1992a,b; Avigad et al., 1992; Candan et al., 1992; Paton, 1992; Erdoğan, 1992, 1993; Erdoğan & Güngör, 1992; Avigad, 1993; Dinter & Royden, 1993; Westaway & Kusznir, 1993; Gautier et al., 1993; Baker & Matthews, 1994; Gautier & Brun, 1994; Westaway, 1994; Patzak et al., 1994; Bozkurt et al., 1995; Dora et al., 1995; Genç, 1995; Feenstra, 1996; Katzır et al., 1996; Bozkurt, 1996; Bozkurt et al., 1993, 1995; Bozkurt & Park, 1993, 1994, 1997; Jacobshagen, 1994; Emre & Sözbilir, 1995* and references therein). In Section 2.3.2 brief information is given about the most important of the above have given references.

2.3.2. General Interpretation

By studying metamorphic rocks, many scientists try to analyse the phases of metamorphism to separate the tectonic regimes in the Aegean region especially on the Attic/Cyclades/Menderes Metamorphic Belt (ACMMB).

The Attic/Cycladic Metamorphic Complex in the main land of central western Greece and the Cyclades area and Menderes Metamorphic Complex in southwestern Turkey which stretches as the most important metamorphic belt in the Aegean region (see also *Ridley, 1984b; Dora et al., 1995; Bozkurt & Park, 1997*) (ACMMB). It is parallel to the main geological units, tectonic features and structures, especially the present day subduction zone (Fig. 2.5). It forms large metamorphic culminations within the Alpine Himalayan belt (*Şengör & Yılmaz, 1981; Okay et al., 1991; Bozkurt & Park, 1997*). These culminations take form under overprinted

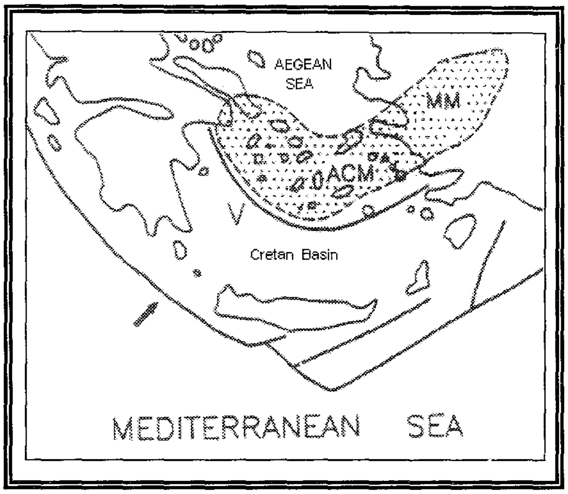


Figure. 2.5. The Attic-Cycladic-Menderes metamorphic belt, which conforms to the arcuate volcanic arc, is indicated by shaded patterns (from *Bozkurt et al., 1995*). ACM-Attic Cycladic Massif; V-volcanic arc; MM-Menderes Massif.

metamorphism events. The belt in western Turkey is overlayed to the North by the İzmir – Ankara ophiolithic (Neo Tethyan) suture zone (see, *Şengör & Yılmaz, 1981*). To the south in Turkey it is overridden by the Lycian Napless of Taurides (*Graciansky, 1972;* see also *Bozkurt et al, 1995*). The massive pre-Alpine metamorphic and magmatic rocks, from the Early Cenozoic, have been basically subjected to two kind of metamorphic events: first a high-pressure-low-temperature (HP-LT) phase, and then a Barrovian type overprint (*Lister et al., 1984; Dora et al., 1995*). The Barrovian type metamorphism is represented by increasing temperature at the existing pressure (HP/↑T) and by HT/LP in later stages. Metamorphosed groups of rocks that are of the same or various types, under specific physico-chemical conditions are named as metamorphic facies. The ACMMB pre-Alpine rocks (various types) were all metamorphosed firstly under the same physico-chemical conditions and are represented by blue schist facies (*Ridley, 1984b*). This facies points to HP/LT condition i.e. compressional tectonic regime related to crustal thickening in the Aegean region (e.g., *Dürr et al., 1978; Şengör & Yılmaz, 1981; Lister et al., 1984; Ridley, 1984b; Wijbrans & McDougall, 1988; Avigad & Garfunkel, 1989; Kessel, 1990; Buick, 1991; Feenstra, 1996*). After these conditions same facies are also seen during the neotectonic regime only along the consumption boundary, south- southwestern edge of the region, although the Aegean region including ACMMB is mainly under an extensional regime that is characterised by HT/LP conditions (*Lister et al., 1984; Feenstra, 1996*). This overprinting phase, which started just after the HP/LT phase is regarded as the main Menderes metamorphism in western Turkey (e.g., *Şengör & Yılmaz, 1981; Dora et al., 1995*).

There are two groups of scientists that have controversial ideas. According to the first group, after the compressional phase that represented blueschist facies, high temperature medium-pressure and low-pressure conditions result from detachment faults that represent high-grade extension (e.g., *Lister et al, 1984; Ridley, 1984b; Seyitoğlu & Scott, 1991, 1992b, Seyitoğlu et al., 1992; Bozkurt & Park, 1993, 1994, 1996, 1997; Verge, 1993; Gautier & Brun, 1994; Jolivet et al., 1994; Dora et al., 1995*). This model is known as a metamorphic core complex of the Cordilleran type (*Lister et al., 1984*). According to the model, this event easily explains the origin of the Early Miocene granodiorite rocks from the granite family in the Menderes Massive. *Lister et al. (1984)* found many similarities between the Cyclades islands and the metamorphic core complexes of the North American Cordillera (*Lister et al., 1984*). In addition, *Bozkurt et al. (1995)* found many similarities between Cyclades and Menderes as metamorphic core complexes. According to the theory, low-angle normal "detachment" faults in the upper crustal layer evolve into major ductile shears at depth (Figs. 2.6., 2.7). Along the

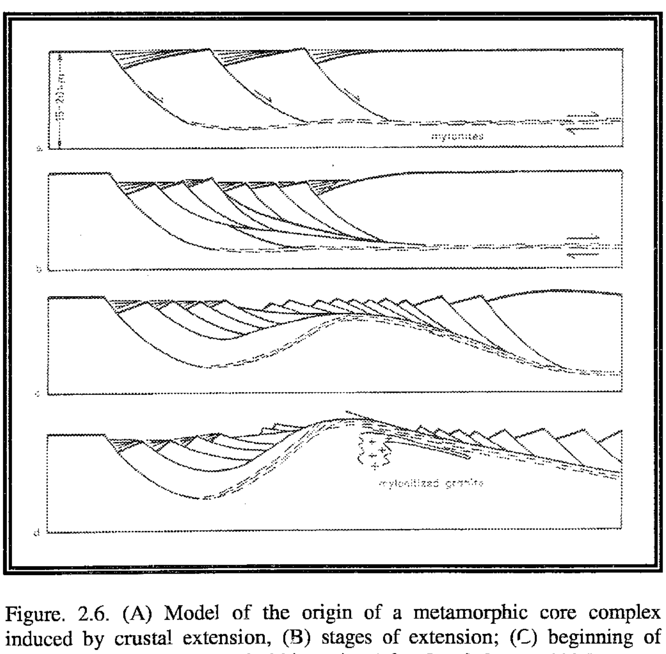


Figure. 2.6. (A) Model of the origin of a metamorphic core complex induced by crustal extension, (B) stages of extension; (C) beginning of crustal up doming; (D) granitoid intrusion (after *Jacobshagen, 1994*).

shear zone, middle-crustal rocks in the lower plate are mylonitized. Further stages of the shear zone include mylonised material dragged out from underneath the detachment faults (*Lister et al., 1984*) (Fig. 2.7.). Thinned crust ie. updoming of the mantle is associated with an elevation of thermal flux which might have caused partial melting that leads to granite emplacement (Fig. 2.6., 2.7). On the other hand, the other group (*Lüttig & Steffens, 1976; Şengör, 1982; Dumont et al., 1979b; Angelier et al., 1981; Şengör & Yılmaz, 1981; Yılmaz, 1997*) claims that extensional features just after blueschist facies are related to further stages of compressional regime. According to this idea, a compressional regime leads to extension that is normal to the compressional direction (e.g., *Candan et al., 1992*). This compression-related extension leads to HT/LP metamorphism because of the well-known granite emplacement in compression directed normal faults (Table 2.3) (e.g., *Becker-Platen, 1970; Benda et al., 1974; Wijbrans & McDougall, 1988; Benda & Meulenkamp, 1990*). According to the this idea the neotectonic (N-S extensional) phase in the Aegean region starts after E-W extension related granite emplacement caused contact metamorphism in the country rocks (*Dora et al., 1995*). Mainly because of these two different ideas different ages are proposed for the beginning of the extensional regime that is ongoing to day in the Aegean region.

Recent studies show that the Gediz and Büyük Menderes grabens are example of huge ductile-brittle deformation and detachment faulting that is located in the central part of the Menderes Massif (western Anatolia) (see *Emre & Sözbilir, 1995*). *Kaya (1982), Emre (1992), Sözbilir & Emre (1996), Emre (1996), Emre & Sözbilir (1995)* study these in detail. *Kaya (1982)* recorded that the Büyük Menderes graben is early Miocene in age while the Gediz graben is Late Miocene in age. *Emre & Sözbilir (1995)* suggest that the Büyük Menderes graben is older than the Gediz graben due to the asymmetric characteristics of the shear zone. *Sözbilir & Emre (1996)* define younger basins as supradetachment basins that are formed above a low-angle normal fault system (detachment fault).

Table. 2.3. Age and type of metamorphism in the Aegean region. GE- granit emplacement; GP- granit pluton; GI- granit intrusion; GS- green schist; BS- blue schist, HP- high-pressure metamorphism, HT- high-temperature metamorphism; HP/LT; high pressure/lowe temperature metamorphism; HT/LP; high-temperature/lowe pressure metamorphism. Extensional phase indicated by light gray, compressional phase by dark gray.

References and Areas Beneath Metamorpic Rocks	CRATEGEUS	CENOZOIC									
		TERTIARY									
		PALEOGENE						NEOGENE			
	Upper Cretaceus		EOCENE			OLIGOCENE		MIOCENE			PLIOCENE
		Paleocene	Lover	Middle	Upper	Lower	Upper	Lover	Middle	Upper	
Aegean Region	**Aegean Region**										
Gautier and Brun, 1994											
Andriessenet al., 1979						HT	HT	HT/MP	HT/MP		
Jacobshagen, 1994											
Andriessen et al., 1979; Altherr et al. 1988; Okay, 1980; Gülen, 1990						HT	HT	HT			
Altherr et al. (1982) (Kessel, 1990)								GI	GI	GI	
Jacobshagen, 1980; Blake et al., 1981; Altherr et al., 1982											
Menderes Massive	**Menderes Massive**										
Hetzel et al., 1995											
Dora et al, 1995			[BS	BS	BS]]	B]					
Dora et al, 1990											
Candan et al., 1992								GP			
Savaşçin, 1990								GP			
Şengör and Yilmaz, 1983								GP			
Bozkurt and Park, 1997							GE				
Şengör et al., 1984; Saur and Friedrichsen, 1986						GE	GE				
Erdoğan and Güngar, 1992											
Cyclades	**Cyclades**										
Feenstra, 1996; Adriessen et al., 1979; Van der Maar and Jansen, 1983; Feenstra, 1985; Dürr, 198				BS			GS	GS			
Wybrans and McDougal, 1986, 1988; Dickson et al., 1987; Schliestedt et al., 1987;				BS			GS	GS			
Okrusch and Bröcker, 1990				BS			GS	GS			
Andriessenet al., 1979				GS							
Baker and Matthews, 1994								HT	HT	HT	
Jolivet et al., 1994							[HT/LP	[HT/LP	[HT/LP	[HT/LP	[HT/LP
Ridley, 1984a, 1982;		BS	BS	BS	BS	BS	BS				
Andriessen et al., 1979;											
Blake et al., 1981											
Katagas, 1984		BS					HT/	HT/MP	HT/MP		
Jacobshagen, 1994											
Andriessenet al., 1979							HT	HT/MP	HT/MP	LT	
Katagas, 1984; Wendt et al. 1977; Altherr et al. 1977;				BS	BS						
Andriessen, 1978; Andriessen el al., 1978											
Bonneau el al., 1980											
Lister et al., 1984											
Crete	**Crete**										
Seidel et al., 1982											
Strymon Valley	**Strymon Valley**										
Dinter and Royden, 1993											

Structural studies on metamorphic rocks are a good tool for direct solutions. But to increase the resolution of the flow directions in the recent tectonic phases there should be more structural study on such detachment zones. The starting age of the neotectonic regime in the Aegean region still subject to debate.

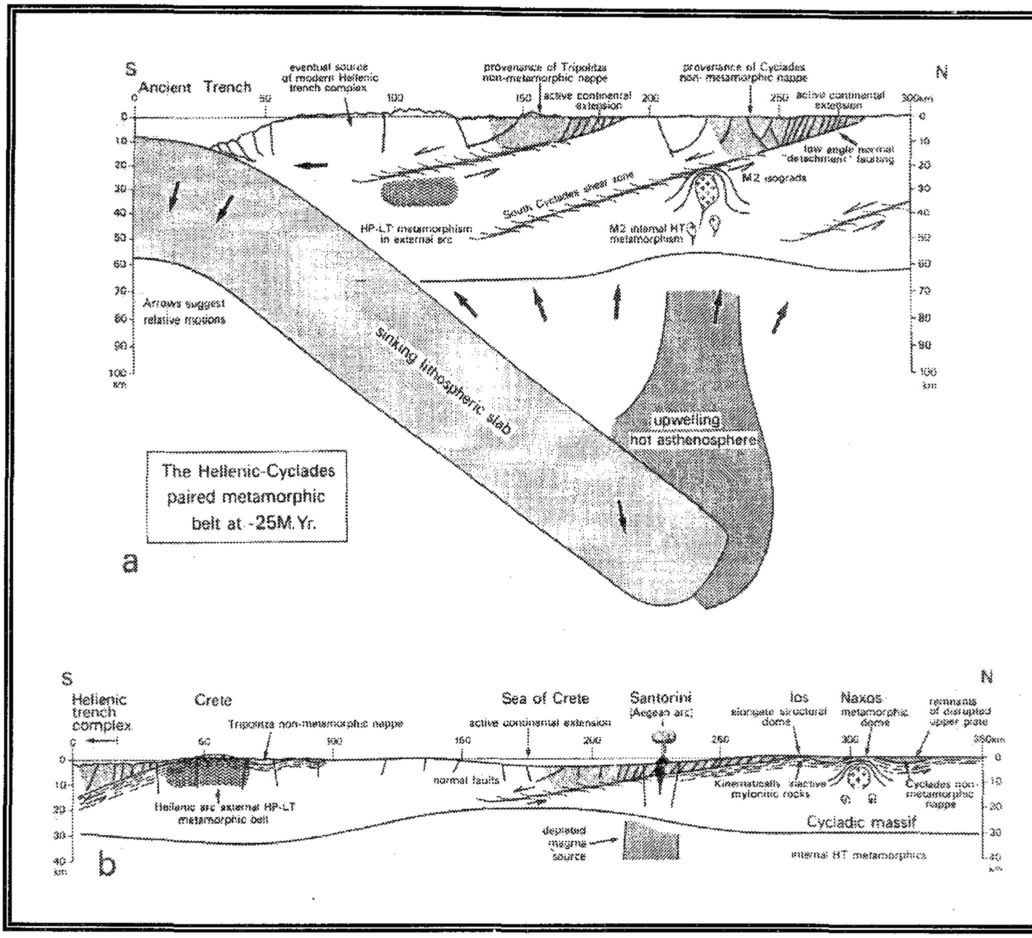


Figure. 2.7. Classic model of paired metamorphic belt, showing mechanism for subsequent exposure of metamorphic rocks-crustal shear zones that evolve upward into low-angle normal faults, a: Position of HP/LT metamorphic rocks that eventually will be exposed on Crete (cross-hatched), and position of HT metamorphic dome that eventually will be exposed on Naxsos. These metamorphic rocks will be dragged out from under upper crustal nonmetamorphic rocks (stippled), g: Crustal extension as a result of operation of these major movement zones. Crustal thickness is based on deep refraction profiles (*Makris, 1982*) (after *Lister et al., 1984*).

Dominantly N-NE lineation patterns exist in metamorphic rocks on many islands that indicate the existence of a major shallow-dipping shear zone in the southern Cyclades e.g. Naxos and Ios (*Jansen, 1973; Lister et al, 1984; Faure & Bonneau, 1988; Gautier et al., 1990; Buick, 1991; Jansen, 1973; Robert, 1982; Gautier & Brun, 1994; Jolivet et al., 1994*) (see e.g., Fig. 2.8.). Detail structural studies of *Jolivet et al. (1994)* on the metamorphic rocks of the Cyclades region show a kinematic situation in the region. They observed that clockwise rotation is added to N-S extension to accommodate the NE-SW direction of ductile flow (Fig. 2.9). The early strain field shown in the metamorphic rocks, which represents deeper parts of

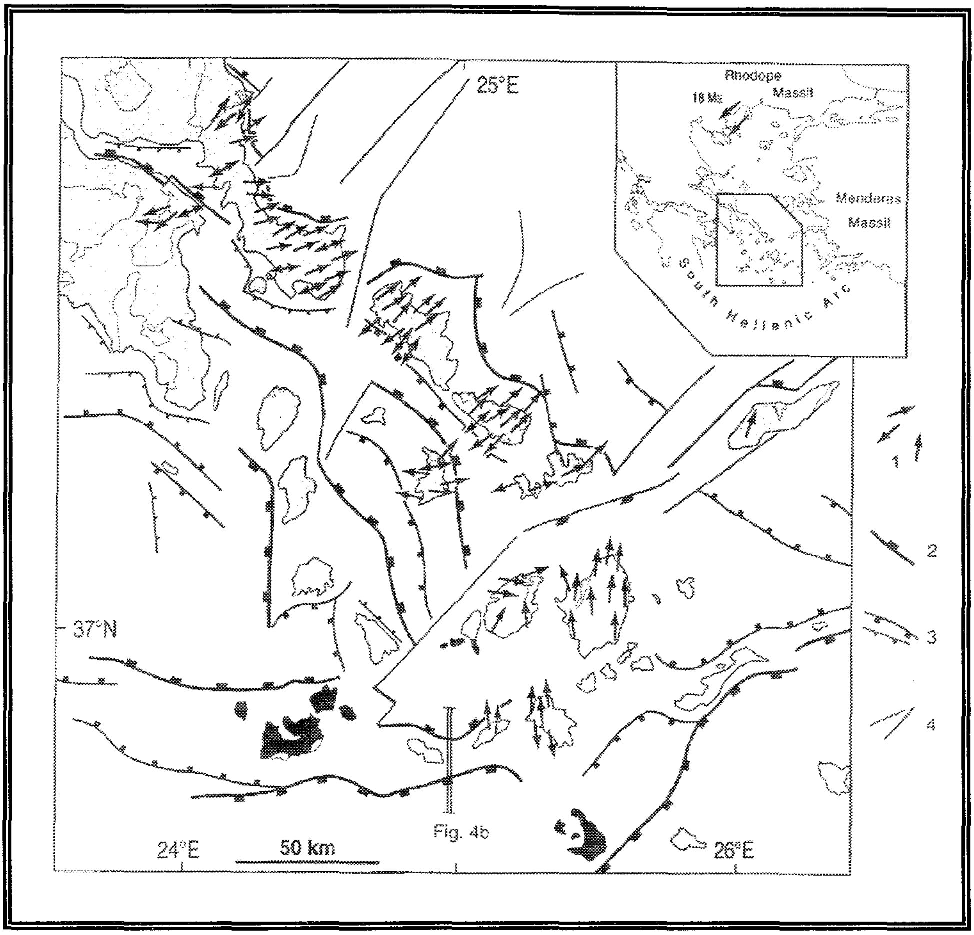


Figure. 2.8. Summary of crustal features in the Cyclades area (after *Gautier and Brun, 1994*). 1- ductile shear-sense data within Lower Unit metamorphic rocks and younger Miocene intrusions; each arrow represents the mean of measurements on several sites; 2, 3, 4- faults which have been active at least during Pliocene-Quaternary time; 2- major normal faults; 3- second-order, commonly antithetic normal faults; 4- strike-slip and/or transform faults (see for more detail *Gautier and Brun, 1994*).

the crust, has been deformed in recent times in such a way that, in a first approach, it fits a same flow field (see *Jolivet et al., 1994*). *Bozkurt et al. (1995)* also investigated lineation patters that dip consistently towards the southwest away from centre of the core complex of the southern Menderes Massive. On the other hand, *Jolivet et al. (1994)* investigated the lineation pattens that dip consistently towards the north away from the core complexes to the North around the Cyclades islands. The observations from these two sites, from south and north respectively, are consistent with the core complex theory (Fig. 2.7). *Jacobshagen (1994)* on the other hand pointed out that several of these shear zones could have originated as overthrusts during nappe stacking.

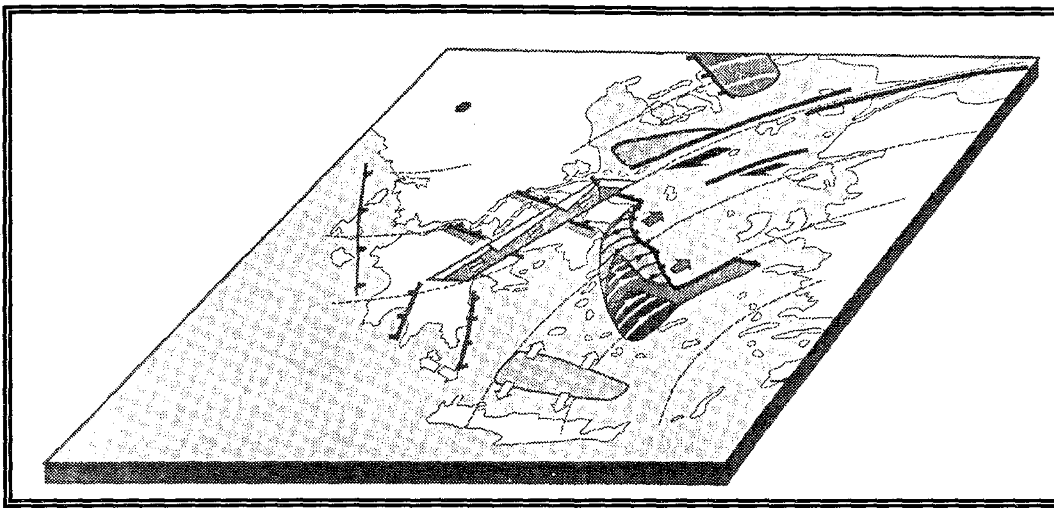


Figure. 2.9. Synthetic diagram of the brittle-ductile relations in the Aegean. Lines with arrows represent shear trajectories in the exhumed metamorphic domains. A gradation of darker shading shows the gradient of high temperature metamorphism from green schist to migmatites from Evvoia to Naxsos which correspond to differential exhumation of deeper and deeper portions of the crust. The window opened in the region of tilted blocks the sense of tilt and the sense of shear at depth (after *Jolivet et al., 1994*).

2.4. PALEOMAGNETIC MEASUREMENTS

2.4.1. Data

Kondopoulou (1985) published some paleomagnetic results in the central Aegean region. *Kissel & Laj (1988)* have postulated important recent rotations for the whole Aegean area using all the available paleomagnetic data. The other data in the region have been obtained from Middle Miocene to Pliocene sedimentary and volcanic formations by *Kissel et al. (1989)*. They reported data from 75 sites in the central part of the Aegean Sea and its eastern and western boundaries (Volos region, Evvoia, Skyros, Lesbos and the İzmir area) and from southern Anatolia (Antalya region). These data will be analysed together with those from northwestern Greece and with the results from the northeastern Aegean and western Turkey reported by some other authors. *Taymaz et al. (1991)* critically reviewed paleomagnetic results in the Aegean region. *Beck Jr and Schermer (1994)* overviewed paleomagnetic studies in the Aegean region and published a measured paleomagnetic pattern in Europe and Mrica. They used inclination data in the region for the first time. Information about other local studies is given below.

The South Aegean Arc:

Previous studies along the South Aegean Island Arc from the Late-Miocene marine clay sequences are presented by *Valet & Laj (1981)*, *Valente et al. (1982)* and *Laj et al. (1982)*. Along the external South Aegean Island Arc as well as in the Oligocene to Middle Miocene Ionian flysch of northwestern Greece, measurements have been made on the Mio-Pliocene sedimentary series by *Kissel et al. (1984, 1985)*. Paleomagnetic results from magmatic rock (2.5-1 My) samples from Melos island in the South Aegean back-arc area have been published by *Kondopoulou & Pavlides (1990)*. Two Late Miocene sections at Potamida and Kotsiana in western Crete have been resample by *Laj et al. (1996)*. At the eastern edge of the arc paleomagnetic study was carried out in Paleocene, Eocene and Lower-Miocene sedimentary formations of the western Taurides belt east of the Isparta (*Kissel et al., 1993*). *Horner & Freeman (1983)* studied the western edge of the South Aegean Arc (SAA) in Greece. They used 2500 paleomagnetic samples from 130 sites of the Early Jurassic to Late

Eocene carbonate sequence from the Ionian zone. For the same area *Kissel et al. (1983)* presented data from Cenozoic Ionian flysch sections. *Kissel et al. (1985)* obtained lower Miocene orientations of Ionian structures. *Mauritsch et al. (1991)* have reported Neogene rotations in the Ionian zone of southern Albania. The Neogene rotation was tentatively interpreted by *Muttoni et al. (1995)* in Albania. A paleomagnetic study of 750 samples obtained from 55 late Eocene to middle Pliocene sedimentary sites in central Albania and north of Greece was made by *Speranza et al. (1995)*. Paleomagnetic studies of older Mesozoic rocks have focused on western and northern Greece (*Marton et al., 1990, Lauer & Kondopoulou, 1991; Edel et al., 1992*). In northwestern Greece, paleodeclination data have been obtained from Eocene and Oligocene formations by *Speranza et al. (1992)*. Magnetobiostratigraphic data are presented from Early/Middle Triassic rocks in northern Albania by *Muttoni et al. (1996)*.

Eastern Part of the Aegean Region:

Previously *Gregor & Zijderveld (1964)*, *Sanver (1968)*, *Van der Voo (1968)*, *Orbay (1979)*, and *Orbay & Bayburdi (1979)* presented paleomagnetic data of Triassic and older rocks from the eastern part of the Aegean region. *Lauer (1984)*, using all the available paleomagnetic results gave tentative reconstructions for the past positions of various parts of Turkey in Triassic times. He used paleomagnetic results Turkey and Cyprus, carried out between 1973 and 1978, with the data of *Gregor & Zijderveld (1964)*, *Sanver (1968)*, *Van der Voo (1968)*, *Orbay (1979)*, and *Orbay & Bayburdi (1979)*. *Pe-Piper & Piper (1977)* made determinations on volcanic rocks over sixty paleomagnetic polarity from Lesbos. From the island of Limnos, *Kondopoulou & Lauer (1984)* report variation of declination in volcanic rocks of 17 to 22 My. In northwestern Anatolia from samples (33-28 My old) of Volcanic rocks they reported some declination values. *Kissel et al. (1987)* reported other paleomagnetic data from the northeastern Aegean region. Paleomagnetic directions obtained from Miocene rocks of western Turkey (*Kissel & Laj, 1988*) are republished and interpreted by *Zanchi et al. (1990)*. In the area paleomagnetic directions of volcanic rocks have been investigated by *Orbay et al, (1993, 1995, 1996)*. *Orbay et al. (1993, 1995)* reveals the paleomagnetic results related to the extension of the Gulfs of Saroz and Edremit. Around Ganos in the north of the area, block rotations were observed from paleomagnetic measurements by *Tapırdamaz and Yaltırak (1995)*. Recently, some paleomagnetic samples collected in Çanakkale-Foça and Bergama from 47 sites (eight of them Oligocene, 35 of them early Miocene, 4 of them Late Miocene) by *İşseven (1995)* (*see also İşseven et al., 1995b*). More recently, *Orbay et al. (1998)* obtained paleomagnetic data from the Pliocene or Late Miocene samples from western Turkey (between Balıkesir-Demirci line and Fetbiye-Burdur) on a large scale. In total, they have obtained 243 oriented drilled core samples from 28 sites in and around Balıkesir, Akhisar, Kula, Simav, Gediz and Tavşanlı. These data were evaluated together with the earlier data of western Turkey, which include 23 sites.

Western Part of the Aegean Region:

Simeakis et al. (1989) obtain Late Miocene-present-day declinations from the North Aegean Trough (NAT). Few data exist from Mesozoic sequences of eastern central Greece (e.g., *Turnell, 1988; Morris, 1995*). *Morris & Anderson (1996)* study the Middle-Late Miocene intrusive rocks of the Cycladic Massif. This contribution provides the first paleomagnetic results from this area.

All the available paleomagnetic measurement results in the Aegean and surrounding area are shown in Table. 2.4.

2.4.2. General Interpretation

Important evidence that block rotations and relative N-S displacements have occurred in the Aegean region comes from paleomagnetic declination and inclination, respectively. Paleomagnetic data of Neogene sediments and volcanics have been used to reconstruct the paleogeographies (e.g., *Kissel & Laj, 1988;* see also *Laj et al., 1982; Kissel et al., 1989*). Paleomagnetic data also have been used as direct evidence in support of various seismological models for the recent deformation.

By considering paleomagnetic data *Kissel & Laj (1988)* reveal the present curvature of the South Aegean Trench by opposite post-early Miocene rotations at the two terminations of the consuming boundary (see for detail *Speranza et al., 1995; Mauritsch et al., 1995, 1991;*

Table. 2.4. Paleomagnetic results in the Aegean region. Clockwise rotation indicated by light gray; counterclockwise rotations indicated by gray; no-rotation indicated by dark gray.

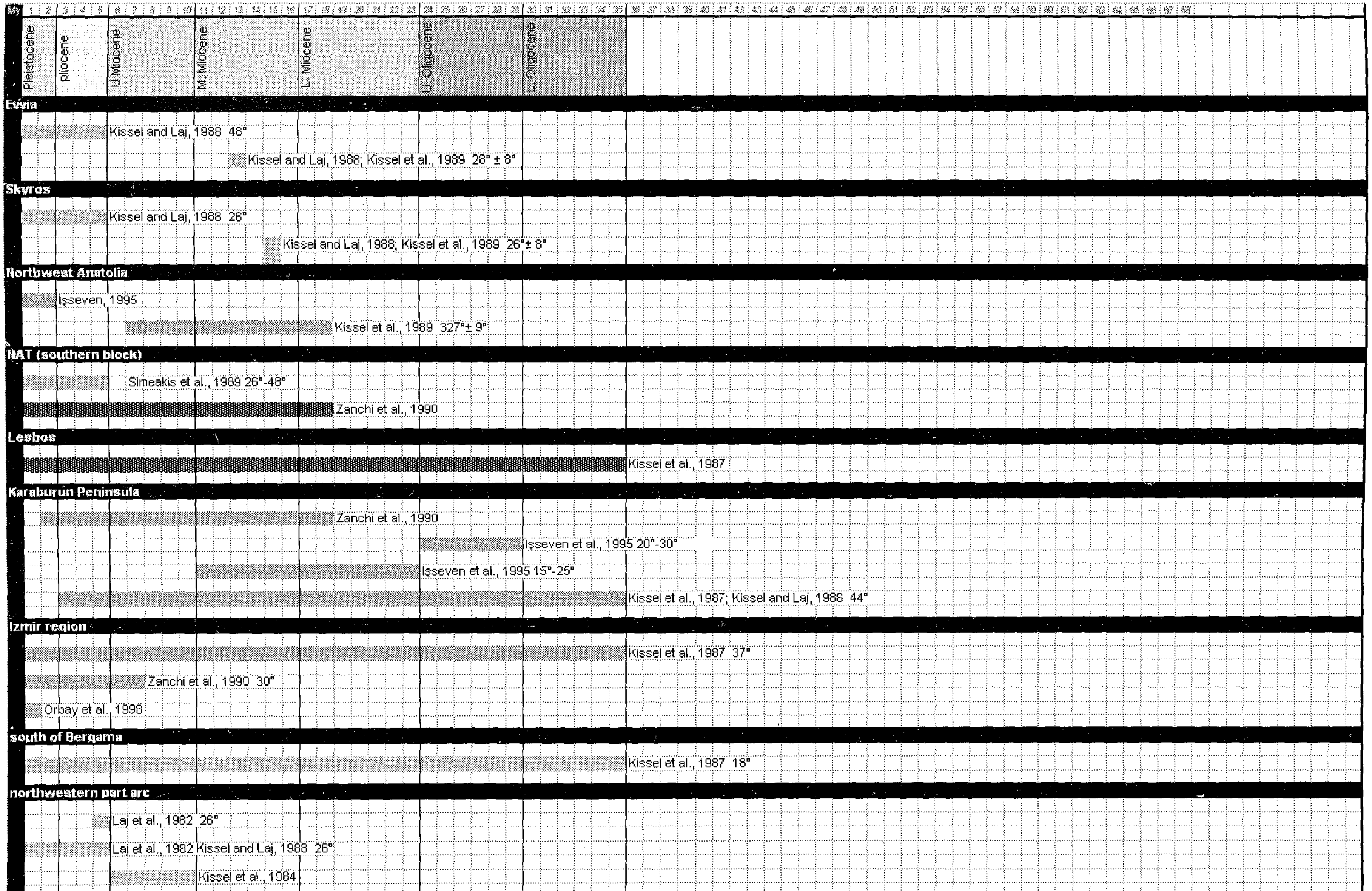
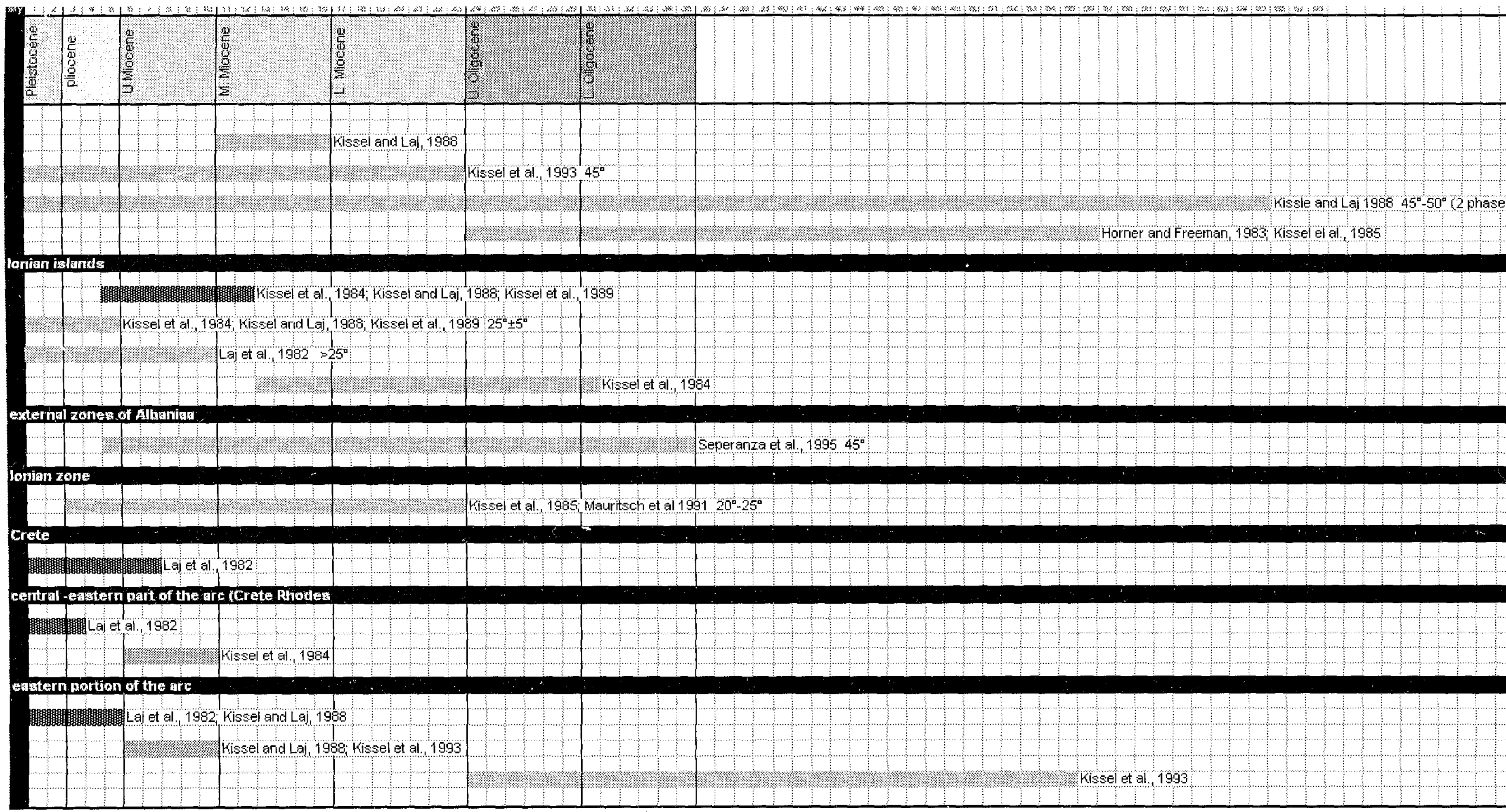


Table. 2.4. (continue)



Muttoni et al., 1996). This interesting proposal suggests, by considering paleomagnetic data that the Lower-Miocene arc was almost rectilinear with an E-W trend and that its curvature has been acquired tectonically in two major phases (see also *Kissel et al., 1993; Speranza et al., 1995*). During the Middle Miocene (13-30 My) a first phase of deformation is characterised by rotations (20-25°) occurring at the two terminations of the arc; clockwise in the west, counterclockwise in the east (Fig. 2.10) (*Kissel et al., 1984*). There was no rotation

Figure. 2. 10. Schematic map showing the geographical extension and the timing of the two rotational phases documented by the paleomagnetic studies (after *Kissel and Laj, 1988*). Dotted symbols- Middle Miocene rotations: shaded symbols- Pilio-Quaternary rotation.

in the Ionian islands during the period 12-5 My (*Kissel et al., 1985*). A second phase of rotation (25°) with clockwise direction occurring in the last 5 My about a pole situated in the southern Adriatic Sea has affected only the northwestern part (Fig. 2.10) (*Laj et al., 1982; Kissel et al., 1984; Kissel & Laj, 1988* and *Kissel et al., 1989*). This rotation started about 5 My ago and has continued since then at an average rate of about 5°/My (*Laj et al., 1982;* see also *Kissel et al., 1993*). A total of about 45° of clockwise rotation has been undergone by the western termination of the arc since Early Miocene times on the internal part of the arc. To the E- NE of Greece results also indicate clockwise rotation similar to the Ionian Zone. Evvoia and Skyros have rotated clockwise by about 48° and 26°, respectively. But the timing of the rotation is uncertain (*Kissel & Laj, 1988; Kissel et al., 1989*). *Kissel & Laj (1988) & Kissel et al. (1989)* suggested that it probably occurred in the last 5 My, but could have occurred over 15 My. Recently *Morris (1995)* deduced a lower clockwise rotation angle of 15° from young lacustrine sediments from the same area of Evvoia. This suggests that at least

15° of the 48° rotation of Evvoia occurred in the last phase. Clockwise rotations of 64°, which are higher than the total of 48°, are found in Mykonos (*Morris & Anderson, 1996*) from Middle and Late Miocene rocks. Results from northern Greece with other observations from the eastern Aegean region suggest that the whole mass beneath Greece has rotated clockwise towards the South Aegean Trench (Fig. 2.11).

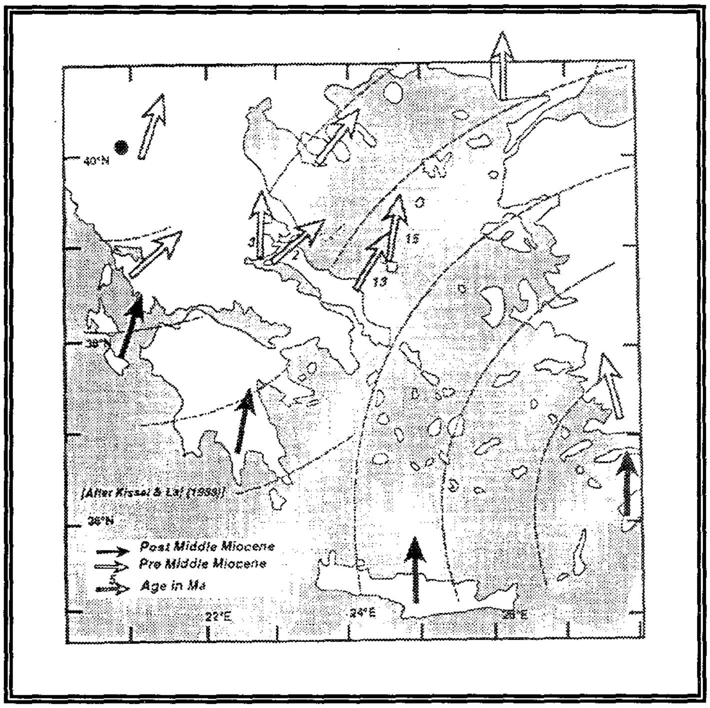


Figure. 2.11. Plot of paleomagnetic vectors in the Aegean region (after *Jolivet et al., 1994*; see *Kissel and Laj, 1988*). Small numbers next to the gray arrows in the center of the figure are the ages of sites in My.

This region could have rotated in the same sense, with a similar magnitude and simultaneously with the arc. During these curving phases, the back of the arc region extended and major grabens such as the Corinth graben were constructed (*Kissel et al., 1993*). On the other hand paleomagnetic results (*Laj et al., 1982; Kissel et al., 1984; Kissel & Laj, 1988*) show that the central and eastern parts of the arc (Crete and Rhodes) have not undergone significant rotation since the Tortonian-Messinian stage (second phase) (see also *Kissel et al., 1993*).

Results of *Kissel et al. (1993)* suggest that the Isparta angle which represents the eastern edge of the arc would result from opposite rotations of the two branches of Isparta angle: first a late Eocene-Oligocene clockwise rotation of the Akseki-Beyşehir Taurides (trending NW-SE) initiating the Tauride arc, and then counterclockwise rotation of the Lycian Taurides (trending NE-SW), probably during Middle Miocene times (*Kissel et al., 1993*).

To the northeast in western Turkey the situation is very complicated (see *Taymaz et al., 1991*). In the İzmir region, the Karaburun peninsula and in the island of Lesbos significant counterclockwise, clockwise or null rotations of several blocks have taken place during the

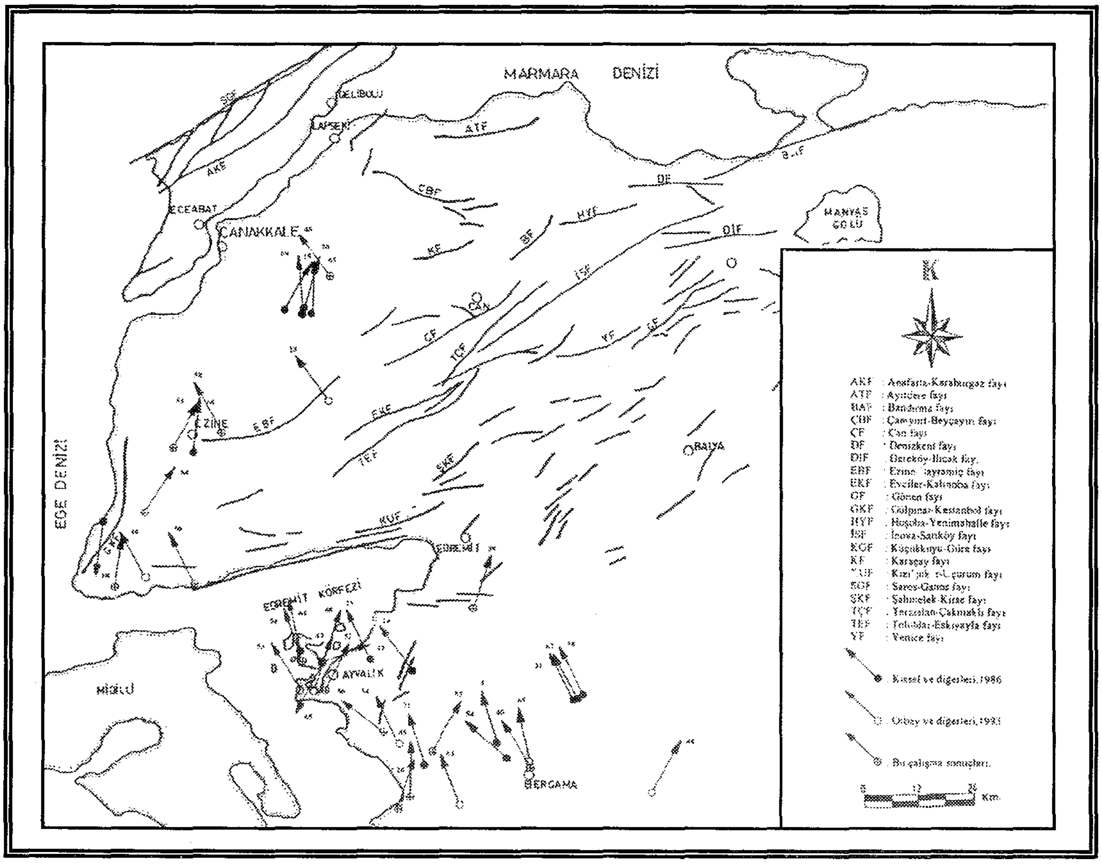


Figure. 2.12. The available paleomagnetic results in western Turkish mainland (after *İşseven, 1995*).

neotectonic extensional regime (Figs. 2.12, 2.13, 2.14) (*Kissel & Laj, 1988; Orbay et al., 1993; İşseven, 1995; Zanchi et al., 1990*). Published paleomagnetic results from western Anatolia are too scarce for the considered epoch, so that they do not allow any definitive conclusion (see *Kissel et al., 1989*). *Taymaz et al. (1991)* suggested the possibility that the counterclockwise rotations are related to the rotation of the big strike faults whereas the clockwise rotations are seen in some smaller blocks caught up in the right-lateral shear between them (see also *Mercier et al., 1991; Nicholson et al., 1986*). The lack of data is also apparent in the central Aegean (*Kissel et al., 1989*). *Kissel et al. (1987)* reported that the paleomagnetic data between the North Anatolian Fault Zone (NAF) and the Alaşehir fault demonstrate significant clockwise and counterclockwise rotations of several blocks, which took place during the neotectonic extensional regime (*Kissel et al., 1987*). They conclude that the brittle upper part of the lithosphere does not everywhere follow the motions of its lower ductile part (see also *Zanchi et al., 1990*). *Zanchi et al. (1990)* found out that in İzmir area extended from Urla in the south to Edremit in the north and from the coast to Akhisar has been affected by counterclockwise rotation of 30°±13 since 7 My. According to results of *Kissel et al. (1987)*, this rotation taking placed during Tertiary time. *İşseven (1995)* who took together all the available results with his results (*Orbay et al., 1993*) in Biga Peninsula concluded that Biga Peninsula has been rotated in counterclockwise direction since Neogene. *İşseven (1995)* suggested that, during counterclockwise rotation of NW Turkey (*Orbay et al., 1993; İşseven, 1995*), south of Edremit Bay rotated faster than the northern part, as a result of this Edremit Bay was opened (*Zanchi et al., 1990; Kissel et al., 1987; Kissel & Laj, 1988*). But in Lesbos just south of the eastern Edremit trough there is no rotation in Tertiary time (*Zanchi et al., 1990; Kissel et al., 1987*). Paleomagnetic data in this eastern region is ambiguous, but consistent with very small or no rotations in the northern part and possibly counterclockwise rotations, relative to Europe, in the south (*Taymaz et al., 1991*). Recent study of *Orbay et al. (1998)* in western Turkey show that western and eastern parts of Balıkesir-Demirci line has shown counterclockwise and clockwise rotations, respectively (Fig. 2.15). All remnant magnetisation directions of Pliocene aged rocks obtained from around Gördes show 44° counterclockwise rotations (*Orbay et al., 1998*). 200.000, 300.000 years aged rocks obtained from Kula Region have also show counterclockwise rotations. This indicates that region has been rotating counterclockwise direction today as detected by GPS results (*Barka & Reileneger, 1997*). Sites of Late Oligocene age in northwestern part of the Turkey also show a counterclockwise rotation 20° to 30° (*İşseven et al., 1995*). This type of rotation (25°-15°) continued in Early and Middle Miocene (*İşseven et al., 1995*).

On the other hand, results of paleomagnetic inclination data that studied by *Beck and Schermer (1994)* show that, the Aegean moved northward with respect to both continents. So

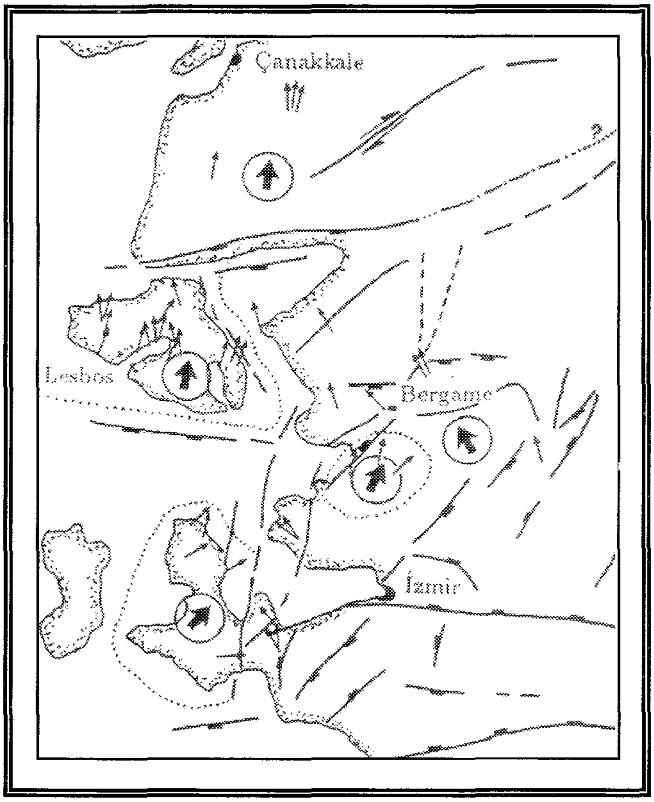


Figure. 2.13. A summary of the mean paleomagnetic declinations in the western part of Turkey (after *Jackson et al., 1992*). Arrows show declinations at sample sites; arrows in circles are the mean directions within roughly coherent areas outlined crudely with dotted lines. The coastlines are stippled; other lines are faults.

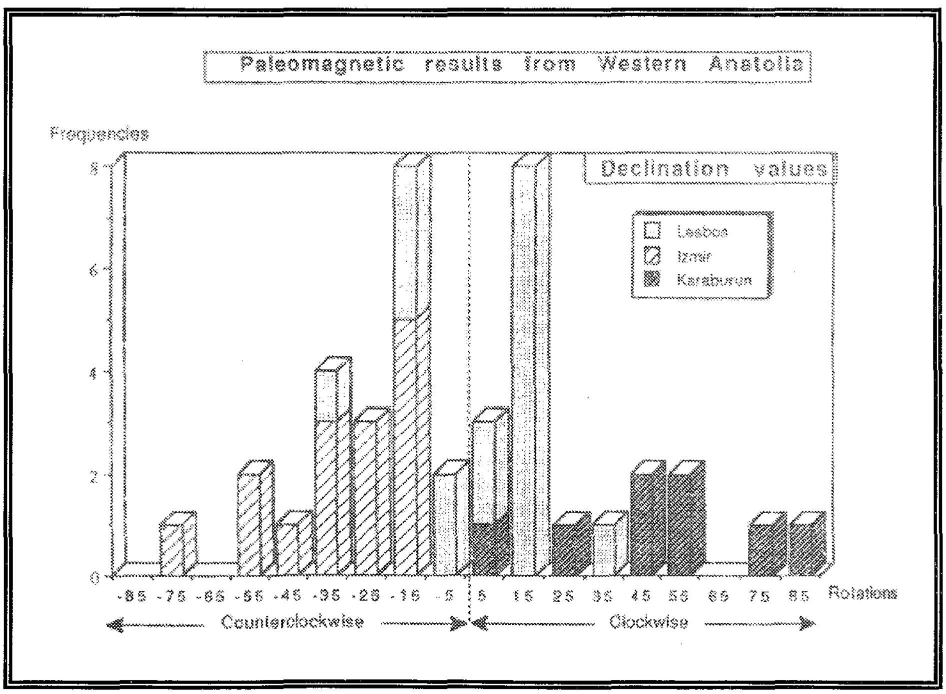


Figure. 2.14. Paleomagnetic results obtained from 38 sites (after *Zanchi et al., 1990*).

they suggested that the entire study area has undergone a northward drift of over 1000 km in the Middle Miocene.

The interpretations of these measurements that are taken place in the Aegean region are not straightforward because of the limited data. Around Aegean coasts and islands there are many paleomagnetic studies that not enough to determine the rotations exactly. Paleomagnetic data especially from the northern Aegean is more sparse and inconclusive yet.

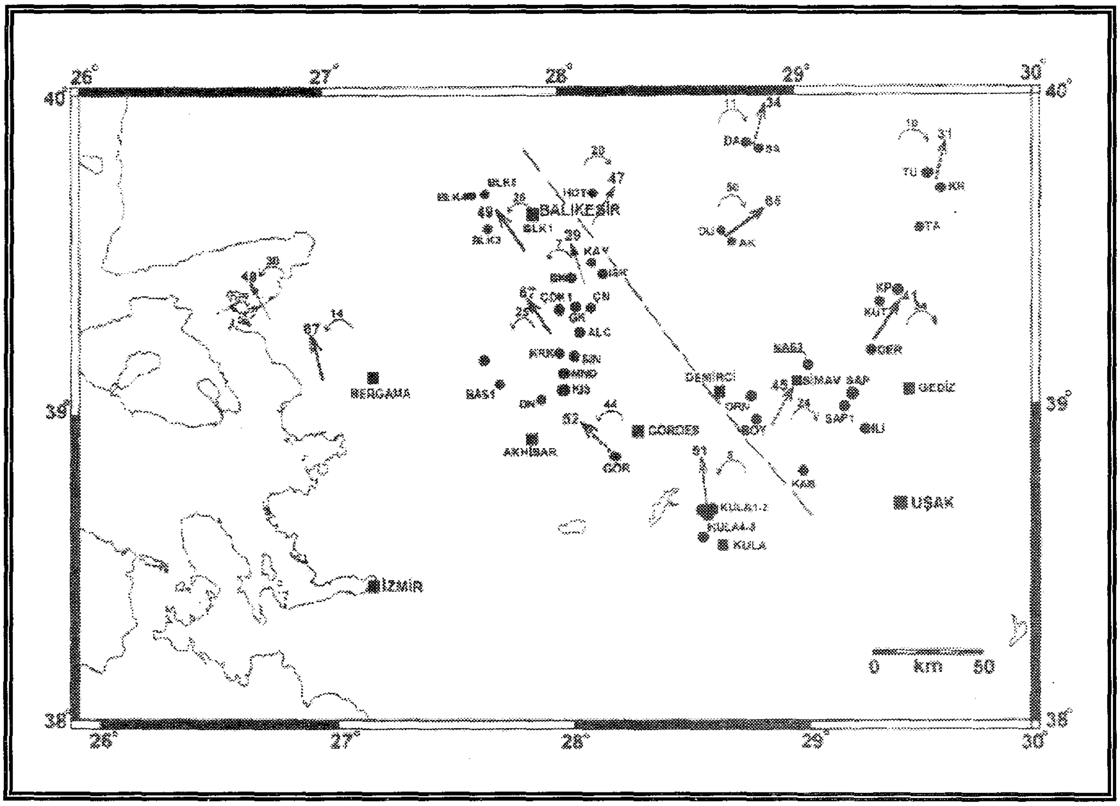


Figure. 2.15. Sampling sites and mean remanent magnetization directions of the same aged samples (after *Orbay et al., 1998*).

2.5. GEOMORPHOLOGICAL STUDIES

2.5.1. Data

Bathymetry

Previous bathymetry data for the Sea of Marmara exist in the Musee Oceanographique de Monaco (1960). For the eastern Mediterranean, previous data sheets were published by U.S. and U.K. Hydrographic offices (*Emery et al., 1966*). During the summer of 1968, the U.S. Naval Oceanographic Office conducted a detailed bathymetric survey of the Aegean Sea. This study, combined with data from other surveys (*Watson & Johnson, 1969; Goncharov & Mikhailov, 1963*), has resulted in a new bathymetric chart by the U.S. Hydrographic Office (*Maley & Johnson, 1971*). *Maley & Johnson (1971)* used high sensitivity "echo sounding" profiles that reflected the asymmetry of the topography better in the Aegean. *Allan & Morelli (1971)* published CONRAD radar data (1963) for the central Aegean Sea. *Allan & Morelli (1971)* and *Heezen et al. (1970)* published a small-scale panoramic bathymetry map. The SACLANT ASW Research Center in La Spezia and the Observatory Geofisico Sperimentale (1961-1965) (*Allan & Morelli, 1971*) surveyed bathymetry (1:750,000) of the Aegean and western part of the eastern Mediterranean. The bathymetric chart of the area south of Crete is based on soundings collected by the British Hydrographic Office up to until 1972, and those collected during 1972 and 1974 on R.R.S. Shackleton (*Jongsma, 1975*, *1977*). Another bathymetric map of the Mediterranean (No. 310) published by the U. S. Defence Mapping Agency Hydrographic Centre, Washington, has been described by *Carter et al. (1972)*. After that *Morelli et al. (1975)* surveyed the Aegean in some detail and a better bathymetric map has been published. In 1977 - 1978, within the framework of the HEAT (The South Aegean Arc and Trench) program by the R.V. "Jean Charcot", a multi-narrow-beam-echo-sounder Sea-Beam (*Renard & Allenou, 1979*) was used to map four zones along the South Aegean Trench. As a part of this program, high-resolution bathymetric maps of four key areas (Fig. 2.16) were made at a scale of 1:20,000 (*Le Pichon et al., 1979*). *Le Pichon et al. (1979)* published small portions of these bathymetric maps. They are published fully in the paper of *Huchon et al. (1982)*. *Jongsma (1975)* used seismic profiles to show detailed morphology of the Cretan Sea. The Intergovernmental Oceanographic Commission (IOC) of UNESCO

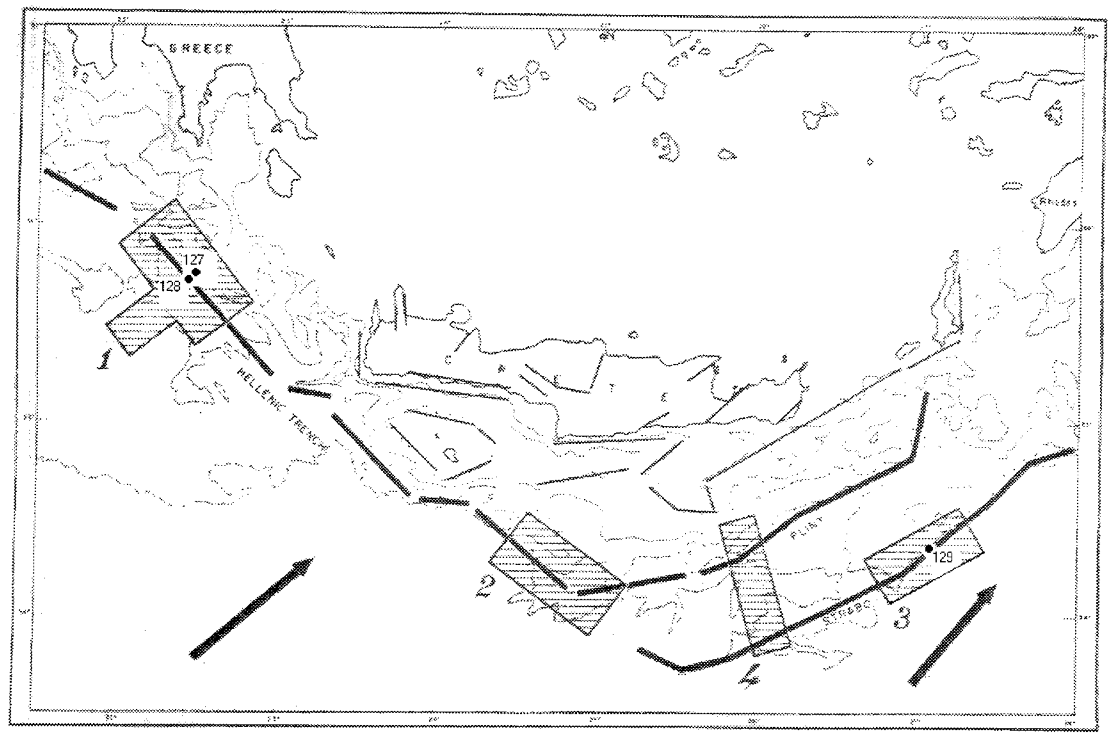


Figure. 2.16. The four boxes are the Seabeam survey areas mapped by R. V. "Jean Charcot" during HEAT cruise. The dots with numbers are locations of DSDP drilling sites. The arrows are the regional slip vectors of the Mediterranean sea floor with respect to the Hellenic arc (after *Le Pichon et al., 1979*).

(IBCM, 1981) published a 1:1,000,000 bathymetric map of the Mediterranean. *Brooks & Ferentinos (1980)* and *Ferentinos et al. (1981)* have studied the bathymetry and shallow structure of North Aegean Trough (NAT) with single channel reflection seismic surveys and similar acoustic techniques. For the Sea of Marmara there are bathymetry data in Hydrographisches Institut (1987). In 1989 (IBCM, 1989) IOC of UNESCO was issued, as an overprint to the International Bathymetric Chart of the Mediterranean (IBCM, 1981) (see *Morelli, 1990; Truffert et al., 1993*). TCB SHOD (Türkiye Cumhuriyeti Bahriyesi Seyir Hidrografi ve Oşenografi Dairesi; the Turkish Navy Hydrography Service) and T.C. Dz.K.K. (Türkiye Cumhuriyeti Deniz Kuvvetleri Komutanlığı; Turkish Commander Army Navy 1990, 1993, 1994a, 1994b) obtained 5 m grid interval bathymetry data from the coasts of Turkey (*Genç et al., 1996b; Güneysu, 1998*). *Smith et al. (1995)* published bathymetric data from the Marmara Sea that have been prepared by TCB SHOD. *Wong et al. (1995)* published 50 m interval data for the Marmara Sea. They modified bathymetry from Musee Oceanographique de Monaco (1960) and Hydrographisches Institut (1987) using their data. *Genç et al. (1996a,b)* digitised the data of coastal bathymetric contours of Turkey from TCB SHOD.

They combined these data with data of IBCM, the bathymetric data of the Marmara Sea from seismic profiles and topographic data of western Anatolia from MTA (1:25,000). *Woodside (1995)* published a detailed bathymetric map for the Anaximander Mountains, which lie at the junction between the South Aegean and Cybrus arcs. There is no bathymetric map of the Aegean and surrounding areas drawn from the completion of various data sources.

Topography

A geomorphplogical map of Turkey (1:1,000,000) has been published by the Turkish General Directorate of Mineral Research and Exploration (MTA, 1991). Global topographic data with 5 minute grid interval resolution are available from NGDS. *Genç et al. (1996b)* collected data from various sources and prepared topographic maps with 2 km grid interval values for western Turkey. To see the morphologic features they showed the relief map drawn from these data illuminated from various directions.

Remote Sensing

Foose (1985) made a satellite survey of the Mediterranenan (Fig. 2.17). Enhanced LANDSAT-5 Thematic Mapper (ITM) images, covering Central Macedonia, in northern Greece, were evaluated digitally for the purpose of mapping various morphostructural and geological features by *Astaras & Soullakel (1990)*. *Angelier et al. (1982)* presented fault traces from aerial images. They compare them with Greek geological maps. Almost all geomorphologists have used remote sensing results.

Submersible Studies

In 1979, a field study using the 3000 m submersible "Cyana" was conducted in the South Aegean Trench system (see *Le Pichon et al., 1979, 1980, 1981, 1982*) (Fig. 2.16). The South Aegean Trench is the first subduction zone in which both Sea Beam and submersible technologies have been used simultaneously.

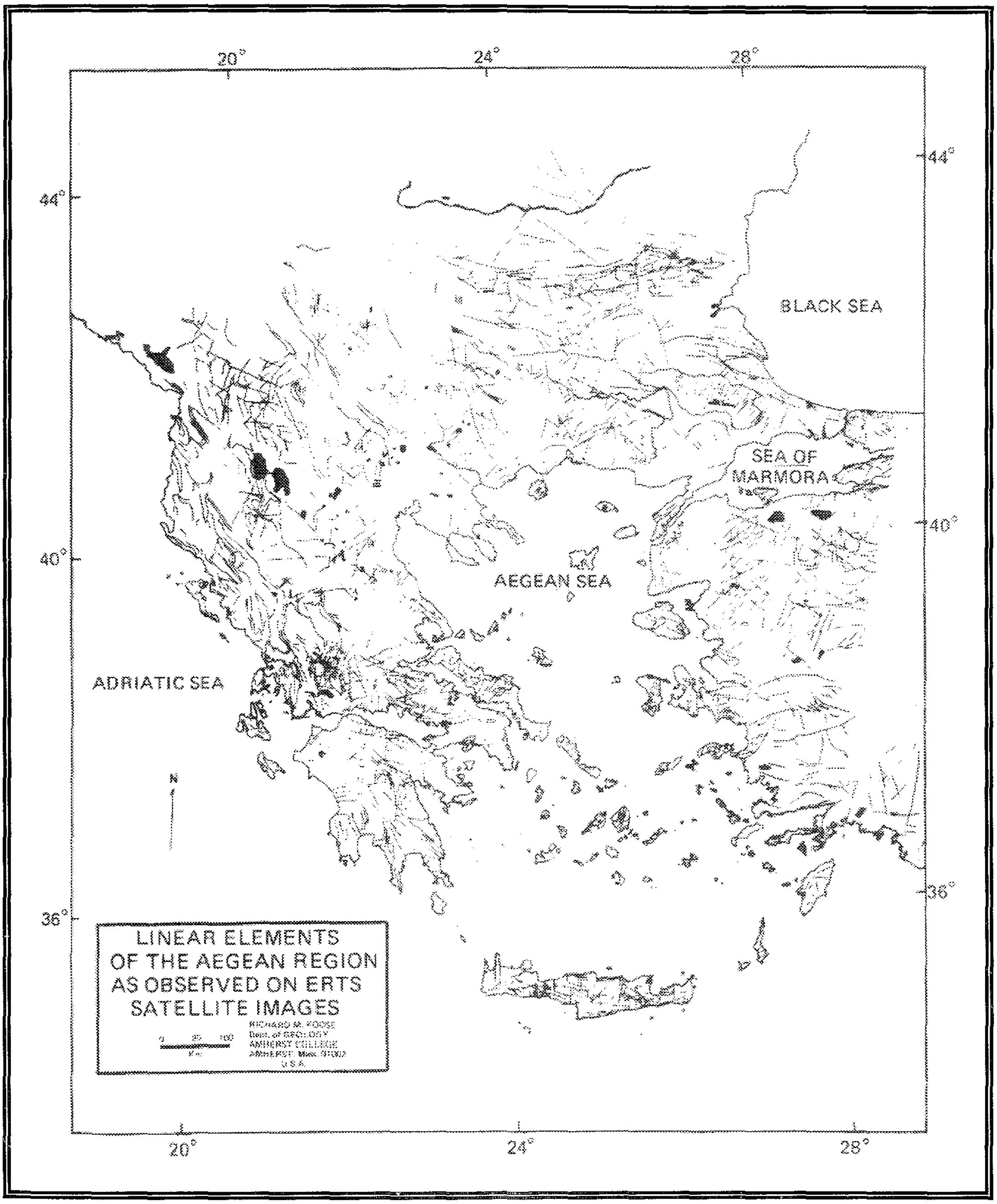


Figure. 2.17. lineament map of the mainlands in the Aegean region (after *Foose, 1985*).

2.5.2. General Interpretation

The Direction of the axis of maximum shortening and extension could be determined from Sea-Beam data, submersible dives, field exploration and geological remote sensing (photogeology) satellite imagery. Lineament orientation, shape and length provide important basic information about the direction and nature of the stress field. Visual angle, artificial effects, the Sun's illumination and light type are important parameters of remote sensing (*Foose, 1985*). For tectonic purposes the light angle is taken low in relief maps. Lineaments and their identification involves the use of raised plastic relief maps, commercially produced from existing bathymetric and topographic maps and aerial photographs (see also *Foose, 1985*).

In the Aegean Sea floor, the morphology and bathymetry are almost completely controlled by the faults only where they are active, because the sedimentation rate is high (*Angelier et al., 1982*). But footwalls under water are protected for a long time because wear and tear does not take place under water. In the Aegean Basin there are some very well-developed preferred orientations of lineaments that reflect radial tensional conditions in the crust (Figs. 2.18) (e.g., *Kronberg & Gunther, 1978*).

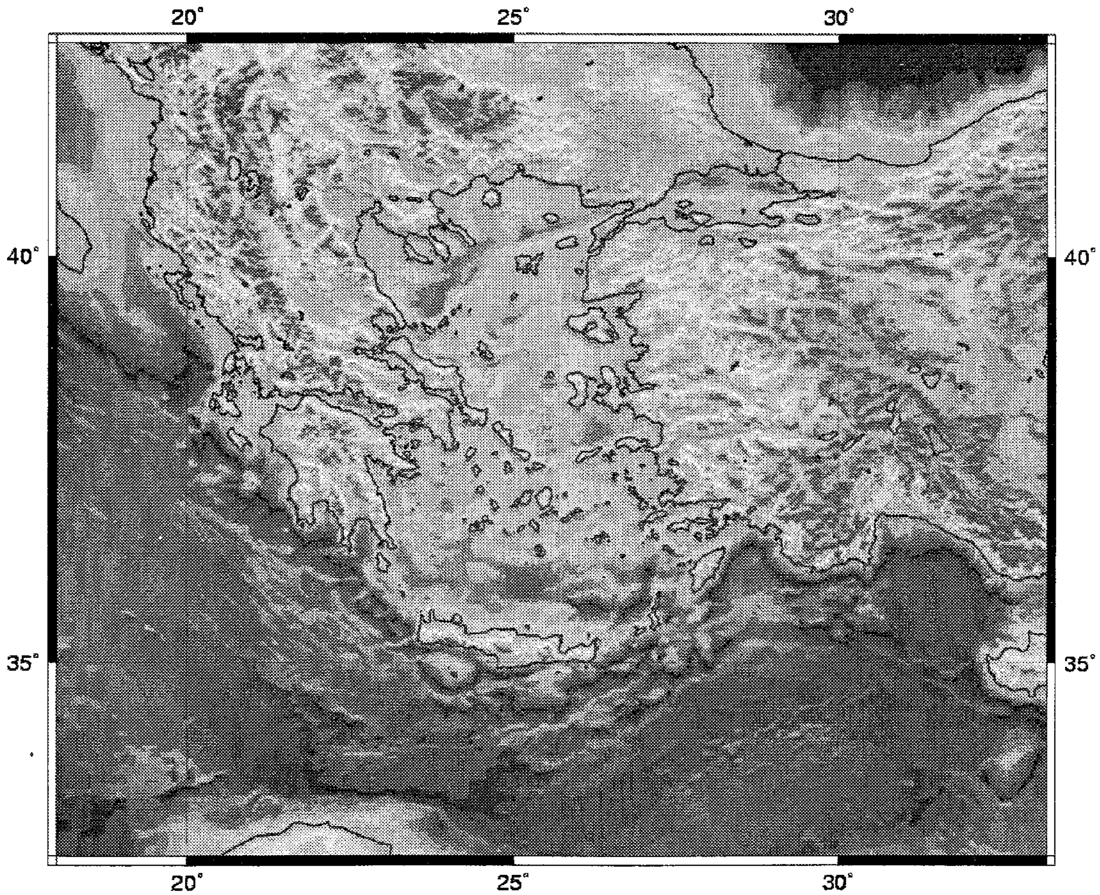


Figure. 2.18. Relief map of the Aegean and surrounding area (after *Genç et al., 1996*).

The dominance of normal faults is also obvious in satellite photographs all over the Aegean region (*McKenzie, 1978*). Thirteen percent of all straight lineaments lie within N-N50W orientation, suggesting a strong system of normal faults in Greece (*Foos, 1985*). Along the coast of Albania 68 per cent of all wavy linements are in the same direction, which suggests strongly that there was a significant NE-SW compressional stress field. Visual examination of satellite images shows normal faults sharply truncating thrusts in this region (*Foos, 1985;* see also *McKenzie, 1978*).

In Crete thirty-two percent of all lineaments and many wavy lineaments that are visible on Landsat photographs define a strongly preferred orientation of thrusting systems that is parallel to the northern coast line (*Foos, 1985*). But it is necessary to check in the field. In particular, some normal faults are almost certainly active, cutting the alluvial plains and the hilly regions, but are not visible on the photographs (e.g., *McKenzie, 1978*).

The fault segmentation and variations in hanging-wall lithology (*Roberts and Jackson, 1991*) control the drainage system in land areas of the Aegean region (*Taymaz et al., 1991*). Over the Menderes Massive the rivers have eroded the surface more vertically than horizontally (*Arpat & Bingöl, 1969*). Some of these rivers run in very narrow valleys having a depth of approximately 500-600 m (*Arpat & Bingöl, 1969*). Considering these findings, *Arpat & Bingöl (1969)* suggested that the Menderes massif has been uplifted as a whole.

The morphology of the sea floor also indicates that the crust type is continental. *Arpat (1976)* pointed out that the Aegean Sea does not have the morphological characteristics of an oceanic crust, especially near coastal areas (Fig, 2.18).

2.6. FIELD WORK ON FAULTS

2.6.1. Data

Geological History of Faulting

A description of sub-phase stress fields of the last extensional tectonic phase can be made by means of fault set analyses (e.g., *Angelier, 1977, 1978, 1979; Mercier et al., 1979, 1987; Angelier et al., 1981, 1982; Barrier & Angelier, 1982; Lyberis et al., 1982*). 4,770 measurements of faults were performed by *Angelier (1978)* in the southern Aegean region, of which most are on Crete and surrounding islands. After that *Angelier et al. (1982)* mapped approximately 4,000 faults around the southern Aegean on land and sea floor, of which 3,700 are reliable (*Angelier et al., 1982*) (Fig. 2.19, 2.20). They combine their new data with the field data of *Angelier (1978)*. They also combine air photograph, satellite imaginary and sea

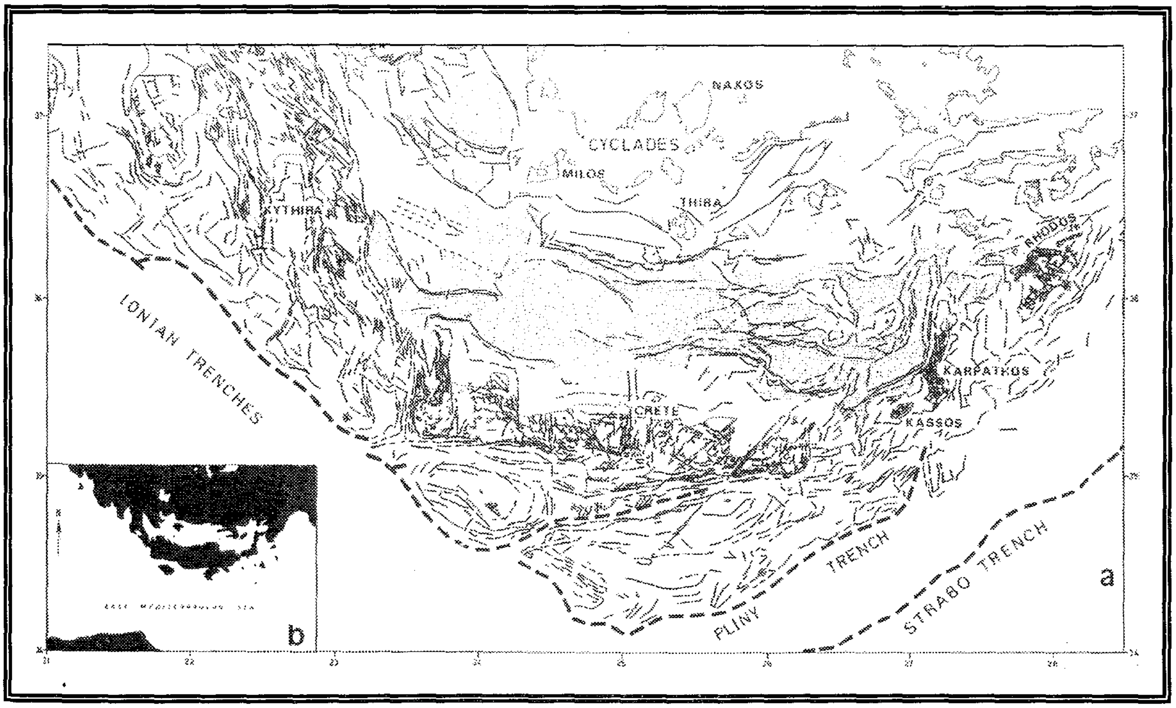


Figure. 2.19. (A) General map of main fault lines of the South Aegean Arc, compiled from various sources (see *Angelier et al., 1982*). Dashed lines-axes of South Aegean trench complex. Depths of more than 800m in the Cretan basin- dotted pattern. (B) Location map (after *Angelier et al., 1982*). Black- shallower than –800 km, white- deeper than –800.

beam data with these data (Fig. 2.19). They used them to draw stress trajectories (fields) for different sub-phases of the last tectonic phase. After that they made deformation reconstruction models for each sub-phase. *Mercier et al. (1989, 1991)* reviewed the structural data from the North Aegean region. The principal tool of their analysis is the construction of an average regional stress tensor from measurements of slip on numerous faults of different orientation (*Carey, 1979*).

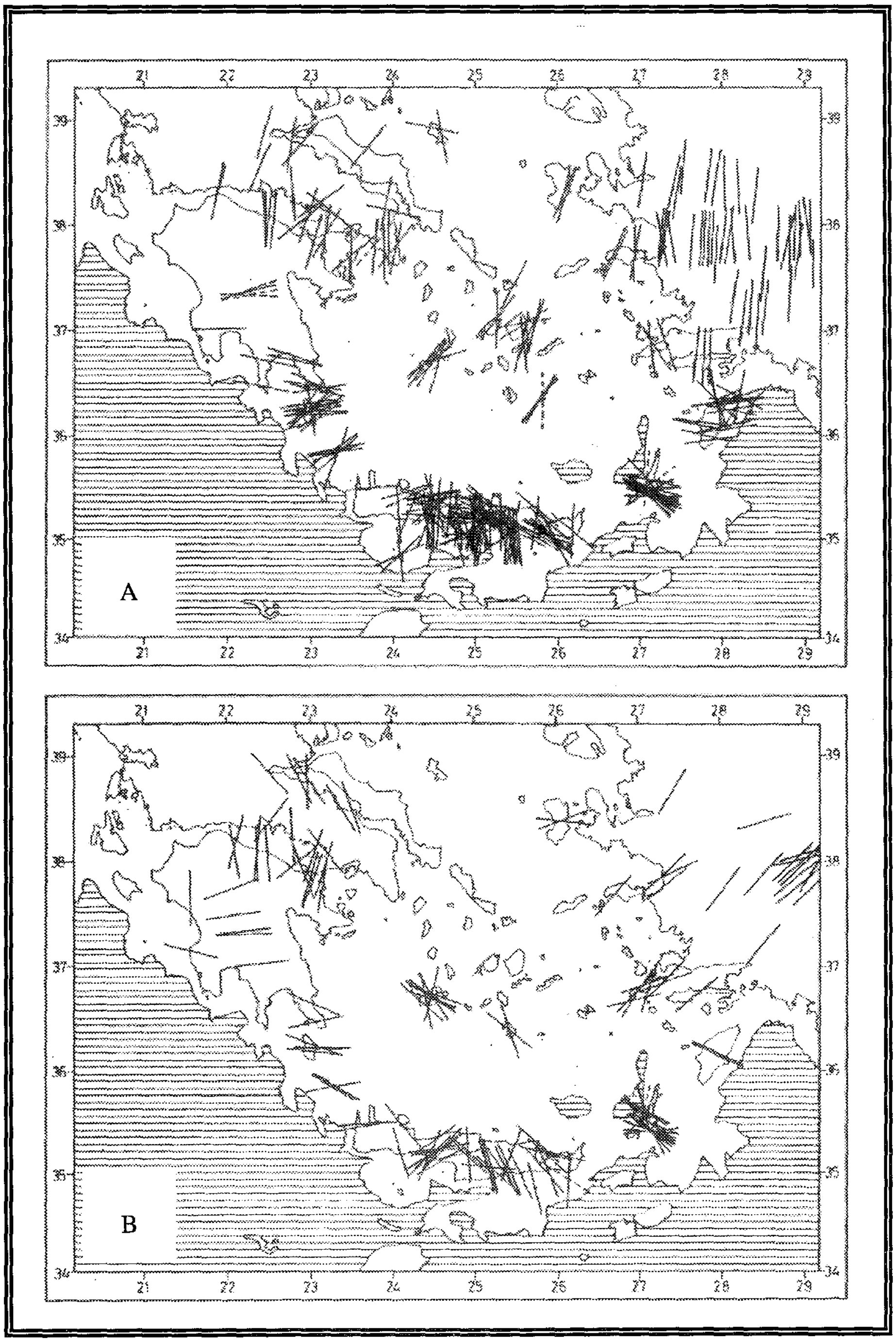


Figure. 2.20. Result of tectonic analysis of fault mechanisms in the central-southern Aegean region (after *Angelier et al., 1982*). Sticks-strikes of the horizontal or sub horizontal minimum principle stress σ_3 (directions of extension). (A) Pliocene and Early Quaternary. (B) on the basis of regional tectonic consistency. All mechanisms purely extensional.

Seismotectonics

In the western Turkish mainland, *Ketin (1969)* investigated the relation between tectonic features and earthquakes. He analysed the active tectonic features of Turkey. *Arpat & Bingöl (1969)* studied the Alaşehir earthquake surface break and the kinematics of grabens of western Turkey and tectonism of the adjacent region in detail. Previously *Pınar (1943)* mapped the active faults around the Marmara Sea. *Ambraseys (1970)* studied the North Anatolian Fault (NAF) zone in detail. *Papazachos & Comninakis (1971)* and *Mulder (1973)* studied on Crete and mapped the stress field by field geology techniques. *McKenzie (1972, 1978)* compared seismicity and active fault studies in the field all over the Aegean region. *Crampin & Üçer (1975)* made a field study in northwestern Turkey. *Dewey & Şengör (1979)*, *Barka & Hancock (1984)* and *Koçyiğit (1990)* made field studies on NAF zone to find the first occurrence of this fault. *Barka & Kandinsky-Cade (1988)* have reviewed and described the geometry of Turkish strike-slip faults (Fig. 2.21). A recent active fault map of Turkey

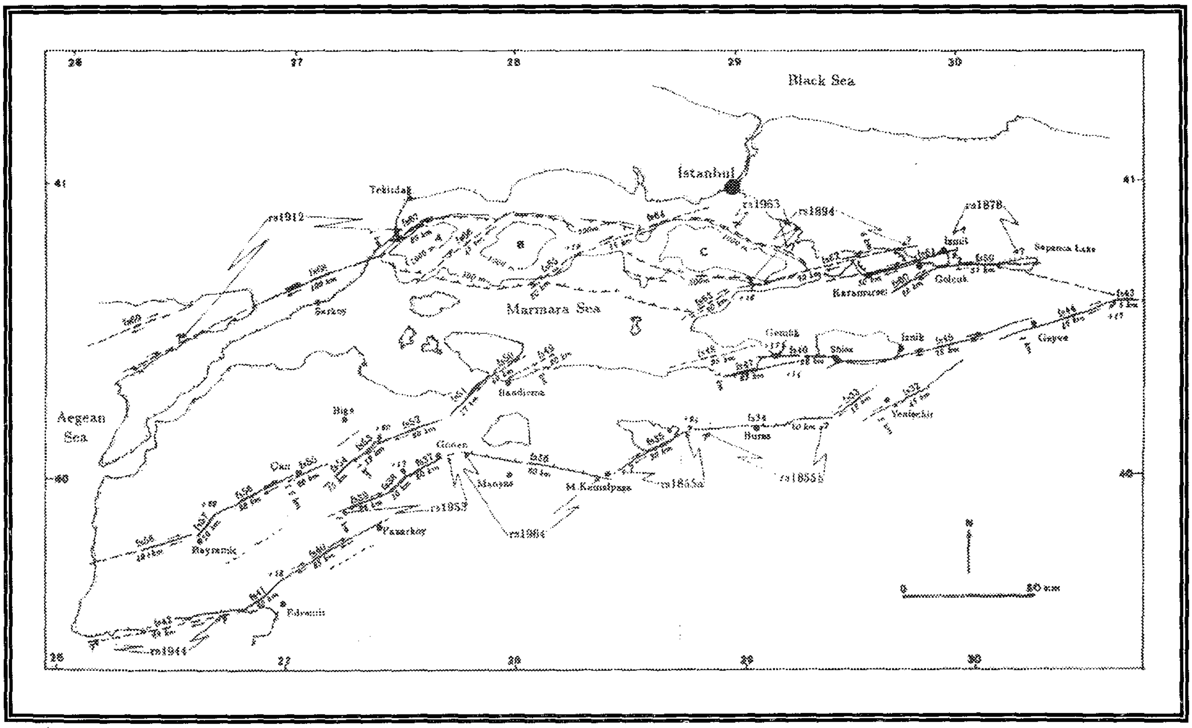


Figure. 2.21. Active fault segments in the northwest Turkey (after *Barka and Kandinky-Cade, 1988*).

(1:1.000.000) for land areas was prepared in 1987 (*Şaroğlu et al., 1997*) and published in 1992 by MTA. An extensive trenching effort was made along the İznik-Mekece section of the middle strand of the NAF by *Ikeda et al. (1989, 1991)* and *Barka (1993)*. *Ambraseys & Jackson (1990)* determined the location, magnitude, and likely style of faulting for all earthquakes with Ms>5.8 of central Greece between 1890 and 1988, using macroseismic and instrumental data. *Armijo et al. (1996)* made a detailed field investigation of the Corinth rift. *Caputo (1990)* made a field study of the NAF. During the study new archeological data were discovered. *Barka (1997)* reviewed the existing fieldwork in the Marmara region. *Greber (1997a,b)* studied stratigraphic evolution and tectonics in the east Marmara Sea. In the same region, field analyses were carried out recently by *Zanchi et al. (1993)*. There are many other regional and local field studies in the Aegean region (e.g., *Ateş & Tabban, 1976; Ambraseys & Tchalenko, 1970, Ketin & Roesli, 1953*), which could not reached.

In some cases, earthquakes have produced impressive surface ruptures several kilometers long. Such structures give important information on the seismotectonic characteristics of the region. Among the surface breaks analysis are the 1912 Saros earthquake (*Ambraseys & Finkel, 1987*), 1928.04.14 (*Richter, 1958*), 1928.04.18 (*Richter, 1958*), 1930.03.28 (*Ambraseys & Tchalenko, 1972*), 1943.06.20 (*Fouche & Pınar, 1943*), 1953 Yenice-Gönen earthquake (*Ketin & Roesli, 1954; Diligan and Hagiwara, 1955*), 1954.04.30 (*Ambraseys, 1975*), 1963.07.26 (*Ambraseys, 1975*), 1963.09.18 (*Ketin, 1966*), 1964.10.06 (*Ketin, 1966*), 1966.10.29 (*Ambraseys, 1975*), 1967.07.22 (*Ambraseys & Zatopek, 1969*), 1967.11.30 (*Sulstarova & Kociaj, 1969; Arsovsky, 1970*), 1969.03.28 Alaşehir earthquake (*Arpat & Bingöl, 1969; Ambraseys & Tchalenko, 1972; Erinç et al., 1971*), 1970.03.28 Gediz earthquake (*Ambraseys & Tchalenko, 1972*), 1971.05.12 Burdur earthquake (*Ambraseys & Tchalenko, 1972; Erinç et al., 1971*), 1986 Kalamata earthquakes' (*Lyon-Cean et al., 1988*). *Barka & Kandinsky-Cade (1988)* mapped the active fault segments in northwestern Turkey in detail (Fig. 2.21). *Barka & Reilinger (1997)* reviewed the earthquake ruptures (1800-1995) in western Turkey and surrounding region (Fig. 2.22).

2.6.2. General Interpretation

In some part of the region a significant part of the deformation occurs aseismically (Fig. 2.26). For this reason seismicity does not accommodate the total predicted deformation (e.g., *Pondrelli et al., 1995*). One can recognise probable aseismic deformation by field, study. In addition, active structures have been effectively aseismic for only the last hundred years. In these cases also, field studies on faults are necessary. It is probable that the maximum length of fault segments restricts the maximum magnitude. The length appears to be 15-20 km (*Ambraseys & Jackson, 1990*). Reliable estimates of scalar moment can be computed from historical catalogues to try to evaluate motions over a longer time interval (e.g., *Jackson & McKenzie, 1988; Pondrelli et al., 1995*).

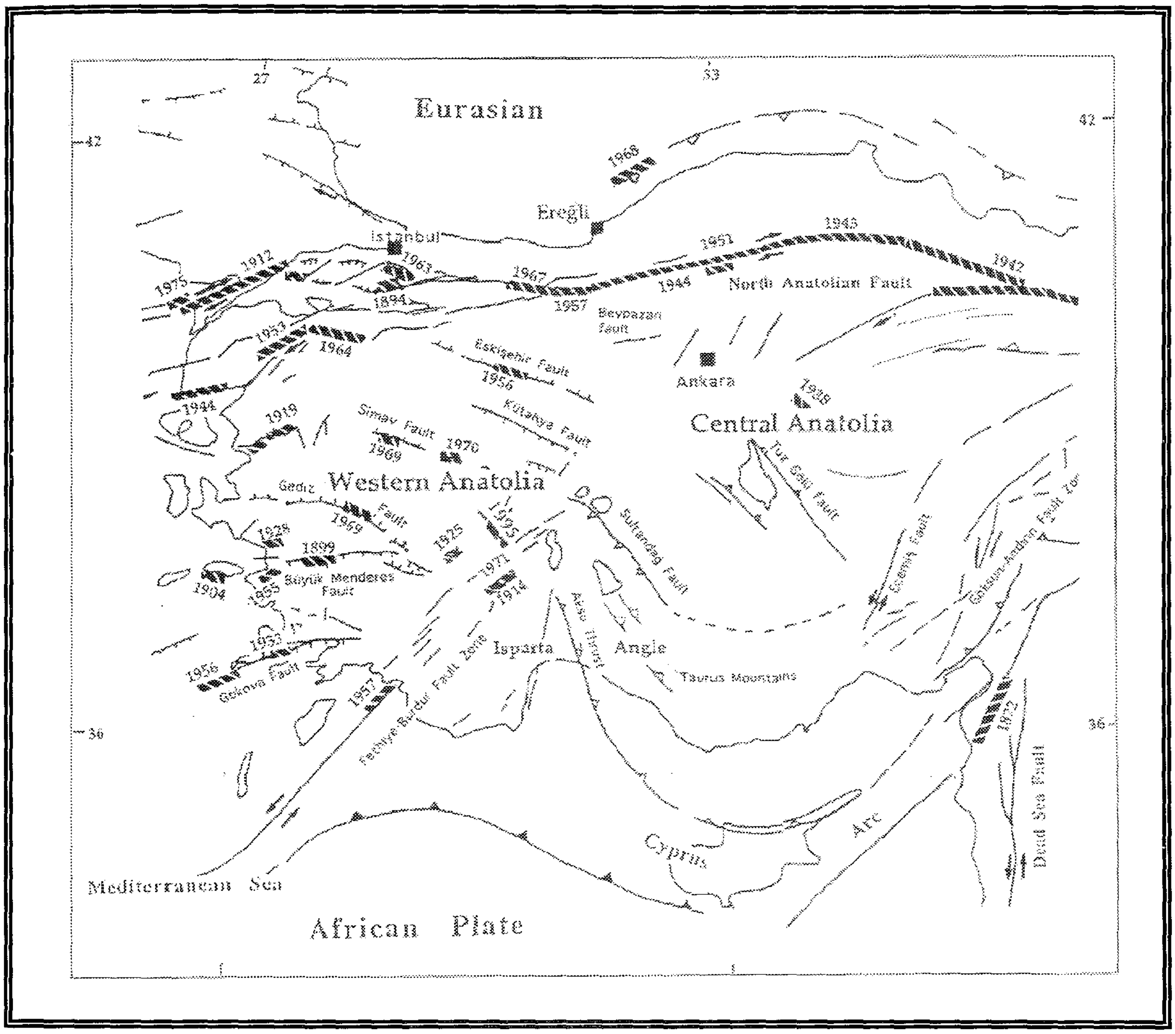


Figure. 2.22. Distribution of earthquake ruptures in the Anatolian region between 1800-1995 (after *Barka and Reilinger, 1997*).

Studies of the kinematics of faults that have been active during the latest extensional phase of the Aegean region offer a detailed view of stress fields. Such structures have given important information on the seismotectonic characteristics of the region at a specific time. *Angelier (1978, 1977), Mercier et al. (1972, 1987)* have shown that the state of stress has changed since the well-known extensional regime of the lower Pliocene time. They also showed that the deformation history of this stage could be divided into sub-phases. Late Miocene and Plio-Quaternary tectonics of the Aegean are poly and complicated; they include both extensional and compressional events (*Angelier, 1978*). In the internal domain, neotectonics appear to be characterized by an alternation of long-lived extensional periods and short-lived compressional events (*Mercier et al., 1987*). *Papzachos & Comninakis*

(1971) and *Mulder (1973)* studied Crete and mapped the ongoing compressional stress field using field geology techniques.

Normal faulting is vigorous in regions of western Turkey, Bulgaria, Greece and Albania. The main dip-slip fault and grabens that are active in Turkey, are: the area between İznik-Bandırma, Adapazarı-Düzce - Hendek depressions, İznik Lake, Sapanca Lake, central part of the Sea of Marmara, gulf of Sarof, Edremit, İzmit, Gemlik and Kerme, the flanks of valleys of Gediz (Alaşehir) and Büyük and Küçük Menderes, Simav, Bergama grabens, Burdur Lakes etc. (*Ketin, 1969*, *Arpat & Bingöl, 1969*). There are also many of smaller dimensions of dip-slip faults all over Turkey (*Ketin, 1969*). Previously *Arpat & Bingöl (1969)* found that western Turkey was covered by listric faults, whose dip angle decreases with depth. In these grabens the sequences' deep angles decrease with decreasing age *(see also McKenzie, 1972)*.

If both focal mechanism solutions and surface breaks are available for the same shock, the ambiguity between the fault and auxiliary plane can be resolved, and hence the slip vector can be determined (see *McKenzie, 1972*). For this reason, detailed maps of surface breaks of large earthquakes are extremely important, but, unfortunately, few such reports exist (*McKenzie, 1978*).

2.7. SHALLOW REFLECTION PROFILES

2.7.1. Data

Mediterranean

The first measurements in the East Mediterranean were obtained by H.M.S. Challenger in 1952 and were interpreted by *Gaskell & Swallow (1953)* and *Gaskell et al. (1953, 1958)*. Long Flexotir seismic profiles in which good penetration was obtained have shown the thickness of the sedimentary sequence in the eastern Mediterranean from Africa to just south of Milos (Profile Kaita) (*Sancho et al., 1973*). The reflection seismic profiling (sounding) was done during 3 marine geophysical cruises undertaken by Cambridge University, U.K., and the Bundesanstalt für Bodenforschung, Hannover, W. Germany. The Osservatorio Geofisico Sperimentale (OGS) shot a vertical reflection seismic line (MS33) off Trieste in 1971 (*Finetti, 1976, 1982*). This seismic line crosses the Sirte abyssal plain from SW to NE, the deformation front, the western Mediterranean Ridge and the South Matapan trench (see also *Lallemant et al., 1994*). The STREAMERS project shot a marine vertical seismic reflection profile ION- 7 with a powerful air gun source running from SW to NE for 180 km from the deep Ionian basin into the western Gulf of Patras, augmented with a few wide-angle seismometer stations on land (*Hirn et al., 1996*).

Aegean

In the South Aegean Trough (SAT) continuous seismic reflection profiling was first carried out in 1959 (see *Hersey, 1965*) to delineate sedimentary basins and pounding effects, particularly south of the South Aegean Arc (SAA) and in the Tyrrhenian and Ionian Seas. Earlier seismic reflection work in the southern Aegean (OSS Oceanographer, 1967) has been published by *Watson & Johnson (1969)*, *Ryan et al. (1969)* and *Wong & Zarudzki (1969)*. Some other previous efforts to obtain information on the sedimentary structure of the area were made by *Ryan et al. (1969)* and *Wong et al. (1971)*. In November 1972, R.R.S. Shackleton ran 200 nautical miles of air gun profiles in the SAT in conjunction with a Heat Flow program (*Jongsma, 1975*). In southern Crete seismic reflection data were collected and published by *Woodside (1975)*. Single channel data along several seismic profiles were collected south and SE of the Crete (*Jongsma, 1977; Jongsma et al., 1983, Jongsma, 1987*). In the Aegean region the Institute of Geophysics of NASA (1981) studied some 20 seismic reflection profiles and two refraction profiles. The study of *Martin (1987)* represents available seismic lines in the Aegean and led to a map of offshore Cenozoic faults inside the basins (see also *Mascle & Martin, 1990*) (Fig. 2.23, 2.24). Multi-channel seismic reflection data were

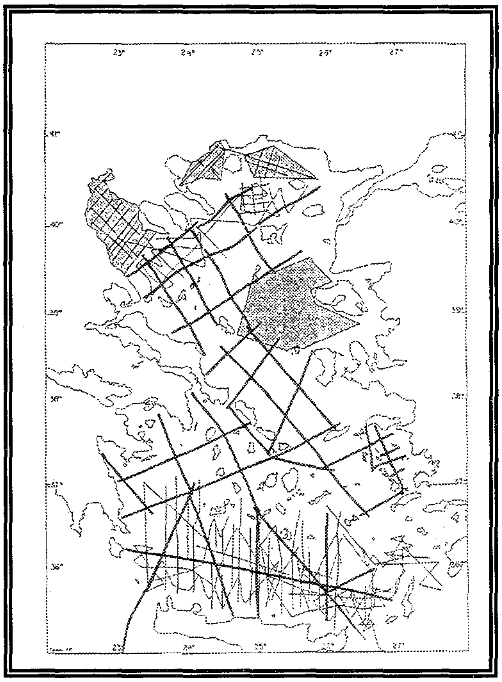


Figure. 2.23. Location of the seismic reflection profiles (After *Martin, 1987; Mascle and Martin, 1990*). Heavy lines indicate multichannel seismic data (MCS); light lines indicate single channel seismic data; stippled areas are where detailed MCS surveys have been conducted.

acquired from the South Aegean Trenches and the eastern Mediterranean Sea during the MSC Prismed survey (March 1993) (*Chaumillon & Mascle, 1995; Chaumillon et al., 1996*). For the region around the Anaximander Mountains at the junction between the South Aegean and Cyprus arcs, *Woodside (1995)* published a bottom reflectivity map. *Eryılmaz (1996)* also studied the basin structures in the Aegean Sea using shallow seismic data.

Several Local Studies

Marathon Oil (1975) provided a N-S trending seismic reflection profile in the Marmara Sea. Several short high-resolution sub-bottom profiles showing neotectonic offsets near the Peloponnesus have been published and interpreted by *Stanley (1974)*. The sedimentary cover, particularly of the Thermaikos, has been studied in connection with oil exploration by multi-

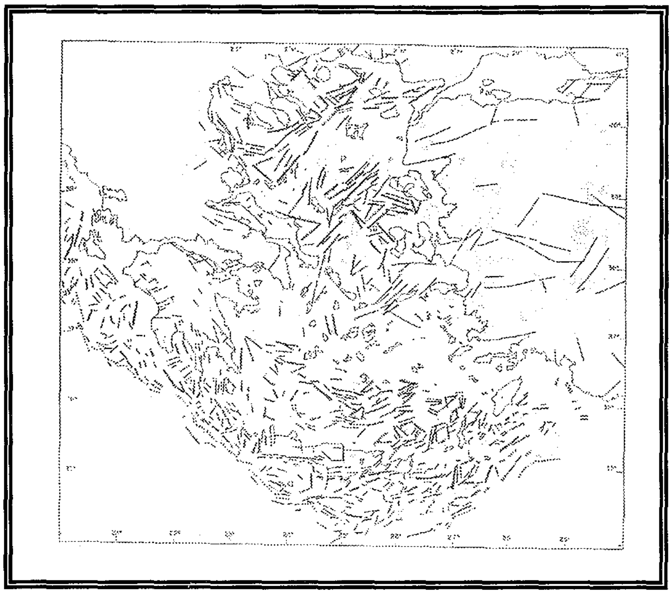


Figure. 2.24. General structural sketch of the Aegean Sea as deduced from seismic reflection study (*Mascle and Martin, 1990;* see also *Martin, 1987*).

channel reflection seismic profiles (*Lalechos & Savoyat, 1977*). In the North Aegean, Jean Charcot 1972 obtained 5 shallow seismic refraction profiles. Three reflection seismic profiles were run in 1972 during the Heat Flow program, which provide information on features in the North Aegean Sea (*Jongsma, 1975*). In 1980, a geophysical study of the North Aegean Trough (NAT), comprising seismic reflection and refraction studies, was conducted by *Ginzburg et al. (1987)*. *Saner (1985)* studied Saros bay by means of seismic profiling. The Gemlik bay area was studied by *Kurtuluş (1985)*, who interpreted MTA Sismik-1 obtained from high-resolution shallow (<300m) seismic reflection profiles obtained in 1984. *Myrianthis (1984)* gave reflection seismic results for the Gulf of Corinth graben. The structure and fault patterns of recent deformation in the Central NAT were studied by *Roussos & Lyssimachou (1991)* based on detailed interpretation of marine reflection seismic lines. Data from continuous seismic profiling, echo sounding, and core sampling of the Late Quaternary sedimentary cover acquired during the 1991 expedition of the RlV Moskovskii Universitet are provided by *Roslyakov & Srokin (1997)* for the northern slope of the Sporades basin. On the basis of multi-channel seismic and the other geophysical data the tectonic and sedimentary evolution of the Saros basin area was be discussed by *Sarı et al. (1995)*. *Ergün & Özel (1995)* studied the shallow seismic reflection data in northern part of the Marmara Sea. *Koral & Öncel (1995)* published İzmit bay seismic profiles. *Smith et al. (1995)* used high-resolution sparker seismic profile data to study the neotectonics and sedimentation of the Marmara region. *Wong et al. (1995)* published seismic reflection studies of the Marmara region. To investigate the marine geology and geophysics of the area, high resolution shallow seismic data were acquired along the coasts of Marmara Sea with the cooperation of Turkish Naval Forces, Department of Navigation, Hydrography and Oceanography (TCB SHOD), İstanbul Technical University (İTÜ) and TÜBİTAK in a joint Marine Geology and Geophysics project. The seismic data were acquired using a high-resolution seismic data collection system (*Kurt, 1994; Yılmaz, 1996*). *Uluğ et al. (1998)* depict the seismic facies and stratigraphy, thickness, and major sedimentary sequences of the Gulf of Gökova, based on high-resolution seismic profiles. The data were combined with those of the Turkish Petroleum Company (TPAO). Recent detailed multi-channel seismic reflection data has been collected in the Marmara Sea (*Çetin et al., 1998*). Seismic data in İzmit bay were obtained in 1994 (Dokuz Eylül University) (*Uluğ et al., 1998)* (see also *Şenöz, 1998*). The northeasten part of the Strait of Çanakkale was investigated by *Demirbağ et al. (1998)*, using seismic data. This study, which was conducted by the Marine Sciences and Management Institute (İstanbul University), Mining Faculty (İstanbul Technical University) and supported by the Earth, Sea, Atmosphere and Environment Exploration Group of TÜBİTAK, was made at the Aegean exit of the strait of Çanakkale (Dardanelles) (*Alpar et al., 1998*).

The collection of the data took 20-25 year and it was collected by various organizations using differing standards and techniques. Because of these, *Saatçılar et al. (1996)* have reviewed and put all the available data together, using one standard for all the available seismic data in Aegean Sea. *Saatçılar et al. (1996)* first studied the Saros graben (Fig. 2.25). *Eryılmaz et al. (1998)* reinterpreted seismic reflection data, which were collected by different institutions, to explore the main structural, stratigraphic and sedimentologic features of the Aegean Sea.

2.7.2. General Interpretation

Seismic exploration beneath the Aegean Sea shows that three layers exist in the basins (e.g., *Martin, 1987; Mascle & Martin, 1990; Eryılmaz, 1996; Eryılmaz et al., 1998*). These are, from bottom to top: the basement, the Messinian deposits with evaporates, and Plio-Quaternary deposits. Widespread evaporates were the most characteristic feature of Late Miocene (Messinian) times (*Woodside, 1975*), called reflector "M" in the Aegean Sea. *Mascle & Martin (1990)* detected late Miocene sediments as the oldest sea sediments on top of the

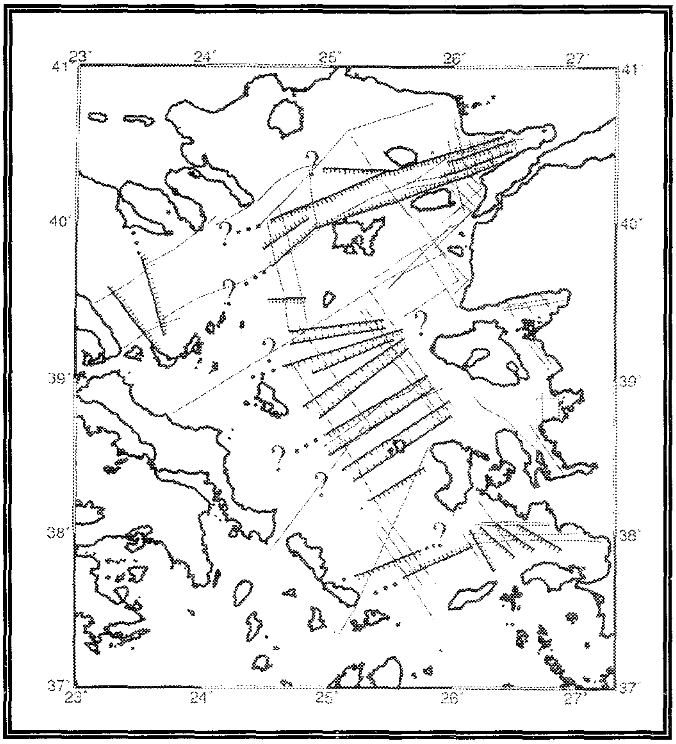


Figure. 2.25. Active normal fault map of the Aegean Sea deduced from seismic profiles (after *Saatçılar et al., 1996*). Thin lines show the seismic lines, the lines with thick marks show the normal faults.

Oligocene-Early Miocene conglomerate. According to *Le Pichon et al. (1984)*, at the base of the Late Miocene-Quaternary sea sediments, a continental conglomerate of probable Lower-Miocene age is assumed to mark the end of the tectonic compressive phase. They correlate the seismic profiles with drilling results to find out the ages of the corresponding layers. From shallow seismic data it is seen that the faulting mechanism continues to the present (e.g., *Mascle & Martin, 1990; Eryılmaz, 1996; Eryılmaz et al., 1998*). Recently, *Uluğ et al. (1998)* suggested that a large part of the Gulf of Gökva basin is subsiding based on seismic reflection data. According to *Sarı et al. (1995)*, the basins of the North Aegean and the Sea of Marmara show characteristics of rapid subsidence accommodated by extension and transform motions (see also *Saatçılar et al., 1996*). *Lalechos & Savoyat (1977)* inferred the distribution of Neogene sediments and showed that their maximum thickness in the NAT is near the Sporades. In the Thermaikos, some 1000 m of Pleistocene and Pliocene sediments were penetrated, followed by some 1600 m of Miocene and 2000 m of Eocene-Oligocene sediments (*Ginzburg et al., 1987*). The sediment thickness becomes thinner in the Central Aegean Sea and their reflection characters are disturbed (*Eryılmaz et al., 1998*). The thinnest sediments are found in the southern Aegean Sea (*Eryılmaz et al., 1998*).

In the eastern Mediterranean Sea the thick sedimentary cover (about 8 km in the Ionian basin and at least 12 km in the Herodotus plaint are known from both reflection and refraction data and are thought to include Mesozoic to Tertiary sequences (e.g., *Finetti, 1976; Ginzburg & Ben-Avraham, 1987*).

In the northern Aegean Sea, faults generally trend NE-SW (Fig. 2.24, 2.25), whereas in the Central Aegean Sea, the general fault trend is in the direction of NW-SE (Fig. 2.24). The Samos and Tharia Basins were interpreted as extensions of the generally E-W trending Menderes Neogene graben system (*Eryılmaz et al., 1998*). Further south, the fault systems developed in the north of Crete are W-E to the east and NW-SE to the west (Fig. 2.24) (see also *Martin, 1987; Mascle & Martin, 1990; Eryılmaz et al., 1998*). In general, faults in the east of the Aegean region strike roughly NE-SW (Fig. 2.24; 2.25) (*Martin, 1987; Saatçılar et al., 1996*) while to the west faults strike NW-SE (*Martin, 1987*) (Fig, 2. 24) (see Section 3 for detail).

Özhan et al. (1985) and *Kavukçu (1990)* studied shallow seismic reflection profiles in the İzmit bay area and pointed out that the area is not a single graben, but that there are two basins. *Akgün & Ergün (1995), Koral & Öncel (1995), Şenöz (1998), Uluğ et al. (1998)* studied seismic reflection profiles and onshore geological structures, and concluded that the İzmit bay area consists of a few en-echelon strike-slip fault segments. They suggest that these form pull-apart basins such as the İzmit and Karamürsel basins (*see also Barka & Kadinsky-Cade, 1988; Barka, 1991, 1992, 1993*). In the İzmit basin, subsidence is greatest in the south so the northern area is being tilted southward (*Uluğ et al., 1998*). *Smith et al. (1995)* found that normal faults and half grabens in the Marmara sea strike approximately E-W, WNW-ESE. This suggested that the faults that occur in the northern half of the Marmara sea have a clearer morphology, since the sedimentation rate is less, relative to the southern Marmara shelf (*Barka, 1997*). The northern Marmara has several pull-apart steps, which form its basins (*Wong et al., 1995; Ergün & Özel, 1995; Barka, 1997*). The fault pattern of the southern shelf region east of Marmara Island appears to be similar to that of the onshore region between Manyas and the Marmara Sea. The studies mentioned are not enough to explain the structure beneath the Marmara Sea.

Saatçılar et al. (1996) compared the fault geometry found by using seismic reflection data with the focal mechanism solutions of the moderate earthquakes that had been investigated.

This comparison did not show any consistency between them especially for the central Aegean region. The focal mechanism solutions in the central Aegean Sea have a large strike-slip component (Fig. 5.4). However, no strike-slip geometry was observed by *Saatçılar et al. (1996).* The focal mechanism solutions of *Pavlides & Caputo (1994)* have little strike-slip component. On the other hand *Martin (1987),* who made an extensive seismic reflection study in the Aegean, suggests that right lateral strike-slip motion occurs on these faults.

2.8. SEISMICITY

2.8.1. Data

Networks

Good determinations of epicenters can be made only for earthquakes, which occurred since 1930, and focal mechanism solutions have become reliable only since the installation of the worldwide network in 1937 (*McKenzie, 1978*). Before 1961 short period polarity measurements alone are available. These are less consistent than long period readings (see for detail *McKenzie, 1972*). Accurate epicenters have been determined since 1961 by the new USCGS worldwide network of seismic stations (World Wide Standard Seismograph Network: WWSSN). Since 1980 Global Digital Seismological Network (GDSN) data have become available.

After 1976, the earthquake detection level was greatly improved through the installation of permanent stations and radio-linked seismic station networks around the Marmara sea (MARNET) and the İznik regions (IZINET) (*Üçer et al., 1997*). The networks in these regions are able to record micro-earthquakes down to a magnitude of M=2.4 (*Üçer et al., 1997*). Kandilli Observatory and Earthquake Research Institute (KOERI) installed a network which is called as IZINET in 1992 around the İznik-Mekece fault (east of Gemlik bay) to detect seismic properties of a possible earthquake as part of a multidisciplinary earthquake study of BÜ KOERI in collaboration with Japan institutions (*Lio et al., 1991*; see also *Dadaşbilge, 1997*). Since 1995, 12 seismic stations (weak and strong motions) have been operating in the Bursa area (*Sellami et al., 1997*). The project "Seismicity and seismotectonics of the Bursa region" was carried out by the following institutions: IGETH: Institute of Geophysics, Federal Institute of Technology Zürich (Switzerland), İTÜ: Department of Geophysics, Mining Faculty of İstanbul Technical University (Turkey), TÜBİTAK: Department of Earth Sciences, Marmara Research Institute of TÜBİTAK, Gebze (Turkey) and KOERI. The sub-project was initially planned for two years (1992-1993) and later extended for another two years (1994-1995). The goal of the project was to install a seismic network to allow detailed monitoring of earthquake activity in the area. *Eyidoğan et al. (1998)* investigated the earthquake activity of the Marmara Sea region by using one and half months operation of a dense network of 52 seismic stations around the Marmara Sea region in 1995, with the collaboration of the institutes, IPGS, TÜBİTAK-Malm BÜ-KOERI and İTÜ. They located the epicenters of micro-earthquakes and determined the focal mechanisms and the stress regime in the region.

After the 1978 earthquake sequence within the Mygdonian graben, the Geophysical Laboratory of the Aristotelian University of Thessaloniki installed a permanent network of 8 seismological stations around the Chalkidiki peninsula (*Christodoulou & Hatzfeld, 1988*). *Hatzfeld et al. (1989)* determined geometry of the west of the slab by using microseismic data collected from 46 temporary stations. They also determined the T axes of focal mechanism solutions on the slab. From these results they map a heterogeneous strain field in the western part of the South Aegean Arc (SAA). To obtain more detailed information about the tectonics of the western SAA, in the summer of 1986 *Hatzfeld et al. (1990)* installed a dense network of 46 temporary stations and conducted a microearthquake survey lasting 7 weeks. During the summer of 1988, *Hatzfeld et al. (1993)* installed 82 seismological stations over the SAA and southern Aegean Sea. Seismological data collected during the summer of 1988 are prepared and analyzed to discuss the microearthquake deformation in western Crete (Greece) (*Chabalier et al., 1992*). The experiment consisted, of a dense network of 34 stations, among them 13 three-component digital stations (*Chabalier et al., 1992*). *Rigo et al. (1995)* made a microseismic investigation in Gulf of Corinth. 35 stations of the temporary network were installed 5 days after the Galaxsidi earthquake (*Hatzfeld et al., 1996*).

Catalogues

Gutenberg & Richter (1954), *Galanopoulos (1960, 1963, & 1968)* and *Ergin (1967)* have published earlier earthquake catalogues for the region. Sources of seismologic data an available for periods and various reliability in *Karnik (1968, 1969)*, *Ergin et al. (1967)*, the Bulletins of the Seismological Institute of the National Observatory of Athens and publications of the BCIS, the United States Coast and Geodetic Survey (USCGS), Kandilli Observatory Database, ISS, ISC monthly catalogues, USGS PDE catalogues

Pınar & Lahn (1952) prepared the first descriptive catalogue of historical earthquakes in Turkey. Later *Ergin et al. (1967)* prepared a catalogue, including earthquakes between 1100-1964 AD. *Soysal et al. (1981)* prepared a catalogue of historical earthquakes that occurred between 2100 BP and 1900 AD by studying Selçuk, Byzantine and Ottoman documents. Recently, *Ambraseys & Finkel (1991, 1995)* have presented a more reliable and complete earthquake catalogue, especially for the destructive earthquakes that occurred in the Marmara sea and surrounding region, for the period between 0- 1900 AD. Completeness analysis for the historical earthquakes has been carried out in the Marmara region by *Üçer (1990)* (see also *Angagi, 1997*). *Üçer et al. (1997)* review the various historical earthquake catalogues for Turkey.

For the Aegean region, the relation between various magnitude scales has been derived in order to produce a statistically homogeneous catalog of historical and present century earthquakes (*Papazachos & Comninakis, 1971; Kiratzi and Papazachos, 1984; Papanastasiou, 1989; Papazachos et al., 1997*). Formulas are also available to calculate the magnitude of strong shocks using macroseismic data (*Galanopoulos, 1961; Drakopoulos, 1978; Tassos, 1984; Papaioannou, 1984; Papazachos et al., 1997; Papazachos, 1992*). *Ketin & Abdüsselamoğlu (1969)*, *Ambraseys & Tchalenko (1972)* and *Ambraseys (1985)* determine the focal depths of the some earthquakes using macroseismic data of faults.

Using joint hypocenter location techniques, *Crampin & Üçer (1975)* and modeling of the long-period body-waves, *Eyidoğan & Jackson (1985)* and *Taymaz et al. (1991)*, improve hypocenter quality in the region. *Makropoulos & Burton (1981)* also have made an effort to improve hypocenter quality in the SAA. *Papadopoulos et al., (1988)* have calculated better locations in the southwestern SAA. There have been several efforts made (*Crampin & Üçer, 1975; Makropoulos & Burton, 1981*) to relocate earthquakes, which occurred pre-1964 (see *Üçer et al., 1997*). All the earthquakes, which occurred during the instrumental period (1900-1995), have been recompiled mainly from *Ayhan et al. (1987)* in Marmara region. *Küleli (1992)* also relocated some selected critical events in the region.

Seismotectonics

Gutenberg & Richter (1954) made an early study of seismicity of the Aegean region. *Galanopoulos (1963)* tried to separate the most seismically active zones in Greece. *Canıtez & Üçer (1967)* prepared a catalogue of focal mechanism diagrams for the earthquakes that occurred in Anatolia and adjoining areas. The study was based on first motions of both P and S waves in the interval 1939-1965. *Ketin (1969)* described epicentral distribution along the general tectonic features in Turkey. *Papazachos & Delibasis (1969)*, using a statistical treatment of focal mechanism solutions of about 90 earthquakes, studied the stress regime along the South Aegean Consumption Boundary (SACB). *Caputo et al. (1970)* and *Papazachos & Comninakis (1971)* studied the Benioff zone using hypocenter distribution of

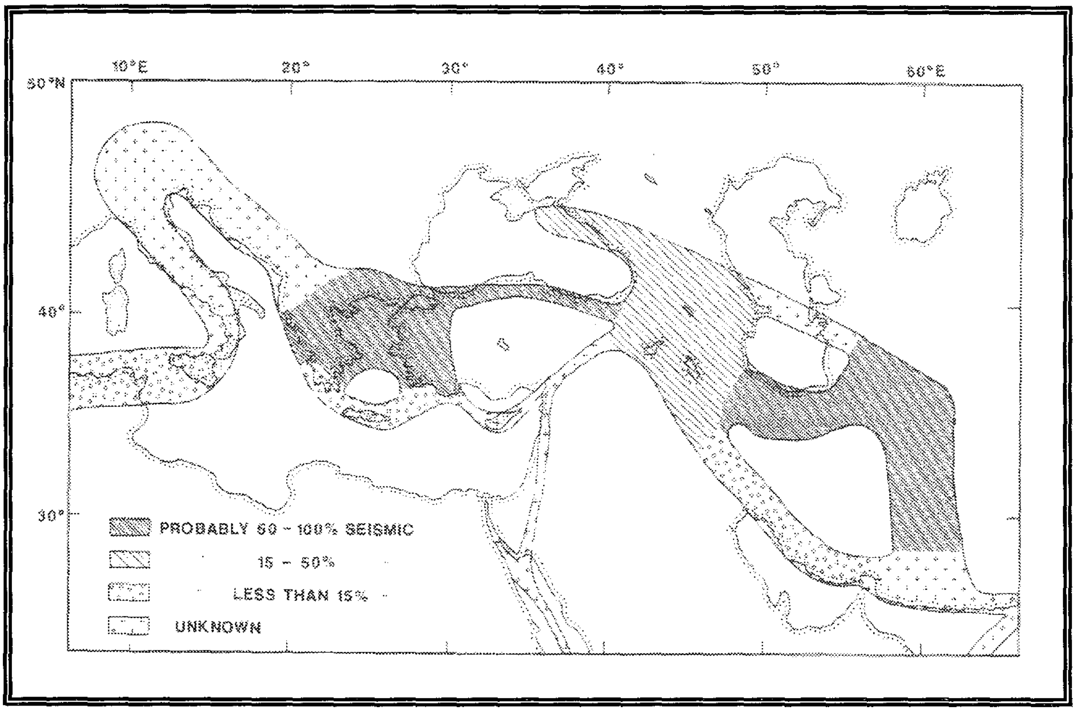


Figure. 2.26. A summary of the observational results of *Jackson and McKenzie (1988)*, showing the areas in which seismicity is likely to occur (after *Jackson and McKenzie, 1988*). The belts marked 'unknown' are those where the seismicity this century is too unrepresentative of longer periods to be conclusive or where there is no independent estimate of the overall deformation rate (*Jackson and McKenzie, 1988*).

local earthquakes. *McKenzie (1970, 1972, 1978)*, by using first motion focal mechanism solutions and seismicitiy distribution, studied the Aegean region in considerable detail. He used the WWSSN long period network that was installed in 1962 to determine the focal mechanisms. Data concerning the focal mechanism and the spatial distribution of earthquakes have been used to investigate the active tectonics of the northern Aegean and surrounding area by *Papazachos (1976)*. *Richter & Stobach (1982)* presented the three-dimensional earthquake distribution and the corresponding tectonic structures in the Aegean region. *Kocaefe & Ataman (1976)* studied Antakya, Finike, Denizli and surrounding area by means of seismicity. *Alptekin (1978)* solved the focal mechanisms of damaging earthquakes that have occurred in the Aegean-Anatolian region from an analysis of P-wave first motions on the long-period seismograms of the worldwide-standardized seismographs (WWSSN). *Alkan (1979)* made focal mechanism solutions for southwestern Turkey. *Le Pichon & Angelier (1979)*, using focal mechanisms of shallow earthquakes occurring along the SAA and the extent of the intermediate seismic belt, made a quantitative estimate of the relative motion between the two plates. The pattern of seismicity and focal mechanism solutions of earthquakes are used to outline the tectonic features of southern Turkey by *Rotstein & Kafka (1982)*. For central Greece, *McKenzie and Jackson (1983)* describe the basic geometry of the deformation by using seismicity data. *Jackson & McKenzie (1984a)* made focal mechanism solutions for the eastern Mediterranean. *Eyidoğan & Jackson (1985)*, by using inversion techniques, investigated source parameters of the earthquakes that occurred in Turkey. *Üçer et al. (1985)* studied the seismicity of western Turkey. They displayed clusterings in time and space. *Yoğurtçuoğlu (1986)* reviewed the focal mechanism data for the whole of Turkey. *Yılmaztürk (1986)* investigated the seismology of the Burdur area (SW Turkey) by using the NOAA catalogue. *Kalafat (1988)* studied the seismicity of southern Turkey and surrounding area. *Eyidoğan (1988)* studied slip rates in western Turkey by using focal mechanisms of the large (Ms >5.5) earthquakes that occurred since 1943. *Martin (1988)* investigated the seismicity of slab by using a careful selection of well-defined ICS data events. *Jackson & McKenzie (1988)* studied actual deformation and seismic deformation using 1901-1981 seismicity data in the Aegean-Anatolian area (Fig. 2.26). *Lyon-Cean et al. (1988)* studied the tectonics of the subduction margin by mean of seismicity data. They give information about western part of the zone. *Extröm & England (1989)* examined how much motion could be accommodated by earthquakes of Ms>6 in the period 1909-1981. They computed the strain rates and obtained the average extension rate of the Aegean region. *Üçer (1990)* studied the seismicity of the Marmara region by using data from the ISK and MARNET networks. *Papazachos (1990)* studied seismicity distribution and its relation with zones of the Aegean and Surrounding area. *Taymaz et al. (1991)* solved focal mechanisms of major earthquakes by using waveform technique for the Aegean region. They compare first motion solutions of some previous studies with their solutions. Three-component long-period seismograms from the Global Digital Seismic Network (GDSN) are used to determine the focal mechanisms of recent large (Mb>5.5) earthquakes in the SAA (*Taymaz et al., 1991*). Some events were relocated. Additional information was obtained from long-period (LP) Worldwide Standard Seismograph Network (WWSSN) records in order to improve coverage of the focal sphere (*Taymaz et al., 1991*). *Udias & Bufron (1991)* selected the focal mechanism solutions of 83 European earthquakes with M>6 from a total of 140 (1935-1983). These were used to derive the directions of the principle axes of stress along the plate boundary between the Eurasian and African plates from Azarodes islands to the Caucasus mountains. *Hatzfeld & Martin (1992)*, using ISC data, determine the geometry of the slab. *Taymaz & Price (1992)* studied the seismotectonics of the Burdur region (SW Turkey). *Papazachos et al. (1992)* made a crustal deformation analysis of the Aegean region based on their previous work that includes the seismicity and the available focal mechanisms. The components of the strain rate tensor and of the velocity tensor are presented for 26 seismic regions identified in the Aegean region by *Papazachos et al. (1992)*. They computed the strain rates and the total seismic energy release for the last century. *Papazachos & Kiratzi (1992)* used recent large earthquake data (observational period of 30 yr) to better define deformation in the region. The seismic activity in the region associated with major structures was determined by *Küleli (1992)*. Local data were obtained from the ISC database for the 23 year interval 1964-1986 recorded by 53 stations in the region (*Küleli, 1992*) (Fig. 2.27). For this region, hypocenters of local earthquakes are displayed on horizontal and vertical cross sections (*Küleli, 1992*) (Fig. 2.28). *Jackson et al. (1992, 1994)* computed a velocity field from the seismic moment tensor of earthquakes of Ms>6 in the period 1909-1983 to estimate the spatial variations in seismic strain rates in the twentieth century for the whole Aegean region. *Mueller & Kahle (1993)* use recent seismic activity in the eastern Atlantic and in the Mediterranean-Alpine region for the period 1970-1980 to describe the motions of the plate boundaries. The available focal mechanism solutions of shallow earthquakes occurring in the central part of western Anatolia were reviewed to determine the active stress directions in this region by *Zanchi & Angelier (1993)*. *Jackson et al. (1994)* compare geodetic (SLR) data with a computed velocity field from the seismic moment, for the whole Aegean region. They used earthquakes of Ms>6 in the period 1909-1983 to estimate the spatial variations in the seismic strain rates that are calculated from the horizontal field. *Pavlides & Caputo (1994)* solve the focal mechanism of moderate earthquakes in the Aegean region. A catalog of published focal mechanism solutions of earthquakes that have occurred between 1963 and 1990 in Turkey and the adjoining area has been prepared by *Özçep & Alptekin (1995)*. This includes 164 focal mechanism solutions. The reliable focal mechanism of the earthquakes (Ms < 4.2) felt in Turkey and the surrounding regions, were determined and their seismotectonic implications were discussed by *Kalafat (1995)* using first motion polarity of P waves method. In addition, only moderate and destructive earthquakes for which various investigators did focal mechanism studies were inclusively compiled by him. Waveform solutions of some 24 large earthquakes that have teleseismic records in Turkey, which occurred between 1964 and 1992, were inverted to their sources by *Pınar (1995)* to get a rupture process. *Ponderelli et al. (1995)* estimated seismic deformation in the Mediterranean area by moment tensor summation. They applied this method to a data set consisting of seismic events that had

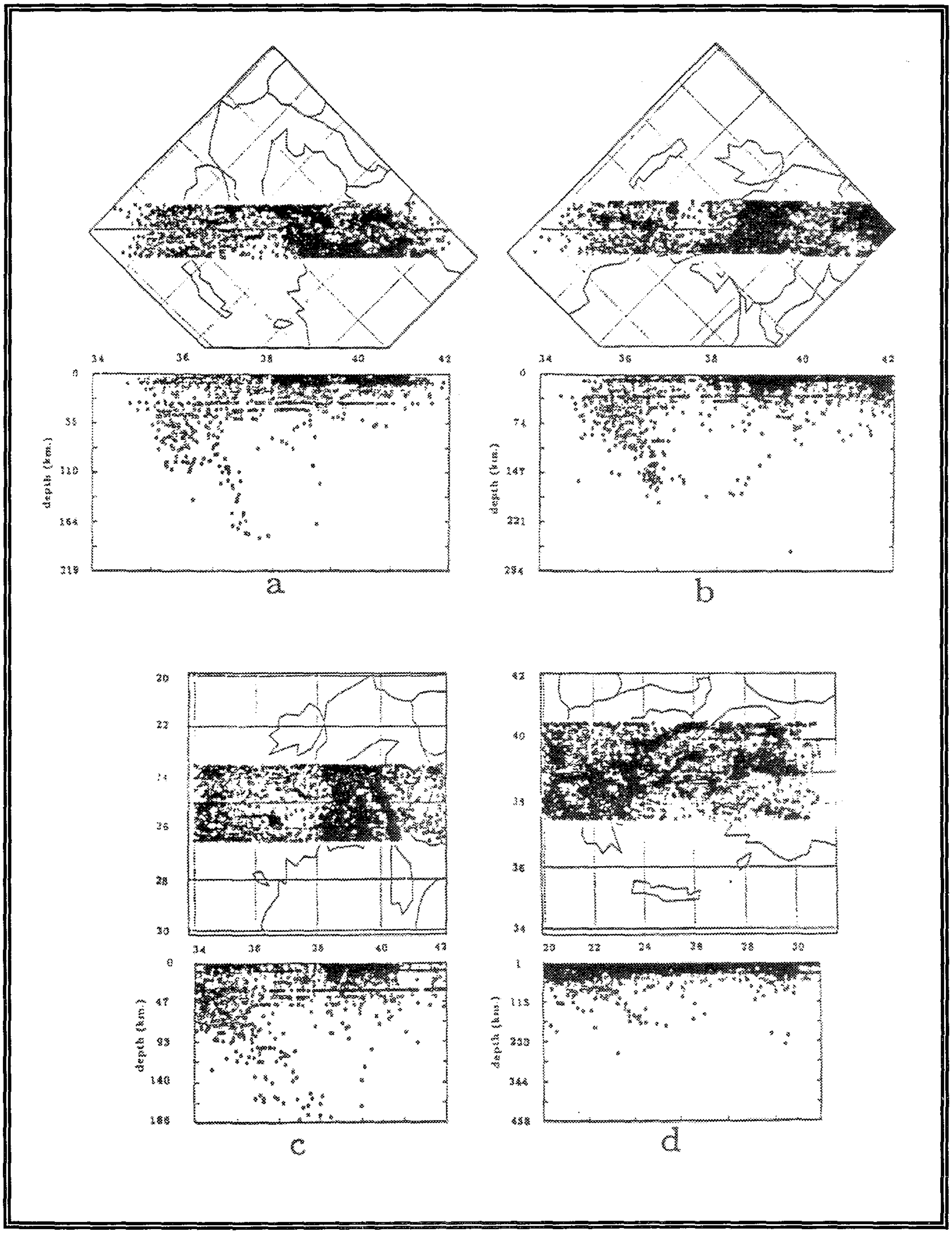


Figure. 2.27. Seismicity shown in the horizontal layers in the Aegean area (after *Küleli, 1997*). (A) The events (all magnitudes) in the 1964-1986 interval recorded by 53 station (ISC), (B) hypocenter distribution in the 0-10 km depth interval for magnitudes (mmag) greater than 3; from (A) to (E) layer thickness was selected according to those from teleseismic inversion model. (C) 0,40 km depth interval, (D) 40, 80 km depth interval, (E) 80 160 km depth interval, (F) deeper than 160 km.

occurred between 1908 and 1992. They use the Centroid Moment Tensor (CMT) Catalog prepared at Harvard University and older data from *Jackson & McKenzie (1988)*. They determine the seismic deformation and compare it to estimates of the overall deformation as obtained from global plate motion and geological studies. *Papazachos et al. (1996)* made a crustal deformation analysis of the Aegean region based on their previous work on the seismicity and available focal mechanisms. The components of the strain rate tensor and of

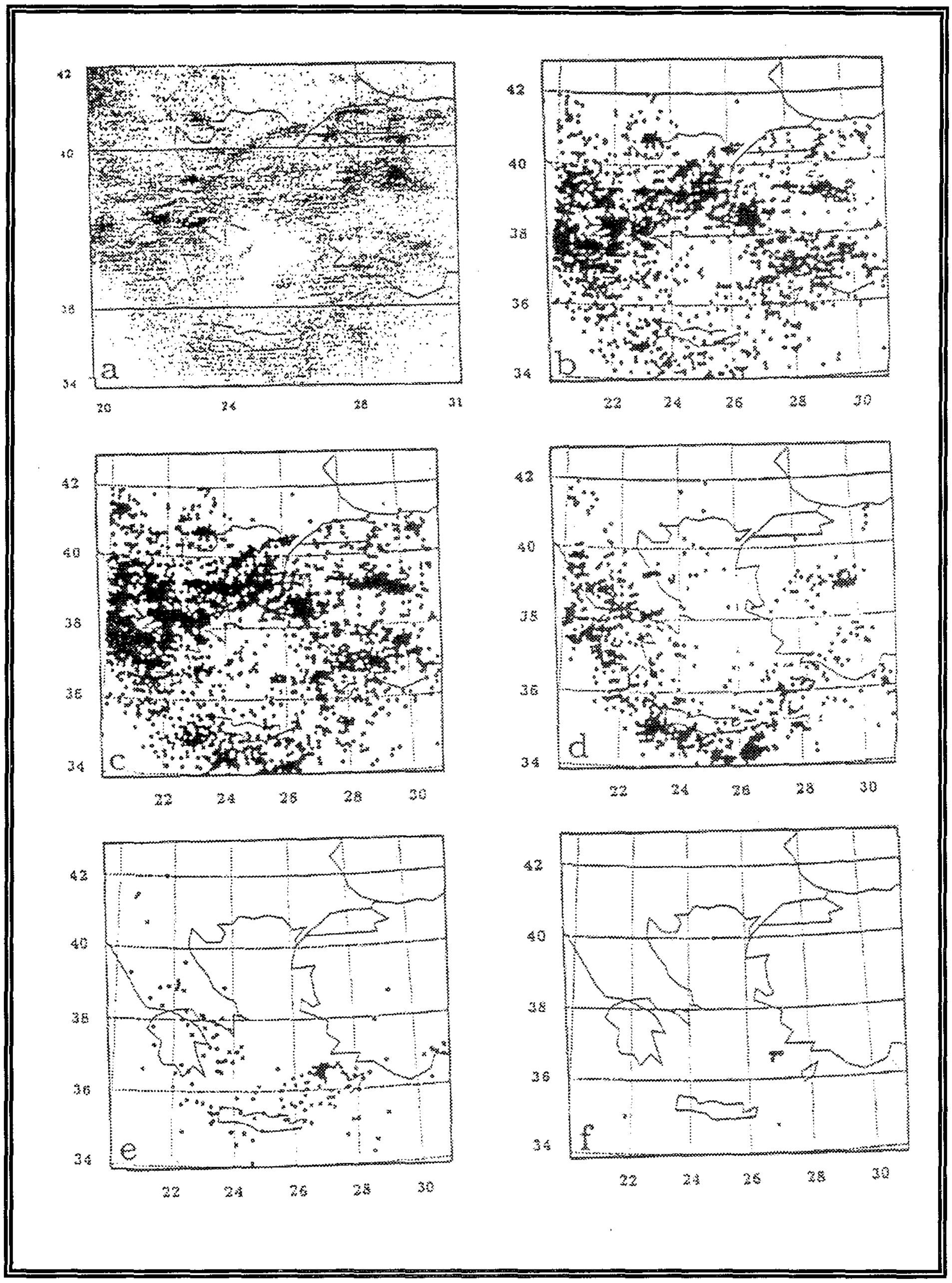


Figure. 2.28. Vertical cross section of the local hypocenters from (A) to (D) (after *Küleli, 1997*). Epicenters bounded on the upper panel for display of a small location map for orientation and latitude and longitude of the epicenters with an angle 90° on the marked line. Hypocenters and epicenters of events from ISC catalogue tape. Hypocenters plotted include all magnitude events.

the velocity tensor are presented for 63 seismic regions identified in the Aegean region by *Papazachos et al. (1996)*. *Dadaşbilge (1997)* displayed the clusterings, locations of maximum magnitudes and energy released per year between the years 1975-1996 in the Aegean region. In the Marmara region *Üçer et al. (1997)* investigated the patterns of the earthquake distribution in time and space for the period of 1976-1995. They also examined the patterns of seismicity for longer periods, including historical (pre-1900) and instrumental (1900-1975) earthquake records for the Marmara region. *Utku (1997)*, using a linear moment tensor inversion method, studied some of the earthquakes from western Turkey.

There are many other studies for the region, which I merely note (e.g., *Delibasis & Galanopoutos, 1965; Hodgsom & Wickens, 1965; Galanopoulos, 1967; Wickens and Hodgson, 1967; Barazangi & Dorman, 1969; Karnik, 1969, 1972; Papazachos & Comninakis, 1971; Papazachos & Comninakis, 1977; Myrianthis, 1984; Scordilis et al., 1985; Caputo, 1990; Ergin & Aktar, 1990*). Available focal mechanism solutions of earthquakes (Ms≥5.5) are displayed in Table. 2.5, giving strike, deep, rake, coordinate, reference and solution method information. In this Table one can see available focal mechanism solutions for a specific earthquake (see Appendix 1).

2.8.2. General Interpretation

It is generally believed that earthquakes can nucleate in continental crust at temperatures of less than 300 °C (*Chen & Molar, 1983; Alptekin et al., 1990*). For this reason the heat structure and crustal and upper mantle structure could be defined by considering the maximum thickness of the seismogenic (brittle) layer, called the upper crust.

The epicenters and fault mechanism solutions are closely associated with major structures in the crust (e.g., *Canıtez & Üçer, 1967; McKenzie, 1972, 1978; Jackson & McKenzie, 1984b; Şengör et al., 1985; Şengör, 1987; Küleli, 1992*), or between the plates. Determination of the main seismic zones in the Aegean region and surrounding area has been made by plotting epicenters of shallow and intermediate depth earthquakes (Figs. 2.27, 2.28).

The relative movement of plates gives rise to earthquakes. The slip on the fault that generates the earthquake is determined by the direction of relative plate motion, and the relative plate velocity controls the average displacement rate. Mechanisms of shallow earthquakes with sufficient magnitude can be used to drive the motion of a plate relative to the surrounding plates. Since the empirical evidence, combined with theoretical considerations, led seismologists to conclude that the 'double-couple' (plane shear stress) system of earthquake equivalent forces is the most appropriate model for the most earthquakes, first-motion studies have usually been interpreted in terms of nodal plane solutions. Some basic discoveries were made by studying the signs of the first arrivals of body waves (e.g., *Sykes, 1967; Isacks et al., 1968*). Further, one seismologist could determine some aspects of the rupture process by waveform modeling using some very broadband P and SH waveforms recorded by global networks (*Ekström and England., 1989*). However, small earthquakes could not modeled teleseismically because of insufficient network of stations. The mechanism of earthquakes determines the direction, but not the rate, of motion between plates (*McKenzie, 1972*). *Brune (1968)* has overcome this problem by using the cumulative scalar moment (M_O) (seismic moment tensor) of earthquakes of a given plate boundary to estimate the total slip which occurred over several decades. Nevertheless, they are obviously correct when the plate boundary consists of a single fault. Most earthquakes happen on preexisting fault planes of weakness, and slips can occur at different angles relative to the principle axes. However, the stress axes that are derived from focal mechanism solutions of large earthquakes may appear as an indication of their general trend for a given region (*Udias & Buforn, 1991*). The foci of the large shocks are better determined because of the better distribution of the recording stations and the larger macroseismic effects. The energy released during microearthquake experiments is small compared to that from earthquakes with a large (>5.5) magnitude (*Hatzfeld et al., 1990*).

After the WWSSN long period network was installed in 1961, focal mechanisms were determined more accurately. But there is always an ambiguity about which plane among two possible nodal planes is the fault plane and which is the auxiliary plane. To avoid the wrong choice, seismologists consider other evidence (see also *Taymaz et al., 1991*):

1. Some of the earthquakes were associated with surface faulting, and the choice of nodal plane was thus straightforward;

2. Some of the offshore events occurred in, or were adjacent to, basins whose clear asymmetry suggests the polarity of the faulting and hence which nodal plane is likely to be the fault plane;

3. The strongest evidence supporting a fault mechanism is the spatial distribution (elongation) of the aftershock zones (locations) of events (*McKenzie, 1972*);

4. Based on the similarity of the dipping slip vector with those in the nearby events;

5. Based on the known dominant tectonic regime.

Much of the deformation must be aseismic in the Aegean region (*Jackson & McKenzie, 1988;* see also *Jackson, 1993*) (Fig. 2.26). Moreover, many new active structures were effectively aseismic for the last hundred years. *Straub & Kahle (1997)* calculate that earthquakes release about 60% of the deformation derived from the geodetic data. Recent investigations of seismic strain rates indicated that they account for 74% of regional SLR rates in NW Turkey (*Jackson et al., 1994*). These results clearly show that a considerable amount of aseismic creep or strain accumulation has been released (*Straub & Kahle, 1997;* see also *Jackson et al., 1992*). However, one must keep in mind that the periods, both of geodetic and seismic data collection, are very short. They might not be representative for a longer time scale (*Straub & Kahle, 1997*). But no reliable catalogue exists for the earlier instrumental times and no catalogue exists of all for the oldest times. For this reason, even though spanning a shorter time interval, the CMT Catalog often gives a better estimate of deformation geometry than the whole data set (85 yr.) (*Pondrelli et al., 1995*). Reliable fault-plane solutions exist only for recent large earthquakes (*Papazachos et al., 1992; Papazachos & Kiratzi, 1992*). *Jackson & McKenzie (1988)* used 70 years data that they calculated to be sufficiently long to be representative of the Aegean and South Aegean Trench, though possibly for NE Iran, and probably not long for the North Anatolian Fault (NAF). Further progress in understanding the geometry and rates of deformation requires additional information, not contained in the seismic moment tensors (*Jackson & McKenzie, 1988*). Such information is likely to come from paleomagnetic and geological observations (*Jackson & McKenzie, 1988* see also *Jackson, 1994*). It is possible to extend the deformation period back in time, by assuming the fault parameters of the past events, and this has been done by some researchers (*Jackson & McKenzie, 1988; Ambraseys & Jackson, 1990; Papazachos et al., 1990; Kiratzi, 1991; Taymaz et al., 1991; Papazachos & Kiratzi, 1992*). No matter how well justified these assumptions are, it is very likely that they introduce considerable error and bias (*Papazachos & Kiratzi, 1992*). The use of all available complete data, which include information on smaller recent shocks and on strong instrumental and historical earthquakes of a much longer period reliable field observations of past strong earthquakes are needed (*Papazachos & Kiratzi, 1992*). The contribution of smaller earthquakes, with Ms <5.8, will increase displacements by around 50% (*Ambraseys & Jackson, 1990; Papazachos & Kiratzi, 1992*). However, according to *Lyon-Caen et al. (1988)*, it is common for small earthquakes to have mechanisms incompatible with a uniform regional strain field principally because they reflect stresses and deformation within the lithosphere, rather than motions between Plates (see *McKenzie, 1972; Taymaz et al., 1991*).

Seismologists try to improve existing data. By modeling of the long-period body-waves *Eyidoğan & Jackson (1985)* determined more accurate focal depths of some earthquakes. The faults inferred by using macroseismic data (*Ketin & Abdüsselamoğlu, 1969; Ambraseys & Tchalenko, 1972; Ambraseys, 1985*) and modeling of the long-period body-waves (*Eyidoğan & Jackson, 1985*) indicate that the major earthquakes in the region are not deeper than 10-15 km (see also *Crampin & Üçer, 1975; Eyidoğan, 1988; Jackson & McKenzie, 1988; Ekström & England, 1989; Taymaz et al., 1991*). The depth of most events in the Bursa area was found to be less than 15km by micro-earthquakes (*Sellami et al., 1997*). They confirm NE-SW to almost N-S oriented crustal extension and E-W dextral shear for the northern part of the area. The focal depths of the earthquakes in western Turkey range from 0 to 50 km according to various bulletins and sources, based on arrival times alone or with those based on crude empirical relationships, but these reports are not reliable (*Eyidoğan, 1988*). The depth control is poorer for all magnitudes (see for detail *McKenzie, 1978*). The hypocenters of very poorly determined events are fixed at 33 km or 10 km depth, which explains the peculiar concentration of events at these levels (*Meulenkamp et al., 1988*) (Fig. 2.27). Small shocks (Mb < 5.0) are much less accurately located. Recent locations from a local network in Greece have shown that several shocks located by NOAA at depths below 70 km are in fact much shallower (*McKenzie, 1978*). *Papadopoulos et al., (1988)* find 60 km misallocation of epicenters along the subduction zone perpendicular to the arc. Given the proximity of the Greek National Network, epicenter misallocation in western SAA appears to be larger than one might have expected (*Papadopoulos et al., 1988; Meulenkamp et al., 1988*). The cause for the misallocation is demonstrated to be the presence of a high-velocity subducted slab (*Engdahl et al., 1982; Frohlich et al., 1982; Papadopoulos et al., 1988*). Seismic rays, which leave the earthquake sources heading north to teleseismic distances, travel in part through the high-velocity slab of the subducted plate. For this reason, these rays will arrive earlier than expected (*Papadopoulos et al., 1988*). Although *Makropoulos & Burton (1981)* have made an effort to improve hypocenter quality in the SAA, calculated locations of earthquakes could further be improved in several ways by *Papadopoulos et al. (1988)*. *Papadopoulos et al. (1988)* made this improvement, based on their results for the SW SAA, or based on relocations using other temporary networks at other parts of the plate boundary. Another means of improving the locations could come from accurate arrival times of North African stations *(Papadopoulos et al., 1988).*

Velocity models that are used at the hypocenter affect the solutions. *McKenzie (1978)* criticized the solutions of *Alptekin (1973)* in this respect. According to *McKenzie (1978)*, and *Alptekin (1973)* probably used too low a velocity at the hypocenter. For these reason his solutions have a much greater component of strike slip motion than the solutions of *McKenzie (1978).*

All three main types of stress distributions (strike-slip, extensional and compressional) and their variations dominate the Mediterranean-Alpine area (see e.g., *Udias, 1980*). But all of them are manifestations of the compressional stress regime governing the diffuse zone of plate contact between the Eurasian and African plates giving an immediate impression of the present state of stress and deformation in the lithosphere (*Mueller & Kahle, 1993*). However, most of the focal mechanisms of the earthquakes along the entire edge of the African plate indicate that N thrusting is the dominant mode of seismic deformation between the Eurasian and African plates (see *Rotstein & Kafka, 1982*).

The Aegean region is one of the most seismically active in the world (e.g., *McKenzie, 1972, 1978; Mercier et al., 1977; Jackson et al., 1982; Armijo et al., 1996*). The region that has continental crust is a part of African-Eurasian boundary, situated north of the northern edge of the Africa plate, and it deforms diffusely (Fig. 2.27). The scatter in the Aegean region is not due to errors in the location of epicenters, which are usually less than 30 km. The intense seismicity and continental deformation attracts a lot of earth scientists' attention. Shocks of large magnitude have occurred in limited zones along known major fractures within the region (Fig. 5.1) (see also *Ryan et al., 1969*). The present shallow seismicity is concentrated in two diffuse belts along the South Aegean Trench and arc and in the northern Aegean (*Galanopoulos, 1967; Caputo et al., 1970; Ryan et al., 1969; McKenzie, 1970, 1972; Papazachos & Comninakis, 1971; Agarwal et al., 1976*) (Fig. 2.27). Large magnitudes (Ms $\geq$ 5.5) recorded in the region have been observed west of North Aegean Trough (NAT) and NAF (Fig. 5.1). The earthquakes recorded around the subduction area in the Cretan basin and Crete, all show similar large magnitudes (Ms $\geq$ 5.5) (Fig. 5.1). The SAT and island arc are seismic regions that contain a broad belt in which seismic activity has occurred at both shallow and intermediate depths (Fig. 2.27). Graben-like basins in western Turkey and Greece are also dominated by high seismicity and large earthquakes (5,5≤Ms) (*Ambraseys & Tchalenko, 1970; Abdüsselamoğlu, 1977; Kocaefe & Ataman, 1982; Akçığ, 1988; Eyidoğan, 1988; Alptekin et al., 1990*) (Fig. 5.1). The other grabens in which large earthquakes have occurred are Corinth and Edremit.

Caputo et al. (1970) and *Papazachos & Comninakis (1971)* provided the first evidence about the subduction using the hypocenter distribution of local earthquakes (Fig. 2.28). The SAT elongation could be identified on a seismicity map as a curved shape (Fig. 2.27). Most of the seismicity stops abruptly about 40 km from the trench. The highest concentration is seen in the western part of the Peloponnesus *(Leydecker et al. 1978)*. Some clusters are located where there are changes in the morphology of the SAT (*Hatzfeld et al., 1989*). The distribution of macroseismic intensities of the earthquakes at intermediate focal depth just behind of the Ionian Trench in Greece is very asymmetrical and strikes parallel to the trench (*Papazachos & Comninakis, 1971*). The same phenomenon is also observed in the some shocks of normal focal depth, but it is not so characteristic as it is to inter-mediate earthquakes *(Papazachos & Comninakis, 1971*). *Barış et al. (1998)*, using an inversion technique, modeled subevents of recent large earthquakes around the Ionian trench. The trend was also investigated parallel to the trench (*Barış et al., 1998*). The 100 km and 150 km isodepth contours of hypocetres of earthquakes obtained from ISC catalogue are not parallel to the South Aegean Trench (*Hatzfeld & Martin, 1992*) (Fig. 5.5). The gently dipping part of the slab is longer in the west than in the east (Fig. 5.5) (*Hatzfeld & Martin, 1992* see also *Papazachos, 1990*).

Reverse faulting is observed from focal mechanisms of earthquakes along the South Aegean Consuming Boundary (e.g., *McKenzie, 1972, 1978*). Published focal mechanisms show that reverse faulting is also observed all along the South Aegean Trench (*Hatzfeld et al., 1990*) (Fig. 5.3). On the other hand, in the internal part of the arc normal faulting is obvious (*Hatzfeld et al., 1990*) (Fig. 5.3). The transition between reverse faulting and normal faulting is rather sharp, but with some overlap of the two families (*Hatzfeld et al., 1990*). In the western part a very shallow dipping aseismic zone, found from ISC catalogue data, extends 200 km NE from the trench and then steepens abruptly at depths of about 50-80 km, eventually reaching dips of 45° (*Hatzfeld & Martin, 1992*). Intermediate microseismicity determined by *Hatzfeld et al. (1989)* also defines a subducted slab dipping gently (10°) towards the northeast for the first 200 km and dipping more steeply (45°) beneath the Gulf of Argols. These results are consistent with the seismicity obtained by a careful selection of well defined events (ISC data) by *Martin (1988).* T-axes of focal mechanism solutions for the deeper events are consistent with the pulling of the cold lithosphere probably towards the north and towards the edge of the slab. They indicate N-S extension similar to the shallow earthquakes above (*Hatzfeld et al., 1989*). Thus this extension is observed on the slab and not only within the overriding crust (*Hatzfeld et al., 1989*).

In the Aegean region a N-S extension is characteristic from focal mechanism solutions. Shallow seismicity is also spread over most of the Aegean area with concentration in some places, such as the Gulf of Corinth and the western Peloponnese (*Hatzfeld et al., 1990*). Several NW-SE trending grabens accommodate NE-SW extension south of the Aegean axial trough to the SAA. However no clear individual faults can be mapped from the seismicity, even in the Gulf of Corinth (*Hatzfeld et al., 1990*). But many large earthquakes are located around the Gulf of Corinth (Fig. 5.1). The focal mechanisms show a consistent pattern of E-W trending normal faulting. This pattern is consistent with microseismic observations obtained by dense temporary networks (*Hatzfeld et al., 1990; Rigo et al., 1995*). Results support a listric type fault dipping to the N.

The most obvious problem in accounting for the observed deformation is the relationship between the thrusting and normal faulting in the northwestern part of the area beneath Albania (*McKenzie, 1978*). Since the thrust and normal faults have similar strikes and dips there is almost no overlap of the dilatational quadrants of the focal mechanism solutions (*McKenzie, 1978*).

Dextral strike-slip faults are obvious in the NAF and NAT in the northern Aegean (*Papazachos & Kiratzi, 1996*). Clear SW-NE striking earthquake lineaments can be traced across the northern Aegean Sea (Figs. 2.27, 5.1). The three earthquakes located to the northwest of the Marmara sea display major strike-slip mechanisms (*Eyidoğan et al., 1998*). Fault · mechanism solutions of micro earthquakes also show a strike-slip mechanism (*Eyidoğan et al., 1998*). The rest of the other micro-earthquakes show that normal fault activity is abundant in the region (*Eyidoğan et al., 1998*). The stress tensor obtained for the former period shows that the extension occurs, but that is still within a shear regime (*Eyidoğan et al., 1998*). The seismicity in western Turkey (eastern Aegean Sea) for large earthquakes (M > 4.5) is considerably lower than that in the northern part of the Aegean. The orientation of the stress tensor calculated the earthquakes during the 1995 microearthquake experiment shows that the crustal extension in NNE or NE direction dominates in the Marmara region (*Eyidoğan et al., 1998*). Pull-apart structures in the Marmara Sea have been identified from the microearthquake epicenter distribution (*Üçer, 1990*). The focal mechanism solutions of large earthquakes and composite focal mechanism solutions of micro-earthquakes in Marmara show oblique mechanisms (*Üçer, 1990*). Slip vectors show SW orientation in the Marmara region (*Evans et al., 1985; Üçer, 1990*). The Sea of Crete and the Cyclades area are relatively aseismic (Fig. 2.27) (*Taymaz, 1996*).

The seismicity in western Turkey is high and displays swarm-type activity with remarkable clustering of low-magnitude earthquakes in time and space (Fig. 5.2) (*Üçer et al., 1985*). South of Bursa (at about 39°N) a striking seismicity feature is located. It has the shape of a NW-SE oriented active cluster extending from the small town of Sındırgı (NE of Akhisar) as far as SE of Afyon (Fig. 5.2) which is obviously related to the Simav-Afyon fault zone and not to the Bursa-Eskişehir fault which has not been significantly affected by major earthquakes in the recent past (*Mueller et al., 1997*). At its western end this zone is cut by a NE-SW directed seismically moderate lineament, which goes past Mudanya (NW of Bursa) region, the Gulf of Gemlik and terminates at Yalova (Fig. 5.2) (*Mueller et al., 1997*). This activity follows the western part of the İzmir-Ankara ophiolite zone lineation (Figs. 3.1 and 5.2) between Adapazarı and İzmir (*Üçer, 1990*).

The stress regime in western Turkey is extensional with dominant normal fault motions and increasing importance of strike-slip faults from south to north (*Zanchi & Angelier, 1993*). The epicenters lie in the grabens of Gediz, Büyük Menderes and Küçük Menderes valleys in western Anatolia, and in the fault zones of the Gulf of İskenderun and Amok Plain in the southeast (*Ketin, 1969*). The dips of at least some of the normal faults in western Turkey become shallower at depth and are called listic faults. The best evidence comes from the 1970.3.28 earthquake whose fault break was mapped by *Ambraseys & Tchalenko (1972)* (see *McKenzie, 1978*).

Zanchi & Angelier (1993) compared stress orientation reconstructed using crustal earthquake mechanisms at different depths (0-20 km and 20-40 km). Trends of extension direction are N-S near the surface and NE-SW (N40°E) with depth. Moreover, the percentage of strike-slip faults increases at depth (*Zanchi & Angelier, 1993*). It looks as if a torque of the stress field was active in the crust beneath western Turkey (*Zanchi & Angelier, 1993*). They interpret this change of the average stress pattern with depth as related to the behavior of pre-existing discontinuities in rock mass, which play a more important role near the surface due to the decrease in lithostatic component of stress. It is well known that a large number of faults in western Anatolia is inherited from older faults with various strikes (*Arpat & Bingöl, 1969; Kaya, 1981; Zanchi et al., 1990*). This is especially the case for large E-W trending faults inherited from the Late Cenozoic extension (*Zanchi & Angelier, 1993*). However, oblique reactivation of such major faults may be difficult in the lower crust (*Zanchi & Angelier, 1993*). In contrast, superficial faulting and the deeper shallow earthquakes give a better account of the regional stress regime, because accommodation is easier at the surface, and confining pressure increases at depth (*Zanchi & Angelier, 1993*). *Pınar (in press)* identified subevents of the Dinar earthquake and found the same difference between subevents of the same earthquake. The results agree with those of *Zanchi & Angelier (1993)*.

The focal mechanisms give the impression that the north and central Aegean Sea is dominated by distributed strike-slip faulting, most of it right-lateral with a NE to ENE strike (Fig. 5.3) (*Taymaz et al., 1991*). The parallel troughs in the east Aegean ends abruptly in front of the South Aegean Axis Line. In front of the NW-SE trending South Aegean Axis Line, NE deeping nodal planes of normal faulting is obvious. For example, the 1967.3.4 earthquake probably represents the Skyros basin NE deeping escarpment (*Taymaz et al., 1991*). Like the NAT, the Edremit trough is also the site of strike-slip faulting, with mechanisms consistent with right-lateral slip on NE to ENE trending nodal planes (*Taymaz et al., 1991*). The aftershock zone following the earthquakes of 19 and 27 December 1981 suggests that the NE-striking nodal planes were the fault planes (*Taymaz et al., 1991*). *Saatçılar et al. (1996)* who studied seismic reflection profiles in the Aegean did not observe any pure shear (strike-slip) evidence beneath the eastern Aegean Sea. Their results are consistent with those of *Pavlides & Caputo (1994)*, who solved focal mechanisms of earthquakes by using local stations. They defined NAT by strike-slip faulting.

Table. 2.5. Focal mechanism solutions of earthquakes (5.5 ≤ Ms) from many studies in the Aegean and surrounding area. All the mechanisms are equal-area projections of lower hemisphere of the focal sphere. On top of each beach-ball the two letters defines authors two first letter of last names; and two-digit date of publication.

Am90	Ambraseys and Jackson, 1990
An87	Anderson et al., 1987
Ba97	Baker et al., 1997
Be97	Bernard et al., 1997
Ca67	Canıtez and Üçer, 1967
Ca84	Canıtez and Büyükaşıkoğlu, 1984
Cl97	Clarke et al., 1997
Dr82	Drakopoulos et al., 1982
Ex89	Extröm et al., 1989
Ey85	Eyidoğan & Jackson 1985
Ey88	Eyidoğan, 1988
Ha96	Hatzfeld et al., 1996
Ha97	Hatzfeld et al. 1997
Ja84	Jackson et al., 1984
Ja88	Jackson et al., 1988
Ja92	Jackson et al., 1992
Kir91	Kiratzi et al., 1991
Kiy87	Kıyak 1987
Li89	Liotier, 1989
Ly88	Lyon-Cean et al., 1988
Mc72	McKenzie 1972
Mc78	McKenzie 1978
Oc60	Öcal 1960
PaB96	Papazachos Basil, 1996
PaC92	Papazachos, 1992
PaE93	Papadimitriou 1993
Pi95	Pınar 1995
Pi96	Pınar et al. 1996
Pi98	Pinar, in press
Sh72	Shirokova (1972)
Ta90	Taymaz et al. 1990
Ta91	Taymaz et al. 1991
CMT H	CMTHarvarth

The final letter gives the source of the earthquake focal mechanism:
(a) from P, SH, or Centroid Moment Tensor waveform inaersions
(b) first motions only
(c) from surface faulting
(d) a guess based on nearby faults
(e) from nearby earthquakes
(G) geodetic method

Some of the solutions used in Figure. 5.4. to derive tectonic scheme of the region. (see Appendix. 1 for more information).

Table. 2. 5. (continue)

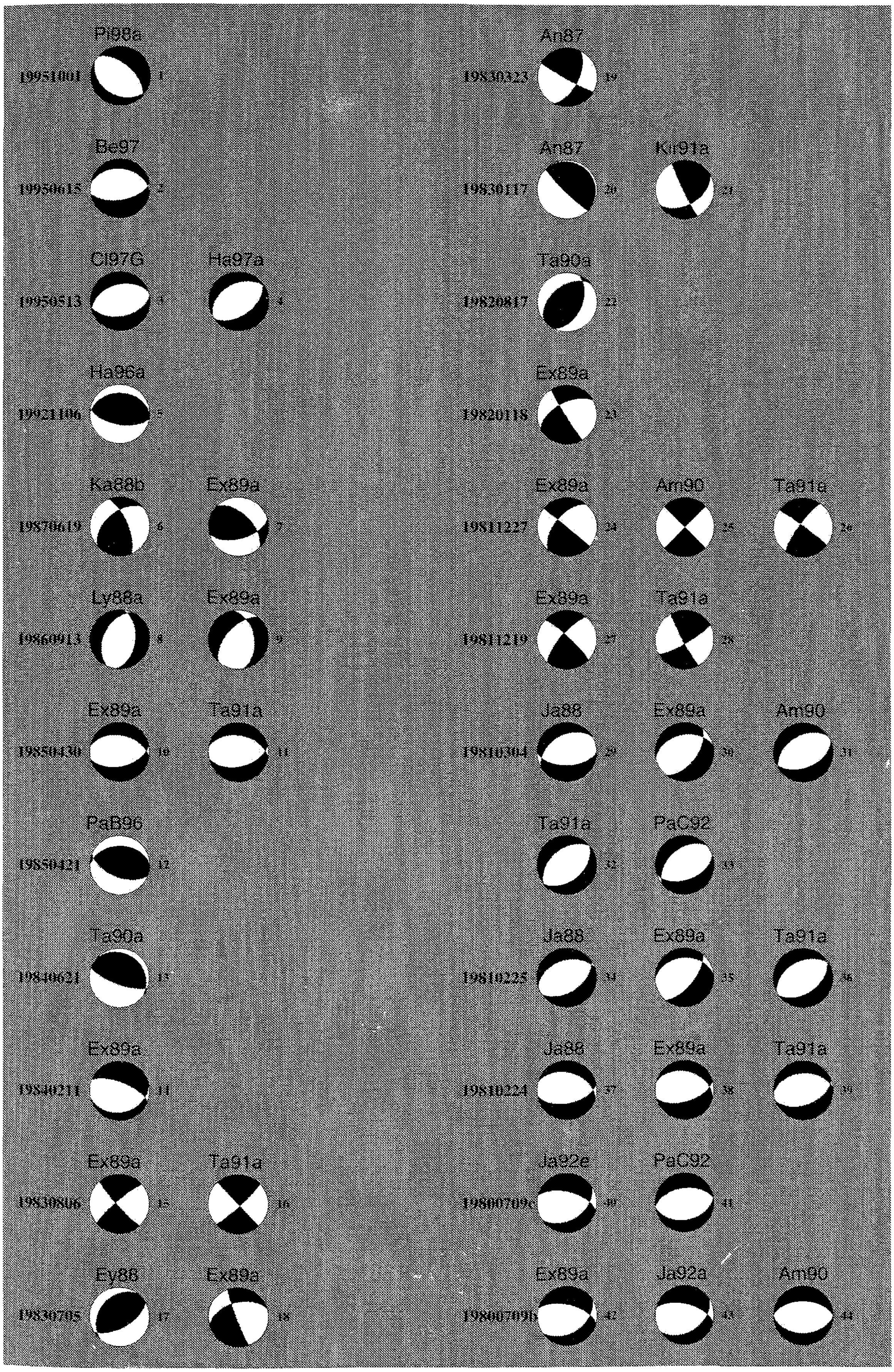
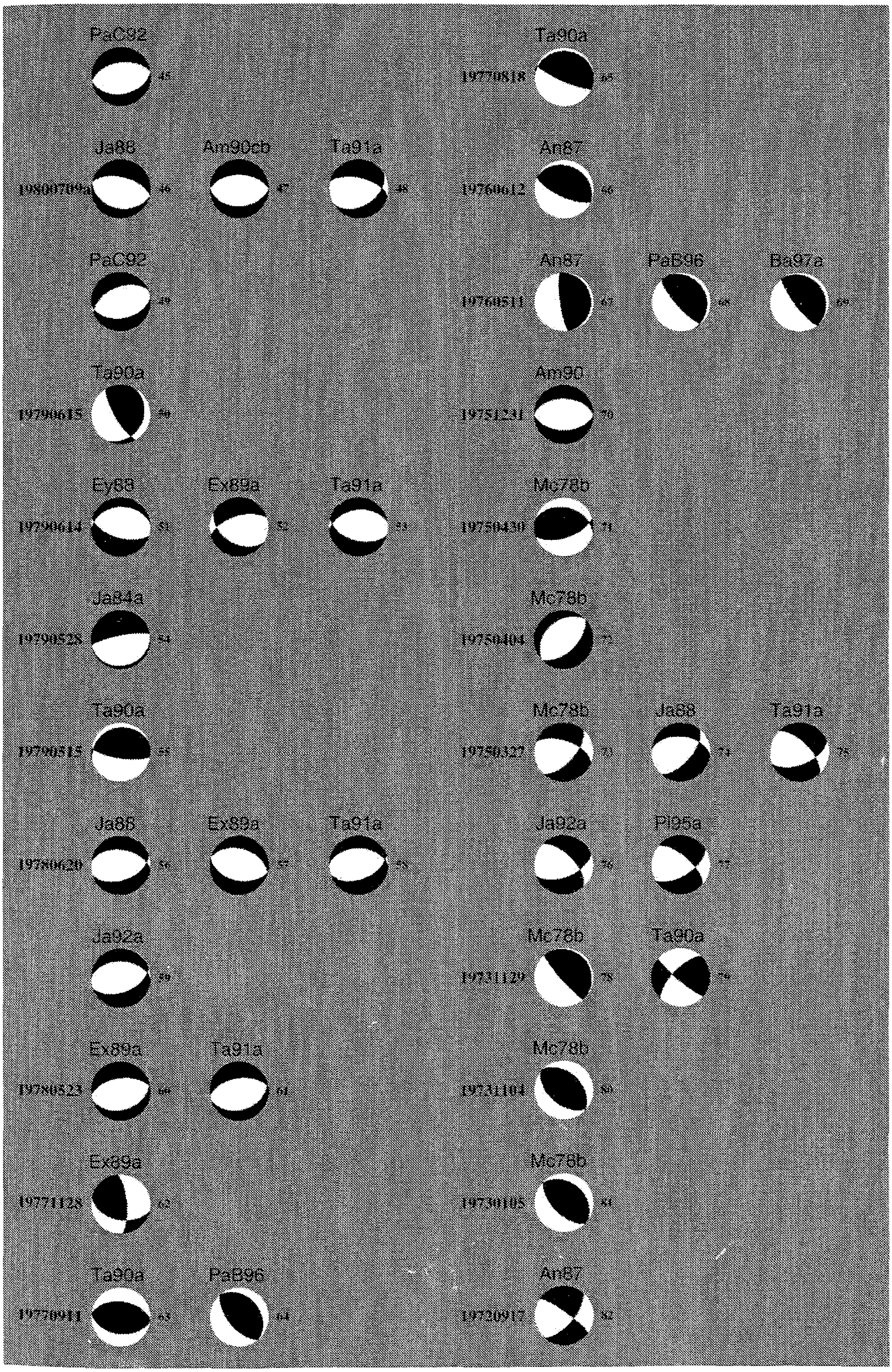


Table. 2.5. (continue)



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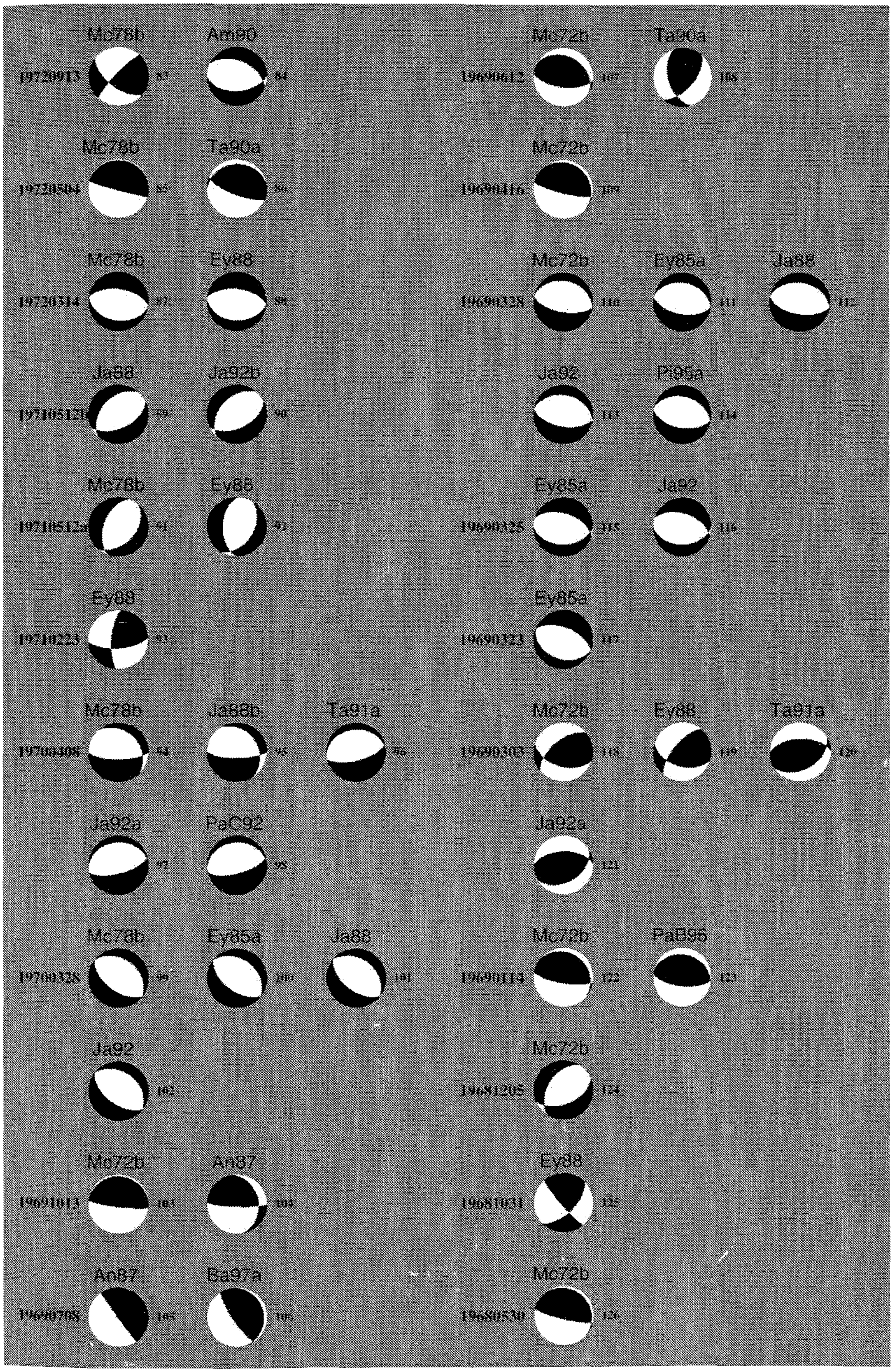


Table. 2.5. (continue)

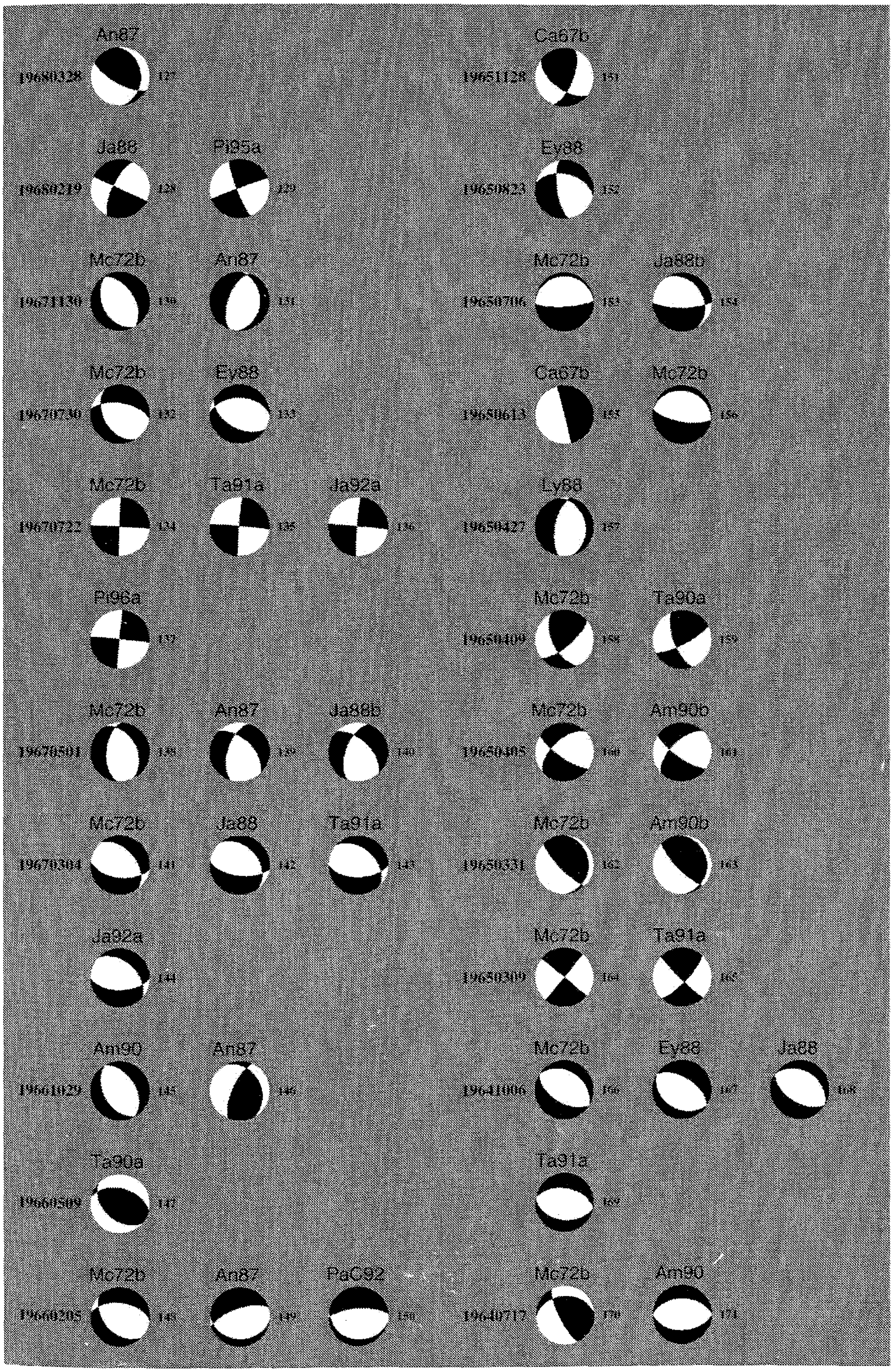
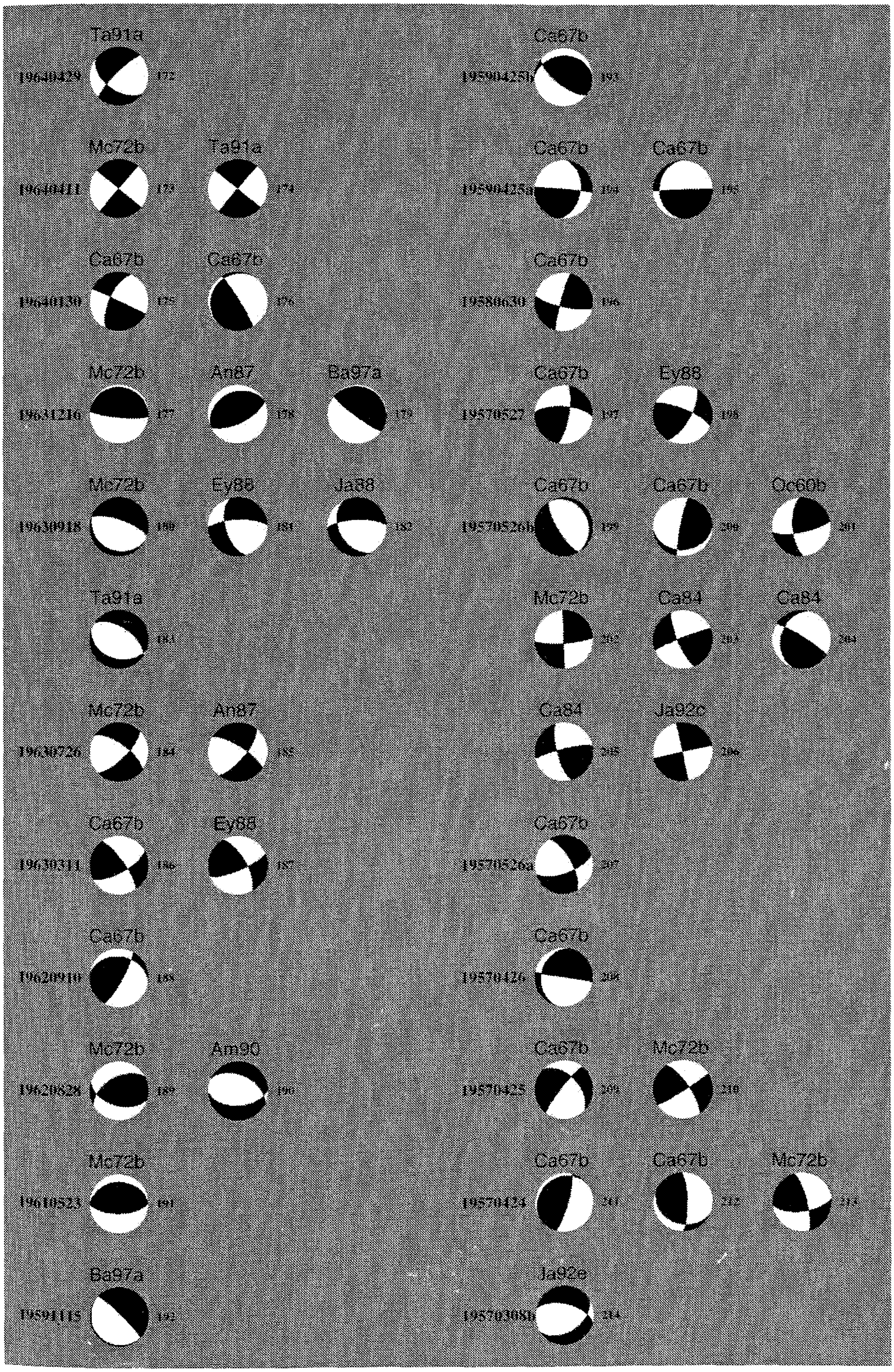


Table. 2.5. (continue)



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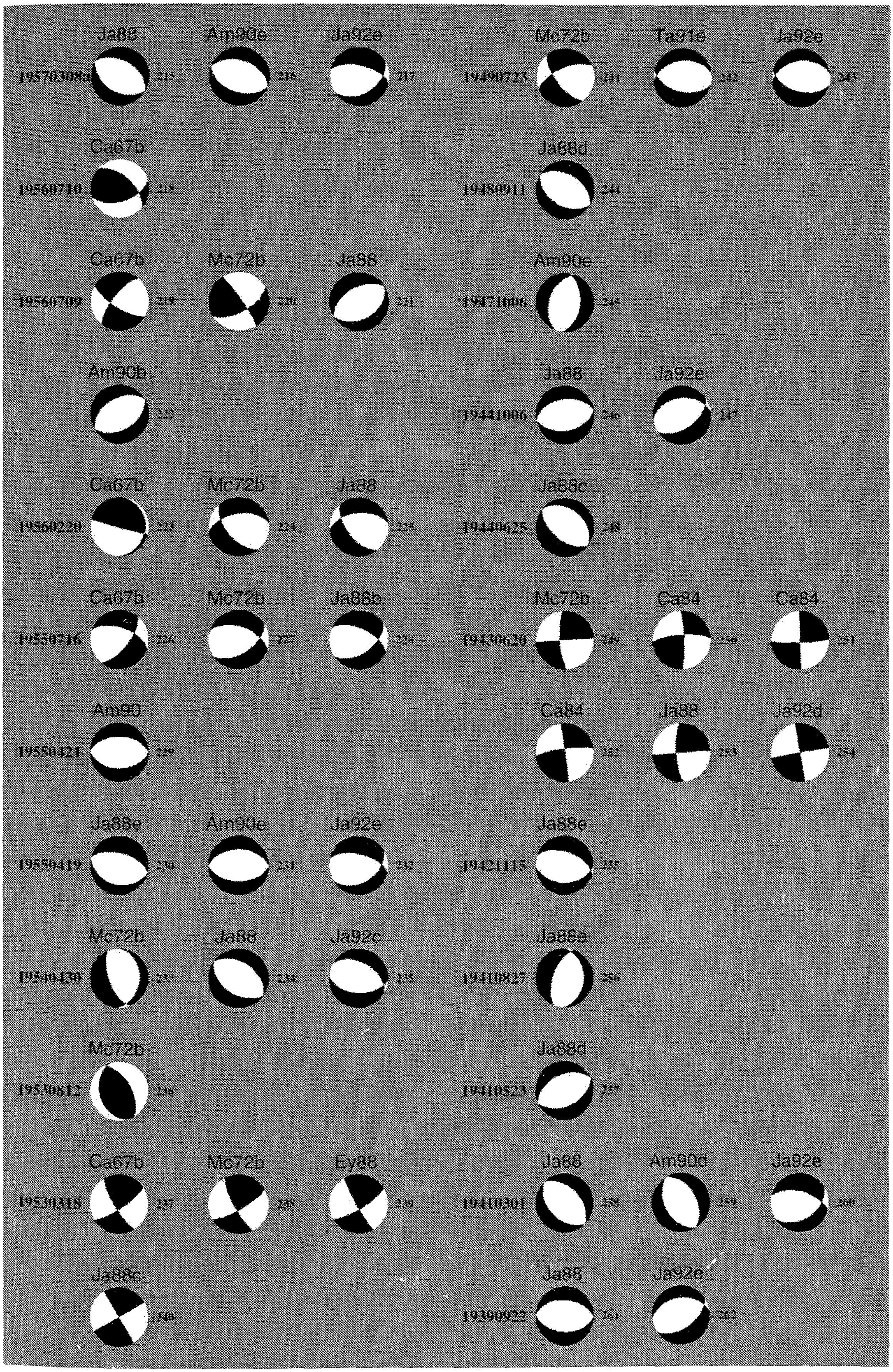

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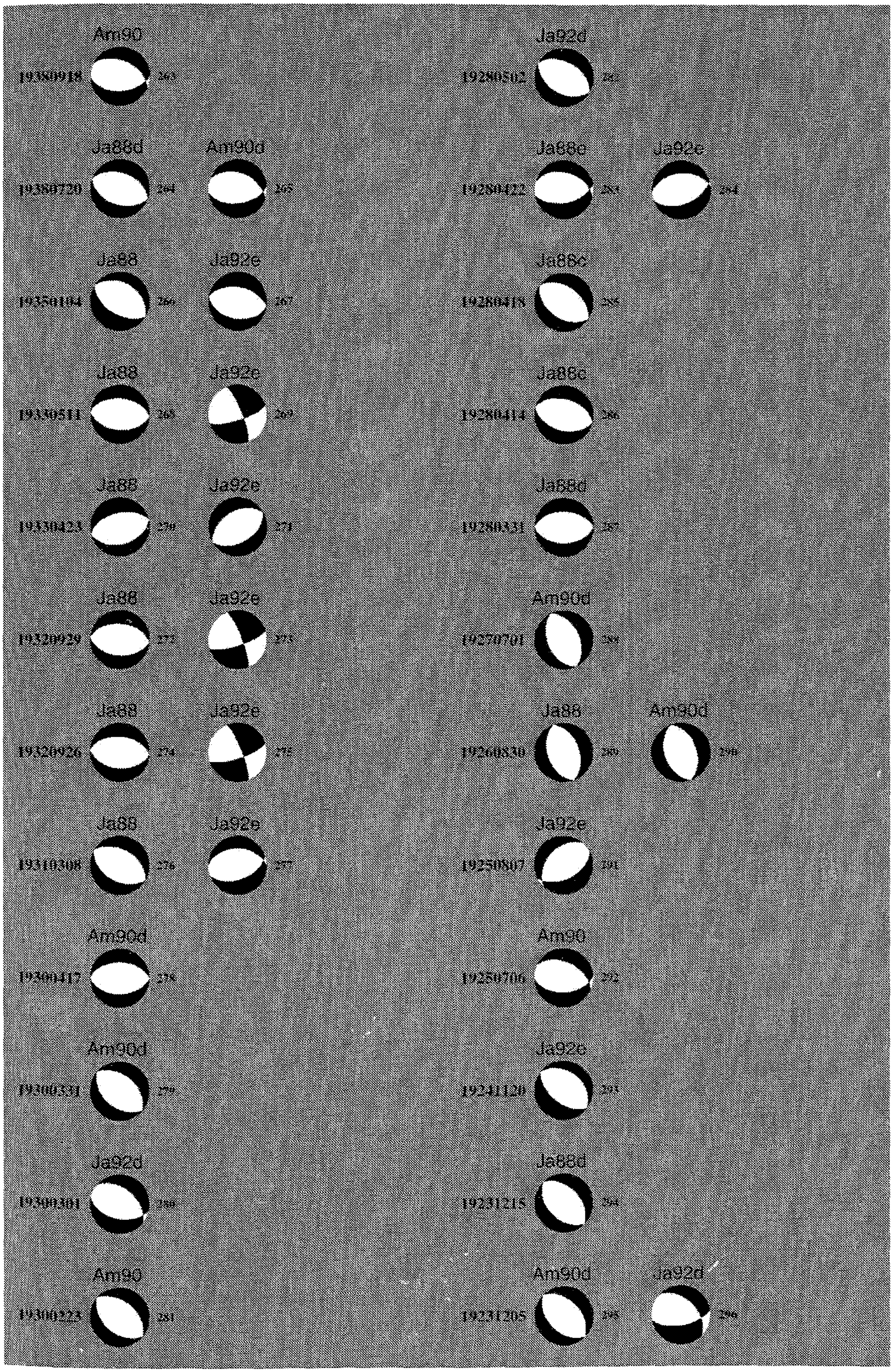

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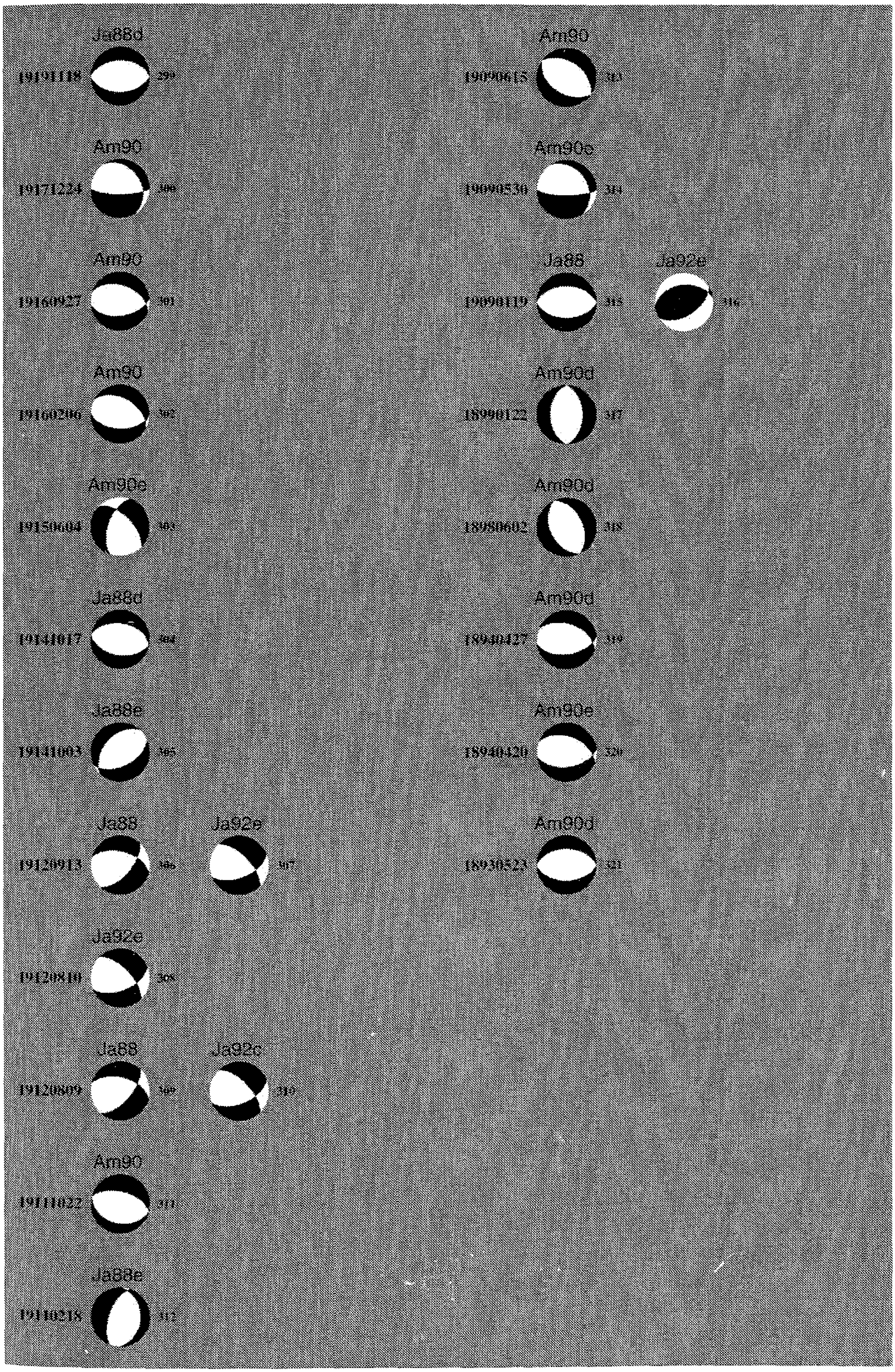

2.9. GEODETIC DATA

2.9.1. Data

Triangulation (Terresial) Network

Greece

Estimation of the velocity across the deforming zone, from the southern side of the Gulf of Corinth and the Sporades islands, comes from *Billiris et al. (1989, 1991)*, who resurveyed the 1890-1900 triangulation network (Fig. 2.29). They also compared results of this survey and a 1988 GPS survey in the region. *Curtis et al. (1997)* used the same terresial data to recover the long-term component of deformation. They compared results from a terresial network with

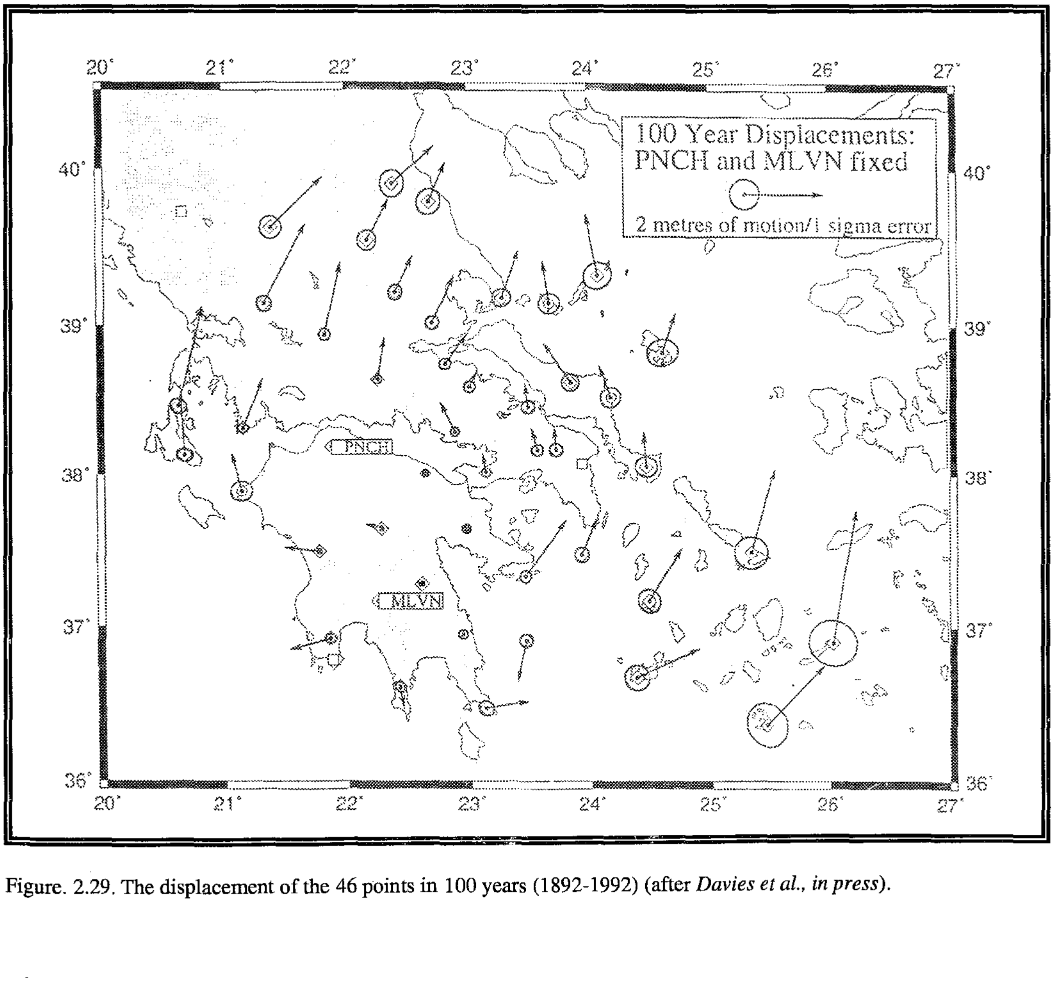

Figure. 2.29. The displacement of the 46 points in 100 years (1892-1992) (after *Davies et al., in press*).

seismicity and GPS measurements. *Davies et al. (in press)* compared seismic strain with measured strain from the terresial network. *Clarke et al. (1997)* estimated the rate of extension of the Gulf of Corinth by using same terresial geodetic data. They also compared them with strain has been released by earthquakes during this century.

Turkey

Local geodetic networks were established around the western section of the North Anatolian Fault (NAF) zone and Marmara region since 1972. The general Directorate of Mapping, İstanbul Technical University (İTÜ) Geodesy Working Group, Office of Mineral Research and Earthquake Research Office of Turkey participated in the establishment of these local networks (see for detail *Aksoy & Deniz, 1997*). After 1984, with the participation of various departments from German Universities, the Turkish national project was enlarged to the German-Turkish Project on Earthquake Prediction Research, as well as geodetic control research. Measurements of the networks, which used to study crustal movements in northwestern Turkey (*Uğur, 1974; Öztürk et al., 1987; Deniz et al., 1993; Eren, 1983*). Recently *Aksoy & Deniz (1997)* reviewed all the data of these networks and display up to date results (Fig. 2.30). Under foundation of Boğaziçi University, three micro-geodetic networks

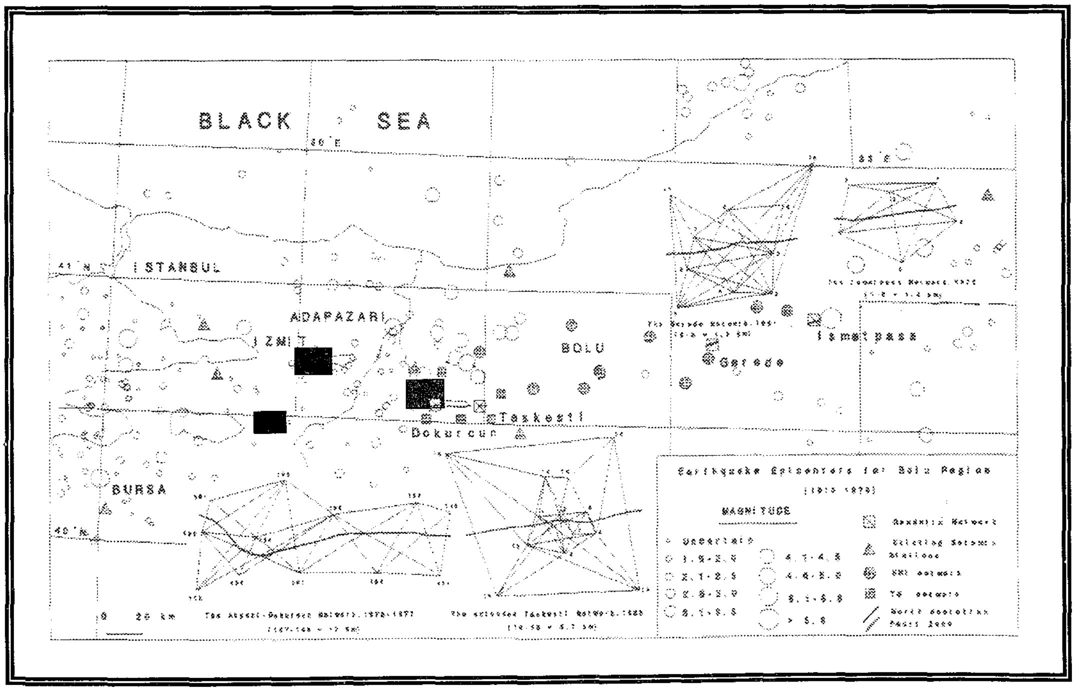

Figure. 2.30. The position of traditional geodetic networks where activities have been carried out (after *Aksoy and Deniz, 1997*) and 3 micro-networks (black filled rectangles) installed by B.U. KOERI.

were installed around the east of the Marmara Sea to detect crustal motions that could be related with NAF by B.Ü. KOERI (Boğaziçi University, Kandilli Observatory and Earthquake Research Institute) (Fig. 2.30).

SLR (Satellite Laser Ranging)

The WEGENER-MEDLAS (Working Group of European Geoscientists for the Establishment of Networks for Earthquake Research - MEDiterranean LASer ranging) Project was started in 1985 by European Institutions (*Wilson & Reinhart, 1993*). On the Eurasian plate a number of sites were occupied by mobile laser systems (SLR) during the WEGENER-MEDLAS campaigns from 1986 to 1989 (Fig. 2.31) (see *Wilson & Reinhart, 1993; Smith et al., 1994; Noomen et al., 1995a,b*). The SLR station coverage in the eastern Mediterranean provides

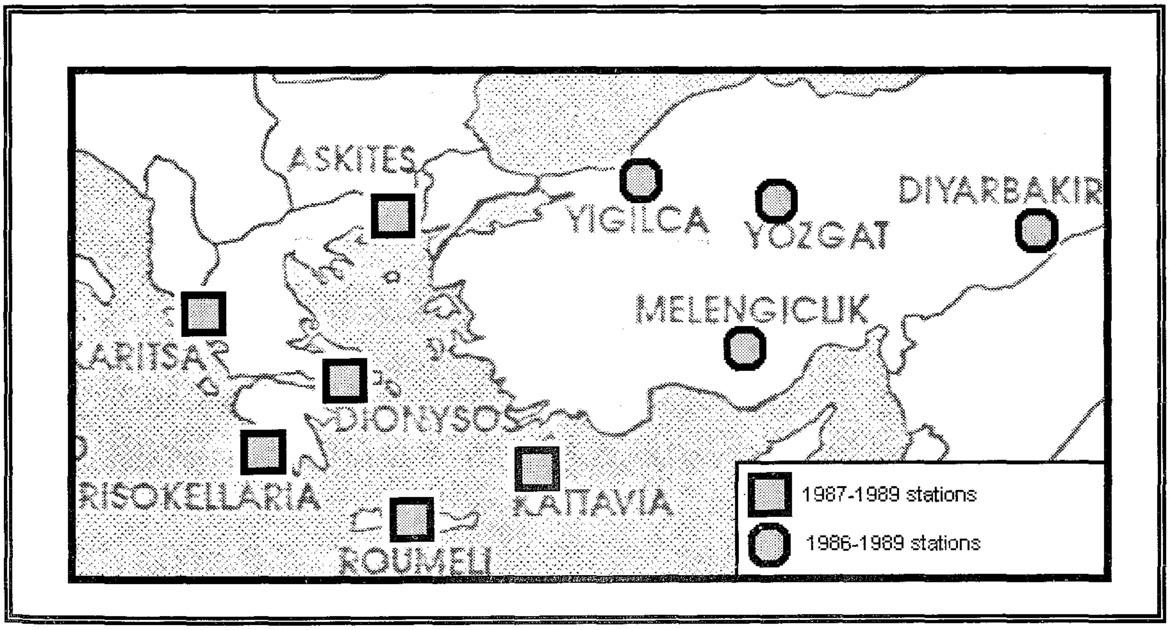

Figure. 2.31. SLR stations of WEGENER-MEDLAS network in Aegean-Anatolian area (*Wilson and Reinhart, 1993*).

some information on broad-scale plate motions (*Noomen et al., 1991, 1995a,b; Cenci et al., 1993*). In Turkey and Greece SLR measurements were taken between 1986-1989 for ten sites (*Wilson & Reinhart, 1993;* see also *Oral, 1994; Smith et al., 1994; Straub & Kahle, 1997*) (Figs. 31, 32). *Jackson et al. (1994)* compared the earthquake moment vectors with the measured velocities of SLR in the Aegean region. Since 1988, GPS experiments have been conducted simultaneously with the SLR observations in the Aegean region. *Oral et al. (1993,* *1995)* and *Oral (1994)* found consistency of the SLR measurements with GPS data for the "Anatolian plate".

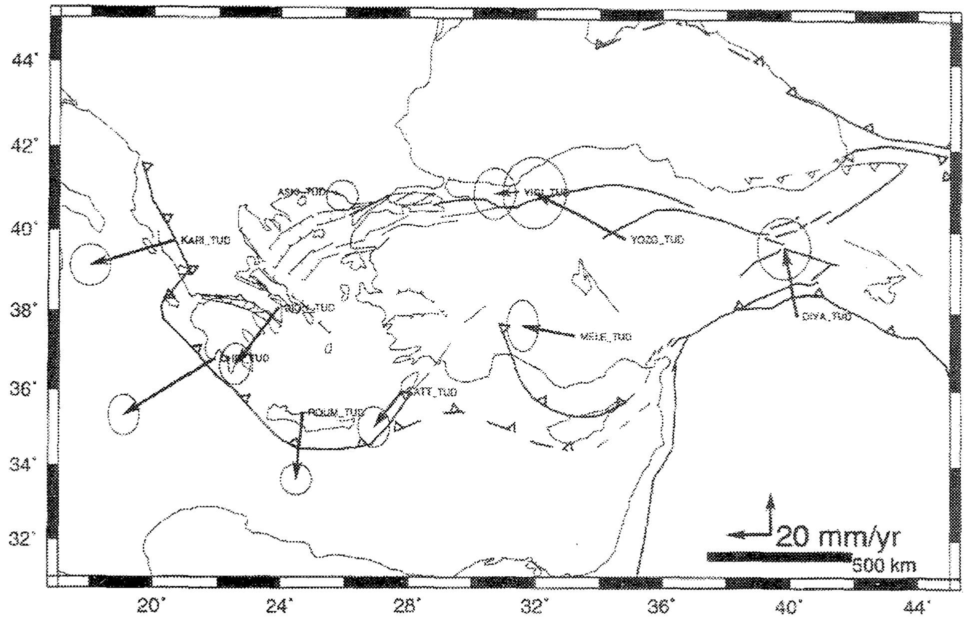

Figure. 2.32. SLR horizontal velocities and their confidence ellipses in a Eurasia-fixed reference frame for the period 1986-1990 (after *Oral, 1994*).

GPS (Global Positioning System)

Greece

A collaborative GPS-Project called West Hellenic Arc and Calaxian Arc Tectonics (WHAT A CAT) was started to determine the crustal strain rates along the western South Aegean Arc (SAA) (*Kahle et al., 1993*). Various institutions from Germany, Greece, Italy and Switzerland are conducting the project. In 1989 the first extensive GPS campaign along the western SAA was carried out by the National Technical University, Athens, Greece, and the ETH, Zurich, Switzerland. GPS stations were also connected to WEGENER-MEDLAS stations (*Kahle et al., 1993*). By using this network at two epochs (1989 and 1993) *Kahle et al. (1995)* described the strain field of southwestern mainland Greece and the Ionian islands (northwest of the western South Aegean Trench) (Fig. 2.33). A permanent network has been installed, which was designed to monitor the ongoing deformation across the KFZ

(Kephelania Fault Zone) and the SAA by *Peter et al. (in press)* (see *Mueller et al., 1997*). Relative displacement vectors for the western Aegean (Greece) were determined by *Davies et al. (1994)* using GPS receivers in 1992. For the southern Aegean region, GPS measurements have been presented by *Kastens et al. (1993, in press)* and *Gilbert et al. (1994)*.

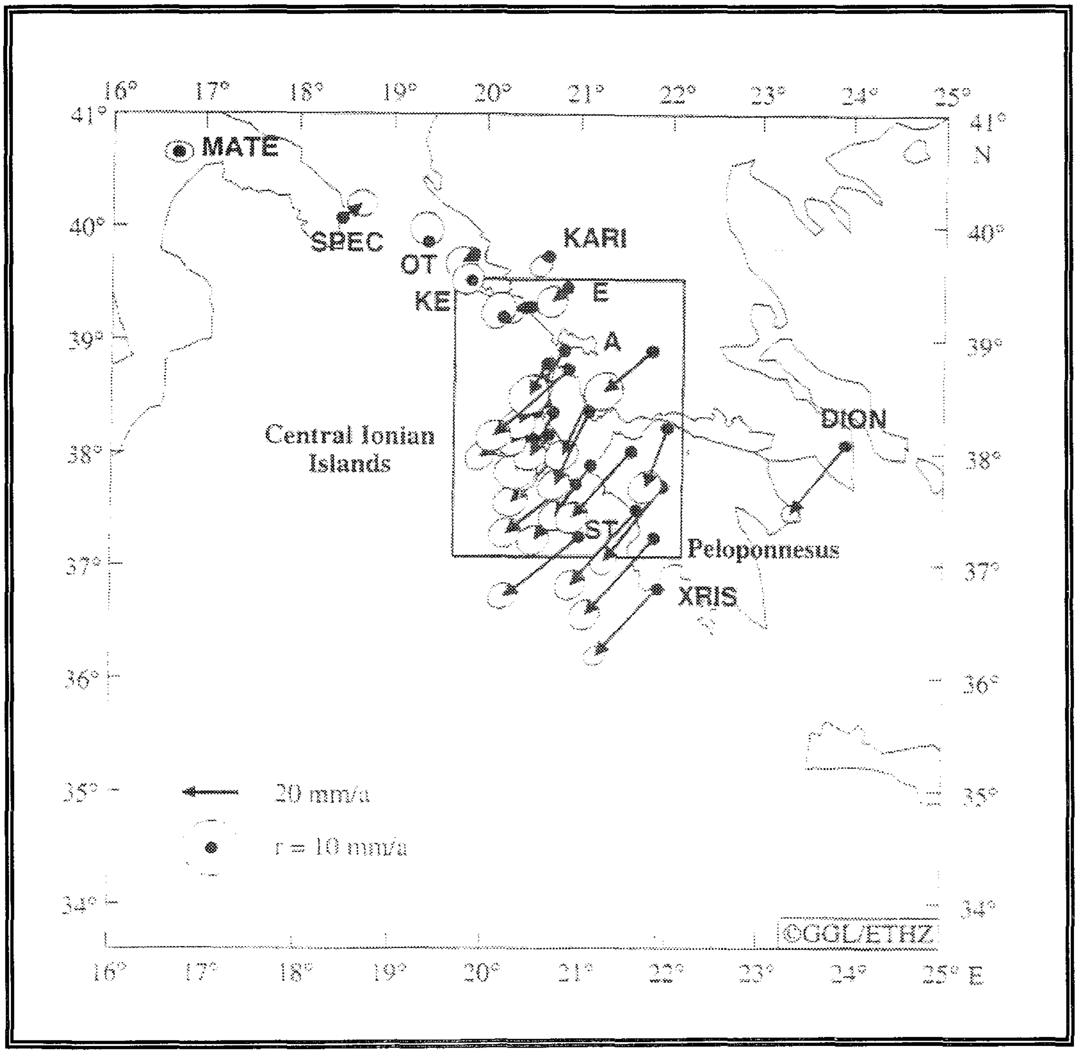

Figure. 2.33. GPS horizontal velocities and their confidence ellipses relative to MATE (Matera) for the period 1989-1993 (after *Kahle et al., 1995*).

Co-seismic displacements of the 1995.05.13 Kozani-Grevena Earthquake, Greece, were measured by *Clarke et al. (1997)*, using a limited number of GPS stations. This method was used for the first time on fault plane determination in the Aegean region.

Turkey

GPS measurements were started in the western Turkish mainland in 1988 (*Oral, 1993*). Surveys carried out in Turkey and surrounding area in 1988, 1989, 1990, 1991, 1992 and 1994 have been most recently published by *Reilinger et al. (1997)* (see also, *Oral et al., 1993, 1995; Oral, 1994*). In 1996 extensive GPS campaigns continued at the previously measured stations (Fig. 2.34) (*Mueller et al., 1997*). Several universities in Turkey (Hacettepe University, Boğaziçi University and İstanbul Technical University) contributed to the collection of data. Durham University (England), IfAG (Germany) and ETH-Z (Switzerland) conducted GPS experiments in Turkey. There is a summary by *Lenk (1995)* of space-based geodetic research activities in Turkey. *Oral et al. (1993)* presented the first results of GPS measurements of regional crustal deformations in Turkey, based on 1988 and 1990 data from western Turkey. They combine the GPS measurements in Turkey with more widely spaced SLR data that provide valuable information for quantifying the diffuse deformation in the eastern Mediterranean. *Oral et al. (1995)* display velocity vectors in both western and eastern Turkey, using 1988, 1990 and 1992 observations (see also *Oral, 1994*). They present crustal velocities of the "Anatolian Plate" by combining their results with independent SLR measurements.

In northwestern Turkey 1990, 1992, 1994 and 1996 GPS campaigns were carried out across a dense network called the Marmara Sea Network (see *Straub & Kahle, 1994, 1997*). The results from the network are summarized by *Straub & Kahle (1997)* (see also *Straub & Kahle, 1995; Straub, 1996*) (Fig. 2.35).

Mueller et al. (1997) summarized all the geodetic studies in the Aegean region and surrounding area. *Mantovani et al. (1995)* critically reviewed SLR and GPS observations in the Aegean region. New GPS campaigns were recently started in the eastern Mediterranean, including also the Balkan region and northern Africa (see *Mueller et al., 1997*).

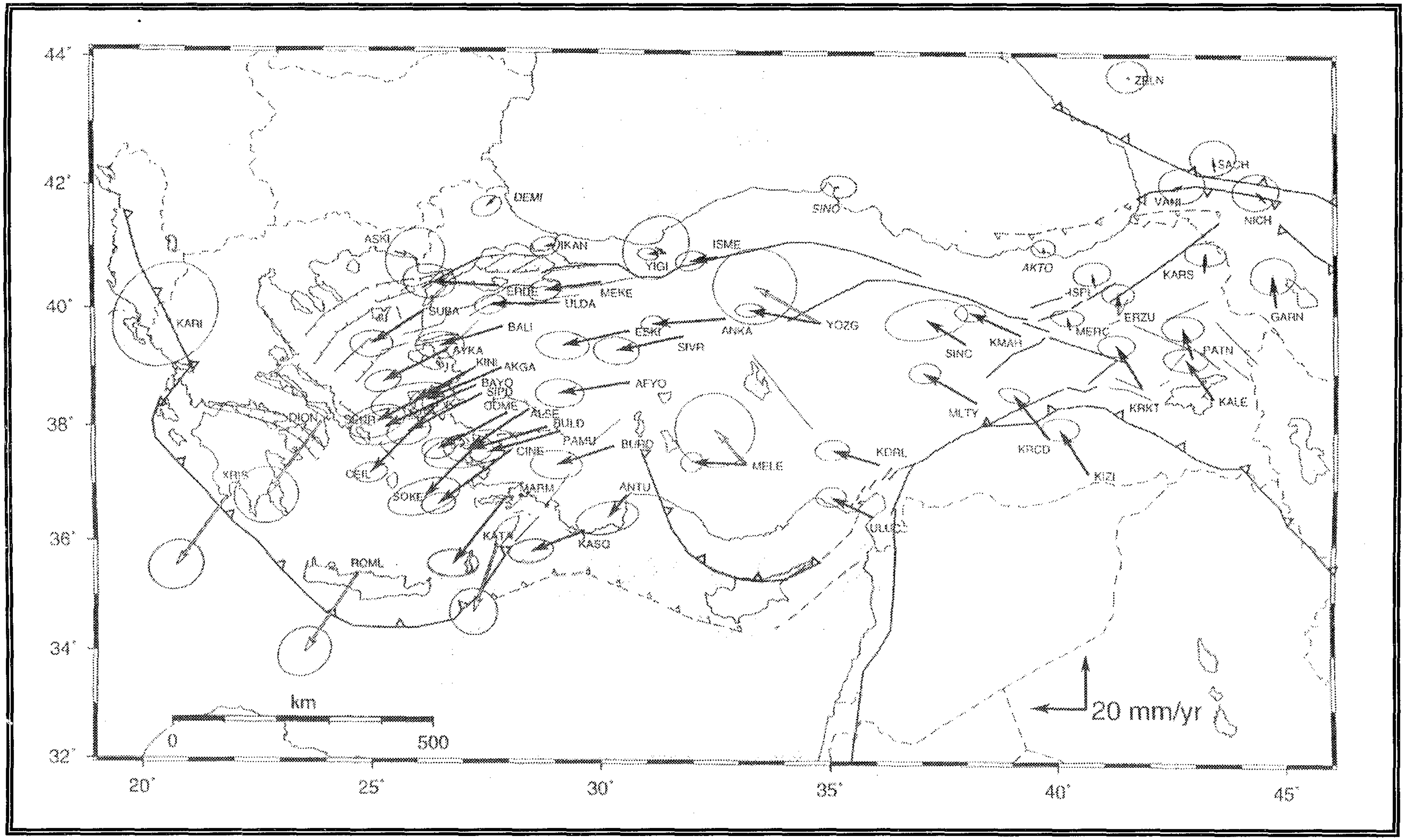

Figure. 2.34. GPS (Solid arrows) and SLR (open arrows) horizontal velocities and their 95% confidence ellipses in a Eurasia-fixed reference frame for the period 1988-1994 (after *Reilinger et al., 1997*). Stations with their codes were denoted.

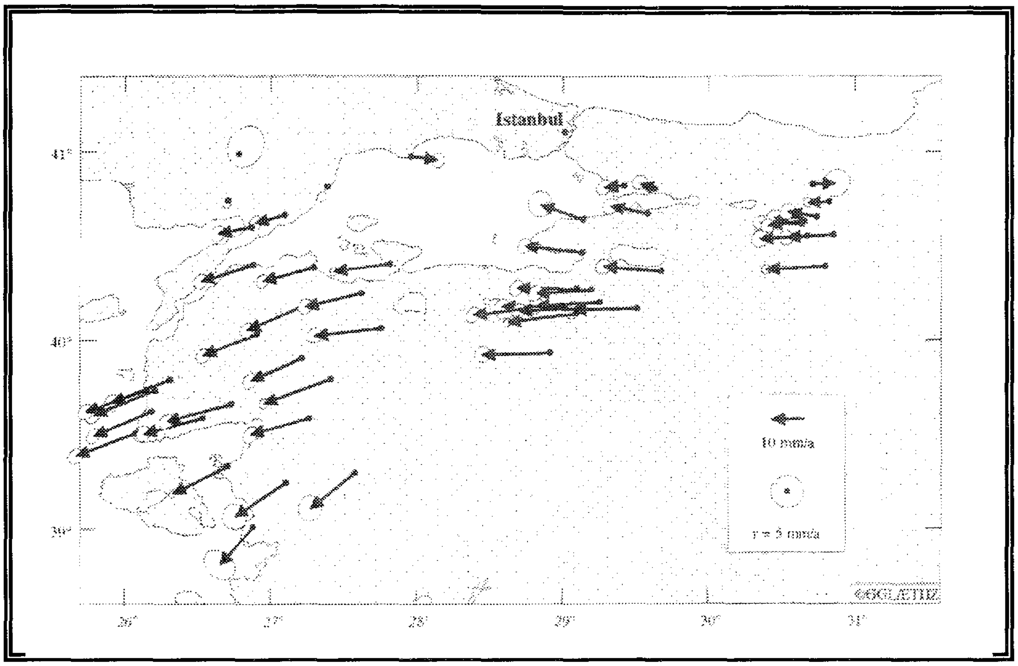

Figure. 2.35. Velocity field derived from the GPS campaigns of 1990, 1992, 1994, and 1996 in northwestern Turkey (after *Straub and Kahle, 1997*). The error ellipses reflect the formal errors.

2.9.2. General Interpretations

The challenge for many earth scientists is to geodetically observe motions across tectonic boundaries and in the interiors of plates, to find out the kinematics of the crust. Modern space geodetic techniques (VLBI, SLR, GPS) have made it possible to measure directly the ongoing rates of the moving plates over long distances (some thousands of kilometers) even though the magnitude of the displacements is relatively small. Until recently, the only way to attempt a reconstruction of the kinematic pattern of such a system was based on the analysis of types and rates of deformation along plate boundaries inferred from geological and geophysical observations, from earthquake focal mechanisms and from terresial geodetic measurements. Geodetic measurements, more recently those using satellite geodesy, provide constraints on the displacement of reference points installed over the Aegean region.

SLR results give a good pattern of the large-scale deformation in the eastern Mediterranean consistent with GPS results (*Oral, 1994; Oral et al., 1995; Le Pichon et al., 1995; Reilinger et al., 1997; Mueller et al., 1997*). However, the SLR network is not sufficiently dense to delineate the details of deformation associated with ongoing continental collision (*Noomen et* *al., 1991, Oral et al., 1993*). The Global Positioning System (GPS) is currently a recent and fast developing geodetic tool (*Straub & Kahle, 1997*). It involves 21 satellites (plus 3 reserves), which emit radio signals (*Straub & Kahle, 1997*). SLR site distribution in the eastern Mediterranean is rather sparse due to the high cost of transporting and operating a mobile system (*Oral, 1994*). On the other hand, GPS is a cheap, easy and sensitive method (*Oral, 1994*). To get better results from satellite geodetic measurements more repeted measurements with more stations must be made. Neither method is useful for estimating vertical movements. It is important to note that the deformation rate found by using geodetic systems is not constant so that the calculated reconstructions may not be true for the longer periods.

From GPS and SLR data *Oral et al. (1995)* and *Le Pichon et al. (1995)* modeled the velocity field of the Anatolian region as a counter-clockwise rigid rotation about an Euler pole located in northern Egypt (near the Sinai Peninsula) relative to Eurasia (see also *Oral et al., 1993; Noomen et al., 1993; Oral, 1994; Le Pichon et at., 1995; Cianetti et al., 1997; Reilinger et al., 1997*). At least 90 per cent of the motion of Anatolia can be accounted for by such rotation (*Reiliniger et al., 1997*). Sites west of the Karlıova triple junction have a tendency to a more westerly directed motion which becomes progressively more pronounced to the W-WSW with stations indicating SW oriented rates in southwestern Turkey (*Reilinger et al., 1997; Mueller et al., 1997*) (Fig. 2.34). However, the lengths of the vectors are not all the same. Residual velocities appear to increase toward the SAA (Fig. 2.34) (*Oral et al., 1995; Reiliniger et al., 1997*). A 2 mm/yr internal deformation is observed on the "Anatolian Plate". On the basis of the GPS and SLR velocities *Reiliniger et al. (1997)* suggest that the motion of the Anatolian-Aegean region may be best described in terms of a single plate (Anatolian Plate) in whose western part (the Aegean region) gradual deformation is taking place internally (south increasing N-S extension) (Fig. 2.34). Even in the Marmara region the highest rates, which are considered as the overall motion of the Anatolian block, are found in the southern part of the Marmara Sea relative to Eurasia (Fig. 2.35) (*Straub & Kahle, 1995, 1997* see also *Mueller et al., 1997*). Further SW the stations in the southern part of the Aegean Sea, south of a line from Evvoia to Chios, are moving at a rate of about 40-45 mm/yr to the SW relative to the Eurasian plate (*Mueller et al., 1997*). This area is also undergoing SW-NE extension in the same direction as the slip vectors on the strike-slip faults in the North and central Aegean (Figs. 2.29, 2.33, 2.34) (*Billiris et al., 1991*). The kinematic field of western Greece, relative to South Italy, is characterized by a negligible motion north of the Kephalonia fault zone

(KFZ). *Clarke et al. (1997)* observed that the extension in the western part of the Corinth rift is greater than that on the eastern part (e.g., Fig. 2.29). This indicates that, in the southwestern part of the Aegean Region, clockwise motion is taking place. *Mueller et al. (1997)* argue that in the area south of the Evvoia-Chios line the motion is essentially homogeneous. On the other hand *Billiris et al. (1991)* and *Clarke et al. (1997)*, who used terresial systems, argue that the area is under going SW-NE extension.

According to the GPS and SLR data, the leading edge of the African plate is being subducted along the SAA at a higher rate than the relative northward motion of Africa (*Oral et al., 1995*). This result is very important because this kind of behavior in subduction systems is an essential indicator of retreat of slab (*Royden, 1993a,b*).

2.10. GRAVITY SURVEY

2.10.1. Data

Source information for gravity data discussed below are listed in Table 2.6.

Table 2.6. Gravity data in the Aegean area

References	***Data source*** *(Numbers from No column. No indicate data source number)*	***Locations*** **EAST MEDITERRANEAN AEGEAN REGION**	***No***
Makris (1973) Makris and Stobbe (1984)	18+14 + data from Greek institutions	Aegean Sea, Turkey and Greek mainland	16 Fig. 2.43, 2.44
Chailas et al. (1992)		Aegean Sea, Turkey and Greek mainland	23
Lagios et al. (1988)	IGME and DEP	Greece and Aegean Sea	21
Morelli (1990)	IBCM, 1989		22
Needham et al. (1973)		North Aegean Sea	17
Vogt and Higgs (1969)		Aegean	13
Genç et al. (1996a)	altimeter data from ERS1	Aegean Sea	26 Fig. 2.39
Allan and Morelli (1965, 1970, 1971) Morelli et al. (1975) Morelli (1990) Finetti and Morelli (1973)	4+15+SACLANT ASW	Aegean Sea	14 Fig. 2.36
Cassinis (1941)		Ionian Sea to Aegean Sea	3
Elbek (1952, 1963)		Turkish mainland	6
Özelçi (1973)	15+MTA	Turkish mainland	18 Fig. 2.41
Akdoğan (1995)	MTA	Turkish mainland	25
Oral (1987)	HGK and MTA	western Turkey	20
Genç et al. (1996b)	MTA	western Turkey	27 Fig. 2.41
Canıtez (1962)		NAF	10
Şalk (1994)	Magsat data	western Turkey and the Aegean Sea	24
Woodside and Bowin (1970)	12+ ACIC	western Turkey and South Aegean Sea	15
Woodside (1975)	12+18+15+C.I.M.	western Turkey and Southeastern Aegean Sea	19 Fig. 2.38
		Mediterranean and South Aegean Sea	
Vening-Meinesz (1932)			1
Cassinis and de Pisa (1935)			2
Fleischer (1964)			4
Cooper et al. (1952)	HMS TALENT		5
De Bruyn (1955), Harrison (1955) Fisher and Hess, (1963)	pendulum data		7
Girdler and Harrison (1957)			8
Worzel (1959)	pendilum data		9
Morelli et al. (1967) Woollard and Rose (1963)			11
Woodside (1968) Rabinowitz and Ryan (1970)	1+2+3+4+5+8+9+W.H.O.I.		12

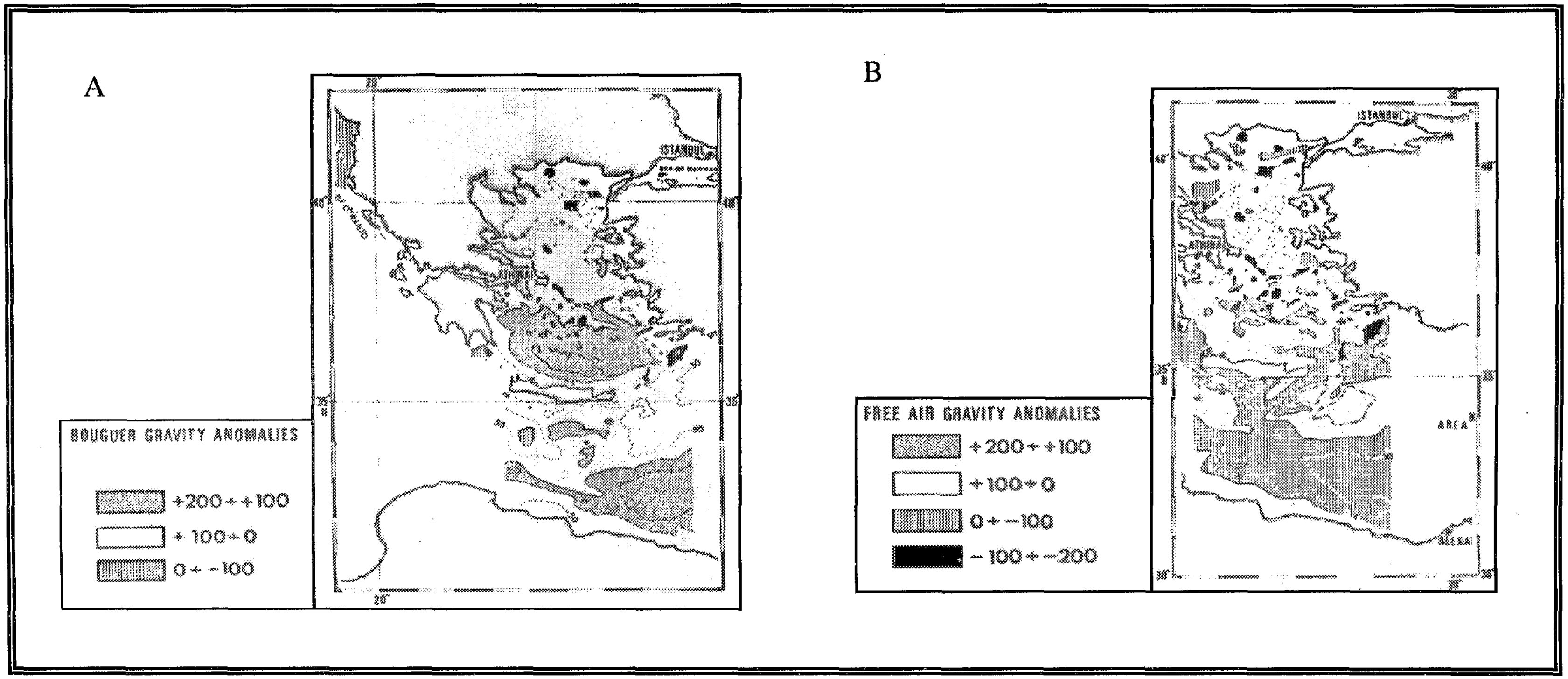

Figure. 2.36. (A) Bouguer and (B) free-air gravity anomaly charts (after *Allan and Morelli, 1971*).

In an early study *Cassinis (1941)* observed the large gravity gradient across the Ionian Sea to

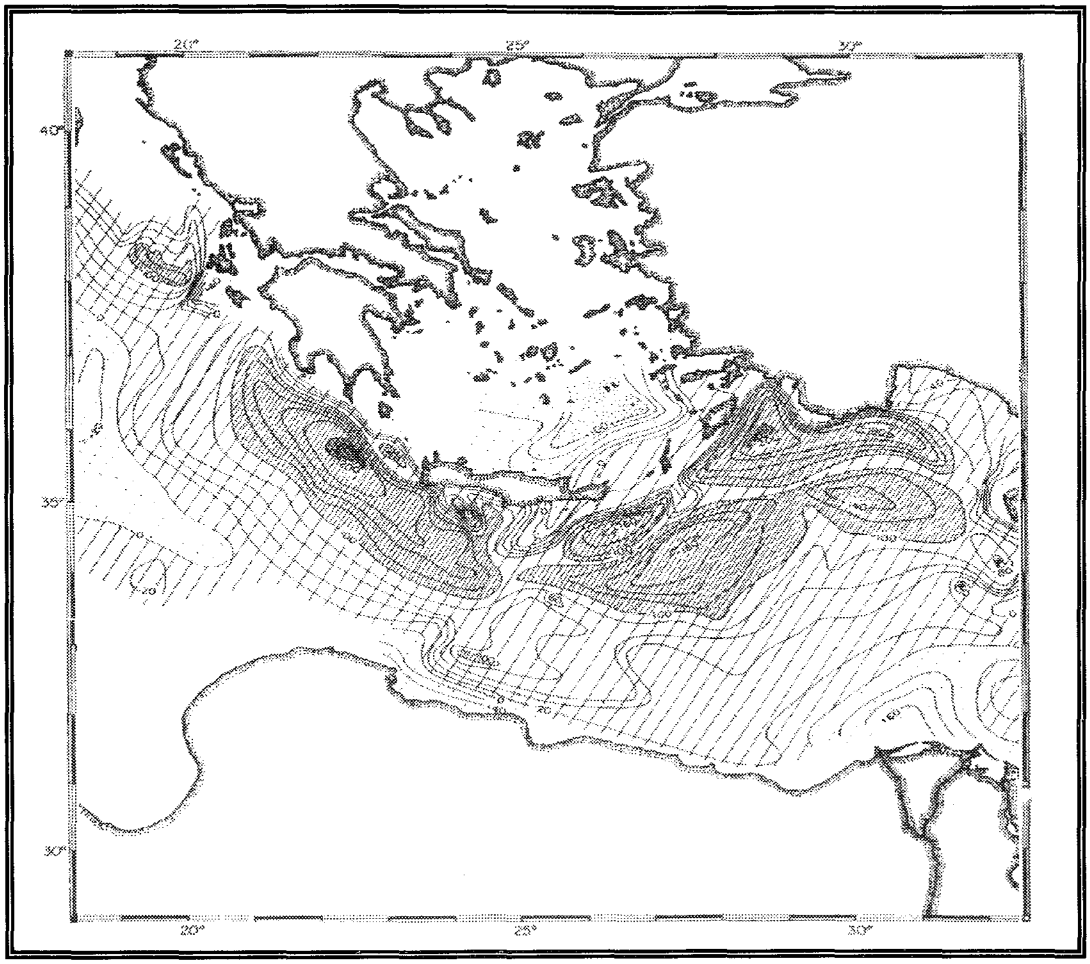

Figure. 2.37. Free-air gravity anomaly map of the south Aegean and surrounding area (from *Rabinowitz and Ryan, 1969*). Diagonal parallel lines represent negative anomalies; the dotted pattern positive anomalies.

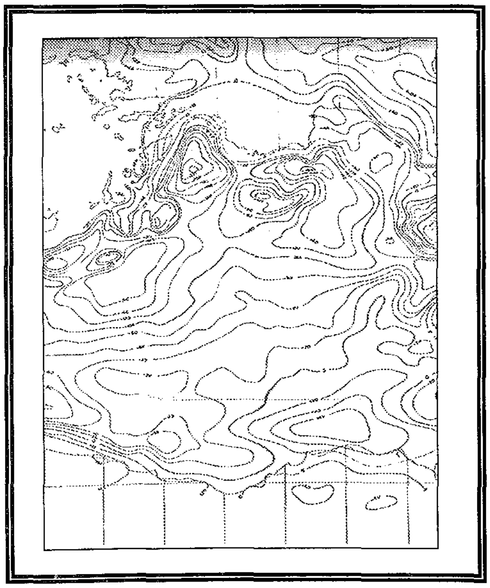

Figure. 2.38. Combined free-air anomaly of Eastern Mediterranean Sea beneath southeastern Aegean region (from *Woodside, 1975*).

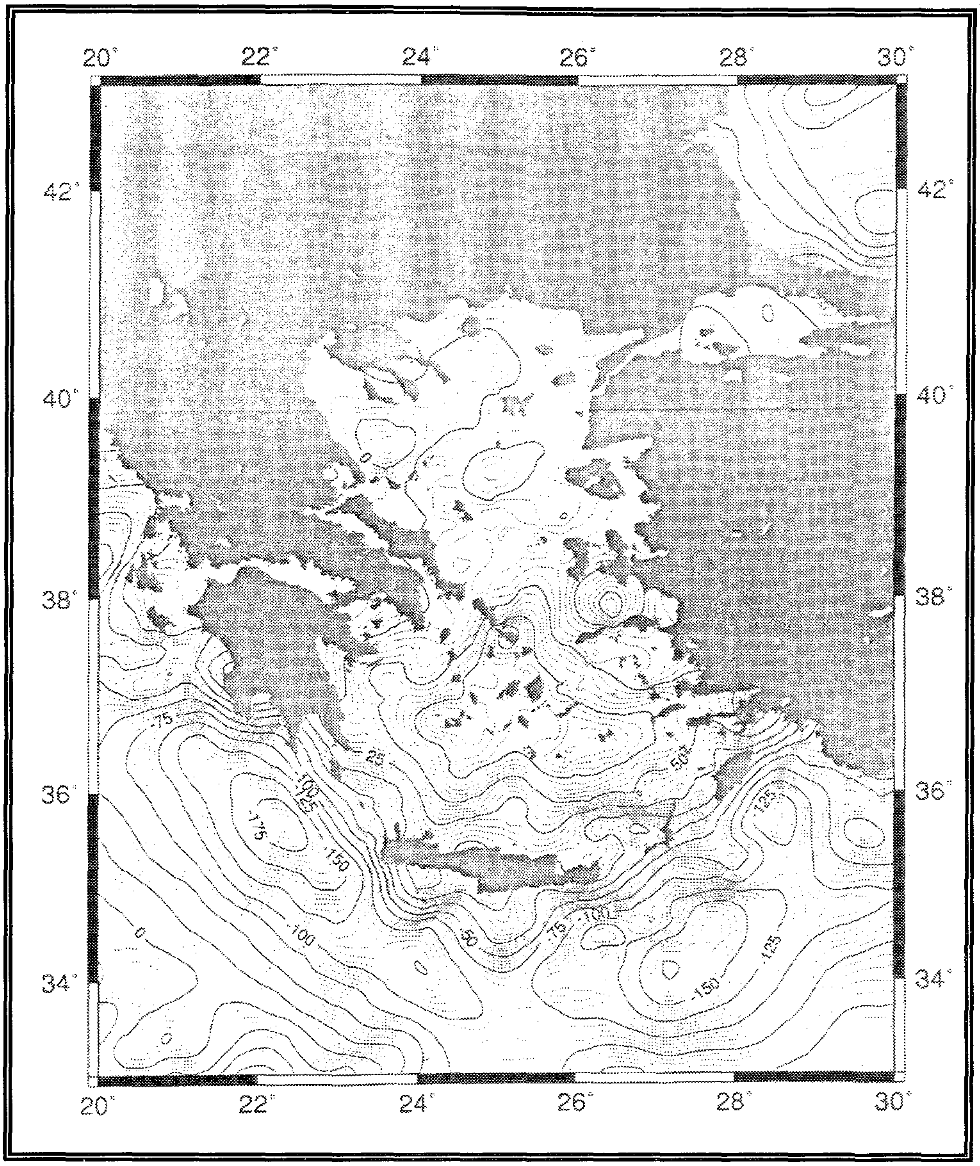

Figure. 2.39. Free air anomaly map of Aegean (after *Genç et al., 1996a*).

the Aegean Sea. The British submarine HMS TALENT (*Cooper et al., 1952*) made further submarine gravity measurements in the eastern Mediterranean that covers the south of Aegean Sea. For the South Aegean, Fleischer constructed a map of free air gravity anomalies in 1946 (see *Fleischer, 1964*). Some gravity measurements were made with the pendulum apparatus aboard submarines and were used in the isostatic gravity maps of the *De Bruyn (1955), Harrison (1955)* and *Fisher & Hess (1963)*. *Worzel (1965)* has summarised and catalogued the pendulum data along the South Aegean Trough (SAT). During the period 1961-1965 co-operative cruises between SACLANT ASW Research Centre, La Spezia, and Osservaterio Geofisico Sperimentale, Trieste, resulted in data that cover most of Aegean Sea (*Allan & Morelli, 1970, 1971; Morelli et al., 1975; Morelli, 1990*). Contour charts of free air and Bouguer gravity anomalies have been constructed a scale of 1:750,000 with 20 miligal contours (Fig. 2.36) (*Allan & Morelli, 1965, 1970, 1971; Finetti & Morelli, 1973*). *Allan & Morelli (1970, 1971)* published the data that cover all the Aegean region. However, the data of *Finetti & Morelli (1973)* also cover part of the Aegean Sea. Another gravimetric study was by *Needham et al., (1973)* in the North Aegean Sea. Some of the surface ship gravity profiles were obtained in 1964 and 1966 aboard the R.V. "Robert D. Conrad" of Lamont-Doherty Geological Observatory, during a reconnaissance survey of the Mediterranean by the Woods Hole Oceanographic Institution (W.H.O.I.). The resulting free air gravity data (*Woodside, 1968, Rabinowitz & Ryan, 1970*) cover part of the South Aegean region (Fig. 2.37). The authors estimate the value of the gravity measurements has an accuracy of 5 miligal. *Woodside & Bowin (1970)* combine preliminary data of Osservatorio Geofisico Sperimentale of Trieste (O.G.S.T.) with the data of *Woodside (1968)*. They also combine these data with land data obtained from the world gravity holdings of the Aeronautical Chart and Information Centre (ACIC) of the U.S. Government. For the purpose of studying the distribution of gravity anomalies which include south and north of Crete. *Rabinowitz & Ryan (1970)* firstly collated their gravity data with data from published pendulum measurements of *Vening-Meinesz (1932), Cassinis & Pisa (1935), Cassinis (1941), Cooper et al. (1952), Girdler & Harrison (1957), Fleischer (1964)* and *Worzel (1959)*. *Allan & Morelli (1971)* included their gravity data (1961-1965 co-operative research) with those of *Fleischer (1964)* and *Woodside & Bowin (1970)*. The density of the survey information that cover most of the Aegean region was designed to be similar to that of the adjacent surveys carried out for the NATO Subcommittee on Oceanographic Research (*Allan & Morelli, 1971*). Cambridge had made an international commitment to the C.I.M. (Cooperative Investigation of the Mediterranean). This study covers the south of the Aegean and western Anatolia (*Woodside, 1975*). *Woodside (1975)* put these results and earlier data (*Woodside, 1968, Rabinowitz & Ryan, 1970*) together (Fig. 2.38) to be issued by the Intergovernmental Oceanographic Commission (IOC) of UNESCO. *Morelli (1990)* printed a Bouguer Gravity Anomaly map (1: 1.000,000, 10 sheets) in 1989 for the Aegean region. A free Air Anomaly (FAA) gravity map derived from the Marine Bouguer Anomaly maps, IBCM (International Bathymetric Chart) (IBCM, 1989) and the bathymetric maps (IBCM, 1981) (see *Truffert et al., 1993*). *Sandwell & Smith (1992)* collected altimeter data from ERS1 and converted them to free-air anomaly, covering all the oceans of the world.

Makris (1976) presented some two-dimensional crustal sections across the Aegean region developed by gravity computations with deep-sea sounding. *Agarwal et al. (1976)* interpreted

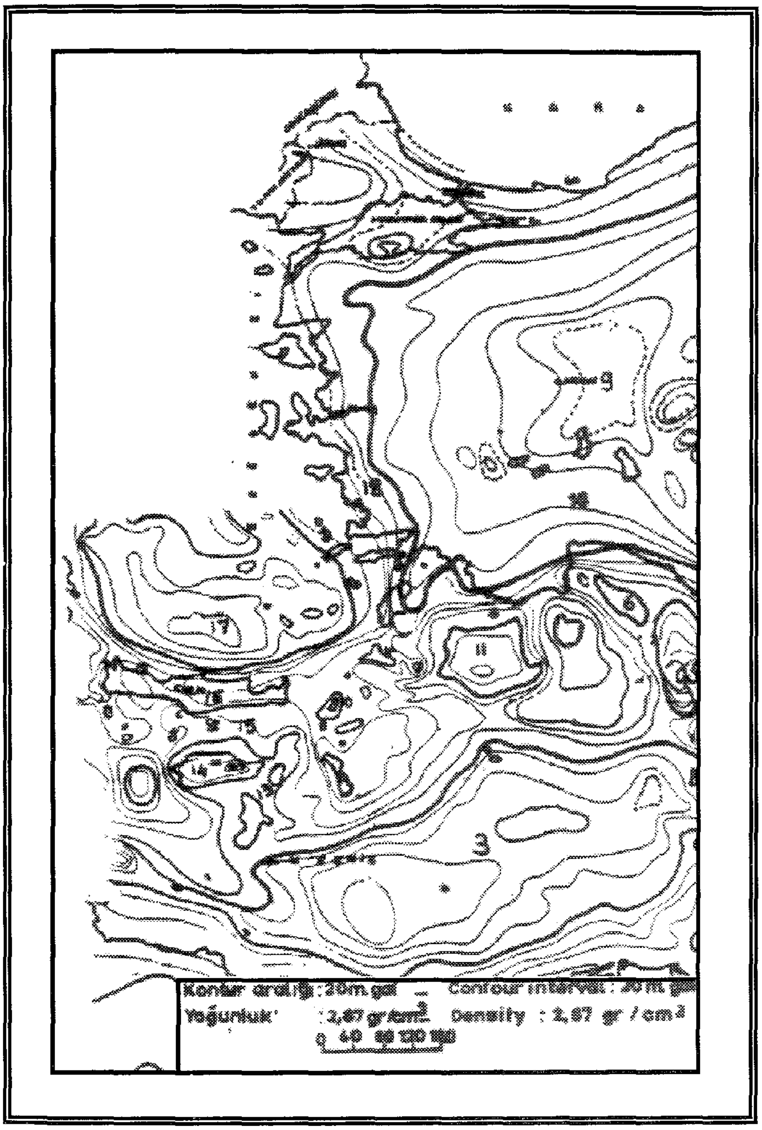

Figure. 2.40. Bouguer anomaly map of south and east Aegean region and Mediterranean region (from *Özelçi, 1973*).

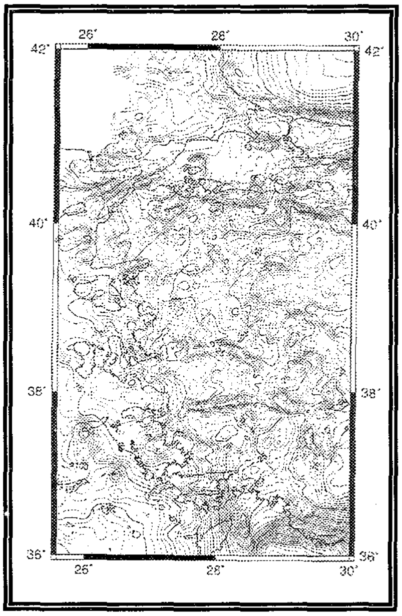

Figure. 2.41. Bouguer gravity map of western Turkey and surrounding area (after *Genç et al., 1996*). Contour interval is 5 mgal.

a two-dimensional gravity profile across the island of Karpathos in terms of a dipping plate in the upper mantle. *Gregersen & Jaeger (1984)* also made a 3-D gravity calculation for a dipping lithospheric plate. Gravity anomalies were also presented for the Central Aegean Sea, Turkey and Greek mainland by *Chailas et al. (1992)*. *Genç et al. (1996a)* published these data for the Aegean Sea area (Fig. 2.39). The gravity map prepared by *Genç et al. (1996a)* has the higher resolution than previous maps for the Aegean (Fig. 2.39). They calculate crustal thickness by using inversion techniques in gravity data from the more detailed regional anomaly map. *Tsokas and Hansen (1997)* also calculate crustal thickness of the central and eastern Aegean region.

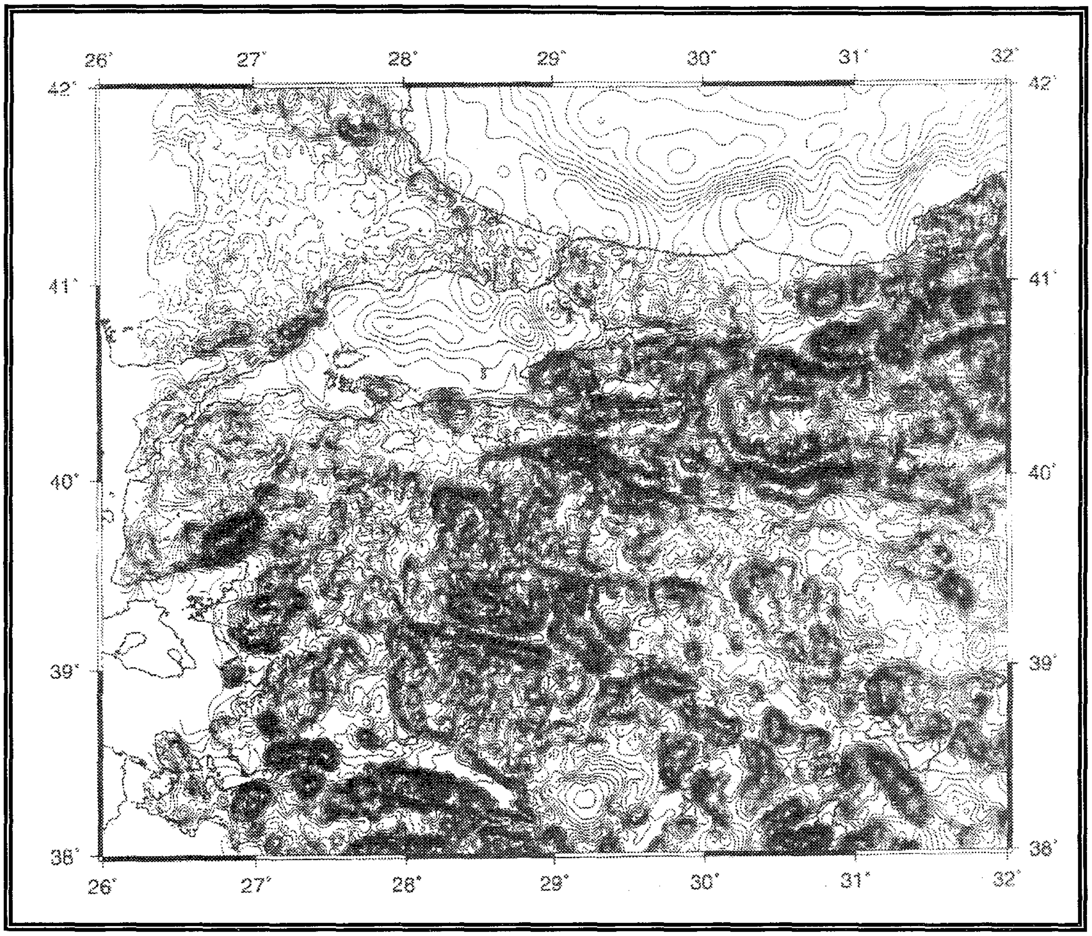

Figure. 2.42. Free-air gravity anomaly map of northwestern Turkey (after *Aygül and Genç, 1998)*. Contour interval is 10 mgal.

Turkish Mainland

Several measurements have been made at major cites (*Elbek, 1952, 1963*), and at a few gravity base stations tied in to the world gravity net (*Woollard & Rose, 1963*). A few measurements have been made in the region of the North Anatolian Fault (NAF) (*Canıtez, 1962*). He used the gravity method to infer the crustal thickness of the NW Anatolia. *Özelçi (1973)* combined *Woodside & Bowin (1970)* data with the data of MTA in Turkey (Fig. 2.40). The Bouguer anomalies in Turkey were obtained from the Cambridge survey (*Woodside, 1975*) during June and early July of 1973, with contours and trends beyond the limits of that survey supplied from the work of *Özelçi (1973)* (*Woodside, 1975*). *Oral (1987)* used maps prepared by the Mapping Service of the Turkish Army (HGK) and MTA. MTA used approximately 5 km station spacing whereas this was 10 km or more in the HGK measurements. A high correlation coefficient (0.96) was found by *Oral (1987)* between HGK and MTA data. *Oral (1987)* stripped the Bouguer anomalies in western Anatolia and took profiles that indicate several surface geological features such as basins filled with Quaternary

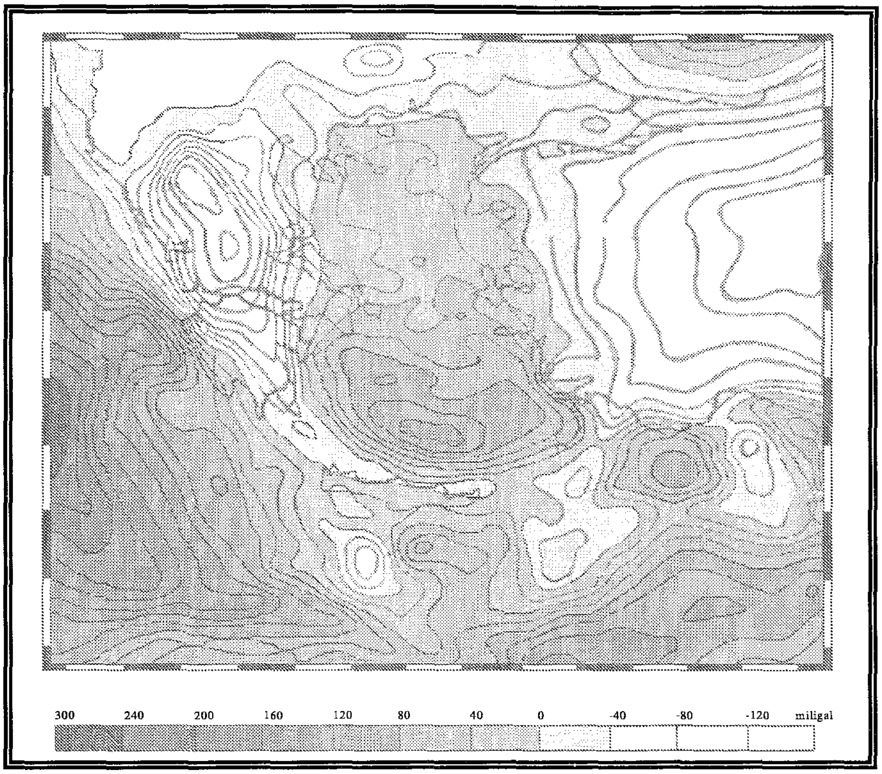

Figure. 2.43 Bouguer gravity map of Aegean and surrounding area (*Makris and Stobble, 1984*).

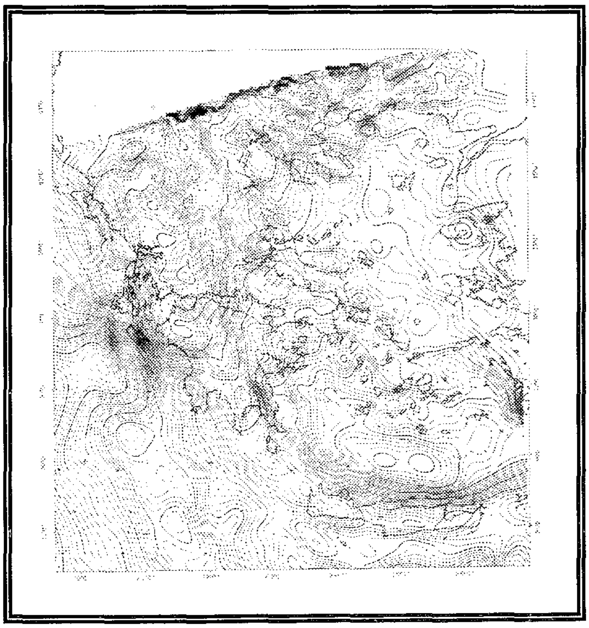

Figure. 2.44. Bouguer gravity map of Greece and surrounding area (after *Makris, 1985*).

and Neogene sediments. *Şalk (1994)* considered Bouguer gravity and Magsat data to prepare a Bouguer gravity map (1 / 5000000) of western Turkey and the Aegean Sea. *Şalk (1994)* calculated crustal thickness over the Aegean Sea and western Turkey and took some profiles. *Akdoğan (1995)* prepared a Turkish Bouguer Anomaly Map from MTA data. *Sarı & Şalk (1995)* also analysed the depth of the metamorphic basement of grabens in West Turkey by 2-

D and 3-D gravity inversion techniques. *Sarı & Ergün (1985)* made a 3D analysis to obtain discontinues in the crust beneath the Saros Bay by using a gravity inversion method (see also *Sarı et al., 1995*). *Genç et al. (1996b)* compiled gravity data from measurements of MTA (65.000) with maximum 5 km interval (Fig. 2.41). *Klingele & Medici (1997)* produced a stripped gravity map of northwestern Anatolia. Gravity data around the Marmara Region have been investigated by *Aygül & Genç (1998)* (Fig. 2.42). They try to minimize artificial effects in merging processes by analyzing data sets. They used land gravity data that were taken from the TÜBİAK YDABCAG 237/G project. Marine gravity data used by them were obtained from *Sandwell & Smith (1992)*, based on altimeter data. *Demirel et al. (1998)*, by using the "Bouguer gravity map of the western Anatolia" that was prepared by the General Command of Mapping and the Directorate of Mineral Research and Exploration, studied the Merle and Enez deltas.

Greek MainlandA Bouguer map with new data, by *Makris (1973)*, covers all the Aegean Sea and Greek mainland (Figs. 2.43, 2.44) *(see also Makris & Stobbe, 1984)*. It includes all available gravity data collected by a number of Greek institutions for the Greek mainland. It has been extended the data for the sea areas around Greece by surveys published by *Finetti & Morelli (1973)* and to western Turkey by *Özelçi, 1973* (see *Makris & Stobbe, 1984*). *Tsokas and Hansen, (1997)* used mainly the data of the Institute for Geology and Mineral Exploration of Greece (IGME) and the Greek Public Petroleum Company (DEP) which were taken from *Lagios et al. (1988)* (Fig. 2.45). *Tsokas and Hansen, (1997)* computed the gravity effect of the crust and extracted it from the Bouguer anomaly. They also extracted the gravity effect of the subducting lithosphere model from the Bouguer anomaly. The remaining anomalies appear to be related to near-surface features and an area of low-velocity mantle in the central Aegean Sea. The map was reproduced from the 4 km grid interval data, which are available to the public.

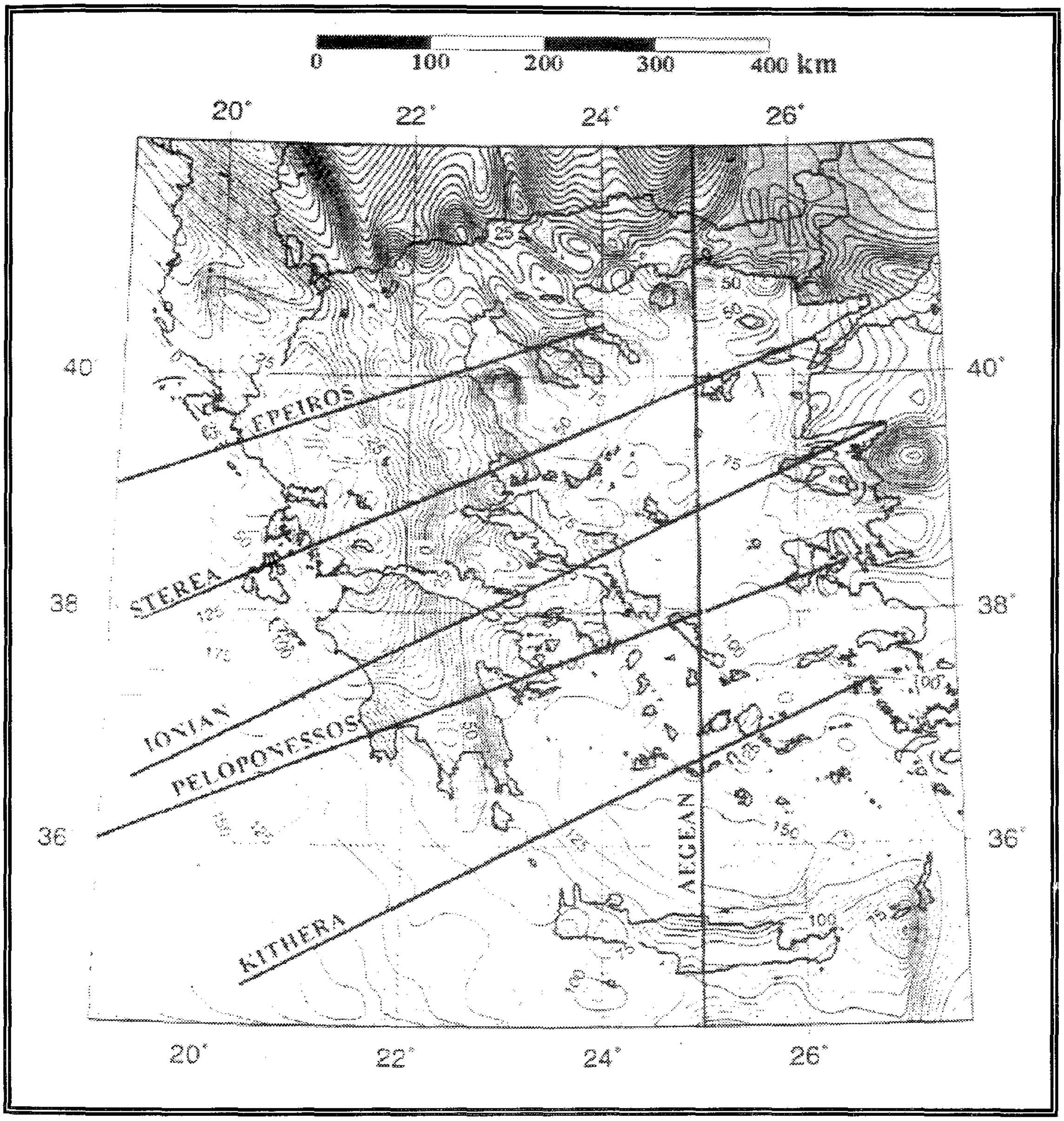

Figure. 2.45. Bouguer gravity anomaly map of Greece and surrounding area (after *Tsokas and Hansen, 1997*).

2.10.2. General Interpretation

There is not yet a map with ERS1 and the land data displayed in the region. One of the most common applications of gravimetry is the determination of lateral density distribution, i.e. the geometry of structures, such as the size of an intrusive body, the thickness of the sediments filling a basin, or the depth of the crust-mantle boundary. In the qualitative interpretation of gravity anomalies, the first rule is to try to separate the long-wave anomalies from the short-wave ones: in the relative sense, the long-wave anomalies are due to deeper mass variations, the short-wave anomalies are on the other hand due to shallower ones. The gravity stripping technique is primarily aimed at modeling the effects of surface geology such as sedimentary basin fills. To do that the long wavelengths are reduced. The short wavelength gravity decreases the effects of the deep structures. The resulting anomaly with short wavelengths the signature of some ultrabasic rocks, ophiolite zones or, sedimentary basins near the surface. The filtering techniques are aimed at reduce by short wavelengths effects. This reveals deeper effects of the mass.

Where the Aegean region has the thinnest crust the Bouguer Gravity values become positive and bathymetric depth maximal (Fig. 2.18, 2.43). Generally, a large positive Bouguer anomaly of gravity is observed in the Cretan basin, while negative anomalies are observed on the Turkish and Greek mainlands (e.g., Fig. 2.43) (*Makris, 1976*). Except for the partly undercompensated coastal area with a positive anomaly in the Turkish mainland (Fig. 2.43) (*Özelçi, 1973; Makris, 1985; Oral, 1987*), the Aegean region is generally in approximate isostatic equilibrium (*Oral, 1987*). The positive Bouguer gravity anomaly that is observed in the western tip of the Turkish mainland (*Rabinowitz & Ryan, 1970; Özelçi, 1973; Şalk, 1994*) is interpreted as the eastern limit of the positive anomaly belt that has been identified as the back of the South Aegean Island Arc. In addition, the Sea of Marmara and north of it gives rise to positive anomalies in contrast to the expected (negative) Bouguer anomaly that would indicate isostatic equilibrium for lands (Fig. 2.43). The zero contour of the Bouguer gravity anomaly follows the trend of NAF at the eastern parts of the Sea of Marmara because of the basin fill (*Şalk, 1994*). The anomaly decreases eastwards with a smooth gradient in Turkey (*Özelçi, 1973; Şalk, 1994*). Negative Bouguer anomalies are observed all over the Greek mainland (Fig. 2.43) (*Makris & Stobbe, 1984*). In the Greek mainland the situation is different from that observed in the western coast of the Turkish mainland (*Şalk, 1994*). The area is exactly in isostatic equilibrium beneath Menderes Massif where relatively higher gravity values are observed as a result of mantle rise and the effects of magmatic rocks near the surface (*Şalk, 1994*). In the Gediz and Menderes Massifs in western Turkey, which have generally E-W trends, large areas are identified with negative Bouguer anomalies (Fig. 2.41) (*Şalk, 1994*). *Lort (1971)* suggested that the Mesozoic granitic bodies exposed in western Turkey possibly continue into the sea, based on gravity anomaly patterns (see also *Ercan & Kılıçer, 1998*). Also in the Marmara region, high gravity anomalies are observed. *Oral (1987)* suggest denser material beneath the axis of the gravity high in the Marmara region (see also *Lyberis, 1984*).

The pattern of the gravity anomalies indicates the existence of a fracture zone running from the northern Evvoia Gulf across the Sperchios Valley and the Greek mainland to the Gulf of Preveza (Fig. 2.44) (*Makris, 1977*). This is also the case along the Gulf of Corinth between the Greek mainland and the northern coast of the Peloponnesos (*Makris, 1977*). From the behavior of the gravity field, *Makris (1977)* separated the ophiolites of the Greek mainland into two groups: one consisting of the ophiolites of the Pindos Mountains and North Evvoia and the second consisting of the Vourinos Complex and the ophiolites of the Vardar Zone along the Chalkidiki Peninsula. These areas are characterised by positive Bouguer values (*Makris, 1977*). Results of *Makris (1977)* indicate that all Greek ophiolites have been moved out of their original position and that their maximum thickness does not exceed 2.5 km (*Makris, 1977*). On the other hand, along the Greek-Bulgarian border, the gravity field becomes negative with values of approximately -60 miligal at the Rhodopi crystalline massive (*Makris, 1977*).

In the Aegean Sea the isogravity lines are less dense than over the surrounding land areas, so the variations are seen more easily (*Makris, 1985*). Strong positive Bouguer anomalies exist over the Aegean Sea that are not seen on land areas (*Fleischer, 1964; Rabinowitz & Ryan, 1970; Allan & Morelli, 1971; Özelçi, 1973; Makris, 1973, 1976; Agarwal et al., 1976; Morelli, 1990; Şalk, 1994*) (Fig. 2.43). Two main zones in the area can be divided according to the gravity anomalies: in the south high Bouguer gravity anomalies (75 miligal) and to the north low gravity anomalies (0 Miligal) (*Genç et al., 1996a*) (Fig. 2.39). The highest values (+175) are measured in the Cretan Basin (*Makris, 1978a*). In the central Aegean Sea values range between +100 and +50 miligal (*Makris, 1985*). *Makris (1977)* found a series of gravity maxima in the Cretan graben. On the other hand, using MAGSAT spacecraft data, *Şalk (1994)* found low gravity anomalies trend ENE-WSW direction.

A narrow zone of large negative (0 to 80 miligal) gravity anomalies marks the southern edge of the Aegean region. It is parallel to the island arc and along the South Aegean and Pliny-Strabo trenches from western Greece to the Rhodes abyssal plain (e.g., Figs. 2.39, 2.43) (*De Bruyn, 1955; Harrison, 1955; Fisher & Hess, 1963; Worzel, 1965; Woodside & Bowin, 1970; Allan & Morelli, 1971; Agarwal et al., 1976*). Relatively high anomalies (free-air anomaly as high as -60 miligal and Bouguer anomaly of +100 to + 140 miligal) centered over the Anaximander seamounts separate the arcuate low south of Crete from a similar trend south of Cyprus (Fig. 2.37) (*Woodside & Bowin, 1970; Rabinowitz & Ryan, 1970; Ergün et al., 1998*). The region over Köyceğiz demonstrates the negative anomalies that appear in contrast to the surrounding (western Turkey) characteristic positive Bouguer gravity field, which could be related to a subduction zone (Fig. 2.41) (*Özelçi, 1973; Şalk, 1994, Woodside, 1975; Oral, 1987*).

The regional Bouguer anomalies are positive in the Mediterranean Sea (*Morelli, 1990; Şalk, 1994*). The strongest positive Bouguer anomalies exist over the Ionian sea (Fig. 2.43) (*De Bruyn, 1955; Fahlquist, 1963; Woodside & Bowin, 1970; Makris, 1976; Morelli, 1990; Şalk, 1994*). According to *Morelli (1990)* a very thin oceanic crust with a very thick sedimentary cover exists in the Ionian abyssal plain. In the Herodoutos abyssal plain the crust was modeled by *Şalk (1994)*, using gravity data, oceanic. On the other hand *Morelli (1990)* pointed out that the Adriatic Promontory has a continental crust.

2.11. MAGNETIC STUDIES

2.11.1. Data

During 1957-1958, the Magnetics Division of the U.S. Naval Oceanographic Office made an aeromagnetic survey to prepare total magnetic field intensity charts of most of the eastern Mediterranean (*Vogt & Higgs, 1969; Lort, 1971*). The first aeromagnetic surveys conducted over the Turkish mainland were by MTA (1960-1961). The data collection was made by Canadian Aeroservice (*Hutchison et al., 1962;* see also *Aydın & Karat, 1995*). Data on total magnetic-field intensity have been obtained from marine surveys made by Saclant ASW Research Centre, La Spezia and the Ossenaterio Geofisico Sperimentalle, Trieste from 1961 to 1965 (*Allan & Morelli, 1965, 1971; Finetti & Morelli, 1973*). It is presumed that marine data are more reliable for detailing features in the Aegean region because of the dense tracking of the ships (see *Lort, 1971*). In 1966 an extensive Ferro field exploration survey was started in western and eastern Turkey (see *Aydın & Karat, 1995*). The intervals between profiles are set according to the lithologic structures (*Genç et al., 1996b*). *Sanver (1975)* prepared a filtered magnetic map for western Turkey. *Ergün (1977)* reviewed the magnetic data obtained by MTA in Marmara and Saros. *Agocs (1977)* obtained an airborne magnetometer profile that crossed the Marmara Sea from Lake Manyas to Yeşilköy Airport (S-N). The magnetic data of Greece and surrounding area have been published by *Makris (1985)* (Fig. 2.46). The Intergovernmental Oceanographic Commission (IOC) of UNESCO issued an overprint of existing Magnetic Anomalies (1993). Between 1978 and 1989 a new aeromagnetic survey was made in Turkey (*Aydın & Karat, 1995*). *Şalk (1994)* prepared magnetic maps from Magsat data obtained from the British Geological Survey covering the area between 25°-45° E longitudes and 35°-45° N latitudes collected by the Magsat Spacecraft (Satellite magnetometer). *Şalk (1994)* used no magnetic data (*Vogt & Higgs, 1969; Allan & Morelli, 1971*) from conventional aeromagnetic surveys. *Genç et al. (1996b)* converted the analog data that had been collected in 1966 in western and eastern Turkey (see above) to digital. The data was sampled with 2.5 m interval from this map (*Genç et al., 1996b*). They mixed this data with previous data for the Aegean Sea (*Finetti & Morelli, 1973*) (Fig. 2.47, 2.48). However, the resolution of the sea data is smaller than the resolution of land data. To adapt the two groups of data *Genç et al. (1996b)* reduce the land data to the sea data interval.

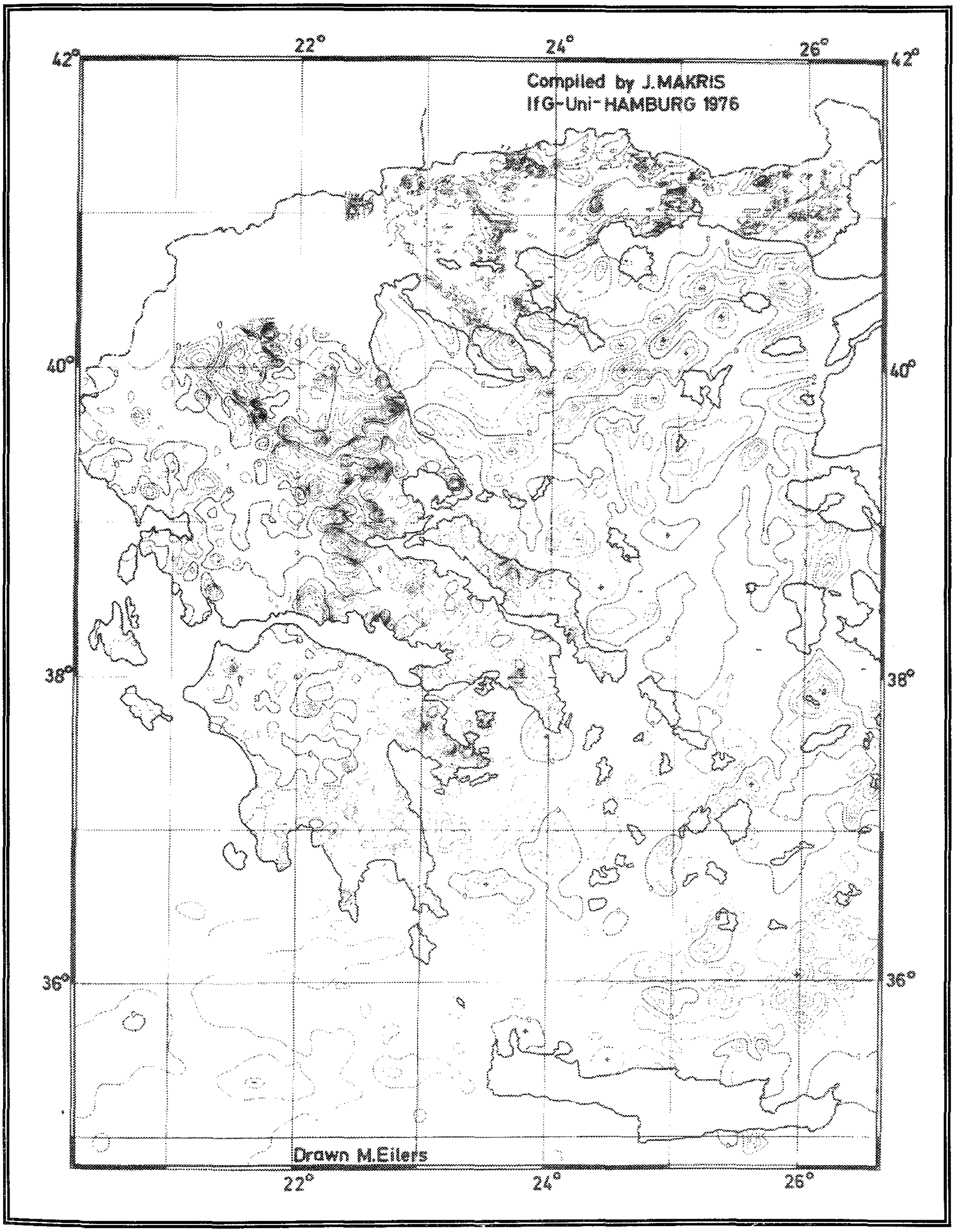

Figure. 2.46. Magnetic map of Greece (after *Makris, 1985*).

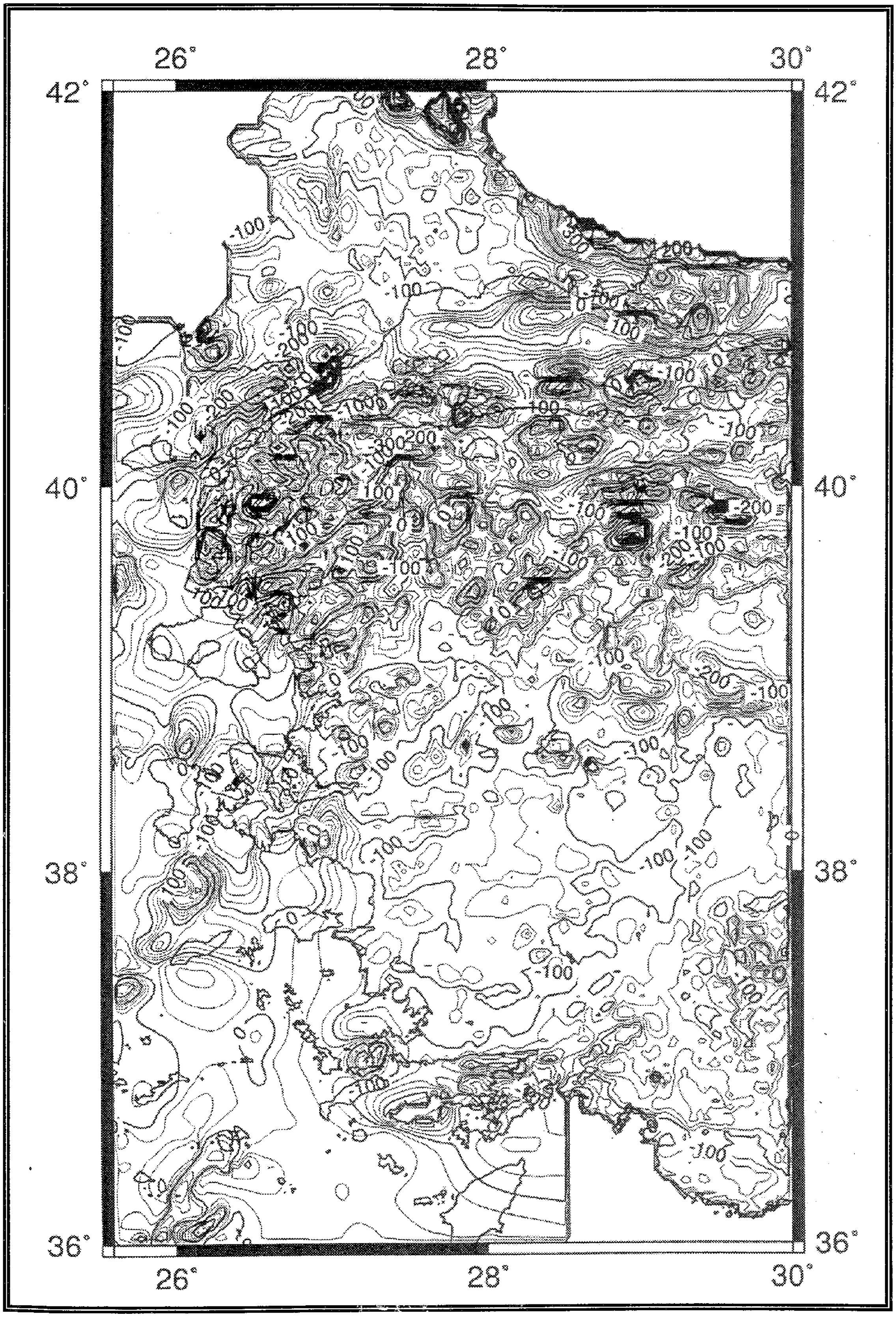

Figure. 2.47. Magnetic anomaly map of western Turkish mainland and surrounding area (after *Genç et al., 1996b*). Contour interval is 2.5 nT.

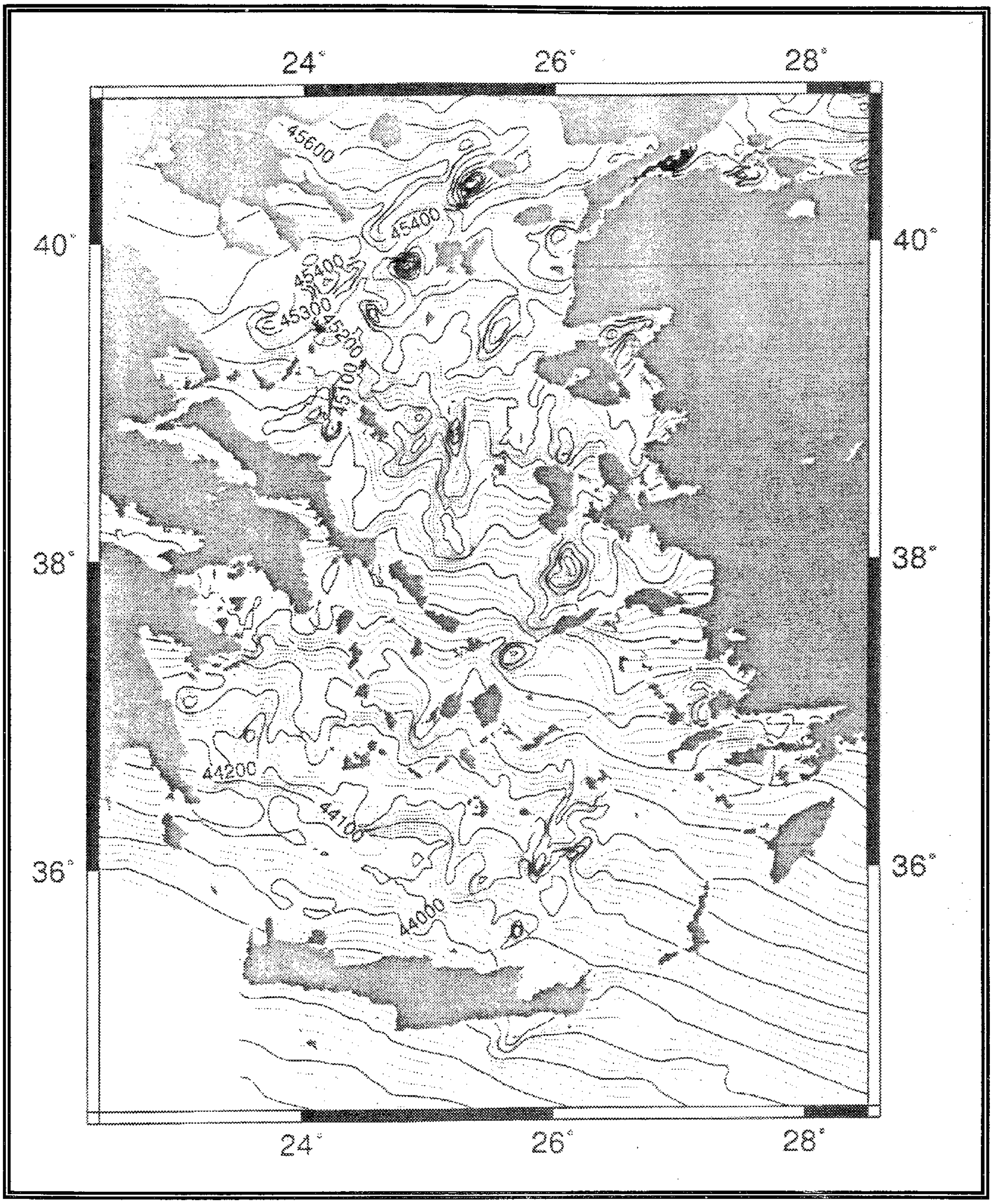

Figure. 2.48. Magnetic anomaly map of Aegean and surrounding area (after *Genç et al., 1996a*).

2.11.2. General Interpretation

An aeromagnetic survey of most of the eastern Mediterranean reveals a large region south of the South Aegean Arc (SAA), which is unusually undisturbed by anomalies (smooth field) and of low amplitude (Fig. 2.48) (*Vogt & Higgs, 1969*). No patterns of magnetic lineation are as distinctive as those over the oceanic ridges (*Allan & Morelli, 1971*). Processes that could have erased magnetisation contrasts are related to the tectonic deformation in the

Mediterranean Sea floor (ie metamorphism or intense fracturing of the magnetic skin on the oceanic crust). *Allan & Morelli (1971)* suggest that this is because a thick sedimentary crust was created by compression between Africa and Europe.

In contrast, the concave side of the South Aegean Volcanic Arc (SAVA) in the Aegean Sea is characterised by regular short-wavelength anomalies (Fig. 2.48) (*Vogt & Higgs, 1969; Allan & Morelli, 1970, 1971; Rabinowitz & Ryan, 1970; Woodside & Bowin, 1970; Finetti & Morelli, 1973*). These anomalies can be related to intrusives of magmatic and granitic masses. (*Caputo et al., 1970; Şalk, 1994*). Linear belts of anomalies coincide with structural trends in the lines of Hercynian granites of the Aegean (*Lort, 1971*). Relatively undisturbed magnetic anomalies occur in the Aegean Sea north of the shallow trough that is between Crete and the SAVA (Fig. 2.48). The magnetic field is smooth (*Vogt & Higgs, 1969*) but relatively high (*Emery et al., 1966*) in the area.

The active volcanic arc zone (SAVA) is remarkable with high magnetic anomalies (*Maley & Johnson, 1971*). The other intense positive magnetic anomalies observed in the Aegean Sea occur along the North Aegean Trough (NAT) (Fig. 2.48) (*Vogt & Higgs, 1969; Şalk, 1994*). This suggests that it may be a tensional graben along which magnetised rock has been intruded (*Vogt & Higgs, 1969*). *Allan & Morelli, (1970)* suggested that this trough probably represents new crust, which is presently being created in a similar way to that of the Red Sea rift (see also *Allan & Morelli, 1971*). But no magnetic lineation like that in mid-oceanic ridges exists along this trough.

The filters eliminate very short wavelengths, sharp anomalies, and rapid variations that depend on surface structures. In all filtered maps of *Şalk (1994)*, the major magnetic anomalies show trend of NE-SW direction at the western Turkish mainland and values increase toward to north (see also *Sanver, 1975*). Important positive magnetic anomalies are observed in NW Turkey (from Biga peninsula to Thrace region), along the İzmir-Ankara ophiolite zone and the Massifs in western Turkey (Fig. 2.47) (Menderes, Gediz, Kazdağ and Uludağ) *Şalk (1994)*. These magnetic anomalies are associated with the granodiorites and the andesitic volcanism, which resulted from active faulting in the region due to extensional tectonics (*Şalk, 1994*). A positive anomaly belt follows the North Anatolian Fault (NAF) zone (*Şalk, 1994*). The filtered magnetic map of *Sanver (1975)* allows comparison of anomalies with volcanic cumulations (see also *Hisarlı, 1995*). Observed NE trending anomalies in the

Aegean Sea may reflect the southwestward extension of Mesozoic igneous complexes from Turkey, and Cenozoic volcanic activity (*Vogd & Higgs, 1969*). A magnetic profile of *Agocs (1977)* in the Marmara Sea shows that a major normal fault is indicated north of and parallel to the south coast of the Marmara Sea.

The Teke domain, indicated by a negative anomaly, which has been considered as an relic of the old oceanic crust of the African plate is placed at the south of the positive anomaly along the Fethiye–Burdur fault zone (*Şalk, 1994*). SW Turkey is considered the transition zone from positive magnetic values to negative magnetic values (Fig. 2.47) (*Şalk, 1994*).

2.12. MAGNETOTELLURIC

2.12.1. Data

Thrace (northwestern Turkey)

In 1976, magnetoteulliric measurements were made of 9 sites in the Thrace peninsula, (1976), from Kırklareli to Tekirdağ (*İlkışık et al., 1996*). The sedimentary and crustal thicknesses were determined from these results (see Table 4.1). In 1980 some additional magnetotelluric soundings were performed in the area (T in Fig. 2.49) (see *İlkışık et al., 1996).* Results show variations of crustal thickness (see Table 4.1).In 1995 changes of the magnetoteulliric field with time were measured at 40 stations in Thrace by the Türkiye Ulusal Deniz Jeolojisi and

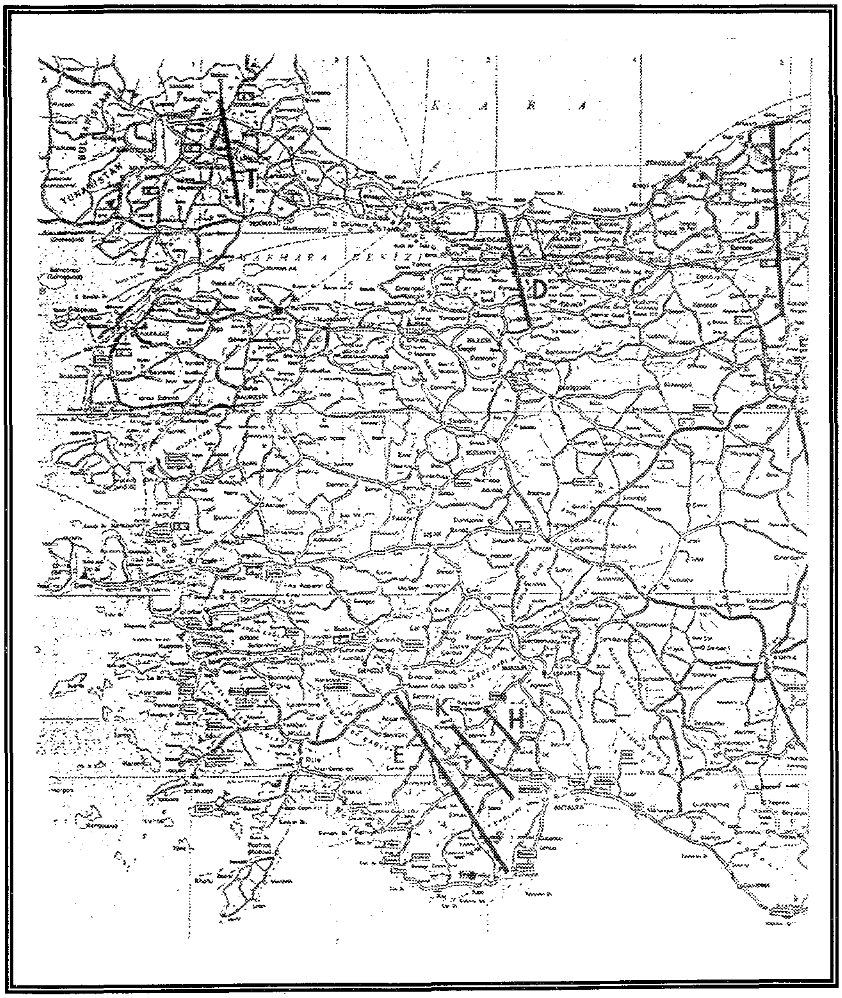

Figure. 2.49. The locations of magnetotelluric profiles in western Turkey (1975-1990) (after *İlkışık et al., 1996*). See text (Section 2.12) for detail.

Jeofiziği Araştırma Programı (National Marine Geology and Geophysics Research Program) (*İlkışık et al., 1996*). The stations are along a 20 km profile (NE-SW) from the Black Sea to the Aegean Sea (Fig. 2.49).

North Anatolian Fault Zone (northwestern Turkey)

In 1983 and 1984 magnetotelluric data were collected at 11 stations in north-northwestern Turkey at two different locations. Six of the sites lie on a profile approximately normal to the North Anatolian Fault Zone (NAFZ) near İzmit (D in Fig. 2.49) (*İlkışık et al., 1996*). The remaining five sites lie to the SE of Zonguldak (J in Fig. 2.49) (*İlkışık et al., 1996*). The measurements cover an approximate depth range of 2 to 100 km (*İlkışık et al., 1996*).

Western Taurudes (southwestern Turkey)

In 1984 and 1985 magnetotelluric measurements were made at 45 sites along 3 profiles in western Tourides, which are approximately normal to the thrust font of the Lycian Naples on the Bey Mountains autochthon and Menderes massive (E, K, H in Fig. 2.49) (*İlkışık, 1990b*).

Western Turkey

In Salihli and Alaşehir magnetotelluric measurements were made for geothermal studies (see *İlkışık et al., 1996*). These measurements in the Gediz graben allow an estimation of crustal thickness (Table 4.1).

Greece

The magnetotelluric sounding method was employed at 15 sites in the Gulf of Corinth, Central Greece, to investigate the electrical properties of the crust and upper mantle.

İlkışık et al. (1996), who reviewed magnetotulliric studies in Turkey and near environs noted that, in 1986 in Milos island in the Aegean Sea, detailed magnetotelluric studies were made by four European Institutes (*İlkışık et al., 1996*).

2.12.2. General Interpretation

Crustal thickness determined from site measurement is included in Table 4.1. The shape of the Lycian Naples and the deep structure of the thrust-front and the structural connection between the Likya Naples and the Menderes massive can be deduced from the 1984-1985 magnetotulliric measurements (*İlkışık, 1990b; İlkışık et al., 1996)* (see Fig. 2.49). The other finding is the shear zone of Fethiye – Burdur (NE-SW) (*İlkışık, 1990b; İlkışık et al., 1996*). The southern edge of the Menderes Masif lies beneath the Gölhisar area (*İlkışık, 1990b; İlkışık et al., 1996*).

2.13. GEOELECTRICAL STUDIES

2.13.1. Data

MTA made geoelectrical studies to determine sediment thickness in Germencik-Aydın and Sarayköy-Denizli areas (*Turgay et al., 1980; Şahin & Ünay, 1981*). Recent geoelectrical studies carried out by MTA were mainly directed to the investigation of Neogen basins for shallow resources (*Şardar, 1995*). The Denizli Kızıldere, Aydın-Sultanhisar-Salavatlı and Aydın-Gemecik-İncirliova geothermal areas were examined by means of geoelectrical studies (*Yücel, 1995*). Recently, to obtain the two dimensional geoelectrical structure of the Büyük Menderes Graben and to illustrate the lateral geoelectircal discontinuities (possible faults), *Çağlar et al. (1998)* have used vertical electric sounding data. The study is continuing. There are many other local studies in the region.

2.14. HEAT FLOW

2.14.1. Data

Europe and Whole Aegean Region:

In 1977, Cermák & Hurtig (197?) prepared a map of the heat flow field in Europe (Fig. 2.50). This effort was followed by other studies (Fig. 2.51) (*Cermák, 1978, 1979; Cermák & Rybach, 1979*). *Stobbe (1980)*, using heat flow map of *Cermák (1979)* (Fig. 2.51), represented some cross sections, which cut the Aegean region in various directions. They made corrections taking account of the sediment thickness and conductivity of the rocks. They displayed them with other correlative geophysical data (see *Makris & Stobbe, 1984; Makris, 1985*). *Hurtig et al. (1990)* prepared a European heat flow map. In this map, the data for Turkey comes from the study of *Tezcan (1979)* (Fig. 2.52).

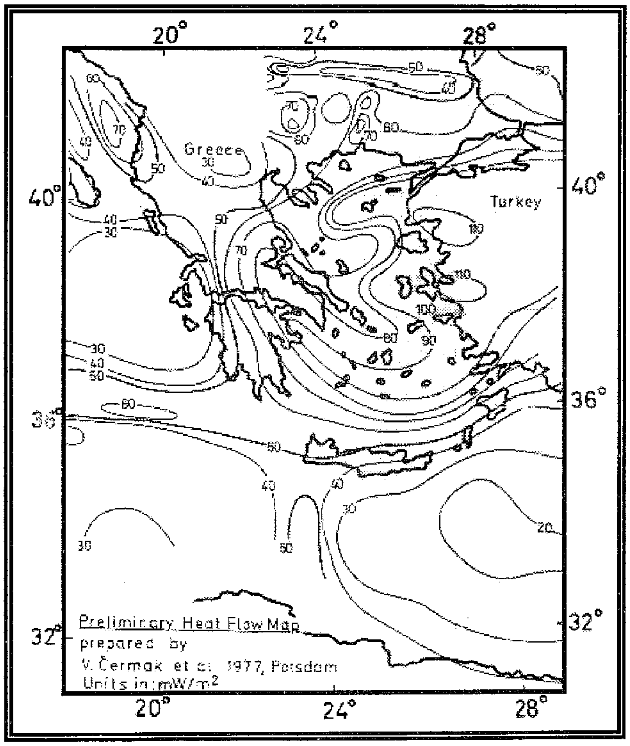

Figure. 2.50. Heat flow map of Aegean and surrounding area.

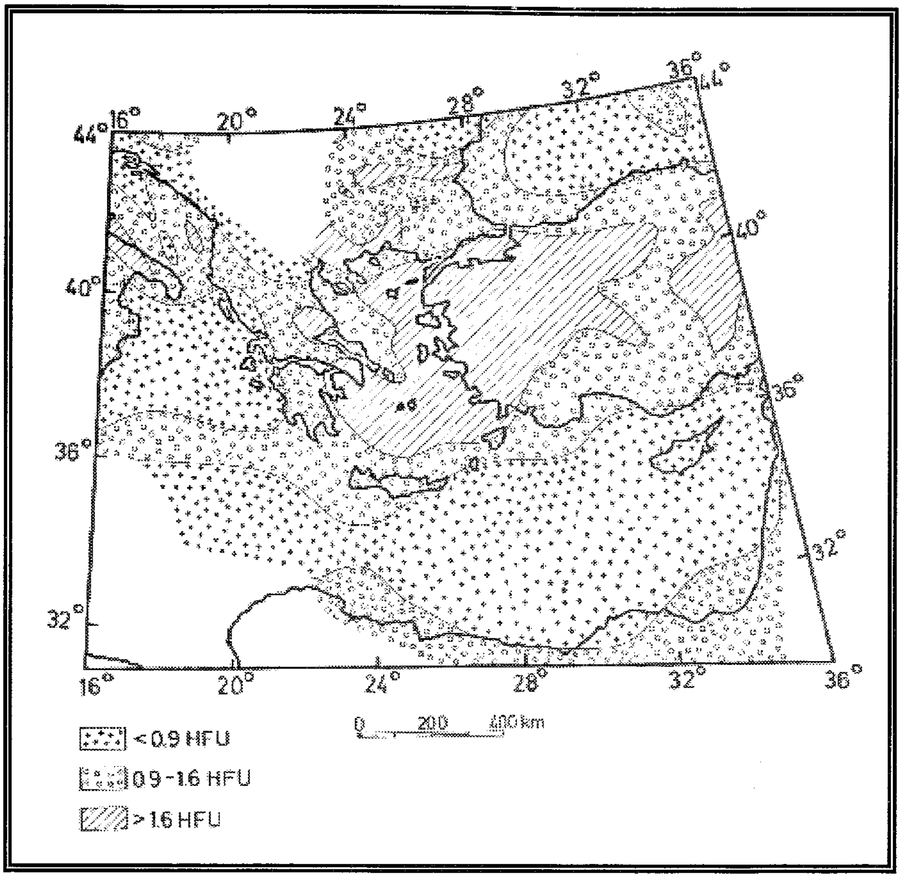

Figure. 2.51. Heat flow map of the Eastern Mediterranean region (after *Makris, 1985*).

Turkey

In Turkey a survey and determination of the chemical composition of thermal springs and mineral waters was first made in 1961. The Mineral Research and Exploration Institute, Ankara, Turkey (MTA) (Çağlar, 1961), published the results. Geological and geophysical systematic studies of geothermal areas were started in 1962 by MTA. They contain more than 600 chemical analyses of hot springs and mineral waters and descriptions of their localities (see *Tezcan, 1979*). A public "Association of Geothermal Energy, Ankara" was founded as late as 1971. A symposium held by this association in 1971 provided a survey of valuable research done in the thermal areas of Turkey (*Demirörer, 1971*). *Tezcan (1979)* published a heat flow map of Turkey using thermal measurements in some bore-holes. In 1975, İstanbul University (İ.Ü.) made thermal water inventory studies in Turkey. In 1979 *Tezcan (1979)* prepared preliminary heat flow maps for Turkey. MTA made thermal water inventory studies in 1980 and investigated all the thermal and mineral water spring groups in Turkey. In 1983

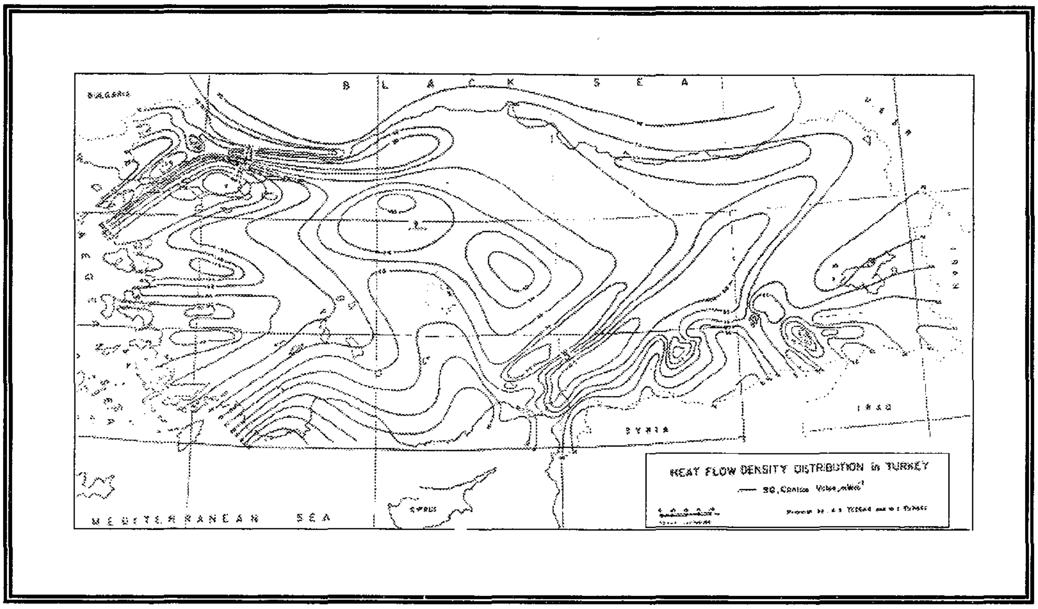

Figure. 2.52. Heat flow density distribution in Tukey (after *Koçak, 1990*).

Canik & Başkan (1983) prepared a map of mineral and thermal waters of western Turkey. Using empirical relationships between the seismic P-wave velocity and the heat production (*Rybach & Buntebarth, 1982*), several heat generation depth sections for different velocity data were determined for western Turkey (see *Alptekin et al., 1990*). These sections suggest that the major portion of heat generation in the crust in western Turkey is in the uppermost 10-15 kilometers (*Alptekin et al., 1990*). *Şimşek (1985, 1988)* investigated total thermal and mineral water spring groups in Turkey. *İlkışık (1989)*, made geothermal studies to find out the heat flow pattern of northwestern Anatolia. *Tezcan & Turgay (1989)* prepared preliminary heat flow maps for Turkey. *Alptekin et al. (1990)* pointed out a relationship between the high heat flow values, the high seismic activity and geological structures in western Turkey. *İlkışık, (1990a)* determined a new heat flow map for western Turkey from silica contents and chemical contents of 187 thermal springs (Fig. 2.53, 2.54). *Koçak (1990)* explained relations between the geothermal systems and the graben structures in western Turkey. *Şimşek (1990)* listed total thermal and mineral water spring groups in Turkey. *Tezcan & Turgay (1991)* displayed a heat flow map of west and central Turkey based on 204 temperature soundings in oil, gas and coal wells and especially in deep geothermal wells. They also made investigations on the terrestrial heat flow density distribution. Rybach & Pfister (*Straub et al., 1992*) have been working on thermal research in the Marmara region (see also *İlkışık, 1995*). The main aim of the interdisciplinary MARMARA Poly-Project of the ETH Zurich (Switzerland) was to study tectonic activity and its relationship to the circulation of normal and geothermal waters, the distribution of zones with elevated seismicity and the spatial heat flow pattern

Figure. 2.53. Heat flow pattern in western Turkey (*İlkışık, 1995*).

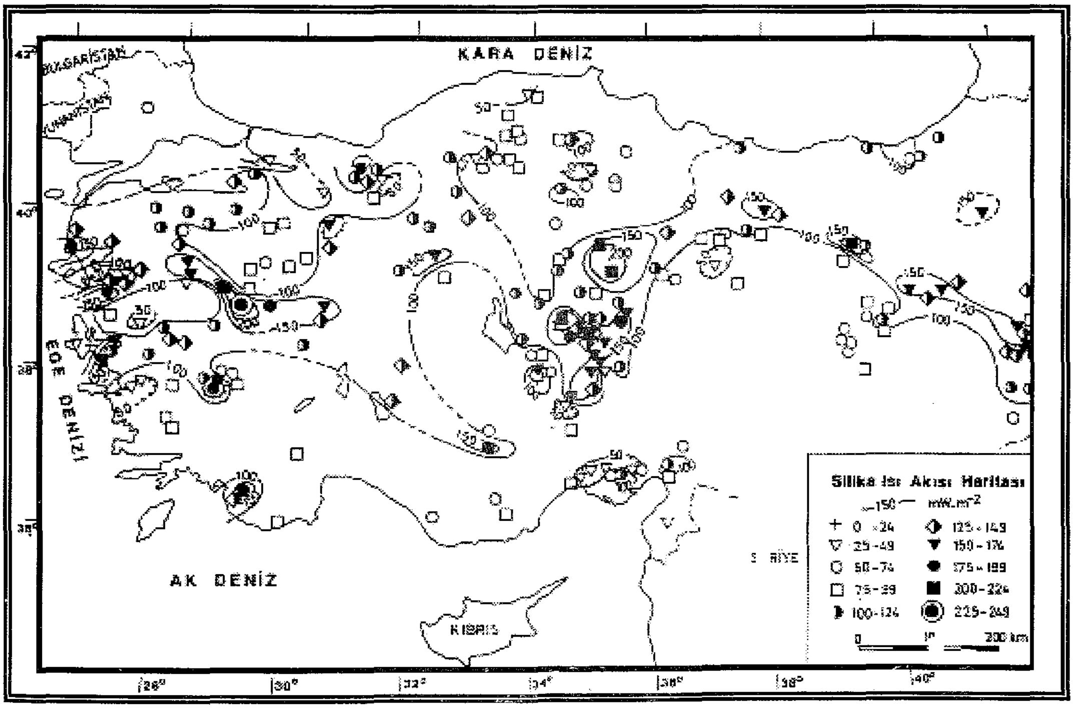

Figure. 2.54. Silica heat flow density distribution in Turkey (after *İlkışık, 1995*).

(*Schindler & Pfister, 1997*). In 1994 MTA Geophysics Department started detailed a heat flow project (Türkiye Isı Akısı Haritası Projesi) to made a detailed heat flow map of Turkey starting from western Turkey. For this project, besides sampling of old and new temperature and gradient logs, the temperature conductivity coefficient (K) of various core samples was measured (see *İlkışık, 1995*). *İlkışık (1995)* and *Tezcan (1995)* made investigations on terrestrial heat flow density distribution. *İlkışık (1995)* critically reviewed the heat flow studies in the Aegean region. *Şimşek (1995)* listed and explained total thermal and mineral water spring groups in Turkey. *Tezcan (1995)* made geothermal studies related to the heat flow in Turkey. *Hisarlı (1995)* determine Curie point depths in Edremit, Susurluk. Using a constant heat conductivity value, he calculated an approximate heat flow map for the area. In Turkey, heat flow values have been collected from thermal measurements taken in deep boreholes (*İlkışık, 1995*). The map of Curie depth points was compared with the gravity, magnetic and geological maps of the area. A seismology group within the MARMARA Poly-Project (1997) recorded hydraulic level changes of the geothermal water system coinciding with the deformation of the earth's crust and seismic signal behavior (*Greber et al., 1997*). *Pfister et al. (1997)* studied the heat flow in the Marmara region to characterize the geothermal situation of this area by detailed regional (terrestrial) heat flow density mapping and thermal spring studies. High-resolution temperature logs and measurements of rock thermal conductivity were used to determine the terrestrial heat flow density and the thermal profiles. The search for available and suitable bore holes was made possible mainly with the help of some national organizations. *Şimşek (1997)* considered hot shallow springs and deep exploration wells in geothermal areas in the Marmara region. He calculated the total potential of natural thermal spring capacity in Marmara. In 1997, Hacettepe University, the International Research and Application Center for Karst Water Resources (UKAM) and the Swiss Federal Institute of Technology (ETH Zurich) signed an agreement concerning "Geothermal investigations in the Marmara region" and "Tectonic activity and its interactions with groundwater circulation, geothermics and seismicity in NW Turkey" under the coordination of the MARMARA Poly-Project. This study also includes a general geothermal potential evaluation of NW Turkey by using various methods and data (see also *Şimşek, 1997*). Recently *Mueller et al. (1997)* compiled a heat flow map of the Anatolian-Aegean region from *Tezcan & Turgay (1991)* and a heat flow map of Europe (*Cermák & Rybach, 1979*) (Fig. 2.55). *Ercan & Kılıçer (1998)* studied the seaward extent of geothermal fields in western mainland Turkey. There are also some local studies concerning geothermal waters, among them *Tchihatcheff (1853), Schliemann (1881), Leaf (1923)* are the earliest authors. *Eşder & Şimşek (1975)*, *İztan &*

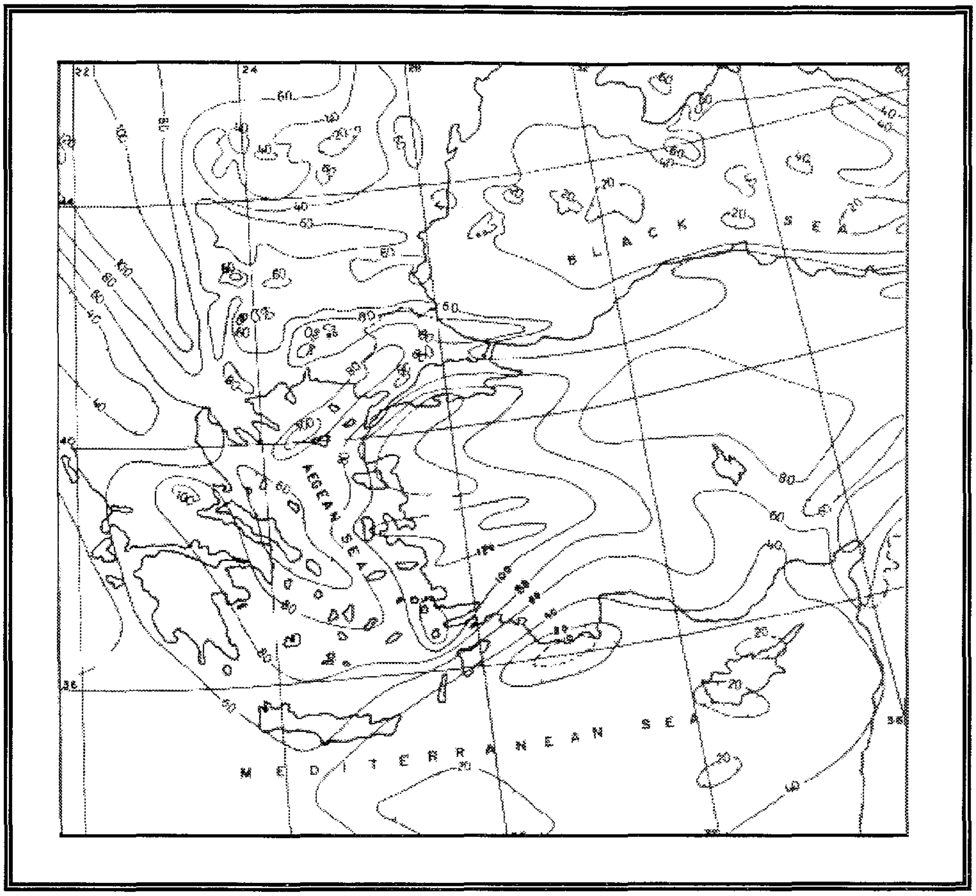

Figure. 2.55. Heat flow in Anatolian-Aegean region (after *Mueller et al., 1997*).

Yazman (1990), *Conrad et al. (1995)*, *Filiz & Tarcan (1995)*, *Eisenlohr et al. (1997)*, *Greber et al. (1997)*, *Yalçın (1997)* and *Şimşek (1997)* are some of the others.

Greece

Fytikas et al. (1976) presented a list of hot springs in Greece. *Fytikas & Kolios (1977)* prepared a heat flow map of the Greek mainland (1:1,000,000), Aegean and Ionian, based on previous heat flow observations obtained in Greece. The whole work is within the frame of the project "Geothermal Energy Investigation in Greece". In the work they have used data that have already been published by other authors on the marine area, and data based on their own measurements and calculations concerning mainly the continental area but also some parts of the marine area.

Aegean Sea and Front of the South Aegean Arc

In the Aegean Sea, first *Erickson (1970)* made the first heat-flow measurement, located in the south of that sea. This measurement is one of the 40 measurements in the eastern Mediterranean. These have been summarized by *Ryan et al. (1969)* and are treated in detail by *Erickson (1970)*. In the Aegean Sea, *Jongsma (1974, 1975)* installed 13 stations, of which nine are continuous. He collected important data, and defined characteristic heat flow structure beneath the Aegean Sea. However, there were not enough stations to find sufficiently reliable results. *Hsu et al.* measured temperature in a drill hole bored by Glomar Challenger 22 (*Fytikas & Kolios, 1977*). In the Aegean Sea, a heat flow value was measured from one of the five boreholes drilled in the eastern and western Mediterranean and Aegean Seas during Leg 42A of the Deep Sea Drilling Project (*Erickson et al., 1977*). This value was compared with 14 conventional values measured previously. *Erickson et al. (1977)* critically reviewed heat flow data from the Mediterranean and Aegean Sea.

2.14.2. General Interpretation

Regional heat flow values give information about the heat flux from the deeper parts of the Earth. There is an inverse relationship between heat flow and the depth of the Moho discontinuity. However, inhomogeneities in rock composition may considerably affect the regional heat flow. For example, the existing fault system, especially normal and strike-slip faults, can be a controlling factor because additional convective heat transport can be assumed here. Large thermal springs occur more frequently, where transtensional and extensional tectonics appear together. This extensional movement tears up the strike-slip faults and causes deep reaching zones of high hydraulic permeability. High permeability permits hot water to rise quickly and to transport heat to the surface from greater depths (*Pfister et al., 1997*). Not surprisingly a close relationship exists between the high heat flow values and the high seismic activity that is a sign of active faults. Nevertheless, the spatial pattern of thermal springs in a region does not coincide with the regional heat flow density distribution (*Pfister et al., 1997*).

Heat flow density is by multiplication of vertical temperature gradient one common method of determining by K (thermal conductivity of the rocks of the specific site). The other method to determine heat flow values that especially from geothermal water contamination of SiO_2.

To prepare a reliable regional heat flow map it is important to know K (thermal conductivity) at each measurement site. If we know K, we can reduce causes that originate from surface and near surface conditions. In continents and also seas, reliable heat flow measurements could just be obtained from deep bore-holes. It is harder to determine reliable results in continents then at sea because of the high probability of ground-water mixing. The neotectonic and Tertiary magmatic (volcanic) activities also provided the basis for significant variations in regional heat flow (*Tezcan, 1995; Şimşek, 1997*). In the Aegean Sea, sedimentation rates, irregularities in sediment cover, sea floor subbottom topography and bottom water temperature variations can cause the heat flow observed near the sea floor and give an additional error to regional anomalies (*Erickson et al., 1977*). Some regional studies show that the region is on a high heat flow zone of Europe (Fig. 2.55). Nevertheless, a detailed heat flow map of the region could not be prepared because of the lack of knowledge about heat conductivity for the estimation sites (e.g., *İlkışık, 1995*). To make reliable estimates, the thermal conductivity coefficient (K) most be known for specific regions. Beside that, knowledge of the ground water circulation system is important to make reliable measurements for a specific site. The data collected on silica contamination of hot springs from western Turkey (*İlkışık, 1990a*) also have some problems (see *İlkışık, 1990a*).

Existing heat flow data indicate a heat flow with a mean value of 2.1 HFU in the northern and central Aegean (*Jongsma, 1974*). Relatively low heat flow values are located in western Greece, in the Ionian Sea and south of Crete (*Erickson, 1970; Ryan et al., 1969; Erickson et al., 1977*), i.e., along the sedimentary arc of the South Aegean Trench (*Fytikas & Kolidos, 1977*) (Fig. 2.55). *Jongsma (1974)* has shown that the mean heat flux through the sea floor is higher than the continental mean and about twice that through the sea floor south of the South Aegean Trench (*Fytikas & Kolios, 1977; McKenzie, 1978*). *Erickson (1970)* had previously taken a heat flow value from a station in the South Aegean, which is one of the Mediterranean stations. The station in the southern Aegean gave a value, which is lower than the worldwide average (*Jongsma, 1975*). Nevertheless, a reliable borehole heat flow measurement that was made in the southern Aegean by Erickson (see *Erickson et al., 1977*) is significantly greater than any of the conventional heat flow measurements (*Jongsma, 1974*) in the southern

Aegean Sea (triangle in Fig. 2.57). The Aegean, because of its shallow average depth, has extremely variable water temperature (*Wüst, 1961; Pollack, 1951*) and it would appear possible that conventional heat flow values (*Jongsma, 1974*) could be seriously affected in the region (*Miller, 1972; Erickson et al., 1977*). However, the general image of the relatively high heat flow must be considered certain. In general the values become higher away from the Trench on the concave side of the arc, similar to other arcs in the World (Fig. 2.57) (*Jongsma, 1975*).

The most outstanding are the areas with high to very high heat flow: the Aegean Sea and western Anatolia (see *Meier et al., 1977*). Studies based on silica content of geothermal waters (*İlkışık, 1990a*) suggest that the mean value of the heat flow for Turkey is 50 per cent higher than the world average (65 mWm) (*İlkışık, 1995*, see also *Alptekin et al., 1990; İlkışık, 1990a*). Western Turkey is characterized by numerous thermal springs (*Tezcan, 1979*). Identified and explored geothermal areas lie mainly along the NE-SW oriented grabens (e.g., Büyük Menderes Graben) (Fig. 2.56) (*Eşder, 1990; Şimşek, 1997; Koçak, 1990*). In western Turkey, at the intersection of grabens, geothermal fields are hotter (*Ercan & Kılıçer, 1998*).

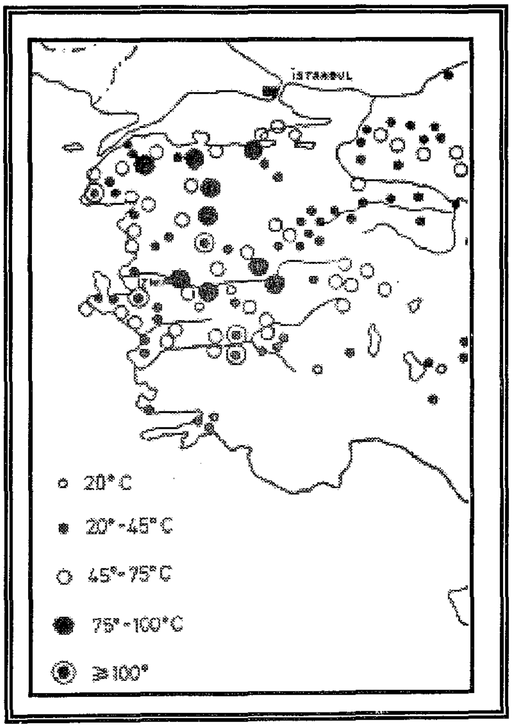

Figure. 2.56. The distribution of hot springs in Turkey (*Tezcan, 1979*).

The tendency to an increase of the geothermal gradients around Hot Springs implies that the regional heat flow also increases in the area around hot springs (*Tezcan, 1979*). However, it should not be unforgotten that the spatial pattern of thermal springs in a region does not coincide with the regional heat flow density distribution (*Pfister et al., 1997*). Northwestern Turkey is also characterized by normal to locally elevated terrestrial heat flow density, compared to a normal value defined by the worldwide continental mean value of 65mWl m (*Pollack et al., 1993*) (see *Tezcan & Turgay, 1991; İlkışık, 1995; Tezcan, 1995; Mueller et al., 1997; Pfister et al., 1997*). The Marmara region (northwestern Turkey) is one of the areas in western Turkey characterized by a high density of geothermal areas (*Eisenlohr, 1997*). Thermal springs are a very common feature of the region. More than 90 hot springs, up to 19 gradient, 24 shallow and 5 deep exploration wells in geothermal areas show the enormous potential of the Marmara region (*Şimşek, 1997*). The high values are concentrated over the NAF zone and extensional faults (*Alptekin et al., 1990, Eisenlohr, 1997, Pfister et al., 1997*). On the other hand, normal heat flow values occur in the northern and eastern part of the area (Bursa, İznik, İzmit and Thrace region) (*Crampin & Evans, 1986; Straub & Kahle, 1997; Pfister et al., 1997*).

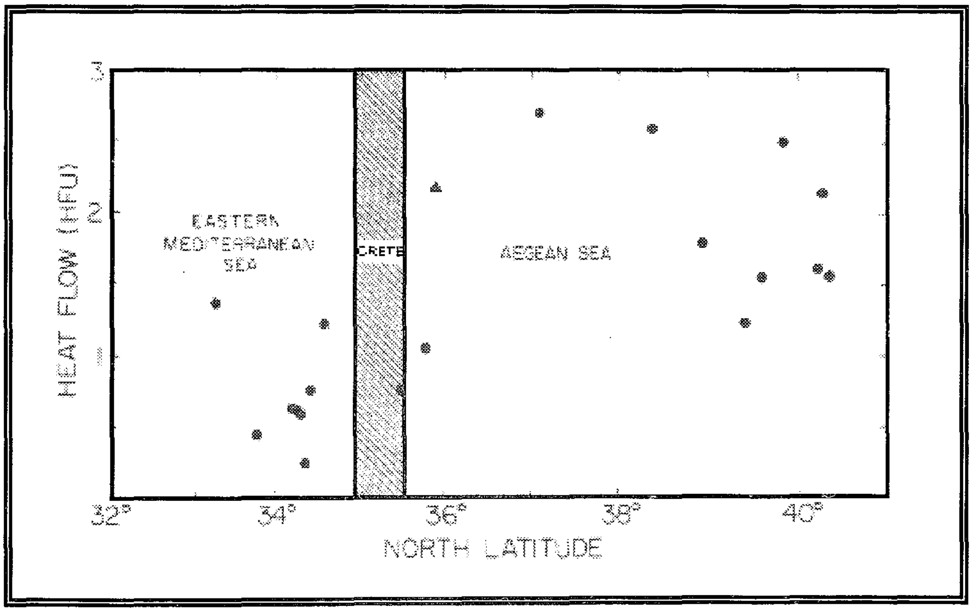

Figure. 2.57. Heat flow values between 20° and 30°E longitude as a function of latitude, showing the abrupt increase in heat flow northward from the Mediterranean across the South Aegean arc into the Aegean Sea (after *Erickson et al., 1977*). Circles show the values obtained using conventional occeanographic techniques; a single triangle shows a value measured in a borehole drilled during the Deep Dea Drilling Project Leg 42A.

He (Helium) anomalies in western Turkey and the presence of 3He enrichments in the crust of western Turkey, are more than expected from normal crustal lithologies (*Koçak, 1990*). He suggests that they have been produced from a mantel source (*Koçak, 1990*). Similarly, Boron found in thermal and mineral waters probably originated from the mantle or during metamorphism from marine sediments. Most of the isotope analyses of C02 gases in thermal and mineral waters in the Gediz Basin also indicate a mantle origin (*Filiz & Tarcan, 1995*). In conclusion, the formation of geothermal areas mainly depends on the mentle source, which means that the heat generated in the deeper parts is transferred by volcanism to a shallower depth in the crust (*Koçak, 1990*).

2.15. TOMOGRAPHY AND SEISMOLOGICAL STUDIES

2.15.1. Data

Moskalenko (1966) used body-waves and surface-waves for calculation of the seismic velocities as well as crustal structure. Travel-time curves for the most prominent phases of the longitudinal (P) and shear (S) waves have been constructed for southeastern Europe by *Papazachos et al. (1966)* by using data for about 1000 earthquakes. The velocity structure of the crust and upper mantle has been studied using dispersion of surface-waves by *Papazachos et al. (1967). Molnar & Oliver (1969)* studied lateral variations in the attenuation of Sn waves in the Aegean region. *Papazachos (1969)* estimated shear-wave (S) velocity layer from phase velocities of Rayleigh waves. *Payo (1967, 1969)* used a large number of observations of phase velocity of Love and Rayleigh waves to determine the crust-mantle structure in the Mediterranean region. *Canıtez (1969)*, studying the Love and Rayleigh waves, found a crustal thickness for the northern and southern Aegean. *Papazachos & Comninakis (1971)*, using travel time residuals of intermediate shock, showed a striking anomaly along the South Aegean Arc (SAA). *Crampin & Üçer (1975)* investigated seismic velocities beneath the Marmara sea by examining four earthquakes recorded at 35 seismic stations. They found the crustal thickness beneath the Marmara Sea. *Canıtez (1975)* determined an average crustal thickness and an uppermost mantle P-wave velocity for the northern Aegean Sea, using dispersion of surface-waves. Teleseismic P-wave travel time residual data for distant earthquakes was observed at 7 Greek seismic stations by *Agarwal et al. (1976).* Their data, however, do not allow a conclusion on the depth of slab penetration (*Papadopoulos, 1997*). P wave seismic travel-time (residual) data for Greek earthquakes observed at distant stations have been interpreted in terms of a dipping plate under the island of Karpathos near 36°N 27°E by *Gregersen (1977). Kenar (1977, 1978)*, using three component seismograms recorded by stations of İTÜ, determined a crustal thickness beneath İstanbul from the spectral ratios of P waves. *Jacoby et al. (1982)* analysed travel time residuals beneath the dipping slab. *Papazachos & Comninakis (1978)* studied recordings of earthquakes from 6 stations in Greece. *Ezen (1979)* examined surface waves that sourced from the South Aegean and was recorded by station of İTÜ. He interpreted qualitatively the group velocities. *Chen et al. (1980)* determined a Pn velocity beneath Turkey from travel time-distance relations. *Canıtez*

& Toksöz (1980) calculate absolute or relative station residuals to determine crustal thickness for the Aegean region. *Romanowicz (1980)* applied tomography techniques to eastern Europe. Using all available long-period Rayleigh wave observations in Europe, *Panza et al. (1980)* and *Suhadolc et al. (1990)* outlined the thickness of the lithosphere in the Europe-Mediterranean region. *Panza et al. (1980)*, using all available long-period Rayleigh wave observations in Europe, outlined the thickness of the lithosphere in the eastern-Mediterranean region. *Hovland & Husebye (1981)* obtained tomographic results on the Upper mantle structure in the northern Aegaen. They used an independent data set consisting of teleseismic P-wave delays of events. *Necioğlu et al. (1981)* determined Pn velocities at permanent earthquake stations in western Turkey. Seismic travel times of teleseismic P-waves and station residuals were used to study the crust/upper mantle structure and its lateral variations in Turkey (*Canıtez & Toksöz, 1982*). They used P-wave travel time residuals from 125 earthquakes received at 18 stations. The travel times were taken from ISC Bulletins. These data are complemented by the available phase and group velocities of surface-waves (*Canıtez & Toksöz, 1982*). The velocity structure of the crust and upper mantle beneath the western Aegean region has been studied using dispersion of surface-waves by *Calganile et al. (1982)*. *Jacoby et al. (1982)* calculate absolute or relative station residuals to investigate changes of the crustal thickness or upper mantle velocity for the Aegean region. Teleseismic tomography has been applied by *Hovland & Husebye (1981)* to eastern Europe. In this study more attention is paid to the Aegean region by including 15 stations from Turkey, Greece, Bulgaria and former Yugoslavia. *Delibasis (1982)* discussed seismic wave attenuation (velocity structure) of the upper mantle beneath the Aegean region. *Ezen (1983)* examined Rayleigh wave propagation beneath the Aegean. A Q-value was obtained in the region by *Tassos (1984)*. *Panagiotopoulos & Papazachos (1985)* calculated Pn velocities based on 23 accurately located very shallow earthquakes (h=1-14 km) in northern (1978) and central (1984) Greece, in central and eastern Greece, south-east of the former Yugoslavia, the Aegean sea, Bulgaria, southern Romania and western Turkey. For Europe and the Mediterranean region an estimate of the 3D-mapping of the earth's seismic P-wave velocity structure from the surface down to a depth of 800km exists, derived from detailed investigation of delay-time tomography (*Spakman, 1985, 1988; Spakman et al., 1988, 1993*) (Fig. 2.58). The dipping high-velocity zone is recognised by *Spakman (1985, 1986)* as the blurred image of the African slab that penetrates deep into the Aegean Upper Mantle. *Spakman (1986)* argued that his data had high enough spatial resolution to permit a significant interpretation. *Spakman & Nolet (1987)* argue that the outline of the slab is resolved with a spatial error between 50-100 km

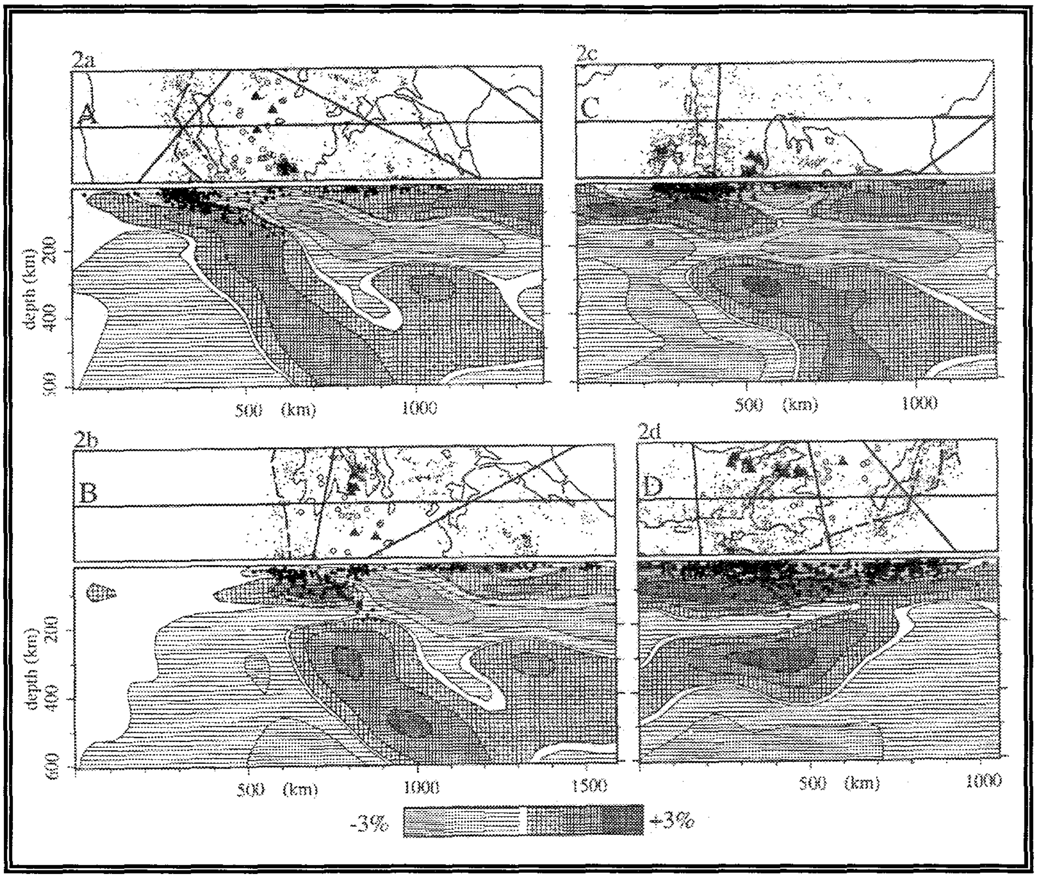

Figure. 2.58. Tomographic images of the Aegean/Eastern Mediterranean Upper Mantle in cross section. The upper panels in a-d display a small location map for orientation (after *Spakman et al., 1988*). The lower panels display the inferred P-Wave velocity heterogeneity. Cross (horizontal) hatching indicates positive (negative) anomalies. Regions of poor spatial resolution are not contoured (large white areas). The horizontal and depth axes are given in km, without vertical exaggeration. Black symbols indicate the projection of hypocenters with M>4 that are located 100 km from the plate. The width of the location map is 3 degrees. For more detail see *Spakman et al. (1988)*.

and that the signs of the velocity anomaly amplitudes are well resolved. The tomograms have better resolution than the image of *Hovland & Husebye (1981)*. Because *Hovland & Husebye (1981)* used an independent data set consisting of P-wave delays of events from outside of this region, whereas *Spakman (1986)* used data coming from events in the Mediterranean region and over half a million teleseismic delay time data reported by ISC. In a more recent study by *Spakman et al. (1993)*, the volume under study encompassed the mantle to a depth of 1400 km and the number of ISC data for inversion increased by a factor of four including data from teleseismic events. *Spakman (1991)* used the same technique for the delay time data selected from the ISC database, from regional events and from stations up to epicentral distance are used. About 18,000 events and 937 stations were used. *Kalafat et al. (1987)* determined the thickness of the crust beneath western Turkey from the travel times by using the Pn phase readings of the stations operated by KOERI in SW Turkey. *Drakatos et al. (1988, 1989)* and *Ligdas & Lees (1993)* made local scale tomographic studies in various parts of the Aegean region. *Ezen (1988, 1991a, b)* studied the effect of crustal thickness in western

Turkey on Rayleigh-wave records. Three-dimensional seismic attenuation has been investigated beneath the Aegean region by *Hashida et al. (1988)*, using the tomographic technique. They found that high-Q values correspond to the distribution of intermediate-depth earthquakes across the SAA, and that low-Q spots near Athens and in the southeastern Aegean sea correspond to volcanoes. *Christodoulou & Hatzfeld (1988)* used a three dimensional inversion technique of the teleseismic travel time residual in the Chalkidiki (Thessaloniki) network to determine velocities for the crust and mantle lithosphere and crustal thickness beneath the Chalkidiki peninsula (Macedonia). *Drakatos (1989)* applied a tomographic technique to determine crust and upper mantle of the region. *Mindevalli & Mitchell (1989)* used the group velocities of Rayleigh and Love waves to determine crustal velocities and thickness in western and eastern Turkey. *Granet & Trampert (1989)* investigated the 3-D velocity structure beneath the Euro-Mediterranean domain down to 1200 km by using teleseismic P-wave arrival times.

Yılmaztürk (1989) used Pn arrivals at seismic stations in the Aegean region, taken from the ISC bulletins, to investigate velocity variations beneath the Aegean region. *Alptekin et al. (1990)* used the observed group velocity of fundamental mode Rayleigh waves from twenty earthquakes (Mb > 4.5) generated in the Burdur region, to find the crust layer and Pn wave velocity between İstanbul and Burdur. They found on optimum structural model for western Anatolia. *Suhadolc et al. (1990)* using all available long-period Rayleigh wave observations in Europe, outlined the thickness of the lithosphere in the eastern-Mediterranean region. *Ligdas et al. (1990)* used P-wave travel time residuals and seismic tomographic techniques to better illuminate the velocity structure of the crust and upper mantle beneath the broad Aegean region (*Ligdas & Main, 1991;* see also *Ligdas, 1990*) (Fig. 2.59). They included teleseismic records between 1976 and 1986 from local stations, which do not routinely report to the ISC, and, therefore, the database for the inversion is substantially different from that of *Spakman et al. (1988)*. Three-dimensional images of the upper mantle in the central Mediterranean were obtained using tomography techniques by *Drakatos & Drakopoulos (1991)*. The data used are arrival times of P-waves of local events. *Ezen (1991b)* determined the thickness of the crust along the N-S direction in western Turkey. *Ezen (1988, 1991a,b)* studied the effect of crustal thickness in western Turkey on Rayleigh-wave records. *Osmanşahin (1989)* used the same method to study the crustal thickness of the Aegean Sea. The most comprehensive study so far of the 3-D S-wave velocity field below Europe based on delay-time and waveform inversions has been carried out by *Zielhuis (1992)*. *Küleli (1992)*

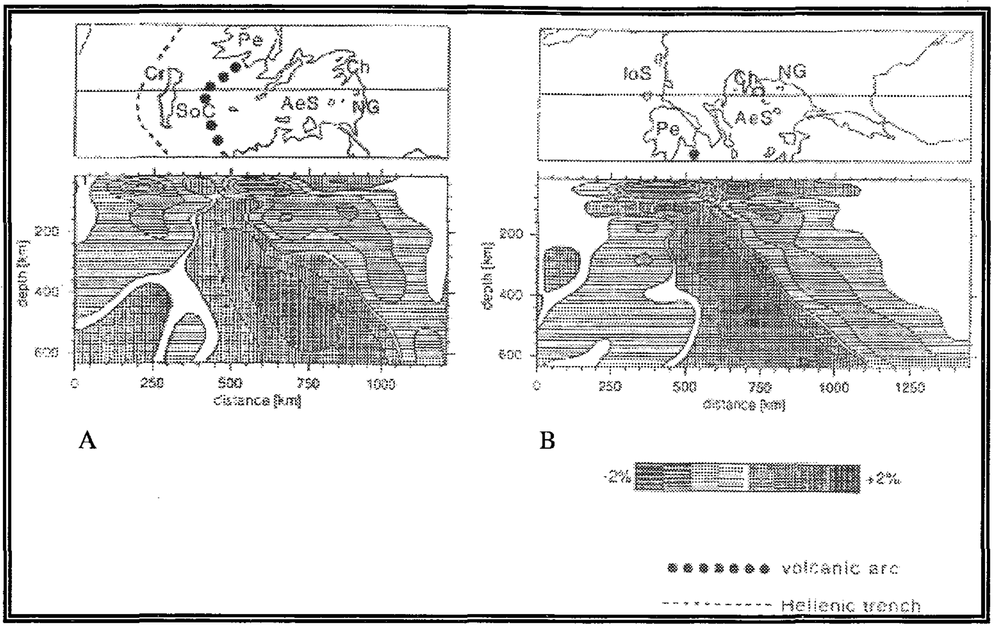

Figure. 2.59. Tomographic images of the Aegean area (after *Ligdas et al., 1990*). Shading is as Fig. 2.58.

using P-wave arrivals applied inversion of teleseismic data from a 23 year interval (1964-1986) for teleseismic and local data events obtained from the ISC and mapped tomographic images in the Aegean region (Fig. 2.60) (see *also Küleli et al., 1995*). *De Jonge et al. (1993)*

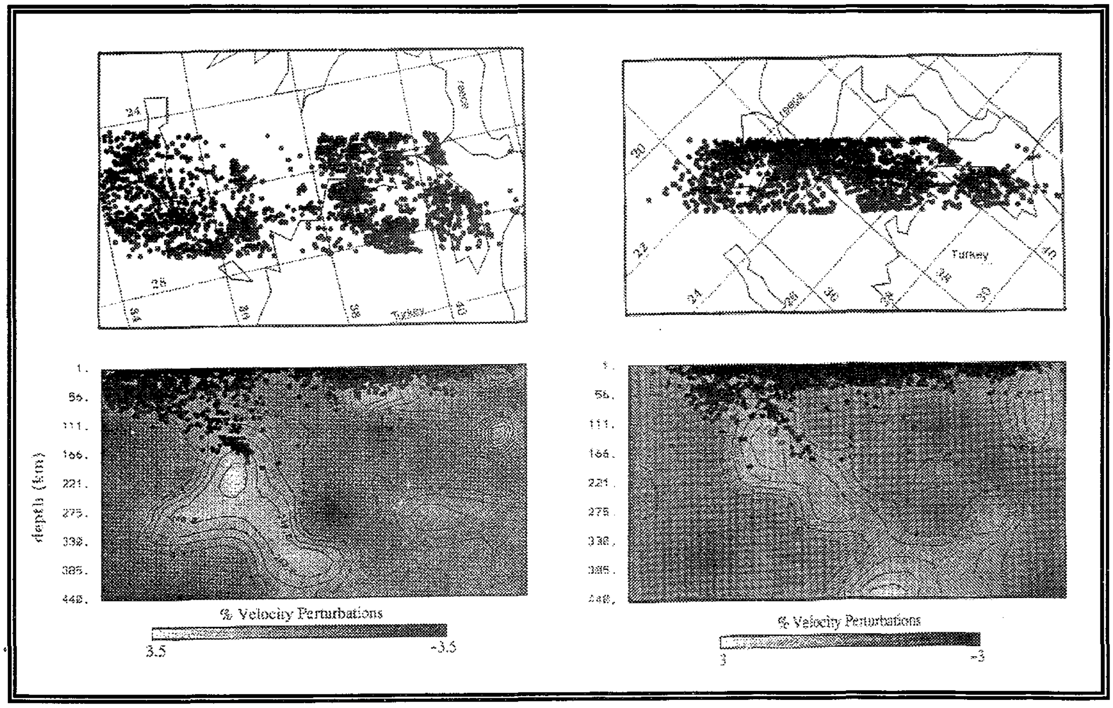

Figure. 2.60. Tomographic images of the Aegean area (after *Küleli, 1992*). The upper panel displays locations of the sections and location of epicenters (M>3) which were included in the tomographic images.

studied the seismic tomography of the region. *Blanco & Spakman (1993)* estimated the present P velocity structure from delay time tomography. The results describe the three-dimensional structure of the lithosphere and mantle from the surface down to a depth of 1400 km. *Kalogeras et al. (1993)* determined the mean dispersion curve for a Dodecaoese to Athens propagation path. The Pn tomography study of *Hearn & Ni (1994)* provides the best estimates of sub-Moho P-wave velocities for the region (see *Rodgers et al., 1997*). Previously, *Akıncı et al., (1994)* studied the seismic attenuation in western Anatolia with limited data sets by using the coda wave nethod. The P-wave tomographic velocity structure of the Tyrhennian (Calabric) and the South Aegean subduction zones were calculated by *De Jonge et al. (1994)*. *Papazachos (1994)* made tomographic studies in the region for the crust and upper mantle. *Alessandrini et al. (1995)* described the P-wave velocity structure of the crust and uppermost mantle of the Aegean and Italian regions obtained through the tomographic inversion of arrival times of regional events. *Horasan & Canıtez (1995)* modelled the crustal structure of western Turkey using complete synthetic seismograms. *Papazachos et al. (1995)* used more than 100,000 P-wave arrivals from 4,229 earthquakes in the Aegean region for the determination of the detailed structure of the crust and the upper mantle in this area. The data they used in the study come from the annual bulletins of the Geophysical Laboratory of the University of Thessaloniki (GLUT) for the time period 1981-1987 for events with surface-wave magnitude 2-3.5. They added data from the ISC bulletins as relocated for the period 1971 -1980 for events with surface-wave magnitude M>5. The original data came from the preliminary monthly bulletins of the National Observatory of Athens (NOA), GLUT, and seismological laboratories of neighboring countries (Italy, Albania, Yugoslavia, Bulgaria and Turkey). The data were enriched with observations from four local experiments in the Thessallaniki basin (northern Greece), Peloponnesus (southern Greece), southern Aegean, and Epirus (northwestern Greece). These data were collected during the summers of 1985, 1986, 1988, and 1989 by the Geophysical Laboratories of the Universities of Thessaloniki and Athens (LGIT) and the Laboratoire de Geophysique Interne et Tectonophysique of the University of Grenoble (France) (*Hatzfeld et al., 1990, 1993; Hatzidimitriou et al., 1992*). *Taymaz (1996)* used S-P wave travel time residuals, which confirmed the lateral velocity variation in the South Aegean region. He found average S and P-wave travel time residuals from 29 earthquakes in the Aegean and the South Aegean Trench near Crete. To reduce the uncertainties in the residuals, *Taymaz et al. (1990)* used synthetic seismograms of P and SH waveforms to constrain depths of earthquakes in the South Aegean Trench near Crete. Recently, a relationship was inreferred between the deep velocity anomaly that extends from beneath southern Asia and the anomaly beneath the Aegean sea by *Widiyantoro & Van der Hilst (1996)*. An analysis of Rayleigh waves (inversion techniques applied to the dispersion of Rayleigh waves) generated by earthquakes in the broader Aegean was undertaken to estimate variations in the S velocity structure in the crust and upper mantle down to about 70 km depth by *Kalogeras & Burton (1996).* The three-dimensional attenuation structure beneath the Aegean sea and the surrounding regions was determined by inversion of seismic intensity data by *Stavrakakis et al. (1997).* In the Mediterranean basin a tomographic study was made by *Martinez et al. (1997)*, based on the fundamental mode of Rayleigh waves. This database was formed from regional and teleseismic events recorded from 1990 to 1993 at very-broadband stations belonging to MEDNET and other cooperative stations, located in the Mediterranean area. *Papadopoulos (1997)* discussed seismic tomography images, presented by other authors' tomography images in the light of an alternative interpretation. Regional seismic tomography of Iberia, Italy, the South Balkans and the Aegean region down to about 400 km are discussed by *Plomerova (1997)* along with results of studies on the anisotropic structure of the lithosphere based on an analysis of spatial variations of P-wave residuals. The three-dimensional velocity structure of the crust beneath the Aegean sea and the surrounding area is investigated by inversion of about 10,000 residuals of arrival times of P-waves from local events reported by the ISC by *Drakatos et al. (1997).* The crustal velocity structure of a 50x50x20 km domain near Patras in the Gulf of Corinth (Greece) was recovered by *Le Meur et al. (1997)* using microearthquake data. These data were recorded during a field experiment in 1991 where a dense network of 60 digital stations was deployed. *Papazachos & Nolet (1997)* presented the detailed characteristics of the P and S velocity structure of the South Aegean Subduction and the Aegean Sea in crust and upper mantle scale in the southern Aegean from the inversion of travel times of local events. Sources of data are the annual and monthly bulletins of GLUT for the period 1980-1994. Many of these travel time data have also been published in the ISC bulletins. A large portion of the data, which comes from the monthly bulletins of seismological observatories of Greece and neighbouring countries, was not routinely reported to ISC but is included in the GLUT final data set of travel times. For the period 1980-1994 almost 4,300 local events were used. *Ekström et al., (1997)* used a large number of globally recorded Rayleigh and Love waves to develop deep structures. *Alessandrini et al. (1997)*, by using Pn, found the regional scale characteristics of the seismic wave velocity in the lithosphere (crust and upper mantle) for the Ionian and Aegean Sea. They used ISC catalogue data for earthquakes from 1964-1992 in the area. All the first P-wave arrivals used were recorded by seismological stations operating in a wider area. *Küleli et al. (1995)* investigated seismic wave velocity distribution in the crust and uppermost mantle by using travel times of P-wave data from earthquakes in western Turkey. They selected Pg and Pn phase arrival times and epicentral distances from ISC bulletins for a 6-year (1981-1986) interval and also used travel times. *Zielhuis & Nolet (1994)* made a three-dimensional model for the S-wave velocity in the upper mantle below Europe. As a result of this modeling Precambrian cratones were found. Using the global body wave tomography technique, *Van der Hilst et al. (1997)* resolved long narrow structures in the lower mantle, some of which can be followed to sites of present-day plate convergence at the Earth' s surface including the Aegean subduction zone. *Rodgers et al. (1997)* presented propagation characteristics of the Sn and Lg phases for the Middle East and Mediterranean (Fig. 2.61). *Horasan et al. (1998)* used a large set of seismic events and stations to determine the average attenuation in the region by using the coda wave method and the coda normalization method and interpreted the results obtained

2.15.2. General Interpretation

Essentially, there are 2 types of seismic waves that radiate from a source (fault):

1) Body waves (P and S)
2) Surface waves (RAYLEIGH and LOVE)

The dispersion of seismic surface-waves is widely used in studying the structure of the Earth's crust. It provides an estimate of the average properties of the crust between the epicentre of the earthquake and the recording station. Differences in travel times, amplitudes, and frequency content of regional phases (Pn, Pg, Sn, and Lg) have proven to be very useful in characterising structural variations in the lithosphere (crust and uppermost mantle). Pn is the first arriving phase at regional distances. For this reason, Pn travel times are sensitive to the uppermost mantle. Seismic tomography, which is a three-dimensional imaging method in Seismology by inverse solution, has been proved an efficient tool to image the interior of the Earth. Since the well-known works of *Aki & Lee (1976)*, or *Crosson (1976)*, many authors have used first-arrival times to recover the 3D seismic velocity structure as well as hypocentre coordinates at local, regional and global scales. In the linearised approach the method inverts for the evaluation of the velocity field from a radially symmetric reference earth using the difference between the observed and calculated travel time (the so-called delay time of travel time residual). Total number of earthquakes, degree of intersection of crossing rays, size of array, position of stations, epicentral distance range, size of cells as well as the inversion

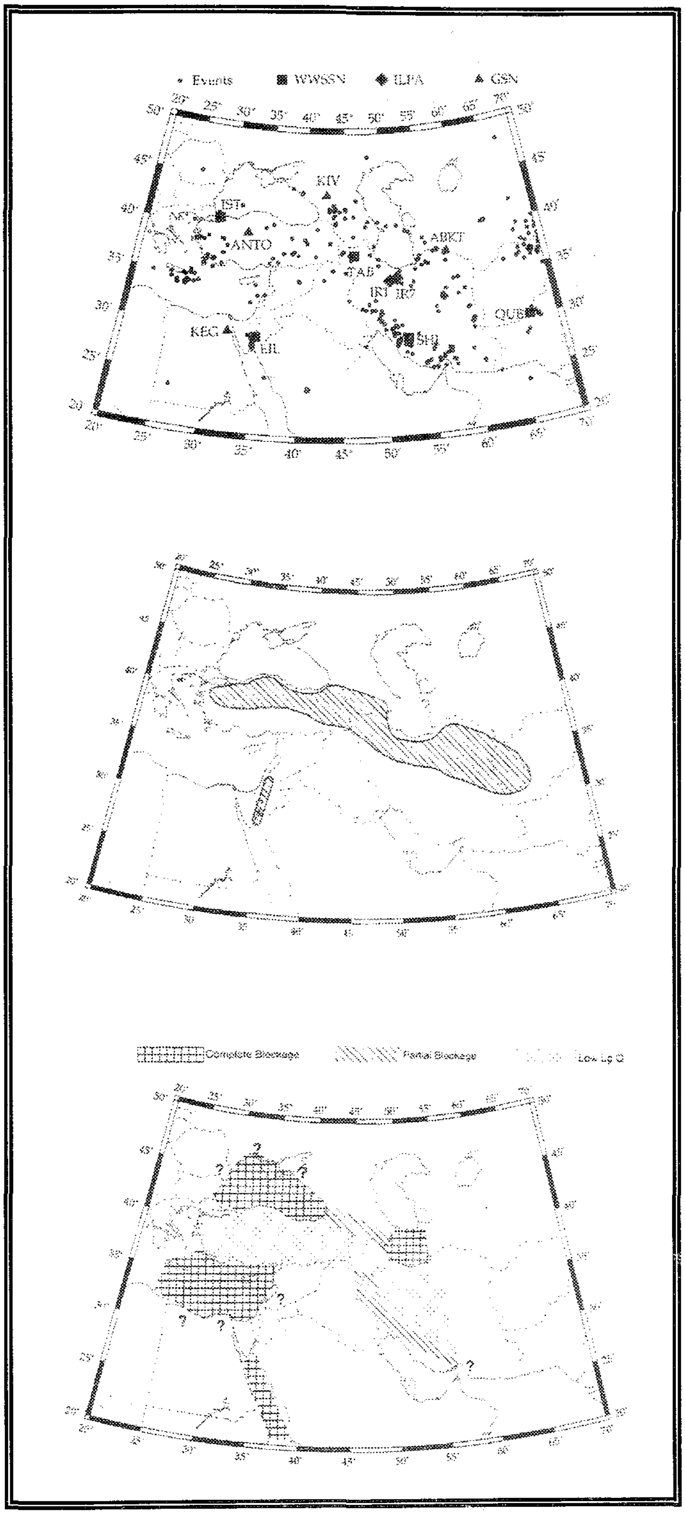

Figure. 2.61. Sn and Lg propogation in the Middle East (after *Rodger et al., 1997*). (a) Events and stations used in the study (b and c). (b) Zones of inefficient Sn propogation (see *Rodger et al., 1997* for detail). (c) Characteristics of Lg propogation.

method applied are the most important factors influencing the reliability of the 3D tomographic images. Teleseismic rays illuminate the deep mantle structure but the resolution in tomography decreases with depth when teleseismic data are used (*Drakatos &*

Drakopoulos, 1991). The tomographic images suggest a close relationship between seismic wave (P, S) velocity pattern and a subduction system. Before thermal assimilation, older subducted oceanic lithosphere exhibits lower temperatures than the surrounding mantle. Hence such subducted slabs will be imaged as high velocity zones in tomography. The upper mantle of the European-Mediterranean region has been the subject of many recent tomographic studies which the aim to determine the velocity-depth distribution of seismic P and S waves in an attempt to obtain quantified images of the lithosphere-asthenosphere system.

A fundamental application of seismology is to study the velocity distribution of the Earth's interior. The Earth's crust may be considered as a layered medium. Each layer has different physical properties, and is influenced by temperature and pressure, which causes seismic wave velocity variations, related to depth. It is not so expensive to get information on the travel times from earthquake data as from DSS, because in this case the energy is offered by nature and the recording permanent seismic stations already exist. However, these travel times are not accurate because the origin times and the foci of the earthquakes are not known with sufficient accuracy, unlike in the DSS method (see section, 2.16). To get better results the number and coverage stations should be improved.

Sn waves in the concave part of the South Aegean Arc (*Molnar & Oliver, 1969*) indicate that the thickness of the lithosphere beneath the inner part of the arc is rather small. Sn is a seismic shear-wave that propagates in the uppermost mantle. It propagates very efficiently across the stable regions of the earth, the continental shields, and deep-ocean basins, but propagation is very inefficient when paths cross the crests of the mid-ocean ridge system or the concave sides of most island arcs. If low attenuation correlates with high strength, the data imply that the uppermost mantle is considerably weaker under the ridge crests and the concave sides of the island arcs than it is elsewhere (*Molnar & Oliver, 1969*). Like Pn, Sn predominantly travels within the uppermost mantle (see e.g., *Stephens & Isacks, 1977*); however, because it is a shear-wave, it is more sensitive to attenuation. Sn generally does not propagate across regions of high heat flow such as spreading centres and back-arcs (*Beghoul et al., 1993*). Lg does not propagate in ocean crust (*Press & Ewing, 1952; Zhang & Lay, 1995*). Accordingly Lg propagates efficiently for the events from the western coast of Turkey (*Rodgers et al., 1997*). Sn propagates efficiently across East Mediterranean (*Molnar & Oliver, 1969; Kandinsky-Cade et al., 1981; Rodgers et al., 1997*). Concordantly Lg propagates inefficiently around the southwestern coast of Turkey (*Rodgers et al., 1997*). These indicate that off the southwestern coast of Turkey beneath the Mediterranean an oceanic type of crust exist.

Earlier *Jacoby et al. (1982)* found a high-velocity slab from the analysis of travel time residuals. In the upper mantle a relevant high velocity anomaly extends below the Greek mainland and the SAA (*Hashida et al., 1988; Drakatos & Drakopoulos, 1991; Babuska et al., 1987; Zielhuis, 1992; Alessandrini et al., 1997; Plomerova, 1997; Drakatos et al., 1997; Papazachos & Nolet, 1997*). A high-Q anomaly also characterises the subducting African slab that represents ongoing subduction (*Hashida et al., 1988; Stavrakakis et al., 1997*).

Between the overridden crust and subducting slab, low velocities are observed at shallow depths in the upper mantle beneath the Aegean back-arc region (e.g., *Jacoby et al., 1982; Spakman, et al., 1988; Ligdas, et al., 1990; Küleli, 1992; Kalogeras & Burton, 1996; Alessandrini et al., 1997; Plomerova, 1997; Drakatos et al., 1997*). The average, S-P residuals from earthquakes in the North and central Aegean are nearly all positive (late arrivals) and some are large (> 3 s) in the crust and upper mantle (*Taymaz, 1996*). This is also supported by the low Q-value in the region (*Tassos, 1984; Hashida et al., 1988; Stavrakakis et al., 1997*). A low velocity zone dominates especially in the inner part of the Aegean volcanic arc and absorbs the greater part of seismic energy (*Drakatos & Drakopoulos, 1991*). This anomaly is associated with the slab dehydration that is well developed under the volcanic arc, especially around a depth of 60 km (*Papazachos & Nolet, 1997*). The low velocity is also observed in western Turkey (*Yılmaztürk, 1989; Drakatos & Drakopoulos, 1991; Drakatos et al., 1997; Stavrakakis et al., 1997*). The NW of Turkey, NAF is recognised also by low Pn velocities (*Yılmaztürk, 1989*). An area of low velocities is also observed, trending parallel to the North Aegean Trough (NAT) (*Papazachos et al., 1995*). These anomalies indicate weak crust. Low-Q zones were obtained (20-40 km) particularly in the southeastern part of the South Aegean Trench because of the thick sediment series in the trough (*Stavrakakis et al., 1997; Hashida et al., 1988*).

The Pn travel time residuals of local earthquakes were used by *Gregersen (1977)*, who interpreted residual anomalies between Crete and Rhodes as indicating a high velocity plate dipping underneath the Aegean arc. *Romanowicz (1980)* used tomography techniques to find a very strong negative residual anomaly at stations in Greece, which represents the presence of a subducting slab of the African plate. *Spakman (1985, 1986), Meulenkamp et al. (1988)*

Spakman et al. (1988), Granet & Trampert (1989) and *De Jonge et al. (1994)* obtained three dimensional images of the upper mantle in the central Mediterranean using tomography techniques. They suggested that the Aegean subduction zone is characterised by a dipping high-velocity zone penetrating to at least 400 km depth and possibly as far as 600 km. *Spakman & Nolet (1987)* argue that the slab's outline is resolved with a spatial error between 50-100 km. Also from the surface to a depth of 400 km *Ligdas et al. (1990), Ligdas & Main (1991)* and to a depth of 480 km *Küleli (1992)* observed the deepening high-velocity zone. *Ligdas et al. (1990)* cautioned that for their tomogram the decreasing resolution with depth should also be taken into account in the interpretation of these results for depths greater than 170 km under mainland Greece and the Aegean Sea (*Ligdas et al., 1990; Ligdas & Main, 1991*). *Van der Hilst et al. (1997)* found that the slab seems to be continuous to approximately 1500 km depth, which is in agreement with the results from a regional study of *Spakman et al. (1993)*. Beneath eastern Europe, the Aegean slab does not seem to have sunk deeper than about 1500km (*Van der Hilst et al., 1997*). In that study, the depth to the leading edge of this slab is well determined (*Van der Hilst et al., 1997*). The interpretation of large-scale positive anomalies by *Spakman et al. (1993)* indicated a slab anomaly down to depths of at least 670 km, possibly down to 800-900 km. They suggested that if the Aegean subduction relates to the destruction of one oceanic basin (the eastern Mediterranean part of the Tethys ocean) then the duration of subduction may possibly be extended to even more than 40 My. *Ligdas & Main (1991)* used a data set of well-controlled local and teleseismic ray paths (*Ligdas, 1990; Ligdas et al., 1990*) to put some bounds to the depth of penetration of the slab. They concluded that the peak amplitude of the tomographic image of a lithospheric slab is found from the inversion of travel-time data to be at depths at or below 400 km. However, no constraints on the maximum depth of penetration could be established with their data set. *Hovland & Husebye (1981)* also imaged a high velocity anomaly beneath the central Aegean down to 550 km, the deepest significant level of mapping, but they did not link this anomaly to The South Aegean Subduction. *Spakman et al. (1988)* and *Granet & Trampert (1989)* found a low velocity zone on a high velocity slab down to 120 km that possible represents a detachment zone along the slab.

2.16. DEEP SEISMIC SOUNDING

2.16.1. Data

Mediterranean

The first seismic-refraction measurements in the eastern Mediterranean were made during the round-the-world cruise of HMS Challenger in 1950-1952 (*Gaskell et al., 1958*). *Moskalenko (1966)* reported three refraction profiles, supplemented by deep-reflection profiles, between Crete and Libya. Seismic refraction surveys in the Ionian (*Ferrucci et al., 1991*) and Levantine (*Ginzburg & Ben-Avraham, 1987*) basins were made to determine the type of the crust in the East Mediterranean. Two expanding spread profiles (ESP) (two-ship refraction and oblique reflection seismic) from the recent Pasiphae cruise (1988 November 28 to December 19) were carried out in the eastern Mediterranean Sea (Fig. 2.62) (*De Voogd et al., 1992*). Results of the cruise provide a well-constrained velocity model for the deep basins of the eastern Mediterranean in the Ionian basin and Sirte abyssal plain down to Moho depth (*De Voogd et al., 1992*). STREAMERS made a study in Ionian Sea (*Hirn et al., 1996*) (Fig. 2.63)

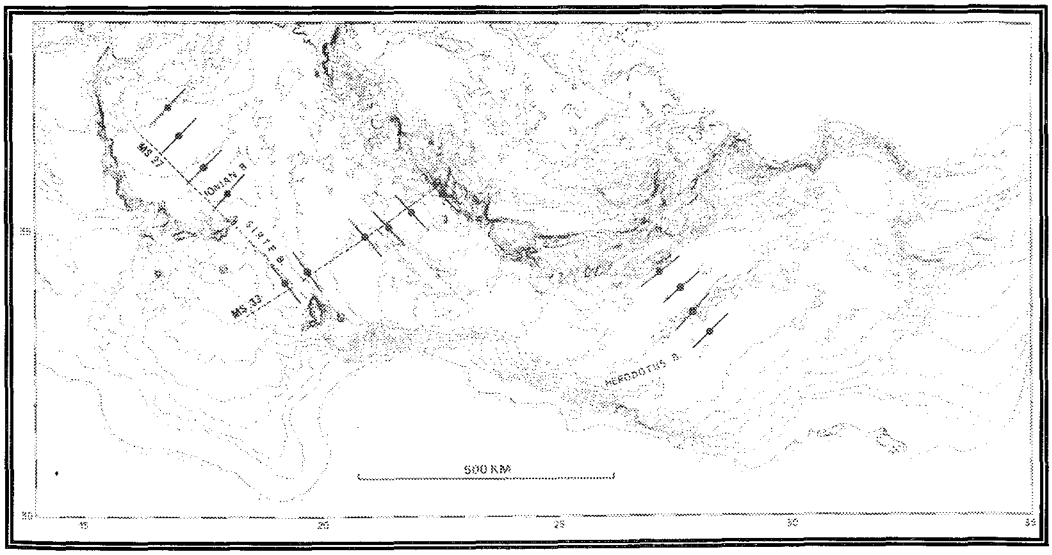

Figure. 2.62. Location map of the expanding profiles (ESP recorded during the Pasiphae cruise; black dots ESP central points, i.e., the places where a velocity model is obtained (after *Voogd et al., 1992*). Dashed lines indicate the seismic profiles (*Finetti, 1976, 1982*).

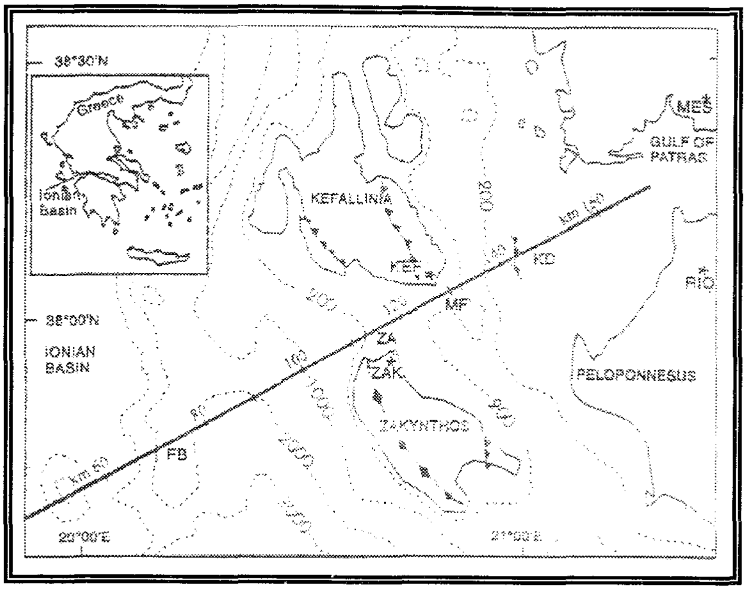

Figure. 2.63. The STREAMERS line (ION-07) shown by a solid line (after *Hirn et al., 1996*).

Aegean

The earliest seismic refraction studies in the South Aegean were made by *Wong et al. (1971)* and *Finetti & Morelli (1973)*. German researchers in the Aegean region took important refraction data. After this came on extensive survey to accurately determine the crust thickness and crust type. As a first stage, deep seismic profiles were taken between 1971 and 1974 by the German research ship Meteor (*Hinz, 1974; Makris, 1973*) (Fig. 2.64, 2.65). In 1976, studies were continued with additional shootings in North Aegean Trough (NAT) (Fig. 2.64). *Makris (1978a)* correlates seismic data with earlier data obtained by other geophysical methods to find out the velocity structure of the upper mantle and crustal thickness. There is good correlation between the intersecting refraction cross-sections (*Makris, 1978b*). A combined near-normal incidence and wide-angle seismic recording program (STREAMERS) was conducted in the western Aegean Sea in 1993 with the principal objective of testing the popular hypothesis of lower crustal deformation (particularity extension) (*Sachpazi et al., 1997*) (Fig. 2.66). The processing and interpretation of the STREAMERS lines in the Aegean is continuing (*Sachpazi et al., 1997*).

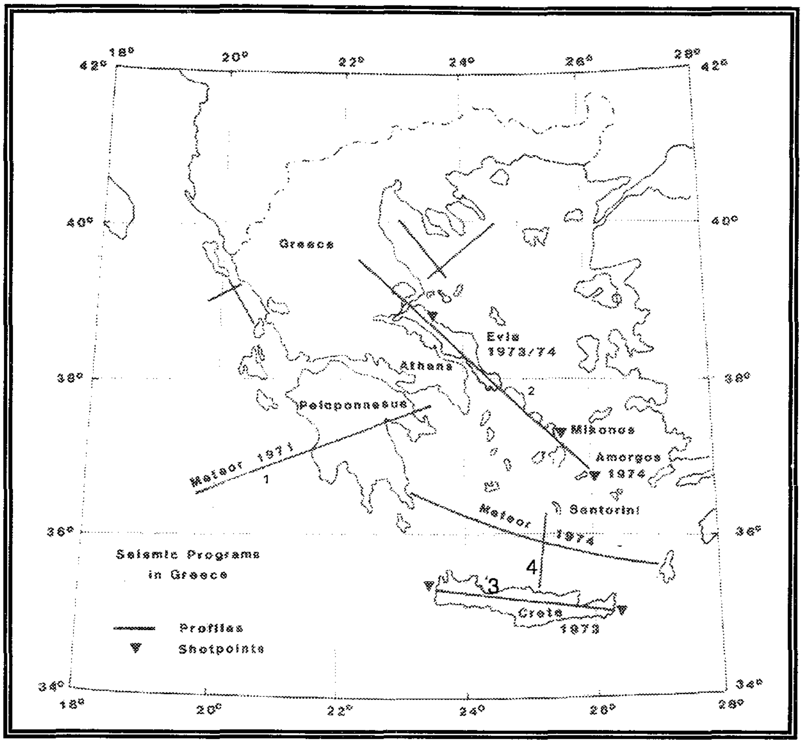

Figure. 2.64. Location of Deep Seismic Soundings (DSS) in Eastern Aegean and surrounding area (after *Makris, 1985*).

Western Turkey

Explosion data have been obtained from quarry blast and conventional explosive sources in the Marmara area. In the last 18 years, four quarry blasts and one controlled source explosion experiment were conducted in northwestern Turkey (mostly in the Marmara region) BÜ Kandilli Observatory and Earthquake Research Institute (*see Küleli et al., 1995*) (Fig. 2.67). Explosions made at Anadolu Kavağı in 1980, were recorded by the seismic stations installed around the Marmara region (*Gürbüz & Üçer, 1985*). First arrival times of P waves were used in the interpretation of the seismic records (*Gürbüz & Üçer, 1985*). A quarry-blast explosion at Adapazarı was recorded along N-S elongated profiles to determine the crustal thickness beneath Adapazarı (*Gürbüz et al., 1980*). In 1997 profiles were constructed with the data from temporary and permanent (portable) seismic stations located along the chosen directions. For a detailed investigation of the velocity distribution in this region, data from four quarry blasts and one controlled source explosion were recently analysed (*Küleli et al., 1995*). Controlled source explosion data were collected with reverse shootings. Associated with the controlled source seismic study, a joint Turkish-German seismic refraction project was organised in

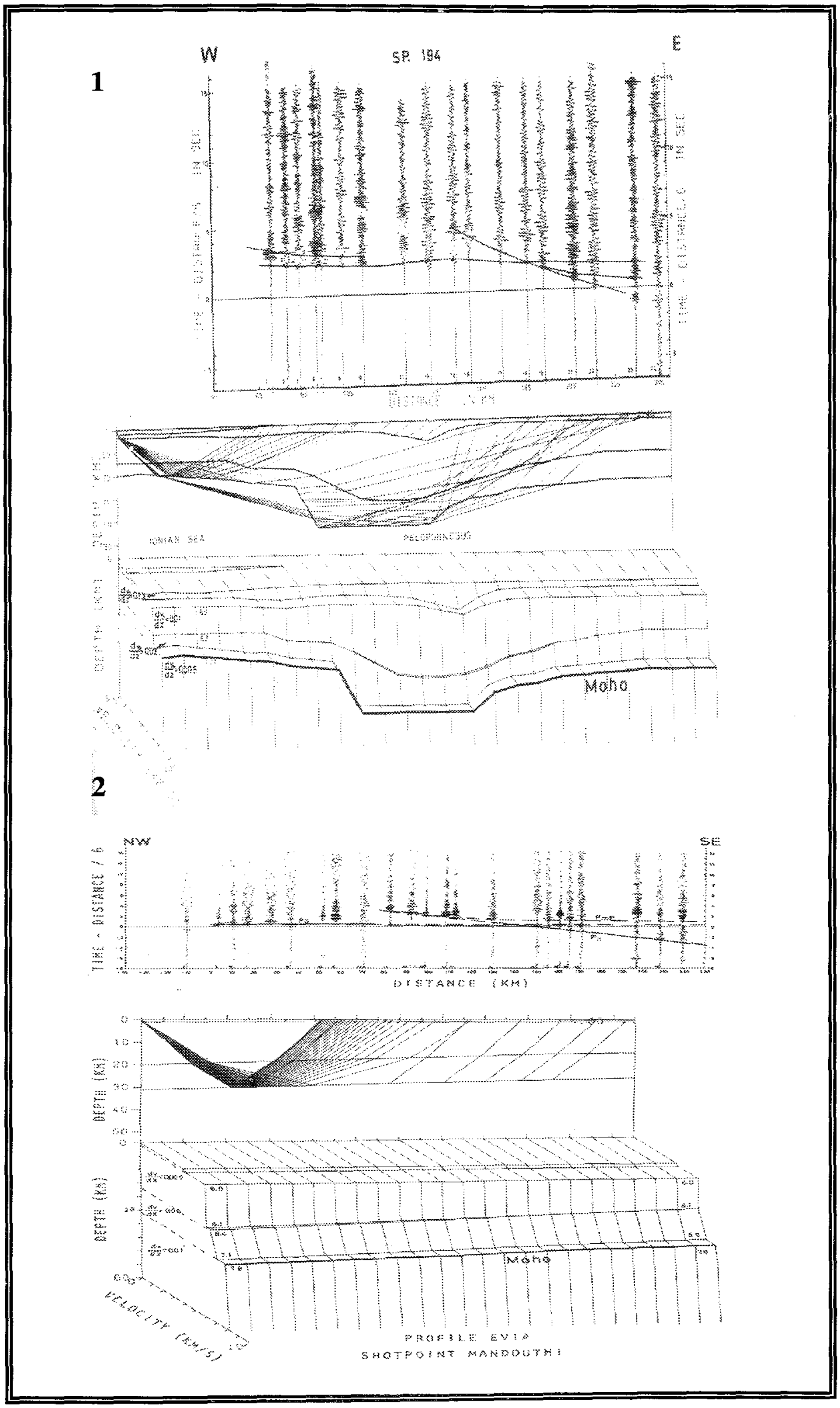

Figure. 2.65. Cross sections shown in Fig 2.64 (after *Makris, 1985*). Solid lines give the computed travel-times as indicated at the middle part of the drawing. The seismic phases to be considered are at Pg, Pn, and PmP.

1991 to study crustal structure in the east of the Marmara sea. Two refraction profiles were chosen perpendicular to each other (N-S and E-W) (Fig. 2.67). There are no controlled source profiles in western and southwestern Turkey reflecting the deep structures.

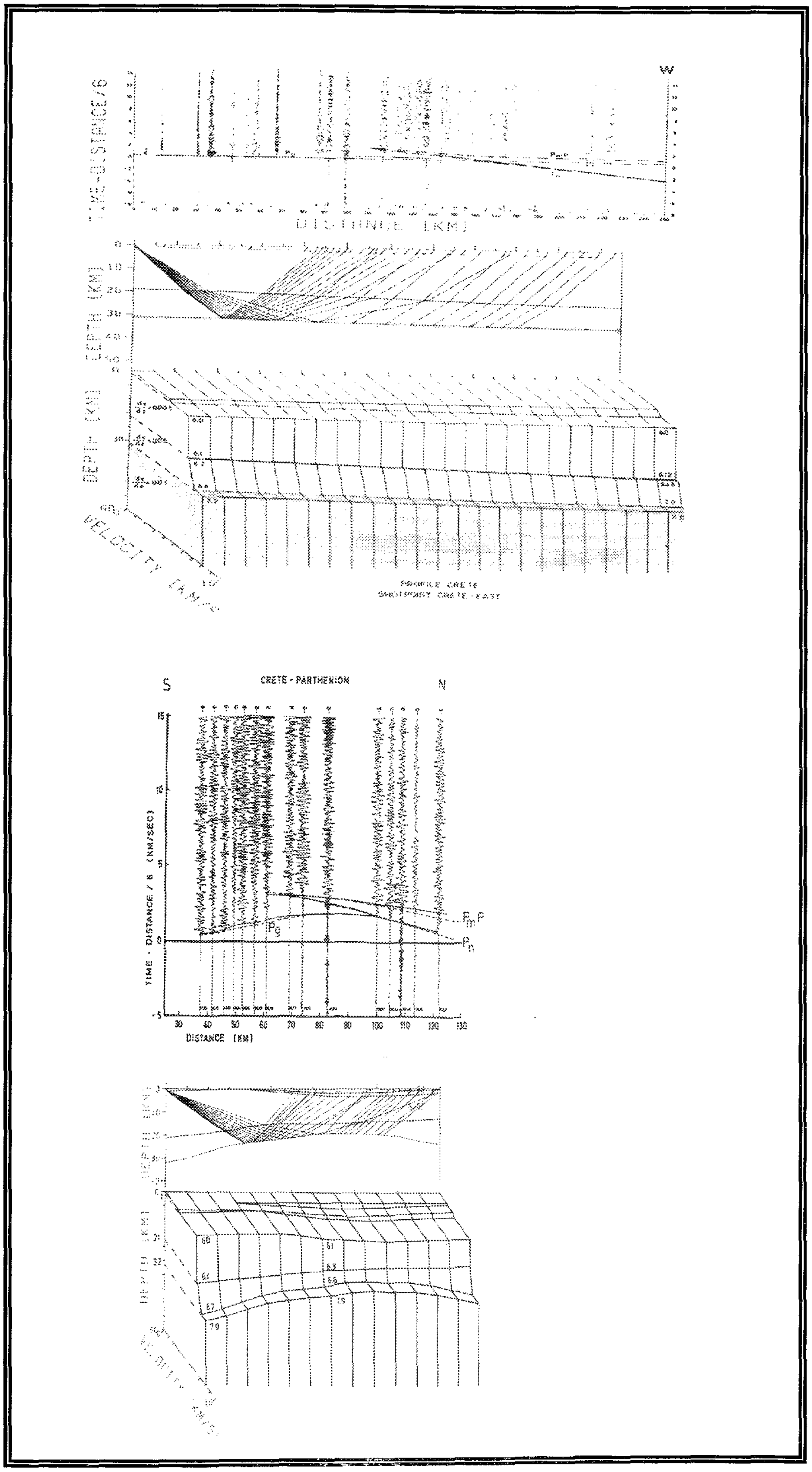

Figure. 2.65. (continued)

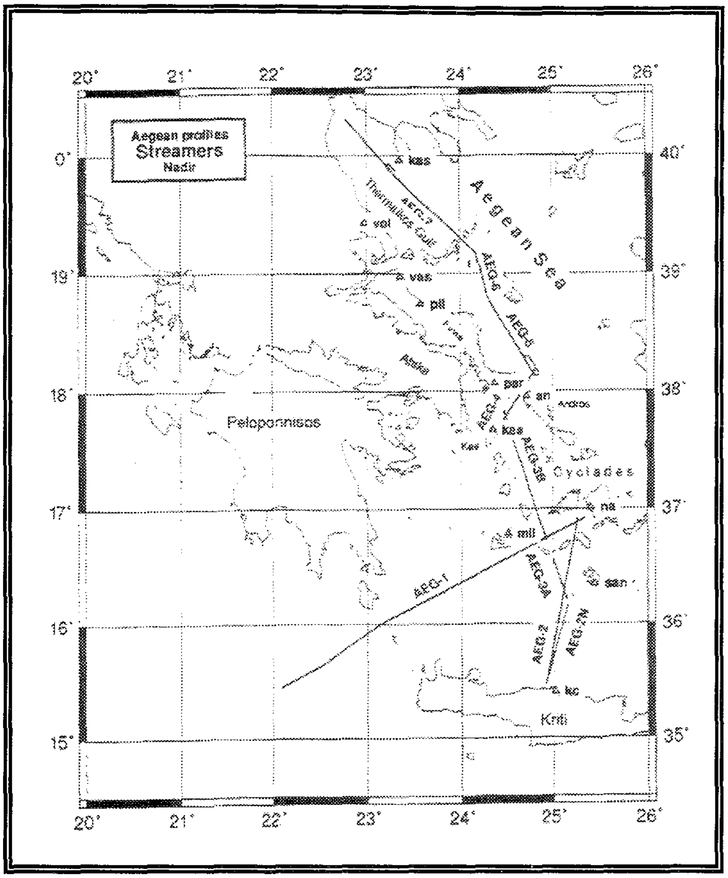

Figure. 2.66. Position map of normal-incidence multichannel lines and land recording stations of the STREAMERS survey in the Aegean (after *Sachpazi et al., 1997*).

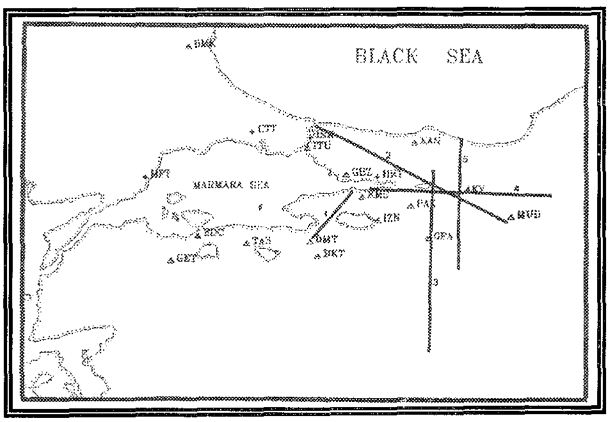

Figure. 2.67. Solid lines show the seismic refraction profiles; 1-Kestel, 2-Anadolu Kavağı, 3-Adapazarı, 4,5-Turkish-German refraction project (after *Küleli et al., 1995*).

2.16.2. General Interpretation

Because of the source is known, deep seismic sounding studies are the most reliable method to determine physical properties of the medium. The main advantage in using explosion data is that the origin times and sites of source waves are accurately known. The main disadvantage of them is their high cost. In the Aegean area, limited DSS profiles were taken. Scientists suffer from lack of DSS data especially in the western Turkish mainland.

Clear information about the crust and upper mantle structure in the eastern and central Aegean region comes mainly from deep seismic sounding data (*Makris, 1973, 1978a,b*). *Makris (1973, 1977)* published a Moho-depth map of the Greek mainland and Aegean areas derived from seismic data and gravity computations.

Below the Aegean Sea, a low velocity zone exists in the uppermost part of the mantle (*Makris, 1978a,b*). The crust in the Aegean Region is purely continental, including the Cretan basin (*Makris, 1978a*). But the Cretan basin shows a minimum crust thickness (20 km) in the central trough zone (*Makris, 1978a*). The crustal thickness increases under the mainlands of Turkey and Greece, the Cretan island arc and the Cyclades. The thickness shows a generally increasing trend from the South Aegean to the North Aegean. The maximum thickness of the sedimentary cover coincides with the thinnest part of the crustal section (*Makris, 1977*). But detailed sediment thicknesses are not known in the Aegean basin. The sedimentary cover above the crystalline basement, which outcrops at South Evvoia and the Cycladic islands, cannot be defined with deep seismic soundings because it is too thin (perhaps 1.0-1.5 km) (*Makris & Vees, 1977*) (see *Morelli, 1985*).

The relatively high velocity at the northern part of Marmara and lower velocity in the central and south of the working area indicate thinning of the continental lithosphere from north to south, or at least presence of a stable mantle lid at NW Turkey (*Küleli et al., 1995*). Pn velocities are lower than the normal mantle velocities here. This suggests the existence of thin (i.e. hot) continental crustal material in this region (*Küleli et al., 1993*)

Modelling of the seismic refraction data of two profiles crossing each other on the NAF zone in the Intra-Pontide suture in the Marmara region indicates rugged shape interfaces. The horizontal velocity variation is identified probably as due to continental depression between the wedges of the southern İstanbul and northern Sakarya zones (*Küleli et al., 1995*).

Seismic refraction surveys *Ferrucci et al. (1991)* by indicate that the Ionian basin is floored by oceanic crust, but the nature of the crust, continental versus oceanic, in the Herodotus basin is still debated (*Lort et al., 1974; Makris & Stobbe, 1984*). The crust is possibly continental in the Herodotus basin (*Lort et al., 1974; Hinz, 1974; Agarwal et al., 1976; Cloetingh et al., 1980; Makris & Stobbe, 1984; Makris et al., 1986*). *De Voogd et al. (1992)* obtained excellent velocity control down to Moho depth in the East Mediterranean. According to the results in the Ionian basin, the crust there is relatively thin and oceanic. In the Herodotus basin, the crust is similarly thin and sediments are thicker (*De Voogd et al., 1992*). The thick sedimentary cover in the Herodotus plain known from both reflection and refraction data (*Finetti, 1976; Ginzburg & Ben-Avraham, 1987*) suggests an old age for the floor (*De Voogd et al., 1992*). The thin crust of the Herodotus basin may be interpreted either as oceanic or thinned continental crust (about 10 km thick). The top of the crust of the Herodotus basin is much deeper than the Ionian crust suggesting that the Herodotus basin is significantly older than the Ionian basin (*De Voogd et al., 1992*).

In December 1988 *Truffert et al. (1993)* performed a two-ship refraction and oblique reflection seismic survey in the eastern Mediterranean Sea (Pasiphae cruise; *De Voogd et al., 1992*). 14 Expanding Spread Profiles (ESP) were shot along three sections: the Calabrian prism, the western Mediterranean Ridge and the eastern Mediterranean Ridge (*De Voogd et al., 1992*).

3. GEOLOGICAL EVOLUTION

It is well known that, since about the beginning of N-S convergence along the Alpine Himalayan system, which covers also Turkey, the Aegean Sea and Greece, many oceanic floors have been consumed between various continental lithospheres (e.g., *Brunn, 1960; Aubouin et al., 1963; Brinkman, 1966; Ketin, 1966; Auboin & Dercourt, 1970*). However, the paleogeography of the oceans called the Tethys and Neo-Tethys and branches of Neo-Tethys (*Şengör & Yılmaz, 1981; Ricou et al., 1984*) are still controversial (*Dilek & Moores*). To day the remnants of the oceanic materials are visible as continuous zones between continental fragments approximately EW in Turkey mainland and NW-SE on the Greek mainland (e.g., *Aubouin, 1973; Bernoulli et al., 1974; Aubouin et al., 1976; Brunn, 1976; Şengör & Yılmaz, 1981; Horvath & Berckhemer, 1982; Jackson McKenzie, 1988; Görür et al., 1992*). These ophiolitic stures and the continental fragments construct an imbricated structure in the area (Fig. 3.1). The zones are parallel each other (Fig. 3.1). These are also parallel to young tectonic features of the Aegean region (e.g., the South Aegean Arc, normal fault system and the South Aegean Volcanic Arc).

Crystalline massifs; the Serbo-Macedonian and Rhodop massifs in Greece and Bulgaria; the Istranca, Uludağ and Kazdağ massifs in western Turkey, form the northern part of the imbricated mass. In northwest Turkey two tectonic zones are located, mainly the southern İstanbul zone and a small part of the northern Sakarya zone, which are separated by a complex suture zone, which is called the Intra-Pontide ophiolitic suture in Turkey. It continues through Greece and it is connected to the Vardar zone (e.g., *Şengör et al., 1985; Okay, 1989*).

Just below these crystalline massifs that represent the Eaurasian continental margin, the second ophiolitic suture zone defining the northern boundary of the Apulian plate follows the Vardar zone in Greece (*Smith, 1971; Dercourt, 1970, 1972; Bernoulli & Laubscher, 1972; Hynes et al., 1972; Boccaletti et al., 1974*) to the İzmir-Ankara-Erzincan (IAE) zone in Turkey which is known for its oceanic material (ophiolite and chert sequences of Jurrasic age) as a clear sture zone (*Brunn, 1960; Mercier, 1966; Brinkman, 1966*). This zone is the remnant of a northern branch of the NeoTethys that was consumed between previously explained crystalline massifs to the north and Median Aegean crystalline belt to the south which consists of metamorphic

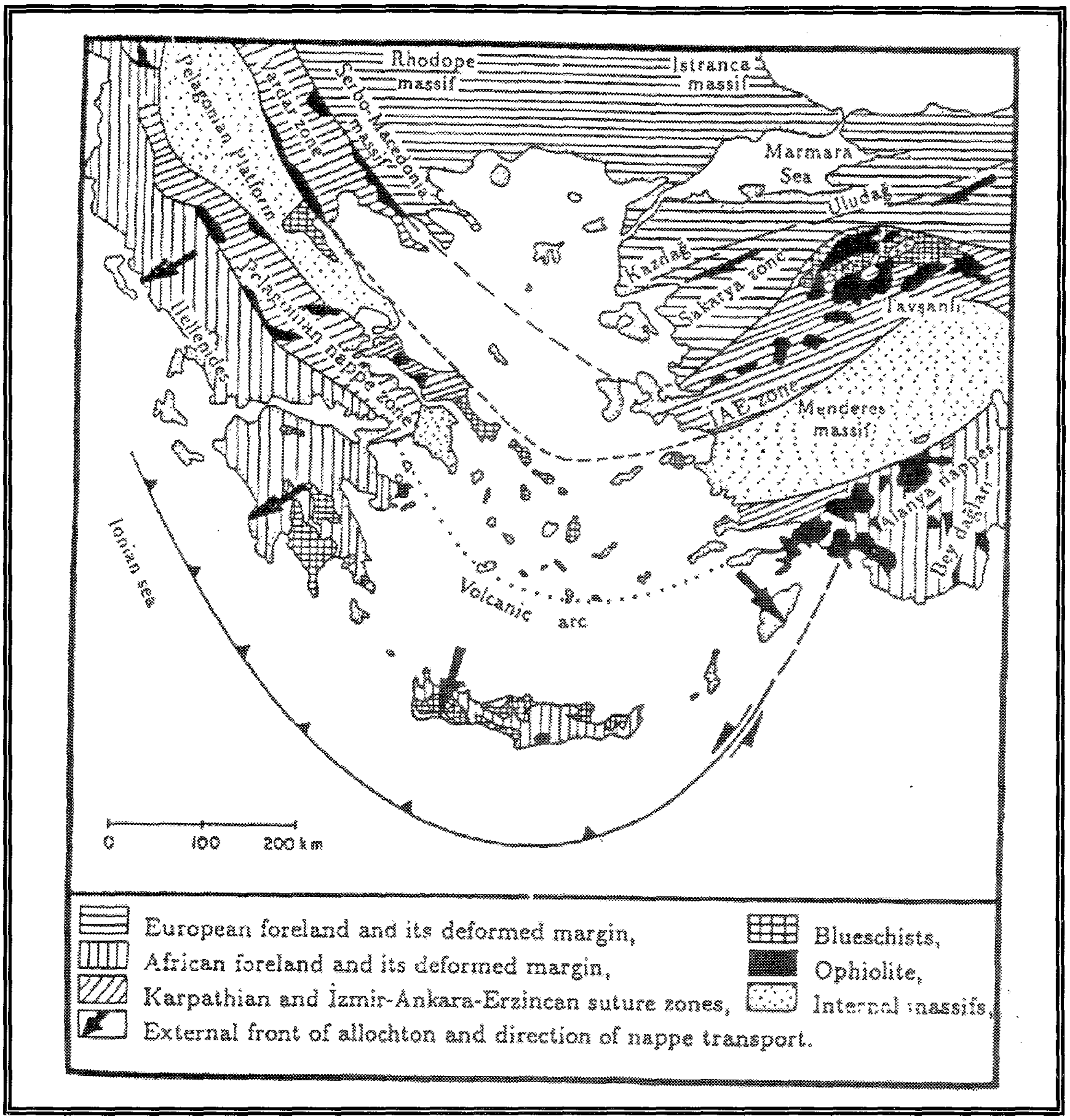

Figure. 3.1. Major Geological units and old oceanic remnants (ophiolites and blue shist) (after *Küleli et al., 1995*). The active volcanic arc is shown by a dotted curve.

and nonmetamorphic Palaeozoic rocks at the base and Mesozoic and Cenozoic rocks (*Şengör & Yılmaz, 1981*). This sture displays HP/LT metamorphism, which is characteristic of suture zones which represent a collisional regime. From the behaviour of the high gravity field one can separate the Pindos Mountains, Othrys and North Evvoia, the Vourinos Complex and the ophiolites of the Vardar Zone along the Chalkidiki Peninsula (*Makris, 1977*). The same behavior of gravity anomaly is observed on the 30-100 km wide İzmir-Ankara-Erzincan suture zone (e.g., *Oral, 1987*).

The Taurid system borders the Vardar and İzmir-Ankara suture zones towards the south, situated. The Pelagonian platform on the Greek mainland, the Attic-Cyladic platform (Naxos, Paros, Mikonos, Ikaria, Tinos) and the Menderes massif in Turkey form the ACMM which represents the northern margin of Gondwanaland as well as the Taurides (*Şengör & Yılmaz,* *1981*). In general this zone contains Paleozoic and Precambrian metamorphic rocks as crystalline schists and marbles. The massif has a complex internal structure and lithological distribution. Age, generation and development of the massif are still widely debated. Continental collision between the Menderes-Taurus block and the Sakarya continent resulted in crustal thickening and shortening in the area beneath the massif, which was structurally overridden by the Lycian ophiolitic napple pile from the south (see for detail *Bozkurt et al., 1995*). The late Cretaceus marked the beginning of a convergent regime that resulted in destruction of the Neo-Tethyan ocean at all fronts in Turkey and was particularly characterised by the emplacement of ophiolite nappes (*Yılmaz, 1989, 1997*). The regional metamorphism (main; HP/LT, Menderes metamorphism), caused by the latest collision across the Neo-Tethys which was taken up by lithospheric shortening and thickening (*Dürr, 1975; Şengör & Yılmaz, 1981; Okay, 1989; Yılmaz, 1989; Yılmaz, 1997*), was followed by thermal doming. The over-thickened dome collapsed and leads to emplacement of augen gneisses and the creation of a wide extensional system all around the massif (*Şengör & Yılmaz, 1981; Şengör, 1982; Bozkurt et al., 1995*). This is thought to have triggered development of the first fault-bounded basins, prior to the E-W oriented Aegean grabens (see Section 2.3).

The third zone, in the Pelagonian nappe zone in Greece and which also consists of ophiolitic material the Alanya nappes (*Horvath & Berckhemer, 1982*) in western Turkey are called as the Lycian zone.

In the southern part of the region, an orogenic belt of Alpidic origin can be traced from the Apulian, the Dinarides and Hellenides in the west, through to the South Aegean Island Arc and the Taurides (Bey Dağları) and Anatorides in the east (*Biju-Duval et al., 1974; Şengör, 1979; Şengör & Yılmaz, 1981; Horvath & Berckhemer, 1982; Mueller & Kahle, 1993*). The present Taurus mountains, which were below sea level until the middle Miocene, underwent a strong uplift in the successive phases, up to 1000 m relative to the Central Anatolian Plate (*Barka & Reilinger, 1997*).

In the southermost (outermost) position, beneath the South Aegean Arc (SAA), an ongoing subduction process has been occurring parallel to the all imbricated units to the north. The subduction zone is composed of sequences of Mesozoic platform carbonates, great allochthonous masses of deep-sea Mesozoic sediments and Cretaceus to Tertiary flysch. To the east of the Aegean region the Bitlis suture zone in eastern Turkey is the manifestation of a convergence process due to continental collision, which is still ongoing (e.g., *Mueller & Kahle, 1993*). The southernmost ophiolite belt (e.g., Troodos in Cyprus, Kızıldağ in southern Turkey) includes relatively complete and undeformed series (*Dilek & Moores*) with respect to the northern ophiolitic series beneath the Aegean region.

The volcanic rocks are also parallel to the geotectonic units and located within the three zones of the region (*Gülen, 1990*). The southernmost volcanic arc is active and linked with the South Aegean subduction processes (Fig. 2.1d). The second oldest volcanic zone is situated along the southern periphery of the Attic-Cyladic-Menderes metamorphic belt, which conforms to the arcuate trend of the subduction zone, extending between Greece and Turkey in the back-arc region (*Küleli, 1992*). This zone is referred as the "inner arc" (*Innocenti et al., 1979*) and consists of Atalanta-Volos volcanics and Antiparos rhyolite (*Küleli, 1992*). The third oldest volcanic zone covers an extensive area in the northern Aegean region where the Attic-Cycladic-Menderes metamorphic belt forms its southern boundary (*Küleli, 1992*).

HP/LT metamorphism is an indicator of a convergent plate boundary. Blueschists are the main product of HP/LT metamorphism represents subduction zones. In the Aegean region, blueschist facieces and ophiolites have been located in two zones. The first zone lies on the south of the IA suture zone in the region located near Tavşanlı in western Turkey, Mount Olympus in mainland Greece and the Cyclades islands in the Aegean sea (*Okay, 1989*). A second 'external blueschist belt' is located along the South Aegean Island Arc and is related to an on-going subduction process (*Okay, 1989*).

The Latest Tectonic Phase

Before the latest extensional regime between the southern margin of the European plate and the northern margin of the African plate, various strands of the Tethys opened and closed. The latest subduction in the Aegean area has been dominated by subduction of the African Plate beneath the SAA (*Le Pichon & Angelier, 1979*). Because of extension on the overriding plate, the crust was considerably thinned. Stretching motion has taken up normal faults that characterise the Aegean region (see for detail Section, 5). This sequence is expected by most of the scientists. But geological arguments for the duration of stretching and subduction are controversial in the Aegean region. To the east, E-W and ENE-WSW trending grabens and to the west, NW-SE trending grabens are the main structures that developed under the latest extensional tectonics in The Aegean region. The most important problem that must be solved is that some grabens to the east, are about normal to the E-W basins. For these grabens Early Miocene age is widely accepted (*Seyitoğlu & Scott, 1991, 1994; Seyitoğlu et al., 1992b, 1994*). However, while some authors suggest that the cross grabens come from an early compressional phase, the others suggest sintectonic evolution for the cross grabens and E-W trending grabens. The following paragraphs lists suggested extension initiation times by various authors who consider various methods and models. These are critically reviewed below.

Based on field data the normal faulting in Greece previously associated with the present movements began in the Miocene (*Aubouin et al., 1963*). This data is consistent with the observations on land in both Greece and Turkey (*Aubouin, 1973; Brunn, 1976*). By studying faults on islands (horsts) in Aegean *Angelier (1978)* dated the extension to the Late Miocene. One group of scientists suggests that older (Early Miocene age) grabens (cross grabens) that are normal to the E-W extensional grabens in western Turkey were created under N-S compressional regime (*Şengör & Yılmaz, 1981; Savaşçın, 1982; Şengör, 1982; Şengör et al., 1985; Yılmaz, 1989; Yılmaz, 1990*). According to them, a compressional orogenic regime continued until the to Late Miocene. They argue that Early Miocene was the time of final emplacement of the Lycian nappes. The average crustal thickness of the overthickened crust resulting from the post-Eocene N-S intracontinental shortening, is estimated to have been over 60 km (*Şengör, 1982*). Another group (e.g., *Savaşçın & Güleç, 1990; Savaşçın, 1990; Yılmaz, 1990*) consider that the compressional and extensional regimes are represented, respectively, by calc alkaline and alkaline volcanism. They dated the extensional deformation initiation time as Early Miocene, by using radiometric methods. Furthermore *Yılmaz (1997)* suggested that between the two phases there is a time gap during which no volcanism was observed.

Dewey & Şengör (1979) find a similarity between the starting time of the Aegean extension and North Anatolian Fault (NAF) and suggest a relationship between them. They suggest that the collision of Arabian and Asian plates in the Middle Miocene, 10 My ago according to their theory, started the neotectonic regime in Turkey (extensional regime in western Turkey). By considering geophysical and geological data *Le Pichon et al. (1984)* studied the subsidence history of the North Aegean Trough (NAT). The drilling results interpreted by *Le Pichon et al. (1984)* in NAT show that in the Early Miocene a compressional regime dominated the deformation. According to the model, the extensional regime started in this basin during about the Tortonian (10 My ago). Tectonic analysis of *Lyberis (1984)* also shows this age for the initiation of the extensional regime.

By considering seismic profiles in the South Aegean region *Jongsma (1975)* determines Middle Miocene extensional features. However, from a study of the reflection profiles it appears that, after the main Alpine movement until somewhere near the end of the Tortonian, the southern Aegean Sea was largely subjected to erosion or non deposition (*Jongsma, 1975*). Based on field data *Mercier et al. (1976)* suggest that extension in the Aegean was widespread by the lower Pliocene (5 My) and started in places during the Middle Miocene (see also *Şengör et al., 1985*). By considering relations between sedimentation and faulting back-arc extension was dated Serrevallian (Middle Miocene) by some authors (*Meulenkamp et al., 1988; Mercier et al., 1979; Le Pichon & Angelier, 1979, 1981; Angelier, 1979; Angelier et al., 1982; Lyberis et al., 1982; Kissel & Laj, 1988; Mercier et al., 1989*). Based on field data *Sözbilir & Emre (1996)* suggest the Middle Miocene age for the creation of the E-W grabens in western Turkey.

A number of different arguments suggest that the stretching started during the early Miocene. Some scientists, by using radiometric dating of volcanic rocks and the modification of the time span of pollen assemblages suggest Late Oligocene-Early Miocene age for the initiation of extensional regime in The Aegean region (*Benda, 1971; Benda et al., 1974; Benda & Mulenkamp, 1979; Seyitoğlu, 1992; Seyitoğlu et al., 1992; Seyitoğlu & Scott, 1991, 1992a,b; Seyitogtu et al., 1994; Jolivet et al., 1994; Dora et al., 1995; Seyidoğlu & Scott, 1996a,b*). According to the some of these scientists (*Seyitoğlu & Scott, 1996a*), E-W grabens and the older grabens (Early Miocene) trending normal to them, developed together in western Turkey. This mechanism can clearly be explained by the cross-graben model of *Şengör (1987)*. *Jolivet et al. (1994)* also suggest the Early Miocene age, based on mineral fabric that is related to extensional motions on metamorphic rocks. They recognised that slip vectors (i.e., present-day displacements) and Miocene ductile lineations (i.e., bulk finite strain) follow the curvature on the scale of the Aegean. This suggests has the kinematic pattern have not been strongly modified since the early Miocene (*Jolivet et al., 1994*). Based on evidence from metamorphic rocks *Bozkurt et al. (1995)* suggest that the over-thickened Menderes Metamorphic dome collapsed in the Late Eocene-Oligocene time, which is older than that previously suggested (Late Miocene) by *Şengör & Yılmaz (1981)*. According to this result, a compressional regime could not have generated the older (Early Miocene) cross grabens.

Seyitoğlu & Scott (1992) argue that, due to the continuation of extension after the Late Miocene, calk alkaline volcanism is observed and the thickness of lithosphere was reduced and consequently alkaline volcanism became dominant in this advanced stage of the extension (Late Miocene and younger). However, an other group suggests a younger age for the initiation of extension (*Savaşçın & Güleç, 1990; Savaşçın, 1990; Yılmaz, 1990*), considering this calk alkaline volcanism as a compressional tectonic indicator. *Zanchi et al. (1990)* suggested a multiphase extension associated with a cross cutting high-angle normal fault and strike slip fault model in western Anatolia.

Variations in the Extensional Regime

There is no doubt that the strain has varied spatially over extensional time (e.g., *McKenzie, 1978; Mercier et al., 1989; Le Pichon et al., 1995*). In the Thermaikos basin two stages of extensional regime were identified by *Brooks & Williams (1982)*. These are NNE-SSW in the Late Miocene-Early Pleistocene and N-S after the Late Pleistocene. According to *Lyberis (1984)*, there are three stages of extensional regime in the Aegean region. These are NW-SE in Late Miocene, NE-SW between Early Pliocene-Early Pleistocene and N-S since the Quaternary extension. *Mascle & Martin (1990)* identified four stages by using seismic profiling data. These are: a compressional regime in the Early Pliocene (see also *Dumont et al., 1979a,b; Angelier et al., 1981; Mercier et al., 1976*); NNE-SSW in central and N-S in southern portions in the Pliocene; after the Pliocene another short extensional regime NNE-SSW in the eastern part of the Aegean region and NNW-SSE in the western part of the extensional regime. They pointed that the tectonic evolution in the region took place in a short time interval of compressional and generally dominant extension. According to the seismic reflection studies of *Jongsma (1975)*, in the Middle Miocene a differential block movement started along E-W and N-S fracture systems, which resulted in some sinking in the Serravallian. From the Late Serravallian to the Late Tortonian the sediments show an increasingly stronger influence of marine conditions (*Jongsma, 1975*). This general submergence resulted in the separation of Crete from the European landmass in the middle Tortonian (*Jongsma, 1975*). Until this time there seems to have been a subaerial connection with the mainland as evidenced by mammalian faunas (*Jongsma, 1975*). Towards the end of the Tortonian and closure to the beginning of the Messinian, an abrupt change to open marine conditions took place, which gave rise to highly calcareous sediments (*Jongsma, 1975*). On the old E-W and N-S faults large displacements took place but younger NW-SE and SW-NE faults also played a pronounced role (*Jongsma,* *1975*). In the western part near Crete it is likely that it corresponds to the period of erosion from somewhere in the Oligocene to the end of Tortonian (*Jongsma, 1975*). According to him Crete was connected to the main European landmass during this period (*Drooger & Meulenkamp, 1973; Jongsma, 1975*). On Crete, *Drooger & Meulenkamp (1973)* recognise two phases, the first (Serravallian to Recent) produced E-W and N-S trending faults. The second phase started in the Pliocene and gave rise to movement along NW-SE and SW-NE faults. On Greece in the Peloponnesus, E-W and N-S high angle normal faults of Pliocene to Holocene age cut the earlier thrust structures (*Smith & Moores III, 1974*). In the eastern part of the southern Aegean similar normal faulting and gentle folding has occurred during the Pliocene (*Mutti et al., 1970, Bernoulli et al., 1974*). *Mercier (1981)* recognises two compressional tectonic phases in the Miocene, Early Pliocene and Pleistocene age. *Angelier (1979)* proposed a series of alternating compressional and extensional tectonic episodes in the area since the Miocene from a study of the fault geometry. *Zanchi et al. (1990)* suggested a multiphase extension associated with a cross cutting high-angle normal fault and strike slip fault model in western Turkey. Based on field data, *Mercier et al. (1976)* suggest that extension in the Aegean was widespread by the lower Pliocene (5 My) and started in some places during the Middle Miocene (see also *Şengör et al., 1985*). Extensional structures are seen in Greece in Miocene time, though *Mercier et al. (1989)* argue that most of the extensional strain occurred in the last 5 My. *Le Pichon et al. (1995)* suggested that the African promontory began to collide with central part of the arc in the Pliocene.

Subduction Initiation Time

Papazachos (1973) calculated a mean consumption rate of 28 mm/yr by considering a 280 km slab length and 10 My subduction initiation time. *McKenzie (1972)* found a 35 mm/yr consumption rate by considering a 350 km slab length and 10 My subduction initiation time.

McKenzie (1978) considered a 200 km length of subducting plate based on the Benioff zone beneath the Aegean region and a 35 mm/yr consumption velocity rate, with the initiation of the subduction at 3 My (see also *Mercier, 1981*). *McKenzie (1978)* and *Kissel & Laj (1988)* have suggested that the onset of subduction is likely to have accompanied the onset of rapid extension in the Aegean and concluded that the subduction began 5 My ago (see also *Taymaz et al., 1991*). *Jackson & McKenzie (1988)* take into account the consumption rate (100 mm/yr) which was estimated from seismicity of overriding lithosphere (60 mm/yr) and 10 mm/yr northward motion of the African plate and deepest seismicity of the slab (300 km) estimated 5 My (see also *Jackson, 1993*). *Le Pichon & Angelier (1979)*, assuming that the Wadati-Beniof zone delineates the geometry of the lithosphere subducted beneath the Aegean region and considering the age of the oldest volcanic rocks on the active volcanic arc, concluded that subduction has been taking place during the last 13 My, with an uncertainty as much as 3 to 5 My. *Mercier et al. (1989)* date the onset of the shortening in the Ionian islands at 16 My, accompanied by subduction of the seafloor. *Spakman (1986), Spakman et al. (1988), Meulenkamp et al. (1988)* and *De Jonge et al. (1994)* have argued that the subduction is much older (26-40 My), based on a tomographically defined subducted slab length much longer than the seismically defined portion. They argue that the maximum depth of earthquakes represents not the deepest extent of the slab, but the deepest extent of material cold enough to generate earthquakes. With considering this length the authors found older ages (26-40 My) than those found previously.

Northeastward subduction of the Ionian and Adriatic Sea floor beneath Greece and Anatolia has accommodated convergence between Europe and Africa throughout much of late Cenozoic time (e.g., *Le Pichon & Angelier, 1979*). After consumption of the all suductible material, in the Late Miocene, a shortening processes that still continues occurred along the eastern Adriatic border (see *Channell & Horvath, 1976; Burchfield, 1980; Horvath, 1984; Babbucci et al., 1997*).

4. CRUSTAL AND UPPER MANTLE STRUCTURE

Considering the complexity of the surface tectonics it is important to investigate the structure of the underlying upper mantle in order to better understand the tectonic evolution of the region (e.g., *McKenzie, 1978; England & Houseman, 1989; Vilotte et al., 1982; England & Jackson, 1989; Jackson et al., 1992*). The global scale tomographic image of the Earth's interior indicates that the structure in the top 100-200 km is closely related to the tectonic regime at the surface. To determine the physical properties of the deep structure, potential (gravity, magnetic, magnetoteulliric, heat flow) or kinetic (seismic waves that generated from earthquakes) energy of the Earth itself and kinetic energy that is created artificially (seismic sounding) have been used in the Aegean region since the 1960' s. To understand the structure of the crust and upper mantle, one should consider all the available data that represent the geophysical properties of the region (see section 2.). Tectonic processes such as subduction change the equilibrium conditions in the Earth's layered medium. As a result of these changes in the mantle, the surface geology is greatly affected or directly formed by material carried from the mantle or the subducting oceanic lithosphere. For example, the anomalous thermal and pressure conditions associated with subduction zones are the most important factors controlling petrogenesis (see *Mueller & Kahle, 1993*). Consequently, the surface features of the region such as magmatism and metamorphism, blueschist emplacement, ophiolites, volcanic island systems, suture zones and arcs, represent evidence about the mantle structure (rheology) of the Earth. But knowledge about the variations in the physical state and chemical composition of the upper mantle is still rather limited (*Mueller & Kahle., 1993*). In addition, source functions of crust and upper mantle structure must be known accurately if more accurate foci of earthquakes are to be determined.

Due to the geodynamic evolution of the Aegean area, the crustal thickness undergoes strong variations over a range of 15-25 km beneath the back-arc basin to 40-60 km beneath the orogenic belts (e.g., *Alessandrini et al., 1997*). In Table 4.1. brief information is given about crustal thickness in various parts and methods are shown for each result. In general the continental lithosphere thins from the north to south (*Makris, 1978a,b; Gürbüz et al., 1980;*

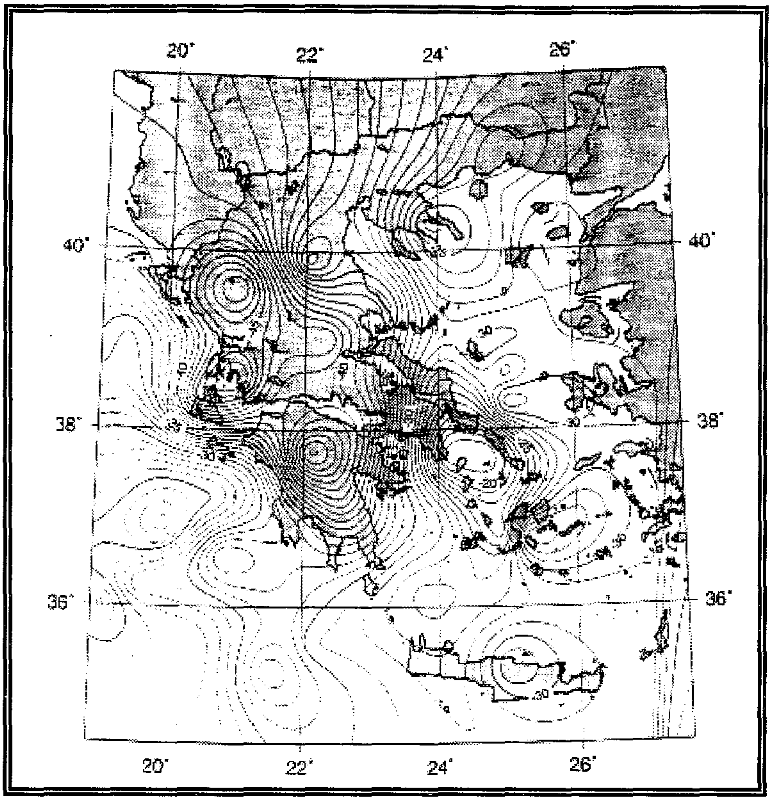

Figure. 4.1. Moho depth map of Greece and surrounding area obtained by a Gravity computation (after *Tsokas and Hansen, 1997*). Contour interval is 1 km.

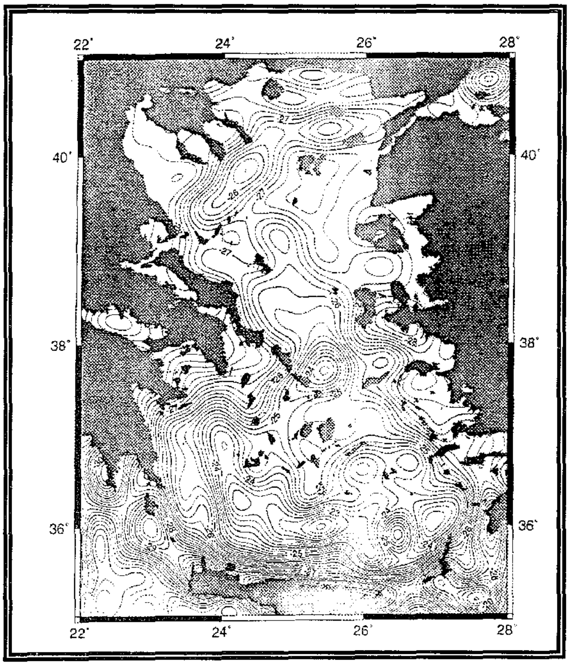

Figure. 4.2. Moho depth map of Aegean obtained by gravity computation (after *Genç et al., 1996a*).

Genç et al., 1996a). After a deep seismic sounding study of *Makris (1973)* it is clearly understood that the crust in Aegean region is continental but thin. In addition, *Rodgers et al. (1997)* found that Lg propagates efficiently for events from the western coast of Turkey. This

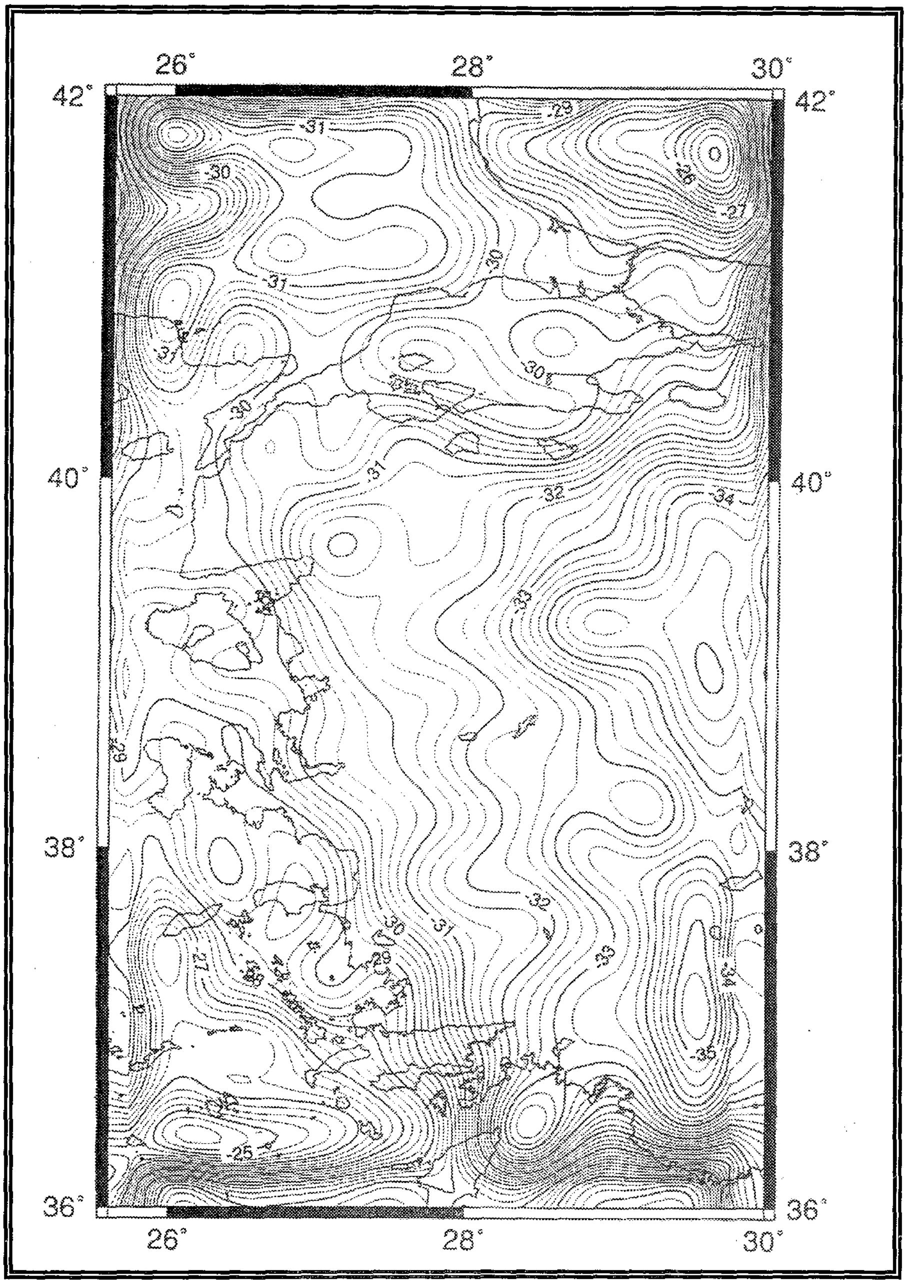

Figure. 4.3. Moho depth map of western Turkey and surrounding area obtained by gravity computation (after *Genç et al, 1996*).

supports the well-known continental type of crust in this coastal region. In the Cretan Basin deep sea drilling under the JOIDES project also indicates absence of oceanic material there. Concordantly, the sea floor morphology of the Aegean does not have any characteristics of that of oceanic crusts (*Arpat, 1976*). A minimum thickness was seen in the north of Crete. Seismic wave velocity contrast is also seen between the mainland and the Aegean Sea (e.g., *Alessandrini et al., 1997*). Lower velocities are remarkable beneath the Aegean Sea over the subduction zone beneath the Cretan basin and Cyclades (*Alessandrini et al., 1997*). Low thickness is also seen behind the Aegean island line and along the North Aegean Trough

(NAT) (Fig. 4.1, 4.2). It is obvious that beneath the Aegean Sea and surrounding area the

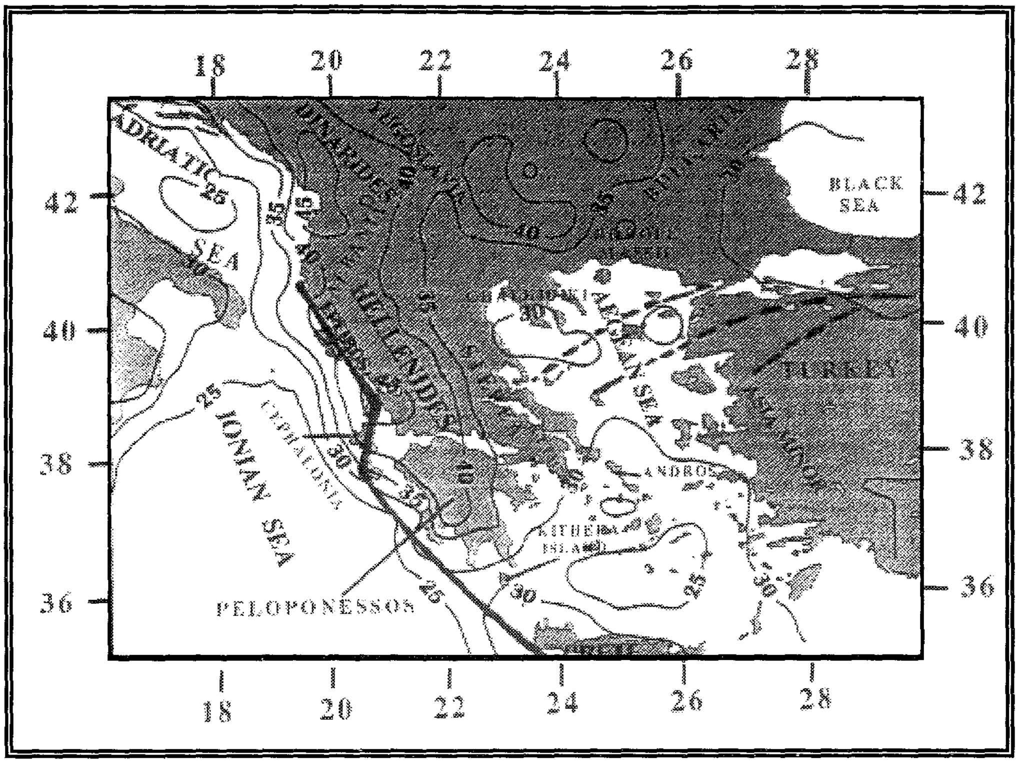

Figure. 4.4. Moho depth map of southeastern Aegean and surrounding area (after *Tsokas and Hansen, 1997*).

crust is very thin and thickens towards the east (central Anatolia), west (Greek mainland) and north (southern Bulgaria) (Fig. 4.2, 4.3, 4.4, 4.5, 4.6). Even the in Marmara region at the NE part of the region, this decrease of crustal thickness can be clearly seen (Fig. 4.7, 4.8). Decreasing Bouguer gravity with smooth gradient is also observed in Turkey towards the east. This indicates a slight increase in crustal thickness *(Özelçi, 1973; Şalk, 1994; Canıtez & Toksöz, 1982; Klingele & Medici, 1997). Oral (1987)* made the assumption that the eastward thickening rate in the western Anatolian crust is about 0.25 km in every 10 km.

Above the slab that is characterised by a high-velocity anomaly, an area of low-velocity anomalies is observed under the Aegean region in the Aegean Sea and western Anatolia because of the melting of the slab and a heating zone that is situated between the slab and the surface (e.g., *Alessandrini et al., 1997*). Because of this, energy of seismic waves is highly absorbed beneath the Aegean Sea (e.g., *Papazachos & Comninakis, 1971; Alessandrini et al., 1997*). The positive Bouguer anomalies beneath the Aegean basin extend in an area of high P-wave velocity (e.g., *Alessandrini et al., 1997*). Probably because the African plate subducts, it triggers upward migration of hot material in the mantle. Because of tectonic faults, permeability and support mass and heat transport increase in the region over the subducting

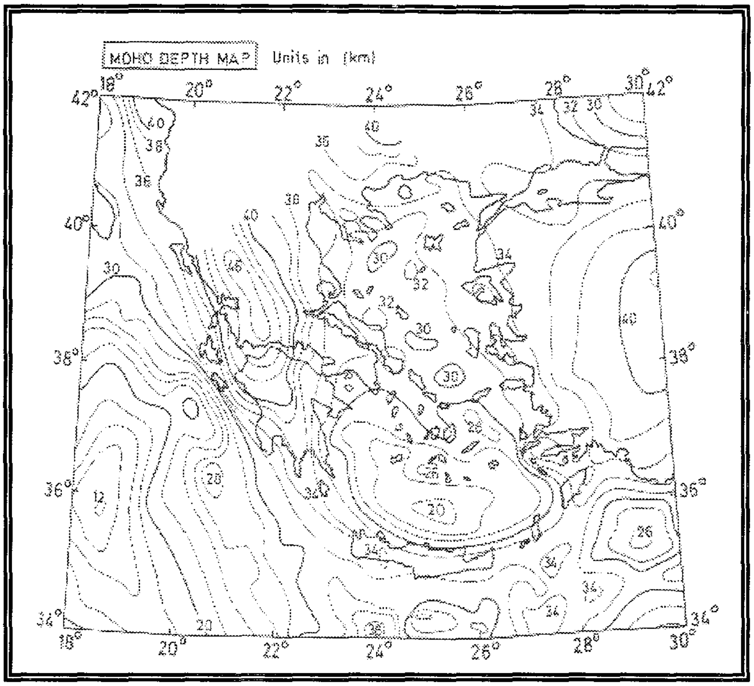

Figure. 4.5. Moho depth map of Aegean and surrounding areas (after *Makris, 1985*). Plotted by combining refraction and gravimetric data.

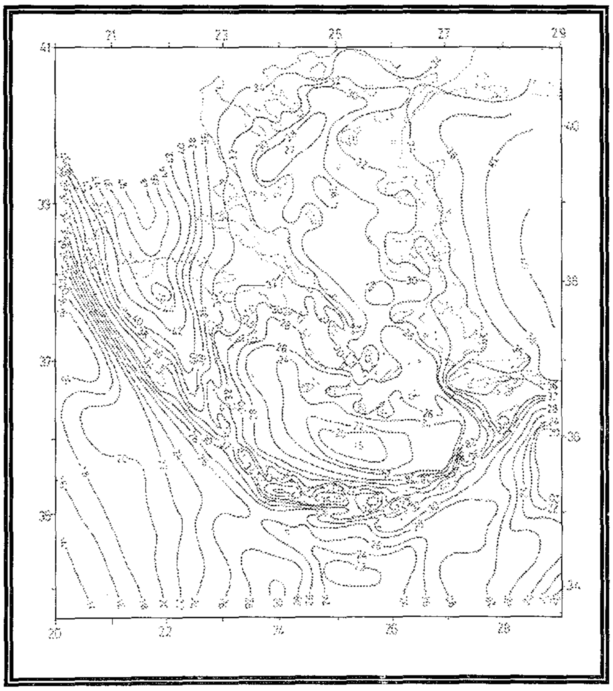

Figure. 4.6. Approximate crustal thickness (km) in the Aegean and surrounding area, derived from Moho-depth map (Fig. 4.5) published by *Makris (1976)*, with corrections including topography and bathymetry (after *Le Pichon and Angelier, 1981*).

slab. The low-velocity (*Papazachos & Comninakis, 1971; Agarwal et al., 1976; Yılmaztürk, 1989; Kalogeras & Burton, 1996; Taymaz, 1996*) and Sn attenuation (*Molnar & Oliver, 1969; Hashida et al., 1988; Rodgers et al., 1997*) anomaly in the Aegean back-arc correlates very well with an area of high heat flow observed predominantly behind the volcanic arc (*Erentöz & Ternek, 1968; Jongsma, 1974; Erickson et al.,1977; Cermák, 1979, Koçak, 1990*). A process of crustal thinning extends over a zone, which has a width of about 400 km both in the E-W and N-S directions (*McKenzie & Yılmaz, 1991*). A positive Bouguer gravity anomaly observed in southern Bulgaria, western Turkey, and Thrace (*Rabinowitz & Ryan, 1970; Özelçi, 1973; Şalk, 1994*) is interpreted as the eastern limits of the positive anomaly belt identified as a concave side of an island arc. The positive gravity anomaly seen inside the coasts of the western Turkey implies a contradiction between topography and gravity. The situation appears to be normal in the Greek mainland (*Özelçi, 1973*) (see *Yılmaztürk, 1989*). The Greek mainland is characterised by a higher seismic wave velocity field (*Yılmaztürk, 1989*) and the thickest crust in the region (*Makris, 1975*). In addition seismic wave velocities in western Turkey, especially in the Menderes Massive, are lower than the velocities in Greece (*Genç et al., 1996b*). This all suggests that the eastern Aegean (western Turkey) region is mostly under intense simple shear of the lithosphere, as is the central Aegean region. This could be caused by regional heating (*Oral, 1987*). This area is also characterised by normal to locally elevated terrestrial heat flow density (*Pfister et al., 1997*) compared to a normal value defined by the world wide continental mean value (*Pollack et al., 1993*). This suggests that beneath the area the crust is thin, or a denser material exists beneath the axis of the gravity high (*Özelçi, 1973*). While some authors suggest uprising of the region with dense upper mantle (*Özelçi, 1973; Yılmaztürk, 1989*) others suggest uprising of the upper mantle because of the extension which results from simple shear (*Wernicke, 1985; Oral, 1987*). Yet others suggest that an asthenospheric plume generates the extension (*Agarwal et al., 1976; Makris, 1976, 1978a,b; Kalogeras & Burton, 1996*).

In contrast to the Aegean Region, the heat flow map shows that low regional heat flow and uniform values are concentrated in the eastern Mediterranean Sea (*Erickson, 1970; Rabnowitz & Ryan, 1970; Ryan et al., 1969; Fytikas & Kolios, 1977*). In the Mediterranean Sea, the regional Bouguer anomalies are almost everywhere positive (*Morelli, 1990; Şalk, 1994*). The strongest positive Bouguer anomalies exist over the Ionian sea (*De Bruyn, 1955; Fahlquist, 1963; Woodside & Bowin, 1970; Makris, 1976; Morelli, 1990; Şalk, 1994*). But this gravity high is not the result of a shallow asthenosphere and low density crustal material, as it is in the Aegean. The crustal structure is of oceanic type for the Ionian (*Ferrucci et al., 1991; De Voogd et al., 1992*), Levantine (*Ginzburg & Ben-Avraham, 1987; De Voogd et al., 1992*) and Sirte basins (*De Voogd et al., 1992*). The Herodotus basin may be interpreted either as an

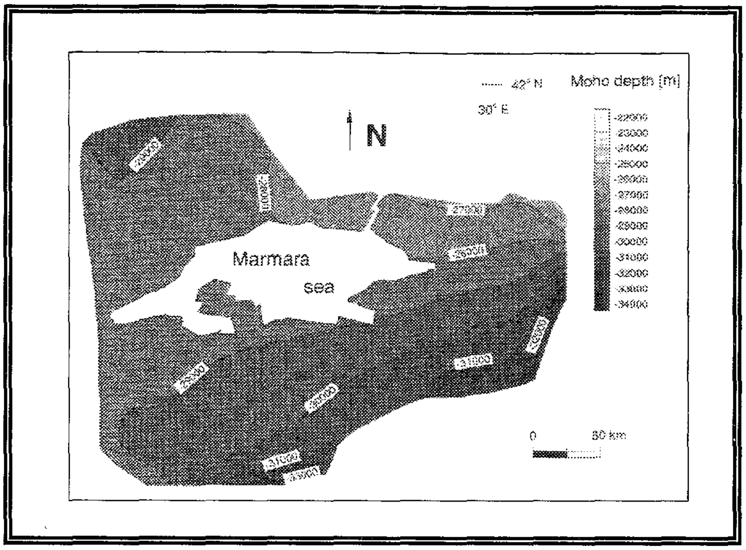

Figure. 4.7. Moho depth map of Marmara and surrounding area obtained by gravity computation (after *Klingele and Medici, 1997*).

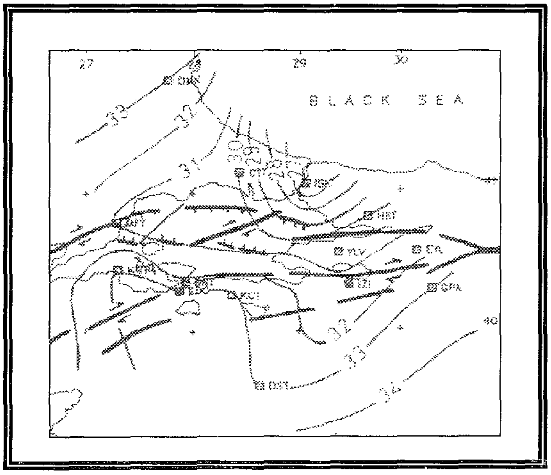

Figure. 4.8. Moho depth map of Marmara and surrounding area (after *Gürbüz et al., 1992*).

oceanic or a thinned continental crust (*De Voogd et al., 1992*). The Lg phase, which propagates in continental crust, was not observed at the southwestern coast of Turkey near Crete (*Rodgers et al., 1997*). This result suggests an oceanic crust for the Herodotus basin. According to the recent results of deep seismic profiles in the Ionian, Sirte and Herodotus abyssal plains, the three basins have a relatively thin crust (8 to 11 km) (*De Voogd et al., 1992;* see also *Lort et al., 1974; Agarwal et al., 1976; Cloetingh et al., 1980; Makris & Stobbe, 1984; Makris et al., 1986*). The Crust of all three basins is overlain by a thick sedimentary cover (8-12 km) (*De Voogd et al., 1992*). The Herodotus basin is covered by thicker sediment than the Sirte and Ionian abyssal plains (at least 10 12 km) (*De Voogd et al.,* *1992*). The Mesozoic to Tertiary sequences suggest an old age for these basins (*Finetti, 1976; Ginzburg & Ben-Avraham, 1987*). The restored water depth suggests that the Herodotus basin is probably significantly older than the Ionian and Sirte basins, and is possibly Triassic (*De Voogd et al., 1992*). For this reason, identification of crustal type is difficult.

The origins of the Mediterranean ridge are one of the most obvious features in the Mediterranean region that is the subject of ongoing debate. In the eastern Mediterranean the magnetic field is strikingly uniform and undisturbed (*Vogt & Higgs, 1969; Emery et al., 1966*) (see section 5.3.). There is no magnetic lineation observed beneath the Mediterranean ridge (*Ryan et al., 1969*). For this reason one cannot say that the crust is oceanic, as it lacks the magnetic lineations that are seen over all the worlds other oceanic crusts.

It is obvious that the African plate is being consumes at the Aegean Subduction Zone. Relatively higher velocities are seen beneath the volcanic arc and in the Cretan Basin under the low velocity zone (*Spakman, 1986; Meulenkamp et al., 1988; Spakman et al., 1988; Granet & Trampert, 1989; Yılmaztürk, 1989; Ligdas et al., 1990; Ligdas & Main, 1991; Küleli, 1992*). The dipping high velocity anomaly (positive P velocity) (*Agarwal et al., 1976*) is consistent with the existence of a cooler, and hence more dense (*Davies & McKenzie, 1969; Sleep, 1973*) and possibly chemically different (*Meulenkamp et al., 1988*), descending old lithosphere. The high-velocity heterogeneity follows the Benioff zone seismicity at about 200 km (e.g., *Papazachos, 1973; McKenzie, 1978; Le Pichon & Angelier, 1981*) and penetrates deeper than the Benioff Zone (*Spakman et al., 1988; Granet & Trampert, 1989; Ligdas & Main, 1991; Küleli, 1992*). The maximum depth of earthquakes can be understood either in terms of heat absorption or rheological deactivation (*Ligdas & Main, 1991*). Thus, the absence of earthquakes does not necessarily mean that the slab does not exist. It could simply be an indication that it is too hot to undergo either brittle failure or plastic instability, while remaining cool enough to the surrounding mantle to produce a velocity contrast which can be picked up by tomography (*Spakman et al., 1988*). In this case subduction velocity is a important parameter. The Mediterranean slab is continuous to approximately 1500 km depth according to a regional tomogram of *Spakman et al. (1993)* (see *Van der Hilst et al., 1997*).

In Köyceğiz is the southwestern Turkish mainland, negative Bouguer anomalies are explained as the eastern extension of the South Aegean Subduction Zone which causes a decrease in density so that Bouguer gravity unusually becomes negative along this corridor (*Woodside,* *1975; Oral, 1987; Şalk, 1994*). This anomaly indicates possible eastward extension of the subduction zone.

Granet & Trampert (1989), who studied seismic wave velocities, indicate that the slab was broken at depth of 250 km. *Spakman et al. (1988)* pointed out that a vertical slab rupture below western Turkey initiated and facilitated slab rollback within the subduction.

Papazachos & Comninakis (1977) observed a compressional focal mechanism in shallow earthquakes and some intermediate depth earthquakes beneath the North Aegean Sea (see also *Boccaletti et al., 1974; Papazachos, 1973, 1976; Papadopoulos, 1997*). They interpreted these results as a second northward dipping subduction zone in the central Aegean region at the back of the South Aegean Arc (SAA). *Papazachos (1976)* further argues that magnetic and gravity anomalies that are associated with a probable second subduction zone. It presumed that the Oligocene-Miocene magmatic activity in the North Aegean region was produced by this subduction system (*Innocenti et al., 1981*) (see also *Fytikas et al., 1984*). The differences between the results for the paths crossing the southwestern, southern and southeastern Aegean are possibly explained by the second subduction zone (*Papazachos et al., 1995*). A cluster of the hypocenters has been located in the same region (*Küleli, 1992*). A high velocity perturbation found almost the same area was interpreted as an old subduction slab by *Küleli (1992)*. But these high velocity perturbations do not show a clear continuity in the vertical and horizontal cross sections (*Küleli, 1992*). Therefore, because of the unreliability of hypocenter locations and poorly estimated velocity perturbations on the tomographic image, this blurred image could not be interpreted as a structural feature (*Küleli, 1992*). *McKenzie (1978)* pointed out the absence of evidence of thrusting on the scale required, or of associated volcanism and lack of control in the depth determination of intermediate shocks (see also *Udias, 1985*). *Küleli (1997)* argues that the depth estimation of two events shows important differences compared with ISC depth estimations and they show rather large standard errors relative to their depths. Therefore, these hypocentres that are determined from local stations were not interpreted as a seismic part of the second active subducted slab in the Aegean region (*Küleli, 1997*). The recent seismic tomographic studies (*Spakman, 1989*) do not corroborate any second subduction zone.

Recently, a surprising new tomographic feature was found which is a deep-reaching mantle plume in the Ionian Sea, which separates the two subduction slabs (see *Mueller et al., 1997*).

Table. 4.1. Crustal thickness estimations in the Aegean area.

CRUSTAL THICKNES ESTIMATIONS

References	***Approx. crustal thickness***	***Method used based on***	***Locations***
			AEGEAN REGION
Drakatos et al. (1997)	30 km		Aegean Region
Papazachos er al. (1995) Drakatos et al. (1997)	40 km		South Aegean Island Arc
			EASTERN TURKEY
Canıtez and Toksöz (1982)	40 (45) km	seismic wave velocity	Eastern Turkey
			MARMARA
Öcal (1960)	>30 km	Rayleigh wave	Marmara
Crampin and Üçer (1975)	24 km	seismic wave velocity	Marmara
Makris (1978a,b)	34 km		Marmara
Canıtez and Toksöz (1982)	30 km	seismic wave velocity	Marmara
Gürbüz and Üçer (1985)	25 km	P wave velocity from a quarry blast experiment	Marmara
Gürbüz et al. (1992)	27-34km	seismic wave velocity	Marmara
Klingele & Medici (1997)	25-32km	gravity computation	Marmara
Kenar (1977)	25-26 km	P wave velocity	Istanbul
Kenar (1978)	30 km	seismic wave velocity	Istanbul (D-B)
Kenar (1978)	25-26 km	seismic wave velocity	Istanbul (K-D)
Kenar (1978)	28-30 km	inversion of body- wave spectra	Istanbul
Gürbüz and Üçer (1988)	28 km	P wave velocity from a quarry blast experiment	Istanbul
Nacioglu et al. (1981)	30-32 km	seismic wave velocity	Istanbul
Gürbüz et al. (1980) and Kenar (1978)	28-29 km	Pn velocity from a quarry blast experiment	Adapazarı, south of the Marmara sea
İlkışık et al. (1996) in1980	40-44 km	magnetotelluric soundings	Kırklareli, Thrace
İlkışık et al. (1996) in1980	28-32 km	magnetotelluric soundings	Central Thrace
İlkışık et al. (1996) in1980	32-35 km	magnetotelluric soundings	Tekirdag, south of Thrace
			BLACK SEA
Canıtez and Toksöz (1982)	30 km	seismic wave velocity	Black Sea
			MEDITERRANEAN
Voogd et al. (1992)	8-11 km	deep seismic sounding	Ionian Sea
Şalk (1994)	20-25 km	gravity computation	Ionian Sea
Makris (1977)	26 km	deep seismic sounding and gravity computation	west coast of the Peloponnesos
Payo (1969)	23 km	short-pedod surface waves	south of Greece
Payo (1996)	23 km	short-pedod surface waves	east of Greece
Voogd et al. (1992)	8-11 km	deep seismic sounding	Herodotus abyssal plain
			WESTERN TURKEY

Table. 4.1. (continue)

Canıtez (1975)	36 km	surface wave dispersion	western Turkey
Canıtez and Toksöz (1982)	35 km	seismic wave velocity	western Turkey
Oral (1987)	40 km	gravity computation	western Turkey
Kalafat et al. (1987)	29-32 km	seismic wave velocity	western Turkey
Alptekin et al., (1990)	35-40 km	seismic wave velocity	western Turkey
Makris and Stobbe (1984), Meissner et al. (1987), Ezen (1993)	28-32 km	various	western Turkey
Horasan and Canıtez (1995)	31 km	SSBDS	western Turkey
Alptekin et al. (1990)	31 km	Dispersion of surface waves	Istanbul – Burdur (N-S Profile)
Makris (1973)	32-34 km	seimic and gravity computations	western coast of Turkey
İlkışık (1980)	35 km	magnetoteulliric sounding	western coast of Turkey
Oral (1987)	36-37 km	gravity computation	western coast of Turkey
Makris (1973)	35 km	seimic and gravity computations	southwestern coast of Turkey
Oral (1987)	30 km	gravity computation	southwestern coast of Turkey
İlkışık (1990)	25 km	magnetoteulliric	Med. Coast of SW Turkey
İlkışık (1990)	20 km	magnetoteulliric	Lycian Nappes
İlkışık (1986) (Oral 1987)	47 km	magnetoteulliric	Teke Peninsula
Şalk (1994)	25 km	gravity computation	Menderes massive
İlkışık (1990)	42 km	magnetoteulliric	Menderes massive
İlkışık (1996)	14 km	magnetoteulliric	Gediz Graben (Salihli-Alaşehir)
Oral (1987)	43 km	gravity computation	central Turkey
			TURKEY-IRAN
Kenar and Toksöz (1980)	40-43 km	Love, Rayleigh wave dispersion	from Iran to Istanbul
			AEGEAN SEA
Papazachos, Polatou & Mandalos (1967) and Papazachos (1969)	35-45 km	seismic wave velocity	Aegaen Sea
Calcagnile et al. (1982)	35 km	surface wave dispersion	Aegaen Sea
			NORH AEGEAN SEA
Makris (1978)	40 km	deep seismic sounding and gravity computation	North Aegean Sea
Makris (1977)	32 km	deep seismic sounding and gravity computation	north of Evia
Papazachos et al. (1966)	33 km	seismic wave velocity	Athens, Norh Aegean Sea
Canıtez (1969)	32 km	Love and Rayleigh waves	Athens, Norh Aegean Sea
			SOUTH AEGEAN SEA
Öcal (1960)	>30 km	Rayleigh wave	southern Aegean Sea
Canıtez (1969)	35 km	Love and Rayleigh waves	southern Aegean Sea
German and Greek teams Morelli (1985)	20 km	deep seismic sounding	southern Aegean Sea
Makris (1976)	20 km	gravity computation	southern Aegean Sea

Table. 4.1. (continue)

Chailas et al. (1992)	20 km	gravity computation	southern Aegean Sea
Papazochos (1993)	20 km	various	southern Aegean Sea
Papazachos (1994)	20 km	gravity computation	southern Aegean Sea
Papazachos (1995)	25km	seismic tomography computation	southern Aegean Sea
Makris (1978)	20-22 km	deep seismic sounding and gravity computation	Cretan Sea
Makris (1975)	20 km	deep seismic sounding and gravity computation	central Cretan Sea
Makris (1975)	28 km	deep seismic sounding and gravity computation	eastern Crete Sea
Makris (1975)	25 km	deep seismic sounding and gravity computation	western Crete Sea
			CENTRAL AEGEAN SEA
Papazachos et al. (1966)	20-35 km	various	central Aegaen Sea
Makris (1973, 1977, 1975), Panagiotopoulos (1984), Kiriakidis (1988) and Kalogeras & Burton (1996) Makris (1978)	26 km		central Aegaen Sea
			GREECE
Makris (1975)	42-46 km	deep seismic sounding and gravity computation	Greece
Papazachos et al. (1966)	43 km	seismic wave velocity	along the Greek Peninsula
Makris (1976)	40 km	gravity computation	along the Greek Peninsula
Chailas et al. (1992)	40 km	gravity computation	along the Greek Peninsula
PapazQchos (1993)	40 km	various	along the Greek Peninsula
PapazQchos (1994)	40 km	gravity computation	along the Greek Peninsula
PapazQchos (1995)	40 km	seismic tomography computation	along the Greek Peninsula
Calcagnile et al. (1982)	36-42 km	surface wave dispersion	along the Greek Peninsula
Makris (1978)a, b	45 km	deep seismic sounding and gravity computation	western Greece
Panagiotopoulos (1984), Kiriakidis (1988) and	45 km	various	western Greece
Chouiaras et al. (1997) Kalogeras & Burton (1996)	28 km	magnetotelluric	Central Greece (Gulf of Corinth)
Papazachos et al. (1966)	32-47 km	short-pedod surface waves	south of Greece
Makris (1977)	46 km	deep seismic sounding and gravity computation	central Peloponnesos
Makris (1977)	34 km	deep seismic sounding and gravity computation	eastern Peloponnesos
Makris (1978)	46 km		Peloponnessos
Şalk (1994)	40 km	gravity computation	Peloponnessos
Makris & Vees (1977)	30-34 km	deep seismic sounding and gravity computation	west of Crete
Kalogeraş & Burton (1996)	45 km	wave velocity	western arc, southwest of Crete
Calcagnile et al. (1982)	31-47 km	surface wave dispersion	**Macedonia, south Yugoslavia**

5. TECTONICS

Beneath eastern Turkey the Arabian plate moves northward with respect to Eurasia at a rate of about 25 mm/yr, based on GPS measurements, global kinematic models, the analysis of oceanic spreadings, fault systems, and earthquake slip vectors (see Table. 5.1. for detail) (*DeMets et al., 1990, 1994; Barka & Reilinger, 1997; Reilinger et al., 1997*). Between these plates continental collision is taking place and the east of the Anatolian mass appears as a buffer. For this reason, thickening and shortening is taking place in the compression direction in the east of Anatolia. To the west, the South Aegean Subduction Zone is formed to accommodate the final stages of convergence of Europe and Africa (*McKenzie, 1970*). Africa moves northward relative to Europe at a rate of about 10 mm/yr (*Chase, 1978; Argus et al., 1989; DeMets et al., 1990, 1994*), but across the South Aegean Trench the relative motion is thought to be approximately 40 to 50 mm/yr towards the southwest (*McKenzie, 1978; Le Pichon & Angelier, 1979; Jackson et al., 1994; Oral et al., 1995*). This was resulted from diffuse deformation (extension) of the Aegean area, behind the South Aegean Arc (SAA).

Because of no boundary to the west and a compressional regime to the east, the thickened mass (Anatolian tampon plate) escapes towards the west from the regime. Further west, the anomalous low elevated extensional area (The Aegean region) allows Anatolia to move with increasing velocity to the W - SW. A possible kinematic model has been derived by *Westaway (1994)* (see also *Oral, 1994; Oral et al., 1995; Le Pichon et al., 1995; Reilinger et al., 1997*) for the motion of Anatolia. According to the model, which is based on geodetic measurements and information from surface geology, the westward movement of the Anatolian mass (26mm/yr along the North Anatolian Fault) occurs as part of an Euler rotation. The rotation pole is situated around the Sinai peninsula. But the lithosphere of Anatolia is not behaving like a typical tectonic plate. This is because the western part of the plate, called the Aegean region, could be described as an internally deforming part of the Anatolian plate (*Reilinger et al., 1997*) and is a diffuse boundary between the Eurasian and African plates. But the velocities measured within the plate are parallel to each other and fit an Euler vector, so it could be considered as a plate. But the lengths of the velocity vectors systematically increase to the west-southwest

Table. 5.1. Obtained velocity rate and direction in various parts of the Aegean region. The numbers in METHOD column indicated the method used to measure rate and/or direction vector: (i) earthquake data, (ii) geodetic data, -s- space geodetic, -c- convensional, (iii) fault observation, (iv) paleomagnetic, (v) kinematic arguments.

Author	Rate	Direction	Method	Measured	Reference
Le Pichon (1968)	28 mm/yr		vii	EURASIA	AFRICA
Chase (1978)	10 mm/yr	180º		EURASIA	AFRICA
DeMets et al. (1990, 1994)	25 mm/yr		vi, i	ARABIA	EURASIA
Reilinger et al. (1997)	24 ±2 mm/yr	N	ii (s)	ARABIA	EURASIA
Barka and Railinger (1997)	23 ±2 mm/yr	N	ii (s)	ARABIA	EURASIA
Papazachos and Kiratzi (1996)	4 mm/yr	49º	i	ADRIATIC	EURASIA
McKenzie (1972)	33 mm/yr	W	vi	ANATOLIA	EURASIA
Jestin et al. (1993)					
Reilinger et al. (1997)	22 mm/yr		ii (s)	ANATOLIA	EURASIA
Eyidoğan (1988)	24 mm/yr		i	MARMARA	EURASIA
Straub and Kahle (1996)	22 mm/yr		ii (s)	MARMARA	EURASIA
Oral et al. (1992)	50± 20 mm/yr		ii (s)	W TURKEY	EURASIA
Jackson and McKenzie (1984a)	32 mm/yr		i, iii	NAF	
Jackson and McKenzie (1988)	39 mm/yr		i	NAF	
Barka (1992)	6-11 mm/yr		iii	NAF	
Oral et al. (1995)	25 mm/yr		ii (s)	NAF	
Le Pichon et al. (1995)	40 mm/yr		ii (s)	NAF	
Straub and Kahle (1995)	15 ±4 mm/yr		ii (s)	NAF	
Reilinger et al. (1997)	30 ±2 mm/yr (ub.)		ii (s)	NAF	
Barka and Railinger (1997)	26 ±3 mm/yr		ii (s)	NAF	
Cianetti et al. (1997)	20 mm/yr		vii	NAF	
Barka and Gülen (1988)	25-35 mm/yr		iii	NAF	
Barka ve Toksöz (1988)	33 mm/yr		iii	NAFZ west of northern branch	
Barka and Railinger (1997)	11 mm/yr		ii (s)	EAF	
Kahle et al. (1995)	25 mm/yr		ii (s)	CF	
Papazachos and Kiratzi (1996)	3 mm/yr	N45ºE	i	CF	
Mueller et al. (1997)	40 mm/yr			CF	
Zanchi and Angelier (1993)		SSW	i	W TURKEY (extension)	
Westaway (1994)	5 mm/yr		iii	W TURKEY (extension)	
Oral (1994), Oral et al. (1995)	10 mm/yr		ii (s)	W TURKEY (extension)	
Barka and Railinger (1997)	15 mm/yr	SW		W TURKEY (extension)	
McKenzie (1972)		220º	i, iii,vi	AEGEAN (extension)	
Tselentis et al. (1986)	20 mm/yr	218-197º	i	AEGEAN (extension)	
Jackson and McKenzie (1988)	20-60 mm/yr		i, vi	AEGEAN (extension)	
Taymaz et al. (1991)	38 mm/yr		vi	AEGEAN (extension)	
Jackson et al. (1992)	30 mm/yr		i	AEGEAN (extension)	
Oral et al. (1995)	30-50 mm/yr		ii (s)	AEGEAN (extension)	
Armijo et al. (1992)	30-50 mm/yr		iii	AEGEAN SEA (extension)	
McKenzie (1978)	40-60 mm/yr			S. AEGEAN	EURASIA
Lepichon and Angelier (1979)	52 mm/yr	227º	vi	S. AEGEAN	EURASIA
Jackson and McKenzie (1988)	20-60 mm/yr		i	S. AEGEAN	EURASIA
Jackson (1993)	40-50 mm/yr		various	S. AEGEAN	EURASIA

Table. 5.1. (continued)

Lepichon et al. (1995)	30 mm/yr		ii (s)	S. AEGEAN EURASIA
Straub and Kahle (1996)	40 mm/yr	SW	ii (s)	S. AEGEAN EURASIA
Reilinger et al. (1997)				
Cianetti et al. (1997)	30 mm/yr		vi	S. AEGEAN EURASIA
Mueller et al. (1997)	40-45 mm/yr	SW	ii (s)	SW AEGEAN EURASIA
Cianetti et al. (1997)	30 mm/yr	SW	vi	SW AEGEAN TURKEY
Tselentis et al. (1986)	0.3 mm/yr	N-S	i	C. GREECE EURASIA
Tselentis et al. (1986)	0.6 mm/yr	E-W	i	C. GREECE EURASIA
Taymaz et al. (1991)	38 mm/yr	SW	vi	C. AEGEAN EURASIA
Billiris et al. (1991)	10-20 mm/yr		ii (c)	C. GREECE EURASIA
Taymaz et al. (1991)		345º±16	i	C. GREECE EURASIA
Roberts and Jackson (1991)		305º±15	iii	C. GREECE EURASIA
Papazachos et al. (1992)	6 mm/yr	205º	i, iii	C. GREECE EURASIA
Clarke et al. (1997)	13 mm/yr			W. Corinth (extension)
Clarke etal. (1997)	6 mm yr			E. Corinth (extension)
Billiris et al. (1991)	12±2 mm/yr	NE-SW	ii (s)	Corinth and Sporades (extension)
Briole et al. (1993)	10 mm/yr			Corinth (extension)
McKenzie (1972)		191º	i, vi	S. AEGEAN AFRICA
McKenzie (1978)	70 mm/yr<	211º	i, vi	S. AEGEAN AFRICA
Jackson and McKenzie (1988)	60 mm/yr<		i	S. AEGEAN AFRICA
Taymaz et al. (1990)	60 mm/yr<	205º	i	SW. AEGEAN AFRICA
Oral (1994)	50±10 mm/yr		ii (s)	SW. AEGEAN AFRICA
Oral (1994)	20 ± 10 mm/yr		ii (s)	SE. AEGEAN AFRICA

(in The Aegean region) because of the "plate" enters in an extensional condition to the west-southwest.

It is well known that the Alpine – Himalayan belt has an amalgamated structure that deformed and faulted in older earlier times. For this reason the Aegean region, as a part of this belt, exhibits a complex pattern with basins of different age and nature, created, overprinted, and obscured by the 180 million year history of the interaction between the African and European plates (see Section 3). Since the starting of the last post-orogenic extensional tectonic (neotectonic) regime of intense deformation, this crustal mass has been failing very easily because of reactivation of the older discontinuities (*Arpat & Bingöl, 1969; Kaya, 1981; Zanchi et al., 1990; Zanchi & Angelier, 1993; Oral et al., 1995*). Because of this, the recent deformation features do not simply show the actual recent tectonic regime and deformation.

All the areas of Greece, western Turkey, Albania, southern Yugoslavia and Bulgaria that are dominated by strike-slip and normal faulting are considered together in this domain, which covers an area of approximately 500 x 500 km (*Jackson & McKenzie, 1988*). Aegean extensional region the most seismically active and rapidly internally deforming region of the entire Alpine-Himalayan belt and the continents in the world today (Fig. 5.1, 5.2) (*McKenzie, 1972; Mercier et al., 1977; Ekström & England, 1989; Jackson et al., 1992*).

It was recognised in 70's that continental lithosphere deformation differs from oceanic lithosphere deformation (e.g., *McKenzie, 1970*). Continental lithospheres are not as compact as oceanic plates because of their chemical composition and physical nature. Oceanic plate is homogenous because it is created along oceanic ridges with fresh and strong material. For this reason oceanic plates are rigid and hardly deform. On the other hand continental crust is not produced from magma as a unique material. It consists of magmatic materials and sedimentary rocks of various ages. In some areas metamorphic rocks of one known age (metamorphism age) cover large areas, but there are also older weakness zones. It does not renew like oceanic crust. Continental crusts in the world have all the evolution of the earth's crust. For this reason, in contrast with oceanic plate boundaries, continental plate boundaries deform diffusely (Figure. 5.1, 5.2). But the appropriate model to describe this kind of deformation in continents (i.e., continuum deformation versus microplate or block behavior) remains the subject of debate (*Thatcher, 1995; Reilinger et al., 1997*).

At present there are two kinematic models thought to describe the diffuse deformation in the Aegean region. Block rotations, which are a common feature of continental tectonics, are driven by forces on their lateral boundaries (*Taymaz et al., 1991*) or by the flow of the continuously deforming (continuum deformation) lithosphere beneath them (*Argand, 1924; Tapponnier, 1977; McKenzie & Jackson, 1983; Jackson & McKenzie, 1984b; Walcott, 1984; England et al., 1985*). Obtaining average stress tensors from the slip directions on distributed faults.is a technique to show stress direction. By using this technique stress trajectories in the Late Cenozoic have been mapped in the Aegean region by *Angelier (1979), Angelier & Le Pichon (1980)* with the consideration of field geology data that is derived from faults on lands

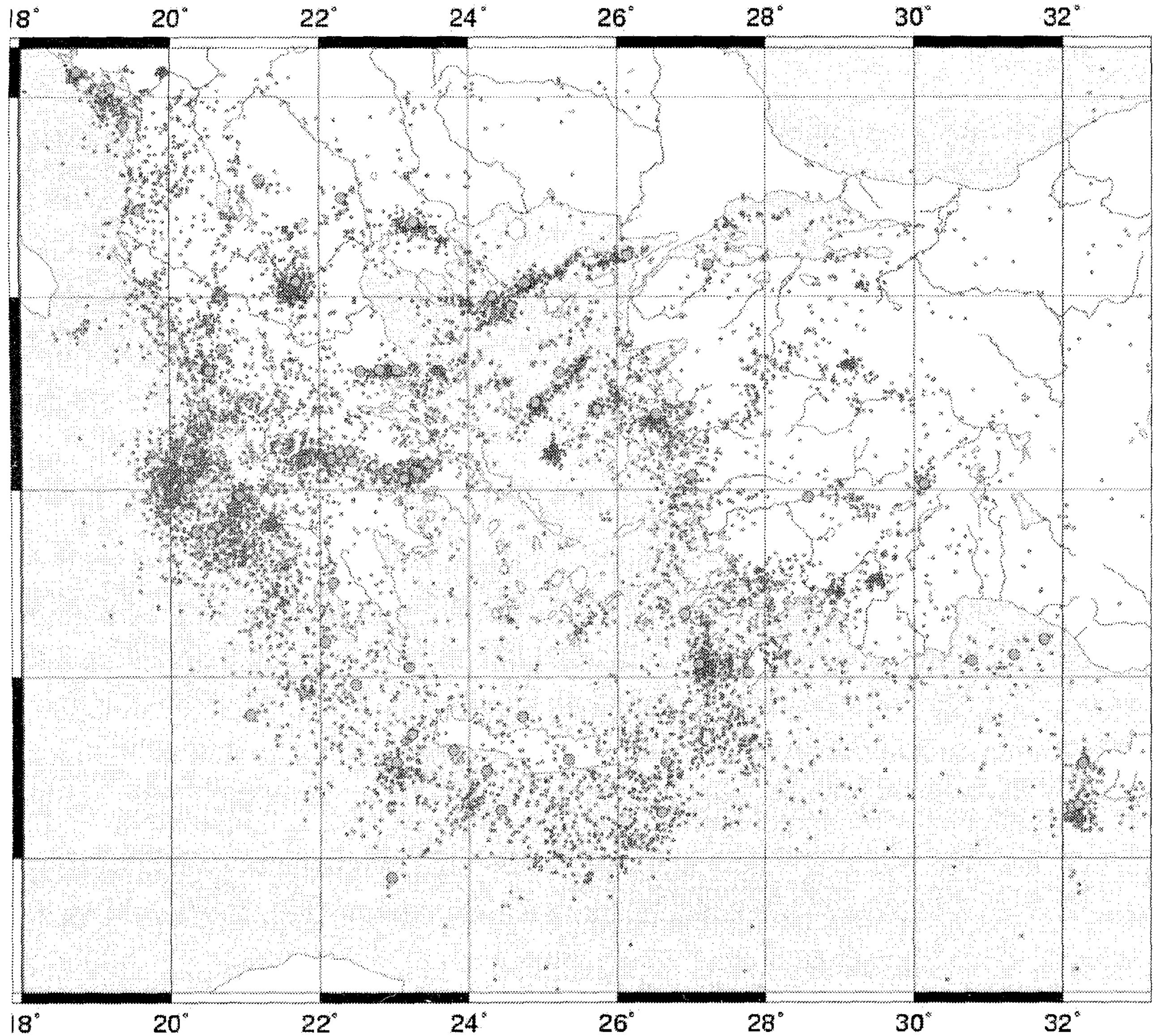

Figure. 5.1. Seismicity of the Aegean. Epicenters of earthquakes with depths between 0 and 50 km reported by the USGS PDE during the period 1973 to 1997. Earthquakes with magnitudes >5.5 are shown by bigger circles.

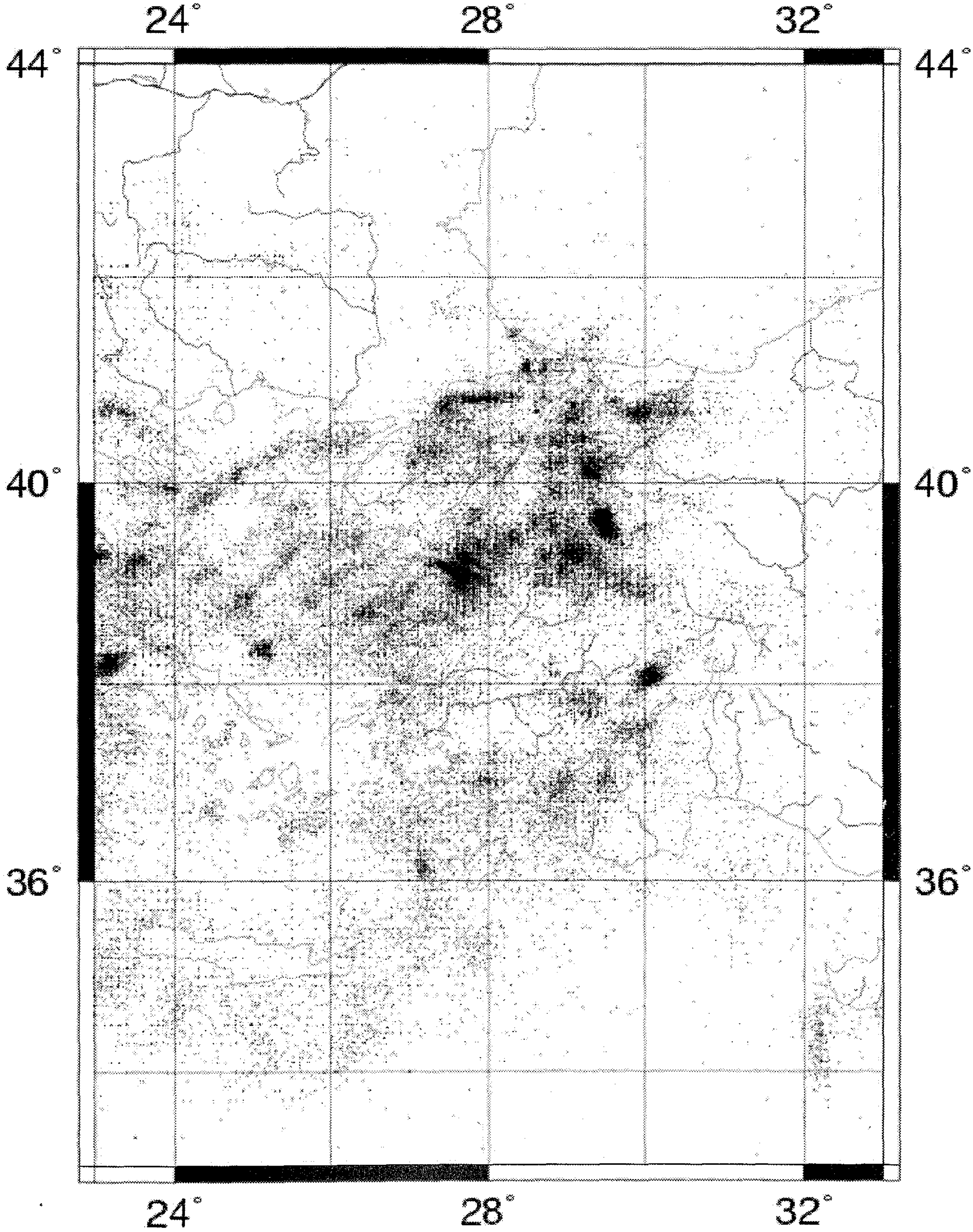

Figure. 5.2. Earthquakes in the central and eastern Aegean from 1976 to 1997. Source: B.U. KOERI.

and slip vectors from earthquake data determined by *McKenzie (1978)* and *Ritsema (1974)* (see also *Carey & Brunier, 1974; Angelier & Goguel, 1979*). The mapped stress trajectories show maximum stress directions on the crust that is assumed to be deforming continuously. They locate a pole of rotation in the southern Adriatic Sea to describe the motion between the southern Aegean and Africa. This model was later modified (*Angelier et al., 1982*) to account for the lack of observed palaeomagnetic rotation in Crete (*Valente et al., 1982*). The amount of subsidence is also used for reconstruction of the deformed southern Aegean region. Results show that, in the Pleistocene, stress trajectories trend in a clockwise direction to the SW. The velocity model constructed by *Le Pichon et al. (1995)*, based on geodetic data, shows that Greek mainland is rotating clockwise with respect to a pole close to Athens. *Jolivet et al. (1994)* similarly drew stress trajectories since the early Miocene by using structural data of metamorphic core complexes, showing stretching direction. The model is subjected to ductile simple shear of lower and leading regime on the upper crust (see also *Arpat & Bingöl, 1969; Wernicke, 1981; Lister et al., 1984; Bozkurt et al., 1993*) (see section 2.3). Ductile simple shear occurring by continuous ductile flow in the lower crust and for a decoupling between the upper and lower crust. *Sonder & England (1989)* used a thin viscous sheet model to investigate thermal dependent rheology on large-scale continental extensional areas. This model is consistent with volcanic data that comes from alkaline volcanics (*Gülen, 1990*) (see Section 2.2.2). The calculated results are also agree with paleomagnetic data in the Aegean region. *Hatzfeld et al. (1997)* presented an analog model of a layer spreading under gravity to describe the Aegean extensional area. A corresponding theory suggests that the lithosphere deforms as a continuum. The laboratory experimental results agree with geodetic measurements, strain pattern from microearthquake studies and paleomagnetic rotations (*Hatzfeld et al., 1997*). But the real stress distribution is not the behaviour of an amalgamated structure because the strain is localised not only by real stress but also with preexisting weakness zones. These zones lead to possible errors in estimating stress distribution.

Jackson & McKenzie (1984b), *Anderson & Jackson (1987)* and *Taymaz et al., (1991)* thought that the correlation between topography and seismicity did indeed imply a relatively rigid behavior of aseismic blocks (see also *McKenzie, 1970, 1972; Nowroozi, 1972; Papazachos, 1973; McKenzie, 1978*). A simple deformation model proposed by *Taymaz et al. (1991)*, using paleomagnetic data, explains the deformation in the Aegean region as block rotations and strike slip faulting (see also *Jackson et al., 1992*). The model consists of a set of broken slats that are pinned at their western and eastern ends to pivoting screws. They rotate to the east counterclockwise and to the west clockwise. These rotations are concordant with kinematic data. The model is subjected to E-W shortening and the reason for this is N-S extension (see also *Brunn, 1976; Dewey & Şengör, 1979*). According to this model, velocities decrease to the south. But geodetic measurements show that the velocities increase to the south towards the South Aegean Consuming Boundary (*Oral, 1994; Oral et al., 1995*). In this respect the model does not work.

According to shallow seismic data and morphology, sediment fill of grabens in the Aegean Sea clearly show evidence of extension (e.g., *Mascle & Martin, 1990*). In the Aegean region the subsidence decrease towards the west and east to the both mainlands of Greese and Turkey, respectively. An extension of 50 % is calculated in the eastern Aegean region (*McKenzie, 1978; Le Pichon & Angelier, 1979* see also *Yılmaz, 1997*), 30 to 80 % in the south Aegean region, and 50 to 70 % in the North and central Aegean (*Le Pichon & Angelier, 1979*). Good correlation exists between the morphology, elevatetion trend and crustal thickness (*Küleli et al., 1993*). The region is covered by listric faults whose dip angles decrease with depth (e.g., *Arpat & Bingöl, 1969; McKenzie, 1978; Şengör, 1979; Jackson & McKenzie, 1984a; Jolivet et al., 1994*). Such faults lead to more extension than is measured in the field. Extensional stresses in the central Aegean region reflect isostatic subsidence and uplift of brittle crustal blocks in response to crustal thinning (*McKenzie, 1978*). The Aegean lithosphere is accommodated by long low-angle shear zones. Stretching indicators on gneiss domes (footwalls of low angle listric faults) in the Cylades and Menderes Massives parallel the maximum extension direction of the region (*Jolivet et al., 1994*). Along such zones a supracrustal slab became detached and began to slide in the extension direction (*Meulenkamp et al., 1988*). The broken slat model did not show these mechanisms.

Tectonic Scheme

From the north to the south the deepest parts of the region appear as intersections of NE trending faults to the east and NW trending faults to the west of the Aegean region. The front of a NW - SE trending elevated morphological feature that traverses west of Aegean Sea almost entirely (Evvoia, Tinos, Naxos, Amargos) the intersections of NW trending faults to the east with NE trending faults to the west appears as NW-SE trending one of the most deepest line (Aegean axial trough) in the Aegean region. This trough is the biggest of the NE trending grabens that represent eastern Aegean morphology. Further north, trending of the deepest parts of the region is related to the North Aegean Trough (NAT) which is one of the most remarkable features of the region. To the south, the remarkable E-W trending Crete basin appears parallel to the Crete.

There is evidence from palaeomagnetism that the eastern Aegean region rotates clockwise relative to stable Europe (*Laj et al., 1982; Kissel et al., 1985; Taymaz et al., 1991; Barka et al., 1995*). On the other hand, in the western Aegean region, clockwise block rotations occur in the Greek mainland and Aegean islands near the mainland (*Taymaz et al., 1991; Barka et al., 1995*) (see also *Le Pichon et al., 1995*). A model that based on geodetic data shows the same pattern of rotation that is obtained from paleomagnetic data (*Le Pichon et al., 1995; Straub & Kahle, 1995*). Normal faulting with some right lateral strike-slip is dominant in the east. Towards the north, in and surrounding the Sea of Marmara, normal faulting with E-W strike and ENE striking right-lateral strike-slip faulting are dominant (*Taymaz et al., 1991*).

5.1. THE EASTERN AEGEAN REGION

In the Aegean region it is obvious from morphology and seismic profiles that active normal fault systems created deep basins known (Sections 2.5., 2.7.). The most continuous geomorphologic escarpments in western Turkey are oriented ENE and appear to be the footwalls of major active normal faults. These escarpments are prominent on satellite images, topographic maps, in the field, and as gravity lows in Bouguer gravity maps (*Oral, 1987; Sarı et al., 1995; Genç et al., 1996b*). Many identified and explored geothermal areas are situated along the faults (*Alptekin et al., 1990; Şimşek, 1997*). Some hot springs that are sourced from deep fracture zones exist along the faults on land and on continuation of these faults in the Aegean Sea. A number of earthquake events also are related to these features. Mainly NE to ENE and, further east, WNW striking basins (Sredna Gora, Strymon, NAT, Gediz, Bakırçay, Edremit, Büyük Menderes and Gökova; Kütahya and Eskişehir faults) characterise the eastern part of the Aegean region (Fig. 5.3). On the western coast of Turkey, the major basins (grabens) and associated mountain chains (horsts) are perpendicular to the Aegean coast, which leads to the deeply indented coastline of the western Turkish mainland. These grabens are the dominant neotectonic structural features (*Ketin, 1968; McKenzie, 1978; Dewey & Şengör, 1979; Jackson & McKenzie, 1984a; Şengör, 1982, 1987; Barka & Reilinger, 1997*). Many

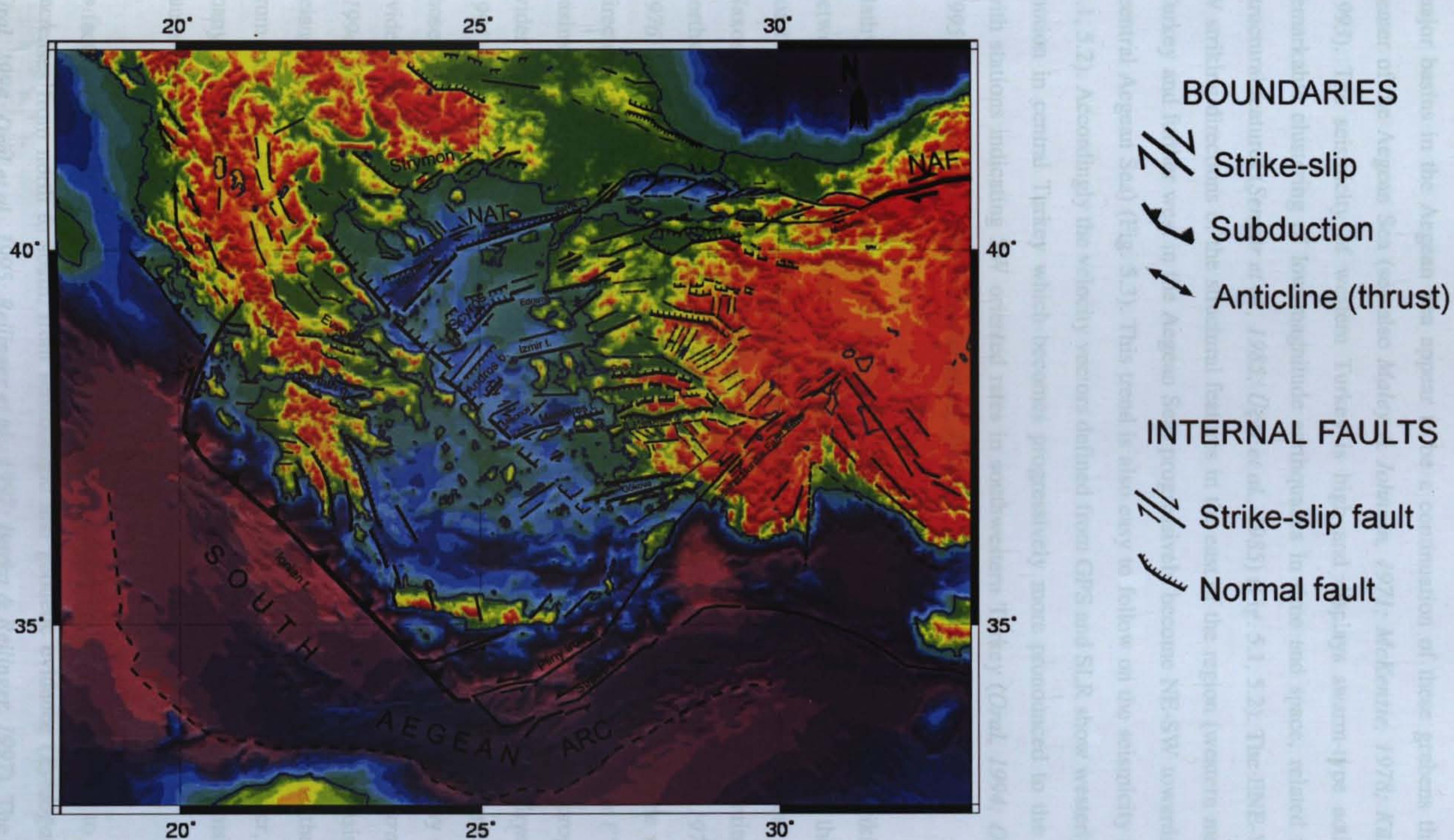

Figure. 5.3. Simplified tectonic map of the Aegean region.

major basins in the Aegean Sea appear to be a continuation of these grabens through the center of the Aegean Sea (see also *Maley & Johnson, 1971; McKenzie, 1978; Küleli et al., 1993*). The seismicity of western Turkey is high and displays swarm-type activity with remarkable clustering of low magnitude earthquakes in time and space, related to diffuse structural features (*Şengör et al., 1985; Üçer et al., 1985*) (Fig. 5.1, 5.2). The ENE-WSW / E-W striking directions of the structural features in the east of the region (western mainland of Turkey and further west in the Aegean Sea) progressively become NE-SW towards the west (central Aegean Sea) (Fig. 5.3). This trend is also easy to follow on the seismicity map (Fig. 5.1, 5.2). Accordingly the velocity vectors defined from GPS and SLR show westerly directed motion in central Turkey which becomes progressively more pronounced to the W-WSW with stations indicating SW oriented rates in southwestern Turkey (*Oral, 1994; Oral et al., 1995; Reilinger et al., 1997; Mueller et al., 1997*).

Bathymetry and shallow seismic data in the Aegean Sea suggest that NE-SW striking basins between the active faults bend abruptly to SW by NW-SE striking faults along the NW-SE coast of the Greek mainland. Islands (Sporades, Skyros, Evvoia, Anros, Tinos, Mikonos, Naxos, Amorgos, Astipalaia, Sirna) give the same impression. The faults continue in the northwestern Greek mainland (Strymon valley) (see also, *Maley & Johnson, 1971; Arpat, 1976; McKenzie, 1978; Taymaz et al., 1991; Jackson, 1993*). At the joints the southerly-directed component of motion (*Oral et al., 1995*) leads to formation of triangular (">" shape) basins (i.e. Scyros, Sporades) (Fig. 2.18). NAT and the other basins become progressively wider towards the west (Fig. 5.3). Fault mechanism solutions of earthquakes (*Taymaz et al., 1991*) reflect dominantly normal faulting with a right lateral strike-slip component related to those basins (*Taymaz et al., 1991*). But *Saatçılar et al. (1996)* did not observe any strike slip evidence in the profiles. According to them, the focal mechanism solutions of *Barakou et al. (1994)*, who used also stations from islands in the Aegean Sea, are more consistent with seismic profiles. But in contrast *Martin (1987)* identified strike-slip evidence in these basins from extensive seismic profiling data (see also *Mascle & Martin, 1990*). Further, there are many seismologists (e.g., *Taymaz et al., 1991*) who find strike slip mechanisms on these faults (see Section 5.5.1).

Is fact that the extension on the grabens of western Turkey decreases from west to east and increases from north to south, from seismologic and geodetic evidence (*Eyidoğan, 1988; Oral, 1994; Oral et al., 1995; Reilinger et al., 1997; Barka & Reilinger, 1997*). The geodetic velocities increase from 24 mm/yr in the southern Marmara Sea to 36 mm/yr to SW at the western flank of the South Aegean Arc (SAA) (*Barka & Reilinger, 1997*). The increase of extension from east to west in Turkey leads to the horizontal westward opening of the ">" shape gulfs along the west coast of Anatolia, such as Saros, Edremit, İzmir and Gökova and further west in the Aegean Sea (see also *Barka & Reilinger, 1997*). The northestern Aegean region is partly under the influence of the right lateral strike-slip motion especially in and around the Marmara Sea. On the other hand southwestern Anatolia is undergoing major extension, resulting in graben formation and probably listric faulting (*McKenzie, 1978*). This could be related to rollback of a slab. Paleomagnetic data are consistent with very small or no rotations in the northern part and possibly anticlockwise rotations, relative to Europe, in the south (see *Taymaz et al., 1991*). But the data are too few to made straightforward interpretations (see also *Taymaz et al., 1991*).

5.1.1. The Northeastern Aegean Region

The area is characterised by a high strike slip component of faults with the N-S extensional regime of the Aegean region. But the kinematics of faulting in the area is not clearly understood yet. The North Anatolian Fault (NAF), Marmara and surrounding basins, Strymon basin, NAT and Skyros basin are the most pronounced and are considered as the major tectonic elements in northwestern Anatolia (Fig. 5.3). Distinct epicentral zones follow these narrow lineations of morphology (Fig. 5.1, 5.2). Young tectonic activity in the region has affected the geomorphology and normal, strike-slip faults and dominantly oblique faults and basins can clearly be seen in the morphology. Strikes of the strike-slip faulting in these places are the most obvious trend in the seismicity map (*Taymaz et al., 1991; Üçer, personal communication*). A good correlation also exists between seismic activity and high heat flow anomalies, because these are deep reaching oblique faults (*Alptekin et al., 1990; Şimşek, 1997*). Gravity anomalies are also well correlated with this major tectonic lines (*Oral, 1987; Sarı & Şalk, 1995*).

To the west of Mudurnu valley there are mainly E-W trending grabens that characterise the Marmara region. NE trending short dextral faults separates these grabens in the Marmara Sea. East of the Marmara Sea, from the epicentre of the 1912.08 (40.75-27.20) Saros earthquake to the Sporades basin, a similar ENE-NE trending right-lateral strike-slip motion with large amount of deep-slip component is dominant, which strikes along the NAT. The Strymon basin to the north of NAT and Edremit trough to the south strike parallel to the NAT. The NAT, Sporades, Edremit and Strymon troughs all terminate to the SW in a similar way (*Lyberis, 1984*). The Edremit trough strikes from the western edge of the Biga peninsula to the west, and further west to the SW. It is bordered at the NAT by NE-SW trending faults. Concordantly, the focal mechanism solution of the 1967.03.04 earthquake shows NE-SW trending normal faulting here. But also aftershocks of events and the focal mechanism solution of the 1981.12.27 event show that dextral motion is taking place as a continuation of the motion of the Edremit trough. Both lineations follow the narrow grabens that can be seen clearly from bathymetry (Fig. 2.18). On the other hand, the area in the Aegean Sea between these zones appears to form aseismic blocks (*Allan & Morelli, 1971*) (Fig. 5.1, 5.2).

This well-known that the purely dextral NAF strikes ENE to the east of the Mudurnu valley. Right-lateral motion of the NAF is clearly expressed by an abundance of physiographic and geologic features (e.g., *Wallace, 1968; Allen, 1969, 1975; Seymen, 1975; Tatar, 1978; Barka, 1981; Barka & Hancock, 1984*), surface breaks of large earthquakes (e.g., *Ketin, 1948, 1969*) and focal mechanism solutions of large earthquakes (Fig. 5.4) (e.g., *Canıtez & Üçer, 1967; McKenzie, 1972, 1978; Alptekin, 1973; Jackson & McKenzie, 1984a*). The narrow NAFZ is the northern boundary of the westward moving Anatolian block and connects the compressional regime in eastern Anatolia with the extensional regime in the Aegean sea area (Fig. 5.3) (*Angagi, 1997*). The boundary becomes diffused around the Marmara region and further west. Extension affects areas to the north in Central and southern Bulgaria; Macedonia and eastern Albania further west. All the mountainous topography in this northern region may be the result of Late Miocene to recent extension (*Burchfiel et al., 1997*). The N-S compression-related westward escape of the Anatolian plate from the Karlıova triple junction is obvious, but where the driving force that is pushing to the east ends and where the driving force that is pulling to the west starts is unclear. In the east of the region the focal mechanism solution and surface break of the 1967.7 .22 earthquake on the NAF are in good agreement and require a slip vector within a few degrees of 90° E (*McKenzie, 1978*). This earthquake represents the NAF pure strike slip motion. The ENE trending pure strike-slip of the NAF ends just west of the Mudurnu valley (see *McKenzie, 1972*). The extensional system begins west of Mudurnu earthquake epicentre (*McKenzie, 1972, 1978; Jackson & McKenzie, 1984a;*

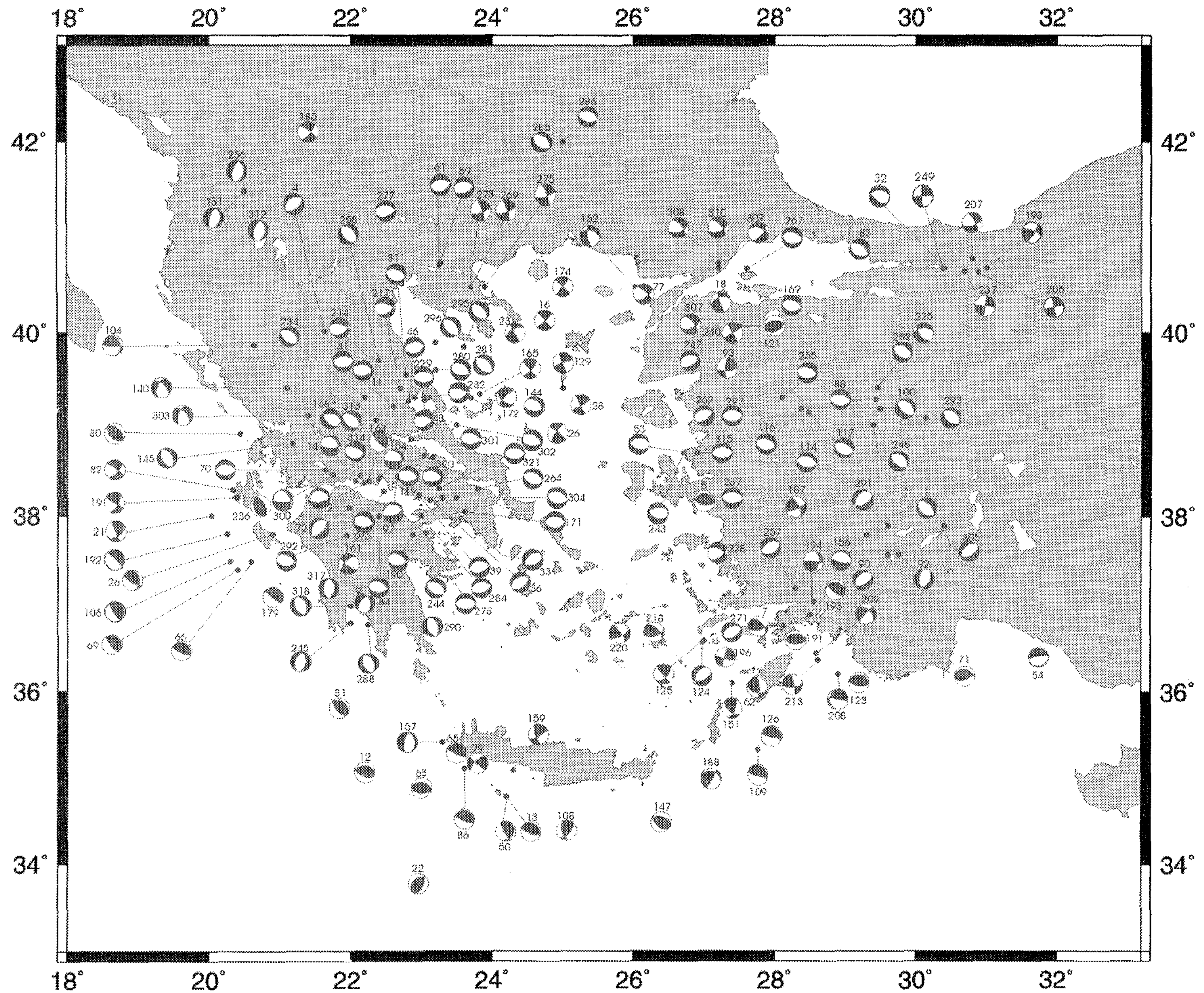

Figure. 5.4. Lower hemisphere projections of the focal mechanism solutions of the earthquakes studied by various seismologists (see Table. 2.5). Compressional quadrants are shaded. Numbers identify the selected solutions in Table. 2.5. The focal mechanisms selected from Table. 2.5. correspond to the solutions found by using the most reliable method and seem to be the most concordant with the tectonic scheme (see Section. 5 and Fig. 5.3).

Lio et al., 1991). Both strike slip and extensional deformation have been distinguished by field geology, geodetic estimations and fault mechanism solutions in the area (*McKenzie, 1978; Dewey & Şengör, 1979; Lyberis, 1984; Crampin & Evans, 1986; Barka & Kandinsky-Cade, 1988, Taymaz et al., 1991; Mueller et al., 1997*). This region is under the influence of both the dextral strike-slip fault regime of the NAF and the tensional regime of the Sea of Marmara and western Turkey. *Barka & Kadinsky-Cade (1988)* and *Barka & Gülen (1988)* introduced a pull-apart model for the Marmara basins (see also *Wong et al., 1995; Ergün & Özel, 1995; Akgün & Ergün, 1995; Koral & Öncel, 1995; Barka, 1997*) to explain the kinematics. In agreement with the model, in the Marmara Sea WNW-ESE trending basins bordered by NE-SW trending right lateral small strike slip segments are obvious from bathymetry.

It is suggested that the right lateral strike slip component of oblique deformation in northwestern Anatolia comes from western continuation of the dextral NAF. According to another idea, the NAFZ consists of mainly a single strand between the Karlıova triple junction and the Mudurnu valley. For example *Allen (1969)*, *Ambraseys (1970)*, *Kıyak (1986)* suggest that, to the west, the Marmara sea and the North Aegean regions are under a purely extensional regime as the northern extent of the extensional regime of the Aegean region. *Jolivet et al. (1994)* argue that the strike-slip deformation is less localised in the Aegean N-S extensional domain than onland in Turkey, and the NAF zone diffuses into a brother zone. This argument suggests that the NAF splays into several segments (3 or 2 branches) in the Marmara region, west of the epicentral location of the Mudurnu earthquake (see also *McKenzie, 1978; Lyberis, 1984; Barka & Kadinsky-Cade, 1988; Taymaz et al., 1991; Barka, 1997; Barka & Reilinger, 1997*).

According to the other idea (e.g., *Ketin, 1969; McKenzie, 1972, 1978, Alptekin, 1973; Dewey & Şengör, 1979; Le Pichon & Angelier, 1981; Lyberis, 1984; Şengör et al., 1985; Barka & Kadinsky-Cade, 1988; Mascle and Martin, 1990; Taymaz et al., 1991; Barka, 1992; Reilinger et al., 1995; Le Pichon et al., 1995; Taymaz et al., 1991; Jackson et al., 1992; Armijo et al., 1996; Mueller et al., 1997; Burchfiel et al., 1997; Barka, 1997; Barka & Reilinger, 1997; Küleli, 1997*), the NAF zone continues through the Marmara and North Aegean Seas (NAT). Furthermore, according to e.g., *McKenzie (1972), Reilinger et al. (1995)* and *Le Pichon et al. (1995)*, the trench system could be connected of the NAF by the Cephalona fault. However, *McKenzie (1972)* also points out the absence of any obvious structure crossing Greece from the Sporades basin to Kephalena fault (see also *McKenzie, 1978; McKenzie & Jackson, 1983; Taymaz et al., 1991*). On the other hand *Dewey & Şengör (1979)* suggested that the strike slip motion of continuation of the NAF is bordered to the west by the Greek shear zone. They suggest a model for the extensional mechanism of the Aegean Region according to the this idea. West of the NW-SE striking Aegean axial trough, the SW-directed motion is taken up by motion perpendicular to the active grabens. In this case, NAT and the other parallel striking faults appear as cross grabens.

The most dramatic bathymetric feature of the North Aegean Sea is the NAT (Fig. 2.18), which consists of a series of deep fault-bounded active basins (*Vogt & Higgs, 1969; Lort, 1971; Lyberis, 1984; Le Pichon et al., 1984; Mercier et al., 1991*). This feature is thought to be a continuation of the NAF by many scientists. Intense seismicity clearly defines the trough (Fig. 5.1, 5.2). It is identified also by magnetic and bathymetric surveys (*Vogt & Higgs, 1969; Lort, 1971*). It had been thought that sea floor was being produced along the trough (*McKenzie, 1970, 1972; Lort, 1971; Allan & Morelli, 1971*), but magnetic surveys have failed to show any linear magnetic anomalies associated with the shallow seismicity (*Vogt & Higgs, 1969; McKenzie, 1972*). The results of geophysical investigations show that the crust underlying the NAT is continental, overlain by up to 5.0 to 5.5 km of sediments (see *Ginzburg et al., 1987*). It seems that the intense SW-NE trending magnetic anomalies, which are parallel to the bathymetric trends, are caused by magnetised rocks that have been intruded by volcanism (*Vogt & Higgs, 1969; Papazachos & Comninakis, 1971*). A heat flow station located over one of the anomalies gave a value of 2.5 H.F.U., showing that recent intrusion of magnetic material probably also contributes to the high heat flow observed in the area (see *Jongsma, 1975*).

The trough is bounded by a major fault to the south and by a faulted flexure to the north (*Ginzburg et al., 1987; Le Pichon et al., 1984; Mascle & Martin, 1990*). Major seismic activity is also located about 50 km south of the deep linear trough, which could related tothe northward deeping major fault (*McKenzie, 1972*). The bathymetry and shallow structure of this trough, in which water depths reach some 1000 to 1400 m (*Maley & Johnson, 1971*), have been studied with seismic reflection surveys and acoustic techniques (*Ginzburg et al., 1987;* see also *Jongsma, 1975*). Some information on the deep sediments (Oligocene to present) has been published by *West (1973)* and *Lalechos & Savoyat (1977)*. The whole sediment column up to and including the seafloor is affected by faulting which points to very recent activity (see *Jongsma, 1975; Roslyakov & Sorokin, 1997*). The NAT could be divided into two basins: the Saros basin, which trends ENE-WSW in the east and the Sporades basin, which trends NE-SW in the west. These changes could be evidence that the western part of the Anatolian Trough contains a larger component of opening (*McKenzie, 1978*). The focal mechanism solutions show that the N-S component of the motion increases westward (*McKenzie, 1978*). North of the trough, basins (Thermaikos, Orfanou, Prinou) strike normal (NNW-SSE) to the trough; they are caused by NE-SW extension (*Mascle & Martin, 1990*).

Focal mechanism solutions in the NAT show mainly right lateral strike-slip faulting (*Taymaz et al., 1991*) (Fig. 5.4). *Roussos & Lyssimachou (1991)* found typical recent strike slip fault patterns in the central NAT, based on detailed interpretation of marine reflection seismic lines (see also *Saatçılar et al., 1996*). They observed transtension (divergent wrenching) with high angle normal faults (see also *Lalechos & Savoyat, 1977*). According to them, the change in fault strike is the probable cause of the development of transtension in that area (see also *Lyberis, 1984*). Focal mechanism solutions of earthquakes larger than M=5.5 show normal faulting and strike slip faulting along the trough. These solutions could easily be associated with bathymetric features of the NAT. Two substantial shocks, 1967.3.4 and 1968.2.19, have occurred south of the NAT and are not associated with any major bathymetric feature (*McKenzie, 1978*). The aftershocks of 1968.2.19 have been relocated by *North (1977)*, who found that they lay on a NE-SW trend to related NAT (*McKenzie, 1978*). This trend follows the expected fault plane and the motion must be right lateral on the fault. All the focal mechanisms for 1975.3.27 (Table. 2.6) give a normal fault with some strike slip component. This mechanism for the 1975.3.27 event together with the description of the surface break of 1912.8 .9 by Allen (1975), are consistent with a strong right lateral strike slip motion on the eastern part of the NAT. The nodal planes of these earthquakes strike parallel to the trough that bounds active faults. *Lyberis (1984)* suggested that displacement on the NAT is greater than along the NAFZ. Strike slip and normal faults are also observed further north over North Aegean coastal areas (*Roussos & Lyssimachou, 1991*). For example, focal mechanism solutions of the 1978.05.23 and 1978.06.20 earthquakes show that normal faulting is also taking place further north. In central and southern Bulgaria, Macedonia and eastern Albania, extension is effective. In the northern Aegean region, *Sokoutis et al. (1993)* show that in the gneiss domes of Kavala and Thassos the direction of extensional ductile shear is NE-SW, which is parallel to the North Aegean Through (see also *Jolivet et al., 1994*). *Jongsma (1975)* observed only local thrusting deformation in sediments. *Saatçılar et al. (1996)* also observed NE-SW directed thrust deformations from seismic sections at deeper levels of sediments. But no evidence of compression has been found by seismotectonic observations (see also *Lyberis, 1984*), or seismic profiles (*Lalechos & Savoyat, 1977*) and the evidence from seismic profiles is not very clear.

5.1.2. The Southeastern Aegean Region

Dominant strike of the major grabens in western Turkish mainland (Edremit, Alaşehir (Gediz), İzmir, Büyük Menderes and Gökova (Bodrum) grabens) is E-W to ENE-WSW. These grabens continue west of the the central (deeper) portion of the Aegean basin (*Mascle and Martin, 1990*). According to the *Barka & Akyüz (1997)*, the E-W and NW-SE faults (e.g., Gökova, Büyük Menderes, Pamukkale and Dinar) are normal faults (e.g., *Şengör et al., 1985*) in western Anatolia. On the other hand NE-SW trending faults (e.g., Fethiye-Burdur fault zone) have a left lateral strike slip component (*Şaroğlu et al., 1987; Barka et al., 1995*). These grabens are closely related to the Aegean neotectonic extensional regime and they are evidence of stress distribution in the area.

The Edremit trough trends ENE-WSW direction from east of Edremit bay to the Skyros basin. The graben is bounded by the same main Northward dipping fault to the south as is the NAT (*Mascle & Martin, 1990; Taymaz et al., 1991*). The İzmir Trough trends ENE-WSW, but it is not clear from south of Midilli to the SW. It is bounded by the same SE-NW trending and northward dipping normal faults as the others to the north (*Mascle & Martin, 1990*). The Menderes Trough is also is a NE-SW trending basin which is parallel to the İzmir, Edremit and North Aegean troughs. Continuation of the Menderes graben could be clearly seen from bathymetry further WSW. NE dipping normal faults (*Mascle & Martin, 1990*) bound the graben to the WSW. There was no evidence of a strike slip component in these grabens from seismic data (*Mascle & Martin, 1990; Saatçılar et al., 1996*). The Gökova graben goes from Gökova bay in a SW direction parallel to the other SW trending grabens in the eastern Aegean sea. This graben also ends in front of the NW-SE, trending long NE dipping normal faulting feature (*Mascle & Martin, 1990*). Southwestern Anatolia is undergoing major extension, resulting in graben formation and probably listric faulting (*McKenzie1978*). The available geodetic 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 peninsula (*Barka & Reilinger, 1997*). Field observations and focal mechanism solutions in western Turkey are consistent with this 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 (*Barka & Reilinger, 1997*). The dips of at least some of the normal faults in western Turkey become shallower with depth, and are called listric faults (see Section 2.3, for detail). This kinematic result could be evidence of the dynamic force that leads to extension. Three shocks are related to listric faulting in western Turkey (*McKenzie, 1978*). The best evidence comes from the 1970.3.28 earthquake whose fault break was mapped by *Ambraseys & Tchalenko (1972)* (see also *McKenzie, 1978*). The surface break for the 1969.3.28 earthquake was mapped by *Arpat & Bingöl (1969)*. Low angle normal faulting and listric geometry were observed on faults in a graben (*Arpat & Bingöl, 1969; Yusufoğlu, 1996; Emre, 1992*). The epicentres of all moderate shocks determined by *McKenzie (1978)*, which have surface breaks, are displaced to the north of their surface breaks (*McKenzie, 1978; Şengör, 1987*). These basins are filled with up to 2 km of Quaternary and Neogene sediments (*Sarı et al., 1995*). The central Menderes Massif is bounded by the northward dipping southern main boundary fault of the Alaşehir (Gediz) graben in the north and the south dipping northern main boundary fault of the Büyük Menderes graben. The gravity profiles of *Sarı et al. (1995)* indicate that the Gediz and Büyük Menderes graben systems have asymmetric features (*Roberts, 1988; Paton, 1992; Sarı et al., 1995)*, which supports the idea of listric fault kinematics. In western Turkey high heat flow was measured along the graben by *Alptekin et al. (1990)*. The anomalies are related to deep reaching detachment faults. There are also many hot springs and young volcanoes in the area. For example, the volcanics of Kula are basaltic and come through huge detachment faults to the surface. In western Turkey there is a metamorphic massif outcrop that is called the Menderes Metamorphic Massif. The Simav, Alaşehir, Küçük Menderes and Büyük Menderes grabens divide the Menderes Massif into four submassifes. This massif is a key to understanding the tectonics of the Aegean region. A complex core mechanism that related to subduction is suggested for the outcropping (upwelling) of the Menderes core complex. The low angle huge detachment faults could have caused updomming of the Menderes massif and its continuation towards the west. The ages of grabens are closely related to the starting time of extension in the area. The start of the formation of the Gediz Graben has been variably assigned to the Miocene (*Yağmurlu, 1987*), Middle Miocene (*Ercan et al., 1978; Seyitoğlu &*

Scott, 1991; Emre, 1992), Late Miocene (*Karamanderesi & Yılmazer, 1982; Arslan, 1984*) and Pliocene (*Ünal & Havur, 1970; Karamanderesi, 1971; Arpat & Bingöl, 1969*).

Şengör (1987) has reported active faults with SW-NE trend that cut across these main grabens, implying that it comprises many elongated blocks with dimensions of 50 to 100 km in the SW-NE and 20 km in the NW-SE directions, which can clearly be identified morphology and on gravity maps (*Oral, 1987; Sarı et al., 1995*). These are older (Early Miocene) than the dominant E-W grabens (*Ketin, 1968; Kaya, 1979, 1982*). Şengör (1987) explained these features as cross grabens (*Şengör et al., 1985; Oral, 1987*). According to the theory, the small grabens that occur in the hanging wall of the main graben join a huge perpendicular normal fault (breakaway fault) of main graben. It is suggested by *Oral (1987)* that the faulting continues deep of and cuts the entire lithosphere. According to this theory, the small grabens and the main graben have evolved simultaneously (see also *Sarı et al., 1995*). If this suggestion is true the starting time of the extension in the region must be before that of the E-W grabens. The other suggestion is that the N-S compressional regime is of Tibet type (*Yılmaz, 1997*) and leads to the north trending E-W extended grabens. Younger grabens on the other hand are thought to belong to the N-S extensional deformation (*Dewey & Şengör, 1979; Şengör, 1982; Şengör et al., 1985*). If this suggestion is accepted the starting time of the extension must be more recent than that of the N-S trending grabens.

5.2. THE WESTERN AEGEAN REGION

The Hellenides-Peloponnisian peninsula is characterised by a NNW-SSE trending mountain belt that constructed of nappe pilling. The belt is indicated by the high negative Bouguer anomaly that strikes NNW-SSE in the western Aegean Region (*Makris, 1977, 1985*). These mountains are cut by NNW to NW striking normal faults to their NW of mountains. The NW-SE trending normal faults to the NE continue step by step to the centre of the Aegean Sea and the last fault zone occurs as eastward dipping asymmetric grabens (e.g., Strymon, Sporades, Skyros, Anros, Mikonos, Sirna) and bounds NE-SW trending basins (e.g., Kavala-Xanthi, NAT, Edremit, İzmir, Menderes and Gökova) to the east (see also *McKenzie, 1978; Roberts & Jackson, 1991; Taymaz et al., 1991*) (see Section 5.1.1). The lineation of faults over the Greek mainland can be seen from bathymetry - topography and geological structure as well as focal mechanism solutions of moderate earthquakes (*Roberts & Jackson, 1991; Taymaz et al.,* *1991*). These grabens were also been recognised from diffuse seismicity distribution (*McKenzie, 1978; Hatzfeld et al., 1990; Rigo et al., 1995; Hatzfeld al., 1996; Mueller et al., 1997*) and a structural geological study (*Mercier et al., 1976*). West of the Aegean region, two huge NW-WNW striking grabens (Corinth and Evvoia) are obvious from bathymetry and topography. Towards the Aegean Sea (east) these two basins become more southerly. These grabens indicate recent extensional structures that accommodate a large part of the internal deformation of Hellenides-Peloponnese (see for detail *Armijo et al., 1996; Hatzfeld et al., 1996*). The Gulf of Corinth, located between Greece and Peloponnese, is a recent N-S extensional structure (*McKenzie, 1978*) not older than 2 My that opens at about 1 cm/yr and accommodates a large part of the extensional deformation (*Armijo et al., 1996; Hatzfeld al., 1996*). In the footwall, 1.3 mm/yr maximum uplifting is found from detailed field geology in this graben by *Armijo et al. (1996).* They suggest elastic banding theory to explain such uplifting. Microseismic surveys support a northward dipping listric shape for faults related to the Corinth graben (*Hatzfeldet al., 1996*). It is suggested that this detachment is due to the uncoupling of the brittle upper crust from a more ductile lower crust (*Melis et al., 1989; Rigo et al., 1995*) (see Section. 2.3). Its western segment is connected to the Ionian trench system by a series of NE-trending transform faults with dextral polarity. This zone is called the Kephalonia (Cappelona) Fault Zone. The rift's eastern segment transverses the volcanic arc of the Aegean orogene and can be considered a first order cross fault (*Hatzfeld al., 1996*). *Taymaz et al. (1991)* suggest that the overall NE-SW motion across the region is due to oblique (left-lateral and normal) slip on these normal faults (*Berckhemer & Kowalczyk, 1978; Myrianthis, 1984*). North of the Kephalonia fault zone, the NW end of the subduction zone is characterised by a negligible motion (see *Mueller et al., 1997*). In northwestern Greece and Albania there is a band of thrusting near the western coasts. The most obvious problem in accounting for the observed deformation is the relationship between the thrusting and normal faulting further east in the northwestern part of the area (see also *McKenzie, 1978*).

East of the Aegean axial trough the strike-slip faults do not cross central Greece to link with the South Aegean Trench (*Taymaz et al., 1991*) with the Kephalonia fault zone. Instead they appear to end in deep basins bounded by normal faults with a NW or WNW strike (see section 5.1). Geodetic displacements measured in Greece indicate that slip vectors parallel to the main basin trend over east and central Aegean. But in the western Aegean region the vectors are perpendicular to the strike of large scale crustal blocks bounded by normal faults (*Billiris et al., 1991; Oral et al., 1995; Jolivet et al., 1994*). On the other hand the vectors further east in the Aegean Sea areapproximately parallel to the bathymetric features which could be thought as cross grabens (e.g., *Şengör, 1987*). *Taymaz et al. (1991)*, by considering paleomagnetic and seismotectonic results; suggest that central Greece is rotating clockwise about a vertical axis as a whole. A homogeneous SW oriented motion of SW Greece reach as rates of 40 mm/yr relative to the Eurasian plate according to GPS and SLR data (e.g., *Mueller et al., 1997*). The Greek mass seems to be pulling the eastern Aegean region towards the south Aegean consumption boundary.

The deformation within the Greek mass occurs as simple shear. Faulting on land was observed dipping north after the 1894 earthquakes along the north Gulf of Evvoia (*Ambraseys & Jackson, 1990*). Fault types of the same type and direction were found for the 1967.03.04 earthquake (*Taymaz et al., 1991*) in the Skyros basin. Skyros island that represents a footwall of the NW striking graben and appears to be tilted to the SW (*Roberts & Jackson, 1991; Taymaz et al., 1991*). Most major normal faults dip northeastward and are responsible for the southwestward tilt of the crustal blocks along the northeastern coast of Evvoia and Attica (see also *Jolivet et al., 1994*). Furthermore, *Papanikolau, et al. (1988)* show that the sediments deposited in the basins are tilted southward. The analysis of structures in the metamorphic domes on footwalls of Evvoia and Attic faults and the granodiorites also show a very consistent fabric with flat foliation and N-S trend of stretching lineation (*Faure & Bonneau, 1988; Gautier et al., 1990; Buick, 1991; Jansen, 1973; Robert, 1982; Jolivet et al., 1994*). Extension is taken up by low angle normal faults, which bound upper crustal blocks such as Evvoia and Attica (*Jolivet et al., 1994*). The Cyclades area is also like a shallow plateau that appear as a footwall of the northward dipping faults to the NE. It is obvious that the NAT and the other NE-SW trending basins end in front of these blocks as cross grabens.

5.3. THE SOUTH AEGEAN SUBDUCTION SYSTEM

Based on earthquake investigations as well as geophysical investigations of gravity, magnetic and seismic wave velocities, *Galanopoulos (1963)*, *Papazachos & Comninakis (1969, 1971)*, *Caputo et al. (1970)*, *Papazachos (1973)*, *Agarwal et al. (1976)*, *Gregersen (1977)*, *Comninakis & Papazachos (1980)*, *Gregersen & Jaeger (1984)*, *Makropoulos & Burton (1984)* and *Spakman et al. (1988)* have shown the existence of dipping lithospheric material towards the concave side of the South Aegean Arc (SAA). Negative low Bouguer gravity anomalies show that there is dense material (a sinkable slab) beneath Crete, Rhodes and the Pelopennese (*Ryan et al., 1969; Makris, 1985; Tsokas & Hansen, 1997*).

Recent volcanism on some islands is that parallel to the South Aegean trough is related to a subducting plate (slab) in the Aegean region. Most of the heat flow values observed over the Cyclades are like those for other island arc systems in the world (*Jongsma, 1974*). Furthermore, heat flow values increase towards the inside of the arc from the trench, as they do in the Kuril and Japan island arcs (*Jongsma, 1974*) (Fig. 2.57). The Benioff zone ends beneath the volcanic arc. At greather deprth the slab could not accumulate stress because it behaves as plastic material above a critical heat and pressure. Limited isotopic data suggest that little of the sediment is incorporated into the volcanic rocks of the Aegean arc (*Barton et al., 1983; Briqueu et al., 1986*), presumably because it is not transported deep enough into the mantle for melting to occur (*Taymaz et al., 1990*). A low-velocity layer is observed in the upper mantle in the middle of the Aegean Sea from earthquake data (*Spakman et al., 1993; Papazachos, 1994; Papazachos et al., 1995*). This remarkable low velocity anomaly may be caused by the dehydration of the subducting lithosphere around a depth of 80 to 90 km (*Papazachos & Nolet, 1997; Tsokas & Hansen, 1997*).

Along the western coast of Albania and Greece, and seaward of a trench from SW Greece through Crete to SW Turkey, the SAA is one of the most remarkable tectonic features of the Mediterranean area as well as of the Aegean region (Fig. 5.3). This is the consuming boundary of the African plate is the region. A narrow zone of negative free air gravity anomalies (Fig. 2.37) (*Woodside & Bowin, 1970, Allan & Morelli, 1971*) and low seismic wave velocities (*Alessandrini et al., 1997; Papazachos & Nolet, 1997*) coincides with this trench. A thick sedimentary prism along the trench causes these anomalies. The detailed morphology in this trough extremely complicated (*Jongsma, 1977*). The South Aegean Trench consists of a number of prominent deep basins that from an arc shape from Kefallinia to Rhodes (Fig. 2.18) (e.g., *Ryan et al., 1969*). To the east of the trough there are narrow linear basins called the Pliny and Strabo trenches. On the other hand, along the Ionian trench, structures west of Crete are not linear line and they extend towards the Ionian Sea (*McKenzie, 1978*). These depressions are collectively called the South Aegean Trough (SAT). To the East based on focal mechanism solutions of earthquakes and shallow seismic profiles *McKenzie (1972)* suggest that the Pliny Trench is the main zone of subduction, rather than the Strabo Trench as supposed by *Ryan et al. (1973)* (see *Jongsma, 1977*). On the other hand the Strabo trench continues further to the NE than the Pliny Trench and probably links up with another trench system to the east. Because the convergent rate is high enough, the deforming structures can be clearly seen in on bathymetry without sedimentary carpet (*Jongsma, 1977*). However these deep basins of the South Aegean Trench are considerably shallower and more highly curved (R=400km) than the other subduction related main trenches elsewhere in the world (*Le Pichon & Angelier, 1979; Jongsma, 1977*). Seismic profiles to the West show that the South Aegean Trench contains as much as 800 m of acoustically stratified recent fill (*Ryan et al., 1969, 1973*), and it may be locally thicker than 1.5 - 2 km (*Biju-Duval et al., 1974; Jongsma, 1977*).

Using Sea-Beam data, geodetic data, oceanic studies and field studies of islands and surrounding mainland, *Angelier et al. (1982)* used detailed investigations of faults to describe the kinematic situation around the area. They recognise three different segments along the South Aegean Trench according to the main fault alignments. In addition, based on microearthquake data focal mechanism solutions of earthquakes, normal faulting is found that indicates NW-SE extension in southern Peloponnese and western Crete and E-W extension in central and eastern Crete (*Hatzfeld et al., 1993*). The Pelaponnises have NNW elongated faults, Crete has E-W, ENE-WSW and NE-SW strike faults and Rhodes has NNE-SSW elongated faults. These faults are mainly normal and strike parallel to the elongation of the South Aegean Trench system. *Armijo et al. (1992)* suggested that the present day E-W extension of the South Aegean region is related to the arc curvature, and the arc parallel extension is possibly related to slabpull force (see section 5.4).

5.3.1. The Eastern Tip of the Arc

Harrison (1955) first drew attention to the similarity of the properties of the South Aegean Island Arc to those of major arcs in the Pacific. This arc has such general characteristics of the arcs as a trench in its convex side, a seismicity belt and negative free air anomaly along the trench, a sedimentary arc in the trench as well as a volcanic arc and a marginal sea (Aegean Sea) in its concave part. In addition, there is convergence between the Aegean lithosphere (in the concave part) and underplating Mediterranean lithosphere (in the convex part). The Aegean region is one of intense seismicity, most of which occurs in a wide belt that roughly follows the bathymetric part of the inner wall of the South Aegean Trench system (e.g., *McKenzie, 1978; Hatzfeld et al., 1993*). Nearly all the earthquakes, including micro-earthquakes deeper than 80 km, however, occurred beneath the Sea of Crete (*Hatzfeld et al., 1993*). No deep seismicity is observed outside the arc, south of Crete. Recent records of microseismic activity study (*Hatzfeld et al. 1997*) show that most of the seismic activity is located along the South Aegean Trench and is related to the active convergent boundary, and most of the larger shocks lie between the deepest part of the trench and the islands.

The trend of the compressional axis of focal mechanisms of earthquakes is perpendicular to the SAA in its western and southern part, keeping the same pattern to become parallel to the arc in the Pliny and Strabo trenches (with left-lateral strike-slip motion) at the eastern end of the arc (fig. 5.4) (e.g., *McKenzie, 1978; Mueller et al., 1997*). These results are compatible with GPS measurements (fig. 2.34) (*Oral et al., 1995*). The compressional axes are nearly horizontal and trending N-S to NNE-SSW, parallel to the suggested SW motion of the Aegean region (e.g., *McKenzie, 1972; Hatzfeld et al., 1993; Jackson et al., 1992; Taymaz et al., 1990*). In Crete by, shoreline change measurements imply by a 2-3 mm/yr uplifting which is related to the compressional regime along the subduction boundary (*Angelier, 1979; Angelier, 1981; Angelier et al., 1982*). Also on Crete there is some field geological evidence of normal faulting (e.g., *Angelier et al., 1982*). If the slab-pull mechanism exists, why is compressional evidence found? Just NE of the thrust zone in the internal part of the arc, normal faulting with similar strikes and deep as the reverse faults on the consuming boundary parallel to the trench, is very well known from the Evvoia to Crete (*McKenzie, 1978; Hatzfeld et al., 1990*). The transition between reverse faulting and normal faulting is rather sharp, but with some overlap of the two families, especially in the NW (*Hatzfeld et al., 1990*). In the area of normal faulting, strike-slip mechanisms are also observed by *Hatzfeld et al. (1990)*. Sea-beam and submersible diving observations and seismic profiles demonstrate that extensional faulting is seen at the sea surface as far south as the inner wall of the South Aegean Trench and north of the inner wall (*Jongsma, 1977; Huchon et al., 1982; Taymaz et al., 1990*). *Jongsma (1975)* also observed normal faulting near the outer wall of the trench in profiles that he interpreted. There is no evidence of normal faulting earthquakes south of the south coast of Crete (*Taymaz et al., 1990*). From morphological studies made by submersible diving, it is observed that the compressional structures affect the sedimentary cover of the outer wall and trench whereas extensional tectonics is present on the inner wall (*Huchon et al., 1982*). The focal mechanism solutions of large earthquakes by *Taymaz et al. (1991)* indicate shortening in the outer wall of the South Aegean Trench south of Crete. These mechanisms and the spatial distribution of the slab show that along this belt the Eurasian lithospheric plate is underthrusted by the African lithospheric plate, In addition there is evidence that most of the upper sedimentary cover is not subducted but rather piles up in front of the trench to form the Mediterranean ridge. And just back of the arc an extensional process is taking place (*McKenzie, 1978*).

The location of the consuming boundary not perfectly known. *Papazochos et al. (1995)* indicate that the lithosphere dives approximately along the line of the south coast of Crete. Recent tomographic results (*Papazachos & Nolet, 1997*) lead to the same conclusion. *Tsokas & Hansen (1997)* also modeled the subduction boundary as this line. *Hatzfeld et al. (1993)*, using microearthquake data, suggested that the boundary between the subducting African plate and the overriding arc cannot be located farther than 50 km south of the South Aegean Trench. However, the model of *Gregersen & Jaeger (1984)* indicates that the subduction process starts about100 km farther to the south. *Küleli (personal communication)* also suggested that the subduction prosesses starts further south. Recent seismic profiles beneath the Ionian islands, interpreted by *Hirn et al. (1996)*, suggest a more seaward position (using the Ionian islands) for the Ionian thrust than was previously assumed. By using gravity modeling and deep seismic data *Truffert et al. (1993)* also locate the consuming boundary more seeward than previously assumed. Their results show that at the trench itself and further outward there is overriding plate material under consolidated sediments which possibly thrust the sheets emplaced during a compressional regime.

The mean slip vector direction of the relative motion between the southern Aegean and Africa found by *McKenzie (1978)* is approximately parallel to the narrow slots in the sea-floor known as the Pliny and Strabo trenches east of Crete. Hence, the eastern part of the South Aegean Trench must be dominantly a strike-slip fault (*McKenzie, 1978; Le Pichon et al., 1979; Huchon et al., 1982*). *Ritsema (1969)* has shown that the motions along the faults are dextral in the eastern part of the region and sinistral in the western part (see *Papazachos, 1973; Le Pichon et al., 1995*). But the available data do not support this model. Some focal mechanism solutions, and the evidence of the sediment disturbance below the inner wall of both Pliny and Strabo trenches, support a thrusting kinematic, but without strike slip component (*Ryan et al., 1969; Jongsma, 1977*). Only the 1957.4.24 and 1957.4.25 earthquakes showed strike slip motion, but the solutions were not well constrained (*McKenzie, 1978*).

The Strabo trench continues further to the NE than the Pliny Trench and probably links up with another trench system and the Isparta Angle, which constitutes the junction between the Cyprus and South Aegean arcs (*Brunn et al., 1971; Barka et al., 1995; Eyidoğan & Barka, 1996; Yağmuroğlu et al., 1997; Barka & Reilinger et al., 1997; Pınar, in press*). *Oral (1987)* observed a high gravity anomaly over the Fethiye-Burdur fault zone (Köyceğiz Corridor) that could probably be explained by continuation of the subduction zone. This anomaly could be evidence of a strand of the slab There is also a high isostatic anomaly observed over the region (*Oral, 1987*). Isparta Angle is a tectonic assemblage, which has a complex tectonic history (*Barka & Reilinger, 1997*). This zone has been studied intensely for the last 25 years (e.g., *Brunn et al., 1971; Graciansky, 1972; Monod, 1977; Ricou et al., 1979; Şengör & Yılmaz, 1981; Yılmaz, 1983, 1984; Poisson et al., 1984; Şenel, 1983; Hayward, 1984; Robertson & Woodcock, 1984; Waldron, 1984; Marcoux, 1987; Akay & Uysal, 1988; Yılmaz, 1989; Robertson, 1990; Kissel et al., 1993; Frizon de Lamotte el al., 1995; Ersoy, 1995; Barka et al., 1995; Barka & Reilinger, 1997*). According to paleomagnetic data (*Kissel et al., 1993*), the angular shape of the Isparta Angle is related to 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; Barka et al., 1995; Barka & Reilinger, 1997*). 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 SW at about 30 mm/yr (*Reilinger et al., 1997*). The eastern flank of the Isparta angle is bounded by the NW-SE trending Sultandağ thrust fault (*Şaroğlu et al., 1983, 1987; Boray et al., 1985*) and the western flank is made up of the transtensional left-lateral NE-SW trending Fethiye-Burdur fault zone (*Dumont et al., 1979a, b; Şaroğlu et al., 1987; Taymaz & Price, 1992; Eyidoğan & Barka, 1996*). This is compatible with GPS results (*Reilinger et al., 1997*). Based on the anomalous tectonic nature of the Isparta Angle, *Barka & Reilinger (1997)* suggest that the eastern flank of the Isparta Angle, which is the northwestern continuation of the Cyprus arc into SW Anatolia, may slow down the westward motion of southern Central Anatolia. This difference is taken up by shortening along the Sultandağ mountains and Aksu thrust (*Barka & Reilinger, 1997*). Retreating of the South Aegean subduction to the south might have followed the incidence of a lateral tear that separated the South Aegean and Cyprian arcs according to *Barka et al. (1995)*. They assume that this lateral tear occurred along the Pliny-Strabo-Fethiye-Burdur shear zone.

5.3.2. The Western Tip of the Arc

There is much evidence to show that the SAA does not extend north of 39° N latitude along the coasts of Albania and Yugoslavia (*McKenzie, 1978; Taymaz et al., 1991; Jackson et al., 1992*). Only one mechanism for a shock beneath Greece (*McKenzie, 1978*) that of 1966.10.29, shows thrusting. To the north of the trench, the seafloor shoals and the seismically active zone run into the northwestern Greece (*McKenzie, 1978; Jackson et al., 1992*). The isobaths in the Ionian Sea (*Morelli et al., 1975*) have a NW-SE trend south of Kefallinia and Zakynthos, but further up, west and SW of Kefallinia island, the isobaths exhibit a NE-SW trend (*Kiratzi et al., 1991*). The latitude also marks the northern limit of intermediate shocks and volcanism (*McKenzie, 1972*). The other recent evidence is that the Bouguer gravity anomaly that shows a smooth pattern north of the island of Cephalonia in the Ionian Sea (*Tsokas & Hansen, 1997*). In map view, the convex arc of the South Aegean Trench resembles a gigantic active transform fold formed at the SW end of a belt of distributed right-lateral shear trending NE-SW (*Brunn, 1976; Anderson & Jackson, 1987; Taymaz et al., 1991*). Both the small-scale bathymetry (*Allan & Morelli, 1971; Stride et al., 1977*) and land geological studies (*British Petroleum, 1971; Brooks et al., 1988; Underhill, 1989; Sorel, 1992*) suggest that the deformation is taken up by a number of right-handed strike-slip faults (Kefalinia Fault Zone) (*McKenzie, 1978; Tsokas & Hansen, 1997*). The zone can be defined by the location of large dextral strike-slip earthquakes (e.g., 1983.01.17) (see *Anderson & Jackson, 1987; Kiratzi et al., 1991*). Micro-earthquakes on Kefalinia show the full range of focal mechanisms, thrust, strike-slip, and normal faults (*Amorese, 1993*) because of the local stress regimes in the wide shear zone (see also *Hirn et al., 1996*). Its importance is emphasised by structural trends on land and offshore (*Finetti, 1982*), by the alignment of earthquakes (e.g., *Makropoulos & Burton, 1984; Melis et al., 1989; Amorese, 1993*) and by displacement rates determined from GPS observations (*Kahle et al., 1995*). Based on geophysical data the Kefalinia Fault Zone (KFZ), Calabrian and SAA possibly form a triple junction in the Ionian Sea (*Kahle et al., 1995*). However, only tomographic results (*Spakman et al., 1988, 1993; Papazachos et al., 1995*) show that the Benioff zone must extend so far to the north. These images may belong to a subducted lithosphere that has finished its evolution.

5.3.3. Adriatic

The Adriatic region (the Adriatic-Apulia platform) represents, in broad terms, the remnant of an African (*Channel & Horvath, 1976*) part of lithosphere currently confined by Tertiary-Quaternary collisional belts between the Apulian block and Albania, Yugoslavia and Greece (*McKenzie, 1978; Jackson et al., 1992; De Alteriis, 1995*). Recent seismic studies of *De Alteriis (1995)* on sediments show that the Adriatic basin is under pure shear to the south and a compressional regime to the north. This result is consistent with earthquake focal mechanism data.

5.3.4. Kinematics of the Slab

In basic view the slab shape gives a previous impression of the kinematic situation within the subduction zone. To the west of the arc, earthquakes deeper than 40 km define a very gentle slope (about 10°) starting at the trench and continuing for 200 km further inside to the east, where the zone abruptly steps to about 45° beneath the gulf of Argolis (*Hatzfeld et al., 1989*). This change occurs beneath an aseismic shallow mass within the crust (*Martin, 1988; Hatzfeld et al., 1989; Papazachos & Nolet, 1997*) Powerful marine vertical-incidence reflection profiling images (*Hirn et al., 1996*) show a major reflector at a depth of about 13 km slightly dipping to the east under the western slope of the Ionian islands then dipping steeply under them.

Papazachos & Comninakis (1971) assumed a symmetrical aphytheatrical shape for the subducting slab. However, based on teleseismic data that was studied by *Hatzfeld & Martin (1992)* and tomographic studies (*Papazachos & Nolet, 1997*) the gently dipping part of the slab is longer in the west than in the east. Acording to recent tomographic results (*Papazachos & Nolet, 1997*) the subducted slab shows a complicated geometry, with a slope that gradually increases from west to east and toward the deeper sections of the slab. In addition the isodepth contours of intermediate earthquakes are not parallel to the South Aegean Trench (Fig. 5.5) (e.g., *Hatzfeld & Martin, 1992; Hatzfeld et al., 1993*). This distribution confirms that the slab is being subducted shallower at the western end than at the eastern end (*Hatzfeld et al., 1992, Hatzfeld et al., 1993; Papazachos & Nolet, 1997*).

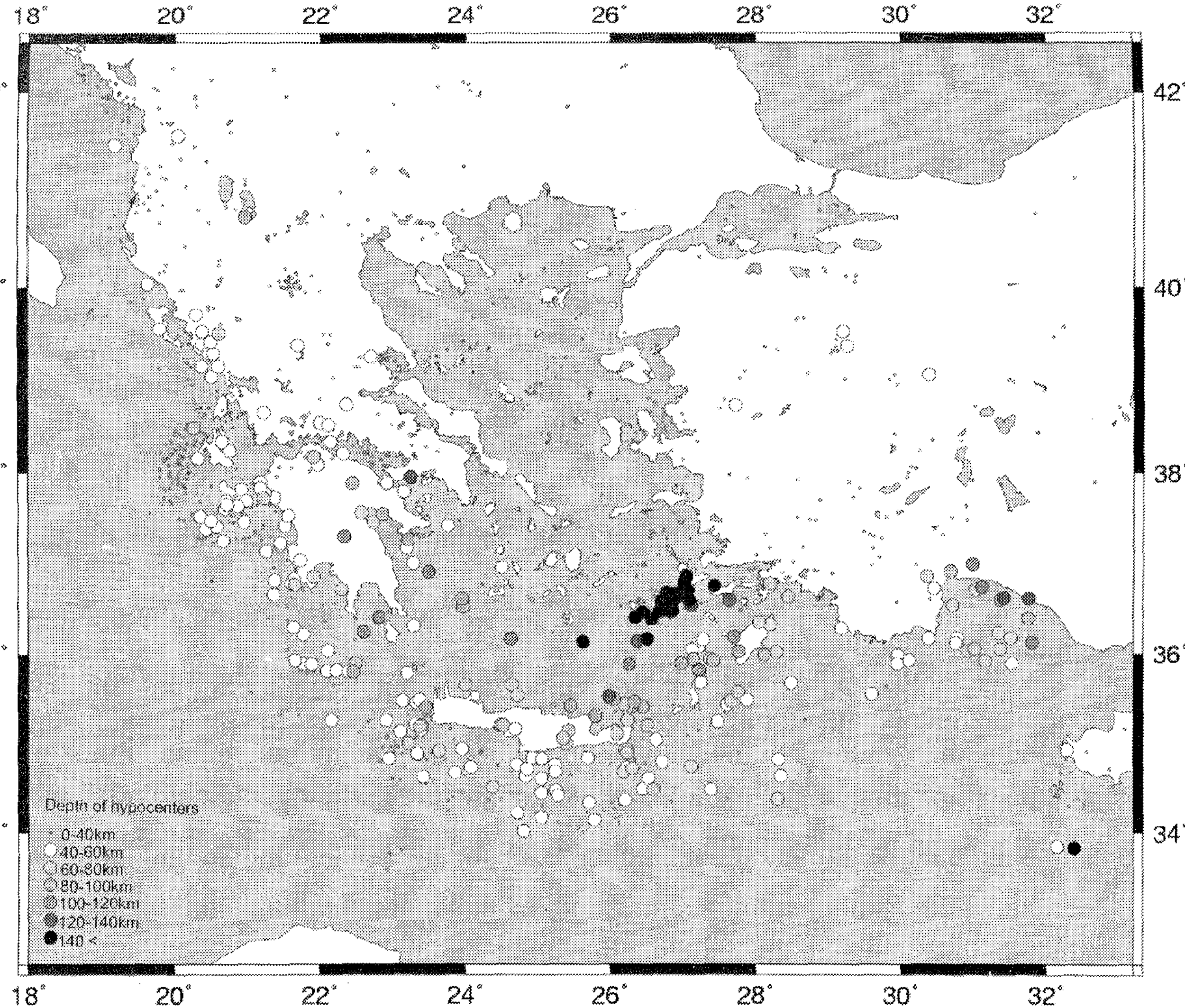

Figure. 5.5. Hypocenter distribution of earthquakes in the Aegean and surrounding area. Data reported by the USGS PDE during the period 1973-1997. The earthquakes with magnitudes >4,5.

There is a sharp boundary between shallowly and steeply dipping sub-crustal seismicity, which can be explained by gravity pulling on the slab. A part of the process could due to the loading of the subducted lithosphere by the overriding plate. It is clear that the plunge of the T-axes of micro-earthquakes' focal mechanism solutions deeper than 30 km, investigated by *Hatzfeld et al. (1989)* support the idea of a pulling of the cold lithosphere towards the north. They located two mechanisms within the mantle beneath the Gulf of Corinth at a depth of 50 km, which show N-S extension, similar to the shallow earthquakes within the overriding crust. Another microearthquake study of *Hatzfeld et al. (1993)* showed that the P axes were nearly horizontal and trending N-S to NNE-SSW, but T axes of earthquakes deeper than 80 km plunge steeply along the downgoing slab direction in the western and eastern parts. They interpret this uncommon behaviour as the downgoing slab being under a dominantly gravitational body force of the slab acting on own body. Based on micro-earthquakes *Hatzfeld et al. (1993)* and on moderate earthquakes *Taymaz et al. (1990)* observed a similar pattern for some large earthquakes located deeper than 40 km beneath Crete. Based on geophysical and geological data analysis *Royden (1993b)* studied some retreating and advancing subduction boundaries. He suggested that, across a retreating subduction boundary in the South Aegean, the system shows that subduction is driven by gravitational forces acting on dense subducted slabs at depths between about 50 and 250 km (see also e.g., *McKenzie, 1972; Le Pichon & Angelier, 1979; Meulenkamp et al., 1988*). At subduction boundaries where the rate of oveall plate convergence is less than the rate of subduction (termed here retreating subduction boundaries), the transmission of horizontal compressive stress across the plate boundaries is small, and regional deformation of the overriding plate is by horizontal extension (*Royden & Burchfiel, 1989; Royden, 1993b*). The tectonic expression of these retreating subduction boundaries includes topographically low mountains, little erosion or denudation, low-grade to no metamorphism, little or no postcollisional convergence, anomalously deep foredeep basins, and a pronounced history of flysch deposition within the adjacent foredeep basin (*Royden & Burchfiel, 1989; Royden, 1993b; Mueller & Kahle, 1993*). *Royden (1993a,b)*, who recognized this evidence more or less in the Aegean region, suggests that the density of the subducting plate plays an important role to drive this kinematic (see also *Royden & Burchfiel, 1989; Mueller & Kahle, 1993*). The old, i.e. cold and dense, crust of the Mediterranean (see Section 4) probably enables activation of gravitational potential forces on the slab. The nappe emplacements seen over Peloponnese, Crete and southwestern Turkey are related to old, i.e. copressional, paleotectonic regimes. Since the starting of the extensional regime that constructed the Aegean region, the thrusting has been taking place only along the Crete, western Pelopennese, Karpathos and Rhosos represents a boundary condition of the SAA system (see also *Angelier, 1981; Le Pichon & Angelier, 1981; Meulenkamp et al., 1988*). Further north at the back of the arc there is the greatest extensional deformation and associated seismicity (called here the Aegean extension) in any of the continents (*Truffert et al., 1993*). Sediment is removed from the upper surface of the subducting lithospheric plate and added to the edge of the overriding plate (e.g., *Karig, 1974; Von Huene, 1984; Moore & Silver, 1987*). This process probably leads to outward thrusting along the trench.

5.3.5. African Plate

The African plate is sinking towards the north beneath the Aegean region. The nature of the lithosphere is important to understand how the slab generates slabpull force. The Mediterranean basins are significantly shallower than the major oceanic basins in the world. The depths in the eastern Mediterranean do not generally exceed 4000 m (*Ryan et al., 1969*). The southern boundary of the Mediterranean is a passive boundary between African continental crust and Mediterranean oceanic (?) crust. These two crusts are parts of the African plate. The northward directed motion of the African plate beneath Eurasia proceeds at about 10 mm/yr (*Minster & Jordan, 1978; Chase, 1978; Argus et al., 1989; DeMets et al., 1990, 1994*). East of the Mediterranean Sea, the Arabian plate moves northward with respect to Eurasia at a rate of about 25 mm/yr (see Table. 5.1 for detail) (*DeMets et al., 1990, 1994; Barka & Reilinger, 1997; Reilinger et al., 1997*). The difference between the velocity vectors in the Arabian plate (25 mm/yr) and Mediterranean Sea (10 mm/yr) is taken up by the left lateral Dead Sea fault. The African plate is subducting to the north along the South Aegean Consumption Boundary (SACB) beneath the Aegean region.

The type of the Mediterranean crust is not well known, although there have been many studies. Seismic profiles show that the Ionian, Sirte and Herodotus basins have a relatively thin crust (8 to 11 km) overlain by a thick sedimentary cover, up to 10 km in the Herodotus abyssal plain. The crustal structure is of oceanic type for both the Ionian and Sirte basins, where typical oceanic layers 2 and 3 are recognised by *De Voogd et al. (1992)* (see also *Ferrucci et al., 1991; Ginzburg & Ben-Avraham, 1987*). But the thin crust of the Herodotus basin may be interpreted either as oceanic or thinned continental crust (about 10 km thick)

(*Lort et al., 1974; Makris & Stobbe, 1984; De Voogd et al., 1992*). The top of the crust of the Herodotus basin is much deeper (*De Voogd et al., 1992*). Therefore, the Herodotus basin is probably significantly older than the Ionian basin, Triassic versus Early Cretaceous in age *(De Voogd et al., 1992). De Voogd et al. (1992)* concluded that the three deep eastern Mediterranean basins mentioned here (Jonian, Sirte, and Herodotus abyssal plains) are most likely relics of a Mesozoic (Neothetis) oceanic crust.

The thickening of sediments above the subducting African plate continues at least 100-150 km south of Crete, and is probably responsible for the shallow bathymetry and deformation of the Mediterranean Ridge, as suggested by numerous authors (e.g., *Ryan & Heezen, 1965; Ryan et al., 1969; Jongsma, 1975; Ryan et al., 1982; Le Pichon et al., 1982; Jackson & McKenzie, 1988; Taymaz et al., 1990*). The deformation of the Mediterranean Ridge is complex and greatly affected by diapiric and dissolution structures (*Ryan et al., 1982; Le Pichon et al., 1982*). Since the work of *Ryan et al. (1969)*, many authors (e.g., *Biju-Duval et al., 1974; Le Pichon et al., 1982; Ryan et al., 1982*) have suggested that this structure is essentially an accretionary prism of material that has been scraped off the downgoing slab in the South Aegean Trench during subduction (see review by *Le Pichon et al., 1982*). It is either thickened in the Mediterranean Ridge or underplates Crete and is partly responsible for its uplift (e.g., *Angelier et al., 1982; Taymaz et al., 1990*). *Le Pichon et al. (1982)* suspected, from the bathymetry and seismicity, that the steep submarine escarpments south of both the Gavdos Rise and the rise farther east are bathymetric expressions of steep northward dipping imbricate reverse faults that reach the surface. *Jongsma (1975)* explained this mechanism as "crumbling and piling up of the sediment carpet" on the Mediterranean Ridge. So the sedimentary cover is deforming independently from the basement, partly by folding. On the other hand the Mediterranean Rise was interpreted by *Woodside & Bowin (1970)* as a crustal thickening in the sediments due to piling in front of the subduction zone at the SAA or to crustal shortening (*Rabinowitz & Ryan, 1970; Finetti, 1976*). But based on recent seismic profiling studies the crust is not too thin to consider this theory. Based on seismic profiling *Hirn et al. (1996)* suggest that a continental backstopped accretionary prism, located well in front of the (SAA), does not seem unlikely as the western end of the subduction zone (Ionian islands). But the sedimentary pill is crushed with African continental crust to the south. Sedimentary wedges (slices) have been thrust upward to the south away from the trench axis, and thus are not the results of gravitational slumping (*Rabinowitz & Ryan, 1970*). There is also an indication that the intensity of folding increases from south to north (*Allan & Morelli,* *1971*). *Kastens (1991)*, based on detailed geological and geophysical studies, found that the Mediterranean Ridge has grown at a rate of 10 about mm/yr. The Mediterranean Ridge rate of growth is faster than that inferred for most other accretionary wedges (*Kastens, 1991*). This could have been caused by the onset of back arc extension in the Aegean and/or by the introduction of a mechanically weak halite layer into the incoming stratigraphic section (*Kastens, 1991*). He further calculates the back arc extension age by considering sediment thickness of the front of the arc. Since the sediment cover in the eastern Mediterranean is considerably thicker (4 km according to *Sancho et al., 1973*) than that of the Pacific Plate (9.5), it seems probable that the amount of material, which has been added to the SAA, is very large (*Jongsma, 1975*). It is here postulated that the uphilling Ptolemy and Stabo Mountains (*Ergün et al., 1998*) are the result of sedimentary accretions to the SAA as a result of subduction (*Jongsma, 1975*).

5.3.6. Cretan Basin

The Cretan basin is situate between the outer arc (Kithera, Andikithera, Crete, Karpathos and Rhodes) and the active volcanic arc (the Cyclades, Thira, Milos etc.). Based on crustal thickness determined from DSS and gravity inversion (*Makris, 1978a*), maximum extension occurs in the Cretan Basin region that is just behind the consuming boundary (see Table. 4.1). It has been stretched perhaps more than any other part of the Aegean (*Le Pichon & Angelier, 1979, 1981; Angelier et al., 1982*), but now it is an undeforming part of the Aegean region (see also *McKenzie, 1978*). *McKenzie (1978)* suggested that the region moves to the south as a whole i.e., not deforming internally. It is characterised by the highest positive gravity anomalies in the Aegean region. The models of some scientists (*McKenzie, 1978; Le Pichon & Angelier, 1979; Angelier & Le Pichon, 1980; Angelier et al., 1982*), based on the relation between crustal thickness and extension, show that Crete is detached from the Cyclades area. The three acoustic units that represent basin filling sediment are also observed in the Cretan basin and in other parts of Aegean Sea by *Mascle & Martin, (1990)*.

The deformation in this region must have slowed down recently, probably because of collision to the south of the South Aegean Trench. By using seismic profiling information *Mascle & Martin (1990)* observed decreasing extension since the Late Miocene in the Cretan basin. In the southern part of the Aegean there is little recent seismic activity clearly identifiable beneath the Cretan Sea (see also *Galanopoulos, 1967; Ergin, 1966*) (Fig. 5.1). This suggests inactivity or no deformation in the region. *Watson & Johnson (1969)* observe 600 m of unconsolidated sediment above layer M (measured by seismic reflection methods) in the Cretan Basin. *Galanopoulos (1967)*, by using the focal mechanism solutions of *Hodgsom & Wickens (1965)*, suggests that the active belt in the northern Aegean and the Cretan Arc formed a conjugate N-E striking fault system in the sense of *Anderson (1951)* and *Hill & Dibblee (1953)* (see also *McKenzie, 1972*). The Cretan basin consists of a series of elongated depressions, which, in general, follow the trend of the volcanic arc. There is a general eastward increase in the depths of the basins (*Maley & Johnson, 1971*). According to *Jongsma (1975)*, immediately south of the Volcanic Arc there is a major SW dipping fault with a displacement of up to 1000 m, which forms the northern boundary of the Cretan basin and is associated with a steep scarp, which locally has an angle of 30° (*Jongsma, 1975*). The southern margin of the Cretan basin is also fault controlled (*Jongsma, 1975*). *Mascle & Martin (1990)* describe this basin as a half graben. A major downvard dipping fault can be traced for 105 km N of the eastern half of Crete (*Jongsma, 1975*). The N-S faults in the region continue to the south with a NNE-SSW direction (*Angelier et al., 1982*). The E-W normal faults in Crete continue to the north in the Cretan Basin (*Angelier et al., 1982*). East of the Mora peninsula, NW-SE trending normal faults occur in the Cretan Basin. *Mascle & Martin (1990)* divide the Cretan basin into three structural regions: east, central and west. In the western Cretan basin (Argoline basin), SSW and NNW trending faults are characteristic. In the North Cretan basin, trends are generally E-W with a depth of 2000m. East of the region, NE-SW trending half grabens are dominant and these surrounds the islands in the west of Cretan Basin (*Mascle & Martin, 1990*).

5.4. DRIVING FORCES

The dynamics of the extensional deformation in the Aegean region are not clearly known. *McKenzie (1978)* suggests 3 main forces that maintained the deformation: forces on the boundaries created by the collision of the African and Arabian plates with the Eurasian plate; the gravitational force, maintained by high topography in eastern Anatolia; and the forces at the base of the lithosphere resulting from the peel-back (or rollback) of the subducted slab over the South Aegean Arc (SAA). *Makris (1977)* suggested that another cause of the extension of the Aegean area is rising of a mantle dome.

The tectonic consequences of northward motion of the Arabian plate - and its push against the Anatolian plate are modeled using a two-dimensional, plane-stress, finite element scheme to find which force is dominant in the Aegean region (*Kasapoğlu & Toksöz, 1983*). Results show that the westward escape of the Anatolian plate to the east (*Ketin, 1968*) is not enough to drive Anatolia from central Anatolia further to the west (*Kasapoğlu & Toksöz, 1983*). Numerical models show that northward convergence of the Arabian and African plates alone cannot explain movements on all known faults (*Kasapoğlu & Toksöz, 1983*). An additional driving force that would produce westward motion of the Anatolian plate is required (*Kasapoğlu & Toksöz, 1983; Royden, 1993a,b*). According to the escape model, the westward moving Anatolian plate is banded to the west and for this reason the Aegean area moves to the south with extension (*McKenzie, 1972; Brunn, 1976; Tapponnier, 1977; Dewey & Şengör, 1979; Şengör & Yılmaz, 1981; Şengör et al., 1985; Taymaz et al., 1991; Yusufoğlu, 1996*). But recent geodetic measurements show that the velocities accelerate southward to the consuming boundary. Extensional tectonics is commonly associated with subduction zones. This is manifested in the marginal sea basins or the west Pacific arcs as well as in the continental basins (Basin and Range, Altiplano) close to the east Pacific Cordilleras. With respect to Eurasia the accelerated motion of the Aegean area (west of the Anatolian block) towards the Ionian and Levantine basins (*Oral, 1993; Oral et al., 1994; Oral et al., 1995; Le Pichon et al., 1995*) needs an independent cause which is most probably the retreat of the subducted Mediterranean slab (slab pull) (*Tapponnier, 1977; McKenzie, 1978; Le Pichon & Angelier, 1979, 1981; Angelier, 1981; Le Pichon, 1982; Mercier et al., 1987; Jackson & McKenzie, 1988; Meulenkamp et al., 1988; Extröm & England., 1989; McKenzie & Yılmaz, 1991; Wortel & Spakman, 1992; Royden, 1993a,b; Oral, 1994; Oral e al., 1995; Le Pichon et al., 1995; Hatzfeld et al., 1997; Davies et al., in press*). This can account for the high heat flow behind the arc (*McKenzie, 1978*). Tomographic imaging shows a large low-velocity anomaly in the region indicating the presence of a hot, partially melted upper mantle above the Aegean slab. The slab can also create an upper mantle flow and asthenospheric upwelling toward the slab – overriding lithosphere interface (*Gülen, 1990*). Mantle convection (thermal circulation) could cause heating (*Kasapoğlu & Toksöz, 1983*). Downbanding and final detachment of the slab could accelerate this process and cause uplift and exposure of metamorphic core complexes such as the Attic-Cycladic-Menderes metamorphic complex and establishment of an extensional stress field as a result of lithospheric thinning (*Gülen, 1990*). Southward migration of the magmatic activity towards the active Aegean volcanic arc is evidence of a retreating consuming boundary towards the south. A progressive steepening of slab angle could cause an additional driving force on overriding material (e.g., *Gülen, 1990*). Subduction could act in the Aegean region by creating the necessary low-stress condition, which allows gravitational collapse (*Jolivet et al., 1994*).

According to the orogenic collapse model (*Dewey, 1988*), overthickened crust would be gravitationally unstable and tend to spread (i.e., collapse) (*Makris, 1985; Seyitoğlu & Scott, 1991, 1996b; Seyitoğlu, 1992*). This could be explained by the thermal collapse model of *Sonder & England (1989)* (see also *Sonder et al., 1987*). Deformation occurs as if the South Aegean Trench acted as a stress-free boundary, which allows radial collapse of the Aegean crust (*Jolivet et al., 1994*).

6. CONCLUSION

From GPS and SLR, *Oral et al. (1995)* and *LePichon et al. (1995)* modeled the velocity field of the Anatolian region as a counterclockwise rigid rotation about an Euler pole located near the Sinai Peninsula, relative to Eurasia. However, the lengths of the vectors are not the same. Residual velocities appear to increase toward the South Aegean Arc (SAA). The lithosphere is not behaving like a plate as described by the plate tectonic concept because the western part of the plate, called the Aegean region, is an internally deforming of the Anatolian plate (*Reilinger et al., 1997*). Africa moves northward relative to Europe at a rate of about 10 mm/yr (*Chase, 1978; Argus et al., 1989; DeMets et al., 1990, 1994*), but across the South Aegean Trench the relative motion is thought to be approximately 40 to 50 mm/yr towards the southwest (*McKenzie, 1978; Le Pichon & Angelier, 1979; Jackson et al., 1994; Oral et al., 1995*). These cause internal deformation (extension) of the Aegean area that is behind the SAA. The N-S compression related westward escape of the Anatolian plate from the Karlıova triple junction is obvious, but where the driving force that is pushing to the east ends and where the driving force that is pulling to the west ends is unclear.

Magmatism related to subduction in the Aegean region exhibits a progressive shifting towards the south (*Borsi et al., 1972; Fytikas et al., 1979, 1984; Savaşçın & Güleç, 1990; Gülen, 1990; Papadopoulos, 1997*). This could be related roughly to southward shifting of subduction since the Early Miocene. There could be more than one subduction process taking place but it is clear that the ongoing subduction is mostly southerly and leads to active volcanism (South Aegean Volcanic Arc) in the South Aegean. This shifting could be related southward to a retreat of the subduction zone. Today, the leading edge of the African plate is being subducted along the SAA at a higher rate than the relative northward motion of Africa. This is more evidence of a retreating slab (*Oral et al., 1995*). In addition, residual GPS and SLR velocities in the Aegean region increase towards the subduction zone. This indicates that slab pull is the driving force in the Aegean region (*Oral et al., 1995*). Microearthquake studies of *Hatzfeld et al. (1989)* support a pulling mechanism of the slab. *Granet & Trampert (1989)*, who studied seismic wave velocities, indicate a break in the slab at a depth of 250 km. *De Jonge et al. (1994)* pointed out that a vertical slab rupture below western Turkey initiated and facilitated slab rollback within the subduction. Paleomagnetic data show clockwise rotation in the western Aegean region, but counterclockwise rotation in the southeastern Aegean region.

Some scientists (e.g., *Kissel & Laj, 1988*) interpret the present curvature of the South Aegean Trough (SAT) in terms of opposite rotations at the two terminations of the consuming boundary. This model was suggested to explain the extensional regime in the region. It considers southward shift of the consuming boundary. But interpretations of these measurements made in the Aegean region are not straightforward because of the limited data (*Taymaz et al., 1991*). To the northeast in western Turkey, the situation is very complicated.

Recent Tomographic images (*Spakman et al., 1993; Van der Hilst et al., 1997*) show that the slab appears to be continuous to approximately 1500 km depth. This result shows that the subduction initiation time was much earlier (26-40My) than previously considered.

There is no reliable evidence indicating a second subduction in the central Aegean.

The region over Köyceğiz has a negative gravity anomaly. This could be related to the eastern part of the subduction zone (*Özelçi, 1973; Şalk, 1994; Woodside, 1975*). This zone is also identified by magnetoteulliric studies of *İlkışık (1990b)*.

Between the high velocity slab and the overriding high velocity crust, a low velocity region dominates the back arc of the SAA. Some regional studies show that the region is situated on a high European heat flow zone. Heat flow values become higher away from the Trench. The low velocity and Sn attenuation anomaly in the Aegean back arc correlates very well with an area of heat flow. These anomalies indicate a shallow Moho depth in the region.

After refraction profiling (*Makris, 1978b*) it is accurately known that the region is a purely continental type with a thin crust. In the Cretan basin deep sea drilling under the JOIDES project also indicated absence of oceanic material. In addition, the sea floor morphology of the Aegean does not have any characteristics of an oceanic crust. The morphology of the Sea Floor also reflects a crustal type of sea floor. Based on sufficient Sn propagation, the Aegean region has a continental crust (*Payo, 1976*). The thickness increases under the mainlands of Turkey and Greece, The South Aegean island arc and the Cyclades. The thickness shows a generally increasing trend from the SAT to the North Aegean Sea. Where the Aegean region has the thinnest crust (*Makris, 1978a*), the Bouguer Gravity values become positive and bathymetry depth maximal (*Makris, 1977; McKenzie & Yılmaz, 1991*). A positive gravity anomaly seen inside the coasts of western Turkey suggests a contradiction between topography and gravity. The situation appears to be normal in the Greek mainland (*Özelçi, 1973*) (see *Yılmaztürk, 1989*). In addition seismic wave velocities in western Turkey, especially in the Menderes Massif, are lower than the velocities in Greece. This all suggests that the eastern Aegean (western Turkey) region is mostly under intense simple shear of the lithosphere, as is the central Aegean region. This area is characterised by also normal to locally elevated terrestrial heat flow density (*Pfister et al., 1997*). In addition, the extent of crustal contamination increases gradually from east to west along the South Aegean Volcanic Arc (SAVA) (*Gülen, 1990*). This also indicates increase of crustal thickness towards the west along the arc.

Efficient Sn and inefficient Lg propogation on the southwestern Mediterranean coast of Turkey represents an oceanic type of crust. The gravity high does not indicate a shallow asthenosphere as it does in the Aegean in the Ionian basin. The cause is the oceanic crustal material of the basin. Seismic refraction surveys indicate that the Ionian basin (*Ferrucci et al., 1991*) is floored with oceanic crust, but the nature of the crust, continental versus oceanic, is still debated for the Herodotus basin (*Lort et al., 1974; Makris & Stobbe, 1984*). The crust of the eastern Mediterranean is overlayed by a thick 8-12 km sedimentary cover (*De Voogd et al., 1992*).

From the north to the south, the deepest parts of the region appear as intersections of NE trending faults to the east and NW trending faults to the west. The front of a NW trending line from Anargos to Evvoia and, further NW, the intersections of NW trending faults to the east with NE trending faults to the west, appear as a NW-SE trend along deepest line in the Aegean region. Southwest of this line grabens appear to be stretched parallel to velocities obtained by geodetic methods and from seismic moment tensor summations. On the other hand northeast of this line the basins are approximately parallel to the velocity field and have a strike slip component according to focal mechanism solutions. These grabens could be thought of as cross grabens. The northeastern Aegean region is characterised by right lateral strike slip component mechanisms. In front of the NW-SE trending Evvoia island and other islands on the same line, NE deeping nodal planes of normal faulting are obvious. The area south of a line from Evvoia to Chios are moving essentially together at about 40-45 mm/yr to the SW relative to the Eurasian plate.

Zanchi & Angelier (1993) observed N-S trends of extension near the surface and NE-SW at depth in western Turkey. Moreover, they observed that the percentage of strike-slip faults increases with depth. *Pınar (in press)* identified subevents of the Dinar earthquake and find the same difference between subevents of the same earthquake.

Geological arguments for the duration of stretching and subduction are controversial in the Aegean reigon. Measured velocity rates based on geodetic and seismological methods could not be constant in time, so the calculated reconstructions may not be true for longer periods. The extension related metamorphism phase is not clear either. Lineation patterns on metamorphic rocks (detachment zones) show paleo stress directions. Some researchers (e.g., *Jolivet et al., 1994; Bozkurt et al., 1995*) show the earlier kinematic situation in the region by considering these lineations. The stretching directions are concordant with ongoing stretching directions (*Jolivet et al., 1994*).

Alkali volcanics exhibit a progressive shift towards the north (*Gülen, 1990*). The alkaline volcanics appear to be related to the extensional tectonic features. This shifting may imply that the extension in the Aegean area might have started first in the south and than propagated northward (*Gülen, 1990; Sonder & England, 1989*). But it should not be forgotten that the data are limited (*Gülen, 1990*).

In this study, results of studies have been compiled into some tables to compare the various results. Magnetic rocks with age and type information, metamorphism phases, and paleomagnetic results are also compiled. A table has been prepared that contains all the available focal mechanisms of earthquakes (5.5<Ms) from many studies in the Aegean and surrounding area, to compare existing solutions and methods. Crustal thickness estimation results based seismic refraction, magnetoteulliric soundings, and gravity in the Aegean region are listed in Table 4.1. Measured velocity rates and directions in various parts of the region by various methods are given in table 5.1 with reference information.

The data used in the Aegean region generally have gaps. Because of this and the complexity of the region there should be multi disciplinary studies. The main problems with the data are listed below.

1- Because the source is known, deep seismic sounding studies are the most reliable to determine physical properties of the medium. Their main disadvantage is their high cost. In the western Aegean region, limited DSS profiles have been taken. But there are no DSS data in the eastern part especially in the western Turkish mainland, which is one of the most important areas for the solution of kinematic problems in the region.

2- It is not expensive to get information on travel time from such earthquake data as DSS. However, these travel times are not as accurate as DSS, because the origin times and the foci of the earthquakes are not known with sufficient accuracy.

3- Many seismologists have used moment tensor summation to find strain in the Aegean region. But there are many handicaps to this method. A significant part of the deformation occurs aseismically in some of the regions. In this case, field work on faults and/or space geodetic measurements could reveal the gap. Seismically active structures have been effectively aseismic for only the last hundred years. Field study on segments of faults gives information about historical earthquakes. Most earthquakes happen on preexisting fault planes of weakness and slip can occur at different angles relative to the principle axes. We can use only large earthquakes, but 50 per cent of the displacements are related to small earthquakes. Fourth, the results might not be representative for a longer time scale. To control this we should made field study on faults and also make paleomagnetic measurements.

4- To get better results from satellite geodetic measurements more repeated measurements with more stations must be made. Moreover, to get a longer data set it is necessary to wait because repeated measurements will be made only once each year. Secondly, they are not useful for the estimation of vertical movements. Some other method should be used to estimate these motions.

5- The historical earthquake catalogues are not complete. And identification of segmentation of faults is not complete. Trenching studies and field studies are needed.

6- To prepare a reliable regional heat flow map it is important to know the thermal K value (heat conductivity) at each measurement site. But few such studies exist. For these reason the existing maps are not very reliable. Ground water mixing is another important factor that influences results.

7- Comparison between reflection and refraction profiling and seismicty and focal mechanisms should be made for all faults.

8- Studies based on detailed geological mapping, establishment of the local volcanic stratigraphy leading to the delineation of volcanic episodes and volcanic/sedimentary rock relationships and the analysis of volcanism by establishing the petrologic character of individual volcanic episodes are surprisingly few (see *Yılmaz, 1989; Gülen, 1990*).

9- Detailed structural studies on the metamorphic rocks (on detachment zones) are not complete enough to identify previous stages of ongoing extensional regime.

10- Paleomagnetic data are limited in the Aegean region especially in the north and northeastern Aegean region.

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APPENDIX. 1.

List of selected earthquakes (5.5<M) in the Aegean region and surrounding area (abravations as in Table. 2.5).

List of earthquakes (5.5<M) in the Aegean region and surrounding area (abravations as in Table. 2.5).

No	Date	Lat.	Lon.	S.	D.	R.	depth (km)	Ref.	Source
1	19951001	38.09	30.15	312	56	-84	10	Pi98a	
2	19950615	38.36	22.20	277	33	-76		Be97	
3	19950513	40.02	21.63	253	43	-95	10	Cl97G	
4				240	40	-85	14,2	Ha97a	
5	19921106	38.16	27.00	270	30	81	10>	Ha96a	
6	19870619			338	68	44	85	Ka88b	
7		36.77	28.13	316	54	137	59,5	Ex89a	
8	19860913	37.06	22.11	204	45	-77	10	Ly88a	
9		37.03	22.20	350	46	-130	15	Ex89a	
10	19850430	39.25	22.79	281	43	-72	6	Ex89a	
11		39.26	22.81	77	50	-105	11	Ta91a	
12	19850421	35.07	22.20	269	36	71	40>	PaB96	(NEIS)
13	19840621	34.80	24.20	110	72	83	39	Ta90a	
14	19840211	38.38	22.07	77	28	-121	15	Ex89a	
15	19830806	40.18	24.73	229	81	-174	10	Ex89a	
16		40.14	24.74	47	83	180	7	Ta91a	
17	19830705	40.33	27.23	218	32	71		Ey88	(Kiy87)
18		40.33	27.23	254	49	-173	15	Ex89a	
19	19830323	38.20	20.40	27	59	175		An87	
20	19830117			135	83	90		An87	
21		38.00	20.03	60	47	174		Kir91a	
22	19820817	33.77	22.96	230	45	109	39	Ta90a	
23	19820118	40.00	24.32	241	57	-176	10	Ex89a	
24	19811227	39.00	24.80	219	57	179	10	Ex89a	
25		38.09	24.92	45	90	180		Am90	
26		38.91	24.92	216	79	175	6	Ta91a	
27	19811219	39.39	25.09	220	73	171	10	Ex89a	
28		39.22	25.25	60	79	175	10	Ta91a	
29	19810304	38.18	23.17	238	42	-115		Ja88	
30		38.21	23.29	257	41	-59	10	Ex89a	
31		38.18	23.17	68	47	-82		Am90	
32		38.24	23.26	230	45	-90	7	Ta91a	
33		38.20	23.30	230	45	-105	7	PaC92	
34	19810225	38.14	23.05	250	42	-80		Ja88	
35		38.12	23.14	264	34	-51	10	Ex89a	
36		38.17	23.12	241	44	-85	8	Ta91a	
37	19810224	38.10	22.84	285	40	-70		Ja88	
38		38.22	22.93	285	37	-64	10	Ex89a	
39		38.23	22.97	264	42	-80	12	Ta91a	
40	19800709c	39.23	22.59	58	41	-128		Ja92e	
41		39.20	22.60	81	40	-90		PaC92	(Papaz83)
42	19800709b	39.27	23.04	58	41	-128	10	Ex89a	
43		39.23	22.76	58	41	-128		Ja92a	
44				270	50	-90		Am90	
45		39.30	22.90	81	40	-90		PaC92	(Papaz83)
46	19800709a	39.29	22.91	284	56	-90		Ja88	
47		39.23	22.76	270	50	-90		Am90cb	
48		39.27	23.04	58	41	-128	8	Ta91a	
49		39.30	22.90	82	42	-79		PaC92	(Papaz83)
50	19790615	34.80	24.20	150	75	70	40	Ta90a	
51	19790614	38.80	26.00	265	32	-109		Ey88	(Dr82)

List of earthquakes (5.5<M) in the Aegean region and surrounding area (abravations as in Table. 2.5) (continued)

No	Date	Lat.	Lon.	S.	D.	R.	depth	Ref.	Source
52		38.81	26.53	121	42	-50	15	Ex89a	
53		38.79	26.57	262	41	-108	8	Ta91a	
54	19790528	36.40	31.75	259	78	-90	98	Ja84a	
55	19790515			253	17	65	65	Ta90a	
56	19780620	40.73	23.25	278	46	-70		Ja88	
57		40.78	23.24	286	43	-88	10	Ex89a	
58		40.78	23.24	271	42	-74	7	Ta91a	
59		40.73	23.25	271	42	-74		Ja92a	
60	19780523	40.76	23.27	74	36	-96	10	Ex89a	
61		40.76	23.27	74	36	-96	9	Ta91a	
62	19771128	36.05	27.76	103	46	24	85	Ex89a	
63	19770911			276	47	89	19	Ta90a	
64		34.90	23.00	320	30	90		PaB96	(Papaz91)
65	19770818	35.30	23.50	114	79	96	38	Ta90a	
66	19760612	37.50	20.60	115	70	90		An87	
67	19760511			172	80	90		An87	
68		37.40	20.40	335	14	106		PaB96	(PaE93)
69				143	77	90	13	Ba97a	
70	19751231	38.50	21.65	90	45	-90		Am90	
71	19750430	36.18	30.70	72	60	74	56	Mc78b	
72	19750404	38.09	21.98	46	54	-90	53	Mc78b	
73	19750327	40.42	26.14	41	60	-142	5	Mc78b	
74		40.42	26.14	41	60	-125		Ja88	
75		40.45	26.12	68	55	-145	15	Ta91a	
76		40.42	26.14	68	55	-145		Ja92a	
77				311	65	-35	18	Pi95a	
78	19731129	35.18	23.80	139	82	90	26	Mc78b	
79				224	67	10	18	Ta90a	
80	19731104	38.90	20.44	130	50	87	8	Mc78b	
81	19730105	35.81	21.84	136	60	94	33	Mc78b	
82	19720917	38.28	20.34	306	80	-26		An87	
83	19720913	37.93	22.39	128	60	170	33	Mc78b	
84		38.00	22.40	295	45	-70	75	Am90	
85	19720504	35.12	23.61	106	36	90	46	Mc78b	
86				112	74	98	41	Ta90a	
87	19720314	39.28	29.42	101	35	-88	33	Mc78b	
88		39.28	29.42	277	55	-88		Ey88	
89	19710512b	37.58	29.60	64	50	-75		Ja88	
90				64	50	-75		Ja92b	
91	19710512a	37.59	29.76	42	42	-73	23	Mc78b	
92		37.59	29.76	32	42	-66		Ey88	
93	19710223	39.62	27.32	86	66	160		Ey88	(Pa86)
94	19700408	38.43	22.66	90	70	-112	17	Mc78b	
95		38.43	22.66	90	74	-115		Ja88b	
96		38.34	22.86	75	67	-94	9	Ta91a	
97		38.43	22.66	75	67	-94		Ja92a	
98		38.30	22.60	75	67	-94		PaC92	(Li89)
99	19700328			128	55	-90	20	Mc78b	
100				308	35	-90		Ey85a	
101				308	35	-90		Ja88	

List of earthquakes (5.5<M) in the Aegean region and surrounding area (abravations as in Table. 2.5) (continued)

No	Date	Lat.	Lon.	S.	D.	R.	depth	Ref.	Source
102		39.18	29.49	308	35	-90		Ja92	
103	19691013	39.86	20.64	97	82	90	8	Mc72b	
104				92	83	61		An87	
105	19690708	37.50	20.30	145	88	90		An87	
106				147	78	86	10	Ba97a	
107	19690612	34.40	25.06	97	63	80	25	Mc72b	
108				163	50	44	19	Ta90a	
109	19690416	35.34	27.77	104	80	85	45	Mc72b	
110	19690328	38.59	28.45	101	61	-90	9	Mc72b	
111				281	34	-90	6-10	Ey85a	
112		38.59	28.45	281	34	-90		Ja88	
113		38.59	28.45	281	34	-90		Ja92	
114				100	52	-93	13	Pi95a	
115	19690325			90	40	-104	8	Ey85a	
116		39.18	28.37	90	40	-104		Ja92	
117	19690323	39.14	28.48	112	34	-90	8	Ey85a	
118	19690303	40.12	27.43	107	50	135	4	Mc72b	
119		40.12	27.43	107	50	147		Ey88	
120		40.08	27.50	60	40	68	5	Ta91a	
121		40.08	27.50	60	40	68		Ja92a	
122	19690114	36.18	29.20	100	74	82	33	Mc72b	
123		36.10	29.20	282	25	95		PaB96	
124	19681205	36.58	26.97	212	46	-117	35	Mc72b	
125	19681031	36.60	27.00	140	82	30		Ey88	(Dr82)
126	19680530	35.49	27.96	104	80	85	21	Mc72b	
127	19680328	37.80	20.90	120	71	65		An87	
128	19680219	39.40	25.00	205	70	180		Ja88	
129				159	82	-3	20	Pi95a	
130	19671130	41.50	20.50	152	40	-87	29	Mc72b	
131				200	58	-80		An87	
132	19670730	40.70	30.40	151	44	-48	16	Mc72b	
133		40.70	30.40	121	50	-70		Ey88	
134	19670722	40.70	30.80	93	90	180	4	Mc72b	
135		40.67	30.69	275	88	-178	12	Ta91a	
136		40.67	30.69	275	88	-178		Ja92a	
137				276	89	-180	10	Pi96a	
138	19670501	39.70	21.30	184	52	-70	15	Mc72b	
139				197	56	-44		An87	
140		39.40	21.10	200	56	-40		Ja88b	
141	19670304	39.20	24.60	98	54	-119	33	Mc72b	
142		39.20	24.60	98	60	-115		Ja88	
143		39.25	24.60	313	43	-56	10	Ta91a	
144		39.20	24.60	313	43	-56		Ja92a	
145	19661029	38.80	21.18	330	45	-90		Am90	
146				204	70	113		An87	
147	19660509	34.50	26.40	132	46	110	16	Ta90a	
148	19660205	39.07	21.72	136	46	-61	22	Mc72b	
149		39.10	21.70	252	66	-100		An87	
150		39.10	21.70	103	23	-75		PaC92	
151	19651128	36.10	27.40	128	58	30		Ca67b	
152	19650823	40.50	26.02	171	70	-48		Ey88	(Ko65)

List of earthquakes (5.5<M) in the Aegean region and surrounding area (abravations as in Table. 2.5) (continued)

No	Date	Lat.	Lon.	S.	D.	R.	depth	Ref.	Source
153	19650706	38.40	22.40	87	76	-90	20	Mc72b	
154		38.40	22.40	90	74	-115		Ja88b	
155	19650613			285	2	28		Ca67b	
156		37.80	29.30	281	20	-90	16	Mc72b	
157	19650427	35.42	23.30	191	64	-79	14	Ly88	
158	19650409	35.10	24.30	50	70	145	51	Mc72b	
159				63	76	157	51	Ta90a	
160	19650405	37.70	21.80	125	74	-36	34	Mc72b	
161		37.48	21.97	125	74	-32	10	Am90b	
162	19650331	38.60	22.40	136	76	78	78	Mc72b	
163		38.38	22.26	136	76	80	45	Am90b	
164	19650309	39.40	24.00	40	90	180	18	Mc72b	
165		39.34	23.82	135	85	15	7	Ta91a	
166	19641006	40.30	28.20	122	54	-90	10	Mc72b	
167		40.30	28.20	122	34	-90		Ey88	
168		40.10	28.00	302	36	-90		Ja88	
169		40.30	28.23	100	40	-90	14	Ta91a	
170	19640717	38.20	23.70	153	74	122	150	Mc72b	
171		38.05	23.61	270	45	-90	155	Am90	
172	19640429	39.30	23.70	225	80	-140	33	Ta91a	
173	19640411	40.50	25.00	130	90	0	33	Mc72b	
174		40.50	25.00	40	90	180	33	Ta91a	
175	19640130	37.30	29.90	296	89	26		Ca67b	
176				329	85	79		Ca67b	
177	19631216	37.10	20.90	96	80	90	15	Mc72b	
178				64	64	90		An87	
179				291	7	74	6	Ba97a	
180	19630918	40.90	29.20	118	20	-82	33	Mc72b	
181		40.71	29.09	268	70	-145		Ey88	
182		40.60	29.00	268	70	-125		Ja88	
183		40.90	29.20	304	56	-82	15	Ta91a	
184	19630726	42.10	21.40	45	70	-156	5	Mc72b	
185				303	74	-21		An87	
186	19630311			61	68	16		Ca67b	
187		38.10	28.30	67	75	23		Ey88	
188	19620910	35.00	27.10	29	79	63		Ca67b	
189	19620828	37.80	22.90	107	48	125	100	Mc72b	
190		37.80	22.88	295	45	-70	95	Am90	
191	19610523	36.60	28.30	89	62	90	72	Mc72b	
192	19591115	37.80	20.26	134		7	13	Ba97a	
193	19590425b	36.90	28.50	266	27	51		Ca67b	
194	19590425a	37.05	28.55	8	40	3		Ca67b	
195				268	89	66		Ca67b	
196	19580630	36.40	27.30	195	82	14		Ca67b	
197	19570527			371	74	23		Ca67b	
198		40.70	31.00	293	74	157		Ey88	
199	19570526b			332	16	-87		Ca67b	
200				191	84	67		Ca67b	
201				79	75	156		Oc60b	
202		40.66	30.89	87	78	179	0	Mc72b	
203				157	77	175		Ca84	

List of earthquakes (5.5<M) in the Aegean region and surrounding area (abravations as in Table. 2.5) (continued)

No	Date	Lat.	Lon.	S.	D.	R.	depth	Ref.	Source
204				196	31	162		Ca84	
205				164	72	171		Ca84	
206				78	90	180		Ja92c	
207	19570526a	40.80	30.80	331	68	-27		Ca67b	
208	19570426	36.20	28.90	192	24	2		Ca67b	
209	19570425			219	83	-42		Ca67b	
210		36.45	28.59	58	85	19	0	Mc72b	
211	19570424			376	78	81		Ca67b	
212				115	27	29		Ca67b	
213		36.37	28.61	83	63	16	50	Mc72b	
214	19570308b	39.40	22.70	58	41	-128		Ja92e	
215	19570308a	39.40	22.70	120	45	-90		Ja88	
216		39.36	22.63	110	45	-90		Am90e	
217		39.40	22.70	58	41	-128		Ja92e	
218	19560710	36.70	26.30	74	54	47		Ca67b	
219	19560709			122	72	-20		Ca67b	
220		36.69	25.81	60	61	8	0	Mc72b	
221		36.60	25.90	60	40	-90		Ja88	
222				60	45	-90		Am90b	(Sh72)
223	19560220			105	88	77		Ca67b	
224		39.86	30.49	140	56	-47	9	Mc72b	
225		40.00	30.10	264	50	-135		Ja88	
226	19550716			289	56	-32		Ca67b	
227		37.66	27.19	55	51	-131	6	Mc72b	
228		37.60	27.20	55	51	-137		Ja88b	
229	19550421	39.30	23.03	90	45	-90		Am90	
230	19550419	39.30	23.10	284	56	-90		Ja88e	
231		39.30	23.03	90	45	-90	6	Am90e	
232		39.30	23.10	58	41	-128		Ja92e	
233	19540430	39.24	22.17	158	62	-103	0	Mc72b	
234		39.30	22.20	305	45	-90		Ja88	
235		39.30	22.20	300	45	-70		Ja92c	
236	19530812	38.11	20.72	163	34	101	0	Mc72b	
237	19530318			149	78	13		Ca67b	
238		40.07	27.39	59	76	166	0	Mc72b	
239		40.07	27.39	59	76	174		Ey88	
240		40.00	27.40	60	90	180		Ja88c	
241	19490723	38.66	26.29	141	65	-31	0	Mc72b	
242		38.66	26.29	262	41	-108		Ta91e	
243		38.70	26.70	262	41	-108		Ja92e	
244	19480911	37.20	23.20	300	45	-90		Ja88d	
245	19471006	36.80	22.00	200	45	-80		Am90e	
246	19441006	39.70	26.80	80	45	-90		Ja88	
247		39.70	26.80	262	41	-70		Ja92e	
248	19440625	39.00	29.40	308	35	-90		Ja88c	
249	19430620	40.70	30.38	176	76	0	0	Mc72b	
250				266	75	172		Ca84	
251				88	83	180		Ca84	
252				262	79	178		Ca84	
253		40.80	30.60	176	76	0		Ja88	
254		40.68	30.47	080	90	180		Ja92d	

List of earthquakes (5.5<M) in the Aegean region and surrounding area (abravations as in Table. 2.5) (continued)

No	Date	Lat.	Lon.	S.	D.	R.	depth	Ref.	Source
255	19421115	39.30	28.10	90	40	-104		Ja88e	
256	19410827	41.70	20.40	200	58	-80		Ja88e	
257	19410523	37.20	28.30	65	45	-90		Ja88d	
258	19410301	39.70	22.40	135	45	-90		Ja88	
259		39.70	22.54	150	45	-90	0	Am90d	
260		39.70	22.40	58	41	-128		Ja92e	
261	19390922	39.10	27.00	93	45	-90		Ja88	
262		39.10	27.00	262	41	-70		Ja92e	
263	19380918	38.27	22.47	295	45	-70	50	Am90	
264	19380720	38.30	23.80	110	45	-90		Ja88d	
265		38.30	23.79	290	45	-70	6	Am90d	
266	19350104	40.70	27.60	302	36	-90		Ja88	
267		40.70	27.60	100	40	-90		Ja92e	
268	19330511	40.50	23.70	95	45	-90		Ja88	
269		40.50	23.70	72	73	-168		Ja92e	
270	19330423	36.70	27.40	256	50	-90		Ja88	
271		36.70	27.40	60	45	-90		Ja92e	
272	19320929	40.50	23.70	95	45	-90		Ja88	
273		40.50	23.70	72	73	-168		Ja92e	
274	19320926	40.50	23.90	95	45	-90		Ja88	
275		40.50	23.90	72	73	-168		Ja92e	
276	19310308	41.30	22.50	120	45	-90		Ja88	
277		41.30	22.50	271	42	-74		Ja92e	
278	19300417	37.82	23.06	270	45	-90	6	Am90d	
279	19300331	39.60	23.17	310	45	-90	10	Am90d	
280	19300301	39.60	23.20	310	45	-70		Ja92d	
281	19300223	39.60	23.17	310	45	-90		Am90	
282	19280502	39.41	29.45	300	45	-90		Ja92d	
283	19280422	38.00	23.00	285	40	-70		Ja88e	
284		38.00	23.00	264	42	-80		Ja92e	
285	19280418	42.00	24.70	125	45	-90		Ja88c	
286	19280414	42.00	25.00	105	45	-90		Ja88c	
287	19280331	38.20	27.40	90	45	-90		Ja88d	
288	19270701	36.78	22.25	335	45	-90	20	Am90d	
289	19260830	36.80	23.20	160	45	-90		Ja88	
290		36.76	23.16	340	45	-90	75	Am90d	
291	19250807	37.90	29.60	64	50	-75		Ja92e	
292	19250706	37.79	21.94	295	45	-70	70	Am90	
293	19241120	39.08	30.14	308	35	-90		Ja92e	
294	19231215	39.90	23.20	135	45	-90		Ja88d	
295	19231205	39.85	23.60	135	45	-90	6	Am90d	
296		39.90	23.20	315	45	-45		Ja92d	
299	19191118	39.10	27.40	270	45	-90		Ja88d	
300	19171224	38.45	21.75	90	74	-115		Am90	
301	19160927	38.85	22.85	290	45	-70		Am90	
302	19160206	39.00	23.50	300	45	-70		Am90	
303	19150604	39.10	21.40	200	56	-40	9	Am90e	
304	19141017	38.20	23.50	100	45	-90		Ja88d	
305	19141003	37.90	30.40	222	42	-107		Ja88e	
306	19120913	40.10	26.80	41	60	-135		Ja88	
307		40.10	26.80	68	55	-145		Ja92e	

List of earthquakes (5.5<M) in the Aegean region and surrounding area (abravations as in Table. 2.5) (continued)

No	Date	Lat.	Lon.	S.	D.	R.	depth	Ref.	Source
308	19120810	40.75	27.20	68	55	-145		Ja92e	
309	19120809	40.70	27.20	41	60	-135		Ja88	
310		40.75	27.20	68	55	-145		Ja92c	
311	19111022	39.55	22.78	284	56	-90		Am90	
312	19110218	41.10	20.70	200	58	-80		Ja88e	
313	19090615	39.05	22.36	305	45	-90		Am90	
314	19090530	38.44	22.14	90	74	-115	10	Am90e	
315	19090119	38.70	26.90	90	45	-90		Ja88	
316		38.70	26.90	262	41	108		Ja92e	
317	18990122	37.20	21.69	180	45	-90	9	Am90d	
318	18980602	37.00	22.00	155	45	-90	80	Am90d	
319	18940427	38.66	23.03	290	45	-70	11	Am90d	
320	18940420	38.65	23.16	290	45	-70	6	Am90e	
321	18930523	38.30	23.25	270	45	-90	10	Am90d	

APPENDIX. 2.

Keyword index of the sited reference titles in alphabetical order. In the keyword index the title is shifted to align each significant word in a column. Index words are sorted alphabetically from center column to the right. Lines are limited in length and a plus sign will appear where the title has been cut. A '#' sign will appear where the titles begins. The keyword index includes the keyword listing, followed by author list and then the publication year. The full reference to a paper can then be found in the references (author index).

	Alphabetical list	Name of Author	Year
ical Study in the Light of New Geoc +	Western Anatolian Area: A Petrolog +	Francalanci, L.+	1990
arte Geologique de la Turquie, -Fai +	'Izmir'# Notes Explicatives de la C +	Egeran, N.+	1944
s	(1/50 000).# Geological Map of Naxo +	Jansen, J. B. H.	1973
pe	(1:5.0 Mio).# Heat Flow Map of Euro +	Cermak, V.	1978
resi Depremlerinin Odak Mekanizmala +	(1963 - 1990)# Türkiye ve Yakın Çev +	Özçep, T.+	1995
ca-Arabia-Eurasia Plate Collision Z +	(1988-1992): Kinematics of the Afri +	Oral, M. B.	1994
iyarbakır ve Padova Yer Çekim (Cazi +	(4 Pandüllü Sterneck Cihaziyle).# D +	Elbek, Y.	1963
ürkiye Baz Cazibe Şebekesi	(4 Pandüllü Sterneck Cihaziyle).# T +	Elbek, Y.	1952
Report of the L Atalante Expedition +	(4-11 August 1995).# ANAXIPROBE 95 +	Woodside, J. M.	1995
of the Strain in Greece in the Int +	(Abstract).# Geodetic Determination +	Billiris, H. et al.	1989
Structure of the Cycladic Blueschi +	(Aegean Sea) - Implications for the +	Avigad, D.	1993
n of High-pressure Metamorphic Terr +	(Aegean Sea).# Uplift and Exhumatio +	Avigad, D.+	1991
# Environmental Isotope Study and 2 +	(Bursa) Area of Northwestern Turkey +	Eisenlohr, T.+	1997
llü Sterneck Cihaziyle).# Diyarbakı +	(Cazibe) Şiddeti Ölçmeleri (4 Pandü +	Elbek, Y.	1963
Seismicity# Rates Crustal Deformat +	(Central Greece) as Determined from +	Tselentis, G. -A.+	1986
l and Instrumental Data.# Average R +	(Central Greece) Based on Historica +	Papazachos, B. C.+	1990
and Associated Seismicity in the Gu +	(Central Greece)# Graben Formation +	Myrianthis, M. L.	1984
Along the Active Nea Anchialos Faul +	(Central Greece)# Recent Tectonics +	Caputo, R.+	1990
Along the Active Nea Anchialos Faul +	(Central Greece).# Recent Tectonic +	Caputo, R.	1990
ectoniques et Sismiques.# Deformati +	(Cephalonie, Greece) - Donnees Neot +	Mercier, J. L.+	1972
termal Alanın Jeofizik Düşey Elektr +	(CSAMT) Etüdü.# Aydın - İmamkçy Jeo +	Yücel, M.	1995
l-Scale Geometry and Kinematics of +	(Cyclades and Evvia Island)# Crusta +	Gautier, P.+	1994
ure and Kinematics of Upper Cenozoi +	(Cyclades Islands, Greece).# Struct +	Gautier, P.+	1993
the Exhumation of a Blueshist Terr +	(Cyclades), Greece.# Cooling During +	Avigad, D.+	1992
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9Ar Age Spectrum Measurements.# Met +	(Cyclades, Greece) Utilizing 40Ar/3 +	Wijbrans, J. R.+	1988
les sur / extension Neogene de I Eg +	(Cyclades, Greece).# Donnees Nouvel +	Faure, M.+	1988
ullürik (CSAMT) Etüdü.# Aydın - İma +	(DES) ve Kontrol Kaynaklı Manyetote +	Yücel, M.	1995
ation Coordinates Computed from SLR +	(DUT) 95C02.# Earth Rotation and St +	Noomen, R.+	1995
arth Rotation and Station Coordinat +	(DUT) 95L02 and SSC (DUT) 95C02.# E +	Noomen, R.+	1995
g and Emplacement Ages of some Ophi +	(Eastern Central Greece).# Spreadin +	Hynes, A. J.+	1972
iz Grabeninin Jeolojisi ve Tektoniğ +	(Geology of the Gediz Graben).# Ged +	Emre, T.	1996
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e, Metamorphisme et Magmatisme des +	(Grece).# Paleogeographie, Orogenes +	Mercier, J. L.	1966
Depuis le Miocene Superieur: I Evo +	(Grece).# Sur les Mauvements Egeens +	Angelier, J.	1977
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Macroseismic Intensity.# A Tomograp +	(Greece) Derived from Inversion of +	Stavrakakis, G. N.+	1997
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cations of Geographic Information S +	(GSI)# Middle East Tectonics: Appli +	Şeber, D.+	1997
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Significance# Microstructures of D +	(Western Turkey) and their Tectonic +	Bozkurt, E.+	1997
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Measurements# Local and Regional Co +	100 Years of Geodetic Displacement +	Curtis, A.+	1995
Measurements.# Local and Regional C +	100 Years of Geodetic Displacement +	Curtis, A.+	1997
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	#12 Mayıs 1971 Burdur Depremi.	Erinç, S.+	1971
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