VELOCITY, DENSITY AND ATTENUATION STRUCTURE OF THE LITHOSPHERE IN THE AEGEAN AND SURROUNDING AREA.

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A B S T R A C T

In the present paper all the available information is presented concerning the velocity, density and attenuation structure of the lithosphere in the Aegean and surrounding area. Reflection and refraction methods and travel time inversion techniques have been applied in the last twenty years for the estimation of the velocity structure. All this information is used in order to construct a map of the Moho discontinuity in the region. The studies concerning the gravity and magnetic fields are reviewed and the most updated information is presented concerning the density structure. Finally, the main results of the studies conducted mainly during the last five years for the Q estimation (attenuation structure) are presented and the dependence of Q on distance is examined.

ΔΟΜΗ ΤΗΣ ΛΙΘΟΣΦΑΙΡΑΣ ΣΤΟ ΧΩΡΟ ΤΟΥ ΑΙΓΑΙΟΥ ΚΑΙ ΤΙΣ ΓΥΡΩ ΠΕΡΙΟΧΕΣ ΜΕ ΒΑΣΗ ΤΙΣ ΤΑΧΥΤΗΤΕΣ ΤΩΝ ΣΕΙΣΜΙΚΩΝ ΚΥΜΑΤΩΝ, ΤΗΝ ΚΑΤΑΝΟΜΗ ΤΗΣ ΠΥΚΝΟΤΗΤΑΣ ΚΑΙ ΤΗΝ ΑΠΟΣΒΕΣΗ ΤΩΝ ΣΕΙΣΜΙΚΩΝ ΚΥΜΑΤΩΝ.

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Π Ε Ρ Ι Λ Η Ψ Η

Στην παρούσα εργασία δίνονται όλες οι πληροφορίες που αφορούν τη δομή της λιθόσφαιρας στο χώρο του Αιγαίου και τις γύρω περιοχές με βάση την κατανομή των ταχυτήτων των σεισμικών κυμάτων, την κατανομή της πυκνότητας και την απόσβεση των σεισμικών κυμάτων. Για τη μελέτη της δομής με βάση τις ταχύτητες των σεισμικών κυμάτων έγιναν έρευνες κατά τα τελευταία είκοσι κυρίως χρόνια εφαρμόζοντας μεθόδους ανάκλασης και διάθλασης καθώς και τεχνικές αντιστροφής των χρόνων διαδρομής των σεισμικών κυμάτων. Ολες οι πληροφορίες που προέρχονται από τις έρευνες αυτές χρησιμοποιούνται για την κατασκευή του χάρτη των βαθών της ασυνέχειας Moho στην περιοχή αυτή. Γίνεται επίσης ανασκόπηση των ερευνών που αφορούν στο μαγνητικό και στο βαρυτικό πεδίο και αναφέρονται συμπεράσματα για την κατανομή της πυκνότητας. Τέλος συνοψίζοντας τα κύρια αποτελέσματα των ερευνών που έγιναν κατά τα τελευταία πέντε κυρίως χρόνια για τον υπολογισμό του παράγοντα απόσβεσης, Q, των σεισμικών κυμάτων και παρουσιάζεται χάρτης χωρικής κατανομής των τιμών του παράγοντα αυτού στην ευρύτερη περιοχή του Αιγαίου.

INTRODUCTION

The structure of the Earth's interior is the distribution of specific physical quantities. These quantities characterize the natural attributes of the materials which comprise the unities of the Earth's interior. In Geophysics the quantities mostly studied are the density and the elastic constants of the materials of the interior of the Earth.

Geophysics mainly studies the vertical variation of physical constants, in order to give solutions to the problems of the structure. This study of the vertical variation can also give valuable information about the lateral heterogeneities.

The study of the structure contributes to the increase of our knowledge about the condition of matter in different depths of the Earth's interior. This is due to the fact that the values of elastic constants depend on the condition of the matter.

It is known that even using contemporary technology we are not able to have rock samples from depths greater than 10-12 km and we cannot study through them the conditions existing in those depths (pressure, temperature, density, conductivity etc.)

Thus, the study of the Earth's structure was based on indirect observation and on the understanding the way in which the shear and longitudinal waves propagate through the unities which comprise the Earth's interior.

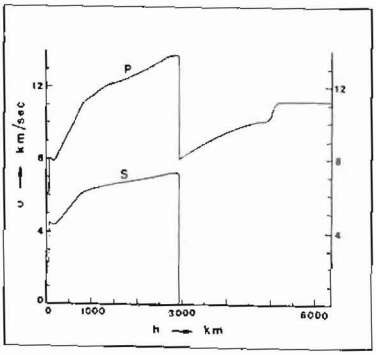

Fig.1. The broad velocity-depth distribution in the earth as deduced by Gutenberg (1959).

Figure (1) depicts the variation of the propagation velocity of the longitudinal waves P and shear waves S, in relation to the depth of the Earth's interior, according to Gutenberg (1959). In fig. (1) it is obvious that the velocity of the seismic waves increases directly with the depth. In fig. (1) we can also see that, there are intervals with normal variation of velocity and intervals with sudden variation corresponding in both cases to specific depths of the Earth's interior. This sudden variation of the velocity is called discontinuity of the velocity and corresponds to a geophysical limit separating two unities of the Earth's interior, which have different structure.

The Earth's interior is divided into three basic unities. These unities are separated by two main discontinuities of the seismic waves velocity and are called the crust, the mantle and the core of the Earth. The first of the two discontinuities, which separates the crust from the mantle is called Mohorovicic discontinuity or Moho or discontinuity "M". This discontinuity has been observed at a mean depth of about 35 km.

The second discontinuity is the geophysical limit between the mantle and the core of the Earth and is called discontinuity Gutenberg. This discontinuity was observed at a depth of 2900 km. Both Moho and Gutenberg are called first order discontinuities. All other discontinuities which divide the three main unities into smaller unities are called second order discontinuities. The second order discontinuities are: 1) discontinuity Conrand, which divides the crust into the granitic and the basaltic layer 2) discontinuity Repetti which divides the mantle into the upper and lower mantle and 3) discontinuity Lehman which divides the core into the external and the internal core.

The development of DSS (Deep Seismic Soundings) projects during the decade of 60's and the relevant results of Soviet scientists as well as the continuation of research during the decade of 70's did not change the scientific point of views concerning the granitic and basaltic layers. If anything they confirm it.

THE LITHOSPHERE IN SE EUROPE

Several models have been proposed for the explanation of the lithoshperic structure in SE Europe. All these models and proposals use as a reference point the subduction of the African plate under the Aegean and the associated phenomena (Papazachos and Comninakis, 1969, 1971; Papazachos and Delibasis, 1969; McKenzie, 1970, 1972, 1978; Lort, 1971; Economides, 1972; Agarval et al., 1976; Papazachos, 1976; Gregersen, 1977; Le Pichon and Angelier, 1979; Dewey and Sengor, 1979; Makropoulos and Burton, 1984; Meulenkamp et al., 1988; Wortel et al, 1990).

The Aegean is considered to be a marginal sea with high volcanic activity (Georgalas, 1962), high heat flow (Jongsma, 1974), magnetic anomalies (Voght and Higgs, 1969), positive isostatic anomalies (Fleischer, 1964; Makris, 1976), high surface seismicity (Galanopoulos, 1963; Comninakis, 1975; Papazachos, 1990) with mainly normal faulting (McKenzie, 1978) and strong attenuation of the seismic energy towards the inner (concave) part of the Hellenic arc (Papazachos and Comninakis, 1971; Delibasis, 1982; Tassos, 1984). Other researchers have proposed different (e.g. Makris, 1976) or complementary (Taymaz et al., 1991) causes, beside the subduction, for the explanation of the geophysical properties of southern Aegean.

For the northern Aegean, the thin (25-30Km) crust (Panagiotopoulos, 1984; Brooks and Kiriakidis, 1986; Chailas et al., 1992), in connection with recent tomographic images (Babuska et al., 1987; Spakman, 1986, Spakman et al., 1992), other geophysical studies (Spasov and Botev, 1987; Shanov et al., 1992) and geochemical data (Soldatos and Christofidis, 1986), suggests the existence of a paleosubduction in this area.

The average thickness of the lithosphere in SE Europe has been calculated to be around 80-90Km (Papazachos et al., 1967; Papazachos, 1969; Papazachos and Comninakis, 1971; Mayer-Rossa and Mueller, 1973). There seems to be a thickening of the lithosphere from 30Km in the area of Crete to 120Km in the area of northern Greece (Calganile et al., 1982). Further north, the lithosphere is probably reaching a thickness of 140-150Km (southern Bulgaria) but it is then thinning to 100-120Km towards northern Bulgaria and Black Sea (Babuska et al., 1986; Shanov et al., 1989) and to 90Km to 110Km towards Ionian and Adriatic Sea (Calganile et al., 1982). Some researches suggest that the lithosphere-asthenosphere discontinuity is identified by the existence of a low-velocity (LVL) layer in S (Papazachos , 1969) or P (Mayer-Rossa and Mueller, 1973) waves. Another LVL possibly exists deeper in the asthenosphere, below 150-160Km (Mayer-Rosa, 1974; Martin, 1988).

Tomographic images using the inversion method (Spakman, 1986, 1988; Spakman et al., 1988) suggest that the dipping lithosphere, in the southern Aegean, is appearing as a high velocity zone, dipping as far as 800Km in the low velocity material of the asthenosphere. This image of the subducting lithosphere is confirmed by other studies using the same (Drakatos, 1989; Ligdas et al., 1990) or different methods (Gregersen and Jaeger, 1984; Hashida et al, 1988; Tassos et al., 1989).

VELOCITY STRUCTURE

The crust in SE Europe

The crust in this area is strongly affected by the tectonic processes. The first systematic study of the crust was that of Papazachos et al.(1966) where an average crustal thickness of 42Km and a P_n velocity of 7.87Km/sec were determined using direct and refracted waves. Grigorova and Sokerova (1967) found thicknesses between 35 and 51Km in two cross-sections in Bulgaria and northern Greece using the same kind of data and Papazachos et al. (1967) and Papazachos (1969) determined crustal thickness between 35 and 45Km for SE Europe using surface waves dispersion data. Similar average thickness (34-39Km) was determined for Greece and the Aegean area, using the same kind of data, by Calganile et al.(1982).

Makris, in a series of papers (Makris, 1972, 1973, 1975, 1976, 1977, 1978a,b; Makris and Moller, 1977; Makris and Vees, 1977; Makris et al., 1977), presented the results of his researches, using travel times from explosions. He calculated average crustal thicknesses of 30 and 44Km for the Aegean and the Hellinides mountain range, respectively and found a thinning of the crust around 20-25Km in the Cretan Sea (fig. 2). He also presented the first Moho-depth map of Greece using a joint interpretation of velocity and gravity data. Later studies using the same methodology (Delibasis et al., 1988; Voulgaris, 1991) found similar results.

Panagiotopoulos (1984) and Panagiotopoulos and Papazachos (1985) determined the crustal thickness under 50 seismological stations of SE Europe. They found thicknesses of 38-45Km for

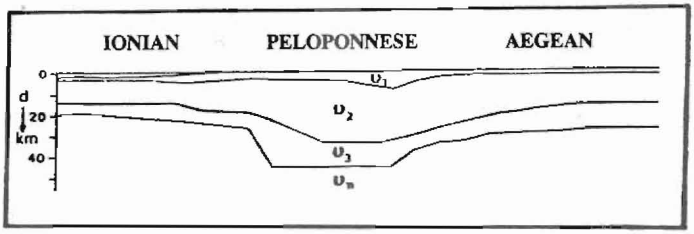

Fig.2. Schematic crustal section across the Greek area. The velocities of P-waves in the sedimentary, the granitic and the basaltic layers, and below the Moho are υ_1=2.5-5.3Km/sec, υ_2=5.4-6.0Km/sec, υ_3=6.2-7.1Km/sec and υ_n=7.7-7.9Km/sec, respectively (Makris, 1977).

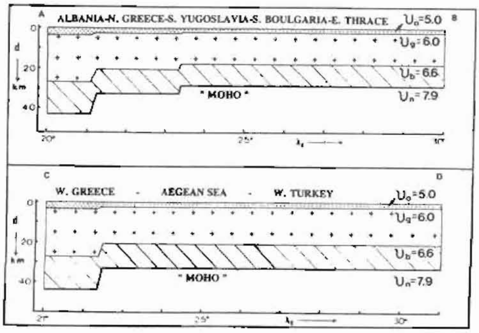

Fig.3. Schematic crustal sections across the E-W direction in northern part (A-B) and southern part (C-D), of Southern Balkan region. The velocities of P-waves υ_o, υ_g, υ_b, υ_n, correspond to the sedimentary, granitic and basaltic layers and below the Moho (Panagiotopoulos, 1984).

western Greece and Albania, 30-35Km for southern Aegean, central Greece, southern Yugoslavia and Minor Asia and 24-30Km for the Rhodope massif (fig. 3). The average crustal thickness they determined was around 31Km.

Drakatos (1989) and Drakatos and Drakopoulos (1991) used the inversion method for the determination of the 3D structure in the Aegean and surrounding areas. They found that low velocities dominate central Greece and Minor Asia, whereas high ones exist in southern Aegean and the Ionian Sea. The same method was applied by Drakatos (1987) and Drakatos et al. (1989) for the central part of northern Greece, Drakatos et al.(1988) for the Axios basin, and Martin (1988) for Peloponnese. These studies revealed a strong correlation of the local surface geology with the velocity anomalies of the upper layers (first 15Km), resulting in low velocities for the tectonically active areas and the sedimentary basins. The latter researcher found an average crustal thickness for the Peloponnese area equal to 40 km.

Important studies have been performed in the eastern Mediterranean and in the Ionian basin. The average crustal thickness is around 20Km (Ewing and Ewing, 1959; Hinz, 1974; Payo 1967, 1969;Lort et al., 1974; Comninakis and Papazachos, 1976;

Makris et al., 1983). The crust is thickening towards the northern part of the Ionian basin and the Adriatic Sea to 22-28Km (Dragasevic, 1974; Locardi and Nicolich, 1988). In the southern Dinarides (eastern coasts of south Yugoslavia and Albania), it is around 43-48Km, reaching its peak value in north-eastern Albania (Dragasevic, 1974; Calganile et al., 1982).

For the area of Bulgaria, Shanov and Kostadinov (1992) used all the previously published data (DSS, seismological, gravity, etc.) and presented the structure of the crust and the upper mantle in the area. According to these researchers, the crust reaches a maximum thickness of 50Km in the south-western Bulgaria but it is constantly thinning mainly towards the Black Sea up to 28-30Km.

In figure (4), a map is presented for the crustal thickness where all the available velocity-structure data have been used (Papazachos, 1993), including the results of some local studies, not mentioned previously (Volvovsky et al., 1991; Ezen, 1991a,b), as well as results from other kind of geophysical data (gravity etc.). We observe that, in agreement with the theory of isostasy, we have a thick crust (40-45Km) under the southern Dinarides and the Hellinides up to the central Peloponnese. A thinning of the crust (20-25Km) is observed in the southern Aegean (Cretan Sea) probably caused by the rise of hot material due to the existing subduction. A similar thinning is observed above the neighbouring Tyrrenean Benioff zone where the crust is thinning localy up to 10Km (Nicolich, 1988; Locardi and Nicolich, 1988). A smaller thinning (25Km) is observed along the north Aegean trough, as also confirmed by gravity data (Brooks and Kiriakidis, 1986). The average thickness for the rest Aegean area is around 30Km. Towards the inner part of the Balkan peninsula the crust is around 35-40Km thick. On the other hand, in the area of the Ionian Sea the crust is quite thin (20Km).

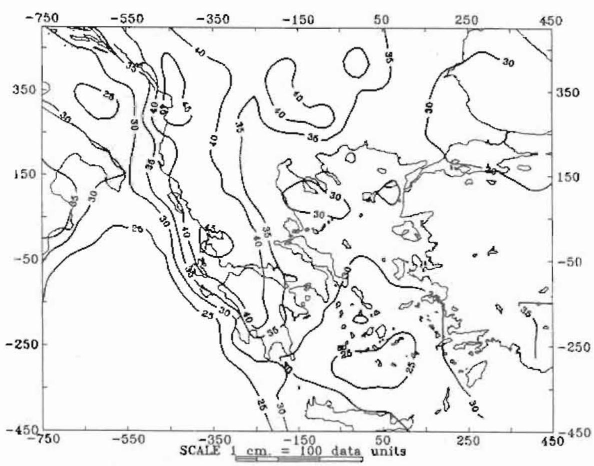

Fig.4. Moho depth map of Greece and surrounding areas (Papazachos 1993).

The Conrad discontinuity seems to be present throughout the whole SE Europe. The available data show that it exhibits the same behaviour as the Moho. The thickness of the overlying granitic layer is approximately 1.5 to 2 times the thickness of the underlying basaltic layer (fig. 2,3) In some cases, the velocity above and below the Conrad is practically the same and therefore it could be ommited without affecting the results of the corresponding studies (e.g. Grigorova and Sokerova, 1967). Also, in some cases a gradual velocity increase without a clear Conrad discontinuity can interpret the travel-time data. However, the clear P_b arrivals in the Hellinides (Papazachos et al., 1966; Makris, 1977) show the presence of the Conrad. The situation is different in the crust of Eastern Mediterranean where there is weak evidence about the existence of a thin (3Km) granitic layer (Moskalenko, 1966; Finetti and Morelli, 1973; Makris et al., 1983) but its existence quite probable only in the Ionian basin (Weigel and Hinz, 1970; Hinz, 1974). This layer is explained by the previous researchers as the extention of the respective layer of the African lithosphere.

The P_n velocity for the whole area seems to be constantly around 7.7-8.0Km/sec (Papazachos et al., 1967; Makris, 1977, Panagiotopoulos, 1984). This value is increased in the Ionian basin up to 8.1Km/sec (Ewing and Ewing, 1959; Hinz, 1974; Makris, et al. 1983) and it is keeping this higher value in the eastern part of southern Italy (Steinmetz et al., 1983). The granitic layer is also showing a consistency about its P_g velocity which is around 6.0-6.2Km/sec (Makris, 1977; Panagiotopoulos, 1984) but lower and higher values have been reported (e.g. Grigorova and Sokerova, 1967). Sonic Logs of deep drillings in north-eastern Greece have shown that the velocity in the metamorphic basement, which is considered to be the top of the granitic layer, is quite constant around 6.0-6.3 Km/sec although this basement is strongly differenciated (Sousounis, 1993). This value increases with depth up to 6.4-6.6Km/sec at the Conrad discontinuity (Makris, 1977; Volvovsky et al., 1991). On the other hand the situation is quite different for the basaltic layer. The P_b velocity shows a large variation with an average value around 6.7-6.9Km/sec. For the top of this layer, the velocity is quite low (e.g. 6.4Km/sec, Makris, 1977) but values as high as 7.1Km/sec (Volvovsky et al., 1991) have been reported. In its deeper parts it approximates the P_n velocity (e.g. Makris, 1977; Voulgaris, 1991). The main difficulty with the P_b velocity is that the corresponding waves are observed as first arrivals only for a small distance range around 120-140Km or not observed at all (Panagiotopoulos, 1984).

Very little information exists for S wave velocities or densities. Usually they are calculated from P velocities using a V_p/V_s ratio and Nafe-Drake's or Birch's law. The average S velocities for the whole SE Europe are 3.3-3.4Km/sec, 3.55-3/75Km/sec and 4.3-4.4Km/sec for the granitic, basaltic and upper mantle layers, respectively (Papazachos, 1969; Martin, 1988; Ezen, 1991a,b). The corresponding densities are about 2.7-2.82gr/cm^3, 2.9-3.0gr/cm^3 and 3.30-3.37gr/cm^3 (Papazachos, 1969; Makris, 1977).

There are indications that there is a low velocity crustal layer, at least locally, in SE Europe. Scordilis (1985) found for the Serbomacedonian massif an increase of the Poisson ratio to

1.78 (from 1.75 of the overlying and underlying layers) for seismic waves arriving at distances between 70 and 100Km (depths of 10-15Km) which he interpreted as evidence for a possible low velocity layer in the crust of this area.

The sedimentary layer in Greece

The sedimentary layer in Greece shows a great complexity. Its peculiarity is due to the fact that it is locally confined and not globally present as the granitic and basaltic layers. The most important results are those from seismic prospecting due to their high resolution, especially when combined with deep drilling data.

For northern Aegean the most important results are those of Lalechos and Savoyat (1977), Lalechos (1986), Martin (1987) and Mascle and Martin (1990) who used seismic prospecting and deep drilling data. Their results, together with some other local studies, wich were based on a variety of geophysical methods (Memou, 1983; Thanassoulas et al., 1987; Kiriakidis et al., 1988; Kiriakidis, 1989; Maltezou et al., 1990; Loukoyiannakis et al., 1990; Sousounis, 1993; Roussos, 1993, pers.comm.) show that the average thickness of the Neogene sedimentary layer is quite high (around 2-3Km). Below the Axios basin, the Thermaikos gulf and the Limnos area there are thick Paleogene sediments. Locally (Prinos-Nestos basin, Orestias basin) the sediments are thicker than 4Km. The average velocities are 2.4-2.8Km/sec for the Neogene and 3.5-3.8Km/sec for the Paleogene sediments reaching 4.1Km/sec in its deeper parts (Martin, 1987; Roussos, 1993, pers.comm.).

For the area of central and southern Aegean, important are the studies of Martin (1987) and Mascle and Martin (1990). Using her seismic profiles and the corresponding velocities for the Neogene sediments, which dominate in the area, we calculated an average thickness of 1.5-2.5Km for the sedimentary layer. However, this layer is locally confined in various basins. The same researcher calculated the depth of sediments in various basins in Cyclades islands and found mean values around 1.5Km. In Myrtoon Sea the sediments reach a maximum depth of 3-4Km (Publ.Petr.Comp., 1993, pers.comm.)

The structure of the sedimentary layer in the outer sedimentary arc is very different. The main study for the structure of this layer in western Greece was made by Monopolis and Bruneton (1982) who used off-shore seismic profiles and deep drillings and studied the whole area from Corfu island up to Kiparissia gulf. Other studies in the same area (Nikolaou, 1986; Kamberis, 1987; Voulgaris, 1991) as well as recent seismic prospecting results (Marnellis, 1993, pers.comm.) completed an initial image for this area. It seems that the Plioquaternary sediments have low velocities and densities (2-3.9Km/sec and 2-2.6gr/cm^3) and correspond to the sedimentary layer that other seismological researches found in the area (Makris, 1977; Delibasis et al., 1988). Below this formation the Eocene-Jurassic limestones have a 3-6Km thicknes and a high velocity (around 6.2Km/sec) and therefore they are included in the granitic layer. Under the limestones, a 1-5Km thick Triassic evapotitic layer exists which is identified as a LVL (e.g. Voulgaris, 1991) due to its small velocity (5.2-5.5Km/sec, Marnellis, 1993, pers.comm.). The sedimentary layer, west of the Ionian islands, is generally around 2-4Km and reaches its peak value, around 5-7Km, between the Ionian islands and the Greek mainland although locally it is quite thinner. Further inland the sediments seem to have a smaller thickness but almost no seismic prospecting data are available due to the lack of hydrocarbon interest.

This structure continues along the whole outer Greek arc (Rousos, 1993, pers.comm.). Makris (1977) found a 1 to 1.5Km thickness for the sediments on Crete (fig. 2). He also observed a thickness of the sediments up to 3Km just above the crustal thinning in the Cretean Sea, as was also confirmed by airgun measurements (Jongsma, 1974). For the rest parts of Greece little information exist about the sedimentary layer. In the Thessaly basin the maximum thickness is around 750-1000m (Fytikas, 1993, pers.comm.). Seismic prospecting studies in Argos-Orestiko area (Publ. Petr. Comp., 1993, pers.comm.) show a sediments' depth varying from 500 to 2000m.

The Eastern Mediterranean is characterised by an extra thick sedimentary layer. The unconsolidated sediments have a thickness of 500-1500m (Gaskell et al., 1958; Moskalenko, 1966; Wong and Zarudski, 1969; Lort et al., 1974; Hinz, 1974; Comninakis and Papazachos, 1976; Makris et al., 1983). Below this layer, a layer with a varying thickness (1-9Km) and P velocity around 3.5-4.0Km/sec (Hinz, 1974; Lort et al., 1974) appears, followed by another layer with P velocity of 4.4-5.2Km/sec and an average thickness of 4-5Km. The mean thickness for the whole sedimentary layer is 7 to 12Km.

DENSITY STRUCTURE

Results of gravimetric campaigns in the major Greek territory have been reported by many researchers. The oldest one is due to Vening Meinesz (1932). All these papers are refered in the review of Papazachos et al. (1987).

With respect to morphology and Bouguer anomaly, Rabinowitz and Ryan (1970) were the first to observe that an excellent resemblance occurs between the Hellenic trough and others occurred elsewhere all over the world. Allan and Morelli (1971), Makris (1972, 1973, 1977 and 1978a,b), Finetti and Morelli (1973), Morelli et al., (1975) and Makris and Moller (1977) conducted systematic surveys offshore and onshore. These results were used by the I.G.M.E. and the University of Hamburg for the compilation of the first map of Bouguer anomaly in Greece. The free air and Bouguer maps in Eastern Mediterranean were also constructed.

The lithosphere subducted beneath the Aegean was modelled by Gregersen and Jaeger (1984) who also tried to fit the produced anomaly to the observed one. Several small scale gravimetric surveys were conducted by I.G.M.E. and the Public Petroleum Company of Greece for various purposes (oil and gas exploration, prospecting for hydrothermal fluids, etc.).

Such kind of surveys were carried out in Grevena and Kastoria (Papanikolaou, 1961), in the basin of Thessaloniki (Stavrou 1961), the basin of Nestos river (Stamou, 1977) and in Milos and Kimolos (Thanassoulas, 1983a; Tsokas, 1985). Kiriakidis

(1984) measured some profiles striking perpendicular to the ophiolitic belt of Axios basin.

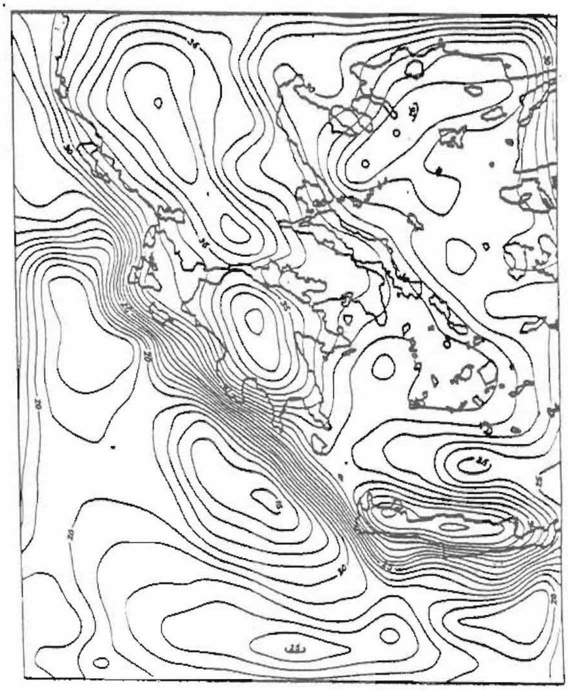

Fig.5. Map showing the variation of the Moho depth (in Km), (Chailas et al., 1992).

Using the existing gravity data along with others (aeromagnetic, seismic profiles, borehole logs, etc.), Loukoyannakis (1981) and Thanassoulas (1983b) studied the structure of Nestos and Anthemounta basin, respectively. Along the same baselines, Memou (1983), Maltezou (1986) and Thanassoulas (1979) studied the Strymon basin and the Rhodope massif, respectively. Thanassoulas et al. (1987), Thanassoulas and Tsokas (1987), Kiriakidis et. al. (1988) constructed the model of the Nestos sedimentary basin. Also Kiriakidis (1989) studied the ophiolitic belt of Axios basin and constructed models of that basin which were deduced from gravity data.

On the basis of spectral analysis of the Bouguer anomaly data Tselentis et al. (1988a) presented a calculation of Moho depth in Epirus. All the known gravimetric data either on-shore or off-shore were combined for the compilation of a new Bouguer anomaly map of the Hellenic peninsula (Lagios et al., 1988). These data were used by Chailas et al. (1992) in order to calculate the Moho depths employing isostatic concepts, shown in figure 5. As it can be seen from that figure, thick crust of up to 40 Km exists along the mainland of Greece up to Peloponnese and in Creete island, while the crust is thin at south Aegean, at the northern Aegean trough and at the eastern Mediterranean. All the other region is characterized by normal crust with mean thickness of about 30 Km.

ATTENUATION STRUCTURE

The attenuation of the seismic waves is a physical property of great importance closely related to the deep tectonics of an area. Therefore the studies on the attenuation structure of the crust and upper mantle have been intensified in the last years. One of the most important parameters in describing the attenuation is the seismic quality factor, Q , which has low values when the attenuation is high and large values when the attenuation is low.

The attenuation structure of the Aegean and the surrounding area is rather complicated, as is expected for an area with active tectonics where a subduction is taking place. The first observation concerning the attenuation characteristics of the area was made by Papazachos and Comninakis (1971) who compared the seismograms of an intermediate depth earthquake occurred in the southern Aegean and recorded at the stations PRK and VAM, northwards and southwards from the epicenter, respectively. The eathquake was recorded at PRK with much smaller amplitude than at VAM and they attributed this difference to the existence of a highly attenuative upper mantle at the southern Aegean region, a result of the subduction of the African plate under the Aegean lithosphere. The same interpretation was also given by Agarwal et al. (1976) and Gregersen (1977) who used travel time residuals to study the deep structure of the southern Aegean and by Delibasis (1982) who studied the attenuation of the seismic waves in the southern Aegean by comparing recorded amplitudes at different seismological stations for seismic rays passing through the S. Aegean.

Studies concerning the estimation of the quality factor, Q , in the Aegean and surrounding area have been made mainly in the last five years by using various methods. Hashida et al (1988) by inversion of seismic intensity data studied the three dimensional attenuation structure of the area. Assuming that the representative frequency of the seismic waves causing the damages is around 1 Hz, they found Q_s values ranging from 60 to about 800 for the crust (upper 40 Km) and they observed that low Q is dominant in the Aegean sea while high Q is dominant in the surrounding land areas.In the upper mantle the high Q (Q around 1000) zones along the Hellenic arc correspond to the subducting African plate and the low Q areas (Q around 40) are in the inner part of the arc. Kovachev et al. (1991) by using OBS data and comparing theoretical S wave amplitude curves versus distance with the observed ones found Q_s values between 200 to 300 for frequency of 8 Hz. The earthquakes occurred at the SE part of the Hellenic arc (between western Crete and Rhodos island) and most of the had focal depths between 1 and 25 Km while a few had depths between 25 and 200 Km, with M_L magnitudes between 1 and 4 and epicentral distances of 10 to 300 Km. Therefore the calculated Q_s values represent an average of the crust and the uppermost part of the mantle for that region.

Table 1. Information on the date, time , geographic coordinates of the epicenter, focal depth, h, and the magnitude, M_L, of the earthquakes used for the estimation of Q_s using acceleration data, the code of the accelerograph station, the epicentral distance, D, and the estimated Q_s value at a mean frequency of 5Hz (Hatzidimitriou et al., 1993).

No	Date	Time	Lat.	Lon.	h	M_L	Station	D	Q_s
1	20 Jun. 1978	20:03:24	40.71	23.27	10	5.9	THE78-1	30	85
2							GEV78-1	80	298
3	11 Aug. 1980	09:15:59	39.30	22.82	3	4.8	ALM80-4	15	87
4	26 Sep. 1980	04:19:18	30.24	22.74	2	4.4	ALM80-5	7	55
5	17 Jan. 1983	12:41:31	37.96	20.23	6	6.1	AGR83-1	128	233
6							LEF83-1	105	148
7							ARG83-1	33	121
8	17 Jan. 1983	16:53:30	38.11	20.37	8	4.9	ARG83-3	13	37
9	31 Jan. 1983	15:27:02	38.14	20.45	5	5.2	ARG83-6	6	33
10							ZAK83-1	56	136
11	20 Feb. 1983	12:42:29	37.76	21.11	2	4.9	ZAK83-2	19	49
12	16 Mar. 1983	21:19:39	38.80	20.88	1	4.9	LEF83-2	16	32
14	23 Mar. 1983	23:51:07	38.22	20.35	1	5.6	ZAK83-3	68	186
15							LEF83-4	75	125
16							ARG83-7	13	115
17	6 Aug. 1983	15:43:53	40.09	24.78	12	5.9	POL83-1	119	322
18							IER83-2	85	364
19	26 Aug. 1983	12:52:10	40.47	23.96	3	4.4	OUR83-2	16	82
20	19 Feb. 1984	03:47:22	40.61	23.40	8	4.2	POL84-1	25	204
21	4 Oct. 1984	10:15:12	37.64	20.85	14	4.5	ZAK84-1	18	117
22	25 Oct. 1984	09:49:16	36.83	21.72	11	4.7	KYP84-2	47	196
23							PEL84-1	27	188
24	22 Mar. 1985	20:37:39	38.98	21.11	2	4.0	AMF85-3	14	68
25	22 Mar. 1985	20:38:54	38.91	21.06	7	4.1	AMF85-4	12	116
26	9 Nov. 1985	23:30:43	41.26	24.02	12	5.0	DRA85-1	17	122
28	13 Sep. 1986	17:24:34	37.11	22.14	8	5.5	KAL86-1	9	37
29	15 Sep. 1986	11:41:30	37.04	22.13	8	4.9	KAL86-7	1	30
30							KAL86-2	1	34
31							MES86-1	10	40
32	21 Dec. 1990	06:57:22	40.93	22.37	10	5.1	EDE90-1	32	76
33							KIL90-1	42	178
34							ABS90-2	70	206

Table 2. Information on the calculated Q_o and n values of the relation $Q_c=Q_of^n$, where Q_c is the coda Q values and f is frequency for different lapse time windows.

Lapse time window	Q_o	n	Ref.
10 - 20 sec	35.7±8.30	0.96±0.10	1
15 - 30 sec	60.0±0.83	0.79±0.01	1
20 - 45 sec	89.1±2.67	0.72±0.02	1
30 - 60 sec	94.4±4.13	0.78±0.03	1
50 - 100 sec	128.5±5.09	0.74±0.02	1
-	73.4±34.8	0.95±0.11	2
-	43.0±5.0	0.81±0.17	3

References: 1. Hatzidimitriou, 1993; 2. Martin, 1988; 3. Tselentis et al., 1988b.

Table 3. Information on the date,the geographic coordinates of the epicenter, the magnitude, M_s, the mean distance of the macroseismic data from the epicenter and the estimated Q value at frequency of 1 Hz (Papazachos 1992).

No	Date	Lat.	Lon.	M_s	Dist.	Q
1	21 Oct 1953	38.60	20.96	6.3	113.9	234
2	19 Apr 1955	39.37	23.00	6.2	104.2	397
3	21 Apr 1955	39.31	23.12	5.8	105.5	269
4	16 Jul 1955	37.55	27.05	6.9	115.6	316
5	09 Jul 1956	36.64	25.96	7.5	153.8	152
6	25 Apr 1957	36.55	28.80	7.2	170.0	361
7	27 Aug 1958	37.40	21.00	6.4	126.7	553
8	14 May 1959	35.00	24.73	6.3	110.3	711
9	15 Nov 1959	37.78	20.53	6.8	176.7	397
10	10 Apr 1962	37.80	20.10	6.3	141.7	448
11	06 Jul 1962	37.81	20.20	6.1	102.9	359
12	29 Apr 1964	39.12	23.64	5.6	121.4	313
13	05 Apr 1965	37.48	21.88	6.1	117.7	326
14	09 Apr 1965	35.03	24.31	6.1	111.6	286
15	06 Jul 1965	38.27	22.30	6.3	107.8	405
16	20 Dec 1965	40.48	25.00	5.6	111.3	276
17	19 Feb 1968	39.50	25.00	7.1	224.9	228
18	29 Nov 1973	35.18	23.75	6.0	163.0	301
19	14 Jun 1979	38.74	26.50	5.9	100.9	649
20	09 Jul 1980	39.27	22.83	6.5	111.0	279
21	24 Feb 1981	38.07	23.00	6.7	100.9	227
22	18 Jan 1982	39.78	24.50	7.0	123.9	340
23	25 Mar 1986	38.38	25.13	5.7	110.9	781

Papazachos (1992) by using a large amount of macroseismic data and taking into account the anisotropic radiation pattern of each earthquake calculated the attenuation coefficient for 92 shallow earthquakes occurred in the Aegean and the surrounding area from 1912 to 1989. By assuming that the dominant frequency of the S waves is around 1 Hz an average Q_s for the upper 20 Km of the crust equal to 350 $\pm$ 140 has been estimated.

Hatzidimitriou et al. (1993) using strong motion accelerograms from 21 earthquakes occurred between 1978 and 1990 in the area of Greece, calculated Q_s values by the spectral decay of the log spectrum of the accelerograms for frequencies between 4 and 10 Hz and epicentral distances between 1 and 128 Km. They interpreted the calculated Q_s values as due to a highly attenuative surface sedimentary layer with 3 Km thickness and a Q_s value of 30 overlaying a halfspace with Q_s equal to 600.

Additional research for the estimation of the Q factor has been also made by using the properties of the coda waves and by applying the single S to S backscattering model of Aki and Chouet (1975). By using data of local earthquakes occurred in Peloponnese and recorded by four digital stations, Martin (1988) calculated coda Q, (Q_c), values for frequencies between 1.5 Hz and 24 Hz. The relation found between the coda Q and the frequency was $Q_c=73f^{0.95}$. For the same area Tselentis et al. (1988b) found the relation $Q_c=44f^{0.81}$ using data of nine aftershocks of the 1986 Kalamata earthquake.

Baskoutas et al. (1992) using analog recordings of Athens Seismological Station found a mean Q_c equal to 182 $\pm$ 77 for the frequency of 1 Hz for earthquakes occurred at distances up to 200 Km with magnitudes 3.7 to 5.2 .

Hatzidimitriou (1993) estimated coda Q values for local earthquakes occurred in northern Greece during the period 1983-1989 and recorded by the telemetric network of the Geophysical Laboratory of the University of Thessaloniki. Coda Q estimations were made for four frequency bands centered at 1.5 Hz, 3.0 Hz, 6.0 Hz and 12.0 Hz and for the lapse time windows 10-20 sec, 15-30 sec, 20-45 sec, 30-60 sec and 50-100 sec. The coda Q values obtained show a clear frequency dependence of the form $Q_c=Q_o f^n$, while Q_o and n depend on the lapse time window . Q_o was found equal to 36 and n equal to 0.96 for the time window of 10 to 20 sec, while for the other windows Q_o was increasing from 60 to 129, with n being stable, close to 0.75 . This lapse time dependence was interpreted as due to a depth dependent attenuation while the high attenuation and the strong frequency dependence found are characteristic of an area with high seismicity, in agreement with studies in other seismic regions.

Finally, the Q factor has been estimated using reflection seismics by Loucoyiannakis et al. (1989) along a line 120 Km long starting from Toroneos gulf in Chalkidiki and terminating in North Sporades Basin. They found Q values between 500 and 1000 for the upper 13 Km of the crust for frequencies between 12.5 and 50 Hz.

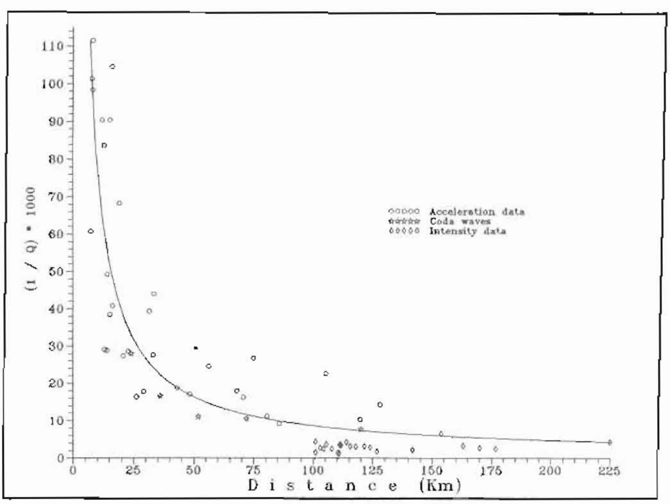

Fig.6. Plot of the calculated Q^{-1} values in the area of Greece, reduced at frequency of 1 Hz, versus distance. Circles denote Q_c values calculated using acceleration data (Hatzidimitriou et al. 1993), stars denote coda Q values (Hatzidimitriou 1993) and diamonds denote Q values as deduced from macroseismic data (Papazachos 1992). The solid line is a least spuares fit to the data of the relation $Q^{-1}=0.00093+0.77246D^{-1}$.

In Tables 1, 2 and 3 we give information on the Q values which have been calculated by using acceleration data, coda waves and intensity data, respectively.

As it is known (Herraiz and Espinoza, 1987) the Q values of the seismic waves depend on the frequency and the hypocentral distance. Therefore in order to be able to compare the Q values from various regions we have to correct for these two factors. In figure (6) we plotted all the Q^{-1} values which have been calculated for the area of Greece and for which information is given on the frequencies and the distances used, versus distance. All values are reduced at frequency of 1 Hz. Circles denote the Q_s values which have been calculated using acceleration data (Hatzidimitriou et al. 1993) plotted versus hypocentral distance. These values have been calculated for frequencies between 3 and 8 Hz and therefore we assumed a mean frequency of 5 Hz. The reduction of the Q_s values at 1 Hz was made by assuming that the frequency dependence of the Q_s is similar to that of the coda waves (Aki 1981), according to the law $Q=Q_0f^n$, and by taking n equal to 0.75 (Hatzidimitriou, 1993). Asterisks denote Q_c values estimated for northern Greece (Hatzidimitriou 1993) and the distance used is derived from the lapse time over which the Q_c was estimated assuming an S wave velocity equal to 3.5 Km/sec. Diamonds denote Q values estimated by using intensity data (Papazachos 1992) plotted versus the mean distances of the used macroseismic data. As we see from that figure there is a systematic decrease of the attenuation with distance. The solid line represents a least square's fit to the data according to the relation $Q^{-1} = A + B.D^{-1}$, where D is distance. The values of A and B found where A=0.00093 and B=0.77246. We see that the Q values deduced from intensity data data are in good agreement with the Q_s and Q_c values even though they show systematically lower attenuation than the average while the Q_s and Q_c values have the same scattering at small and big distances.

It is clear from figure (6) that the Q values that have been calculated for the area of Greece cannot be compared and therefore a regionalization is not possible because they have been calculated for a wide range of distances which means that they reflect attenuation characteristics of different parts of the lithosphere. Obviously more research is needed for the Q regionalization by using homogeneous data sets and paying special attention to the frequency and distance dependence of the Q factor.

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