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P- and S-wave velocity model of the southwestern margin of the Precambrian East European Craton; POLONAISE'97, profile P3

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Abstract

The large-scale seismic experiment POLONAISE'97 investigated the seismic structure of the crust and the uppermost mantle in the Polish region of the Trans-European Suture Zone. This paper covers the interpretation of seismic data along the 300-km-long P3 profile, which is located in the Precambrian East European Craton (EEC) parallel to the Teisseyre–Tornquist Zone. The recordings were of high quality, with seismic energy detectable out to 300 km offsets. The crustal model developed by two-dimensional raytracing forward modelling and waveform analysis is characterized by a subhorizontal uniform structure, although the boundaries of several large basement features are crossed. The crystalline crust consists of three parts: the upper, middle and lower crust with P-wave velocities of 6.1–6.4 km/s, 6.55–6.7 km/s and 7.05–7.15 km/s respectively. The P_mP wave can usually be correlated at distances beginning at about 100 km, but it has variable reflection character. Moho depth increases from 38 km at the NW end of the profile to 44 km at the SE end. The P_n velocity of 8.05–8.1 km/s is less than that previously found in neighbouring areas of the EEC. The V_P/V_S ratio was determined separately for the upper/middle and lower crust to be 1.67 and 1.77 respectively. Following the P_n wave, another phase with an apparent velocity of about 8.3 km/s is interpreted as a weak reflection from a low-contrast discontinuity in the uppermost mantle.

Keywords: Precambrian East European Craton; POLONAISE'97; velocity model

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1. Introduction

The POLONAISE'97 (POlish Lithospheric ONsets — An International Seismic Experiment) large-scale seismic experiment was carried out in May 1997 in Poland, Germany and Lithuania. Its main purpose was to investigate the structure of the crust and the uppermost mantle in the region of the Trans-European Suture Zone (TESZ), the Palaeozoic Platform and the Precambrian East European Craton (EEC) (Fig. 1). A total of 64 shots of charge between 100 and 1000 kg were detonated along five profiles of a total length of about 2000 km. The seismic energy was recorded by 613 seismic stations, located at 792 receiver locations in two deployments.

This paper covers the interpretation of seismic data from the P3 profile (Fig. 1), which was located on the EEC parallel to its SW margin and the Tornquist–Teisseyre Zone (TTZ). This location was chosen in order to provide a reference model of the crust for the SW part of the EEC. The length of the profile, with 11 shotpoints and 100 receivers (mainly one-component PRS stations), was 300 km. This density of shotpoints and receivers provides a more detailed model of crustal and upper mantle structures than those from previous studies in the region. The seismic stations also recorded off-line shots from other profiles. In this paper we describe two-dimensional (2-D) modelling of the in-line recordings.

2. Geological and geophysical characteristics of the study area

Three main geotectonic provinces are located on the Polish territory: the Precambrian EEC in the NE part, the Palaeozoic Platform in the SW part and the Carpathians in the south. The SW boundary of the EEC is the Polish Trough (Guterch et al., 1996). Profile P3 is located near the margin of the EEC, parallel to the axis of the

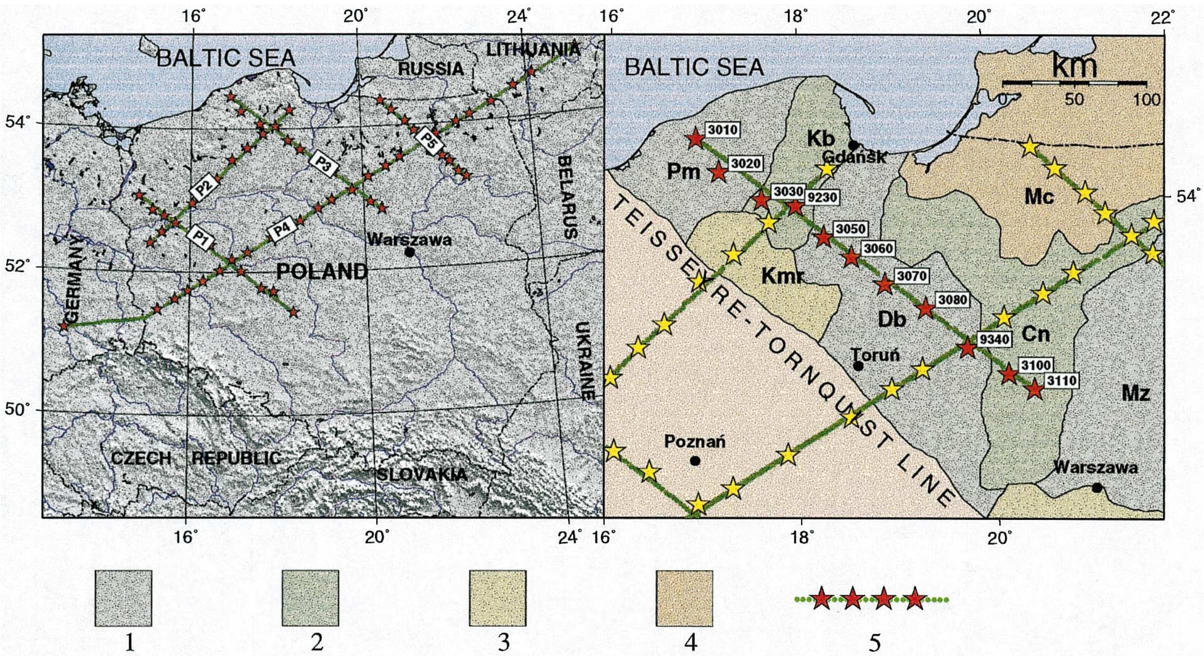

Fig. 1. Location of the P3 profile and main tectonic units of the crystalline basement of the Precambrian Platform in Poland. (1) Pre-Karelian granitoid massifs: Pm, Pomorze; Db, Dobrzyń; Mz, Mazowsze. (2) Pre-Karelian metamorphic belts: Kb, Kaszuby; Cn, Ciechanów. (3) Karelian metamorphic–magmatic complexes. (4) Gothian metamorphic–magmatic complex (Mc, Mazury). (5) Location of shotpoints and receivers along the profile.

Polish Trough and to the TTZ, which is a tectonic inversion zone.

The geological structure and physical parameters of the sedimentary cover of the East European Craton in Poland have been extensively studied (e.g. Znosko, 1968; Skorupa, 1974; Grad et al., 1991b; Młynarski, 1984). The thickness of the sedimentary cover in the Polish area of the EEC is small in northeasternmost Poland, where the Cenozoic–Mesozoic sediments directly overlie the crystalline Precambrian basement at a depth of 200–500 m. The maximum thickness of sediments (8 km) has been detected in the marginal zone of the platform, where two sedimentary complexes can be distinguished: the lower unit of Cambrian to Silurian age, and the upper unit of Permian to Cenozoic age. Along the P3 profile the sediment thickness varies from 4 to 6 km.

The dominant units of the Precambrian crystalline basement in Poland are three large granitoid structures of Pre-Karelian age (>2.6 Ga): the Dobrzyń, Pomorze and Mazowsze massifs (Fig. 1). They were metamorphosed in several phases, mainly during the Gothian (<1.7 Ga), are characterized by subdued magnetic anomalies (Kubicki and Ryka, 1982; Ryka, 1984; Królikowski and Wybraniec, 1996), and are supposed to be the oldest units of the crust. The main rocks of the massifs are the reddish orthoclase granites that contain small amounts of oligoclase, muscovite and biotite (Ryka, 1984). Pre-Karelian metamorphic–magmatic belts (Kaszuby, Podlasie and Ciechanów zones) separate the massifs from each other. Their composition differs from the composition of the granitoid massifs. For example, the crystalline rocks of the Podlasie zone can be divided into a lower (granulite) group, consisting of two-pyroxene granulites, enderbites, pyroxene gneisses and pyroxene amphibolites, and an upper (gneiss) group, consisting of biotite gneisses and sillimanite–andalusite gneisses (Znosko, 1968; Kubicki and Ryka, 1982; Ryka, 1984). The shape and location of the metamorphic–magmatic belts have been inferred from correlative strong positive magnetic anomalies (Królikowski and Wybraniec, 1996). Other, younger units of the EEC in Poland are the Karelian metamorphic–magmatic complexes, adjoining the TTZ, and the Gothian metamorphic–magmatic zone in NE Poland (Ryka, 1984).

In the area of the Precambrian Platform and TTZ, a variety of seismic investigations have been executed. Shallow refraction and reflection measurements in a dense network of profiles provide a detailed description of the sedimentary sequences and maps of the depth of the crystalline basement in the platform area. These studies show that the P-wave velocities for the upper sedimentary complex vary from less than 2 km/s in the Cenozoic layers to 2.5–3.5 km/s in the Mesozoic and about 4.8 km/s in Permian layer. For the second complex, of Silurian to Cambrian age, the P-wave velocities are in the range 4.5–5.6 km/s (Grad, 1987; Grad and Ryka, 1996). Measured refraction P-wave velocities in the crystalline basement in the EEC area are in the range 5.8–6.3 km/s (Młynarski, 1984; Grad et al., 1991a,b).

Several deep seismic profiles cross the TTZ (LT-4, LT-5 and LT-7) and extend into the EEC, providing information on seismic parameters of the crust and the depth of the Moho boundary (Guterch et al., 1986, 1994). They show that the crystalline complex of the crust in the EEC area consists of three layers characterized by P-wave velocities of 6.1–6.3 km/s, 6.7–6.9 km/s and 7.2–7.3 km/s, reaching depths of 22–25 km, 33–36 km and 42–47 km respectively. The thickness of the crust is 42–47 km, and the V_P velocity in the upper mantle is 8.1–8.2 km/s (Guterch et al., 1986).

The P3 profile crosses the following basement units: Pomorze granitoid massif at the NW end, Kaszuby metamorphic belt, Dobrzyń massif and Ciechanów metamorphic belt at the SE end (Fig. 1). The profile is located along the margin of a prominent gravity low, reaching −40 mGal in its centre (Guterch et al., 1999). Grabowska et al. (1998) attribute the gravity low to the combined effects of thickening of the sedimentary cover at the edge of the platform, crustal thickening, and density variations in the mantle.

3. Data acquisition and processing

The seismic data were recorded by 85 single-channel PRS stations and 15 three-component

RefTek stations with a 10-ms sampling interval. Data processing included introduction of time corrections where needed, removal of high-amplitude spikes, notch filtering of traces with high-amplitude monochromatic (industrial?) noise and filtering of the whole dataset with a 2–15 Hz band-pass filter in order to remove low- and high-frequency noise. The seismic energy is concentrated between 2 and 12 Hz frequency, with the maximum at 6–8 Hz (4–6 Hz for P_mP arrivals).

Recordings were sorted into shot gathers and plotted in seismic sections. Seismograms were normalized to maximum amplitude in the plotted time window and plotted with a reduction velocity of 8 km/s.

4. The seismic wave field

The seismic record sections are presented in Figs. 2–7. The P-wave field is of high quality, with seismic energy visible up to 300 km offsets, although the traces at offsets of more than 250 km are noisy and first arrivals are questionable or difficult to correlate in some sections.

At small (up to 10 km) offsets, first arrivals with low apparent velocity below 5 km/s represent waves refracted in the upper sedimentary layers. In each section, there were only a few such traces because of the seismometer spacing of about 3 km; thus, thin sedimentary layers with highly variable velocity could not be detected. These arrivals were

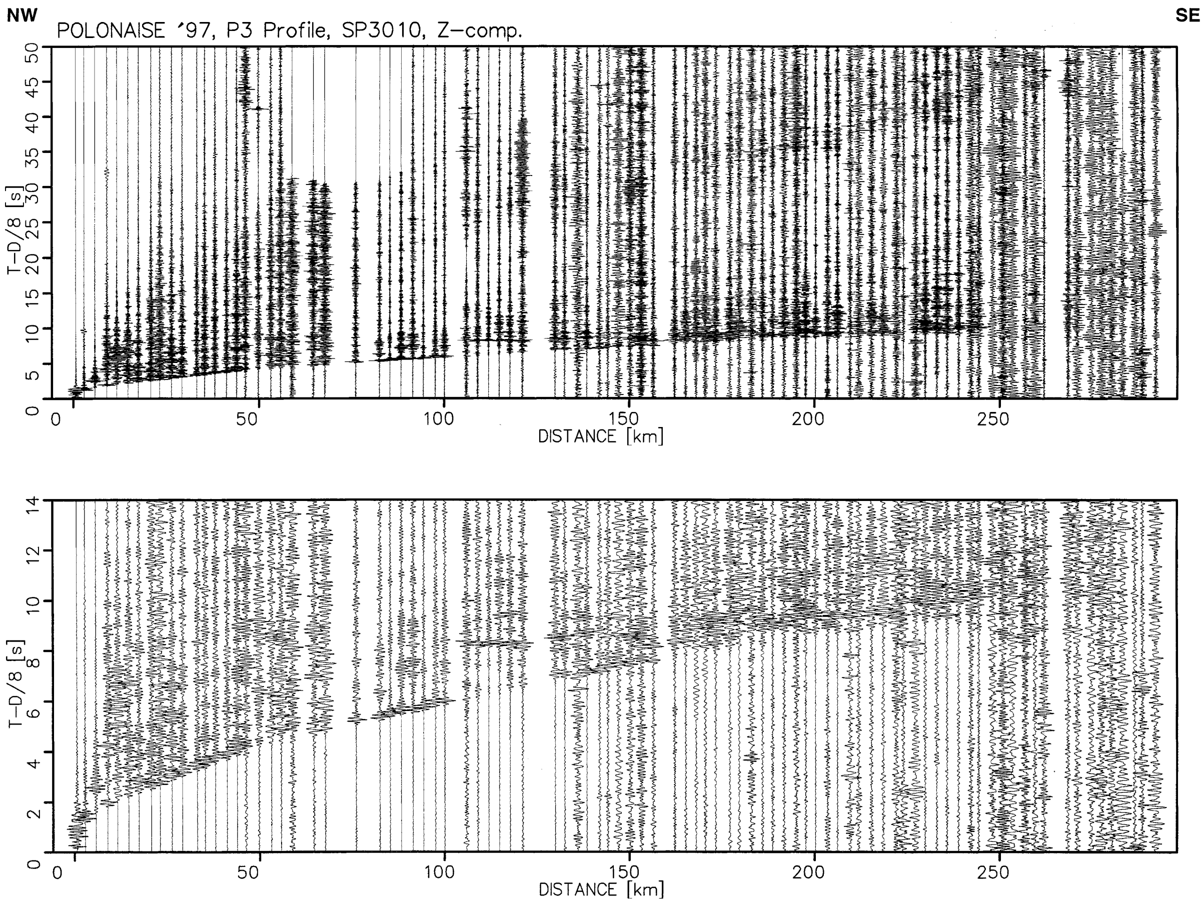

Fig. 2. Seismic record sections for SP 3010 (top: time window to 50 s, showing P and S arrivals; bottom: only P arrivals). Reduction velocity is 8 km/s.

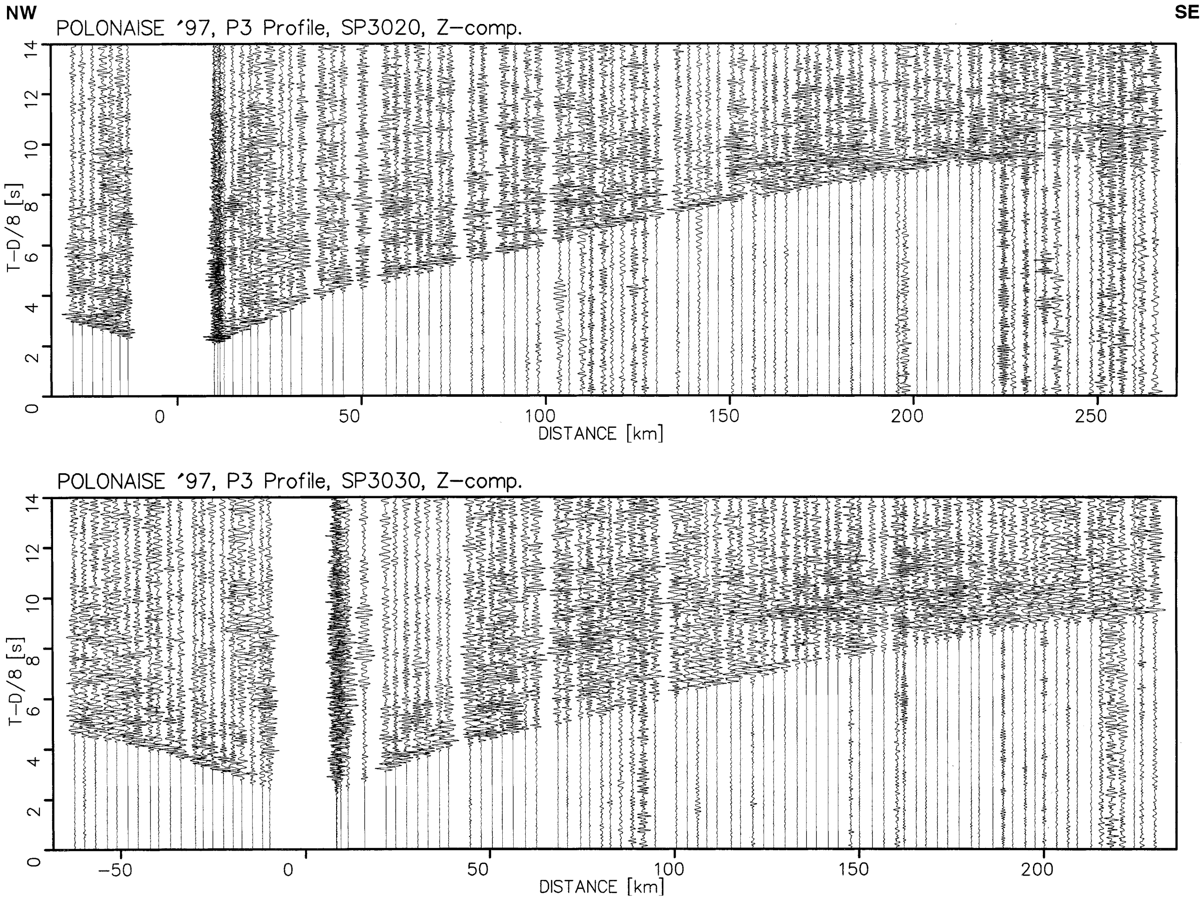

Fig. 3. Seismic record sections for SP 3020 and SP 3030. Reduction velocity is 8 km/s.

used for overall control of velocity distribution in sediments, which was determined from other sources of information such as velocity profiles in boreholes and earlier high-resolution seismic surveys.

At offsets from 10 to 40 km, the first arrivals have an apparent velocity of 5.6 km/s. They represent a refraction from a horizon in consolidated (Silurian–Ordovician) sediments. These waves are correlated to the largest offsets in record sections from SP 3010, 3020, etc. at the NW end of the profile. They are not identified in sections from the opposite end of the profile, showing that the horizon is present only below the NW part of the profile.

The crustal refracted waves, identified to 250 km offset, were clear in the whole distance interval. They can be generally divided in two segments: the first one with apparent velocity from 6.1 to 6.3 km/s representing the Pg wave refracted from the top of the crystalline basement (upper crust), and the second one with an apparent velocity of 6.5–6.6 km/s, representing refracted waves from the middle crust. The changes of the apparent velocities of individual crustal branches are gradual, which shows that velocity contrasts at any first-order discontinuity in the upper/middle crust are small. On one section, from SP 9340, a phase with apparent velocity >7.0 km/s is a first arrival in a narrow offset interval (175–200 km). Despite being short, it can indicate a lower crustal layer with P-wave velocity higher than 7.0 km/s.

At offsets larger than 200 km, first arrivals with apparent velocities higher than 8.0 km/s are

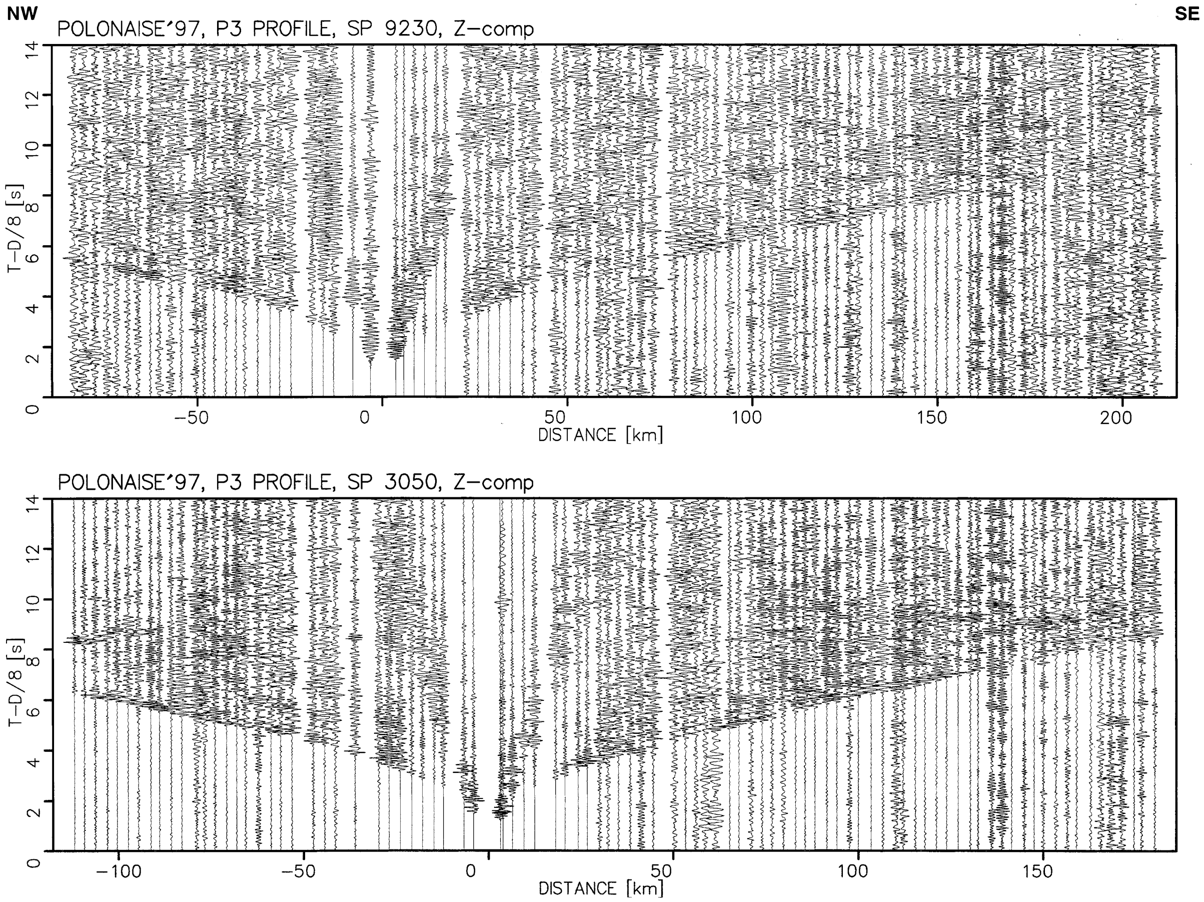

Fig. 4. Seismic record sections for SP 9230 and SP 3050. Reduction velocity is 8 km/s.

observed on most sections. The best examples are found on sections from SP 3100 and SP 3110. Even on these sections, their amplitude is low and rapidly decreases with distance compared with later arrivals. We suggest that part of these arrivals are P_n waves with an apparent velocity of 8.0–8.1 km/s. Secondary arrivals, with an apparent velocity of about 8.3 km/s, are interpreted as weak reflections from a low-contrast discontinuity in the upper mantle (Fig. 8). This later phase appears as a first arrival because of the pronounced amplitude decay of the P_n phase.

Other secondary arrivals include short (10–15 km) fragments of waves reflected from discontinuities in the upper crust, close to the first arrivals. A later, weak reflected wave is correlated on four sections and was modelled as a reflection from the top of the lower crust.

The P_mP wave is usually observed at distances greater than 100 km. Its character is highly variable, on some sections (e.g. SP 3010 and SP 3110), these arrivals are strong and coherent, whereas on others they are weak, scattered and only fragmentarily observable (e.g. SP 9340 and SP 3100) (Fig. 9).

The shear wave field has a lower *S*/*N* ratio than the P-wave field. The S_g wave is often not observable, even near the shotpoint. The S_g arrivals of best quality are correlated on the section for SP 3110. The S_mS wave is identified on several seismic sections as an envelope of increased amplitude, rather than clear arrivals that can be phase corre-

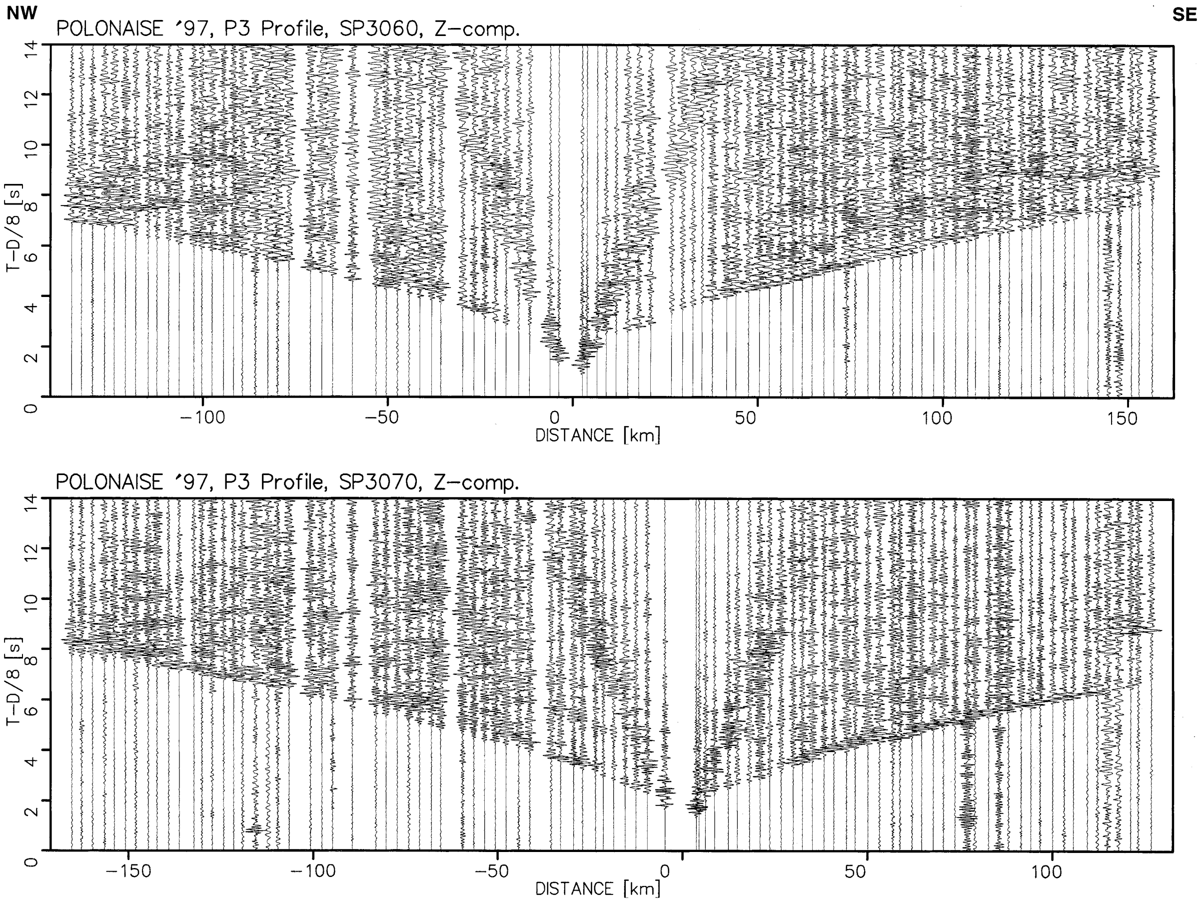

Fig. 5. Seismic record sections for SP 3060 and SP 3070. Reduction velocity is 8 km/s.

lated. The S_mS travel-time curve has been picked along the onset of the increased amplitudes. The time uncertainty is estimated to be ± 0.3 s.

5. Data interpretation

The first step of the interpretation was identification and picking of phases corresponding to first and later arrivals. For picking of the first arrival times we used non-filtered sections, whenever possible, to avoid false first arrivals produced by non-causal filtering. Picked times from different subgroups of our team were compared, and the most doubtful picks were removed from the final data set. Estimated average uncertainty of picked times for Pg is 0.05–0.1 s and for P_n and P_mP is 0.15–0.20 s. The picked travel times were the input for the 2-D seismic modelling with the raytracing package SEIS83 (Červený and Pšenčík, 1983) and with a graphical interface by Komminaho (1997).

The starting model for the sedimentary layers was compiled from a previous interpretation of the shallow refraction and reflection profiling and borehole data in the vicinity of the profile. The velocity information for the sediments was slightly modified in our model in order to obtain general agreement for calculated and observed travel times of sedimentary and Pg phases. We attempted to obtain acceptable travel-time differences with minimum change of the a priori velocity distribution. The major change was an introduction of a low-velocity layer in the Silurian–Ordovician complex under the 5.6 km/s layer, in order to compensate

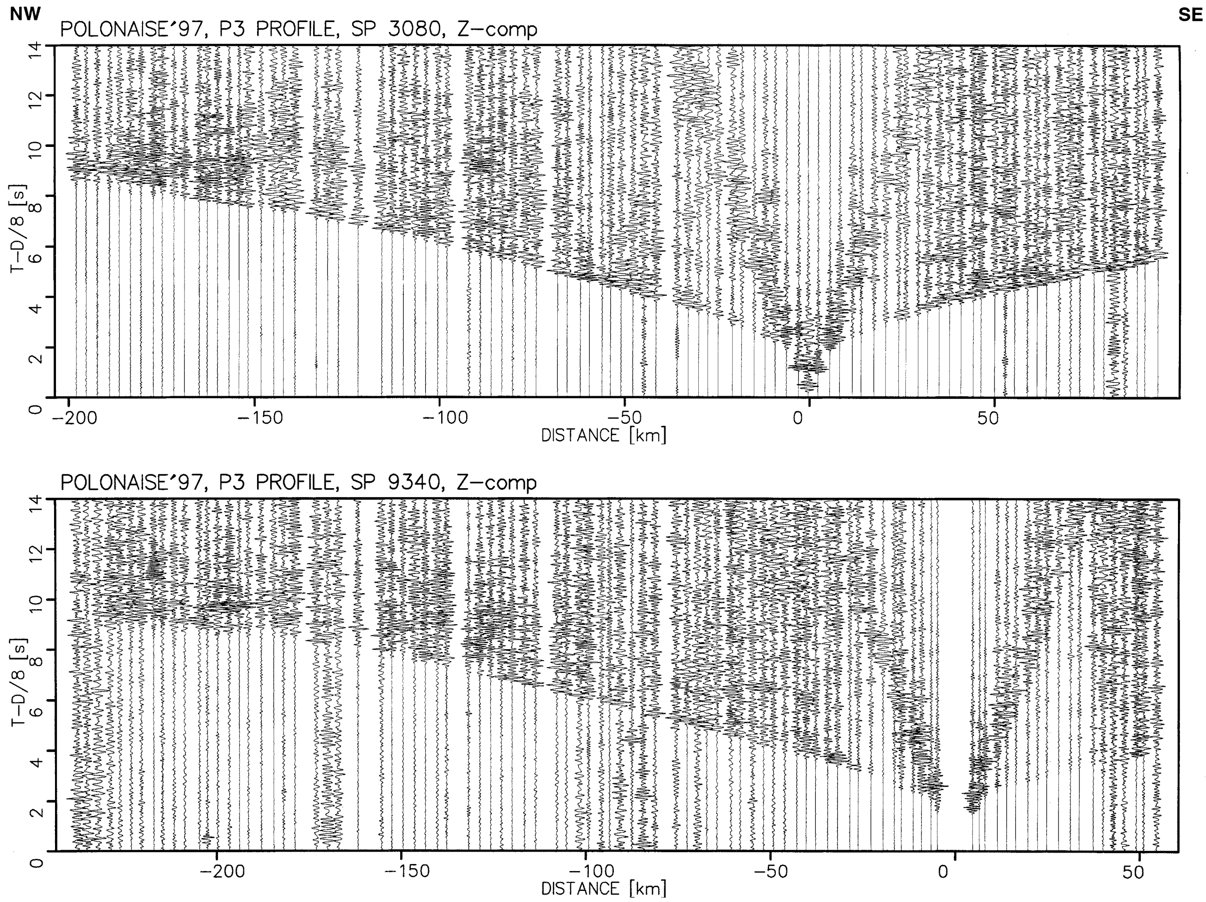

Fig. 6. Seismic record sections for SP 3080 and SP 9340. Reduction velocity is 8 km/s.

for a misfit in arrival times of seismic phases from the crystalline basement. We found this change acceptable, as the deeper sedimentary complex is not constrained as well as the shallow layers.

With the model of the sedimentary layers held fixed, the modelling of the remainder of the crust was carried out by trial-and-error iteration. It involved changing of the structure in order to obtain the agreement between calculated and experimental travel times (Figs. 10–12). In the late phases of modelling, synthetic record sections were calculated and qualitatively compared with the data, taking into account the relative amplitude of phases. This waveform modelling provides additional constraints on velocity contrasts and gradients (Figs. 10–12).

Using the derived P-wave velocity distribution and interfaces, the V_P/V_S ratio value was also estimated by fitting the calculated and observed S-wave travel times (Fig. 13). Because of the uncertainties of the S-waves picks, we decided to model only the overall V_P/V_S ratio in the upper/middle and lower crust.

6. Velocity models

The velocity model that produces a theoretical wavefield consistent with the experimental one within a satisfactory accuracy is shown in Fig. 14. The basic features of the model are as follows:

(1) The sedimentary cover consists of seven layers with a total thickness of 4–6 km. The P-wave velocities range from 2 to 5.7 km/s.

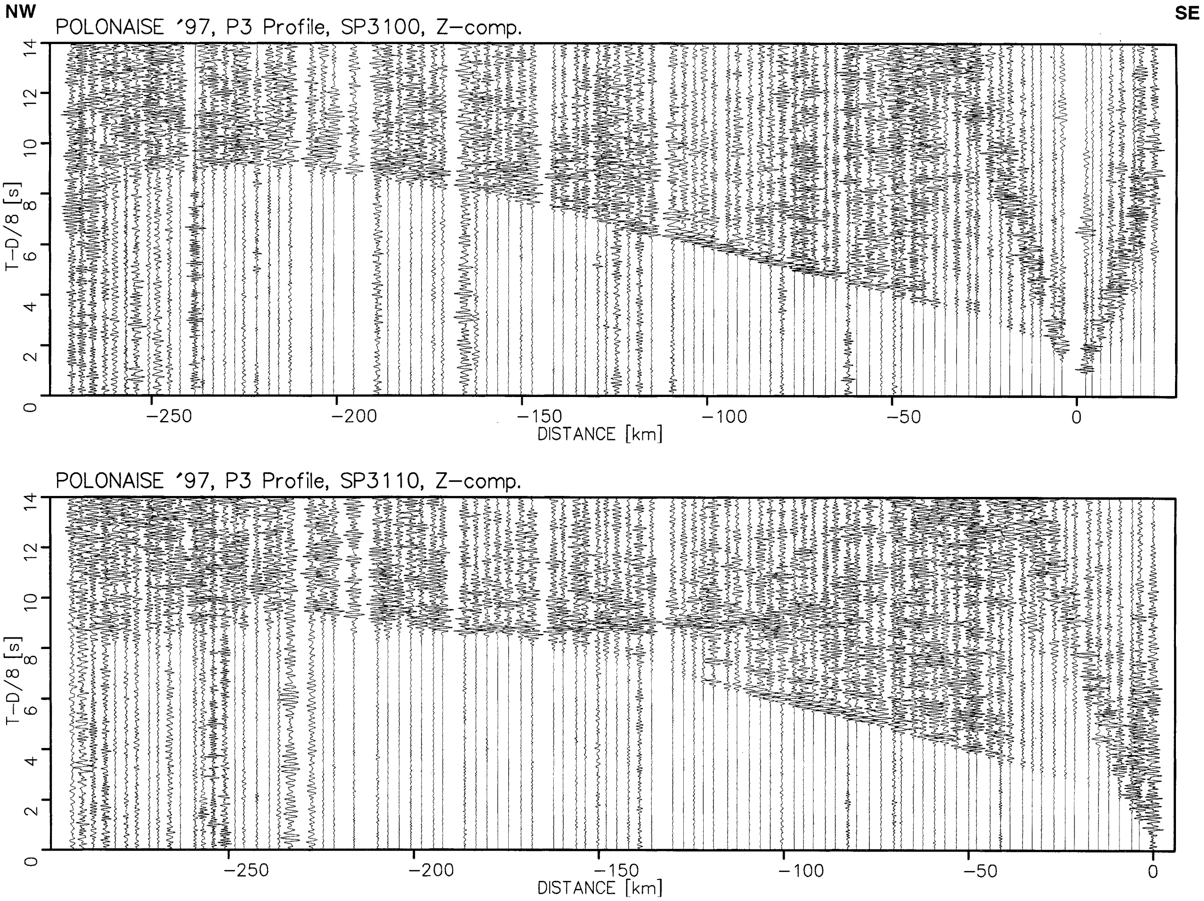

Fig. 7. Seismic record sections for SP 3100 and SP 3110. Reduction velocity is 8 km/s.

(2) The upper crust consists of two layers: the first one is described by a P-wave velocity of 6.1–6.15 km/s extending from depths of 4–6 km to 8–9 km and constituting the top of the Precambrian basement; the second one, with a P-wave velocity of 6.3–6.4 km/s, reaches depths of 15–20 km. The physical parameters describing the layers were constrained using refracted waves (first arrivals), as no reflections from these layers were observed. Because of the low velocity contrast at the interface between these two layers, it could be argued that the data require one layer with V_P changing from 6.1 to 6.4 km/s, but our tests showed that the calculated travel-time curves do not agree with data as well as in the case of the two-layer model. We did not find any evidence for significant horizontal velocity variations or contrasts in either the shape of the travel-time curves or the amplitude characteristics of the seismograms. This shows that the geological block structure of the basement is not reflected in detectable changes of physical properties in the upper crust.

(3) A middle crustal layer with a P-wave velocity of 6.55 to 6.7 km/s extends to depths from about 20 to 30 km. This layer was also constrained by the information from first arrivals, representing the refraction at its top. This layer is characterized by horizontal velocity variations with a high-velocity (+0.1 km/s) area at about 100 km distance from the NW end of the profile. This high-velocity area is needed to explain the high apparent velocity of refractions from shotpoint 3070, but the high velocity is only constrained in the upper few kilometres of the layer.

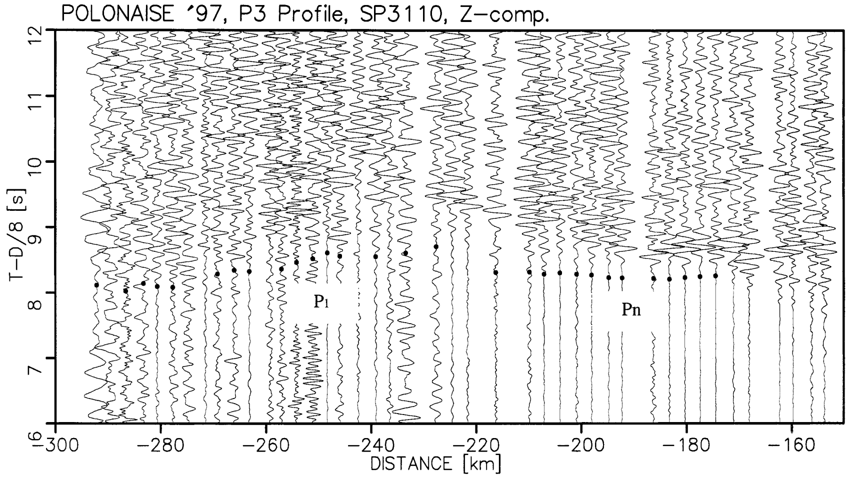

Fig. 8. Part of seismic record section for SP 3110 showing the interpreted P_n first arrivals and a reflection from mantle discontinuity (P_1).

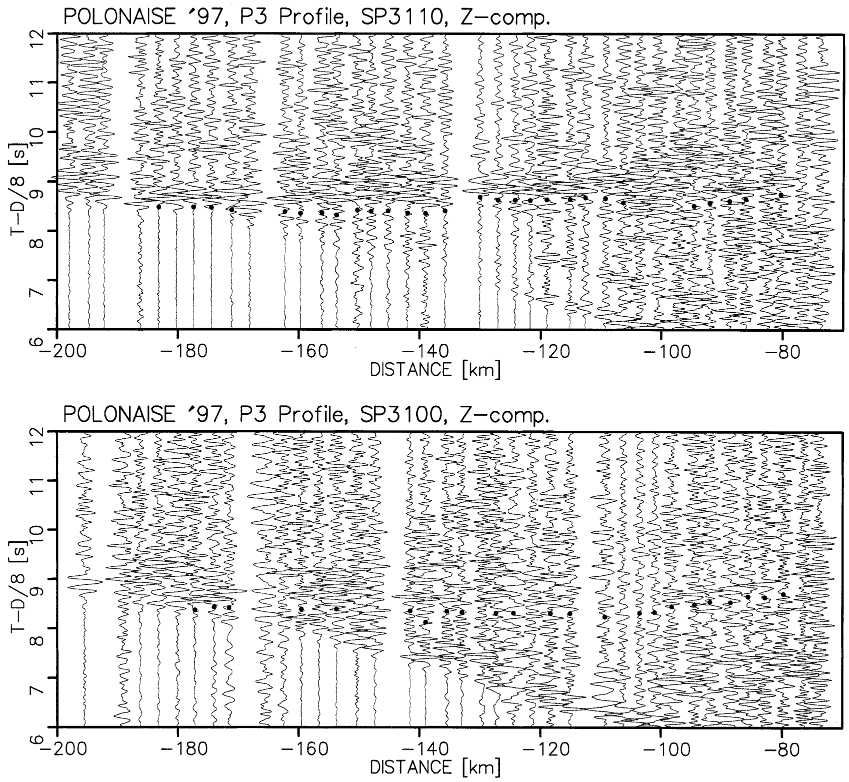

Fig. 9. Parts of seismic record sections for SP 3110 (top) and 3100 (bottom), showing different character of reflections from the Moho boundary. Both sections show the same offset interval.

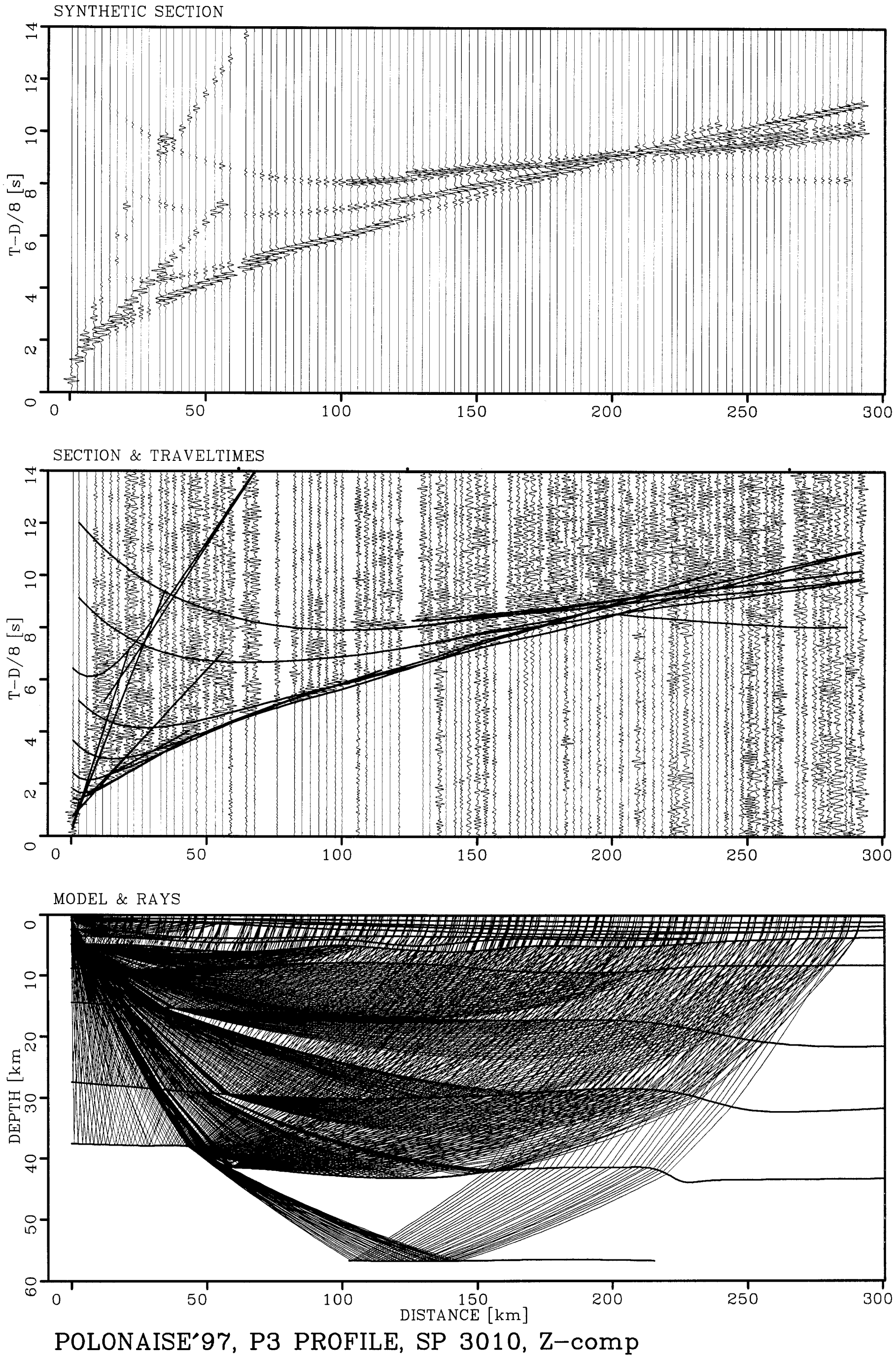

Fig. 10. A 2-D seismic modelling example for SP 3010. Bottom: model and ray diagrams; middle: seismic record section and calculated travel-time curves; top: synthetic record section.

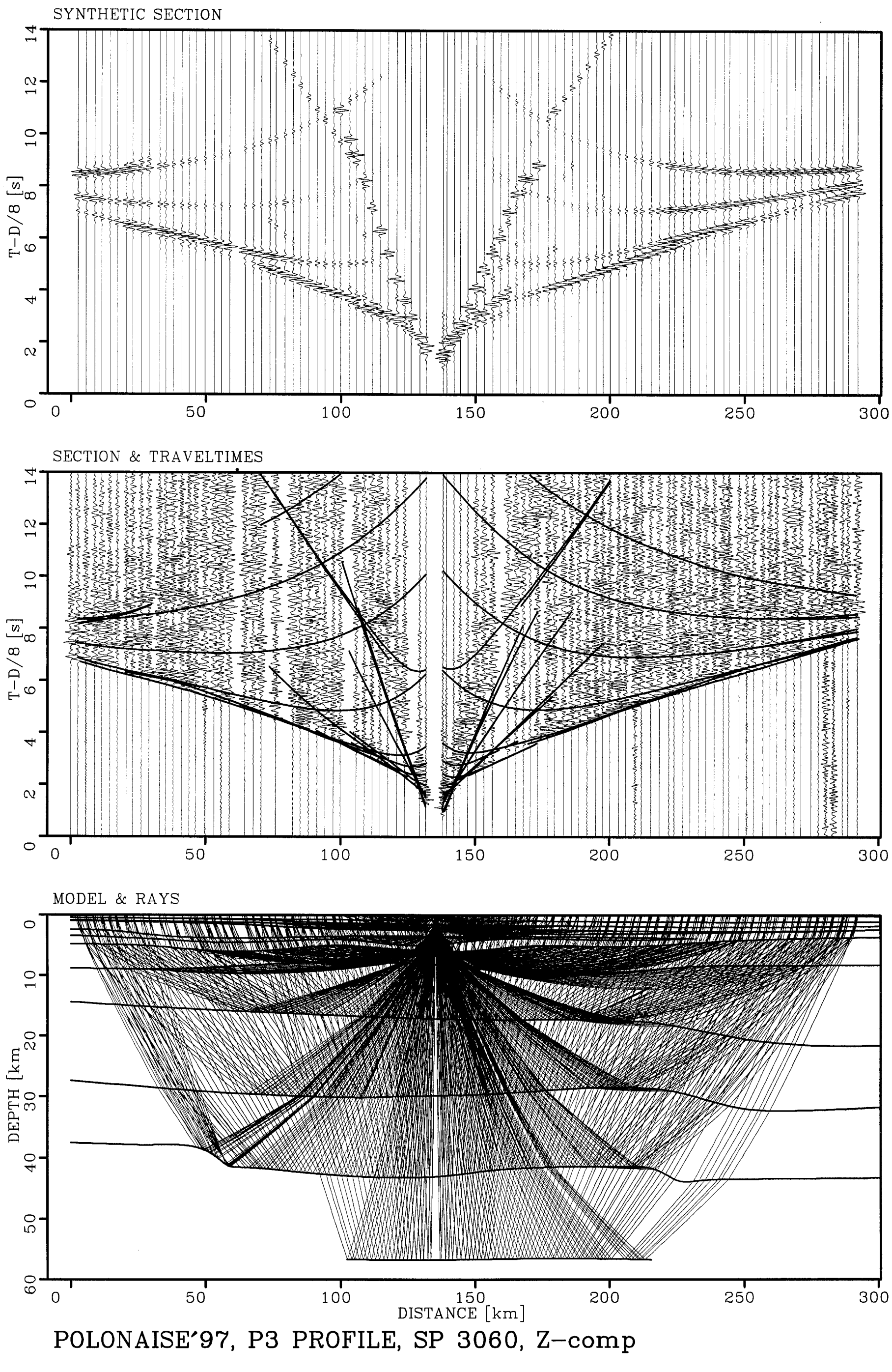

Fig. 11. A 2-D seismic modelling example for SP 3060. Bottom: model and ray diagrams; middle: seismic record section and calculated travel-time curves; top: synthetic record section.

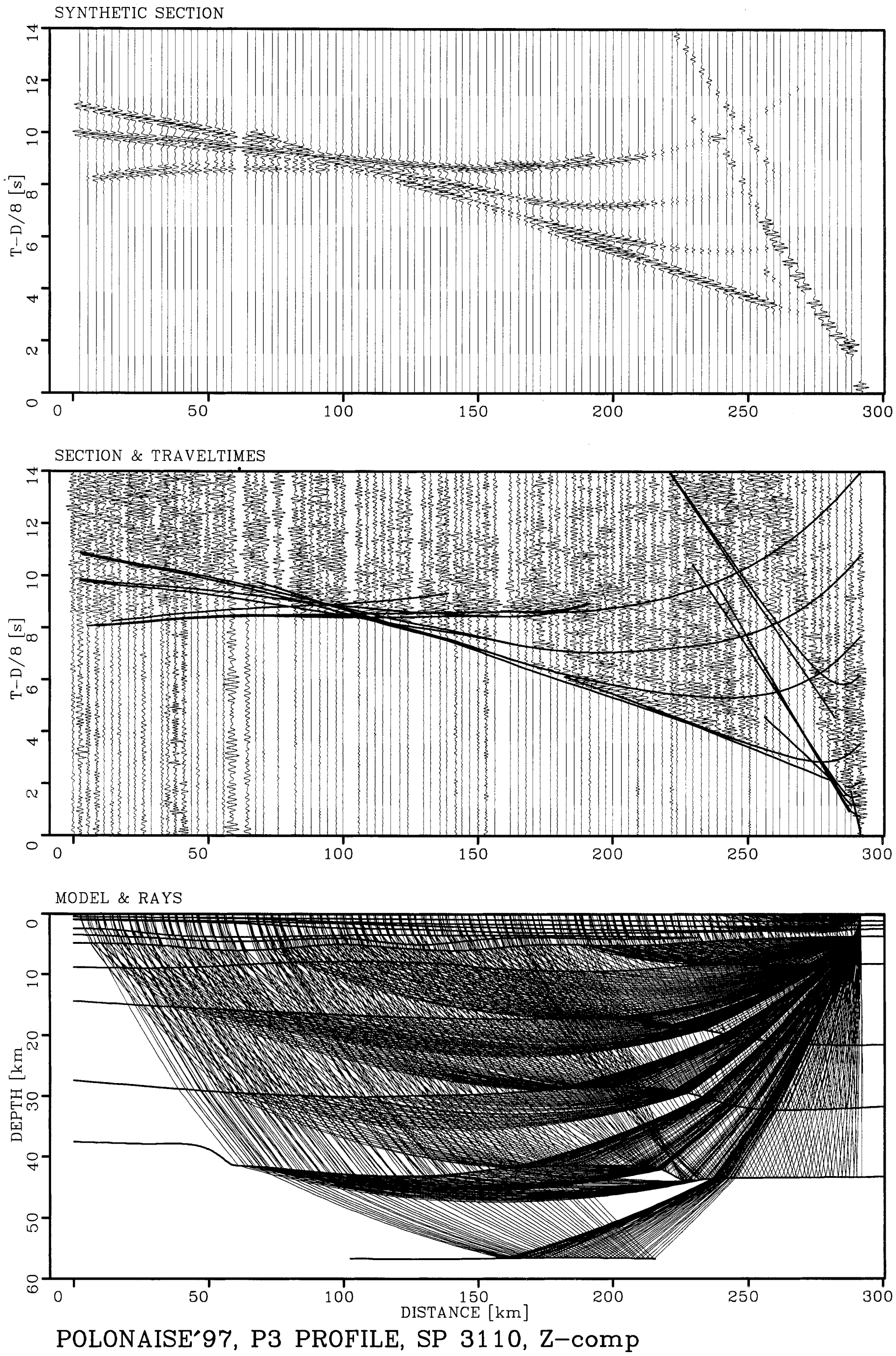

Fig. 12. A 2-D seismic modelling example for SP 3110. Bottom: model and ray diagrams; middle: seismic record section and calculated travel-time curves; top: synthetic record section.

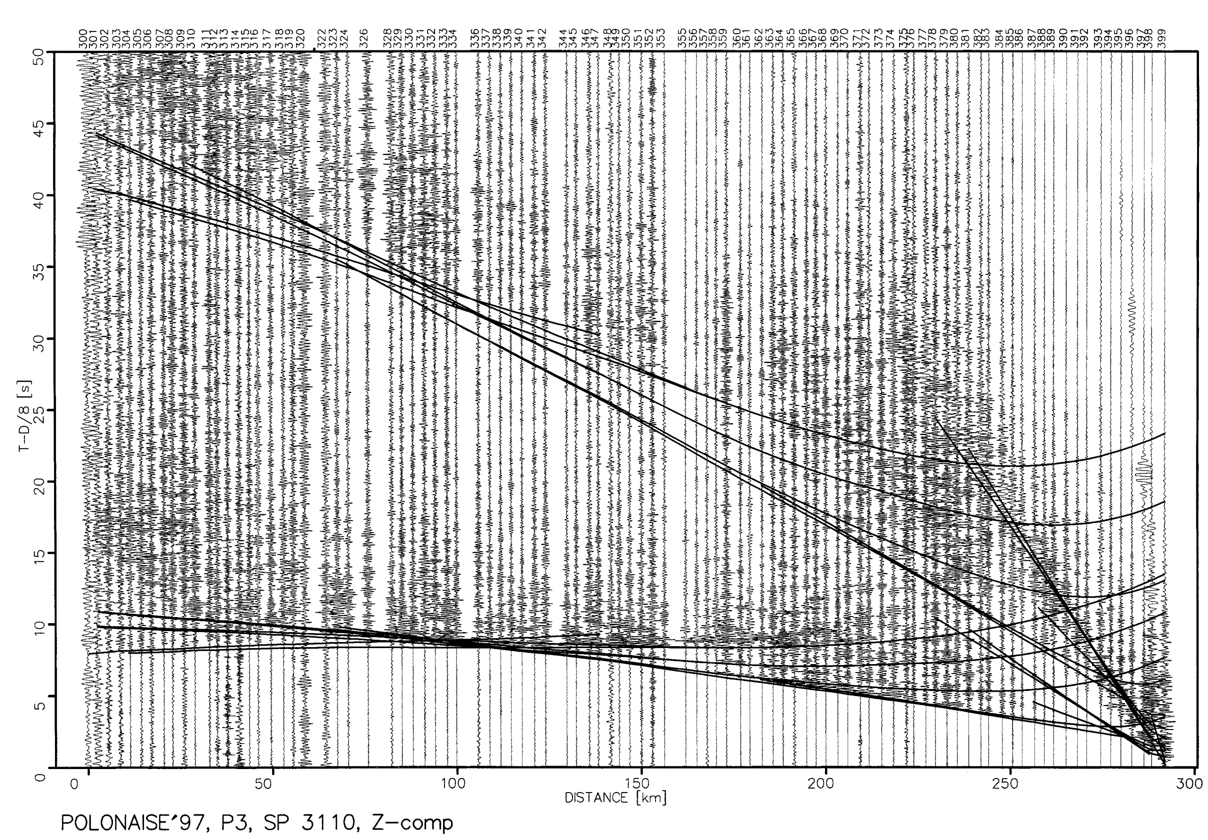

Fig. 13. Travel times for modelled P and S waves superimposed on the record section for SP 3110.

(4) A lower crustal layer characterized by a P-wave velocity of 7.15 km/s extends from a depth of about 30 km to the Moho boundary. The depth to the top of the layer was determined from reflected waves, which were not as strong as the P_mP phase, but were observed on several sections.

The uncertainty of the velocity value in this layer is significantly higher than the values for upper layers, where the velocity is determined from refracted waves. We found only one phase in the first arrivals, with an apparent velocity >7.0 km (SP 9340), and it was only observed over a short offset interval of 25 km. This phase may be interpreted as a lower crustal refraction, providing a direct constraint for the P-wave velocity of the lower crust. Indirect constraints on the velocity of the lower crust were provided by modelling the amplitudes and travel times of P_mP reflections. The curvature of the P_mP phase is an indication of velocity, and our analysis of the amplitudes of lower crustal reflections and P_mP was helpful in determining velocity contrasts between the middle and lower crust and between the lower crust and mantle.

(5) The Moho discontinuity is interpreted to lie at depths of 38 km at the NW end of the profile, 42 km in the central part, and 44 km at the SE end. At distances of 50 km and 220 km from the NW end the Moho depth changes abruptly. These features were proposed in order to explain the discontinuities of the P_mP phase identified on the sections for SP 3010 and SP 3110 (Figs. 2 and 7).

(6) The uppermost mantle has a P-wave velocity of 8.05–8.1 km/s. Low relative amplitudes of the P_n phase imply a low-velocity gradient in the mantle under the Moho.

(7) A discontinuity in the upper mantle at 56 km depth explains high-velocity arrivals

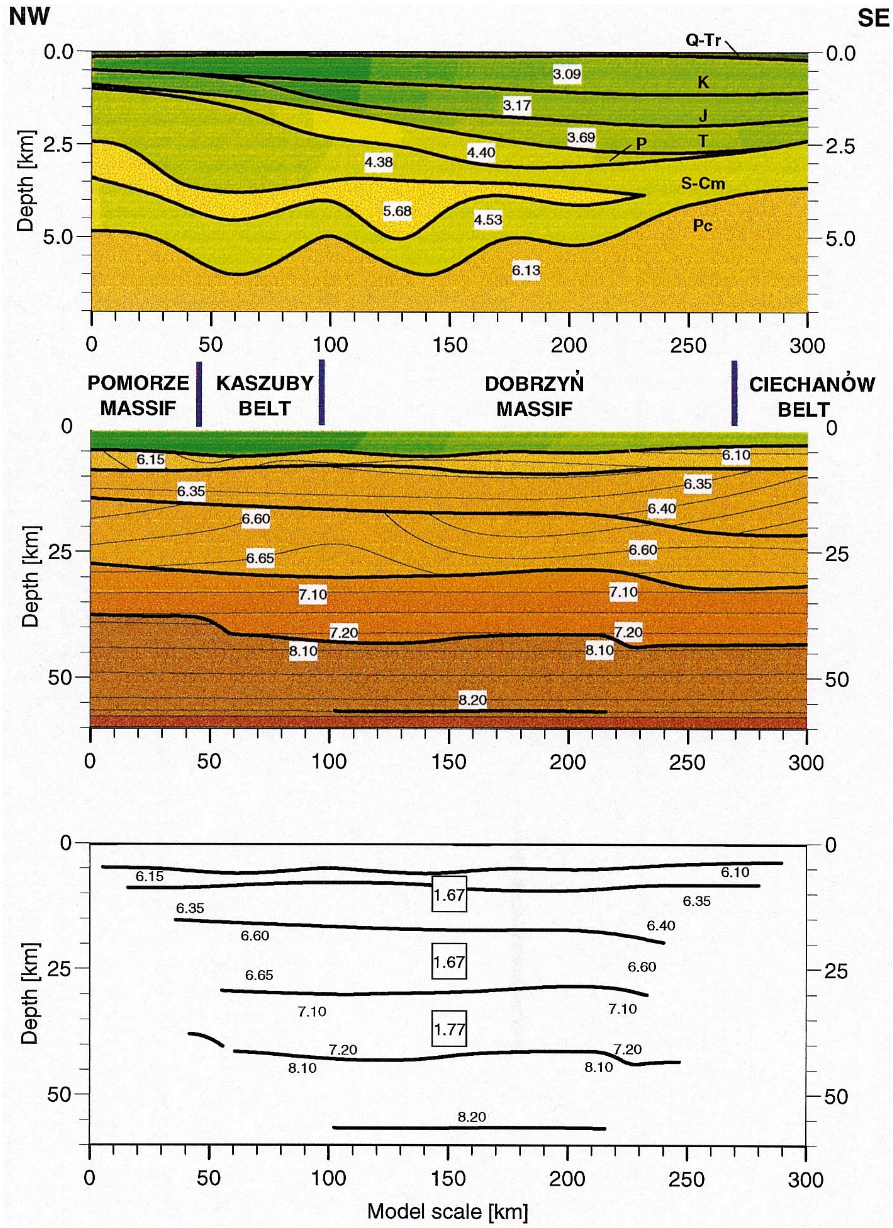

Fig. 14. The velocity model for profile P3. Top: cross-section of the sedimentary cover derived from other studies: Q-Tr, Quaternary and Tertiary; K, Cretaceous; J, Jurassic; T, Triassic; P, Permian; S-Cm, Silurian to Cambrian; Pc, Precambrian. Middle: the P-wave velocity distribution (kilometres per second) and location of basement tectonic units along the profile. Bottom: diagram showing only the constrained parts of the model; V_P values in kilometres per second, V_P/V_S values in squares (bottom). Velocity isoline interval is 0.05 km/s.

observed on sections for SP 3100 and 3110 (Figs. 7 and 8). However, these arrivals were observed over only short offset intervals. The velocity contrast at the discontinuity is small (on the order of 0.1 km/s), as required by amplitude modelling. We have no firm indication of whether this contrast is positive or negative, and a low-velocity layer may exist below the reflector (Fig. 14). However, its appearance at only far offsets may indicate a positive velocity contrast.

The V_P/V_S ratio was determined separately for the upper/middle and lower crust to have values of 1.67 and 1.77 respectively. For the sedimentary succession an average value of 1.80 was assumed based on the data presented in Christensen (1982). The V_P/V_S for the upper mantle has not been calculated owing to lack of S_n arrivals.

7. Discussion of errors

Uncertainties of model parameters are estimated from the uncertainty of picked travel times, the misfits between calculated and observed travel times, and the ray coverage in the model. Uncertainties due to erroneous interpretation of arrivals cannot be estimated, but their probability increases with decreasing data quality. We estimate that the uncertainties of the depths to the interfaces do not exceed 1.0 km for the upper crust and 1–2 km for the crust–mantle boundary in areas with reasonable ray coverage. Uncertainties in the P-wave velocities are 0.1 km/s for the upper/middle crust, 0.15–0.2 km/s for lower crust, and 0.1 km/s for the upper mantle. The estimated uncertainty in the V_P/V_S ratios is about ± 0.03. However, there may be additional uncertainties due to erroneous correlation of S-waves because of poor data quality.

8. Summary and discussion

The crustal structure presented in the model (Fig. 14) is horizontally uniform. The crystalline crust consists of four layers with P-wave velocities of 6.1–6.2, 6.3–6.4 (upper crust), 6.55–6.7 (middle crust) and 7.15 km/s (lower crust). This is generally consistent with other models from shield and platform areas. For example, a similar velocity structure was observed by the EUROBRIDGE Seismic Working Group (1999) in Lithuania 500 km NE of P3, except for their low-velocity zone and much higher velocity contrast at the Moho discontinuity. However, the average P-wave velocity in the crystalline crust of our model (6.6 km/s) is significantly higher than the global average for shields and platforms (6.45 km/s) reported by Christensen and Mooney (1995). In the area coinciding with the Kaszuby metamorphic belt a slight P-wave velocity increase at a depth of about 20 km was modelled. This can be explained by the presence of more mafic material than in the surrounding massifs.

The extensive database of velocities for various rock types presented by Christensen (1982) and Christensen and Mooney (1995) provides a framework for interpreting the velocity structure of our model. The upper crustal velocities correlate with a variety of felsic igneous and metamorphic rocks. The middle crust has velocities indicative of the presence of rocks such as diabase and diorite. The lower crust has relatively high velocities that suggest the presence of gabbro or granulites. The only rock type tabulated that truly matches the modelled velocities (7.15 km/s) for the appropriate depths and temperatures is mafic garnet granulite. Additional data suggestive of this composition are provided by the V_P/V_S ratios. The ratios we calculated from the velocity data in Christensen (1982) and Christensen and Mooney (1995) were 1.75 for granulites and 1.9 for gabbros. Thus our observed ratio in the lower crust (1.77) also indicates a granulitic composition.

Depth to the Moho ranges from 38 km at the NW end of the profile to 44 km at the SE end. Local changes of depth of the otherwise horizontal Moho, resulting in crustal thickening in the SE direction, may be associated with tectonic segmentation of the crust (Figs. 1 and 14). Areas of the depth change occur close to tectonic unit boundaries. The thickening of the crust in the SE direction is consistent with the trend observed in the EUROBRIDGE Profile (EUROBRIDGE Seismic Working Group, 1999). In the upper mantle a P-wave velocity of 8.05–8.1 km/s is observed, which is less than that previously found in neighbouring areas of the EEC. However, this value is consistent with global averages reported by Christensen and Mooney (1995).

It is interesting to compare these results with models from other seismic experiments located close to the Tornquist Zone. The BABEL experiment profiles A and B, located across the Tornquist Zone in the Baltic Sea 400 km NW from P3, reveal the structure of the crystalline crust, which is in some places almost identical to P3 (BABEL Working Group, 1993). A rather different structure is presented by models of the crust on the SW margin of the Baltic Shield in southern Sweden. Here the crust is slightly thinner, generally less differentiated and lacking significant regions with P-wave velocities over 7.0 km/s (EUGENO-S Working Group, 1988; Thybo, 1997).

An upper mantle discontinuity with a small velocity contrast is proposed at a depth of about 56 km. A similar feature has been proposed by Grad et al. (1991a) in the EEC area near the Baltic coast, based on long-offset (up to 650 km) recordings along the EUGENO-S profile 4. In the Lithuanian part of the EEC, such a mantle discontinuity was found by the EUROBRIDGE Seismic Working Group (1999) at a depth of about 70 km. The BABEL Working Group (1993) also noted the existence of a similar reflected phased but did not model it. Thus, such mantle reflectors may be typical for large parts of the Baltic Shield and EEC.

Our analysis of relative amplitudes of the seismic data suggests that the lack of coherent reflections from the upper and middle crust results from the small velocity contrasts between individual crustal layers. Variation in the wave field of the Moho reflections (in some places of high amplitude and coherent, in other places scattered and irregular) suggests a horizontally variable character for this boundary, with a complex transition zone present in some places and a sharp interface in other places. Similar variability has been used to distinguish between tectonic features in the Tornquist Fan area to the NW (Thybo et al., 1998). In future studies, details of such differences should be tested, e.g. using waveform-modelling techniques such as the reflectivity or finite differences methods.

The crustal velocity model is generally consistent with the results of previous deep seismic measurements, although the high density of shots and receivers used in this project make the current results more accurate and detailed than the previous ones. Hence, the model presented (Fig. 14) is the best constrained velocity structure in the Polish part of the EEC.

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