Crustal and upper mantle structure of the Western Carpathians from CELEBRATION 2000 profiles CEL01 and CEL04: seismic models and geological implications

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SUMMARY
CELEBRATION 2000 was a large international experiment that focused on the lithospheric structure in Central Europe. It consisted of several wide-angle reflection and refraction seismic profiles. Profiles CEL01 and CEL04 are located in the transition from the Precambrian and Palaeozoic platforms to the Carpathian orogen. The data were modelled using 2-D tomographic and forward modelling techniques. The profile CEL01 (about 900 km long) extends from the southwestern margin of the East European Craton to the Pannonian basin system. The depth of Moho boundary varies along the profile from 27–33 km under the Pannonian basin system area, 30–39 under the Carpathians and within the TESZ to 40–45 km under the East European Craton. The CEL04 profile (630 km long) starts in the Polish trough (PT) and reaches the Pannonian basin system. Along the CEL04 profile, the Moho interface shallows from 40 km beneath the PT to about 35 km in the Maopolska unit. The crustal thickness is 43 km beneath the Carpathians and 25 km beneath the Pannonian basin system. The results obtained are consistent with a commonly accepted view of the Carpathians, thought to be formed largely due to the interplay between Palaeogene to Middle Miocene subduction of oceanic or suboceanic crust and subsequent collision. A feature that we interpret as the subduction-related orogenic root is at least partially preserved in both profiles and is manifested by a crustal thicknesses of 39 km (CEL01) and 43 km (CEL04). The boundary between the Malopolska unit and the Bruno–Silesian unit is discernible north of the Carpathian frontal thrust along the CEL01 profile, but is not observed farther to the southeast (CEL04) beneath the Carpathians. In the Pannonian basin system, no substantial differences in the crustal structure across the Mid-Hungarian Line were observed. Upper mantle reflections observed along both profiles originate from a north-dipping mantle discontinuity, probably representing a shear zone related to collision and possibly ongoing convergence between the European plate and ALCAPA (Alps-Carpathians–Pannonian).

Key words: Carpathians, crustal structure, explosion seismology, orogeny, seismic modelling, subduction.

INTRODUCTION
A consortium of 28 institutions from Europe and North America, undertook a large active source seismic experiment focused on Central Europe in 2000. This experiment, called CELEBRATION 2000 (Central European Lithospheric Experiment Based on Refraction, 2000), targeted the structure and evolution of the complex area including Trans-European Suture Zone (TESZ), the southwestern portion of the East European Craton (EEC), the Carpathians, the Pannonian basin system (PBS) and the Bohemian massif (Guterch et al. 2001, 2003). The experiment was carried out within the framework of the EUROPROBE Program—TESZ, EUROBRIDGE and
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Figure 1. (a) Geological map and location of the CEL01 and CEL04 seismic profiles. Carpato–Pannonian area after Kovač et al. (1994), European platform after Belka et al. (2002) and Verniers et al. (2002, modified), Mid-Hungarian Line after Fodor et al. (1999), BCH—Branisko-Cierna Hora Mts., BSU—Bruno–Silesian Unit, CSVF—Central Slovak Volcanic Field, HCM—Holy Cross Mts., KD—Kosice depression, LB—Levoča basin, SR Mts.—Slovenske Rudohorie Mts., TD—Túrka depression, TESZ—Trans-European Suture Zone. (b) Location of shot points along CEL01 and CEL04 profiles as well as positions of other seismic profiles discussed in the text.

PANCARDI projects. The layout of the experiment was a network of interlocking wide-angle reflection and refraction seismic profiles. In this paper, we present the results of seismic modelling based on data from profiles CEL01 and CEL04, crossing each other near the northern boundary of the Carpathian Foredeep in Poland, and located in the transition from the Precambrian and Palaeozoic platforms to the Mesozoic–Cainozoic Carpathian orogen (Fig. 1). The northeast–southwest trending profile CEL01 (about 900 km long) starts at the EEC; crosses the Lysogóry unit (LU) and the Malopolska unit (MU); the Bruno–Silesian unit (BSU); the Outer and Inner Carpathians, separated by the Pieniny Klippen Belt (PKB), and ends in the area of the PBS, where it crosses the Danube basin and passes close to the Transdanubian Range (TR). The CEL04 profile (630 km long) trends NW–SE, starts in the Polish trough (PT) and crosses the LU, MU, BSU, ALCAPA (Alps-Carpathians–Pannonian) unit and Tisza unit (TU). The EEC is composed of Proterozoic magmatic and metamorphic rocks covered by Vendian and Palaeozoic strata. The LU, MU and BSU are composed of Proterozoic and Palaeozoic rocks overlain by upper Palaeozoic to Mesozoic strata.

The Western Carpathians belong largely to the ALCAPA unit (Fig. 1b) (Fodor et al. 1999) which reaches the BSU in the north and the Tisza unit in the south. The last contact follows the Mid-Hungarian Line, which is a Tertiary strike-slip fault zone (Plašienka et al. 1997, and references therein). The architecture and origin of the contact between the ALCAPA unit and the BSU are the main topics of the present paper and will be discussed in detail. The Western Carpathians comprise, from the north to the south: Carpathian Foredeep, Outer Carpathians, PKB and Inner Carpathians. The Inner Carpathians and the Tisza unit are partly covered by strata filling the PBS. The Western Carpathians have been formed since the Late Jurassic (Plašienka et al. 1997) through Quaternary (Tokarski & Święczewska 2005) times largely due to the convergence of European and African plates. The convergence resulted in diachronous collisions of the plates as well as smaller intervening continental blocks (e.g. Dewey et al. 1973).

The Inner Carpathians comprise of numerous nappes composed of low- to high-grade metamorphic and plutonic Palaeozoic rocks and largely unmetamorphosed Palaeozoic to Cretaceous strata (cf. Biely et al. 1996; Plašienka et al. 1997, and references therein). The nappes are largely north-verging—except in the southernmost part of the Inner Carpathians, where the structural polarity is to the south. They were formed during Late Jurassic through Late Cretaceous times.

GEOLoGICAL SETTING

The CEL01 and CEL04 seismic lines cross East European Craton, LU, MU, BSU, ALCAPA (Alps-Carpathians–Pannonian) unit and Tisza unit (TU). The EEC is composed of Proterozoic magmatic and metamorphic rocks covered by Vendian and Palaeozoic strata. The LU, MU and BSU are composed of Proterozoic and Palaeozoic rocks overlain by upper Palaeozoic to Mesozoic strata.

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Subsequently, the Inner Carpathians were covered by deposits of the Late Cretaceous, Palaeogene and Early Neogene intermontane basins, filled largely by the flysch strata up to 3400 m thick (Janoško & Jacko 1999; Soták et al. 2001, and references therein).

The Outer Carpathians are composed of several north-verging nappes consisting largely of the Lower Cretaceous to Lower Miocene flysch and formed during Early Palaeogene (Świerczewska & Tokarski 1998; cf. Zytko 1985) through Late Miocene (Zuchiewicz et al. 2002, and references therein) times. To the north, the Outer Carpathians are thrust over the Carpathian Foredeep filled by Neogene strata. To the south, the Outer Carpathians are separated from the Inner Carpathians by the PKB, which is a narrow zone of extreme shortening and wrenching (Birkennäger 1986). The PKB was folded twice, during Late Cretaceous and during Tertiary, thus it represents the innermost part of the Outer Carpathians and the outermost part of the Inner Carpathians.

There is a general agreement that the main tectonic features of the Outer Carpathians and the Carpathian Foredeep were largely formed during Tertiary times. This is due to the southward-directed subduction of the oceanic or suboceanic crust, intervening between the continental crust of the southwest margin of the European plate (the BSU) (descending plate) and the continental crust of the ALCAPA unit (hanging-wall plate) and their subsequent collision (for review see: Konečný et al. 2002). According to this interpretation, during Tertiary times, the Carpathian Foredeep was a peripheral foredeep formed due to regional flexure of the descending plate (Krzywicz 1997), whereas the Outer Carpathians were an accretionary prism with back-stop located at the PKB. At the same times the PBS was formed (Horváth 1993; Tari et al. 1993) within the Inner Carpathians and the Tisza unit, due to backarc stretching and mantle upwelling (Konečný et al. 2002). The PBS is filled by more than 2 km of Palaeogene and up to 7 km of Neogene and Quaternary strata (Royerden et al. 1983).

Products of subduction-related volcanism and mantle upwelling related volcanism penetrate and cover the Inner Carpathians in places, the PKB and the innermost part of the Outer Carpathians (Szábo et al. 1992).

The subduction-related nappe stacking in the Outer Carpathians was followed by regional collapse (Zuchiewicz et al. 2002, and references therein) resulting in the formation of intermontane basins filled by Neogene and Quaternary strata. During Quaternary to Recent times, the stress arrangement in theOuter Carpathians has been largely compressional with \( \sigma_1 \) oriented roughly perpendicular to the belt (Zuchiewicz et al. 2002, and references therein).

**PREVIOUS GEOPHYSICAL INVESTIGATIONS**

Crustal and upper mantle structure of the Carpathians was the subject of many studies. Their results, based on seismic, gravity and heat flow data, are summarized below.

Bouguer anomalies in the region are dominated by a linear minima (up to \(-60\) mGal) that extend from the eastern Alps through the Vienna basin along the Carpathians and their northern foreland. The zone of negative gravity anomalies includes: Carpathian Foredeep, Outer Carpathians, PKB and northern part of the Inner Carpathians. The belt of positive anomalies of \(20–30\) mGal is located mainly in the southern and eastern territory of Slovakia (Bielik 1999; Bielik et al. 2004).

The large-scale feature of heat flow data is an increase from the northernmost units (EEC, MU) towards the PBS. North of the Carpathians, the heat flow value is in the range of \(40–60\) mW m\(^{-2}\). In the Carpathians, it increases to about \(70\) mW m\(^{-2}\), except of the Central Slovak Volcanic Field, where a local maximum occurs. The highest heat flow (over \(110\) mW m\(^{-2}\)) is observed in the area of East Slovakian basin (Zeyen et al. 2002; Bielik et al. 2004).

The first seismic results from the Polish and Slovak Western Carpathians were published by Uchman (1975). In this study, based on 1-D seismic modelling, crustal thickness along the N–S-trending international profile V (Fig. 1b) was determined to be 40 km in Poland outside the Carpathians. Beneath the Outer Carpathians north of the PKB a 70-km-wide and 50-km-thick crustal root was observed. Further to the south, crust beneath the Inner Carpathians and the PBS is 35–27 km thick (Bielik et al. 2004).

Tomek (1993) described the seismic reflection results along the 170 km long NW-NW SSE trending profile 2T running from the Outer Carpathians through the PKB to the Inner Carpathians. The dominant features of this data are mostly south-dipping bands of reflectors in the upper to lower crust. In Outer Carpathians, they are interpreted as an image of Tertiary southward subduction of the European plate (BSU). Reflections from the Moho boundary were found 37 km deep in the Outer Carpathians part, and 32 km deep in the Inner Carpathians. No crustal root was found along this profile.

Interpretations of data from several deep reflection profiles across the Slovak Outer and Inner Carpathians (3T, 6HR, 2T and G1-2, Fig. 1b), as well as gravity models along coinciding lines, were described by Vožár et al. (1998) and Vožár et al. (1999).

A 3-D model of the crustal structure in the Western Carpathians based on gravity and seismic modelling was presented by Bojdys & Lemberger (1986). In the region corresponding to location of CEL01 and CEL04 profiles, the depth of the crystalline basement is maximal (>10 km) near the PKB and decreases to the north to a few kilometres. Beneath the Inner Carpathians, a 48–54-km-thick crustal root has been modelled, while to the north and south the crust is about 38 km and 27 km thick, respectively. The authors interpret crustal thickening as a result of south-verging subduction of oceanic part of the European plate and following collision.

Geophysical models of the lithosphere for the Slovak territory were compiled by Bielik (1995, 1999). Crustal thickness is largest near the boundary between Inner and Outer Carpathians, reaching 36 km, and decreases to 25–30 km towards the PBS. Lithospheric thickness decreases from over 150 km in the Polish Outer Carpathians and 120 km under the PKB to about 60 km beneath the PBS. The models show the existence of a 110–130 km thick lithospheric root beneath the Carpathians along the transects 2T and KP-X (Fig. 1b). Zeyen et al. (2002), Bielik et al. (2004) and Szabó et al. (2004) performed integrated modelling of the Western Carpathian lithosphere based on gravity, topography, xenolith, and heat flow data. The reported lithospheric thicknesses range between 115–140 km beneath the MU, 90–120 in the area of the Bohemian massif and 75–100 km beneath the PBS. A lithospheric root (130–150 km) beneath the Western Carpathians was interpreted as a remnant of the subducted slab of the European plate.

**DATA ACQUISITION AND PROCESSING**

During the field work, seismic energy was generated using chemical explosions of 100–1200 kg charges at 22 shot points along the profile CEL01 and at 18 shot points along the profile CEL04 (Guterch et al. 2003). The average distance between the shots was 30 km and the station spacing was 2–5 km. Shot timing and positioning of receivers was done using GPS. Seismic data were recorded mainly...
by one-component stations REFTEK-125 (TEXAN) and sampled at 10 ms. The station sensors were 4.5 Hz vertical component geophones. Data processing included shot time corrections (if any) and bandpass filtering (usually 2–15 Hz) of the whole data set in order to remove low- and high-frequency noise. The frequency content of the seismic data was variable for different shot points, probably due to the varying local environment. Thus, the filter window was determined interactively during data interpretation. Recordings were sorted into shot gathers; seismic sections were trace-normalized to the maximum amplitude along the trace and reduction velocity of 6 or 8 km s$^{-1}$ was applied. In several sections, strong surface waves have been muted to enhance $P_{\text{sed}}$ and $P_g$ arrivals.

SEISMIC DATA

Based on acquired seismic data, several $P$-wave phases were correlated and used for modelling. First arrivals include refractions from the upper crustal sedimentary layers ($P_{\text{sed}}$), upper/middle crystalline crust ($P_g$), and refractions from the upper mantle ($P_P$). The first arrivals can usually be correlated up to 250–300 km offsets.

In later arrivals, the strongest phase represents reflections from the Moho discontinuity ($P_{P\text{mP}}$). Reflections from mid-crustal discontinuities ($P_{1\text{P}}$) and from the top of the lower crust ($P_{2\text{P}}$) are also observed. At large offsets, a reflection from upper mantle discontinuities ($P_{\text{gP}}$) can be identified for several shot points.

Crustal phases

Data from shots located in the EEC show clear and continuous $P_g$ phase with apparent velocities of 5.8–6.2 km s$^{-1}$ (Fig. 2). In the marginal area of the EEC, the apparent velocity of $P_g$ is unusually high (6.8–7.2 km s$^{-1}$), reflecting anomalously high $P_g$ velocity in the middle crust. The $P_{\text{sed}}$ phase occurs in short offset intervals near the shot locations, due to a relatively thin sedimentary cover. In most of the sections, the $P_g$ phase is visible in the offset range of ~200–350 km. Reflections from the Moho are of high amplitude, but usually the onsets are not sharp. Their critical point occurs at 130–150 km offset, suggesting a large crustal thickness. Intracrustal reflections are only observed in some sections, as they are mostly obscured by long $P_g$ coda.

In several seismic sections, outside the EEC (MU, LU, BSU, ALCAPA, Tisza), first arrivals comprise a prominent $P_{\text{sed}}$ phase, originating in sedimentary layers of considerable thickness. It can be distinguished from the $P_g$ phase due to lower apparent velocity — 4.5–5.5 km s$^{-1}$ in the LU and MU, and 2.5–5.5 km s$^{-1}$ in the ALCAPA and Tisza units.

A characteristic feature of both phases, when compared to the EEC, is a rapid increase of their apparent velocity with offset, caused by a vertical velocity gradient due to lithology changes and of degree of consolidation with depth, as well as fluctuations of the apparent velocity due to a laterally inhomogeneous upper crustal structure.

The $P_{\text{sed}}$ and $P_g$ phases are usually of good quality and can be observed up to offsets of 100–150 km except for the Inner Carpathians, where a seismic signal, also including later arrivals, is not recorded at offsets exceeding 30–70 km (Fig. 4). Poor propagation of seismic energy in this area may be due to attenuation of the seismic waves by
discontinuities ($P_{\text{1P}}$) and from the top of the lower crust ($P_{\text{2P}}$) are also observed. At large offsets, a reflection from upper mantle discontinuities ($P_{\text{gP}}$) can be identified for several shot points.

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Additionally, there is a significant difference in the apparent velocity of the $P_g$ phase south and north of the PKB (Fig. 7), which suggests a strong contrast in the upper crustal composition of the Inner and Outer Carpathian units.

In the area of the MU and LU, Moho reflections are of high amplitude. Intracrustal reflected phases are also observed, but usually they cannot be traced consistently in more than one section, which makes their interpretation ambiguous. Particularly clear intracrustal reflections are observed for shot 21150, CEL01, in the offset range of 50–100 km. They are thought to originate from the bottom of the thick upper crustal metasedimentary layer (Fig. 8). The $P_g$ phase of variable quality can be observed with apparent velocity 8.0–8.2 km s$^{-1}$. An example of a strong $P_g$ arrival is shown in Fig. 9 (shotpoint 21180, CEL01).

The data from the Carpathians (in particular from the Inner Carpathians) are largely characterized by at best weak intracrustal reflected arrivals (Fig. 5), and the $P_{P\text{mP}}$ phase quality is variable and often hard to correlate. Similarly, $P_g$ phase correlation was problematic. Nevertheless, the determination of the deep crustal structure beneath this area was made possible using the recordings performed in the Carpathians but originating from shots outside the Carpathians, where deep crustal and mantle phases were of better quality.

Very good quality data were recorded in the area of the PBS (Tisza unit and partially ALCAPA unit) at the southern end of the profile CEL04 (SP 21440–21460, Fig. 6). In the first arrivals, the $P_{\text{sed}}$ phase with relatively low apparent velocity (2.5–4.5 km s$^{-1}$) can be correlated and represents reflection in Palaeogene to Quaternary infill of the PBS. At larger offsets, refraction from crystalline basement ($P_g$) shows apparent velocity of 5.9–6.1 km s$^{-1}$. Very clear $P_g$ phases are visible as first arrivals starting at offsets of only 100 km, proving the existence of an unusually thin crust in this area. Apparent velocity of the $P_g$ phase (7.7–7.8 km s$^{-1}$) suggests unusually low $V_p$ velocity of the ALCAPA upper mantle (Figs 6, 11 and 12b). Moho reflections are observed with critical point at small offsets of 60–70 km, due to the thin crust. Clear intracrustal reflections in this area are mostly interpreted as originating from the top of the lower crust.

The $P_g$ phase at the SW end of the CEL01 profile (SP 21010–21030, e.g. Fig. 3a, offsets ~40 to 0 km) has relatively high apparent velocity (6.2–6.3 km s$^{-1}$), but terminates abruptly at ~40–50 km offset. This may be due to the presence of slower material beneath the uppermost higher velocity rocks, forming a local low-velocity layer.

Reflections from the upper mantle discontinuities

An outstanding feature of the seismic data, from both CEL01 and CEL04 profiles, is a set of high-amplitude reflections from upper mantle discontinuities (Figs 2 and 3b). The reflections originate from several shot points located north from the Carpathians, in MU and EEC. However, independently of the shot location they are consistently observed in roughly the same area, namely in northernmost part of the Inner Carpathians at 270–400 km distance along profile CEL01 (Figs 9 and 10) and 240–340 km along profile CEL04 (Fig. 13). Therefore, they occur in a wide range of offsets, ~300 to 600 km. Confinement of this phase to a particular geographic location, and its high apparent velocity (~9.0–9.2 km s$^{-1}$), suggests that it originates from a N-dipping mantle discontinuity, located in a
Figure 2. Seismic sections from the profile CEL01, shot points 21190 and 21230 with the labels of the main seismic phases. Reduction velocity is 8.0 km s$^{-1}$. Bandpass filter 2–15 Hz had been used. Data were normalized to maximum trace amplitude. Note clear $P_g$ wave within EEC (NE part of sections), and far offset recordings of $P_1P$ wave. Strong surface waves have been muted to enhance $P_{sed}$ and $P_g$ arrivals. CLF—Cracow–Lubliniec Fault, HCF—Holy Cross Fault, PKB—Pieniny Klippen Belt, TTL—Teisseyre–Tornquist Line.
Figure 3. Seismic sections from the profile CEL01, shot points 21030 (recordings from PB and Carpathians) and 21130 (recordings from Carpathians, MU and EEC). Reduction velocity is 8.0 km s\(^{-1}\). Bandpass filter 2–15 Hz had been used. Data were normalized to maximum trace amplitude. CLF—Cracow–Lubliniec Fault, HCF—Holy Cross Fault, PKB—Pieniny Klippen Belt, TTL—Teisseyre-Tornquist Line.

Figure 4. Examples of seismic data from the profile CEL01, shot points 21090 and 21100, illustrating decrease of the seismic signal at the Pieniny Klippen Belt. Reduction velocity is 8.0 km s\(^{-1}\). Bandpass filter 2–15 Hz had been used. Data were normalized to maximum trace amplitude. PKB—Pieniny Klippen Belt.
limited area below the Outer and Inner Carpathians. Other geometries (i.e. long subhorizontal discontinuity) would produce reflections in approximately the same offset range, but in different locations along the profiles.

Another set of mantle reflections is observed at the SE end of profile CEL04 for shots located in the Tisza unit (Fig. 6b), suggesting existence of another, and smaller south-dipping discontinuity beneath the Inner Carpathians. Also, a subhorizontal reflection from beneath the PBS was tentatively interpreted in the southeastern part of the CEL04 sections, although the quality of the corresponding phase is quite low.

**SEISMIC MODELLING METHODS**

The interpretation of the seismic data consisted of two steps. First we applied tomographic inversion of first arrivals in order to obtain generalized models of the crustal structure based on most objective part of the data, first arrivals. For the next step, trial and error forward modelling was performed, using all available P-wave phases.

Tomographic inversion by Hole (1992) uses the first arrival travel-times for determining a smooth 2-D or 3-D, P-wave velocity model. The procedure uses the backprojection algorithm (Humphreys & Clayton 1988), based on the linearization of the relation between...
the traveltimes and the slowness. The model is parametrized in an equidistant rectangular grid and the $P$-wave velocities are defined at the grid nodes. In the forward step, the traveltimes and residuals are calculated using a finite difference algorithm for solving the eikonal equation (Vidale 1990). In the inverse step, the slowness perturbations are calculated by uniformly distributing the traveltime residual along a ray. The perturbations are then summed up for all rays, smoothed and added to the original model. The procedure is repeated iteratively until a model with satisfactory traveltim e residuals is obtained.

The data for inversion consisted of 1410 $P_{sed}$, $P_g$ and $P_n$ traveltime values from the profile CEL01 and 984 values from the profile CEL04, shot points 24160 and 24150. Reduction velocity is $8.0 \text{ km s}^{-1}$. Bandpass filter 2–15 Hz had been used. Data were normalized to maximum trace amplitude. CLF—Cracow–Lubliniec Fault, MHL—Mid-Hungarian Line, PKB—Pieniny Klippen Belt.
CEL04. The initial 1-D model for the upper crust was calculated by 1-D inversion of an average traveltime curve. The 2-D model grid spacing was 1 x 1 km. The computation consisted of several steps. In subsequent steps, data with increasing offset were involved in order to gradually enlarge the depth of ray penetration. At each step, several iterations were made with decreasing size of the smoothing operator to gradually increase the resolution of the model and to stabilize the inversion result. The final RMS traveltime residual amounted to 0.10 s for the CEL01 model and 0.12 s for the CEL04 model. These values are close to our average picking error. The ray coverage, traveltimes and residuals for both models are shown in Fig. 14. Similar diagrams of residuals and of ray density for the forward models are shown in Fig. 15. The final tomographic models are presented in Figs 16 and 17.

The 2-D tomographic model was further refined by trial-and-error forward modelling using the SEIS83 program package (Červený & Pšenčík 1983) with the graphical interface MODEL (Komminaho 1997). In this approach, we used all refracted and reflected P-wave phases. This provided more constraints on the velocity field and on the location of discontinuities. The initial velocity model was created based on obtained tomographic result; additionally constraints provided by borehole data and previous shallow seismic studies were used, when available, for assessment of the velocity distribution in the uppermost crust.

The SEIS83 algorithm calculates ray paths, traveltimes and synthetic seismograms (e.g. Fig. 9) of P and S waves in the high-frequency approximation. The model consists of layers separated by velocity discontinuities. In each layer, the P-wave velocity is parametrized in an irregular rectangular grid and interpolated by bicubic splines. The solution was sought iteratively as the traveltimes were calculated for the current \( V_p \) model and compared with the observed traveltimes. Then the \( V_p \) model was changed in order to minimize the misfit. The modelling also involved the calculation of synthetic sections and qualitative comparison of amplitudes of synthetic and observed seismograms. As relative amplitudes of seismic waves depend on velocity gradients and velocity contrasts at discontinuities, analysis of synthetic seismograms provides additional constraint on the velocity distribution.

**SEISMIC MODELS**

The forward modelling results and tomographic inversion models along the profiles CEL01 and CEL04 are presented in Figs 16 and 17. Tomographic and forward models are largely similar in terms of overall \( V_p \) distribution. Nevertheless it should be noted that tomographic models, due to the nature of the inversion algorithm,
Figure 9. Example of forward modeling along the profile CEL01, shot point 21180—region of Carpathians and MU. Top: synthetic seismograms, middle: seismic data with theoretical traveltimes, bottom: model with ray paths. Reduction velocity is 8.0 km s$^{-1}$. Note good quality of the $P_1P$ phase constraining the upper mantle reflector. CLF—Cracow–Lubliniec Fault, HCF—Holy Cross Fault, PKB—Pieniny Klippen Belt.

represent a generalized and smoothed image of the earth’s crust structure. Moreover, large fragments of tomographic models are unconstrained due to the lack of sufficiently good quality first arrivals in some regions (e.g. middle and lower crust in the area of the PBS along the CEL01 profile and in the Outer Carpathians along the CEL04 profile). In the forward models, data coverage is better as the reflected phases, which sample otherwise unconstrained regions of the crust, were also included in modelling. Thanks to the large data set used, as well as suitable model parametrization, forward modelling allows the delineation of tectonically significant features such as the Moho discontinuity, crustal boundaries or shear zones. In the tomographic model, due to the nature of the algorithm applied, the smoothing performed during the inversion and model parametrization, the velocity discontinuities are manifested as broad gradient zones and thus features such as the Moho boundary cannot be located with the precision justified by the total data set. In spite
of these differences, the results of both methods reveal a consistent image of the main crustal features and they therefore will be discussed jointly.

**CEL01 profile**

**East European Craton**

Along the CEL01 profile (Fig. 16), the EEC is covered by a thin (<2 km), undeformed layer of sediments, except at the Lublin trough, where sedimentary rocks with $V_p = 4.0–5.5$ km s$^{-1}$ reach approximately 10 km depth. The crystalline crust of the EEC consists of three laterally homogeneous layers, with total thickness of 43–45 km. The $V_p$ in the upper, middle and lower crustal layers is about 6.0, 6.4 and 6.7–7.0 km s$^{-1}$, respectively. The EEC upper mantle is characterized by $V_p$ of about 8.05 km s$^{-1}$ (see also Grad et al. 2003). Beneath the Lublin trough, unusually high velocities of 7.1 km s$^{-1}$ are observed in the upper and middle crust, starting from a depth of 16 km. This anomaly seems to be a continuation of a high velocity/high density intrusion located further SE, found by previous seismic and gravity modelling (Betlej et al. 1967; Perchu´c 1984; Guterch et al. 1986; Grabowska & Bojdys 2001).

**Lysogóry and Małopolska units**

Beyond the EEC margin, in the area of the MU and LU, the upper/middle crust consists of a thick layer, up to 20 km, with relatively low $V_p$ of 5.5–5.9 km and high vertical $V_p$ gradient. In the
Figure 11. Example of forward modelling in the area of Pannonian Basin and Carpathians—profile CEL04, shot point 24170. Top: synthetic seismograms, middle: seismic data with theoretical traveltimes, bottom: model with ray paths. Reduction velocity is 8.0 km s$^{-1}$. CLF—Cracow–Lubliniec Fault, MHL—Mid-Hungarian Line, PKB—Pieniny Klippen Belt.
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Figure 12. Examples of seismic data along the profile CEL04, shot points 24010 (recorded in MU) and 24140 (recorded in PB and Carpathians) with theoretical traveltimes from forward modelling. Reduction velocity is 8.0 km s\(^{-1}\). CLF—Cracow–Lubliniec Fault, MHL—Mid-Hungarian Line, PKB—Pieniny Klippen Belt.

In the MU and LU area, the lower crust with \(V_p\) of 6.8 km s\(^{-1}\) seems to be only 6–8 km thick. In the former unit, crustal thickness changes over a distance of 50 km: from 44 km NE from the Holy Cross Fault to about 32 km SW from it.

Bruno–Silesian and ALCAPA units

The uppermost crust in the area of the Outer Carpathians and the Carpathian Foredeep comprises an asymmetric depression filled by low-velocity rock complexes (velocities 4.5 and 5.6 km s\(^{-1}\)). The low-velocity material forms a south-dipping layer, delimited in the north by the outer limit of the Carpathian Foredeep and from the south by the PKB, where it reaches maximal thickness of about 10 km. Beneath the Carpathian Foredeep, at an approximately 480 km distance, the subvertical contact between low \(V_p\) of the MU and higher \(V_p\) (6.1 km s\(^{-1}\)) basement to the SW, probably belonging to the BSU, may represent a continuation of the Cracow–Lubliniec Fault (CLF). In the ALCAPA unit, in the uppermost crust of the Inner Carpathians, a 3–5-km-thick layer with \(V_p\) values of 5.3–5.6 km s\(^{-1}\) was found. The crystalline basement is characterized by the \(V_p\) velocity of 6.1 km s\(^{-1}\). The information about the structure of the Outer Carpathian crystalline crust is relatively poor due to low signal-to-noise ratio in this area and scarce reflected phases, therefore no \(V_p\) anomalies are discernible. For the same reason, the \(V_p\) velocity in the lower crust beneath BSU was only approximately determined to be about 6.8 km s\(^{-1}\). In the area of the PBS, two areas of low velocities (\(\sim 3\) km s\(^{-1}\)) occur at depths down to 3–5 km, at distances 0–50 km and 150–220 km, corresponding to the location of the Palaeogene to Quaternary Zala and Danube basins, respectively. Between them, the uppermost crust with \(V_p\) of 6.2–6.3 km s\(^{-1}\) overlies a layer with lower velocities (5.8–6.0 km s\(^{-1}\)) forming a velocity inversion. The middle and lower crust are characterized by \(V_p\) values of 6.3 and 6.55 km s\(^{-1}\), respectively. The Moho discontinuity under the Carpathians was found at a depth of 30–39 km, with maximum beneath the PKB, and beneath the Pannonian basin at 27–30 km depth. The \(V_p\) of the upper mantle is lowest in the PBS and Inner Carpathians area (7.9 km s\(^{-1}\)), while further north beneath the Outer Carpathians, it is higher (8.05–8.1 km s\(^{-1}\)). In the upper mantle beneath the Inner and Outer Carpathians, a north-dipping reflector with high impedance contrast was found. It is based on the interpretation of the strong \(P_1\) reflected phase, observed for several shot points, therefore it is a well determined structure. To the south its depth is 45 km, while in the north it reaches 70 km.
Figure 13. Example of forward modelling along the profile CEL04, shot point 21130 (region of Carpathians and MU). Top: synthetic seismograms, middle: seismic data with theoretical travel times, bottom: model with ray paths. Reduction velocity is 8.0 km s$^{-1}$. CLF—Cracow–Lubliniec Fault, MHL—Mid-Hungarian Line, PKB—Pieniny Klippen Belt.

**CEL04 profile**

*Polish trough and Małopolska unit*

The uppermost crust in the area of the PT along profile CEL04 (Fig. 17) consists of a 1–2 km thick sedimentary layer with $V_p$ velocities of 3.0–3.5 km s$^{-1}$ and 3–6-km-thick layer with $V_p$ of 4.0–5.0 km s$^{-1}$. $V_p$ in the upper crust of the PT (5.6–5.7 km s$^{-1}$) seems to be lower than in neighbouring MU, where it amounts to 5.8–5.9 km s$^{-1}$. The lower crust in the PT area at the NW end of the profile contains a layer with elevated $V_p$ (7.0 km s$^{-1}$). Beneath the MU, the lower crustal velocity is 6.6–6.8 km s$^{-1}$. The upper mantle at the NW end of profile CEL04 shows elevated velocities (8.2–8.4 km s$^{-1}$). High mantle and lower crustal velocity was proposed mainly based on the seismic results from the profile CEL03 (Janik et al. 2005), crossing the CEL04 profile. CEL04 data did not allow modelling of these features reliably. The crustal thickness beneath this part of the profile changes from 35–38 km in the area of the PT to 30–32 beneath the MU, where upper mantle $V_p$ is 8.0–8.1 km s$^{-1}$.

*Bruno–Silesian and ALCAPA units*

The uppermost crust in the Outer Carpathians and the Carpathian Foredeep comprises a depression filled by low-velocity sedimentary rocks (4.0–5.3 km s$^{-1}$) with thicknesses varying from 0 km at the outer limit of the Carpathian Foredeep to 8 km. Below this depression, in the upper part of the crust, there occurs south-dipping (8$^\circ$) discontinuity. It continues to SE as a boundary that separates higher-velocity rocks (5.65 km s$^{-1}$) from underlying lower-velocity rocks (5.2 km s$^{-1}$). Along CEL04, the presumed location of CLF continuation beneath the Carpathian Foredeep is not manifested in the model by any velocity contrast, unlike along the CEL01 profile. Further to the south in the Inner Carpathians, the upper crustal velocities are higher and reach 5.7–6.0 km s$^{-1}$ at depths of 0–10 km. A discontinuity with 5$^\circ$ southerly dip represents the top of the lower crust in this part of the model. Along this surface, the middle crust with $V_p$ of 6.25–6.35 km s$^{-1}$ overlies the higher-velocity lower crust. Lower crustal velocities are estimated to be 6.8–6.9 km s$^{-1}$ in the Outer Carpathians and 6.5–6.6 for the Inner Carpathians, although these values have relatively higher uncertainty than
Figure 14. Diagrams showing theoretical and observed travel times of first arrivals (top), ray coverage (middle) and traveltime residuals (bottom) from tomographic inversion along the profiles CEL01 and CEL04.
Figure 15. (a) diagrams showing theoretical and observed traveltimes of first arrivals (top), ray coverage (middle) and traveltine residuals (bottom) from forward modeling along the profiles CEL01 and CEL04. Green points—refracted arrivals, red points—reflections, black circles—theoretical traveltimes. Yellow lines—fragments of discontinuities (crustal and mantle reflectors) constrained by reflected phases. The red points plotted along the interfaces mark the bottoming points of the modelled reflected phases (every third point was plotted) and their density is a measure of the positioning accuracy of the reflectors. (b) Seismic section for SP 24170 with $P_{mP}$ and $P_n$ traveltimes calculated for the Moho depth different from the final model by $\pm 1$ km. Reduction velocity is 8 km s$^{-1}$.
elsewhere in the model due to lower data quality in this area. Under the central part of the Carpathians (260–340 km distance), a prominent crustal root is observed in an 80 km wide area where thickness of the crust reaches 43 km. To the south and north of the root, the crust is 32 km thick and shallows to 25 km beneath the East Slovakian basin. Beneath the Carpathian crustal root, a north-dipping reflector was found. Its location and depth range (50–70 km) are consistent with results from the CEL01 profile. Additionally, a short south-dipping reflector, located south of the PKB, was modelled at 35–45 km depth based on upper mantle reflections, observed e.g. for shot 24150 (Fig. 6).

Tisza unit

There are no substantial differences in the crustal structures between the Tisza Unit to the SE and the neighbouring ALCAPA to the NW. In the uppermost crust, the Palaeogene to Quaternary infill of the East Slovak basin is represented by very low $V_p$ velocity (2.3–2.5 km s$^{-1}$), at depths 0–3 km. Beneath this layer, the upper crust with $V_p$ of 5.9–6.0 km s$^{-1}$ was found. Characteristic features of the Tisza unit (and the southernmost part of ALCAPA) are unusually low velocity of the lower crust (6.4–6.6 km s$^{-1}$), small crustal thickness (22 km beneath the TU) and low upper mantle velocity (7.8 km s$^{-1}$). In the upper mantle at the depth of 35–40 km, a north-dipping reflector was proposed, but it is constrained with less confidence than the above-mentioned Carpathian mantle reflectors.

The most prominent crustal tectonic feature within the Carpathians is the zone of deformation occurring at 240–320 km (CEL04) and at 400–480 km (CEL01) beneath the Outer Carpathians and the Carpathian Foredeep. Its structure differs significantly between the models. In the CEL04 profile, the zone affects the upper and lower crust and the Moho, while within the CEL01 profile, it is observed in the uppermost crust and in the shape of the Moho discontinuity, which is deepest in the southern part of the zone.

In both models, the Moho within the West Carpathians dips northwards from 22–25 km under the PBS down to 39 km (CEL01) and 43 km (CEL04) under the Outer Carpathians, shallowing farther northwards. North-dipping mantle reflectors (3–20°) occur in both models beneath Carpathians. A south-inclined, moderately dipping (30°) reflector occurs beneath PKB on CEL04 profile. Beneath the ALCAPA–Tisza contact, a north-inclined, shallowly dipping (3–13°) reflector is observed.

**Uncertainty of the Seismic Models**

The errors of the seismic forward model result from several factors: data timing errors, misidentification of the seismic phases, travel-time picking errors, inaccuracy of modelling (misfit between data and modelled traveltimes), non-uniqueness of the relation between the data and the model (e.g. trade-off between reflector depth and the velocity above it), geometry of the experiment and simplification of the method which does not account for 3-D structure or anisotropy. The errors like phase misidentification, introduced by the interpreter during correlation of the seismic phases, are subjective and impossible to quantify. Other error sources, like seismic anisotropy of the crust, are impossible to evaluate without suitable geometry of the experiment. Therefore, it is not possible to perform a full and systematic error analysis. In this study, we attempted to evaluate the errors resulting from data (picking) inaccuracy and from the misfit between the model and the data. In the first case, an empirical relation between signal-to-noise ratio at the onset of given phase and pick uncertainty, described by Zelt & Forsyth (1994), has been used. The measure of the model misfit has been obtained by calculating the traveltime residuals for all data points. Based on that, we determined the RMS residual and average residual for each phase. Figs 14 and 15 present the traveltime residuals, as well as diagrams of ray coverage and density of observed reflections along modelled velocity discontinuities. The values of the residuals are summarized in Table 1. The average residuals are close to zero, which means that there is no systematic deviation of the model parameters with respect to the data. The pick uncertainties are in the range of 0.06–0.10 s, while RMS residuals are larger (mostly 0.13–0.20 s). Assuming that the pick uncertainties are a component of RMS residuals (a relatively smooth model does not fit random data errors), the latter value has been used for estimating the uncertainty of the model parameters corresponding to evaluated traveltime errors. In this order, several sets of residuals were calculated for models with slightly perturbed $V_p$ values and depth of interfaces. The parameters of a particular layer/interface have been assigned an uncertainty value equal to the magnitude of parameter perturbation resulting in a change of the traveltime misfit of the same order as the average and RMS residual for the unperturbed (final) model. As an example, Fig. 15(b) shows the traveltimes of $P_{pP}$ and $P_\ell$ phases, calculated for the Moho depth different from the final model by ±1 km, which corresponds to the
Figure 16. The 2-D models of the P-wave velocity along the CEL01 profile developed by forward ray-tracing modelling (top) and by tomographic inversion (bottom). The grey colour covers the unconstrained parts of the model. Bold lines mark boundaries constrained by reflections and well constrained interfaces in the uppermost crust; dashed bold lines mark layer boundaries where no reflections were observed. Thin lines represent velocity isolines at intervals of 0.05 km s\(^{-1}\) in the forward model and 0.2 km s\(^{-1}\) in the tomographic model. Black triangles mark positions of the shot points. Red lines mark main geological boundaries. Vertical exaggeration of the models is 4.5. CLF—Cracow–Lubliniec Fault, HCF—Holy Cross Fault, HCM—Holy Cross mountains, KF—Kock Fault, PKB—Pieniny Klippen Belt, TTL—Teisseyre–Tornquist Line.

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Figure 17. The 2-D models of the P-wave velocity along the CEL04 profile developed by forward ray-tracing modelling (top) and by tomographic inversion (bottom). The grey colour covers the unconstrained parts of the model. Bold lines mark boundaries constrained by reflections and well constrained interfaces in the uppermost crust; dashed bold lines mark layer boundaries where no reflections were observed. Thin lines represent velocity isolines at intervals of 0.05 km s\(^{-1}\) in the forward model and 0.2 km s\(^{-1}\) in the tomographic model. Black triangles mark positions of the shot points. Red lines mark main geological boundaries. Vertical exaggeration of the models is 4.5. CLF—Cracow–Lubliniec Fault, MHL—Mid-Hungarian Line, PKB—Pieniny Klippen Belt.

Traveltime error of about ±0.15 s. The error of depth to the upper crustal interfaces is ∼±0.2 km, for lower crustal interfaces and Moho depth ±1–1.5 km, and for mantle reflector ±3 km. \(V_p\) velocity error for upper crust is ±0.05 km s\(^{-1}\), for lower crust ±0.15 km s\(^{-1}\) and for upper mantle ∼±0.05 km s\(^{-1}\). These estimates do not account for the velocity–depth trade-off and for the influence of variable ray density. In well constrained areas the uncertainty may be lower, otherwise it may increase. Moreover, as they are based on perturbing the parameters of the whole layer/interface, they are a measure of the accuracy of the average (large-scale) layer or interface properties, while uncertainty for the small-scale anomalies may be larger.

Quality of constraints imposed by the calculation of amplitudes of synthetic seismograms is even harder to determine as this process consists of qualitative (visual) comparison of seismograms only. Although in general satisfactory fit was achieved, in some cases, as e.g. for shot point 21130 (Fig. 13), the relative amplitudes of reflected phases are incorrect. Here, a
Table 1. Summary of pick uncertainties, average residuals and RMS residuals for CEL01 and CEL04 forward models. The $P_g$ phase denotes here crustal refracted phases in first and later arrivals.

| Phase        | Number of data points | Uncertainty (s) | Average residual (s) | RMS residual (s) |
|--------------|-----------------------|-----------------|----------------------|-----------------|
| $P_{\text{sed}} + P_g$ | 1350                  | 0.06            | -0.004               | 0.13            |
| $P_{i1}$     | 198                   | 0.08            | 0.030                | 0.16            |
| $P_{i2}$     | 174                   | 0.10            | -0.002               | 0.17            |
| $P_{mP}$     | 599                   | 0.10            | -0.030               | 0.18            |
| $P_s$        | 158                   | 0.09            | -0.040               | 0.17            |
| $P_1P$       | 188                   | 0.09            | 0.070                | 0.19            |
| $P_{\text{sed}} + P_g$ | 826                   | 0.06            | -0.007               | 0.14            |
| $P_{i1}$     | 50                    | 0.08            | -0.030               | 0.13            |
| $P_{i2}$     | 258                   | 0.09            | 0.030                | 0.17            |
| $P_{mP}$     | 502                   | 0.09            | 0.040                | 0.19            |
| $P_s$        | 261                   | 0.08            | 0.060                | 0.17            |
| $P_1P$       | 137                   | 0.10            | 0.150                | 0.40            |

Better fit could be obtained by increasing $V_p$ in the lower crust by about 0.4 km s$^{-1}$. This, however, would contradict the traveltime constraints, which we consider to be of primary importance. Therefore the model has been accepted in spite of the amplitude misfit.

**GEOLOGICAL INTERPRETATION**

Interpretation of the uppermost crustal lithologies within the Carpathians is straightforward as all upper crustal complexes are either exposed or have been probed in boreholes.

We believe that the zone of crustal deformations discernible on the CEL04 (240–320 km) and CEL01 (400–440 km) profiles relates to the suture—evidence of Tertiary collision between the ALCAPA and Bruno–Silesia. The suture is better preserved in the CEL04 profile, where it is discernible in the upper and lower crust. Here, the uppermost crust is formed by the Outer Carpathian accretionary prism that is composed largely of Early Cretaceous through Early Miocene flysch (velocity 4.3 km s$^{-1}$). The prism, up to 6 km thick, overlies a stack of slices composed of Palaeozoic and Mesozoic sedimentary, metamorphic and crystalline rocks of the Inner Carpathians (velocity 5.6–5.9 km s$^{-1}$). Both the Inner and Outer Carpathians are thrusted northwards over the Bruno–Silesian crust composed of Proterozoic to Mesozoic metamorphic and sedimentary rocks (velocity 5.9–6.1 km s$^{-1}$). Similar north-verging overthrust occurs within the lowermost crust. Along this overthrust, the ALCAPA crust (velocity 6.00–6.60 km s$^{-1}$) is thrusted over the Bruno–Silesian lower crust (velocity 6.75–6.90 km s$^{-1}$), which forms a prominent crustal root. We believe that this crustal root, as well as the south-dipping reflector, penetrating mantle in an extension of the root, correspond to a remnant of the subducted slab.

In the CEL01 profile, the suture discussed above is delineated only in the uppermost and the lowermost crust. The Outer Carpathian accretionary prism (up to 8 km thick) is an outstanding feature there, overlying the Proterozoic to Mesozoic metamorphic and sedimentary rocks (velocity 5.9–6.1 km s$^{-1}$) of the Bruno–Silesian crust. Due to scarce seismic data, the contact between the ALCAPA and Bruno–Silesia is not discernible in the deeper crust below the accretionary prism, and no structural details are observed. A notable feature is a depression of the Moho boundary below the accretionary prism, which may correspond to a remnant of the crustal root.

In the descending plate (Bruno–Silesia), crustal units dip southwards, within the whole crust (CEL04) or only in the upper crust (CEL01). The subvertical contact observed in the CEL01 profile within the descending plate (approximately 480 km) may correspond to the CLF separating the Bruno–Silesia and MU.

In the upper plate (ALCAPA), the uppermost crustal complexes correspond to:

1. a stack of slices composed of Palaeozoic and Mesozoic sedimentary, metamorphic and crystalline rocks (velocity 5.6–5.9 km s$^{-1}$) in Slovenske Rudohorie Mts. (CEL01) and Branisko–Čierna Hora Mts. (CEL04);

![Figure 18](image-url). Cartoons illustrating hypothetical development of the lithospheric structure within the Tertiary suture zone of the Western Carpathians, based on the CEL04 model (Fig. 17). Similar interpretation applies to CEL01 profile. (a) at the end of the subduction phase; (b) during collision of ALCAPA and European plate; (c) present day architecture; (1) Outer Carpathians accretionary prism and sedimentary infill of the Levoča basin; (2) upper crust; (3) middle crust; (4) lower crust; (5) upper mantle; (6) ALCAPA asthenosphere; blue line—reflector in the upper mantle. For further explanations see text.
(2) stack of Palaeozoic to Tertiary sedimentary, metamorphic and crystalline rocks (velocity 6.1 km s\(^{-1}\)) of the Transdanubian Range (CEL01);
(3) Tertiary strata (\(V_p = 4.3\) km s\(^{-1}\)) of the Levoča basin (CEL04) and
(4) Tertiary and Quaternary strata (velocity ca. 3.0 km s\(^{-1}\)) of the Danube and Zala basins (CEL01) as well as Tertiary and Quaternary strata (\(V_p = 2.3–2.5\) km s\(^{-1}\)) of the East Slovak basin (Turňa and Košice depressions) (CEL04).

The latter basin also extends over the Tisza unit (CEL04). The ALCAPA upper mantle is characterized by low \(V_p\) (7.8 km s\(^{-1}\)). As the seismic velocity is inversely related to the temperature, possible explanation can be the high temperature at the Moho in this area.

**DISCUSSION**

The most prominent tectonic feature within the West Carpathians is the Tertiary suture between the ALCAPA and BSU (European plate). The Outer Carpathian accretionary prism is nicely depicted on both models. A major overthrust is discernible beneath the prism on the CEL04 profile. Along this overthrust, upper ALCAPA crust as well as the Outer Carpathians are thrust northwards over the Bruno–Silesian crust. Similarly, on the CEL01 profile, the accretionary prism is thrust over the Bruno–Silesian crust.

Another prominent feature, observed in both models, is a crustal root related to the Tertiary subduction. The results of previous research show that the Tertiary crustal root is lacking in some parts of the Western Carpathians. According to Tomek (1993) and Tomek & Hall (1993) in the western part of the Western Carpathians (Fig. 1), the plunging slab became completely destroyed and the Moho boundary became subhorizontal. This has been confirmed by gravity modelling (Szaﬁan & Hall 1997). However, Bielik (1999) suggests the presence of a crustal root in some parts of the Western Carpathians, which corresponds with the results of Uchman (1975), Bojdys & Lemberger (1986) and with our own data. The lack of the crustal root in some parts of the Carpathians can be explained by the complete detachment of the plunging slab and its sinking in the asthenosphere (cf. Gvirtzman 2002). Another plausible explanation for the absence of the crustal root is post-subduction phase changes (Tomek & Hall 1993; cf. Szaﬁan et al. 1997). Moreover, the evidence presented by Costa & Rey (1995, and references therein) show that the post-subduction (post-thickening) collapse of an orogen may involve a rejuvenation of the crust. Such rejuvenation may erase the record of the events that occurred prior to the collapse. This model fits well with the structural history of the Outer Carpathians, where creation of the subduction-related orogen was followed by a collapse (Zuchiewicz et al. 2002, and references therein). We believe, therefore, that rejuvenation may be responsible for the lack of the suture in the middle and lower crust in the CEL01 model. The CEL04 area seems to be less affected by this process as the suture in the lower crust was preserved there—possibly due to laterally varying degree of rejuvenation. It is also possible that fragments of the suture zone are not resolved by available data. The geometry of the suture zone fragments as interpreted from the CEL04 model suggests that at the lower crustal level, the suture was shifted northwards after cessation of the subduction (Fig. 18), most likely in a compressional stress regime during collision and (?!) post-collision continuing convergence between the ALCAPA and the European plate.

According to current geological interpretations (e.g. Zuchiewicz et al. 2002), the Bruno–Silesia is cut from NE by the CLF—a major tectonic feature separating the BSU from the MU (Malinowski et al. 2005). The CLF is an outstanding feature in the CEL01 profile, affecting the upper crust. Moreover, in the CEL02 profile (Fig. 1b), 35 km to NW from CEL01, the CLF affects the whole crust (Malinowski et al. 2005). However, in the CEL04 profile, 20 km SE from CEL01, and in the CEL05 profile 100 km to SE (Grad et al. 2006), the fault is not discernible. Summing up, the CLF is a prominent crustal feature in the NW part of the discussed region whereas it is not discernible in its SE part.

In part of the ALCAPA unit, there is a velocity inversion in the upper crust of the Transdanubian Range. The uppermost crust (velocity of 6.1 km s\(^{-1}\)) overlie there a complex with velocity 5.8–6.0 km s\(^{-1}\). This architecture nicely fits the geological interpretation of this area (Dudko et al. 1990; see also Szaﬁan et al. 1997). In this interpretation, dense (2760 kg m\(^{-3}\)) Cretaceous strata (mostly carbonates) overlie less dense Triassic marls (2570 kg m\(^{-3}\)) and metamorphic Palaeozoic rocks (2630 kg m\(^{-3}\)). A characteristic feature of the PBS is its low crustal thickness (22–30 km), with the thinnest crust beneath the Tisza unit. This well known crustal thinning is believed to be related to backarc extension and related mantle upwelling (for review see: Konečny et al. 2002). The major tectonic feature in the PBS, the Mid-Hungarian Line separating the ALCAPA and Tisza is not discernible (CEL04).

One of the most prominent features in both models (in terms of high amplitude of the reflection events they produce) is the presence of north-dipping reflectors in the upper mantle beneath the Carpathians. They are very well constrained thanks to the very good quality of the reflections, observed on several sections and therefore modelled with high confidence. Similar reflections are also observed in CEL05 data (Grad et al. 2006). Since in both models (CEL01 and CEL04) the reflectors have a similar inclination, depth range and geographical location, they are likely to represent one prominent discontinuity in the upper mantle, extending over a wide area and dipping to the north, approximately perpendicularly to the strike of the Carpathian arc. Such dipping mantle reflectors are usually interpreted as images of subducted lithospheric slabs (e.g. Balling 2000), where seismic velocity differs from the surrounding mantle, or shear zones (Abramovitz et al. 1998; Krawczyk et al. 2002), where alteration of rock structure produces impedance contrast. As discussed earlier, the existing geological data strongly support concept of south-dipping subduction, opposite to the dip of the reflector. Therefore, we believe that the presence of a north-dipping subduction zone is unlikely, and we prefer the following, alternative interpretation, as more consistent with other data. As this part of the lithosphere is likely to have been subject to heavy deformation due to ALCAPA–European convergence, the discontinuity may represent a shear zone or set of shear zones, which originated in a compressional stress regime during collision of the continental lithospheric plates after the intervening oceanic or suboceanic lithosphere was subducted (Fig. 18). The lack of south-dipping reflections from the subduction zone can be explained by slab detachment or phase changes (Tomek & Hall 1993; cf. Gvirtzman 2002). The amplitude of the relative movement along the shear zone and the affinity of the underlying mantle cannot be determined from the data. The simplest solution would assume a small-amplitude movement and a European origin of the underlying lithosphere. However, an alternative scenario, involving large-amplitude translation and underthrusting of the delaminated ALCAPA uppermost mantle beneath the wedge of rigid and strong European lithosphere, cannot be excluded (Meissner et al. 2002; Snyder 2002). This scenario might provide another possible mechanism of destruction of the subduction zone. Such shear zones may contain strongly altered and mylonitized rocks,
low-velocity melt fractions or water-saturated material. According to Warner & McGearry (1987), mantle shear zones can contain mafic rocks crystallized from partial melts originating from the mantle, or partially hydrated peridotite, which both have sufficient velocity/density contrast compared to neighbouring mantle to produce high reflection coefficients. Fractured mantle material in a shear zone can also produce velocity contrast, compared to unfractured rock, which is detectable by seismic measurements (Meissner 1996). Hansen & Balling (2004) mention mylonitization as one of the factors that decrease seismic velocity in shear zones due to grain size reduction and mineral reactions (they estimate that high amplitude wide-angle reflections from low-velocity shear zones require velocity contrast at least 10 per cent). Therefore, all these factors can be responsible for decrease of the seismic velocity in the reflecting zone, creating high impedance contrasts.

CONCLUSIONS

Modelling of the seismic data from the CELEBRATION 2000 profiles CEL01 and CEL04 provided models of the crustal structure in the Western Carpathians and, we believe, enhanced the knowledge about the tectonic evolution of the Carpathian orogen.

(1) Obtained models of the crustal and mantle structure support the commonly accepted view of the West Carpathians as being largely formed due to the interplay between: (i) the Palaeogene to Middle Miocene southward subduction of the oceanic or suboceanic crust intervening between the ALCAPA unit and the BSU (European plate) and subsequent Middle Miocene collision, (ii) Early to Middle Miocene mantle upwelling beneath the PBS, (iii) Late Miocene to Recent collapse and (iv) Late Miocene to Recent convergence of the ALCAPA unit and the European plate.

(2) The architecture of the collision-related crustal suture was partly reshaped and partly destroyed during collision and by post-collapse convergence between the ALCAPA unit and the European plate (Fig. 18).

(3) The subduction-related orogenic root is at least partially preserved in both profiles (CEL01, CEL04) where crustal thickness reaches 39 and 43 km, respectively.

(4) The CLF, a prominent tectonic feature corresponding to the boundary between the MU and the BU, is discernible north of the Carpathian frontal thrust along the CEL01 profile, but is not observed farther to SE (CEL04) beneath the Carpathians.

(5) We have not observed any substantial differences in the crustal structure between the Tisza unit and the ALCAPA unit. Therefore, the Mid-Hungarian Line separating these units is not discernible (CEL04).

(6) Prominent reflections from the upper mantle observed along both profiles are the evidence for a north-dipping mantle discontinuity, which may represent a shear zone related to a collision between the ALCAPA unit and the European plate and possibly ongoing convergence between these units.

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