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3-D Crustal Velocity Structure in Northwestern Greece

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Abstract—The three-dimensional crustal velocity structure in the area of the northwestern Greek mainland was determined by *P*-wave travel time inversion, applying a two-step tomography procedure. The data set consists of the travel-time residuals of 584 well located earthquakes. In order to improve the initial (reference) velocity model, before the inversion of travel times, the "minimum 1-D" model was determined. Several tests were conducted to estimate model stability and hypocenter uncertainties. The velocity distribution in the shallow layers (4 and 7 km) is strongly affected by the crustal thickness variation and the complex tectonics. A first, well-defined velocity discontinuity appears at a depth of 3–6 km, along the Hellenides Mountain chain. A second low velocity anomaly is detected at a depth of 9–12 km and may be connected with the Alpidic orogenesis. Another interesting feature appears beneath the Amvrakikos Gulf (horstgraben structure), where relatively low velocities (<6.0 km^{-1}) appear to a depth of 20 km. Finally, a well-pronounced velocity boundary is found at a depth of 16 km. In general, low velocities are predominant along the Dinarides-Hellenides Mountain chain, rather typical for the upper crust.

Key words: Greece, crustal structure, tomography.

Introduction

The Aegean region represents a marginal basin behind an active subduction zone located along the outer Hellenic arc (Fig. 1, in the frame). Rapid and intense deformation characterizes the Aegean region, which is between the two major lithospheric plates of Africa and Europe (McKenzie, 1978; Le Pichon and Angelier, 1979). Extensional process and consequent crustal stretching, presumably which started in the Miocene, involves the Aegean backarc basin (Malinverno and Ryan, 1986). The extensional tectonics in the Aegean basin have not yet led to the creation of oceanic lithosphere. Moreover, the convergence of the Eurasian and the African plates has induced crustal shortening and thickening of the Dinaric and Hellenides chains (Patacca *et al.*, 1990).

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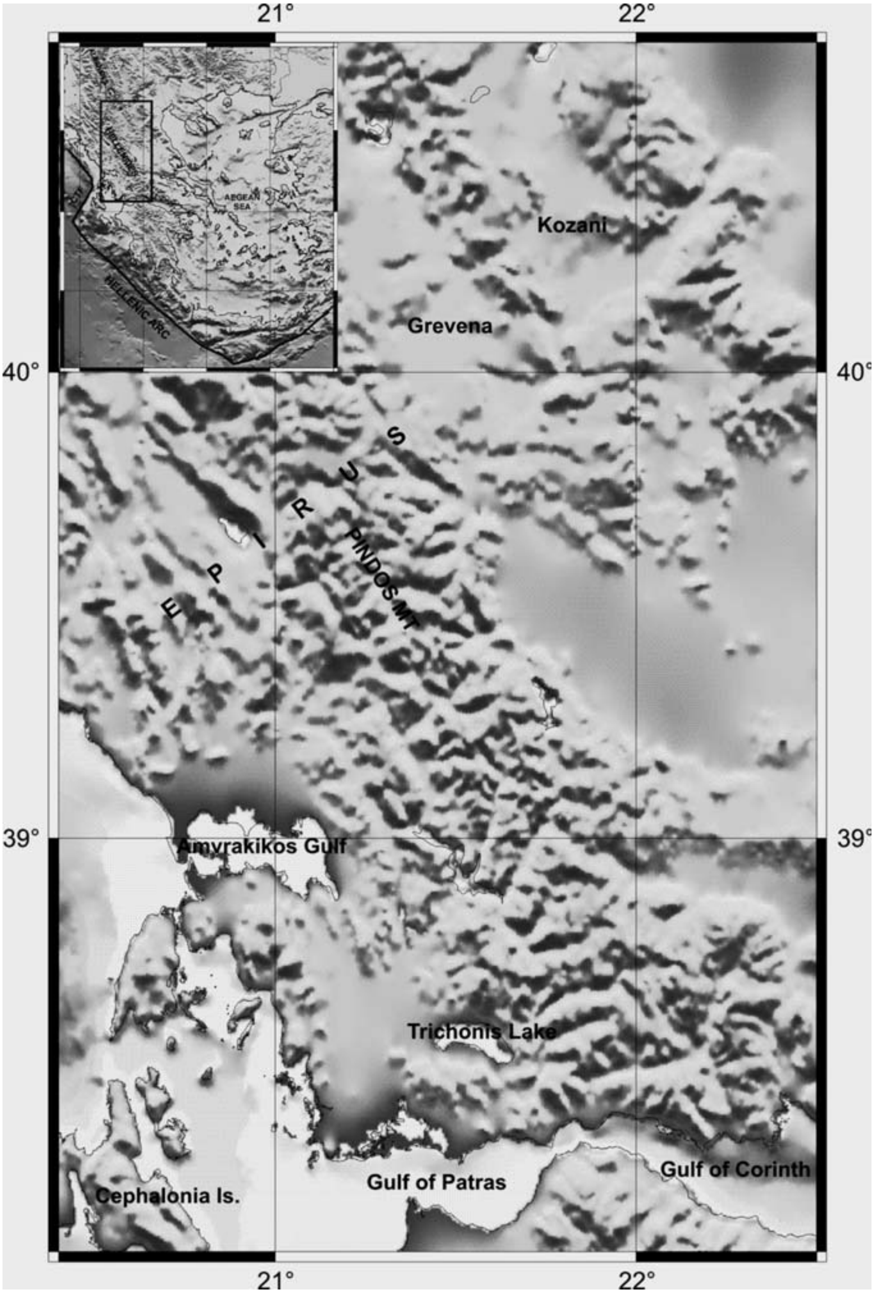

Figure 1
Location map of the study region with the localities cited in the text.

Northwestern Greece (Fig. 1) plays a pivotal role in most of the geodynamical models of current deformation in the Aegean (HATZFELD *et al.*, 1995). It is close to the pole of rotation proposed for the Hellenic trench (LE PICHON and ANGELIER, 1979; Le Pichon *et al.*, 1995) and therefore the deformation should vary spatially.

Limited or low detail information exists about the 3-D fine structure of the northwestern Greek mainland (Fig. 1). Most of this information results from tomography investigations of regional scale (SPAKMAN, 1986; SPAKMAN *et al.*, 1988; DRAKATOS and DRAKOPOULOS, 1991; LIGDAS and MAIN, 1991; PAPAZACHOS *et al.*, 1995; ALESSANDRINI *et al.*, 1997; DRAKATOS *et al.*, 1997; MELIS and TSELENTIS, 1998). In general, these studies outline the predominant features of the eastern Mediterranean region. Due to their large-scale parameterization, it is impossible to extract detailed information about the fine crustal structure of the area of interest. Moreover, on a local scale, few tomography studies exist concerning very restricted areas, which are part of the Greek mainland (DRAKATOS *et al.*, 1998; PAPAZACHOS *et al.*, 1998; HASLINGER *et al.*, 1999). Therefore, the primary aim of this study is to determine the crustal fine structure of the northwestern Greek mainland and to derive a reliable velocity model.

Tectonic and Geological Regimes

The tectonic activity in northwestern Greece is rather complex since the Miocene time (MERCIER *et al.*, 1979). Compression is seen in Epirus at or near the middle Miocene time and is likely to have been associated with the jump of the active thrusting from the Pindus to the Ionian zone (HATZFELD *et al.*, 1995). Compression occurred again around the lower Pliocene time and during the middle Pleistocene time (UNDERHILL, 1989). From paleomagnetic observations it is demonstrated that no significant rotation has affected the region during the lower Cenozoic time. On the contrary, the region has rotated as a whole since upper Oligocene time (KISSEL *et al.*, 1985).

The neotectonic regime of western Greece is governed by the relative motions between the Eurasian and African plates and the Aegean and Adriatic micro plates. Africa, Adriatic and Aegean plates form a triple junction in the Ionian Sea, where the dextral strike-slip Cephalonia transform fault (Fig. 1, CTF, in the frame) strikes the front of the Hellenic subduction zone (SACHPAZI *et al.*, 2000). The southern part of the investigated region is characterized by the existence of approximately E-W trending grabens, like the Gulf of Patras and the Gulf of Corinth (BROOKS *et al.*, 1988; ARMIJO *et al.*, 1996). As a result, large earthquakes ($M > 6.0$) are located around the Ionian Islands and along the grabens, where reverse and normal faulting are predominant, respectively. In the other regions of the western and northern Greek mainland the large earthquakes occur mainly around the pull-apart basin of the Amvrakikos Gulf and Trichonis Lake

(King *et al.*, 1993) and in the Kozani-Grevena region (Figs. 1 and 3). Finally, moderate to strong earthquakes occur at the Albanian–Greek border region.

The Hellenides Mountain chain marks the southern continuation of the Albanides and Dinarides (Auboin, 1959). Within the Albanides and Hellenides, the Pindus Mountain is located between southern Albania and the Gulf of Corinth (Fig. 1). The Hellenides, which consist of rocks mainly formed during the Mesozoic and Cenozoic times, are considered to be a typical fold and thrust belt due to the shortening of a passive continental margin. Sediments are of considerable thickness and the basement in places can be located at a depth of 10 km or more (Auboin, 1965). In general, along the Dinarides-Hellenides Mountain chain, the crust is 35–40 km thick. The thickness decreases from north to south and from west to east (Panagiotopoulos and Papazachos, 1985).

Method

In order to investigate the three-dimensional crustal velocity structure of the region (Fig. 1), the two-step tomography procedure proposed by Kissling *et al.* (1994) has been applied. The inverse problem of three-dimensional local earthquake tomography is formulated as a linear approximation to a nonlinear function (Pavlis and Booker, 1983). In general, solutions are obtained by linearization with respect to a reference earth model (Aki and Lee, 1976; Nolet, 1978). Thus the solutions obtained and the reliability estimates depend on the initial reference models. Inappropriate models may result in artifacts of significant amplitude. To overcome these problems, Kissling *et al.* (1994) proposed a two-step procedure to obtain 3-D tomographic results with minimal dependence on the reference model.

As a first step the travel-time data are inverted jointly with revised hypocenter coordinates and station corrections, to obtain a 1-D tomographic solution. This model is called the "minimum 1-D model" (Kissling, 1988). The determination of this model is a trial and error process that ideally starts with the collection and selection of *a priori* information about the subsurface structure (Makris, 1978; Calganile *et al.*, 1982; Panagiotopoulos and Papazachos, 1985). Since this process can lead to ambiguous results, particularly when more than one *a priori* 1-D model has been established, several parameters which control the inversion must be varied and the corresponding results must be evaluated (Kissling *et al.*, 1994).

As a second step, the 3-D tomography inversion is determined using the minimum 1-D model (Fig. 2) as the starting model. In this step, the tomography technique initially introduced by Thurber (1983) and later improved by Eberhart-Phillips (1986,1990) was applied. The method performs an iterative simultaneous inversion for 3-D velocity structure and hypocenter parameters using

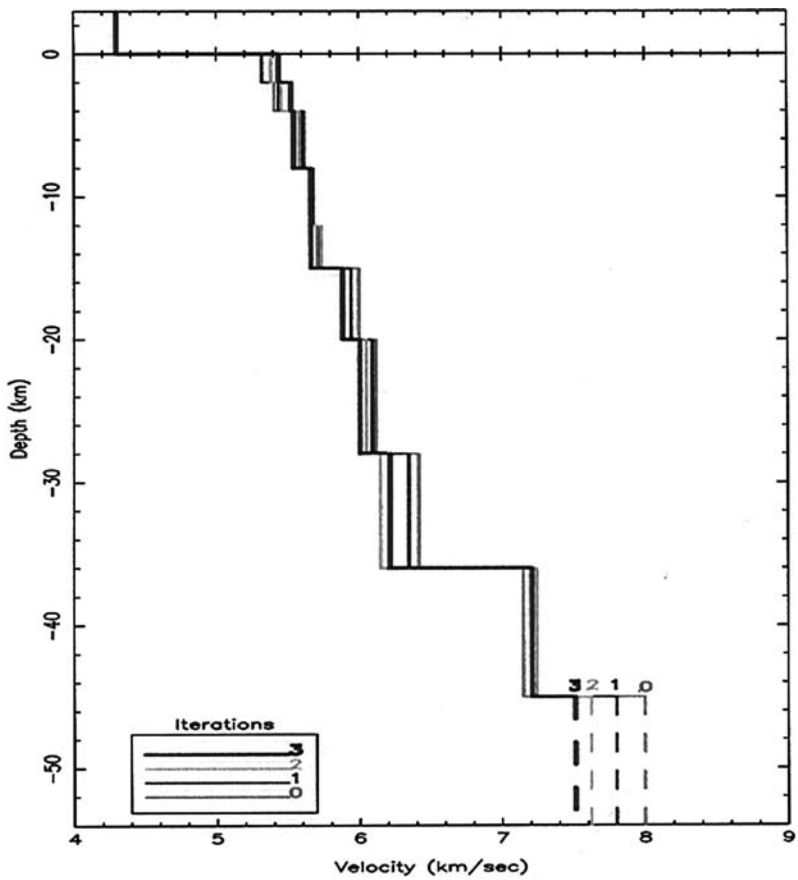

Figure 2
The obtained 1-D minimum velocity model: The numbers denote the three iterations and the final accepted model (number 3) is shown.

travel-time residuals from local earthquakes. The velocity of the medium is parameterized by assigning velocity values at the intersections (grid points) of a nonuniform three-dimensional grid. The step of the grid is selected by trying to have enough ray paths near each grid point such that its velocity is well resolved. The values of the velocity and its partial derivatives in a point along a ray path are computed by linear interpolation between the velocity values at the surrounding grid knots. Thus the obtained velocity distribution will not show sharp discontinuities, typical of refraction models or of block-type parameterizations. The approximate ray-tracing technique proposed by THURBER (1983) is used. This technique defines the ray path as the arc with the fastest travel time out of a group of circular arcs connecting the source and receiver.

Data Selection and Model Configuration

The resolution and the reliability of the tomographic results strongly depend on the density of crossing rays. This presumes a dense station network with good distribution throughout the investigated area. In order to overcome this difficulty, the data are selected from different time periods and from different networks installed in the study region. On May 13, 1995, a strong earthquake (M_s = 6.6) took place at Kozani-Grevena region (Fig. 3). Immediately after the earthquake the Institute of Geodynamics (National Observatory of Athens) and the University of Thessaloniki (Geophysical Laboratory) installed portable networks in the region. Therefore, for the time period 1995–1996, the data are provided from the above-mentioned networks, as well as from the permanent network of the Public Power Corporation (PPC) of Greece. For the time period 1997–1999 the data used are those recorded by the PPC network. The combined use of data from different seismic networks does not affect the results of studies concerning the structure of the Earth (e.g., DRAKATOS and DRAKOPOULOS, 1991; PAPAZACHOS *et al.*, 1995; DRAKATOS *et al.*, 2002). The networks operated with instruments of the same type and, therefore, there are no problems of data compatibility. 1,500 earthquakes were initially selected for the

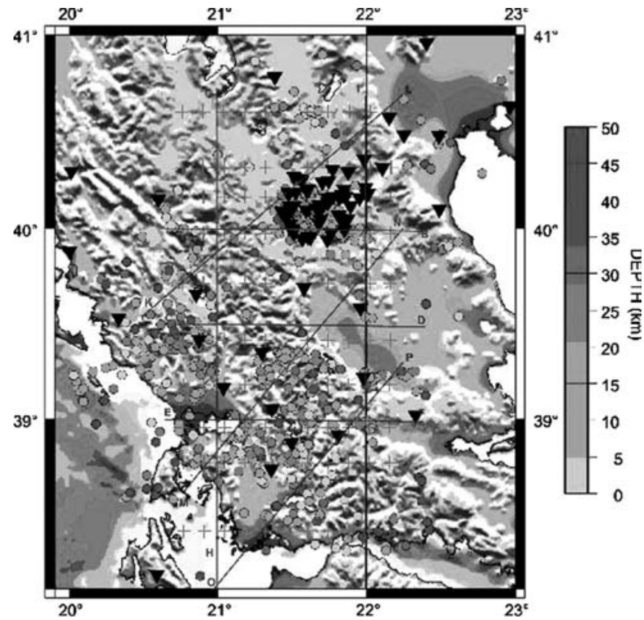

Figure 3
Distribution of epicenters (shown with solid circles and a gray scale for the associated depth) and station sites (shown with black solid triangles) are plotted. The crosses represent the tomography grid used in this study. The bold lines are the cross sections of Figure 6. The epicenter (star) of Kozani-Grevena earthquake is also shown.

study region and a final set of 584 very well located events was concluded following the criteria: each one of them is recorded by at least ten stations and the RMS is less than 0.5 (Fig. 3). The relatively large number of events concentrated in the central part of the region outlines the aftershock area of 13th May, 1995 Kozani-Grevena earthquake.

For the implementation of the method, seven horizontal depths at 0.5, 4, 7, 10, 13, 16 and 20 km have been considered in the area between 38.30 ^{o}N–40.70 ^{o}N and 20.65 ^{o}E–22.25 ^{o}E. At each depth the grid consists of ten nodes both in the E-W and N-S directions (Fig. 3 and Table 1). The distance between two consequent nodes is 12 km in the E-W direction and varies from 16 km to 30 km in the N-S direction. Therefore, the dimensions of the grid are 120 $\times$ 245 km in E-W and N-S directions, respectively. The initial velocity values assigned at each horizontal plane of nodes are 5.44, 5.54, 5.65, 5.74, 5.82, 5.89, and 6.01 km/s, respectively. A total number of 6,981 *P* arrivals was used. Some of the events are located outside the modeled area. The inclusion of earthquakes and /or stations located outside the study area is necessary to improve the ray-path distribution, the quality of the hypocenter location and the resolution of the velocity (EBERHART-PHILLIPS, 1986, 1990).

A relocation of the events was made before the application of the inversion procedure. With the purpose to achieve the best depth determination, in the relocation step we included not only *P*-wave but also *S*-wave arrivals. Totally, 10,626 arrivals (6,981 *P*-wave and 3,645 S-wave readings) were considered. The data set has been inverted three times to create a stable solution.

Results

After the final relocation of the events the RMS values were improved significantly and most of the events have RMS value varying from 0.1 to 0.3 s (Fig. 4a). The magnitude and the focal depth distributions are also shown in Figure 4b and c. The results of the inversion are displayed in Figure 5, in terms of velocity values at each node and only for the well-resolved nodes. Figure 6 shows the

Table 1

Grid Geometry

Depth (km)	Nodes (E-W)	Nodes (N-S)	Initial Velocity (kms^{-1})
0.5	10	10	5.44
4.0	10	10	5.54
7.0	10	10	5.65
10.0	10	10	5.74
13.0	10	10	5.82
16.0	10	10	5.89
20.0	10	10	6.01

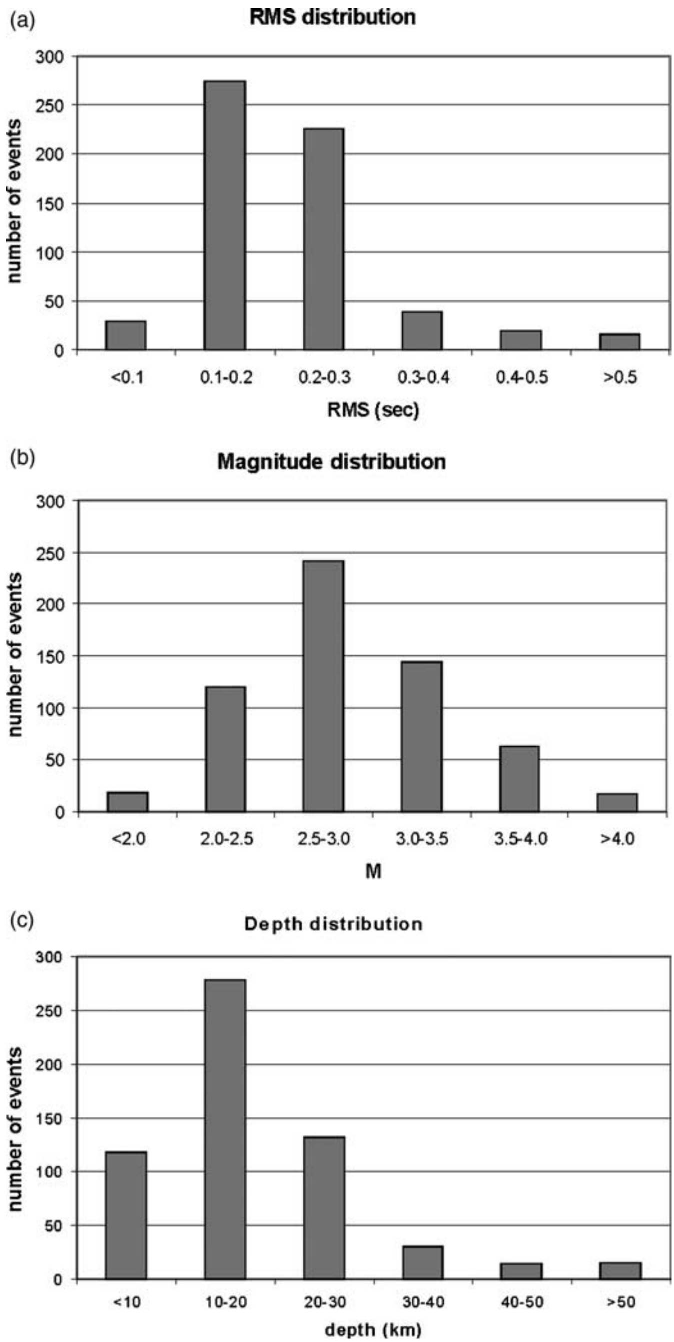

Figure 4
Histograms of RMS (a) magnitude (b) and depth (c) distributions.

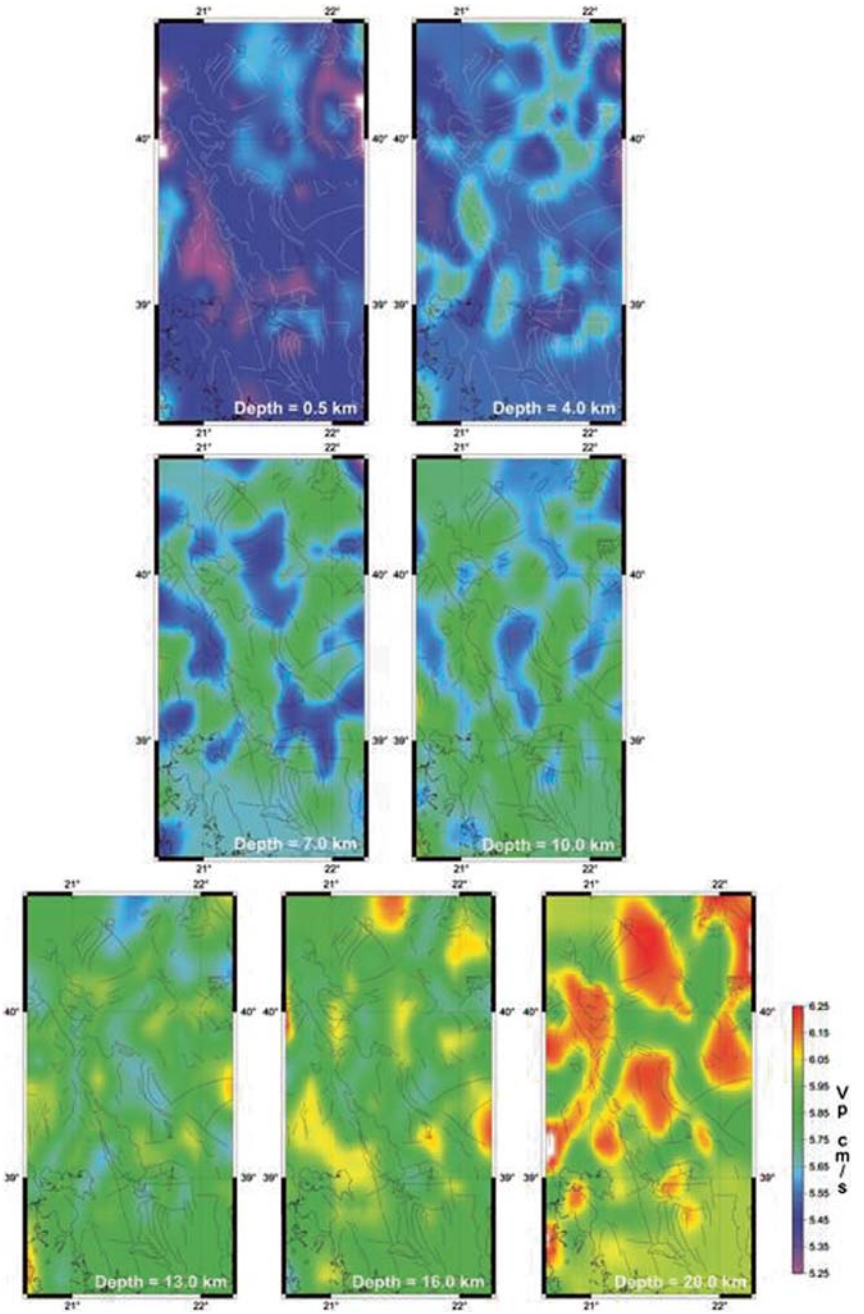

Figure 5
P-wave velocity distribution obtained after 3-D simultaneous inversion. Depth (0.5, 4, 7, 10, 13, 16 and 20 km) denotes the layer where grid nodes were applied. Major tectonic elements are placed over the velocity distribution at each layer.

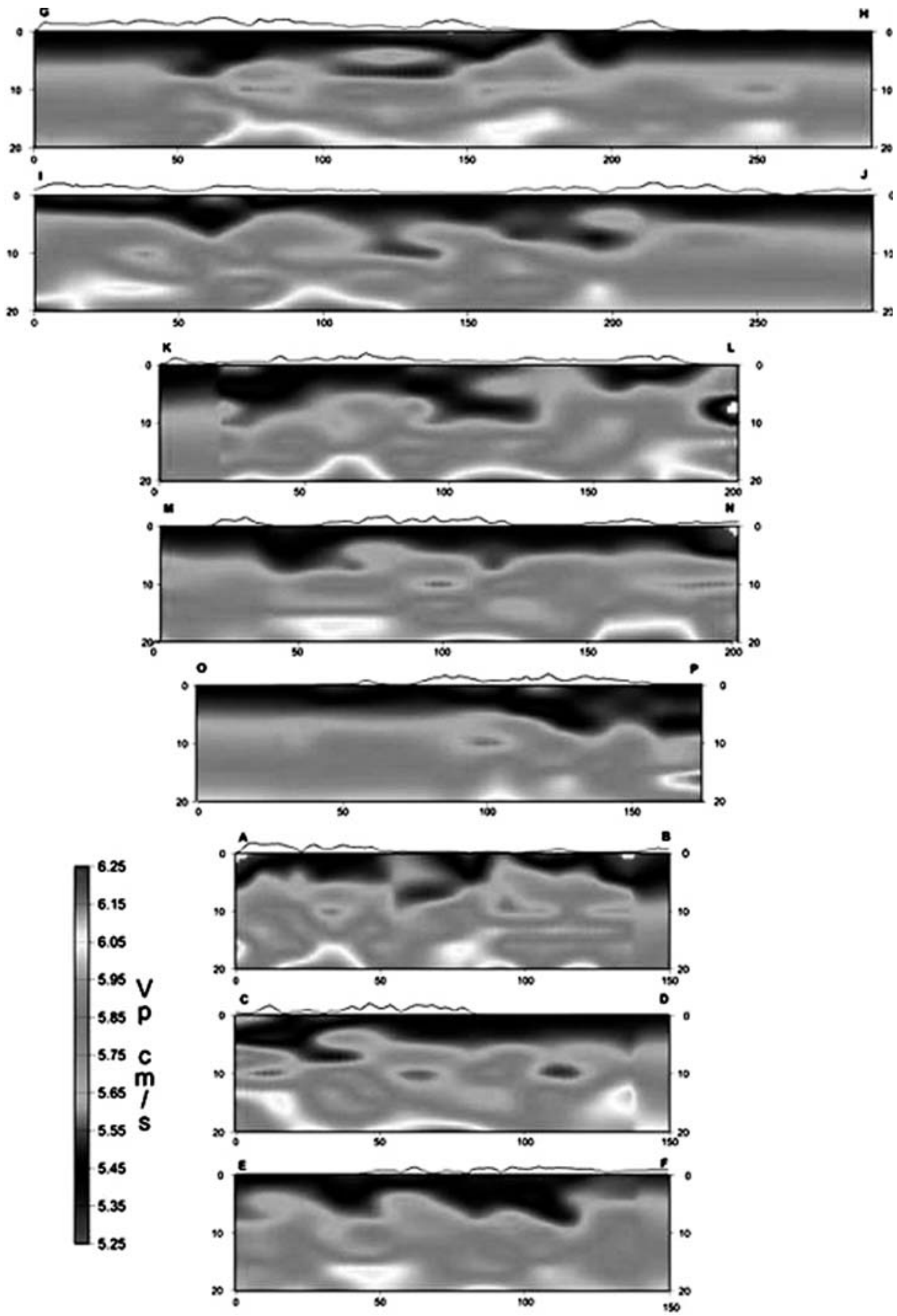

Figure 6
P-wave velocity distribution along the cross-sections, shown in Figure 3.

velocity distribution along the cross sections (bold lines) of Figure 3. Several parameters, as for example the number of hits at each block and the sum of derivatives, can be used to check the reliability and stability of the solution: the most reliable is the diagonal element of resolution matrix. The resolution, the reliability and the stability of the solution strongly depend upon the degree of intersection of crossing rays. This presumes a dense station network with good distribution over the investigated area. Despite the relatively low number of stations and their uneven distribution, the resolution values are high (Fig. 7) due to the fine distribution of the selected events and the small area of the investigated region. As a result, a stable and unique solution in the major part of the investigated region is found.

The results (Figs. 5 and 6) show a rather complicated velocity distribution. The velocity at the shallower depths seems to be affected from the surface geology. Low velocities are predominant along the Dinarides-Hellenides Mountain chain, trending NE-SW. At the southern part of the region (roughly below the parallel of 39 °) the velocity increases gradually with respect to the depth, ranging from 5.25 to 6.0 km/s. These velocity values are typical for the upper crust in the region (PAPAZACHOS *et al.*, 1995). A first well-defined velocity discontinuity appears at a depth of 7 km (Fig. 6, cross sections KL, MN, OP, EF). The velocity remains almost stable down to a depth of 10 km and then a pronounced decrease is shown at a depth of 16 km. At a depth of 20 km, the velocity values become typical for the upper crust. At the northern part of the investigated region high velocities are predominant, ranging from 5.25 to 6.25 km/s.

Discussion – Conclusions

Data from several field experiments were combined to compute for the first time a reliable three-dimensional *P*-wave velocity model for the crust of the northwestern Greek mainland. The region is characterized by varying crustal thickness, which decreases from NW to SE (PANAGIOTOPOULOS and PAPAZACHOS, 1985). The presence of three different geological zones (from west to east: Ionian, Gavrovo, Pindos) underline the complex seismotectonic and geological regime of this region.

The obtained "minimum 1-D model" for the *P*-wave crustal velocity is rather similar to the one obtained by HASLINGER *et al.* (1999), taking into consideration that the last one refers to a restricted area (Amvrakikos Gulf), which is part of the study region. The velocities of the obtained model are typical for the crust.

The velocity distribution in the shallow layers (4 and 7 km) is strongly affected by the crustal thickness variation and the complex tectonics. A first, a well-defined velocity discontinuity (5.65 km^{-1}) appears at a depth of 3–6 km along the Hellenides Mountain chain (Fig. 6, cross sections GH and IJ). A second low velocity anomaly ($<$ 6.0 km^{-1}) is detected at a depth of 9–12 km. A similar anomaly was determined by PAPAZACHOS *et al.* (1995). According to them, the existence of a low velocity layer in

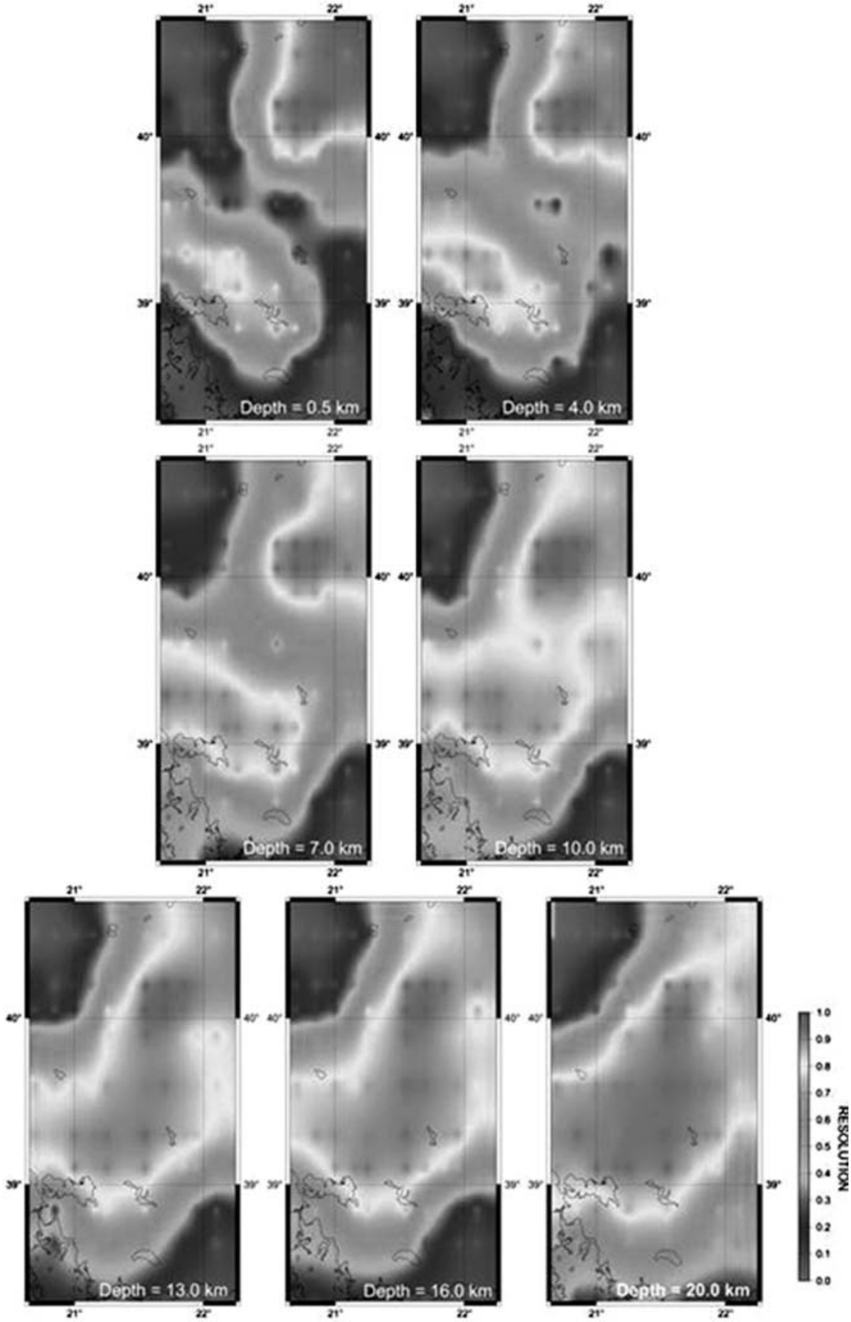

Figure 7
The values of the diagonal elements of the resolution matrix, corresponding to the layers where the grid nodes were applied.

this region indicates a possible connection with the Alpidic orogenesis. In areas of crust thickening and vertical uplifts, such as the mountain chains, the low velocity layer is due to granitic or other intrusions in the crust during the uplift (MUELLER, 1977; LABHART, 1975). MONOPOLIS and BRUNETON (1971) note the existence of thick Triassic formations in the Ionian geological zone. These evaporates toward the northeast under the inner geological zones. This could explain the alternations of low and high velocity zones (in an west - east direction) shown in the cross sections AB and CD in Figure 6.

Another interesting feature appears beneath the Amvrakikos Gulf, where low velocities are predominant (cross sections GH, MN and EF). The same velocity distribution was determined by HASLINGER *et al.* (1999) in the southern part of the Amvrakikos Gulf, characterized by a horst graben structure.

Finally, the present study confirms that the *P*-wave crustal velocity structure in the northern and western parts of the Greek mainland projects strong horizontal variations as a result of the complicated crustal structure.

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