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Lithospheric density model along the CEL09 profile and its geological implications

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Abstract: We present a new 2D lithospheric density model along the seismic profile CEL09 crossing the Bohemian Massif, the Western Carpathians, and the Pannonian Basin. The resulting model consists of five principal layers: sediments, upper crust, lower crust, lower lithosphere, and asthenosphere. The thicknesses of the Neogene sedimentary basins vary from 0 to ~5.5 km while the Paleogene flysch sediments dip to a depth of ~6.5 km. The most complex upper part of the upper crust in the Bohemian Massif is represented mainly by low-density granitoid plutons (~2.60–2.68 g cm^{-3}), metamorphic rocks (~2.69–2.74 g cm^{-3}) and high-density basic and ultrabasic bodies (~2.78–2.79 g cm^{-3}). In the Western Carpathians, this layer is formed by the crystalline Malé Karpaty Mts. (2.66–2.67 g cm^{-3}), Trans-Danubian range (2.73–2.74 g cm^{-3}), and the pre-Cainozoic basement of the sedimentary basins (2.67–2.74 g cm^{-3}). The densities of the lower part of the upper crust range from 2.78 g cm^{-3} (in the Western Carpathian–Pannonian region) to 2.77–2.80 g cm^{-3} (in the Bohemian Massif). In the lower crust, four different sectors were modelled. In the Saxothuringian, they are divided into two layers, the upper layer (2.84–2.85 g cm^{-3}) and the lower layer (3.12 g cm^{-3}). The Moldanubian has the thickest lower crust (~20 km) with a density of 2.98 g cm^{-3}; the lower crust in the Moravo–Silesian has a density of 2.97 g cm^{-3}. The Western Carpathian–Pannonian region is represented by slightly lower densities of 2.94–2.96 g cm^{-3}. The gravity modelling indicates that the Western Carpathians were overthrusted by ~30 km onto the Bohemian Massif resulting in a neo-transformation of the crust/mantle and related lithosphere after subduction.

Keywords: complete Bouguer anomaly, 2D gravity modelling, CELEBRATION 2000 seismic profile CEL09, Western Carpathians, Bohemian Massif, Pannonian Basin.

Introduction

The study of the crust and lithosphere in the Bohemian Massif, Western Carpathians, and Pannonian Basin was always of great interest to geologists and geophysicists (e.g., Mayerová et al. 1985; Šefara et al. 1987; Ibrmajer et al. 1989; Matte et al. 1990; Stráník et al. 1993; Balla 1994; Chlupáč & Vrána 1994; Krejčí & Jurová 1997; Plašienka 1997; Kováč 2000; Matte 2001; Schulmann et al. 2009; Vozár et al. 2010; Kováč et al. 2016; Šamajová et al. 2018). Important geophysical knowledge has been obtained by interpretations of the reflection and refraction seismic (e.g., Tomek et al. 1979, 1987, 1989; Beránek & Zátopek 1981a,b; Posgay et al. 1981; Tomek & Hall 1993; Vozár et al. 1999), gravimetric (Bielik 1988; Lillie et al. 1994; Bielik et al. 2005; Šimonová et al. 2019), magnetometric (Mašín & Jeleň 1963; Filo & Kubeš 1994; Bucha & Blížkovský 1994; Kubeš et al. 2010), magnetotelluric (Červ et al. 1994; Bezák et al. 2014; Majcin et al. 2018), and geothermic (Čermák 1994; Majorowicz et al. 2003) measurements.

For the last twenty years, the results of the international seismic experiment CELEBRATION 2000 (**C**entral **E**uropean **L**ithospheric **E**xperiment **B**ased on **R**ef**r**action **2000**) (Guterch et al. 2003a,b) have significantly increased our knowledge of the continental lithosphere in Central Europe. Among them, the refraction and wide-angle reflection profile CEL09 (Hrubcová et al. 2015), see Fig. 1, played an important role, since it provided constraints for seismic velocities of the crust and upper mantle in the Bohemian Massif, Western Carpathians and Pannonian Basin and enabled more intensive studies of these tectonic units.

The Bohemian Massif represents the largest exposure of rocks deformed during the Variscan orogeny in Central Europe. The Western Carpathians form an arc-shaped mountain range related to the Alpine orogeny, and the Pannonian Basin is associated with Neogene back-arc extension. These geological units developed in different times and space; therefore, they offer an outstanding opportunity to study not only their different lithospheric structure, but also their current mutual tectonic interaction.

In our study, we concentrate on gravity modelling along the seismic profile CEL09 to provide additional constraints on composition, structure, and tectonics of the lithosphere in the Bohemian Massif, Western Carpathians, and Pannonian Basin. We accentuate modelling of the upper crustal structure

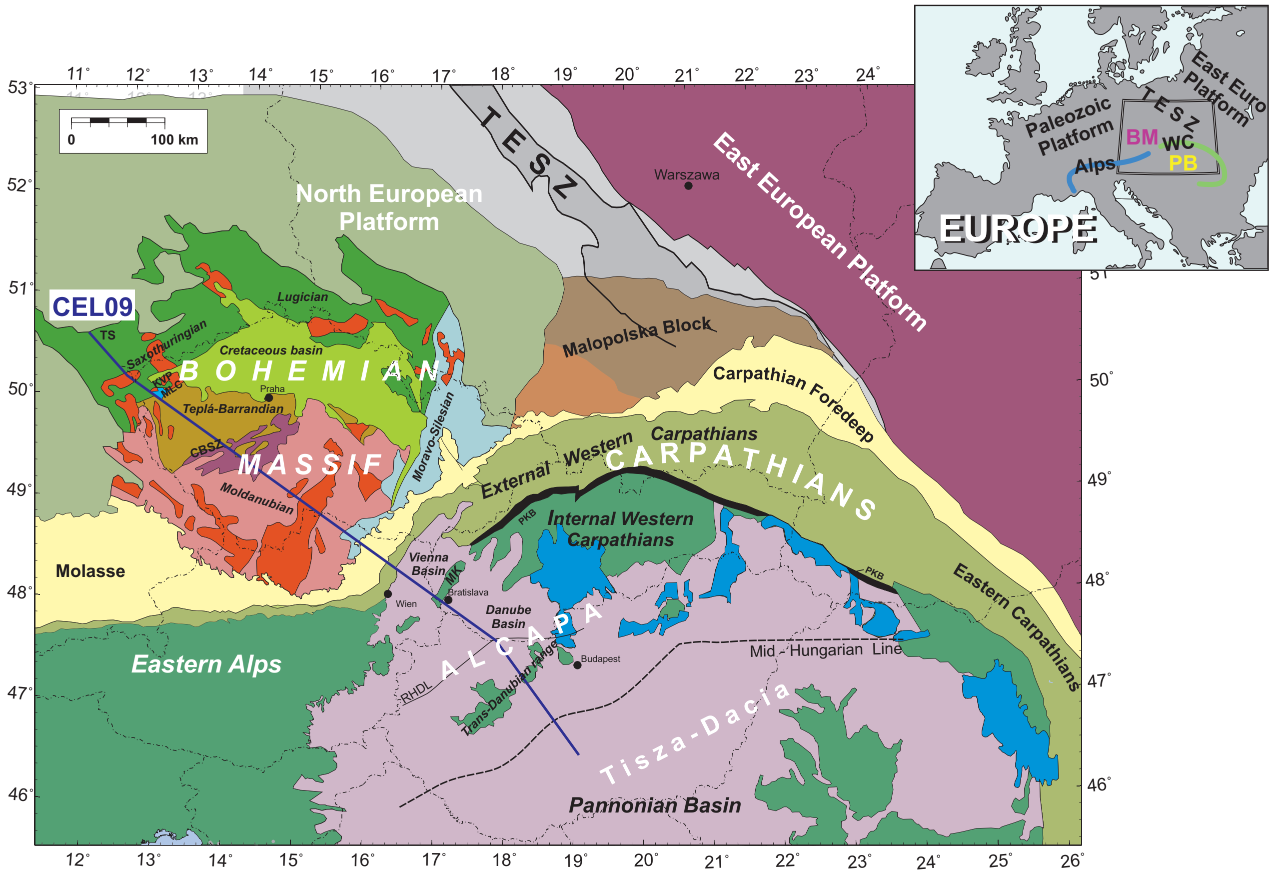

Fig. 1. Simplified tectonic map of the Bohemian Massif, Western Carpathians, Pannonian Basin, and their surrounding tectonic units with the localization of the CELEBRATION 2000 refraction seismic profile CEL09 (modified after Hrubcová et al. 2010; Hók et al. 2014; Grygar 2016; Šujan et al. 2021). TS – Teuschnitz syncline, KVP – Karlovy Vary Pluton, MLC – Mariánské Lázně amphibolite Complex, CBSZ – Central Bohemian Shear Zone, MK – Malé Karpaty Mts, PKB – Pieniny Klippen Belt, RHDL – Rába–Hurbanovo–Diósjenö lineament, TESZ – Trans European Suture Zone.

constrained by additional geological and geophysical knowledge not encompassed by velocity modelling. We also consider the gravity effect of the lithosphere–asthenosphere boundary (LAB). We model deep crustal density contacts between the key tectonic units and their tectonostratigraphic subunits. Since the profile CEL09 represents 2D velocity results, we restricted our gravity modelling to two dimensions.

Tectonic setting

Bohemian Massif

The Bohemian Massif represents the easternmost termination of the Paleozoic Variscan orogenic belt in Central Europe and is a complicated terrane consolidated during the Paleozoic. Its current structure is the result of convergence and collision of the Laurentia, Baltica, Avalonia, and Gondwana continents after the closure of various oceanic basins including nappe thrusting, and strike-slip faulting between 500 and 250 Ma (Matte et al. 1990; Dallmeyer et al. 1994). 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 the Cadomian and Variscan orogeneses, the area of the Bohemian Massif can be subdivided into four main regional tectonostratigraphic units from NW to SE (Dallmeyer et al. 1994): the Saxothuringian, the Teplá–Barrandian, the Moldanubian, and the Moravo–Silesian, which are all separated by faults, shear zones or thrusts (Fig. 1). In the south-east, the Moravo–Silesian crystalline rocks consist of a Cadomian basement known as the Brunovistulian. It is covered by Devonian to Carboniferous sediments and submerges to the east beneath the Carpathian Foredeep, where it forms the basement reactivated during the Alpine orogeny (Kalvoda et al. 2008).

Vienna Basin

The Vienna Basin represents a Neogene structure superimposed on rock sequences of the Bohemian Massif, the Eastern Alps and the Western Carpathians. The main subsidence period of the Vienna Basin was during the Miocene (Hölzel et al. 2008). The initial basin structure was a wedge-top zone (DeCelles & Giles 1996) on the frontal parts of the north-west propagating thrust sheets of the Eastern Alps (Early Miocene; ~18–16 Ma). Within this zone, mainly fluvial sediments filled several small basins that later merged into one large basin. The Vienna Basin was filled with clastic sediments up to 5.5 km thick (Middle to Late Miocene; ~16–7 Ma). The basin-floor section below the Neogene fill consists of the Alpine–Carpathian imbricated system. From the north to the south, the individual thrust piles are represented by the Waschberg–Zdanice zone, the Flysch zone, the Calcareous Alps (including its Paleozoic base, the Grauwacken zone), the Central Alps, and the Tatrides. All these units lie on top of the Miocene Molasse, a Mesozoic series, and the crystalline basement (Arzmüller et al. 2006).

Western Carpathians

The Western Carpathians represent the northernmost part of the Alpine orogenic belt in Europe. They form a northward-convex arc that was consolidated as a result of a series of Jurassic to Tertiary subduction and collision events during the Alpine orogeny (McCann 2008a, b). They can be divided into the External (Outer) and Internal (Inner) Western Carpathians (Fig. 1) separated by the Pieniny Klippen Belt (PKB), which is a first-order tectonic structure composed of several successions of mainly deep- and shallow-water limestones mostly of Jurassic to Cretaceous age (Plašienka et al. 2020; Hók et al. 2014). Tectonic units of the Western Carpathians are arranged in a belt-like order with external units in the north and internal units in the south. The External Western Carpathians include the Carpathian Flysch Belt, which is composed of several north-west, north, and north-east verging nappes, as well as the Carpathian Foredeep filled by the Neogene strata. The Internal Western Carpathians were subject to extensive crustal shortening (Plašienka 1997) and include various pre-Tertiary units that are partially covered by Tertiary sediments and Neogene volcanic complexes. Tectonically, the Western Carpathians have a complicated geological structure that has been continuously forming since the Paleozoic era. The oldest Paleozoic rocks experienced the first stage of deformation during the Variscan orogeny; however, the younger Alpine overprint is common. The Alpine orogeny affected the area in several stages from the Jurassic to the Neogene. During this period, parts of the Tethys Ocean were subducted under the African plate, and the Western Carpathian blocks (microplate ALCAPA) were thrust over the margin of the Eurasian plate.

Pannonian Basin

The Pannonian Basin is a typical back-arc basin, which is a result of the subduction and subsequent underthrusting of the European platform under the ALCAPA and Tisza–Dacia microplates (Horváth 1993). The boundary between the Eastern Alpine–Western Carpathian Domain and the North Pannonian Domain is represented by the Rába–Hurbanovo–Diósjenö lineament (RHDL) and is considered to be a projection of the flatlying boundary (Hók et al. 2014). The evolution of the Pannonian Basin is generally related to rollback subduction and collisional processes taking place at the exterior of the Carpathians (Cloetingh et al. 2006; Horváth et al. 2006). The back-arc extension in the Pannonian Basin system in the middle-late Miocene period was coeval with the late stages of thrusting in the adjacent Carpathian belt. The east-west extension within the basin can be related both to the arrangement of continental lithospheric fragment boundaries outside of the Pannonian area (which prohibited continued convergence of the Pannonian fragment with Europe) and to the continued subduction and shortening beneath the Eastern Carpathians at the same time. The shallow course of the LAB boundary indicates the back-arc stretching and mantle upwelling during the Neogene extension in the Pannonian Basin region (Konečný et al. 2002). These processes were accompanied by the Upper Miocene basic and alkaline volcanism (Konečný et al. 2002; Harangi & Lenkey 2007). The Pannonian Basin system is filled by more than 2 km of Paleogene and up to 7 km of Neogene and Quaternary sediments (Bielik 1988; Kilényi & Šefara 1989; Vakarcs et al. 1994; Bielik et al. 2005).

Lithosphere asthenosphere boundary

The lithosphere–asthenosphere boundary (LAB) in Europe was interpreted by Plomerová & Babuška (2010), Jones et al. (2010), Dérerová et al. (2006), or Alasonati Tašárová et al. (2016). Beneath the Bohemian Massif, the LAB is quite flat with moderate thickening to the south and the estimated depth range of 130–150 km. In the Western Carpathian region, a significant feature of the LAB topography is a broad lithospheric thinning with distinct and sharp LAB, shallowing to the depths of ~60–80 km beneath the Pannonian Basin.

Seismic velocity structure along profile CEL09

The NW–SE oriented CEL09 profile of the CELEBRATION 2000 seismic experiment was designed to cross the key tectonic units of the Bohemian Massif, Western Carpathians, and Pannonian Basin (Fig. 1). It is 720 km long and intersects (from NW to SE): the Saxothuringian, the Mariánské Lázně amphibolite Complex (MLC), the Teplá–Barrandian, the granitoid intrusions extending along the Central Bohemian Shear Zone (CBSZ), the Moldanubian, the Moravo–Silesian, the Carpathian Foredeep, the External Western Carpathians, the Vienna Basin, the southern part of the Malé Karpaty Mts, the Danube Basin, the Trans-Danubian range, and the Pannonian Basin (Guterch et al. 2003a, b; Růžek et al. 2003; Hrubcová et al. 2005).

The refraction and wide-angle reflection data along this profile were interpreted by seismic tomographic inversion and by 2D forward ray-tracing modelling of P and S waves enhanced by amplitude modelling of further phases in later arrivals (reflections and available refractions; Hrubcová et al. 2005, 2015; Hrubcová & Geissler 2009). The retrieved crustal structure (Fig. 2) shows the upper crust of the Bohemian Massif with a relatively high near-surface P-velocity gradient and velocities of 5.8–6.0 km s^{-1} at 2–3 km depths. Deeper parts of the upper crust exhibit a very low vertical gradient and the velocity of ~6.0 km s^{-1}. The sedimentary cover of the Carpathian Foredeep, the Western Carpathians and the Pannonian Basin is characterized by lower velocities (3.8–4.2 km s^{-1}) down to depths of 3–6 km. The highest velocity gradient is at the contact of the Western Carpathians with the Bohemian Massif (velocities of 2.5–5.5 km s^{-1}) where the Western Carpathian Flysch extends to the depth of ~6.5 km. The alternation of higher and lower velocities at the contact of the Western Carpathians with the Pannonian Basin represents the andesitic and rhyolitic rocks of the Tertiary volcanic edifices followed by volcano-sedimentary complexes. Very low velocities (2.2–3.2 km s^{-1}) down to a depth of 2 km reflect the Neogene and Quaternary sediments of the Pannonian Basin.

In the upper and middle crust, two reflectors at depths of 8–13 km and 16–21 km are identified through the whole profile. In the central part of the Moldanubian and the Moravo–Silesian, it is characterized by a depth of 10–12 km, shallowing to 8 km beneath the Western Carpathians and to 7–8 km beneath the Pannonian Basin. The lower reflector is the deepest near the contact of the Saxothuringian and Moldanubian, in the Moravo–Silesian it is almost horizontal at 18 km depth. In the Western Carpathians and the Pannonian Basin, this reflector is shallower, undulating at depths of 16–18 km.

Different units of the Bohemian Massif show different characters of the crust-mantle transition and the Moho. In the Saxothuringian and partly beneath the Teplá–Barrandian, a lower-crustal high-velocity layer (velocities 6.9–7.5 km s^{-1}) was inferred above the Moho with the top at a depth of 26–28 km. The Moho was interpreted at the bottom of this layer with a small velocity contrast masked by reflectivity within the layer (Hrubcová et al. 2013, 2017). The Moldanubian shows the Moho as the first-order discontinuity with the velocity increase from 6.9 to 8.1 km s^{-1}, and the maximum Moho depth of 39 km. The upper mantle reflector is detected at depths of ~55 km. The crust-mantle transition at the contact of the Western Carpathians with the Bohemian Massif shows pronounced lateral variations with the step-like Moho anomaly at depths from 28 to 34 km (Hrubcová et al. 2008, 2015). The Moho in the Western Carpathians is in a depth range of 28–31 km with the upper-mantle velocities smaller than in the Bohemian Massif. Beneath the Pannonian Basin the Moho is shallower with the first-order discontinuity at a depth of ~25 km. It shows a sharp velocity increase from 6.5 to 7.8 km s^{-1} in agreement with other geophysical interpretations in this area (Posgay et al. 1981; Grad et al. 2006; Środa et al. 2006).

2D gravity modelling

The 2D gravity modelling is a suitable method that complements the results from 2D refraction and reflection seismics. It results in a 2D (vertical section) structural density model. The quantitative interpretation of the gravity data acquired along the CEL09 profile was performed with the GM-SYS (Gravity and Magnetics Modelling System) software, an interactive tool for calculating the 2D gravity and magnetic effect of geological models with the fast calculation of the response (GM-SYS User's Guide for version 4.9 2004).

The forward modelling is based on the principle that the observed gravity data are equal to the gravity effect of the subsurface density distribution. Alternatively, the topographically corrected gravity anomalies/disturbances are equal to the gravity effect of the subsurface anomalous density distribution (Vajda et al. 2020). The complete Bouguer anomaly (CBA) serves as the input for modelling the topographically corrected anomalous gravity. The modelling is done in iterations until the optimal fit between the gravity effect of the model density distribution and the observed gravity data is achieved.

Topography and gravity

The topography and the CBA along the CEL09 profile are shown in Fig. 3. The topographic data was taken from the SRTM (Jarvis et al. 2008) with 2-km resolution and reported vertical accuracy better than ±16 m. The gravity values were digitized from a newly compiled digital data set of the CBA map in the pan-Alpine area (Zahorec et al. 2021) with the 2-km resolution. In the Czech Republic and the Slovak Republic this compilation was based on the gravity data published by Ibrmajer (1963), Švancara (2004), Bielik et al. (2006), and Švancara et al. (2021). The accuracy of the gravity data is better than ±1 mGal (Pašteka et al. 2017). Generally, the Bohemian Massif area is characterized by higher topography (~500 m a.s.l.) compared to the Western Carpathian–Pannonian Basin region (~230 m a.s.l.). The gravity values change from −48.4 up to +23.3 mGal and show several relative gravity highs and lows separated by significant horizontal gravity gradients (Fig. 3).

Initial density model

The selection of a starting (initial) density model is cruicial for successful iterative modelling process. Our initial model consisted of individual bodies with constat densities and, except for the uppermost parts, it was based on the distribution of seismic velocities along the refraction and wide-angle reflection profile CEL09 (Hrubcová et al. 2015), see Fig. 2. In this model, we preserved following boundaries: (a) the Tertiary basement of the sedimentary basins, (b) the boundary between the upper and lower part of the upper crust, (c) the boundary between the upper and lower crust, and (d) the Moho.

The densities of the surface and near-surface (<5 km) anomalous bodies were defined based on the results from laboratory measurements on surface and borehole rock samples, and from well-logging (e.g., Eliáš & Uhmann 1968; Šefara et al. 1987; Ibrmajer et al. 1989; Szabó & Páncsics 1999). The rock

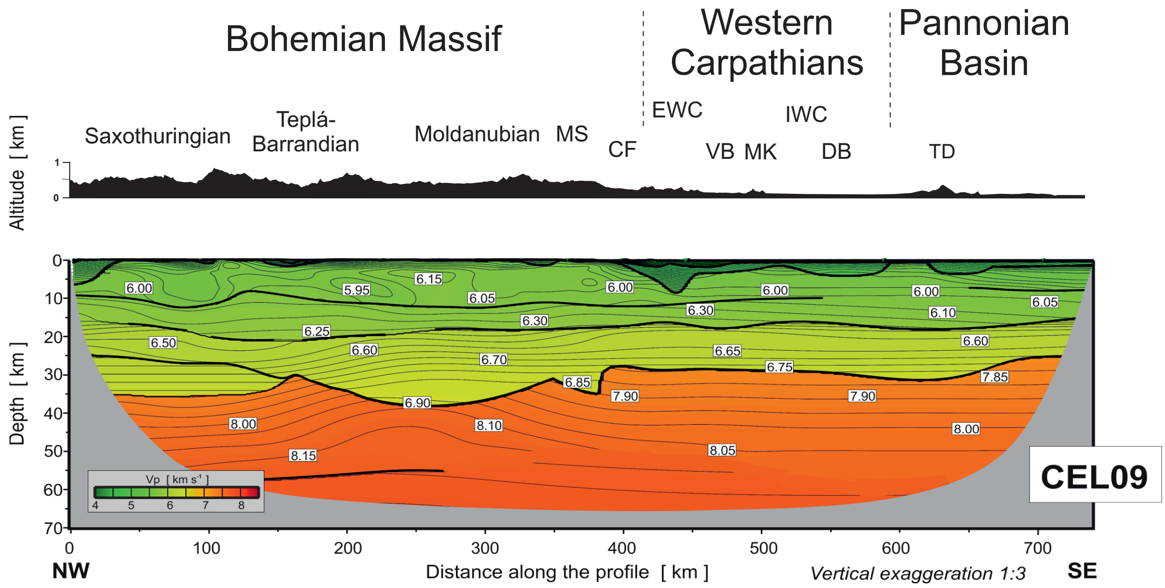

Fig. 2. P-wave velocity model derived from ray-tracing modelling along the CEL09 profile (modified after Hrubcová et al. 2015). MS – Moravo–Silesian, CF – Carpathian Foredeep, EWC – External Western Carpathians, VB – Vienna Basin, IWC – Internal Western Carpathians, MK – Malé Karpaty Mts., DB – Danube Basin, TD – Trans-Danubian range. Vertical exaggeration is 1:3.

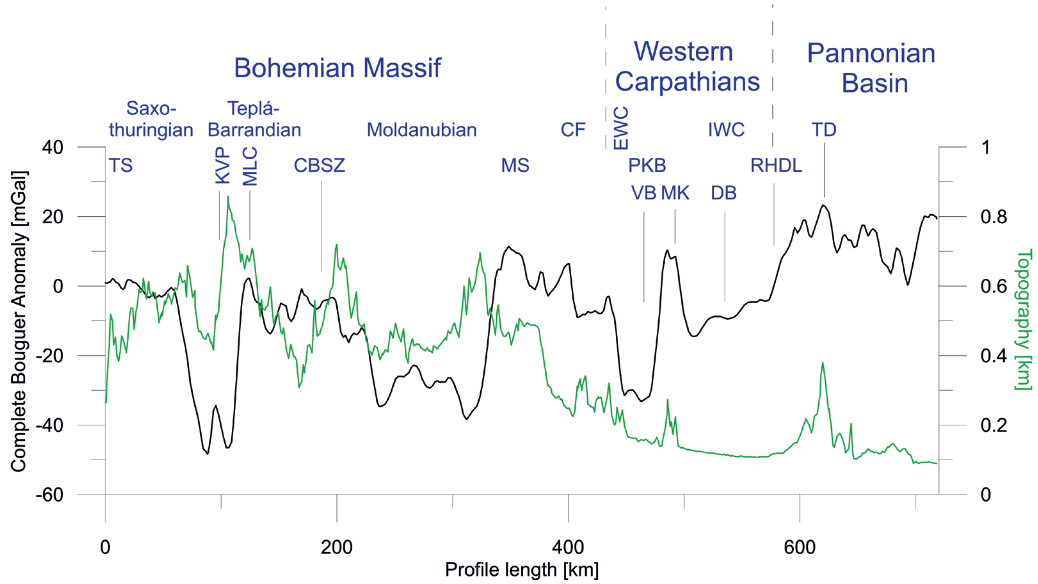

Fig. 3. Complete Bouguer anomaly (black, Zahorec et al. 2021) and topography (green, Jarvis et al. 2008) along the CEL09 profile. TS – Teuschnitz syncline, KVP – Karlovy Vary pluton, MLC – Mariánské Lázně amphibolite Complex, CBSZ – Central Bohemian Shear Zone, MS – Moravo–Silesian, CF – Carpathian Foredeep, EWC – External Western Carpathians, PKB – Pieniny Klippen Belt, VB – Vienna Basin, IWC – Internal Western Carpathians, MK – Malé Karpaty Mts., DB – Danube Basin, RHDL – Rába–Hurbanovo–Diósjenö lineament, TD – Trans-Danubian range.

densities at depths >5 km were transformed from modelled seismic Vp velocities after Hrubcová et al. (2015) to in situ densities ρ in a 50-km horizontal and 5-km vertical grid (Table 1). The transformation was carried out by two different approaches: (1) for the upper and lower crust, the Sobolev–Babeyko's (Sobolev & Babeyko 1994) and Christensen–Mooney's (Christensen & Mooney 1995) formulas were applied; (2) for the lower lithosphere, the formulas of Christensen & Mooney (1995) as well as Lachenbruch & Morgan (1990) were used. The Vp velocities were considered only to a depth of 50 km, although the velocity model indicated the seismic discontinuity within the lower lithosphere at ~55 km depth (120–270 km along the profile). However, the velocities beneath this reflector were not constrained as

Table 1: Summary of the resulting in situ densities (ρ) from Christensen & Mooney's (1995) and Sobolev & Babeyko's (1994) formulas.

Average densities (g cm^{-3})		Profile (km)													
		50	**100**	**150**	**200**	**250**	**300**	**350**	**400**	**450**	**500**	**550**	**600**	**650**	**700**
Depth (km)	10	2.75	2.69	2.78	2.68	2.68	2.69	2.70	2.70	2.68	2.74	2.69	2.69	2.70	2.70
	15	2.78	2.80	2.80	2.78	2.78	2.78	2.79	2.78	2.79	2.79	2.78	2.78	2.75*	2.75*
	20	2.88	2.84	2.79	2.84	2.89	2.89	2.89	2.92	2.92	2.92	2.92	2.92	2.93	2.94
	25	2.91	2.90	2.90	2.93	2.96	2.95	2.95	2.96	2.96	2.96	2.96	2.96	2.98	3.18*
	30	3.10	3.10	2.96	2.99	3.02	3.02	3.02	3.28	3.28	3.28	3.28	3.28	3.18*	3.26
	35	3.30	3.30	3.29	3.30	3.05	3.04	3.30	3.30	3.30	3.29	3.28	3.28	3.27	3.26
	40	3.30	3.29	3.30	3.32	3.32	3.32	3.31	3.30	3.31	3.30	3.30	3.30	3.28	3.28
	45	3.30	3.30	3.32	3.33	3.34	3.33	3.32	3.32	3.32	3.32	3.30	3.30	3.29	–
	50	–	3.32	3.32	3.33	–	–	–	–	3.33	3.32	3.32	3.32	3.30	–

– No or problematic Vp data, * problematic determination.

no arrivals were observed from below (Hrubcová et al. 2005, 2015), so the densities were not calculated.

As far as concerns deeper parts, Lillie et al. (1994) found that the configuration of the LAB is an important component in modelling long-wavelength gravity anomalies in the Alpine–Carpathian–Pannonian region. Therefore, this boundary was introduced into our modelling. The LAB topography (Fig. 4) was defined based on seismic (Jones et al. 2010; Plomerová & Babuška 2010) and magnetotelluric (Praus et al. 1990) data, and integrated geophysical modelling (Dérerová et al. 2006; Alasonati Tašárová et al. 2016). Transformed densities ρ for the lower lithosphere changes from 3.27 to 3.33 g cm^{-3} along the profile. Thus, we determined the average density for this layer to be 3.30 g cm^{-3}. The average asthenosphere density (3.27 g cm^{-3}) was estimated based on the results of Lillie et al. (1994) who assumed differential density of ~ −0.03 g cm^{-3} in the asthenosphere compared to lower lithosphere for the Alpine–Carpathian–Pannonian region. The calculated gravity effect of the LAB is −45 mGal, which clearly justifies the interpretation of the observed gravity data with this boundary.

The initial density model is shown in Fig. 4 and its gravity response indicates only a partial correlation with the CBA along the profile (e.g., in the Moldanubian, the Moravo–Silesian, the Carpathian Foredeep and the Trans-Danubian range). However, large root-mean-square (RMS) differences of ~28 mGal appear in different locations (e.g., the Saxothuringian, the Karlovy Vary Pluton, the Danube Basin). Above, in the Western Carpathian–Pannonian region, the calculated gravity anomalies seem to be shifted compared to the observed data (e.g., in the Vienna Basin, the Malé Karpaty Mts., and in the Danube Basin).

Resultant density model

To achieve the resultant (final) density model (Fig. 5), we modified interactively the position, geometry and density of the anomalous bodies by trial-and-error approach until a reasonable fit was obtained. Considerable changes in the final density model were related to the sedimentary layers in the Western Carpathian–Pannonian region, the outcrops of the Malé Karpaty Mts., the Trans-Danubian range and the upper part of the upper crust. Considering constraints on boreholes (e.g., Biela 1978a, b), sedimentary densities (e.g., Eliáš & Uhmann 1968; Šefara et al. 1987; Ibrmajer et al. 1989), sedimentary thicknesses (e.g., Fusán et al. 1987; Kilényi & Šefara 1989; Krejčí & Jurová 1997; Kováč 2000; Bielik et al. 2005), and surface geology (Kodym et al. 1967; Geological map of Slovakia 2013), the upper part of the upper crust was modelled in more detail compared to seismic modelling. At modification of the anomalous bodies in the upper crust of the Bohemian Massif we also used the results obtained by Hrubcová et al. (2005).

The RMS difference between the calculated and observed gravity for the final model amounted ~1.00 mGal, which represents a reasonable fit considering the accuracy of the input gravity data (CBA).

Results and interpretation

The final density lithospheric model (Fig. 5) consists of five principal layers: sediments, upper crust, lower crust, lower lithosphere, and asthenosphere.

The sedimentary cover is formed by the Tertiary sediments. Note that the occurrence of this layer, except for a small profile section (80–90 km distance) in the Bohemian Massif, is exclusively in the Western Carpathian–Pannonian region. The Neogene sediments in the Carpathian Foredeep have an average density of 2.30 g cm^{-3} and their thickness reaches the maximum values of ~3.0 km. The Flysch sediments of the External Western Carpathians have a higher average density of 2.51 g cm^{-3} and their thickness is slightly more than 6 km. In the Vienna, Danube and Pannonian Basins the Neogene sediments are characterized by an average density of 2.37 g cm^{-3}. The sedimentary thickness of the Vienna Basin varies from 0 km to >5 km. In the Danube Basin, and especially in the Pannonian Basin, the pre-Cainozoic basement relief is broken into partial elevations and depressions. The maximum depth of the Danube Basin is ~3.6 km, while in the Pannonian Basin it is ~4.5 km.

The pre-Paleogene upper crustal layer consists of two parts: the upper and lower part. The composition and structure of the upper part is the most varied and needed the most

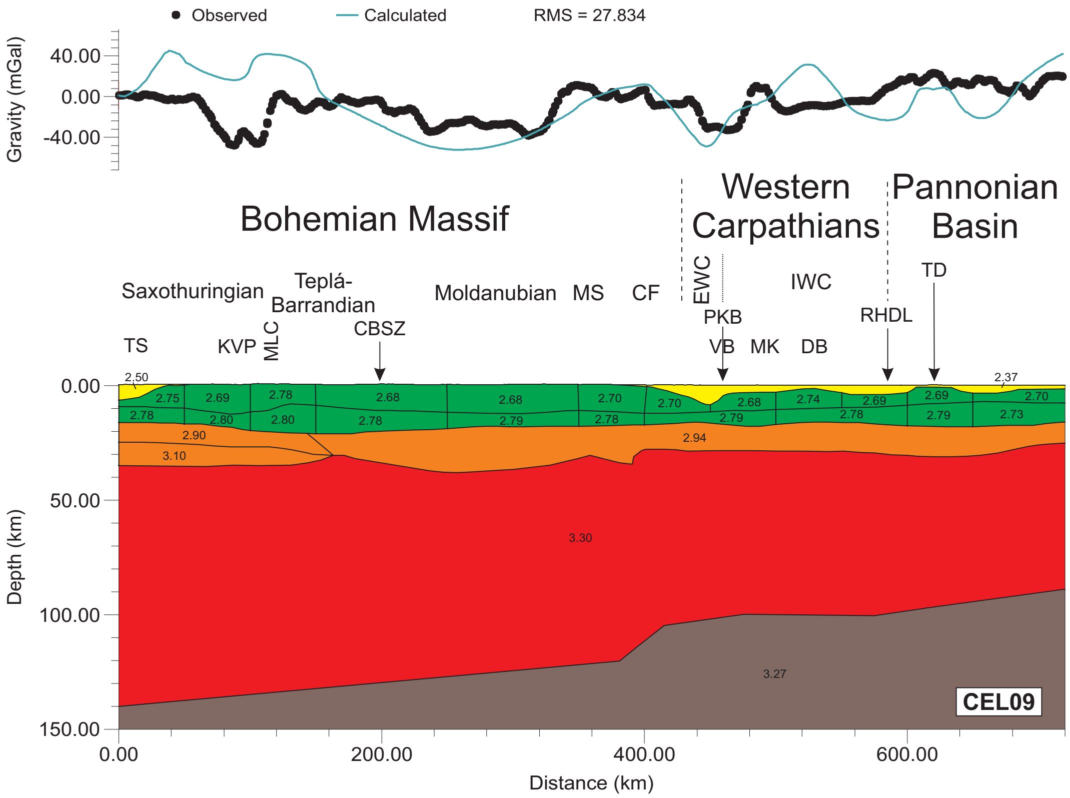

Fig. 4. Initial density model based on the seismic results, and the in situ densities ρ transformed from modelled Vp. For explanations, refer to Fig. 3. Root-mean-square (RMS) between the observed and calculated gravity indicated. Vertical exaggeration is 1:2.

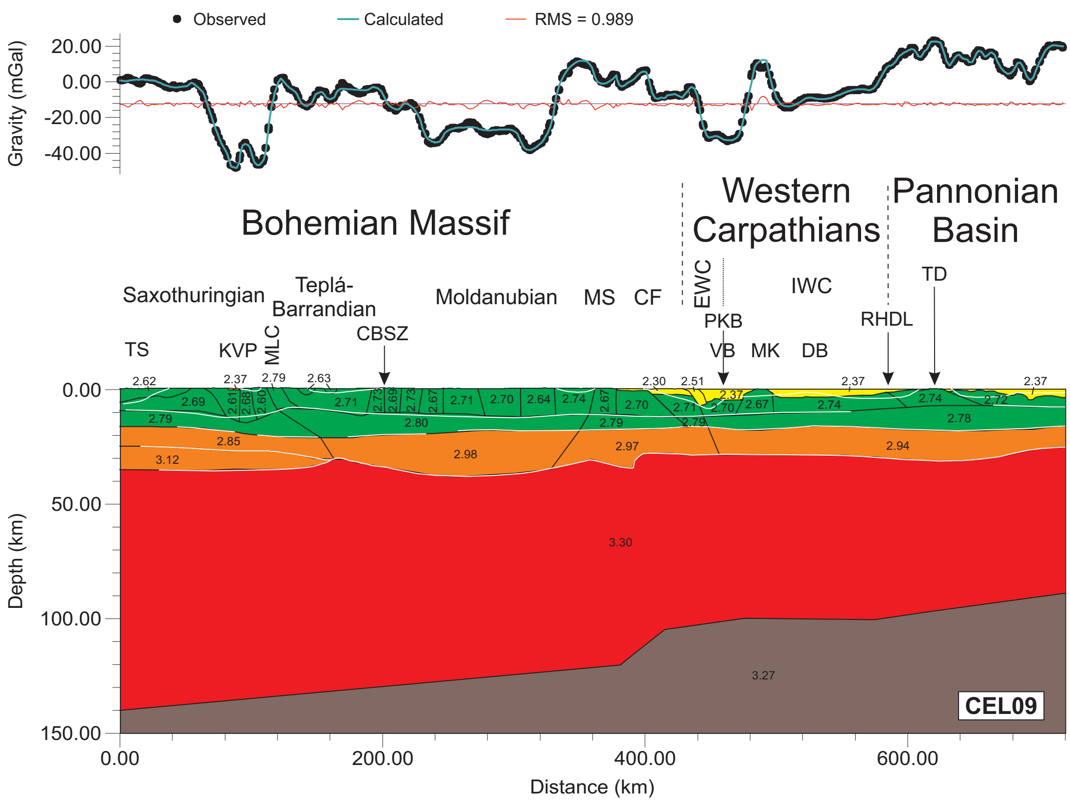

Fig. 5. Final 2D density lithospheric model of the CEL09 profile. White lines represent the boundaries of the seismic model calculated by Hrubcová et al. (2015). For explanations, refer to Fig. 3. Vertical exaggeration is 1:2.

modifications due to sources of short-wavelength gravity anomalies as discussed also by Hrubcová et al. (2005). The densities in this part range from 2.60 to 2.79 g cm^{-3}.

In the Bohemian Massif, the largest gravity low (almost −50 mGal) is interpreted as the low-density Karlovy Vary pluton (2.60–2.68 g cm^{-3}). The bottom boundary of this pluton is modelled at a depth of ~14.5 km. This result is in agreement with Ibrmajer et al. (1989), who hypothesized that the pluton may reach to a middle part of the crust. The granitoid pluton is part of the SE margin of the Saxothuringian and at the same time it forms the border between the Saxothuringian and the Teplá–Barrandian. On the contrary, the north-western part of the Saxothuringian is formed by the Carboniferous Flysch of the Teuschnitz syncline (Hrubcová et al. 2005) with a density of 2.62 g cm^{-3}. The Teplá–Barandian unit is represented by a gravity high, which is due to the Proterozoic spilite zones with accumulations of basic volcanites and other mafic rocks (Ibrmajer et al. 1989; Bielik et al. 2006). In the westernmost part of the Teplá–Barrandian, the high-density MLC (2.79 g cm^{-3}) is interpreted. The Moldanubian gravity low is caused mostly by huge bodies of low-density granitoids (2.64–2.71 g cm^{-3}) belonging to the Central Bohemian and Moldanubian plutons. The south-easternmost part of the Bohemian Massif is represented by a narrow band of the Moravo–Silesian unit, which is formed by metamorphic rocks with densities of 2.67–2.74 g cm^{-3}.

In the Western Carpathian–Pannonian part, the crystalline Malé Karpaty Mts. and the Trans-Danubian range cause significant gravity highs. Moreover, the contacts of both mountains with the surrounding sedimentary basins are very steep and accompanied by significant horizontal gravity gradients. From this point of view, these contacts are probably of tectonic origin. The pre-Cainozoic basement of the Carpathian Foredeep, the External Western Carpathians, the Vienna Basin, and the Pannonian Basin has a slightly lower average density (2.70–2.72 g cm^{-3}) compared to the Danube Basin (2.74 g cm^{-3}). This effect can probably be explained by the influence of the high-density Kolárovo anomalous body (e.g., Bielik et al. 1986; Prutkin et al. 2011, 2014), which is in the Danube Basin upper crust and it is formed by mafic or ultramafic rocks (Šujan et al. 2021 and references therein).

The lower part of the upper crust is divided into four sectors. The first three sectors are assigned to the Bohemian Massif and the last one to the Western Carpathian–Pannonian region. Beneath the Saxothuringian, the Moravo–Silesian and the Carpathian Foredeep, it is characterized by a density of 2.79 g cm^{-3} and under the Teplá–Barrandian and Moldanubian units it is 2.80 g cm^{-3}. The Western Carpathian–Pannonian lower part of the upper crust has a density of 2.78 g cm^{-3}. Its thickness along the whole profile is quite variable and varies from 4 to 12 km. The largest thicknesses can be observed under the Teplá–Barrandian and the Trans-Danubian range. On the contrary, the thinnest is under the Karlovy Vary pluton and the External Western Carpathians.

The lower crust consists of five different zones. The first two zones form the lower crust beneath the Saxothuringian and partly the Teplá–Barrandian. The upper zone is characterized by the average density of 2.85 g cm^{-3} (average Vp velocity is ~6.5 km s^{-1}) and the lower zone by a higher density of 3.12 g cm^{-3} (Vp velocities range from 7.1 km s^{-1} at the top to 7.9 km s^{-1} at the bottom of this zone). The lower zone represents a laminated high-velocity lower crustal layer, which is typical for the Variscan areas (DEKORP Research Group 1994) and it may be caused by mafic intrusions or underplating (Hrubcová et al. 2005, 2015, 2017). The Moldanubian lower crust has a density of 2.98 g cm^{-3}. The fourth zone represents the lower crust of the Moravo–Silesian and the Carpathian Foredeep with the density of 2.97 g cm^{-3}. The fifth zone belongs to the Western Carpathian–Pannonian region with the average density of 2.94 g cm^{-3}.

Discussion

In gravity modelling, the errors of the resultant models come from a combination of several factors: the inaccuracy of topographical and gravity data, uncertainties in determination of the rock densities, uncertainties in the seismic velocity model, inaccuracies of the Moho and LAB topographies, amount of data, or inaccuracy of modelling (misfit between observed and calculated gravity values, 2D method of interpretation not accounting for 3D structure). Since the errors introduced by the interpreter during the trial-and-error gravity modelling are subjective and impossible to quantify, it is not possible to perform a systematic error analysis. Therefore, we calculated many models before finding the ones presented here.

Different transformation formulas determine densities with different accuracy. The average error in densities calculated according to Sobolev & Babeyko's (1994) formula is ±0.05 g cm^{-3}; in the case of Christensen & Mooney (1995) it varies from ±0.05 to ±0.12 g cm^{-3}. Our resultant densities are in the range of inaccuracy of applied transformation.

One of the basic premises of our modelling was that the lower part of the upper crust, the lower crust, and the Moho of the density models were defined exclusively by seismic model (Hrubcová et al. 2015) because its accuracy is higher than the accuracy of the density model. However, to obtain an acceptable fit between measured and calculated data from gravity modelling, especially under the Karlovy Vary Pluton, the Teplá–Barrandian and the Moldanubian, we adjusted boundaries of these layers. Nevertheless, these changes were only of ±1–2 km, which is roughly in the range of inaccuracy of seismic modelling (Hrubcová et al. 2005, 2015).

The LAB in our model is based on previous interpretations. Jones et al. (2010) and Plomerová & Babuška (2010) estimated the LAB depth variations to be of ~±10 km. According to Grinč et al. (2013) and Šimonová et al. (2019), integrated modelling involves estimated uncertainty of thermal lithospheric thickness of ~10–20 km and/or 10–15 %. Since the gravity modelling of the LAB is much more inaccurate due to its large depth and small density contrast, we did not change the LAB topography. Nevertheless, its implementation together with its density contrast explain the long-wavelength gravity anomalies as suggested also by Lillie et al. (1994).

The distribution of seismic velocities in the lower lithosphere (Fig. 2) indicates that the average lithospheric density under the Bohemian Massif is slightly higher compared to the Western Carpathians. Therefore, we also calculated the second (optional) lithospheric density model (Fig. 6), where the average lower lithospheric density of the Bohemian Massif is 0.01 g cm^{-3} higher (3.30 g cm^{-3}) than in the Western Carpathian–Pannonian region (3.29 g cm^{-3}). This is also consistent with the results of the geophysical–petrological modelling (Alasonati Tašárová et al. 2016) detecting zone of lower velocities and densities (~3.29 g cm^{-3}) in the lower lithosphere below the Danube and Pannonian Basins. When comparing the optional density model (Fig. 6) with the original model (Fig. 5), they differ in the crustal and lower lithosphere densities, while the densities of sedimentary basins and the position and geometry of individual bodies are preserved. Above, the density changes do not exceed ±0.03 g cm^{-3}, which means that they are significantly smaller than the inaccuracy of the modelling.

We provide our gravity modelling along the 720 km-long profile CEL09 based on seismic velocities after Hrubcová et al. (2015). This model (Fig. 2) results not only from the first arrivals but also from the amplitude modelling of further phases in later arrivals (reflections and available refractions). Compared to the first arrival inversions (e.g. Růžek et al. 2007), which belong to the class of the simplest and smoothest velocity models, Hrubcová et al. (2015) provide more constraints on seismic velocities (especially in the middle and lower crust, which is usually not constrained by the first arrival inversions), as well as discontinuities (crustal and/or mantle reflectors, and the Moho). This is very important for gravity modelling. Further velocity models in the area, such as the ambient noise results (Ren et al. 2013; Schippkus et al. 2018) are confined to the S-wave velocities and thus not suitable for density recalculations.

The seismic velocities in the Bohemian Massif were also interpreted by Hrubcová et al. (2005), who, on top of that, recalculated the velocities into densities to verify their seismic model. When comparing both density models, there is a general agreement in the western and central Bohemian Massif; however, there are also several differences: (1) Our gravity model is calculated for the entire profile (720 km length) expanding the interpretation to the Western Carpathian-Pannonian region. (2) The Tertiary sediments in our modelling are constrained by results from surface and borehole laboratory measurements, and from well-logging. (3) We provide gravity modelling through the whole lithosphere, while Hrubcová et al. (2005) refined their densities mainly in the upper crust. (4) In the lower crust, at the contact of the Bohemian Massif and the Western Carpathians, our densities are calculated from the longer seismic velocity model of Hrubcová et al. (2015), which is better constrained due to implemented reciprocity from the Pannonian Basin and incorporation of the off-line modelling (for further details, see Hrubcová et al. 2015). (5) We implement the course of the LAB, calculate its gravity effect, and provide realistic densities in the lower lithosphere.

In the lower lithosphere, our results show a density difference at ~460 km along the profile implying the contact between the Bohemian Massif and the ALCAPA microplate (Fig. 7). The contact of these two units at this place complies with the overthrusting of the Western Carpathians onto the Bohemian Massif in the NW direction, also supported by the location of the alkaline volcanism (Vass et al. 1988), the pull-apart Vienna Basin (Royden et al. 1983) in the SE, and deep position of the PKB.

The contact of the Bohemian Massif and the ALCAPA plate at ~460 km distance also agrees with the interpretation of Tomek & Hall (1993), who integrated similar Moho anomaly and shallower crust/mantle boundary (depths of 28–30 km) located south-easterly at the reflection profile 8HR to the Bohemian Massif plate. Their explanation involves a “neo-Moho” that developed after subduction either by a stress relaxation or by a phase change of the formerly mafic lower crust into the gabbro-eclogite.

Thus, we can delineate a whole lithospheric zone affected by the Miocene subduction delimited by (1) the change of the topography at the crust/mantle Moho boundary, (2) the change of the topography at the LAB, and (3) the density difference in the lower lithosphere (Figs. 6, 7). This zone was probably formed after subduction and can represent “neo-lithospheric” contact of the subducting Bohemian Massif and the ALCAPA plates.

Seismic reflection data along the reflection profile 9HR (Tomek & Hall 1993) indicate that the Saxothuringian was underthrusted along a SE dipping thrust zone beneath the Teplá–Barrandian/Moldanubian. The seismic CEL09 model (Hrubcová et al. 2005) supported this idea by locating the Saxothuringian/Teplá–Barandian contact at the lower crustal level. Its location was inferred from the differences between a high-velocity strongly reflective lower crust, which was attributed to the Saxothuringian unit, and a moderate-velocity unreflective lower crust and sharp Moho characteristics for the Barrandian/Moldanubian unit. In our density models (Figs. 5, 6) two Saxothuringian lower crustal layers are characterized by the densities of 2.84–2.85 and 3.12 g cm^{-3}, while the Barrandian/Moldanubian lower crust is represented by only one layer with a density of 2.98 g cm^{-3}. Though the average lower crustal densities under both these tectonic units are the same, they differ in nature, since the lower zone of the lower crust in the Saxothuringian (density 3.12 g cm^{-3}) corresponds to the high-velocity laminated lower crust detected from seismic modelling (Hrubcová et al. 2005) and represents the mafic composition not detected beneath the Moldanubian. Another crustal thrust zone (at 328–360 km distance) is characterized by the density contrast of 0.01 g cm^{-3} in the lower crust (Figs. 5, 6) assuming the Moldanubian overthrusting onto the Moravo–Silesian during the Variscan collision as discussed by Hrubcová et al. (2005).

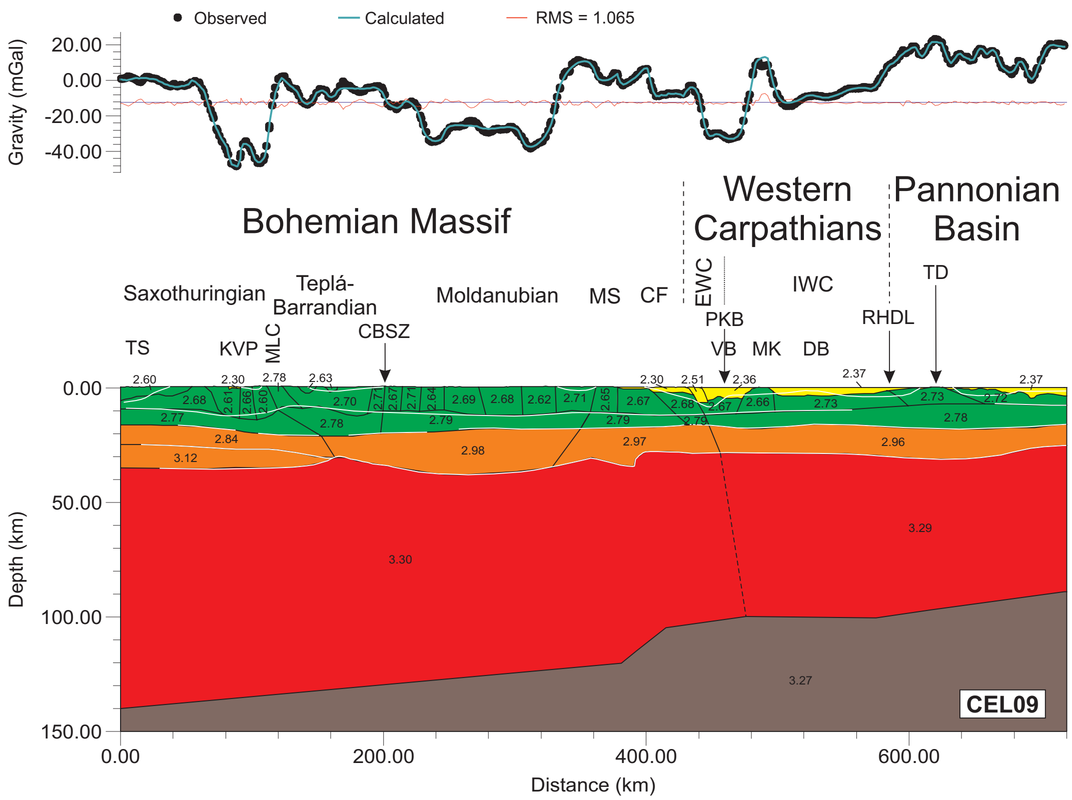

Fig. 6. Optional 2D density lithospheric model of the CEL09 profile. Variant of the resulting 2D density model, where a density contrast of 0.01 g cm^{-3} between the Bohemian Massif and the Western Carpathian–Pannonian lower lithospheres was assumed. White lines represent boundaries of the seismic model calculated by Hrubcová et al. (2015). For explanations, refer to Fig. 3. Vertical exaggeration is 1:2.

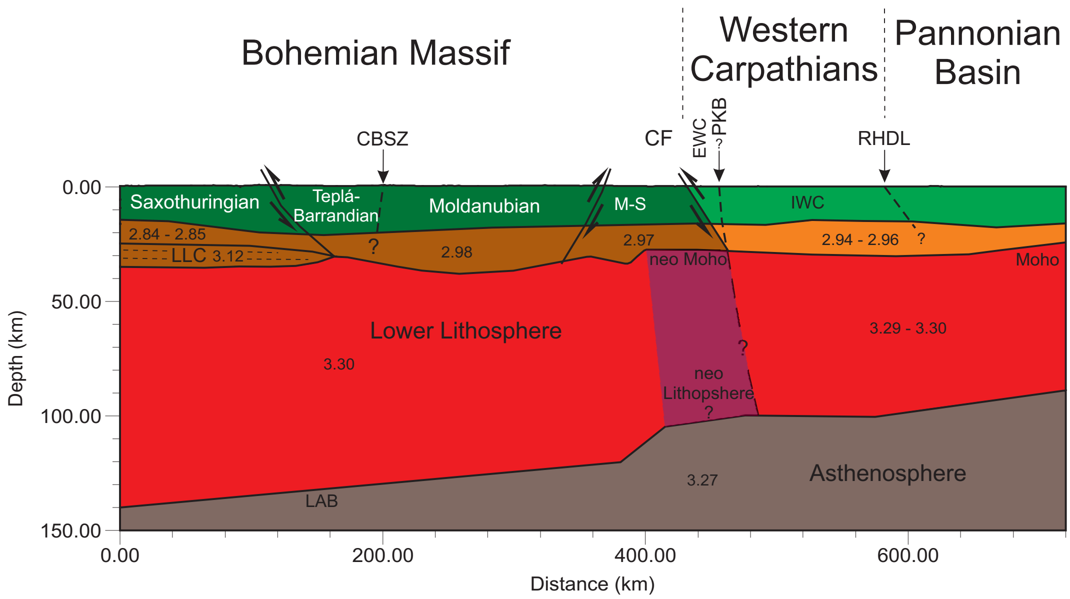

Fig. 7. Schematic tectonic model of the CEL09 profile summarizing the main interpreted density features of the lithosphere with their possible tectonic interpretation (the Bohemian Massif is modified after Hrubcová et al. 2005). The density characteristics illustrate differentiation in the lower crust for different parts of the Bohemian Massif and the Western Carpathian–Pannonian region. Values are in g cm^{-3}. LLC – laminated part of the Saxothuringian lower crust. For explanations, refer to Fig. 3. Arrows indicate relative movement along the contact zones. Note the "neo-lithospheric contact" of the Bohemian Massif and the Western Carpathians marked by purple zone. Vertical exaggeration is 1:2.

Conclusions

We provide 2D gravity modelling along the CELEBRATION 2000 seismic refraction and wide-angle reflection profile CEL09 crossing the Bohemian Massif, the Western Carpathians, and the Pannonian Basin. Our resulting 2D density models provide an improved interpretation of the upper crust with incorporated geological and geophysical constraints, and shed light on the composition, structure, and tectonics of the lithosphere in the area. The obtained results can be summarized as follows:

- The lithospheric density model consists of five principal layers: sediments (2.30–2.51 g cm^{-3}), upper crust (2.60–2.80 g cm^{-3}), lower crust (2.84–3.12 g cm^{-3}), lower lithosphere (3.29–3.30 g cm^{-3}) and asthenosphere (3.27 g cm^{-3}).
- The Tertiary sedimentary layer in the Western Carpathian–Pannonian region is formed by the higher-density Paleogene (2.51 g cm^{-3}) and the lower-density Neogene sediments (2.30–2.37 g cm^{-3}). The thickness of the Neogene sediments in the Carpathian Foredeep, the Vienna Basin, the Danube Basin, and the Pannonian Basin ranges from 0 to ~5.5 km. The Paleogene sediments of the External Western Carpathian flysch zone dip to a depth of ~6.5 km.
- The pre-Paleogene upper crustal layer is split into two layers: the upper and lower part. The boundary between them is located at 7–15 km depths. The upper part of the crust is divided into inhomogeneities with varying densities. In the Bohemian Massif, the positive and negative gravity anomalies largely reflect on the one hand the gravity effects of the light granitoid plutons (~2.60–2.68 g cm^{-3}) and metamorphic rocks (~2.69–2.74 g cm^{-3}), on the other hand the heavy basic and ultrabasic bodies (~2.78–2.79 g cm^{-3}). In the Western Carpathians, this layer is built by the crystalline Malé Karpaty Mts. (2.66–2.67 g cm^{-3}) and Trans-Danubian range (2.73–2.74 g cm^{-3}), and the pre-Cainozoic basement of the sedimentary basins (the Carpathians Foredeep, the Vienna Basin, the Danube Basin, and the Pannonian Basin), which have densities varying from 2.67 to 2.74 g cm^{-3}.
- The densities of the lower part of the upper crust range in narrow interval from 2.78 g cm^{-3} (in the Western Carpathian–Pannonian region) to 2.77–2.80 g cm^{-3} (in the Bohemian Massif).
- Four (five) different sectors of the lower crust result from the first/optional density models. The Saxothuringian sector consists of the upper (2.84–2.85 g cm^{-3}) and lower (3.12 g cm^{-3}) layers; its total thickness varies from 9 km to 19 km. The Moldanubian lower crust is characterized by the density of 2.98 g cm^{-3} with the maximum thickness of ~20 km. The 13-km-thick Moravo–Silesian lower crust dipping under the Carpathians Foredeep and the External Western Carpathians have a density of 2.97 g cm^{-3}. The Western Carpathian–Pannonian region is represented by slightly lower-density of 2.94–2.96 g cm^{-3} compared to the Bohemian Massif with the average thickness from 9 to 13 km.
- The comparison of surface and deep (lithospheric) structure suggests that the Western Carpathian are overthrust by ~30 km onto the Bohemian Massif.
- The detected lithospheric zone (in Fig. 7 marked by purple colour) is delimited by: (1) the change of the topography at the crust/mantle Moho boundary; (2) the change of the topography at the LAB; and (3) the density difference in the lower lithosphere. This zone was probably formed after the Miocene subduction and can represent the "neo-lithospheric contact" of the subducting Bohemian Massif beneath the Western Carpathians that developed either by stress relaxation or by phase changes.

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