Three-dimensional gravity modelling of the European Mediterranean lithosphere

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Accepted 1997 January 3. Received 1996 December 11; in original form 1995 December 13

SUMMARY
Residual gravity anomalies characterizing the density heterogeneities in the upper mantle of the Alpine belt of Western Europe are determined. Residual anomalies were calculated on a 1° × 1° grid by subtracting the gravity effect of the density model for the Earth's crust from the observed gravity field. Our 3-D density model consists of two regional layers of varying thicknesses with a lateral variation in average density: the sedimentary cover and the crystalline crust. Offshore, the model is supplemented by a sea-water layer. This 3-D density model is based on a generalized velocity model represented by structure maps of the main seismic horizons (the 'seismic' basement and the Moho boundary) and a map of the average P-wave velocity in the consolidated crust. The density distribution within the model layers was obtained using the correlation functions between P-wave velocity and density. For sediments, sediment consolidation with depth was taken into account. The gravity effect of the model, approximated by parallelepipeds 1° × 1° in planar size, was calculated by a program designed for solving 3-D gravity problems. The program takes into account the spherical configuration of the Earth. This method also permits the estimation of the isostatic state of the crust. A mantle origin of residual gravity anomalies is confirmed by their close correlation with upper-mantle velocity heterogeneities, established by both seismic-tomography and thermal-regime data. The residual gravity field is inverted into the distribution of anomalous density within the uppermost mantle layer (between the Moho and a bottom level at 200 km depth). The most important anomalies are the high-density domains caused by the thick lithosphere of the Adriatic plate and by the lithosphere 'roots' beneath the Alps and the Calabrian Arc. Negative density anomalies over the Pannonian Basin and the Western Mediterranean basins reach -0.04 to -0.05 × 10³ kg m⁻³ due to thermal expansion of the asthenosphere.

Key words: 3-D gravity modelling, density heterogeneity, isostasy, lithosphere, mantle.

INTRODUCTION
The availability of data on subcrustal density heterogeneities is of great importance for the understanding of the nature of endogenic processes operating in the upper mantle. It is known that the gravity influence from such subcrustal heterogeneities may reach several hundreds of milligals (Bur'yanov et al. 1987; Artemjev et al. 1994; Yegorova et al. 1995).

The commonly used approach for the estimation of subcrustal density heterogeneities was developed in the 1970s. This approach obtains the mantle gravity anomalies by subtracting the gravity influence of the Earth's crust from the observed gravity field. The reliability of the resulting anomalies depends on the authenticity of the initial information, i.e. the quality of seismic data and the method of interpretation of both seismic and gravity data. Rather significant discrepancies, caused by differences in the methods of seismic-data interpretation used by different authors, were demonstrated by Kissling (1993), using the example of the lithospheric structure of the Alps. Ambiguity in the results obtained may be considerably reduced using data from another method, such as gravity. The most widely used approach is gravity modelling, which operates on an initial model based on velocity cross-sections along seismic profiles. Such 2-D density models have been constructed along most of the DSS (deep seismic sounding) profiles in Europe. It should be noted that along these profiles different versions of the density model have been obtained by various authors. The basic reasons for these discrepancies are: (1) the use of different velocity–density relations; (2) differences in the calculation of anomalous densities; and (3) an inadequate
account of the three-dimensionality of the medium. The divergence of different, often contradictory, density models of the profiles impedes their analysis and generalization with the aim of both formulation of the spatial structure of the Earth's crust and obtaining information on subcrustal density heterogeneities.

The correct method for estimation of subcrustal density heterogeneities is 3-D gravity modelling, based on calculations of the gravity influence from spatial mass distribution in the crust. Due to the large amounts of initial data, 3-D gravity modelling is a more labour-consuming process than 2-D modelling, and until now it has not been widely used. Currently, the efficient methods for solving direct and inverse gravity problems from complex mass distributions have been elaborated by Talwani (1973), Strakhov, Lapina & Yefimov (1987), Starostenko & Manukyan (1987), Götze & Lahmeyer (1988) and Starostenko (1990).

The 3-D regional gravity modelling is based on the formulation of generalized velocity models from structure maps for the main crustal horizons. At present, small-scale maps of the Moho topography for the continent are available (Meissner, Wever & Fluh 1987; Giese & Pavlenkova 1988; Ziegler 1990; Giese & Buness 1992). Seismic investigations, carried out on the EG T profile, permit specification of the Moho pattern over the Alpine-Mediterranean area (Ansorge, Blundell & Mueller 1992). 3-D gravity modelling of large regions only became feasible at the beginning of the 1990s. Artemjev et al. (1994) have constructed the density model for the lithosphere of Eurasia, where the crystalline crust was adopted as homogeneous with a constant-density jump of $0.43 \times 10^3$ kg m$^{-3}$ at the Moho. As a result of gravity isostatic modelling and cross-spectral analysis, these authors obtained the distribution of gravity residual anomalies caused by the effect of both the heterogeneous mass distribution within the crystalline crust and the heterogeneities in the subcrustal upper mantle.

The different methods for regional gravity modelling, taking into account the heterogeneous-density crystalline crust, were tested using the 3-D model for the crust of the southern part of Eastern Europe (Yegorova et al. 1995). Residual anomalies characterizing the heterogeneous density subcrustal layer were obtained by the subtraction of the crustal gravity effect from the averaged observed field. The most significant mantle anomalies have been revealed below the eastern part of the Alpine orogenic belt, the Caucasus and the South Caspian Depression. Yegorova et al. (1995) proposed a simplified method for deriving mantle gravity anomalies based on the correlation between the P-wave Moho traveltimes and the crustal gravity influence. This approach allows a rapid estimate (without solving the direct gravity problem) of the crustal gravity effect and of the residual anomalies for the whole European continent, which are mostly due to subcrustal density inhomogeneities. Differences in the deep structure of the two major geoblocks of the continent, that is the West and East European platforms, have been confirmed. Regions of relatively light upper mantle have been distinguished beneath the eastern and northwestern parts of the East European Platform. A heavier upper mantle has been outlined below the Alps, the Calabrian Arc, the Caucasus and the South Caspian Depression (Yegorova et al. 1995).

This paper presents new results of 3-D gravity modelling accomplished over the western (Mediterranean) part of the Alpine belt. Particular attention is given to the analysis of the correlation between the resulting mantle gravity (density) anomalies and the data on both the upper-mantle velocity structure and the thermal state of the upper mantle obtained by seismic tomographic and geothermic methods.

**INITIAL DATA**

The initial observed gravity field that we obtained, averaged on $1' \times 1'$ grid, was based on the Bouguer anomaly (BA) and free-air anomaly (FAA) data obtained by ZNIIGAiK (the Central Research Institute of Geodesy, Airsurvey and Cartography of the former Soviet Union). Gravity field maps for Northern Eurasia have been compiled by Artemjev et al. (1994), who used Bouguer reduction, and Kogan & McNutt (1993) employing free-air reduction. Unlike these compilations, we have used a combined reduction joining the BA on land and the FAA offshore (on the coastal line, values of the BA and the FAA became equal). Accordingly, offshore a layer of sea water was included in the model. A map of the observed gravity field using such a combined reduction for the whole European continent and the Northern Atlantic was published by Yegorova et al. (1995).

Fig. 1 shows a fragment of this map for the southern part of Western Europe, including the Alpine-Mediterranean region. The background of the observed gravity field ($g_{ob}$) is of the order of $10-20 \times 10^{-5}$ m s$^{-2}$ (1 milligal = $10^{-5}$ m s$^{-2}$). The Alps are clearly distinguished by a rather intense wide low drop to $-100 \times 10^{-5}$ m s$^{-2}$ with two branches of a southeastern strike: along the Apennines and along the Dinarides. A vast domain of negative $g_{ob}$ values embraces the Iberian Peninsula and the Pyrenees. Sedimentary basins are marked by slightly positive $g_{ob}$ values. BA values exceed $+20 \times 10^{-5}$ m s$^{-2}$ over the Pannonian Basin, and positive

**OBSERVED GRAVITY FIELD**

![Figure 1. Schematic map of the observed gravity field (contour interval $10 \times 10^{-5}$ m s$^{-2}$) for the European Mediterranean region in a combined reduction of Bouguer anomalies on land and free-air anomalies offshore, averaged over a $1' \times 1'$ grid [fragment of the map for Europe and North Atlantic, Yegorova et al. (1995)].](https://academic.oup.com/gji/article-abstract/129/2/355/591399)
domains of FAA values up to $+40 \times 10^{-5}$ m s$^{-2}$ distinguish the Western Mediterranean Basin.

We based our gravity model on the information about the crustal structure of Europe (thickness of the sedimentary cover, crustal thickness, average crustal seismic velocity and characteristic crustal types) generalized by Giese & Pavlenkova (1988). These data have also been used in the compilation of the Geothermal Atlas of Europe (Hurtig et al. 1992). In the Alpine–Mediterranean region we have taken into account recent seismic data on Moho topography obtained by investigations of the EGT profile (Blundell, Freeman & Mueller 1992).

**PARAMETRIZATION OF THE MODEL**

The two regional layers (the sedimentary layer and the crystalline crust) in our 3-D crustal model for the study region were assigned laterally varying average density values. Offshore, the model was supplemented by a sea-water layer with a density $\rho = 1.04 \times 10^3$ kg m$^{-3}$. Anomalous densities of layers were obtained by the subtraction of the average density of the layer ($\bar{\rho}$) from the density of the upper mantle, an accepted value of $3.3 \times 10^3$ kg m$^{-3}$. $\Delta \rho = (\bar{\rho} - 3.3) \times 10^3$ kg m$^{-3}$.

Since the initial data on the velocity structure of the model (Giese & Pavlenkova 1988; Hurtig et al. 1992) have no information on seismic-velocity distribution in the sedimentary cover, we paid particular attention to data acquisition on the density and velocity distributions in this layer. Determination of the average density in the sedimentary layer was evaluated by different approaches. The first approach consists of $P$-wave velocity ($V$) conversion into its density equivalent ($\rho$) by using appropriate correlation functions $\rho(V)$.

For the Western Mediterranean, where sediments are distinguished as 'uncompacted' ($\bar{V} = 2.0$ km s$^{-1}$), 'normal' or 'semi-compacted' ($\bar{V} = 3.0$ km s$^{-1}$) and 'compacted' ($\bar{V} = 4.5$ km s$^{-1}$) (Moskalenko 1975), we have estimated the average $P$-wave velocity ($\bar{V})$ for the whole sedimentary-cover thickness. These data were recalculated into density values using the correlation function $\rho = 1.882 + 0.0871V + 0.1104V^2 - 0.0132V^3$ of Balavadze et al. (1975). The second approach utilizes sediment compacting with depth. As a principle for density distribution of the continental sediments of Western Europe, we have used the results obtained for the East European Platform (Podoba & Ozerskaya 1975). We have recalculated these data into average density values for the whole sediment thickness taking into account 100 per cent water saturation. The relation obtained is shown in Fig. 2 by slanted shading. For depths $H \geq 4$ km this curve transects the $\rho = 2.6 \times 10^3$ kg m$^{-3}$ line, i.e. at this depth the sediment density reaches the values usual for crystalline rocks. This distribution coincides quite well with the function accepted by Granser (1987) in gravity modelling for the sedimentary cover of the Pannonian Basin (Fig. 2, dotted line). The average density of uncompacted sediments over the Bay of Biscay was estimated using the dependence obtained by Rusakov (1990) in his study of young oceanic sediment compacting with depth. This curve, shown in Fig. 2 by a solid line, is displaced to the left, relative to the sediments of the EEP, on the $\rho$ axis as depth approaches more than 6 km and intersects the $\rho = 2.4 \times 10^3$ kg m$^{-3}$ line.

Two domains of uncompacted sediments with $\rho \leq 2.3 \times 10^3$ kg m$^{-3}$ ($\Delta \rho \geq -1.0 \times 10^3$ kg m$^{-3}$) are distinguished on the map of $\Delta \rho$ distribution in the sedimentary layer (Fig. 3d). The first embraces the basin of the Bay of Biscay and the Aquitaine

![Figure 2. Dependence of sediment density on depth. 1: distribution of average density with depth for sediments of the East European Platform (Podoba & Ozerskaya 1975) adopted as the basic distribution for gravity modelling; 2: the function used by Granser (1987) for the Pannonian Basin; 3–5: areas of density variations obtained by Kinsman (1975) for sandstones (3), clays (4), sandstones and limestones (5); 6: dependence used for sediment density distribution in the Bay of Biscay basin according to Rusakov (1990).](https://academic.oup.com/gji/article-abstract/129/2/355/591399)
Figure 3. The 3-D density model for the crust of the European Mediterranean region. Data averaging is over a 1° x 1° grid. The model consists of a sea-water layer with $\rho = 1.03 \times 10^3$ kg m$^{-3}$ and two regional layers with varying average density: a sedimentary cover and a crystalline crust. (a) Sea-bottom topography, km; (b) depth of the sedimentary layer (Giese & Pavlenkova 1988), km; (c) depth of the crystalline crust (the Moho), km, according to Giese & Pavlenkova (1988), Ansorge et al. (1992) and Giese & Buness (1992); (d) distribution of the anomalous density $\Delta \rho$ of the sedimentary layer (negative values $10^3$ kg m$^{-3}$): $\Delta \rho = (\rho - 3.3) \times 10^3$ kg m$^{-3}$, where $\rho$ was obtained with the use of Fig. 2; (e) $\Delta \rho$ distribution in the crystalline crust layer (negative values $10^2$ kg m$^{-3}$); $\rho$ values were calculated from $V_p$ data (Giese & Pavlenkova 1988) by using the conversion function $\rho = 2.7 + 0.27(V_p - 6.0)$. 

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lateral changes in average density: the sedimentary cover (Fig. 3d) and the crystalline crust (Fig. 3e), divided by the surface of the ‘seismic’ basement (Fig. 3b). The lower constraint of the model is the Moho boundary (Fig. 3c). A general view of the model structure, presented in Fig. 4, is compiled from the main boundaries of the model.

**METHOD AND RESULTS OF THE MODELLING**

The model was approximated by blocks $1^\circ \times 1^\circ$ in plan. The gravity effect was computed using a program designed for solving direct 3-D gravity problems, taking into account the spherical configuration of the Earth (Starostenko & Manukyan 1987) and operating with elementary bodies as spherical parallelepipeds defined by geographic coordinates. The gravity influences of each of the model layers, the sea-water layer $g_w$ (Fig. 5a), the sedimentary layer $g_{sd}$ (Fig. 5b) and the crystalline crust $g_{cc}$ (Fig. 5c) were calculated. The total gravity effect of the model $g_{rt}=g_w+g_{sd}+g_{cc}$, obtained by summing the effects of all the model layers (Fig. 5a-c), is shown in Fig. 5(d). The quality of the initial data in the 3-D model (based on both the contour interval values of the velocity model layers and the accuracy of the derivation of average densities within these layers) is estimated to have an rms accuracy of $20 \times 10^{-5}$ to $25 \times 10^{-5}$ m s$^{-2}$. This makes it possible to obtain maps of the gravity effect with a contour interval of $50 \times 10^{-5}$ m s$^{-2}$ (Figs 5, 6 and 7). To account for the effect of the masses beyond the area studied, the model was supplemented with a $2^\circ$ strip. The width of this fringe is about 220 km, that is, more than five times wider than the average vertical model size (35 km).

Since the range of anomalous density of the model layers varies from $-2.27 \times 10^3$ kg m$^{-3}$ (sea-water layer) up to $-0.47 \times 10^3$ kg m$^{-3}$ (crystalline crust), the values of the calculated gravity effect reach large negative values (up to $-900 \times 10^{-5}$ m s$^{-2}$ over the Alps; see Fig. 5d). Therefore, it is necessary to choose a certain background level and subtract it from the $g_{rt}$ values. Since the model presented for the European Mediterranean crust is part of the 3-D model for the whole European continent, and proceeding from our experience of 3-D gravity modelling (Yegorova et al. 1995), we have used the average value of gravity influence $g_{rt}$ for the crust of the East European Platform (EEP) as that level. The EEP is the stable Precambrian core of the European continent, which is characterized by an observed gravity field with anomalies up to $20 \times 10^{-5}$ m s$^{-2}$. For the 3-D crustal model considered and the $\rho(V)$ function, the average value of $g_{rt}$ for the EEP is equal to $-790 \times 10^{-5}$ m s$^{-2}$. The distribution of the residual gravity anomalies $\Delta g_{rt}$, obtained as $\Delta g_{rt}=g_{rt}-g_{bob}-g_{cc}+790 \times 10^{-5}$ m s$^{-2}$, is shown in Fig. 6(a).

Along with these gravity computations, the isostatic equilibrium state of the Earth’s crust (Fig. 6b) was evaluated by weight calculations over $1^\circ \times 1^\circ$ columns. These columns comprise the crustal layers of the model and a subcrustal layer between the Moho and a bottom level within the upper mantle at $H=80$ km.

Because the calculated gravity effects from the model layers reach rather high negative values (up to $900 \times 10^{-5}$ m s$^{-2}$, see Fig. 5c), it is expedient to consider not only the absolute values of the calculated anomalies, but their relationships as well. As shown in Fig. 5(a), the sea-water layer is the strongest contributor of the total model effect $g_{rt}$ in the Bay of Biscay and in the Tyrrhenian Sea ($-250 \times 10^{-5}$ m s$^{-2}$) and the weakest in the Ionic Sea ($-150 \times 10^{-5}$ m s$^{-2}$). Maximum values of the gravity effect of the sedimentary layer $g_{sd}$ (Fig. 5b)—greater than $-200 \times 10^{-5}$ m s$^{-2}$—are also characteristic of the Bay of Biscay. The anomaly of this amplitude embraces the sedimentary basin over the central southern part of the Adriatic Sea. The Balearic and the Pannonian sedimentary basins are distinguished by $g_{sd}$ anomalies with values less than $-150 \times 10^{-5}$ m s$^{-2}$. Taking into account the gravity effect value of the crystalline crust $g_{cc}$ (Fig. 5c), one may establish the following set of values, depending on the thickness of the crystalline crust (or depth to the Moho): the Bay of Biscay and the Balearic Basin ($-200 \times 10^{-5}$ m s$^{-2}$), the Tyrrhenian Sea ($-250 \times 10^{-5}$ m s$^{-2}$), the Pannonian Basin ($-450 \times 10^{-5}$ m s$^{-2}$), and the Ionic Sea ($-500 \times 10^{-5}$ m s$^{-2}$).

Orogener are characterized by maximum negative $g_{sd}$ anomalies reaching $-900 \times 10^{-5}$ m s$^{-2}$ over the Alps, and $-800$ and $-700 \times 10^{-5}$ m s$^{-2}$ over the Dinarides and the Pyrenees, respectively.

The most peculiar feature of the model effect $g_{rt}$ pattern (Fig. 5d) is the anomalous domain adjoining the Alps and the Adriatic plate. The Pyrenees and the Bay of Biscay are clearly reflected by $g_{rt}$ anomalies of $-750 \times 10^{-5}$ m s$^{-2}$. Over the Balearic Basin and the Tyrrhenian Sea, $g_{rt}$ values form domains with amplitudes of $-550 \times 10^{-5}$ m s$^{-2}$. An isometric $g_{rt}$ anomaly of $-600 \times 10^{-5}$ m s$^{-2}$ has been obtained over the

**Figure 4.** Block-diagram of the 3-D crustal density model for the Alpine–Western Mediterranean area composed of (from top to the bottom) the sea-water bottom (Fig. 3a), the sedimentary-layer bottom (Fig. 3b) and the Moho boundary (Fig. 3c). A: the Alps; BB: the Bay of Biscay; D: the Dinarides; P: the Pyrenees; PB: the Pannonian Basin; TS: the Tyrrhenian Sea; BS: the Balearic Sea.
Figure 5. Gravity influence of the 3-D density-model layers, derived by solving the direct 3-D gravimetric problem (Starostenko & Manukyan 1987) over a 1° x 1° grid: (a) the sea-water layer; (b) the sedimentary cover; (c) the crystalline crust; (d) the total gravity effect of all three density-model layers. Isoline interval (negative values) $-50 \times 10^{-5}$ m s$^{-2}$.

Pannonian Basin, while the Paris Basin is outlined by a $-700 \times 10^{-5}$ m s$^{-2}$ isoline. Most parts of the region are characterized by gravity residual anomalies ($\Delta g_r$) with an amplitude of about $-100 \times 10^{-5}$ m s$^{-2}$ (Fig. 6a). Maximum negative anomalies ($<-200 \times 10^{-5}$ m s$^{-2}$) have been obtained over the Balearic Basin and the Tyrrhenian Sea. The Pannonian Basin is marked by a $\Delta g_r$ minimum of $-150 \times 10^{-5}$ m s$^{-2}$. Positive residual anomalies have been determined in the areas of Alpine orogeny. The Alps and the Adriatic plate are outlined by two $\Delta g_r$ highs of $50 \times 10^{-5}$ m s$^{-2}$ amplitude and the Calabrian Arc and the Pyrenees are marked by local $\Delta g_r$ highs.

Fig. 6(b) shows the isostatic state of the main units of the region and has an isoline pattern similar to those in Figs 5(d) and 6(a). Depending on the type of compensation, two groups of units are clearly distinguished in Fig. 6(b). The first includes the orogens of the Alps and the Pyrenees, the Adriatic plate and the Calabrian Arc. It is necessary to introduce additional masses either within the crust (most probably in its lower part) or in the subcrustal layer to maintain the equilibrium state of the crust of these units. The second consists of the so-called overcompensated units of the Western Mediterranean and the Pannonian Basin, where the crust or the uppermost mantle, proceeding from isostatic estimates, should be expanded.

The residual anomalies in Fig. 6(a) contain errors due to the modelling procedure, the effect of unknown crustal heterogeneities that were not taken into account by modelling, and the influence of the heterogeneous density distribution within the subcrustal upper mantle layer. It is impossible to separate the $\Delta g_r$ anomaly into the components mentioned above. The same concerns apply to the isostatic anomalies in Fig. 6(b), which reflect the degree of equilibrium of the Earth's crust. However, the value of the mantle component exceeds by one order of magnitude the anomalies due to the modelling procedure and the crustal component. Therefore, we can assume that the $\Delta g_r$ anomalies are mainly of mantle origin.

The following step is the transition from mantle gravity anomalies, $\Delta g_m$, to mantle density anomalies, $\Delta \rho_m$. It is necessary to obtain a density distribution within the subcrustal layer such that a gravity effect comparable with the $\Delta g_r$.
anomalies as shown in Fig. 6(a) is produced. The top of the layer containing the anomalous subcrustal masses is the Moho (Fig. 3c). Which level should be taken as the bottom of this layer? According to seismic tomographic investigations (Spakman, van der Lee & van der Hilst 1993), the mantle has a heterogeneous velocity structure down to a depth of 1400 km, but the most contrasting and strong V anomalies are concentrated above a depth of H = 200 km. Therefore, this level was adopted for the lower constraint of the layer containing the most anomalous masses within the upper mantle. The gravity effect from the initial density distribution in this subcrustal layer, calculated with the program of Starostenko & Manukyan (1987), is compared with the residual gravity field Δg (Fig. 6a). The final density distribution, obtained as a result of the iteration process, and the gravity effect are shown in Fig. 7. It is seen that the upper mantle below the study region is generally discompacted using \( \Delta \rho_m = -0.01 \) to \(-0.02 \times 10^3 \text{ kg m}^{-3} \) relative to the \( \rho_m = 3.3 \times 10^3 \text{ kg m}^{-3} \) adopted for the upper-
Subcrustal density anomalies ($10^3$ kg m$^{-3}$, 3-D gravity modelling)

Velocity inhomogeneities $\Delta V$ (%) on $H=95$ km (Spakman et al., 1993)

Calculated temperatures (°C) on $H=50$ km (Bur'yanov et al., 1987)

Thickness of the lithosphere (km, S-waves, Mueller and Panza, 1984)
mantle density below the East European Platform. Below the Western Mediterranean, the Pannonian Basin and the Rhine Graben the discomapcted uppermost mantle density has \( \rho_m = 3.25 \times 10^3 \text{ kg m}^{-3} (\Delta \rho_m = -0.04 \text{ to } -0.05 \times 10^3 \text{ kg m}^{-3}) \).

In the upper mantle below the Alps and the Adriatic plate, there are high-density bodies with \( \Delta \rho_m = 0.05 \times 10^3 \text{ kg m}^{-3} \) and >0.03 \( \times 10^3 \text{ kg m}^{-3} \), respectively. The uppermost mantle Beneath the Calabrian Arc and the Pyrenees has a slightly increased density up to \( 3.31 \times 10^3 \text{ kg m}^{-3} (\Delta \rho_m = 0.01 \times 10^3 \text{ kg m}^{-3}) \).

**COMPARISON OF MANTLE GRAVITY ANOMALIES WITH OTHER DATA**

At present, substantial amounts of various data on structure, state and composition of the upper mantle are available. In our analysis we only consider the relationship between mantle density anomalies (Figs 6a, 7 and 8a) and some representative data on seismic tomography and geothermy (Figs 6b and c).

As a result of one of the first studies of the subcrustal S-wave velocity structure beneath most parts of the European continent, the lithosphere thickness was estimated by Mueller & Panza (1984). High S-wave velocity domains, along with a thick lithosphere, correspond to high-density bodies in the upper mantle below the Alps and the Adriatic plate. Low-density upper-mantle domains beneath the Western Mediterranean Basin correlate with both low-velocity values in the subcrustal layer and an asthenosphere rising up to a depth of 30 km (Fig. 8d).

Seismic tomographic investigations using surface S waves (Snieder 1988) have shown high-velocity domains beneath the Adriatic plate and the Calabrian Arc, as well as low-velocity heterogeneities below the Western Mediterranean (with the most pronounced in the Balearic Basin) at a depth of 150 km. These velocity heterogeneities also correlate well with mantle density anomalies.

As a result of seismological investigations using P waves, the lithosphere thickness for central Europe was compiled by Babuska & Plomerova (1992). Against the background of typical values of thickness for Variscan lithosphere of 100–200 km there are clearly distinguished lithospheric ‘roots’ as deep as 200 km beneath the Alps. A striking coincidence is noted between the configuration of the lithospheric ‘roots’ of the western and eastern Alps (Fig 8c; Babuska, Plomerova & Granet 1990; Babuska & Plomerova 1992) and the shapes of the mantle density anomalies (Fig. 8a). The lithosphere thinning (or asthenosphere rising) of up to 60 km found by Babuska & Plomerova (1992) below the Rhine Graben and Pannonian Basin agrees with our residual anomalies showing the zones of lower density in the subcrustal layer.

The most complete and detailed information on the P-wave velocity structure of the European Mediterranean mantle down to a depth of 1400 km was obtained by Spakman et al. (1993) using seismic tomography. Analysing their velocity sections, obtained at 50 km depth intervals, we draw the conclusion that the main gravity effect could be created by mass heterogeneities located at the subcrustal interval of 150–200 km. The highest contribution is made by masses located within the interval corresponding to the velocity section at \( H = 95 \text{ km} \) depth (Fig. 8b). From a comparison of Figs 8(a) and (b), we notice the direct correlation between \( \Delta \rho_m \) and \( V_p \) (see also Table 1 and Fig. 9b), which clearly manifests itself as the positive mantle density anomaly embracing the Alps and the Adriatic plate. Beneath these units the high-velocity heterogeneity of a similar configuration is observed. Below the Alps this heterogeneity is defined down to a depth of 350–400 km. At depths of more than 200 km, the high-velocity domain of the Adriatic lithosphere is replaced by the two high-velocity zones (beneath the Apennines and the Dinarides) extending down to a depth of 400 km. Subcrustal upper mantle of low density below the Pannonian Basin (Fig. 8a) coincides with the low-velocity zone of complex configurations (Fig. 8b) down to a depth of 200–250 km. A low-density upper-mantle domain shown below the Western Mediterranean (with its separation into the Tyrrhenian and Balearic anomalies) generally correlates with low-velocity areas. In the Paris Basin and the Bohemian Massif, the \( \Delta \rho_m \) anomalies correspond in location and sign with the \( \Delta V_p \) anomaly. The established correlation between \( \Delta \rho_m \) and \( V_p \) is shown in Fig. 9(b).

Differences in the configuration of anomalies demonstrated in Figs 8(a) and (b) occur in the vicinity of the northeastern coast of the Tyrrhenian Sea, where, according to the results of seismic tomography, an intense \( \Delta V_p \) minimum of NE strike is distinguished. In the \( \Delta \rho_m \) pattern this low-velocity domain corresponds with the low-density anomaly over the Tyrrhenian Sea and the Northern Apennines. At a depth of \( H = 51 \text{ km} \) (Spakman et al. 1993) an anomaly of north-easterly trend, which embraces the Apennines, has been identified. However, at depths \( H = 145 \) and 195 km below these anomalies, strong positive \( \Delta V_p \) anomalies have been observed. Such a relationship of velocity anomalies indicates that \( \Delta V_p \) anomalies in the subcrustal uppermost mantle are compensated by velocity heterogeneities located below.

It should be noted that a distinct agreement between mantle density anomalies and velocity structure was obtained for this region by a tectonic reconstruction method (De Jonge, Wortel & Spakman 1993). In particular, correlations were confirmed between: (1) low-density zones and low-velocity domains in the upper mantle below the Western Mediterranean and Pannonian basins; and (2) high-density bodies and high-velocity domains in the upper mantle below the orogens of the Alps, Dinarides and Calabrian Arc.

Clear correlation has been established between mantle density anomalies and the temperature regime in the uppermost subcrustal layer (Figs 8c, 9c and d). For example, the Western Mediterranean and the Pannonian basins with mantle density anomalies \( -0.04 \text{ to } -0.05 \times 10^3 \text{ kg m}^{-3} \) are marked by the 1000 °C isotherm at a depth \( H = 40 \text{ km} \) (Čermak 1984) and the 1200 °C isotherm at a depth \( H = 50 \text{ km} \) (Fig. 8c) according to Bur'yanov et al. (1987). In comparison with this high-

**Figure 8.** Upper-mantle heterogeneities of the European Mediterranean region. (a) Density distribution in the subcrustal layer (between the Moho and a base level at \( H = 200 \text{ km} \)) according to 3-D density-modelling results in Fig. 7; the scale shows \( \Delta \rho_m \) values in \( 10^3 \text{ kg m}^{-3} \). (b) P-wave velocity distribution of model EUR 89B at \( H = 95 \text{ km} \) (Spakman et al. 1993); the scale shows \( \Delta V_p \), %. (c) Distribution of calculated temperatures, °C, at \( H = 50 \text{ km} \) (Bur'yanov et al. 1987). (d) Lithosphere thickness, in km, obtained by Mueller & Panza (1984) in their analysis of S waves. (e) Thickness of the lithosphere in central Europe, obtained by Babuska & Plomerova (1992) in their analysis of P-wave residuals.

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Table 1. Geological and geophysical characteristics of subcrustal inhomogeneities below the Alpine-Western Mediterranean region.

<table>
<thead>
<tr>
<th>Region</th>
<th>Type of endogenic regime (1)</th>
<th>Thickness of the crust, km (2, 3)</th>
<th>Mantle gravity anomaly, $10^{-5}$ m s$^{-2}$</th>
<th>Mantle density anomaly, $10^3$ kg m$^{-3}$</th>
<th>Heat regime on the surface, mW m$^{-2}$ °C</th>
<th>Calculated $r'$ at $H = 40$ km, km s$^{-1}$</th>
<th>Calculated $r'$ at $H = 95$ km, km s$^{-1}$</th>
<th>Velocity distribution in the upper mantle</th>
</tr>
</thead>
<tbody>
<tr>
<td>Balearic Basin</td>
<td>strong taphrogenic</td>
<td>15-20</td>
<td>$-150$ to $-200$</td>
<td>$&lt; -0.04$</td>
<td>70-80</td>
<td>$&gt; 1000$</td>
<td>4.35</td>
<td>7.9-7.95</td>
</tr>
<tr>
<td>Tyrrhenian Sea</td>
<td>medium taphrogenic</td>
<td>15-20</td>
<td>$-150$</td>
<td>$&lt; -0.04$</td>
<td>150</td>
<td>1200</td>
<td>4.2-4.3</td>
<td>7.9</td>
</tr>
<tr>
<td>Ionic Sea</td>
<td>weak taphrogenic</td>
<td>25</td>
<td>$-50$</td>
<td>$&lt; -0.015$</td>
<td>30</td>
<td>300-500</td>
<td>4.5-4.8</td>
<td>8.0-8.1</td>
</tr>
<tr>
<td>Pannonian Basin</td>
<td>rift</td>
<td>26-30</td>
<td>$-150$</td>
<td>$-0.04$ to $-0.05$</td>
<td>80-100</td>
<td>1000</td>
<td>4.35</td>
<td>7.9</td>
</tr>
<tr>
<td>Rhein Graben</td>
<td>tectonomagmatic activation</td>
<td>26</td>
<td>$-150$</td>
<td>$-0.04$</td>
<td>100-120</td>
<td>---</td>
<td>4.35</td>
<td>---</td>
</tr>
<tr>
<td>Armorican Massif</td>
<td>orogenic</td>
<td>35</td>
<td>$-100$</td>
<td>$-0.01$</td>
<td>60-70</td>
<td>700-800</td>
<td>4.5-4.6</td>
<td>8.1</td>
</tr>
<tr>
<td>Bohemian Massif</td>
<td>orogenic</td>
<td>32-36</td>
<td>$-50$</td>
<td>$-0.01$</td>
<td>50-60</td>
<td>700</td>
<td>4.5-4.65</td>
<td>8.0</td>
</tr>
<tr>
<td>Iberian peninsula</td>
<td>orogenic</td>
<td>30-32</td>
<td>$-150$</td>
<td>$-0.03$ to $-0.04$</td>
<td>90-100</td>
<td>1000</td>
<td>4.5</td>
<td>8.0</td>
</tr>
<tr>
<td>Pyrenees</td>
<td>orogenic</td>
<td>35-45</td>
<td>$-100$</td>
<td>$0.01$</td>
<td>70-80</td>
<td>900-1000</td>
<td>8.1</td>
<td>---</td>
</tr>
<tr>
<td>Apennines</td>
<td>orogenic</td>
<td>30-36</td>
<td>0 to $-100$</td>
<td>--</td>
<td>30-60</td>
<td>700-900</td>
<td>4.5-4.65</td>
<td>7.9-8.0</td>
</tr>
<tr>
<td>Dinarides</td>
<td>orogenic</td>
<td>40-45</td>
<td>$-50$</td>
<td>$0.02$ to $-0.03$</td>
<td>30-60</td>
<td>500</td>
<td>8.0-8.15</td>
<td>---</td>
</tr>
<tr>
<td>Adriatic Plate</td>
<td>orogenic</td>
<td>36-40</td>
<td>50</td>
<td>$0.01$ to $-0.03$</td>
<td>30-50</td>
<td>500-700</td>
<td>4.5-4.65</td>
<td>8.0-8.15</td>
</tr>
<tr>
<td>Calabrian Arc</td>
<td>orogenic</td>
<td>34</td>
<td>0</td>
<td>$0.01$</td>
<td>40-60</td>
<td>700</td>
<td>---</td>
<td>8.0</td>
</tr>
<tr>
<td>Alps</td>
<td>orogenic</td>
<td>40-50</td>
<td>50</td>
<td>$0.04$ to $-0.05$</td>
<td>70-90</td>
<td>1000</td>
<td>4.5-4.65</td>
<td>8.0-8.15</td>
</tr>
</tbody>
</table>

(1) Belousov (1990); (2) Giese & Pavlenkova (1988); (3) Hurtig et al. (1992); (4) Čermák (1984); (5) Mueller & Panza (1984); (6) Spakman et al. (1993).

temperature background of the Alpine–Mediterranean region, the subcrustal uppermost mantle below the Adriatic plate (with the adjoining part of the Apennines) and the Calabrian Arc have extremely low temperatures of less than 600 °C. According to the distribution of calculated Moho heat flow (Čermák 1993), the hyperthermal areas of the Western Mediterranean and Pannonian basins have a high Moho heat flow of $\sim 40-50$ mW m$^{-2}$; reaching the highest values (60 mW m$^{-2}$) in the Tyrrhenian Sea, whereas it is between about 20 and 30 mW m$^{-2}$ beneath the Adriatic plate, the Ionic Sea, and the Alps.

On the basis of the generalization of the above data (see Table 1 and Figs 8 and 9), we distinguish the following regimes:

(1) a taphrogenic and rift regime (the Western Mediterranean region, the Pannonian Basin, the Rhine Graben) with thin (15–25 km thick) crust, elevation of the top of the asthenosphere, low-velocity inhomogeneities in the upper mantle, high heat-flow values and $\Delta \rho_m = -0.04 \times 10^3$ kg m$^{-3}$;

(2) massifs of Hercynian orogenesis with 30–35 km thick crust, increased values of heat flow and $\Delta \rho_m = -0.01$ to $-0.02 \times 10^3$ kg m$^{-3}$;

(3) an orogenic regime (the Alps, the Pyrenees, the Calabrian Arc, the Adriatic plate, the Dinarides) with thick (35–50 km thick) crust, thick lithosphere and high-velocity inhomogeneities in the upper mantle, moderate heat-flow values and mantle density anomalies $\Delta \rho_m = 0.02-0.05 \times 10^3$ kg m$^{-3}$.

This subdivision does not strictly hold for the geothermal parameters for the Ionic Sea, with a low heat-flow density (30 mW m$^{-2}$), and the Alps, with a generally increased heat-flow density at the surface (about 80 mW m$^{-2}$).

**DISCUSSION**

In the plate-tectonic framework, the Alps are considered as a structure formed as the result of the collision of the Eurasian and African plates, which led to the conservatism of a rigid, cool, high-velocity and -density lithospheric block in the upper mantle (Kissling 1993).

At present, as a result of continent-wide geophysical investigations (seismic-reflection and -reflection studies, joint investigations along the EGT profile, modeling of gravity data), two models of the deep structure of the Alps have been proposed. As a first approximation they may be described as subsidence with thickening of either (1) the crust or (2) the lithosphere. The common feature of these models is their shape and function as a tectonic 'wedge', as well as their increased velocity and density values. As a result of the analysis of available geophysical data, Kissling (1993) arrived at the conclusion that the structure of the Alps may be ascribed either to the average crustal density exceeding the normal density or to a combination of both. On the basis of our density-modelling results, we prefer the second version, the high-density heterogeneity in the upper mantle beneath the Alps, interpreted as a lithospheric heterogeneity (Mueller 1993; Babuska et al. 1990; Panza & Plomerova 1992). According to seismic-tomography data (Spakman et al. 1993), this heterogeneity has been defined down to the top of the transition zone between the upper and lower mantle. The mantle density anomaly below the Alps can be subdivided into two parts on the basis of shape and orientation. The western part is slightly extended in a sub-
Figure 9. Correlation of mantle density anomalies with the following parameters: (a) crustal thickness (Fig. 3c, Table 1); (b) P-wave velocity inhomogeneities at H = 95 km (Spakman et al. 1993); (c) heat-flow density at the surface (Hurtig et al. 1992); (d) calculated temperatures at 40 km depth (Cermak 1994). Filled circles indicate the regions of taphrogenic regime of Belousov (1990) (1: the Balearic Basin; 2: the Tyrrhenian Sea; 3: the Ionian Sea; 4: the Pannonian Basin); empty circles indicate the rift regime of the Rhine Graben; empty rectangles indicate regions of tectonomagmatic activation regime (1: the Armorican Massif; 2: the Paris Basin; 3: the Bohemian Massif; 4: the Iberian Peninsula); filled rectangles indicate regions of orogenic regime (1: the Pyrenees; 2: the Apennines; 3: the Dinarides; 4: the Adriatic plate; 5: the Calabrian Arc; 6: the Alps).

The lithosphere below the Adriatic plate and adjoining parts of the Apennines and Dinarides has a more complicated structure. Most of the positive mantle density anomaly obtained here may be interpreted as being due to a thick lithosphere (up to 150 km) beneath the Adriatic plate (Spakman et al. 1993). The non-activated regime of the upper mantle below the Adriatic plate manifests itself in: (1) the temperature pattern at 40 km depth (Cermak 1984), similar to that below the East European platform; (2) generally reduced heat-flow density at the surface; and (3) low values of calculated temperatures in the subsurface layer (Hurtig et al. 1992). Two high-density domains were identified as subduction zones below 200 km on both sides of the Adriatic lithospheric plate (Spakman et al. 1993). Selvaggi & Amato (1992) assumed the existence of a recent subduction zone situated to the west of the Adriatic plate. According to their analysis of recent seismicity, intermediate and deep earthquake hypocentres define a zone dipping at an angle of 35°–45° from the Adriatic Sea to the Tyrrhenian Sea. Taking into account the above information, we consider that the mantle gravity anomaly embracing the Adriatic plate can be explained by the total influence of its lithospheric layer of 150 km thickness and...
by the high-density domains identified as subduction zones situated at 200–400 km depth on both sides of the plate, below the adjacent areas of the Apennines and Dinarides.

A mantle density anomaly of a similar nature is inferred below the Calabrian Arc. Here, a subvertical slab in the upper mantle down to a depth of 400 km has been defined by Granet & Trampert (1989). According to Spakman et al. (1993), high-velocity heterogeneity below this arc is defined down to depths of 200–250 km. Below this level it joins a high-velocity anomaly embracing the Apennines.

It is also possible to view the above tectonic regimes in the framework of the concept of endogenic regimes of Belousov (1990). This concept has much in common with plate tectonics in the explanation of basin origin in general and the Pannonian and Western Mediterranean basins in particular. It is assumed that the upwelling of heated asthenospheric material into the subcrustal layer plays an important role in the formation of these structures. The negative density anomalies distinguished here are the result of the influence of heated expanded asthenospheric material whose top is elevated to a depth of 30 km (Mueller & Panza 1984; Panza & Suhadole 1990). In accordance with seismic tomographic studies (Spakman et al. 1993) this low-velocity domain starts to manifest itself at \( H = 95 \text{ km} \). Low-velocity and -density domains in the upper mantle correlate well with high temperatures exceeding 1000 and 1200 °C at depths \( H = 40 \) and 50 km, respectively (Cermák 1984; Bur'yanov et al. 1987). Thermal activity is also manifested in the terrestrial heat-flow density exceeding that for adjacent units by 30–40 mW m\(^{-2}\). The bottom of the asthenospheric layer below the Pannonian and Western Mediterranean basins was estimated as 200 km (Spakman et al. 1993).

Gravity-modelling results may give additional useful information about the state of the upper-mantle material. For example, we have estimated the density anomalies of the level. On these grounds we suggest that the negative mantle gravity anomalies over the Pannonian and Western Mediterranean basins correspond with low-velocity mantle domains, and also with high temperatures at the subcrustal level. They are caused by the cold and thickened (up to 150 km) lithosphere of the Adriatic plate and the Calabrian Arc. The explanation of these anomalies is based on high-velocity domains in the upper mantle below these structures and on reduced temperatures at the subcrustal level. For the rest of the area, the upper mantle of Western Europe, as compared with the mantle below the East European Platform, is characterized mainly by minor negative residual anomalies. Residual gravity lows with \(-200 \) to \(-250 \times 10^{-5} \text{ m s}^{-2}\) amplitudes over the Pannonian and Western Mediterranean basins correspond with low-velocity mantle domains, and also with high temperatures in the upper mantle, as well as with terrestrial heat-flow density. These anomalies are due to thermal expansion, \( \Delta \rho = -0.04 \) to \(-0.045 \times 10^4 \text{ kg m}^{-3}\), of the asthenospheric layer, the top of which is elevated up to the Moho boundary and the base of which is defined at about 200 km depth.

CONCLUSIONS

Progress in studying the Earth's deep structure with gravity data on the basis of 3-D regional gravity modelling depends on two main points: (1) the availability of a regional generalization of seismic data (deep seismic sounding) with the objective of creating the 3-D velocity model for the crust of the study region; and (2) the development of appropriate algorithms for the calculation of the gravity effect from differently approximated mass distributions. At present both these problems have been practically solved. As a base for the 3-D density model of the Earth's crust of the European continent, approximated by two regional layers (the sedimentary cover and the crystalline crust; offshore, a sea-water layer was added), we have used the generalized velocity model of Giese & Pavlenkova (1988), and have taken into account the data recently obtained on the Moho topography for this part of Alpine–Mediterranean region from the seismic study of the EGT profile (Ansorge et al. 1992). The gravity effect of the model was calculated with the program developed by Starostenko & Manukayan (1987) for solving the 3-D direct problem, taking into account the spherical form of the Earth, which also estimates the equilibrium state of the lithosphere. Residual gravity anomalies, obtained by subtraction of the gravity influence from the crustal model from the observed field, reach some hundred \( 10^{-5} \text{ m s}^{-2}\), which exceeds the value of the possible anomalies due to disregarded crustal heterogeneities by one order. With this program we calculated the density-anomaly distribution in the subcrustal layer (between the Moho and the base level at 200 km) responsible for the residual gravity field.

A mantle nature of the main part of the residual gravity anomalies is confirmed by clear correlation with velocity heterogeneities of the subcrustal upper mantle obtained by both seismic tomographic and geothermic data. The most peculiar feature of the mantle gravity and density anomalies of the region is the positive anomalies over the Alps, the Adriatic plate and the Calabrian Arc. The explanation of these anomalies is based on high-velocity domains in the upper mantle below these structures and on reduced temperatures at the subcrustal level. They are caused by the cold and thickened (up to 150 km) lithosphere of the Adriatic plate and the lithospheric 'roots' beneath the Alps and the Calabrian Arc.

For the rest of the area, the upper mantle of Western Europe, as compared with the mantle below the East European Platform, is characterized mainly by minor negative residual anomalies. Residual gravity lows with \(-200 \) to \(-250 \times 10^{-5} \text{ m s}^{-2}\) amplitudes over the Pannonian and Western Mediterranean basins correspond with low-velocity mantle domains, and also with high temperatures in the upper mantle, as well as with terrestrial heat-flow density. These anomalies are due to thermal expansion, \( \Delta \rho = -0.04 \) to \(-0.045 \times 10^4 \text{ kg m}^{-3}\), of the asthenospheric layer, the top of which is elevated up to the Moho boundary and the base of which is defined at about 200 km depth.

ACKNOWLEDGMENTS

The authors are grateful to the anonymous referees whose helpful remarks have resulted in a thorough revision of an earlier version of the manuscript. Our appreciation goes to Dr H. de Boorder (Utrecht University, the Netherlands) for his useful suggestions and help with English. The research described in this publication was made possible in part by Grant No. USD200 from the International Science Foundation.
Northern Eurasia as derived from the gravity data and isostatic models of the lithosphere, *Tectonophysics*, 240, 249–280.


