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Stripped image of the gravity field of the Carpathian-Pannonian region based on the combined interpretation of the CELEBRATION 2000 data

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Abstract: The Carpathian-Pannonian region is one of the areas of Central Europe with good coverage of geophysical and geological data. This is due to its complicated evolution that attracted scientific interest already in the past. In addition, several international seismic experiments were conducted here in the last 10 years. The model to be presented uses most of these available data to perform a combined gravity–seismic interpretation. The analysis of the gravity anomalies is performed in order to identify the sources of the anomalies, separate their effects and localize the lithospheric inhomogeneities. The gravity stripped image of the region reveals significant differences in the nature of the Microplates ALCAPA and Tisza-Dacia from the surrounding regions.

Key words: Carpathian-Pannonian region, Bouguer anomaly, 3-D density modelling, interpretation of the gravity field, lithospheric structure.

Introduction

The Carpathian Mountain belt, extending over 1300 km, is surrounded by the Pannonian Basin System, Eastern Alps, Bohemian Massif, Trans-European Suture Zone (TESZ) and the south-western part of the Precambrian East European Craton (EEC) (Fennosarmatian Craton). The present structural pattern of the Carpathians was formed by the Late Jurassic to Tertiary subduction-collision orogenic processes in the Tethyan mobile belt between the stable European Platform in the NE and the Apulia/Adria-related continental blocks (ALCAPA and Tisza-Dacia) drifting from the SW (e.g. Kováč 2000).

Recently, several international seismic refraction experiments (POLONAISE'97, CELEBRATION 2000, ALP 2002, SUDETES 2003), aimed at investigation of the lithospheric structure in this area, have been performed (e.g. Guterch et al. 2003a).

The work presented uses most of these data to perform a combined gravity and seismic 3-D modelling. For this purpose, complete Bouguer anomaly newly compiled (Bielik et al. 2006) was used for the forward modelling by means of the Interactive Gravity and Magnetic System (IGMAS) (e.g. Götze 1976; Götze & Lahmeyer 1988). All additional geological and geophysical data available were combined into a 3-D structural image of the Western Carpathians and Pannonian Basin. It is impossible, however, to perform 3-D modelling of the Carpathian-Pannonian region without including the surrounding units. Therefore, this large-scale lithospheric model also comprises the TESZ, EEC, Bohemian Massif and Eastern Alps. It extends down to a depth of 220 km and is developed along 31 parallel cross-sections cutting the above named units.

Geological setting

The East European Craton (EEC), formed during the Precambrian, is composed of Proterozoic igneous and metamorphic rocks covered by Vendian and Paleozoic strata (Dadlez et al. 2005). It is divided from the younger Paleozoic platform to the SW by the Trans-European Suture Zone (TESZ). The TESZ is a broad (up to 200 km) zone, crossing Europe from the North Sea to the Black Sea. The north-eastern boundary of the TESZ in Poland is the fault zone, called the Teisseyere-Tornquist-Zone (TTZ) (Dadlez et al. 2005 and references therein). In Poland, the TESZ consists of several suspected terranes accreted to the south-western margin of the EEC during the Paleozoic (e.g. Winchester et al. 2002; Guterch & Grad 2006). These terranes are referred to as Bruno-Silesian (called also Upper-Silesia), Małopolska and Łysogóry blocks. The latter two are exposed in SE Poland in the Holy Cross Mountains (HCM) and are separated by the Holy Cross Fault. The SW part of the HCM belongs to the marginal parts of the Małopolska block (Kielce Unit) and the NE part of the HCM belongs to the Łysogóry block (Łysogóry Unit) (e.g. Schätz et al. 2006). Both blocks are interpreted as Baltica derived terranes (e.g. Malinowski et al. 2005; Janik et al. 2005). The Bruno-Silesian block seems to differ from the Małopolska. These two units have different stratigraphic development and are separated by a 500 m wide zone, referred to as the

Kraków-Lubliniec Zone (Buła et al. 1997). Hence the Bruno-Silesian block is interpreted as a fragment of Gondwana (e.g. Malinowski et al. 2005). However, some authors (e.g. Schätz et al. 2006 and references therein) interpreted all three terranes as exotic terranes of Gondwanan provenance.

The Bohemian Massif, forming the easternmost part of the Variscan belt, is the largest stable outcrop of pre-Permian rocks in Western Europe. It consists mainly of metamorphic rocks, granites, and subordinate fossiliferous Paleozoic rocks (Hrubcová et al. 2005 and references therein).

The Western Carpathians belong to the ALCAPA microplate (e.g. Kováč 2000), which reaches the Pieniny Klippen Belt (PKB) in the north, and the Tisza-Dacia microplate in the south. The ALCAPA microplate is thrusted over the Tisza-Dacia microplate along the Mid-Hungarian Line, which is a Tertiary strike-slip fault zone (e.g. Plašienka et al. 1997). The Western Carpathians comprise, from north to south: the Outer Western Carpathians, Pieniny Klippen Belt (PKB) and the Central Western Carpathians. The Carpathian Foredeep is located in front of the Outer Carpathians along the entire Carpathian orogen (Fig. 1).

The Pannonian Basin System (PBS) was formed as a back-arc system due to the lithospheric extension and mantle upwelling behind the Carpathian arc during two stages (e.g. Horváth 1993; Kováč 2000). The driving mechanism for the extension is thought to be, traditionally, the subduction roll-back of the European Platform (e.g. Royden et al. 1983; Konečný et al. 2002; Szabó et al. 2004). The PBS is filled with Tertiary and Quaternary strata that reach, in some places, more than 6 km (e.g. Kováč 2000; Makarenko et al. 2002).

Review of results of the previous geophysical investigations

The former geophysical investigations provided information on the crustal thickness, revealing the crust-mantle boundary (Moho) to be very shallow in the Pannonian Basin. The Moho deepens towards the Carpathians to the north and east, as well as towards the Bohemian Massif and the Eastern Alps in the west (e.g. Horváth 1993; Šefara et al. 1996). Similarly to the Moho, also the lithosphere-asthenosphere boundary (LAB) in the Pannonian Basin region is very shallow. According to the seismological data, magnetotelluric sounding and geothermal measurements (e.g. Babuška et al. 1987; Praus et al. 1990; Horváth 1993; Čermák 1994), the lithosphere-asthenosphere boundary in the Pannonian Basin is at depths of 60 to 80 km. More recent data based on the 2-D integrated

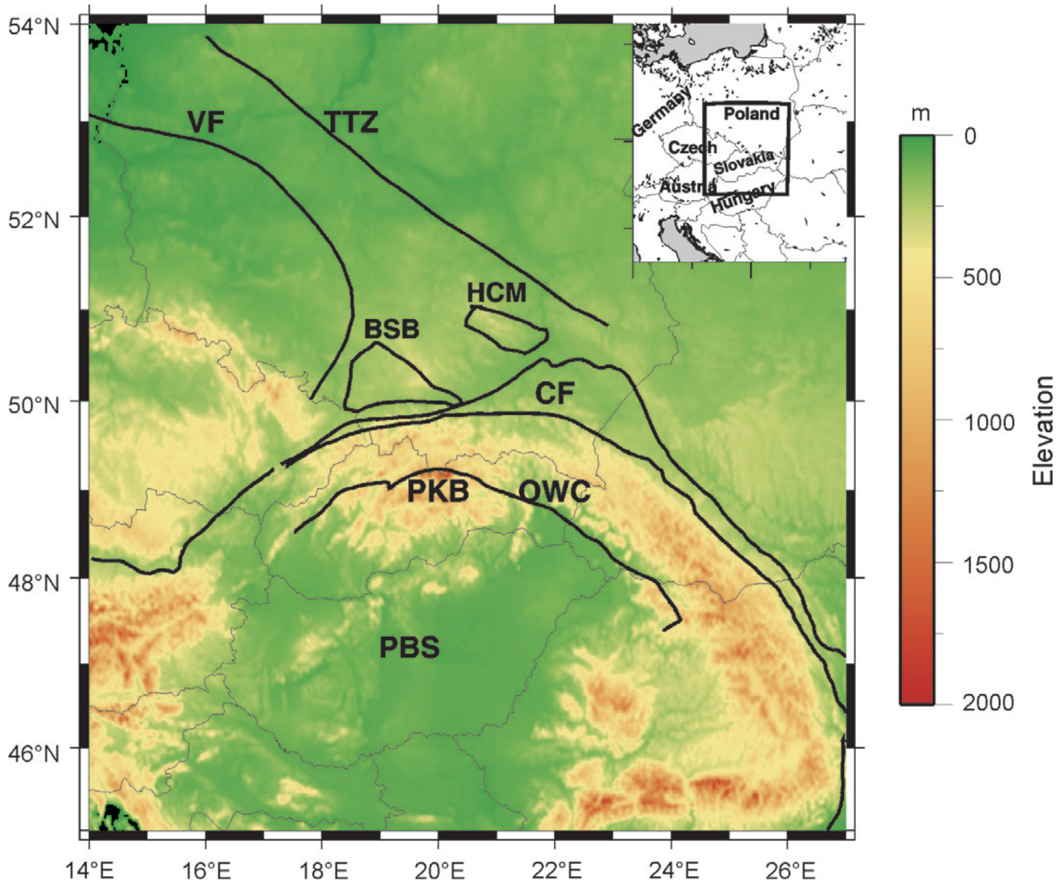

Fig. 1. Location map. Acronyms stand for: **PBS** — Pannonian Basin System, **PKB** — Pieniny Klippen Belt, **OWC** — Outer Western Carpathians, **CF** — Carpathian Foredeep, **BSB** — Bruno-Silesian Block, **HCM** — Holy Cross Mts, **TTZ** — Teisseyre-Tornquist Zone, **VF** — Variscan Front.

modelling combining the heat flow density distribution, absolute topographic elevation, gravity data and geoid (Zeyen et al. 2002; Dérerová et al. 2006) show the Pannonian Basin LAB to be at a depth of 80 km. The S-wave receiver function method reveals that the LAB in the northern edge of the Pannonian Basin is at a depth of 75 km (Geissler et al. 2007). Petrological analysis of the upper mantle xenoliths also confirms that a significant mantle uplift (50–60 km) occurred beneath the Pannonian Basin (Falus et al. 2000). Additionally, according to the global thermal model for the continental lithosphere of Artemieva (2006), the lithosphere of the Pannonian Basin region is 50 to 100 km thick.

Maps of heat flow density distribution show a clear difference between the Pannonian Basin System and the surrounding units. While the Bohemian Massif, Carpathian Mountains and the European Platform are characterized by medium values of 40 to 70 mW/m^2, the heat flow in the PBS reaches 80–130 mW/m^2. Higher heat flow values of 80–100 mW/m^2 occur partly also in the Eastern Alps (e.g. Pollack et al. 1993; Čermák 1994; Lenkey et al. 2002).

The region of Central Europe is also well covered by gravimetric and magnetic measurements. The Bouguer anomaly used in this work was compiled based on the nationally acquired data of Slovakia, Poland, Hungary, the Czech Republic and Austria by Bielik et al. (2006). The Bouguer anomalies are characterized by low values of some −20 to −65 mGal (1 mGal = 1×10^{-5} m/s^2) along the Western Carpathians (Central and Outer) and drop down to less than −120 mGal above the Eastern Alps and the Eastern Carpathians. The Pannonian Basin, Bruno-Silesian block and so-called Małopolska High in southern Poland, have positive values of 0–20 mGal (Fig. 2). The Małopolska High (e.g. Grabowska & Bojdys 2001 and references therein) has two parts, distinctive also in the Bouguer gravity anomaly. The NE part, located on the south-western edge of the EEC is called the Lublin High. The SW part belongs to the

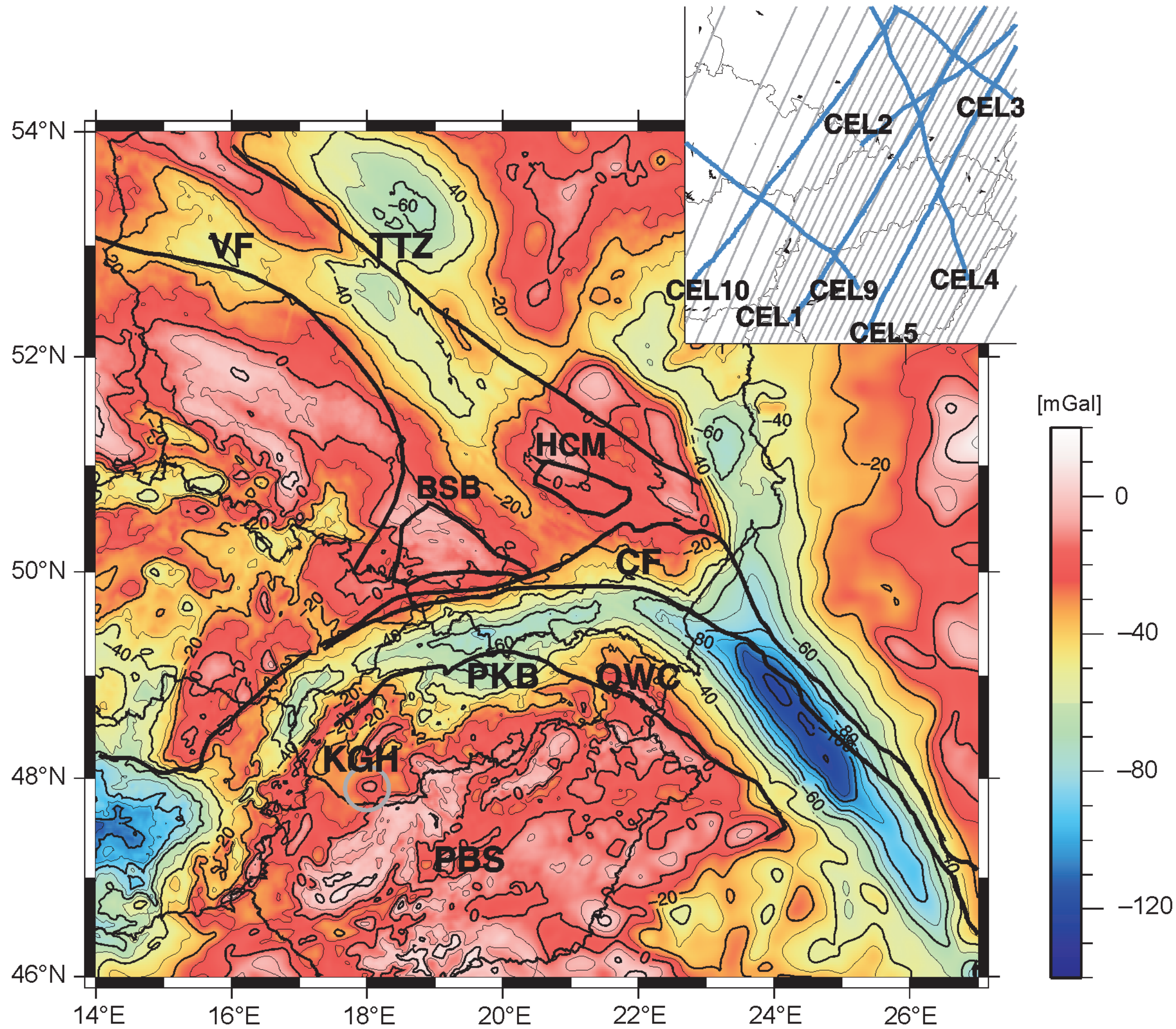

Fig. 2. Bouguer anomaly modified after Bielik et al. (2006). Acronyms stand for: **PBS** — Pannonian Basin System, **PKB** — Pieniny Klippen Belt, **OWC** — Outer Western Carpathians, **CF** — Carpathian Foredeep, **BSB** — Bruno-Silesian Block, **HCM** — Holy Cross Mts, **TTZ** — Teisseyre-Tornquist Zone, **VF** — Variscan Front. The Małopolska High, stretching from the HCM to the southern edge of the EEC, is separated by the TTZ into two parts. The Kolárovo gravity high (**KGH**) is marked by a circle. The location of the cross-sections of the 3-D model (thin grey lines) and of the CELEBRATION 2000 experiment (thick black lines) are also shown (upper right corner).

Łysogóry and Małopolska blocks and partly overlaps with the Holy Cross Mts (HCM, Fig. 2).

The station complete Bouguer gravity anomaly is tied to the IGSN71 datum (e.g. Torge 1989). However, although this datum has been conventional for more than 30 years, some works (e.g. Grabowska & Bojdys 2001; Malinowski et al. 2005; Janik et al. 2005) still use data in the Potsdam Gravity System, shifted with respect to the IGSN71 by ~14 mGal. For this reason, their values of the Bouguer gravity anomaly in Poland are higher than those shown here in Figs. 2 and 3.

Constraining data and indirect density determination

The large-scale international **C**entral **E**uropean **L**ithospheric **E**xperiment **b**ased on **R**ef**ra**ction (**CELEBRATION 2000**) was conducted in June 2000, as a joint experiment of 28 institutions in Europe and North America (Guterch et al. 2003b). The data were collected along ~17 profiles, giving a total length of 8,900 km. So far, 7 of these profiles have been processed and published. These are profiles CEL01, CEL02, CEL03, CEL04, CEL05, CEL09 and CEL10 (Fig. 2) (Malinowksi et al. 2005; Janik et al. 2005; Hrubcová et al. 2005; Środa et al. 2006; Grad et al. 2006 and Růžek et al. 2007). The CELEBRATION data together with the results of the previous investigations of the region (e.g. Bielik et al. 2004 and references therein) and information from 2-D integrated modelling of Dérerová et al. (2006) were the primary input for the 3-D modelling performed within this study. All of these data constrain the geometry of the structures modelled (the depth to major boundaries, such as sediments, Moho, lithosphere-asthenosphere boundary). In addition, the P-wave seismic velocities from the CELEBRATION 2000 experiment were converted into densities using the empirical relationships of Christensen & Mooney (1995) and Sobolev & Babeyko (1994).

There are significant differences reaching 0.05–0.13 Mg/m^3 (1 Mg/m^3 = 1 g/cm^3 = 1000 kg/m^3) in the calculated density values obtained from the two empirical relationships (Table 1). Thus, there is a wide range of densities corresponding to the observed seismic velocities. Additionally, the empirical velocity-density curves, or linear relationships, provide only a mean density of a particular rock unit, which internally may be quite variable. It has been demonstrated by various studies (e.g. Hacker & Abers 2004) that both velocities and densities strongly depend on the composition of rocks. Thus, rocks with similar velocities may have significantly different densities and vice versa. The determination of densities based only on a velocity-density relationship is hence one of the major problems related to developing combined gravity and seismic models. Therefore, all information related to the composition of the structures modelled should be considered while gravity modelling is performed. However, a density model always approximates the real structures in a simplified way, with larger units extending sometimes over tens of kilometers (in distance and depth). These units are usually characterized by a single (constant) density value. For this reason, although a significant scatter around the mean value is associated with empiri-

Table 1: Densities for particular depths derived from the P-wave seismic velocities (vp) using the approach of Sobolev & Babeyko (1994) (S&B) and Christensen & Mooney (1995) (C&M). The temperatures for the approach of Sobolev & Babeyko (1994) were calculated assuming surface heat flow of 40–50 mW/m^2 for cold, 60 mW/m^2 for medium, 70 mW/m^2 for warm and 80–90 mW/m^2 for a hot region. The relationship of Christensen & Mooney (1995) is a nonlinear relationship derived for all rock types (including the upper mantle rocks) and is recommended for crust-mantle sections.

Depth [km]	vp [km/s]	rho (S&B) cold [Mg/m^3]	rho (S&B) medium–warm [Mg/m^3]	rho (S&B) hot [Mg/m^3]	rho (C&M) [Mg/m^3]
10	**6.05**	2.64	2.65	2.65	2.73
10	**6.1**	2.66	2.67	2.68	2.75
10	**6.2**	2.70	2.71	2.72	2.79
20	**6.3**	2.76	2.78	2.79	2.82
20	**6.4**	2.81	2.82	2.84	2.85
20	**6.5**	2.85	2.87	2.88	2.89
30	**6.6**	2.91	2.93	2.95	2.94
30	**6.7**	2.95	2.97	3.00	2.97
30	**6.8**	2.99	3.02	3.04	3.00
30	**6.9**	3.04	3.07	3.09	3.03
30	**7.0**	3.09	3.12	3.14	3.06
30	**7.15**	3.16	3.19	3.21	3.11
40	**7.1**	3.15	3.18	3.21	3.12
40	**7.2**	3.20	3.23	3.26	3.15
40	**7.3**	3.24	3.28	3.31	3.18
40	**7.4**	3.29	3.33	3.36	3.20

Table 2: Observed P-wave seismic velocities (vp) from the CEL-profiles and densities employed in the 3-D density model. The densities of an alternative model reproducing the Małopolska High are marked by an ^{A}. The asterisks mark maximum values, which occur only locally and should not be taken as average values.

Units	P-wave velocity [km/s]	rho-employed [Mg/m^3]
sediments	2.3–4.5	2.45
upper crust PB	5.8–6.0 (6.2–6.3*)	2.7
middle crust PB	6.0–6.4	2.8
lower crust PB	6.25–6.55	2.9
sediments (IWC)	< 4.0	2.62
upper crust WC	5.65–6.15	2.67
middle crust WC	5.95–6.4	2.73
lower crust WC	6.4–6.75	2.9
sediments (OWC)	< 3.0–5.6	2.6
sediments (CF)	< 4.0	2.5
sediments TESZ	2.0–5.3	2.5 and 2.62
upper crust TESZ	5.5–6.2	2.7–2.71
middle crust TESZ	6.2–6.6 (6.88*)	2.96 (2.91^A)
lower crust TESZ	6.6–7.1 (7.2*)	3.13 (3.17^A)
middle crust TESZ (15–30 km depth)	7.15	3.12 (3.1^A)
sediments	2–5.4	2.6
upper crust EEC	6–6.4	2.78
middle crust EEC	6.4–6.65	2.9
lower crust EEC	6.7–7.05 (7.2*)	3.05
sediments	2.5–5.4	2.54
upper crust BM	5.8–6.4	2.71
middle crust BM	6.3–6.7 (7.2*)	2.9
lower crust BM	6.8–7.4 (7.8*)	3.1
sediments	≤ 5.5	2.55
upper crust EA	5.8–6.2	2.7
middle crust EA	6.4–6.8	2.88
lower crust EA	6.8–7.4	3.1

cal velocity-density relationships (Barton 1986 and references therein), they may be applied to larger-scale lithospheric models. Moreover, existing seismic models or boreholes also constrain the depth to the major boundaries, such as sedimentary basins and the Moho. The structures modelled are, therefore, a trade-off between the seismic models (extent of the structures and their velocities) and gravity anomalies.

The relationship of Sobolev & Babeyko (1994) requires P/T conditions in order to calculate the in situ densities from the seismic velocities. This is very convenient for the modelling of Central Europe because here the various units are characterized by extremely different temperature conditions. Therefore, this relationship was found more appropriate for the determination of densities from seismic velocities. The temperatures at various depths are calculated on the basis of the known surface heat flow values (Pollack et al. 1993; Majorowicz et al. 2003) according to the heat conduction equation (e.g. Turcotte & Schubert 2002). In situ pressure is estimated as a function of depth, standard density and overpressure factor.

The summary of the observed P-wave velocities (Table 2) shows significant differences between the units considered in the model (the Western Carpathians, Pannonian Basin, Bohemian Massif, TESZ, EEC and Eastern Alps).

Discrepancies in the interpretation of the CELEBRATION 2000 data

The crustal root of more than 50 km in the TESZ area interpreted by Guterch et al. (1986) and shown also in the maps of the Moho depth of Bielik (1999 and references therein) was not approved by the CELEBRATION results (e.g. Dadlez et al. 2005; Guterch & Grad 2006). Moreover, there are also some discrepancies among the CELEBRATION profiles themselves, since 3-D seismic interpretation has not yet been performed and all of the above mentioned CEL-profiles were processed in 2-D. The most significant differences (exceeding the errors estimated) occur at the intersection points of the following profiles:

— CEL01 and CEL02: thickness of sediments differ by 2 km, the thickness of the upper crust and its velocity by 3–4 km and 0.35 km/s, respectively.

— CEL01 and CEL03: velocity at 15 km depth differs by 0.68–0.88 km/s.

— CEL01 and CEL04: the upper-crustal velocity is different by 0.15 km/s and the depth to the interface between the upper and middle crust differs by at least 1 km; the depth to the Moho varies by 3 km.

— CEL02 and CEL03: the lower-crustal high-velocity body is located at different depths, causing a velocity-difference of ~0.3 km/s at depths of 30–40 km.

— CEL02 and CEL04: the depth to the boundary between the upper and middle crust and to the Moho differs by 2 km.

— CEL03 and CEL05: the thickness of the sediments varies by 6 km and the velocities at ~10 km depth are different by 0.5 km/s and 0.15 km/s.

— CEL04 and CEL10: the velocities in the upper and lower crust differ by 0.2 and 0.3 km/s, respectively; the difference to the depth of the Moho is 4 km.

Due to these discrepancies, we believe a 3-D approach using these data as constraining information is very useful in order to bring more insights into the structural image of the area.

Density modelling

The 3-D modelling is performed using IGMAS, where the modelled geological bodies are approximated by polyhedra of constant density (e.g. Götze 1976). The structures are defined along 2-D cross-sections that are connected via triangulation into a 3-D volume. Thus, the geometry of the geological bodies between the cross-sections is interpolated. Therefore, in order to obtain more reliable results, a greater number of 2-D planes must be included. The model presented is developed along 31 cross-sections, separated by 20 km across the Western Carpathians and Pannonian Basin and 40 km in the Bohemian Massif and Eastern Alps. They are parallel to each other and are almost identical with the direction of the CEL05 profile. The model consists of 41 bodies, representing the above mentioned units, thus giving a rather complex structure. Constraining data can be visualized in IGMAS interactively along the modelled cross-sections by means of the GIS functions (e.g. Schmidt & Götze 1999). This is a great advantage during the modelling of large areas including various datasets.

In general, each of the units consists of four crustal layers: sediments, upper, middle and lower crust (Fig. 3). The densities employed in the 3-D model are summarized in Table 2. The densities of the sediments in the Pannonian Basin System and Western Carpathians are based on the previous investigations (e.g. Makarenko et al. 2002; Bielik et al. 2005 and references therein). The densities of the sediments in the other units are mainly based on the P-wave velocities of the CEL-models and relationships of Gardner et al. (1974) and Wang (2000), determined for sedimentary rocks. The upper mantle is in general divided into two parts, the lithospheric and asthenospheric mantles. The densities of the lithospheric mantle underneath the ALCAPA and Tisza-Dacia microplates and the European Platform (EP) are not constant. The colder EP upper mantle was assigned a density of 3.36 Mg/m^3, which is by 0.03 Mg/m^3 greater than the density of the ALCAPA's lithospheric mantle (Fig. 3). This is due to the temperature difference of some 200–300 °C at depths of 50 and 100 km (e.g Artemieva 2006). According to the data of Kuskov & Kronrod (2006), the density decrease associated with higher temperatures (for constant pressure and composition) is in the order of −0.013 Mg/m^3 for +100 °C (meaning that −0.03 Mg/m^3 corresponds to ~230 °C difference in temperature). The asthenosphere, having a different composition than the lithospheric mantle, has lower Mg# (100×Mg/(Mg+Fe)) and higher density than the lithosphere (e.g. Poudjom Djomani et al. 2001 and references therein; Kuskov & Kronrod 2006). The asthenosphere was assigned a density of 3.38 Mg/m^3. The Pannonian Basin is, additionally, a region where a mantle upwelling takes place. The average density of the upwelling asthenospheric mantle at depths of 80 to 180 km has a low value of 3.3 Mg/m^3. This value was determined based on the calculations of Cella & Rapolla (1997), and it also agrees with previous models (e.g. Lillie et al. 1994; Bielik 1999).

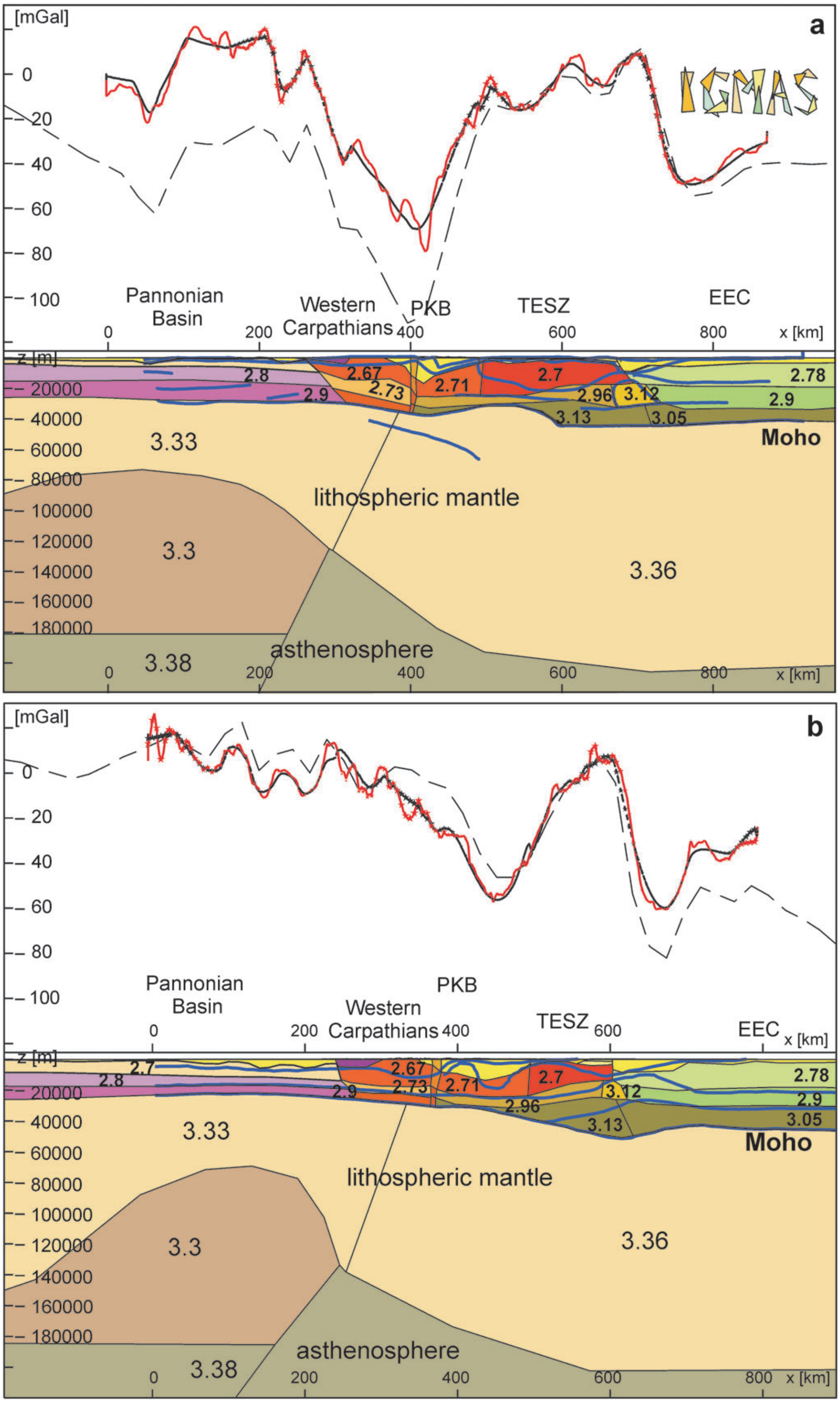

Fig. 3. Two of the cross-sections from the 3-D model, corresponding to the seismic profiles CEL01 (**a**) and CEL05 (**b**). The upper box of each frame shows the observed (red) and modelled (black) gravity anomaly. The 2-D gravity effect along each profile is marked by a dashed line. The lower box shows the structures modelled and densities assigned. The blue lines are boundaries from the seismic models after Środa et al. (2006) for the CEL01 and Grad et al. (2006) for the CEL05.

The employed densities are not relative to one reference density, but a three-layered reference model with negative densities is chosen instead. The reference model used for this area has two crustal layers. The upper crust (at depths 0–15 km) has a density of –2.67 Mg/m^3, while the lower crust (at depths 15–35 km) has a density of –2.9 Mg/m^3. These values are consistent with the velocity model IASP91 of Kennett & Engdahl (1991) and the global data of Christensen & Mooney (1995). The upper mantle (at depths of 35–220 km) has a density of –3.36 Mg/m^3, which is consistent with values for the subcontinental lithospheric mantle given by Poudjom Djomani et al. (2001).

Last but not least, it is important to mention differences between the 2-D and 3-D modelling. A geological structure can be treated as two-dimensional, if its length is much greater with respect to its width (e.g. Blakely 1996). This might be often the case in reality (e.g. rift and fractures zones etc.). However, due to the arcuate shape of the Carpathians, this condition in the Western Carpathians is usually not fulfilled. There is a possibility in IGMAS to compute 2-D gravity effect of the structures modelled along a particular cross-section. The differences between the calculated 2-D and 3-D effects are significant. If only the 2-D gravity modelling was performed for the profile CEL01, with the Moho depth fixed according to the seismic model, the density of the intra-crustal structures would be overestimated in the area of the microplate ALCAPA and the TESZ crust underlying the Outer Carpathians (Fig. 3a). In contrary, the crustal densities along the CEL05 profile would be underestimated for the microplate ALCAPA and overestimated for the EEC (Fig. 3b). 3-D gravity modelling thus provides more realistic results for regions characterized by complicated and non-linear geological structures. Therefore, it is more adequate than a 2-D modelling for interpretations of the gravity anomalies of such regions.

Results and interpretation

The microplate ALCAPA is separated from the platform by the Pieniny Klippen Belt (PKB), extending through the whole crust. The ALCAPA (Pannonian Basin System and Western Carpathians) has a constant density lower crust, but the two-layered upper crust and sediments have different densities (Table 2). North of the PKB, the Outer Carpathians and the Carpathian Foredeep are underlain by the crust of the TESZ. According to the seismic interpretations, the TESZ and EEC have similar structure of the middle and lower crust, but significantly different upper crust (low-velocity TESZ upper crust) (e.g. Dadlez et al. 2005; Guterch & Grad 2006). This is also included in the 3-D density model presented (Fig. 3). However, the middle and lower crust of the TESZ in the density model have higher densities than the EEC middle and lower crust (Table 2). These higher-density crustal layers are required to reproduce the Małopolska High, as it was also suggested by Grabowska & Bojdys (2001). Alternatively, if the middle crust of the TESZ is modelled with a density almost identical to the EEC middle crust, the density of the TESZ lower crust must be increased (Table 2, values indicated by [A]). Additionally, a high-velocity body was interpreted in the transition area between the TESZ and EEC in the middle crust underneath the Lublin High (EEC) along the CEL01, CEL02, CEL03 profiles (Malinowski et al. 2005; Janik et al. 2005; Środa et al. 2006). The velocities along the CEL05 (Grad et al. 2006) are also elevated in this region. Similarly, a high-density body in the middle crust is required in the density model in order to reproduce the gravity anomaly. It is a 3-D structure, stretching from the CEL05 profile in the south to ~60 km north of the CEL01 profile. Its high velocity of 7 to 7.15 km/s and density of 3.12 Mg/m^3 indicates mafic composition, similar to olivine gabbro or garnet granulite (Sobolev & Babeyko 1994; Christensen & Mooney 1995). Grabowska et al. (submitted) assume this intrusion to be due to the metamorphic processes, resulting in the increase of the density and variations of magnetic properties of rocks forming the crystalline crust of this unit.

Fig. 4a,b shows the depth to the bottom of the sediments and to the Moho obtained from the 3-D modelling with the available constraining data. The thickness of the sediments (Fig. 4a) is modelled according to the data compiled by Makarenko et al. (2002) and Bielik et al. (2005). However, slightly different densities are employed for the sediments and upper crust in the 3-D model presented and, therefore, the results of 3-D modelling differ from the above mentioned results. The Moho along the CEL-profiles is consistent with the seismic data. The minimum crustal thickness of ~22 km is located along the CEL05 profile (Grad et al. 2006) and its vicinity, which corresponds to the centre of the Pannonian Basin (Fig. 4b). This observation is also consistent with the xenolith data (Szabó et al. 2004). The xenoliths from the central part of the basin are significantly deformed because the active rifting and lithosphere thinning mostly took place here. In contrary, the xenoliths from the margins of the basin are only slightly deformed or undeformed. The Danube Basin is characterized by a crustal thickness of 28–30 km, increasing to 35 km toward the west. The Central Western Carpathians have 28–35 km thick crust, while the crust beneath the Outer Western Carpathians and the Carpathian Foredeep is 35 to 43 km thick. The maximum crustal thickness of ~50 km is modelled beneath the TESZ along the CEL05 profile (e.g. Guterch & Grad 2006) and Eastern Alps (e.g. Behm et al. 2007).

The gravity stripping is performed in order to analyse the different components of the gravity signal. The gravity effect of the sediments (Fig. 4c) was calculated in IGMAS using the density differences of the sediments with respect to the density of the upper crust employed in the model for each unit (Table 2). The gravity effect of the Moho was calculated using the Parker algorithm (Parker 1972), with a constant density difference of 0.3 Mg/m^3 at the Moho for the whole region (Fig. 4d).

The sediment stripped map (Fig. 5a) in the area of the Central Western Carpathians shows the negative effect of the thick low-density upper and middle crust (according to the 3-D model the thickness reaches ~25 km). In contrast, the Pannonian Basin is generally characterized by a positive anomaly of ~20 mGal. In the eastern part of the PBS, the gravity high reaches even 40 to 50 mGal, reflecting the extremely shallow Moho in this region (Fig. 4b). A complete stripped map (Fig. 5b), however, clearly shows similarities between the PBS and Western Carpathians. When the effect of the Moho

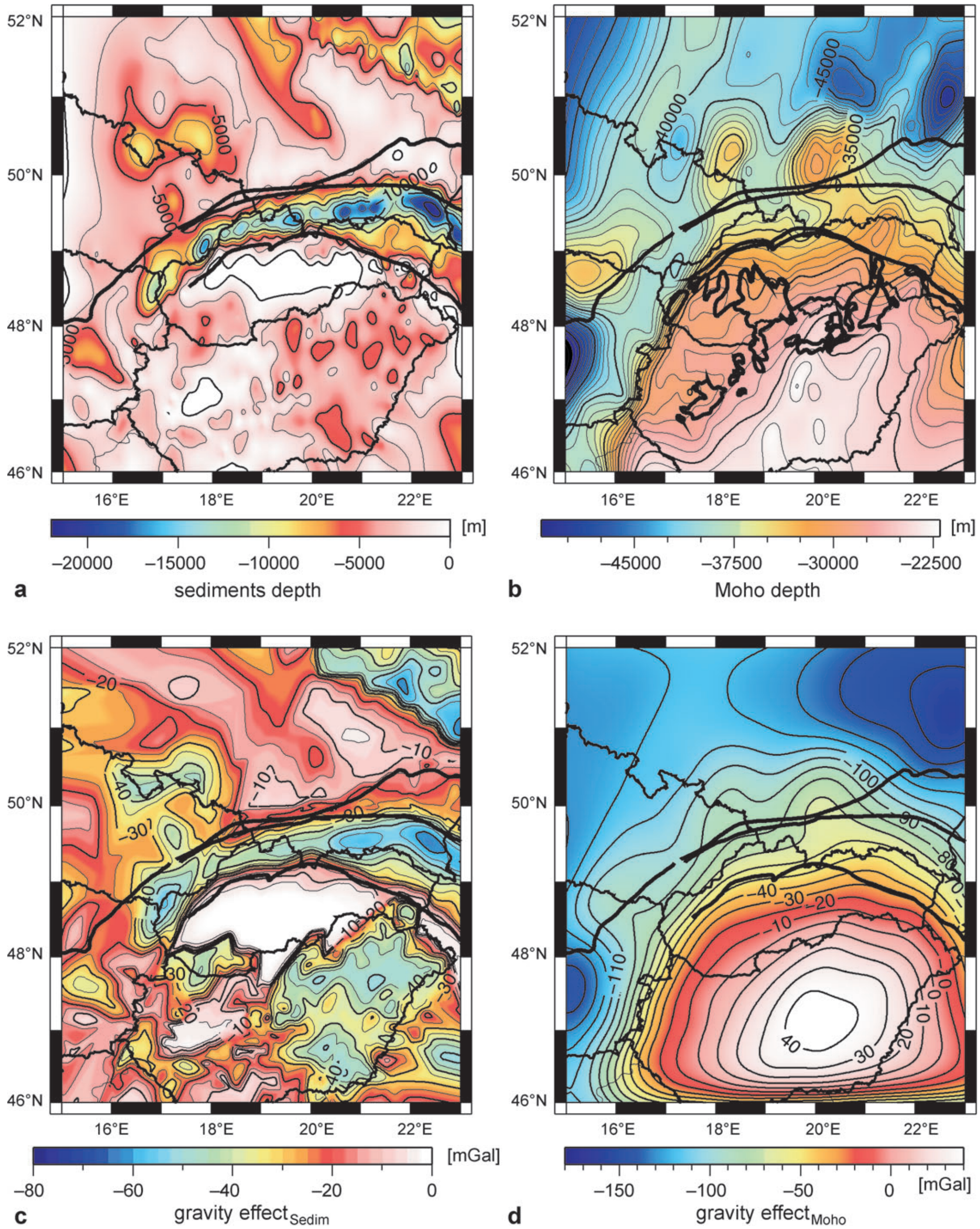

Fig. 4. The depth to the bottom of the sedimentary basins (**a**), to the Moho (**b**) and the calculated gravity effects of the sediments (**c**) and Moho (**d**). The thick lines mark the units of the Western Carpathians also shown in Figs. 1 and 2.

(shallow in the PBS, deeper in the Western Carpathians) is removed, the residual "lithospheric" anomaly reveals the lithosphere of the microplates ALCAPA and Tisza-Dacia to be characterized by remarkably lower anomalies than the surrounding regions. The greatest gradient coincides with the location of the PKB (Fig. 5b), separating the microplate ALCAPA from the platform. This indicates that the lithospheric structure of the microplates ALCAPA and Tisza-Dacia in terms of density distribution is very different from the European Platform and the Eastern Alps.

The sedimentary basins with maximum infill of some 6 km in the Pannonian Basin are associated with moderate gravity lows of the Bouguer anomaly, reaching 0 to −12 mGal (Fig. 2). This is also the case in the Eastern Slovak Basin, where areas with sediment infill of ~5 km are characterized by gravity anomalies of +5 to −7 mGal. However, the gravity effect of these basins filled with sediments of low density (2.45 Mg/m^3) reach ~−45 mGal (Fig. 4c). This negative effect is partly compensated by positive effect of the shallow Moho that is in the order of 0–40 mGal. Additionally, lower

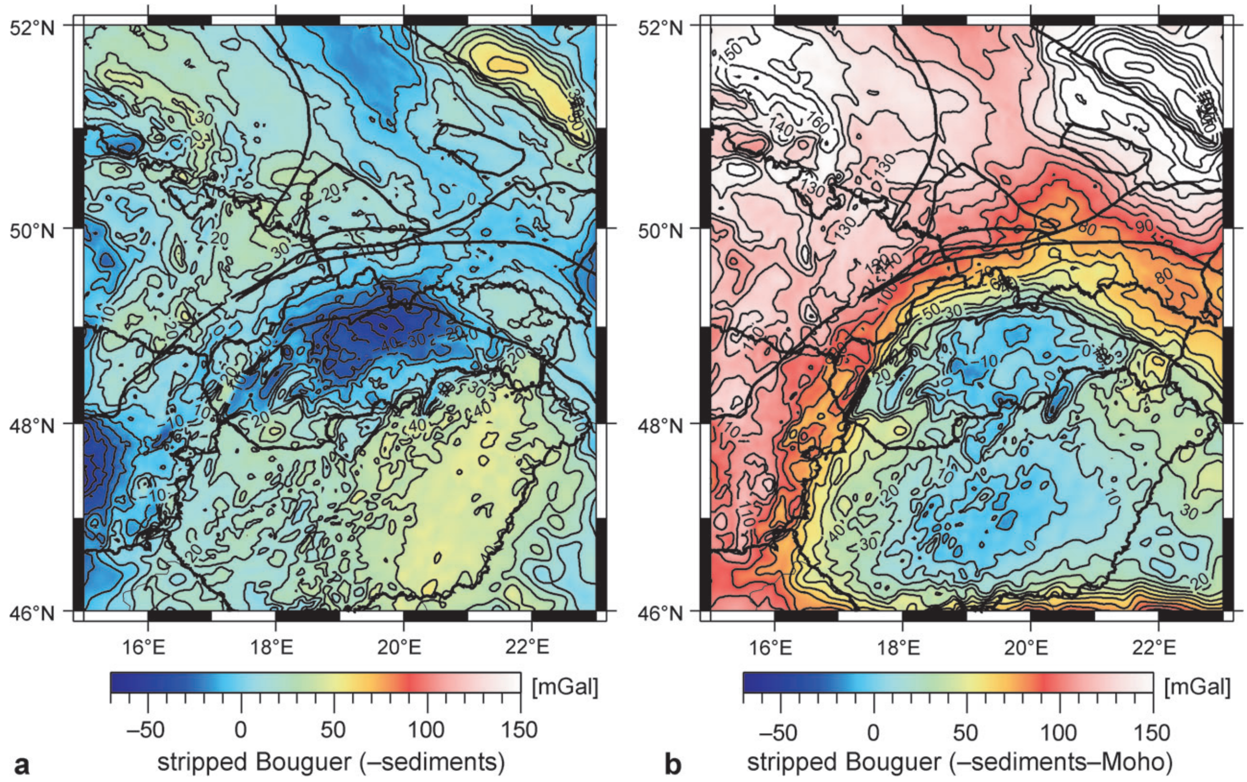

Fig. 5. Sediment stripped map (**a**) and the complete stripped map (**b**), also referred to as a residual "lithospheric" anomaly. The thick lines mark the units shown also in Figs. 1 and 2.

crustal intrusions, reaching some 10 km in depth (e.g. Ádám & Bielik 1998), also compensate for the negative effect of the sediments. Similarly, Kolárovo gravity high is reproduced by a dense lower crust (Bielik et al. 1986), reaching depths of 9 km and a width of some 10 to 20 km that is included along 2 cross-sections in the 3-D model.

Conclusions and outlook

A 3-D forward modelling was performed for the Western Carpathians, Pannonian Basin System and the surrounding units. The model uses mainly data collected recently during the CELEBRATION 2000 experiment. It brings them into one structural image in order to study the lithospheric structure of this region. By means of the combined 3-D modelling, preliminary estimates of the density distribution of the crust and upper mantle, as well as the depths of the sedimentary basins and the Moho were derived. These data allowed the performance of gravity stripping, which is in the area of the Pannonian Basin crucial for the analysis of the gravity field. In this region, two pronounced features, namely the deep sedimentary basins and shallow Moho, with opposite gravity effects hinder the interpretation of the gravity field by means of filtering (e.g. in the wavenumber domain), estimating isostatic regional and residual fields or performing the gravity anomaly inversion. The results of the gravity stripping revealed the lithosphere of the ALCAPA and Tisza-Dacia microplates to be very similar and much less dense than the surrounding lithosphere.

The upper mantle of the Pannonian Basin, where an asthenospheric upwelling takes place, significantly differs from the surrounding regions. The upper mantle characterized by an asthenospheric upwelling is, with respect to the "normal" upper mantle, anomalous in terms of lithospheric thickness, temperature and density distribution. As it has been proved in the course of the 3-D modelling, the influence of the different upper mantle densities of the units modelled on the crustal structures is pronounced. Thus, the upper mantle density must be determined as precisely as possible, considering all available information. Therefore, in order to determine the upper mantle densities, the combined geophysical–petrological approach of Afonso (2006) will be applied. The mantle densities better constrained should improve the estimations of the densities and composition of the crust and enhance the localization of the lithospheric inhomogeneities.

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