Pergamon

J. Geodynamics Vol. 26, No. 2-4, pp. 341-351, 1998

PII: S0264-3707(97)00061-6
0264-3707/98 $19.00+0.00

OBSERVATIONS ON THE 3-D CRUSTAL VELOCITY STRUCTURE IN THE KOZANI–GREVENA (NW GREECE) AREA

G. DRAKATOS,[1] D. PAPANASTASSIOU,[1] N. VOULGARIS[2] and G. STAVRAKAKIS[1]

[1]National Observatory of Athens, Institute of Geodynamics, P.O. Box 20048, GR 118 10 Athens, Greece
[2]University of Athens, Department of Geology, Geophysics-Geothermy Division, Panepistimiopolis, 157 84 Ilissia, Greece

(Received 18 April 1997; accepted 13 August 1997)

Abstract—Three-dimensional velocity structure of the upper crust was determined by inversion of P-waves travel time in the area of the Aliakmon river basin, where a strong earthquake of Ms = 6.6 occurred on May 13, 1995. The area is located at the northwestern part of the Greek mainland.

To investigate the 3-D crustal velocity structure of the region, a two-step tomographic procedure has been applied. The data set consists of the travel time residuals of 366 very well located earthquakes, recorded by at least seven stations of a portable network, which was installed at the Kozani–Grevena broader region by the Institute of Geodynamics, immediately after the shock of 13 May, 1995. In order to improve the initial (reference) velocity model, before the inversion of the travel times, the "minimum 1-D" initial velocity model was obtained.

The results show that at shallow depths low velocities are predominant in the whole region and there is no sharp horizontal velocity variation. The velocity at shallow depths seems to be affected from the surface geology. Despite the absence of large scale seismotectonics features in the region there is a remarkable vertical increase of the velocity below the depth of 15 Km, supporting the existence of a second order discontinuity in the crust of the region. © 1998 Published by Elsevier Science Ltd. All rights reserved

GEOLOGICAL–SEISMOTECTONIC REGIME

The Kozani–Grevena broader region is located in the northwestern part of Greek mainland (Fig. 1) and has been considered as an area of low seismicity. The main geological formations of the area are the metamorphic rocks of Mesozoic age uncomformably covered by molassic sediments of the Meso–Hellenic basin, while over them exist sediments of Neogene and younger age (Robert *et al.*, 1985). Aliakmonas valley crosses two different geological zones, the Pelagonic zone at the west and the Vardar zone at the east. The several normal faults

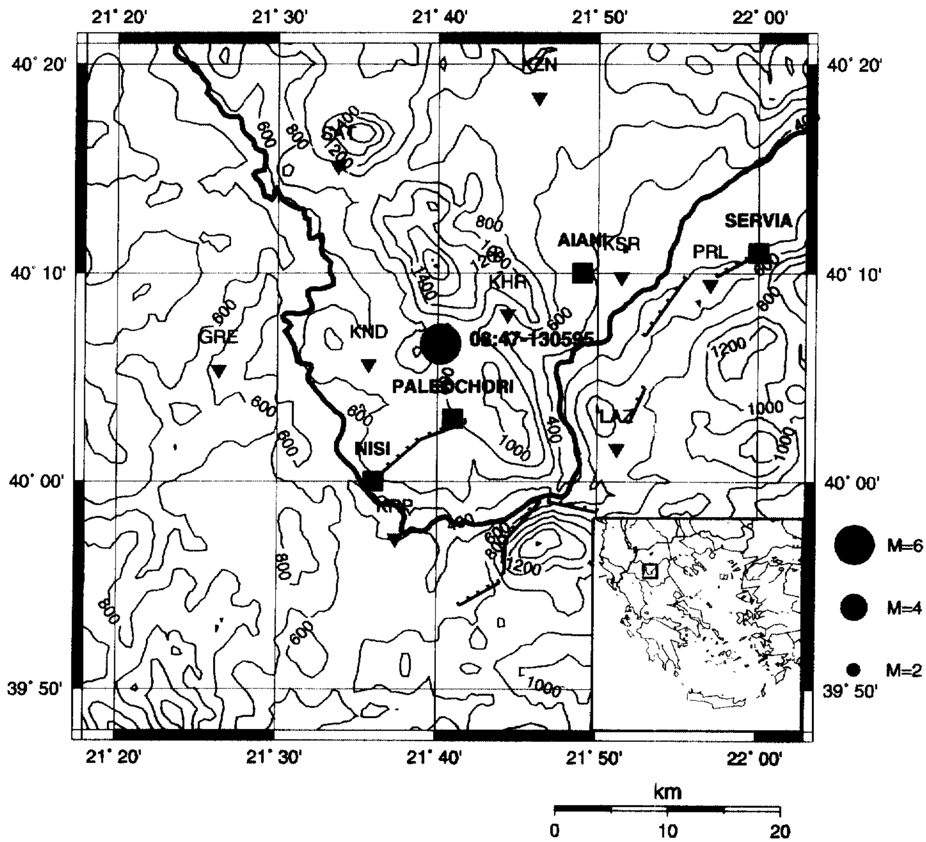

Fig. 1. The Kozani–Grevena region is shown. The triangles denote seismological stations. The epicenter of the May 13, 1995 earthquake is shown.

in the area are characterized by a predominant orientation NNE–SSW (Mountrakis, 1984; Papazachos *et al.*, 1986; Pavlides and Mountrakis, 1987; Caputo and Pavlides, 1993).

The overall crustal and upper mantle structure in the eastern Mediterranean region was studied by many researchers, who mainly focused on the lithosphere-asthenosphere system and on the upper mantle. The crustal thickness beneath the Aegean Sea is estimated, through the analysis of surface wave dispersion, to about 35 km, under the Greek peninsula between 36 km to 42 km and under Macedonian and southern Yugoslavia between 31 km and 47 km (Calcagnile *et al.*, 1982). The lower part of the crust exhibits higher velocities than the upper crust (Makris, 1978a). The Bouguer anomalies strike parallel to the NNW–SSE trend of the Greek mainland and positive values are predominant in the broader area of the Aliakmon river basin (Makris, 1978b).

In the last decade, many investigators studied the eastern Mediterranean area using tomography techniques (Spakman, 1988; Spakman *et al.*, 1988, 1993; Hashida *et al.*, 1988; Wortel *et al.*, 1990; Drakatos and Drakopoulos, 1991; De Jonge *et al.*, 1993; Papazachos, 1994; Papazachos *et al.*, 1995; Alessandrini *et al.*, 1997; Stavrakakis *et al.*, 1997; Drakatos *et al.*, 1997).

In all the above mentioned studies the major features of the Aegean region have been obtained, but in general very few papers can be found on the local scale characteristics of the seismic waves velocity in the first tens kilometers. Drakatos *et al.* (1988) investigated for the first time the velocity structure of the region, in terms of tomography technique, using regional data. The present study is a first attempt to investigate the fine structure of the upper crust of the region included between 39.90°N–40.40°N and 21.30°E–22.00°E, using time residuals of P-wave arrivals of local data.

METHOD

In order to investigate the three-dimensional crustal velocity structure of the region, 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 model. 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 of the reference model.

As a first step, the travel time data are jointly inverted to obtain a 1-D tomographic solution, together with revised hypocentre coordinates and station corrections. 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. Since this process can lead to ambiguous results, particularly when more than one a-priori 1-D models have been established, several parameters which control the inversion need to be varied and the corresponding results need to be evaluated (Kissling *et al.*, 1994).

As a second step, the 3-D tomographic inversion is determined using the minimum 1-D model as the starting model. In this step, the tomographic technique initially introduced by Thurber (1983) and later improved by Eberhart-Phillips (1986, 1990) was applied. The velocity of the medium is parameterized by assigning velocity values at the intersections (grid points) of a nonuniform, three-dimensional grid. The spacing within the grid is selected by trying to have enough ray paths near each grid point so that its velocity may be well resolved. The spacing need not be uniform throughout the study area. The velocity for a point along a ray path and the velocity partial derivatives are computed by linearly interpolating between the surrounding grid points. Thus, the velocity solution will show gradational changes in velocity rather than the sharp discontinuities shown in typical refraction models or block-type parameterizations. As a ray tracing technique, the approximate ray tracing proposed by Thurber (1983) is used. This technique selects the ray paths as the arc with the fastest travel time out of a suite of circular arcs connecting the source and receiver. This is a useful technique but it puts limits on the application of the inversion method. At large distances where a refracted wave may be the first arrival, a circular arc is a poor approximation and overestimates the path length. This technique performs very well for station to source distances less than 100 km (Thurber, 1983; Eberhart-Phillips, 1986, 1990), therefore is the most adequate to be applied in the case of Kozani–Grevena region.

DATA AND RESULTS

Immediately after the main shock of May 13th 1995, the Institute of Geodynamics (I.G.), deployed a portable network of nine stations in the region. Together with the permanent station of Kozani, the network covered the greater part of the aftershock area. Hundreds of aftershocks have been recorded. The data set consists of the arrival times of 366 very well located earthquakes (Fig. 2) each one of them recorded by at least seven stations of the network.

For the implementation of the method, a set of nodes has been considered, in the area between 39.90°N–40.35°N and 21.30°E–22.00°E, distributed in four horizontal layers at a depth of 2, 5, 11 and 17 km (6 nodes in the E–W and 5 nodes in the N–S direction, respectively). The distance between two consequent nodes is 10 and 14 km in EW and NS direction, respectively. The initial velocity values assigned at each horizontal plane of nodes are 4.25, 4.99. 5.34 and 6.17 km s^{-1}, respectively. The velocity model used is the "minimum

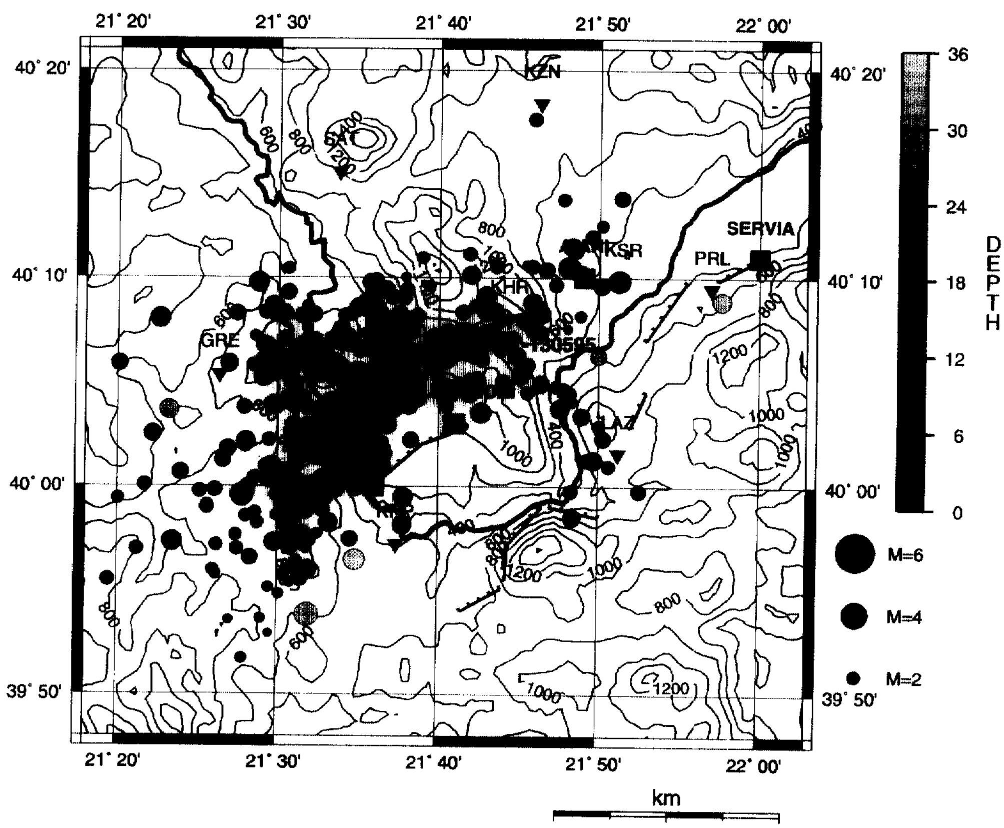

Fig. 2. The epicenter distribution of earthquakes located in the investigated area.

1-D model" (Fig. 3), which has been obtained following the procedure proposed by Kissling *et al.* (1994). The damping factor assigned to the velocity during the inversion procedure was set to be 15%.

The total number of observations used is 4033. Some of the events are located outside the modeled area. The inclusion of earthquakes and (or) stations outside the modeled area is necessary to improve the ray path distribution within the modeled area. If only events and stations within the modeled area were included, the quality of hypocentre location and

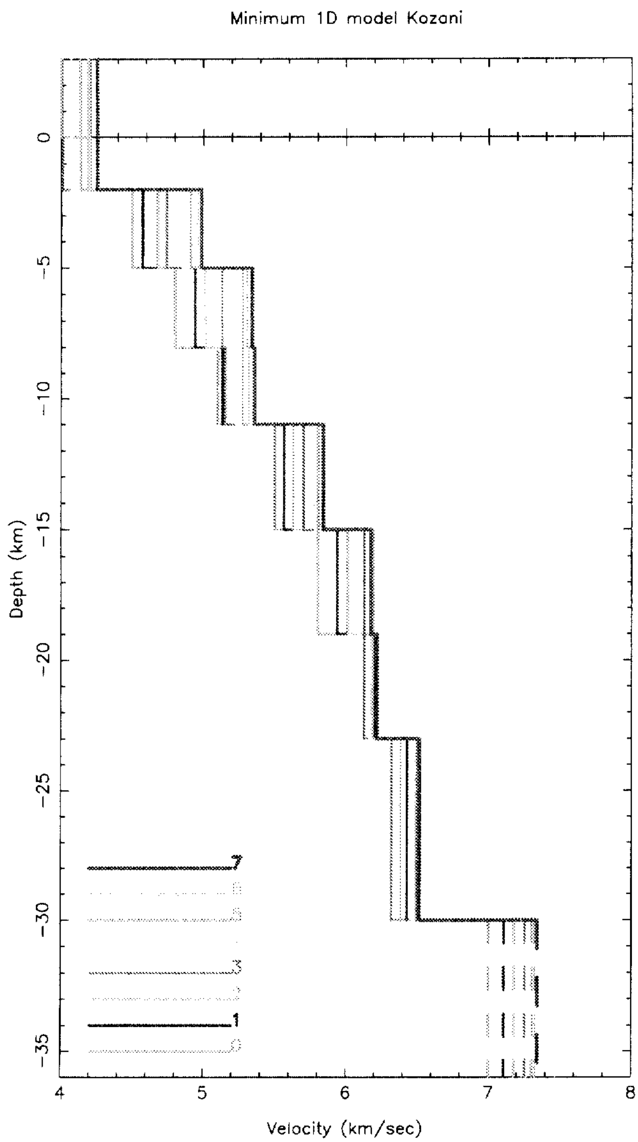

Fig. 3. The "minimum 1-D model" is shown (No 7).

resolution of velocity would be significantly reduced. However, it must be remembered in interpreting the results that the peripheral velocity gridpoints include ray paths from the surrounding area.

Before the inversion procedure, the relocation of events has been made. In order to achieve the best depth determination, we included in the relocation step except the P-wave arrival times and the S-wave arrivals too. The data set has been inverted five times to get a stable solution.

The effective number of nodes included in the inversion procedure is 67. The computed velocity value of the other nodes was constrained since the number of observations in those nodes was small.

At the depth of 2 Km low velocities cover the whole area (Fig. 4) but the peripheral regions are covered by relatively higher velocities. At the depth of 5 and 11 Km (Fig. 4) the velocity values are higher than that of the first layer, while at the depth of 17 Km (Fig. 4) higher velocities cover the investigated area.

To check the reliability and the stability of the solution, the resolution matrix is calculated. The results are shown in Fig. 5. Solutions with a diagonal element value greater than 0.60 are considered stable and reliable. Therefore, the results of this study are supported from the high resolution values. Even in some peripheral nodes the values of diagonal elements are high. This is due to the inclusion in the data set of several events out of the investigated area.

An other numerical test of the solution "quality" is the covariance matrix. So, in Fig. 6 the standard errors are shown.

DISCUSSION—CONCLUSIONS

The 3-D velocity structure of the upper crust of the Kozani–Grevena region is determined, by inversion of the arrival times of local earthquakes. Due to the small extent of the investigated region, no sharp variations of velocity are expected.

At the depth of 2 Km, an extended low velocity area covers the most of the region. The velocity values range from 4.4 to 4.7 Km s^{-1}. The low velocity values at this depth seem to be typical for a sedimentary basin like the investigated area.

The situation changes at the depth of 5 Km. At the central part of the region relatively high velocities are predominant. The velocity values range from 5.3 to 5.7 Km s^{-1}, while in the eastern part lower velocities appear as well as in the southeastern and northwestern parts of the region. Thus, the velocity fluctuations are affected not only by the vertical gradient but also from lateral heterogeneities. The relatively high velocities in the central part are affected from Vourinos massif. In the third grid of nodes, at the depth of 11 Km, the velocity values in the central part of the region are exactly the same as in the depth of 5 Km. Similar velocity values are also proposed, for the upper continental crust of the region by Makris (1978b) and Drakatos *et al.* (1988).

Due to the very small area of the investigated region and the different model parameterization, a direct comparison between the results of this study and previous tomography investigations is not adequate. However, the computed velocity values are in general close to those determined for the upper crust of northwestern Macedonia by Papazachos *et al.* (1995); Alessandrini *et al.* (1997) and by Drakatos *et al.* (1997).

The velocity distribution determined at the depth of 17 Km should be examined carefully, because due to the lack of deep events and therefore the small number of seismic rays

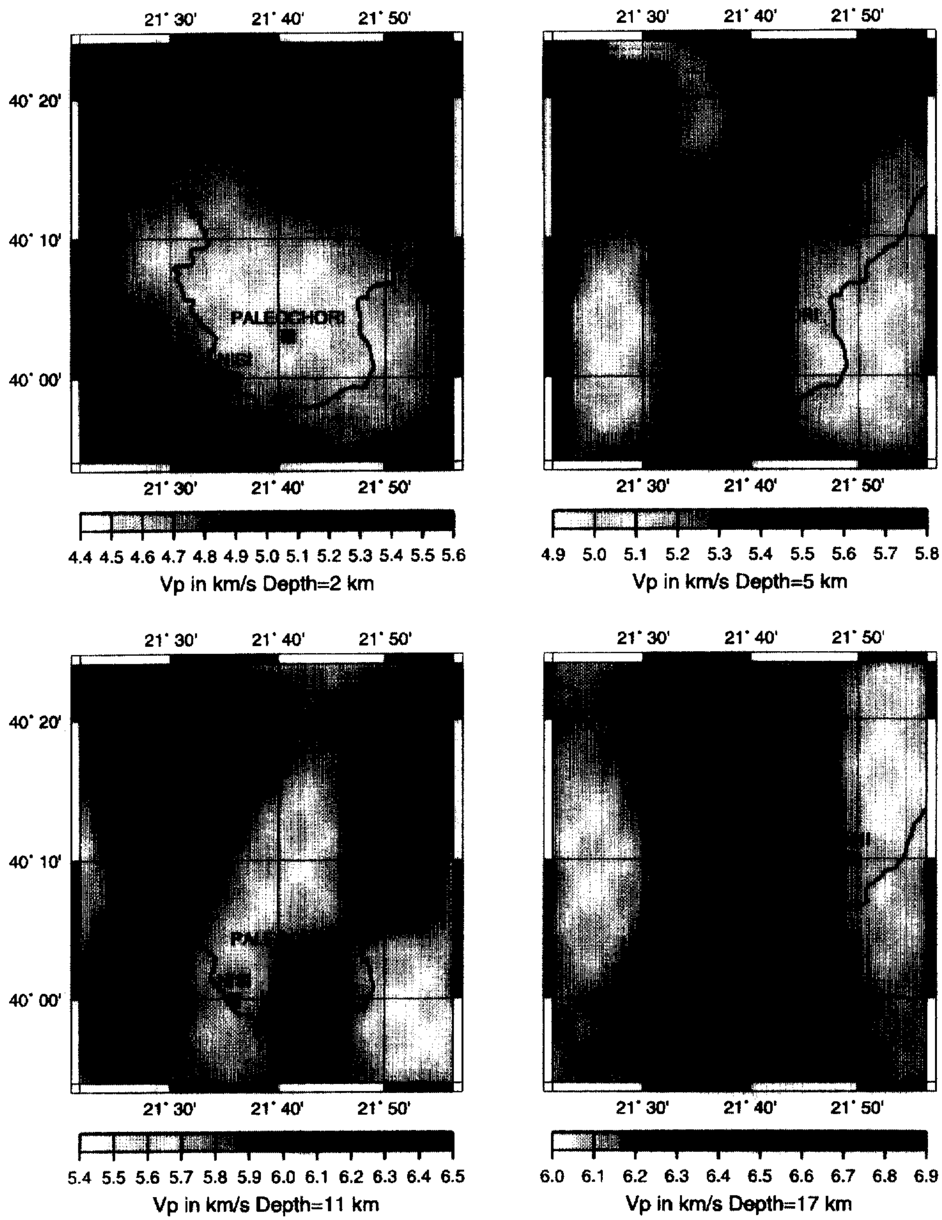

Fig. 4. The 3-D structure of the crust is shown, in terms of velocity values at the four grids of nodes.

crossing the peripheral nodes, could be artifacts. The most remarkable feature at this depth is the sharp velocity variation with respect to the previous layers. There is a sharp velocity discontinuity at this depth. The same discontinuity has been determined by Drakatos *et al.* (1988). They claimed the existence of a Conrad discontinuity in the region. According to Makris (1978a), the lower part of the crust exhibits higher velocities than the upper part,

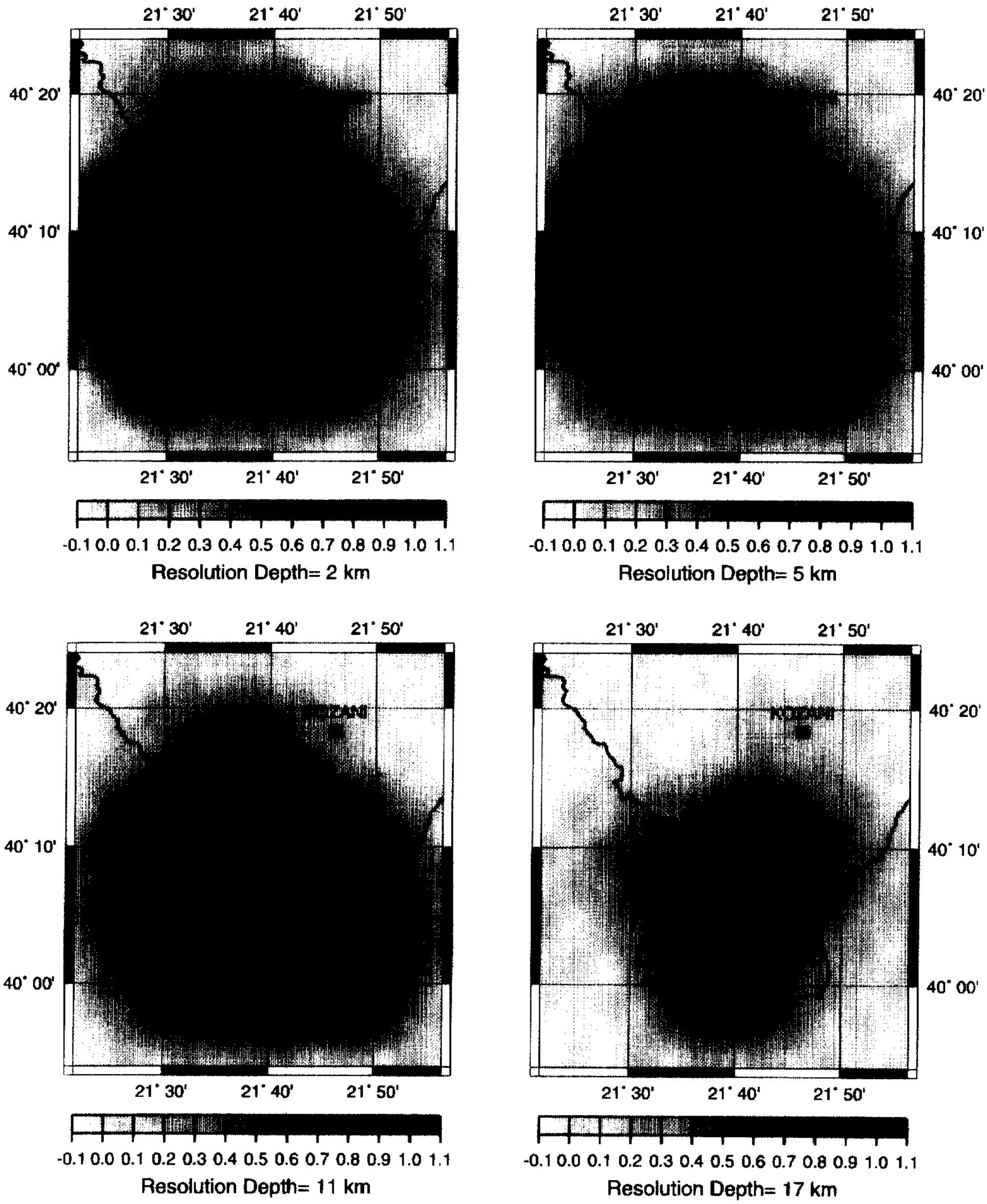

Fig. 5. The diagonal elements of resolution matrix are shown for each grid of nodes.

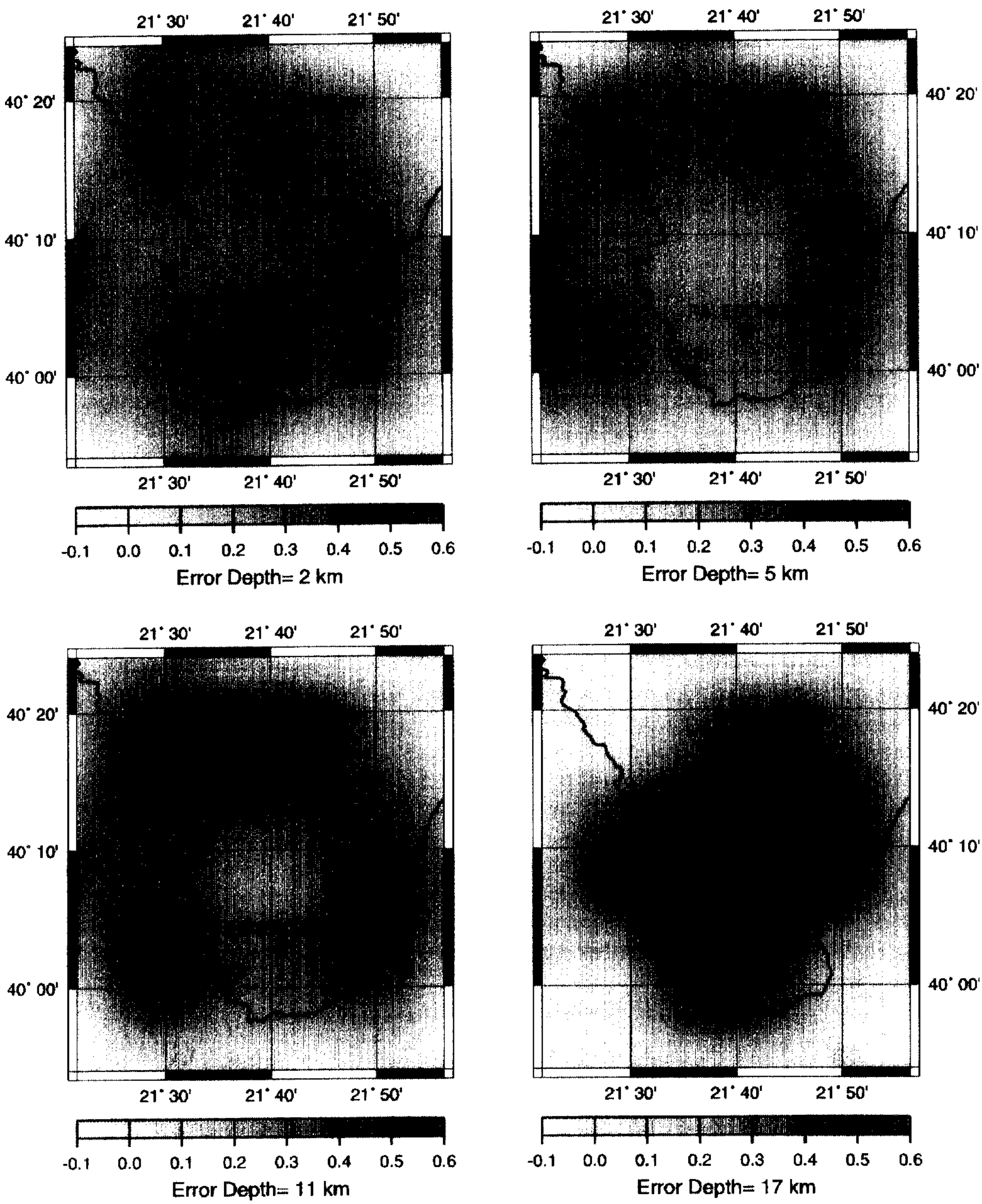

Fig. 6. The covariance matrix is shown.

but a Conrad discontinuity has not been observed as a first order discontinuity within the crust of the region.

REFERENCES

Aki, K. and Lee, W. H. K. (1976) Determination of three-dimensional velocity anomalies under a seismic array using first P arrival times from local earthquakes, 1. A homogeneous initial model. *J. Geophys. Res.* **81,** 4381–4399.

Alessandrini, B., Beranzoli, L., Falcone, C., Mele, F., Drakatos, G., Karantonis, G. and G. Stavrakakis (1997) Back arcs basins and P-wave crustal velocity in the Ionian and Aegean regions. *Geophys. Res. Lett.* **24,** 527–530.

Calcagnile, G., D'Ingeo, F., Farrugia, P. and Panza, G. F. (1982) The lithosphere in the central-eastern Mediterranean area. *Pure Appl. Geophys.* **120,** 389–406.

Caputo, R. and Pavlides, S. (1993) Late Cainozoic gedynamic evolution of Thessaly and surroundings (central-northern Greece). *Tectonophysics* **223,** 339–362.

De Jonge, M. R., Wortel, M. J. R. and Spakman, W. (1993) From tectonic reconstruction to upper mantle model: An application to the Alpine-Mediterranean region. *Tectonophysics* **223,** 53–65.

Drakatos, G., Latoussakis, J., Stavrakakis, G., Papanastassiou, J., Drakopoulos, J. and Bourboulis, A. (1988) Examination of a velocity model using 3-D velocity structure technique. *Proceed. of the Workshop on Induced seismicity and associated phenomena.* Czech. Acad. Sci., Liblice, 1988. pp. 403–421.

Drakatos, G. and Drakopoulos, J. (1991) 3-D velocity structure beneath the crust and upper mantle of the Aegean sea region. *Pure Appl. Geophys.* **135,** 401–420.

Drakatos, G., Karantonis, G. and Stavrakakis, G. (1997) P-wave Crustal tomography of Greece with use of an accurate two-point ray tracer. *Annali di Geof.* **XL**(1), 25–36.

Eberhart-Phillips, D. (1986) Three-dimensional velocity structure in northern California coast ranges from inversion of local earthquake arrival times. *Bull. Seism. Soc. Am.* **76**(4), 1025–1052.

Eberhart-Phillips, D. (1990) Three-dimensional P and S velocity structure in the Coalinga region, California. *J. Geophys. Res.* **95,** 15343–15363.

Hashida, T., Stavrakakis, G. and Shimazaki, K. (1988) Three-dimensional seismic attenuation structure beneath the Aegean region and its tectonic implication. *Tectonophysics* **145,** 43–54.

Kissling, E. (1988) Geotomography with local earthquake data. *Rev. Geophys.* **26,** 659–698.

Kissling, E., Ellsworth, W. L., Eberhart-Phillips, D. and Kradolfer, U. (1994) Initial reference models in local earthquake tomography. *J. Geophys. Res.* **99**(B10), 19635–19646.

Makris, J. (1978) Some geophysical considerations on the geodynamic situation in Greece. *Tectonophysics* **46,** 251–268.

Makris, J. (1978) The crust and upper mantle of the Aegean region from deep seismic soundings. *Tectonophysics* **46,** 269–284.

Mountrakis, D. (1984) Structural evolution of the Pelagonian Zone in Northwestern Macedonia, Greece. In *The Geological Evolution of the Eastern Mediterranean*, ed. J. E. Dixon and A. H. F. Robertson, Vol. 17, pp. 581-591. Spec. Publs geol. Soc. Lond.

Nolet, G. (1978) Simultaneous inversion of seismic data. *Geophys. J. Roy. Aston. Soc.* **55,** 679–691.

Papazachos, B. C., Kiratzi, A. A., Hatzidimitriou, P. M. and Karakostas, B. G. (1986) *Seismotectonic Properties of the Aegean Area that restrict Valid Geodynamic models. 2nd Wegener Conference*, Dionysos, Greece 14-16 May, 1986, pp. 1–16.

Papazachos, C. B. (1994) Structure of the crust and upper mantle in SE Europe by inversion of seismic and gravimetric data (in Greek). Ph.D Thesis, 208 pp. , Univ. of Thessaloniki, Thessaloniki.

Papazachos, C. B., Hatzidimitriou, P. M., Panagiotopoulos, D. G. and Tsokas, G. N. (1995) Tomography of the crust and upper mantle in southeast Europe. *J. Geophys. Res.* **100**(B7), 12405–12422.

Pavlides, S. B. and Mountrakis, D. M. (1987) Extensional tectonics of northwestern Macedonia, Greeece, since the late Miocene. *J. Struct. Geol.* **9,** 385–392.

Pavlis, G. L. and Booker, J. (1983) A study of the importance on nonlinearity in the inversion of eartquake arrival time data for velocity structure. *J. Geophys. Res.* **88,** 5047–5055.

Robert, S., Rassios, A., Wright, L., Vacondios, I., Vrachatis, G., Grivas, E. and R. Nesbi H. (1985) *Structural controls on the location and form of the Vourinos chromite deposits Mineral Deposits within the European Community*, ed. J. Boissonas and P. Omenetto. Spec. Publ. Soc. Geol. Appl. to Min. Dep., 6, pp. 246–249.

Spakman, W. (1988) Upper mantle delay time tomography with an application to the collision zone of the Eurasian, African and Arabian plates. Ph. D. thesis, 200 pp. , Univ. of Utrecht, Utrecht, Netherlands.

Spakman, W., Wortel, M. J. R. and Vlaar, N. J. (1988) The Hellenic subduction zone: A tomographic image and its dynamic implications. *Geophys. Res. Lett.* **15,** 60–63.

Spakman, W., Van del Lee, S. and Van der Hilst, R. D. (1993) Travel-time tomography of the European-Mediterranean mantle down to 1400Km. *Phys. Earth Planet. Inter.* **79,** 3–74.

Stavrakakis, G., Drakatos, G., Karantonis, G. and Papanastassiou, D. (1983) A Tomography image of the Aegean region (Greece) derived from inversion of macroseismic intensity data. *Annali di Geof.* **XL**(1), 99–110.

Thurber, C. H. (1983) Earthquake locations and three-dimensional structure in the Coyote Lake area, Central California. *J. Geophys. Res.* **88,** 8226–8236.

Wortel, M. J. R., Goes, S. B. D. and Spakman, W. (1990) Structure and seismicity of the Aegean subduction zone. *Terra Nova* **2,** 554–562.