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Velocity models inferred from p-waves travel time curves in south Aegean

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VELOCITY MODELS INFERRED FROM P-WAVES TRAVEL TIME CURVES IN SOUTH AEGEAN

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Abstract

The seismicity recorded during 1st January to 31st August 2005 from a new telemetry network installed and operating on the island of Crete, is used in an effort to obtain new velocity models for the area of south Aegean. The models are constructed from the P-waves travel time curves and are later used for the events relocation with the HYPOINVERSE algorithm and station delays calculation. Furthermore, results are discussed and compared with the ones derived from other significant previous works presented the last years. We anticipate by combining all the available information from the literature and the analysis of our data set to contribute to the seismotectonic modeling of the study area and to construct a most complete image of the geometry of the subducted plate.
Key words: *velocity structure, earthquakes, relocation.*

Περίληψη

Με σκοπό τη δημιουργία μοντέλων ταχύτητας στην περιοχή του νοτίου Αιγαίου, χρησιμοποιούμε τις καταγραφές σεισμών κατά τη χρονική περίοδο από 1η Ιανουαρίου έως 31η Αυγούστου 2005 από ένα νέο τηλεμετρικό δίκτυο που εγκαταστάθηκε και λειτουργεί στην περιοχή της Κρήτης. Τα μοντέλα ταχύτητας κατασκευάζονται από τις καμπύλες χρόνων διαδρομής των επιμήκων κυμάτων και χρησιμοποιούνται σε συνδυασμό με τις χρονικές διορθώσεις στο χρόνο άφιξης των σεισμικών κυμάτων σε κάθε σεισμολογικό σταθμό του δικτύου για τον ακριβή προσδιορισμό των εστιακών παραμέτρων των σεισμών που έχουν καταγραφεί στην περιοχή του νοτίου Αιγαίου με τη χρήση του προγράμματος HYPOINVERSE. Συνδυάζοντας όλες τις διαθέσιμες πληροφορίες από τη βιβλιογραφία και τα αποτελέσματα της παρούσας μελέτης προσδοκούμε να συμβάλουμε στην αποσαφήνιση του σεισμοτεκτονικού προτύπου της περιοχής καθώς και της γεωμετρίας της καταδυόμενης λιθόσφαιρας της ανατολικής Μεσογείου.
Λέξεις κλειδιά: μοντέλα ταχύτητας, σεισμοί, επαναπροσδιορισμός εστιακών παραμέτρων.

1. Introduction

Over the past four decades, Aegean and its surroundings have attracted the interest of the scientific community worldwide since it is one of the most seismically active and one of the most rapidly deforming parts of the Alpine-Himalayan mountain belt. Its most prominent tectonic feature is the subduction of the eastern Mediterranean lithosphere under the Aegean Sea (Papazachos and Comninakis 1970) along the Hellenic Arc. The Aegean Sea is bounded to the north by the stable continental Eurasia, to the south by the African plate, to the east by the Anatolian microplate and to the west by the Adriatic microplate. The convergence between Eurasia and Africa is emerged due to the different rate of sea-floor spreading between the southern and northern part of the Atlantic Ocean, though the tectonics of the back-arc Aegean region is dominated by strike-slip and extensional motions. Even though the northward movement of the African plate is of about 1 cm/y, the convergence across the Hellenic Trench is approximately equal to 4.5 cm/y. This high rate is due both to the westward motion of Anatolia relative to Africa and to the rapid internal deformation of the Aegean (McKenzie 1988, Jackson 1994, Papazachos and Kiratzi 1996 among others) that affects the region itself.

Fundamental goal of numerous intensive studies has been the determination of the extremely complex plate dynamics in the broader Aegean area. The determination of focal mechanisms (e.g. Jackson and McKenzie 1988, Hatzfeld *et al.* 1989, Papazachos *et al.* 1991, Papazachos and Kiratzi 1996, Taymaz *et al.* 1990, Bohnhoff *et al.* 2005), and tectonic observations for the Miocene and the Quaternary time (Mercier *et al.* 1987, Armijo *et al.* 1996) have shown that along the Hellenic Arc thrust faulting prevails with P-axes trending roughly NE-SW, which is consistent with the underthrusting of the eastern Mediterranean oceanic lithosphere beneath Aegean. The complicated stress field is attributed to the westward motion of north Anatolia, to external compressional forces, exerted by the eastern Mediterranean and Apulian lithospheres and to intense internal deformation of the Aegean microplate.

Active reflection and refraction experiments yielded high resolution images of the top 20 km of the crust, while detailed tomographic studies provided information on the structure of the mantle (Spakman *et al.* 1993) and the velocity distribution in the crust (Papazachos *et al.* 1997). All this information is enhanced by magnetic and gravity measurements (e.g. Tsokas and Hansen 1997, Tirel *et al.*, 2004; Snopek, 2005) and travel time analysis from local and regional earthquakes (e.g. McKenzie 1972, 1978, Le Pichon and Angelier 1979, Jackson and McKenzie 1988, Spakman *et al.* 1988, Taymaz *et al.* 1996). Accurate geodetic data from triangulation and GPS measurements (e.g. Oral *et al.* 1995,

Figure 1 – Main tectonic structures of the Hellenic Arc and Trench system

Reilinger 1997, Kahle *et al.* 1998,; Papazachos 1999, McClusky *et al.* 2000) and geologic field observations provided important information on the regional active tectonics. Papazachos *et al.* (2000) identified from seismological observations a well-shaped Benioff zone to a depth up to 180 km below central Aegean, which starts at a depth of 20 km under the convex side of the sedimentary part of the arc and dips towards the back-arc area, with a dip angle about 30º for shallow depths (20km<h<100km) and of about 45º for greater depths (100km<h<180km).

In the present study we use the arrival times of P-waves of local earthquakes recorded at the stations of a new digital telemetry network operating on the island of Crete and other three regional networks in order to obtain the crustal structure of south Aegean. At the current stage of this work, we have focused on using well-constrained earthquake locations to produce velocity
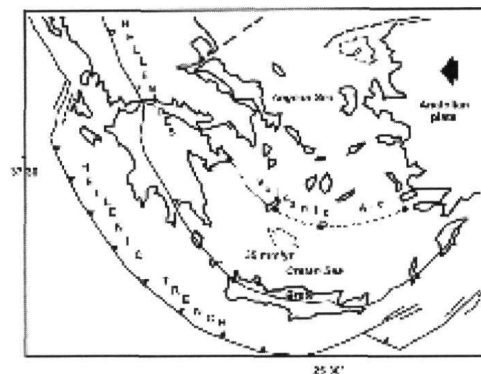
models and, by this way to relocate the whole available data set. The results from the above analysis will be furthermore used in order to relocate the earthquakes recorded in the last several decades in the area.

2. Previous work on the determination of crustal structure of the Aegean

2.1. Introduction

A better knowledge of the crustal structure in a region as complex as the Aegean is important to better understand its past and present tectonic and geodynamic evolution, as well as to improve the accuracy of local and regional earthquakes location. The crustal structure and thickness of the Aegean has been extensively investigated by a variety of studies using deep seismic soundings (e.g. Makris 1978, Makris and Stobbe 1984, Bohnhoff *et al.* 2001), gravity data (e.g. Tsokas and Hansen 1997, Tirel *et al.* 2004, Snopek 2005), surface wave data (e.g. Papazachos 1967, Karagianni *et al.* 2002) and receiver function inversion (Sodoudi *et al.* 2005). The mantle structure has been studied mostly by P-wave travel time inversion using local and teleseismic events (Spakman *et al.* 1988, Papazachos *et al.* 1995, Papazachos and Nolet 1997). In the next paragraph we shortly review the most significant results regarding the crustal structure of the Aegean.

2.2. Main results of Seismic surveys on the crustal structure of Aegean

Five seismic lines along the Cretan and Libyan Seas (Bohnhoff *et al.* 2001, Makris and Brönner 2001, Brönner 2003, Meier *et al.* 2004) as well as results from Expanding Spread Profiles analysis in the western Mediterranean (de Voogd *et al.* 1992, Truffert *et al.* 1993, Fruehn *et al.* 2002, Jones *et al.* 2002) provided detailed information about the structure of the deep basins and crust below Crete and at the collision front between the Aegean microplate and the oceanic lithosphere of the African plate. Additionally, the 20 reflection profiles which were shot along a 50 km-wide corridor across the western Mediterranean Ridge in the framework of the International Mediterranean Ridge Seismic Experiment, IMERSE (Fruehn *et al.* 2002) and the Seismic study of the Mediterranean tectono-sedimentary prism, PRISMED surveys (Chaumillon and Mascle 1997, Huguen *et al.* 2001) provided information about the sedimentary structure of the Mediterranean Ridge.

More specifically, the results from the above deep seismic studies can be summarised in the following: Below the north part of central Crete the maximum thickness of the continental crust is estimated and found equal to 32.5 km, whereas at the eastern part the crust is thinner (24 to 26 km). In the upper crust P-wave velocity varies from 5.8 to 6.3 km/s whereas at the lower crust from 6.4 to 6.9 km/s. Below central Cretan Sea the crust is thinner (about 15 km) with maximum sediments thickness around 3.5 km and its maximum P-wave velocity equals to 4.35 km/s. Onshore central Crete the thickness of sediments is found to be up to 6 km consisting mainly of Mesozoic limestones (P-wave velocity 5.8 to 6.0 km/s) (Bohnhoff *et al.* 2001). To the south offshore the continental crust thins gradually to 17 km and extends about 100 km from the south of central Crete, 140 km south–west of the island and about 60 km south-east from the eastern coast. To the south it comes to contact with the thick sediments of the Mediterranean Ridge, which is in good agreement with the results from the IMERSE and PRISMED experiments. The southern border of the Aegean plate is located close to the border between the inner and the central units of the Mediterranean Ridge (e.g. Mascle *et al.* 1999, Huguen *et al.* 2001) about 100 to 120 km south of Crete.

Along the onshore southern profile, the crust is continental with a thickness of about 28 km at the central south Crete and 17 km of its southeast part. The point from decoupling of the dipping slab from the continental crust is located below SE Crete. The upper mantle velocities vary from 7.7 km/s to 8.0 km/s in the Aegean and African lithospheres, respectively.

Receiver function (e.g. Knapmeyer and Harjes 2000, Li *et al.* 2003, Endrun *et al.* 2002) and surface wave studies (Meier *et al.* 2004) reveal an average depth of the oceanic Moho within the NE-dipping subducting African lithosphere about 55 km below Crete. The discontinuities found at about 30 to 35 km depth below central Crete are interpreted as Aegean continental Moho. Consequently, an about 10 to 15 km thick Aegean mantle wedge is assumed to be present beneath central Crete.

Table 1 - Moho depths from various studies (modified after Sodoudi 2005)

Author(s)	Methodology	Crete	Cretan Sea minimum	Cyclades
Papazachos (1967)	seismic wave velocity	-	-	22-35
Makris (1975, 1976, 1977, 1978)	deep seismic soundings and Gravimetry	30-32	20	28
Makris and Vees (1977)	refraction	30-32	20	26
Calcagnile (1982)	surface wave dispersion	-	35	-
Panagiotopoulos *et al.* (1985)	Earthquake Seismology	-	24-30	-
Papazachos *et al.* (1995)	Seismic tomography	30-40	25	25-30
Tsokas and Hansen (1997)	Gravity	28-30	26	19-32
Bohnhoff *et al.* (2001)	reflection/refraction	24-32.5	15	-
Vigner (2002)	Reflection	-	-	26-25
Karagianni *et al.* (2002)	surface wave dispersion	-	22-24	-
van der Meijde *et al.*(2002)	receiver function	-	29	25
Li *et al.* (2003)	receiver function	31-39	-	25
Tirel et al. (2004)	Gravity	28-31	23	25
Karagianni *et al.* (2005)	surface wave tomography	-	20-22	24-26
Sodoudi (2005)	P and S receiver functions	28-32	22-24	24-30
Snopek (2005)	gravity	30-35	15-25	25-30

Tirel *et al.* (2004) created a new map for the Aegean crustal thickness inferred from inversion of marine gravimetric measurements. They observed a thin crust of about 22 km in the Cretan Sea, whereas the Cyclades are marked with a rather flat Moho of about 25 km. Other scientists (e.g. Makris and Vees (1997), Li *et al.* (2003) estimate a depth of about 25-26 km of Moho beneath Cyclades, whereas they propose a rather thinner crust of about 15-20 km below the Cretan Sea.

Makris and Yegorova (2005) combining the results of Bohnhoff *et al.* (2001) and Brönner (2003) and by inverting gravity measurements concluded that the crust below central Crete is thick (32-34 km) and at the eastern part is thinner (24-26 km). The Aegean crust south of Crete is covered by sediments having a thickness of 6 to 7 km, and then 10 km of continental crust follows plus 6.5 km of subducted oceanic crust. Below western Crete the 21-25 km continental crust is followed by 6.5 km of oceanic crust resulting in a Moho depth at 30-32 km as previously defined by Makris (1977), Makris and Vees (1977), Tsokas and Hansen (1997), and Bohnhoff *et al.* (2001).

Sodoudi (2005) applied the P and S receiver function analysis observed beneath the southern Aegean Basin to conclude a very thin crust, reaching 26 km eastwards, and a thicker crust at the Cyclades region, and to estimate the Moho depth at about 25-30 km. In Table 1 the results concerning the Moho depths from the aforementioned studies, are summarized.

3. Present work on the crustal structure of south Aegean

3.1. Description of the SASN network

Accurate earthquake locations are an imperative prerequisite in order to seek for active structures in a study area and to investigate their geometric and kinematic properties. To this end a major requirement is the installation and operation of a local seismological network to provide optimum station coverage of the study area. Aiming to explore this information from the seismicity occurring along the Hellenic Arc, the data recorded at the digital telemetry network of the Geophysical & Seismological Laboratory (GSL) of the Technological Educational Institute of Crete were used. The South Aegean Seismological Network (SASN) consists of nine short-period three-component seismic stations along the island of Crete, and one broadband station located in the city of Chania (Fig. 2). Network installation started in spring of 2004 and it was in full operation (10 stations) at the end of the same year. Recently this network expanded to the east with the installation of three new stations (SIT, IER and KLD). Table 2 provides information on station characteristics, as their names, geographical coordinates, altitude and type of seismometers. Additionally to these data all the available waveforms from the seismological stations at south Aegean operated by the GEOFON network (their sites are shown on Fig. 2) were analysed (http://www.gfz-potsdam.de/geofon/www_req/gfn_data.html). Moreover, the time arrivals from those earthquakes recorded also at the seismological stations of the National Observatory of Athens and the Seismological Network of Geophysics Department, Aristotle University of Thessaloniki were used in the present study.

Table 2 – Information on the stations characteristics

STATION LOCATION	CODE NAME	LAT °N	LONG °E	ALT. (m)	DAS	SEISMOMETER
Chania (BB)	CHNb	35.519	24.042	34		Guralp CMG-40T
Chania	CHN	35.519	24.042	76		Sercel L-4-3D 1Hz
Palaiochora	PLC	35.234	23.686	25		Sercel L-4-3D 1Hz
Rethymno	RTH	35.365	24.502	51		Sercel L-4-3D 1Hz
Gramvousa	GRM	35.620	23.605	133		Sercel L-4-3D 1Hz
Sfakia	SFK	35.201	24.138	79		Guralp CMG-40T-1
Hrakleio	HRK	35.304	25.071	70	Reftek 130A	Guralp CMG-40T-1
Ag. Barbara	AGB	35.140	25.132	629		Guralp CMG-40T-1
Ag. Galini	AGL	35.096	24.688	26		Guralp CMG-40T-1
Kantanos	KAN	35.326	23.740	473		Guralp CMG-40T-1
Kaloudiana	KLD	35.487	23.690	100		Sercel L-4-3D 1Hz
Sitia	SIT	35.209	26.106	10		Guralp CMG-40T-1
Ierapetra	IER	35.014	25.733	33		Guralp CMG-40T-1

3.2. Data analysis

After the phase picking using the PQL program a preliminary location of the earthquakes was made. The velocity structure is very crucial since it provides the basis for the determination of the ray paths and hence the travel times. For the initial location, we considered the velocity structure proposed by Papazachos and Nolet (1997) and then we explore how it can be estimated in different regions from the travel times of P-waves. More than 1500 earthquakes were recorded by the stations used and located preliminary using the HYPOINVERSE (Klein 2000) computer program.
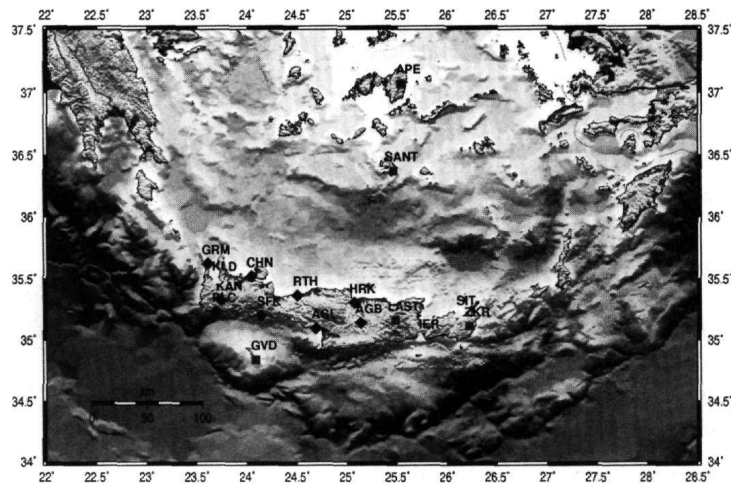
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Figure 2 – Sites of the stations of SASN (diamonds represent the station sites during the time period of the present study, whereas the stations installed lately are represented by triangles) and GEOFON (squares), seismological networks

Having in mind that the use and interpretation of earthquake data especially for seismotectonic purposes strongly depend on accurate earthquake locations we are trying to select from the initial data set the earthquakes whose hypocentral parameters are well-determined. To achieve this goal we define some selection criteria adequate enough to allow us to end up with a set of adequately constrained earthquake locations. According to these, we include in our data set the earthquakes recorded by at least 7 stations (the minimum number of phases is fixed to 8, which means at least seven P and one S arrival time) with small time residuals. It has been shown in the literature, from experiments conducted in the same area (e.g. Hatzfeld *et al.* 1989), that earthquake locations with significant number of phase arrivals with small time residuals do not differ much if various reasonable velocity structures are adopted during the location process, and consequently they can be considered reliable. We set the error at the origin time (RMS) to be less than 0.5s, and the errors at the determination of the hypocenter (ERH) less than 5 km and (ERZ) less than 10 km. Although the values of these errors have only statistical meaning, it has been shown in seismic sequences studies that aftershocks with similar error estimates represent satisfactorily the geometry of the fault, which caused the main shock (e.g. Karakostas *et al.* 2003). Applying these criteria we ended up with a set of 420 well constrained earthquake locations.

Aiming to improve the initial locations derived using a laterally homogeneous model we tried to develop velocity models for smaller regions and calculate residuals at individual stations, treating them later as station corrections. By this way we can avoid errors due to the different ray paths and different epicentral distances if the model and time residuals have been determined for large areas. Taking into account the spatial distribution of the epicenters and the specific tectonic features, we divided our study area into 15 smaller sub-regions, having in mind to keep a satisfactory number of well-located earthquakes in each one of them. After several trials and mapping of seismicity we ended up in the separation depicted in Figure 3. In an effort to determine the velocity structure of the area, i.e. evaluate the P-wave velocity as a function of depth, we constructed the travel time curves i.e. the travel times of P-waves versus the epicentral distance in each sub-region. With the appropriate algorithms based on the theory of refraction we calculated the velocity of P-waves and the thickness of each layer. The models derived after applying the aforementioned method consist of two or three layers over a half space. The value of the S-waves velocity is calculated indirectly through the value of the V_P/V_S ratio, a crucial parameter for the depth control of the earthquakes. We selected those events with more than five reliable S arrivals and we constructed individual Wadati plots. From the slope of the $\tau_S-\tau_P=f(\tau_P)$ curve we calculated a mean V_P/V_S ratio equal to

1.78±0.01. This value was used later at the location process and it is in agreement with the results from other studies [e.g. 1.79 obtained by Hatzfeld *et al.* (1990), 1.80 by De Chabalier *et al.* (1992)].

For each sub-region separately we located the selected events using the new velocity model and examined the travel time residuals. In some cases, we obtained satisfactory results, whereas in others we had to try different values either in velocity or the thickness of a layer in order to achieve minimization of the mean RMS. These velocity models were further used to relocate all the recorded events using the HYPOINVERSE algorithm and calculate the stations delays.

Moreover, continuing the procedure of the relocation, time delays were determined and used as time corrections applying the following procedure. We relocated each event several times using the stations delays

Figure 3 – Earthquakes used to define velocity models in 15 sub-regions shown on the map

calculated at every iteration, eliminating at the same time the arrivals with large residuals. This iterative process comes to an end when there is no significant change at the earthquake location, which means that the mean time residual is less than 0.1 s. For the present study the final mean time residual in all cases was less than 0.03 s. The calculation of station delays is of great importance, since they contain the information about the lateral velocity variation. Groups of positive, negative or mixed station delays observed in different regions bear information about the crustal structure: positive station delays are mainly indicative of thick or/and slow crust compared to the assumed model and negative station delays are indicative of a thin or/and fast crust compared to the assumed model. The incorporation of station delays in the location process has resulted in a construction of a 3D velocity model from the 1D model, since the horizontal heterogeneities are taken into account. In Figure 4 the finally adopted velocity models determined for each sub-region are presented.
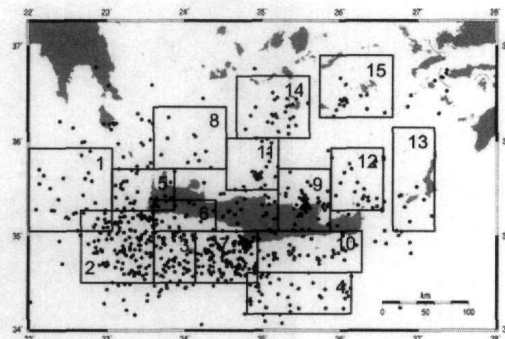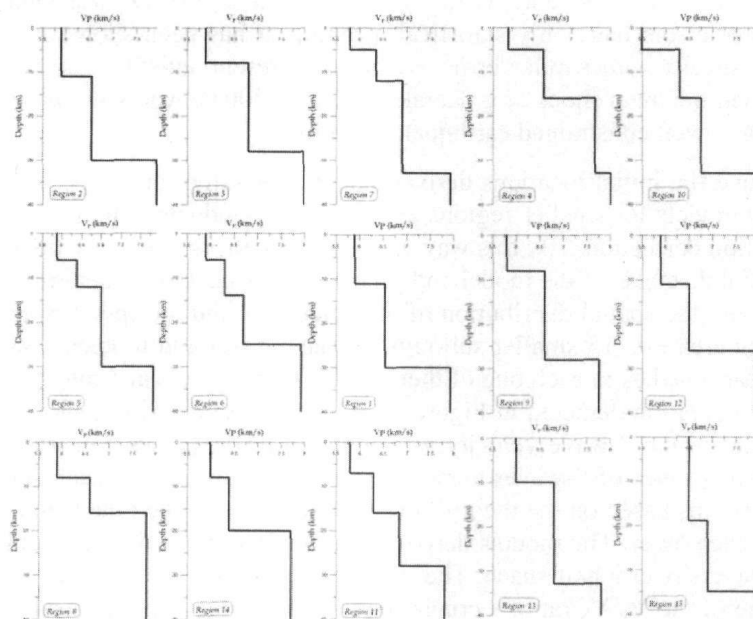

Figure 4 – 1D velocity models for each sub-region

In regions 2,3,7,4 and 10 located south of Crete, we determined a crustal thickness of about 30-33 km and the P-waves velocity varying from 6.30 to 6.90 km/s. We observe a thin crust of about 17-20 km in the central Cretan Sea (regions 8 and 14), with P-waves velocity 6.40-6.60 km/s, whereas in region 11 the thickness of the crust reaches the 27 km. In western Crete (regions 5, 6), we observe that the thickness of the crust varies from 25-30 km, and at the eastern part of the island (regions 9 and 12) the crust is thinner, 26-28 km. Finally, in regions 13 and 15 we observe a thick crust of about 33 km.

4. Results and Discussion

Velocity models for the southern part of the Hellenic Arc have been determined using data from a local telemetry network on the island of Crete and three permanent regional networks, based on the travel time curves of reliably located earthquakes. In addition to these models time delays at all stations from events at different regions were calculated and applied as time corrections to improve the reliability of earthquake location in the area. Our final models are consistent with previous results concerning the crustal structure in the south Aegean. The back-arc area is generally characterized by a thin crust, while a thicker crust is observed under the accretionary prism. Based on these models and the calculated time delays we are working on the compilation of a new catalogue of seismic events with well-constrained hypocentral parameters. Further work will be done using all the available phase arrivals from the International Seismological Centre Bulletin, as well as from other local or regional experiments conducted at the area, in an effort to examine in more detail the geometry of the Benioff zone in the Hellenic arc, as well as the properties of the seismogenic layer, which overthrusts the Mediterranean lithosphere.

5. Acknowledgments

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