Lithospheric model along transect HT-1 across Western Carpathians and Pannonian Basin based on 2D integrated modelling

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Abstract: 2D integrated modelling approach was applied to determine the lithospheric structure along transect HT-1 located in the Carpathian-Pannonian Basin–European platform region. Our approach combines simultaneous interpretation of surface heat flow, topography, gravity and geoid data. All available geophysical and geological data were used to create an initial model that has been afterwards modified by trial and error method until reasonable fit was obtained between input data and model predictions. The main focus of our study was the position and shape of the lithosphere-asthenosphere boundary (LAB). In the Pannonian Basin the modelled LAB is at depths of about 80–90 km and rapidly dips towards the Western Carpathians where its depth reaches values 145 to 150 km. Beneath the European platform the LAB depth is about 135–140 km. We can observe a slight lithospheric root under the Western Carpathians. This lithospheric thickening is interpreted as a small remnant of a subducted slab. This result is in a good agreement with the previous lithospheric models in the Carpathian-Pannonian Basin.

Key words: integrated 2D modelling, heat flow, topography, gravity, geoid, lithosphere, asthenosphere, Carpathian-Pannonian–European platform

1. Introduction

The Carpathian-Pannonian–European platform area with its geological complexity provides a great opportunity to study the structure of the lithosphere, the asthenospheric and lithospheric processes taking place within it and their mutual interaction during the continental collision, the orogeny,

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the volcanic arc and related fore-arc and back-arc basin development. For the study of the lithospheric structure we utilize the 2D integrated modelling method that simultaneously interprets different geophysical data. This method has already been used to estimate the lithospheric structure along several transects passing through the Western and Eastern Carpathians and the Pannonian Basin (*Zeyen et al., 2002; Dérerová et al., 2006; Hlavňová et al., 2015*). In this paper we present a lithospheric structure model along the Carpathian-Pannonian–European platform transect HT-1 that passes through the High Tatras mountains.

Transect HT-1 (Fig. 1) starts in the Drava depression in Croatia, then goes along the eastern margin of the Zala Basin and continues northeast through the Trans-Danubian central range. After leaving this area, the profile enters the easternmost part of the Danube Basin, continues through the Central Slovakia Volcanic Field, the Internal Western Carpathians, the High Tatras, the Pieniny Klippen belt, the External Western Carpathians, the Carpathian Foredeep, the Trans-European Suture Zone (Malopolska Block) and ends in the European Platform. The length of the transect is approximately 760 km.

2. Geological settings

The Drava depression is a part of the southwestern part of the Pannonian Basin system, which had been formed as a back-arc system due to lithospheric extension and mantle upwelling behind the Carpathian arc (e.g. *Csontos et al., 1992; Horváth, 1993; Kováč, 2000*). The Drava depression is filled with Tertiary and Quaternary strata. The thickness of the sedimentary filling varies between 0 and 6 km (*Bielik, 1988; Kilényi and Šefara, 1989; Bielik et al., 2005*).

The Transdanubian Range (*Haas et al., 2001*) consists of the Palaeozoic, Triassic and Mesozoic formations. In some parts the Carboniferous granite forms part of the crystalline basement of this tectonic unit. Its northeastern part is made up of Eocene andesite.

The Central Slovakia Volcanic Field is an extensive volcanic region of the Carpathian arc and the Pannonian Basin. Its origin is related to processes of subduction and back-arc extension during Neogene evolution of the arc (*Vozár et al., 1999*).

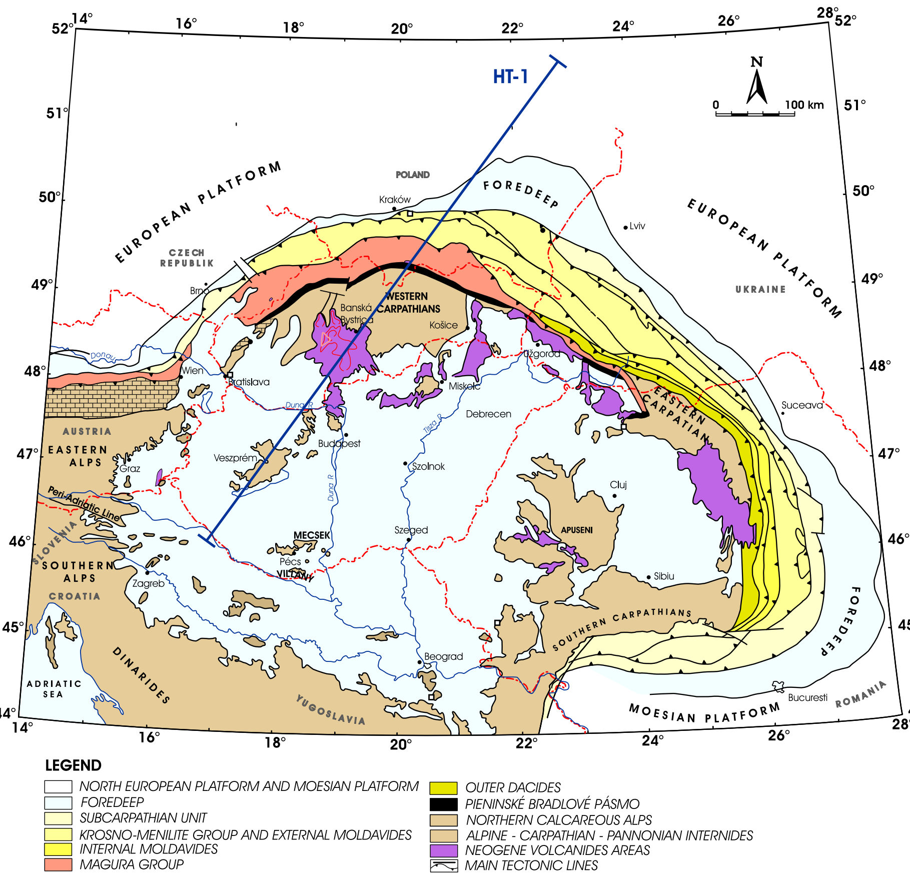

Fig. 1. Location of profile HT-1 on the map of the Carpathian-Pannonian–European platform region.

The Western Carpathians result from a series of Jurassic to Tertiary subduction and collision events during the Alpine orogeny (*McCann, 2008a, 2008b*). Tectonically, the Western Carpathians have a complicated geological structure formed since the Paleozoic era. The oldest Paleozoic rocks experienced the first stage of deformation during the Hercynian orogeny, but younger Alpine overprint is common. Alpine orogeny affected the area in several stages from Jurassic to Neogene. During this period, parts of the Tethys Ocean were subducted under the African plate, while the Western Carpathian blocks (microplate ALCAPA) were thrust over the margin of the Eurasian plate. They are divided into the External and Internal Western Carpathians. The boundary between them is formed by the Pieniny Klippen Belt (PKB). The PKB is a highly compressed narrow zone with layers of Jurassic to Lower Cretaceous limestone "klippen" surrounded by more plastic Cretaceous marl (*Hók et al., 2014; Plašienka et al., 2020*). The Internal Western Carpathians were subject to extensive crustal shortening (*Plašienka et al., 1997*) and include various pre-Tertiary units. The External Western Carpathians include the Carpathian Flysch Belt composed of several north-west, north, and north-east verging nappes, and the Carpathian Foredeep filled by Neogene strata.

The East-European Platform (EEP) formed during Precambrian. It is composed of Proterozoic igneous and metamorphic rocks covered by Vendian and Palaeozoic strata (*Dadlez et al., 2005*). It is separated from the younger Paleozoic Platform to the SW by the Trans-European Suture Zone (TESZ, e.g. *Pharaoh, 1999*). The zone is a broad (up to 200 km) zone, crossing Europe from the North Sea to the Black Sea.

3. Method

2D integrated modelling method is an approach that combines joint interpretation of several geophysical fields, namely gravity anomalies, topographic heights, geoidal heights and surface heat flow data. A detailed description of the method can be found in (*Zeyen and Fernàndez, 1994; Zeyen et al., 2005*). A 2D finite element algorithm calculates the two-dimensional temperature distribution in the lithosphere, given its thickness, defined as the 1300 °C isotherm, and the distribution of heat production and thermal conductivity solving the steady state heat conduction equation:

$$\lambda \nabla^2 T = A \,, \tag{1}$$

where λ is the thermal conductivity ($\mathrm{Wm^{-1}K^{-1}}$), T the temperature (°C) and A the heat production ($\mathrm{Wm^{-3}}$).

Based on temperatures that are calculated at every node, densities are determined at the same nodes, depending on temperature and pressure, as well as on predefined densities at room conditions. In the upper crust, with relatively low temperatures and high porosities, pressure and temperature effects are supposed to balance each other. In the lower crust and lithospheric mantle, the density decrease due to temperature is usually supposed to be stronger than the increase due to pressure except for very low temperature gradients. In our calculations, we assumed a thermal expansion coefficient of $3 \times 10^{-5}\,\mathrm{K}^{-1}$. The obtained density distribution serves for calculation of the gravity (Bouguer or free air) anomalies along the transects (*Talwani et al., 1959*) and, for every column of the model, the topography under the assumption of local isostatic equilibrium based on the formulas given by *Lachenbruch and Morgan (1990)*. The formulas used to calculate geoid have been published by *Zeyen et al. (2005)*.

The joint use of gravity, topography and geoid data enables us to distinguish between density variations at different depths. Shallow (crustal) density variations are better controlled by the gravity data, especially if the crustal structure is known. Density variations in the deeper lithosphere are supposed to be mainly due to temperature variations and have a strong influence on long wavelength component of topographic heights, but relatively little effect on gravity. The geoid, reflecting variations in the undulation of the equipotential surface of the actual potential at sea level, above the equipotential surface of the normal potential (the normal reference ellipsoid) vanishes with inverse distance. Therefore, the geoid, compared to gravity anomalies, which vanish with the inverse square distance, is more sensitive to density variations in the lower lithosphere than in the crust. In addition, to suppress the effect of sub-lithospheric density variations on the geoid, the long-wavelength component of the geoid in terms of spherical harmonics up to degree and order 8 was removed (cf., *Bowin, 1991; Zeyen et al., 2005*).

4. Geophysical data and starting model

4.1. Input geophysical data

The surface heat flow data were compiled from the worldwide data set of *Pollack et al. (1993)*. Topography data were taken from the GTOPO30 database (*Gesch et al., 1999*). The free air gravity anomalies were taken from the TOPEX 1-min gravity data set (`ftp://topex.ucsd.edu/pub` (*Sandwell and Smith, 1997*)). Geoid data were taken from the EGM-2008 global model (*Pavlis et al., 2008*).

Mantle heterogeneities produce long-wavelength, high-amplitude anomalies that dominate the geoid. Consequently the observed geoid anomaly is the combination of mass anomalies in both the lithosphere and the sub-lithospheric mantle. In order to use the geoid for the study of the distribution of masses in the lithosphere, we remove the spectral part of the geoid assumed to be generated by the lower mantle, by means of removing the low degree and order spherical harmonics *Bowin (1991)* and *Chase et al. (2002)* advocate the removal of harmonics up to degrees 7 or 11. For the lithosphere of the Carpathian-Pannonian–European platform region with estimated thickness varying in the range 80–220 km, the removal of spherical harmonics up to degree and order 8 is recommended (*Zeyen et al., 2005; Dérerová et al., 2006*).

For each geophysical dataset, we have extracted several parallel profiles to calculate the lateral variability of the data.

4.2. Creation of starting model

The starting (initial) model iteratively modified in our 2D integrated modelling to fit jointly all the geophysical data was constructed based on previous studies with their interpretations. The LAB for the starting model at transect HT-1 was extrapolated from the map of the lithospheric thickness published by *Dérerová et al. (2006).* The Moho boundary for the starting model was taken from the map published by *Bielik et al. (2018).* The depth of the boundary between upper and lower crust for the starting model was taken from (*Bielik, 1995*). The sedimentary layer for the starting model was constructed based on the data published by *Kilényi and Šefara (1989)*, *Krejčí and Jurová (1997)*, and *Makarenko et al. (2002).*

5. Results

The starting 2D lithospheric model described above was in the manual iterative process of the 2D integrated modelling modified to obtain the best possible fit, jointly, for all modelled input geophysical data. The shape and the depth of the structural density interfaces and the modelled lithospheric bodies were modified wherever needed, along with the thermal and density-related parameters. Since the near-surface structures, the sedimentary layers and the upper crust, are quite well known, the more significant modifications focused on deeper structures like the Moho and the LAB. The model was modified until we reached a reasonable fit between the given input geophysical data and the model predictions (the same data obtained as the output generated by the subsurface structure). The resultant model is presented in Fig. 2 along with the final set of densities and thermal parameters listed in Table 1.

Table 1. Densities and thermal properties of the different bodies used in our modelling along transect HT-1. No.: reference number in Figure 2, HP: heat production (μW m^{-3}), TC: thermal conductivity (W m^{-1} K^{-1}), ρ_0: density at room temperature (kg m^{-3}).

No.	**Unit**	**HP**	**TC**	$\boldsymbol{\rho_0}$
1	Pannonian Basin sediments	3.5	2.5	2500
2	Flysch sediments	1.0	2.0	2610
3	Pannonian upper crust	3.0	2.5	2750
4	Carpathian/European platform upper crust	2.0	2.5	2750
5	Lower crust	0.2	2.0	2950
6	Lower (mantle) lithosphere	0.05	3.4	3200

The main focus of our study is the depth and shape of the LAB. In the Pannonian Basin the modelled depth is about 80–90 km and rapidly increases towards the Western Carpathians, where it reaches values of 145–150 km, and 135–140 km beneath the European platform. We can observe the occurrence of a weakly pronounced lithospheric root under the Western Carpathians. This lithospheric thickening is interpreted as a small remnant of a subducted slab. It has been previously detected by *Spakman et al. (1993)*, *Lillie et al. (1994)*, and *Wortel and Spakman (2000)*. It is also in agreement with our previous models in the Western Carpathians presented by *Zeyen et al. (2002)* and *Dérerová et al. (2006)*.

The Moho depth varies between 25 and 27.5 km beneath the Pannonian Basin and about 37–40 km beneath the Western Carpathians. It reaches about 30–37 km beneath the European platform. Figure 2 indicates that thickening of the Moho boundary underneath the Western Carpathians follows the thickening of the lithosphere in this area. These values are in correlation with our previous modelling (*Zeyen et al., 2002*), but differ somewhat from the data taken from the map published by *Bielik et al. (2018)*, which served as input data for our initial model. The main difference was observed

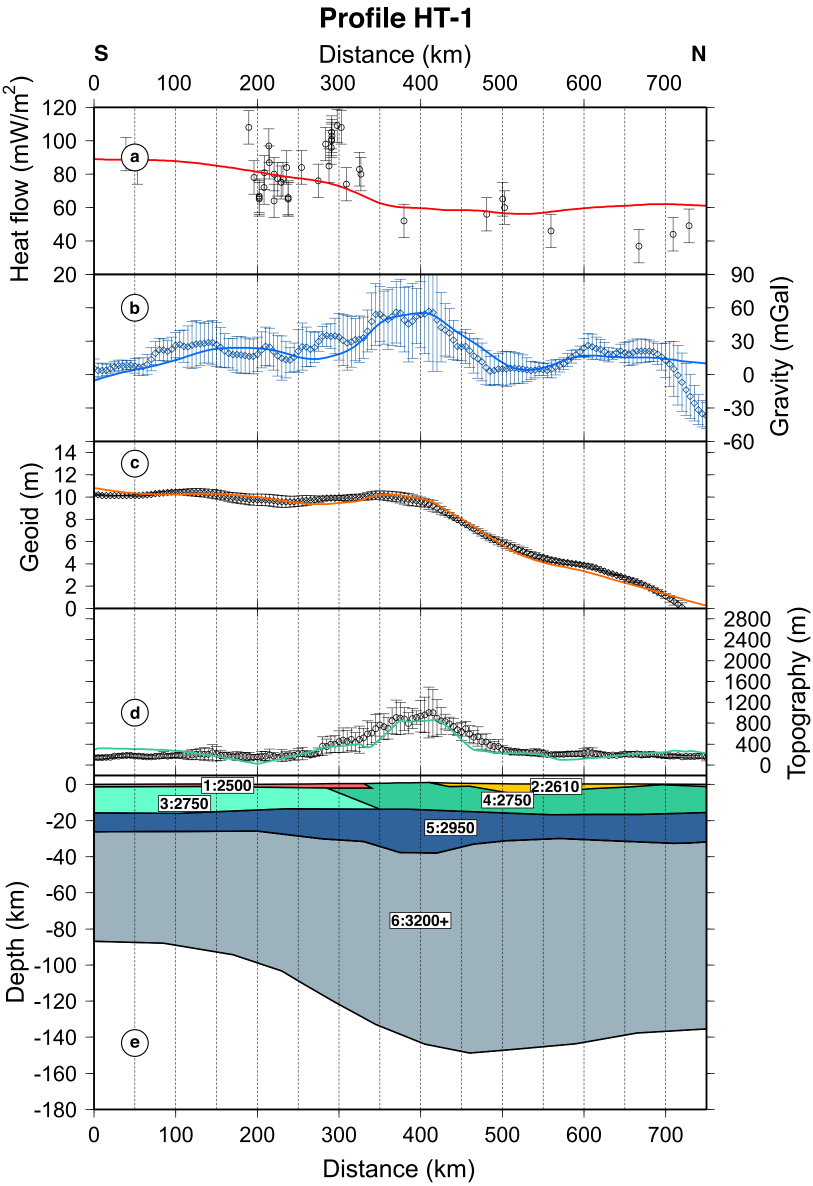

Fig. 2. Lithospheric model along transect HT-1. (a) Surface heat flow, (b) free air gravity anomaly, (c) geoid, (d) topography with dots corresponding to measured data with uncertainty bars and solid lines to calculated values. Numbers in (e) correspond to material number in Table 1.

along our profile beneath the European platform, where *Bielik et al. (2018)* suggest values up to 45 km, while we modelled significantly lower values with a maximum around 37 km.

The depth of the boundary between upper and lower crust changed minimally and varies between 17 and 20 km, which is in correlation with data published by *Bielik (1995)*.

In order to fit the surface heat flow data, the upper crust had to be divided into two units that differ in their thermal parameters, while the densities remain the same (Table 1). Although the heat flow data show a high degree of scatter, from Fig. 2 it is obvious that the Pannonian Basin is hotter and with much higher values of surface heat flow. To fit this feature, the thermal parameters for the upper crust had to be modified accordingly.

The sedimentary layer changed minimally, only certain corrections regarding the depth of flysch sediments were made.

6. Conclusions

We have used the available geological data to construct the initial model, and the available geophysical data to perform joint 2D integrated modelling that combines surface heat flow data, topographic heights, gravity anomalies and short-wavelength (stripped of degree and order 8 components) geoidal heights data. We focused mainly on the deeper parts of the lithosphere, the Moho boundary and the lithosphere-asthenosphere boundary (LAB). Our final model shows that the lithospheric thickness increases from its minimum 80–90 km beneath the Pannonian Basin up to 145–150 km in the Western Carpathians, and 135–140 km beneath the European platform. The maximum values and the shape of the LAB beneath the Western Carpathians suggest the formation of a lithospheric root. The lithospheric thickening is accompanied by crustal thickening (Moho) beneath the Western Carpathians.

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