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Discussions

Traces of the crustal units and the upper mantle structure in the southwestern part of the East European Craton

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Abstract

The presented study is a part of the passive seismic experiment PASSEQ 2006–2008 which took place around the Trans-European Suture Zone (TESZ) from May 2006 to June 2008. The dataset of 4195 manually picked arrivals of teleseismic *P* waves of 101 earthquakes (EQs) recorded in the PASSEQ seismic stations deployed to the east of the TESZ was inverted using the non-linear teleseismic tomography algorithm TELINV. Two 3-D crustal models were used to estimate the crustal travel time (TT) corrections. As a result, we obtained a model of *P* wave velocity variations in the upper mantle beneath the TESZ and the EEC. In the study area beneath the craton we observed 5 to 6.5 % higher and beneath the TESZ about 4 % lower seismic velocities compared to the IASP91 velocity model. We found the seismic lithosphere-asthenosphere boundary (LAB) beneath the TESZ at a depth of about 180 km, while we observed no seismic LAB beneath the EEC. The inversion results obtained with the real and the synthetic datasets indicated a ramp shape of the LAB in the northern TESZ where we observed values of seismic velocities close to those of the craton down to about 150 km. The lithosphere thickness in the EEC increases going from the TESZ to the NE from about 180 km beneath Poland to 300 km or more beneath Lithuania. Moreover, in western Lithuania we possibly found an upper mantle dome. In our results the crustal units are not well resolved. There are no clear indications of the features in the upper mantle which could be related with the crustal units in the study area. On the other hand, at a depth of 120–150 km we possibly found a trace of a boundary of proposed palaeo-subduction zone between the East Lithuanian Domain (EL) and the West Lithuanian Granulite Domain (WLG). Also, in our results we may have identified two anorogenic granitoid plutons.

1 Introduction

The East European Craton (EEC) (Fig. 1), the palaeocontinent Baltica, has been not tectonically reworked for at least 1.45 Ga (Bogdanova et al., 2006). The EEC includes a mosaic of tectonic structures. It has formed during the collision of three palaeocontinents: Sarmatia, Volgo-Uralia and Fennoscandia 2–1.7 Ga (Bogdanova et al., 2001; Artemieva, 2007). The EEC in the east is terminated by Uralides orogen and in the west by the Trans-European Suture Zone (TESZ) – the boundary between Proterozoic Eastern Europe and Phanerozoic Western-Central Europe (Nolet and Zielhuis, 1994). The inner major sutures in the EEC are the Central Russia Rift System and the Pachelma Rift which mark amalgation of Baltica in the north, Sarmatia in the west and Volga-Uralia in the east during the Proterozoic period (Gorbatschev and Bogdanova, 1993). During long evolution the EEC resulted in a complex structure of the crust and the upper mantle, which were intensively investigated during a number of studies (e.g. Guterch et al., 1999; Grad et al., 2002, 2006; Guterch et al., 2004; EUROBRIDGE Seismic Working Group, 1999; Pharaoh et al., 2000; Wilde-Piórko et al., 2010; Knapmeyer-Endrun et al., 2013).

Our study is focused on the SW part of the EEC. The study area covers the NW part of the territory of the passive seismic experiment PASSEQ 2006–2008 (Wilde-Piórko et al., 2008) which was carried out around the TESZ in order to study the lithosphere and asthenosphere beneath the area. The aims of our study are to define (1) whether the crustal units can be traced in the upper mantle, whether they follow the shape of the Moho boundary and how they terminate at the TESZ, and (2) to estimate the seismic P wave velocity structure of the upper mantle and the lithosphere thickness beneath the study area using the data acquired during the PASSEQ 2006–2008 project and the method of non-linear teleseismic tomography.

2 Crust and lithosphere structure

The deep seismic sounding (DSS) projects such as EUROBRIDGE (EUROBRIDGE Seismic Working Group, 1999), POLONAISE'97 (Guterch et al., 1999), CELEBRATION 2000 (Malinowski et al., 2008), etc. (Fig. 2) carried out around the TESZ in the SW part of the EEC provided crucial information about the crustal and upper mantle structure in the area to the depth of about 80 km. The structure of the upper mantle extending to several hundreds of km has been modeled during other studies (e.g. Artemieva et al., 2006; Majorowicz et al., 2003).

2.1 Crustal units in Lithuania

The NE part of the EEC is composed of several Svecofennian crustal domains (Figs. 2 and 3). Grad et al. (2006) and Motuza et al. (2000) summarized the results of the DSS projects conducted in the region and distinguished different tectonic domains in the upper lithosphere along the EUROBRIGDE profile: the Vestervik-Gotland block (the Trans-Scandinavian Igneous Belt), the West Lithuanian Granulite Domain (WLG), the East Lithuanian Domain (EL), and the Belarus–Podlasie Granulite Belt (BPG). The Moho boundary in the WLG is 42–44 km, while in the EL and the BPG it varies from 50 to 57 km. The 35–40 km wide zone in between the WLG and the EL with abrupt change in crustal thickness, seismic velocities and other physical parameters is known as the Middle Lithuanian Suture Zone which is considered as a palaeosubduction zone along which the terrain in the east subducted under the terrain in the west. Motuza (2005) also interpreted the rocks of the crystalline crust of the WLG as a back-arc complex, rocks of the Middle Lithuania Suture Zone as a volcanic island arc complex, and the rocks of the EL as an accretionary complex. The contact between the EL and the BLG further to the NE is not so prominent (Motuza, 2005). In the WLG the seismic velocities in the uppermost mantle vary from 8.65 to 8.9 km s^{-1} and increase along the EUROBRIDGE profile from the west to the east (Motuza et al., 2000). The crustal features of the EL

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show lineaments extending the NE–SW direction which coincide with the direction of collision with Sarmatian palaeocontinent (Motuza, 2005; Bogdanova et al., 2001).

The anorogenic magmatism took place around Lithuania and the adjacent areas 1.6–1.5 Ga resulting in a number of granitoid intrusions (Bogdanova et al., 2006). Two large granitoid bodies of rapakivi-type are present in our study area: the Riga Pluton in western Latvia and the Mazury Complex in the Kaliningrad District of Russia and NE Poland (Rämö et al., 1996; Dörr et al., 2002).

2.2 Crustal units in Belarus

The junction between Fennoscandia and Sarmatia is significant in Belarus (e.g. Bogdanova et al., 1996) (Figs. 2 and 3). The crustal pattern in the area shows crustal units with alternating granulite and amphibolite facies which vary in age and origin. The structural features suggest that the accretion was driven by several events of subduction and collision, and the accretionary tectonics prevailed 2.0–1.8 Ga (Bogdanova, 1999; Claesson at al., 2001).

The Volhyn-Orsha Aulacogen (VOA) of Meso- to Neoproterozoic age follows the junction of Fennoscandia and Sarmatia while the Osnitsk-Mikashevichi Igneous Belt (OMIB) represents an active continental margin along the NW edge of Sarmatia (Bogdanova et al., 1996). The 200–250 km wide OMIB consists of various grades of amphibolite facies (Aksamentova and Naydenkov, 1991) and contain large batholiths of age 2.02–1.95 Ga which are only slightly metamorphosed and deformed and younger rapakivi-type granites of age 1.0–1.75 Ga (Skobelev, 1987).

At the edge of Sarmatia there are the Central Belarus Belt (CB) and the Vitebsk Granulite Domain (VG) of the Palaeoproterozoic age (about 2.0 Ga). The VG adjoins the CB in the east and NE. Bogdanova et al. (1996) and Stephenson et al. (1996) indicated the complex crustal structures along the Fennoscandia-Sarmatia junction with the VG and the CB slightly dipping to the SE direction beneath the edge of Sarmatia. The CB consists of bodies of amphibolite and granulite facies (Bogdanova et al., 2001) with significant tectonic faults separating the units of different composition. The study

of Claesson et al. (2001) showed that the subcrustal rocks of the VG are similar to the ones of the southeastern CB.

2.3 Crustal units in Poland

The before mentioned (see Sects. 2.1 and 2.3) crustal units of prolonged shape (or "belts") from Lithuania and Belarus continue to the SW direction into Polish territory (Bogdanova et al., 2006) and terminate at the TESZ (Fig. 2). The results obtained during the POLONAISE'97 (Guterch et al., 1999) and CELEBRATION 2000 DSS projects provided detailed models of the crust and the upper mantle structure in Poland (e.g. Czuba et al., 2001; Malinowski et al., 2008). The boundary between the EEC and the TESZ in NW Poland is near vertical and suggests a strike–slip character (Grad et al., 2005). The westernmost part of the EEC adjoining the TESZ has thick continental crust of average thickness of 40–50 km (Grad et al., 2006; Guterch et al., 2004). Dadlez et al. (2005) and Grad et al. (2006) discussed in details the structure of the crust and the uppermost mantle in the SW part of the EEC obtained by different DSS projects (Fig. 3). There were reported some steep changes in the Moho depths and the seismic velocities along some profiles, e.g. a "step" with the increase in the Moho depth from 42 to 44 km (which is comparable to the resolution of the method) was found at the P5 profile between the Mazury Complex and the Mazowsze Massif (Czuba et al., 2001). Dadlez et al. (2005) summarized that not all Moho "steps" occur exactly at the places of the proposed terrain boundaries. Moreover, no clear boundaries are visible in the crust between Precambrian terrains postulated by Bogdanova et al. (1996).

2.4 Upper mantle structure in the study area

The cratonic lithosphere has been shown to extend to depths of about 200–250 km (Plomerova et al., 2002; Eaton et al., 2009) which is deeper than that of the younger continental regions (e.g. Shomali et al., 2006; Gregersen et al., 2010). Artemieva (2003) found thickness of the thermal lithosphere of about 250–275 km in the EEC

for the Archean Kola-Karelian province and some parts of Volga-Uralia. However, the seismic lithosphere is systematically thicker about 50 km than the thermal lithosphere (Artemieva, 2007). Sandoval et al. (2004) indicated the high-velocity anomaly extending to at least 250 km depth beneath the central part of the Fennoscandian Shield using the method of body-wave tomography. Hjelt et al. (2006) also reported that in the Fennoscandian Shield the seismic velocity anomalies extend to the depths of at least 250–300 km. The study of Artemieva et al. (2006) showed the thickness of the thermal lithosphere of about 180 km for the EEC, while the results of geothermal modeling obtained by Majorowicz et al. (2003) indicated thermal lithosphere thickness of 200 km for the EEC. The study of Artemieva et al. (2006) showed thickness of the seismic lithosphere more than 250 km beneath the SW part of the EEC.

The reflectors in the upper mantle just beneath the Moho boundary in Fennoscandia have been found by Czuba et al. (2001), Yliniemi et al. (2004) and Grad et al. (2002). A major southwards dipping reflector has been found beneath the EUROBRIDGE'97 profile, extending from the Moho boundary down to the depth of about 75 km (Thybo et al., 2003), while a steeply SW direction dipping mantle reflector reported below the OMIB and the VB correlates with a subhorizontal reflector in the EUROBRIDGE'96 profile. Similar subhorizontal lithospheric reflectors were observed beneath the TESZ (Grad et al., 2002; Guterch et al., 2004) and the Baltic Sea (Hansen and Balling, 2004). Beneath the WLGD in the upper mantle there were reported reflectors at a depth of 73–82 km which originated possibly due to delamination processes (Motuza et al., 2000; Motuza, 2005). Moreover, there was found a locally increased heat flow ranging between 55 and 100 mW m^{-2} in the WLGD (Kepezinskas et al., 1996; Rastiene et al., 1998).

3 Dataset

We used data recorded during the PASSEQ 2006–2008 project (Wilde-Piórko et al., 2008) which took place around the TESZ from June 2006 to July 2008 (Fig. 4). Using

seismological bulletins of the USGS (http://earthquake.usgs.gov/) and the ISC (http://www.isc.ac.uk/) we prepared a list of 101 earthquakes (EQs) with epicentral distances from 30 to 92° (Artlitt, 1999; Sandoval, 2002) with respect to the central point at the Lithuanian–Polish border (coordinates 23° E and 54° N) and the magnitude range from 5.5 to 7.2 (Fig. 5). The higher and the lower values of the epicentral distance ensure that the first observed arrivals are the direct P waves and that they hit the target area from below steeply enough. The relatively large magnitudes ensure better quality of the observed seismic signals; on the other hand, the magnitudes should not be too large, because it is difficult to interpret the seismic signals generated by large-scale seismic sources.

We used Seismic Handler Motif (SHM) program package (http://www.seismic-handler.org/) to perform analysis and manual picking of the teleseismic P wave arrivals. During the data analysis we applied the World Wide Standardized Seismographic Network – Short Period (WWSS-SP) filter which includes both simulation filtering and instrument response with autocut of 0 s, and picked the P wave arrivals on seismograms of vertical components (Fig. 6). Every P wave arrival was assigned with a quality (weighting) factor depending on the time error (Table 1). The data with quality factor > 3 was not used in the inversions. We compiled a dataset of 4195 P wave arrivals from the data of 94 seismic stations.

We used Seismic Handler (SH) program package and location information of the listed 101 EQs from the ISC seismological bulletins to calculate the theoretical travel times (TT) of the first teleseismic P wave arrivals. Then we applied a subtraction procedure in order to obtain the TT residuals for every picked arrival:

$$T_{\text{picked}} - T_{\text{theoretical}} = T_{\text{residual}} \tag{1}$$

where T_{picked} is the observed TT, $T_{\text{theoretical}}$ is the theoretical TT calculated with SH, and T_{residual} is the TT residual.

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4   Inversion procedure

4.1   Teleseismic tomography inversion

We used TELINV code (Voss et al., 2006) to perform inversion of the compiled dataset. The program utilizes a nonlinear inversion method and can either (1) calculate propagation of rays through a 3-D velocity model and output TT, raypaths and synthetic relative TT, or (2) invert teleseismic relative P wave residuals for 3-D velocity structure. The ray tracing is performed computing the 3-D minimum TT raypaths assuming a constant slowness in each cell (Steck and Prothero, 1991). The ray coverage of the cell blocks is affected by horizontal and vertical grid spacing (Arlitt, 1999). For the full description of the inversion procedure see Thomson and Gubbins (1982), Thuber (1983), Menke (1984), Koch (1985) and Aki et al. (1997).

4.2   Crustal travel time corrections

In teleseismic tomography it is very important to use reliable crustal TT corrections in order to eliminate the effects which are created by the Earth's crust, while the crust is much more heterogeneous compared to the deeper layers of the Earth. The variation of thickness of the sedimentary cover is significant in the study area ranging from several tenths of meters in the Belarus-Mazurian High to almost 20 km in the Polish Basin, while the Moho variation is from 35 km beneath the TESZ to almost 60 km beneath the NW Belarus. The teleseismic tomography inversion performed not correcting the crustal effects (Fig. 7) show significant influence from the thick sedimentary cover beneath the TESZ.

The crustal TT corrections which we used in our study have been compiled using two 3-D crustal models by Majdański (2012) for Poland (Fig. 8a left) and by M. Budraitis (unpublished) for Lithuania (Fig. 8a right). Both models have been compiled using results of available DSS projects (e.g. EUROBRIDGE, CELEBRATION, POLONAISE, BABEL, Sovietsk – Kochtla-Jarve, etc.) carried out around Poland and Lithuania. We

calculated the crustal TT corrections using the following equation:

$$TT_{model} - TT_{iasp} = TT_{diff} \tag{2}$$

where TT_{model} is TT through the crustal velocity models by Majdański (2012) or by M.
Budraitis (unpublished), TT_{iasp} is TT through the IASP91 velocity model, and TT_{diff} is
TT difference. Although the crustal TT corrections for individual seismic stations do not
take into account the bending of the seismic rays in the crust, the result is reliable as
the rays hit the surface almost vertically and the crust is thin compared to the entire
velocity structure.

4.3 Model parameterization

In the teleseismic tomography inversion as an input model we used the IASP91 velocity
model (Kennett and Engdahl, 1991) and transformed it into the 3-D velocity model with
16 layers of different thicknesses down to 700 km. As the resolution of the inversion
is governed by spacing between seismic stations, frequency content of the seismic
signals and seismic ray geometry, we used spacing of 50 km between the nodes of the
model grid in horizontal directions (Fig. 4). We performed a number of inversions with
different values of smoothing and damping in order to assess the optimal parameters of
the inversion. After careful analysis we found that the same value as spacing between
the grid nodes in horizontal directions (i.e. 50) is applicable for the diagonal elements
of the smoothing matrix, while the damping value was determined investigating the
trade-off curve between the data variance and model variance (Fig. 9).

The inversions with both the synthetic and the real datasets were performed using
the crustal TT corrections and the defined optimal parameters of smoothing and damp-
ing for the layers between 60 and 350 km.

5 Resolution

The resolution assessment includes calculation of spatial resolution and standard deviations of the model parameters and helps to evaluate the precision of inversion results. In our study we used the hit matrix method to assess the resolution, and the synthetic checkerboard test with the real station configuration in order to indicate the parts of the study area which can be reasonably resolved. The hit matrix is based on a calculation of the number of rays which transverse a particular cell (Fig. 10).

The synthetic checkerboard velocity model contains blocks of 200 km wide in the horizontal directions and four layers thick with ±4 % velocity difference compared to the IASP91 velocity model (Fig. 11a). The inversion results show that the synthetic velocity structure is fairly well resolved in the areas with good station coverage (Fig. 11b).

6 Synthetic "geologiocal" model

We compiled a synthetic "geological" 3-D velocity model using as a base the velocity model by Wilde-Piórko et al. (2010), but we modified both thicknesses of different layers (because of different model grid) and some values of the seismic velocities, because some values seemed to be too high or too low according to other studies (e.g. Griffin et al., 2003). In our synthetic model (Fig. 12a) we introduced the seismic P wave velocities 2 to 6 % higher compared to the IASP91 velocity model at different depths beneath the craton. In the TESZ area we introduced the shape of a ramp-type of the lithosphere-asthenosphere boundary (LAB) dipping to the NE direction with seismic velocity values close to those of the cratonic part but up to 2 % smaller in the upper layers down to about 180 km. At the depths between 270 and 350 km we introduced velocities 2 to 4 % lower compared to the IASP91 velocity model, and the higher velocity area (about 4 % higher P wave velocities compared to the IASP91 velocity model) in the NE part, which implies that we expect the deeper cratonic roots in this part of the study area.

The results obtained with the synthetic dataset (Fig. 12b) show reasonably resolved the ramp-type shape of the LAB and the higher velocity area at the lower inverted layers. The lower velocity area in the results in the cratonic part (in the middle of the study area) down to about 120 km is an artificial feature (Fig. 12b).

7 Results and discussion

We resolved structure of the upper mantle from 60 km down to 350 km in our study area (Figs. 12c and 13).

7.1 Lithosphere structure

In general, beneath the EEC in the study area (Fig. 12c) we obtain 5 to 6.5 % higher seismic velocities compared to the IASP91 velocity model. The higher velocities in the upper mantle can be traced going down to the depth of about 180 km beneath NE Poland which coincides with the results of Wilde-Piórko et al. (2010) and Majorowicz et al. (2003). Going further to the NE the lithosphere thickness increases and beneath Lithuania it is at least 300 km or more. Thick lithosphere was previously reported for other cratonic areas, i.e. the Fennoscandian Shield (Sandoval et al., 2004), but there was found no evidence of the seismic LAB anywhere within the depth of 300 km beneath the Fennoscandian Shield (Bruneton et al., 2004). The shear wave studies of Legendre et al. (2012) showed no deep cratonic roots below about 330 km in the EEC. There is a good correlation between our results obtained with the real dataset and the synthetic dataset (Fig. 12), which imply that the lithosphere thickness may increase going from the TESZ towards the NE and could be larger than 300 km in the EEC. Morever, the results with the synthetic dataset show that the P wave velocities beneath the craton down to 180 km could be at least 5 % higher compared to the IASP91 velocity model.

In the EEC down to 90 km we observe the lower seismic velocities beneath western Lithuania (i.e. the WLG) (Fig. 13) which could be related with an upper mantle dome. Motuza et al. (2000) proposed that the mantle dome could be related with delamination processes, because beneath the WLG the heat flow which is significantly higher com-
pared to the adjacent areas was observed (Kepezinskas et al., 1996; Rastiene et al., 1998) and the high density reflectors in the upper mantle have been found (Giese, 1998; Motuza et al., 2000). These high density bodies can potentially represent delaminated slices of the crust which sank into the mantle (e.g. Defant and Kepezhinskas, 2002). In our results (Figs. 12 and 13) we do not find any well defined high velocity
reflector in the upper mantle, on the other hand, below the discussed low velocity area (i.e. the proposed upper mantle dome) we observe area of velocities which are significantly higher than those of the surroundings. As the delamination processes occur locally, the lower and the higher velocity areas observed in our results beneath the WLG could be possibly related to the local upper mantle dome and the delaminated
high density rocks.

Beneath the TESZ we find about 4 % smaller seismic velocities compared to the IASP91 velocity model, except for the northern TESZ (northern Poland) where we observe the seismic velocity values close to those of the craton down to about 150 km. Knapmeyer-Endrun et al. (2013) also found a high velocity anomaly beneath the north-
ern TESZ and proposed that it could have been formed either as a low-temperature remnant of former subduction which penetrated into the mantle transition zone (MTZ), or this anomaly is due to increased water content. As we find no evidence of the anomaly extending to the MTZ, we assume that it may be due to higher water content. In the northern part of the TESZ we find the seismic LAB at a depth of about
180 km. We also indicate the seismic LAB of a ramp-type dipping to the NE direction, which coincides with the inversion results obtained with the synthetic dataset (Fig. 12b). Hansen and Balling (2004) also reported on a number of mantle reflectors beneath the Baltic Sea along the TESZ dipping to the N–NE direction.

The velocity model by Wilde-Piórko et al. (2010) proposed the higher *P* wave velocity values compared to the IASP91 velocity model for the TESZ and the cratonic area for depths more than 250 km. Our results with the real and the synthetic datasets (Fig. 12) indicate that the seismic velocities at these depths could be possibly 1 to 3 % smaller
5 compared to the IASP91 velocity model.

7.2 Traces of the crustal units

The crustal units are not well resolved in our results. There are no clear indications of the structures (Fig. 2) in the upper mantle (the uppermost inverted layers of the velocity model) which could be related with the crustal units in the study area (Figs. 12c, 13). We
10 may infer only one possibly resolved boundary between the EL and the WLG beneath Lithuania which could be related with the local lower velocity areas at the depths from more than 100 km to about 150 km. This area was interpreted by Motuza (2005) as a palaeosubduction zone. The other possible explanation for the lower velocity area beneath the southernmost Lithuania – NE Poland at a depth of 100–150 km is an effect
15 due to an anorogenic granitoid massif, the Mazury Complex (Fig. 7), which is 40 km wide and 6.5 km thick extending 200 km from the Baltic Sea through the Kaliningrad District of Russia into NE Poland. A number of studies (e.g. Bruneton et al., 2004; Beller et al., 2013) showed that the upper mantle beneath anorogenic granitoid massifs inside cratonic crust is different from that of the surrounding cratonic mantle. There is
20 another anorogenic granitoid massif, the Riga Pluton (Fig. 7), in western Latvia which in our results could be related to the lower velocity area down to about 150 km in the NE part of our study area. As the granitoid massif lies on the edge of the study area where resolution is quite poor, we cannot assert its effects on our results. Both the Mazury Complex and the Riga Pluton are of Rapakivi-type and has formed 1.6–1.5 Ga
25 (Rämö et al., 1996; Dörr et al., 2002).

8   Conclusions

– Beneath the EEC we obtain 5 to 6.5 % higher seismic velocities compared to the IASP91 velocity model. The lithosphere thickness increases towards the NE from about 180 km beneath NE Poland to at least 300 km or more beneath Lithuania.

– Beneath the TESZ we find the seismic velocities about 4 % smaller compared to the IASP91 velocity model, and only in the northern TESZ we observe higher seismic velocities down to about 150 km, which show that northern part of the TESZ is craton-like.

– The seismic LAB beneath the northern part of the TESZ is at a depth of about 180 km, and most likely it is of a shape of a ramp. We did not find the seismic LAB beneath the EEC.

– The seismic velocities in our study area at the depths more than 250 km could be 1 to 3 % smaller compared to the IASP91 velocity model.

– We possibly observe the upper mantle dome beneath western Lithuania.

– In our results we did not find strong correlation between the crustal units and the upper mantle, on the other hand, we may have possibly resolved the trace of boundary between the EL and the WLG beneath Lithuania.

– Possibly in our results we may have identified the Riga and the Mazury anorogenic granitoid plutons.

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Table 1. Dataset compiled during the manual picking procedure of the *P* wave arrivals.

Weighting factor	Time error	Number of picks
1	< 0.2 s	2808
2	0.2–0.3 s	958
3	0.3–0.4 s	429
In total:		4195

Fig. 1. Tectonic settings of the EEC (after Artemieva et al., 2006).

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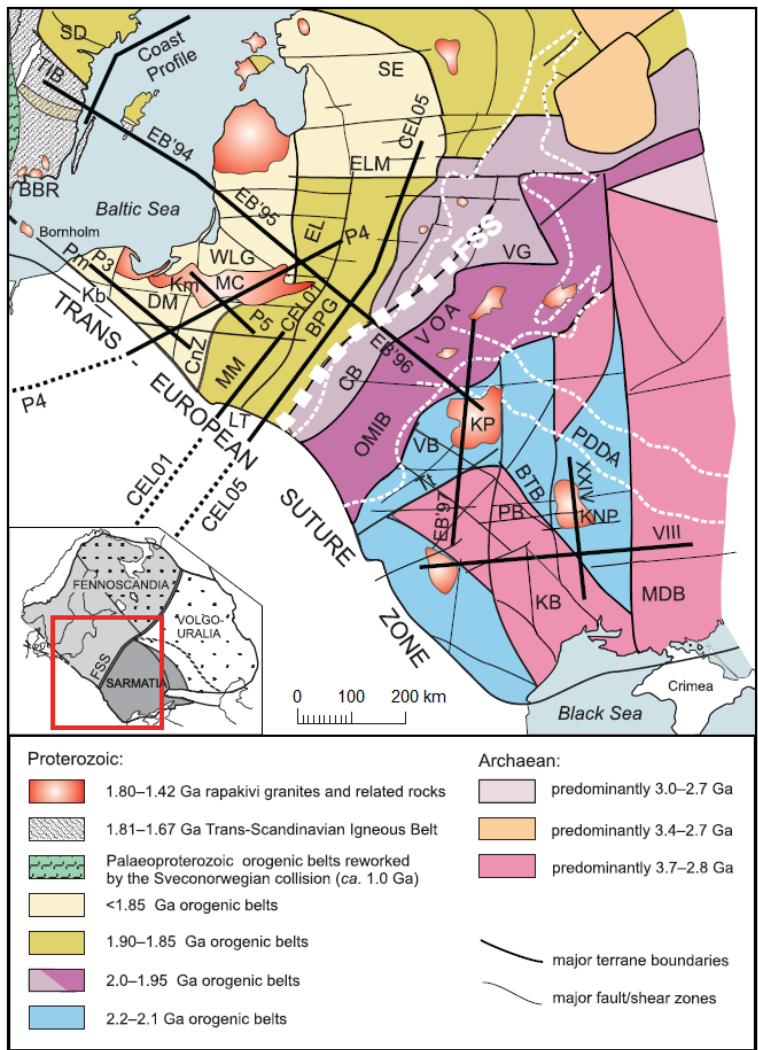
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Fig. 2. Simplified tectonic map (after Bogdanova et al., 2001) of the SW margin of the EEC and locations of refraction and wide-angle reflection deep seismic sounding (DSS) profiles. Solid straight lines – DSS profiles EUROBRIDGE (EB'95, EB'96 and EB'97), POLONAISE'97 (northern part of P4, P3 and P5), VIII and XXIV profiles; dashed lines – parts of profiles in the TESZ and the Carpathians; and white dashed lines show boundaries of aulacogens. Units: BBR, Blekinge–Bornholm region; BPG, Belarus–Podlasie Granulite Belt; BTB, Belaya–Tserkov Belt; CB, Central Belarus Belt; CnZ, Ciechanów Zone; DM, Dobrzyň Massif; EL, East Lithuanian Domain; ELM, East Latvian Massif; FSS, Fennoscandia–Sarmatia Suture; KB, Kirovograd Block; Kb, Kaszuby Block; Km, Kêtrzyn Massif; KNP, Korsun–Novomirgorod Pluton; KP, Korosten Pluton; LT, Lublin Trough; MDB, Middle Dnieper Block; MM, Mazowsze Massif; MC, Mazury Complex; OMIB, Osnitsk–Mikashevichi Igneous Belt; PB, Podolian Block; Pm, Pomorze Massif; PDDA, Pripyat–Dnieper–Donets Aulacogen; SD, Svecofennian Domain; SE, South Estonian Granulites; TIB, Trans-Scandinavian Igneous Belt; Tt, Teterev Belt; VB, Volyn Block; VG, Vitebsk Granulite Domain; VOA, Volyn–Orsha Aulacogen; WLG, West Lithuanian Granulite Domain.
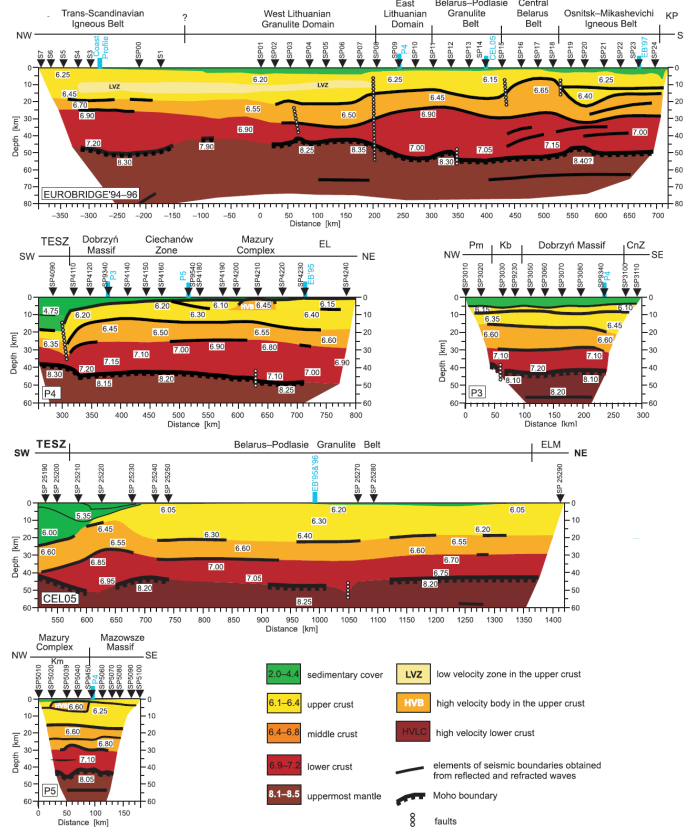
Fig. 3. Models of the crust and the uppermost mantle along the EUROBRIDGE transect (EB'94 and EB'96), the POLONAISE'97 profiles P4 (northern part), P5 and P3, and CELEBRATION 2000 profile CEL05 (after Grad et al., 2006). Values of the *P* wave velocities are given in km s^{-1}. Arrows indicate positions of the shot points; the crossing points with other profiles are marked in blue. For other explanations see Fig. 2.

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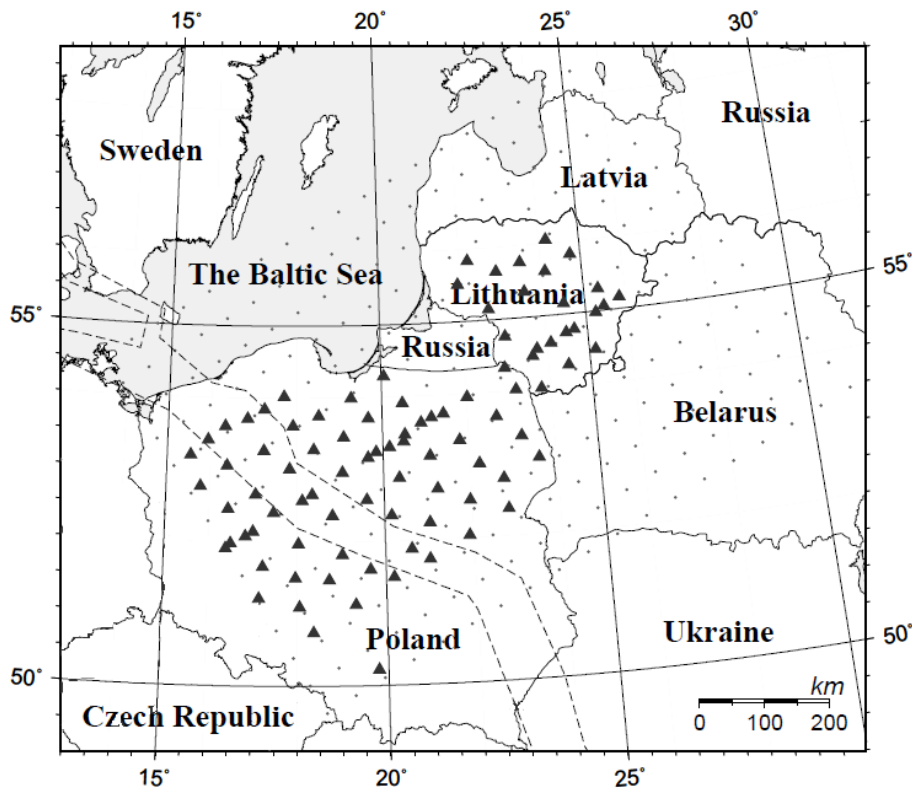
Fig. 4. Map of the seismic stations (triangles) used in the study, and locations of nodes of the model grid (dots). The area in between the dashed lines indicates the Teisseyre–Tornquist Zone (TTZ).
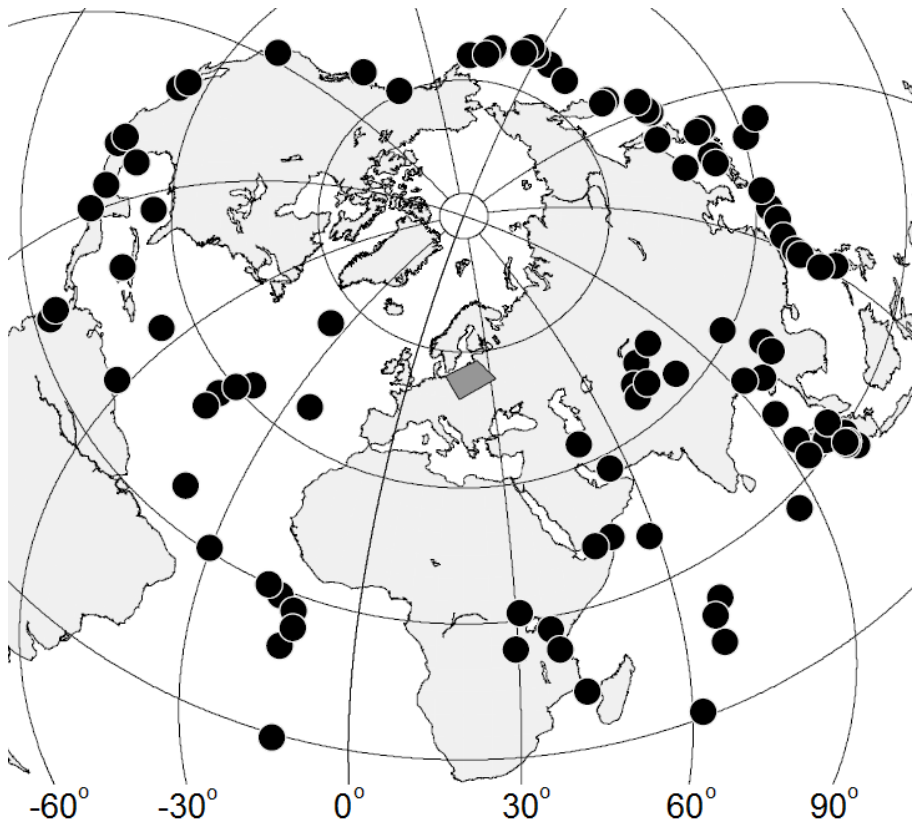
Fig. 5. Map of the epicenters of 101 EQs used in teleseismic tomography inversion. Grey rectangle indicates the study area.
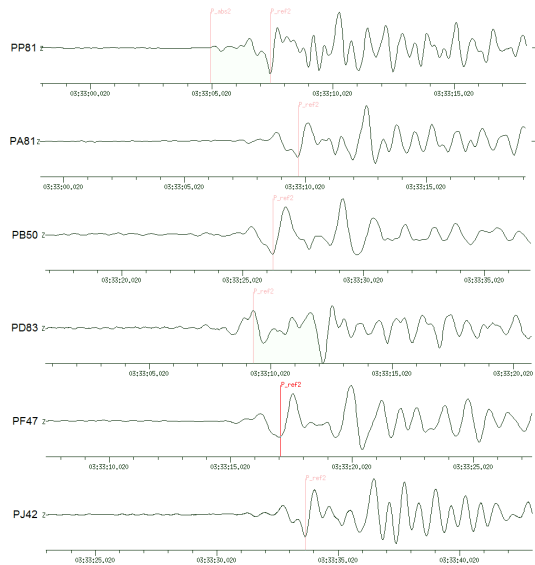
Fig. 6. Example of manual picking of the *P* wave arrivals. Filtered seismograms of an EQ on 02.08.2007 at 3:21 UTC. From all seismograms of one-event file we picked the best trace (the reference station) with relatively low signal-to-noise ratio (SNR) and picked the absolute *P* wave arrival (P_abs) (the onset of the *P* wave) and the relative *P* wave arrival (P_ref) of some well-expressed minimum or maximum of the seismic signal on the same trace (i.e. station PP81). Then we compared the waveform of the reference seismogram with the waveforms of other seismograms, and picked the relative *P* wave arrivals there. For some EQs we observed more than one type of the waveform, thus, we grouped the events with similar waveforms and picked absolute and relative *P* wave arrivals for each group separately. Every pick was assigned with a quality factor from 1 (best quality) to 3 (poor quality). In the picking procedure with SHM the purple picks (i.e. stations PP81, PA81, PB50, PD83 and PJ42) indicate *P* wave arrivals of quality factor 1, while the red ones (i.e. station PF47) indicate *P* wave arrivals of lower quality (i.e. either 2 or 3). In the data of some stations we indicated inversed polarities (i.e. station PD83).
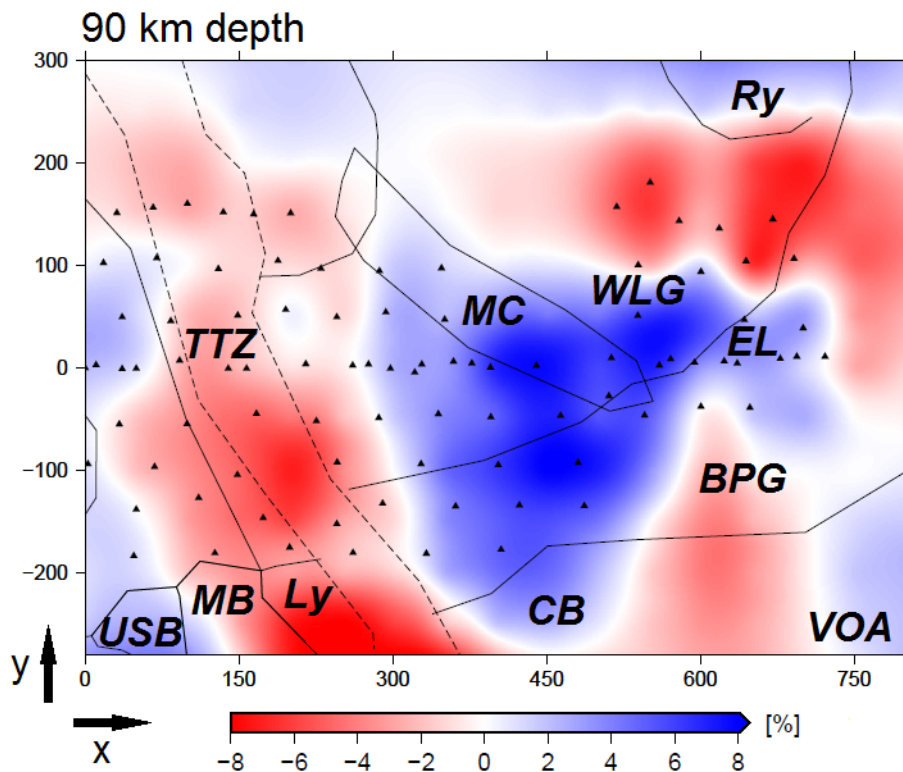
Fig. 7. *P* wave velocity variations obtained during the teleseismic tomography inversion with the real dataset without crustal TT corrections. The lines indicate tectonic units: BPG, Belarus–Podlasie Granulite Belt; CB, Central Belarus Belt; EL, East Lithuanian Domain Ly, Lysogory; MB, Malopolska Block; MC, Mazury Complex; Ry, Riga granitoid pluton; TTZ, Teyssere – Tornquist Zone; USB, Upper Silesian Coal Basin; VOA, Volyn – Orsha Aulacogen; WLG, West Lithuanian Granulite Domain. The BPG, the EL and the WLG form the Baltic-Belarus Belts.
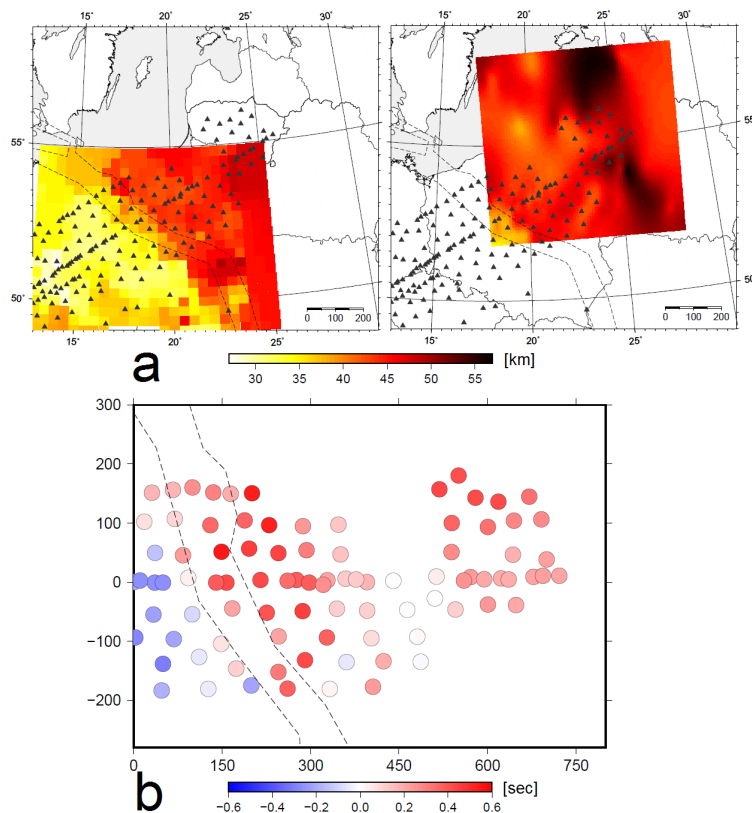
Fig. 8. (a) Moho maps compiled by Majdański (2012) (left) and M. Budraitis (unpublished) (right) used to estimate the crustal TT corrections. The Moho depth in the depicted area vary from 27 to 57 km. **(b)** Estimated crustal TT corrections for individual seismic stations. Values are expressed in seconds relative to the IASP91 velocity model.

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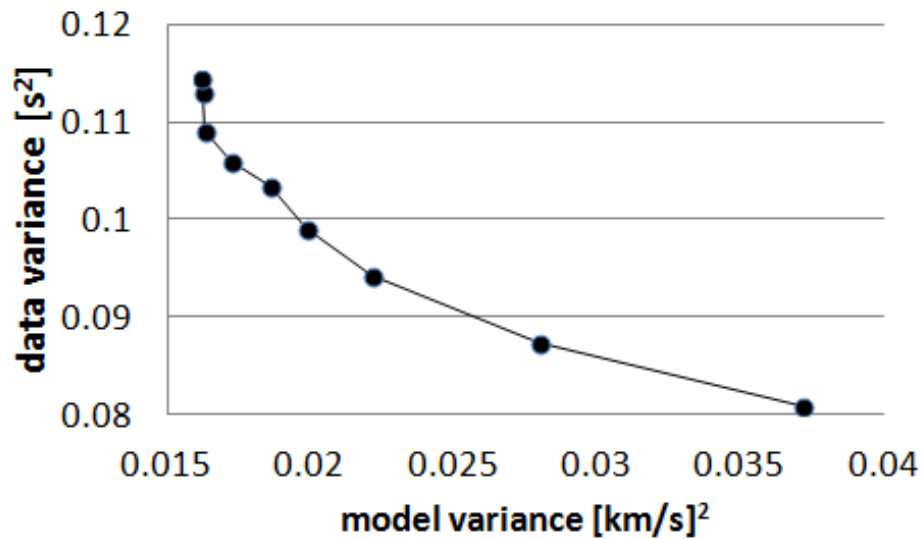
Fig. 9. Data variance vs. model variance obtained during inversions with damping values from 10 to 360.

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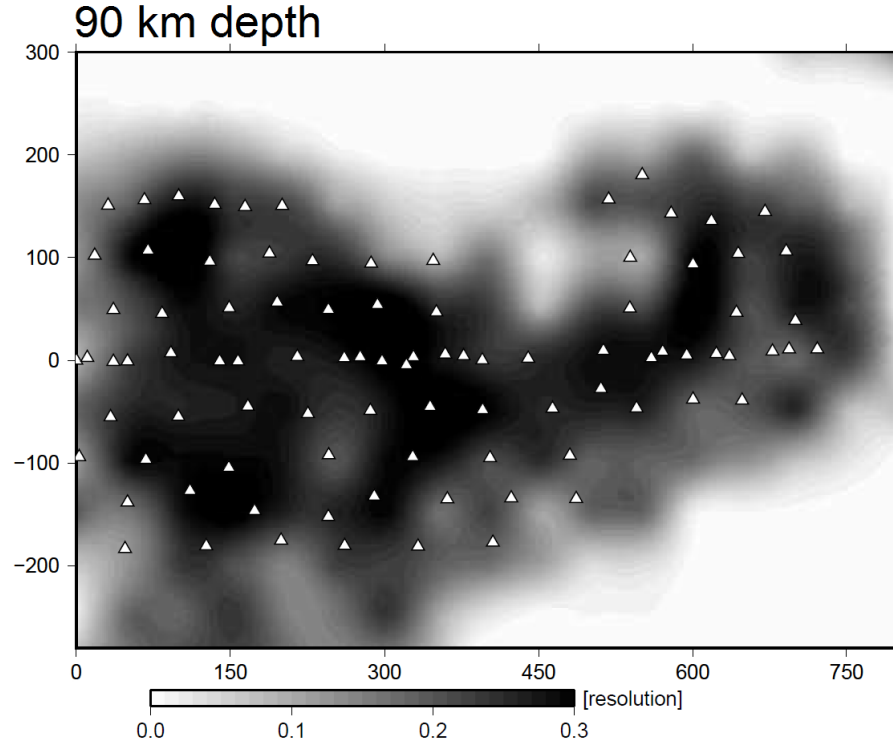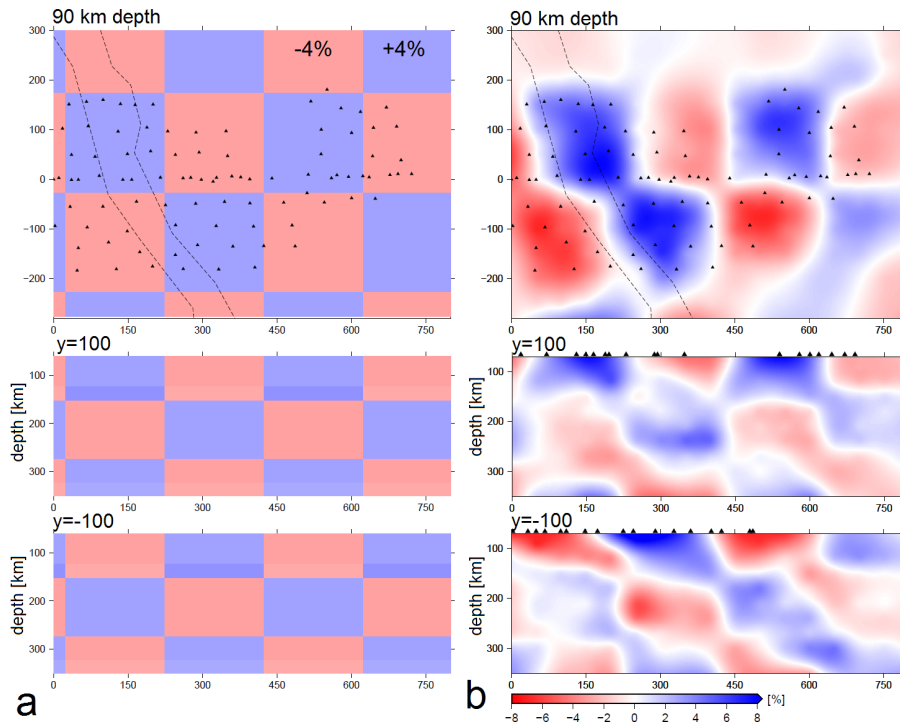Printer-friendly Version

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90 km depth

Fig. 10. The hit matrix of the horizontal slice at a depth of 90 km obtained with the real dataset and station configuration. White triangles mark the seismic stations. The scale shows relative amount of the number of rays which transverse a particular cell.

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Fig. 11. Results of the synthetic checkerboard test. Horizontal slice at a depth of 90 km and two vertical slices parallel to the main PASSEQ transect of the target area. **(a)** Initial velocity model with synthetic blocks of 200 km wide in the horizontal directions and ±4 % P wave velocity difference compared to the IASP91 velocity model. **(b)** Inversion results with the synthetic dataset. Dashed lines indicate the TTZ. Triangles indicate the seismic stations, and on the vertical slices they indicate seismic stations ±50 km around the depicted transects.
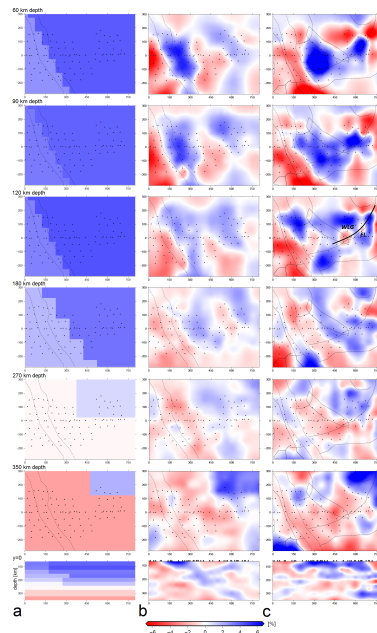
Fig. 12. The initial synthetic "geological" velocity model **(a)**, the inversion results with the synthetic dataset **(b)**, and the inversion results with the real dataset **(c)**. The *P* wave velocity perturbations on the horizontal slices at different depths and the vertical slice along the main PASSEQ transect. The bluish and reddish areas show correspondingly the higher and the lower *P* wave velocities compared to the IASP91 velocity model. The thick line at the depth slice of 120 km indicates possibly resolved boundary between the EL and the WLG. The thin dashed lines on the horizontal slices indicate the TTZ. Triangles indicate the seismic stations, and on the vertical slices they indicate seismic stations ±50 km around the main PASSEQ transect. The solid thin lines on the horizontal slices (right side) indicate boundaries of different tectonic units (for detailed explanation see Fig. 7.).
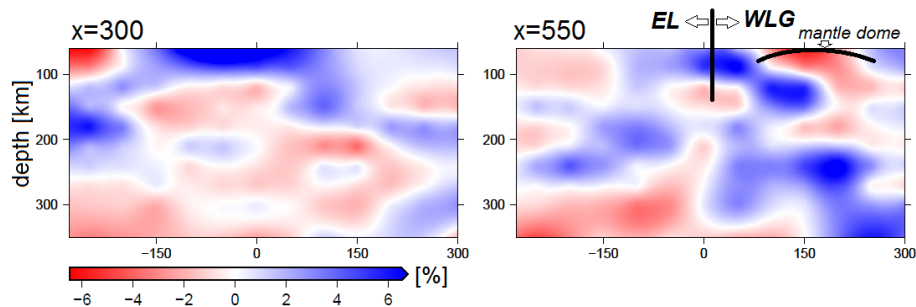
Fig. 13. Vertical slices perpendicular to the main PASSEQ transect close to the eastern edge of the TESZ (left), and about 350 km to the NE from the TESZ (right). The thick lines indicate possibly resolved boundary between the EL and the WLG, and the mantle dome beneath the WLG.