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Review article

A review of geophysical studies of the lithosphere in the Carpathian–Pannonian region

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Abstract: Here, we revisit the most prominent features of the complete Bouguer anomaly map and their interpretation, along with the current knowledge of the lithospheric thickness in the Carpathian–Pannonian region. The stripped gravity map, i.e., the sediment-stripped complete Bouguer anomaly map, was used to interpret the most prominent highs and lows of the gravity field. The complete Bouguer anomaly data were used in structural density modelling and integrated geophysical modelling to determine or revise the previously known sources of the most pronounced gravity features of the region. The Carpathian gravity low was divided into three sub-lows: the Western, Eastern, and Southern. The Western Carpathian gravity low consists of the clearly distinguishable External and Internal lows, which are due to different causes. The source of the External Western Carpathian gravity low reflects the low-density sediments of the External Western Carpathians (2.49–2.59 $g\,cm^{-3}$) and the Foredeep (~2.43 $g\,cm^{-3}$), while the Internal Western Carpathian gravity low is explained by the upper crustal deficit mass, which is formed by the rocks of the Alpine Tatric and Veporic units. These tectonic units are built mainly from granites and crystalline schists, of which the average density (~2.70 $g\,cm^{-3}$) is lower than the average density of the lower crust of the Internal Western Carpathians (~2.90 $g\,cm^{-3}$). The main sources of the Eastern and Southern Carpathian gravity lows are the gravity effects of the crustal roots created by continental collision, the Foredeep, and the surface sediments of the External Carpathians. The Pannonian gravity high is caused by the expressive Moho elevation (24–26 km). Since the Pannonian Basin upper mantle, which is built by high-density peridotites or dunites, is located several kilometres closer to the surface, this rock material represents a great excess mass (high-density anomalous bodies). Based on the calculated stripped gravity map, several local gravity highs (>+50 mGal) have been recognised, and they are all located in the Danube Basin, the Transcarpathian Basin, the Békés Basin, as well as the Makó trough. Their sources are high-density crustal bodies (Eo-Alpine metamorphic complexes), whose apical parts reach depths of only 7 to 12 km. Finally, the expressive different depths of the lithosphere-asthenosphere boundary in the Western and Eastern Carpathians were explained by the different Neo-Alpine development of both orogens. The mantle lithospheric root (~240 km) in the Eastern Carpathians is results from the sinking of the upper part of the broken slab during the frontal continental collision. On the contrary, no thickening of the mantle lithosphere was observed in the junction zone of the Western Carpathians and the Bohemian Massif. The typical thickness of the continental lithosphere (~100 km) in this zone was explained by the oblique continental collision. The Pannonian Basin system is characterised by one of the thinnest continental crusts (~25 km) and lithospheres (~75 km) in the world.

Keywords: complete Bouguer anomaly, stripped gravity map, integrated geophysical modelling, lithosphere, gravimetric interpretation, Carpathian–Pannonian region

Introduction

The geology of Central Europe is very complex and consists of the Western European Paleozoic Platform (including the Bohemian Massif), the Precambrian Eastern European Craton, the Trans European Suture zone (TESZ), the Carpathian Orogen, and the Pannonian Basin System. The mosaic of platforms, orogenic arc, and related fore-arc and back-arc basins offer an exceptional opportunity to study the structure and composition of the lithosphere, as well as the interaction of lithospheric and asthenospheric processes during formation of these structures (e.g., Alasonati Tašárová et al. 2016; Šimonová et al. 2019). In addition, the Carpathian–Pannonian region, together with its neighbouring tectonic units, represents a natural laboratory that allows geoscientists to study simultaneously not only the continental collision, but also its extension.

Therefore, the geology of Central Europe has always been the focus of geoscientists, and the results of geodynamic research have been published in many papers (e.g., Balla 1984; Royden & Horváth 1988; Royden & Burchfiel 1989; Ratschbacher et al. 1991a, b; Horváth 1993; Csontos 1995; Fodor et al. 1999; Pharaoh 1999; Tari et al. 1999; Kováč 2000; Plašienka 2002, 2018; Golonka 2004; Bezák et al. 2004; Schmid et al. 2004; Horváth et al. 2006, 2015; Bada et al. 2007; McCann 2008a, b; Vozár et al. 2010; Matenco & Radivojević 2012; Hók et al. 2014; Hetényi et al. 2015; Balázs et al. 2017).

Geophysical research has primarily been based on the seismic deep reflection and refraction (e.g., Beránek & Zátopek 1981; Tomek et al. 1987, 1989; Posgay et al. 1995, 2006; Vozár & Šantavý 1999; Hrubcová et al. 2005, 2010; Grad et al. 2006; Środa et al. 2006; Janik et al. 2009, 2011; Malinowski et al. 2009; Brückl et al. 2010; Brixová et al. 2018a, b), seismological (e.g., Babuška et al. 1987; Babuška & Plomerová 2006; Plomerová & Babuška 2010), gravimetric (e.g., Tomek et al. 1979; Bielik 1988a, b; Lillie et al. 1994; Szafián et al. 1997; Bielik et al. 2006; Kiss 2006; Królikowski 2006; Szafián & Horváth 2006; Sumanovac 2010; Grabowska et al. 2011; Alasonati Tašárová et al. 2016; Pánisová et al. 2018; Šamajová et al. 2019; Godová et al. 2021; Zahorec et al. 2021), geothermal (e.g., Čermák & Hurtig 1979; Majcin et al. 1998; Lenkey 1999; Majorowicz et al. 2019) and magnetotelluric measurements (e.g., Jankowski et al. 1977; Varga & Lada 1988; Praus et al. 1990; Ádám et al. 2008; Majcin et al. 2018; Bezák et al. 2020; Vozár et al. 2021), as well as their interpretations.

Despite the great efforts of geologists and geophysicists, many questions concerning the deep structure, composition, and tectonics of the lithosphere in the Carpathian–Pannonian region and its surrounding geological units remain unanswered. Therefore, the main goal and mission of the paper is not only to highlight, but also attempt to find solutions to some of these problems in the light of the current geoscientific knowledge. The paper deals first with pointing out the most prominent dominant gravity anomalies in the Carpathian–Pannonian region, and then tries to find the optimal answer as to the sources of these anomalies. For this purpose, the stripped gravity map was calculated. Finally, the paper addresses the problem of determining the important boundary between the lithosphere and asthenosphere (LAB) and its implications for Neo-Alpine evolution of the study area.

Tectonic evolution of Central Europe

The complex tectonic evolution of Central Europe includes three geological periods of orogenic processes linked with folding during the Caledonian, Hercynian (Variscan), and Alpine orogenesis.

The European Platform was formed during the Precambrian and consists of the Precambrian Eastern European Craton in the NE and a younger Western European Paleozoic Platform in the NW (Fig. 1). The Eastern European Craton is formed by Proterozoic igneous and metamorphic rocks covered with the Vendian and Paleozoic strata (Dadlez et al. 2005). Both tectonic units are separated by the TESZ (e.g., Pharaoh 1999), which is ~200 km wide, and runs through Europe from the North Sea to the Black Sea. The north-eastern boundary of this zone, which is located in Poland, is formed by the Teisseyre–Tornquist zone (Dadlez et al. 2005). It consists of several interesting terranes accreted to the southeastern border of the Eastern European Craton during the Paleozoic (Winchester et al. 2002).

The Bohemian Massif represents the easternmost termination of the Paleozoic Variscan orogenic belt in Central Europe and is a complicated terrane that was consolidated during the Paleozoic. Its current structure is the result of the convergence and collision of the Laurentia, Baltika, Avalonia, and Gondwana continents after the closure of various ocean basins, followed by nappe thrusting and continental collision with strike-slip movements that took place between 500 Ma and 250 Ma (Matte et al. 1990; Dallmeyer et al. 1994; Schulmann et al. 2009; Guy et al. 2011). The Bohemian Massif consists mainly of low- to high-grade metamorphic and plutonic Paleozoic rocks exposed on the surface. Based on the respective effects of Cadomian and Variscan orogeneses, the Bohemian Massif area (Dallmeyer et al. 1994; Hrubcová et al. 2005) is subdivided into several regional tectonostratigraphic units, which are separated by faults, shear zones, or thrusts.

In the Eastern Alps, several tectonic units can be recognised: the Molasse Basin, the Flysch Zone, the Northern Calcareous Alps, and the Central Eastern Alps, which are distinguished by their age, rock composition, and their lithostratigraphic affiliation (Alasonati Tašárová et al. 2009).

The Carpathian–Pannonian region consists of the Carpathian orogen and the Pannonian back-arc basin system. The geological picture is the result of Neogene evolution, at the beginning of which the Inner Carpathian region consisted of two independently-moving microplates known as the ALCAPA (Alps–Carpathians–Pannonian Basin) and Tisza–Dacia megatectonic units (Fig. 1). The structural and paleogeographic development contains the elements reflecting the collision of the orogen with the platform, as well as the consequences of stretching the overthrusted plates, accompanied by rifting, mantle upwelling, and Neogene volcanism.

The tectonic evolution of the Carpathian–Pannonian region in the present-day of its knowledge still offers us a lot of space for discussion (Csontos 1995; Alasonati Tašárová et al. 2009, 2016; Janík et al. 2011). According to Alasonati Tašárová et al. (2016), one group interprets the evolution of the Carpathian–Pannonian Basin region in terms of gravitational collapse of the continental lithosphere (Alasonati Tašárová et al. 2016). This interpretation excludes the existence of subduction and favours active continental lithospheric delamination under the Carpathians (e.g., Knapp et al. 2005; Gemmer & Houseman

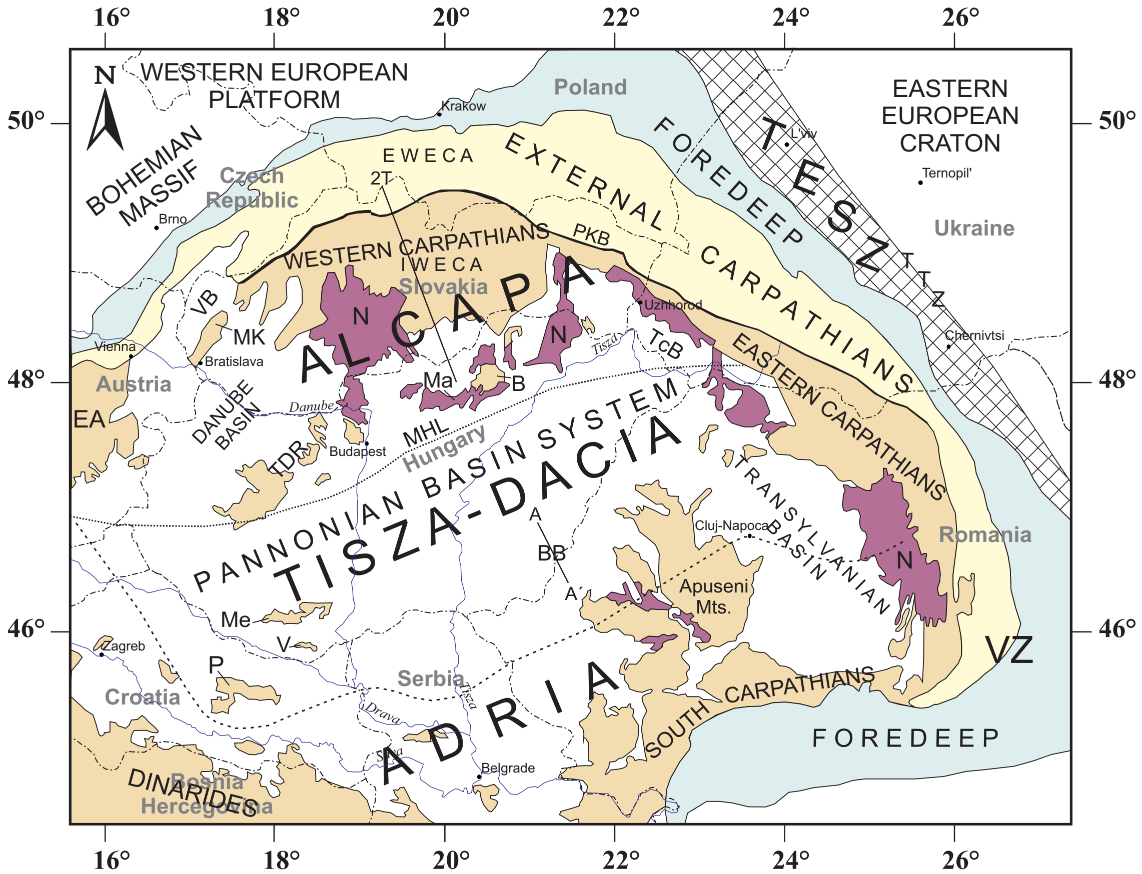


Fig. 1. Tectonic sketch of the Carpathian–Pannonian region (modified from Lillie et al. 1994; Kováč 2000; Geological map of Slovakia 2013; Hók et al. 2014). Locations of the profiles 2T and A–A´. Abbreviations: EWECA=External Western Carpathians, IWECA=Internal Western Carpathians, PKB=Pieniny Klippen Belt, TESZ=Trans European Suture Zone, TTZ=Tornquist–Teisseyre zone, EA=Eastern Alps, VB=Vienna Basin, MK=Malé Karpaty Mts., MHL=Middle Hungarian Line, Ma=Mátra Mts., B=Bükk Mts., TcB=Transcarpathian Basin, TDR=Transdanubian Range, N=Neovolcanics, Me=Mecsek Mts., V=Villany Mts., P=Papuk Mts., BB=Békés Basin.

2007; Göğüş et al. 2016; Bracco Gartner et al. 2020). The second group, however, includes the subduction and associated sub-lithospheric mantle uplift as a key process in the tectonic development of the Carpathian–Pannonian region (e.g., Ratschbacher et al. 1991a, b; Csontos et al. 1992; Horváth 1993; Tomek & Hall 1993; Linzer 1996; Kováč et al. 1998; Kováč 2000; Konečný et al. 2002).

Geology of the Western Carpathians and the Pannonian Basin System

The recent structure of the Western Carpathians contains several different allochthonous tectonic units moved during the two phases of Alpine orogeny (Hók et al. 2016). The Paleo-Alpine phase is characterised by the subduction, collision, and stacking of groups of nappes in the Internal Western Carpathians (IWECA), accompanied by the extention of oceanic realms in the External Western Carpathians (EWECA) during the Cretaceous (Plašienka 1995, 1999; Hók et al. 2014). The Neo-Alpine phase is characterised by oblique diachronous subduction of the EWECA basement along the periphery of the IWECA (Kováč 2000; Hók et al. 2014). In the collision zone, the rootless Flysch Belt nappes were thrusted onto the European platform margin. During this phase, basin formation and back-arc type volcanism were active in the IWECA. Oblique collision of the IWECA with the European platform, in combination with the rollback of the European slab, caused a large, counterclockwise rotation of the IWECA during the Neogene (e.g., Márton & Fodor 2003). Simultaneously, lateral extrusion of the Carpathians from the Eastern Alpine area (Ratschbacher et al. 1991a, b), as well as escape of the Transdanubic and Bükkic terranes from the Southern Alpine and Dinaride realms and their accretion to the IWECA occurred (Haas et al. 2001; Márton & Fodor 2003).

The Western Carpathians were divided into two main parts: the Externides and the Internides (Mišík et al. 1985; Froitzheim et al. 2008; Hók et al. 2014), which are referred to by their acronyms, EWECA (External Western Carpathians) and IWECA (Internal Western Carpathians). These two main parts are separated by the Pieniny Klippen Belt (PKB), which is a narrow zone with a complex Paleo-Alpine structure that has been predominantly affected by the younger Neogene deformation (Hók et al. 2014; Plašienka et al. 2020). The EWECA lie in the north with a dominant Tertiary deformation. The IWECA are in the south, where the Mesozoic (Cretaceous) deformation dominated. The IWECA consist of Paleo-Alpine (Cretaceous) crustal units (Tatricum, Veporicum, Gemericum) and a cover in the form of Mesozoic nappe units (Fatricum, Hronicum, Meliaticum, Turnaicum, Silicicum). Sedimentary basins with Upper Cretaceous, Paleogene, and Neogene infill and neovolcanic complexes represent a Neo-Alpine formation superimposed on the Paleo-Alpine nappe system (e.g., Kováč et al. 2016, 2017). The overthrusting of the IWECA was completed before the Cenozoic Era (approximately 65 Ma), while the EWECA were folded during the Cenozoic (30–12 Ma). The morphological and tectonic locations of the Western Carpathians were mostly influenced by Tertiary (Neo-Alpine) tectonics. The Tertiary accretionary prism of the EWECA is a common, and at the same time, unifying element of the entire Alpine–Carpathian Mountain range (Tomek et al. 1979). It consists of several nappe systems that have been overthrust onto the European platform. The final process of the accretionary prism formation was connected to the flexure of the platform margin onto which the Foredeep was developed (Kováč 2000; Alasonati Tašárová et al. 2009). The Western Carpathian Foredeep is mostly filled with marine sediments of the Middle Miocene (Oszczypko 1998).

The Pannonian Basin System formed as a typical continental back-arc basin, in which there was a roughly east-west oriented ~220–290 km Miocene extension that was accompanied by rollback flexure of the Carpathian or Dinaric lithospheric plates (e.g., Matenco & Radivojević 2012; Horváth et al. 2015) and mantle upwelling behind the Carpathian arc (Csontos et al. 1992; Horváth 1993; Royden 1993a, b; Kováč 2000). The general extensional geometry of the basin is characterised by individual sub-basins filled by ~1–3.5 km of lower to lowermost Late Miocene syn-kinematic deposits and overlain by a 1.5–3.5 km thick post-extensional sedimentary cover (Balázs et al. 2016). The thickness of the sedimentary infill ranges from 0 to 7 km (e.g., Bielik 1988a; Kilényi & Šefara 1989; Csato et al. 2007; Alasonati Tašárová et al. 2009). The formation of the extension basin, as well as the subsequent post-rift thermal development, has been accompanied by intense volcanism (e.g., Lexa et al. 1993; Konečný et al. 2002; Harangi & Lenkey 2007).

Complete Bouguer anomaly map

The complete Bouguer anomaly (CBA) map is the essential gravity map for gravimetric or integrated geophysical studies of the structure and composition of the lithosphere. The CBA data are topographically-corrected gravity anomalies (Vajda et al. 2020), i.e., gravity anomalies corrected for the gravitational effect of topographic masses, as well as for the effect of the (negative) density contrast of seawater (e.g., Tenzer et al. 2009; Pašteka et al. 2017; and references therein). The subsurface structural positive density contrasts produce gravity highs (positive anomalies) in the CBA map, while negative density contrast structures produce gravity lows (negative anomalies). All 2D linear geological structures (such as faults and lineaments), if accompanied by their respective density contrasts, are manifested by linear horizontal gravity gradients.

Since the observed values of the CBA represent the sum (superposition) of gravity effects of all density inhomogeneities (anomalous structural sources) located within the lithosphere, the interpretation of individual gravity anomalies (their decomposition) is often very difficult. The complete Bouguer anomaly map of the Carpathian–Pannonian region (Fig. 2a) shows several positive and negative anomalies of various amplitudes, sizes, and origin.

The most prominent regional features of the CBA are the Carpathian gravity low (CGL) and the Pannonian gravity high (PGH), which are indicated in Fig. 2a. The CGL is a continuation of the Alpine gravity low, and it is characterized by gravity values ranging from −30 to −110 mGal (1 mGal $=10^{-5}\,\mathrm{m\,s^{-2}}$). It is divided into the Western Carpathian gravity low (WCGL), Eastern Carpathian gravity low (ECGL), and Southern Carpathian gravity low (SCGL) (Fig. 2b). These expressive gravity lows correlate with the Western, Eastern, and Southern Carpathians. The Eastern and Southern Carpathians are accompanied by the highest (−110 mGal), while the Western Carpathians by the lowest (−60 mGal) amplitude. A second prominent feature is represented by the Pannonian gravity high (PGH, Fig. 2a), which extends over the Pannonian Basin System. It is characterised by the gravity values that vary at a relatively narrow interval from −10 to +20 mGal. It is worth noting that there is a well-established anti-correlation between the Carpathian gravity low and the Pannonian gravity high, including the topography. The high Carpathian topography is characterised in the complete Bouguer anomaly map by a regional gravity low, while the low Pannonian topography by an overall regional gravity high.

In general, orogenic belts are typically represented in the complete Bouguer anomaly map by belts of negative anomaly (gravity lows). This is caused by the fact that the roots of orogens isostatically compensate the topographic masses.

Several local gravity highs can also be observed within the regional Pannonian gravity high (Fig. 2b): the Transdanubian (TdGH), the Mecsek (MeGH), the Papuk (PaGH), the Transcarpathian (TcGH), and the Békés (BGH). The TdGH, the MeGH, and the PaGH are caused by the pre-Cenozoic rocks that build the Transdanubian Range, the Mecsek Mts., and the Papuk Mts. The reason for the Transcarpathian and Békés gravity highs is different, which we will explain below.

Even though considerable attention had been paid to the interpretation of the sources of the major gravity anomalies in the Carpathian–Pannonian region (e.g., Ibrmajer 1981; Steinhauser et al. 1990; Lillie et al. 1994; Królikowski & Petecki 2001; Švancara 2004; Bielik et al. 2006), their interpretation remains problematic. In order to be able to better answer the questions regarding the sources of the dominant gravity features in the studied area, the so-called stripped gravity map was calculated.

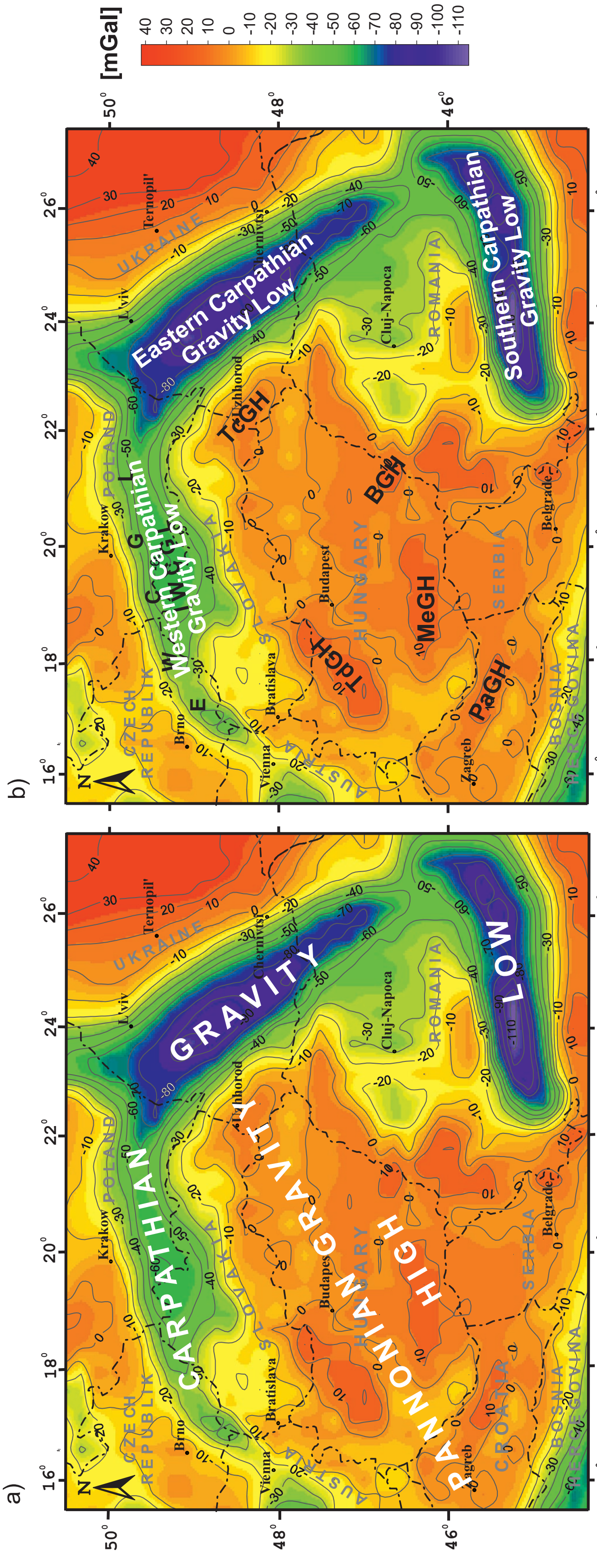
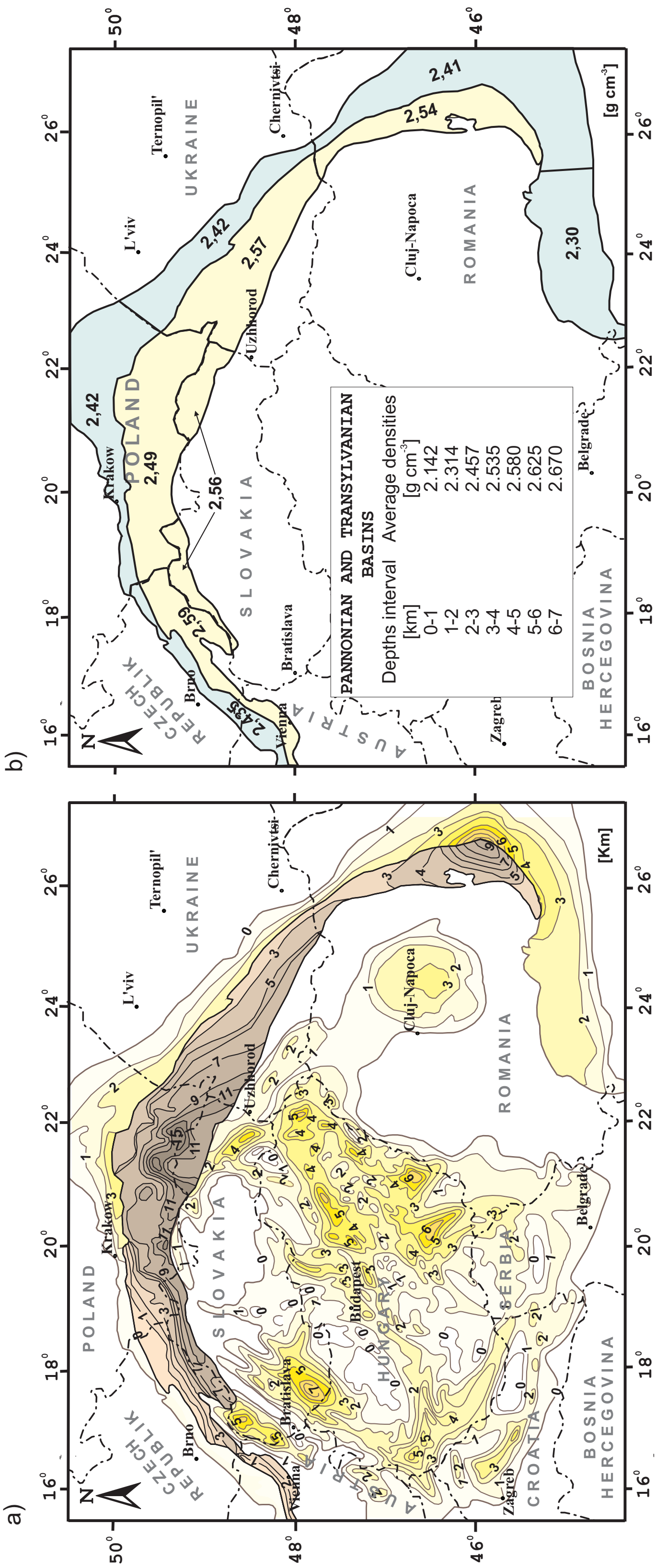


Fig. 2. Complete Bouguer anomaly (CBA). Compilation based on gravity data published by Ibrmajer (1963), Švancara (2004), Bielik et al. (2006), Švancara et al. (2021). **a** — The most prominent regional gravity anomalies of the CBA. **b** — The most prominent local gravity anomalies of the CBA. Abbreviations: EWCGL = External Western Carpathian gravity low, IWCGL = Internal Western Carpathian gravity low, TcGH = Transcarpathian gravity high, TdGH = Transdanubian gravity high, PaGH = Papuk gravity high, MeGH = Mecsek gravity high, BGH = Békés gravity high.

Stripped gravity map

The gravity-stripping procedure has proved to be useful in revealing deeper inhomogeneities or morphology of deeper density interfaces in the lithospheric structure that might be masked in the CBA map by shallower structures with relatively stronger gravity signal (Hammer 1963; Bielik 1988a; Bielik et al. 2013). Stripping is applied to crustal or lithospheric structures that are already known from previous or independent geophysical studies (e.g., Vajda et al. 2008; Tenzer et al. 2009). Stripping is based on forward computation of the gravity effect of the known subsurface 3D structure (such as sedimentary fill) and the removal of this effect from the CBA (Bielik et al. 2013). Stripping requires that the 3D geometry of the structure and its density contrast are known. The density contrast is taken relative to the density of the surrounding rock environment (in the case of sediments relative to the topographic masses or to the upper crust). The resultant stripped gravity map (SGM), such as the sediments-stripped CBA, is calculated by subtracting the gravity effect of the sediments (of their negative density contrast) from the CBA. The benefit of interpreting the SGM as opposed to the CBA map is due to the inhomogeneities deeper beneath the sedimentary basement that were masked by the effects of sediments in the CBA map and are now uncovered in the SGM.

To calculate the SGM in the Carpathian–Pannonian region, we had to compile both the thickness (Fig. 3a) and the density (Fig. 3b) models of the sediments which cover the Pannonian Basin System and the Transylvanian Basin, the External Carpathians, and the Foredeep. The thickness of the Neogene-Quaternal sediments (Fig. 3a) varies between 0–7 km in the Pannonian Basin System, 0–3 km in the Transylvanian Basin (Bielik 1988a; Kilényi & Šefara 1989; Makarenko et al. 2002; Bielik et al. 2004, 2005), and 0–9 km in the Foredeep (Poprawa & Nemčok 1989; Matenco 1997; Kováč 2000; Bielik et al. 2004, 2005 and the references therein). The largest 15 km thickness of the Paleogene sediments (Mocanu & Radulescu 1994; Krejčí & Jurová 1997; Kováč 2000; Bielik et al. 2005; Rylko & Tomaš 2005), which forms the accretionary prism of the External Carpathians, can



PANNONIAN AND TRANSYLVANIAN BASINS

Depths interval [km]	Average densities [g cm^{-3}]
0-1	2.142
1-2	2.314
2-3	2.457
3-4	2.535
4-5	2.580
5-6	2.625
6-7	2.670

Fig. 3. a — Thickness of the Carpathian–Pannonian sediments in km; **b** — Density model of the Carpathian–Pannonian sediments in g cm^{-3} (modified from Bielik et al. 2004, 2005).

be observed in the eastern part of the EWECA, while in the seismic Vrancea zone it is about 9 km (Fig. 3a).

The densities of the Pannonian Basin System and Transylvanian Basin sediments (e.g., Khomenko 1971; Granser 1987; Bucha & Blížkovský 1994; Szafián et al. 1997; Šefara & Szabó 1997; Makarenko et al. 2002; Bielik et al. 2005 and the references therein) change with depth (Fig. 3b). Due to less information on density data (e.g., Sovchik 1976; Tomek et al. 1979; Ibrmajer & Suk 1992; Królikowski & Petecki 2001; Bielik et al. 2004, 2005 and the references therein), the External Carpathian and Foredeep density model is simpler. The densities here are considered constant with depth and vary only laterally (Fig. 3b).

The gravity effect of the 3D model of the Carpathian–Pannonian sediments (Fig. 4) was calculated by means of the algorithm, which was developed by Starostenko et al. (1997). The gravity effect values vary from −0 to −80 mGal. The highest values of −80 mGal are observed in the eastern part of the External Western Carpathians, and a little less (−60 mGal) in the Vrancea zone. In the Pannonian Basin System and the Transylvanian Basin, the largest amplitudes of the gravity effect are around −45 to −50 mGal. The resulting stripped gravity map is presented in Figure 5.

At first glance, the stripped gravity map appears similar to the complete Bouguer anomaly map. However, the differences are significant; not only in the size of the amplitudes, but also in the characteristics of the gravity anomalies. On the SGM, the Pannonian gravity high is characterised by much more pronounced positive values, which reach up to +70 mGal. On the other hand, the maximum negative values of the Carpathian gravity low on the SGM are lower, although still accompanied by high negative amplitudes. They are −50 mGal in the Western Carpathians, −80 mGal in the Eastern Carpathians, and −100 mGal in the Southern Carpathians. These differences themselves enabled us to improve the interpretation of the sources of the most dominant and prominent gravity features (anomalies) in the Carpathian–Pannonian region.

Western Carpathian gravity low

Tomek et al. (1979) were the first authors to realize that the Carpathian gravity low (CGL, shown on map Fig. 2a) represents one of the most significant gravity anomalies in the Carpathian Mts. In their paper, they also interpreted the depth of the source of the Western Carpathian gravity low (WCGL). Their conclusion was that the source of the WCGL is unusually shallow (maximum lower boundary of this source reaches a depth of 8.5 km) represented by the low-density sediments of the External Western Carpathians (EWECA) and the Foredeep. However, this interpretation was soon challenged by Pospíšil & Filo (1980). Based on the stripped gravity map features, the Western Carpathian gravity low must be divided into two gravity sub-lows: the External Western Carpathian gravity low (EWCGL) and the Internal Western Carpathian gravity low (IWCGL), since their sources are completely different. On the SGM, we can see that the External Western Carpathian gravity low has practically disappeared. Based on this and the results of our gravity modelling (Fig. 6, Bielik 1995), we agree with the opinion of Tomek et al. (1979) that the source of the External Western Carpathian gravity low is the low-density sediments of the External Western Carpathians and the Western Carpathian Foredeep. However, it is impossible to apply a source of the same type to explain the existence of the Internal Western Carpathian gravity low. Firstly, in this area, there are almost no low-density sediments on the surface. This zone of the Internal Western Carpathians is built mostly by the pre-Cenozoic crystalline basement rocks in the Tatric, Veporic, and Gemeric tectonic units. The results of our gravity modelling (Fig. 6, Bielik 1995) clearly show that the source of the Internal Western Carpathian gravity low is the Internal Western Carpathian upper crust with a density contrast of −0.20 g cm^{-3} (1 g cm^{-3}=1000 kg m^{-3}). This upper crust is built largely by granites and crystalline schists (see Geological map of Slovakia, Vozár et al. 2021). The average density of these rocks (~2.70 g cm^{-3}) represents a significant deficit mass against the average high-density (~2.90 g cm^{-3}) of the Internal Western Carpathian lower crust (e.g., Šimonová & Bielik 2016; Šimonová et al. 2019). This means that the Internal Western Carpathian gravity low can be explained by the upper crustal deficit mass, which is predominantly formed by the rocks of the Alpine Tatric and Veporic units.

Eastern and Southern Carpathian gravity lows

The Eastern and Southern Carpathians in the stripped gravity map (Fig. 5) are still characterised by significant gravity lows, even after correction of the complete Bouguer anomaly for the (negative) gravity effect of the sediments. Therefore, the sources of the Eastern and Southern Carpathian gravity lows (ECGL and SCGL) cannot be due to just the low-density sediments of the External Carpathians and the Foredeep. And so, it is necessary to look for additional sources to explain the cause of both the gravity lows. When taking into consideration the calculations of the gravity models (e.g., Dérerová et al. 2006, 2021; Grinč et al. 2013) and the crustal thickness (Fig. 7, Bielik et al. 2018 and references therein), we can state

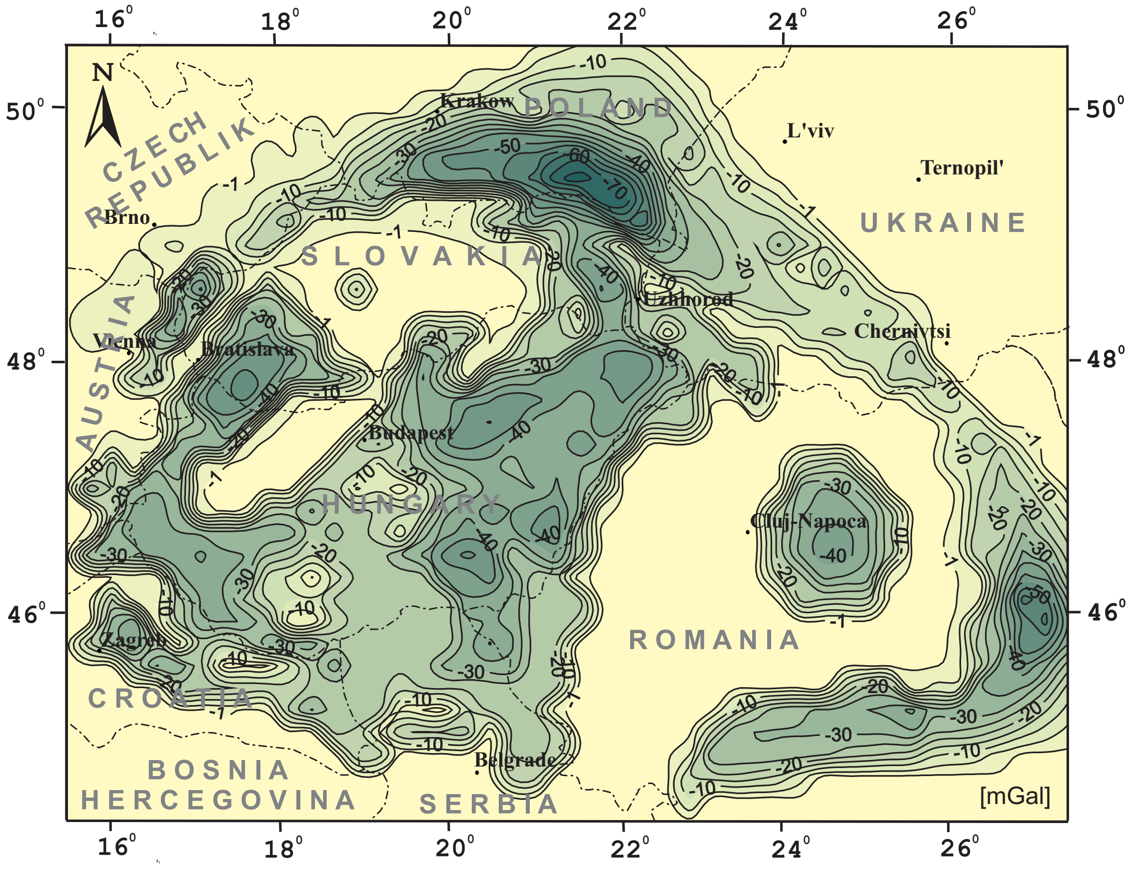


Fig. 4. 3D gravity effect of the Carpathian–Pannonian sediments. Abbreviation: VZ=Vrancea zone.

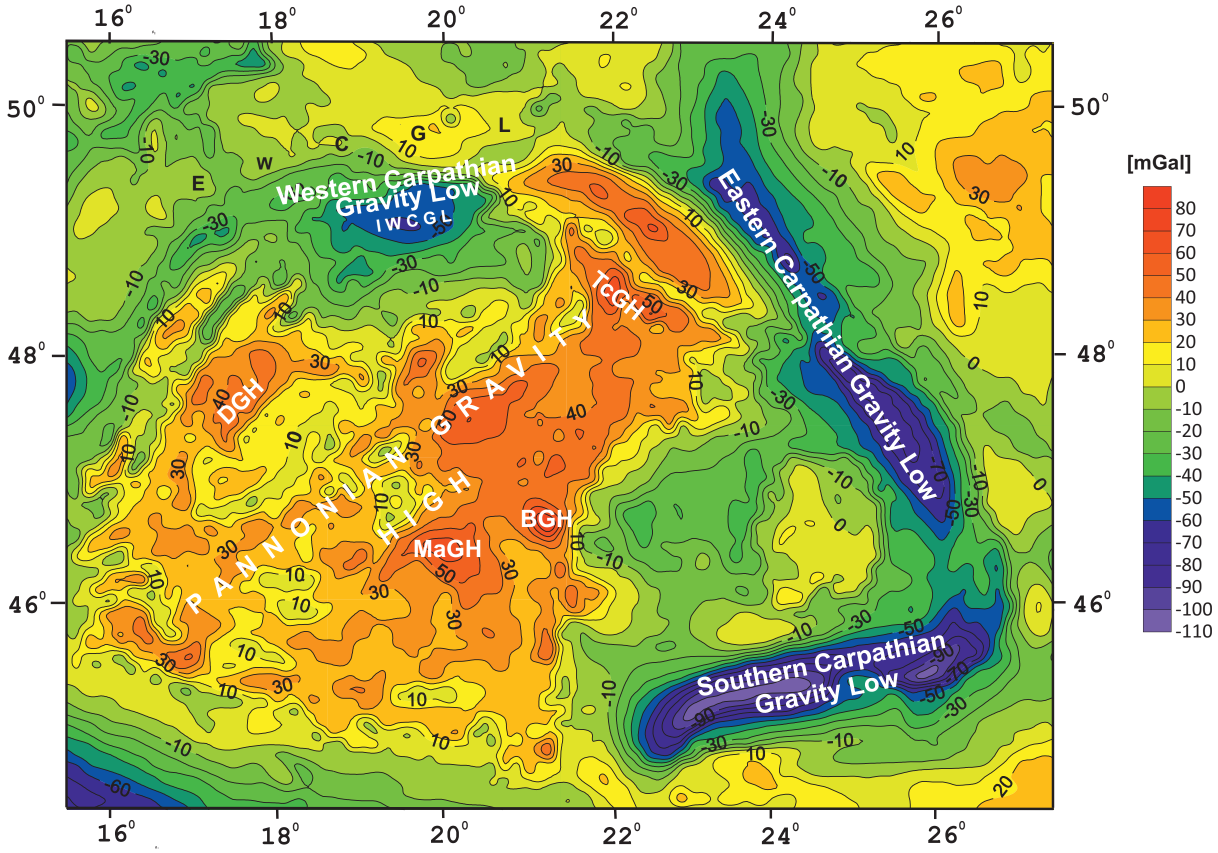


Fig. 5. Stripped gravity map of the Carpathian–Pannonian region. Abbreviations: EWCGL=External Western Carpathian gravity low, IWCGL=Internal Western Carpathian gravity low, TcGH=Transcarpathian gravity high, DGH=Danube gravity high, MaGH=Makó gravity high, BGH=Békés gravity high.

that the main source of the Eastern and Southern Carpathian gravity lows are crustal roots that were formed by continental collision below the external parts of the Eastern and Southern Carpathians. It follows that the observed amplitude of the ECGL and the SCGL (Figs. 2 and 5) is the sum of the negative gravity effects of both surface sediments and thick crust (crustal root). Both surface and deep anomalous zones represent a deficit mass. The gravity effect of the Cenozoic surface sediments of the External Eastern Carpathians and the Foredeep is negative, since their average densities are less in comparison with the average density of the upper crust of the Eastern and Southern Carpathians. Similarly, the densities of the Eastern and Southern Carpathian crustal roots are less than average density of the upper mantle beneath these orogens, and therefore their gravity effects are also negative. In terms of percentage, the observed amplitudes of the Eastern and Southern Carpathian gravity lows account for 70 % of the crustal roots and 30 % of the sedimentary gravity effects (Dérerová et al. 2006).

Pannonian Basin gravity high

The Pannonian Basin represents a significant regional gravity high, not only on the complete Bouguer anomaly map (Fig. 2b), but also on the stripped gravity map (Fig. 5). It may appear strange in the gravity field of the Pannonian Basin System because, even though it is covered with low-density sediments that reach a thickness of several kilometres, it is accompanied by positive instead of negative values of the gravity field. The positive gravity field can only be explained by the fact that under the sedimentary basement, there must be an anomalous body whose positive gravity effect is greater than the negative gravity effect of the sediments. This notion is strongly supported by the calculated SGM (Fig. 5), because the amplitude of the Pannonian gravity high (PGH) is significantly larger in comparison with the CBA (Figs. 2a, b). On the basis of many 2D density (e.g., Lillie et al. 1994; Szafián et al. 1997; Zeyen et al. 2002; Dérerová et al. 2006; Šimonová et al. 2019) and 3D density models (e.g., Alasonati Tašárová et al. 2009, 2016), the regional PGH is caused by the expressive Moho elevation (Fig. 7). Since the Pannonian Basin upper mantle is built by high-density peridotites or dunites and is located several kilometres closer to the surface, they represent a great excess mass (high-density anomalous bodies). The Moho (upper mantle) depth here is only 24–26 km (e.g., Grad et al. 2009; Bielik et al. 2018 and references therein). On the SGM, the sub-basins of the Pannonian Basin System, such as the Danube Basin, the Békés Basin, the Transcarpathian Basin, and the Makó Through, which have the thickest sedimentary infill, are accompanied by significant local gravity highs (values >+50 mGal). They correlate very well with the areas of major extension (Horváth & Royden 1981). The sources of these local gravity highs are high-density crustal bodies (Eo-Alpine metamorphic complexes) with the density contrast of +0.30 g cm^{-3}, which were modelled in the upper and lower crust beneath the Békés Basin (Fig. 8a) by Nemesi & Slomfai (1992), Ádám & Bielik (1998), Bielik & Ádám (2006); the Danube Basin by Bielik (1998), Prutkin et al. (2011, 2014); and the Transcarpathian Basin by Pospíšil (1980) and Bielik (1998). Based on these modelling results, Bielik & Ádám (2006) suggested the scheme of a narrow rift (sub-basin) model of continental extensional tectonics for the Pannonian Basin System (Fig. 8b). The apical parts of these bodies reach depths of only 7 to 12 km.

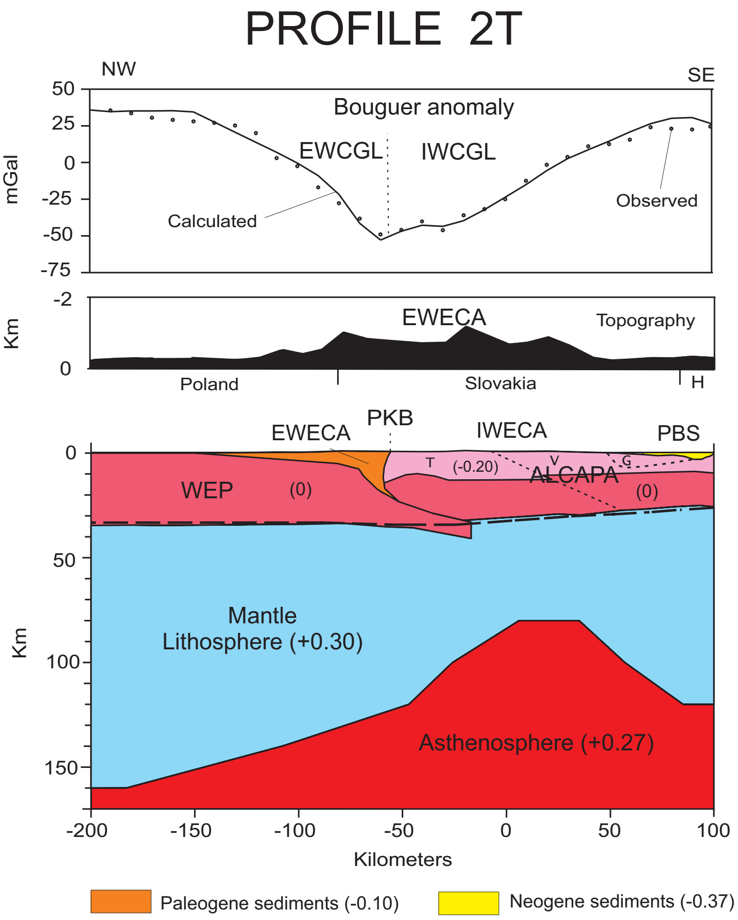


Fig. 6. Density model along the seismic 2T profile. The density contrast values are in g cm^{-3}. Abbreviations: WEP=Western European Platform, EWCGL=External Western Carpathian gravity low, IWCGL=Internal Western Carpathian gravity low, H=Hungary, EWECA=External Western Carpathians, IWECA=Internal Western Carpathians, PKB=Pieniny Klippen Belt, PBS=Pannonian Basin System.

2D and 3D integrated modelling

One of the goals of the 2D and 3D integrated geophysical modelling (Zeyen et al. 2002; Dérerová et al. 2006; Alasonati Tašárová et al. 2009, 2016; Grinč et al. 2013, 2014; Šimonová et al. 2019) was the determination of the lithosphere–asthenosphere boundary (LAB). It lies between a cooler, rigid lithosphere and the warmer, ductile asthenosphere and represents a mechanical interface between both of these layers. The actual depth of the LAB is still a topic of debate and study, although it is known to vary according to the environment.

The first LAB calculations in the Carpathian–Pannonian region were performed using seismological (Babuška et al. 1987, 1988), later magnetotelluric (Praus et al. 1990; Horváth 1993; Ádám 1996), and geothermic (Čermák 1993) data. However, the resulting lithosphere thicknesses determined by these geophysical fields disagree by as many as 50–60 km in some areas (e.g., in the South Slovak Basin).

To eliminate these discrepances, systematic 2D (Zeyen et al. 2002; Dérerová et al. 2006; Grinč et al. 2013; Šimonová et al. 2019) and 3D (Alasonati Tašárová et al. 2009, 2016) integrated geophysical modelling was applied to determine the LAB morphology more accurately. In the 2D solution, we applied the CAGES program (Zeyen & Fernandez 1994). This software is capable of a combined interpretation of heat flow, absolute topographic elevation, geoid, and gravity. For the 3D solution, we first used the IGMAS software (Interactive Gravity and Magnetic Application System) and then the LiTMod 3D. The IGMAS software (Götze & Lahmeyer 1988; Schmidt & Götze 1999) enables to model geological bodies, which are defined along several parallel vertical cross-sections (profiles). The software connects the profiles via triangulation, thereby generating a 3D structure. LitMod 3D (Afonso et al. 2013a, b) was developed to perform integrated geophysical–petrological LIThospheric forward MODelling of the lithosphere and the sub lithospheric mantle down to the top of the transition zone at a depth of 410 km (Fernàndez et al. 2010; Fullea et al. 2010; Grinč et al 2014; Alasonati Tašárová et al. 2016). All codes (CAGES, IGMAS and LitMod 3D) allow for the modelling of several geophysical data sets simultaneously. All the mentioned integrated approaches significantly reduce the uncertainties related to modelling the geophysical data sets individually (separately).

The lithosphere thickness in the Carpathian–Pannonian region (Fig. 9) shows important differences across the chain, as well as along the strike of the Carpathian arc. Lithosphere thickness varies from 240 km in the Eastern Carpathians to 75–110 km, under the Pannonian Basin System. Along the Carpathian Mts., the lithosphere thickness increases from the Western (~100 km) to the Eastern Carpathians (~240 km). The Southern Carpathians are characterised by a lithospheric thickness of ~180 km.

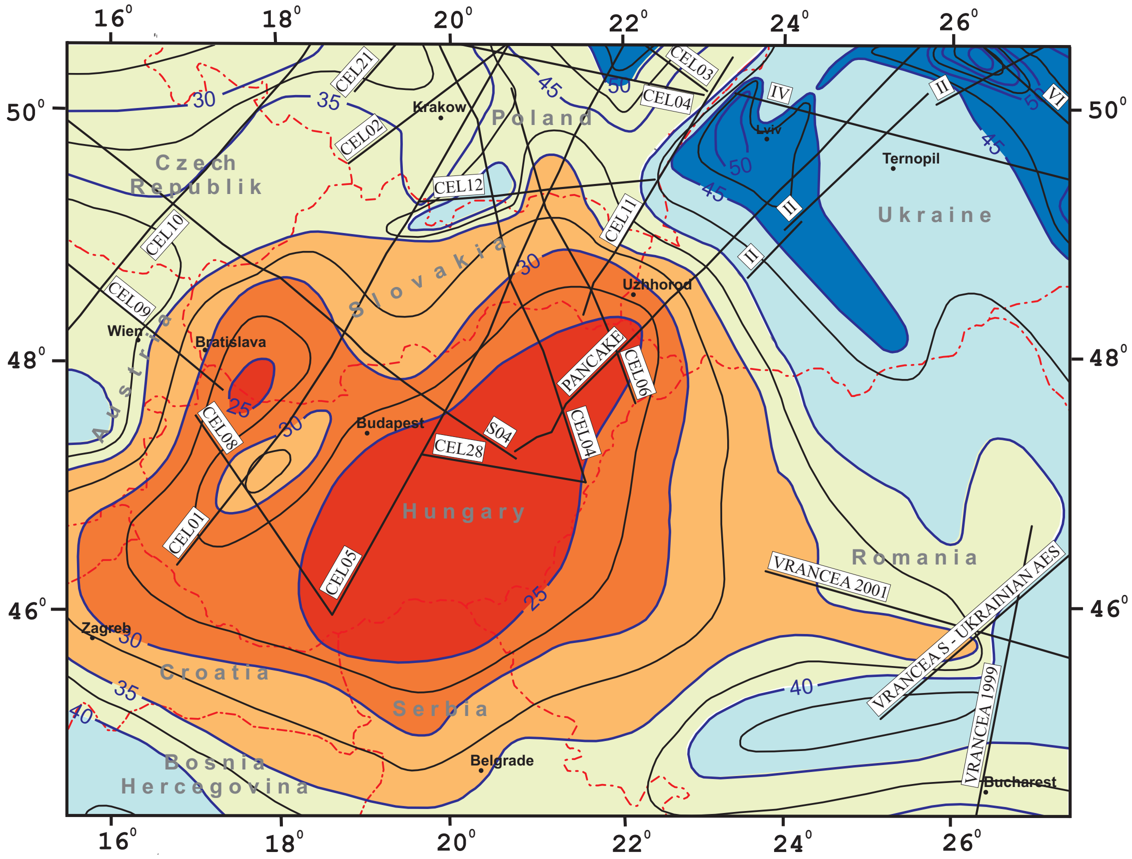


Fig. 7. The Moho depth map in the Carpathian–Pannonian region (modified from Bielik et al. 20018 and references therein).

The first most prominent feature of the lithosphere thickness map (Fig. 9) is the existence of the lithosphere root in the Eastern Carpathians, which reaches a depth of >240 km. Based on the results of seismic tomography (Wortel & Spakman 2000), we suggest that the lithospheric root represents the remnant of a slab detachment. At the beginning of the Neo-Alpine evolution of the Carpathian–Pannonian region, the subduction process had been accompanied by roll-back and slab detachment (Fig.10, Wong et al. 1997; Wortel & Spakman 2000). According to Wortel & Spakman (2000), the lower part of the broken slab gradually sank into the deeper mantle and formed a distinct horizontal P-waves velocity anomaly beneath the Pannonian Basin System. This anomaly represents the positive percentage deviation from average mantle velocities and extends to depths between 410 to 660 km. The seismic Pannonian anomaly is only part of the giant seismic anomaly that can be observed at these depths below the Mediterranean region (Wortel & Spakman 2000). Recent continental lithosphere root in the Eastern Carpathians is the result of a gradual sinking of the upper part of the broken slab to depths greater than 240 km by continuation of the frontal collision (convergence) in this region. According to Lillie et al. (1994) and Wortel & Spakman (2000), the seismic Vrancea zone (Bokelmann & Rodler 2014; Bala et al. 2021; Petrescu et al. 2021) could represent the last stage of the lateral migrating slab detachment process along the Carpathian orogen.

When comparing the lithosphere thickness in the Western and Eastern Carpathians, we can observe that the Western Carpathians are characterised by a much thinner lithosphere. The junction zone of the western part of the Western Carpathians and the Bohemian Massif has a lithosphere thickness of 100–120 km and, unlike the Eastern and Southern Carpathians, no lithospheric root is observed here. We assume that this fact can be explained by different Neo-Alpine evolution of this region. During the Tertiary collision, the ALCAPA microplate, which represents a wedge-shaped mega-unit, was strongly squeezed into its Eastern Alpine part between the Adriatic indenter and the southern spur of the Bohemian Massif. The Western Carpathian part of the ALCAPA was extruded from this collision towards the unconstrained eastern area, occupied by an extended lithosphere underlying complex of the present Carpathian Flysch Belt (e.g., Balla 1984; Ratschbacher et al. 1991a, b; Csontos et al. 1992; Kováč et al. 1994; Schmid et al. 2008). The ALCAPA lithosphere fragment moved towards the east along the left-lateral strike-slip Salzachtal–Ennstal–Mariazell–Puchberg fault zone in the Eastern Alps and along the similar Mur–Mürz–Leitha fault zone toward the northeast at the Alpine–Bohemian–Carpathian boundary (Ratschbacher et al. 1991a, b; Fodor 1995; Linzer 1996; Lankreijer et al. 1999). This means that the relative movement of the ALCAPA lithosphere fragment changed from E to NE because of lateral extrusion (oblique collision).

This scenario could explain the absence of lithosphere thickening in this transition zone, since the movement was mainly strike-slip along a deep reaching fault following the contact of the European platform (Bohemian Massif) and the microplate ALCAPA (the Western Carpathians).

The second most prominent feature of the lithosphere morphology is the very thin lithosphere beneath the Pannonian Basin System (only 75–90 km). The young and warm continental back-arc Pannonian Basin System is characterised by not only upper mantle (Fig. 7), but also asthenosphere

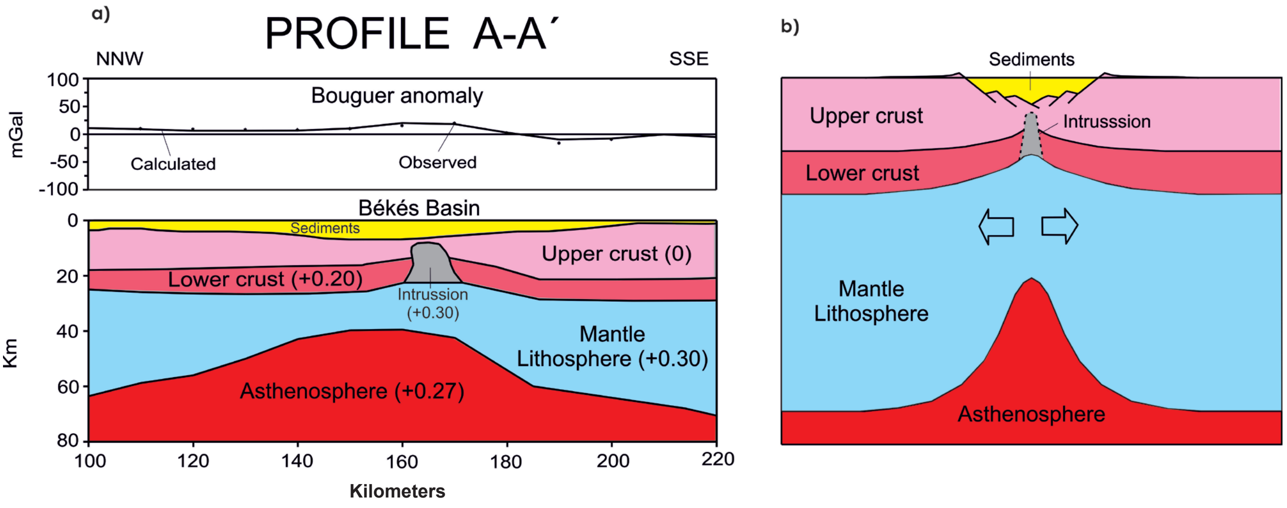


Fig. 8. a — Density model along the profile A–A´. The density contrast values are in g cm^{-3}. **b** — Scheme of a narrow rift model (sub-basin) of continental extensional tectonics for the Pannonian Basin System (modified from Bielik & Ádám 2006).

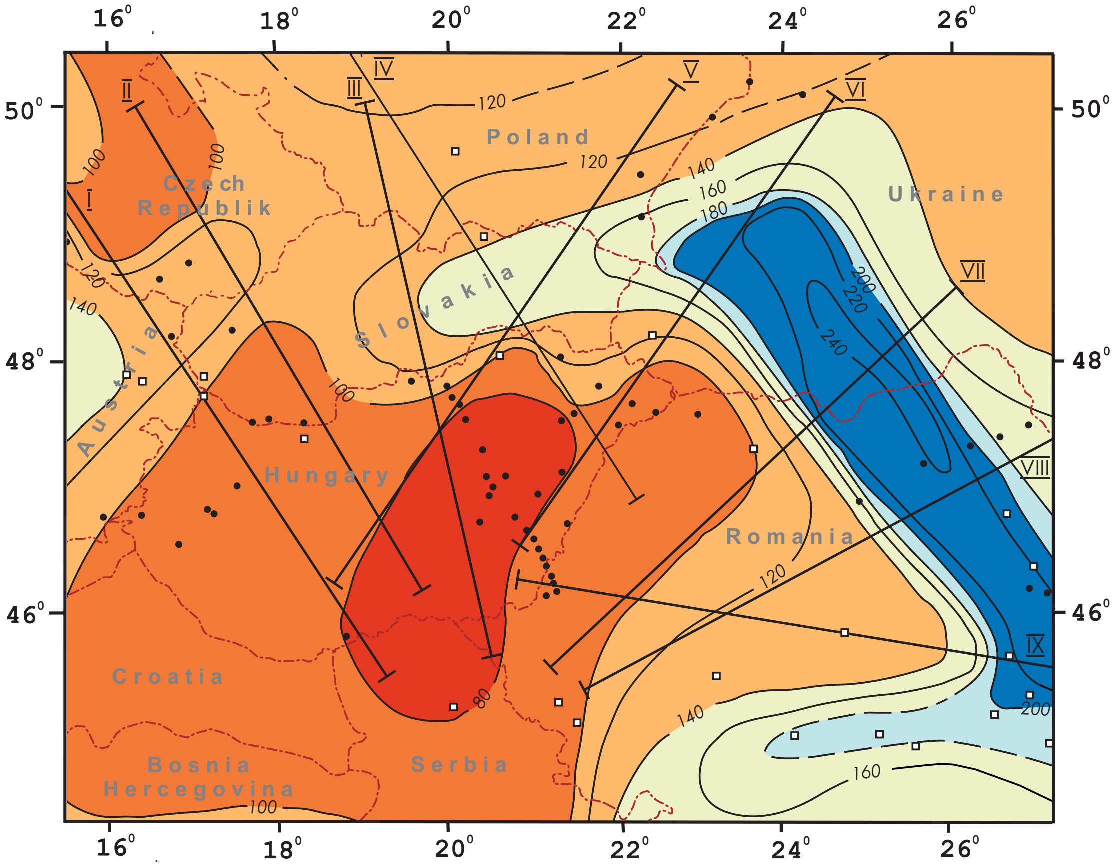


Fig. 9. Map of the lithosphere–asthenosphere boundary depth in the Carpathian–Pannonian region (modified from Dérerová et al. 2006).

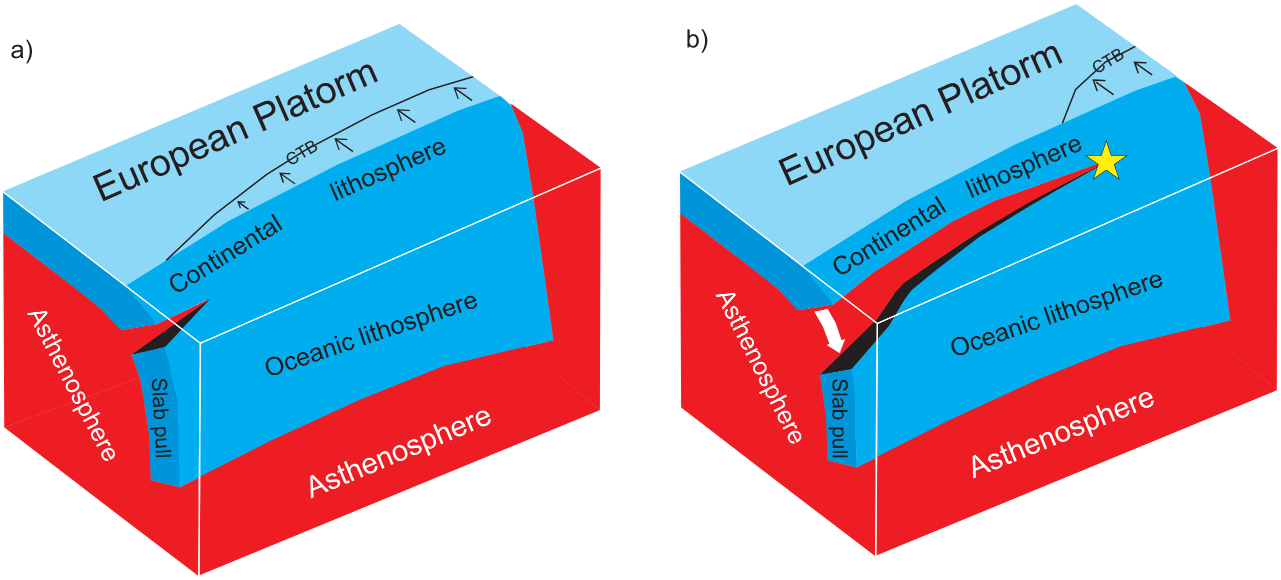


Fig. 10. Lateral migration of slab detachment (modified from Wortel & Spakman 2000): an initially small tear in the slab propagates approximately horizontally (**a**) and develops into a large tear (**b**). The yellow star indicates seismic activity in the stress concentration region. Abbreviation: CTB=Carpathian thrust boundary.

upwelling (Fig. 9). With its crustal (24–26 km) and lithospheric (75–90 km) thicknesses, the Pannonian Basin System belongs to the continental basin, which is characterised by one of the thinnest crusts and lithospheres worldwide. Based on geophysical features, it could be compared to the Basin and Range Province in North America (Lillie 2005).

Conclusion

More than fifty years of detailed geophysical and geological research in the Carpathian–Pannonian region and its surrounding tectonic units have resulted in sufficient knowledge of their subsurface geology, crustal and lithospheric structure, and mantle processes. Data and results from interpretation of gravity data and integrated geophysical modelling have been reviewed in this paper with the aim of offering an optimal quantitative interpretation of the most significant gravity anomalies, as well as the explanation of varying lithospheric thicknesses in the studied area. And so, the most important observations and their interpretation have led to the following main conclusions:

- The most prominent feature of the complete Bouguer anomaly map and the stripped gravity map is the Carpathian gravity low and the Pannonian gravity high. The Carpathian gravity low consists of the Western Carpathian gravity low, the Eastern Carpahian gravity low, and the Southern Carpathian gravity low, which correlate with the Western, Eastern, and Southern Carpathians.
- The Western Carpathian gravity low consists of two different gravity sub-lows: the External Western Carpathian gravity low and the Internal Western Carpathian gravity low. The source of the External Western Carpathian gravity low is low-density sediments of the External Western Carpathians and the Western Carpathian Foredeep. The source of the Internal Western Carpathian gravity low is the Internal Western Carpathian upper crust (~18 km thick), which is built mostly by low-density granites and crystalline schists of the Alpine Tatric and Veporic units
- The Eastern and Southern Carpathian gravity lows are caused by gravity effects of two different sources: the near-surface and the deep one. The near-surface source is represented by the low-density sediments, which belong to the External Carpathians and the Foredeep. The deep source is formed by the crustal roots, which are located below the external parts of the Eastern and Southern Carpathians. The gravity effect of the deep source is approximately twice the size of the near-surface source.
- The source of the regional Pannonian gravity high is the Moho elevation.
- The local Transdanubian gravity high, Papuk gravity high, and Mecsek gravity high observed on the complete Bouguer anomaly map reflect the pre-Cenozoic rocks that build the Transdanubian Range, the Papuk Mts. and the Mecsek Mts. On the other hand, the stripped gravity map discovered considerable local gravity sub-highs: the Danube gravity high, Makó gravity high, Békés gravity high, and Transcarpathian gravity high over the deepest sub-basins of the Pannonian Basin System. The sources of these local gravity sub-highs are high-density crustal bodies (Eo-Alpine metamorphic complexes) with density contrast of +0.30 g cm^{-3}, whose apical parts reach depths of only 7 to 12 km.
- The Eastern Carpathian lithosphere root is the result of a sinking upper part of the broken slab to a depth of >240 km during the ongoing convergence between the European platform and ALCAPA and Tisza–Dacia microplates,

which is characterised by frontal continental collision. It is assumed that the slab is formed by the continental lithosphere.

- The junction zone of the Western Carpathians and the Bohemian Massif wasn't accompanied by thickening of the mantle lithosphere, since the oblique continental collision dominated here. The collision probably took place along a deep reaching fault following the contact of the European platform (Bohemian Massif) and the microplate ALCAPA (the Western Carpathians).
- The Pannonian Basin system is characterised by one of the thinnest continental crusts (~25 km) and lithospheres (~75 km) in the world.

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