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Upper lithospheric seismic velocity structure across the Pripyat Trough and the Ukrainian Shield along the EUROBRIDGE'97 profile

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Abstract

We present new results on the structure resulting from Palaeoproterozoic terrane accretion and later formation of one of the aulacogens in the East European Platform. Seismic data has been acquired along the 530-km-long, N–S-striking EUROBRIDGE'97 traverse across Sarmatia, a major crustal segment of the East European Craton. The profile extends across the Ukrainian Shield from the Devonian Pripyat Trough, across the Palaeoproterozoic Volyn Block and the Korosten Pluton, into the Archaean Podolian Block. Seismic waves from chemical explosions at 18 shot points at approximately 30-km intervals were recorded in two deployments by 120 mobile three-component seismographs at 3–4 km nominal station spacing. The data has been interpreted by use of two-dimensional tomographic travel time inversion and ray trace modelling. The high data quality allows modelling of the P- and S-wave velocity structure along the profile. There are pronounced differences in seismic velocity structure of the crust and uppermost mantle between the three main tectonic provinces traversed by the profile:

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(i) the Pripyat Trough is a ca. 4-km-deep sedimentary basin, fully located in the Osnitsk–Mikashevichi Igneous Belt in the northern part of the profile. The velocity structure is typical for a Precambrian craton, but is underlain by a ca. 5-km-thick lowest crustal layer of high velocity. The development of the Pripyat Trough appears to have only affected the upper crust without noticeable thinning of the whole crust; this may be explained by a rheologically strong lithosphere at the time of formation of the trough. (ii) Very high seismic velocity and Vp/Vs ratio characterise the Volyn Block and Korosten Pluton to a depth of 15 km and probably also the lowest crust. The values are consistent with an intrusive body of mafic composition in the upper crust that formed from bimodal melts derived from the mantle and the lower crust. (iii) The Podolian Block is close to a typical cratonic velocity structure, although it is characterised by relatively low seismic velocity and Vp/Vs ratio. A pronounced SW-dipping mantle reflector from Moho to at least 70 km depth may represent the Proterozoic suture between Sarmatia and Volgo–Uralia, the structure from terrane accretion, or a later shear zone in the upper mantle. The sub-Moho P-wave seismic velocity is high everywhere along the profile, with the exception of the area above the dipping reflector. This velocity change further supports a plate tectonic origin of the dipping mantle reflector. The profile demonstrates that structure from Palaeoproterozoic plate tectonic processes are still identifiable in the lithosphere, even where younger metamorphic equilibration of the crust has taken place.

Keywords: Crustal accretion; Aulacogens; Secular lithospheric evolution; East European Craton; Seismic velocity; Ukrainian Shield

1. Introduction

There is considerable debate about the existence and type of plate tectonic processes during the Precambrian evolution of the Earth. Tectonic structure has survived in the continental lithosphere since the Proterozoic (e.g. BABEL Working Group, 1990). It is still an open question whether the tectonic style is similar to present-day tectonic style or later metamorphic and other processes may have changed the inherited structure. A thin layer of sedimentary rocks often covers the basement of Precambrian cratons and erosion has levelled the surface topography. The basement of the East European Platform is largely unknown, and it may therefore be thought of as relatively homogeneous. However, recent research has shown that there are pronounced changes in rock types which may be related to Proterozoic and Archaean plate tectonic processes, such as collision between major plates, amalgamation of terranes, and docking of island arcs. The main changes in rock types take place along lineaments, which may be traced by means of magnetic data, as in an analysis of major terrane boundaries in the East European Craton by Bogdanova et al. (1996). Another characteristic feature of the East European Platform is the existence of aulacogens, which are identified as large, elongated structures with a sedimentary fill of mainly Late Precambrian or Palaeozoic age. They are often assumed to have formed by rifting processes, although compared to more recent rift zones, they are generally much wider and the available information indicates a different type of deep crustal structure. A specific feature of aulacogens appears to be a flat Moho and a lower crust with very high seismic velocity, which may be related to magmatic activity.

Acquisition of seismic data along the EUROBRIDGE deep seismic sounding (DSS) transect is a EUROPROBE project (Bogdanova et al., 2001). The approximately 1500-km-long, NW–SE and N–S trending profile segments span the western part of the East European Craton extending from the Palaeoproterozoic accretionary belt of Fennoscandia to the western part of the Ukrainian Shield (Fig. 1). The principal scientific objective of the EUROBRIDGE seismic profiling is to establish the seismic velocity structure of the lithosphere beneath the sediment-covered platform between the exposed Proterozoic and Archaean complexes of the Baltic and Ukrainian Shields. The EUROBRIDGE international seismic group has carried out three major phases of onshore data acquisition. EUROBRIDGE'95 was recorded in a NW–SE direction in Lithuania to image Fennoscandian deep structure (EUROBRIDGE Seismic Working Group, 2001). Coverage was extended to the southeast across Lithuania and Belarus by EUROBRIDGE'96 to image the regional setting of the Fennoscandia–Sarmatia suture zone, which is the most important tectonic boundary in the Belarus–Baltic region of the East European Craton (EUROBRIDGE Seismic Working Group, 1999). In the third phase, the N–S trending EUROBRIDGE'97 profile was recorded between Belarus and Ukraine to image the northwestern part of Sarmatia and the Pripyat Rift. The EUROBRIDGE'97 profile is 530 km long and crosses EUROBRIDGE'96 in southern Belarus (Fig. 1).

This paper presents a detailed interpretation of the P- and S-wave velocity structure along the EUROBRIDGE'97 profile. The northern end of

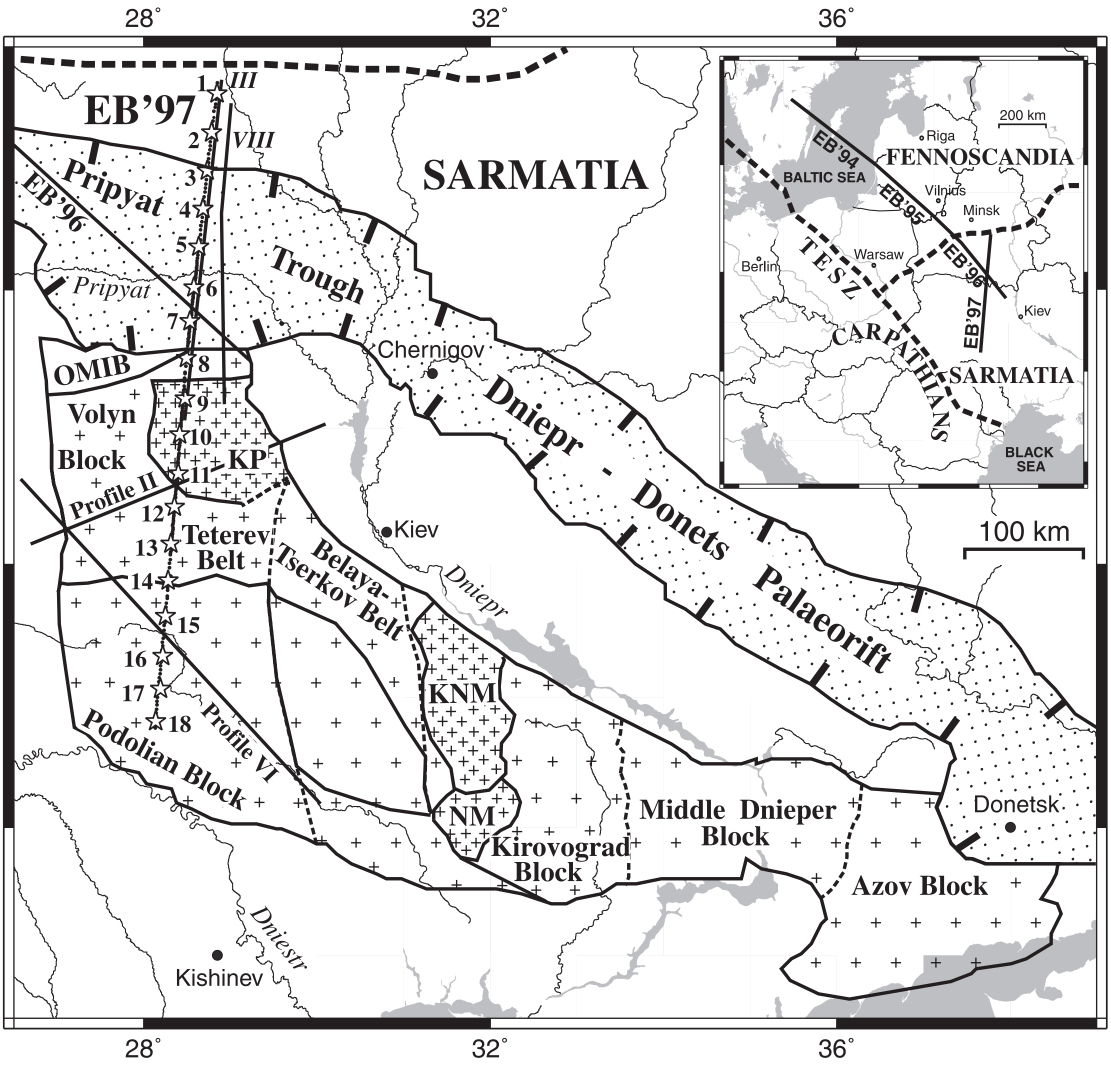


Fig. 1. Location map of the EUROBRIDGE'97 DSS profile and its intersection with the earlier EUROBRIDGE'96 profile. Earlier deep seismic sounding profiles are also shown, including the parallel and close profile VIII (illustrated in Figs. 3 and 10). Relevant regional units of the Ukrainian Shield ('+' hatching) and the overlying Pripyat–Dnieper–Donets Palaeorift system are shown. Inset map shows the location of all the EUROBRIDGE DSS profiles. The thick dashed lines indicate the near-surface location of the suture zones separating the Fennoscandia from Sarmatia and the East European Craton from Palaeozoic Europe (TESZ). Asterisks represent the shot points of EUROBRIDGE'97. Abbreviations: KNM—Korsun–Nowomyrgorod Massif; KP—Korosten Pluton; NM—Novoukrainka Massif, TESZ—Trans European Suture Zone.

the profile line is located within the ca. 2.0 Ga Osnitsk–Mikashevichi Igneous Belt, which marks the northwestern margin of the Sarmatian crustal segment. The profile crosses the Pripyat Trough, which is a part of the extensive Phanerozoic Pripyat–Dnieper–Donetsk palaeorift. The major basement components traversed in the Ukrainian Shield are the Palaeoproterozoic Volyn Block, which has been intruded by the gabbro–anorthosite–rapakivi Korosten Pluton, and the Archaean Podolian Block. The main findings are (1) the Korosten Pluton shows high seismic velocity throughout the crust and most pronounced in the upper crust which is related to several mafic intrusions, (2) an indication that the rifting activity at the Pripyat Trough was largely shallow, and (3) a dipping mantle reflector may be related to Precambrian plate tectonic accretionary processes.

2. Regional geology

The East European Craton comprises three major segments: Fennoscandia, Sarmatia and Volgo–Uralia (Bogdanova et al., 1996). The segments are large composite terranes, each with an independent Archaean and Palaeoproterozoic history, brought together during the Late Palaeoproterozoic (ca. 1.80–1.75 Ga). This process was characterised by a long-lasting "soft" collision, in which newly formed island arcs successively accreted to the edges of the Archaean crustal nucleus of the craton. Major differences in crustal structure, lithology and time of crust-forming processes characterise the segments. The suture zones between the segments are fundamental features of the East European Craton, and they probably underwent reactivation during Meso- to Neoproterozoic rifting episodes. Most of the craton is covered by platform sediments. The fill of the Pripyat Trough is mainly of Phanerozoic age. Many important present-day regional structures were created during basement formation and amalgamation, and others were formed in association with the later episodes of sedimentation.

EUROBRIDGE'97 traverses the northwestern part of the Sarmatia segment, including the Pripyat Trough (Fig. 1). The geology of the main basement units (Fig. 2) of the upper crust is summarised below.

2.1. Osnitsk–Mikashevichi Igneous Belt (OMIB)

This belt constitutes the northwestern zone of Sarmatia as a ca. 2.0 Ga active margin. Its northwestern border forms the contact between the Palaeoproterozoic terranes related to Fennoscandia and Sarmatia (Bogdanova et al., 1996). The OMIB has a SW–NE trend and is about 100–150 km wide, probably limited by deep listric faults (EUROBRIDGE Seismic Working Group, 1999). It consists of Palaeoproterozoic igneous complexes, essentially free of regional metamorphism. The complexes include metagabbro–diabase (2.02 Ga), large batholiths of the dominant diorite–granodiorite–granite (2.0–1.97 Ga), and quartz–syenite–granite (1.8–1.75 Ga) (Aksamentova, 1990). The OMIB assemblages are chemically similar to those of present-day active continental margins. The OMIB is partly concealed by the sediments of the Pripyat Trough.

2.2. Pripyat Trough (PT)

The PT is part of the regional Palaeozoic–Phanerozoic Pripyat–Dnieper–Donets Palaeorift. The PT is about 280 km long and 150 km wide and contains up to 6 km of sediment fill. The oldest formations are 200–300 m of Middle Devonian terrigenous and carbonate strata. The most widespread formations range from Upper Devonian to Middle Triassic age and include terrigenous, carbonate, saliferous and volcanogenic lithologies. These rocks form strata of several kilometres' thickness, whereas the youngest strata (Upper Triassic to Quaternary) are only 150–200 m thick (Aizberg et al., 1987; Garetsky and Klushin, 1989). The subsidence history and its tectonic implications have been discussed by Kusznir et al. (1996). Juhlin et al. (1996) and Stephenson et al. (2001) present detailed interpretations of reflections seismic profiles across the Pripyat Trough.

2.3. Volyn Block (VB)

The Palaeoproterozoic gneiss complexes, particularly seen in the SE and SW parts, are metamorphosed in amphibolite to epidote–amphibolite facies with ages of formation ranges from 2.4 to 2.1 Ga (Shcherbak et al., 1989; Skobelev et al., 1991). Younger rocks are represented by anatectic granites in numerous

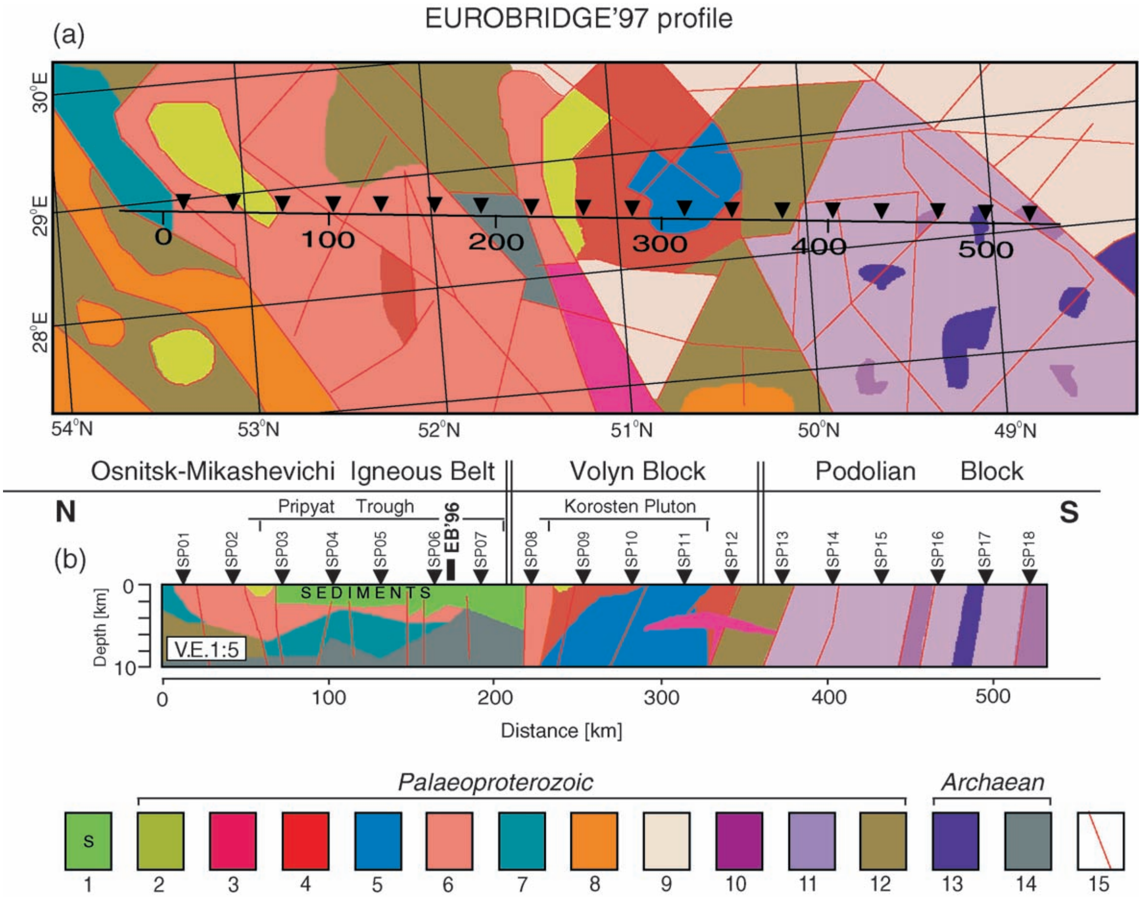


Fig. 2. Geological map (a) and upper crustal cross-section (b) along the EUROBRIDGE'97 DSS profile. Arrows show the positions of the shot points. Vertical exaggeration, 1:5. Legend: (1) sedimentary rocks, Palaeoproterozoic; (2) Tolckachy Suite, quartz sandstones and schists (1.7–1.6 Ga); (3) Perga complex, subalkaline leicogranites (1.76–1.75 Ga); (4) Korosten complex, rapakivi granites (1.77 Ga); (5) Korosten complex, anorthosites and gabbros (1.8–1.76 Ga); (6) Osnitsk complex, granodiorites and subalkaline granites (2.0–1.98 Ga); (7) Bucky complex, gabbronorites, monzodiorites, quartz monzonites (2.05–1.98 Ga); (8) Zhitomir and Uman complexes, granites (2.06–2.02 Ga); (9) Paleoproterozoic anatectites and migmatites; (10) Berdichev complex, charnockites (2.1–2.0 Ga); (11) Berdichev complex, high-aluminous granites and migmatites (2.1–2.0 Ga); (12) Vasilyevka strata, plagiogneisses (2.5–2.4 Ga), Archean; (13) Gaivoron (3.1 Ga) and Litin (2.8 Ga) complexes, enderbites; (14) Dniester–Bug complex, granulites, gneisses, quartzites (>3.4 Ga); (15) deep-reaching faults.

small massifs and bodies in the southern and western parts of the domain (Zhitomir complex, 2.06–2.02 Ga. Slightly younger (2.02–1.98 Ga) intrusive rock complexes may have formed along deep faults, probably associated with synchronous tectonic events further to the NW in the OMIB. Many types of rock assemblage have been distinguished which were formed from different magma sources and by varying degrees of melting. These intrusive complexes are not metamorphosed and are always discordant with older lithologies. During the subsequent ca. 170 Ma of tectonic stabilisation, only denudation products were deposited, accumulating in depressions. Quartz sandstone, conglomerates and clay shale occur in the northwest of the region.

2.4. Korosten Pluton (KP)

This distinctive, layered plutonic complex was previously thought to extend to a depth of 6 km. Seismic reflection and gravity interpretation indicate that the crust has been extensively intruded by mafic melts (Bucharev, 1992). Melting of lower crust may have been an important factor for the generation of both mafic and felsic rocks in the KP, as based upon Sm–Nd values for these rocks Dovbush et al. (2000).

Diverse magmatic activity took place between 1.8 and 1.73 Ga (Verkhogliad, 1995), probably caused by the approach of the Fennoscandia and Sarmatia terranes (Stepanyuk et al., 1999). The igneous rocks consist of the gabbro, anorthosite and rapakivi granite of the Korosten Pluton, and subalkaline leucogranite (1.76–1.74 Ga) to the west of the pluton. Volcanic subalkaline andesites and rhyolites (1.73 Ga) and sedimentary rocks (quartz sandstones and slates) overlie the northern part of the Korosten Pluton. No rocks of later Proterozoic age are known in the western area of the Ukrainian Shield, which became a stable platform toward the end of the Palaeoproterozoic.

2.5. Podolian Block (PB)

The PB is in the southern part of the western Ukrainian Shield. Its oldest geological units are the >3.4 Ga supracrustal Dniester–Bug Group (Lesnaya et al., 1995). Mafic and intermediate granulites of this complex are widespread in the central part of the PB. The Dniester–Bug Group is intruded by enderbites of 3.1 Ga and 2.8 Ga age (Shcherbak et al., 1989; Lesnaya et al., 1995). The Bug complex represents a younger (ca. 2.6 Ga) stratified unit observed in narrow synforms mainly in southern and central parts of the PB. The Bug and Dniester–Bug complexes are quite similar in composition. Various anatectic granitoids of 2.08–2.02 Ga were formed in the north, close to the VB during the Palaeoproterozoic (Shcherbak et al., 1989; Skobelev et al., 1991; Lesnaya et al., 1995).

3. Previous geophysical studies

The EUROBRIDGE'97 profile crosses the EUROBRIDGE'96 profile (EUROBRIDGE Seismic Working Group, 1999) between shot points SP06 and SP07 in the Pripyat Trough (Fig. 1). At the cross point, the latter profile shows an overall seismic velocity–depth profile which is close to the expected three-layered crustal structure of a Precambrian craton (Meissner, 1986), although relatively complicated. The velocities of individual layers of the crystalline crust below the ca. 4-km-thick sedimentary sequences are: 6.25 km/s to 12 km depth, 6.4 km/s to 15 km depth, 6.65 km/s to 28 km depth and 7.0 km/s to 49 km depth. The middle crust appears to consist of two main layers and the lower crust is strongly reflective in wide-angle incidence, some of which may be attributed to the existence of the Pripyat Trough at the cross point. A pronounced subhorizontal mantle reflector has been interpreted at a depth of ~60 km over a length of 250 km along the profile; it is unknown whether it represents a positive velocity contrast or a localized reflective band in the uppermost mantle. Kozlovskaya et al. (2002) present the results of gravity modelling for density and the petrophysical implications along the EUROBRIDGE'95 and '96 profiles.

The study area has been covered relatively densely by DSS profiles, mainly to offsets smaller than 150 km; we refer to Sollogub (1982 and references therein) for a review. The EUROBRIDGE'97 profile is located in the vicinity of four previous DSS and deep reflection seismic profiles in Belarus. These earlier profiles are: Grodno–Starobin, and CDP seismic reflection profiles III and VIII across the Pripyat Trough (Karatayev et al., 1993). Interpretations of the deep crustal structure along profiles III and VIII (Fig. 3) show a system of deep listric faults in the vicinity of the Pripyat Trough and two reflectors at depths of 35 and 47–50 km (Juhlin et al., 1996; Stephenson et al., 2001). Further, zones of unusually low seismic velocity in the lower crust and beneath the Moho have been interpreted (Garetsky and Klushin, 1989). The zone between the two reflectors is seismically reflective, much in the same way as often observed in extensional areas. We find from the new refraction data that the Moho is the lower of the two reflectors.

Within the northern region of the Pripyat Trough the average P-wave velocity of the sedimentary cover is described by the relationship: $V = 2250$ m/s $+ z$ 0.397 s^{-1}, where z is the depth (Belinsky et al., 1984). The southern region of the Pripyat Trough is more complicated due to the presence of saliferous deposits, which usually have strong velocity and density contrasts to the surrounding rocks. The depth to the salt surface ranges from 400 to 2600 m with steep gradients. There is evidence of salt tectonics, which complicates the interpretation of DSS data. The lateral dimension of salt domes is often less that 4 km, which is comparable to the distance between obser-

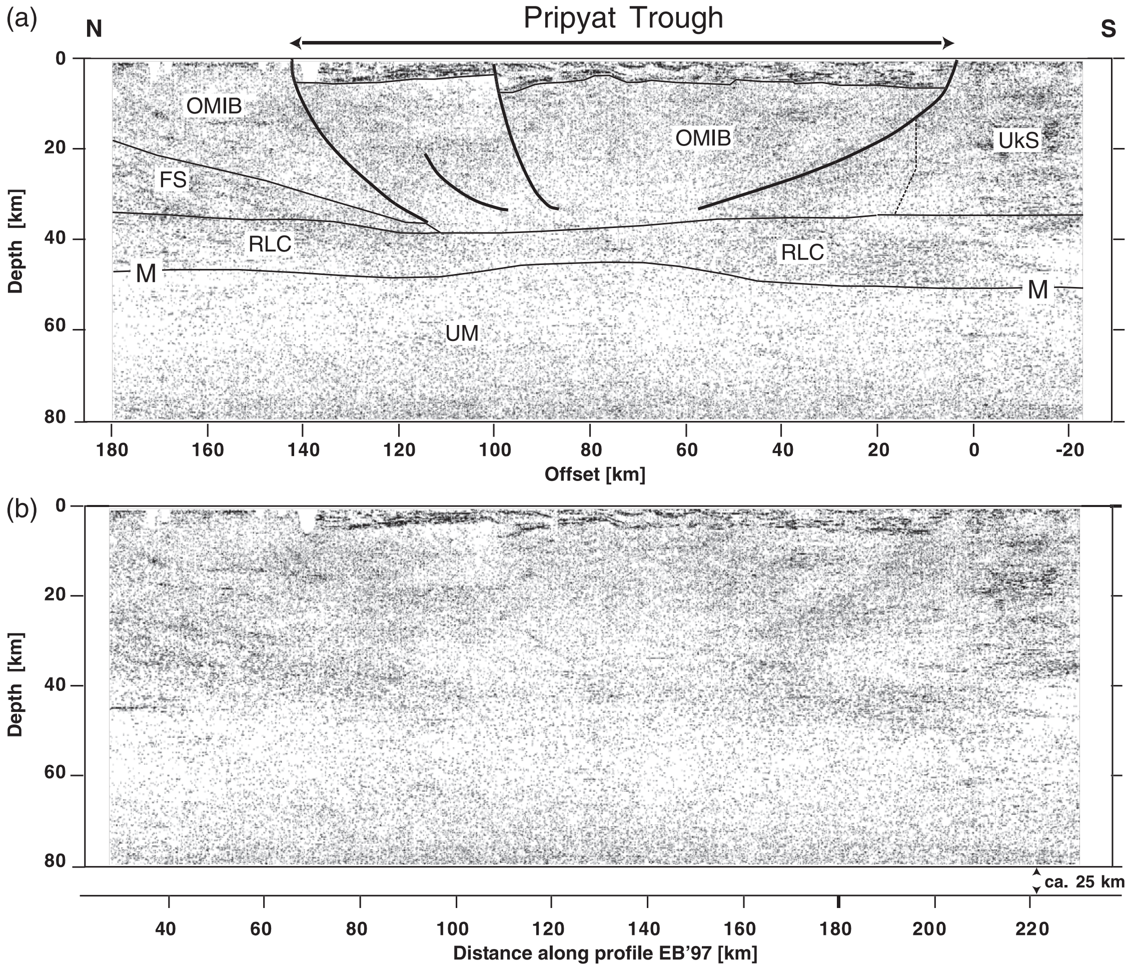


Fig. 3. Normal-incidence reflection seismic CDP section along line VIII converted into a depth profile (after Juhlin et al., 1996; Stephenson et al., 2001). The profile line is parallel to the EUROBRIDGE'97 profile at a distance of ca. 25 km. (a) Overlay showing interpretation of the main crustal blocks, the Pripyat Rift, putative steep deep crustal discontinuities, and a reflective lower crust above the reflection Moho, which is defined as the lower termination of crustal reflectivity. (b) Raw seismic section. Abbreviations: OMIB—Osnitsk–Mikashevichi Igneous Belt; RLC—reflective lower crust; UkS—Ukrainian Shield; UM—upper mantle.

vation sites. The velocity of the strata between the salt surface and the basement is 3800–4500 m/s. The basement surface is tilted in a mosaic of blocks with different inclination. An estimate of the average velocity in the sedimentary cover as a whole cannot be provided because of the salt structures and basement topography. For the post-salt deposits, Belinsky et al. (1984) estimate the average velocity to be $V = 1800$ m/s $+ z$ 0.42 s^{-1}, where z is the depth.

The EUROBRIDGE'97 profile crosses DSS profiles II and VI at km 313 and km 441 (Fig. 1). The recording frequencies of 7–30 Hz and seismograph spacing of 100–200 m adopted for these profiles provide high resolution of the seismic velocity models. The vertical component of ground movement was recorded to maximum offsets of 220 km.

Profile II intersects the EUROBRIDGE'97 profile in the southern part of the Korosten Pluton (KP). The P-wave velocity model suggests that the crust of the KP region has significantly higher velocities than the adjacent regions (Ilchenko, 1985; Ilchenko and Kaluzhnaya, 1998). A zone of alternating layers of low and high velocity was modelled down to 20 km depth using sharp delays of the travel time curves. The interpreted low-velocity layers are thicker (1–4 km, velocity 5.8–6.2 km/s) than the high-velocity layers (1–2 km, velocity 6.5–6.95 km/s), and they are laterally more variable in thickness. The lateral positive velocity contrast of the KP to the surrounding blocks is 0.6 km/s at the top of crystalline basement and 0.35 km/s at 20 km depth. The zone widens within this depth interval from 33 to 85 km. The high-velocity layers may consist of mafic rocks. The velocity model below 20 km depth is inferred from a strong reflection only, assumed to be the PmP reflection from the Moho. There is indication for a 70-km-wide zone of increased, relatively constant velocity of 6.95–7.0 km/s compared to the surrounding areas with velocities of 6.6–6.95 km/s in the same depth range. The Moho is modelled at a depth of 37–39 km and the P-wave velocity of the topmost mantle is assumed to be 7.9–8.2 km/s. Several uppermost mantle reflectors are interpreted between 2 and 6 km below the Moho. However, our new EUROBRIDGE'97 results from modelling refracted waves from the same depth interval, show that the assumed Moho may instead be the top of a lower crustal layer with high velocity (>7.0 km/s), and that the "mantle reflectors" may reside in the lowest crust (c.f. the discussion in the following chapters).

Profile VI intersects the EUROBRIDGE'97 profile within the Podolian Block (PB). The velocity model obtained from refraction data of Profile VI shows a low vertical velocity gradient where the P-wave velocity increases from 6.1 to 6.2 km/s at the top of the basement to 6.4 at a first-order discontinuity at 21 km. Below this discontinuity, a high vertical velocity gradient is proposed for the lower crust from 6.5 to 7.2 km/s immediately above the Moho at 52–54 km depth, based on indications from the PmP reflection only. The sub-Moho velocity of 8.4–8.5 km/s is constrained by uppermost mantle refractions.

4. Data acquisition and description of the observed seismic wave field

EUROBRIDGE'97 was an international collaborative project with participation from institutions in Belarus, Denmark, Finland, Germany, Poland, Russia, Sweden, UK and Ukraine. Seismic data acquisition was undertaken in August and September 1997 along a 530-km-long N–S transect within the Sarmatian segment of the East European Craton (Fig. 1). Designed to investigate crustal and upper mantle structure, the EUROBRIDGE'97 profile crosses terranes of Sarmatia, with particular attention to achieve a high recording density in the vicinity of the Korosten Pluton.

The scheme of shot points and recording stations is similar to that employed for the EUROBRIDGE'95 and EUROBRIDGE'96 profiles. Seismic energy was recorded from shots fired at 18 locations (SP01–SP18; Fig. 1; Table 1) located at nominal intervals of 31 km. Each source consisted of explosives distributed in an array of drillholes, with total shot sizes varying between 250 and 1000 kg (Table 1). Each borehole was 40 m deep and loaded with a maximum explosive charge of 85 kg. Shots SP01–07, 14–15 and 17 were fired with high precision by use of a GPS system. Data acquisition was undertaken in two deployments with about 120 mobile three-component seismometers deployed at a typical station spacing of 3–4 km. Overlapping coverage between the two deployments in the centre of the profile ensured double density recording in the region of the Korosten Pluton of shots from several locations.

In Phase 1 of the field program, shots SP01–07, 09, 11, 13 and 14 were recorded in the segment from SP01 to the vicinity of SP14, an interval of approximately 400 km length. Phase 2 coverage extended from the vicinity of SP06 to SP18, such that seismic waves from shots SP04, 06, 08, 10 and 12–18 were recorded over a distance of about 380 km. Seismometers in the overlapping centre part of the profile, from approximately SP06 to SP14, were moved by half a station spacing (about 1.5 km) to obtain a high density coverage in the area of the Korosten Pluton for the repeated shots SP04, 06, 13 and 14. These shots were recorded in both phases of EUROBRIDGE'97 and therefore along the whole profile.

Each recording group carried out initial processing of the data acquired by them, using in-house or other software, to extract 130-s-long seismograms starting at each shot time. The extracted traces were collated into a common data set of shot gathers at GFZ-Potsdam. Geometry information was applied to the common data set, which was then distributed to all participants.

Table 1
Details of explosive sources used

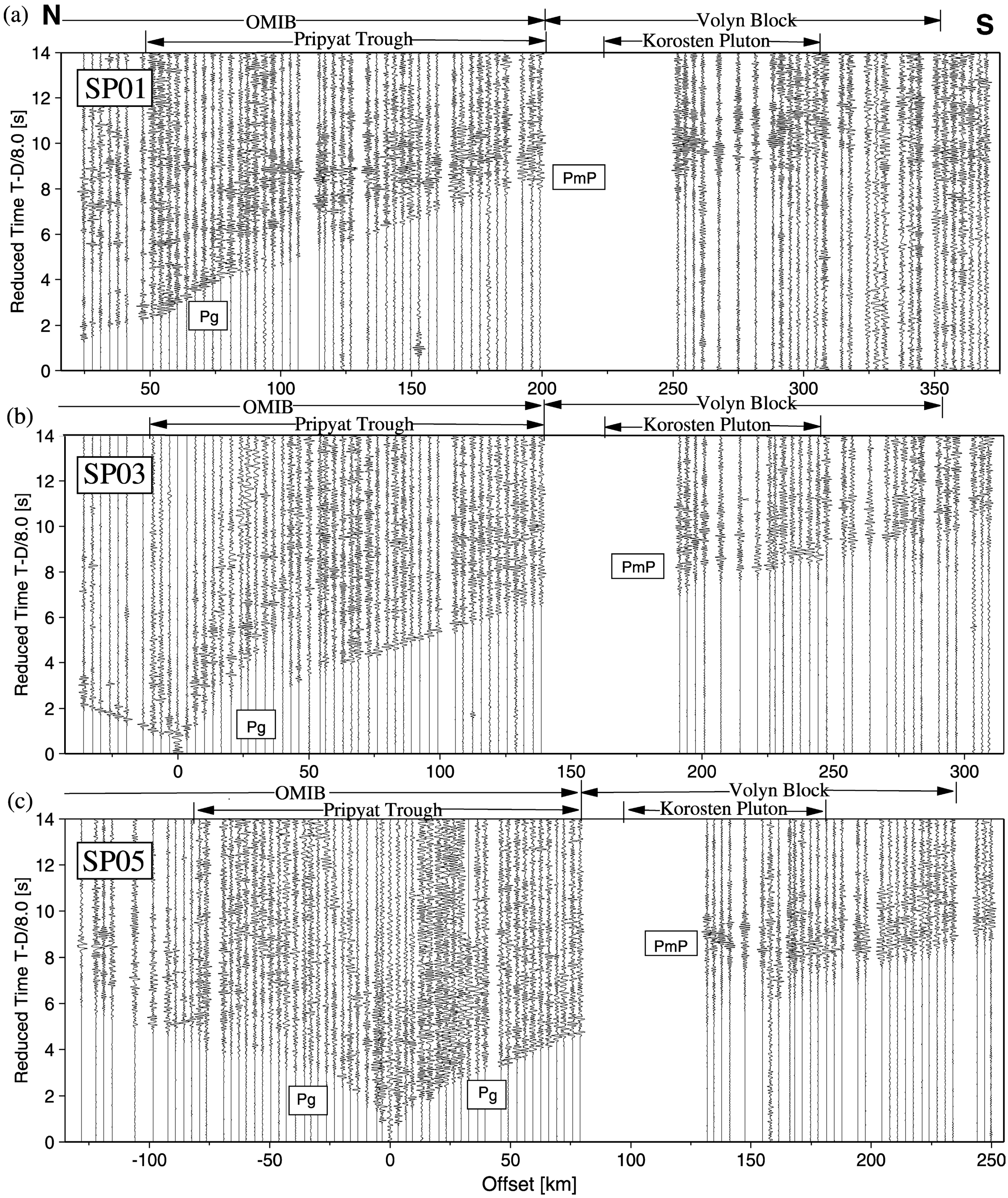
Shot point number	Latitude N (φ)	Longitude E (λ)	Altitude h (m)	Time (UTC) (y:d:h:m:s)	TNT charge (kg)
SP01	53.3828	28.8520	158.0	1997:243:22:30:00.000168	800
SP02	53.1070	28.7882	155.0	1997:244:22:30:00.000171	300
SP03	52.8470	28.7364	144.0	1997:243:23:30:00.000202	700
SP04	52.5800	28.6832	131.0	1997:244:23:30:00.000204	300
SP04				1997:250:23:00:00.000193	800
SP05	52.3097	28.6410	130.0	1997:244:21:00:00.000165	600
SP06	52.0185	28.5818	136.0	1997:243:21:00:00.000163	300
SP06				1997:249:22:30:00.000194	800
SP07	51.7823	28.5362	138.0	1997:244:01:00:00.000191	400
SP08	51.5158	28.4934	161.0	1997:249:21:30:00.7964	700
SP09	51.2220	28.4796	182.5	1997:243:23:00:00.0911	250
SP10	50.9663	28.4183	193.0	1997:250:00:30:00.9060	250
SP11	50.6708	28.3956	220.0	1997:244:23:00:03.6228	700
SP12	50.4284	28.3556	238.0	1997:250:23:30:00.4192	400
SP13	50.1551	28.3130	232.0	1997:244:01:30:00.3194	639
SP14	49.8862	28.2828	265.0	1997:249:23:30:00.000165	400
SP15	49.6188	28.2509	262.0	1997:250:21:30:00.000162	250
SP16	49.3153	28.2190	314.0	1997:251:00:00:00.6587	700
SP17	49.0638	28.1918	310.0	1997:251:00:30:00.000163	250
SP18	48.8240	28.1372	280.0	1997:250:21:00:00.5329	1000

All record sections show distinct P-wave arrivals along the whole covered part of the EURO-BRIDGE'97 profile (Fig. 4). Interpretation has been hindered by the 50 km wide data gap at km 210 to 260, where Russian equipment was not deployed and the alternative instruments suffered from a timing system failure.

First arrivals with apparent velocities of ~ 4 km/s, associated with the sedimentary cover, are recorded at offsets up to ~ 10 km in the northern part of the profile. The intercept time of the crystalline Pg phase is 0.5–1.5 s in this region (SP01–SP07, Figs. 4 and 7). For SP09–SP18 in the southern, shield part of profile, the Pg phase is observed basically from the shot point. There is pronounced variation in apparent velocity of the Pg phase along the profile. Typically, the apparent velocity is 6.0–6.2 km/s for offsets up to 100–150 km, but apparent velocities up to about 6.4–6.6 km/s are observed in the central part of the profile (SP06–SP11 in Figs. 4 and 7). These high velocities, observed in both directions with overlapping reciprocal travel times, are clearly associated with a shallow high-velocity zone located in the central part of profile around the Korosten Pluton (Fig. 7a).

There is pronounced difference in seismic character between the northern and southern segments of the profile regarding intracrustal reflectivity and reflections from the Moho (c.f. the two recording directions for SP07, Fig. 4d). Only few of the intracrustal reflections can be correlated over any significant distance in the northern part of the profile, although the crust is strongly reflective and the sections show seismic arrivals from slightly after the Pg to the complicated PmP (e.g. SP03, SP05 and SP07, Fig. 4). However, there is indication for distinct, mainly lower crustal reflectivity in the northernmost sections (e.g. SP01, Fig. 4a and SP02, Fig. 7b). In the southern part of the profile, a clear Pg and a complicated PmP are identified on the background of a weakly reflective or non-reflective crust (SP10–SP17, Fig. 4e–i). This illustrates that the crust is strongly heterogeneous in the northern part of the profile and more homogenous (transparent) with few distinct reflectors (interfaces between the main layers or small-scale velocity fluctuations) in the southern part. Further evidence can be seen in the sections from SP16 and SP18 (Fig. 9b and f).

The strongest observed seismic arrivals are reflections from the lower crust and PmP reflections from the Moho (e.g. SP06 at the distance interval of km 260–360, 7–9 s reduced time, Fig. 9a). Together, they usually show a remarkably "ringing" wave field of


(a)
N
OMIB
Volyn Block
S
Pripyat Trough
Korosten Pluton
SP01
PmP
Pg
Reduced Time T-D/8.0 [s]
(b)
SP03
(c)
SP05
Offset [km]

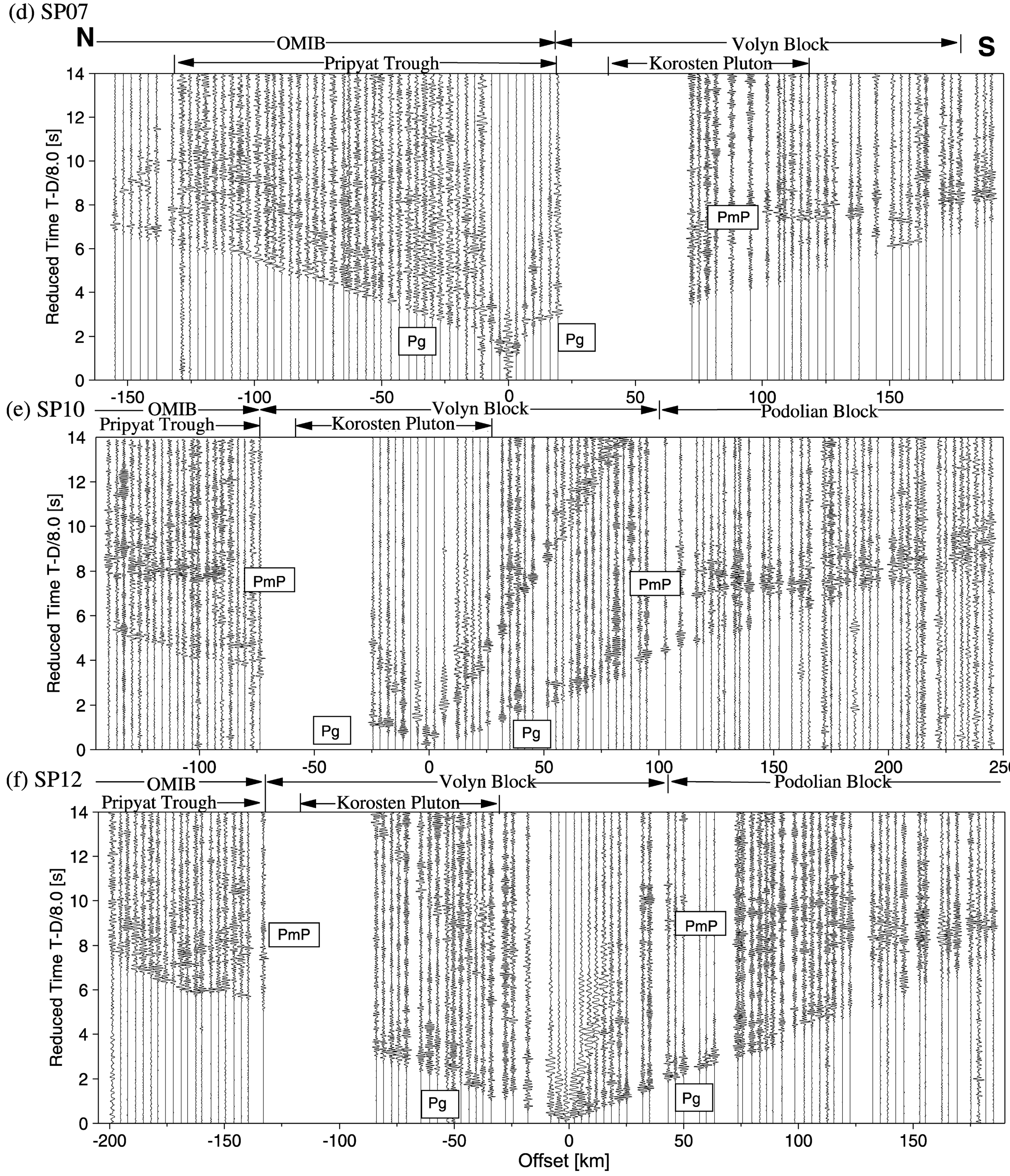


Fig. 4. Examples of amplitude-normalized vertical component seismic sections for a selection of shot points along the EUROBRIDGE'97 profile. The data have been filtered using an Ormsby band-pass filter of 2–18 Hz and displayed using a reduction velocity of 8.0 km/s. Locations of major basement tectonic units and the Pripyat Trough are indicated. Abbreviations: Pg—Seismic refraction from the uppermost crystalline crust; PmP—Seismic reflections from the Moho; Pn—Refracted arrivals from the sub-Moho uppermost mantle. (a) Shot SP01, (b) shot SP03 and (c) shot SP05, (d) shot SP07, (e) shot SP10, (f) shot SP12, (g) shot SP13, (h) shot SP15, (i) Shot SP17.

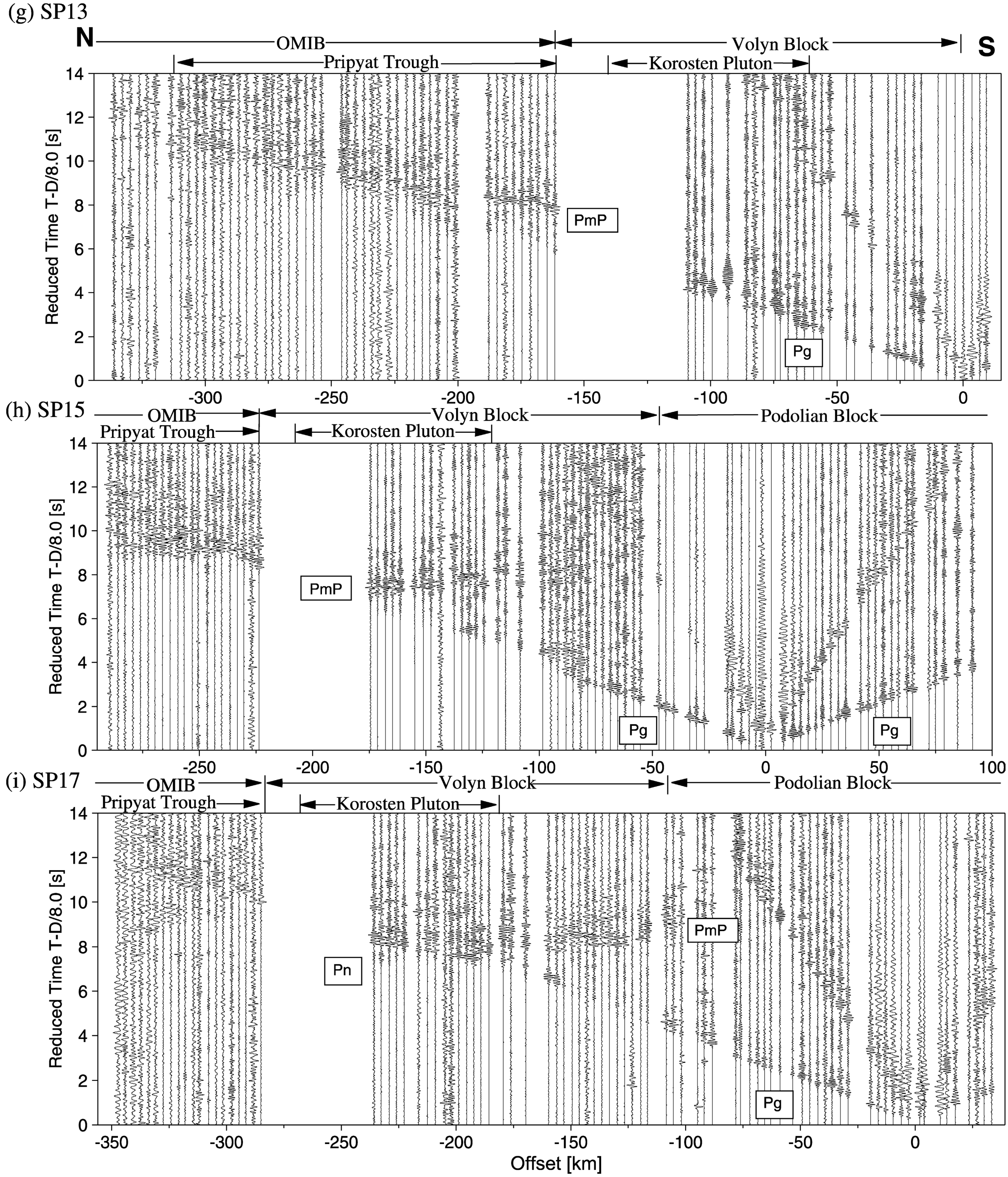


Fig. 4 (*continued*).

1.5–2.5 s duration; the lower crustal arrivals mainly in the northern part of the profile and the PmP along most of the profile. These phases are observed at offsets of 90–200 km and the two series of reflectivities are often merging, although usually a distinct onset of the PmP can be identified. In some record sections, however, a distinct PmP reflection is observed as a sharp and short phase with a duration of less than three periods (e.g. SP07 southward and SP18 northward, Figs. 4d and 9f).

A strong seismic phase from the lower lithosphere (P1) is observed as a late arrival in several record sections (SP16, 17 and 18 at offsets of 250–350 km, illustrated in Figs. 7e, 9b and f). This phase is characterised by a high apparent velocity of >9 km/s and a ~ 1-s-long coda, indicating that it originates from a dipping reflector in the uppermost mantle, which may either be a first-order discontinuity or an inhomogeneous zone with a complicated fine structure. In the opposite direction, the strongest long-range energy arrives very late (around 9 s reduced time in the offset interval 260 to the end of the section at 360 km) with an apparent velocity of 8.2 km/s in the section for SP06 (Fig. 9a). This energy may easily be mistaken for Pn arrivals, but there are clear, although weak, distinct earlier arrivals on several individual seismograms. This time corresponds to the arrival time for the mantle phase in the sections for SP14–18, indicative of a strong southward dip of the mantle reflector.

5. P-wave modelling

5.1. Seismic tomographic inversion of first arrivals

Initial two-dimensional models of seismic P-wave velocity were obtained by application of three different methods for *tomographic inversion*, using first arrivals only. A first model of the upper 20 km (Fig. 5a) was calculated with the commercial ProMAX program. The model is parameterized with a regular grid of 5 × 1.25 km, horizontally and vertically, respectively. The root mean square (RMS) residual difference between observed and calculated travel times for this model is 120 ms. The uppermost 5–10 km of the model is best resolved with up to 80 rays per unit grid size. Beneath this depth, values decrease to 20 rays per unit grid size at 20 km depth.

The second tomographic inversion model (Fig. 5b) was calculated with the method described by Hole (1992) to the maximum penetration depth of the rays. In the first iterations, only picks with offsets smaller than 40 km were inverted. In the following iterations, this offset threshold was stepwise increased to 300 km in steps of 40 km. This stepwise increase in maximum offset ensures that the shallow part of the model is constrained before the deep part, which has the least dense ray coverage and for which the calculated travel times are also influenced by the upper part of the model. The model is parameterized in a 1 × 1 km grid, which is finer than the obtainable resolution. Therefore, smoothing filters were applied in each iteration to the calculated derivatives and model perturbations, but not to the model itself. The initial model was chosen with a high average velocity at all depths in order to prevent artificial low-velocity zones from developing during the inversion procedure. At each chosen maximum offset, the size of the smoothing filter was decreased stepwise from 120 × 30 to 8 × 4 km to increase the resolution. The total number of iterations was 175 and the final RMS residual travel time difference was 89 ms.

The third P-wave velocity distribution (Fig. 5c) was calculated with the FAST tomographic inversion program package (Zelt and Barton, 1998) to the deepest ray penetration. The tomographic method applies regularized inversion, which allows the user to constrain the flatness and smoothness of the perturbations. The grid size was 1 × 1 km for forward calculations of the rays and 15 × 2 km for inverse computations. An initial one-dimensional model, based on previous studies was subject to different flattening and smoothing parameters, and the starting model for the inversion was calculated as the average of the nine models with the smallest RMS travel time misfit. The resulting model was obtained after 20 iterations. It provides a RMS misfit of 96 ms between the observed and computed travel times. The three resulting velocity models based on tomographic inversion show the same overall structure, although details vary at scale below the resolution. The differences are primarily caused by differences in processes and the choice of smoothing parameters, as well as different choices of criteria for terminating the inver-

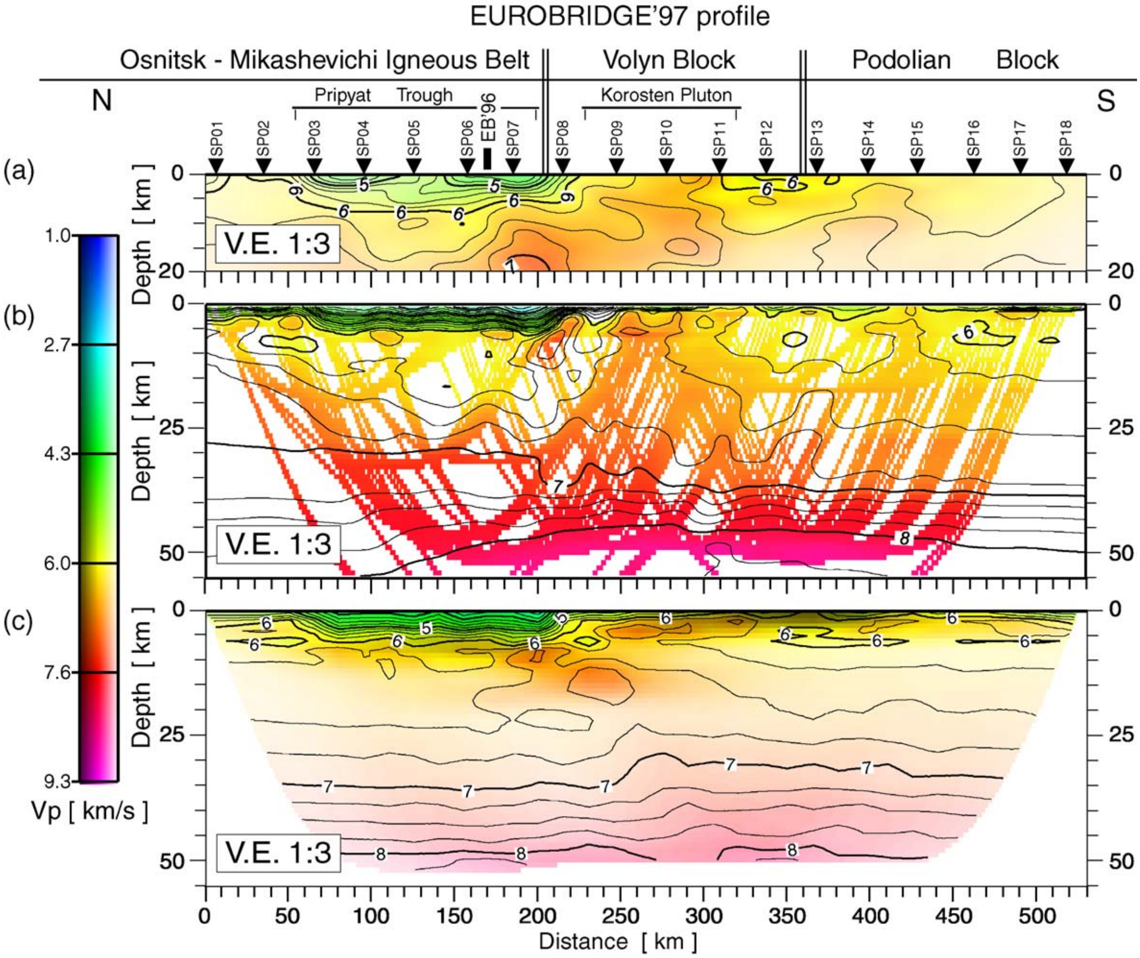


Fig. 5. Two-dimensional P-wave tomographic velocity models along the EUROBRIDGE'97 profile. Contoured velocity isolines (km/s) are shown on the plots. (a) Model computed with the program from ProMAX. (b) Model computed with the program by Hole (1992). (c) Model computed using program FAST by Zelt and Barton (1998). In panels a and c, the color intensity indicates ray density which reflects the degree of resolution. In panel b, the actual ray paths are shown to indicate the ray coverage. The RMS residual difference between observed and calculated travel times are 120, 89 and 96 ms, respectively, in the three cases.

sion procedures, as reflected in the differences in RMS travel time differences of 120, 89 and 96 ms. In any case, small details in the iso-velocity contours are below the obtainable resolution and should be disregarded. All models delineate the shape of a ca. 4-km-deep sedimentary basin at the Pripyat Trough as well as very high velocities at the Korosten Pluton, surrounded by areas with low velocity in the near-surface. The resulting models of the two latter procedures indicate that relatively high velocities at the Korosten Pluton may be traced to depths of at least 20 km, perhaps even to the Moho, which in both cases is around 45 km depth as indicated by the 7.75 km/s contour. There is indication for a lateral split of the high velocities in the near-surface at the Korosten Pluton, which appears to be present over a 100–150-km-wide interval, probably widening vertically to 20 km depth.

5.2. Two-dimensional ray-tracing modelling

Variations in amplitude, shape and time duration of both reflected and refracted seismic phases from the crust and uppermost mantle imply variations in the nature of the boundaries and transition zones. Based on all such phases that could be correlated with some certainty in the seismic record sections, we derived a two-dimensional model of the lithospheric structure down to 60–80 km depth (Fig. 6). The method used was iterative, manual travel time fitting, using the ray-tracing package SEIS83 (Červený and Pšenčík, 1983), supported by the programs MODEL and XRAYS (Komminaho, 1998). The basis for the kinematic modelling was a consistent series of travel times, which was picked using traditional seismic phase correlation for refracted and reflected phases. The time differences at reciprocal points do not exceed 50 ms, which is about the uncertainty of picks for the upper crustal refraction Pg. The uncertainties for deeper seismic phases are higher; we assess them to be on the order of 100 ms for crustal refractions and the PmP reflection from the Moho, and 200 ms for intracrustal reflections and the Pn refraction from the sub-Moho mantle. The model includes first-order discontinuities in seismic velocity which may represent major transitions in the lithosphere.

The initial model of velocities was based on the resulting models from the tomographic inversion (Fig. 5), and the overall lithospheric layering was derived from the number of seismic reflection phases (Fig. 7) together with results from previous high-resolution interpretations of seismic velocity in the area. The latter were particularly important for constraining the top basement and the sedimentary sequences, since EUROBRIDGE'97 is designed to image primarily the deep structure.

In the ray-tracing procedure, travel times are successively recalculated for a sequence of models which is constructed by the interpreters in order to improve the fit between observed and calculated travel times for one seismic phase at a time. In the initial modelling, the uppermost velocity structure is determined using the appropriate seismic phases. This is followed by stepwise modelling of phases from layers at deeper levels, usually in a sequence in which the velocity is determined by refracted phases from the relevant layer, after which the thickness of the layer is determined by reflected phases from the interface to the next, deeper layer together with refractions from the deeper layer. Usually only super-critical reflections are strong enough to be identified in the seismic record sections. Changes to the model are assessed by comparison of calculated and observed travel times. This iterative procedure is continued until a reasonable fit has been obtained, as judged from the assessed uncertainty of each modelled seismic phase. Additionally, synthetic seismograms were calculated to control vertical velocity gradients within the layers and velocity contrasts across the seismic discontinuities. The synthetic seismograms calculated for the final model are in qualitative agreement with the observed seismograms, in particular for the relation in amplitude between first arrival, lower crustal reflectivity and the reflection from the Moho (PmP) (Fig. 7). However, some difference is also evident for the sub-Moho refraction Pn as well as some of the upper mantle reflections. In some of the presented examples, the Pg to PmP relation is not reproduced very well in certain offset intervals, generally between 110 and 160 km. We ascribe the discrepancies mainly to the use of ray-tracing techniques for the calculation of the synthetic seismograms in a complex model. The detailed velocity model of the lithosphere along profile EUROBRIDGE'97 from ray-trace modelling is shown in Fig. 6. Fig. 8 shows a simplified sketch of the lithospheric model, which also indicates which parts of the model that are constrained by raypath coverage. It is based on the ray-tracing model (Fig. 6), the tomographic inversion models, and qualitative implications from the observed wavefield.

6. P-wave velocity model

There is significant lateral variation of seismic velocity along the EUROBRIDGE'97 profile (Fig. 6a,b). Pronounced lateral transitions were modelled in the upper crystalline crust at 150–210 km (beneath the southern half of the Pripyat Trough) and 300–350 km. The exact form and nature of these zones cannot be resolved with the DSS technique, which relies on modelling of wide-angle reflected and refracted waves. The zones separate crustal units (blocks) of different velocity structure in the north, centre and south of the profile, which can be spatially correlated with the Osnitsk–Mikashevichi Igneous Belt, the Volyn Block including the Korosten Pluton, and the Podolian block.

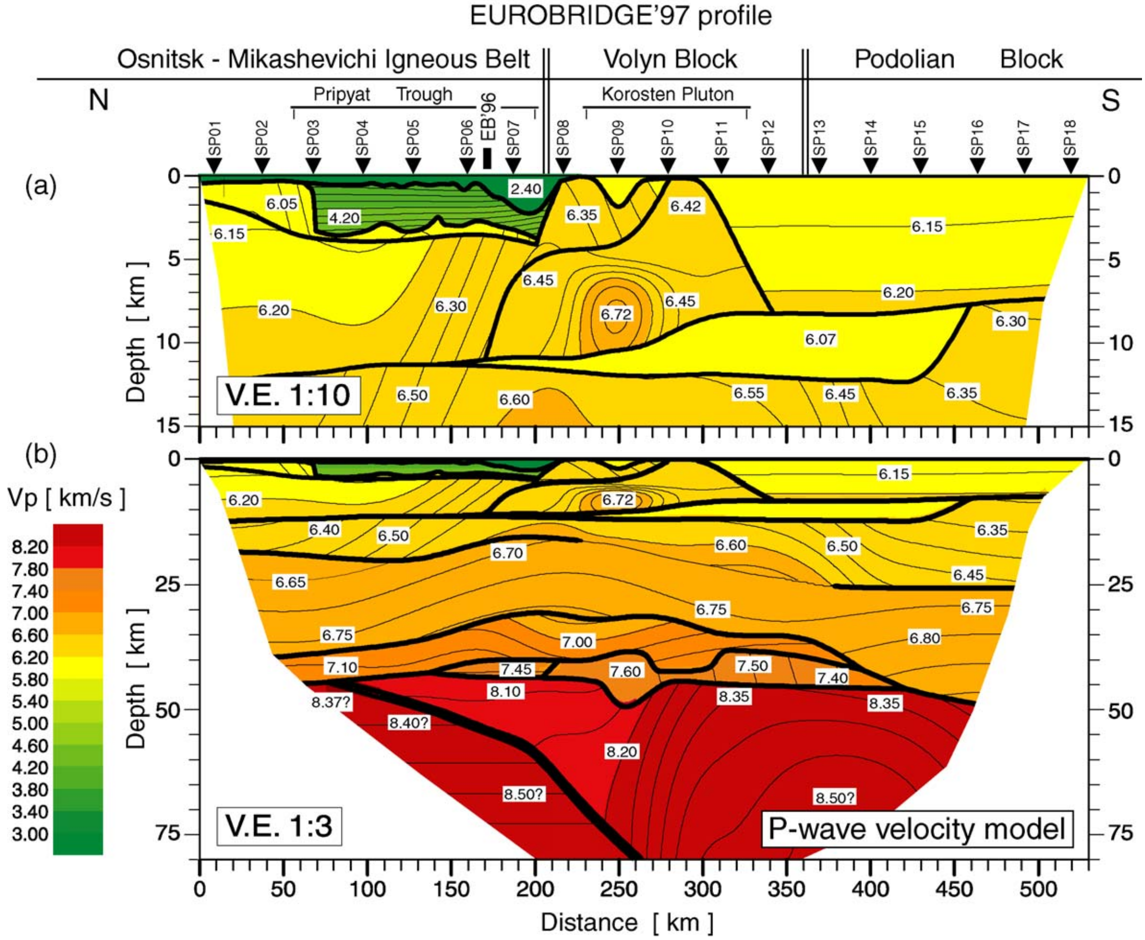


Fig. 6. Two-dimensional models of seismic P- and S-wave velocity along the EUROBRIDGE'97 profile in Belarus and Ukraine. P-wave velocity model: (a) upper crustal details; (b) the complete lithospheric model. Thick, black lines represent major velocity discontinuities (interfaces). The dipping reflector in the mantle is constrained by reflections only, hence the velocity below this feature is unknown. (c) S-wave velocity model. Those parts of the first-order discontinuities that have been constrained by reflected and/or refracted arrivals of S-waves are marked by thick lines. (d) Model of Vp/Vs ratio distribution. Those parts of the first-order discontinuities that have been constrained by reflected or refracted arrivals of P- or S-waves are marked by thick lines. Thin lines represent velocity isolines with values (km/s) shown in white boxes. Locations of major basement tectonic units and the Pripyat Trough are indicated. Arrows show positions of shot points. The location where EUROBRIDGE'97 crosses the EUROBRIDGE'96 profile is marked as EB'96.

The most complicated and significant structures are in the upper crust down to 10–15 km depth, as well as in the lower crust below a depth of 35 km. In the northern part of the profile (km 60–220), low velocities of ~ 2.4–4.2 km/s in an approximately 4-km-thick layer represent the sedimentary fill of the Pripyat Trough. The centre of the profile, in the vicinity of the Korosten Pluton (Fig. 1) is characterised by upper crustal velocities of 6.4–6.7 km/s, which are extremely high given the shallow depth interval (0–10 km) in which they occur (Fig. 7a). In the northern and southern parts of the profile, top crystalline basement velocity values are 6.1–6.2 km/s. In the south, a weak low-velocity zone (LVZ) underlies this basement and part of the high-velocity zone of the Korosten Pluton. The LVZ has been modelled with a rather small

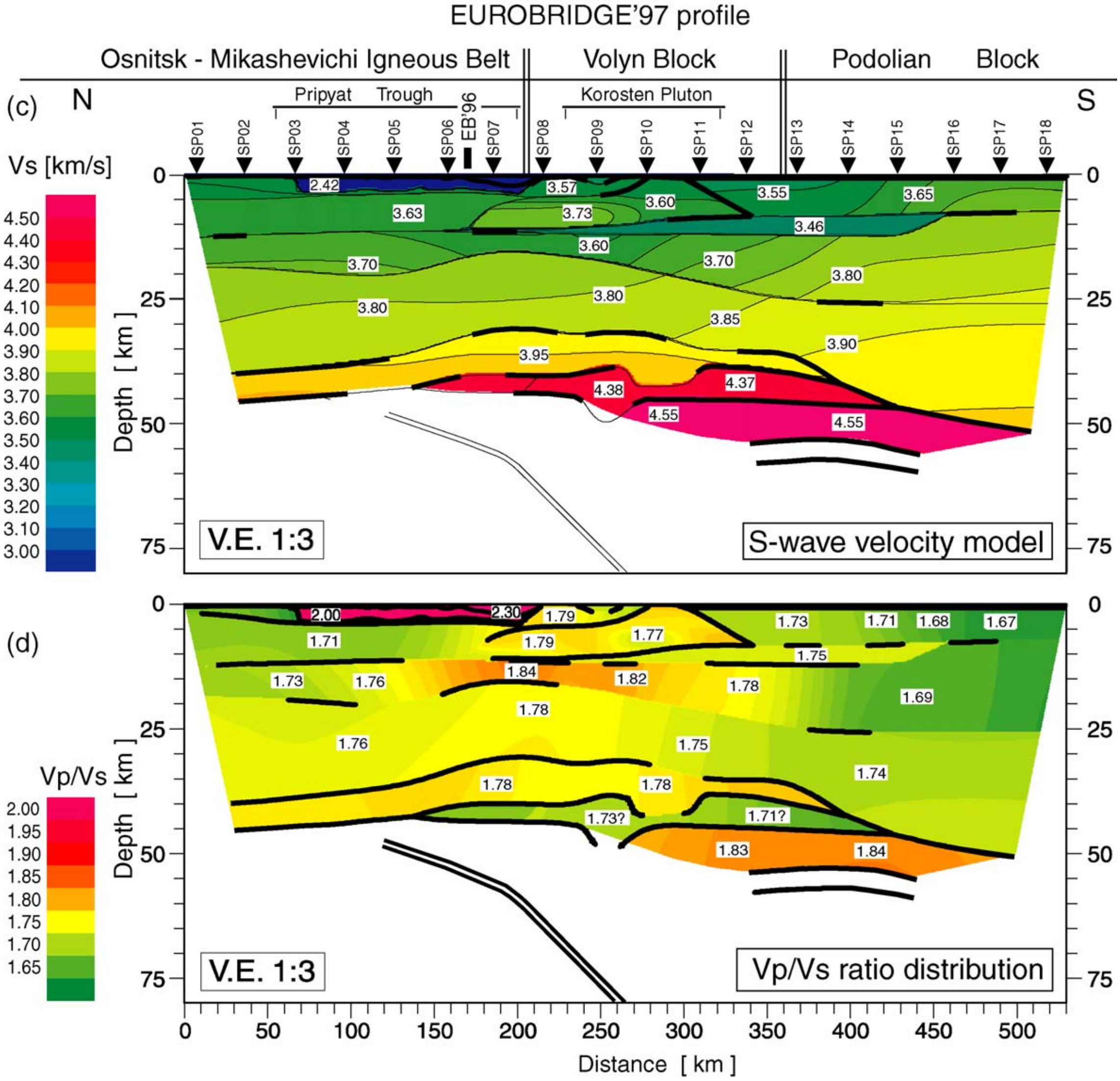


Fig. 6 (*continued*).

velocity contrast of 0.1 km/s to the surrounding rocks, but the exact velocity is unconstrained by the available data.

Below ~ 15 km depth the middle to lower crust is represented by two layers with velocities of ~ 6.4 and ~ 6.7 km/s. We only observe sporadic reflections between these layers (Fig. 8). Deeper crustal units are absent at the southern end of the profile (at distances of more than 450 km), where a relatively simple, ''normal type'' of Precambrian crustal structure with a three-layered crust (e.g. Meissner, 1986) exists within the Podolian Block. A deeper, third zone of high velocity is found elsewhere, ca. 5 km thick within the Osnitsk–Mikashevichi Igneous Belt and thicker and more complicated in the Volyn Block. This lowest crustal layer is characterised by very high velocities of 7.0–7.4 km/s. At the base of the crust, particularly in the central part of the profile within the Volyn Block, there is a complicated structure with velocities exceeding 7.4 km/s. This lowermost layer of the crust is associated with large contrasts in P- and S-wave velocities which may explain the observation of strong PmP and SmS arrivals from the centre part of the profile. However, the exact velocities are relatively uncertain, as they are only constrained from amplitude modelling of reflections from the interval, and not from refracted arrivals. The observation of a very strong reflection from the Moho shows that the velocities of this thin lowermost crustal layer must be well below the velocities in the sub-Moho mantle. Within the northern and southern parts of the profile, the lowermost crust and uppermost mantle include zones of pronounced reflectivity at km 60–130 and km 310–420, respectively, as interpreted from reverberative reflectivity in front of and behind the PmP reflection (Fig. 9). In contrast, the reflectivity indicates rather sharp velocity contrasts in the central part of profile, where a complicated lower crustal transition has been modelled.

The crust is about 46 km thick, with a slight thickening to about 50 km in the southern part of the profile (around km 450). The sub-Moho velocity

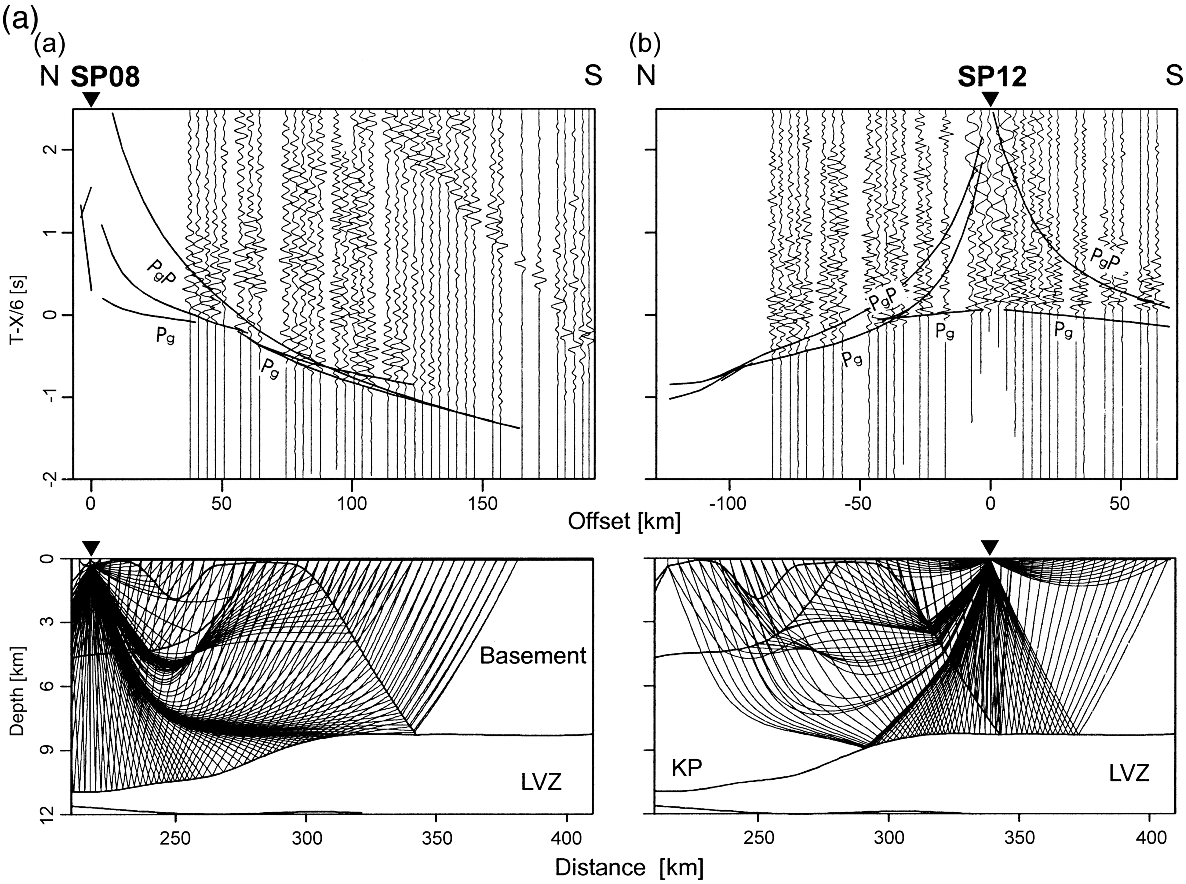
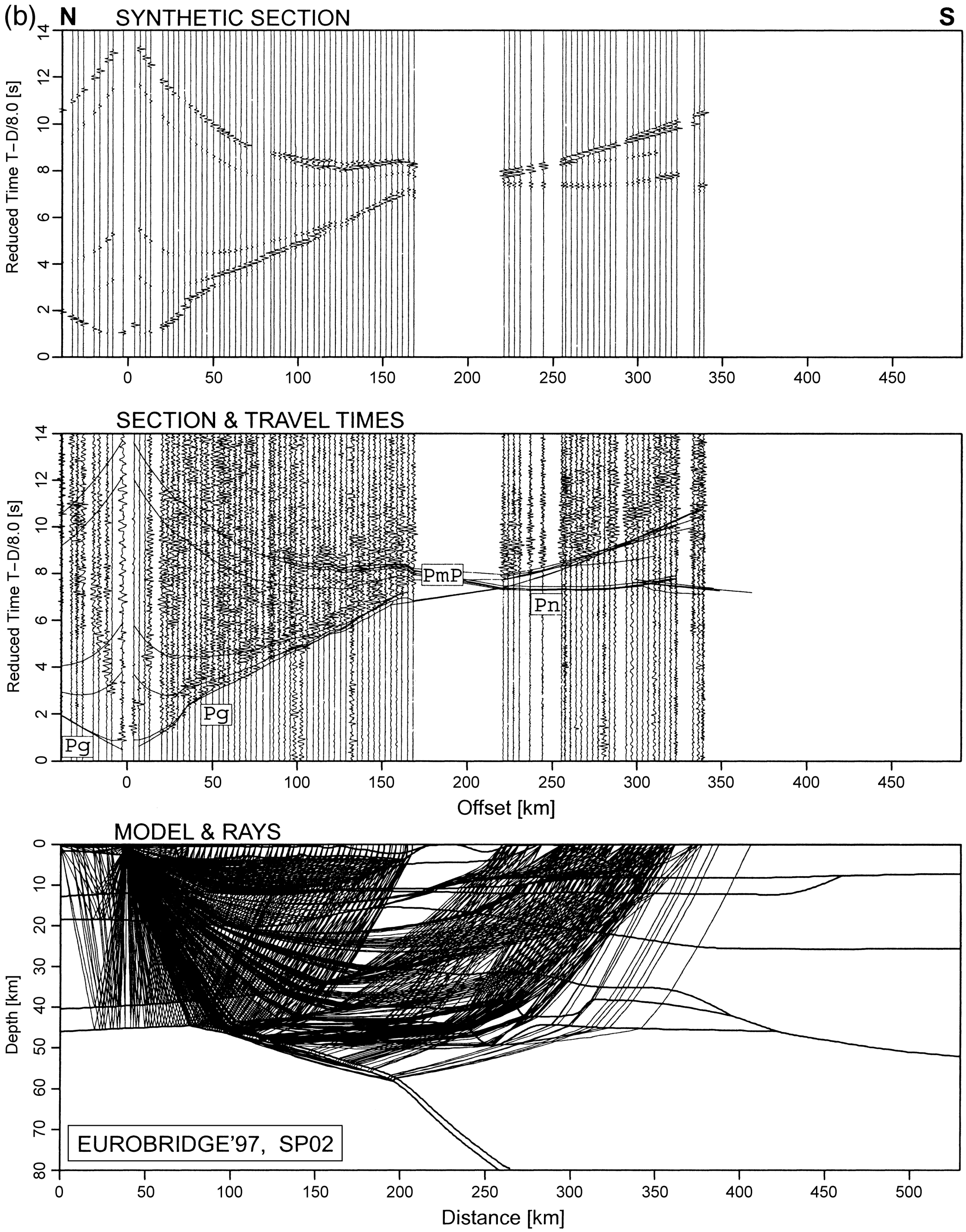


Fig. 7. Ray-trace modelling results for selected shot points of EUROBRIDGE'97 using the P-wave model of Fig. 6. Abbreviations: Pg—seismic refraction from the upper crystalline crust; PgP—seismic reflection from an intra-basement horizon; LVZ—low-velocity zone; PmP—seismic reflection from the Moho; Pn—refraction from the sub-Moho upper mantle; P1—arrivals from the lower lithosphere. Reduction velocity is 8.0 km/s except in (a) where it is 6.0 km/s. (a) Documentation of the high P-wave velocity in the Korosten Pluton (KP) for shot points SP08 and SP12. (b) Ray-trace modelling results for shot SP02. The upper diagram shows the synthetic sections, which should be compared with the data in terms of the relative amplitudes within a trace. The middle diagram shows the amplitude-normalized, vertical component seismic section, overlain by the computed travel time curves. The lowest diagram shows the ray coverage for the shots. (c) Ray-trace modelling results for shot SP11. (d) Ray-trace modelling results for shot SP14. (e) Ray-trace modelling results for shot SP18. See panels a and b for further explanation.



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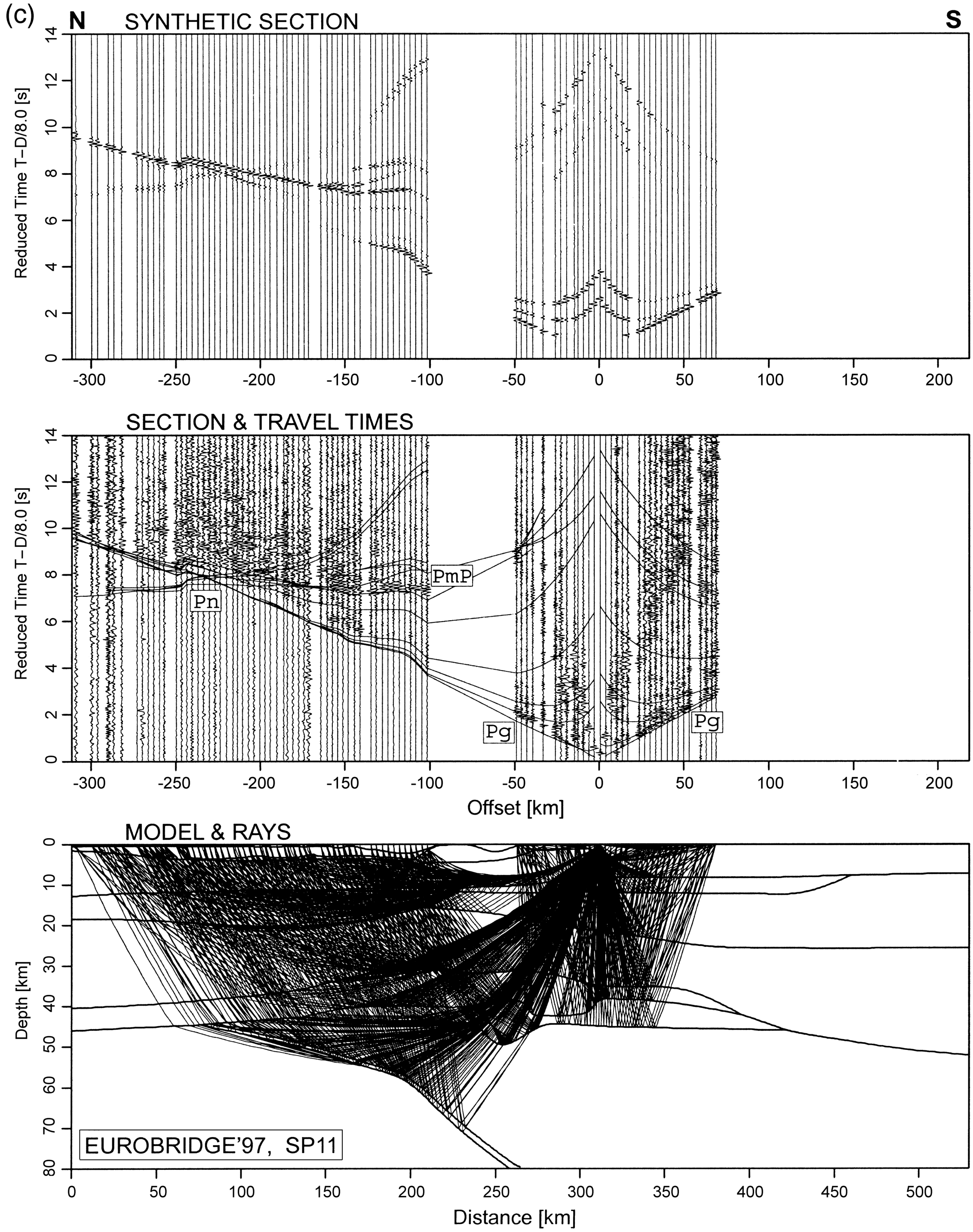

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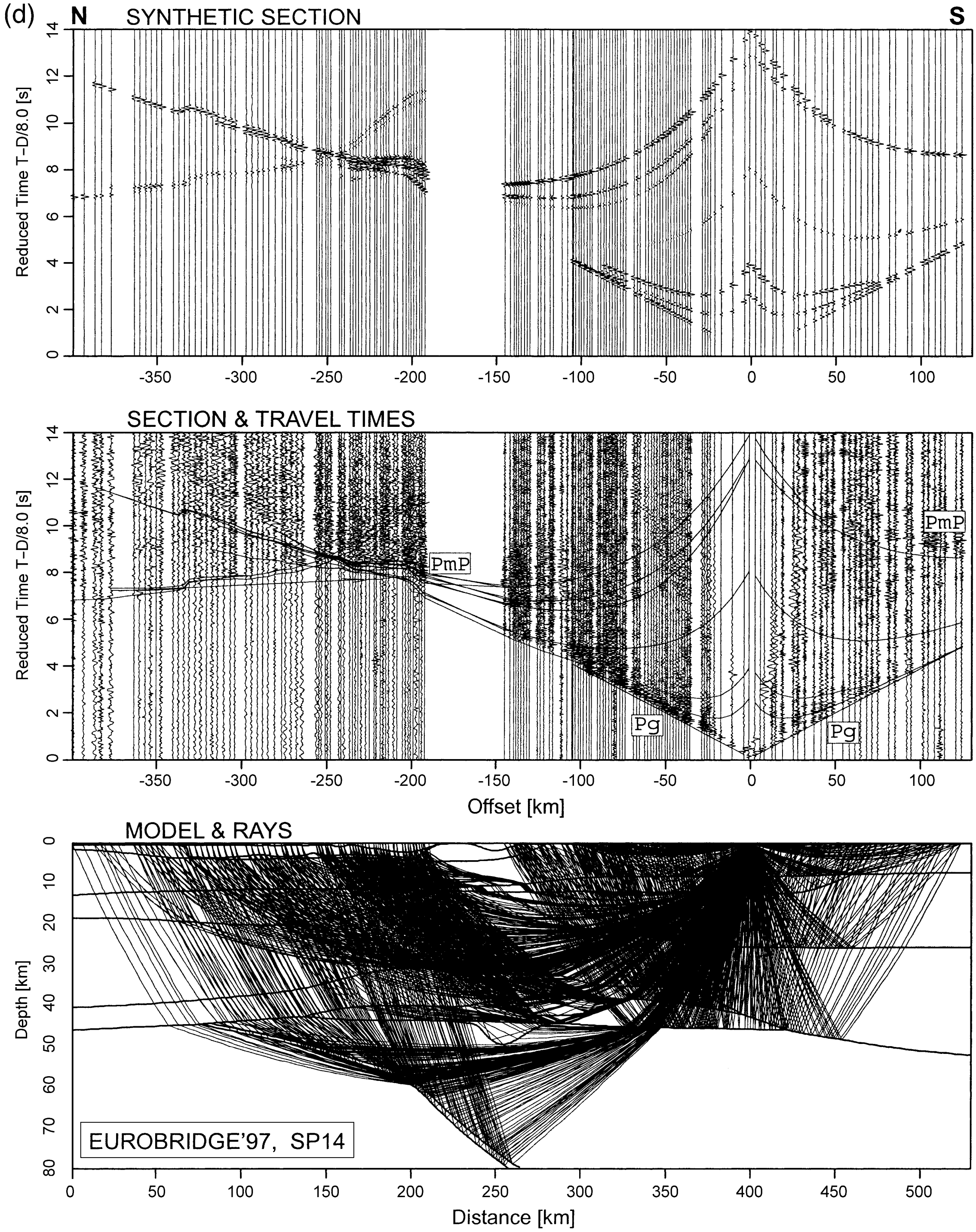

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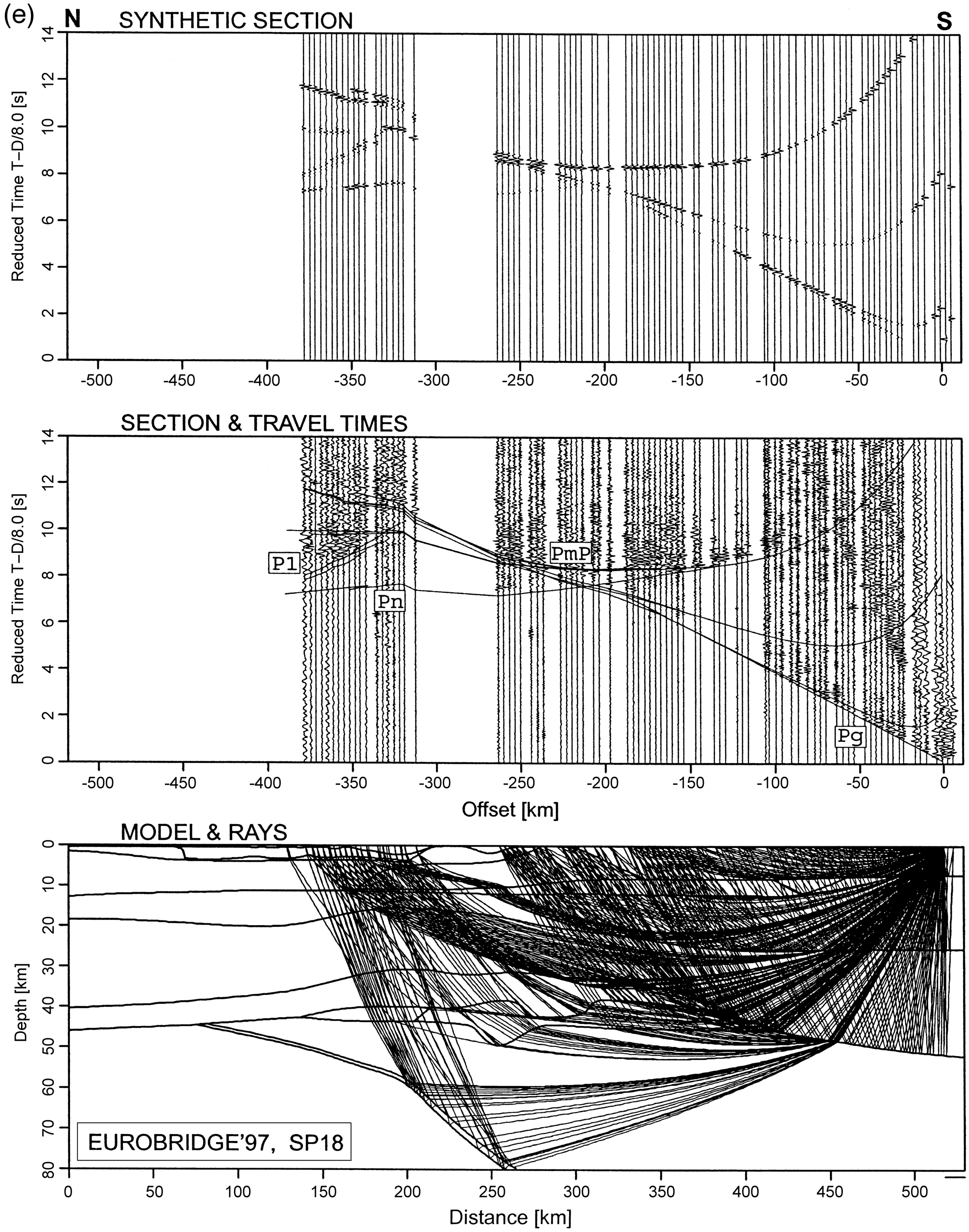


Fig. 7 (*continued*).

(Pn velocity) is as high as 8.3–8.4 km/s along most of the profile, except for the central part of the profile, where it is ~ 8.1 km/s between km 80 and 260.

The cause of the P1 phase is a dipping mantle interface. It may either represent a change in lithology, i.e. a first-order seismic discontinuity between layers with different velocity, or it may represent a relatively narrow mantle zone of high-velocity gradient and inhomogeneity without any noticeable change in seismic velocity across the interface, e.g. a tectonic shear zone. Both possibilities produce rather similar synthetic seismograms. The southward apparent dip from ~ 46 km depth at km 120 to ~ 80 km depth at km 260 along the EUROBRIDGE'97 profile is well constrained. The complex wave field indicates increased reflectivity at the dipping reflector at about 60 km depth at km 170–200.

7. Analysis of S-waves

7.1. S-wave field description

The recorded seismic data shows S-wave signals of remarkable high quality and signal/noise ratio along the whole profile, best in the southern shield part. Both primary and secondary refracted and reflected S-wave phases have been correlated in all 18 vertical

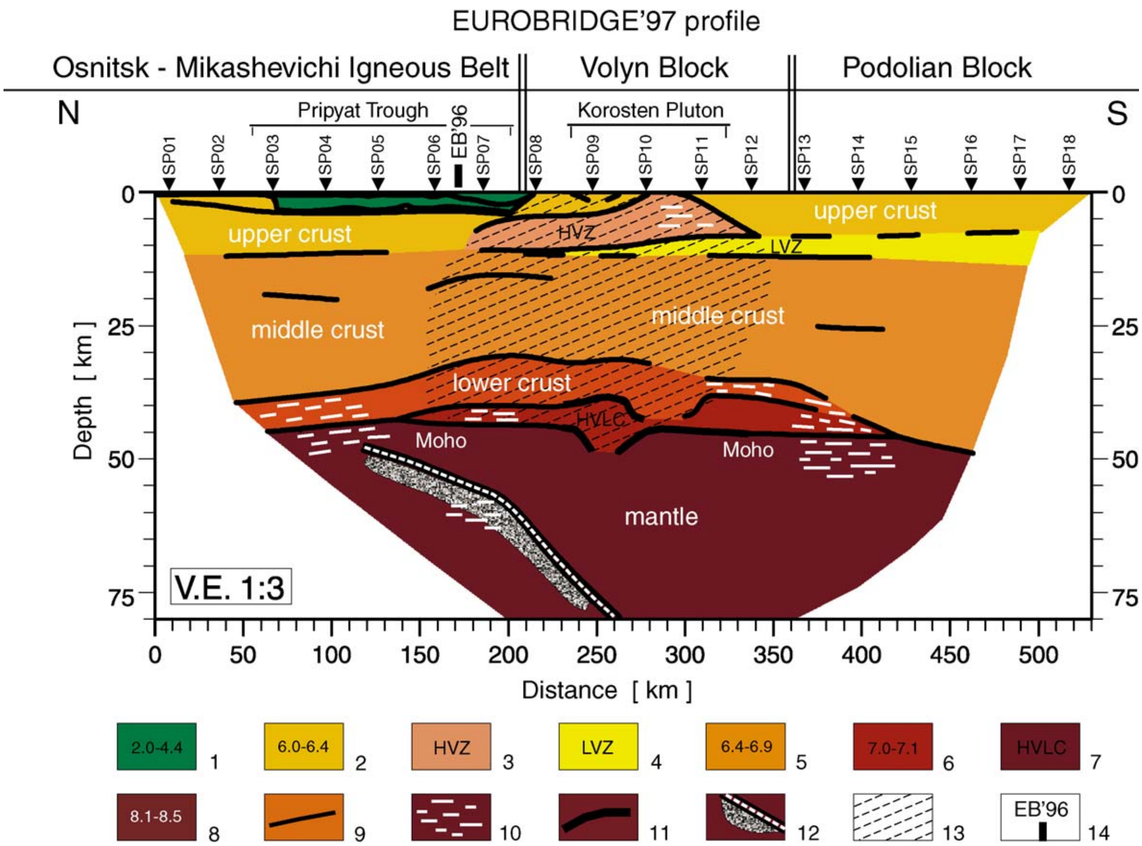


Fig. 8. Simplified sketch of the lithospheric structure derived along the EUROBRIDGE'97 profile. (1) Sedimentary cover (Vp = 2.0–4.4 km/s); (2) crystalline uppermost crust (6.0–6.4 km/s); (3) high-velocity zone associated with the Korosten Pluton; (4) upper crustal low-velocity zone; (5) middle crust (6.4–6.9 km/s); (6) lower crust (7.0–7.1 km/s); (7) high-velocity lower crust; (8) uppermost mantle (8.1–8.5 km/s); (9) sections of major boundaries interpreted from P-wave refractions and reflections; (10) zones of high reflectivity; (11) strong reflectors; (12) mantle reflector or zone of high-velocity gradient; (13) zone of anomalously high velocity, probably associated with the Korosten Pluton; (14) point of intersection with the EUROBRIDGE'96 profile. Shot point locations are shown by triangles above the profile.

component record. From previous experience (EUROBRIDGE Seismic Working Group, 1999), we find the vertical component as good as the horizontal components for detection of S-waves in the record sections. This may be surprising, but it is often the case for crustal refraction data. Record sections with a reduction velocity of 4.52 km/s were prepared for correlation of S-wave phases using a band-pass filter of 1–8 Hz and the axes scaled by a factor of $3^{-0.5}$ to match the corresponding P-wave sections with a reduction velocity of 8 km/s (Fig. 9).

In the northern part, in the interval of km 0–200, the observed initial ("first") Sg arrivals are weak compared to the background P-wave coda and strong secondary S-wave arrivals (Fig. 9a,c,d). S-wave reflections from the lower crust are clearly identifiable beyond offsets of approximately 50 km in record sections from the northern shots (SP01–07; Fig. 9a,c). In the southern part of the profile, high-quality, clear S-waves are observed at all offsets (SP08–18; Fig. 9b,d–f).

The high-quality data to the south of km 200 allows exact phase correlation of refracted and reflected S-waves on most record sections. The differences in picked travel times between the two directions usually do not exceed 100 ms at the reciprocal points. Remarkably, Sg arrivals are strong enough, with very high signal/noise ratio, to be comparable to Pg arrivals in the southern part of the profile (Fig. 9). In the northern part, Sg arrivals are weaker than the corresponding Pg arrivals and it is usually only rarely possible to correlate the Sg phase in the background seismic noise and P-wave coda.

S-wave reflections from the middle crust are only sparsely identified. Reflections from the lower crust and from the Moho discontinuity (SmS) are the strongest S-waves observed in all sections. These waves are recorded in many sections over long offset ranges from 50–80 to 300–350 km. In the southern part of the profile, strong, well-correlated SmS arrivals indicate a large velocity contrast across the Moho (e.g. SP16 and SP18; Fig. 9b,f). In contrast, arrivals in the northern part of the profile are only identified as an envelope around an amplitude increase (e.g. SP02; Fig. 9c), indicative of a smaller S-wave velocity contrast across the Moho than in the south. S-wave reflections from the upper mantle were detected at offsets of 320–480 km (e.g. SP16, Fig. 9b). Direct comparison of the travel times of P- and S-waves show large deviations of the Vp/Vs ratio from the average value of 1.73 for the whole depth range covered.

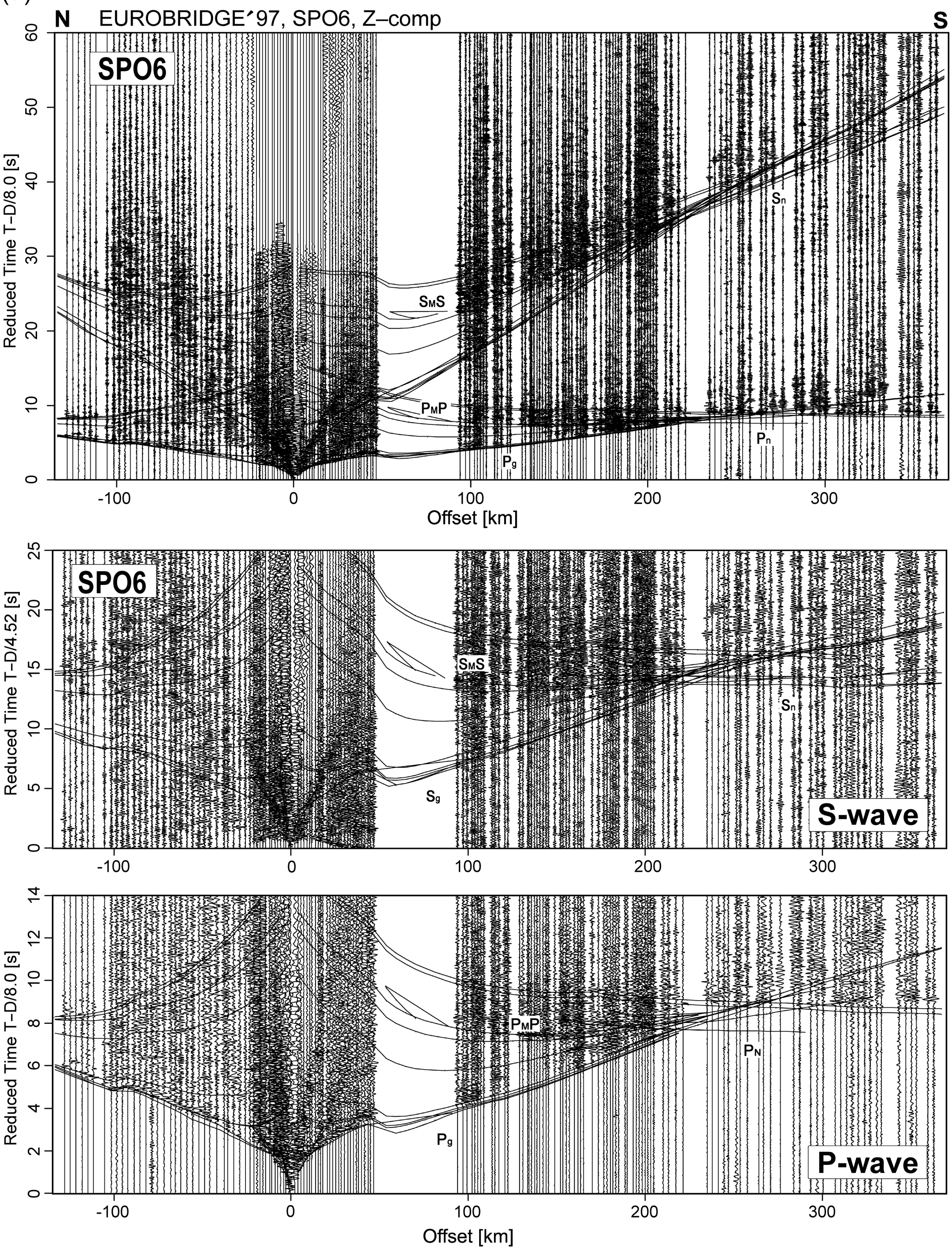
7.2. S-wave velocity and Vp/Vs ratio models

The quality of S-wave arrivals in the northern 200 km of the profile allows only a limited number of travel time picks to be made. Nevertheless, the main secondary and some primary S-wave phases may be picked, although with uncertainty. This precludes modelling of an independent S-wave model there. Instead, the best branches of correlated S-wave travel times were used to estimate the S-wave velocity distribution model (Fig. 6c), from which the distribution of Vp/Vs ratio is calculated for the principal layers of the crystalline crust and upper mantle (Fig. 6d). The geometry of discontinuities in the S-wave velocity model was inherited from the P-wave velocity model (Fig. 6a,b). The travel time picks to the south of km 200 are of sufficient quality for modelling of an independent S-wave velocity model. Also for this modelling, the first-order discontinuities were inherited from the P-wave model.

Pronounced lateral changes in Vp/Vs ratio were modelled in the upper and middle crystalline crust at distances of km 150–210 and km 300–350 (Fig. 6c), where the main changes in P-wave velocity also occur (Fig. 6a,b). These zones divide the model into three blocks. There are coincident, lateral changes in the S-

Fig. 9. Data examples that illustrate P- and S-wave seismic sections, with modelled arrival times superimposed. All data examples are shown as amplitude-normalized vertical component record sections with a band-pass filter 1–8 Hz applied. Two examples show both P- and S-wave arrivals in the same sections using a reduction velocity of 8.0 km/s (upper panels of a and b). All examples show P- and S-wave sections with reduction velocities of 8 and 4.52 km/s, respectively. The ray coverage for P-waves of the seismic model for the respective shot points is shown at the bottom of panels c–f. Theoretical travel time curves calculated for the models of seismic velocity in Fig. 6 are superimposed on the sections. By use of a relative scaling factor of 1.73 between P- and S-wave record sections, the travel times are directly comparable for a Poisson's ratio of 0.25. Abbreviations: Pg, Sg—seismic refractions from the upper crystalline crust; PmP, SmS—seismic reflections from the Moho; Pn, Sn—refractions from the sub-Moho upper mantle; P1—P-wave phase from the lower lithosphere. (a) SP06, (b) Shot SP16, (c) Shot SP02, (d) Shot SP08, (e) Shot SP10, (f) Shot SP18; see Fig. 8a for further explanation.

(a)


N
EUROBRIDGE'97, SPO6, Z–comp
S
SPO6
Reduced Time T-D/8.0 [s]
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Offset [km]
SPO6
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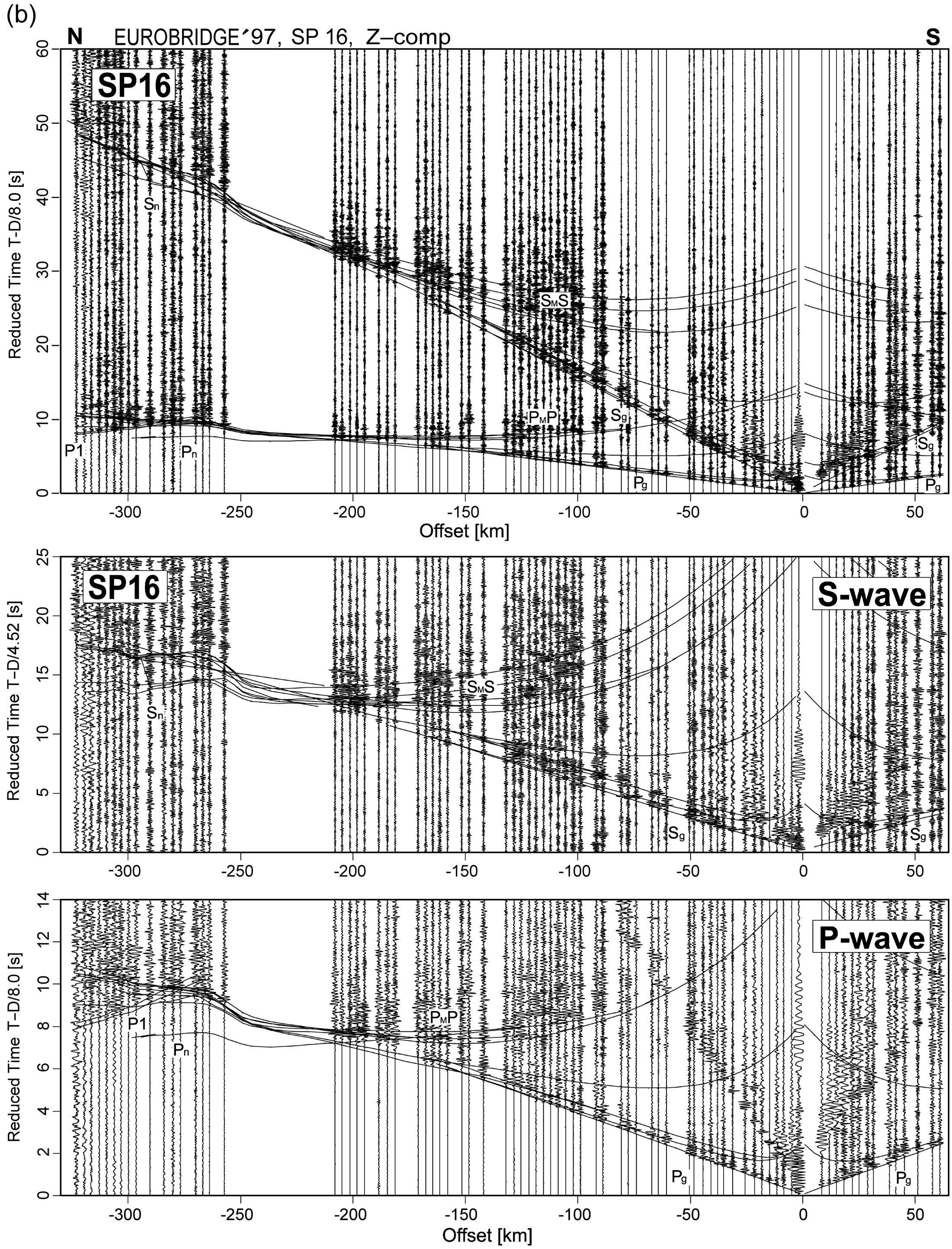

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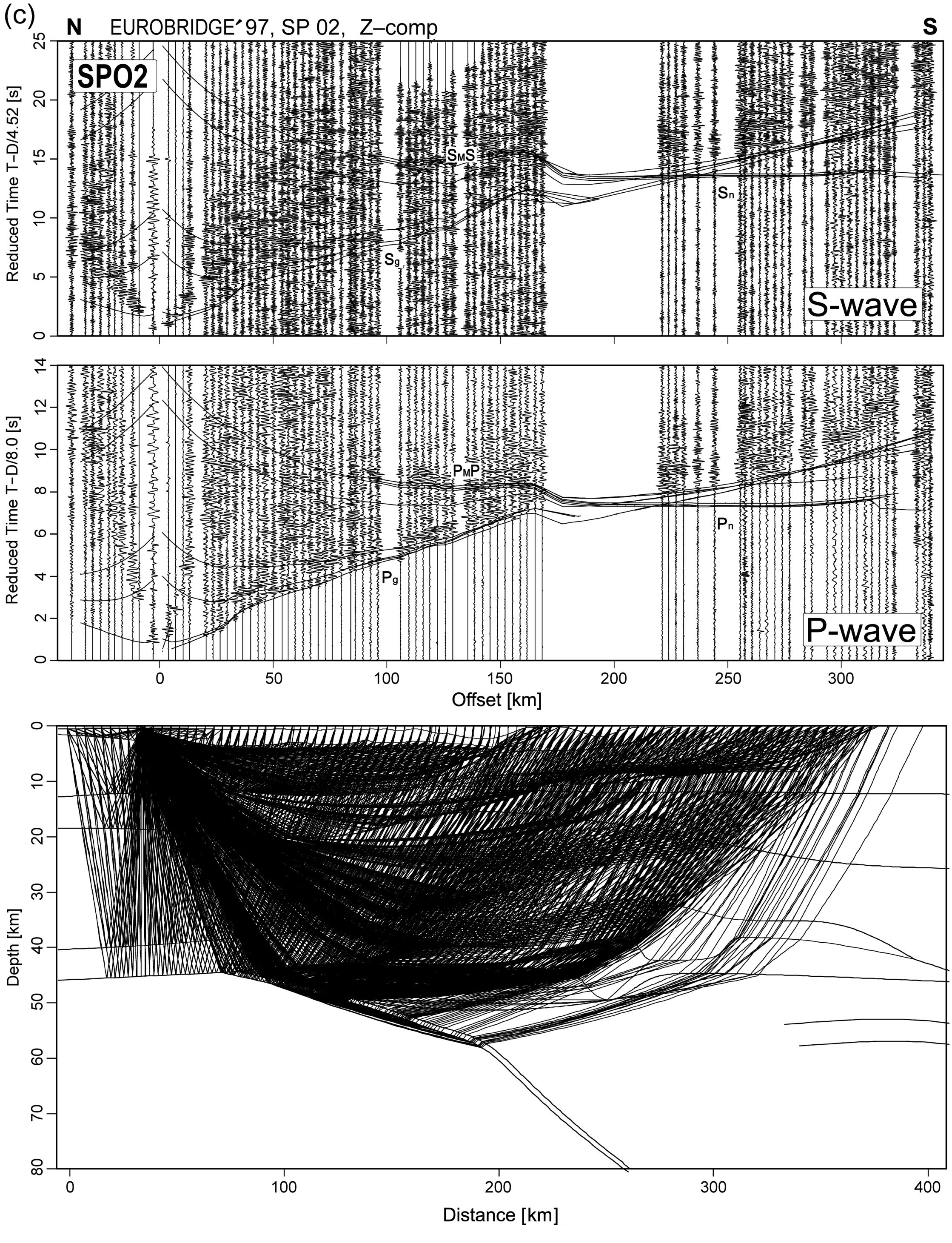

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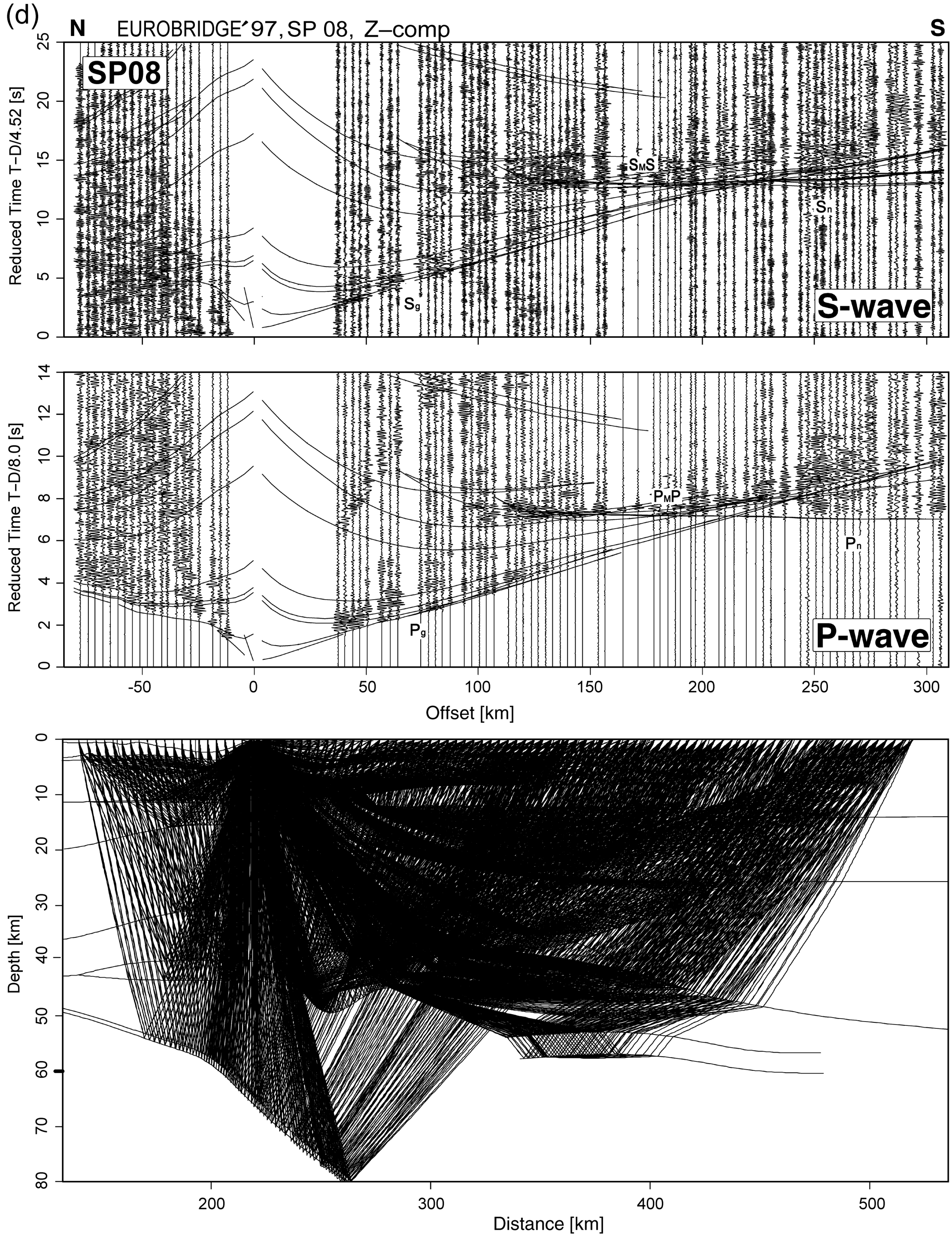
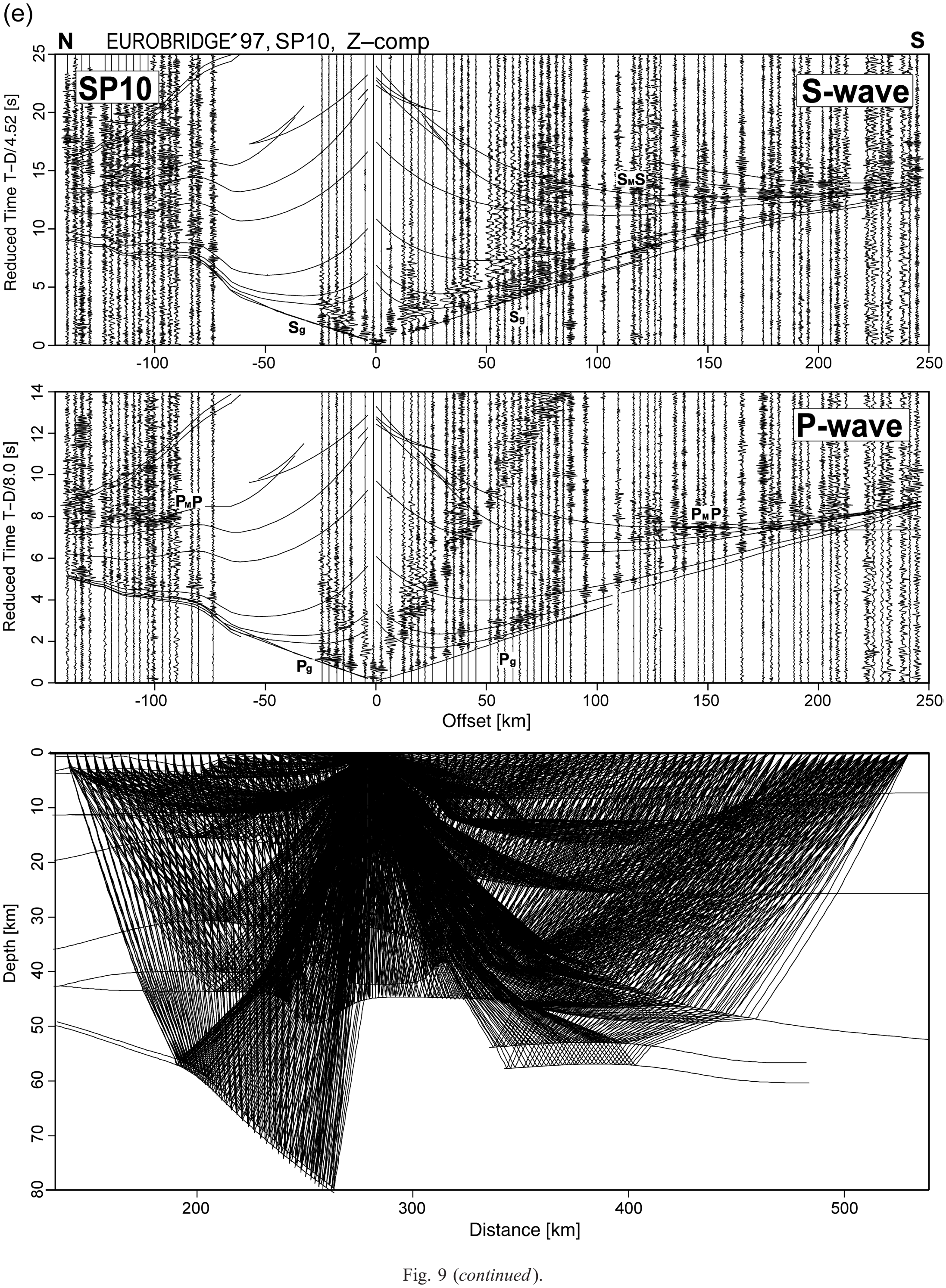


Fig. 9 (*continued*).



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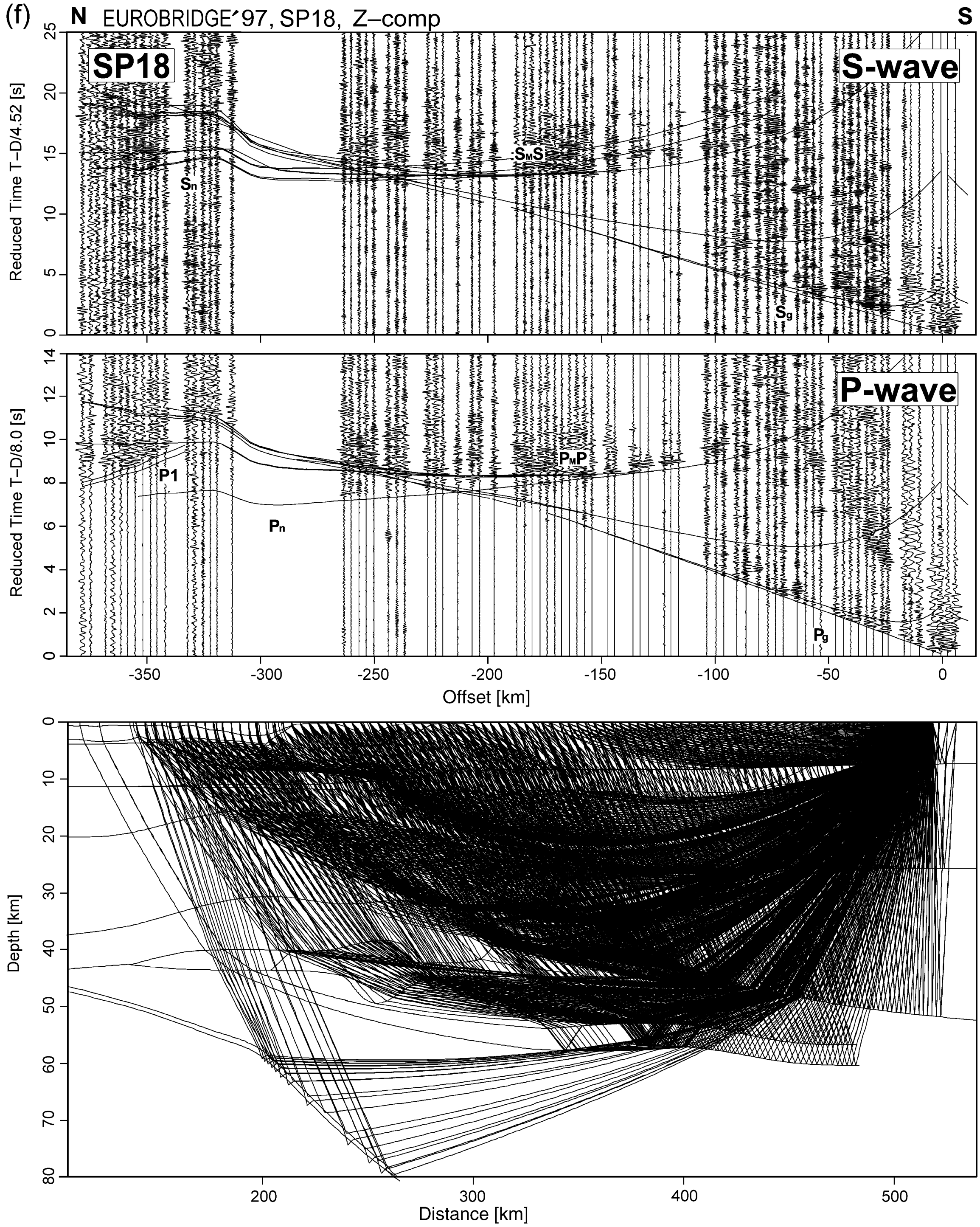

Fig. 9 (*continued*).

wave velocity (Fig. 6d), although the variation is smaller than for the P-wave velocity. The sediments of the Pripyat Trough in the northern block have Vp/Vs ratios as large as 2.0–2.3. These values were adapted from a detailed study by Molotova and Vasilev (1960). In the crystalline crust of the northern Osnitsk–Mikashevichi Igneous Belt, the Vp/Vs ratio increases with depth from 1.72 to 1.78, although not well constrained because of the low signal/noise ratio for most crustal S-wave arrivals in this part of the profile. The model values provide the correct mean velocity for the whole crust, as can be checked from the high-quality SmS reflections. The S-wave velocity increases from 3.63 km/s in the upper crust to 3.80 km/s near the Moho.

The "high-velocity body" in the central part of the profile, associated with the Korosten Pluton, has much higher Vp/Vs ratio in the upper and middle crust than the neighbouring blocks: 1.77–1.79 for the two main upper crustal layers, corresponding to S-wave velocity values of 3.57–3.73 km/s; 1.75 was chosen for the LVZ, although non-constrained by the data (in the central and southern blocks); 1.78–1.84 at depths of 15–20 km (Vs ~ 3.60 km/s). Such extraordinary high values of Vp/Vs were also measured in samples of biotite-bearing gneisses and amphiboles from similar depth in the deep drill hole on the Kola Peninsula in the northeastern part of the East European Craton (Kern et al., 2001).

The southern block of the Ukrainian Shield has the lowest values of Vp/Vs ratio in the model: in the upper crust it varies from 1.71–1.73 (Vs ~ 3.55 km/s) at 300–430 km distance to 1.67–1.68 (Vs ~ 3.65 km/s) at 430–500 km distance. This variation is significant because the P-wave modelling indicates a homogeneous upper crust. Below a depth of 12 km, Vp/Vs exhibits no lateral variation with a vertical increase from 1.69 to 1.74 at the Moho (Vs ~ 3.75–3.95 km/s).

The S-wave velocities in the lowest crust and uppermost mantle of the central block are about 3.95, 4.25–4.45 and 4.55 km/s. These values are high compared to the surrounding areas, similarly to the high P-wave velocities. A high Vp/Vs of 1.78 was also found for the main lower crustal layer, whereas the lowest, thin transition layer into the uppermost mantle has been modelled with a low Vp/Vs ratio, in order to explain the high amplitude of the reflectivity around the SmS phase. The sub-Moho mantle exhibits extremely high values of Vp/Vs of ~ 1.83 in the part of the model, where a reliable estimate can be given based on Sn arrivals.

There are some discrepancies between the P- and S-wave models. At a depth of 55 km between km 340 and 480 along the profile, the S-wave record sections indicate the existence of a strong reflector in the upper mantle (c.f. the last reflection at 90–200 km offset for SP16; Fig. 9b). There is no indication for a similar structure from the P-wave sections, which instead indicate a highly reflective uppermost mantle in high frequencies at these distances from reverberative arrivals (SP16). The S-wave data further indicates that the Moho is flat around km 255, where the P-wave data indicates a distinct flexure.

8. Discussion

8.1. Crustal blocks and palaeo-terranes

The seismic models of the EUROBRIDGE'97 profile define three blocks (or segments of palaeo-terranes of different affinities), based on strong lateral contrasts in the P- and S-wave velocities and the Vp/Vs ratio. The centre block corresponds to the basement area of the Volyn Block, in which the profile mainly traverses the Korosten Pluton. The southern part corresponds to the Podolian Block of the Ukrainian Shield. The northern part is fully within the Osnitsk–Mikashevichi Igneous Belt of the Palaeoproterozoic active margin of Sarmatia, where the profile mainly traverses the area of the Pripyat Trough. Despite the changes in surface conditions from a thin platform cover and a sedimentary trough in the north to exposed crystalline basement in the shield area, the vertical velocity profiles in the northern and southern blocks are very close to the average velocity profile of typical Precambrian cratons. A ~ 45-km-thick crystalline crust is expected in such areas, consisting of three main components with P-wave velocities of ~ 6.1, ~ 6.5, and ~ 7.0 km/s (Meissner, 1986). The EUROBRIDGE'97 profile shows similar velocities, although the deep crustal velocity may be slightly low in the Osnitsk–Mikashevichi Igneous Belt where a thin layer above Moho instead has a velocity substantially higher than 7.0 km/s. The occurrence of P-wave velocities higher than 7.0 km/s and relatively thick crust (~ 50 km) appear to characterise this part of the East European Craton as the 'Proterozoic type' described by Grad and Luosto (1987), BABEL Working Group (1993), Grad and Tripolsky (1995), EUROBRIDGE Seismic Working Group (1999), and Środa and POLONAISE Profile P3 Working Group (1999).

The EUROBRIDGE'97 profile crosses the boundaries between three distinct basement areas or blocks of the Sarmatian crustal segment of the East European Platform. The velocity model shows pronounced differences between these blocks. However, the main reason for these changes may be the presence of the Korosten Pluton which was emplaced in the Volyn Block and constitutes a large part of this block along the profile. It appears evident that this overprinting of the original velocity structure of individual terranes is responsible for the main contrasts. The combined EUROBRIDGE'95 and '96 profiles also show distinct lateral changes in the velocity structure between all the Palaeproterozoic orogenic belts along these profiles, although changes are less pronounced than along the current profile. The contact zones between the three main blocks in the EUROBRIDGE'97 profile are pronounced, but the applied DSS method does not allow assessment of the width of the transitions; the best estimate obtainable with the current resolution is that they are less than 50 km wide. The profile clearly demonstrates that the transitions between the individual blocks have been preserved since the Palaeproterozoic as changes in physical properties. However, younger processes have modified the velocity structure without destroying the fine-scale reflectors in the upper to middle crust, which are probably related to faulting, either during crustal assemblage or during the subsequent formation of the Pripyat Trough.

At the cross point, the velocity structure is in agreement between the two EUROBRIDGE profiles, except for the depth to the lower crustal layer characterised by a P-wave velocity of ~ 7.0 km/s (~ 27 km in the EUROBRIDGE'96 profile and ~ 31 km in the EUROBRIDGE'97 profile). This discrepancy may be caused by the uncertainty in interpreting the lateral extent of the elevated velocities at the Korosten Pluton along the EUROBRIDGE'97 profile. The EUROBRIDGE'97 model shows slightly increased velocity associated with the Korosten Pluton at the cross point, contrary to the EUROBRIDGE'96 model. This indicates that the Korosten Pluton anomaly is sharper than shown in the velocity model, i.e. that the lateral velocity gradient is too small in the model at the Korosten Pluton. Hence, the elevated velocity may be underestimated and extend too far to the north in the present model. However, the lateral data coverage along EUROBRIDGE'97 is much better than along EUROBRIDGE'96, such that the current profile expectedly provides the best resolution of the seismic velocity structure. The EUROBRIDGE'96 interpretation includes a pronounced zone of lower crustal reflectors below ~ 20 km depth in the Osnitsk–Mikashevichi Igneous Belt. The EUROBRIDGE'97 data also shows lower crustal reflectivity from the same block, but only below ~ 35 km depth and the character of the reflectivity in this direction indicates a fine-scale layering, rather than distinct reflectors as along the other profile. Seismic reflection sections in the proximity of the profiles also show lower crustal reflectivity in near-normal-incidence. EUROBRIDGE Seismic Working Group (1999) suggests that the reflectivity may be either a characteristic of the Osnitsk–Mikashevichi Igneous Belt or related to the formation of the Pripyat Trough. With the new velocity model along the EUROBRIDGE'97 profile, we find it unlikely that the reflectivity can mainly be caused by structure from the development of the Pripyat Trough. We prefer the first explanation, namely that the elevated seismic velocity and pronounced reflectivity in the lower crust are characteristics of the Osnitsk–Mikashevichi Igneous Belt. The change in reflectivity characteristics and depth between the two profiles indicates anisotropy in the fine structure of the lower crust.

The general intracrustal, wide-angle reflectivity is high at all crustal levels in the northern part of the Osnitsk–Mikashevichi Igneous Belt, although reflections are not observed from any pronounced first-order discontinuities. It could be speculated that the reflectivity is only apparent and caused by the platform cover instead of intracrustal reflectors. However, the strongly changing conditions along the EUROBRIDGE'97 from normal platform cover to a trough make it unlikely that the sedimentary sequences may be the primary explanation for the reflectivity. Hence, dispersed crustal reflectivity is another characteristic of the Osnitsk–Mikashevichi Igneous Belt, as in all the seismic sections from the northern part of the EUROBRIDGE'97 profile.

The crustal structure of the Ukrainian Shield proper in the southern part of the EUROBRIDGE'97 profile in the Podolian block, is relatively uniform and without pronounced lateral variation, except for a pronounced change in Vp/Vs ratio which is not understood at present. Most of the lateral change in velocity appears to be related to the evolution of the Korosten Pluton, although a seismic low-velocity zone at around 10 km depth is only interpreted in the southern part of the profile and not in the southernmost part of the profile. It is remarkable that the crustal thickness increases away from the Korosten Pluton to ∼ 50 km, which may be the average thickness of the crust in the western Ukrainian Shield.

8.2. Pripyat Trough and magmatic intrusions

Comparison of the velocity model with the parallel seismic reflection profile VIII (Juhlin et al., 1996; Stephenson et al., 2001) shows some correspondence between the velocity layers and intracrustal reflectivity (Fig. 10). There is clear correspondence between the two profiles at the Pripyat Basin, and the Moho corresponds roughly to a change between strong crustal reflectivity and an almost transparent upper mantle. The dipping mantle reflector is not evident in the reflection profile, although there is a tendency for a change in reflection character. Dipping crustal, normal-incidence reflections cross some of the velocity interfaces, which indicates that the velocity distribution is a younger feature than the high-frequency reflectors. The change in velocity across the reflectors may be caused by metamorphic changes which post-date the tectonic creation of the normal-incidence reflectors. If so, these transformations have largely left the small wavelength, pre-existing reflectivity unchanged.

The crustal thickness is relatively constant along the EUROBRIDGE'97 profile: ∼ 46 km in the central part of the profile and deepening to the sides to ∼ 50 km. The Moho is at a constant depth and there is apparently no crustal thinning associated with the Pripyat Trough, as also noted by the EUROBRIDGE Seismic Working Group (1999) at the northern part of the trough. One may speculate that crustal thinning associated with the formation of

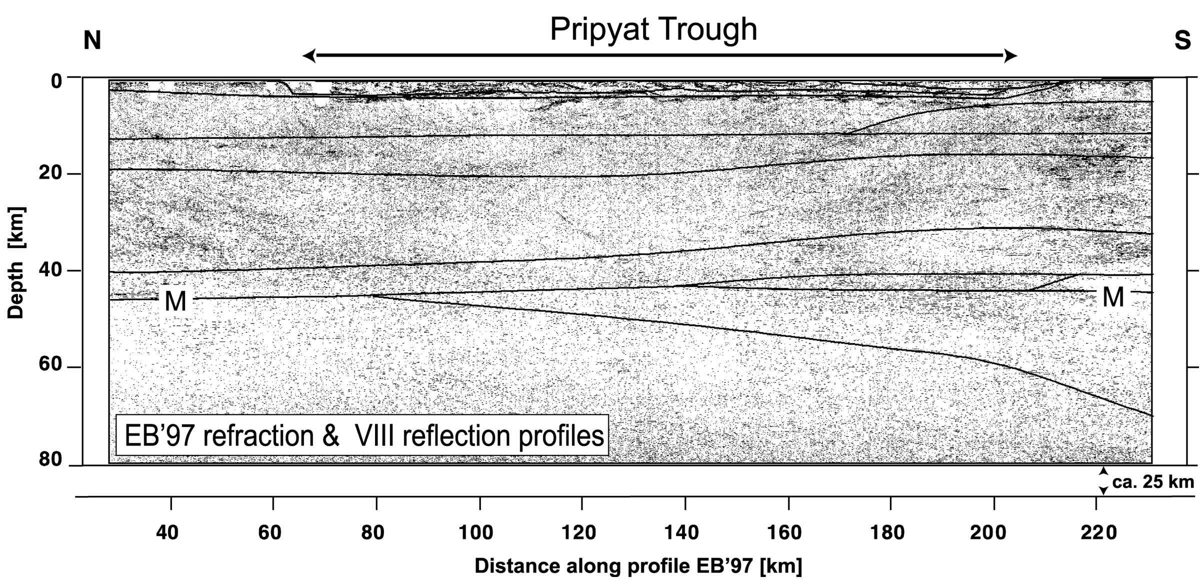

Fig. 10. Seismic reflection profile VIII with the velocity model along the EUROBRIDGE'97 profile superimposed. There is general agreement between the two models, regarding the shape and thickness of the Pripyat Trough and the Moho depth. Further, dipping intracrustal reflections tend to sole out at the top of the lowest crustal layer. The two profiles are parallel but are located ca. 25 km apart; hence, one should not expect complete correspondence.

the trough may have been hidden by magmatic ponding at the base of the crust (magmatic underplating), as indicated by the high seismic velocities in the centre part of the profile. Such masking of crustal thinning has recently been suggested under the currently active Kenya Rift Zone (Thybo et al., 2000). However, these velocities are spatially coincident with the Korosten Pluton, and are most likely associated with the emplacement of this magmatic body. Otherwise, the seismic velocity models show no evidence for magmatic activity in the crust along neither the EUROBRIDGE'97 profile nor the EUROBRIDGE'96 profile. Wilson and Lyashkevich (1996) interpret the presence of volcanic rocks from the Pripyat–Dnieper–Donets rift system as an indication of strong volcanic activity caused by a mantle plume. However, the relatively constant seismic velocities around the Pripyat Trough are normal to low for a platform area. Our velocity model shows no sign of magmatic intrusions in the crust below the Pripyat Trough, except for the apparent northward extension of the high-velocity anomaly from the Korosten Pluton to under the trough. It appears likely that the Devonian volcanic rocks originated directly from the mantle without affecting the crust along the EUROBRIDGE'97 and EUROBRIDGE'96 profiles.

The layer of P-wave velocity ∼ 6.4 km/s has been observed to the southeast from the Fennoscandia–Sarmatia suture along the EUROBRIDGE'96 profile to beneath the Pripyat Trough. The EUROBRIDGE'97 profile indicates that it only extends to the southern edge of the Pripyat Trough and the transition into the Volyn Block. The EUROBRIDGE Seismic Working Group (1999) suggested that this layer may have acted as a decollement for the extension that created the Pripyat Trough and that the extension only affected the upper crust. The flat Moho may be explained by significant tectonic involvement of the middle crust in the rift process because of a rheologically strong crust. A flat Moho has also been detected below Proterozoic basins in the Baltic Shield (BABEL Working Group, 1993). From comparisons to typical rock velocities (Christensen, 1996), the high-velocity gradient indicates that metamorphic or compositional changes from felsic to mafic granulites may take place in this interval.

8.3. Compositional constraints from the Vp/Vs ratio

Changes in P-wave velocity values may be used to discriminate between different tectonic/magmatic areas, although these values are not sufficient to constrain the rock types. The high-velocity gradient in the middle crust indicates a gradual change from felsic to mafic granulites. Further constraints may be derived from knowledge of the Vp/Vs ratio distribution (e.g. Musacchio et al., 1997). High Vp/Vs values usually indicate low quartz content of the rocks or the presence of pore fluids at high pressure (Christensen, 1996). The high-quality S-wave energy observed along most of the EUROBRIDGE'97 profile has allowed direct modelling of the S-wave velocity and Vp/Vs ratio distribution along the profile. The Vp/Vs ratio and P-wave velocity encountered in the northern and southern blocks support a model of the crust with an increasing content of basic rocks with depth, from felsic rocks at the surface to intermediate, mafic rocks in the lower crust. However, the very low Vp/Vs ratio in the lower crust of the Podolian Block of the Ukrainian Shield indicates unusually high quartz content in the lower crust.

The most outstanding aspect of the seismic velocity model is that the Vp/Vs ratio is distinctly higher in the upper and middle crust around the Korosten Pluton than in the surrounding blocks. These high Vp/Vs ratio values extend along the profile underneath the surface expression of the Korosten Pluton and further to the north, roughly coinciding with a high P-wave velocity anomaly in the same depth interval. The model shows Vp/Vs values of 1.77–1.79 coinciding with P-wave velocities larger than 6.70 km/s down to a depth of 10 km, and up to 1.84 associated with P-wave velocities of 6.60–6.70 km/s to 15 km depth. These values indicate a basic composition of the rocks. The highest Vp/Vs ratio is compatible with a mafic, gabbroic rock with a high content of plagioclase, pyroxene or amphibole (Christensen, 1996), consistent with the presence of a magmatic body composed of anorthosite and gabbro–norite with remnants of granite–gneiss. This may be explained by the presence of a plutonic body that formed from mantle-derived melts with additional melts from a gabbroic lower crust (Dovbush et al., 2000). The internal layering as seen from the shingling of refracted seismic phases may be caused by layering from several phases of melt injection into an original granitic–gneissic upper crust or, alternatively, fractional crystallization during cooling (Thybo et al., 2000). Our results show that the pluton extends deeper than previously believed, to depths of at least 15 km, and that there is internal layering as well as lateral differentiation. The highest content of anorthosites in the body appears to occur in the depth interval between 10 and 15 km at distances between km 170 and 260 along the profile. The presence of anorthosite and gabbro–norite to large depths is further supported by a pronounced, relative gravity low over the Korosten Pluton (Wybraniec et al., 1998), as high velocity and low density are characteristics of such rocks. The high-velocity lower crust also exhibits high Vp/Vs values in the northern and central part of the profile, except for the lowest thin layer of velocity >7.40 km/s, which is only interpreted in the central part of the profile. These values indicate a mafic lower crust with a lower content of plagioclase than in the upper body. The upper main layer of the lower crust may have a high content of pyroxenes, compatible with the high Vp/Vs ratio and relatively high P-wave velocity. The lower layer may be in a metamorphic grade corresponding to garnet-granulites, compatible with the very high P-wave velocity and relatively low Vp/Vs ratio. The internal stratification indicates that melting and underplating processes may have been active during the emplacement of the Korosten Pluton. However, the high velocities in the lowest crust may also have their origin in metamorphic processes that transformed original granulitic lower crustal rocks into felsic eclogites (Mengel and Kern, 1992; Abramovitz et al., 1998). Such processes create rock types with very high seismic velocity for felsic rocks. They may have been active during a period of nearby subduction that may have provided water to catalyse the metamorphic processes and which may have cooled the lowest crust.

There is pronounced lower crustal layering in the central part of the seismic velocity model beneath the Korosten Pluton and further to the north. This layering may be explained by intrusion of mantle-derived melts into the lower crust and around the original Moho, where several series of intrusion and substantial fractionation of the melts during cooling may have occurred. The layering may have formed during the emplacement of the Korosten Pluton or during the formation of the Pripyat Trough. Outside this zone, there appears to be substantial seismic reflectivity or scattering, which originate from finer heterogeneity than in the central part of the profile. The origin of this reflectivity is not understood; it may be speculated that it may be related to edge effects at the presumed intrusions into the lower crust.

Another remarkable feature of the Vp/Vs ratio distribution is the very high values of 1.85–1.84 in the sub-Moho mantle, wherever constrained, coinciding with a remarkably high P-wave velocity of >8.35 km/s. The sub-Moho Pn velocity is high along the whole profile, except in the distance interval km 80–260, where the velocity is around 8.10 km/s, which is a Pn velocity value often encountered in the southern Baltic Shield along the BABEL profile and East European Platform along the POLONAISE profiles (BABEL Working Group, 1993; Grad et al., 2002). The extremely high Pn velocities are encountered along the whole EUROBRIDGE transect from the Baltic Sea into the Ukrainian Shield. The pronounced difference in velocity between the EUROBRIDGE profiles and other profiles in the region may be because the profiles sample different geographical areas. However, it is remarkable that the Pn velocities are mainly sampled in a SW–NE direction along the BABEL and POLONAISE profiles, whereas the EUROBRIDGE profiles have a NW–SE direction. Hence, substantial anisotropy cannot be excluded, although further profiling is needed before any firm judgement can be made.

8.4. Upper mantle features

A strongly southward dipping reflector has been interpreted in the uppermost mantle, extending from the Moho at km 80 to a depth of 75 km at km 250. The reflector coincides with a subhorizontal reflector in the EUROBRIDGE'96 profile at the cross point with the EUROBRIDGE'97 profile. This indicates that the true dip direction is approximately SW to SSW, which is approximately perpendicular to the dominating strike directions of the individual terranes in the Palaeoproterozoic accretionary belt to the northwest. The seismic data along the EUROBRIDGE'96 and EUROBRIDGE'97 profiles can only constrain the position and approximate dip-direction of the reflector, but the velocity below the reflector is unknown. From the structure along the EUROBRIDGE'97 profile, it is tempting to relate the dipping reflector to Palaeoproterozoic accretion of a series of terranes, including the continental margin of the Osnitsk–Mikashevichi Igneous Belt, which were accreted to the Archaean nucleus of Sarmatia. However, the SW–NE strike direction of the terrane boundaries is not favourable for such interpretation, although the strike direction is NW–SE for the main suture between the Sarmatian and Volgo–Uralian plates (Bogdanova et al., 2001). The boundary between the Archaean Podolian block and the Paleoproterozoic Teterev–Belaya Tserkov Belt and its continuation to the north is NNW–SSE striking in the immediate vicinity to the profile. The near-surface location of this suture is approximately 500 km to the northeast of the dipping mantle reflector. We suggest that the dipping reflector is the trace of the suture or, alternatively, a collision-related shear zone in the upper mantle. The relatively low mantle velocities to the south of (above) the dipping mantle reflector lead us to favour a compositional change across the reflector, which may represent the structural outline of the suture in the uppermost mantle. The basis and conclusion for this interpretation is similar to the inference of a crustal and upper mantle suture further north in the Baltic Shield (Abramovitz et al., 1997).

The uppermost mantle generally shows substantial occurrences of subhorizontal reflectors in the sub-Moho mantle of the Baltic Shield and East European Platform (e.g. Grad, 1992; BABEL Working Group, 1993; EUROBRIDGE Seismic Working Group, 1999; Grad et al., 2002). These reflectors are often found around 10 km below the Moho. Remarkably, the EUROBRIDGE'97 profile only shows one such reflector in the S-wave model and none in the P-wave model. Because the reflectors are generally identified along single profiles, one cannot exclude that some of these are dipping in a direction away from the profiles, c.f. the above discussion, but the abundant occurrence indicates that some of the reflectors are actually subhorizontal. Some of the possible causes that are being discussed include magmatic intrusions, shear zones from tectonic movement, seismic anisotropy, variation in fluid content, and metamorphic transitions, e.g. between a felsic lower crust in eclogite facies (REF) and a basic upper mantle (e.g. Ryberg et al., 1995; Thybo and Perchuc̀, 1997; Abramovitz et al., 1997, 1998; Steer et al., 1988). The mantle reflectors are also observed as subhorizontal features beneath substantial Moho undulations. This indicates that they are younger features than the undulations.

9. Conclusions

Seismic data of high quality for interpretation of both the P- and S-wave velocity structure was acquired during the EUROBRIDGE'97 project along a 530-km-long, N–S-striking profile in the East European Platform. The profile extends from the Pripyat Trough in the Osnitsk–Mikashevichi Igneous Belt, across the Korosten Pluton and the Volyn Block into the Podolian Block of the Ukrainian Shield. Seismic energy from 18 shot points at approximately 30 km distance along the profile were recorded by 120 mobile three-component seismographs in two deployments. The data has been interpreted by seismic tomographic inversion of the travel times of first arrival P-waves, by two-dimensional ray-tracing of travel times of primary and secondary P- and S-waves, as well as by calculation of two-dimensional synthetic seismograms to match the P-wave arrivals. The models of seismic P-wave velocity and Vp/Vs ratio are indicative of a number of features of the crust and uppermost mantle along the profile:

1. There are pronounced differences in velocity structure between the three main crustal blocks traversed by the profile. These blocks are related to Palaeoproterozoic amalgamation of the crust by terrane accretion to an Archaean nucleus. Hence, tectonic structure and compositional difference has survived for approximately 1.8 Ga.
2. Contrasts in seismic velocity cross dipping reflectors as identified in seismic normal-incidence reflection sections. This indicates that the seismic velocity structure has changed since the formation of the crust, although pronounced contrasts in seismic velocity has also survived at the terrane boundaries.
3. The Korosten Pluton is imaged as a high-velocity anomaly to depths of at least 15 km instead of 6 km as previously interpreted, possibly connected to a lower crustal, high-velocity anomaly. The Vp/Vs

ratio is high, indicative of a basic composition, consistent with a mafic body of mantle-derived melts with contribution from the lower gabbroic crust.
4. The Ukrainian Shield shows a typical cratonic velocity structure, however, with relatively low seismic P-wave velocity and very small Vp/Vs ratios.
5. The northern part of the Osnitsk–Mikashevichi Igneous Belt is characterised by a typical velocity structure of the Baltic–Belarussian part of the East European Platform area, i.e. with a typical platform structure, which remarkably, is underlain by a thin layer of very high velocity.
6. The Pripyat Trough is outlined as a rift type sedimentary basin. The Moho is basically flat across the trough, which indicates that the extension only affected the upper to middle crust, e.g. by stretching of a rheologically strong lithosphere, or, alternatively, that processes of magmatic underplating have masked crustal thinning.
7. A steeply southwesterly dipping mantle reflector is present in the profile. This reflector correlates with a subhorizontal reflector in the NW–SE-striking EUROBRIDGE'96 profile. We suggest that this reflector is the mantle expression of the suture between Sarmatia and Volga–Uralia or a mantle shear zone related to later tectonic processes.
8. The sub-Moho mantle has very high seismic velocity along most of the profile, with the exception of a 200-km-wide belt above the dipping mantle reflector.

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