

CHAPTER 2

In this chapter, the geological and tectonic settings of the study area as well as the seismicity and previous studies are introduced.

2. Geology and tectonics of the Aegean

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 belts are situated in the Mediterranean region. Four of them have island arc system characteristics, with marginal seas. The southern Aegean region is one of them (Fig. 2.1)

In general view, 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 by the oceanic material beneath the Mediterranean Sea (northern edge of the African plate). The Aegean Sea floor (about 350 m mean depth) is seen as a high plateau between the deeper Black Sea floor (mean depth 1300 m) and Mediterranean Sea floor (mean depth 1500 m). The Aegean is characterized by a relatively thicker crust (25-30 km) than a typical oceanic crust, which might be interpreted as a thinned continental crust.

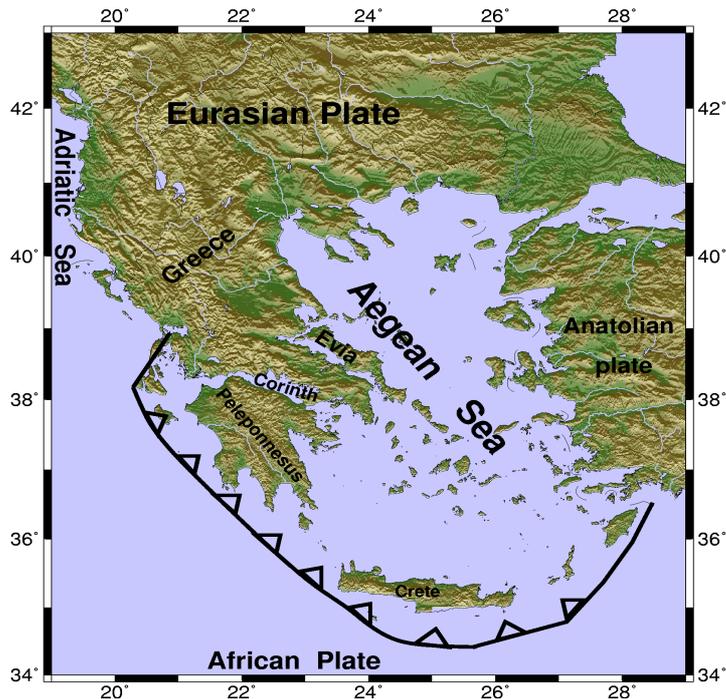


Fig. 2.1. Aegean Sea and its surrounding area. The Aegean micro plate located in the convergent region between the Eurasian and the African plates and bounded to the west by the Adriatic Sea and to the east by central Anatolia.

The Aegean is also situated in the convergent boundary between the African plate and Eurasian plate (Fig. 2.1). The crustal and mantle structures of the Aegean are the result of a complex history that started probably during the late Cretaceous (e.g. McKenzie, 1972). The African plate has rotated counter-clockwise with respect to Eurasian plate during the last 92 My (Müller & Kahle, 1993). 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 (African plate beneath Eurasian plate). However, in the Aegean a roughly N-S directed lithosphere shortening rate is increasing from west to east (Müller et al., 1997). The region is also characterized by high heat flow, which is related to thinned and deformed (stretched) continental crust. This thinning is continuing till now. For this reason, it is the world wide most seismically active and rapidly internally deforming area of the entire Alpine-Himalayan belt and at of all

continents (McKenzie, 1972; Mercier et al., 1977; Jackson et al., 1994).

2.1 Tectonic setting

The tectonic framework of the eastern Mediterranean is dominated by the collision of the Arabian and African plates with Eurasia (McKenzie, 1970; Jackson and McKenzie, 1984, 1988). Plate tectonic models (DeMets et al., 1990) based on analysis of global seafloor spreading, fault systems, and earthquake slip vectors indicate that the Arabian plate is moving in a north-northwest direction relative to Eurasia at a rate of about 18 mm/yr (McClusky et al., 2000), averaged over about 3 My (Fig. 2.2). These models also indicate that the African plate is moving in a northerly direction relative to Eurasia at a rate of about 6 mm/yr (McClusky et al., 2000). Different motions between Africa and Arabia (~ 8-15 mm/yr) are thought to be taken up predominantly by left-lateral motion along the Dead Sea transform fault. This northward motion results in westward extrusion of the Anatolian plate (McKenzie, 1970).

According to this model, the Anatolian plate is rotating counter-clockwise, relatively to Eurasia, about an Euler pole located north of the Sinai peninsula (31.1°N, 33.4°E), which results in a slip velocity of 24 mm/yr in the North Anatolian Fault (McClusky et al., 2000). This westward motion of Anatolia does not result in a compressional field in the Aegean area, which on the contrary, is dominated by an extensional stress field. Moreover, the Aegean moves as a more or less coherent unit in respect to Eurasia with much high velocity (~30 mm/yr) than the corresponding velocity of the Anatolian plate (Kahle et al., 1998) to the southwest. For this reason the Aegean area can be considered as separate microplate. This high rate convergence along the Hellenic arc, rather than the relative northward motion of the African plate itself, requires that the arc moves southward (~ 30 mm/yr) relative to Eurasia proper (e.g. Sonder and England, 1989; Royden, 1993). The corresponding velocity direction changes from ENE-WSW in the central coast of western Turkey to NNE-SSW in the southern Aegean. It also shows that additional forces (except from Anatolia westward motion) exist in the Aegean, related either to the slab pull or the gravitational collapse

Chapter 2. Geology and tectonics of the Aegean

of the Hellenides, which are responsible for this velocity acceleration and change of direction from almost E-W to almost N-S.

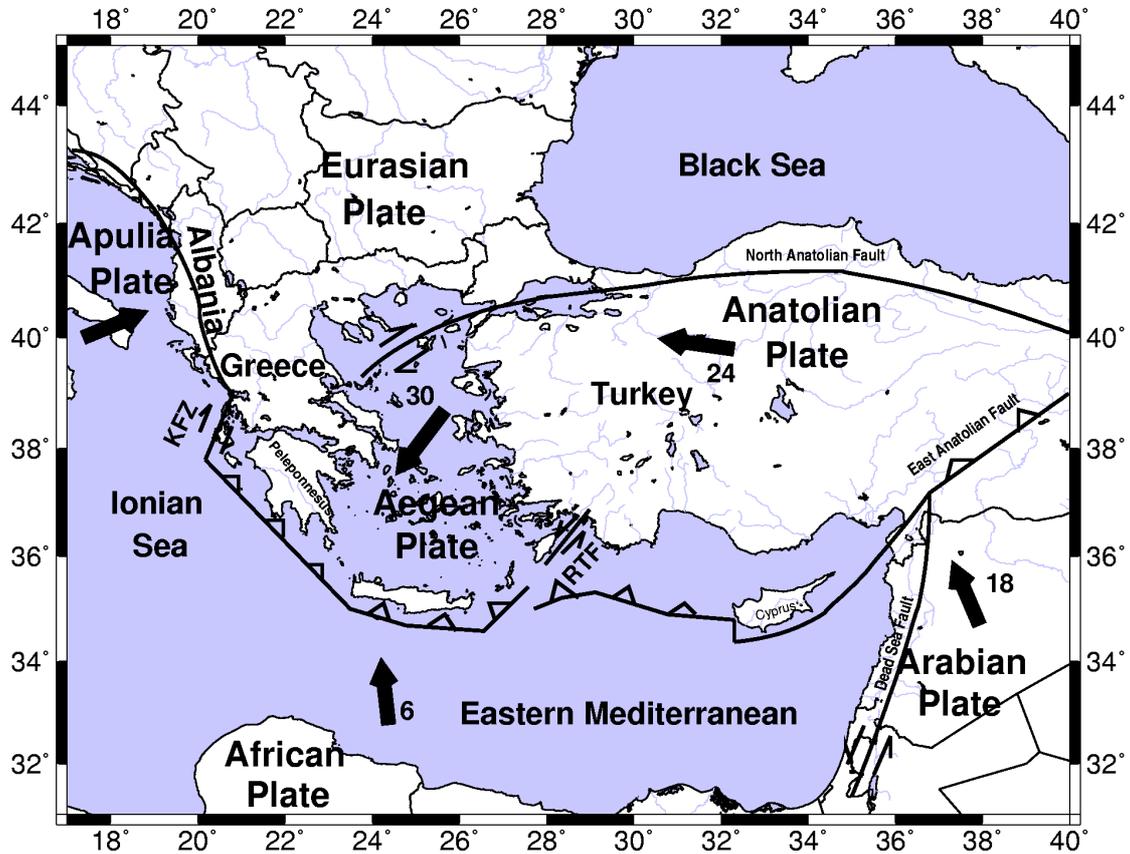


Fig.2.2 Schematic illustration of the principal motions affect active tectonics in the Aegean and surrounding area. Arrows indicate the direction of the motion relative to Eurasia. The values (mm/yr) are shown near each arrow (McClusky et al. 2000).

However, the GPS measurements (McClusky et al., 2000) indicate that relative to Eurasia, the southwestern Aegean-Peleponnesus moves toward the SSW in a coherent fashion with low internal deformation (< 2 mm/yr). The southeastern Aegean region deviates significantly from this coherent motion, rotating counter-clockwise and moving toward the Hellenic trench (i.e. toward the SE) at ~ 10 mm/yr relative to the southwestern Aegean. The Africa-Aegean relative motion is approximately normal to the arc along the western arc and parallel to the arc along the eastern segment. Rapid motions in the SE Aegean are a response to rapid sinking of the downgoing plate and

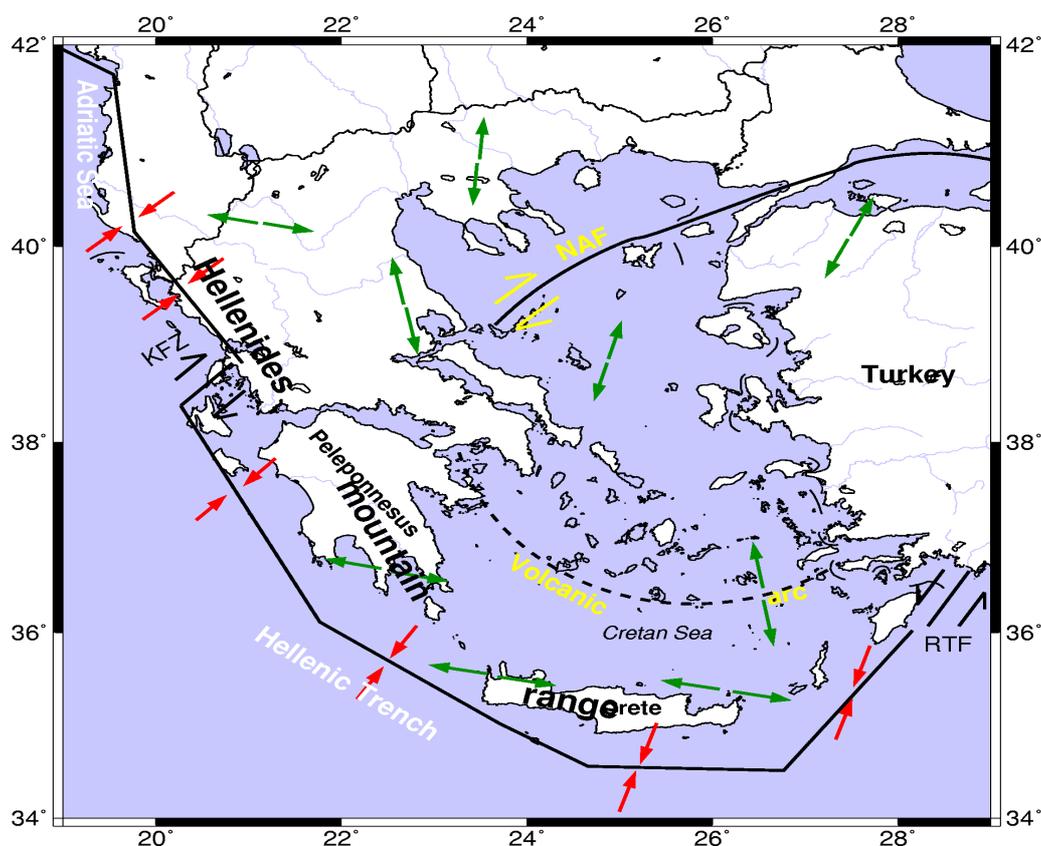


Fig. 2.3. The compression and extension zones in the Aegean. The main zone of compression with thrust faults observed along the Hellenic arc and western coast of Greece is related to the subduction of the African plate beneath Eurasian plate and continental type collision between Adriatic microplate and western Greek-Albanian coast, respectively. The back arc area is dominated by N-S extension. A narrow zone of E-W extension is observed in the forearc area. The situation of volcanic arc associated with the subduction is marked with dashed line. NAF = North Anatolian Fault; KFZ = Kefallonia Fault Zone; RTF = Rhodes Transform Fault.

allows the arc to override more easily the African plate. Right lateral strike-slip deformation associated with the North Anatolian Fault extends into the north Aegean terminating near the Gulf of Corinth or continuing to the Kefallonia fault zone (KFZ in Figure 2.2). The dextral sense of strike-slip motion along the KFZ indicates an abrupt change in the Hellenic subduction zone (Kahle et al., 1998). The north Aegean trough and Gulf of Corinth form the principal northern boundary of the southwestern Aegean plate. The main zone of compression observed along the Hellenic arc and along the western coast of northern Greece and Albania (Fig. 2.3), is associated with

Chapter 2. Geology and tectonics of the Aegean

the subduction of the eastern Mediterranean beneath the Aegean and the continental-continental type collision between Adriatic (Apulia) microplate and the western Greek-Albanian coasts (Anderson & Jackson 1987). The largest part of the backarc Aegean is dominated by normal faults with an E-W trend, suggesting an N-S extension. Fault plane solutions (Kirtazi et al. 1987, Papazachos et al. 1998) and GPS measurements (McClusky et al. 2000) indicated recently a narrow zone of E-W extension in the forearc, which lies between the thrust faults of the outer Hellenic arc and the N-S extension field in the backarc area.

The eastern limit of the Hellenic arc is defined by the sinistral Rhodes Transform Fault (RTF in Figure 2.3), which is the only main strike-slip fault in the Aegean area, which is not dextral. All data mentioned previously exhibit a consistent geotectonic setting, which is characterized by a dominant extension regime in the Aegean after Lower Miocene. Figure 2.3 shows the schematic illustration of the different zones in the Aegean area. The kinematic status of the Aegean shows that the bulk of the motion of the southern Aegean/Hellenic arc appears to be associated with coherent rotation of the southern Aegean plate. It is likely concentrated on SW striking, right-lateral strike slip and NW oriented normal faults in the northern Aegean Sea. It shows also a gradual increase of deformation velocities, with respect to Europe, from 10 mm/yr in the North Aegean Trough up to 35-40 mm/yr in the southernmost Hellenic arc.

2.1.1 The northern Aegean region

The area is characterized by a high strike slip component of faults with the N-S extensional regime of the Aegean region (Fig. 2.4). The Strymon basin, North Aegean Trough (NAT) and Skyros basin are the most pronounced and are considered as the major tectonic elements in northern Aegean region, as shown in Figure 2.4. Young tectonic activity in this region has affected the geomorphology and normal strike-slip

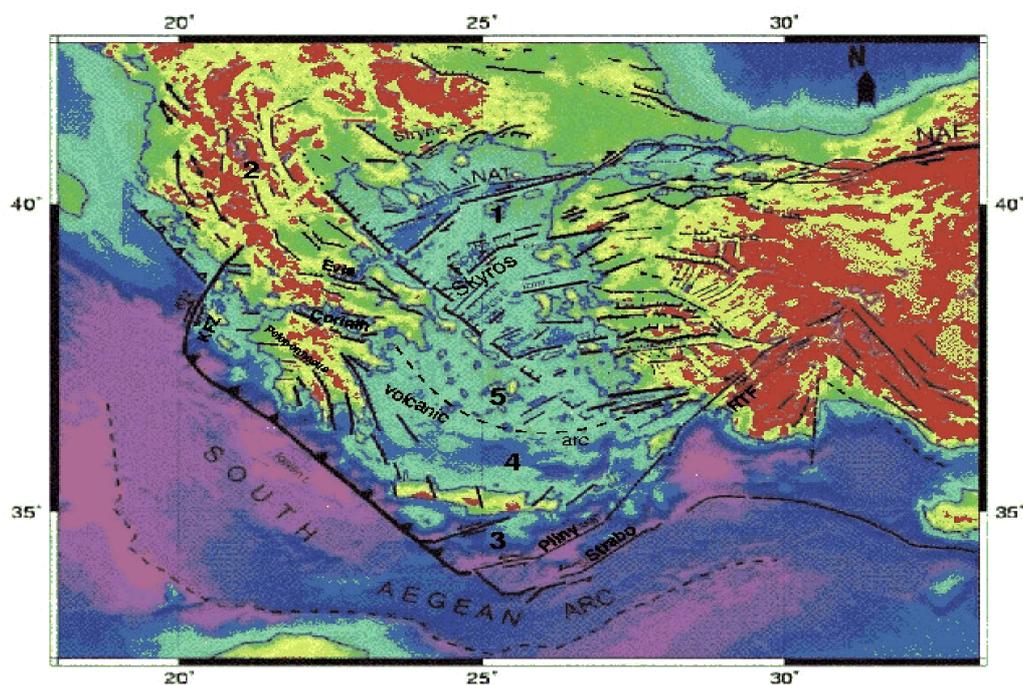


Fig. 2.4. Detailed tectonic features in the Aegean. Aegean is dominated by normal faulting. Major normal faults are demonstrated in the central and northern Aegean Sea. The most strike-slip faults are dextral in the Aegean area, the only main fault, which is not dextral is situated in the eastern part of the Hellenic arc (RTF). 1 = northern Aegean region; 2 = western Aegean region; 3 = south Aegean subduction; 4 = Cretan basin; 5 = Cyclades.

faults. Therefore, dominantly oblique faults and basins can be clearly 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). A good correlation also exists between seismic activity and high heat flow anomalies, because these are deep reaching oblique faults (Alptekin et al., 1990; Simsek, 1997). Gravity anomalies are also well correlated with these major tectonic lines. The narrow North Anatolian Fault Zone (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. As Figure 2.4 shows, the boundary becomes diffused further west.

The most dramatic bathymetric feature of the North Aegean Sea is the NAT (Fig. 2.4), which consists of a series of deep fault-bounded active basins (Vogt & Higgs, 1969; Lort, 1971; Le Pichon et al., 1984). This feature is thought to be a continuation of the

Chapter 2. Geology and tectonics of the Aegean

NAF by many scientists. Intense seismicity clearly defines also the trough. The trough is bounded by a major fault to the south and by a faulted flexure to the north (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 be related to the northward dipping major fault (McKenzie, 1972). 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 indicate 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). Focal mechanism solutions in the NAT show mainly right lateral strike-slip faulting (Taymaz et al., 1991). Large principal values of extension are also found in the northern Aegean Sea, along the NAT.

2.1.2 The western Aegean region

The Hellenides-Peloponnian peninsula is characterized by a NNW-SSE trending mountain belt that constructed of nappe piling. The belt is indicated by the high negative Bouguer anomaly that strikes NNW-SSE in the western Aegean Region (Makris, 1977, 1985). The NW-SE trending normal faults to the NE continue step by step to the center 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 like NAT. 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 recognized 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 Evia) are obvious from bathymetry and topography. Towards the Aegean Sea, these two basins become more southerly. These grabens indicate recent extensional structures

that accommodate a large part of the internal deformation of Hellenides-Peloponnesus (see for detail Armijo et al., 1996; Hatzfeld et al., 1996). The Gulf of Corinth, located between Greece and Peloponnesus, is a recent N-S extensional structure (McKenzie, 1978). It is not older than 2 My, it opens at about 1 cm/yr and accommodates a large part of the extensional deformation (Armijo et al., 1996; Hatzfeld et 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). Its western segment is connected to the Ionian trench system by a series of NE-trending transform faults with dextral polarity called the Kefallonia Fault Zone (KFZ). The rift's eastern segment transverses the volcanic arc of the Aegean orogeny and can be considered as a first order cross fault (Hatzfeld et 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; Myriantthis, 1984). North of the Kefallonia fault zone, the NW end of the subduction zone is characterized by a negligible motion (see Mueller et al., 1997). In northwestern Greece and Albania, there is a band of thrusting near the western coasts (Fig. 2.4). 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).

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 (Oral et al., 1995; Jolivet et al., 1994). On the other hand the vectors further east in the Aegean Sea are approximately parallel to the bathymetric features which could be thought as cross grabens (e.g., Sengör, 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. Furthermore, some evidences show that the sediments deposited in the Evia and Attica are tilted southward. The analysis of structures in the metamorphic domes on footwalls of Evia and Attica faults also show a very consistent fabric with flat foliation and N-S trend of stretching lineation (Faure

& Bonneau, 1988; Gautier et al., 1990; Jolivet et al., 1994). Extension is taken up by low angle normal faults, which bound upper crustal blocks such as Evia 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.

2.1.3 The southern Aegean subduction - Hellenic Subduction Zone

Based on earthquake investigations as well as geophysical investigations of gravity, magnetic and seismic wave velocities, Papazachos & Comninakis (1969, 1971), Caputo et al. (1970), Papazachos (1973), 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 southern Aegean arc (Hellenic arc). Negative low Bouguer gravity anomalies show that there is dense material (a sinkable slab) beneath Crete, Rhodes and the Pelopennesus (Ryan et al., 1969; Makris, 1985; Tsokas & Hansen, 1997).

Recent volcanism on some islands parallel to the South Aegean trough is related to a subducting plate in the Aegean region (Fig. 2.4). 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. The Hellenic arc has such general characteristics of the arcs as a trench in its convex side, a seismicity belt, 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.

However, the Benioff zone ends beneath the volcanic arc (~ 180 km), only tomographic results (Spakman et al., 1988, 1993; Papazachos et al., 1995) show that the Benioff zone must extend further to the north. At greater depths the slab could not accumulate stress because it may behave as plastic material above a critical heat and pressure. Limited isotopic data suggest that little of the sediment is incorporated into

Chapter 2. Geology and tectonics of the Aegean

the volcanic rocks of the Aegean arc (Barton et al., 1983; Briquieu et al., 1986). This means that it is not transported deep enough into the mantle for melting to occur. A low-velocity layer is observed in the upper mantle in the middle of the Aegean Sea from earthquakes 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 Hellenic arc is one of the most remarkable tectonic features of the Mediterranean area (Fig. 2.4). This is the consuming boundary of the African plate in the region. However, the location of the consuming boundary is not perfectly known. Recent tomographic results (Papazachos & Nolet, 1997) indicate that the lithosphere dives approximately along the line of the south coast of Crete.

A narrow zone of negative free air gravity anomalies (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 is extremely complicated. The South Aegean Trench named Hellenic trench, consists of a number of prominent deep basins that form an arc shape from Kefallonia to Rhodes (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 linearized 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) suggested that the Pliny Trench is the main zone of subduction, rather than the Strabo Trench as supposed by Ryan et al. (1973). 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 (Fig. 2.4).

Because the convergent rate is high enough (~ 35 mm/yr), the deforming structures

Chapter 2. Geology and tectonics of the Aegean

can be clearly seen in bathymetry without sedimentary carpet. However these deep basins of the Hellenic trench are considerably shallower and more highly curved ($R=400\text{km}$) than the other subduction related main trenches elsewhere in the world (Le Pichon & Angelier, 1979; Jongsma, 1977).

Based on microearthquake data, focal mechanism solutions of earthquakes and normal faulting in the region, indicate NW-SE extension in southern Peloponnese, western Crete and E-W extension in central and eastern Crete (Hatzfeld et al., 1993). The Peloponnese 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 Hellenic trench system. It was also 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.

2.1.4 The Cretan Basin

The Cretan basin (see Figure 2.4) is situated between the outer arc (Kithera, Andikithera, Crete, Karpathos and Rhodes) and the active volcanic arc (the Cyclades, Thira, Milos etc.). Based on crustal thickness determinations, maximum extension occurs in the Cretan Basin region. This area just behind the consuming boundary has been stretched perhaps more than any other part of the Aegean (Le Pichon & Angelier, 1979, 1981; Angelier et al., 1982), but now is an undeforming part of the Aegean region. McKenzie (1978) suggested that the region moves to the south as a whole i.e., not deforming internally. It is characterized 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 deformation in this region must have slowed down recently, probably because of collision to the south of the Hellenic 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 (Galanopoulos, 1967; Ergin, 1966). This would suggest that there is no deformation in the Cretan Sea. The Cretan basin consists of a series of elongated depressions, which follow the trend of the volcanic arc. There is a general eastward increase in the depths of the basins (Maley & Johnson, 1971). The southern margin of the Cretan basin is also fault controlled (Jongsma, 1975). Mascle & Martin (1990) described this basin as a half graben. A major downward dipping fault can be traced for 105 km N of the eastern half of Crete. 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).

2.2 Geological setting

The African plate (Tethys Ocean) is being consumed by subduction towards the north. The crust of Alpine Himalayan belt is constructed of approximately parallel trending units whose configuration developed in paleotectonic times. To both sides of the Aegean Sea (Greece and Turkey), the units that are massifs and suture zones (ophiolite and blue schists) are linked to each other (Fig. 2.5). 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 center of the Aegean Sea indicates that during the Mid-Eocene the region was characterized by a thick continental crust (> 50 km) because of a compressional regime. The timing and cause of transition from compressional to extensional regime in the Aegean region are not well defined.

Chapter 2. Geology and tectonics of the Aegean

some evidences suggest that extension may intensified during the Pliocene (~ 5 My) (Westaway, 1994b). This extension are attested by the creation of deep basins (Masclé & Martin, 1990), the exhumation of metamorphic core complexes mostly in the Cyclades islands, the crustal thinning, and the existence of major active crustal faults in the whole Aegean. If back arc extension predates the initiation of the NAF as hypothesized (Angelier et al., 1982; Armijo et al., 1996), it is at least plausible that the southern Aegean was extended prior to NAF deformation propagation into the north Aegean. This earlier mode of extension was greatest in the Sea of Crete where it reaches a factor of two (McKenzie, 1978; Angelier et al., 1982), and it is responsible of the curvature of the Hellenic arc. This extension is consistent with paleomagnetic rotations of rock units in the opposite sense observed on the east and west sides of the Aegean (e.g. Kissel & Laj, 1988). Unlike the present episode of deformation, the North Anatolian Fault was not involved in this earlier deformation. This earlier episode of deformation was discontinuous both in time and space (Mercier et al., 1989). Propagation of the NAF into the north Aegean concentrated the more recent (~ 5 My) phase of extension in the northern Aegean and in the Gulf of Corinth, leaving the southern Aegean to translate SW as a coherent unit. The relatively young age inferred for the Gulf of Corinth ~ 1 My (Armijo et al., 1996) is consistent with this scenario.

The Cretan Basin that is the most aseismic area in the region possibly had been stretched parallel to Hellenic arc. However, possible collision of the Hellenic arc with the African promontory might stop further stretching of the Cretan basin. After that event, the Aegean region underwent a new stage of the extensional regime or the second stage of extension is related to the combined effects of the still active migration of the African slab and of the westward migration of Anatolia. 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).

2.3 Seismicity

The Aegean region is one of the most seismically active areas in the world (e.g., McKenzie, 1972, 1978; Mercier et al., 1977; Jackson et al., 1982; Armijo et al., 1996). The scatter in the seismicity of the Aegean region is not due to errors in the location of epicenters, which are usually less than 30 km. Figure 2.6 shows that earthquakes are not homogeneously distributed over the Aegean but are primarily located around the Hellenic Trench and western Greece and therefore related either to the active subduction process or to the continental collision with Apulia. Earthquakes are also located in the north Aegean Sea and related to the north Anatolian fault.

Most of the seismic energy release from the north Aegean Sea. The released seismic energy is not homogeneous over the Aegean. The largest is in the north Aegean Sea, where several earthquakes with magnitude greater than 7 have been occurred since the beginning of the 19th century (Ambraseys and Jackson 1990). The less energy is in the Sea of Crete, where only one earthquake of magnitude 7 has been occurred. This difference in seismicity is also observed in the historical seismicity records. The present shallow seismicity is concentrated in two diffuse belts along the Hellenic trench and in the northern Aegean (Galanopoulos, 1967; Caputo et al., 1970; Ryan et al., 1969; McKenzie, 1970, 1972; Papazachos & Comninakis, 1971). The Hellenic trench and volcanic arc are seismic regions that contain a broad belt in which seismic activity has occurred at both shallow and intermediate depths (Fig. 2.6). Graben like basins in Greece are also dominated by high seismicity and large earthquakes ($M_s = 5.5$) (Ambraseys & Tchalenko, 1970).

As Figure 2.6 shows, the Hellenic trench elongation can be identified on a seismicity map as a curved shape. Most of the seismicity stops abruptly about 40 km from the trench. The highest concentration is seen in the western part of the Peloponnese. Some clusters are located where there are changes in the morphology of the Hellenic trench (Hatzfeld et al., 1989).

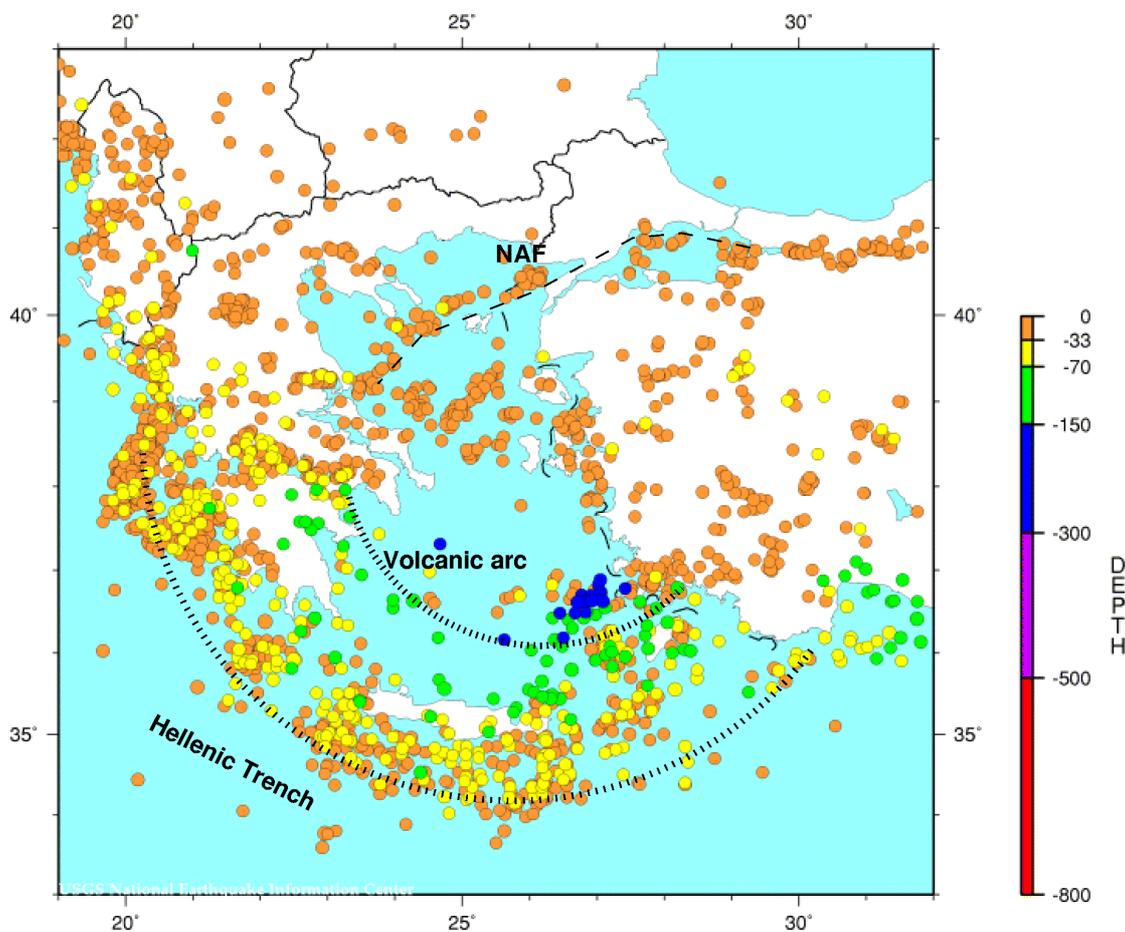


Fig. 2.6. Seismicity in the Aegean area ($m_b > 4.5$, 1973-2005, NEIC catalogue). Earthquakes are mostly located either along the Hellenic trench and western Greece associated with the subduction process and continental collision, respectively or in the north Aegean Sea related to the north Anatolian fault system. The deepest earthquakes occur in SE Aegean in the depth of 180 km.

The 100 km and 150 km isodepth contours of hypocenters of earthquakes are not parallel to the Hellenic arc (Hatzfeld & Martin, 1992) and show that the gently dipping part of the slab is longer in the west than in the east (Hatzfeld & Martin, 1992; Papazachos, 1990). In other parts of the world, convergent-plate tectonic environments similar to that of the Hellenic arc have produced earthquakes of magnitude 8 and larger.

The Cyclades islands and southern Aegean area are observed almost strain-free. There is also low seismicity in this area, although intermediate-depth earthquakes (depths greater than 50 km) occur within the subducting Africa plate beneath central Greece,

Chapter 2. Geology and tectonics of the Aegean

SE Aegean and Cyclades Islands. The seismicity map shows that the most intermediate-depth earthquakes occur in the southeastern part of the Aegean.

Reverse faulting is observed from focal mechanisms of earthquakes along the Hellenic arc, on the other hand, in the internal part of the arc normal faulting is obvious (Hatzfeld et al., 1990). The transition between reverse faulting and normal faulting is rather sharp, but with some overlap of the two families (Hatzfeld et al., 1990). 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 Cyclades. These results are consistent with the seismicity obtained by a careful selection of well defined events (ISC data) by Martin (1988).

2.4 Previous geophysical studies

The lithospheric structure in the Aegean area has been the focus of many intensive geological and geophysical studies. Due to the tectonic complexity in this area, which includes the subduction of the African plate beneath Aegean as well as the internal extension within the whole Aegean, studies are generally divided into two main parts. One of them concentrates on the structure of the continental Aegean plate, while the other deals with the oceanic lithosphere of the subducted African plate.

2.4.1 The continental Moho of the Aegean plate

Due to the geodynamic evolution of the Aegean area, the crustal thickness undergoes strong variations. This variations determined to be over a range of 15-25 km beneath the back-arc basin (Aegean Sea) to 40-60 km beneath the orogenic belts (mainland Greece)(e.g., Alessandrini et al., 1997). In Table 2.1, brief information is given about crustal thickness in various parts. In general, the continental lithosphere thins from the north to south (Makris, 1978a,b).

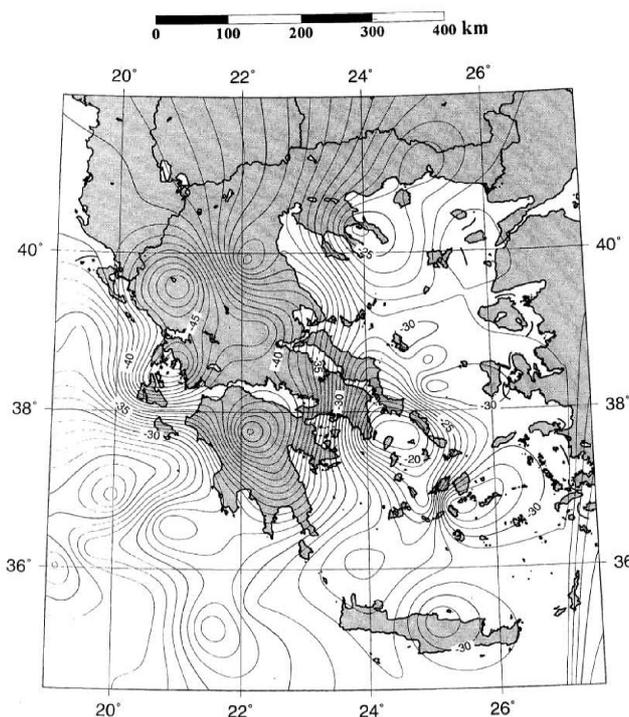


Fig. 2.7. Moho depth map obtained from gravity data by Tsokas & Hansen (1997).

After a deep seismic sounding study of Makris (1973), it is clearly understood that the crust in Aegean region is continental but thin. Results from seismic refraction experiments (Deep Seismic Profiles - DSP) were presented in a series of papers by Makris (1973, 1976, 1978). Makris in his pioneering work, proposed an average crustal thickness for the Aegean and the Hellenides mountain range that varied between 30 and 44km, respectively, whereas for the southern Aegean (Cretan sea) between 20 and 25km and less than 20km for the Ionian Sea. Makris (1977) also compiled the first map of the Mohorovicic discontinuity depth in the broader Aegean area by combining data from these seismic experiments with gravimetric data. Similar results were also obtained from DSP data by Makris & Stobbe (1984), Delibasis et al. (1988), Bohnhoff et al. (2001) and Clément et al. (2004). These results were also confirmed later by travel time analysis from local and regional earthquakes (Panagiotopoulos & Papazachos, 1985; Papazachos, 1994; Papazachos et al., 1995).

Chapter 2. Geology and tectonics of the Aegean

Maps of the Moho topography derived from gravity data imaged the variations of the crustal thickness in the area of Greece and Aegean Sea (Lepichon & Angelier, 1979; Makris & Stobbe, 1984; Tsokas & Hansen, 1997; Tiberi et al., 2001; Tirel et al., 2004). The more complete Moho map including Aegean, Greece and the island of Crete was obtained from gravity data along six profiles within Aegean by Tsokas & Hansen (1997). They estimated the mean crustal thickness for the area of Greece to be 32 km, whereas it thins to 20 km and 25 km in the central Aegean Sea and north Aegean region, respectively. Their results showed the thickness of 40-49 km beneath the Hellenides in western Greece (Fig. 2.7).

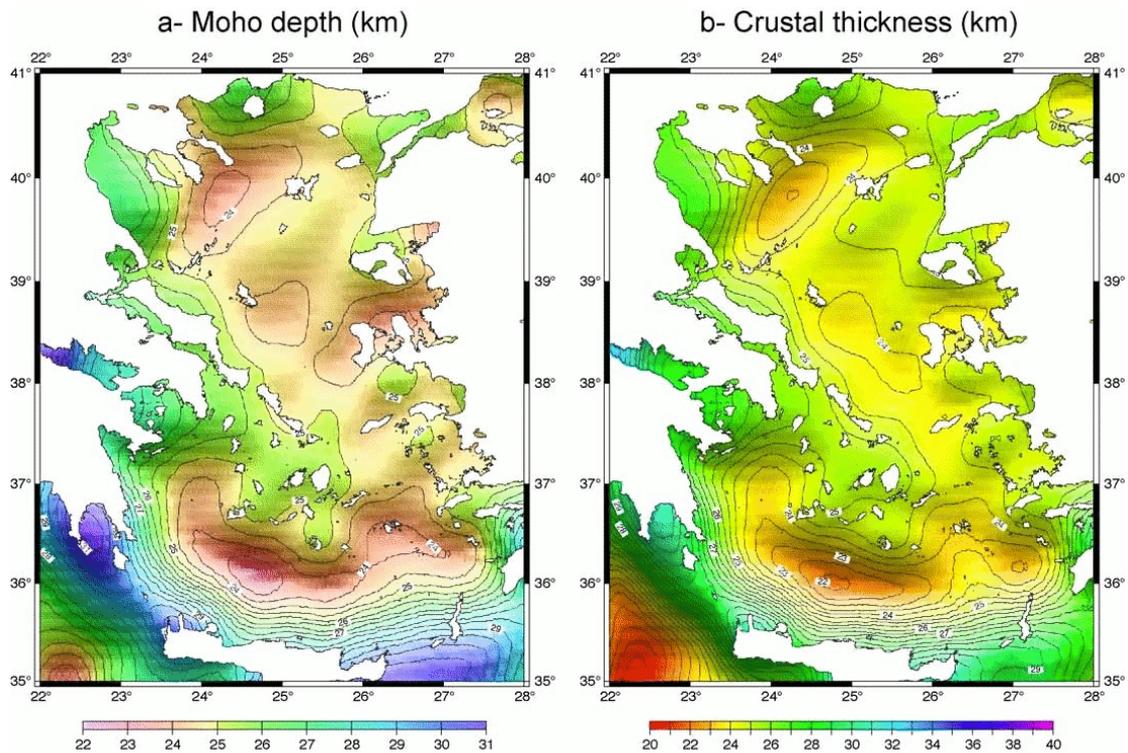


Fig.2.8. Moho depths and crustal thicknesses in the Aegean region deduced from gravity inversion by Tirel et al. (2004).

The more recent work of Tirel et al. (2004) inferred from gravity inversion across the Aegean Sea revealed a homogeneous and relatively thin crust about 25 km in this region, with variations of only +2 km near continental Greece and Anatolia, and -2

km across the North Aegean trough and Cretan Sea (Fig. 2.8). The most homogeneous image of the Aegean velocity structure is deduced from tomographic studies (Spakman, 1986, Christodoulou & Hatzfeld 1988, Spakman et al., 1988; Ligdas et al., 1990; Drakatos & Drakopoulos 1991; Ligdas & Main 1991; Ligdas & Lees 1993; Spakman et al., 1993; Papazachos et al., 1995; Papazachos & Nolet, 1997; Karagianni et al., 2003, 2005). Karagianni et al. (2005) used the Rayleigh wave group velocities to infer a 3-D tomographic image of the shear wave velocity structure of the crust and upper mantle in the Aegean Sea using regional events (Fig. 2.9). This work is complemented with a group and phase velocity study at longer period using teleseismic events (Bourova et al., 2005).

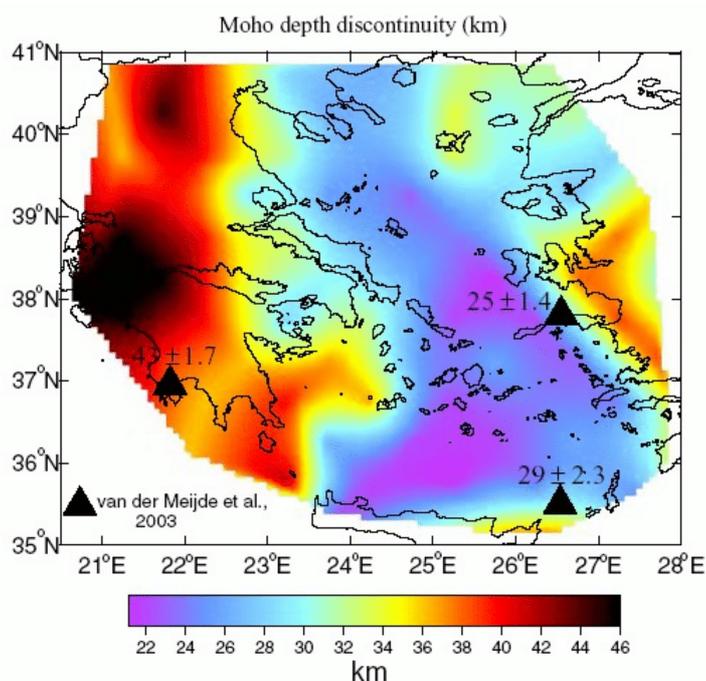


Fig. 2.9. Moho depth map estimated from dispersion of Rayleigh wave group velocities by Karagianni et al. (2005).

They found different crustal thicknesses in the Aegean Sea and Greece, which are in agreement with the general characteristics estimated from previous works in this area. According to their works, a thick crust (40-45 km) has been found for the western Greece and the Peloponnesus along the Dinarides-Hellenides mountain range, while in

Chapter 2. Geology and tectonics of the Aegean

Tabel 2.1. Moho depths obtained from previous works in the Aegean region.

Authors	Method	Crete	Cretan Sea	Cyclades	North Aegean	Greece	Hellenides
Papazachos (1966)	seismic wave velocity	-	-	20-35	33	-	43
Makris (1975)(1976)(1977)	deep seismic sounding and gravimetry	30-32	20	28	-	30-32	46
Makris and Vees (1977)	refraction	30-32	20	26	-	-	-
Calcagnile (1982)	surface wave dispersion	-	35	-	35	-	36-42
Panagiotopoulos et al. (1985)	earthquake seismology	-	24-30	-	-	-	30-35
Papazachos et al. (1995)	seismic tomography	30-40	25	25-30	25-30	-	35-40
Tsokas and Hansen (1997)	gravity	28-30	26	19-32	26-28	-	40-49
Tiberi et al. (2001)	gravity	-	-	-	-	25	40
Bohnhoff et al. (2001)	reflection refraction	24-32.5	15	-	-	-	-
Vigner (2002)	reflection	-	-	26-25	25	-	-
Karagianni et al. (2002)	surface wave dispersion	-	22-24	-	32	-	42-46
van der Meijde et al. (2002)	receiver function	-	29	25	-	-	43
Li et al. (2003)	receiver function	31-39	-	25	-	-	-
Tirel et al. (2004)	gravity	28-31	23	25	24-26	-	-
Karagianni et al. (2005)	surface wave tomography	-	20-22	24-26	26-32	30-34	40-46

the eastern part of Greece crustal thickness has been estimated to be 30-34 km. In the southern Aegean Sea, the crust is approximately 20-30 km thick, whereas the northern Aegean Sea exhibits a crustal thickness of about 25-30 km. The complexity of the crustal structure beneath the island of Crete, probably due to its location in the forearc of the subduction zone has been shown by many studies (Makris, 1978; Knapmeyer, 2000; Bohnhoff et al., 2001; Li et al., 2003; Endrun et al., 2004). The average crustal thickness has been estimated to be 28-30 km. Li et al. (2003) applied P receiver function method for the stations on the island of Crete and observed significant converted phase with a negative amplitude for most stations. It has been interpreted as a reversed Moho velocity contrast, caused by the hydration and serpentinization of

mantle mineral in the forearc.

2.4.2 The oceanic Moho of the subducting African plate

The dipping high-velocity zone was recognized by the early work of Spakman (1985, 1986) as the blurred image of the African slab that penetrates deep into the Aegean upper Mantle.

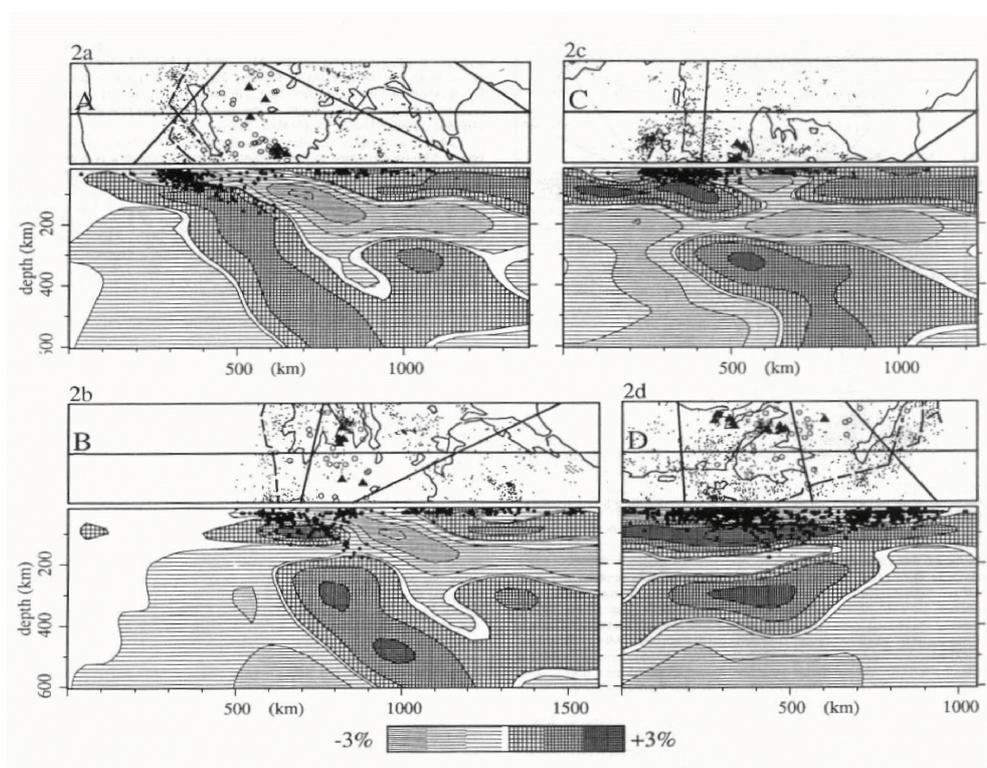


Fig. 2.10. Results of the early tomographic study in the Aegean by Spakman et al. (1988) show reliably the existence of the subducted African plate beneath Aegean to depths of at least 600 km.

Tomographic images of the Hellenic subduction zone by Spakman et al. (1988) demonstrated slab penetration in the Aegean upper mantle to depths of at least 600 km (Fig. 2.10), while the seismicity shows a maximum depth of 150-180 km. They imaged the descending slab up to northern Greece. Their results indicated a continuous slab beneath Crete. Although under Greece, on the western segment of the

Chapter 2. Geology and tectonics of the Aegean

subduction (approximately section of the arc which is located NW of the Kefallonia island-SW Peloponessus area), it seems to be intersected by a low velocity zone at depths around 200 km. They interpreted this zone as a level where the detachment of the lower part of the slab occurred. Papazachos et al. (1995) inverted residuals of the first P arrivals from earthquake across Aegean and showed that the velocity anomaly in the mantle under the southern Aegean extends much further and deeper to the northeast than the Benioff zone of the intermediate depth earthquakes indicates. However, Bijwaard & Spakman (1998) could image the oceanic African lithosphere up to depths of 1200 km beneath Greece. A large number of regional- or local-scale tomographic results have been also presented for several sub-regions or the broader Aegean area (e.g. Martin, 1988; Christodoulou and Hatzfeld, 1988; Drakatos, 1989; Ligdas et al., 1990; Drakatos and Drakopoulos, 1991; Ligdas and Lees, 1993; Papazachos et al., 1995; Papazachos & Nolet 1997).

Recently, Papazachos and Nolet (1997) presented new tomographic results for the P and S-wave velocity distribution in the Aegean area using non-linear inversion of travel times through 3-D ray tracing, which allowed the detailed description of the regional 3-D structure of the crust and uppermost mantle (up to ~160km).

Figure 2.11 shows the spatial distribution of P-velocities along three cross-sections, where strong crustal thickness variations and mid-crustal low velocity layers are found. Moreover, the signature of the relatively thin subducted slab in the upper mantle, as well as its detailed geometric characteristics are clearly identified. The subducted African lithosphere resolved from seismicity (Makropoulos & Burton, 1984; Papazachos, 1990; Hatzfeld & Martin, 1992) shows an unusual shape with an increasing dip from west to the east. The derived Benioff zone, which was identified through the accurate location of intermediate-depth ($60\text{km} \leq h \leq 180\text{km}$) earthquakes, defines the boundary between the Eurasian and African lithospheric plates along the Hellenic arc.

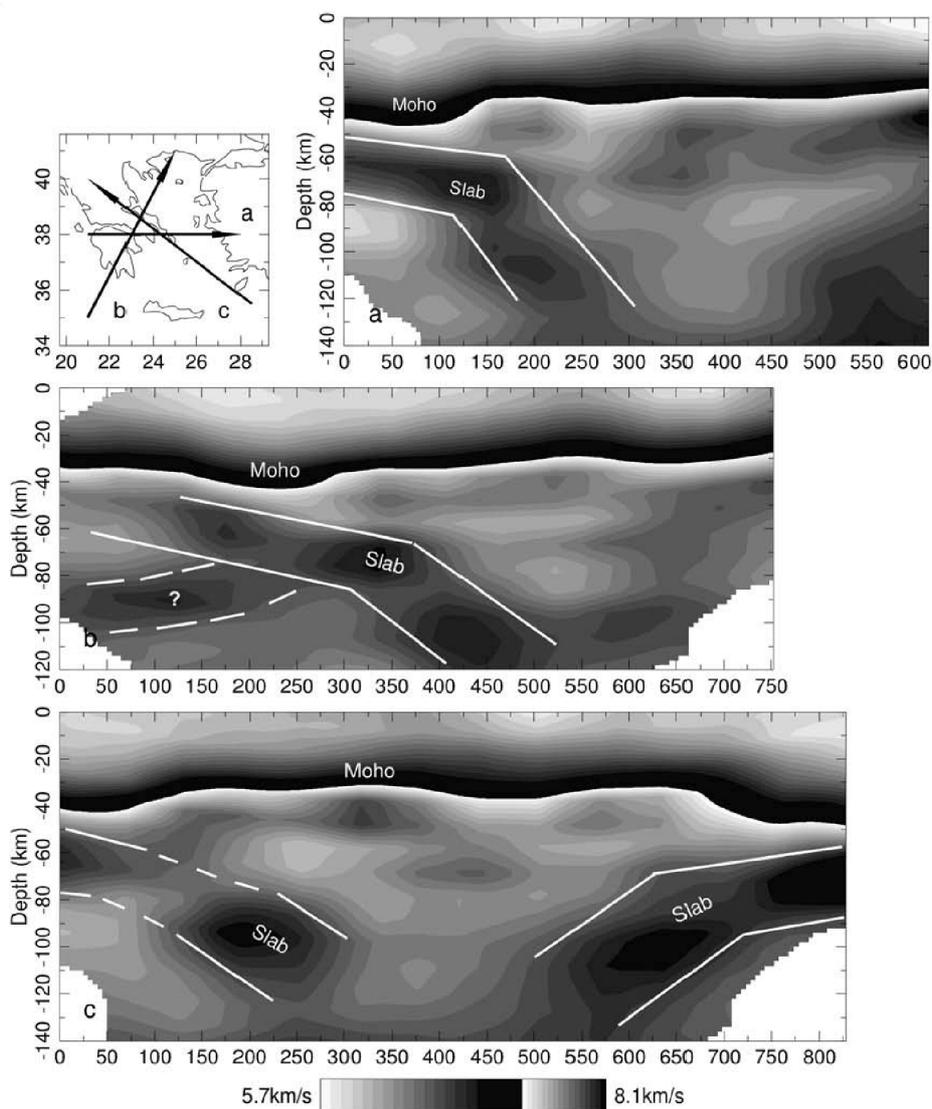


Fig. 2.11. Tomographic images of the subducted African plate along three profiles in the Aegean by Papazachos & Nolet (1997).

It has been shown that the Benioff zone consists of the shallow part ($h < 100 \text{ km}$), which dips at a low angle where earthquakes with magnitudes up to 8.2 occur and of the deeper part ($100 \text{ km} \leq h \leq 180 \text{ km}$), which dips at a larger angle where earthquakes with magnitudes up to 7.0 occur (Papazachos, 1990; Kiratzi and Papazachos, 1995; Papazachos, and Nolet, 1997; Papazachos et al., 2000). Figure 2.12 shows a map with well-located epicentres of shallow and intermediate-depth earthquakes in the southern Aegean area (Papazachos et al., 2000) as well as three cross-sections where the foci of

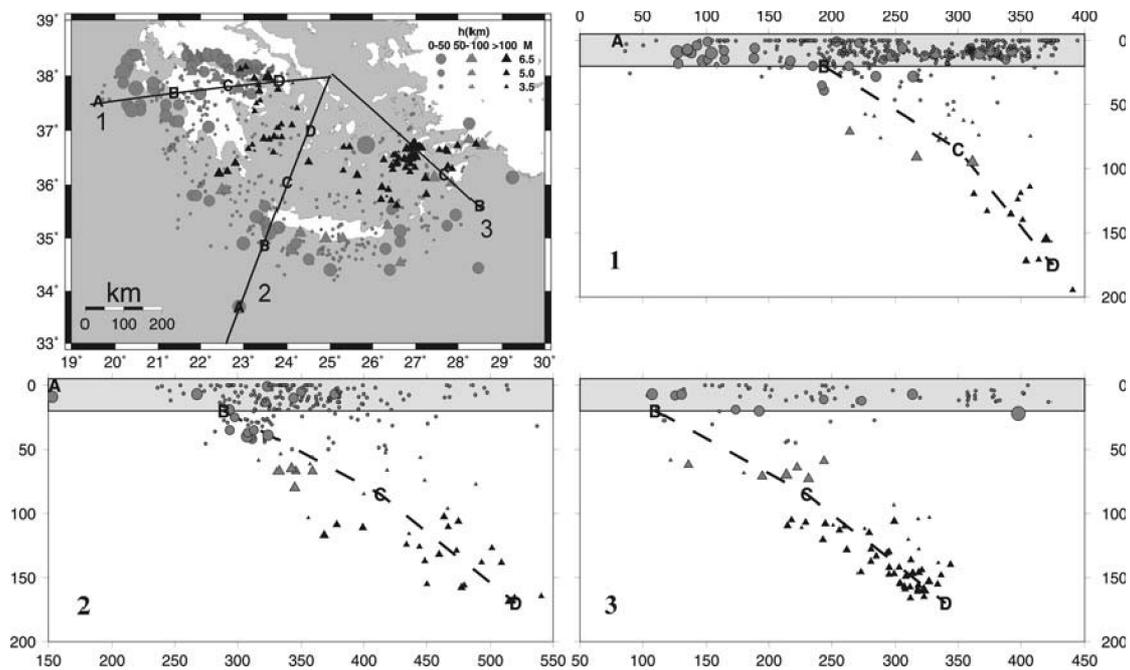


Fig. 2.12. Maps of accurate located epicenters of earthquakes in the southern Aegean Sea along three profiles shown on the top-left map by Papazachos et al. (2000).

these shocks have been projected on vertical planes (along the lines shown on the map). Receiver function analysis of the lithospheric structure in the Aegean area and Greece are more limited. The early work of Knapmeyer (2000) using P receiver functions in western Crete showed a downgoing oceanic Moho at a depth of 40-70 km, while Li et al. (2003) mapped it down to a depth of 100 km beneath Santorini island in the volcanic arc.