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TOMOGRAPHY OF THE CRUST AND UPPER MANTLE IN SE EUROPE

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ABSTRACT

Compressional velocity structure of the crust and the upper mantle in SE Europe (broader Aegean area) is studied by inverting residuals of the first P arrivals from shocks in this region (16o -31°E, 34° - 43°N). The data used come from regional events recorded by the permanent network of stations during the period 1971-1987, enriched with data from experiments with portable seismographs in four regions of this broad area. This study confirms the strong variations of the crustal thickness in this area, as well as the subduction of the eastern Mediterranean lithosphere under the Southern Aegean and gives further detailed information on the crustal and upper mantle structure of the area. Such important new information is the existence of a low velocity crustal layer in western Greece and Albania and 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.

ΣΥΝΟΨΗ

Στην παρούσα εργασία παρουσιάζεται η δομή ταχύτητας των επιμήκων κυμάτων στο φλοιό και τον άνω μανδύα στη ΝΑ Ευρώπη με την αντιστροφή των χρονικών υπολοίπων των πρώτων αφίξεων P κυμάτων από σεισμούς αυτού του χώρου (15° - 31°E, 34° - 43°N). Τα δεδομένα που χρησιμοποιήθηκαν προέρχονται από σεισμούς της περιοχής οι οποίοι καταγράφηκαν από το μόνιμο δίκτυο των σεισμολογικών σταθμών κατά την περίοδο 1981-1987, εμπλουτισμένα με δεδομένα από πειράματα με φορητούς σεισμογράφους σε τέσσερις περιοχές αυτού του χώρου. Η μελέτη αυτή επιβεβαιώνει την έντονη μεταβολή του πάχους του φλοιού καθώς και την κατάδυση της λιθόσφαιρας της ανατολικής Μεσογείου κάτω από το νότιο Αιγαίο και δίνει παραπέρα λεπτομερείς πληροφορίες για τη δομή του φλοιού και του πάνω μανδύα αυτού του χώρου. Τέτοιες πληροφορίες είναι η ύπαρξη ενός στρώματος χαμηλής ταχύτητας στη δυτική Ελλάδα και την Αλβανία και ότι η ανωμαλία ταχύτητας στο μανδύα κάτω από το νότιο Αιγαίο εκτείνεται οριζόντια και βαθύτερα προς τα βορειοανατολικά από την περιοχή ανωμαλίας ταχύτητας που υποδεικνύει η ζώνη Benioff των σεισμών ενδιαμέσου βάθους.

INTRODUCTION

The velocity structure of the Southeastern Europe exhibits a great complexity (e.g. Spakman et al., 1993). This complexity is a result of the geodynamic evolution of this area which plays an important role in the formulation of the presently observed tectonic structure. In fact, the velocity field is often used for the *reconstruction of the image* of the tectonic regime (and

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therefore testing of models of the geological evolution) and vise-versa (de Jonge et al., 1993).

The main feature of this area (fig.1.) is the subduction of the African plate beneath the Eurasian, especially in the Hellenic arc (Papazachos and Comninakis, 1969/70; Papazachos and Delibasis, 1969; McKenzie, 1970, 1972, 1978; LePichon and Angelier, 1979, Wortel et al., 1990). As a result of this subduction, the Aegean exhibits the characteristic structure of a marginal sea with volcanic activity (Georgalas, 1962), magnetic anomalies and positive isostatic anomalies (Fleicher, 1964; Vogt and Higgs, 1969; Makris, 1976), high heat flow (Fytikas et al., 1985) and high attenuation of the seismic energy (Papazachos and Comninakis, 1971, Hashida et al., 1988). The high shallow seismicity is expressed with reverse faults in the external part and nornal faults in the concave (inner) part of the Hellenic arc (Comninakis, 1975; McKenzie, 1978) while the intermediate depth seimicity has a well-defined Benioff zone (Papazachos and Comninakis, 1969/70). Of major significance is also the westward motion of Turkey, expressed with the right lateral motion of the North Anatolia Fault and its continuation in the Northern Aegean together with the North Aegean trough (fig.1). The isodepths of 100 and 160Km for the intermediate depth earthquakes in the southern Aegean (Papazachos, 1990) are also shown in figure (1).

The velocity structure in this area was studied using classic methods for travel times from earthquakes (Papazachos et al., 1966, Panagiotopoylos and Papazachos, 1985) or explosions (Makris, 1972, 1978; Voulgaris, 1991), dispersion of surface waves (Papazachos et al., 1967, Calganile et al., 1982) and gravity data (Makris, 1973; Chailas and Lagios, 1992). Only recently the first tomographic images were presented and allowed a more detailed description of the lithosphere and the mantle. These results (Spakman, 1986, 1988; Spakman et al., 1988, 1993; Ligdas et al., 1990; Ligdas and Main, 1991) were mainly trying to establish the principal features of this lithosphere-upper mantle system, up to the depth of 800Km or more. Other studies were concentrated on a more or less local scale (Drakatos, 1989; Drakatos et al., 1988, 1989; Martin, 1988; Christodoulou and Hatzfeld, 1989, Ligdas and Lees, 1993). The structure of the crust and of the uppermost mantle have not been studied in detail. In all the regional

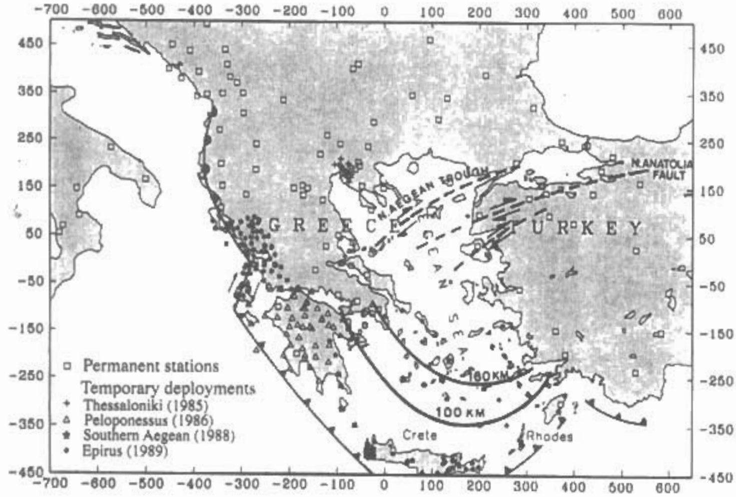

Fig. 1: Tectonic features of the studied area. The seismological station networks from which data were used in this study, both regional and local, are also presented. In the Southern Aegean, two thick solid lines indicate the isodepths of 100Km and 160Km for the intermediate depth events of the Benioff zone.

studies, the crust is dealt as one layer (Drakatos and Drakopoulos, 1991; Ligdas et al., 1990) or is a part of the first layer (Spakman, 1988). Moreover, the number of stations and the number of arrivals used is limited. Generally, the tomographic images are dominated by a high velocity zone (presummably the African lithosphere) dipping beneath the Aegean in low velocity material.

In the present work, more than 100000 P arrivals from 4229 earthquakes in the region 16oE-31oE and 34oN-43oN are used for the determination of the detailed structure of the crust and the upper mantle in this area. The resulting structure shows a great complexity but is mainly controlled by the strong variations of the crustal thickness and the subduction of the African plate. Detailed features can be recognized, both in the crust (e.g. low velocity layer under the Hellinides) and in the upper mantle (extension of the velocityanomaly under the southern Aegean further and deeper to northeast).

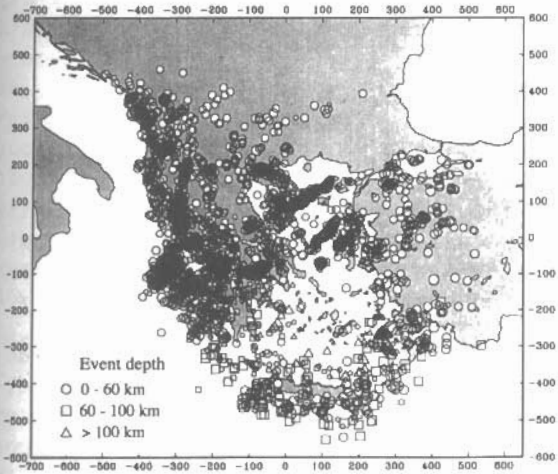

Fig. 2: Map of the finally determined epicenters of the events for which data were used in this study. The events have been separated in three groups according to their focal depth.

THE DATA

The data come from the annual bulletins of the Geophysical Laboratory of the University of Thessaloniki for the time period 1981-1987 for events with surface wave magnitude MS≥3.5 and for the period 1971-1980 for events with surface wave magnitude MS≥5.0. This data set was enriched with data from four local experiments (fig.1) in Thessaloniki basin (N. Greece), Peloponnesus (S. Greece), Southern Aegean and Epirus (NW.Greece), conducted during the summers of 1985, 1986, 1988 and 1989, respectively, by the Geophysical Laboratories of the Universities of Thessaloniki, Athens and Grenoble (France). These experiments, especially the last two, had very dense networks which recorded a significant number of events at areas where the permanent network is very sparse. Moreover, the Southern Aegean network recorded many intermediate depth events which are very usefull in the illumination of the velocity structure of the upper mantle.

The initial data set consisted of approximately 20000 events and 200000 P arrivals. After a detailed study of this set, quality criteria were established, regarding mainly the number of observations and the distance of the closest recording station from the epicenter of each event. The filtering process, based on these criteria, resulted in a final data set of 4229 events with 109655 P arrivals recorded at 327 stations. The final epicenters of the events for which data were used are shown in the map of figure 2. The locations of the stations are depicted in figure 1. In figure 3, the frequency histogram of the P arrivals used for each earthquake are shown. It is observed that for the majority of the earthquakes more than 14 arrivals were used.

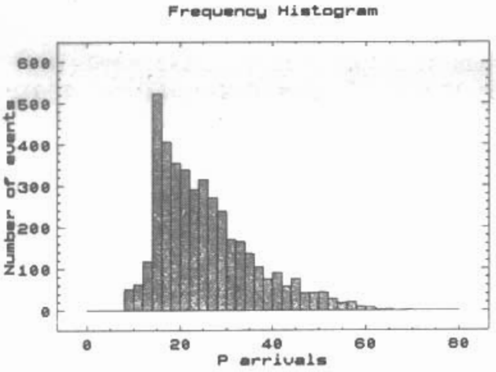

Fig.3: Frequency histogram of the P arrivals used for each event in the present study.

This data set was separated into two subsets, consisting of events with depths less and above 60 Km, respectively. This separation, was based on the fact that almost all events with depths larger than 60 Km are the intermediate-depth events in the Southern Aegean Benioff zone.

MODEL AND RAY GEOMETRY CONFIGURATION

Usually, in inversion studies, the flat-earth layered model is used. However, as the scale of the survey increases, the errors due to the spherical shape of the earth can not be neglected. In our case, the dimensions of the studied area is approximately 1200Km X 1000Km and therefore the earth's curvature should be taken into account. However, we would still like to use a "cartesian" system of coordinates for convenience (as in a flat earth model). In the present study, we adopted the Tranverse Mercator Projection (TMP) which establishes a curvilinear cartesian system on the earth's surface. If we choose the center of our coordinate system in the center of our area, then the maximum angular error of the TMP is less than 0.4o and the maximum distance error is around 0.4%. However, the distance error was calculated for each ray and travel times were corrected accordingly. Using this coordinate system, we created a 3-dimensional grid of cells of equal size with constant velocity. The dimensions of each cell were 100Km X 100Km.

For a given ray, the travel time residual is a function (in a first order approximation) of the corrections to the hypocenter parameters and the velocities of the cells hit by the ray. The whole data set forms a linear system of equations $\mathbf{\Delta T = A\, \Delta X + B\, \Delta V}$. The hypocenter matrix of this equation, **A**, can be separated (Pavlis and Booker, 1980) from the slowness matrix, **B**, resulting in a linear system of the form:

$$\Delta T' = B'\ \Delta V \qquad (1)$$

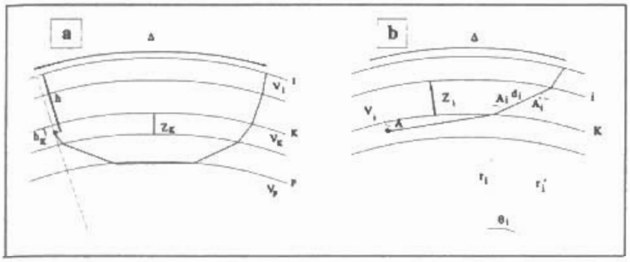

Fig.4: Geometry of the refracted (a) and direct (b) waves in the spherical layered earth. See text for explanation of the quantities presented in the figure.

where $\mathbf{\Delta T'}$ and $\mathbf{\Delta V}$ are the altered residual and velocity correction vectors and **B'** is altered velocity derivative matrix.

For the ray tracing, we incorporated the approximate ray tracing scheme of Thurber and Ellsworth (1980). In this aproach, an equivalent 1-dimensional structure is calculated from the 3-dimensional model for the epicenter-station path. The direct and refracted waves are calculated in this "equivalent" model and the fastest ray is selected after the travel time is recalculated in the 3-dimensional medium. The main disadvantage of this kind of approximate ray tracing is that the spatial mapping of the velocity anomalies is not always quite accurate.

In order to perform the ray tracing, some modifications need to be made in the geometry of the refracted and direct waves. In figure 4, the plots of direct and refracted rays are shown in a spherical earth. For the refracted waves, the travel time for an event at the earth's surface is given by:

$$T=\frac{r_P\Delta}{r_0}+2\sum_{i=1}^{p-1}\int_{r_P}^{r_0}\frac{\left(V_P^2-\left(\frac{r_P}{r}\right)^2V_i^2\right)^{\frac{1}{2}}}{V_iV_P}dr\approx\frac{r_P\Delta}{r_0}+2\sum_{i=1}^{p-1}\frac{\left(V_P^2-a_i^2V_i^2\right)^{\frac{1}{2}}}{V_PV_i}z_i \tag{2}$$

where $a_i=r_p/r_i$, r_p is the radius of the top of the P layer and r_i' is the average radius of the i layer. Generally, for an event at depth h (figure 4a), the travel time is given by a formula, similar with the flat earth case :

$$T=\frac{r_P\Delta}{r_0}+\sum_{i=1}^{p-1}\frac{\left(V_P^2-a_i^2V_i^2\right)^{\frac{1}{2}}}{V_P\ V_i}z_i+\sum_{i=k}^{p-1}\frac{\left(V_P^2-a_i^2V_i^2\right)^{\frac{1}{2}}}{V_P\ V_i}z_i-\frac{\left(V_P^2-a_K^2V_K^2\right)^{\frac{1}{2}}}{V_P\ V_K}h_K \tag{3}$$

For the direct waves, the false position and secant method are used. Some adaption of the methods is necessary since the tangent of the take-off angle, used in the calculations, can be infinite in certain pathological cases. The length, d_i, of the ray in each layer, i, is given by the relation:

$$d_i=\sqrt{4\sin^2\frac{\theta_i}{2}r_i(r_i-z_i)+z_i^2}=\sin\theta_i\frac{r_i}{\sin A_i}=\sin\theta_i\frac{r_i'}{\sin A_i'} \tag{4}$$

The travel time is then simply calculated as:

$$T=\sum_{i=1}^{K}\frac{d_i}{V_i} \tag{5}$$

Equations (3) and (5) are used for the calculation of the travel times but also for the calculation of the spatial derivatives of the hypocenter matrix, **A**. For example, the spatial derivatives for the refracted waves are calculated from equation (3):

$$\frac{\partial T}{\partial\Delta}=\frac{r_P}{r_0}\frac{1}{V_P}\quad\text{και}\quad\frac{\partial T}{\partial h}=\frac{\partial T}{\partial h_K}=-\frac{\left(V_P^2-a_K^2V_K^2\right)^{\frac{1}{2}}}{V_P\ V_K} \tag{6}$$

For the direct waves, the derivative, with respect to the the epicentral distance, Δ, is equal to:

$$\frac{\partial T}{\partial\Delta}=\sum_{i=1}^{k}\frac{\partial\Delta_i}{\partial\Delta}\frac{r_i}{r_0}\frac{1}{V_i}\sin(A_i-\theta_i)=\frac{r_K}{r_0}\frac{\sin(A-\theta_K)}{V_K}\frac{\partial\sum_{i=1}^{k}\Delta_i}{\partial\Delta}=\frac{r_K}{r_0}\frac{\sin(A-\theta_K)}{V_K} \tag{7}$$

The vertical derivative can be calculated numerically or be approximated by:

$$\partial h \qquad 2d_K \qquad V_K \qquad (8)$$

INVERSION OF THE DATA

In the present work, the linear system (1) was solved using the damped least squares' method where a damper, ϵ^2, is added to the diagonal of the normal equations matrix. Massive computer memory was available, hence resolution and errors were formally calculated using the Singular Value Decomposition (SVD). Various damping factors were tried in order to "trade" between resolution and error and finally the value of $\epsilon^2 = 300$ was selected. The effect of large residuals was smoothed out by applying an exponential decaying weight to residuals between 3 - 6 sec. After the calculation of the velocity structure, the events were relocated in this new structure. This procedure was repeated in an itterative scheme. At each step, the velocity model was computed for a second grid which had half a grid offset in both horizontal dimensions with respect to the original one and the final velocity structure resulted as the average of the two.

Initially, a 1-dimensional model was calculated. The final velocity model is presented in Table 1. The majority of the discontinuites is not arbitrarily chosen. The 20Km and 30Km approximately correspond to the bottom of the granitic and the basaltic layer for the average model of the area (Panagiotopoulos and Papazachos, 1985), while the 40Km is about equal to the maximum crustal thickness in the area. The last one (160Km) is near the maximum depth of the intermediate depth events in the Southern Aegean Benioff zone.

Starting from this 1-dimensional model, we calculated the final 3-dimensional model. After four (4) iterations we practically had no variance reduction. The final variance improovement was 36.7% (23.0%, 11.8%, 4.7% and 2.1% for the four iterations, respectively). At each iteration, all events were relocated in the new model. The average total horizontal change was 4.8 ± 6.5 Km and the average total vertical change was 2.9 ± 7.6 Km, while very few foci changed more than 20Km in either horizontal or vertical direction. The average change in the origin time is -0.50 ± 1.11 sec and in almost all cases between -3.5 sec and 2 sec. The frequency histogram of the final standard devations (figure 5) is a superposition of two histograms, one for the regional events, recorded by the permanent stations, and one for the events of the local experiments, recorded by portable stations, with average values around 0.9 sec and 0.3 sec, respectively.

Table 1: Final 1-dimensional velocity model

Depth (Km)	Velocity (Km/sec)
0-10	5.96
10-20	6.09
20-20	6.74
30-40	7.31
40-60	7.77
60-90	7.97
90-120	8.10
120-160	8.12
160-	8.20

If we compare the epicenters finally determined (fig. 2) with the initial ones, we can not detect any obvious changes, horizontal or vertical, for the surface events. However, for the intermediate depth events with depths larger than 100Km, severe depth changes are observed in the outer (convex) part of the Hellenic arc. In this part of the arc, we do not expect events with such depths (>100Km), since the subduction takes place further to the North. After the relocation, only two events (from the initial nineteen) still maintain such a large

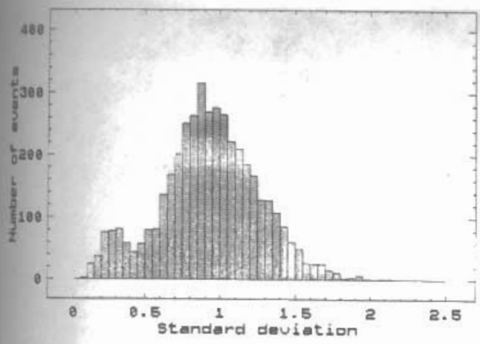

Fig. 5: Frequency histogram of the final standard deviations of each shock for which data were used in the present study. A superposition of the histograms for the permanent and portable stations is observed, with average deviations around 0.9 and 0.3 sec, respectively.

depth, in this area. This hypocentral improvement stregthens the importance of the relocation in a 3-dimensional medium, especially in a subduction zone, where strong velocity changes are observed.

TOMOGRAPHIC IMAGES

In figures 6a to 6h the velocity distribution within each layer is shown. In the first layer (fig. 6a), which corresponds roughly to the sedimentary and the upper part of the granitic layer, the Aegean and Asia Minor generally exhibit low velocities (5.2-5.8 Km/sec). North Greece shows high velocities within a zone of NW-SE direction where velocities above 6.2 Km/sec are detected. This zone almost coincides with the geologic Servomacedonian zone, consisting of metamorphic formations. The Dinarides - Hellenides mountain chain (S. Yougoslavia, Albania, Epirus, Peloponnesus, Crete, Rhodes, SW.Turkey), which will be referred as DHMC onwards, shows typical granitic layer velocities (5.9-6.1 Km/sec). However, a low velocity anomaly is builting in the western part of Albania and Epirus (NW Greece). This anomaly seems to dip to the east and appears to extend along the DHMC in the second layer (Fig. 6b). This practically results in a Low Velocity Layer (LVL) at depths around 10-20 Km, connected to the Alpidic mountain chain in this area. In the inner part of the Aegean typical granitic (6.0-6.2Km/sec) and basaltic (6.5-6.7Km/sec) velocities are observed.

In the third layer (Fig.6c), the velocities along the DHMC are around 6.1-6.3 Km/sec, showing that we are still in the granitic layer. However, in the S.Aegean we have basaltic velocities (6.5-6.7 Km/sec), as a result of the crustal thinning in the volcanic arc above the S.Aegean Benioff zone. This effect is also pronounced in the N.Aegean, along and north of the N.Aegean trough, where even higher velocities are observed. Although, the extend of the crustal thinning in this area has been a matter of debate (Brooks and Kiriakidis, 1986), it seems, that the crust is thinner in this area than in the S.Aegean.

The fourth layer (fig. 6d) exhibits the higher velocity contrasts. This is due to the fact that along the DHMC (up to Crete), the crust is around 40 Km thick (or more). Therefore, the velocities in this area are typical of the lower basaltic crust (6.5-6.9 Km/sec). However, in the N.Aegean (along and north of the N.Aegean trough), S.Aegean, Asia Minor and S.Adriatic, the crust is thinner (around 30 Km). As a result, for this layer, the velocities in these areas are much higher (7.5-7.9Km/sec), representing the upper mantle. This relatively low sub-moho velocities (around 7.8 Km/sec) is characteristic for this area (Papazachos et al., 1966).

The fifth layer (fig. 6e) shows upper mantle velocities in all its extent. Along the DHMC, where the overlying crust is thick, relatively lower velocities, around 7.6Km/sec, are observed. This image is completely

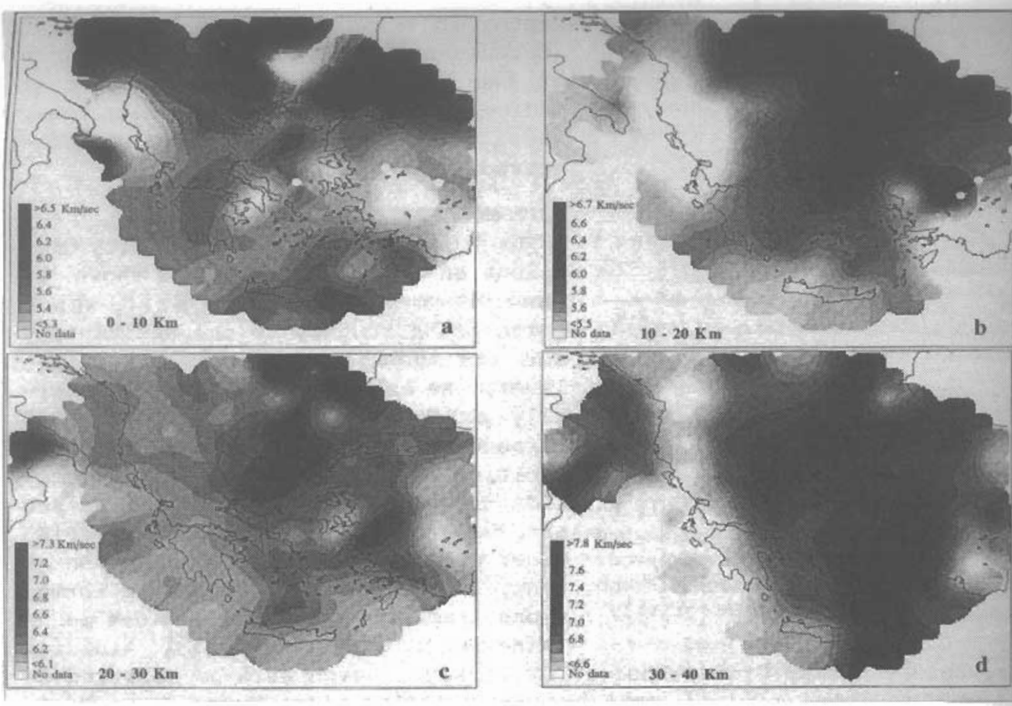

Fig. 6(a-d): Tomographic images of the four top layers in 16-level grey scale. Depth range is indicated in the figures.

changed in the next layer (60-90Km; fig.6f). The velocity structure is now controlled by the subduction process in the S.Aegean and not by the crustal thickness variations. Along the outer Hellenic arc, a high velocity zone appears, corresponding to the dipping African lithosphere. This zone extends further to the north, along the western part of Epirus, Albania and S.Yougoslavia. In the inner part of the Hellenic arc, a low velocity zone is observed, just below the volcanic arc. The amphitheatrical shape of this zone coincides very well with the shape of the dipping lithosphere. This indicates that the low velocities in this area are simply a result of the high heat flow due to the partial melting of the underlying subducting lithosphere (see also the isodepths of the intermediate depth events in fig. 1).

This image is continued in the next layer (fig. 6g). Both the high (outer) and the low (inner) velocity zones move further to the inner part of the Hellenic arc. The outer limits of the subducting lithosphere are starting to show as relatively lower velocity areas in the outer edges of the high velocity zone. In the inner part of the arc, the low velocity zone covers the whole Aegean sea and extends even further to the northeast (N.Greece, NW Turkey). This development of the velocity anomalies is almost identical in the last layer (120-160Km; fig. 6h), although its spatial extent is limmited due to the lack of seismic rays wich penetrate it.

In all tomographic images, it should be noted that velocity anomalies in the edges of its image can sometimes be ficticious and misleading. This is

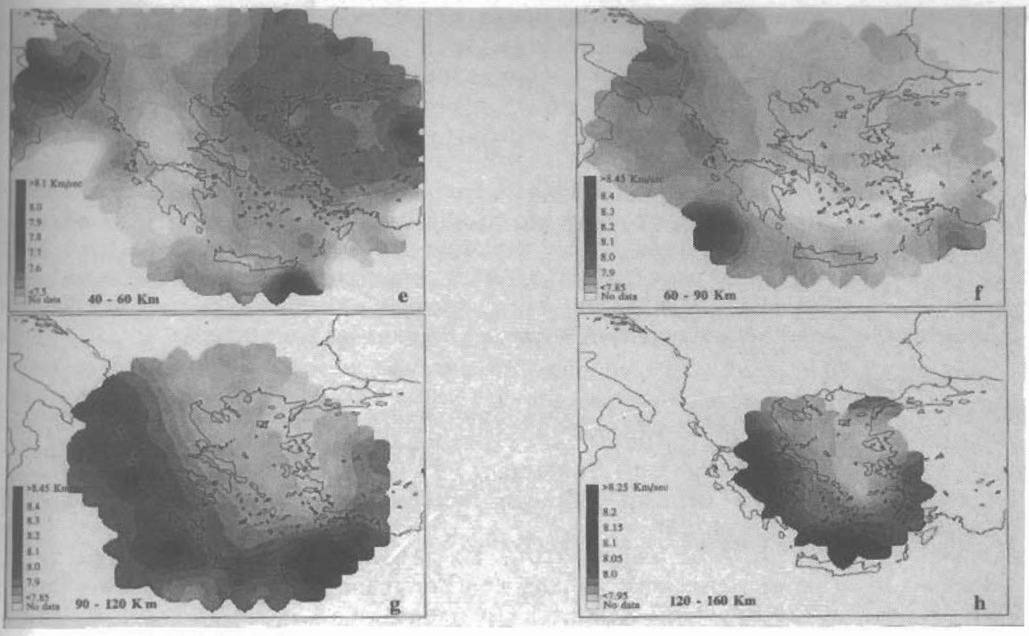

Fig. 6(e-h): Tomographic images of the four bottom layers in 16-level grey scale.

due to the fact that peripheral shells are not well covered, spatially or azimuthally, by seismic rays. A typical example are the anomalies in the first four layers (fig. 6a-d) in the northeast edge of the picture (eastern coasts of Bulgaria) where both the earthquakes and the stations' network are quite sparse. This results in an almost uniform velocity (6.3-6.8Km/sec) for these first four crustal layers.

DISCUSSION - CONCLUSIONS

The present study confirms that the P-wave velocity structure of the crust and upper mantle in SE Europe shows strong horizontal variations as a result of the complicated crustal structrure and the subduction process in the S.Aegean.

The velocity distribution in the shallow layers presented in the present study (0-10Km, 10-20Km, 20-30Km, 30-40Km, 40-60Km) is strongly influenced by the variation of the depth of the Conrad and Moho discontinuities. Of special interest is the identification of a low velocity layer at a depth between about 10 and 20 Km in western Greece and Albania. The observation that the crust in the northern Aegean is even thinner than the crust in the southern Aegean is also quite significant.

The velocity structure of the deeper layers (60-160Km) reveals the subduction process in the S.Aegean. The African lithosphere appears as a high velocity dipping zone, having an amphitheatrical shape in the S.Aegean. A low velocity zone appears in the inner part of the subduction, possibly as a result of the high heat flow due to the partial melting of the subdicting lithophere. Both these velocity anomalies extend much further to the north than the present subduction, as indicated by the intermediate depth events.

This propably indicates that the subduction has gradually migrated from the N.Aegean to its present position, possibly since Eocene (DeJonge et al., 1993), leaving the velocity anomalies as a trace of its migration. These observations, along with other geophysical data (Shanov et al., 1986), indicate the possible existence of an old subduction in the N.Aegean.

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