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2-D and 3-D velocity patterns in southeastern Europe, Asia Minor and the eastern Mediterranean from seismological data

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Summary. 2-D group velocity and 3-D *P*-wave velocity patterns have been obtained from analyses of the group velocity dispersion of Rayleigh waves in south-eastern Europe and Asia Minor and of the *P*-wave travel-time residuals along ray-paths with a penetration depth of 100–300 km in southeastern Europe, Asia Minor and the eastern Mediterranean. The inversion procedure used is based on Backus–Gilbert formalism for linear inversion of travel-time data extended to 2-D and 3-D inhomogeneous media. Group velocity distributions have been obtained for periods $T = 10, 20, 30$ s and are in a very good agreement with the well-known characteristics of the crust and upper mantle structure. They can be used for the construction of models of the crust in south-eastern Europe and Asia Minor. The *P*-wave velocity patterns obtained for the depth interval 50–300 km are discussed in terms of geophysical (regional isostatic anomaly, heat flow) data for the region. It is proposed that the correlation found to exist between the different mapped geophysical parameters can be explained by compositional changes in the low-velocity layer (the increase of the iron content in minerals).

Key words: 2-D and 3-D velocity, crustal models, southeastern Europe, Asia Minor, eastern Mediterranean

1 Introduction

The region of south-eastern Europe, Asia Minor and the eastern Mediterranean has a very complex geological structure. This, in turn, reflects the collision of the two major African–Arabian and European plates and many microplates. The presence of intermediate-depth earthquakes suggests the existence of subduction zones in the Hellenic arc and in the Vranchea area. Nevertheless, there is no generally accepted geological concept of the tectonic evolution of the region. This can be explained to a certain extent by insufficient information regarding its deep structure. Extensive geophysical studies and especially seismic profiling (Sollogub, Guterh & Prosen 1980; Makris 1977) carried out in recent years have permitted the development of some ideas about the structure of the crust and the upper mantle down to a few tens of kilometres beneath the Moho discontinuity. These investigations were made along several profiles mainly in south-eastern Europe and were not sufficient for building up an image of the lateral heterogeneities in the region.

Calcagnile *et al.* (1982) by means of Rayleigh waves dispersion analysis, and Hovland & Husebye (1982) using 3-D inversion of *P*-wave teleseismic travel-time residuals, made a considerable advance in mapping the velocity inhomogeneities in the lithosphere and asthenosphere. The seismograph network configuration and insufficient data on the dispersion properties of surface waves made studies of Asia Minor and the eastern Mediterranean impossible. The territory of the interaction areas between African–Arabian and European plates is of particular interest since almost no information about lateral velocity inhomogeneities is available yet.

This more detailed investigation has been made possible by the large number of earthquakes which have occurred in 'intermediate' distances in this and adjacent regions, by the improve seismograph network developed in recent years, and by a new improved inversion procedure.

2 Method

The method used in this study is based on Backus–Gilbert formalism for linear inversion of travel times, extended for 2-D and 3-D inhomogeneous media by Yanovskaya (1980, 1984) and Gobarenko & Yanovskaya (1983). The method is intended for estimation of average velocity corrections relative to the starting model from a finite set of travel times along ray-paths crossing the area under investigation. It may be applied to the inversion of surface-wave as well as body-wave data. To complete and clarify the inversion procedure some basic formulae in 3-D case are given.

The lateral variations of the Earth's structure are assumed to be small. In case of body-wave data a velocity depth curve $v_0(z)$ is taken as a starting model, and the unknown velocity corrections are $\delta v(x, y, z) = v(x, y, z) - v_0(z) \ll v_0(z)$. Since these corrections are assumed to be small, the travel-time residuals are linear functionals of $m(x, y, z) = -\delta v/v_0(z)$

$$\delta T_i = T_i - T_{0i} = \int_\Omega G_i(x, y, z)\, m(x, y, z)\, dx\, dy\, dz, \tag{1}$$

where Ω is an area which contains all the ray paths ($0 < x < X$, $0 < y < Y$, $0 < z < Z$); $G_i(x, y, z)$, the data kernel is a generalized function equal to 0 except on the ith ray-path where it tends to infinity.

The problem is to estimate $m(x, y, z)$ from a finite set of the travel-time residuals $\delta T_i (i = 1, 2, \ldots, N)$.

Following the Backus–Gilbert approach, a linear average of $m(x, y, z)$ is determined

$$\langle m \rangle_{x,y,z} = \iiint_\Omega m(x, y, z)\, A(x, y, z, x', y', z')\, dx'\, dy'\, dz'. \tag{2}$$

The averaging function $A(x, y, z, x', y', z')$ must be subjected to the normalization condition

$$\iiint_\Omega A(x, y, z, x', y', z')\, dx'\, dy'\, dz' = 1. \tag{3}$$

From equations (1) and (2) it follows that $\langle m\rangle_{x,y,z}$ is a linear combination of the data,

$$\langle m\rangle_{x,y,z} = \sum_{i=1}^{N} a_i(x, y, z)\, \delta T_i \tag{4}$$

and the averaging kernel, of the data kernels

$$A(x, y, z, x', y', z') = \sum_{i=1}^{N} a_i(x, y, z)\, G_i(x', y', z'). \tag{5}$$

In order that $\langle m\rangle_{x,y,z}$ resembles the unknown function $m(x, y, z)$ as closely as possible, the averaging kernel $A(x, y, z, x', y', z')$ should be concentrated in a small vicinity of point (x, y, z).

A simple criterion of 'δ-ness' of a 3-D kernel consists of the minimization of the sum

$$\begin{aligned} s(x, y, z) = & \int_0^X \left(H(x - x') - \int_0^{x'} \int_0^Y \int_0^Z A(x, y, z, \xi, y', z')\, dy'\, dz'\, d\xi\right)^2 dx' \\ & + \int_0^Y \left(H(y - y') - \int_0^{y'} \int_0^X \int_0^Y A(x, y, z, x', \eta, z')\, dx'\, dz'\, d\eta\right)^2 dy' \\ & + \int_0^Z \left(H(z - z') - \int_0^{z'} \int_0^X \int_0^Y A(x, y, z, x', y', \zeta)\, dx'\, dy'\, d\zeta\right)^2 dz'. \end{aligned} \tag{6}$$

The expression for the unknown coefficient a_i is derived after the minimization of (6) under condition (3) and is similar to that obtained by Johnson & Gilbert (1972) for the 1-D case (see Gobarenko & Yanovskaya 1983). A disadvantage of the proposed criterion is that the solution $\langle m\rangle_{x,y,z}$ is found to be a sum of three functions, each of them depending on one coordinate only. It is evident that an approximation of an arbitrary function of three variables by a sum $f_1(x) + f_2(y) + f_3(z)$ is inadequate in general. Therefore, some peculiarities of the unknown velocity function will be smoothed. Moreover the solution depends on the orientation of the coordinate system. This fact leads to a further smoothing of the solution in ill-posed coordinate systems and optimal orientation of the system will correspond to the minimum smoothed solution.

However, if in some coordinate system the unknown function $m(x, y, z)$ can be represented as $m_1(x) + m_2(y) + m_3(z)$, and besides it is the system in which a solution is determined by the described method, the results of the inversion turn out to be satisfactory. There are several examples of the efficiency of this method (see e.g. Gobarenko & Yanovskaya 1983; Yanovskaya 1984, and all referred to there).

3 Surface waves

3.1 DATA

The group velocities of Rayleigh waves have been determined for periods $T = 10, 20, 30$ s in the region of south-eastern Europe and Asia Minor from observations at 10 seismological stations from sources also located in the region under investigation. The traces are shown in Fig. 1.

The total number of traces varied at each period because of the different length of the trajectories, but never fell below 40. The traces for all periods covered the area uniformly.

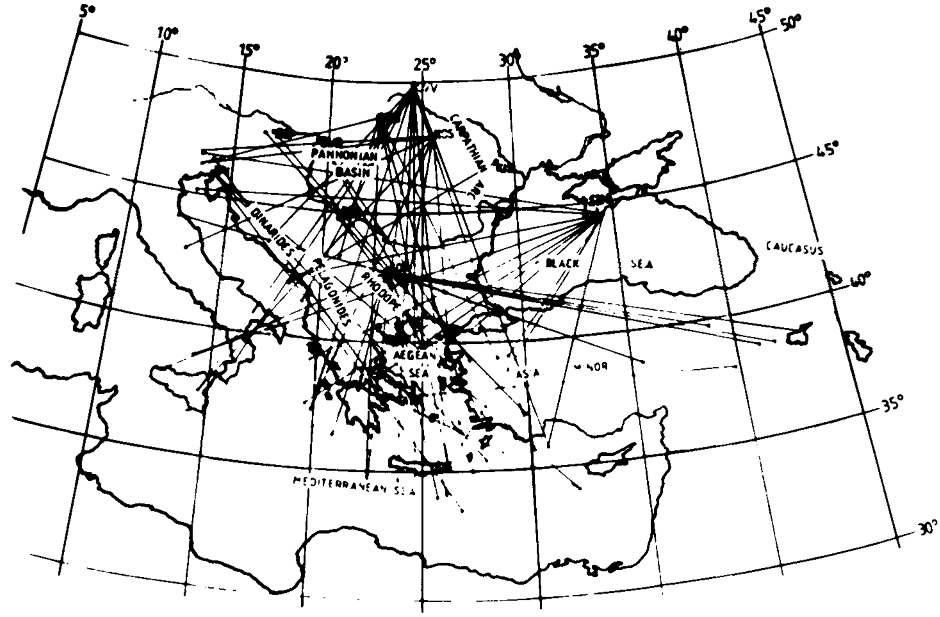
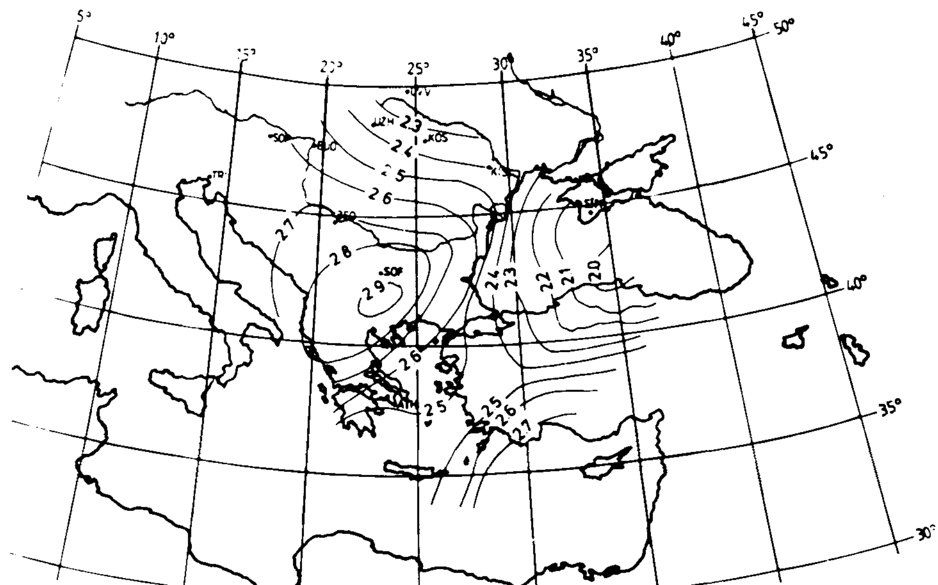


Figure 1. The distribution of the tracks for Rayleigh waves.

The errors in the estimation of group velocities were about 0.08 km s^{-1} and were caused by uncertainties in the epicentre location and in the determination of peaks and troughs on the records.

As mentioned above, a disadvantage of the method used for 2-D and 3-D inversion is the dependence of the solution on the orientation of the coordinate system. The criterion we



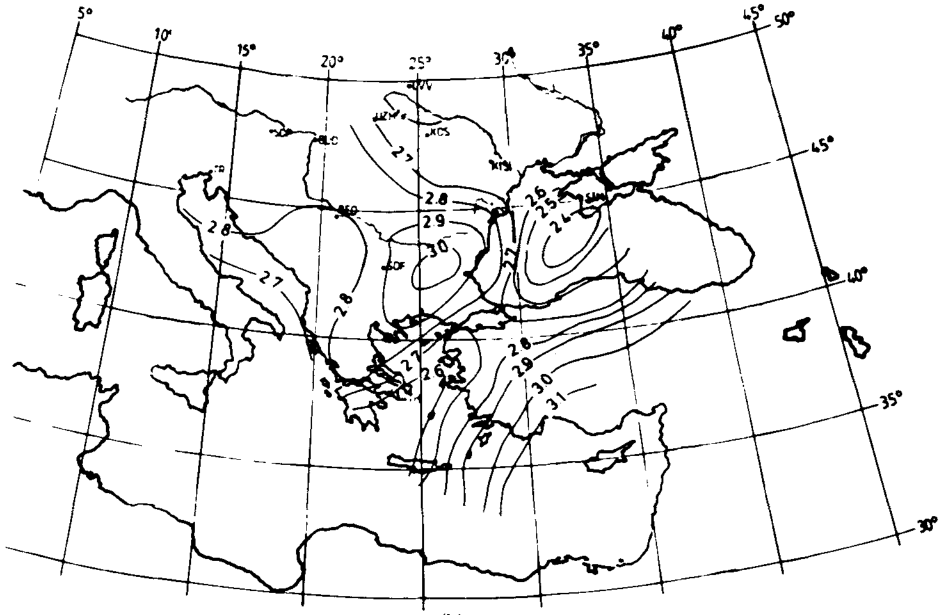
(a)

Figure 2. The group velocity (in km s^{-1}) pattern for Rayleigh waves at (a) $T = 10$ s, (b) $T = 20$ s, (c) $T = 30$ s.

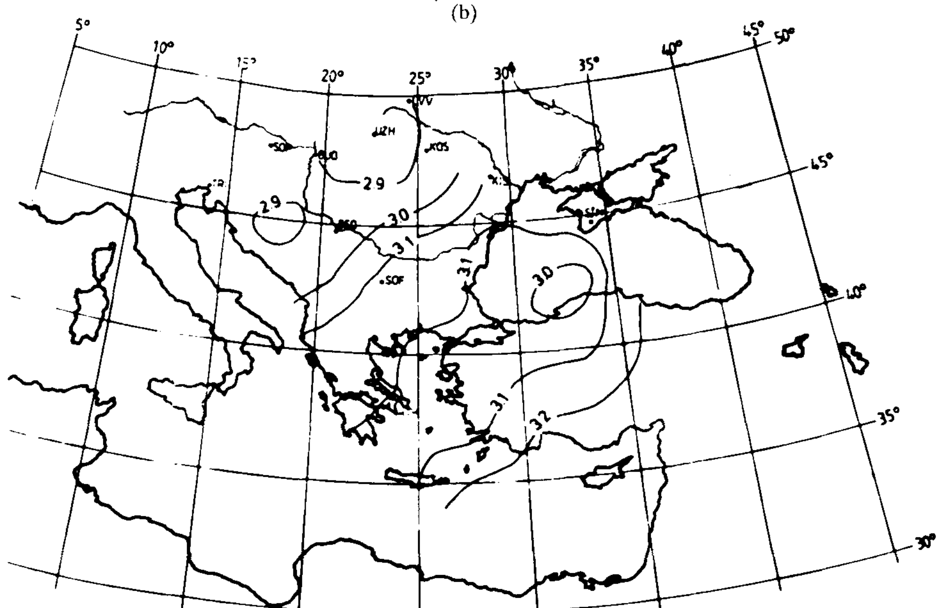
used for determination of the best orientation of the coordinate system was the minimum smoothness of the solution, i.e. the results with the highest contrast. The best results in our case were obtained if the coordinate axes were taken along and perpendicularly to the Adriatic Sea.

3.2 RESULTS

The results of the inversion are shown in Fig. 2. For $T = 10$ s the isolines of 'local average' group velocity outline well-known peculiarities of the sedimentary layer in the region. A



(b)



(c)

Figure 2 – *continued*

minimum in the group velocity patterns is situated over the Black Sea margin, a zone with a thick low-velocity sedimentary layer (up to 14–16 km, Goncharov & Neprochnov 1976). The second minimum is observed over the Carpathian foredeep and is caused also by the existence of a thick sedimentary layer (up to 18–20 km in Fockshan depression, Sollogub, Guterh & Prosen 1980). Another group velocity low is located over the Aegian Sea and reflects the influence of the 3–4 km thick sedimentary layer (Makris 1977) as well as that of the liquid layer. Maxima of the group velocity pattern are observed over the Rhodope massif and Asia Minor where a very thin or no sedimentary layer exists.

For $T = 20$ s a similar pattern of group velocities is observed. One feature should be emphasized: the smoothing of the differences between the maximum and minimum velocities. A rapid increase is easily seen in the zone of the Black Sea margin which is probably caused by the existence of a thin crust and high velocities in the upper mantle.

At a period of $T = 30$ s a further smoothing of the values is observed. The highest velocities are noted over the central part of the Balkan Peninsula and over Asia Minor. The Carpathian foredeep is a velocity low at all periods. This velocity low is connected with the existence of a very thick crust in the zone. The second minimum is located over the southern part of the Pannonian basin where evidence of a velocity low in the upper mantle exists (Belyaevsky 1981).

We have observed similar group velocity patterns for Love waves at periods of $T = 10$–30 s (Yanovskaya & Nikolova 1984). The very good agreement between the group velocity distributions for Love waves at period T and those of Rayleigh waves at $T + 10$ s should be underlined. This is probably caused by a different penetration depth of these two types of surface waves.

The following concluding remarks can be made on the basis of the analysis of the group velocity patterns: (1) the distribution at $T = 10$ s is in good agreement with the well-known peculiarities of the structure obtained from other geophysical investigations and is mainly influenced by the thickness of the sedimentary layer in the region, (2) at $T = 20$ s and 30 s the group velocity patterns reflect the thickness and the velocity parameters of the crust and upper mantle quite well, (3) we propose a very thin or no sedimentary layer as well as high crust and high upper mantle velocities in the region of Asia Minor on the basis of the good agreement between 'local average' group velocity values in Asia Minor and in the central part of the Balkan Peninsula.

4 Body waves

4.1 DATA

The good results obtained for 2-D inversion encouraged us to carry out further investigations of the velocity structure in the region. *P*-wave travel-time residuals as reported by ISC for stations in Europe, Asia Minor and North Africa from earthquakes located in the investigated region for 1970–81 were used for a 3-D inversion. These data contained information only about the properties of the media under investigation: thus no *a priori* assumptions were necessary. The following selection criteria were adopted: the earthquake magnitude was more than 4 and the minimum number of stations reporting the event was set to 50; the depth of penetration of the ray-path was required to be about 100–300 km.

It was convenient to divide the whole region into 18 epicentral zones; within each the coordinates of the earthquakes did not vary by more than 1° latitude, 1° longitude and 40 km in depth. Each earthquake zone included at least 25 earthquakes, so it was possible to make a statistical analysis of residuals at the stations. The minimum number of the reliable *P*-wave onsets at a station had to be more than 10. The mean and the dispersion

(σ^2) of *P*-wave residuals for the ray-paths were calculated. Regardless of the above-mentioned selection criteria, sometimes great differences in *P*-wave travel-time residuals were observed. That is why in addition the values of the residuals exceeding $2\sigma^2$ were eliminated. The new mean and dispersion were calculated for the ray-path and were used for the 3-D inversion.

A primary analysis of the residual values showed the existence of considerable velocity anomalies in the region. For example, the *P*-wave travel-time residuals observed for North African stations were strongly negative (exceeding -5 to -6 s for an individual event) while these for Asia Minor stations had big positive values ($+3$ to $+4$ s). The residuals obtained for different ray paths were compared with the station corrections calculated by Dziewonski & Anderson (1984). The former residuals were either of opposite sign or were considerably smaller than the station corrections.

The anomalies in *P*-wave travel times could also be caused by velocity anomalies close to the earthquake source but the above-mentioned values were observed from different epicentral zones, so that they must reflect mainly upper-mantle structure peculiarities.

As mentioned in Section 2 the solution depends on the orientation of the coordinate system; thus it was more convenient to divide the whole data set into smaller subsets. Three subsets were arranged to cover the different parts of the region and to overlap in the boundary zones so that a control on the solutions obtained by subsets was available. The fourth data subset was formed partly from the data of the first three and partly from ray-paths with a deeper depth of penetration. This data subset covered the whole region. The primary orientation of the coordinate system was chosen using criteria described in Gobarenko, Nikolova & Yanovskaya (1986), to which a further improvement was made using the criterion of the highest contrast of the solution.

4.2 RESULTS

The earthquake zones and stations are shown in Fig. 3. The total data set was composed of 270 ray-paths. The first data subset covered the region of south-eastern Europe and the

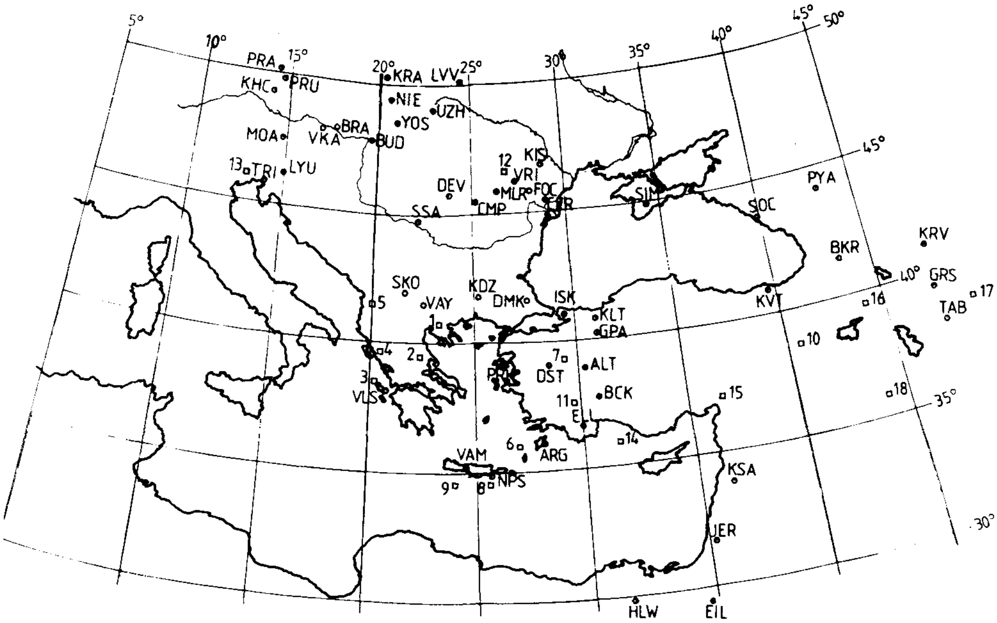
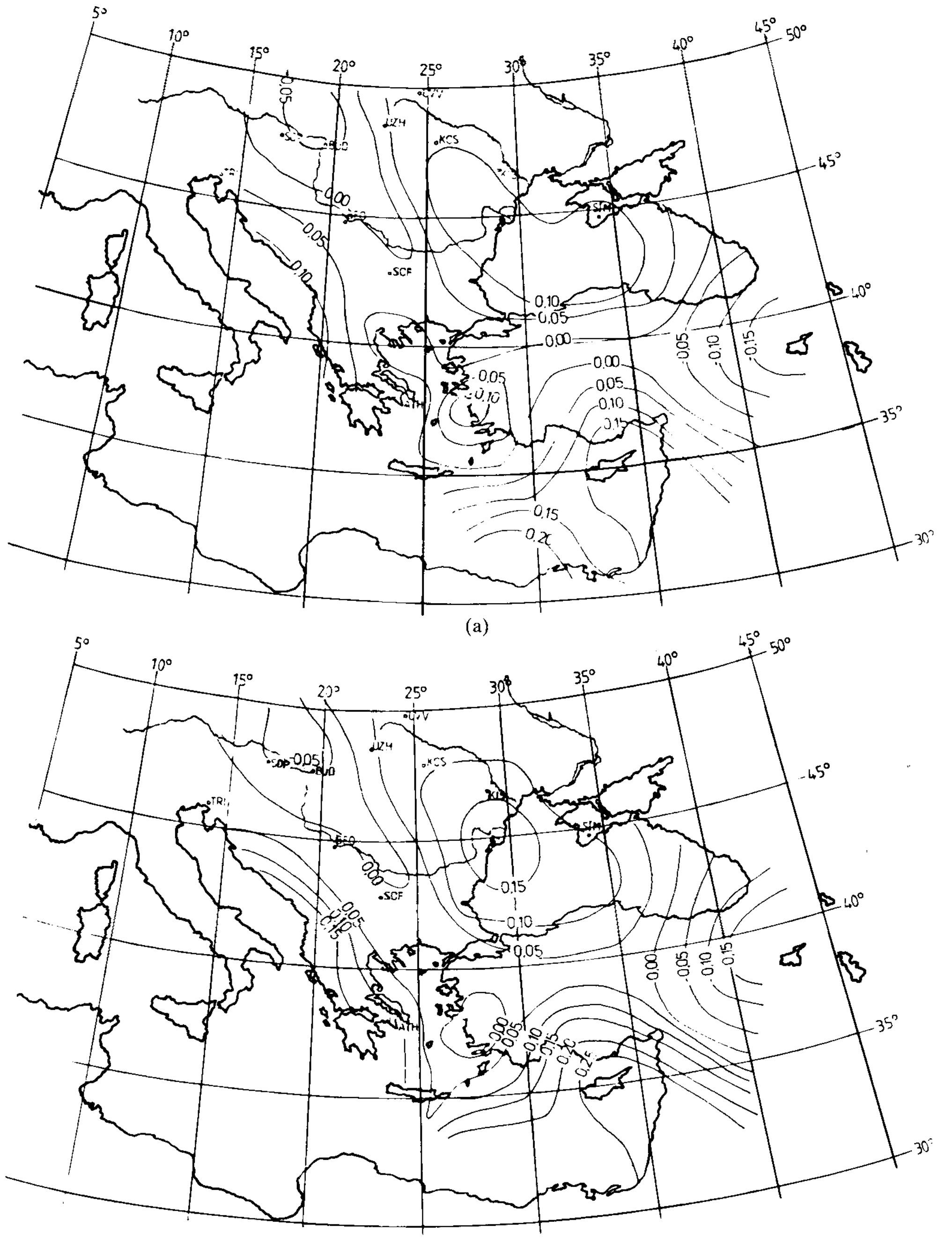


Figure 3. The distribution of the earthquake zones and seismic stations.

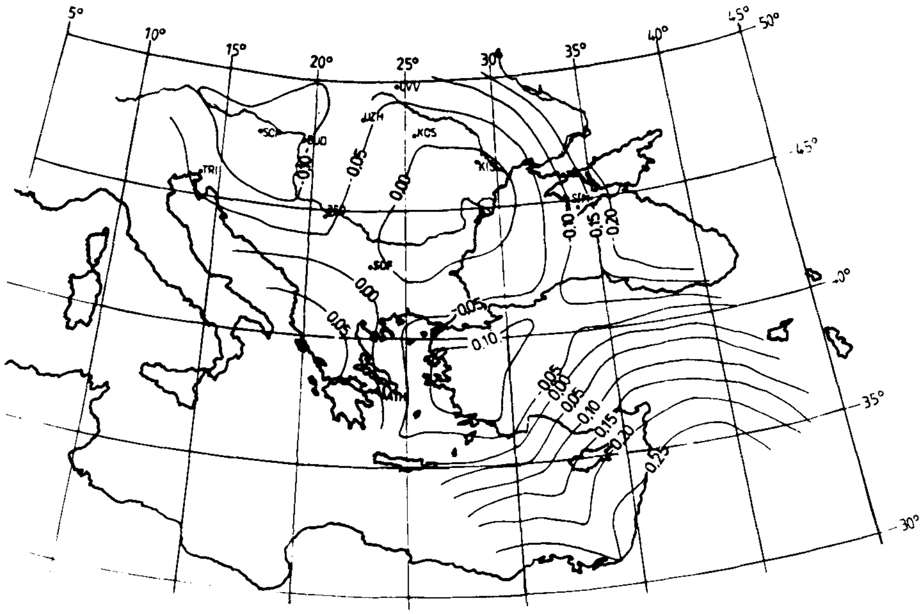
Aegean Sea, the second one Asia Minor and the eastern Mediterranean, the third one the Black Sea and the zone located to the south of the Caucasus, and the fourth one south-east Europe, Asia Minor and the eastern Mediterranean. Each data subset contained 100 ray-paths except the third one with 90 paths.



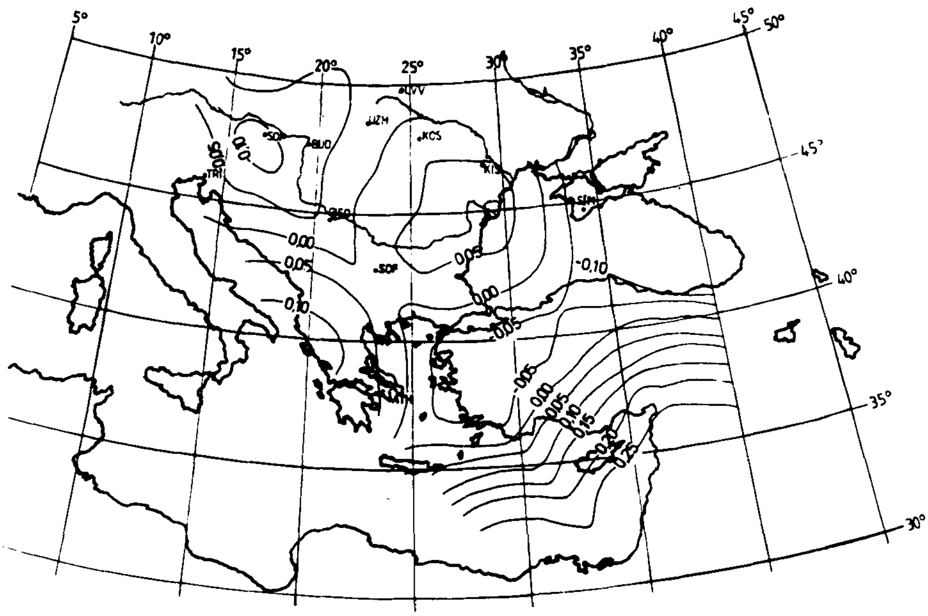
(b)

Figure 4. The *P*-wave velocity (in km s^{-1}) pattern for (a) depth 50 km, (b) depth 100 km, (c) depth 200 km, (d) depth 300 km.

The 'local average' velocity corrections relative to Jeffreys–Bullen velocity model were determined at depths from 50 to 300 km at 50 km spacing. For this depth range (50–150 km) the distributions of the velocity corrections were plotted according to the results of the inverse problem solutions for the first three data subsets. The results in the overlapping zones were averaged. For the depth range 200–300 km, the results for the fourth data subset were preferred. The distributions obtained are shown in Fig. 4.



(c)



(d)

Figure 4–*continued*

The following comments on the velocity anomaly patterns for the region under investigation can be made. The distributions are quite similar for all depth ranges. Velocity lows are located over west Turkey, the Pannonian basin and to the south of Caucasus. Very high velocities are observed in the eastern Mediterranean, the Black Sea and the Carpathians, the Dinarides and the Pelagonian massif. Some changes in the velocity patterns for different depth ranges are clearly seen.

At a depth of 100 km a rapid decrease in the size of areas characterized by negative velocity corrections is observed. Velocity lows are noted over the west part of Asia Minor, the Pannonian basin and the zone located to the south of the Caucasus. With the depth increasing further, the absolute values of the negative velocity corrections increase, velocity lows expand and connect with each other. This tendency can be seen quite well for the depth levels 200–250 km for which positive velocity corrections are obtained over the eastern Mediterranean, the Carpathians and the Pelagonian massif only. Velocity highs occur over the eastern Mediterranean and the south part of Asia Minor.

The local velocity values can exceed the 'local average' ones and that is why the existence of the low velocity zone at the depths of 200–250 km in south-eastern Europe and in central Turkey would be expected. There is no such tendency in the eastern Mediterranean and in southern Turkey: thus in these regions the structure of the lithosphere has to be of the shield type. At the 300 km depth level a decrease in the area of the velocity lows is observed and values of the corrections are diminishing.

5 Discussion

Extensive geophysical investigations have been carried out across various parts of the region during recent years. The results of these studies have enabled us to compare the *P*-wave velocity anomaly images with the heat-flow distribution obtained by Cermak *et al.* (1979), and with the regional isostatic anomaly pattern (Artemiev 1975). Regional isostatic anomaly highs are observed over the Pannonian basin, the Aegean Sea and to the south of the Caucasus (Fig. 5). These highs coincide with the velocity lows obtained from the inversion. The regional isostatic anomalies are caused by anomalous masses at depths of 100–200 km. Thus a comparison of these two distributions is possible. Positive isostatic anomalies of smaller values are observed over the central part of the Balkan Peninsula and coincide with slightly negative velocity corrections at depths of 200–250 km. Negative isostatic anomalies are located over the eastern Mediterranean and correspond to the velocity highs.

The heat-flow pattern determined from numerous investigations in Europe and adjacent areas is shown in Fig. 5. It can be seen clearly that the velocity highs correspond to the heat-flow and vice versa.

The correlation obtained between different geophysical patterns is somewhat confusing. If a direct relationship between seismic wave velocity and density exists, then in conformity with Birch's law we should observe velocity highs coinciding with the zones of positive isostatic anomalies.

Just the contrary tendency is found to exist in Pannonian basin, the Aegean Sea and in the zone located to the south of the Caucasus and it may be explained in the following manner.

Birch (1972), in experiments investigating the properties of olivine under high pressures and temperatures, found out that compositional changes in this mineral (decrease of the forsterite concentration and increase of the fayalite's one, i.e. increasing of the iron content) cause the increasing of the density and decreasing of the seismic waves velocities

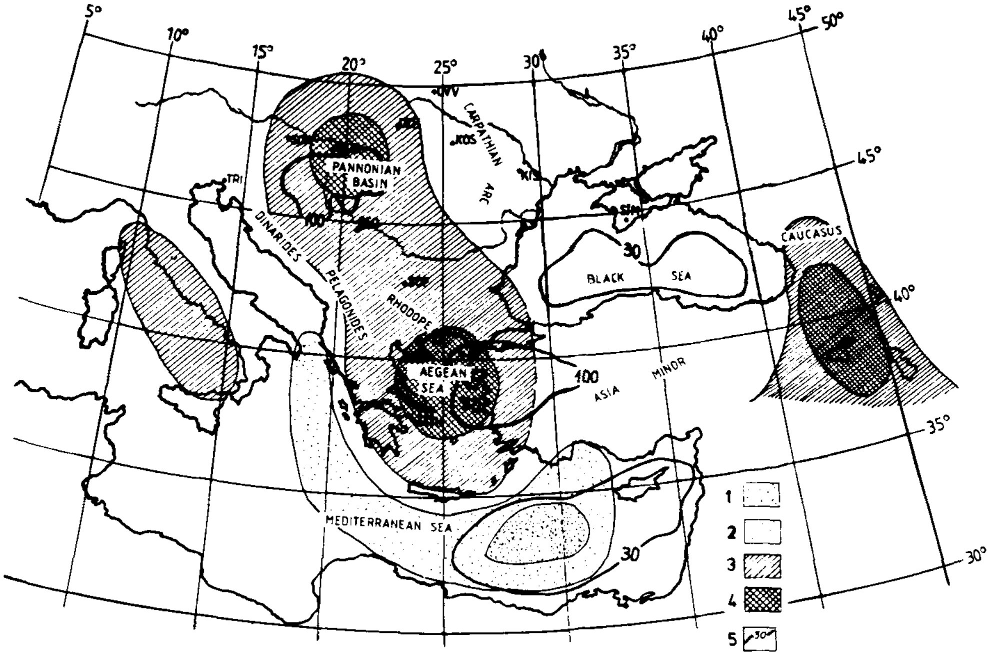


Figure 5. The regional isostatic anomalies (Artemiev 1975) and the heat flow (Cermak & Ryback 1979) patterns. 1 – intensive negative isostatic anomalies, 2 – negative isostatic anomalies, 3 – positive isostatic anomalies, 4 – intensive positive isostatic anomalies, 5 – isolines of the heat flow.

simultaneously. It should be mentioned also that under this proposition the low-velocity layers for P- and S-waves coincided.

If this phenomenon is accepted to take place in the low-velocity layer, then the coincidence of the positive isostatic anomalies, velocity lows and high heat flow is due to the upwelling of the low-velocity layer to the surface.

The coincidence of the velocity highs, negative regional isostatic anomalies and low heat-flow in the eastern Mediterranean is caused by a draw off of the low-velocity layer due to the sinking of the lithosphere asthenosphere boundary close to a subduction zone.

Conclusions

(1) In the region of south-eastern Europe, group velocity distributions of Rayleigh waves were obtained for periods $T = 10$ s, 20 s, 30 s. These patterns are related to the peculiarities of the crust structure and may be used for construction of the models of the crust.

(2) P-wave velocity patterns for south-eastern Europe, Asia Minor and for the zone located to the south of the Caucasus have been obtained for the depth range 50–300 km. The Jeffreys–Bullen velocity lows are observed over the Pannonian basin, the Aegean Sea, the west part of Asia Minor and the zone situated to the south of the Caucasus. Velocity highs are located over the eastern Mediterranean and the southern part of Asia Minor, the Black Sea and the Carpathians, the Dinarides and the Pelagonian massif.

(3) The correlation found to exist between regional isostatic anomalies, the heat-flow data and the P-wave velocity anomaly patterns is explained on the basis of the compositional changes in the asthenosphere (increase of the iron content in the low-velocity layer).

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