

THE CRUST-MANTLE TRANSITION AND THE MOHO BENEATH THE VOGTLAND/WEST BOHEMIAN REGION IN THE LIGHT OF DIFFERENT SEISMIC METHODS

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ABSTRACT

The structure of the crust and the crust-mantle boundary in the Vogtland/West Bohemian region have been a target of several seismic measurements for the last 25 years, beginning with the steep-angle reflection seismic studies (DEKORP-4/KTB, MVE-90, 9HR), the refraction and wide-angle experiments (GRANU'95, CELEBRATION 2000, SUDETES 2003), and followed by passive seismic studies (receiver functions, teleseismic tomography). The steep-angle reflection studies imaged a highly reflective lower crust (4 to 6 km thick) with the Moho interpreted in a depth between 30 and 32 km and a thinner crust beneath the Eger Rift. The refraction and wide-angle reflection seismic studies (CELEBRATION 2000) revealed strong wide-angle reflections in a depth of 26–28 km interpreted as the top of the lower crust. Long coda of these reflections indicates strong reflectivity in the lower crustal layer, a phenomenon frequently observed in the Caledonian and Variscan areas. The receiver function studies detected one strong conversion from the base of the crust interpreted as the Moho discontinuity at a depth between 27 and 37 km (average at about 31 km). The discrepancies in the Moho depth determination could be partly attributed to different background of the methods and their resolution, but could not fully explain them. So that new receivers function modelling was provided. It revealed that, instead of a first-order Moho discontinuity, the observations can be explained with a lower crustal layer or a crust-mantle transition zone with a maximum thickness of 5 km. The consequent synthetic ray-tracing modelling resulted in the model with the top of the lower crust at 28 km, where highly reflective lower crustal layer can obscure the Moho reflection at a depth of 32–33 km.

Key words: Bohemian Massif, Vogtland/West Bohemia, crustal structure, Moho, refraction and wide-angle reflection, receiver function, seismic methods, Eger Rift

1. INTRODUCTION

The crust-mantle boundary, the Moho (*Mohorovičić, 1910*), has been a major target for the Earth scientists for the last one hundred years, since its first observation from an

earthquake in 1909. Since then, our knowledge of this prominent boundary has improved. From seismic reflection profiling the crust-mantle transition can be divided into three categories: no distinct reflections at the Moho level, one or more sub-horizontal reflections, and reflections that project from the crust into the mantle (Cook, 2002). On the other hand, the reflection Moho can differ from the crust-mantle boundary obtained by other geophysical surveying. In petrologic conception, the Moho can be attributed either to changes in the bulk chemical composition or to the transition in mineral phases (Mengel and Kern, 1991). It is clear that various methods sample this boundary in a different way, measure different physical properties and have different spatial resolution (e.g. Hammer and Clowes, 1997). Thus some elements of the Moho obtained by several methods are not seen, as well as contrasting definitions may lead to different understanding of this prominent boundary.

This is just the case for the Vogtland/West Bohemian region in the western part of the Bohemian Massif. The area is situated in the transition between different Variscan structural units: the Saxothuringian zone in the northwest, the Teplá-Barrandian and Moldanubian zones in the southeast (see Fig. 1). The whole region was affected by the Permo-Carboniferous post-orogenic extension as well as the alkaline magmatism during the Cenozoic evolution of the Eger Rift. This geodynamically active zone belongs to the European Cenozoic Rift System (Prodehl *et al.*, 1995). The most recent expressions of the geodynamic activity are represented by the widespread CO₂ emanations and Quaternary volcanism (Bräuer *et al.*, 2005a,b). It is also a region of a frequent occurrence of seismic activity manifested by the intraplate earthquake swarms with magnitudes up to 4.6 (Horálek *et al.*, 1996, 2000; Fischer and Michálek, 2008). Fischer and Horálek (2003) locate the hypocenters of these swarms to the upper and middle crust down to about 20 km depth with the majority between 5 and 15 km.

Geographically it is a border region between Saxony, Bavaria and Bohemia and its deep structure was a target of several deep seismic measurements in the last 25 years, beginning with the steep-angle reflection seismic studies as DEKORP-4/KTB, MVE-90, 9HR (DEKORP Res. Group, 1988; Behr *et al.*, 1994; DEKORP and Orogenic Processes Working Groups, 1999; Tomek *et al.*, 1997) as well as the refraction and wide-angle reflection experiments as GRANU'95 (Enderle *et al.*, 1998), CELEBRATION 2000 with CEL09 profile (Hrubcová *et al.*, 2005) or SUDETES 2003 with S01 profile (Grad *et al.*, 2008). Passive seismic experiments with permanent and temporary seismic stations were carried out to study the major lithospheric discontinuities using the receiver function method (Geissler *et al.*, 2005; Wilde-Piórko *et al.*, 2005; Heuer *et al.*, 2006). Different seismic methods, namely the receiver function and the refraction and wide-angle reflection measurements, sampled the crust-mantle boundary, which resulted in the interpretation of the depth and nature of this major discontinuity with respect to the method applied. The aim of this study is to summarize the existing seismic results and discuss them in the light of the theoretical and methodological constraints of the methods together with their resolution. Synthetic modelling of both methods was performed with the aim to find some indicators for joint interpretation.

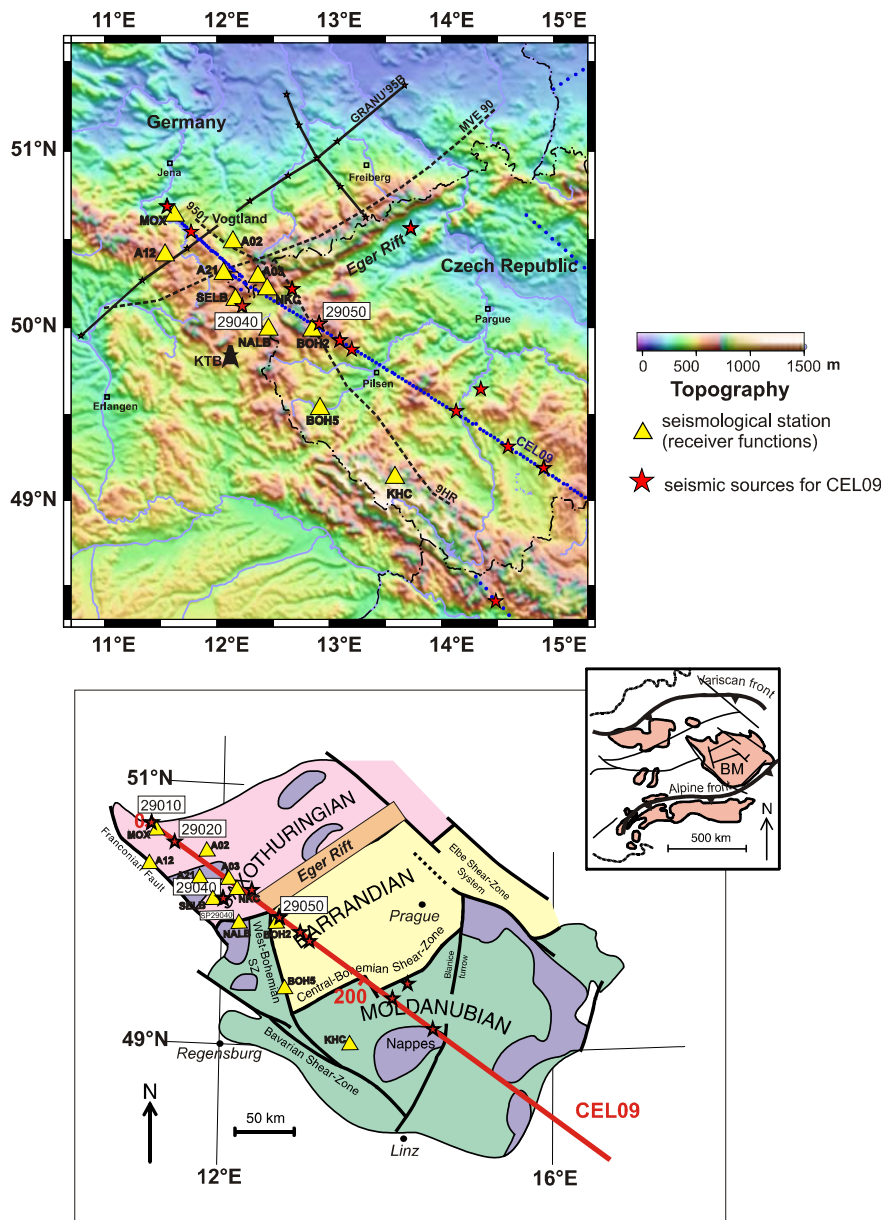


Fig. 1. Topographic and tectonostratigraphic maps of the study area with the location of seismic profiles and seismological stations (yellow triangles). Red stars represent shot points for CEL09 profile. BM, Bohemian Massif (after Pitra *et al.*, 1999; Franke *et al.*, 2000).

2. GEOPHYSICAL INVESTIGATION IN THE VOGTLAND/WEST BOHEMIAN AREA

2.1. Refraction and Reflection Studies

The crustal and uppermost mantle velocity structure in the western part of the Bohemian Massif can be inferred from the interpretation of seismic data along the refraction and wide-angle reflection profile CEL09 of the CELEBRATION 2000 experiment (*Guterch et al., 2003; Hrubcová et al., 2005*). This profile traverses the Bohemian Massif in the NW-SE direction and was interpreted by trial-and-error forward modelling using a ray-tracing algorithm (*Červený and Pšenčík, 1984*). In this modelling approach, the P-wave velocity distribution was derived using not only first arrivals, but also further phases, i.e. reflected waves and available refractions in later arrivals. The modelling was enhanced by the calculation of synthetic sections and qualitative comparison of the amplitudes of the synthetic and observed seismograms. Since the amplitudes of seismic waves are very sensitive to the velocity gradients and the velocity contrasts at the discontinuities, synthetic seismograms of both reflected and refracted seismic waves can give an additional constraint on the velocity distribution.

The ray-tracing velocity studies of the CEL09 profile in the NW part of the Bohemian Massif (*Hrubcová et al., 2005*) revealed a lower crustal layer with a velocity gradient from 6.9 to 7.5 km s⁻¹ above the Moho (Fig. 2). Strong reflection from the top of this layer in a depth of 26–28 km was not attributed to the Moho discontinuity though this reflection was the strongest reflected phase in the NW part of the CEL09 profile and based on its amplitude and shape it might have been interpreted as a reflection from the Moho (PmP phase). But as discussed in *Hrubcová et al. (2005)*, the arrival time of the observed refraction from the upper mantle (Pn phase) did not fit the critical point of the mentioned reflected phase as it should fit in case of refraction and reflection from the same discontinuity (Fig. 2). The Pn mantle refraction occurred 1 s later and did not fit a strong reflection from the top of the lower crust. Moreover, this strong reflection masked a relatively weak PmP Moho reflection arriving later. The character of the reflection from the top of the lower crust showed long and irregular coda, which indicated that the lower crustal layer was highly reflective probably due to a presence of thin layers of the material with contrasting seismic velocities.

The differences in amplitudes and coda length of the reflected phases along the CEL09 profile that were not possible to analyze by the ray-tracing modelling were also studied by the reflectivity method (*Fuchs and Müller, 1971*). The idea was to simulate a variability of the lower crust and the Moho reflections assuming a 1-D seismic velocity-depth structure and especially to test the existence of laminated layers with alternating high and low velocities at the lower crustal level. Section 29040 in Fig. 3 (according to *Hrubcová et al., 2005*) displays a high amplitude reflection from the top of the lower crust, with a long coda suggesting strong reflectivity of the lower crustal layer in the Saxothuringian zone. The coda obscures a relatively weak PmP phase as a reflection from the Moho. The Pn mantle refraction is weak, but observable. The proposed 1-D velocity model explains it by the existence of the lower crustal layer with a background V_p velocity of 6.9–7.3 km s⁻¹,

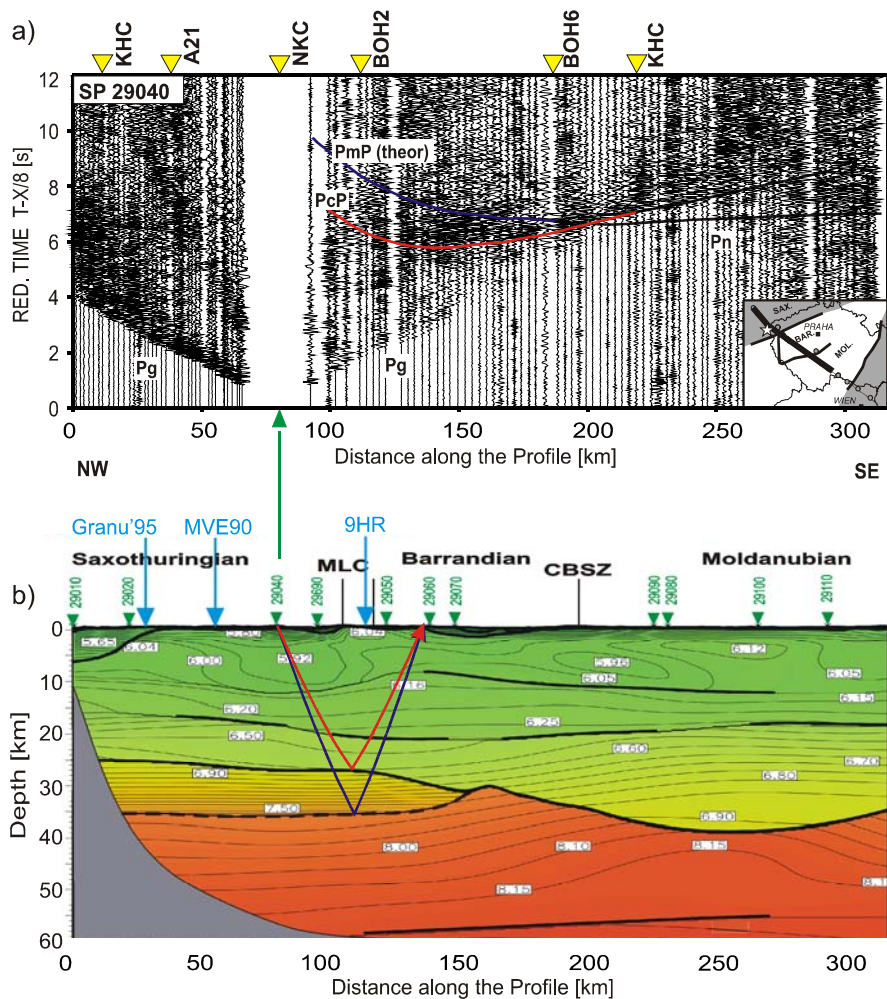


Fig. 2. a) Amplitude-normalized vertical-component seismic section of the shot point 29040 in the Saxothuringian with calculated travel times. Strong PcP (reflection from the lower crust) with long coda masks PmP (theor) phase (a theoretical reflection from the Moho) that is not visible in the data. The P_g and P_n are refractions from the crust and mantle, respectively (after Hrubcová *et al.*, 2005). Yellow triangles show projections of broad-band stations indicated on the top. Sax., Saxothuringian; Bar., Barrandian; Mold., Moldanubian. b) The 2-D model of P-wave velocity along the CEL09 profile developed by forward ray-tracing modelling. Bold lines mark boundaries constrained by the reflections and well constrained interfaces; dashed bold lines mark layer boundaries where no reflections were observed. Thin lines represent velocity isolines spaced at intervals of 0.05 km s^{-1} . Triangles show projections of the shot points. Arrows show locations of the other refraction and reflection profiles. Red arrow represents ray reflected from the top of the lower crust; dark blue arrow represents theoretical ray reflected from the bottom of the lower crust. MLC, Mariánské Lázně Complex; CBSZ, Central Bohemian Shear Zone. Vertical exaggeration is 1:2.

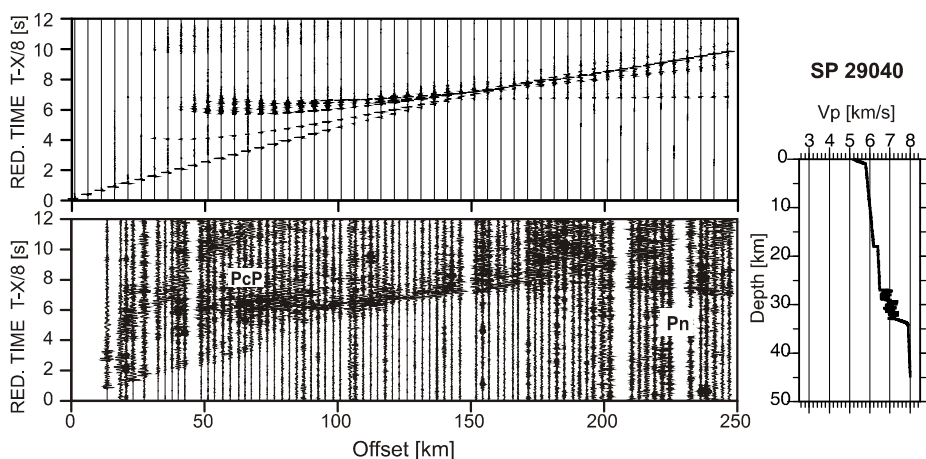


Fig. 3. Modelling of the lower crust and the Moho using the reflectivity method (*Fuchs and Müller, 1971*) for the SP 29040 along the CEL09 profile of the CELEBRATION 2000 experiment. Top - synthetic reflectivity seismogram, bottom - seismic data, right - 1-D velocity model. The synthetic and seismic data sections show amplitude-normalized vertical component plotted with the reduction velocity of 8.0 km s^{-1} . Data have been band-pass filtered by 2–15 Hz. Strong reflections form the top of the lower crust (PcP phase) and from the laminated lower crust with the coda masking the reflections from the Moho (PmP phase is not visible), weak refraction from the mantle (Pn phase) (according to *Hrubcová et al., 2005*).

consisting of layers of randomly alternating high and low velocities with a standard deviation of 4% and correlation length of 300 m. Moho was interpreted as a 1 km thick gradient zone at a depth of 33 km with velocities increasing gradually from 7.3 to 7.9 km s^{-1} , which produced a refracted phase weak enough to fit the data.

Similar results were reported by the wide-angle refraction and reflection profile GRANU'95 and the reflection profile MVE-90. Being perpendicular to the CEL09 profile, they imaged a highly reflective lower crust with a thickness of 4 to 6 km and with an uppermost mantle more or less without reflections. The Moho was interpreted at 10 s of the two-way travel time, which corresponds to a depth from 30 to 32 km, where strong reflectivity dies out. The GRANU'95 results indicate the existence of a high-velocity layer at the base of the crust in the Saxothuringian zone in SE Germany with a velocity increase from 6.5 km s^{-1} in the middle crust to an average value of 7.0 km s^{-1} at a depth of 24 km interpreted as the top of the lower crust (*Enderle et al., 1998*). The MVE-90 profile shows a highly reflective layer in this area at 8–10 s two-way travel time (Fig. 4) corresponding to a depth range of 24–32 km (*DEKORP Research Group, 1994*), which may be viewed as the laminated lower crust. Average P wave velocities in the lower crust and the upper mantle along the CEL09 profile (7.1 and 7.9 km s^{-1} , respectively) are consistent with the velocities along the GRANU'95 and MVE-90 profiles (7.0 and 7.9 – 8.1 km s^{-1} , respectively). However, the CEL09 crustal thickness of 34 km is slightly bigger than in case of GRANU'95 and MVE-90 where they interpreted the thickness of 30

and 33 km, respectively. The seismic reflection profile 9HR (Tomek *et al.*, 1997) indicated a thinner crust of about 29 km with the Moho at 9.2 s the two-way travel time beneath the Eger Rift deepening towards SE and imaged reflectors in the uppermost mantle beneath the rift axis (Fig. 5).

2.2. Receiver Function Studies

The receiver function analysis is based on teleseismic recordings of the three-component broad-band seismological stations and show the relative response of the Earth's structure near the receiver (e.g., Vinnik, 1977; Kind *et al.*, 1995; Yuan *et al.*, 1997; Geissler *et al.*, 2008). Seismic phases converted from P-to-S (Ps conversions) at discontinuities underneath the receiver are particularly useful for studying seismic structure of the lithospheric discontinuities (Fig. 6). The strongest P-to-S conversions with positive polarity (indicating a velocity increase with depth) are often attributed to the Moho discontinuity that is usually the sharpest seismic velocity contrast near the crust-mantle boundary. The delay time of these P-to-S converted phases in relation to the

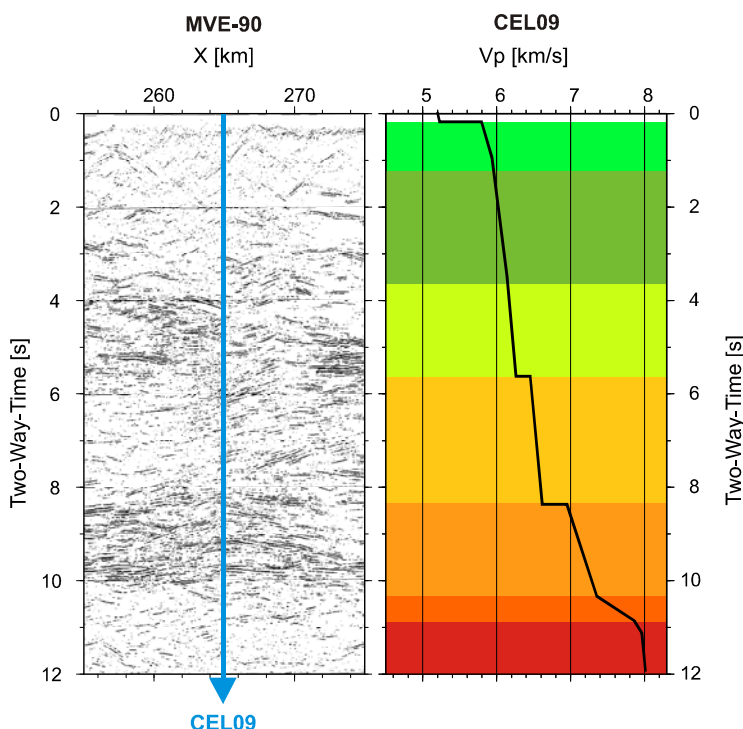


Fig. 4. Comparison of the reflections along the seismic reflection profile MVE-90 (DEKORP Research Group, 1994) with the 1-D velocity model from the refraction and wide-angle reflection profile CEL09 (converted to two-way travel time) at the crossing point. Note band of reflectors between 8.2 s and 10 s of two-way travel time at the MVE-90 profile corresponding to the high gradient lower crust in the CEL09 profile.

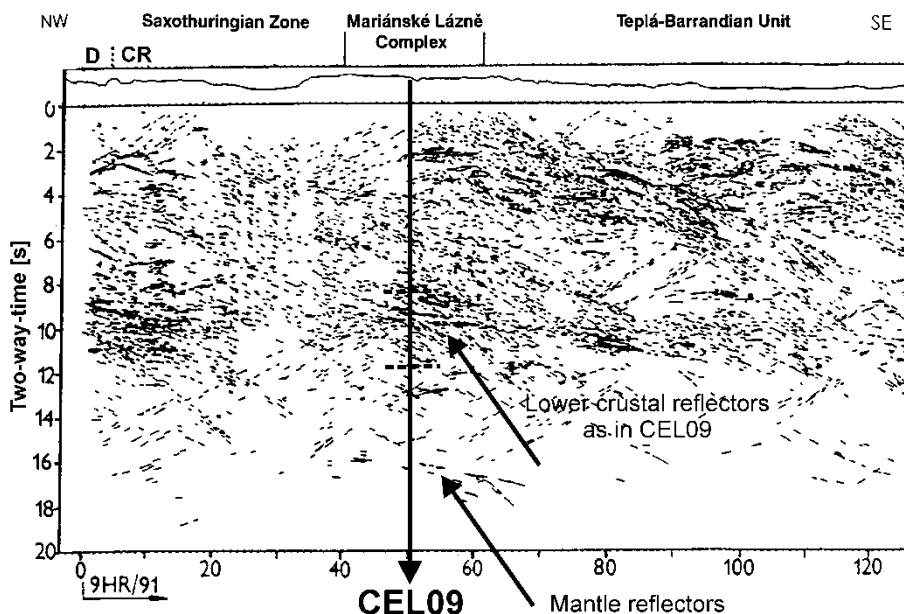


Fig. 5. The reflection seismic profile 9HR (Tomek *et al.*, 1997). Note the crossing point with the refraction and wide-angle seismic profile CEL09 and interpreted lower crust.

P waves depends on the depth of the discontinuities and the S wave velocities above them. Their amplitudes depend on the contrast of seismic velocities, densities, and the incidence angle of the impinging P wave.

If a velocity model of the crust is known, the crustal thickness can be calculated from the measured delay times of the Moho P-to-S conversion. However, the crustal thickness estimated only from the delay time of the Moho P-to-S converted phase trades off strongly with the crustal V_p/V_s ratio. The ambiguity can be reduced significantly by incorporating the later multiple converted phases, namely, the PpPs and PpSs and PsPs (Fig. 6, see Geissler *et al.*, 2005 for references).

To separate different types of waves the receiver function studies start with a rotation from the Z, N-S and E-W (ZNE) components into the P, SV, and SH system (the LQT components). For the rotation of the horizontal components, theoretical values of back azimuths are used. The angles of incidence are determined by minimizing the energy on the SV component (Q) at a time of the P signal. A time-domain deconvolution method is used to remove the source signal and source-side reverberations from the records in order to allow records from different events to be stacked. The amplitudes of the SV and SH components are normalized in relation to the incident P wave. The arrival times of converted phases are measured as a relative travel time, where zero is the equivalent to the P wave onset. For moveout correction, the IASP91 Earth reference model (Kennett and Engdahl, 1991) is used to reduce the time scale of records at any distance to the fixed reference epicentral distance of 67° (slowness: 6.4 s per degree).

The receiver function studies in the Vogtland/West Bohemian region observe converted phases from the base of the crust at about 3.7 s after the P wave onset (Fig. 7, see also *Geissler et al., 2005; Heuer et al., 2006*). Multiple conversions follow at about 13 s (positive, PpPs conversion) and 16 s (negative, PpSs and PsPs conversions). The depth to a discontinuity can be estimated from the P-to-S delay times assuming a constant average crustal velocity V_p of 6.3 km s^{-1} and V_p/V_s ratio of 1.73 (*Geissler et al., 2005; Heuer et al., 2006*) by multiplying the delay times by a factor of 8.3. Observed delay times of about 3.7 s for the primary P-to-S conversions indicate the depth of this discontinuity at about 31 km. Underneath the western part of the Eger Rift, the delay time of only 3.0–3.3 s indicate the thinning of the crust to about 27 km. As could be shown by *Heuer et al. (2006)* with a dense seismic network, this area of interpreted Moho updoming is restricted to the Cheb basin at the Eger Rift axis and its southwards continuation. Later converted P-to-S arrivals of about 4.3–4.5 s are observed at the stations in the SE part of the Vogtland/West Bohemian region, which suggests the deepening of the Moho in this area to about 36–37 km (station KHC; see *Geissler et al., 2005*).

As was shown by *Geissler et al. (2005)* the primary and multiple Moho conversions observed in the study area could be modelled assuming a rather simple crust-mantle transition with a first-order Moho discontinuity or with a thin gradient zone between the crust and upper mantle. Pronounced lower crustal high-velocity layers seemed not to be adequate to explain the observed waveforms.

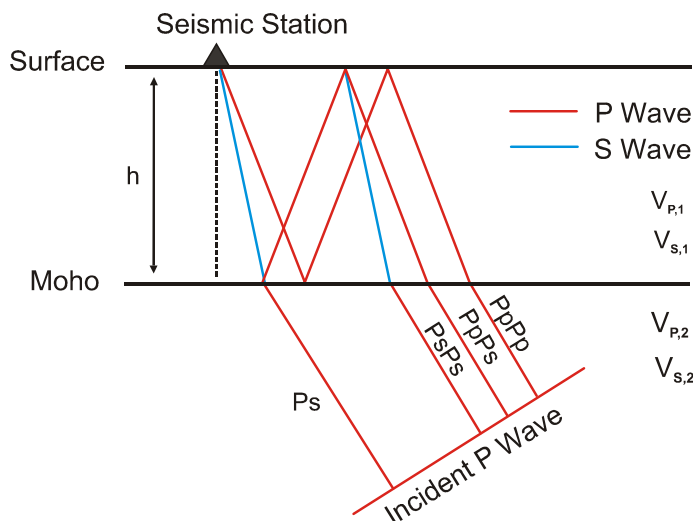


Fig. 6. Schematic model for the receiver function with a horizontal layer at a depth of the Moho over a half space. Plane P wave is incident upon the Moho from below and generates P-to-S converted waves (Ps) and many reverberations between the free surface and the Moho (e.g. PpPp, PpPs, PsPs).

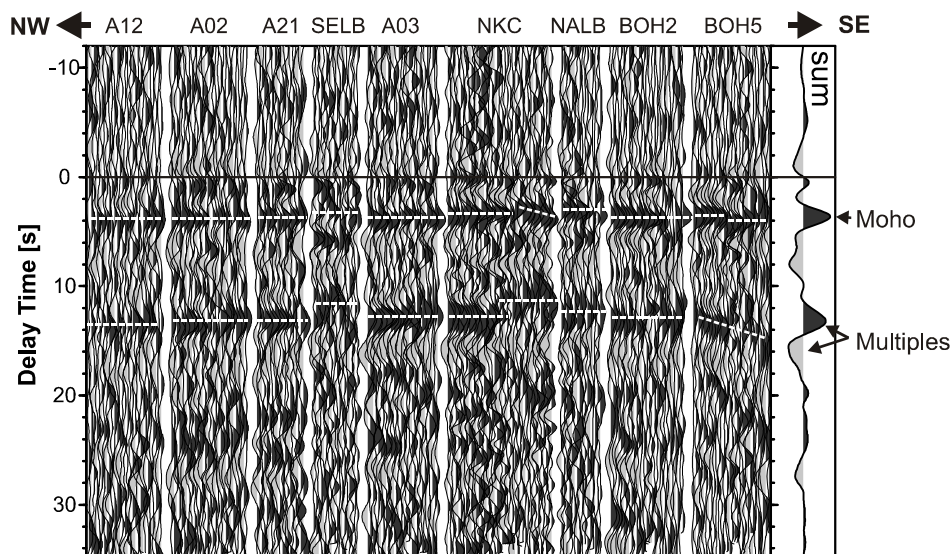


Fig. 7. Single receiver functions for individual broadband seismic stations located close to the CEL09 seismic profile and sorted by backazimuth for each station starting from N (left trace) clockwise. Primary and multiple Moho conversions are indicated. The Moho is observed at about 3.0 to 3.7 s with respect to the arrival time of the teleseismic P wave (indicated as zero), which corresponds to the Moho depths ranging from 27 to 31 km. The differences in the Moho Ps delay times can hardly be seen in the primary conversions (with the exception of stations SELB, NALB and partly NKC; upper dashed lines), but they are obvious from the multiple phases (as indicated for the PpPs phase by lower dashed lines).

2.3. Synthetic Tests for the Receiver Functions

Based on the existing results for Vogtland/West Bohemia, the Moho determination indicated by the active and passive seismic methods show differences in the depth as well as in the character of the transition from the crust to the mantle. To evaluate these differences, first of all we studied the synthetic receiver functions. We extracted 1-D velocity profiles each 10 km along the CEL09 ray-tracing model and calculated their synthetic receiver function response following the plane wave approximation approach of *Kind et al. (1995)*. For such calculation, we assumed a V_p/V_s ratio of 1.73 for the upper and middle crust, 1.76 for the lower crust, and 1.79 for the mantle. Then we compared these synthetics with the observed receiver functions (Fig. 8). The first 8 seconds delay time showed more or less good fit with the exception for the amplitudes of the strongest conversion. The observed converted phases showed higher amplitudes than the calculated synthetics, which would indicate a stronger velocity increase at the crust-mantle transition than originally modelled by the CEL09 ray tracing. Another and even bigger discrepancy exists in the amplitudes and delay times of the multiple phases (Fig. 8), where the receiver function data do not fit the synthetics calculated from the CEL09 velocity profile. A prominent feature in the synthetic receiver functions is related to the modelled

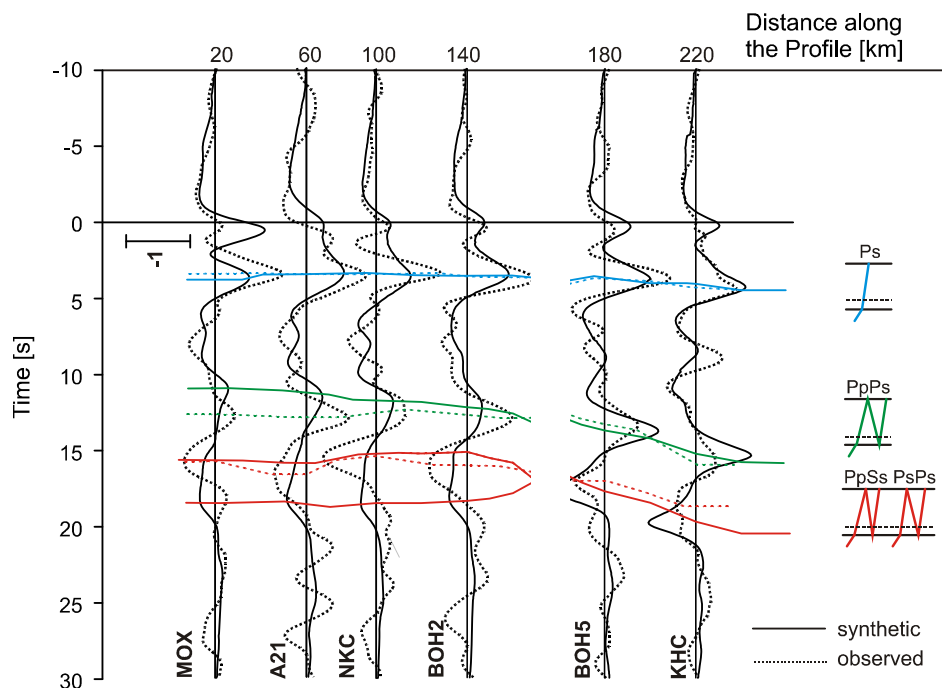


Fig. 8. Comparison of the observed receiver functions (dotted black lines) with the synthetic response of 1-D velocity models extracted from the CEL09 profile (solid black lines) at places close to the stations for the receiver functions. Scale shown corresponds to 10% of the amplitude of the primary P signal. Blue lines mark primary P-to-S conversions from the Moho, green lines are PpPs multiples, red lines represent PpSs and PsPs multiples (for each colour, the dotted lines indicate the observed data, solid lines the responses from the CEL09 model). Note that the responses from the discontinuities at the top and bottom of the CEL09 model almost coincide with the observed data in primary P-to-S conversions (blue lines), but do not coincide for the PpPs multiples (green lines) and even split for PpSs and PsPs multiples (red lines).

discontinuities on the top and at the bottom of the lower crustal layer. For the primary P-to-S conversions the receiver function responses from these two discontinuities almost coincide (Figure 8, blue lines). But they do not coincide for the PpPs multiples (Fig. 8, green lines) and even split for PpSs and PsPs multiples (Fig. 8, red lines). This can be explained in a way that there are interfering multiples from the top and bottom of the lower crustal layer. Looking at the amplitudes, the PpPs multiples seems to be more influenced by the velocity contrast on top of the lower crust whereas the PpSs and PsPs multiples seem to be more influenced by the bottom of the lower crustal layer.

Since we were not able to get a reasonable fit of the receiver function data with the response from the original CEL09 ray-tracing model, we started to alternate the velocity structure. As a reference we look at the station A03 situated at a distance of 70 km along the CEL09 profile. The station A03 showed relatively simple and coherent receiver functions in contrast to some nearby stations (e.g. NKC, Figs. 7 and 8). The upper and

middle crustal velocities were fixed to the CEL09 model. We only changed the lower crustal velocity structure to find better fit with the observed receiver functions. The original CEL09 model converted to S-wave velocity produced reasonable fit in the travel time of the primary conversion but there was no agreement in the amplitudes nor in travel times for the multiples (Fig. 9a, violet line). A model with the first-order Moho discontinuity at a depth of 30 km revealed too strong multiple phases (Fig. 9b, black line). Generally, strong multiples can be reduced by changing the density structure and/or introducing velocity gradient zones. Since we did not have good further constraints to change densities, we tried to get a better fit with further changes of the velocity structure. The best fit in the travel times and amplitudes showed the model with a discontinuity on the top of a velocity gradient zone at the base of the crust (Fig. 9c, orange line). A very good fit was also achieved for the model with a thin lower crustal layer between a depth of 26 and 30 km and a thin Moho transition layer at a depth from 30 to 32 km (Fig. 9d, blue line).

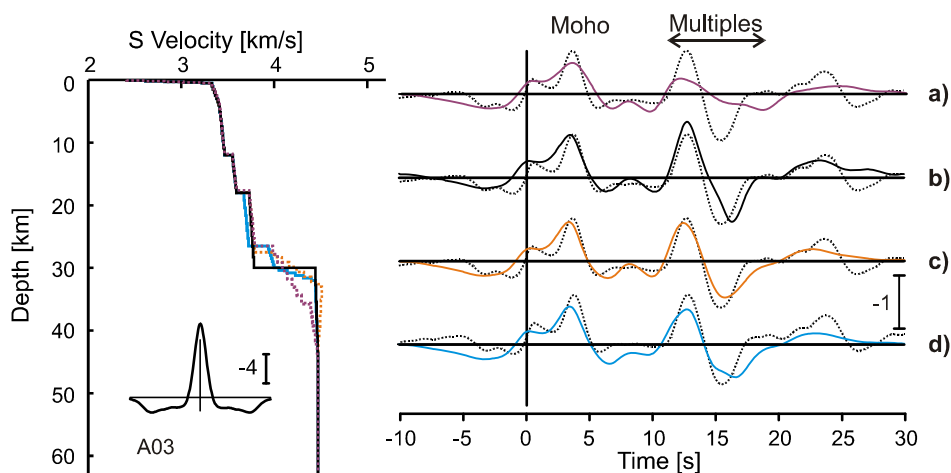


Fig. 9. Receiver functions calculated for different 1-D velocity models using the plane wave method (after *Kind et al., 1995*). The arrival time of teleseismic P wave is indicated as zero. A03 represents a form of the initial pulse. Dotted lines show the summed receiver functions for the reference station A03. Coloured lines show the receiver function response of the different velocity models. The upper and middle crustal velocities are the same for all models and are close to the CEL09 model at a distance of 70 km. The lower crustal velocities are changed to get a better fit of the data (see Table 1 for model values). Violet line - CEL09 velocity model at 70 km; black line - sharp Moho velocity contrast at a depth of 30 km; orange line - thinner lower crustal layer with velocity gradient to the Moho level (model with the best fit of the receiver function data); blue line - thin lower crustal layer at a depth of 26–30 km and thin Moho layer at a depth of 30–32 km (model that fits quite well both the receiver function and wide-angle data).

Table 1. Input seismic velocity models for modelling. n - number of layers within each interval ($n = 0$, discontinuity). Density was calculated after *Birch (1961)* from V_p .

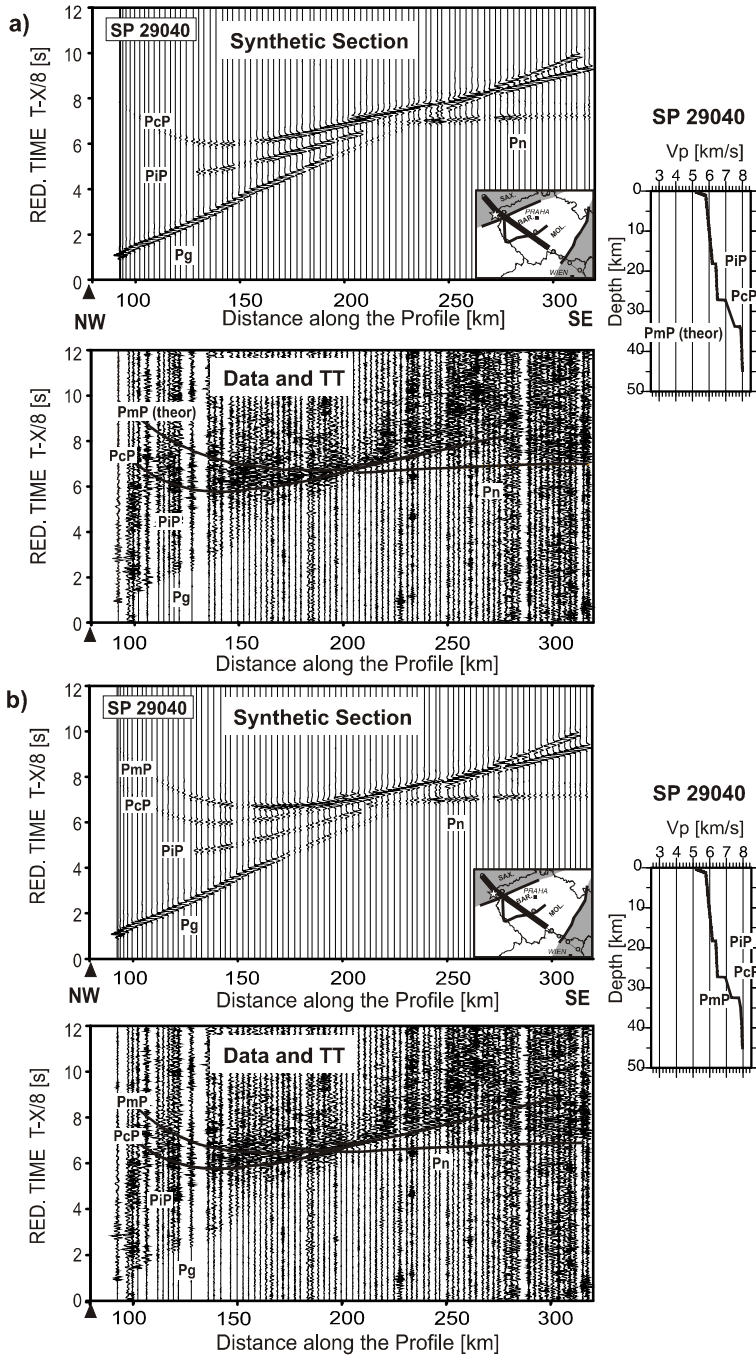
CEL09 original velocity model at km 70			n
Depth [km]	V_p [m/s]	V_p/V_s	
0.0	4.00	1.73	5
0.5	5.75	1.73	5
3.8	5.90	1.73	5
11.8	6.00	1.73	10
11.8	6.14	1.73	0
17.6	6.22	1.73	10
17.6	6.45	1.73	0
26.5	6.56	1.73	10
26.5	6.85	1.73	0
35.8	7.60	1.76	7
35.8	7.85	1.79	0
44.0	8.00	1.79	7
100.0	8.00	1.79	1

2.4. Synthetic Tests for the Ray Tracing

Synthetic receiver function modelling for the station A03 showed that the delay times and amplitudes of the primary and multiple conversions can be explained not only by the first-order discontinuity but also by a maximal 5 km thick gradient zone/lower crustal layer at the base of the crust. Since the ray tracing originally modelled the wide-angle data with a thicker lower crustal zone, pronounced discontinuity on the top of it and less pronounced Moho at a depth of 35 km (see Fig. 2), we tried to change the ray-tracing velocity model and see how it can fit the data.

For such modelling, we did not consider the NW end of the CEL09 profile where the top of the lower crust was interpreted at a depth of 26 km, because the crust-mantle transition in this area was mainly constrained by the crossing profile (GRANU'95) and not that much from the CEL09 refraction data themselves. Instead, we concentrated on Vogtland/West Bohemia, keep the top of the lower crust at a depth of 28 km and tried to model a thinner lower crust. To obtain a reasonable fit in the travel times we ended up with the Moho at a depth of about 32–33 km and the lower crustal zone with the velocities from 6.9 to 7.3 km s⁻¹ (Fig. 10).

This model has thinner lower crust and is much closer to the result of the receiver function synthetic modelling (Fig. 9d, blue line). In return, in this approach, the Moho is more pronounced with a velocity increase of 0.5 km s⁻¹. For this reason the Moho reflection is visible in the synthetic sections where we calculate also the amplitudes. This does not fully correspond to the observed wide-angle data where only a response from the top of the lower crust is visible (see Fig. 10). On the other hand, a long coda of the reflection from the top of the lower crust suggests strong reflectivity within this layer as



previously modelled with the reflectivity method by *Hrubcová et al. (2005)* and as shown in Fig. 3. The velocity fluctuations in the lower crust correspond to such variations reported by other authors in the Variscan areas, e.g., in Germany (*Sandmeier and Wenzel, 1990*) or in Poland (*Jensen et al., 1999*). *Hrubcová et al. (2005)* simulated the reflectivity of the lower crust and showed that the reflectivity can mask a weak reflection from the Moho. If we follow such approach, we may infer that some reflectivity may also obscure a more pronounced reflection from the Moho.

3. DISCUSSION

In the presented study we concentrated on Vogtland/West Bohemia with the focus on the lower crust and crust-mantle transition in this region. We tried to compare the outcomes from two different seismic methods: the receiver functions and the refraction and wide-angle reflection modelling, and constrained them by previously known results from other seismic investigations. Above, we performed new modelling for the two data sets, receiver function data and refraction and wide-angle data, to get closer to a unified solution. Previously, all methods found indications for a major discontinuity near the base of the crust but differed partly in the depth determination and in the characterization of the crust-mantle transition.

The major question at the beginning was why the original interpretations differed so much. One very important factor addressing this question is the different theoretical background and methodology of both methods as well as their spatial resolution. The receiver function provides near-vertical information on the velocity contrasts and the depth of the seismic discontinuities beneath the stations similar to that of the reflection seismics. In return, the refraction modelling gives smoothened sub-horizontal information on the velocities and depth of discontinuities along large profiles. The comparison of the wide-angle and “reflection” Moho is documented e.g. by *Barton et al. (1984)* or *Braile and Chaing (1986)*. *Jones et al. (1996)* show the differences between near-vertical reflection Moho and the Moho modelled from wide-angle data sets, where they are usually offset from each other, with the wide-angle Moho being shallower to the near-vertical Moho from reflection.

Another important factor is that each method samples the boundary at slightly different place. For the receiver function, the direct P-to-S conversions sample the Moho at a distance of about 5–10 km away from the station and its crustal multiples sample the Moho over a distance of 5–30 km from the station in the direction dependant on the back

Fig. 10. (Facing page) Comparison of the ray-tracing forward modelling along the CEL09 profile for SP 29040 illustrating the different velocities in the crust-mantle transition. In each subfigure: top - synthetic sections, bottom - amplitude-normalized seismic sections with calculated travel times, side - 1-D velocity models. Reduction velocity is 8 km s^{-1} , locations of major tectonic units and shot points are indicated. Other description as in Fig. 2. **a)** The velocity model as according to *Hrubcová et al. (2005)*, (see also Fig. 2). Note missing PmP (theor) in the synthetic section. **b)** The velocity model with the top of the lower crust located in the same position. The Moho is shallower and more pronounced. Note PmP phase visible in the synthetic section.

azimuth of the respective incoming teleseismic phase. On the other hand, the refraction and wide angle reflection give the information about the sub-horizontal velocities and discontinuities along the profile mostly from a mid-point between the shot and the receiver. For deeper crustal interfaces and the Moho it is usually more than 30 km away from the shot. From this point of view, the dipping of the reflective lower crustal boundary towards the SE and the whole lower crust edging away in the SE direction may contribute to the different results.

Another important factor there is the V_p/V_s ratio and the fact that each method applies this ratio in a different way. For calculation of the depth of the discontinuity from the observed Ps delay times we have to know the average P and S-wave velocities (e.g., average crustal V_p and V_p/V_s). But the receiver function technique provides information on the near-vertical velocity contrasts constrained by the S-wave velocity distribution in the crust. Thus, it is complementary to the controlled source profiles, in which the P waves are primarily recorded and the S waves are weak or absent.

As far as concern the V_p/V_s ratio in West Bohemia, we tried to calculate it from the WEBNET broadband data (e.g. Fischer and Horálek, 2003) and wanted to find some depth dependence, but its scatter for the upper and middle crust was quite wide. It ranged from some 1.6 to 1.8, and did not show any simple pronounced relation to the depth. Above, Kolář and Boušková (2003) tried to show some indications for its azimuthal variations, which made the task even harder.

The dispersion in the V_p/V_s values can indicate possible uncertainty in the depth determination. Geissler *et al.* (2005) interpreted the depth of the Moho at an average depth of 31 km from the P-to-S delay times assuming constant average crustal V_p of 6.3 km s^{-1} and V_p/V_s ratio of 1.73. If higher values of the V_p/V_s were applied then the interface would be shallower, and vice versa. In detail, when considering the V_p/V_s of 1.8 we could get the depth of 28 km and for the V_p/V_s of 1.6 we get the depth of 37 km. For the delay time of 3.3 s in the Eger Rift, where Geissler *et al.* (2005) interpreted the Moho at 27 km, the depth values can range in-between of 25 km and 33 km for the V_p/V_s of 1.8 and 1.6, respectively.

Also, the frequency band of teleseismic body waves covers the range of one to several seconds and thus differs from the controlled source methods with the frequency of several Hz. This determines the resolution of each method and gives the ability to see the details in the determination of the discontinuity. To prove this, we examined the model, which was previously used to explain the reflectivity from wide-angle observations. We calculated a receiver function response for a lower crust with thin (500 m thickness each) internal layers of alternating velocities. However, the frequency content of teleseismic phases does not allow imaging such thin layering and only the average velocity structure in the lower crust can be resolved.

There are some other factors that have not been taken into consideration so far, and which to some extent can influence the achieved results. The effect of anisotropy, reported in this region (Vavryčuk, 1993; Vavryčuk and Boušková, 2008) and not considered in the above studies, may play an important role in the different interpretation of the crustal thickness as shown by e.g. Jones *et al.* (1996).

If we look at all the above mentioned aspects, they are quite important though they are inadequate to explain the differences of the results from the active and passive seismic

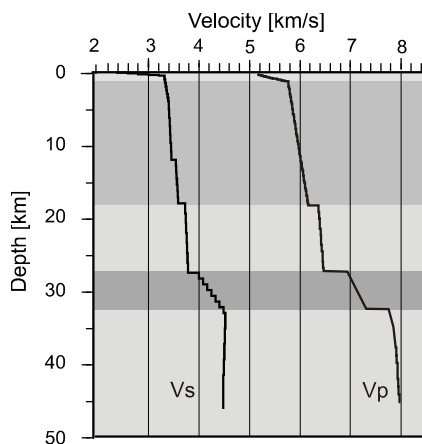


Fig. 11. Final 1-D joint velocity model. V_s - S-wave velocity from the receiver functions as for the A3 station, V_p - P-wave velocity from the ray-tracing data as for the SP 29040.

methods. For this reason, a special synthetic modelling of both the receiver function and the ray-tracing data was performed with the aim to find the indicators for some joint interpretation. The synthetic modelling of the receiver function observed at the station A03 revealed that an acceptable fit to the data could be achieved for the model with a thin lower crustal layer at a depth of 26–30 km and a thin Moho layer at a depth of 30–32 km (Fig. 9d, blue line). The consequent synthetic ray-tracing modelling showed that such thinner lower crustal layer can fit the data in the travel times, but more pronounced Moho produces a stronger reflection that is not supported by the observations (Fig. 10). On the other hand, a reflectivity of the lower crust recorded by the data and reported also in some other Variscan areas may actually obscure the Moho reflection.

The joint model (Fig. 11) ended up with the top of the lower crust at 28 km, where highly reflective lower crustal layer can obscure the more pronounced Moho at 32–33 km depth. Such model would be in accordance with the wide-angle refraction and reflection profile GRANU'95 and the reflection profile MVE-90 with a highly reflective lower crust of 4–6 km thick and with an uppermost mantle more or less without reflections. The Moho in the MVE-90 data is interpreted at 10 s of the two-way travel time and corresponds to a depth of 30 to 32 km, where the strong reflectivity dies out.

4. CONCLUSIONS

In the presented study we concentrated on Vogtland/West Bohemia with the focus on the lower crust and the crust-mantle transition in this region. We used the outcomes from different seismic methods, namely the receiver functions and the refraction and wide-angle reflection modelling. At the beginning the results of these methods differed since the refraction interpretation attributed the strong reflector at the depths of 26–28 km to the top of the lower crust with a high reflectivity. The Moho was modelled at a depth of

35 km to get a weak contrast on the Moho, compared to the top of the lower crust, in a place where the reflectivity dies out. The receiver function indicated a strong first-order converter at a depth of 31 km, with a local updoming beneath the western Eger Rift to some 27 km.

As discussed, the discrepancies can partly be attributed to different aspects of both methods and their resolution. Among them, the uncertainty in the knowledge of the V_p/V_s ratio and the different frequency band of both methods play an important role. Another fact is that the methods sample the boundary at slightly different places. However, these methodological aspects are not enough to explain the observed discrepancies.

New receiver function modelling for a reference station showed that there might be a lower crustal layer (or gradient zone) of maximal 5 km thickness instead of a single first-order discontinuity, which can explain the observed receiver functions. The consequent ray-tracing modelling along the CEL09 profile showed that such thinner lower crust can fit the data in the travel times. The reflectivity within the lower crust recorded by the data and reported also in some other Variscan areas may obscure the Moho reflection. The new model shows the top of the lower crust at a depth of 28 km, where high reflectivity obscures the Moho reflection at a depth of 32–33 km. The new findings of this study provide the base for a future joint mapping/interpretation of the crust-mantle boundary in the whole region, which is known for its complex lithospheric structure. It also shows the necessity for the joint interpretation of different data sets with regard to the resolution and background of each method applied.

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