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ORIGINAL PAPER

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Crustal structure in the Carpatho-Pannonian region: insights from three-dimensional gravity modelling and their geodynamic significance

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Abstract A three-dimensional gravity modelling of the Carpatho-Pannonian region was carried out to get a better image of the Moho boundary and the most prominent intra-crustal density heterogeneities. At first, only the major density boundaries were considered: the bottom of the Tertiary basin fill, the Moho discontinuity and the lithosphere to asthenosphere boundary. Density contrasts were represented by relative densities. The improved density model shows a transitional unit of high density at the base of the crust along the Teisseyre-Tornquist Zone. In the Western Carpathians, an extensive, relatively low-density unit was inferred in mid-crustal levels. The border zone between the Southern Carpathians and the Transylvanian basin is characterized by a sharp, step-like contact of the two crustal units. The Moho configuration reveals important information on the tectonic evolution of the region. Zones of continental collision are represented by thick Moho roots (Eastern Alps, Eastern Carpathians). Transpressional orogenic segments, however, are different: in the Western Carpathians, the Moho is a flat surface; in the Dinarides, a medium Moho root is observed; the Southern Carpathians are characterized by a thick crustal root. The differences are explained with the presence or absence of "subductible" oceanic crust along the Carpathians during the extrusion of Pannonian blocks.

Keywords Gravity modelling · Crustal structure · Carpathians · Pannonian basin · Teisseyre-Tornquist Zone

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Introduction

The crustal thickness of the Carpathian-Pannonian area has been a subject of extensive research from the 1950s, using the two standard geophysical methods for the determination of the depth to the Moho discontinuity: seismic reflection and refraction measurements. The first results of reflection seismic surveys gave a clear indication that the Pannonian basin had a thin crust (Gálfi and Stegena 1959). It was rather surprising, because the widely accepted view at that time was that the Pannonian basin represented a type example of the rigid "median masses", an outgrowth of the "Zwischengebirge" concept (Kober 1928). A significant improvement of the regional picture was brought about by a systematic deep seismic sounding (DSS) survey in Central and Southeastern Europe (Szénás 1972; Sollogub et al. 1973). These measurements provided further important information on the crustal structure of the Pannonian basin and the surrounding mountains: the Carpathians and Dinarides. The crust of this mountain arc proved to be remarkably thick, although rather variable, along strike of the different orogenic segments.

Two-dimensional gravity modelling was utilized to constrain the seismic results in the Carpathians for a long time. Deep crustal root (48 km) below the flysch zone of the Western Carpathians (see Figs. 1, 2), bounded by subvertical faults both from the north and the south provided a good agreement with observed gravity data in the model calculations of Vyskočil (1972). Horváth and Stegena (1977) arrived at a conclusion that the Eastern Carpathian gravity minimum can be adequately explained by a sharp, 65 km deep crustal root and a short wavelength component which derived from the well-developed foredeep sedimentary rocks. Polish gravity studies (Bojdys et al. 1983; Bojdys and Lemberger 1986) also confirmed the crustal root indicated by DSS profiles and concluded that in the Polish Carpathians, the large-scale pattern of gravity anomalies was controlled by the Moho topography.

More recent deep seismic measurements (Tomek et al. 1987, 1989; Tomek and Thon 1988; Tomek 1993) ruled out a deep crustal root beneath the Western Carpathians, in fact they revealed a rather flat Moho, at a normal depth (33–34 km). These data were incorporated in the latest gravity models of the region (Bielik et al. 1990; Vyskočil et al. 1992; Szafián et al. 1997; Vozár et al. 1998), concluding that the observed gravity minimum is a superposition of the gravity effects caused by low-density upper crustal materials, such as the foredeep molasse, flysch wedge, granites and crystalline schists. Further 2-D studies utilised a wide range of geophysical and geological data in order to reveal the deep structure and tectonic evolution of the Western Carpathians (Lillie et al. 1994; Bezák et al. 1997; Bielik 1998; Bielik et al. 1998; Šefara et al. 1998).

New seismic measurements were also carried out in the Pannonian basin (e.g. Posgay et al. 1995, 1996; Hajnal et al. 1996) which have led to a more detailed knowledge on the thickness and structure of the crust. Posgay et al. (1991) collected the Austrian, Czechoslovakian and Hungarian seismic data and published a contour map of the Mohorovičić discontinuity, which was completed and updated by Horváth (1993), Lenkey et al. (1998) and Lenkey (1999).

After decades of strict confidentiality, gravity data have become gradually available from more and more countries of the region. The first regional Bouguer anomaly map was published by Tomek (1988), followed by the more detailed maps of Szafián et al. (1997) and Tari et al. (1999).

The aim of the present study was to carry out 3-D gravity model calculations using the Moho depth map constructed from seismic data. Comparison of the observed and calculated gravity anomalies can result in some corrections to the Moho depth map and identification of the most prominent intracrustal density heterogeneities.

Three-dimensional gravity modelling has obvious advantages over 2-D model calculations, therefore a few previous studies have already used this approach to calculate the gravitational effects of low-density sediments of the Pannonian basin (e.g. Granser 1987; Bielik 1988; Meskó 1988; Papp and Kalmár 1995; Szabó and Páncsics 1999), to constrain the Moho topography (Szabó and Páncsics 1999), or to resolve the spatial extension of local density inhomogeneities (e.g. Bielik et al. 1992). Yegorova et al. (1995, 1997) and Yegorova and Starostenko (1999) utilized large-scale 3-D gravity modelling to identify density heterogeneities in the upper mantle beneath Europe.

Tectonic review

In the present study, three major tectonic units of Europe are discussed: the Trans-European Suture Zone, the Carpathian arc and the intra-Carpathian area, including the Vienna, the Pannonian and the Transylvanian basins (see Figs. 1, 2).

The Trans-European Suture Zone (TESZ) is the most prominent geological boundary in Europe. It stretches in a northwest–southeast direction from the North Sea to the Black Sea, separating mobile Phanerozoic terranes in the west from the Precambrian Platform to the east. In offshore Denmark and Poland, the TESZ is represented by the Teisseyre-Tornquist Zone (TTZ), a 70–130 km wide belt characterized by thickened crust with near vertical block margins (e.g. Guterch et al. 1986, 1999; Stephenson et al. 1995). Seismic surface wave tomography (Zielhuis and Nolet 1994) revealed high S-wave velocities below the Precambrian Platform while the western areas are characterized by low velocities: the remarkable contrast between the two parts can be traced down to 200 km depths.

The Eastern Alps are a zone of active continental collision (e.g. TRANSALP Working Group 2002) between the European and Adriatic plates. As the recent deep seismic reflection images reveal, the internal structure of the Eastern Alps is defined by upper/lower crustal decoupling along transcrustal faults with opposite thrust directions of both the European and the Adriatic plates. Continental collision and crustal thickening in the Alps led to lateral extrusion and gravitational collapse of the ALCAPA block (Ratschbacher et al. 1990, 1991a, 1991b), assumed to be active during Early to Middle Miocene times. The extruded ALCAPA and the independent Tisza-Dacia blocks (see Fig. 1; after Balla 1984; Csontos et al 1992; Csontos 1995) moved towards a "free boundary" offered by the subductible (mostly oceanic) lithosphere of the Carpathian flysch basin. Retreating subduction along the Carpathian arc (Royden et al. 1983a, 1983b; Linzer 1996) combined with slab detachment (Wortel and Spakman 1992, 2000), together with gravitational collapse and extension of the overthickened lithosphere of the ALCAPA were the dominant driving forces behind the formation of the Pannonian basin system and the Carpathians (Horváth 1993; Tari 1994, 1996; Tari et al. 1999). Both ALCAPA and Tisza-Dacia blocks went through significant deformation and rotation upon intruding the Carpathian embayment, whose entrance had a "bottleneck" shape due to the rigid indenters of the Bohemian Massif and the Moesian Platform (Bada 1999). Palaeomagnetic data suggest that ALCAPA suffered 30–50°counter-clockwise rotation while the Tisza-Dacia block experienced a clockwise rotation of 60–80° (e.g. Márton 1993, 2001; Márton and Márton 1978, 1996; Pătraşcu et al. 1992, 1994; Márton et al. 2003). These rotations, however, do not necessarily imply that these blocks rotated as rigid and coherent lithospheric blocks. Instead, it is more reasonable to interpret palaeomagnetic data in terms of differential rotation of smaller crustal flakes, most probably detached at a mid-crustal level (Horváth 1990, 1993; Tari et al. 1999).

The main collision occurred between the ALCAPA and Tisza-Dacia blocks and the East European Platform along the Outer Eastern Carpathians. As soon as the external nappes of the Eastern Carpathians reached the

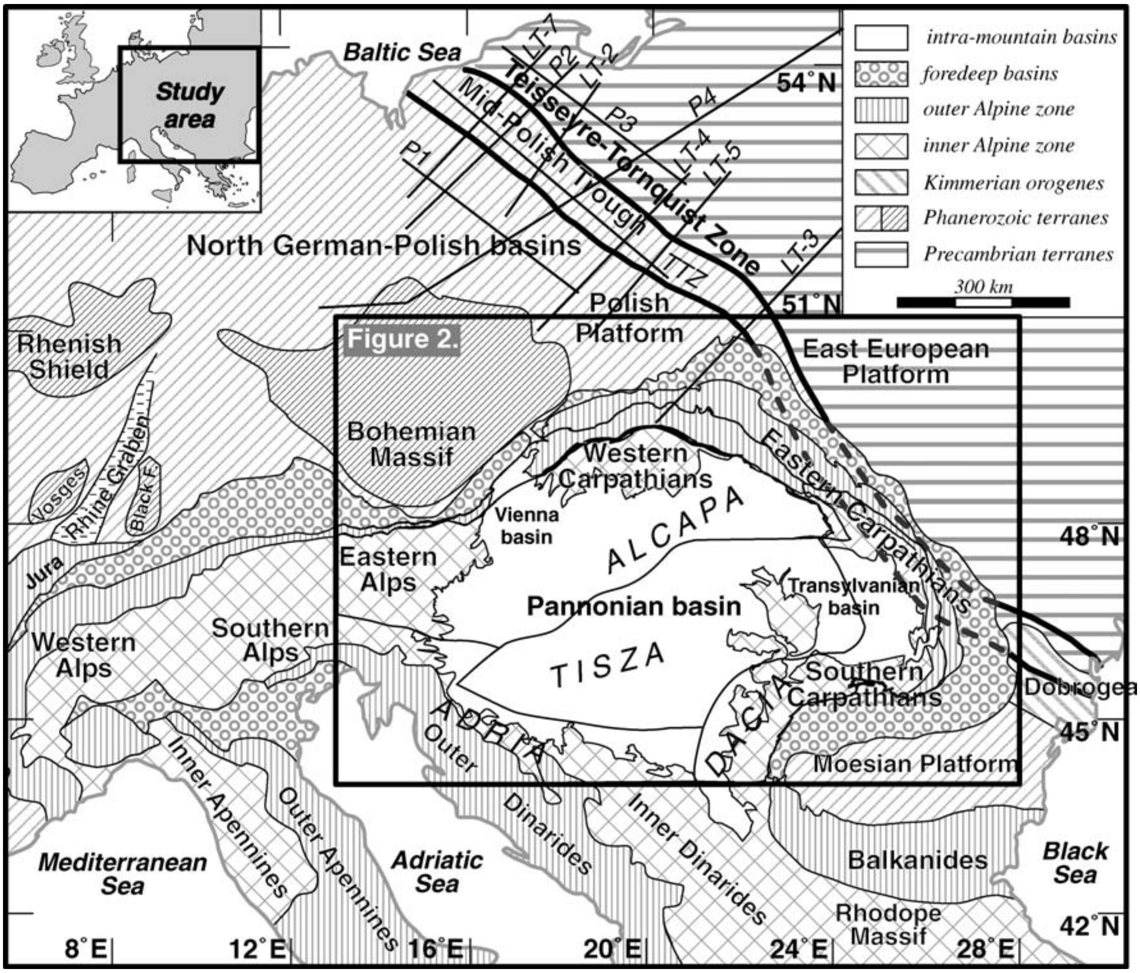

Fig. 1 Major tectonic units of Central Europe. *Rectangle* indicates the studied area (modified after Csontos 1995; Bada et al. 1999; Pharaoh 1999; Guterch et al. 1986, 1999)

East European block along the TTZ (Fig. 1) in the north during the Late Miocene (Matenco et al. 2003), convergence stopped because underthrusting of the up to 50-km-thick continental crust and the East European lithosphere is not possible due to their strong buoyancy. An immediate and most important consequence was the onset of substantial uplift in the rear part of the orogenic wedge (Sanders et al. 1999). Rollback of the subducted slab, however, continued until the slab became subvertical and, eventually, broke off (Wortel and Spakman 2002). In the Southern Carpathians, the substantially thinner and weaker Moesian platform was still involved in underthrusting and stacking of crustal material, while further to the south and west eastward motion of the Inner Carpathians was accommodated by large-scale E–W directed dextral transpressional motion (Matenco et al. 2003).

Meanwhile, in the NW corner of the system, the sinistral opening of the Vienna basin during late Early Miocene times accommodated the northeastward step of the Carpathians with respect to the Alpine thrust belt (e.g., Royden 1985; Lankreijer et al. 1995; Fodor et al. 1999). Strain partitioning occurred along the oblique front: sinistral shear zone along the Vienna Basin but thrusting almost perpendicular to the arcuate thrust front in the external flysch and molasse zones (Fodor et al. 1995).

The major structural formations of the Dinarides were created by the collision of the Apulian and the European plates, through distinct geodynamic stages (e.g. Lawrence et al. 1995; Tari and Pamić 1998). During Late Jurassic to Early Cretaceous times subduction of oceanic crust between Apulia and Eurasia occurred. In the Mid to Late Cretaceous, direct contact locked the continents in the east, but convergence continued in the northern zone. Late Cretaceous to Paleogene times were characterized by continuing continental collision, even attempted subduction of continental crust could be documented (Lawrence et al. 1995). Following the Late Eocene–Early Oligocene, uplift of the Dinarides terminated the compression process and post-orogenic collapse initiated transtensional faulting.

The last phase in the evolution of the Carpathian-Dinarides-Pannonian area is the development of a new compressive stress field during the latest Pliocene and Quaternary, following the locking of the intra-

Carpathian basin system (Horváth 1995; Horváth and Cloetingh 1996). The present stress field of the study area was discussed in details by Bada et al. (1998, 2001) and Gerner et al. (1999). Their results indicate that the alignment of the largest horizontal stress exhibits a radial pattern from the Adriatic indenter towards the Alpine-Carpathian arc. In the Southern Alps and the north-western Dinarides, thrust faulting is dominant although dextral strike-slip motion can also be observed along NW–SE directed faults. Along the southern Dinarides and the Dalmatian coast thrusting with strike-slip components is identified, with NE–SW maximum horizontal stress direction. Thrust faulting in the Vrancea region is distinctly different from the compressive Adriatic regime, while both the Western and Southern Carpathians are characterized by strike-slip faulting.

Modelling concept and input data

There are two basic approaches in gravity modelling—regardless whether it is two or 3-D. The traditional method is to set up a reference crust or lithosphere with constant thickness and average density values for each unit. The assumed densities of the structures to be modelled are subtracted from the reference values, and gravity anomalies are calculated using these density differences. If enough structural, seismic and density data are available for a study area, the reference crust can even have three distinct layers (see e.g. Holliger and Kissling 1992), representing the depth dependent variations of density. In lack of such detailed data sets for the whole study area, average densities are frequently used in a simple two-layer model (crust and mantle). However, in this case, the depth-dependent density increase is neglected and we get unrealistic density differences either at the base of the sediments and/or along the Moho, and, as a consequence, exaggerated gravity anomalies. Let us consider a three-layer density model: sedimentary basin fill, crust and mantle. If the crust is characterized by a density of 2,670 kg m^{-3}, a typical value for Bouguer corrections, we get reliable gravity signal from the basement, but the unrealistic density difference along the Moho gives rise to an extremely high positive anomaly. If the crustal density is increased in order to reach an acceptable density difference along the Moho, the basement of the sedimentary basin becomes dominant in the modelled gravity.

In such cases, another approach namely the use of differential densities relative to a standard crust provides more reliable and interpretable results (Lillie et al. 1994). In this type of modelling, the different boundaries are only characterized by density differences. Take the above-mentioned sedimentary basin as an example. We can set the density differences to acceptable values both along the basement and the Moho. This implies that the density of the crust increases with depth from a near-surface value up to a lower crustal value, however, the vertical density increase within the crust does not result

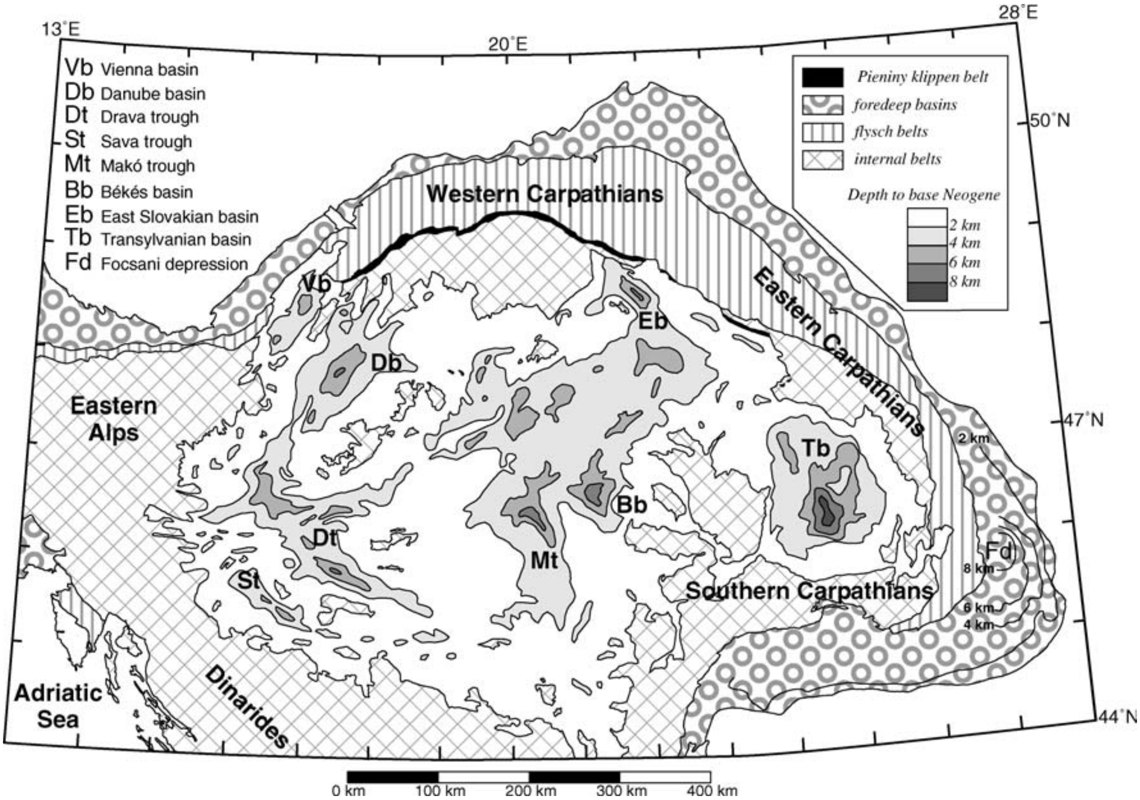

Fig. 2 Deep subbasins of the intra-Carpathian area (after Horváth 1988; Royden and Sandulescu 1988). The deepest part of the Carpathian foreland basin, the Focşani depression, is also indicated

in a gravity anomaly—only lateral inhomogeneities are detected by Bouguer anomalies. There is one important consequence of using relative densities. The amplitudes of the modelled gravity are correct, however, the absolute levels of the measured and calculated gravity fields are not necessarily similar. The measured field in our case is the Bouguer anomaly, while the average level of the modelled field significantly depends on the depth range of the model. As a consequence, the modelled field has to be shifted—with a single constant value for the whole model—in order to compare it to the measured Bouguer anomaly values. Shifting is calculated in such a way that the mean value of the modelled anomaly is identical to the mean value of the measured anomaly.

In the present modelling study, we utilized the second approach for the following reasons. The large scale of the model and the rather heterogeneous seismic data available for the crustal structure of the modelled region did not allow building up a detailed density model of the crust. Furthermore, our intention was to compare our results to those of Lille et al. (1994), who studied more or less the same region, and attempted the simulation of the large scale features of the gravity anomaly field. Moreover, the resolution of the gravity data set was not enough to give constraints on a very detailed model. Therefore, we decided to treat both the crust and the sedimentary sequences as single units (i.e. no density boundaries were set within the crust or the sediments). In our model, the crust was considered as a density reference and the following relative density values were assigned to the major units: 0 kg m^{-3} for the crust, 300 kg m^{-3} for the basin fill, +300 kg m^{-3} for the lithospheric mantle and +270 kg m^{-3} for the asthenosphere (after Lillie et al. 1994). The lithosphere to asthenosphere boundary—which, by its nature, is not a sharp density contrast—was taken into account in order to evaluate the gravity effect of this surface. The geometry of these surfaces were defined after Mahel' (1973), Royden and Săndulescu (1988), Mueller and Panza (1984), Panza (1985), Babuška et al. (1987, 1990), Ziegler (1990), Horváth (1988, 1993) and Dimitrijević (1995); see Figs. 2, 3 and 4. Let us emphasise again—as it is vital for the understanding of our results—that the above-mentioned density values represent the density differences between each unit and the crust, e.g. the density of the asthenosphere is less than that of the lithospheric mantle.

In order to obtain a valid model anomaly field, the gravity effects of the Moho and the lithosphere to asthenosphere boundary were calculated well outside of the study area (Fig. 2). Where the available maps allowed, these far-field effects were calculated between 0°E and 32°E; however, data east of 28°E were sparse.

It should be mentioned that different authors used different methods in determining the depth of the lithosphere to asthenosphere boundary: dispersion of seismic surface waves (Mueller and Panza 1984; Panza 1985) and average travel-time residuals of P waves (Babuška et al. 1987, 1990). The two approaches resulted in misfits at places that had to be smoothed. Please note, that in the central part of the modelled region, homogeneous data sets were used, the above-mentioned "misfits" were observed in Western Europe, far from the target area of the present study, therefore they did not significantly influence the validity of the results. Along the southern, western and northern margins of the model, the depths of the Moho and the lithosphere to asthenosphere boundary are 35 km and 100 km, respectively; along the eastern boundary, in the modelled area (44°N–51°N) the thickness of the crust is 45 km, while the lithosphere to asthenosphere boundary is set at the depth of 200 km, representing the East European platform. The bottom of the model was set at the depth of 250 km. It is to be noted that the curvature of the Earth was neglected in this model calculation.

The 3-D modelling software package used in the present study (Götze and Lahmeyer 1988; Götze et al.

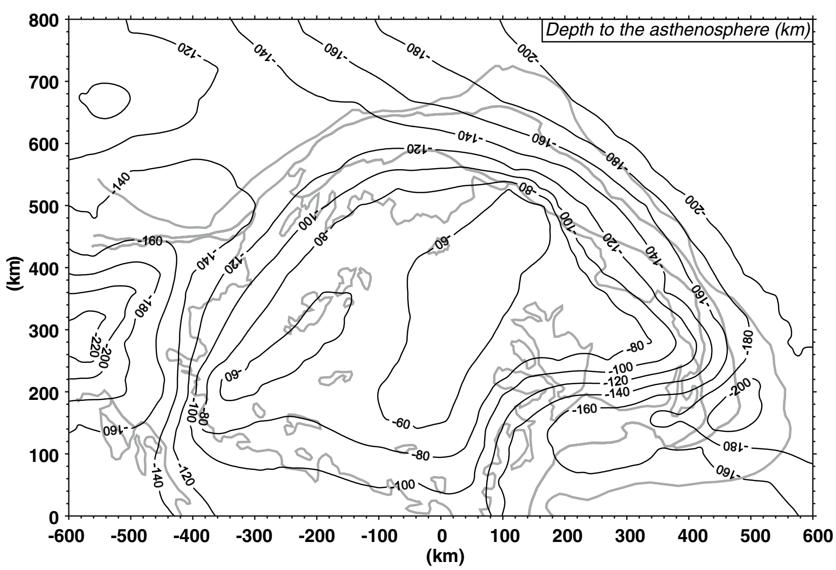

Fig. 3 Depth to the base of the lithosphere in the Carpatho-Pannonian area (after Horváth 1993). *Light grey lines* indicate outlines of the foredeep basins, flysch belts and internal belts of Fig. 2

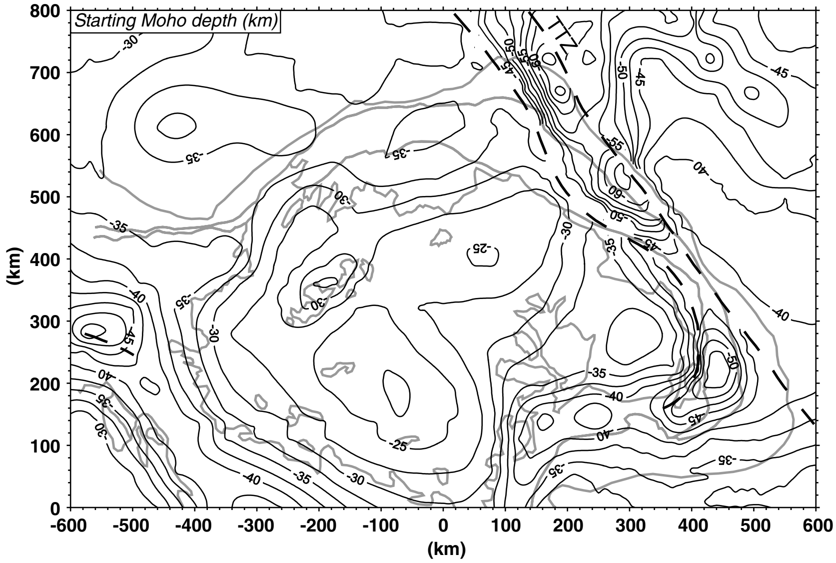

Fig. 4 Depth to the base of the Moho in the Carpatho-Pannonian area (after Horváth 1993). Spacing of isolines is 2.5 km. *Thick dashed lines* indicate deep crustal faults. *Light grey lines* indicate outlines of the foredeep basins, flysch belts and internal belts of Fig. 2

1990) requires the input geometry organized along vertical, parallel, not necessarily equidistant sections. The 33 sections of the model were organized parallel to the lines of latitude. Locations of Sections 6–27 are shown in Fig. 5. Sections 2–5 are located along 40°N–43°N, with a distance of 1°, Sections 28–32 run along 52°N–56°N, with the same spacing. There are two more sections, one in the south and one in the north, that are far away (in the distance of 3,000 km) from the modelled area, in order to avoid edge effects. For the same reason, the length of each section is 6,000 km, centred (with their zero points) along 20.5°E meridian.

The gravity data cover the 44°N13°E–51°N28°E spherical rectangle (Fig. 5). The data of the Bouguer anomaly map compiled by Szafián et al. (1997), Szafián (1999) and Tari et al. (1999) were used, the spacing of the data points along the sections is approximately 10 km. Data in the Polish part of Sections 28, 29, and from the Dinarides along Section 5 are also included, with 20 km spacing. As it is discussed in the above-mentioned publications, the Bouguer anomaly map of the study area was compiled from many published sources, all but one at least 1:1 million in scale. The accuracy of this regional data set is affected basically by two factors: (1) recalculation of different data sets in order to use a unique international gravity formula and (2) interpolation between different data sets along their boundaries. Therefore, the resulting data set is very suitable for large scale, regional studies, but not for detailed local modelling.

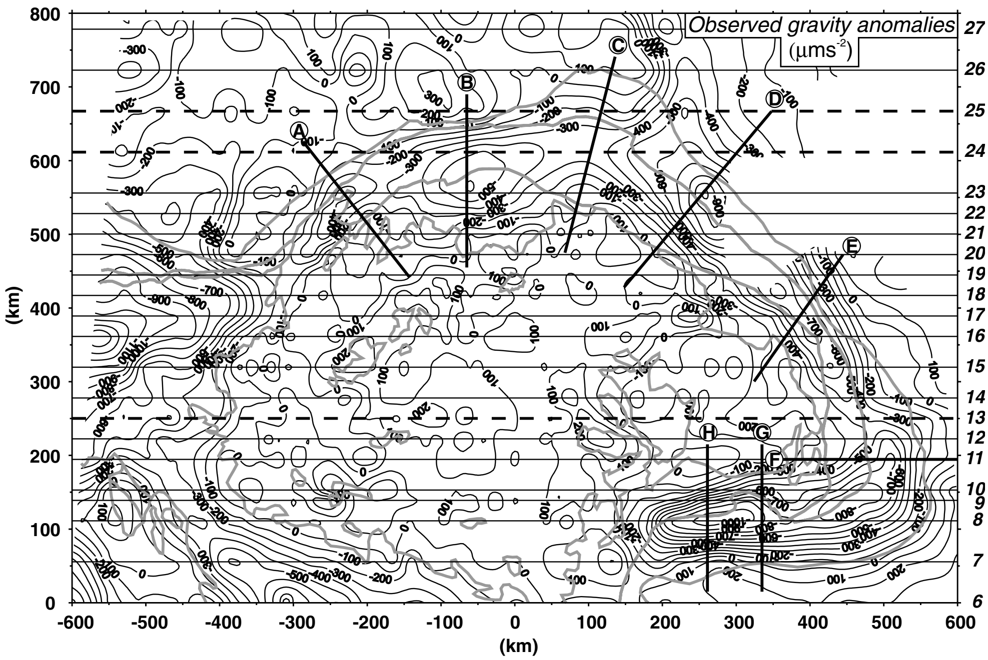

Fig. 5 Observed Bouguer anomaly field of the 3-D study area (after Szafián 1999; Tari et al. 1999). *Thin parallel lines* (6–27) denote the modelled sections, while *thick lines* denote sections discussed in the text. *Light grey lines* indicate outlines of the foredeep basins, flysch belts and internal belts of Fig. 2

Discussion of the results

Figure 6 shows the misfit of the starting model, in other words the difference between the observed and modelled gravity anomalies. Dark grey denotes areas where the model values are too high, while light grey shading represents areas where the model predicted too low anomalies. The map highlights the main areas where the initial model needed further attention. Obviously, the initial density model could not adequately predict the complicated anomaly pattern of the European foreland (Bohemian Massif and Polish Platform) and the negative anomalies of the Eastern Alps, the Vienna basin and the Western Carpathians. Furthermore, the initial model gives too high values for the southern part of the Pannonian basin, and too low values for the TTZ, the Southern Carpathians and the deep subbasins of the southeastern Pannonian basin.

The Alps

The Alpine gravity low was the subject of several papers; these studies were reviewed in details by Tomek (1988). It is remarkable that there is a rather good correlation in the Alps between the topographic crest and the axis of the gravity low. Therefore, it seems to be evident that the excess mass of the mountains is compensated by an Airy-type local isostatic process. However, the Alps turned out to be overcompensated (e.g. Götze et al. 1991; Kissling 1993). Furthermore, the axis of the gravity minimum stretches north of the area of the thickest crust (e.g. Ebbing 2004). Thus, other explanation had to be found for the observed gravity pattern. A group of authors (e.g. Götze et al. 1978; Tomek 1988; Ebbing et al. 2001; Ebbing 2004) concluded that shallow sources of low density—granite plutons, light metamorphites and molasses—contributed to the negative anomaly.

When the gravity effect of the shallow sources and the varying thickness of the crust are accounted for, a broad positive residual anomaly of some 800 μm s^{-2} is obtained (Schwendener and Mueller 1990; Blundell et al. 1992) immediately south of the Alps. This long-wavelength anomaly was explained by two different ways. Schwendener and Mueller (1990) presumed the presence of a high-density body beneath the Southern Alps, representing lithospheric material. On the other hand, Cassinis et al. (1990) rejected this model of lower lithospheric delamination, and concluded that changes in Moho topography are responsible for the observed anomalies, and the upper mantle was homogeneous. They also proposed two density models that equally satisfied the observed gravity pattern: one model featured a single, sharp crust-mantle boundary, while the other included a transitional, high-density layer above the Moho interface. The latest 3-D density model studies (Ebbing et al. 2001; Ebbing 2004), based on the results of the TRANSALP research project, concluded that subcrustal or sub-lithospheric density heterogeneities played an important role in the gravity field of the Alps. They also pointed out that at first order, two main sources of the Bouguer field could be identified: the density contrast at the Moho and density inhomogeneities within the uppermost 10 km of the crust. Unfortunately, the TRANSALP project failed to resolve so far the complicated internal crustal structure in the central part of the Eastern Alps (TRANSALP Working Group 2002).

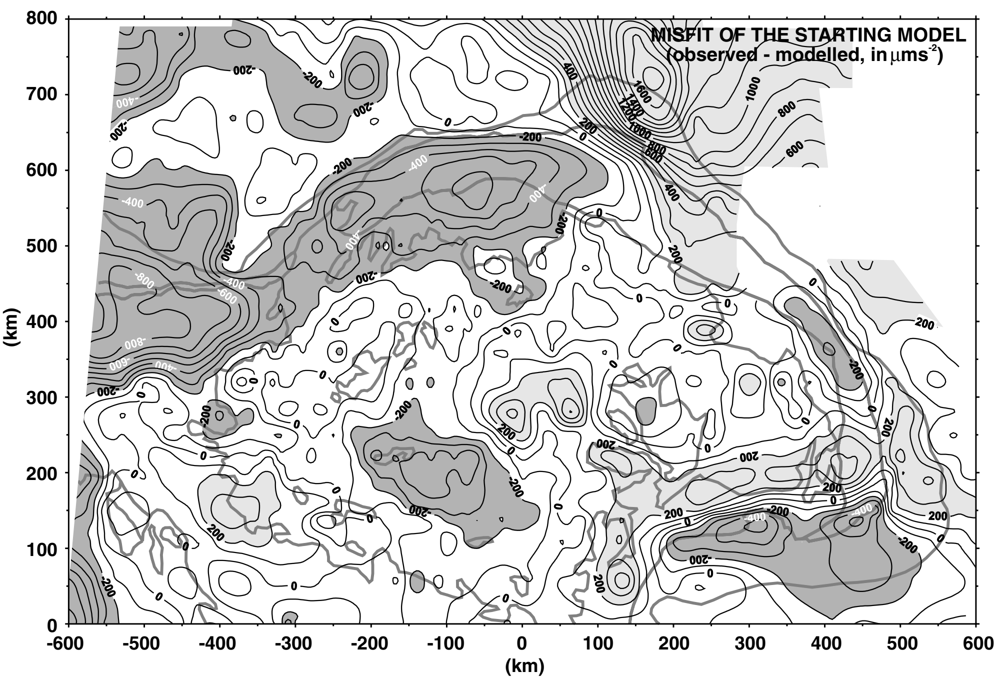

Fig. 6 The difference between the observed and modelled gravity fields in the initial model. *Dark grey shading* denotes areas where the initial model predicted too high values, while *light grey* indicates too low values. *Light grey lines* indicate outlines of the foredeep basins, flysch belts and internal belts of Fig. 2

As Fig. 6 shows, the initial Moho root was not deep enough to reproduce the observed negative anomaly; therefore this boundary was moved downwards. However, the overall 10–15 percent increase in the crustal thickness cannot be verified completely—although the maximum 50–53 km crustal thickness of the density model is in accordance with seismic data and detailed gravity modelling results (see TRANSALP Working Group 2002; Ebbing 2004)—thus it has to be considered as a manifestation that mass deficit is required in the shallow parts of the crust.

The European foreland

The relatively short wavelength gravity anomalies of the Bohemian Massif and the Polish Platform show a rather diverse pattern. Stripes of positive and negative anomalies are oriented NW–SE in the Polish Platform, while in the Bohemian Massif they are prevailingly arranged in NE–SW direction. Blížkovský et al. (1994a, 1994b) constructed a series of stripped anomaly maps of the area: they calculated the gravitational effect of different units and subtracted it from the observed anomaly field. With this method, they corrected the Bouguer anomaly map of the region for the effects of the Moho and the lithosphere–asthenosphere boundary, and concluded that the anomalies can be explained by sources located in the upper 15 km of the crust.

Teisseyre–Tornquist Zone

Several deep seismic surveys have been carried out since the 1960s in order to study the crustal structure of the TTZ (Guterch et al. 1986, 1999; Grad et al. 1999). The earlier DSS surveys (Guterch et al. 1986) concluded that crustal thickness along the TTZ was 50–55 km, and revealed an interesting feature: in the northwestern and central portions of the TTZ the lowermost, approximately 10 km thick layer of the crust was characterized by relatively high seismic velocities (7.5–7.7 km s^{-1} upper boundary velocity), therefore it was considered as a transition layer between the crust and the mantle. In the southeastern part of the TTZ, close to the Carpathians, this transition layer was not observed. Guterch et al. (1986) explained it as a consequence of tectonic disturbances in this stretch of the TTZ, which could be identified only in the deeper parts of the crust (from depths of about 12 km down to the Moho). They suggested that these disturbances were caused either by wide shear zones or by intrusions of upper mantle material.

Dadlez et al. (1995) carried out tectonic subsidence modelling of the Mid-Polish Trough—a deep basin above the NW part of the TTZ, filled with Permian-Mesozoic sediments—and found that their model implying crustal thinning as the driving mechanism for the subsidence was inconsistent with the observed crustal thickening. They concluded that the above discussed, seismically defined Moho showed the position of the Moho before the Permo-Triassic rifting and the present day Moho coincided with the top of the transition layer. They proposed that the high-velocity transition layer was produced by dense, mafic mantle material, which had intruded into the original lower crust during lithospheric thinning.

A more recent set of seismic lines—the LT-7 profile (Guterch et al. 1994), the TTZ profile (Grad et al. 1999) and the POLONAISE'97 seismic refraction-wide angle reflection experiment (Guterch et al. 1999)—was shot in the northwestern part of Poland and brought new insights into the complex crustal structure of this tectonic zone (for location of seismic lines see Fig. 1). LT-7 and TTZ profiles modified the earlier Moho

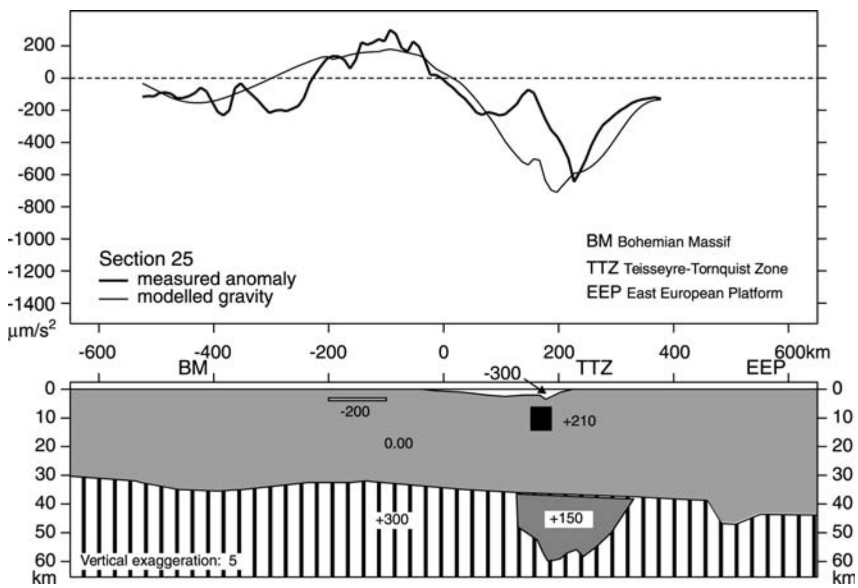

Fig. 7 Density model along Section 25 of the final model. Note the high-density crustal root under the Teisseyre-Tornquist Zone, and the high-density body in the shallow crust (probably related to basic intrusion). Numbers are densities (in kg m^{-3})

topography: the maximum crustal thickness in the TTZ was slightly more than 35 km in the northwestern part of Poland, increasing to a bit above 41 km—instead of 50 km—towards the SE. The P4 profile of the POLONAISE'97 experiment revealed a very complex crustal structure across the TTZ: the Moho depth was about 50 km under the Zone, decreasing to 32–39 km towards the SW (Palaeozoic Platform) and 43–45 km to the NE (Precambrian Platform). There is a remarkable asymmetry between the maximal thickness of the sedimentary cover in the Polish Basin (16–20 km) and the crustal root to the NE (Guterch et al. 1999). Jensen et al. (2002) presented an integrated Moho map of the northern TESZ, a part of which is represented by the TTZ. Their map confirms the findings of Grad et al. (1999) and Guterch et al. (1999). It also shows that an up to 50 km deep SE–NW trending Moho trough can be observed on the eastern flank of the TTZ. Furthermore, their analysis revealed that the lower crust was reflective in the TTZ and concluded that this phenomenon was initiated during the Carboniferous, Permian and Triassic lower crustal extension of the Palaeozoic Platform, accompanied by contemporaneous intrusions of basaltic dykes and sills into the weakness zones of the lower crust.

The largest misfit between the observed and modelled anomalies of the starting model is in the vicinity of the TTZ (Fig. 6). This is not surprising: a very thick crustal root should cause a significant and wide negative anomaly instead of a narrow belt of anomalies in the order of $-600\ \mu\text{m s}^{-2}$.

Following the premise that higher seismic velocities correspond to higher densities (e.g. Rybach and Buntebarth 1984), the model crust was subdivided in this region and a high density layer was introduced into the model, which might represent the transition between old, pre-Permo-Triassic Moho and a new, post-stretching Moho (Dadlez et al. 1995; Stephenson et al. 1995). The top of this unit was set in the depth of 35–43 km (after Beránek et al. 1972; Sollogub et al. 1973; Guterch et al. 1986), and a density contrast of $+130\ \text{kg m}^{-3}$ was assigned to it (Fig. 7). Note that Fig. 14 shows the lower boundary of this unit as Moho. A similar solution was proposed by Horváth and Stegena (1977), with even higher density contrast, and by Grabowska and Raczyńska (1991) for the northwestern part of Poland. In another density model, Grabowska et al. (1998) suggested that below the narrow zone of the northwestern part of TTZ the density of the mantle was higher than in the surrounding regions.

As Fig. 14 shows this "transitional" unit is approximately N–S directed, it coincides with the TTZ in the transition zone between the Western and Eastern Carpathians, however, it is shifted towards the Precambrian Platform north of the Carpathians (cf. Jensen et al. 2002).

Another unpredicted feature is a positive anomaly belt that accompanies the TTZ from the southwest. Intrusions of upper mantle material into the upper crust can produce such positive anomalies (see Grabowska and Perchuć 1985; Królikowski et al. 1996), therefore a high-density body was incorporated into the upper crust along the TTZ (Fig. 7).

Some parts of the Precambrian platform area northeast of the TTZ are characterized by large misfits between measured and modelled anomalies (see Fig. 13). Unfortunately, lack of gravity and seismic information makes it difficult to analyse and model the crustal structure of the East European Platform.

Western Carpathians

The gravity anomaly field of the starting model is characterized by significant discrepancies with the observed anomalies in the territory of the Western Carpathians. As Fig. 6 shows, the maximum of this discrepancy is $-600\ \mu\text{m s}^{-2}$. Since a flat Moho was confirmed by deep seismic reflection surveys (Tomek et al. 1987 1989; Tomek and Thon 1988)—even if its origin is a much debated question (see e.g. Tomek 1993; Roure et al. 1996; Szafián et al. 1997; Zoetemeijer et al. 1999)—the sharp horizontal gradients on both slopes of the Western Carpathian gravity minimum and source depth calculations show that the mass deficit must be located in the shallow crust (Tomek et al. 1979) and has to occupy a relatively large area.

Based on the results of the 2-D density model studies in the area (Tomek et al. 1979; Pospíšil and Filo 1980; Bielik et al. 1990; Vyskočil et al. 1992; Szafián et al. 1997) a low-density body was inserted in the upper part of the crust, with a density contrast of $-200\ \text{kg m}^{-3}$ (Figs. 7, 8, 9).This body can be related to the low-density sedimentary complex of the accretionary wedge created during subduction (Tomek et al. 1989), combined with relatively low-density granites and crystalline rocks (Pospíšil and Filo 1980; Bielik et al. 1990). The results of Lillie et al. (1994) and Bielik (1995) show that the short wavelength changes of the Moho depth in the collision zone of the Western Carpathians also contribute to the total gravity low of the region.

Eastern and Southern Carpathians

In the Eastern Carpathians, deep seismic measurements (Sollogub et al. 1988a; Zajac et al. 1988; Răileanu et al. 1994) and the results of hydrocarbon exploration (Sovchik and Vul 1996) have documented that the accretionary wedge, associated with Tertiary subdction, may reach the thickness of 12 km. This sedimentary complex is characterized by surprisingly high porosities and low densities, even in great depths (Bortnitskaya et al. 1977; Sollogub et al. 1988b). More to the south, at the Focşani depression, the substrata of the Neogene basement are at a depth of about 9 km (Dumitrescu and Săndulescu, 1970; Royden and Săndulescu 1988). All these thick sedimentary units are accounted for in the

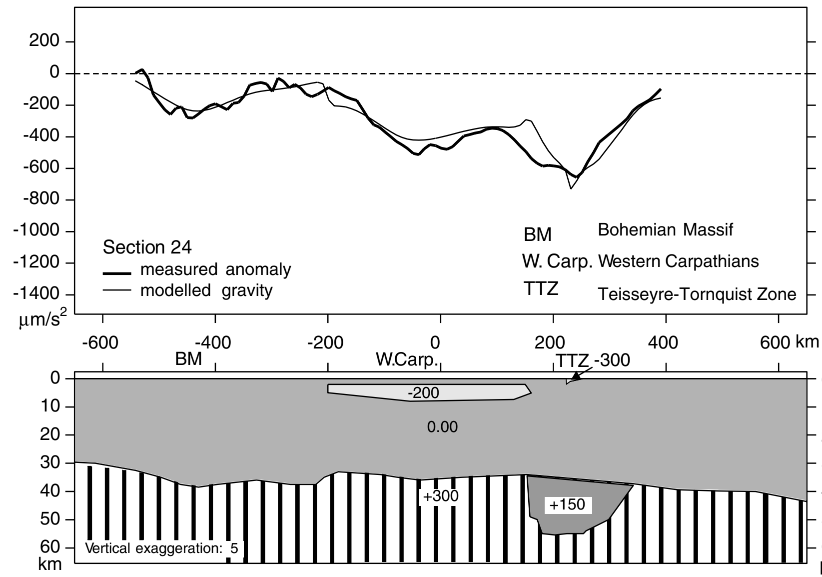

Fig. 8 Density model along Section 24 of the final model. Note the low-density unit in the shallow part of the Western Carpathian crust and the high-density crustal root under the Teisseyre-Tornquist Zone. With these modifications, the 3-D model provided a good fit between observed anomalies and calculated gravity. Numbers are densities (in kg m^{-3})

final model by an elongate foredeep wedge, with a density contrast of -280 kg m^{-3} (Fig. 10).

In the foreland region of the Southern Carpathians it was relatively easy to resolve the misfit. Geological data (e.g. Ştefănescu 1984; Dicea 1996) and flexural modelling studies (Matenco et al. 1997) also indicate that the depth of the foredeep basin may reach 7–8 km below the thrust front, while in the initial data set only 2–2.5 km was assumed after Royden and Săndulescu (1988). This foredeep geometry is enough to reach a good fit between the modelled and the observed anomalies (Fig. 11).

In the inner parts of the Southern Carpathians and the Carpathian bend, Fig. 6 shows significant misfits. As

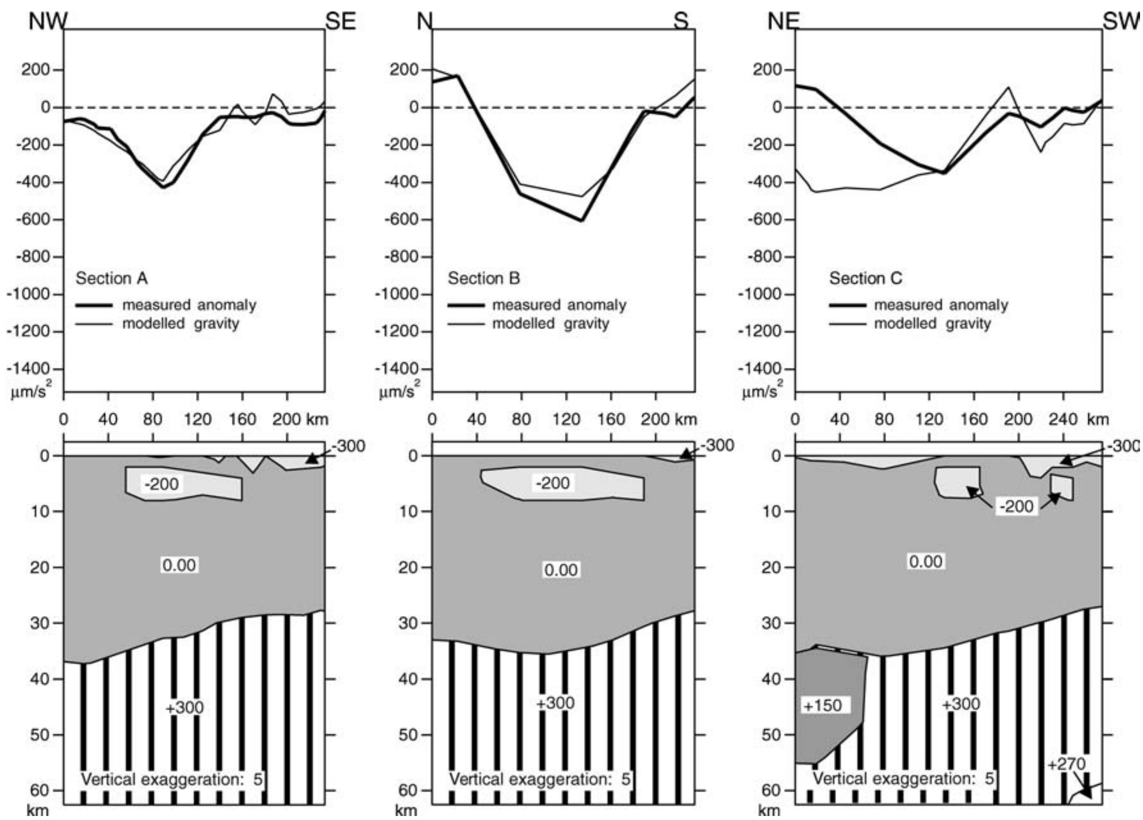

Fig. 9 Cross-sections of the final model in the Western Carpathians. The minimum zone of the Bouguer anomaly field is well reproduced after the low-density unit was introduced in the upper crust. Numbers are densities (in kg m^{-3})

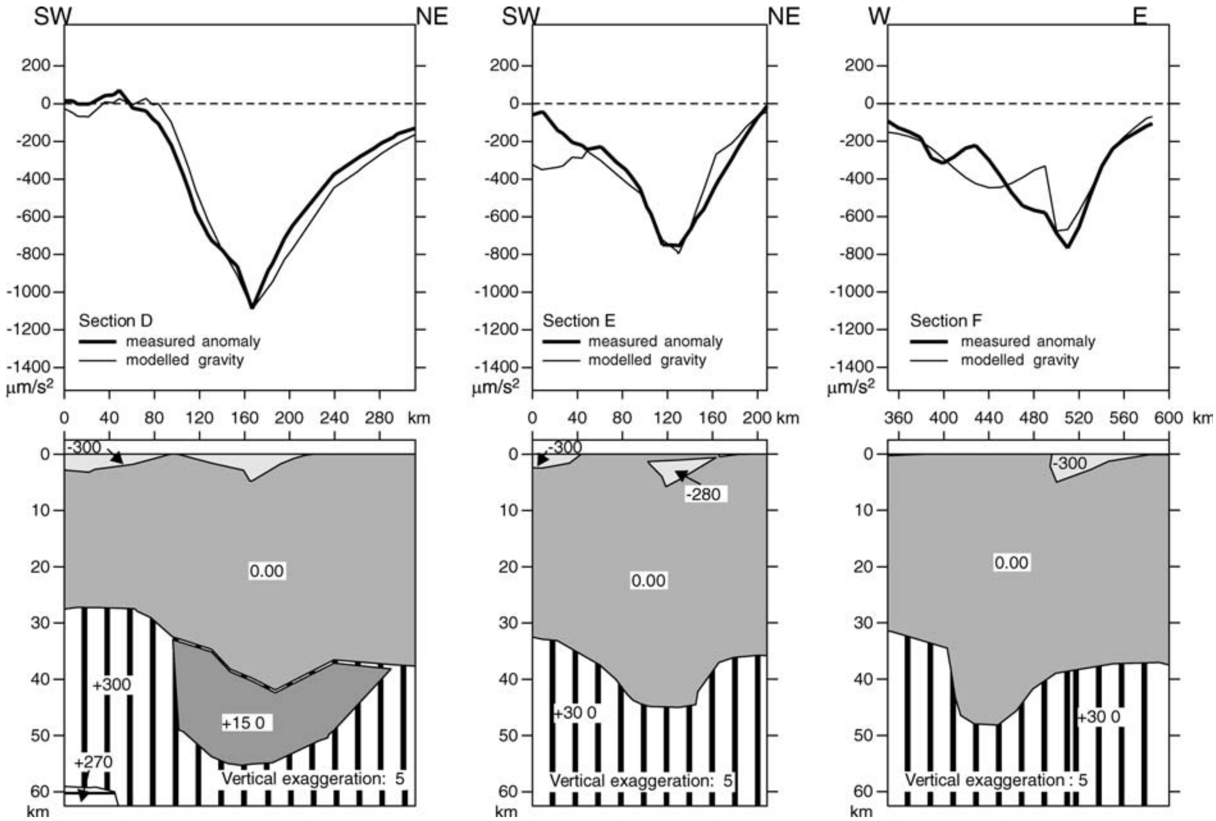

Fig. 10 Cross-sections of the final model in the Eastern Carpathians. Numbers are densities (in kg m^{-3})

the topography of the Mohorovičić discontinuity is not well known in this area (Lăzărescu et al. 1983; Rădulescu 1988; Sollogub et al. 1988b) we have significantly departed from the initial model. Horváth (1993) suggested that the rise of the Moho under the Southern Carpathians towards the Transylvanian basin is very steep. 2-D gravity modelling also suggested this Moho configuration (Szafián et al. 1997). In the present modelling it was necessary to emphasize this steep transition and produce an almost step-like contact between the two crustal units (Figs. 11, 12).

The other area of significant changes in the initial Moho topography is the Vrancea zone, the southernmost part of the Eastern Carpathians. The 50–53 km crustal thickness of Horváth (1993) was inferred from one deep seismic section (Sollogub et al. 1988b), which had imaged only discontinuous reflectors. Other studies (Rădulescu et al. 1976; Cornea et al. 1981) locate the base of the crust in the depth of 40–47 km.

The gravity anomalies produced by the initial model, which took into account the 50–53 km thick crust in the Vrancea zone, were 300–400 $\mu m\ s^{-2}$ lower than the observed ones (Fig. 6). In order to obtain a better fit, the model Moho had to be displaced upwards by 4–6 kilometres. This conclusion is at odds with the results of Bielik and Mocanu (1998), who carried out 2-D model calculations and found that the observed gravity anomalies can be reproduced even with a 53-km thick crustal root. However, in our opinion, the present 3-D model calculation provides a more reliable approximation of the real geometry of the major density boundaries (see discussion below). Our conclusions are further supported by seismic refraction data in the SE Carpathians (Hauser et al. 2001), which indicate 41 km as a maximum Moho depth along a section NE of the maximum Moho depth of our model.

Southern part of the Pannonian basin

The crustal thickness minimum (less than 22 km) of the starting model was located in the southern part of the Pannonian basin, following the interpretation of Aljinović (1987). However, the position of this feature is uncertain, because another study (Dragašević 1987) shifts this minimum approximately 50 km to the northeast, while the most recent Moho depth map of Serbia (Dimitrijević 1995) shows this minimum shifted even more to the north.

Figure 6 shows that this crustal thickness minimum is untenable: a significant Moho updoming would result high positive gravity anomalies that could not be compensated by the relatively thin sedimentary cover. As the re-evaluation of deep seismic lines (Lenkey et al. 1998; Lenkey, 1999) did not confirm the existence of this minimum either, it was disregarded. The modified crustal density structure is in a much better agreement with the observed gravity anomalies.

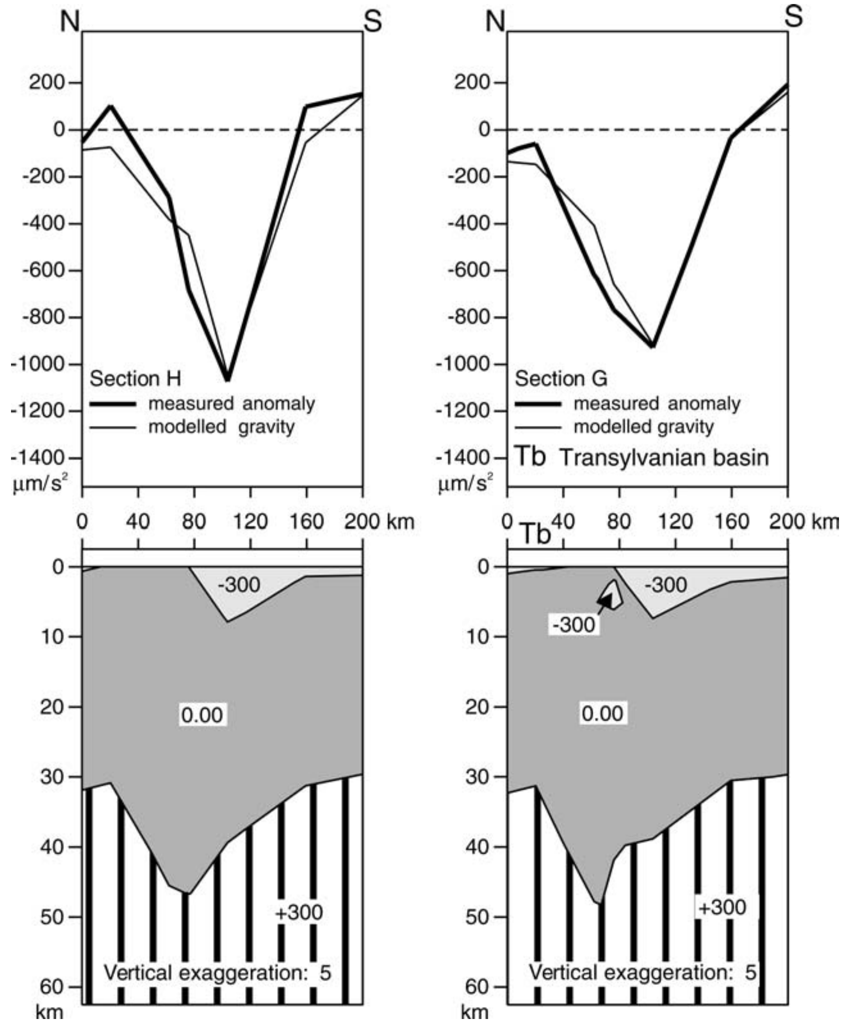

Fig. 11 Cross-sections of the final model in the area of the Southern Carpathians. The minimum zone of the Bouguer anomaly field is well reproduced after a deep foreland basin and a steep rise of the Moho towards the Transylvanian basin were introduced. Numbers are densities (in kg m^{-3})

Another area which deserves attention in the southern part of the Pannonian basin is the region of the two deep subbasins: the Makó trough and the Békés basin (Fig. 2). The surprising positive gravity anomaly of these subbasins has received a significant attention and was explained by (1) Moho updoming (Nemesi and Stomfai 1992; Posgay et al. 1995, 1996), (2) large, very high density intrusions into the lower crust (Bielik 1991; Grow et al. 1994), (3) the combination of these two phenomena (Posgay et al. 1995; Szafián et al. 1997), (4) or even a fragment of obducted oceanic material in the upper crust (Grow et al. 1994).

Three-dimensional gravity modelling revealed that small scale Moho undulation cannot adequately explain the observed gravity anomalies, and some crustal mass surplus is indeed required. However, this extra mass is much smaller in size and relative density than the one inferred from the 2-D modelling. The reason for this is in the 2-D approach, where we presume that the density boundaries extend infinitely perpendicular to the section. Thus the compensation of a deep sedimentary basin of infinite size requires extreme mass surplus. With 3-D approach the gravity effects of the subbasins are reduced, therefore their compensation can be solved in a more reasonable way. Most probably the solution for the positive gravity anomalies above deep subbasins is hidden in the complex internal structure of the pre-Tertiary basement, and a detailed, combined analysis of gravity and seismic data—similarly to the Danube basin (Szafián et al. 1999)—completed with magnetic model calculations is required.

Figure 13 shows the difference between the observed anomalies and the Bouguer anomaly field generated by the final density model. The correlation coefficient (0.89) of the two fields indicates that the model calculations were successful in reproducing the pattern of the gravity anomalies in the Carpatho-Pannonian region. The standard deviation (~135 µm s^{-2}), however, shows that the local details are not modelled adequately.

Figure 14 summarizes the changes of the model structure relative to the starting model. Light grey shading denotes areas of low-density masses, dark shading points out high-density bodies, while horizontal lines show the position of the high-density crustal root under the TTZ. Low-density bodies represent mostly

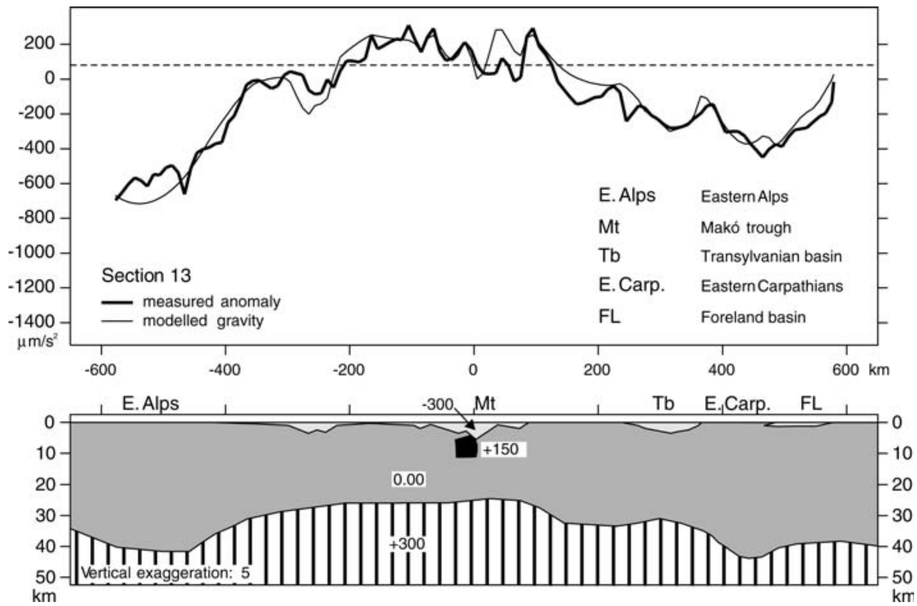

Fig. 12 Density model along Section 13 of the final model. Shallowing Moho and a high-density body under the deep subbasins in the central part of the plane could reproduce the positive anomaly, however the gravity minimum over the basement high suggests that a low-density unit is hidden in the basement rocks. Numbers are densities (in kg m^{-3})

sedimentary rocks, granites, crystalline complexes and light volcanics, high-density bodies probably originate from basic intrusions.

Regarding the lithospheric thickness, there was no need for major modifications in the position of the lithosphere to asthenosphere boundary. Although its topography significantly contributes to the gravity anomalies (Lillie et al. 1994), and errors in the position of this boundary affect the modelled gravity, it is a long wavelength component, and the relatively small scale differences between the modelled and observed field cannot be resolved with changes in the position of the lithosphere to asthenosphere boundary.

Crustal thickness and geodynamics

The previous chapter of this paper presented the variation of Moho topography in the Carpathian arc, the Dinarides and the Pannonian basin. The present Moho geometry reflects the end result of multi-phase geodynamic history of the studied area, which includes subduction, continental collision, large-scale strike-slip motions and lithospheric extension, as discussed above.

Differences in the geodynamical history of each tectonic unit might be responsible for the widely var-

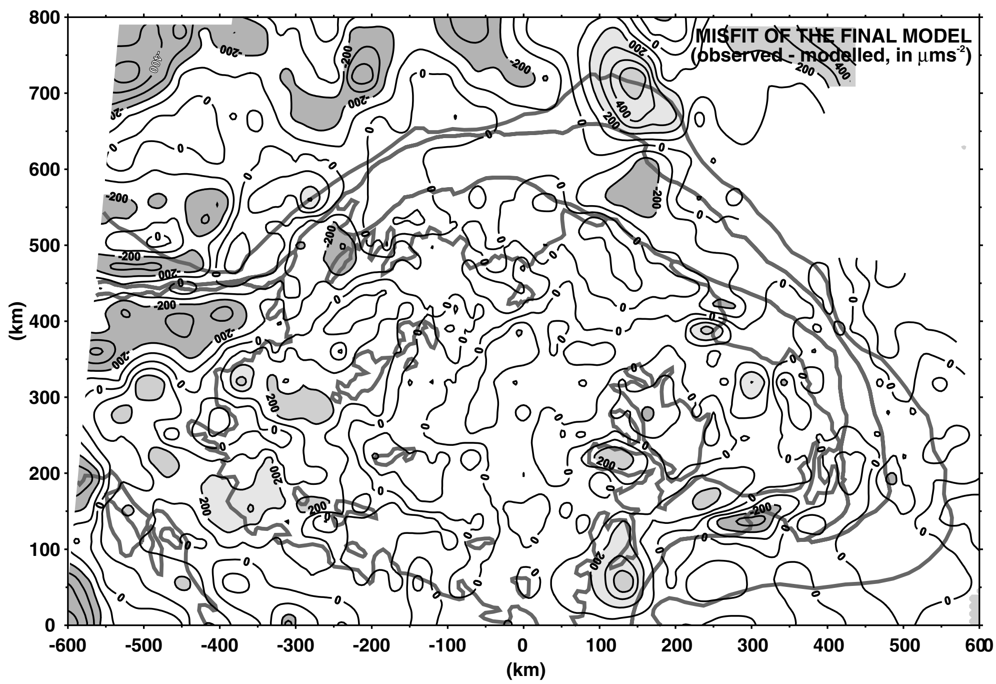

Fig. 13 The difference between the observed and modelled gravity fields in the final model. Spacing of isolines is 100 μm/s^2. *Dark grey shading* denotes areas where the final model predicted too high values, while *light grey* indicates too low values. The map clearly shows that the final model provides a much better fit than the initial one, but the complicated anomaly pattern of the Bohemian Massif and the Polish Platform, and the negative anomaly of the Eastern Alps are still not reproduced. *Light grey lines* indicate outlines of the foredeep basins, flysch belts and internal belts of Fig. 2

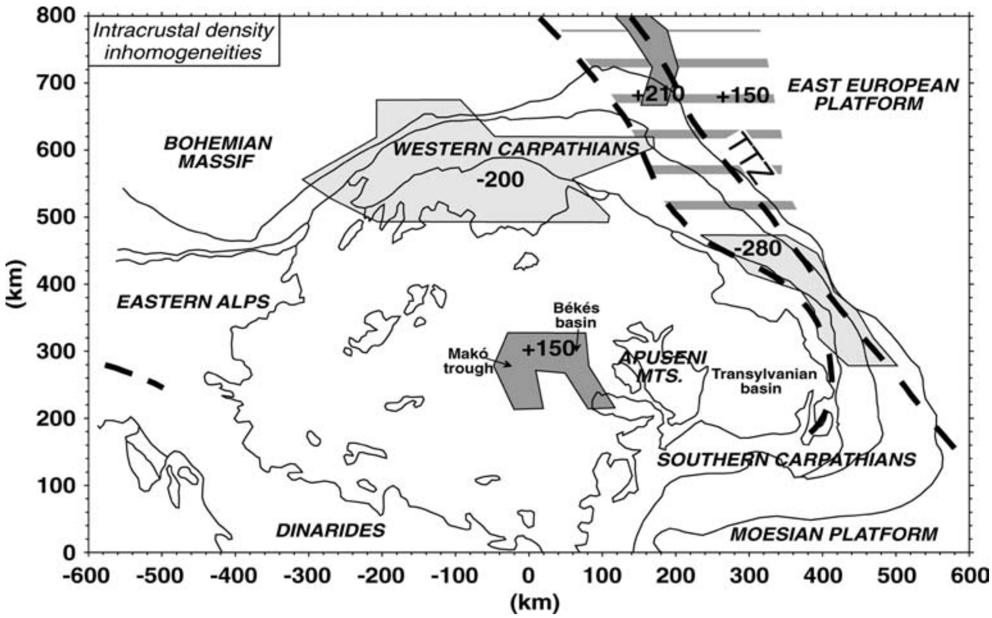

Fig. 14 Depth to the Moho in the final model together with locations and densities of the different units that were incorporated during modelling. Spacing of isolines is 2.5 km. *Horizontal lines* denote the extent of the high-density transitional unit under the Teisseyre-Tornquist Zone: the lower boundary of this unit is considered as Moho in this map. Numbers on isolines are kilometres below sea level. *Light grey shading* indicates low densities (negative density contrasts), *dark grey* represents high densities (positive density contrasts). Densities are given in kg m^{-3}. *Light grey lines* indicate outlines of the foredeep basins, flysch belts and internal belts of Fig. 2

iable Moho topography (Fig. 14). The Eastern Alps are a place of active continental collision, therefore a thick crustal root, which is actually a stack of European and Adriatic fragments, is justified. A similarly thick crust characterises the Eastern Carpathians, where collision with Pannonian continental fragments, together with inherited features of the TTZ have created a thick crustal root. During the last 30 million years, the Southern Alps and Dinarides were a transitional area from pure collision to transpression, accordingly they exhibit an overthickened to nearly normal continental crust. The Western Carpathians are characterized by transtensional-transpressional plate motions with no or minor collision, at least since the Middle Miocene, thus no crustal root can be observed. In contrast, the Southern Carpathians have an emphasised crustal root, despite the fact that it is also a transpressional orogen (Linzer 1996; Schmid et al. 1998). The corner effect of Moesia and the subsequent clockwise rotation and deformation of the Tisza-Dacia block gave rise to significant crustal stacking (e.g. Danubian units, see Matenco and Schmid 1999) in lack of subductible Carpathian flysch basin in this stretch of the European foreland, especially that the Moesian plate was thin and weak enough to get involved in subduction during Late Miocene times (Lankreijer et al. 1997; Matenco et al. 2003). Sperner et al. (2004) concluded that in the transition zone between the Southern and Eastern Carpathians soft collision took place, where only the transition zones of the Tisza-Dacia block and the European foreland thrust over each other but no excessive continental subduction occurred.

Regarding the Pannonian basin, Fig. 14 clearly shows that the eastern part of it is characterized by a uniform, NE–SW oriented mantle updoming, with the thinnest crust under the deepest subbasins. This direction is also valid for the Danube basin. Comparison of the lithospheric and the crustal thickness maps reveals that the detached extension of the crust and the subcrustal lithosphere in the Danube basin, inferred by 2-D gravity modelling (Szafián et al. 1999), is also supported by 3-D gravity modelling studies.

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