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Lithospheric structure beneath trans-Carpathian transect from Precambrian platform to Pannonian basin: CELEBRATION 2000 seismic profile CEL05

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[1] In 2000, a consortium of European and North American institutions completed a huge active source seismic experiment focused on central Europe, the Central European Lithospheric Experiment Based on Refraction or CELEBRATION 2000. This experiment primarily consisted of a network of seismic refraction profiles that extended from the East European craton, along and across the Trans-European suture zone region in Poland to the Bohemian massif, and through the Carpathians and eastern Alps to the Pannonian basin. The longest profile CEL05 (1420 km) is the focus of this paper. The resulting two-dimensional tomographic and ray-tracing models show strong variations in crustal and lower lithospheric structure. Clear crustal thickening from the Pannonian basin (24–25 km thick) to the Trans-European suture zone region (~50 km), together with the configuration of the lower lithospheric reflectors, suggests northward subduction of mantle underlying Carpathian-Pannonian plate under the European plate. This, however, conflicts with strong geological evidence for southward subduction, and we present three tectonic models that are to not totally mutually exclusive, to explain the lithospheric structure of the area: (1) northward "old" subduction of the Pannonian lithosphere under the East European craton in the Jurassic–Lower Cretaceous, (2) a collisional zone containing a "crocodile" structure where Carpatho-Pannonian upper crust is obducting over the crystalline crust of the East European craton and the Carpathian-Pannonian mantle lithosphere is underthrusting cratonic lower crust, and (3) lithosphere thinning due to the effects of Neogene extension and heating with the slab associated with "young" subduction southward in the Miocene having been either detached and/or rolled back to the east. In the last case, the northwestward dipping in the lithosphere can be interpreted as being due to isotherms that could represent the lithosphere/asthenosphere boundary in the Pannonian region.

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1. Introduction

[2] The complex tectonic history of central Europe (Figure 1) reflects breakup of the Rodinian supercontinent and the subsequent growth of continental Europe beginning with the Caledonian orogeny. Caledonian and younger Variscan orogenesis involved accretion of Laurentian and Gondwanan terranes to the rifted margin of Baltica (East European craton, EEC) during the Paleozoic. From central Poland northward, the region also experienced volcanic activity during the Permian and tectonic inversion during the Alpine orogeny, which in the south continues today. The Trans-European suture zone (TESZ) is a term used to refer to the suite of sutures and terranes that formed adjacent to the rifted margin of Baltica [e.g., *Pharaoh et al.*, 1997], and these features extend from the British Isles to the Black Sea (Figure 1). The tectonic evolution of this region shares many attributes with the Appalachian/Ouachita orogen [*Keller and Hatcher*, 1999; *Golonka et al.*, 2003a, 2003b] and is certainly of global importance to studies of terrane tectonics and continental evolution. The Bohemian massif (Figure 1) is mostly located in the Czech Republic and is a

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Figure 1. Location of the CELEBRATION 2000 profile CEL05 on the background of simplified tectonic map of the central Europe. HCM – Holy Cross Mountains; LT – Lublin trough; MM, Małopolska massif; TESZ, Trans-European suture zone; USB, Upper Silesia Block. Red rectangular in the insert shows the study area; Carp., Carpathians; EEC, East European craton; V.T.E., Variscan terranes of Europe.

large, complex terrane whose origin can be traced to northern Gondwana (Africa). In southern Poland, several structural blocks such as the Małopolska massif (MM in Figure 1) are located adjacent to Baltica and were probably transported laterally along it, similar to Cenozoic movement of terranes along the western margin of North America. The younger eastern Alps, Carpathian arc and Pannonian back-arc basin form interrelated components of the Mediterranean arc basin complex (Figure 1).

[3] Beginning in 1997, central Europe has been covered by a network of seismic refraction experiments to investigate its complex lithospheric structure [*Guterch et al.*, 2003a]. These experiments (POLONAISE'97, CELEBRATION 2000, ALP 2002, and SUDETES 2003) have only been possible due to a massive international cooperative effort. The total length of all profiles is about 20,000 km, and during these four experiments, a total of 295 large explosions provided the seismic sources. As a result of these

Figure 2. Location of profile CEL05 together with CELEBRATION 2000 profiles (CEL01–CEL10). Large yellow stars refer to the location of 26 shot points along profile CEL05 with shots numbers in boxes (25010–25290). Table 1 provides detailed information about the shot points. Small red dots refer to recording positions. Other seismic refraction and wide-angle reflection profiles (POLONAISE'97 profiles P1–P5, LT-7, TTZ, EUROBRIDGE profiles EB'95&96 and EB'97) are all marked by black solid lines.

experiments, a network of seismic refraction profiles now extends from the East European craton, along and across the Trans-European suture zone (TESZ) region of Poland and the Bohemian massif, through the Carpathians and eastern Alps to the Pannonian basin, the Dinarides and Adriatic Sea [*Guterch et al.*, 1998, 1999, 2000, 2001, 2003a, 2003b; *Brueckl et al.*, 2003; *Grad et al.*, 2003a, 2003b].

[4] The focus of this paper is the Central European Lithospheric Experiment Based on Refraction 2000 (CELEBRATION 2000), which involved a consortium of European and North American institutions, 28 in all (Figure 2). This international collaborative project involved geophysical groups from Poland, the United States, Hungary, the Czech Republic, Slovakia, Austria, Canada, Denmark, Turkey, Russia, Belarus, Germany, and Finland. CELEBRATION 2000 targeted the tectonic features along the TESZ region, as well as the southwestern portion of the East European craton, the Carpathian Mountains, the Pannonian basin, eastern Alps and the Bohemian massif. The purpose of this paper is to present a detailed analysis of the main CELEBRATION 2000 profile CEL05 (Figures 1 and 2).

2. Previous Geophysical Investigations

[5] The area of the CELEBRATION 2000 experiment has been studied employing a variety of geophysical methods at relatively low resolution. Early deep seismic sounding (DSS) studies were performed in the Carpathian Mountains

Table 1. Location and Parameters of Shot Points for Profile CEL05

Shot	Latitude, °N	Longitude, °E	Height, m above sea level	Date	Time, UTC	TNT, kg	Offset, km	Time Correction
25010	46°06.819′	18°42.486′	110	25 Jun 2000	2115:00.000	750	6.33	0
25020	46°23.277′	18°54.814′	90	25 Jun 2000	2215:00.000	400	40.67	0
25040	47°09.340′	19°32.480′	140	24 Jun 2000	2130:00.000	750	138.55	0
25050	47°24.991′	19°46.992′	101	7 Jun 2000	2115:00.000	300	172.81	0
25070	47°50.033′	20°10.386′	145	8 Jun 2000	0130:00.000	400	227.67	0
25080	48°05.836′	20°20.119′	330	7 Jun 2000	2130:00.000	500	259.06	0
25090	48°20.252′	20°30.842′	270	8 Jun 2000	2145:00.000	500	288.84	0
25100	48°28.840′	20°38.875′	272	8 Jun 2000	2130:00.000	300	307.58	0
26710	48°34.833′	20°48.310′	300	8 Jun 2000	2245:00.000	1200	323.22	0.321
25110	48°45.160′	20°53.807′	609	8 Jun 2000	2230:00.000	800	343.05	0.578
25120	48°52.972′	20°59.827′	480	10 Jun 2000	0130:00.000	1200	359.30	0.796
25130	49°13.653′	21°04.972′	686	9 Jun 2000	2245:00.000	250	395.74	0.255
25140	49°27.668′	21°15.493′	601	7 Jun 2000	2330:00.000	900	424.61	0
25150	49°40.168′	21°24.628′	291	8 Jun 2000	0100:00.000	180	450.24	0
25170	49°58.248′	21°40.881′	360	8 Jun 2000	2330:00.000	150	488.99	0
25180	50°09.055′	21°46.659′	222	9 Jun 2000	0000:00.000	250	509.90	0
25190	50°19.350′	21°59.650′	196	9 Jun 2000	0030:00.000	360	534.04	0
25200	50°28.395′	22°06.888′	168	9 Jun 2000	0100:00.000	400	552.87	0
25210	50°44.858′	22°24.231′	257	10 Jun 2000	0015:00.000	350	589.49	0
25220	51°04.549′	22°38.794′	248	9 Jun 2000	0015:00.000	200	629.68	0
25230	51°27.748′	23°03.876′	169	8 Jun 2000	0015:00.000	300	681.50	0
25240	51°45.698′	23°22.981′	159	9 Jun 2000	2345:00.000	500	721.36	0
25250	51°57.718′	23°28.332′	144	8 Jun 2000	2345:00.000	600	743.71	0
25270	54°11.301′	26°39.142′	180	7 Jun 2000	2100:00.000	500	1067.42	0
25280	54°23.888′	26°49.758′	154	8 Jun 2000	2100:00.000	1500	1093.13	0
25290	56°54.450′	29°18.650′	198	10 Jun 2000	0115:00.000	15000	1411.23	0

region of southeast Poland and Slovakia along international profiles V and VIII [*Uchman*, 1975; *Sollogub et al.*, 1976] and regional profiles LT-3 and LZW [*Guterch et al.*, 1986a, 1986b] (Figure 2). These low-resolution profiles indicate that the crustal thickness in southeast Poland varies from ~48 km within the EEC, to ~45 km in the Holy Cross Mountains area (Figure 1), to ~35 km in the Paleozoic terranes, to ~55 km in the Teisseyre-Tornquist zone (TTZ) that lies along the southwest margin of the EEC (Figure 1). In the Carpathians and their foredeep, a crustal thickness of ~40 km was found.

[6] North of the Carpathian region, the East European craton (EEC) was investigated along early DSS international profile VII, as well as, modern profiles LT-7 and TTZ, and those recorded as part of the EUROBRIDGE and POLONAISE'97 projects [*Guterch et al.*, 1986a; *Grad*, 1976, 1986; *EUROBRIDGE Seismic Working Group*, 1999, 2001; *Środa and POLONAISE P3 Working Group*, 1999; *Środa et al.*, 2002; *Wilde-Piórko et al.*, 1999; *Czuba et al.*, 2001, 2002; *Janik et al.*, 2002; *Grad et al.*, 1999, 2002a, 2002b, 2003a; *Kozlovskaya et al.*, 2004; *Dadlez et al.*, 2005; *Majdański and Grad*, 2005]. The average depth of the crystalline basement in the EEC is ~2 km. In the region of Mazury-Suwałki elevation (NE Poland), the basement depth is only 0.3–1 km and increases toward the southwest to 7–8 km along the margin of the EEC. The mean velocity of the sedimentary cover increases with thickness and varies from ~2.5 km/s where these strata are 1 km thick to about 4.3 km/s where they are 8 km thick [*Grad*, 1986]. The velocity of the crystalline basement is generally uniform being 6.1–6.2 km/s, but it is higher (6.4–6.6 km/s) where intrusions of rapakivi-like granite, gabbro and anorthosite are found in the upper crust [e.g., *Czuba et al.*, 2002; *Grad et al.*, 2003a]. In the EEC, velocities of 6.5–6.7 and 6.9–7.2 km/s characterize the middle and lower crust, respectively; the depth to the Moho discontinuity ranges from 40 to 55 km; and the sub-Moho velocity is 8.05–8.2 km/s.

[7] In the western Carpathians, a program of deep reflection seismic profiles was undertaken in Slovakia [*Vozár et al.*, 1999; *Šantavý and Vozár*, 1999]. However, an old international profile [*Uchman*, 1975] was the only DDS profile made in this area prior to the CELEBRATION 2000 experiment. In the Pannonian basin, both refraction and deep reflection profiles were recorded in 1970s [*Posgay et al.*, 1981, 1986, 1995]. These studies discovered a thin crust (25–30 km) and a low-velocity layer in the upper mantle (lithosphere-asthenosphere boundary?) the top of which is at a depth of ~55 km [*Posgay et al.*, 1981].

[8] Many nonseismic studies have been undertaken in the region. For example, heat flow measurements indicate a major change in thermal regime across central Europe [*Čermák and Bodri*, 1998]. Using over 3200 heat flow measurements, *Čermák and Bodri* [1998] explained a number of prominent heat flow anomalies (e.g., Pannonian basin, French Massif Central, the Alps) as a product of the deep-seated lithospheric processes. In the area of profile CEL05, the TESZ separates "cold" lithosphere of the EEC with low heat flow 30–40 mW/m^2 from "hot" lithosphere with higher heat flow of 40–70 mW/m^2 in the Paleozoic terranes and Carpathians, and even higher values (80–110 mW/m^2) are found in Pannonian basin [*Majorowicz and Plewa*, 1979; *Čermák et al.*, 1989; *Maj*, 1991a, 1991b; *Plewa*, 1998; *Čermák and Bodri*, 1998; *Zeyen et al.*, 2002; *Majorowicz et al.*, 2003]. The characteristic Moho temperature for the EEC has been estimated to be 500°C [*Majorowicz et al.*, 2003] and 590–620°C [*Čermák et al.*, 1989]. In the TESZ region, the Moho temperature

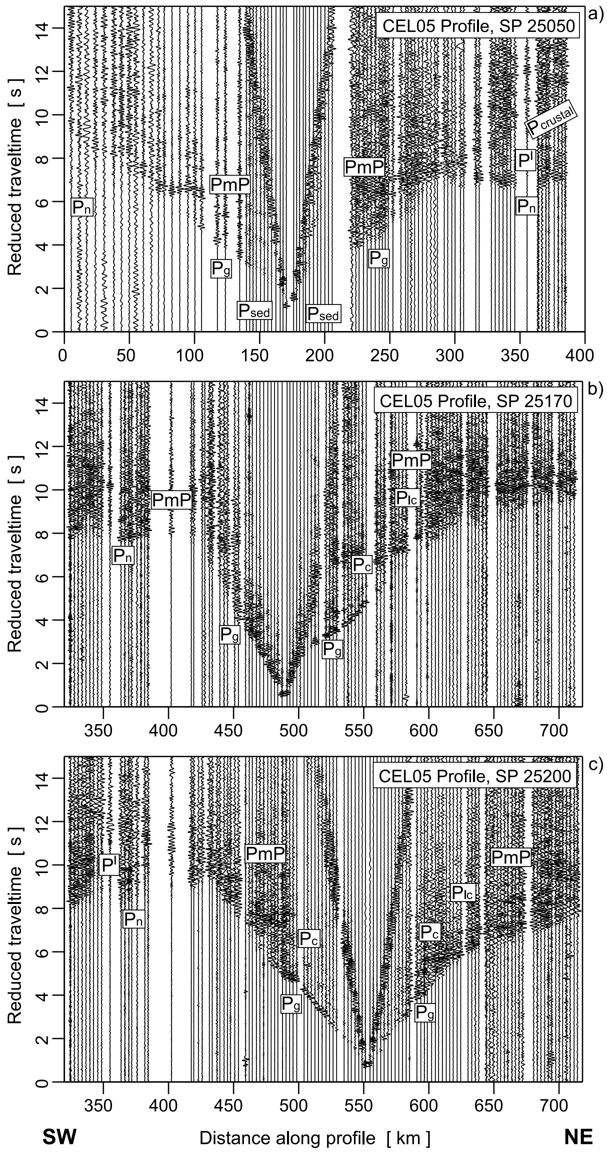

Figure 3. Example of trace-normalized, vertical component seismic record sections for (a) SP25050, (b) SP25170, and (c) SP25200. A band-pass filter (2–12 Hz) has been applied. *P*sed, waves refracted/reflected in sediments; *Pg*, waves refracted from the basement; *Pc*, waves reflected from midcrustal discontinuity; *P*lc, waves reflected from discontinuity in the lower crust; *PmP* and *Pn*, reflected and refracted waves from the Moho; P^{I}, waves from the lower lithosphere. Reduction velocity is 8.0 km/s.

increases to 650–750°C and beneath the Carpathians and Pannonian basin it even increases to 800–900°C [*Dövényi et al.*, 1983; *Čermák et al.*, 1989; *Majorowicz et al.*, 2003].

[9] In the area of profile CEL05, gravity and magnetic anomalies indicate significant variations in lithospheric structure [*Królikowski and Petecki*, 1995; *Karaczun et al.*, 1978; *Wybraniec et al.*, 1998; *Grabowska and Bojdys*, 2001; *Petecki et al.*, 2003]. Bouguer anomalies in the area of the EEC are relatively homogeneous (0 ± 20 mGal) and decrease in the area of the EEC margin down to about −40 mGal. The anomaly field becomes more complex in this region with a large high in the Holy Cross Mountains area (Figure 1) and lower values to the northwest along the EEC margin. In the Carpathians, Bouguer anomalies reach values of about −80 mGal, and increase into the Pannonian basin (20–30 mGal) where the crust thins.

[10] Magnetic anomalies along the cratonic part of profile CEL05 contain many short-wavelength variations from −1500 to +1500 nT that correlate well with tectonic features and intrusions in the Precambrian basement. Farther to the southwest in the TESZ, the Carpathian foredeep and Carpathians, magnetic anomalies are subdued (±100 nT) presumably due to the deeply buried magnetic basement. Except in areas of Tertiary volcanism, magnetic anomalies are also subdued in the Pannonian basin because of the thick cover of sedimentary rocks.

3. Data Acquisition, Processing, and Seismic Wave Field

[11] The layout of the CELEBRATION 2000 experiment was a network of interlocking recording profiles whose total length was about 9000 km and along which the station spacing was 2.8 or 5.6 km. Thanks to IRIS/PASSCAL, Canadian, and European resources, the total number of instruments deployed was 1230. Shots (147 in all) were fired along most of the recording profiles, so that in addition to forming an array, about 5400 km of traditional profile data were obtained. The sources ranged in size from 15 t to 90 kg, with the average being ~500 kg.

[12] Profile CEL05 is the longest recorded in this experiment. Its length is 1420 km, and it begins in Hungary, crosses Slovakia, Poland and Belarus, and ends in northwestern Russia (Figure 2). The southwestern part of the profile (0–200 km) lies in the Pannonian basin. The profile then crosses a tectonically complex zone that includes the Carpathians and their foredeep (200–500 km) and the TESZ (500–700 km), including the Małopolska massif (MM), Lublin trough and Teisseyre-Tornquist zone (TTZ). The northeastern part of the profile (700–1420 km) crosses the East European craton (Figures 1 and 2). Data from 26 explosions made along profile CEL05 were collected using over 360 modern seismic recorders, with a nominal station spacing 2.8 km in Hungary, Slovakia and Poland and 4.6 km in Belarus and Russia. More details about the layout of the experiment are provided by *Guterch et al.* [2001, 2003b]. Detailed information about the shot points are provided in Table 1.

[13] The *P* wave field on the CEL05 record sections has high signal-to-noise ratio, particularly for southern (Pannonian basin) and northern cratonic parts of the profile. Identification and correlation of seismic phases was done manually on a computer screen using software that allows flexible use of scaling, filtering, and reduction velocity [*Zelt*, 1994; *Środa*, 1999]. Clear arrivals of refracted and reflected waves from sedimentary layers, the crystalline crust and the upper mantle were typically observed up to offsets of 200–300 km and for some shots, even over 900 km. Examples of record sections for profile CEL05 from five shot points (Figure 2) are shown in Figures 3–5, and many more examples of record sections are shown in the subsequent figures that illustrate the results of our

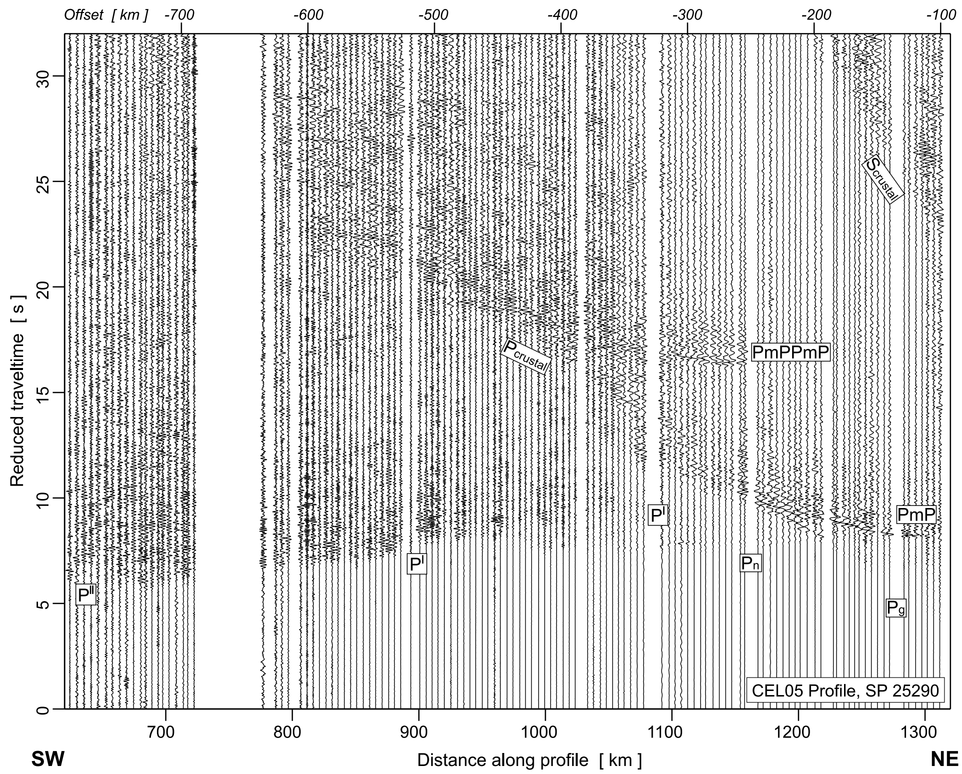

Figure 4. Example of trace-normalized, vertical component seismic record section for SP25290. A band-pass filter (2–12 Hz) has been applied. Reduction velocity is 8.0 km/s. *PmPPmP*, wave-reflected twice from the Moho discontinuity and the free surface; *P*crustal and *S*crustal, over critical crustal *P* and *S* waves traveling in the crust; P^{I} and P^{II}, waves from the lower lithosphere. Other phases are labeled as in Figure 3.

modeling efforts. As discussed in detail below, our analysis began with tomographic inversion of travel times of refracted (diving) waves observed as first arrivals (Figure 6). Then, the velocity model shown in Figure 7 was derived by ray tracing and synthetic seismogram modeling. The calculations of travel times, ray paths, and synthetic seismograms were made using the ray theory package SEIS83 [*Červený and Pšenčík*, 1983], enhanced by employing the interactive graphics interfaces MODEL [*Komminaho*, 1997] and ZPLOT [*Zelt*, 1994] with modifications by *Środa* [1999]. The initial velocity model based on the tomographic inversion results was successively altered by trial and error, and travel times were recalculated many times until agreement was obtained between observed and model-derived travel times (within a misfit of the order of 0.1–0.2 s; e.g., Figures 8–16). In this modeling, first arrivals, post critical waves, reflections and even multiple reflections were employed.

[14] In addition to matching travel times, synthetic seismograms were calculated to control velocity gradients within the layers and the velocity contrast at the seismic boundaries. The final synthetic seismograms show good qualitative agreement with the relative amplitudes of observed refracted and reflected waves (e.g., Figures 11 and 12). This combined approach allowed us to identify midcrustal boundaries, and provided a particularly precise determination of the shape of the Moho.

[15] General observations about the data are that waves from sedimentary cover (*P*sed) are observed as first arrivals in the vicinity of shot points up to offsets of only 1 km on the EEC while they are observed to offsets of 10–20 km in the TESZ region and even ~30 km in the Carpathians (e.g., Figures 3, 8, and 15). Their apparent velocities range from 2.5 to 5.5 km/s. After these arrivals, the *Pg* phase that travels in the crystalline basement (upper crust) is recorded to highly variable offsets of 1–250 km (e.g., Figures 3 and 8). The complexity of the upper crustal structure is well illustrated in Figure 5, which shows seismic sections and *Pg* wave travel times for shot point SP25210 recorded in four directions along profiles CEL05 and CEL03 [*Janik et al.*, 2005]. The slowest arrivals occur to the southwest toward the Carpathian foredeep and Carpathians, and at offsets of 100–150 km, they are ~2 s late compared to other directions. As discussed below, the highly variable crustal structure along the profile results in mantle arrivals also becoming first arrivals at highly variable offsets (e.g., Figure 10).

[16] In the Pannonian basin region, the seismic wave field is characterized by strong *Pg* arrivals to offsets of ~50 km from the shot point (e.g., SP25010 in Figure 8 and SP25040 in Figure 9). For offsets out to 150 km, the relative amplitude of the *Pg* wave is small relative to the dominant reflection from the Moho (*PmP*). Reflections from the top of the middle crust (Pc) are well correlated and observed throughout the whole "Pannonian" part of profile (e.g., SP25010 in Figure 8 and

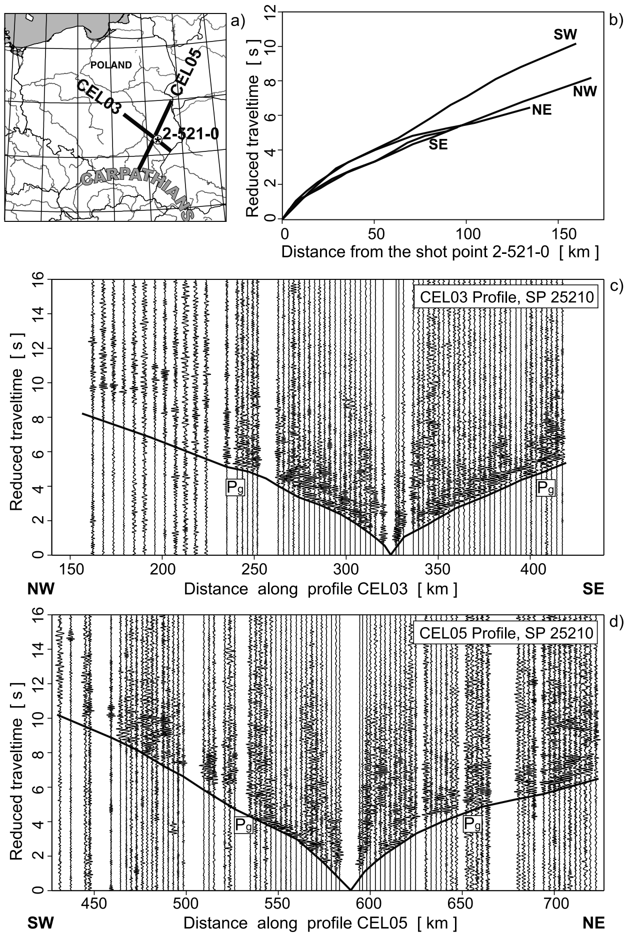

Figure 5. Example of differentiation of *Pg* wave travel time for shot point SP25210 recorded along profile CEL03 and CEL05 (a). Location of SP25210 at crossing point of profiles CEL03 and CEL05 in SE Poland; (b) reduced travel times of *Pg* wave for SP25210 recorded in NW, NE, SE, and SW directions; (c) seismic record section for SP25210 along profile CEL03; and (d) seismic record section for SP25210 along profile CEL05. Note the late arrivals toward the southwest.

SP25040 in Figure 9). Reflections from the top of the lower crust (*P*lc) are well correlated and observed throughout the whole "cratonic" part of profile (e.g., SP25190 in Figure 10 and SP25250 in Figure 11). Refractions from the middle and lower crust are correlated as secondary arrivals in the seismic sections, primarily at long offsets where refractions merge with wide-angle reflections to form the over critical phases *P*crustal and *S*crustal, which in some cases are well recorded to offsets of ~300 (e.g., SP25050 in Figure 3a, SP25010 in Figure 8a). These phases are important for determining the velocities in the middle and lower crust, particularly for the Pannonian crust where the velocities of these layers are 6.2–6.6 km/s. The phase diving in the uppermost mantle (*Pn*) has relatively low amplitudes and low apparent velocities (7.9–8.0 km/s). The intercept time for *Pn* is only ~6 s (for reduction velocity 8 km/s), and its cross-over distance with the crustal (*Pg*) arrivals is only 100–120 km. These observations clearly show that the Moho in the Pannonian basin is shallow. Starting at offsets of ~200 km, arrivals from the mantle lithosphere (P^{I} and P^{II}) were observed for some shots (Figures 8–10).

[17] The wave field observed in Carpathians is complex as is its known geologic structure. First arrivals (*Pg*) are distinct up to ~100 km offset and are characterized by large variations in apparent velocity and amplitude (Figures 3b and 3c; SP25180 in Figure 9a). Midcrustal reflections (*Pc*) are usually recorded at short distance intervals (20–50 km) and are also characterized by variations in apparent velocity and amplitude (e.g., distances of 520–560 km for SP25170 in Figure 15b). *Pn* arrivals were only fragmentarily recorded in the Carpathians. This characteristic attribute of the wave field testifies to the complex structure of the crust–upper mantle transition in this region. The observed wave field is similar in the Carpathian foredeep adjacent to the northeast. However, waves with low apparent velocities (4.0–5.0 km/s) are observed up to 30–50 km offsets (Figure 13), and midcrustal phases are even more complex.

[18] In the TESZ area, the character of the wave field is more uniform, and the travel times of both refracted first arrivals (*Pg*) and waves from the Moho and uppermost mantle (*PmP* and *Pn*) show clear evidence that the crystalline basement and crust–upper mantle boundary are deep. Compared to the Pannonian area, the *Pn* wave intercept time is much later (reduced time 8.0–8.5 s for reduction velocity 8 km/s). The crossover distance between crustal (*Pg*) and mantle refractions (*Pn*) is also much larger (200–220 km; Figure 10). Waves generated in this region are well recorded, particularly toward the northeast in the EEC (Figure 10a).

[19] As observed in the POLONAISE'97 data [*Grad et al.*, 2003a], the *P* wave field recorded on the EEC is uniform with low signal-to-noise ratios (e.g., Figures 11 and 12). Because of the thin sedimentary cover, the *P*sed phase produces first arrivals only to offsets of 1–10 km from the shot point. The *Pg* phase has a uniform character and is observed in the first arrivals up to ~200 km offsets, with apparent velocities increasing with distance from 6.0–6.2 to 6.6–7.0 km/s. These crustal waves continue to overcritical distances of up to 250–350 km (e.g., Figure 11). Relatively small amplitudes characterize waves from the middle crust, which indicates small contrasts at seismic boundaries and small velocity gradients within layers. Waves reflected from the Moho (*PmP*) are the dominant arrivals in the offset interval of 90–120 to 200–250 km. *Pn* arrivals have small amplitudes (e.g., Figures 11 and 12), and their apparent velocities are 8.1–8.2 km/s. The intercept time for *Pn* occurs at a reduced time of ~8 s, which is ~2 s more than for the Pannonian crust. For a number of shot points, well-recorded mantle lithospheric waves were observed up to 400–500 km offsets (e.g., Figures 10a, 11, and 12) and even to 800–1200 km. For the northern shot points, an unusual variety of *P* and *S* wave phases was recorded, and these phases are shown in Figures 4, 13, and 14. For shot points SP25280 and SP25290, *P* and *S* waves were doubly reflected from the Moho, and good quality lower lithosphere phases P^{I} and S^{I} are observed (Figures 13 and 14).

4. Derivation of Crustal Models Using Tomographic Inversion

[20] The initial model for profile CEL05 was derived using two-dimensional (2-D) tomographic inversion of the

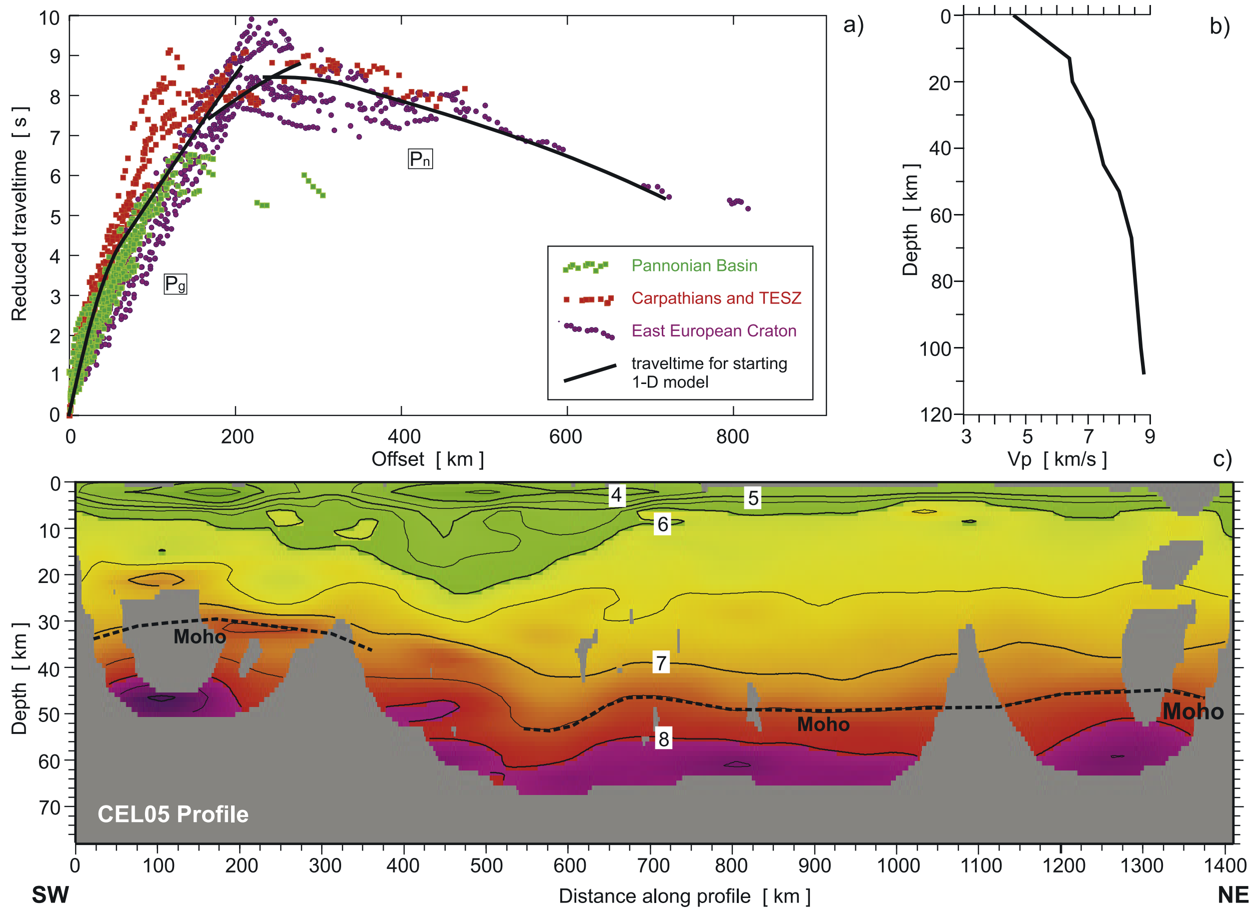

Figure 6. Result of two-dimensional tomographic inversion of *P* wave first arrival travel times for profile CEL05. (a) *P* wave first arrival travel time picks for the Pannonian basin (green), Carpathians and TESZ (red), and East European craton (pink) with first arrival travel time (black solid line) for (b) the starting 1-D velocity model. The complexity of the crustal structure along this profile is illustrated by 3–4 s of deviation of first arrival travel times along most of the observed range of offsets. Such a deviation within continents is large and reflects the difference between a crustal thickness of 25 and 50 km. Reduction velocity is 8.0 km/s. (c) Two-dimensional *P* wave velocity model along profile CEL05 obtained using the tomographic inversion program package FAST [*Zelt and Smith*, 1992]. Numbers are *P* wave velocities in km/s. The Moho is represented by velocity isoline of 7.5 km/s. Vertical exaggeration is ~5.7.

first-arrival travel times. Phases picked were not included in the inversion if there was any reasonable doubt about them being first arrivals. Over 1530 picks from 26 shot points were selected for inversion. All picks are shown in Figure 6a, together with the mean travel time, and the 1-D starting model (Figure 6b). The data set is divided into three groups of picks, corresponding to the Pannonian basin, Carpathians and TESZ, and East European craton (Figure 6a). The complexity of the crustal structure beneath profile CEL05 is illustrated by the ~4 s deviation of first-arrival travel times along most of the offset interval with observed arrivals.

[21] The tomographic inversion program FAST [*Zelt and Barton*, 1998] was employed. The forward calculation of travel times and ray paths uses the scheme of *Vidale* [1990], with modifications by *Hole and Zelt* [1995]. The velocity distribution derived from first-arrival travel time tomography is shown in Figure 6c (the RMS error for this model was 0.193). Tomographic inversion does not define velocity discontinuities well. Thus a discontinuity like the Moho should be represented by the isoline that is the average of the velocities above and below it. Therefore we believe that the Moho is best represented by the ~7.5 km/s velocity isoline not the actual value expected for the uppermost mantle (~8.1 km/s). Using this approach, the depth of the Moho beneath the Pannonian basin (0–300 km along profile) is only ~30 km while beneath EEC it reaches 40–45 km, and it is even over 50 km in the TESZ region (500–600 km along the profile). The average velocity for the deepest sediments (~5.8 km/s) and crystalline basement (~6.1 km/s) is close to 6 km/s, so this velocity isoline approximately delineates the sediment – basement contact. As a consequence, a deep sedimentary basin is indicated in the distance interval 250–650 km along the profile, where

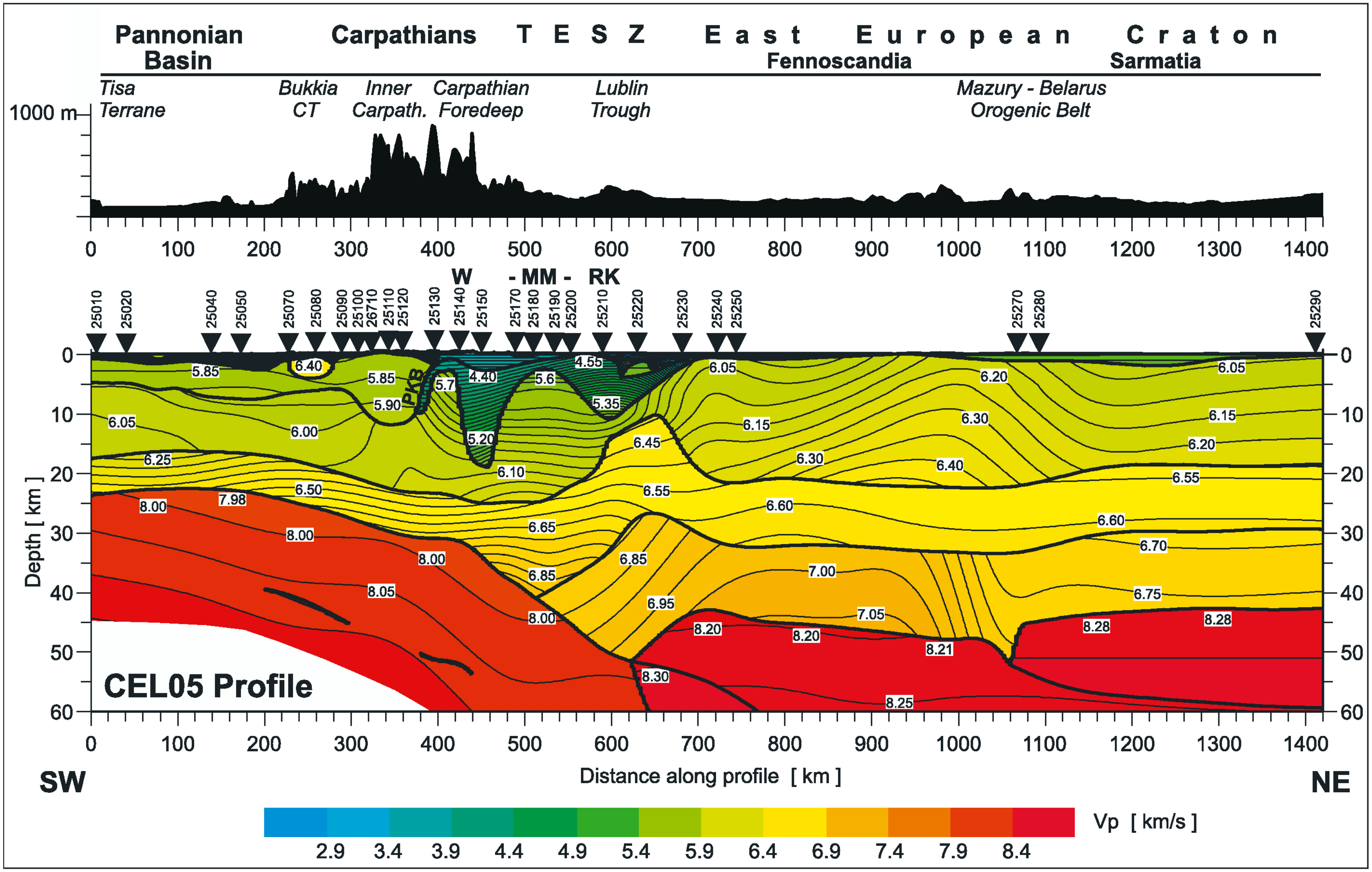

Figure 7. Two-dimensional *P* wave velocity model for CELEBRATION 2000 profile CEL05 obtained by forward ray-tracing modeling using the SEIS83 package [*Červený and Pšenčik*, 1983]. The thick solid lines are layer boundaries, and thin lines are isovelocity contours in km/s; numbered triangles refer to shot points. At the top of the model, topography along the profile is shown. Bukkia CT, Bukkia Composite terrane; RK, Radom–Kraśnik uplift (at 580–600 km of profile); W, Wysowa (at ~420 km of profile). Crossing points with CEL03 and CEL04 profiles are at SP25210 and SP25110, respectively. Vertical exaggeration for the model is ~6.8.

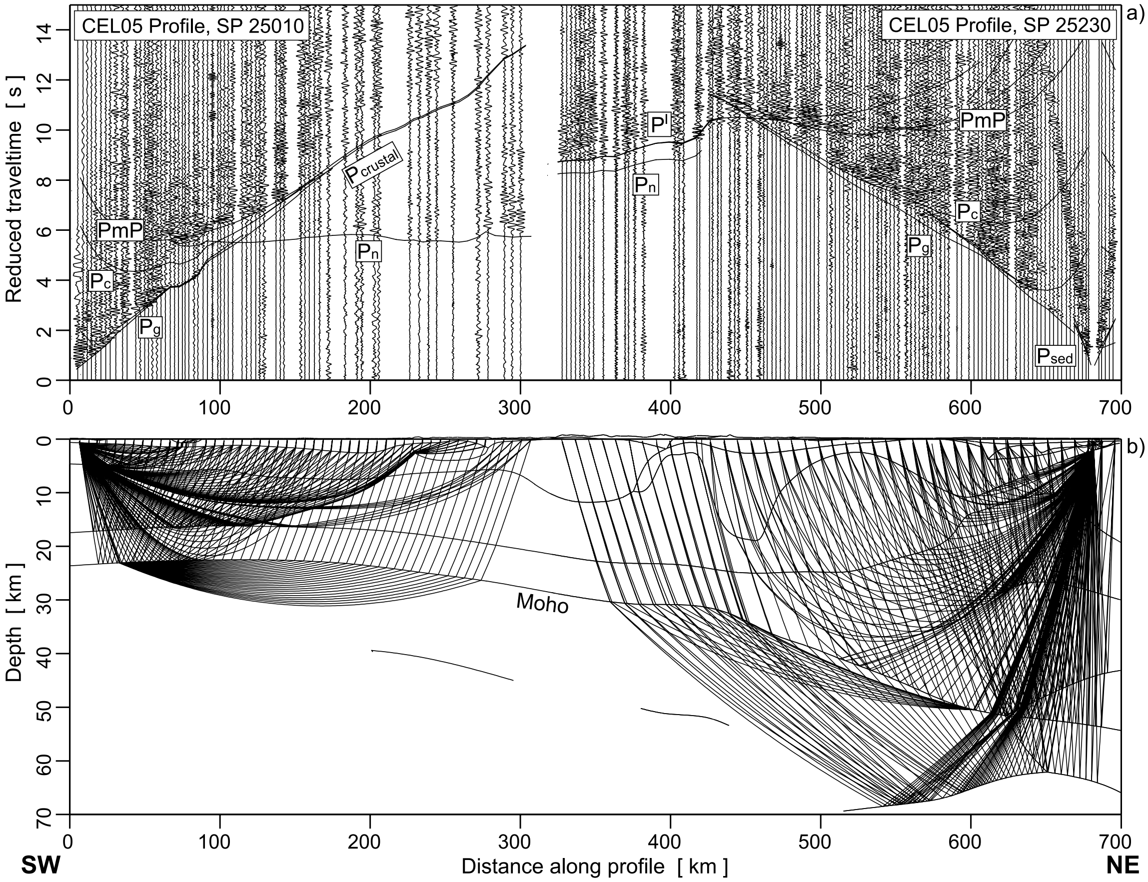

Figure 8. (a) Amplitude-normalized seismic record sections for SP25010 and SP25230 with theoretical travel times of P waves calculated for the model of the crust derived using the SEIS83 ray-tracing technique and (b) ray diagrams calculated for the model of the crust. Phases labeled as in Figures 3 and 4. Note an almost 2 s difference in Pn wave arrivals between the SW (Pannonian) and NE (Carpathians) parts of the section.

the 6.0 km/s isoline reaches ~20 km in depth. The model obtained in this inversion was used as a guide as we modeled the full wave field using ray tracing and synthetic seismograms.

5. Forward Modeling by Ray Tracing and Synthetic Seismograms

[22] Thanks to years of petroleum exploration and geologic studies, the initial model for the sedimentary cover and shallow basement was constrained by borehole information and earlier geophysical studies, including high-resolution seismic reflection surveys. This information provides a more detailed model of the uppermost structure (up to 3–7 km depth) than can be obtained from the CELEBRATION 2000 refraction data alone. With the uppermost crust constrained and with the tomographic model as a guide, all refracted and reflected phases identified in the correlation process were modeled in detail using 2-D ray-tracing and synthetic seismograms. The velocity model derived for the structure along profile CEL05 is shown in Figures 7 and 17 and reveals large variations in the structure of the crust and lithospheric mantle. Figure 7 includes a vertically exaggerated depiction of the topography traversed by the profile. Along most of its length, the profile passes through lowland areas with an average elevation of 200 m. In the Carpathians (200–500 km along profile), elevations range from 400 to 900 m.

[23] In the Pannonian basin region to the southwest (0–350 km), the crustal structure is relatively simple. However, in the distance interval of ~240–280 km along the profile, a shallow high-velocity body was delineated (V_p ~ 6.4 km/s). Beneath the upper sedimentary layer and a layer with velocities of ~5.8 km/s, two almost homogeneous (very small velocity gradient) crustal layers are observed. The velocity increases from ~6.1 to 6.2 km/s in the depth interval 5–18 km in the first layer and from 6.3 to 6.5–6.6 km/s to a depth of ~25 km in the second one. This part of the crust is well documented in the record sections shown in Figures 3a, 8a (left part), and 9a (left part). In particular, overcritical crustal arrivals at offsets of 120–300 km in Figure 8a document the relatively low velocities in the lower crust of the Pannonian basin region. The Moho here lies at a depth of only 24–25 km, and the uppermost mantle beneath it is characterized by velocities of 7.95–8.0 km/s.

[24] The most complicated structure is observed at distances of 350–700 km in the transition from the Pannonian basin the EEC. This transition includes the Carpathians and their foredeep, the Małopolska massif,

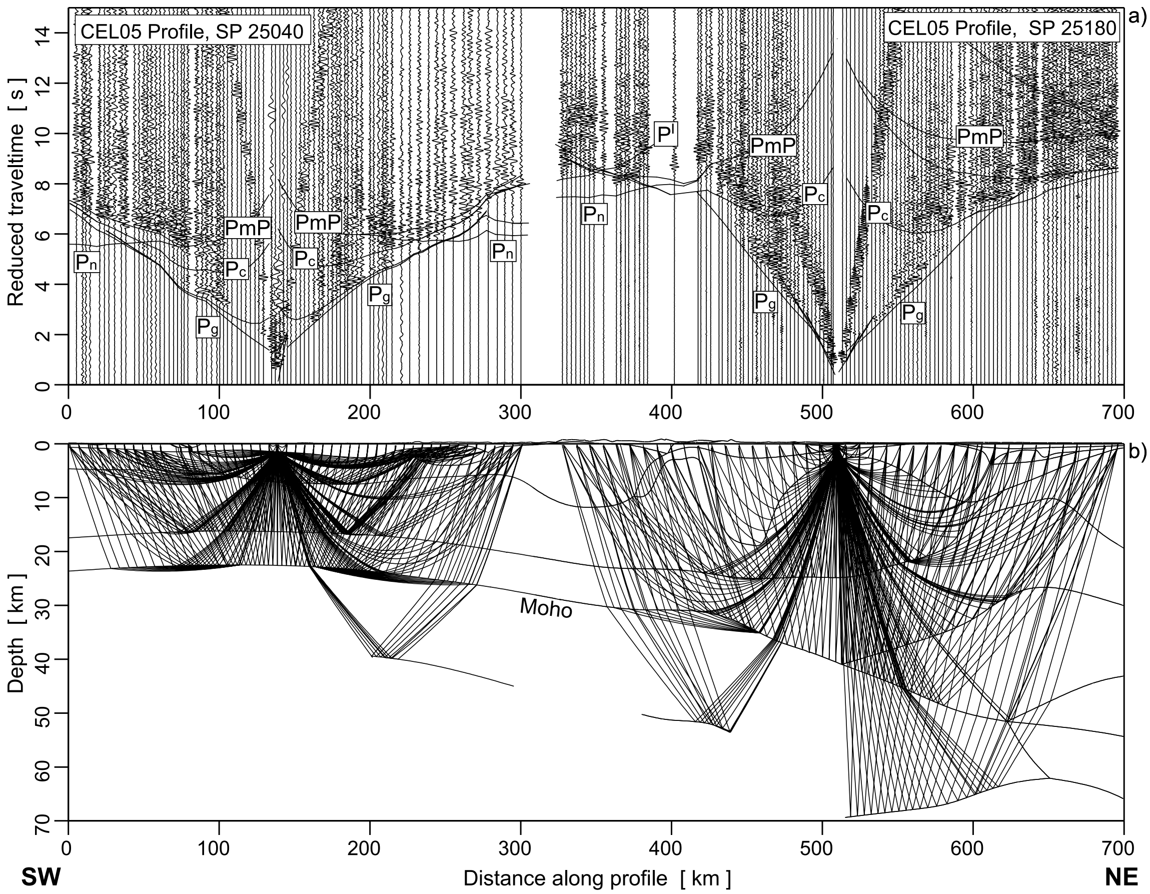

Figure 9. (a) Amplitude-normalized seismic record sections for SP25040 and SP25180 with theoretical travel times of *P* waves calculated for the model of the crust derived using the SEIS83 ray-tracing technique and (b) ray diagrams calculated for the model of the crust. Phases labeled as in Figure 3. Note an almost 2 s difference in *Pn* wave arrivals between the SW (Pannonian) and NE (Carpathians) parts of the section.

the rifted margin of the EEC, and the Lublin trough. This is an area of complex topography, complicated structure in the sedimentary cover (where rocks with velocities <5.5 km/s reach depths of ~20 km), complex structure within the crystalline crust, and deep Moho (~50 km). In the distance interval 370–400 km, a slab of low velocity sediments a few kilometers thick dipping to the southwest can be traced down to ~ 10 km. Beneath the Carpathians, a two-layer crystalline crust is present with velocities of 5.9–6.2 and 6.5–6.8 km/s, respectively, while beneath the TESZ and EEC margin farther to the northeast, a three-layer crust is present. A midcrustal reflector beneath the Carpathians at a depth of 22–26 km is well documented by the reflected phase *Pc* (Figures 15a, 15b, 16a and 16b). In the distance interval 600–670 km along the profile, a dome of relatively high velocities (~6.4 km/s) protrudes into the upper crust and may represent an intrusion. In the same area, an uplift of the lower crust is also observed. The maximum crustal thickness along the profile (52 km) is observed in the area of the crossing point with profile CEL03 (at distance 600–650 km), where a similar velocity distribution in the crust was also found [*Janik et al.*, 2005]. The complicated structure between the Pannonian basin and the EEC is documented in Figures 8 and 9 where the difference between arrival times of *Pn* waves in the southwest and the northeast exceeds 2 s. The location of the deepest Moho is based on arrivals from shot points SP25230 and SP25180, which are shown in Figures 8a and 9a, respectively. The uppermost mantle velocity in this area is ~8.0 km/s.

[25] Farther to the northeast (700–1420 km), the crustal structure of the EEC is typical for cratonic areas, with a thin sedimentary cover and three-layer crystalline crust similar to other areas in the western part of Baltica [*Grad and Luosto*, 1987, 1994; *Grad and Tripolsky*, 1995; *EUROBRIDGE Seismic Working Group*, 1999, 2001]. The upper crust has a thickness of 15–20 km and velocities of 6.0–6.4 km/s. The middle crust is ~10 km thick and has typical velocities of 6.5–6.7 km/s. The lower crust has a thickness of 12–15 km and velocities of 6.7–7.0 km/s [e.g., *Grad et al.*, 2003a]. The depth of the Moho varies in the interval 42–48 km over the entire ~700 km long cratonic part of profile, and velocities in the uppermost mantle vary only slightly. They are ~8.20 km/s in the southwestern part of the craton (700–1000 km) while in the northeast they are

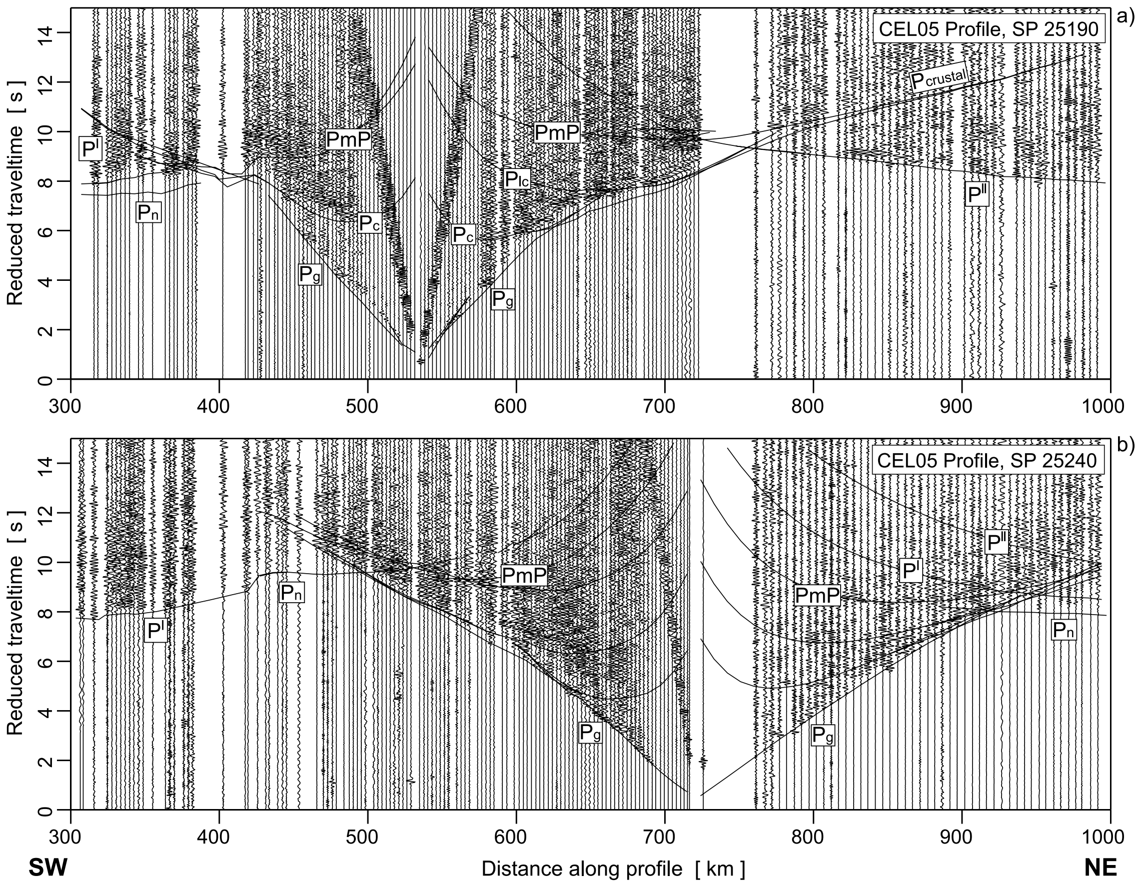

Figure 10. Amplitude-normalized seismic record sections for (a) SP25190 and (b) SP25240 with theoretical travel times of *P* waves calculated for the model of the crust derived using the SEIS83 ray-tracing technique. Note the lower lithosphere phase P^{I}. Other phases are labeled as in Figures 3 and 4.

slightly higher (~8.25 km/s). The location of this small change at ~1050 km along the profile correlates with small changes in the structure of the entire crust. In the southwestern part of the EEC, the upper crust is slightly thicker (2–4 km), and velocities in both the upper and lower crust are slightly higher (by 0.15–0.20 km/s) than to the northeast. The thickness of the middle crust, the velocity of which is 6.55–6.65 km/s, is almost constant across the EEC.

[26] Along the profile, many shot points produced well-recorded waves reflected in the lower lithosphere (P^{I} and P^{II} phases), and the corresponding reflectors were modeled as shown in Figures 7 and 17. Examples of these phases are shown in Figures 4 and 8–14. In the Pannonian-Carpathian area, the P^{I} wave was recorded at offsets of 150–450 km. This wave is usually much stronger than the *Pn* wave in the same distance range (e.g., SP25230 in Figure 8a at 330–420 km; SP25040 and SP25180 in Figure 9a at 250–400 km; SP25190 in Figure 10a at 300–380 km). The corresponding reflector in the lower lithosphere is subparallel to the Moho but is ~15 km deeper. Thus it dips northward from the Pannonian basin under the Carpathians (Figures 7 and 17). In the TESZ area (520–650 km), a very pronounced, south dipping reflector was found at a depth of 60–70 km that indicates a large positive velocity contrast (8.1 and 8.45 km/s between the Pannonian and Baltica mantle, respectively). In the EEC area, two phases (P^{I} and P^{II}) reflected in the lower lithosphere were clearly observed (Figures 4 and 10–12). The depths of the corresponding reflectors are 75–80 and ~95 km. Some short reflecting horizons were also found beneath the EEC at depths of ~55 and 60–70 km.

[27] In order to facilitate the estimation of errors in the velocity structure derived, Figures 8–16 show seismic record sections with calculated travel times for the final model of the structure (Figures 7 and 17) superimposed. In most cases, theoretical travel times fit the observed travel times for both refracted and reflected waves to within ±0.10–0.15 s. In addition, synthetic seismograms show good qualitative agreement with relative amplitudes of observed refracted and reflected waves (Figures 11 and 12). The high signal-to-noise ratio of the data, the number of shot points, and the agreement between the kinematic and dynamic properties of the main phases observed provides confidence that the uncertainties of the velocities and depths to discontinuities shown in Figures 7 and 17 are low relative to typical studies of this type. However, the limitations of 2-D modeling must be kept in mind.

[28] Bases on many iterations in the modeling and sensitivity analysis, we estimate that for the upper crust

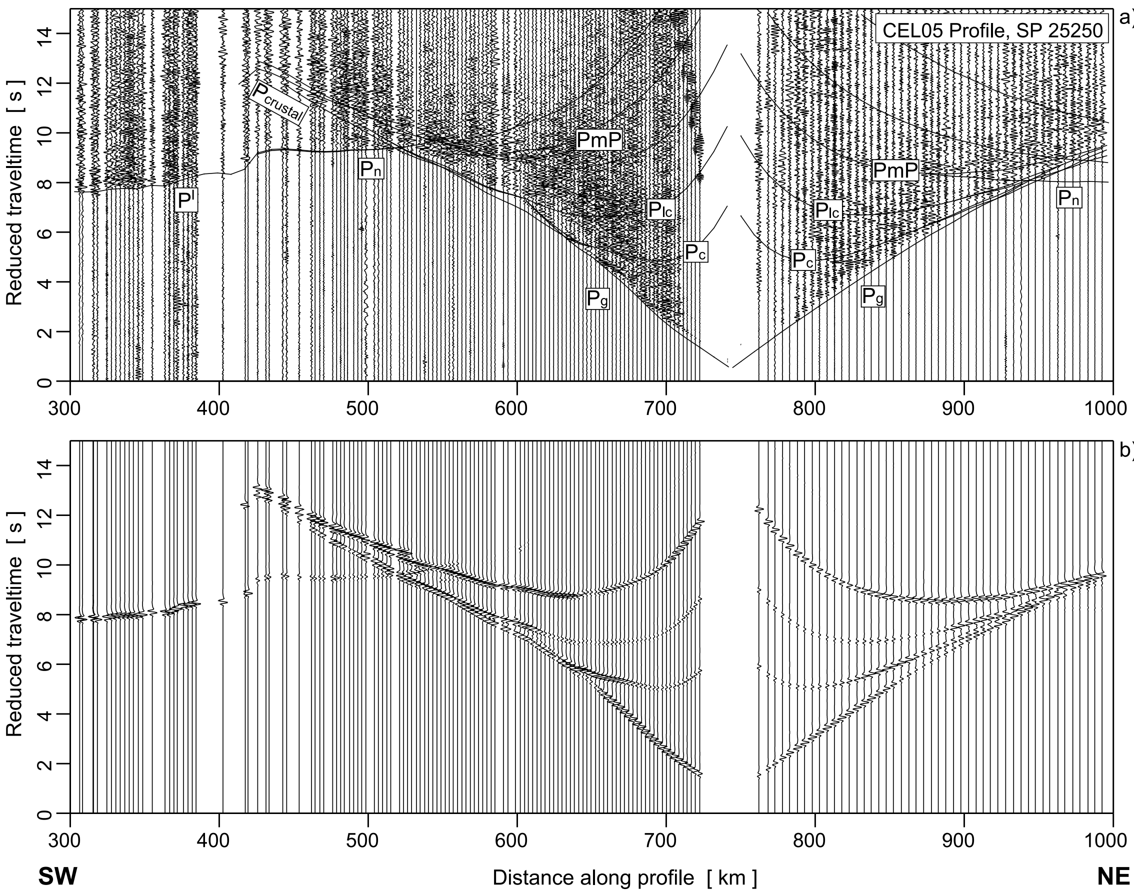

Figure 11. (a) Amplitude-normalized seismic record section for SP25250 with theoretical travel times of *P* waves and (b) synthetic seismograms calculated for the model of the crust derived using the SEIS83 ray-tracing technique. Note the overcritical crustal phase *P*crustal and the lower lithosphere phase P^{I}. Other phases are labeled as in Figures 3 and 4.

(crystalline basement) where the coverage by *Pg* waves is highest, the velocity uncertainty is ±0.1 km/s, and we estimate that the uncertainty for the velocity of the uppermost mantle to only be slightly higher in most regions of the model. Because waves refracted from the lower crust are rarely observed in the first arrivals, the uncertainty of the velocities determined here is lower, namely ±0.15–0.20 km/s. In many cases, this situation is improved because of well-recorded overcritical crustal waves (*P*crustal) that penetrate the lower crust. The depths of midcrustal boundaries are usually determined within ±2–3 km, and better ray coverage for the Moho constrains it to ±1–2 km. Similar estimates of uncertainty were obtained earlier from the analysis of POLONAISE'97 profiles, which are characterized by similar methodology, source and receiver density, and data quality [*Janik et al.*, 2002; *Grad et al.*, 2003a], and are in line with estimated values based on many such studies [e.g., *Mooney*, 1989].

6. Geological Interpretation, Tectonic Models, and Discussion

[29] Profile CEL05 is important for understanding the tectonic relationships between the Carpathian-Pannonian region and the East European craton (Baltica). This region is part of a very complex collisional environment between the European and Adriatic plates, which involved a variety of microcontinents and oceanic areas [e.g., *Golonka et al.*, 2003a, 2003b]. The identification and determination of the extent of these terranes is controversial, because they have been subducted, obducted, and/or translated during subsequent events.

[30] Recent studies of the Carpathian-Pannonian region have interpreted it to consist of the ALCAPA (Alps-Carpathian-Pannonian) and Tisza-Dacia blocks that are separated by the Mid-Hungarian line (Figures 1 and 18) [e.g., *Bielik et al.*, 1998, 2004; *Szabó et al.*, 2004]. The southern part of profile CEL05 primarily crosses the ALCAPA block (distances of 0–390 km, Figure 7) that extends up to the Pieniny Klippen belt (PKB), which represents the structural boundary between Inner and Outer Carpathian units [e.g., *Kováč et al.*, 1993]. The first 250 km of the profile lies in the Pannonian basin where the crust is thin (∼25 km), and our velocity model indicates the area is covered by 2–3 km of Tertiary sedimentary strata, which is in agreement with estimates based on drilling and seismic reflection data [e.g., *Royden and Dövényi*, 1988]. The remainder of the crust has a low average velocity (∼6.1 km/s), which indicates that it is not oceanic in nature. The crust gradually thickens under the Carpathians, and from the Carpathians northward to

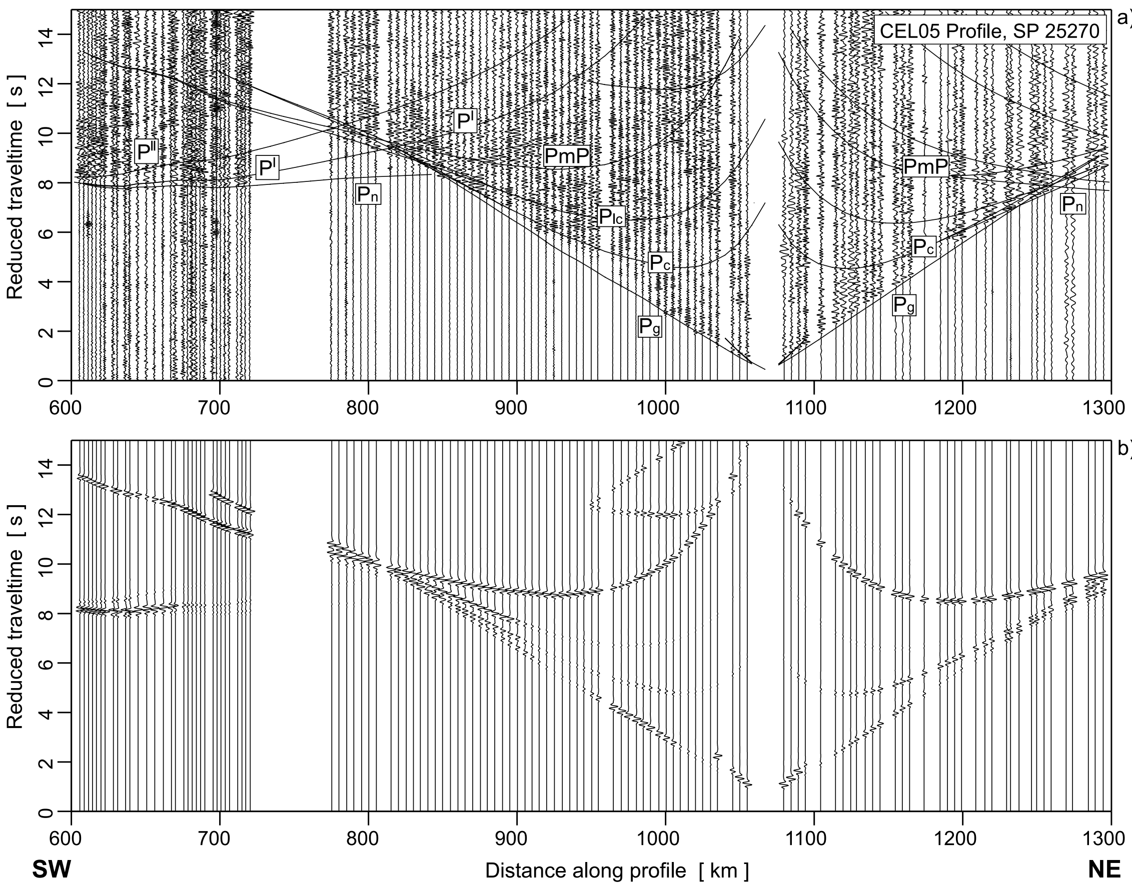

Figure 12. (a) Amplitude-normalized seismic record section for SP25270 with theoretical travel times of P waves and (b) synthetic seismograms calculated for the model of the crust derived using the SEIS83 ray-tracing technique. Note the lower lithosphere phases P^{I} and P^{II}. Other phases are labeled as in Figure 3.

the margin of the EEC (550 km, Figure 7) the profile primarily crosses the Małopolska massif (MM). This massif consists of folded metasediments of Vendian–Early Cambrian age, covered by younger Paleozoic to Neogene platform sediments. The origin of the MM is much debated, but it is probably a lithospheric terrane that was once part of Baltica [*Dadlez et al.*, 1994; *Janik et al.*, 2005]. Our results show that the southern part of the MM is bounded to the south by the Carpathian foredeep and on the north by the margin of the EEC adjacent to the Lublin trough (Figures 7 and 17). Beginning at ~650 km, the profile traverses the Precambrian East European craton (EEC) that formed from the accretion of three major lithospheric terranes: Fennoscandia, Sarmatia and Volgo-Uralia. These features are interpreted as large composite terranes, each with an independent history during Archean and Early Proterozoic times [*Bogdanova*, 1996; *Bogdanova et al.*, 2001]. The boundaries between them are marked by Mesoproterozoic to Neoproterozoic intracratonic rift systems. In western Belarus, profile CEL05 approximately follows the suture between Fennoscandia and Sarmatia. Specifically, it crosses the Belarus pre-Cadomian granulite orogenic belt (1.9–1.85 Ga), Central Belarus orogenic belt, and Vitebsk granulite zone (ca 2.0–1.9 Ga).

[31] The portions of the Pannonian basin and the Carpathians traversed by profile CEL05 have been divided into the following structural units based on geological data (Figure 18) [*Kovács et al.*, 2000; *Rakus et al.*, 1998]: (1) the Tisia block south of the Mid-Hungarian line; (2) a central unit consisting of the Pelso structural block and the Alpine-type nappe structures of the Inner Carpathians (Tatricum, Veporicum, Zemplinicum, and Gemericum); and (3) a northern unit consisting of the Outer Carpathians and their foredeep north of the Pieniny Klippen belt. The bottom of the nappe structures is probably coincident with the bottom of the second crustal layer in Figure 7 (distance 340–360 km) at the depth of 19–20 km. However, as expected, these complex units cannot be resolved individually by CEL05 data. The presence of these structures was inferred by *Vozár et al.* [1999] based on the interpretation of deep seismic reflection profiles. To the south in the Bukkia Mountains (~250 km, Figure 7), the lens of higher velocity rocks (6.45 km/s) can be connected to the occurrence of subvolcanic bodies associated with the Mátra volcano. *Hajnal et al.* [2004] found similar high-velocity bodies in the region to the east in a more detailed seismic study that was part of the CELEBRATION 2000 effort. Just north of this lens, the increase in the thickness of the first layer (V_p = 5.80–5.90 km/s) at a distance of ~300 km is probably linked with the Diósjenő line and reflects the internal structure of the Inner Carpathian (IC) units. The northern boundary of the IC is delimited by the Pieniny Klippen

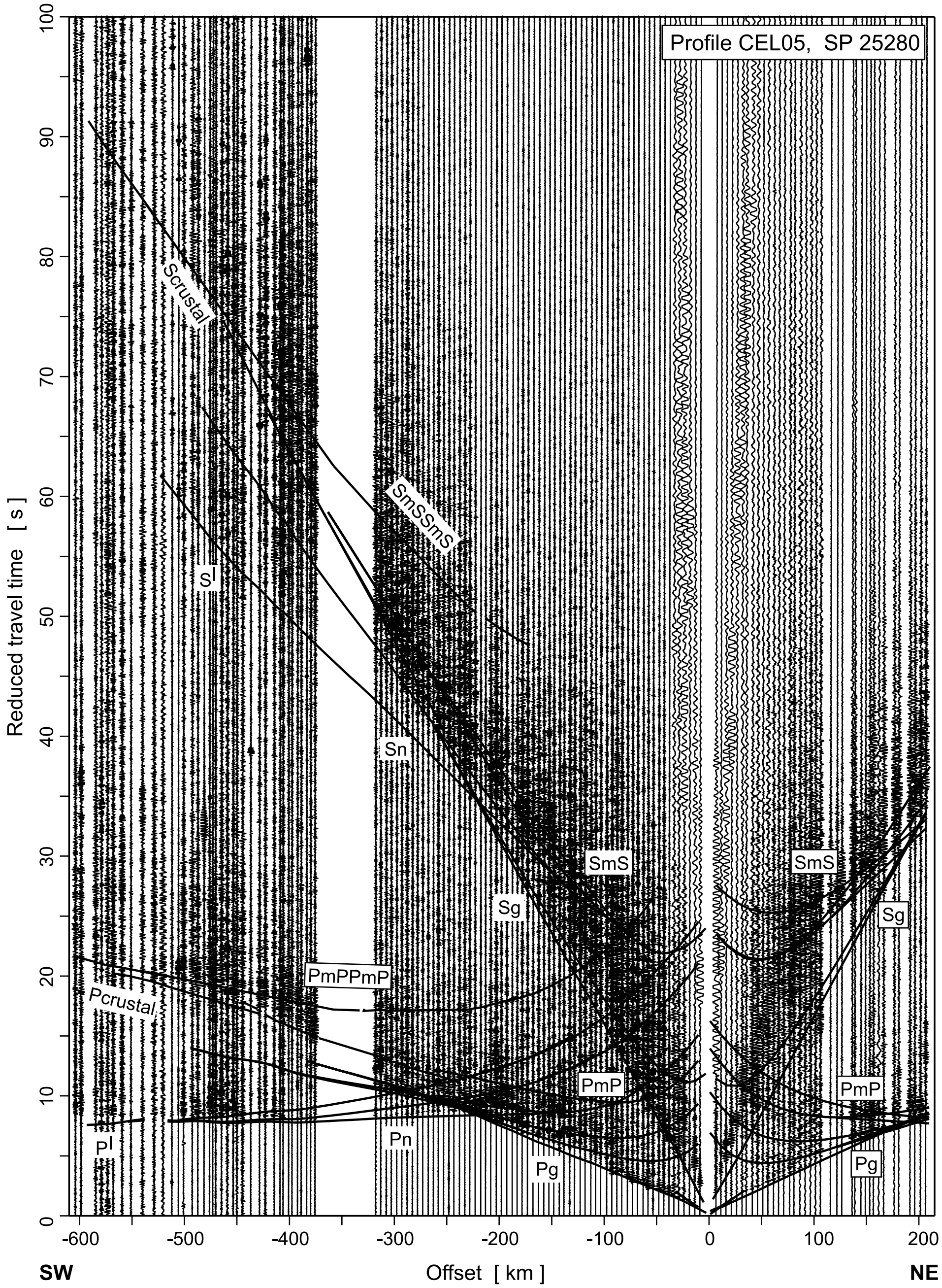

Figure 13. Amplitude-normalized seismic record section for SP25280 showing the full wave field, including both *P* and *S* waves with theoretical travel times calculated for the model of the crust derived using the SEIS83 ray-tracing technique. Note the *PmPPmP* and *SmSSmS* waves that are reflected twice from the Moho discontinuity and free surface and high-quality lower lithosphere phases recorded to long offsets. Reduction velocity is 8 km/s.

belt, which is a prominent regional detachment (shear) zone that dips southward at 35–40° and has also experienced left-lateral displacement [e.g., *Royden*, 1988]. The PKB can be traced to a depth of 10 km as a narrow belt of low velocities on Figure 7. The continuation of the PKB can be interpreted to be approximately parallel to the 6.05 km/s velocity isoline within the second layer and reaches the boundary between second and third layers at a depth of ~20 km.

[32] The PKB has been interpreted to be related to a southward dipping suture zone [e.g., *Birkenmajer*, 1986; *Šefara et al.*, 1998]. There is no indication of south dipping interfaces below a depth of ~20 km (300–400 km distance in Figure 7), but the detachment could be nearly horizontal beginning at this depth. The presence of fragments of oceanic crust is another issue of tectonic significance, and besides the high-velocity body (6.45 km/s at ~250 km) discussed above, the only feature that could be

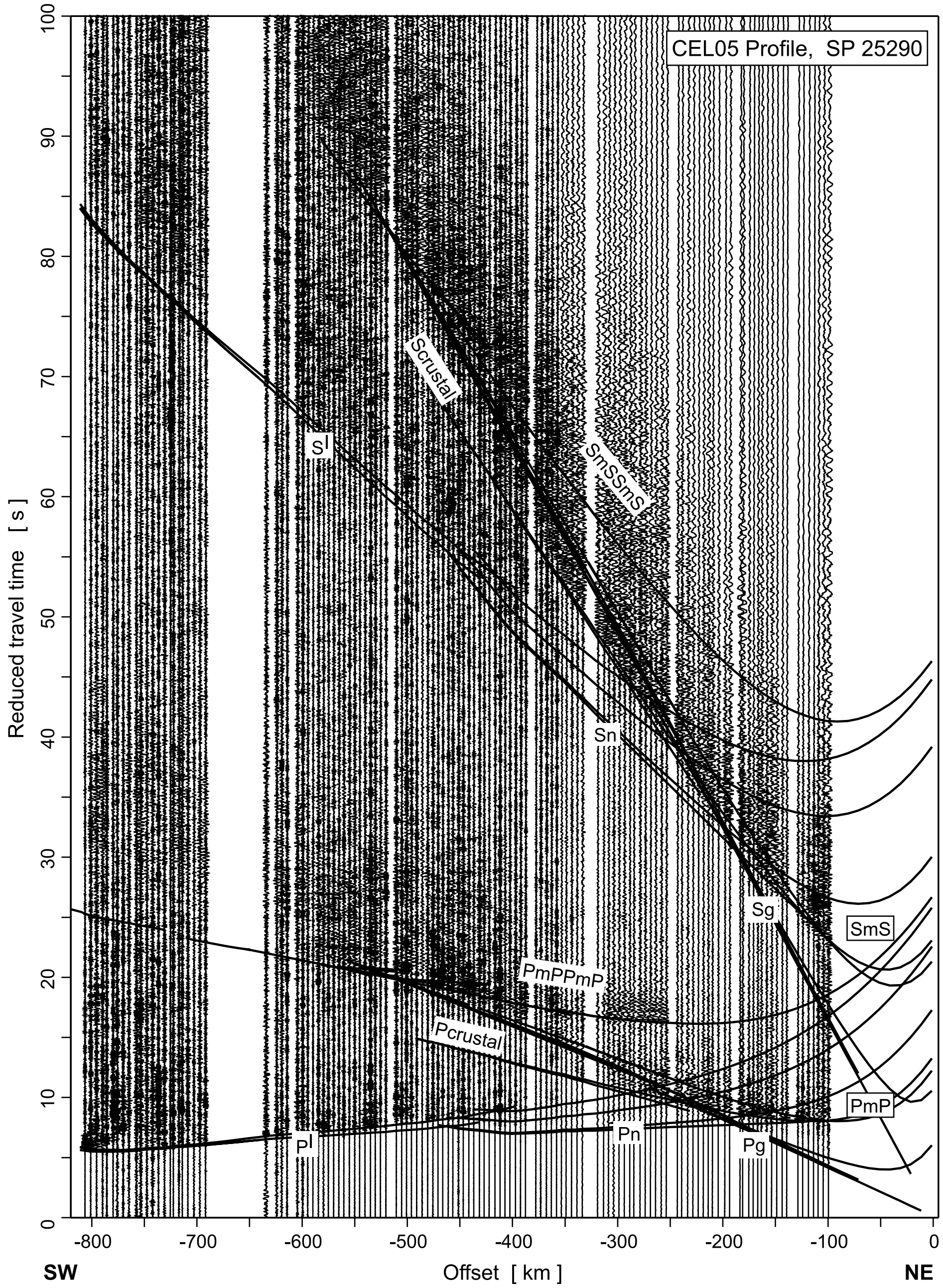

Figure 14. Amplitude-normalized seismic record section for SP25290 showing the full wave field, including both *P* and *S* waves with theoretical travel times calculated for the model of the crust derived using the SEIS83 ray-tracing technique. Note the *PmPPmP* and *SmSSmS* waves reflected twice from the Moho discontinuity and free surface and high lower lithosphere phases recorded to long offsets. Reduction velocity is 8 km/s.

interpreted as a remnant oceanic crust is a wedge-shaped body at the base of the crust under the Małopolska massif (500–600 km distance in Figure 7). This body could represent relict Penninic Ocean crust subducted northward during the Late Cretaceous, as was proposed by *Książkiewicz* [1977] and *Golonka et al.* [2003a, 2003b].

[33] An interesting observation is the clear crustal thickening from the Pannonian basin to the TESZ region. Taken at face value, the configuration in Figures 7 and 17 suggests northward subduction of a Carpatho-Pannonian plate toward the north under the European plate. However, this direction of subduction conflicts with strong geological evidence from surface mapping and shallow boreholes for southward subduction during the Cenozoic [e.g., *Birkenmajer*, 1976, 1986; *Royden*, 1988; *Plašienka et al.*, 1997; *Bielik et al.*, 2004]. The smooth velocity variations in Figure 7 suggest that the lithosphere under

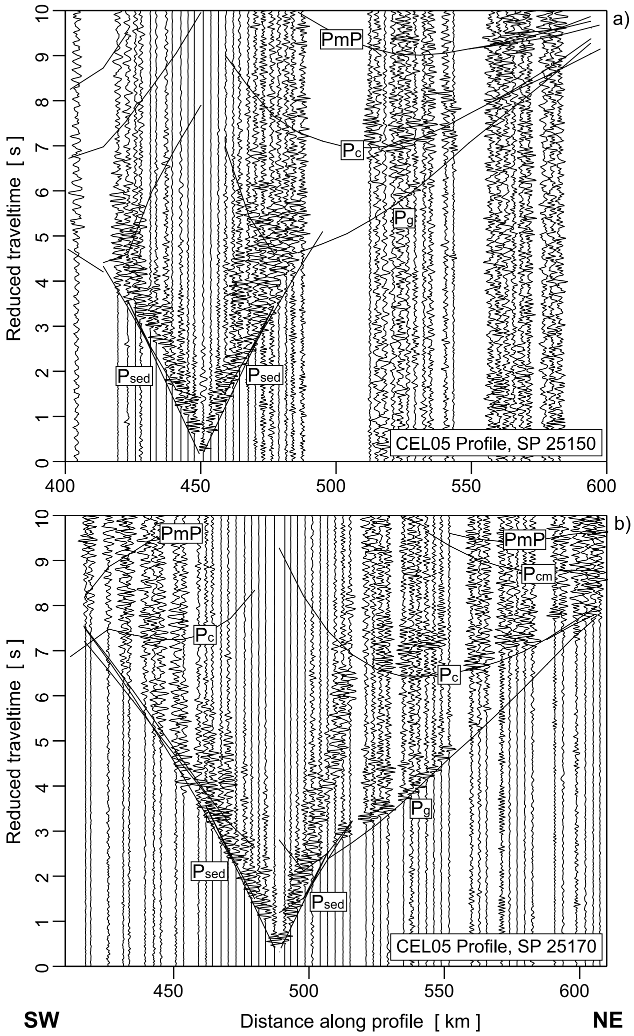

Figure 15. Amplitude-normalized seismic record sections for (a) SP25150 and (b) SP25170 with theoretical travel times of *P* waves calculated for the model of the crust derived using the SEIS83 ray-tracing technique. Note long branches of the *P*sed phase refracted in the sedimentary cover and the strong reflected phase *Pc* from the middle crust. Other phases are labeled as in Figure 3.

the Pannonian basin region has thinned due to the effects of extension and heating, which is documented by high heat flow in the Carpathian/Pannonian area. In this case, the northwestward dip in the upper mantle in this region (Figure 7) can be interpreted as being due to thermal effects such as small amounts of partial melt following isotherms and could even represent the lithosphere/asthenosphere boundary. Recent studies have integrated the data available prior to the CELEBRATION 2000 experiment to produce maps of the lithospheric thickness in the region [*Bielik et al.*, 2004; *Szabó et al.*, 2004] that indicate northward lithospheric thickening from the Pannonian basin region. However, the Pannonian basin lithosphere is ~60 km thick in these maps, which is 10–20 km deeper than the mantle reflectors we have identified.

[34] The structure of the upper crust is complex in the region between the Carpathians and the EEC. The layers with velocities between 4.00 and 5.35 km/s represent the sediments of the Carpathians, the foredeep, and the Mesozoic-Paleozoic cover of the southern EEC and adjacent Małopolska massif. The boundary between units with velocities of 4.00 and 4.50 km/s visible in this area has disputable origin, since it cuts across different tectonic units. An uplift of basement rocks (velocities ≥5.7 km/s; 400–420 km distance in Figure 7) rises to a depth of less than 5 km in Slovakia near SP25130, and is ~9 km deep near Wysowa, beneath SP25140. An uplift of high-resistivity material near Wysowa was also observed in magnetotelluric data [*Ryłko and Tomaś*, 1999; *Żytko*, 1999], and its summit was at a depth of 10 km. In addition, a deep (5319 m) borehole nearby (Smilno 1 [*Leško et al.*, 1987]) does not penetrate Oligocene-Eocene rocks. Flysh rocks penetrated by another deep borehole (Hanusovce 1; 5833 m) near the PKB [*Leško*, 1985] had velocities ranging from 4.1 up to 6.0 km/s. Therefore it is probable that a part of this uplift, probably down to ~10 km, represents

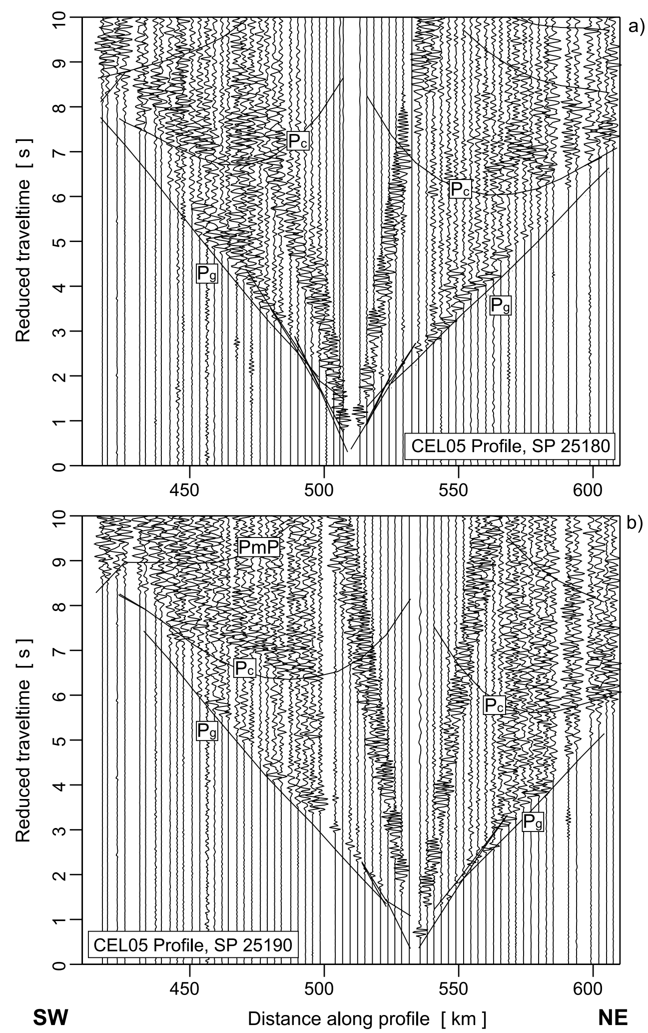

Figure 16. Amplitude-normalized seismic record sections for (a) SP25180 and (b) SP25190 with theoretical travel times of *P* waves calculated for the model of the crust derived using the SEIS83 ray-tracing technique. Note the strong reflected phase *Pc* from the middle crust. Other phases are labeled as in Figure 3.

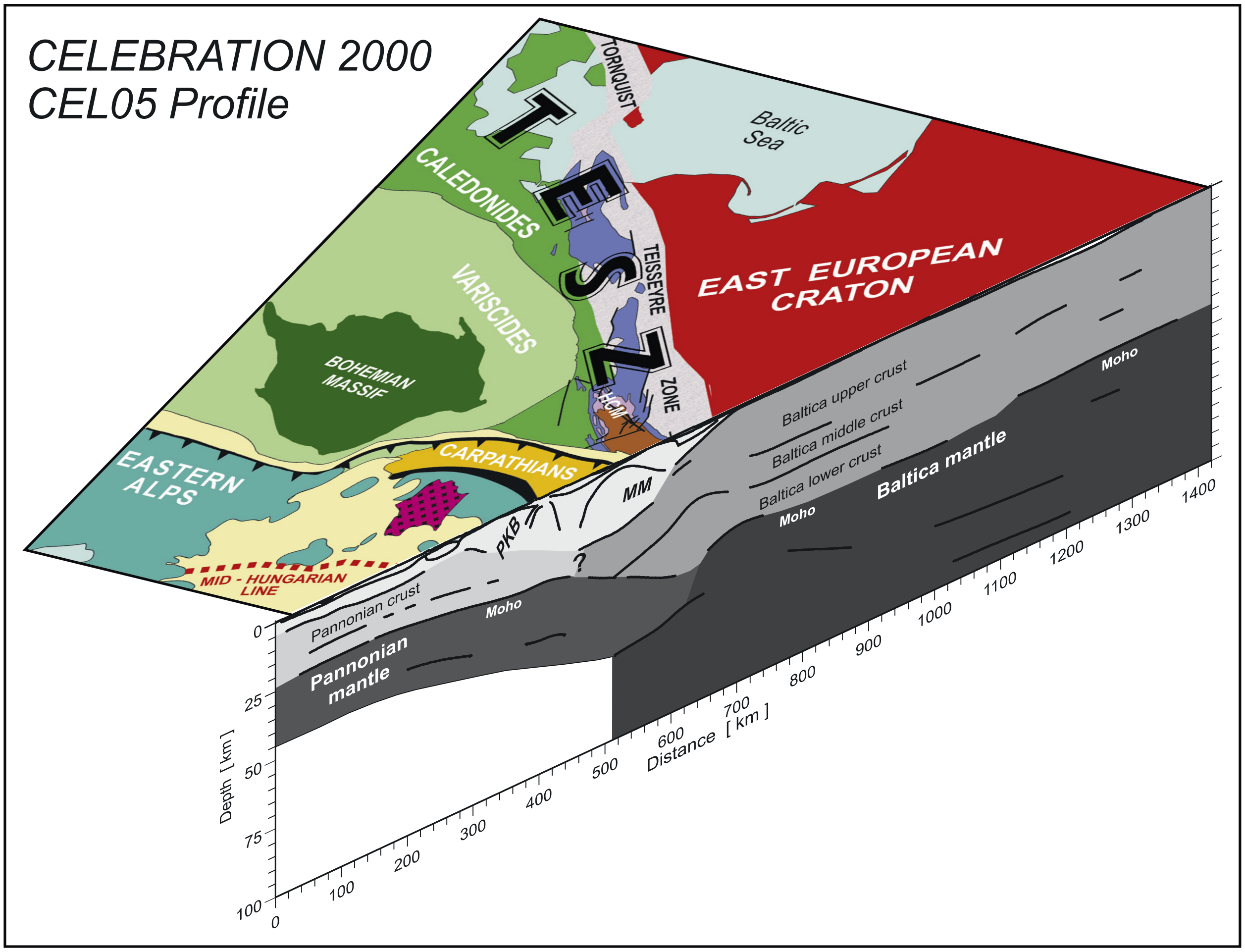

Figure 17. Lithospheric model beneath profile CEL05 down to 100 km depth together with tectonic map of central Europe. Note mantle lithospheric reflectors about 15 km deeper than the Moho, up to 75–80 and ~95 km depth. HCM, Holy Cross Mountains; MM, Małopolska massif; PKB, Pieniny Klippen belt. Vertical exaggeration for the model is ~3.75.

Carpathian flysch (the Magura Nappe) and not the crystalline basement beneath the Carpathians. The lower part of this uplift probably represents a more internal massif than the MM, and the foredeep (430–470 km) filled by sedimentary rocks marks the boundary between the massifs.

[35] Thickening of the sedimentary layer (V_p = 4.5–5.5 km/s) at ~530 km on the profile (Figure 7) reflects the Late Caledonian structures of the Łysogóry unit that is bounded to the north by the Radom–Kraśnik elevation (the uplift at ~590 km in Figure 7) [*Żelichowski and Kozłowski*, 1983]. The northern boundary of this unit coincides with an increase in thickness of Variscan units within the Lublin trough. A new interpretation of this area [*Antonowicz et al.*, 2003] indicates that the upper crustal features are all allochtonous and characterized by thin-skin tectonics with series of duplexes within the Radom–Kraśnik elevation lying structurally inboard the Lublin trough. At the bottom of the Lublin trough (620–640 km), the lens of rocks with velocities distinctly higher than in surrounding rocks cannot be confirmed by the relatively abundant deep boreholes in this area. The local uplift of the Moho and deeper crustal layers could be the result of bulging associated with a north dipping subduction zone or the crustal thinning originally associated with the rifting that broke up Rodinia and formed this margin of Baltica.

[36] Starting at ~650 km, profile CEL05 crosses into the Precambrian East European craton approximately along the suture zone between Fennoscandia and Sarmatia. This major lithospheric boundary appears to be manifested as gentle undulations of velocity isolines within the upper and lower crust, as well as by a change in uppermost mantle velocity (~8.2 and 8.3 km/s) at a distance of ~1050 km.

[37] The new results from profile CEL05 are important in studies of the tectonic evolution of the Carpathian-Pannonian area although they show only the present-day velocity structure. Our analysis indicates that there are several possible explanations for how tectonic events in the past are reflected in today's lithospheric structure. Three models and a possible tectonic interpretation of the area of profile CEL05 are shown in Figure 19.

[38] One major result of this study is that deepening of the Moho and the reflectors in the lower lithosphere from Pannonian basin through Carpathians to the EEC is suggestive of northward subduction under the EEC (Figure 19a). *Książkiewicz* [1977] postulated that subduction of the

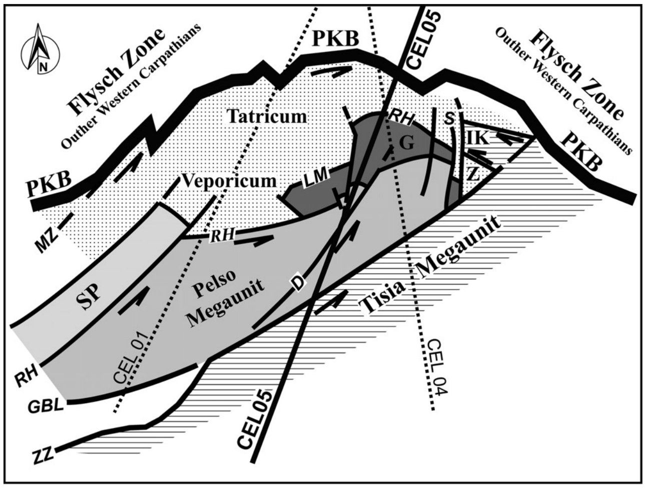

Figure 18. Distribution of the main tectonic units of the Inner Western Carpathians, Pelso block, and Tisia terrane. Alpine type nappes of the Inner Western Carpathians: Tatricum, Veporicum, Zemplinicum (Z), and Gemericum (G); SP, interpreted southern Penninicum below the Tertiary/Quaternary fill of the Danube basin; IK, Iòa-čovce/Kričevo tectonic units (below the Tertiary/Quaternary fill of the East Slovakian lowland) interpreted as northern part of Tisia terrane. Tectonic lines indicate PKB, Pieniny Klippen belt; MZ, Mur/Mürz/Záhorie seismic zone; RH, Rába/Hurbanovo/Diósjenő line; LM, Lubeník/Margecany structure zone; D, Darnó (or Diósjenő) line; S, Slaná/Sajó faults; ZZ, Zágreb/Zemplín (Mid-Hungarian line) tectonic zone [*Csontos and Nagymarosy*, 1998; *Kovács et al.*, 2000].

Pannonian lithosphere under the East European craton occurred during the Jurassic–Early (Lower) Cretaceous. In their paleogeographic reconstruction of the circum-Carpathian area *Golonka et al.* [2003a] also proposed that north-northwestward subduction of the Meliata-Halstatt Ocean crust was completed by the end of the Jurassic, ∼140 Ma, and that the location of this closure corresponds to the Mid-Hungarian line (Figure 18). However, we must note that the remnants of the Meliata Ocean exposed in the Carpathian area are not in situ. In fact, their emplacement occurred during large-scale Tertiary strike-slip movements [e.g., *Csontos and Vörös*, 2004] and does not represent an ocean between the Pannonian basin and the Inner Carpathians. Thus although upper crustal data have consistently been interpreted to favor southward subduction in the Cenozoic, a Mesozoic episode of northward subduction is also consistent with our geophysical and some geological data. This event would probably have been short-lived and the convergence was probably oblique because of the lack of a volcanic arc of this age.

[39] A model that emphasizes collisional features instead of subduction is shown in Figure 19b. The lithospheric-scale "crocodile" structure in this model would be the result of the Carpathian-Pannonian upper crust obducting over the crystalline crust of the EEC and Carpathian-Pannonian mantle lithosphere underthrusting the cratonic crust. This model is consistent with the upper crustal geology in that it produces south dipping, north vergent features (detachments) beginning at the EEC margin and extending to the PKB and beyond. Near the EEC margin, some of these structures could even be reactivated Variscan features. This model could be applicable to a phase of modest Cenozoic convergence and thus would not be inconsistent with Mesozoic northward subduction.

[40] The third model (Figure 19c) explains the observed structure as the result of "young" (Miocene) southward subduction of Variscan and EEC lithosphere and of Neogene thinning of the Pannonian lithosphere due to extension. Our data show no evidence of a south dipping slab in the upper mantle or south dipping features in the lower crust from the Carpathian foredeep southward. However, one could imagine the south dipping features adjacent to the EEC (500–700 km) being attached to a south dipping slab that broke off and either sunk to great depth and/or was absorbed.

[41] Given the complexity of the tectonic evolution of the region (Figure 18), more analysis of other CELEBRATION 2000 data and additional integration will be required before a clearer understanding of the implications of these data will emerge. We presently favor a model in which the area of thick crust between 500 and 700 km along the profile was produced by northward, probably Mesozoic, convergence via thickening of the lower crust, while the Moho geometry to the south is primarily due to Neogene extension and heating in the Pannonian basin region. Profile CEL05 crosses the western Carpathians and suggests that the convergent component of the largely oblique movements there may be less than in the eastern Carpathians.

7. Conclusions

[42] CELEBRATION 2000 profile CEL05 was designed to study the main tectonic features associated with the Pannonian basin, the Carpathians, the TESZ, and the EEC. The seismic model for this 1420 km long profile reveals a diverse and complex lithospheric structure (Figures 7 and 17). Because of the high quality of the seismic data, the variety of structures associated with specific tectonic features is well documented, and will become the basis for further integrated geophysical and tectonic analysis.

[43] 1. The crustal structure of the Pannonian basin consists of a ∼5 km thick sedimentary layer with velocities of V_p < 5.8 km/s, two simple layers in the crystalline crust (V_p = 6.1–6.2 km/s and V_p = 6.3–6.6 km/s) and a Moho depth of ∼25 km. Near the northern margin of this basin, a shallow high-velocity body was found that correlates with the Bukkia Mountains, a the Miocene volcanic intrusion. The uppermost mantle is characterized by *P* wave velocities of 7.95–8.0 km/s.

[44] 2. The most complicated structure is observed in the central part of the profile in the transition between the Pannonian basin and East European craton. The sedimentary cover with velocities V_p < 5.5 km/s reaches a maximum depth of ∼20 km in this region. In the Pieniny Klippen belt (the border between the Inner and Outer Carpathians), a low-velocity slab of sedimentary rocks dips southward at 35–40° to a depth of 10 km. We interpret this dipping feature to be associated with a regional detachment zone. A

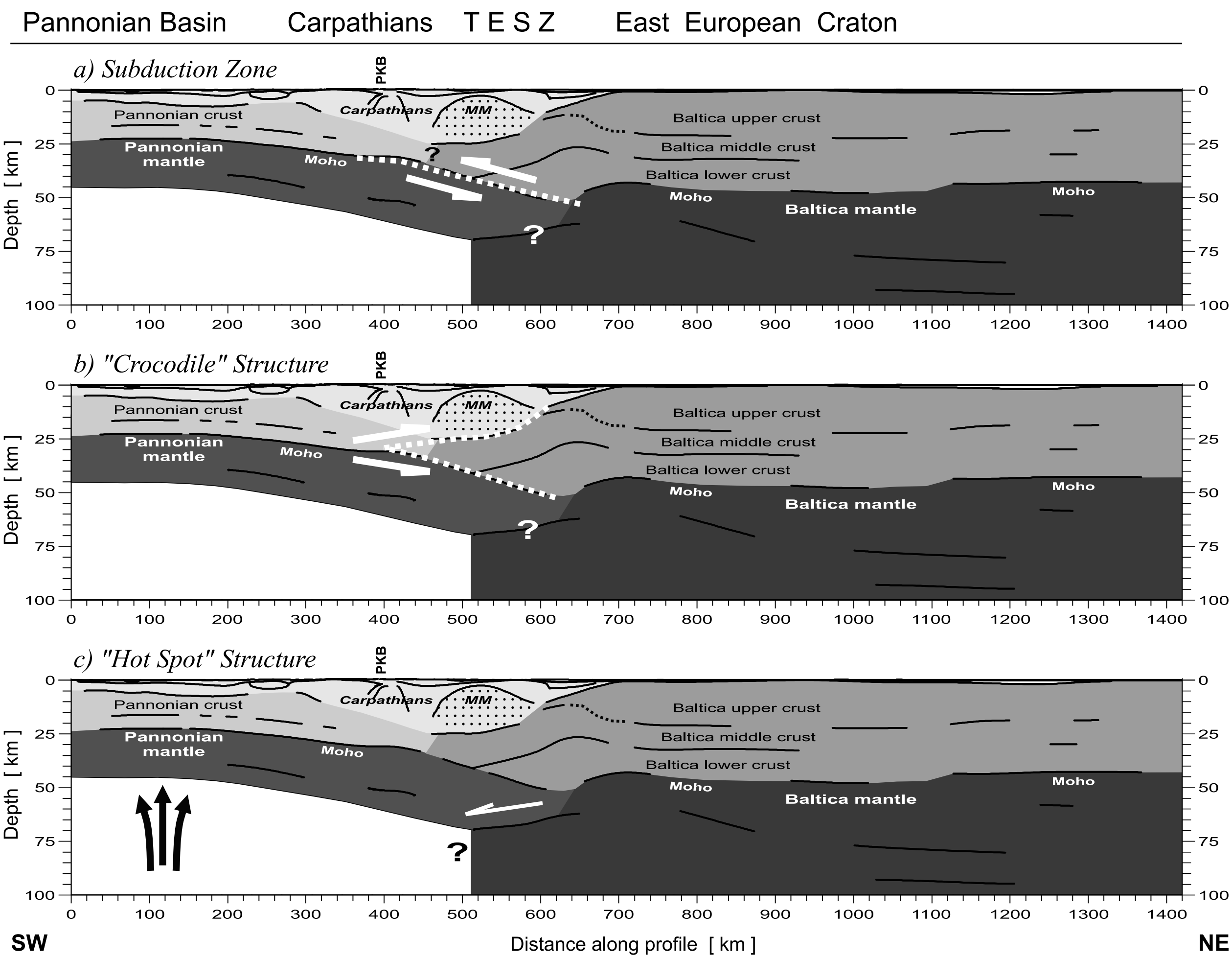

Figure 19. Possible tectonic interpretations of the features observed in the velocity model derived in this study. Vertical exaggeration for these models is ~2.75. (a) Model of the "old" (northward) subduction of the Pannonian lithosphere under East European craton in the Jurassic–Lower Cretaceous [*Książkiewicz*, 1977] or by the end of the Jurassic, ~140 Ma [*Golonka et al.*, 2003a]. (b) Collisional model forming a "crocodile" structure where Carpatho-Pannonian upper crust is obducted over crystalline crust of Variscan-EEC origin and Carpathian-Pannonian mantle lithosphere underthrusted the cratonic crust. (c) Thinning of the Pannonian lithosphere due to extension and high heat flow, and "young" (southward) subduction of the EEC lithosphere in the Tertiary (Miocene). PKB, Pieniny Klippen belt; MM, Małopolska massif.

structural massif is found just north of the PKB and is separated from the Małopolska massif by the 20 km deep Carpathian foredeep. Beneath the Carpathians, a complex two-layer crystalline crust is observed (with velocities 5.9–6.2 and 6.5–6.8 km/s), while beneath the TESZ and EEC margin farther to northeast, a three-layer crust was found with the Moho deepening to ~50 km depth. In the TESZ area, the deep crustal layers are upwarped producing high velocities (~6.4 km/s) at a depth of only ~10 km.

[45] 3. The crustal structure of the EEC is typical for Precambrian cratonic areas, with a thin sedimentary cover, and three-layer crystalline crust. The depth of the Moho varies within the interval of 42–48 km over the entire ~700 km long cratonic part of the profile. Velocities in the uppermost mantle of the EEC are 8.20–8.25 km/s. A small change in the lithospheric structure correlates with the suture between Fennoscandia and Sarmatia, two major terranes of the EEC (Baltica).

[46] 4. Over the whole length of profile CEL05, reflectors in the lower lithosphere were delineated. A reflector in the lower lithosphere of the Pannonian/Carpathian area follows the shape of the Moho ~15 km below it and thus dips to the north. In the area of the EEC, two reflectors at 75–80 km and ~95 km, and some short reflecting elements at depths of ~55 and 60–65 km were found.

[47] 5. Three possible tectonic interpretations of the area of profile CEL05 in the transition between the Carpathian-Pannonian plate and East European craton are consistent with the observed velocity structure. These models are not mutually exclusive because of the complex Mesozoic and Cenozoic tectonic evolution of the region. The first model invokes "old" (northward) subduction zone of the Pannonian lithosphere under the East European craton in the Jurassic–Lower

Cretaceous, although we know that the subduction of the Meliata Ocean did not occur between the Pannonian basin lithosphere and the Inner Carpathians. The second model invokes lithospheric-scale collision to form a "crocodile" structure where Carpathian-Pannonian upper crust was obducted over the crystalline crust of the EEC and Carpathian-Pannonian mantle lithosphere was underthrusted beneath the cratonic crust. The third model invokes Neogene thinning of the Pannonian lithosphere due to extension together and "young" (Miocene) southward subduction of the EEC lithosphere. We suspect that elements of all three of these models will be required to explain the complex tectonic evolution of this region.

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