



Three-dimensional S-wave velocity model of the Bohemian Massif from Bayesian ambient noise tomography

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ABSTRACT

We perform two-step surface wave tomography of phase-velocity dispersion curves obtained by ambient noise cross-correlations in the Bohemian Massif. In the first step, the inter-station dispersion curves were inverted for each period (ranging between 4 and 20 s) separately into phase-velocity maps using 2D adjoint method. In the second step, we perform Bayesian inversion of the set of the phase-velocity maps into an S-wave velocity model. To sample the posterior probability density function, the parallel tempering algorithm is employed providing over 1 million models. From the model samples, not only mean model but also its uncertainty is determined to appraise the reliable features. The model is correlated with known main geologic structures of the Bohemian Massif. The uppermost low-velocity anomalies are in agreement with thick sedimentary basins. In deeper parts (4–20 km), the S-wave velocity anomalies correspond, in general, to main tectonic domains of the Bohemian Massif. The exception is a stable low-velocity body in the middle of the high-velocity Moldanubian domain and high-velocity body resembling a promontory of the Moldanubian into the Teplá-Barrandian domain. The most pronounced (high-velocity) anomaly is located beneath the Eger Rift that is a part of a Tertiary rift system across Europe.

1. Introduction

Earth's ambient seismic noise, generated mainly by processes in oceans and atmosphere, has been present in seismic records to distress and vex seismologists for many years. It was only recently that the recordings of the ambient noise were recognized to be useful: by cross-correlating long series of noise recordings between two stations the Green's function between them may be obtained (Campillo and Paul, 2003; Shapiro and Campillo, 2004). The Greens' functions represent response in a given location to an impulsive source in another point and thus contain purely information about the seismic properties of the media between the two points. Therefore, it is natural to use them in tomography.

In the first ambient noise tomography applications, the results were presented by a set of 2D group (and later phase) velocity dispersion maps obtained by the inversion of traveltimes between station pairs for each period separately. The dispersion maps for various regions were estimated: US – Shapiro et al. (2005) and Lin et al. (2008), New Zealand – Lin et al. (2007), Korean peninsula – Cho et al. (2007), Europe – Yang et al. (2007), etc. They show good correlations with known geologic structures and may be used for preliminary interpretations.

The phase/group velocity dispersion maps at a set of periods can be also translated into a 3D velocity model. This has been done on various scales, from global (Haned et al., 2015; Nishida et al., 2009) to regional (Badal et al., 2013; Li et al., 2010; Pang et al., 2016), up to local networks, for example, around volcanos (Matos et al., 2015; Obermann et al., 2016; Ryberg et al., 2016; Spica et al., 2015). The ambient noise traveltime dataset may be also combined with another dataset, for instance, with teleseismic traveltimes to increase the resolution at greater depths (e.g., Yang et al., 2008; Ouyang et al., 2014; Guo et al., 2016; Rawlinson et al., 2016), or with receiver functions beneath the stations to improve sensitivity to interfaces in the velocity structure (e.g., Bodin et al., 2012; Guo et al., 2015; Shen et al., 2012; Růžek et al., 2012). The ambient noise tomography has proved useful also for broader applications, such as tomography of ocean ridges using a network of ocean-bottom seismometers (Mordret et al., 2014; Zha et al., 2014) or assessment of seismic anisotropy (Guo et al., 2012; Shirzad and Shomali, 2014).

Inversion of surface wave dispersion data into a 3D velocity model traditionally employs a two-step approach. In the first step, the period-dependent traveltimes based on dispersion curves measured between two points are inverted into phase/group velocity dispersion maps. In

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the second step, the dispersion curves on a regular grid are extracted and inverted for each grid point independently into a set of 1D velocity models. These 1D models are then assembled into a final 3D velocity model. In this second step, various techniques are employed. Standard deterministic approach is based on iterative linearized least square inversion (e.g. Li et al., 2010; Luo et al., 2012; Matos et al., 2015; Porritt et al., 2016). However, stochastic approaches based on Monte Carlo (MC) methods are becoming more popular. The great advantage of the MC methods is that the result is not represented by a single model, but by a rather large set of models sampling the posterior probability density function. This is essential in the case of multimodal probability density functions, when the solution is nonunique and should not be represented by one particular model. The MC methods are able to explore several areas of model parameter space where the probability density function attains significant values. One may either extract coherent properties of the models (e.g., by model averaging), or assess uncertainties and correlations between the retrieved model parameters. Moreover, the MC solutions are numerically more stable – one does not have to deal with matrix inversion as in the case of the linearized inversion. Various MC methods have already been employed in ambient noise tomographic applications, for example, simulated annealing (Spica et al., 2014, 2015), neighborhood algorithm (Gao et al., 2011; Mordret et al., 2014) or other MC search algorithms (Guo et al., 2015; Jiang et al., 2016, 2014). In our application, we employ another MC technique called parallel tempering, which was introduced into geophysical problems only recently by Sambridge (2014). The method is well balanced between fast convergence and avoiding entrapment in local maxima of the probability density function.

In the traditional two-step inversion, a 3D model is compiled from separately inverted 1D vertical models defined on a regular horizontal grid. The 1D models may be defined by a large number of velocity layers (Li et al., 2010; Pang et al., 2016; Porritt et al., 2016) but also in a transdimensional way, where the number of model parameters is regarded as a hyperparameter (e.g., Young et al., 2013a,b; Pilia et al., 2015; Galetti et al., 2017). However, we take a different approach: instead of performing 1D inversion for each point of the dispersion map separately, the 3D model is parametrized using fixed layers and the inversion is carried out for all the model parameters simultaneously.

We apply the methodology to the Bohemian Massif – a remnant of the Variscan orogen with a complex structure and history (see Fig. 1a) summarized in Section 2. The input data consist of phase velocity dispersion curves between the station pairs in periods 4–20 s obtained from

ambient-noise cross-correlations – see Section 3. In Section 4, we outline the methods applied in our problem. Then, we present the results in a form of dispersion maps and S-wave velocity model in Section 5. As a result of the limited period range of the dispersion data, the resulting model is bound to recover only the top 25 km of the Bohemian Massif crustal structure. In Section 6, the models are interpreted in terms of known geologic structures and compared with models obtained by other authors. The work represents the first 3D ambient noise tomography of the Bohemian Massif. Moreover, since the inversion is performed in the Bayesian framework, the model uncertainty is estimated as well.

2. Bohemian Massif

2.1. Tectonic setting

The Bohemian Massif (Fig. 1a) is a relic of the European Variscan orogenic belt that formed ~400–300 Myr ago as a result of convergence between Gondwana and Laurussia (Franke, 2000; Matte, 2001). It consists of several major tectonic domains with different history and dominant rock types (for overview see Schulmann et al., 2009). Most of them originally formed a part of the Gondwanan active margin, and separated as continental micro-plates during the Late-Cambrian–Ordovician times. When the motion of the plates changed and the region between Gondwana and Laurussia was closing, the Saxothuringian domain was a part of the subducting plate (e.g., Franke, 2000). It recorded significant deformation and metamorphism with intensity increasing towards its south-eastern boundary where the oceanic suture was located. The most prominent relic of the suture is the Mariánské Lázně Complex that contains mafic rocks buried along a cold geotherm in a sequence typical for closure of an oceanic domain (Beard et al., 1995).

The Teplá-Barrandian domain was a part of the upper plate during the orogeny and preserved the pre-Variscan upper crust including sedimentary sequences affected only by low-grade deformation and metamorphism (Drost et al., 2004). The south-eastern margin of the Teplá-Barrandian domain was intruded by magmatic rocks together forming the Central Bohemian Plutonic Complex. The composition of these rocks corresponds to melting of mantle variably enriched by crustal component, which is typical for a magmatic arc above a subduction zone (Janoušek et al., 2000).

The Moldanubian domain was a continuation of the Teplá-

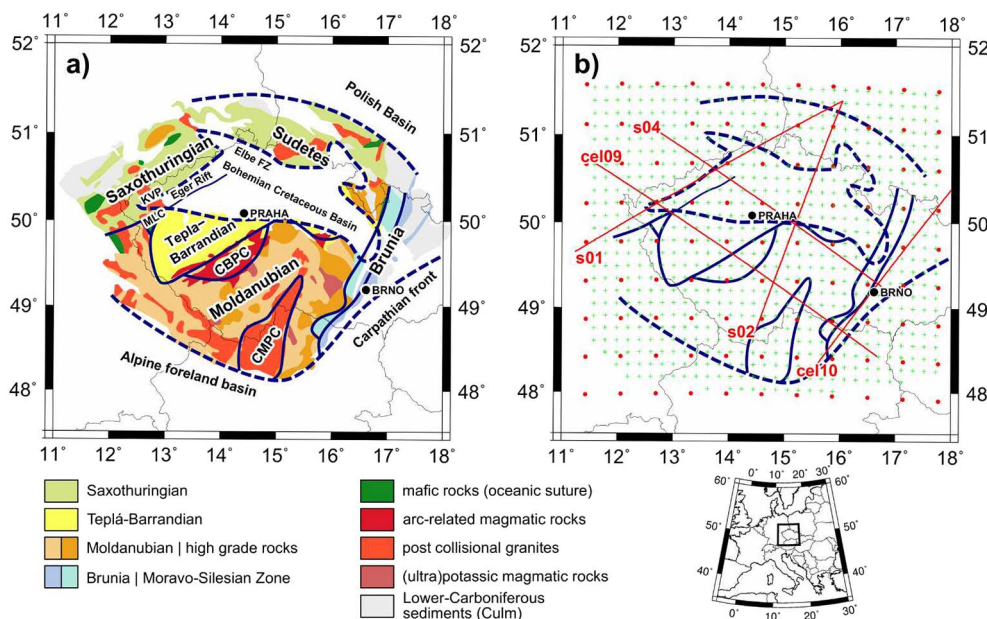


Fig. 1. Map of the studied region: a) Main geologic structures of the Bohemian Massif modified after Franke (2000) and Schulmann et al. (2014). MLC – Mariánské Lázně Complex, CBPC – Central Bohemian Plutonic Complex, CMPC – Central Moldanubian Plutonic Complex, KVP – Karlovy Vary Pluton, FZ – fault zone. b) Distribution of points where the models are defined: Green pluses – data points where phase velocity dispersion maps are determined by 2D adjoint localization (first step). Red dots – model control points from which 3D velocity model is interpolated in the 3D inversion (second step). Selected seismic profiles measured during seismic experiments CELEBRATION 2000 (CEL09, CEL10) and SUDETES 2003 (S01, S02 and S04) across the studied region, are shown by red lines. Blue outlines show main geologic structures (solid – tectonic domains, dashed – post-Variscan sedimentary cover). Black thin outlines show state borders. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

Barrandian domain during the orogeny, but it was affected by medium-to-high grade metamorphic conditions (Schulmann et al., 2009). The contrasting character of the two adjacent domains results from their respective vertical displacement at a shear zone that developed in the weakened magmatic arc region (the Central Bohemian Plutonic Complex, Dörr and Zulauf, 2010). Erosion of the elevated Moldanubian surface then led to exposure of its middle and lower crust. The central part of the Moldanubian domain also contains a large accumulation of plutonic bodies that formed by crustal melting at the late stage of the orogeny – the Central Moldanubian Plutonic Complex (Finger et al., 2009).

The easternmost part of the Massif is the Brunia (or Brunovistulian) domain. Along the margin with the Moldanubian domain, the Brunia-derived rocks were strongly deformed and metamorphosed within the so-called Moravo-Silesian Zone as a result of thrusting of Brunia underneath the Moldanubian domain (see references in Schulmann et al., 2009). The interpretation of the Bouguer gravity anomaly suggests that the Brunia basement continues 50–70 km west of its surface margin and is covered only by a thin layer of Moldanubian rocks (Guy et al., 2011). In addition, westward continuation of Brunia on the level of the mantle lithosphere was inferred from the seismic anisotropy (Babuška and Plomerová, 2013). The Brunia basement is covered by Variscan as well as younger sedimentary sequences (Kalvoda et al., 2008).

During the collapse of the Variscan orogen, its elevated topography was subject to extension and gradual erosion which resulted in formation of several Permo–Carboniferous sedimentary basins. Later on in the Permian, subsidence of the Polish basin started eventually leading to formation of a several kilometers thick sedimentary cover north of the Bohemian Massif.

At ~70 Myr ago, the onset of the Alpine orogeny led to activation of the Elbe Fault Zone, subsidence of the surrounding area and formation of the Bohemian Cretaceous Basin (Uličný et al., 2009). Its today's thickness reaches 1 km and it covers a significant part of the Teplá-Barrandian domain. North-east of the Elbe Fault Zone, the Sudetes domain is located showing similar characteristics as the Saxothuringian domain.

The Alpine orogeny further reactivated existing fractured zones across Europe by rifting and related volcanism and sedimentation (Dèzes et al., 2004). In the Bohemian Massif the Eger Rift was activated along the Saxothuringian–Teplá-Barrandian boundary showing recent regular earthquake swarms, CO₂ emanations and elevated surface heat flow (Čermák, 1994; Cloetingh et al., 2010). Various seismic studies pointed to anomalous character of the crust and to asthenospheric updoming beneath the Eger Rift (e.g., Heuer et al., 2006; Geissler et al., 2005; Plomerová et al., 2007; Hrubcová and Šroda, 2015). The Alpine orogeny also shaped the southern margin of the Bohemian Massif that is now covered by Alpine-Carpathian foreland basins and overthrust by the Western Carpathians in the east.

2.2. Previous tomographic studies

Due to its complex history and structure, the region of the Bohemian Massif has been subject to many seismologic studies starting from the 80s (see Novotný and Urban, 1988). Most of the recent crustal studies use active experiments along profiles, such as CELEBRATION 2000 and SUDETES 2003. The applied tomographic methods result in 2D vertical mainly P-wave velocity models. The crustal models along the active experiment profiles (see also Fig. 1b) were determined: CEL09 crossing most of the tectonic domains (Hrubcová et al., 2005; Novotný, 2012), CEL10 along the Moravo-Silesian Zone (Hrubcová et al., 2008), S01 along the Eger Rift (Grad et al., 2008; Novotný et al., 2009), S04 almost parallel with CEL09 (Hrubcová et al., 2010). Models along several profiles concentrating on the Bohemian Massif were acquired by Růžek et al. (2007) and in the Sudetes by Majdański et al. (2006). The models show some common basic characteristics of the Bohemian Massif: firstly, there is very thin or no sedimentary coverage on the Bohemian

Massif in the seismic data. Secondly, one can divide the crust into upper and lower one with P-wave velocity contrast of ~0.5 km/s in almost all tectonic domains. Usually, both upper and lower crust is strongly homogenized with low vertical gradient. In the subsurface parts, the high-velocity bodies are correlated with mafic intrusions whereas the low-velocity anomalies with granitic structures. In the upper crust, the distinct anomalies were found in the area of the Eger Rift. The Moldanubian domain shows higher P-wave velocities compared to the other domains representing the high-grade rocks. The Moho beneath the Bohemian Massif reaches to depths 30–40 km.

Thanks to numerous experimental data, tomographic inversion into 3D models was performed as well, for example, focused on the Moravo-Silesian region by Růžek et al. (2011), or over larger regions: P-wave velocity model for the Sudetes region by Majdański et al. (2007) or S-wave velocity model of the Bohemian Massif and Alps by Behm (2009). These models show good correlation of the subsurface structures with known geology. 3D P- and S-wave local velocity model of West Bohemia using not only experimental but also local earthquake data was acquired by Růžek and Horálek (2013). In this model, the areas with low Poisson ratio correlate with focal zones of the West-Bohemia earthquakes. Several crustal models were compiled together into a 3D model using different interpolation techniques by Karousová et al. (2012). The map of the Moho depth extracted from this model shows significant crustal thickening in the southern part of the Moldanubian domain and in the Brunia domain.

The crustal domains of the Bohemian Massif were found to continue also in the mantle lithosphere where they are characterized by different P-wave anisotropy patterns (Babuška and Plomerová, 2013; Plomerová et al., 2005).

1D S-wave velocity models beneath the stations in the Bohemian Massif are provided by receiver function studies (e.g., Wilde-Piörko et al., 2005; Geissler et al., 2012; Heuer et al., 2006). Receiver function studies combined with ambient noise cross-correlation data focused on the crust of the Bohemian Massif were performed by Růžek et al. (2012). Beneath the stations located in different tectonic domains, the coherent properties in the receiver functions may be found. Except for the Moho deepening under the Moldanubian, the Moho updoming was found not only beneath the Eger Rift, but also in the area of the Bohemian Cretaceous Basin where it was connected to the tectonic activity along the Elbe Fault Zone.

The stepping stone for our work are the ambient noise cross-correlation studies by Růžek et al. (2016) who investigated the S-wave velocity properties for different domains of the Bohemian Massif. The domains show slightly different characteristics, but the variability within each domain is comparable with the differences among them.

In our work, we utilize these ambient noise cross-correlation data to obtain 3D S-wave crustal velocity model of the whole Bohemian Massif.

3. Data

We adopt inter-station dispersion curves estimated independently for all components (transverse T, radial R and vertical Z), which were extracted from ambient noise cross-correlations by Růžek et al. (2016). The stations used in the processing are a) permanent stations belonging either to the Czech Regional Seismological Network (CRSN) or the Virtual European Broadband Seismological Network (VEBSN); b) temporary stations operating within experiments BOHEMA I–III or PASSEQ (Babuška et al., 2005; Plomerová et al., 2003; Wilde-Piörko et al., 2008); c) stations belonging to adjacent regional networks (Saxonian and Bavarian). The total number of stations is 72. The selected stations are equipped with broadband sensors and the noise measurements span over a time period of 12 years. To extract dispersion curves between the stations, the cross-correlations of the preprocessed noise between pairs of stations were calculated. More details on the method and resulting inter-station dispersion curves can be found in Růžek et al. (2016). The resulting inter-station dispersion curves range between periods 4 and

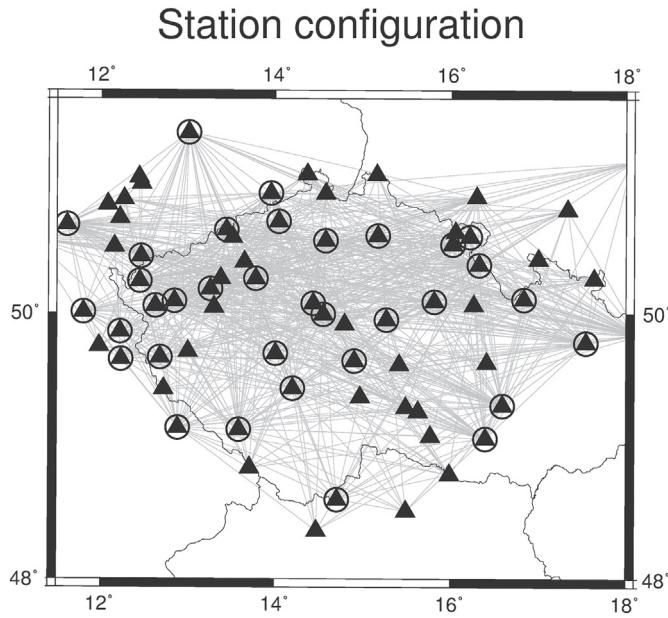


Fig. 2. Station configuration of the investigated area. The station-pairs with measured phase-velocity dispersion curve for the R component are connected by lines. The selected stations (marked by circle) serve as source for the adjoint inversion.

20 s. In the inversion, we employ only phase velocity dispersion curves with high signal-to-noise ratio in the cross-correlation function.

Fig. 2 shows the station and data coverage for the R component. The number of measured traveltimes is almost 700. Note, that for the other two components, the data coverage is even better due to generally higher-quality cross-correlation functions.

4. Methods

The inversion method is derived from the traditional two-step approach described in Section 1. In the first step, we invert the inter-station dispersion curves separately for the selected periods into a 2D regular grid (so-called phase velocity dispersion maps) using 2D adjoint method. For the second step, we perform the inversion of the dispersion maps into a 3D S-wave velocity model in a Bayesian framework.

4.1. Adjoint localization

To obtain phase velocity dispersion maps, we apply 2D adjoint inversion (Fichtner et al., 2006; Tape et al., 2007; Tromp et al., 2005) to inter-station dispersion curves for selected periods. The station coverage as well as the number of dispersion traveltimes in the studied region may be considered sufficient for the tomographic problem (see Fig. 2).

The adjoint method belongs to iterative gradient methods of the misfit minimization. The misfit considered in our problem is the L2 norm of the cross-correlation traveltime differences. The misfit gradient (i.e., the derivative of the misfit with respect to the model parameters) is calculated using one forward and one so-called adjoint calculation. In our problem, the adjoint calculation is represented by backpropagation of so-called adjoint wavefield from the receivers back to the sources (Peter et al., 2007; Tape et al., 2007). The forward and adjoint wavefields are then combined into finite-frequency or sensitivity kernels, which emphasize areas with increased values of the misfit gradient.

The greatest advantage of the adjoint method is that it accounts for the finite-frequency effects of the wave propagation. Nevertheless, the method requires regularization, for example, by means of Gaussian smoothing that is applied to the calculated sensitivity kernels (Peter et al., 2011; Tape et al., 2007, 2010). The width of the Gaussian

smoothing function as well as the optimal number of iteration steps was determined using preliminary synthetic tests with actual data noise (Valentová et al., 2015). More details are to be found in Section 4.3.

4.2. Bayesian inversion

We apply a Bayesian approach to solve the second stage of the inverse problem. The result of the inversion is represented by a set of model samples obtained by a random walk according to the posterior PDF. The method is appropriate for non-linear problems, such as inversion of dispersion curves. Moreover, from the model samples one may estimate not only the best or the mean model, but also its uncertainty.

The posterior PDF is defined as conditional PDF on model parameter space after measurement of data $\mathbf{d}$ is acquired (e.g., Tarantola and Valette, 1982; Mosegaard and Tarantola, 1995; Tarantola, 2005). This is usually expressed using the Bayes theorem:

$$p(\mathbf{m}|\mathbf{d}) = \frac{p(\mathbf{m})p(\mathbf{d}|\mathbf{m})}{p(\mathbf{d})}, \quad (1)$$

where $p(\mathbf{m})$ is the model parameter PDF which is independent on the data measurement (i.e., prior PDF, denoted p_{prior}). Conditional probability of observing data given model $\mathbf{m}$, $p(\mathbf{d}|\mathbf{m})$, is so-called likelihood function. The likelihood function for measured data $\mathbf{d} = \mathbf{d}_{\text{obs}}$ contains statistical information on the data measurement error, but may also include modeling error. However, the information on the modeling error is usually difficult to estimate or negligible compared to the measurement error, and is therefore not assumed. The posterior PDF as a function of model parameters $\mathbf{m}$ for measured data $\mathbf{d} = \mathbf{d}_{\text{obs}}$ can be rewritten as

$$p(\mathbf{m}) = k p_{\text{prior}}(\mathbf{m})p(\mathbf{d}_{\text{obs}}|\mathbf{m}), \quad (2)$$

where k is a PDF normalization constant.

Usually, the data PDF is considered in a form of Gaussian distribution, then

$$p(\mathbf{d}_{\text{obs}}|\mathbf{m}) \propto \exp(-S(\mathbf{m})), \quad (3)$$

where $S(\mathbf{m})$ defines misfit between measured data $\mathbf{d}_{\text{obs}}$ and synthetics calculated using a theoretical relation $\mathbf{g}(\mathbf{m})$ with Gaussian covariance matrix $\mathbf{C}_d$:

$$S(\mathbf{m}) = (\mathbf{d}_{\text{obs}} - \mathbf{g}(\mathbf{m}))^T \mathbf{C}_d^{-1} (\mathbf{d}_{\text{obs}} - \mathbf{g}(\mathbf{m})). \quad (4)$$

To draw samples from the model space according to the posterior PDF, the Markov chain MC random walker is employed. In each step of the chain, the model parameters are randomly perturbed considering Gaussian probability function. The new (proposed) model is accepted or rejected based on the Metropolis algorithm: if the posterior PDF of the proposed model is higher, the model is accepted; if the PDF is lower, the model may be still accepted with some probability. The acceptance rate depends on the width of the Gaussian generating the perturbations, i.e., the perturbation of the misfit. To increase the efficiency of the sampler we apply a method called parallel tempering (PT, Sambridge, 2014). The PT algorithm is similar to the better-known simulated annealing, as it introduces modification of the PDF by an additional parameter called temperature T . The modified PDF $p(\mathbf{m}, T)$ is given by

$$p(\mathbf{m}, T) = k p_{\text{prior}}(\mathbf{m}) \exp\left(\frac{-S(\mathbf{m})}{T}\right). \quad (5)$$

The samples are drawn following this modified PDF assuming multiple values of the temperature T .

For high temperatures, the PDF becomes smooth (note that for $T \rightarrow \infty$, $p \rightarrow p_{\text{prior}}$). For low temperatures, the PDF maxima are more pronounced and as $T \rightarrow 0$, PDF converges to δ function located in the global maximum (i.e., global minimum of $S(\mathbf{m})$). For $T = 1$, the modified PDF equals to the original PDF. During the simulated annealing,

temperature T gradually decreases from high values to lower values so that the sampler gradually concentrates into areas with higher PDF values. In the PT method, multiple Markov chains each with different temperature sample the model parameter space simultaneously. The chains with lower temperature values sample locally areas of PDF maxima, whereas chains with higher temperatures are able to escape the local maxima of PDF. Moreover, two chains can exchange their temperature values between the chain advances. The probability that two chains (denoted $[m_i, T_i]$ and $[m_j, T_j]$) swap their temperatures (resulting in $[m_i, T_j]$ and $[m_j, T_i]$) is given by the Metropolis-Hastings criterion. This condition ensures, that the subset of samples for temperature $T = 1$ represents sampling of the original untempered PDF $p(\mathbf{m})$. The PT algorithm is well balanced between efficiency and stability for complex multimodal PDFs (Sambridge, 2014). Another advantage is straightforward parallelization of the sampling computer code.

4.3. Implementation details

In the 2D adjoint localization, the regularization is applied primarily by means of Gaussian smoothing of the sensitivity kernels and by the number of iterations. The regularization parameters are estimated by synthetic tests using a smooth and a complex model to examine reliability of both long-scale and short-scale features. For the 20 s Love data inversion, the width of the Gaussian smoothing function is set to 100 km with iterations stopping at 6th step (for more information and related synthetic tests see Valentová et al., 2015). For other periods (both Love and Rayleigh datasets), the size of the Gaussian function scales with the corresponding wavelength.

The phase velocity dispersion maps are calculated for periods (4, 6, 8, 10, 12, 16, 20) s in the first stage and serve as input data in the second step of the inversion in which the Bayesian approach is applied. We extract dispersion curves in a selected 2D regular grid of data points (green pluses in Fig. 1b). The 3D velocity model is represented by a set of vertical 1D layered models on a regular horizontal grid of model control points (red dots in Fig. 1b). The spacing of the data and model control points is 16 km and 50 km, respectively. In each model control point, we assume 7 layers above a halfspace with interfaces at depths of 2, 4, 8, 12, 18, 24, and 32 km. Although the model is parametrized in 3D, the synthetic dispersion curves are calculated for computational reasons assuming 1D layered model at each data point. In the MC methods, the forward problem is calculated numerous times and thus a fast solver (such as 1D) is necessary. We employ the code VDISP which is based on a matrix method using Thomson-Haskell and Watson's matrices for Love and Rayleigh waves, respectively (Novotný, 1999). To obtain a 1D layered model for the synthetic dispersion curve calculation, the model is interpolated in each layer from the model control points into the data grid points by cubic spline interpolation.

The Bayesian inversion is regularized by the adopted parametrization. Firstly, the number of layers is rather low and with fixed thicknesses; the number of model control points is kept lower than the number of data points. Furthermore, the model is interpolated into the data points using smooth functions (cubic splines).

The Bayesian inversion employs the PT algorithm to sample the posterior PDF. As the model prior information $p_{\text{prior}}(\mathbf{m})$ we use homogeneous PDF on a selected interval for all model parameters, in particular between 1 and 15 km/s. Another constraint on the model parameters is the requirement of a nonnegative velocity gradient with depth for each model control point. For simplicity, the data covariance matrix C_d is assumed to be diagonal with standard deviation corresponding to error of dispersion maps (see Section 5.1).

We run the computations in parallel on 12–24 CPUs, the temperature T ranges between 1 and 50. To remove the dependency of Markov chains on the starting model, we introduce long burn-in phase (10k steps) when the accepted models are not saved. The main (post-burn) phase is stopped when the sampler appears to have converged on the model space (i.e., no improvement in model space sampling occurs with

further thousands of chain steps). Only the samples from chains at temperature $T = 1$ accepted every 100th step are saved and further processed. The resulting number of samples from the posterior PDF exceeds 1 million.

5. Results

In the first step, the inter-station dispersion curves were inverted for each period separately by the 2D adjoint method. The results are represented by a set of phase velocity maps in Section 5.1.

In Section 5.2, we present the results of the Bayesian inversion of the dispersion maps into a 1D S-wave velocity model (and v_p/v_s) using different datasets and assumptions. The 1D Bayesian model of the Bohemian Massif reflects its average structure. Moreover, with the help of synthetic tests we discuss to what extent the variance in the 1D velocity model parameters contains also variability due to the 3D structures.

In Section 5.3, the results of the Bayesian inversion of the dispersion maps into a 3D model is presented as 2D depth-slices for both absolute and relative S-wave velocities, as well as horizontal variations of the v_p/v_s ratio. The correlations of the results with geology are discussed in Section 6.

5.1. Phase velocity maps

Fig. 3 shows the inverted phase velocity dispersion maps for each period as obtained from the three components individually: Love (T) and both Rayleigh (R and Z). The perturbations in phase velocities are shown with respect to average values specified on the right side of each map.

We treat the dispersion curves for R and Z component as two independent datasets, so that by comparing the resulting models we get insight into the accuracy of the first part of the inversion. Regarding the average models, both components give very similar results. Contrarily, the phase-velocity perturbations of the R and Z component, although similar in some features, differ in some details as well.

We use differences between both Rayleigh dispersion maps to estimate the error of the dispersion maps. The RMS of the differences ranges between 0.10–0.15 s for different periods, which agrees well with the value of 0.14 s estimated from Rayleigh phase velocity differences between inter-station pairs by Růžek et al. (2016).

5.2. 1D velocity model

Here, we apply our stochastic MC inversion to obtain a representative 1D layered velocity model for the real dataset. We assemble data from all phase velocity maps, excluding points where the initial velocity remains unchanged (i.e., there is no sensitivity of the data on the velocity structure).

1D inversion is computationally cheap (≈ 4 h on 8 CPUs) and thus may be used with the help of synthetic 1D tests to investigate properties of the inverse problem. The 1D inversions were performed employing different datasets: only Love phase velocities or all-component phase velocities. The 1D model is composed of 2 km thick layers down to 32 km and a halfspace. Furthermore, the addition of the v_p/v_s model parameter to 1D S-wave velocity parameters was examined.

Fig. 4a shows results of Love phase velocity inversion into a 1D S-wave velocity model. The model appears to have a relatively wide but unique maximum. This shows that although the result is stable, either the data sensitivity to the S-wave velocity model is low or the 1D model is inappropriate to represent the dataset due to the actual higher spatial variability of the true velocity model, or both.

In Fig. 4b, we show the results of inversion of Love and Rayleigh phase velocity dispersion data into a 1D layered S-wave velocity model with fixed $v_p/v_s = 1.57$. By increasing the number of data, the variance of the model parameters decreases. This suggests that the data from all components are well represented by a common 1D S-wave velocity

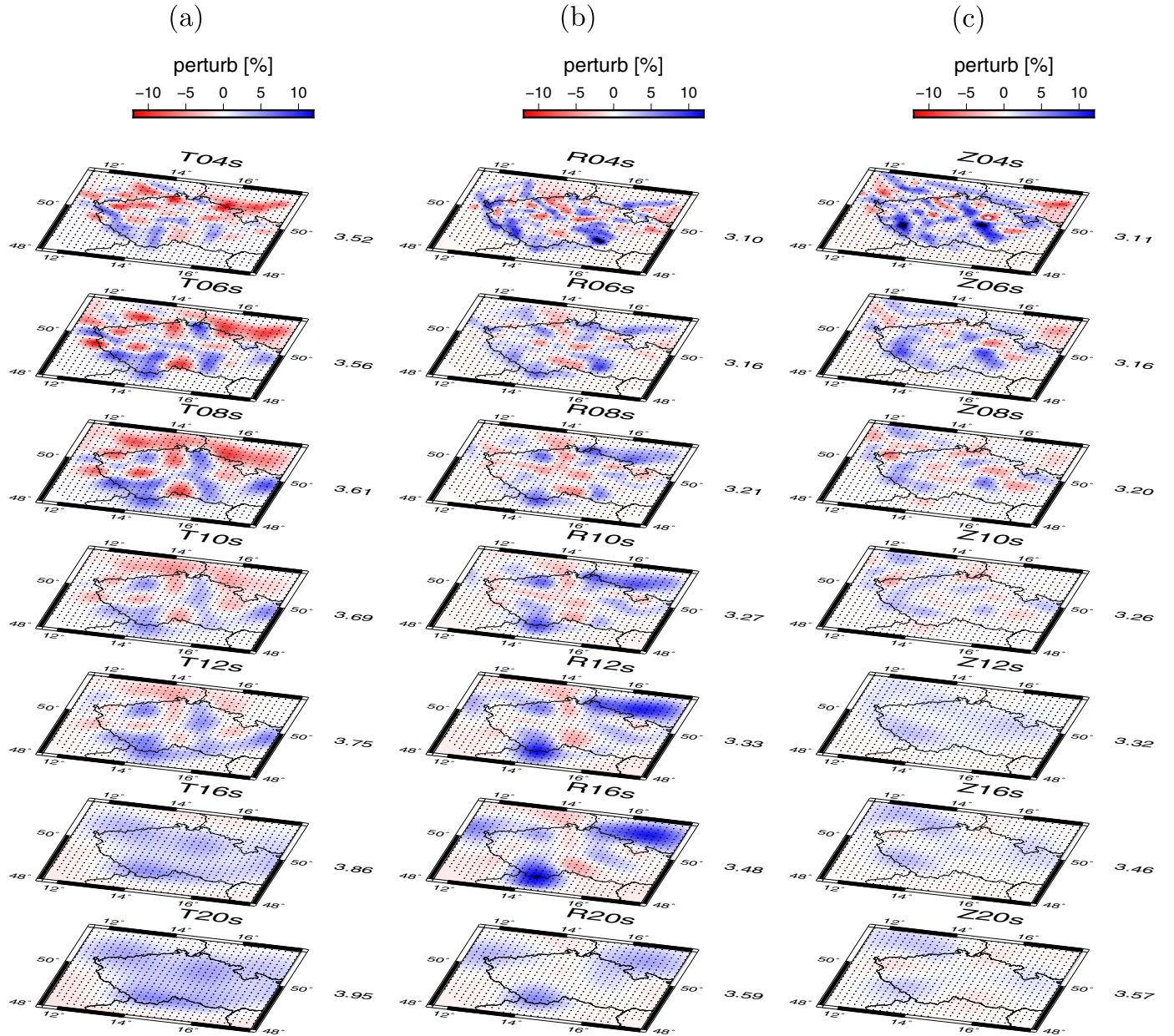


Fig. 3. Phase velocity dispersion maps obtained by the 2D adjoint inversion for each period from all components: a) transverse T (i.e., Love wave), b) radial R c) vertical Z (both Rayleigh waves). The period increases from top to bottom (see legend). The model perturbations (color scale) are presented relative to mean phase velocity model, denoted for each map to the right in km/s. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

model.

Fig. 4c displays models obtained from the inversion of all components into a 1D S-wave velocity model and an additional parameter – depth-independent v_p/v_s . The resulting S-wave velocity model is similar to the previous one (Fig. 4b), both in terms of the best model and the variance. This shows that the inversion of the v_p/v_s parameter does not affect the inversion of the S-wave velocity model much, except for slightly amplifying the velocity increase at ~ 20 km. The obtained v_p/v_s ratio has low variance and appears to be well resolved. However, by assuming depth-independent v_p/v_s , the inversion estimates v_p/v_s corresponding to the average over all depths. In general, the averaged parameter is better constrained than v_p/v_s at a given depth. Therefore, it must not be interpreted as a low variability of the v_p/v_s of the real 3D structure.

From the 1D velocity models, it is difficult to distinguish between the effects of data resolution and inherent variability due to the 3D

structure. To separate these two effects, we performed several synthetic tests. The synthetic tests with both noiseless and noisy data were performed with similar setting as the real data inversion. The resulting variance of the real data inversion is distinctly larger than that of the synthetic tests. This suggests that the most of the 1D model variability is due to the lateral inhomogeneity of the real crust.

5.3. 3D velocity models

The result of our Bayesian inversion consists of > 1 million PDF samples (i.e., 3D models). To present the results, we display the mean model calculated from all models and the best model. The advantage of the mean model is that it presents only stable features. Therefore, the mean model is typically smoother than any single model drawn by the MC sampler. The best model is shown as a representative model to examine differences in properties between a single model and the

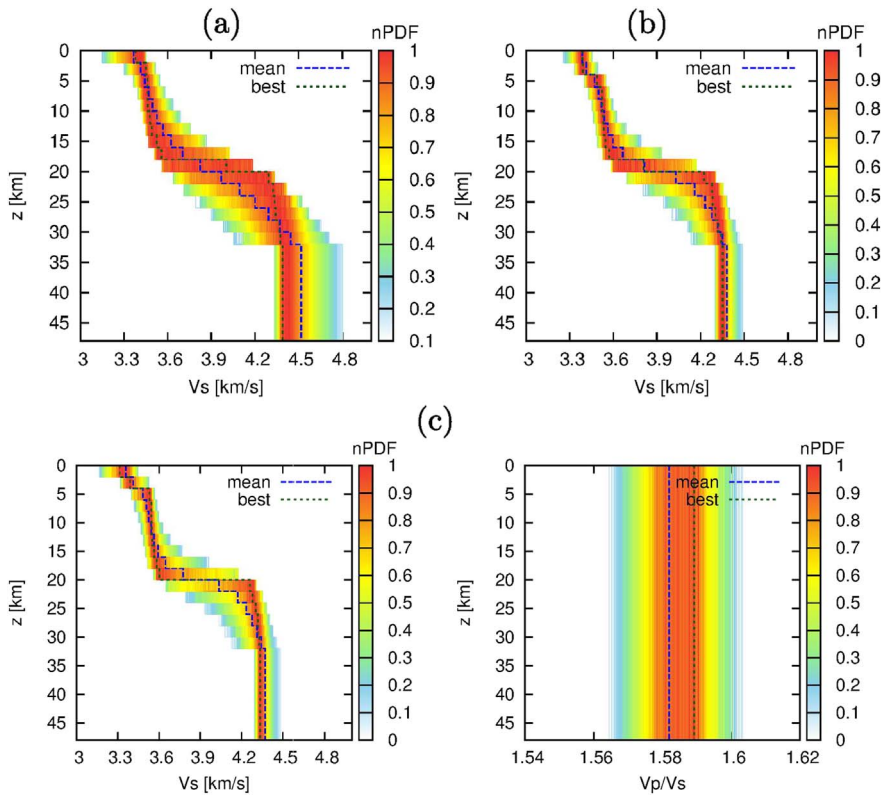


Fig. 4. Results of the inversions of the phase velocity dispersion maps into 1D layered models by the PT algorithm. The color scale corresponds to the normalized PDF (nPDF) scaled by the PDF value of the best model (black dashed line). a) Inversion of Love dispersion maps into a 1D S-wave velocity model. b) Inversion of all Love and Rayleigh dispersion maps into a 1D S-wave velocity model, v_p/v_s being fixed. c) Inversion of all Love and Rayleigh dispersion maps into a 1D S-wave velocity model (left) and depth-independent v_p/v_s ratio (right). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

averaged one.

Fig. 5 presents the depth-slices of the mean and the best 3D S-wave velocity model. To better visualize its lateral variations, Fig. 6 shows the same models but in terms of perturbations relative to the horizontally averaged model.

All models show increase of S-wave velocity with depth. Regarding the lateral variations, both mean and best model share the most significant structures. The differences between them appear on smaller scales.

Simultaneously with S-wave velocities, we have performed the PT exploration also for the depth-independent v_p/v_s ratio. The reasons for the v_p/v_s depth-independence are: a) poor resolution of v_p/v_s and b) to reduce the parameter space. The resulting mean model and its standard deviation are shown in Fig. 7. Although, the range of the v_p/v_s ratio is wide (1.5–1.8), the standard deviation in recovered areas lies mostly below 0.05, suggesting strong horizontal structural variability of this parameter.

5.4. Uncertainty of the 3D model

The greatest advantage of employing MC methods to solve the inverse problem lies in the plurality of models representing the solution. As an example, Fig. 8a shows vertical 1D models in a model control point located in the middle of our domain. From this example, we see that the best resolved part in the inversion lies at depths of 2–18 km. At greater depths, the variance in the S-wave velocities is very high. It also appears that the PDF of the S-wave velocities in deeper parts as well as v_p/v_s ratio have 2 local maxima. We ascribe this to the undersampling of the PDF in the particular parameter domain. Also note that for model control points located at the boundaries of our domain, the overall uncertainty increases.

Uncertainty of the model along a profile can be estimated by standard deviation of the mean model (Fig. 8b right). In general, the lowest uncertainty (as indicated also by the 1D models in Fig. 8a) is achieved down to ~20 km. However, the uncertainty changes also laterally

along the profile (between 1–2% for the well resolved part). Alternatively, we visualize these changes via standard deviation of selected S-wave velocity isolines (Fig. 8b left).

We emphasize that we should be extremely cautious when interpreting the imaged structures below 25 km, where the model variance increases rapidly. In particular, our models are not suited for the search of the Moho. The following geologic interpretation should be confined to large-scale structures only. This limitation is a consequence of the method applied, namely: a) employment of surface wave data, which is inherently sensitive to the averaged (smoothed) structures both horizontally and vertically, making it impossible to obtain velocity interfaces; b) model parametrization – horizontal grid with relatively large (50 km) spacing; and c) averaging of great amount of single models generated by the MC inversion which produces stable but smooth structures.

6. Discussion

6.1. Geological interpretation of dispersion maps and 1D S-wave velocity profile

The result of the first part of the inversion – the phase-velocity dispersion maps – usually correlate well with known geology and are used for preliminary interpretation (e.g., Saygin and Kennett, 2010; Nicolson et al., 2012). In our dispersion maps (see Section 5.1, Fig. 3), there is a high velocity structure in the southern part of the domain present in almost all maps, which may be related to the Moldanubian domain. Another stable high velocity anomaly, located in the center of the north-west border of the Czech Republic is found easily on maps for components T and R for periods 8–16 s, where it is surrounded by low velocities. Moreover, this anomaly can be also tracked for the Z component. This anomaly is situated beneath the Eger Rift zone, where a high velocity body is usually found in the tomography (Alexandrakis et al., 2014; Grad et al., 2008; Mousavi et al., 2015; Růžek et al., 2007). For the shortest periods, the structures are much more complex and

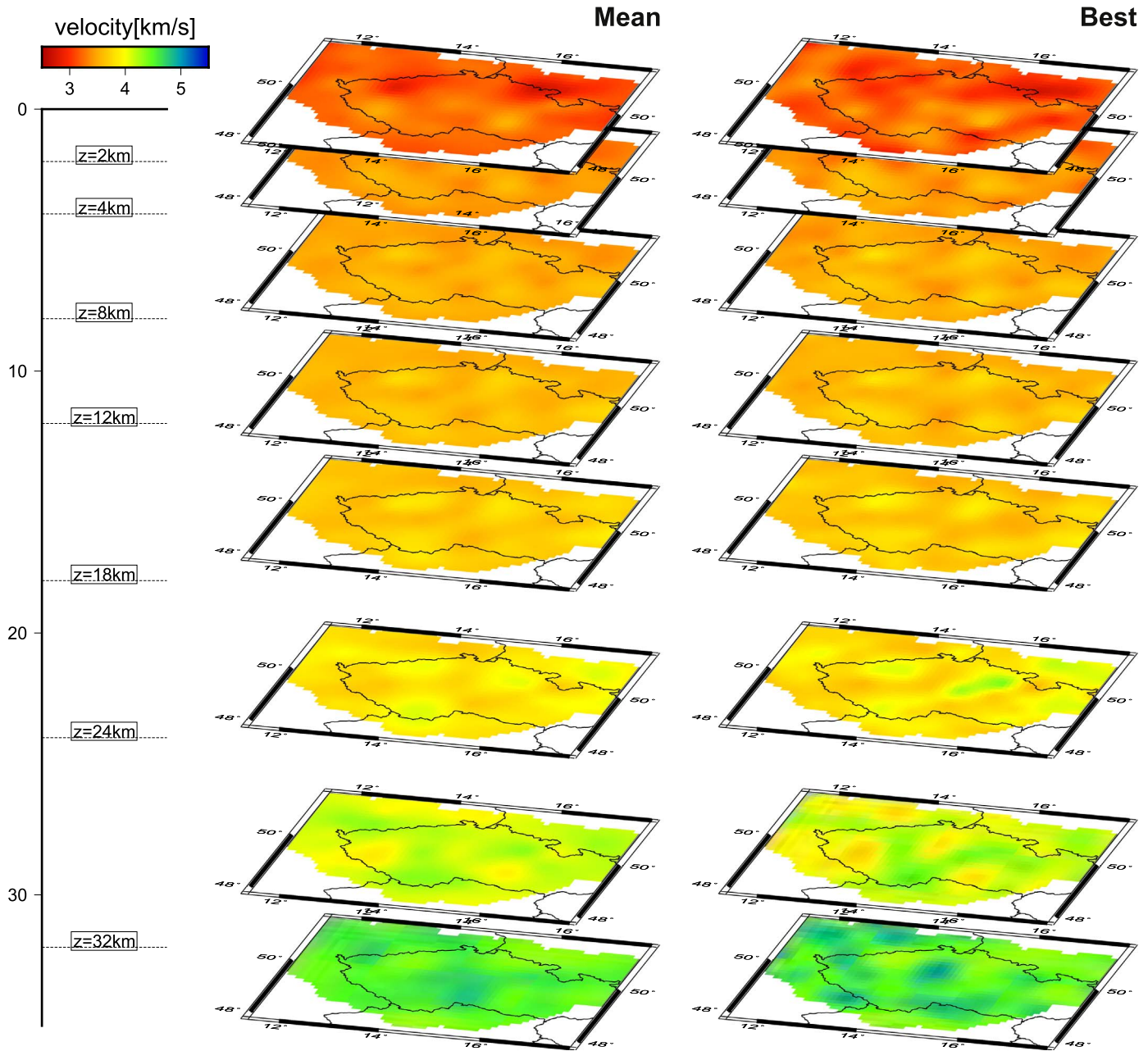


Fig. 5. Depth slices through the mean (left) and best (right) S-wave velocity models obtained by our 3D inversion using the PT algorithm. The layers are shown in the left. The areas with no station coverage are masked.

stable features present for all components are more difficult to determine. In the Elbe Fault Zone, one may observe a narrow low velocity anomaly for all maps at periods 4–6 s.

Before the inversion into 3D model, we performed Bayesian inversion of the dispersion maps into 1D S-wave velocity model and v_p/v_s ratio (see Section 5.2, Fig. 4). This 1D model may be considered as a representative model of the Bohemian massif. In the near-surface part (depth less than 4 km), there is a moderate velocity gradient and relatively large variance. This may point out to uneven sedimentary cover of the Bohemian Massif. The rest of the upper crust (from 4 to ~ 12 km) shows very low velocity gradient and a very low variance. This suggests strongly homogenized upper crust across the whole domain.

Around 20 km depth, there is a strong S-wave velocity increase indicating significant structural difference between the upper and lower crust. The increased variance in these depths may be associated with the high velocity gradient being laterally heterogeneous in the real

structure.

Below 25 km, there is no significant velocity jump corresponding to the Moho. Moreover, the variance in these depths decreases, namely in the halfspace. This is due to the fact that the variance estimated for the S-wave velocity at these depths is only formal as it represents variance of the S-wave velocity averaged over all depths below 32 km. Therefore, the obtained variance reflects neither vertical nor lateral variability of the real 3D structure, and thus the Moho is not resolved in our model.

The Bohemian Massif shows low v_p/v_s value ~ 1.6 indicating rather rigid, consolidated material. The interpretation of the v_p/v_s horizontal perturbations in Fig. 7 is rather ambiguous as the uncertainty of this parameter may be poorly estimated. In similar sense as the S-wave velocity of the halfspace velocity mentioned above, the estimated v_p/v_s corresponds to the average over all depths. The increase in the northern part (Sudetes), may be attributed to the presence of the thick

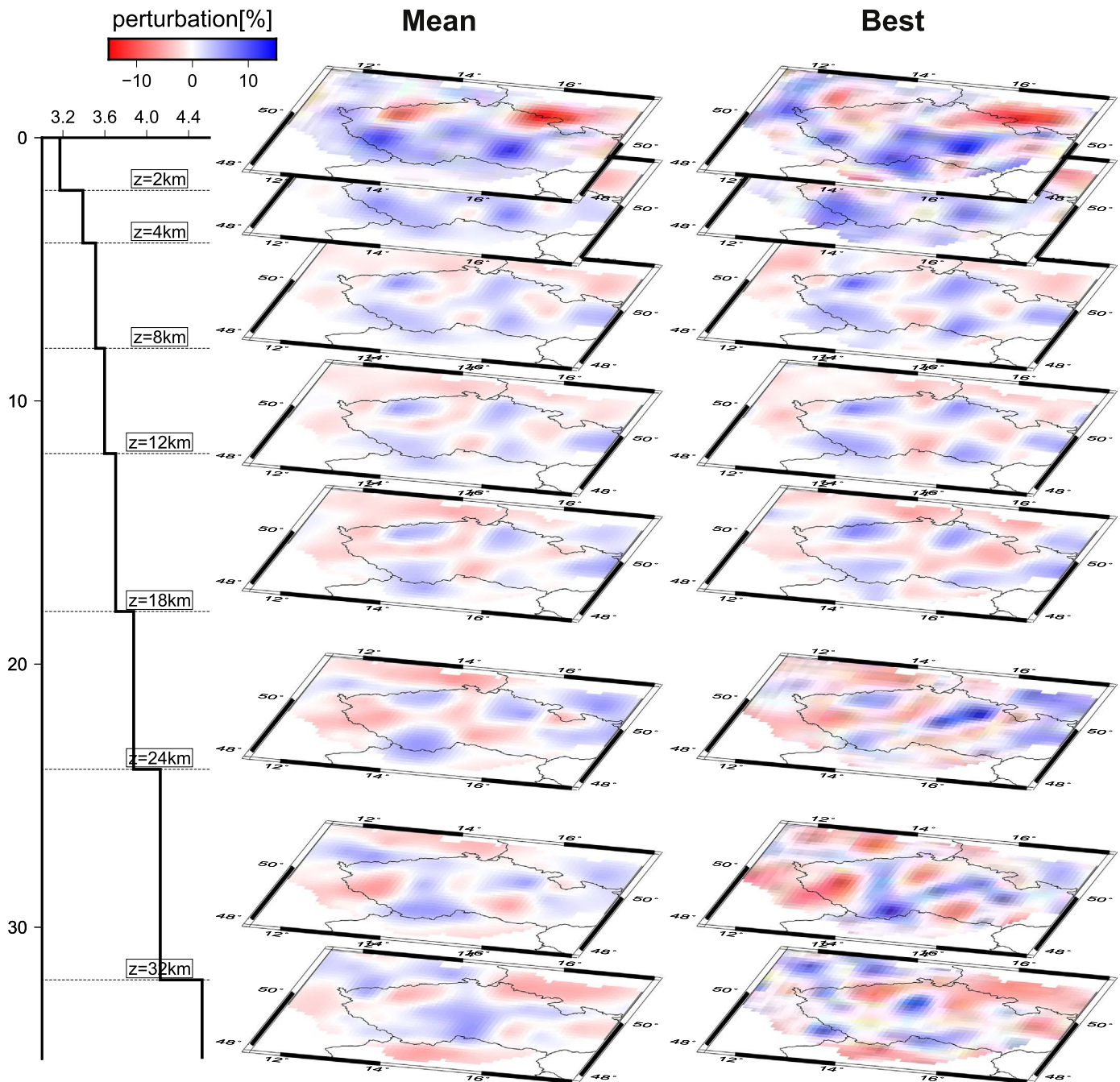


Fig. 6. The same as Fig. 5 but for S-wave velocity perturbations calculated as relative differences from horizontally averaged velocities (left panel).

sedimentary coverage.

6.2. Geological interpretation of 3D S-wave velocity model

Here, we discuss the correlation of our 3D mean S-wave velocity model with known geology with the help of vertical cross-sections (Figs. 9, 10), corresponding to selected profile measurements of active seismic experiments CELEBRATION 2000 (CEL09, CEL10) and SUDETES 2003 (S01, S02, S04). We also compare the cross-sections with 2D models obtained along these profiles by other authors, but with great caution. Indeed, the latter models were interpreted in terms of P-wave velocities and were based on other kind of data than in our case (body waves, frequency ranges).

Fig. 9 shows the vertical cross-sections through the mean and the best 3D models interpolated along the profiles. The mean isovelocity

lines of 3.5 and 4.0 km/s are delineated together with standard deviation representing their uncertainty. Furthermore, Fig. 10 shows the velocity model perturbations relative to their horizontal averages to better distinguish different structures.

Although the Bohemian Massif is composed of several main tectonic domains, its overall structure is much more complicated. As our results also imply, the correlation with geologic domains is not that obvious. This was already suggested by Růžek et al. (2016) who obtained rather heterogeneous models for each domain.

For the following discussion, to better distinguish the horizontal extent of the velocity structures, Fig. 11 displays maps of two isovelocity topographies – for values 3.25 and 3.6 km/s. The values were chosen to show the near-surface structures (0–4 km) and the deeper parts (4–16 km), respectively.

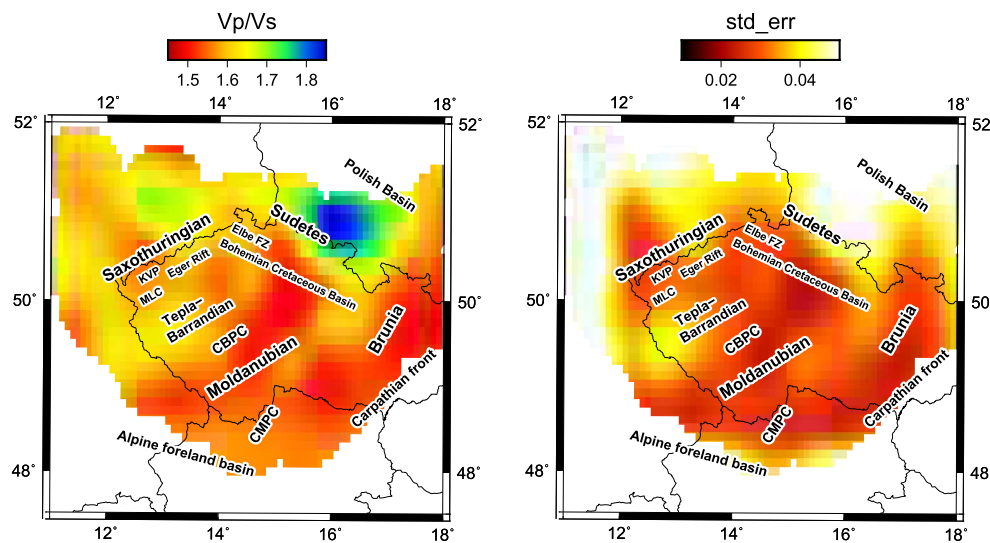


Fig. 7. Mean model and standard deviation of v_p/v_s value obtained by our 3D inversion. The v_p/v_s ratio is assumed to be depth-independent.

6.2.1. Sedimentary basins

For the shallowest layers (0–2 km in Figs. 5 and 6, and Fig. 11a), we find good correlation of low velocity perturbations with sedimentary basins. The extensive Bohemian Cretaceous Basin is visible in our model mainly where the sedimentary cover is presumably thicker. Such area is located at the northern rim of the Bohemian Cretaceous Basin in the Sudetes (including the Intra-Sudetic Basin) continuing further to the Polish Basin, where the pronounced low-velocity anomaly is present. Thicker lower-velocity layer in this area is also visible on profile S02 at distances > 250 km (see Figs. 9, 10), and can be also found in S02 velocity profiles by Růžek et al. (2007) and Majdański et al. (2006). Other such area corresponds to the Eger Rift where the low near-surface

velocities are retrieved in a ~ 50 km wide zone (see also profiles S01 and S04 in Figs. 9 and 10).

The anomaly corresponding to the sedimentary cover of Brunia is rather variable with its largest amplitude in the north-western part, where the thickness of the sedimentary layer increases. Similar anomalies were identified on CEL10 profile by Hrubcová et al. (2008) and Růžek et al. (2007).

6.2.2. Tectonic domains

Deeper parts of our model (depths 4–24 km in Figs. 5 and 6, Fig. 11b) reflect the main geologic domains.

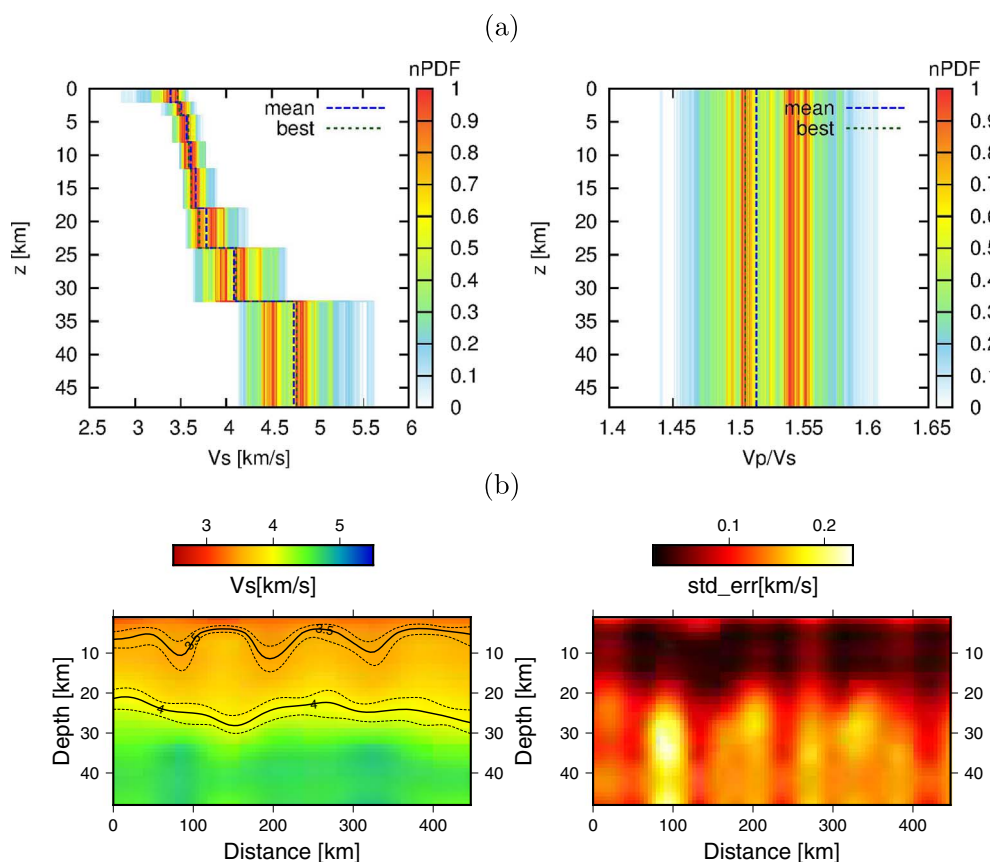


Fig. 8. a) 1D vertical models in a selected model control point in the middle of the domain. Color palette shows nPDF – PDF normalized to its maximum (best model, see legend). b) 2D mean model and its variance interpolated along the CEL09 profile. Two velocity isolines are shown with their standard deviation. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

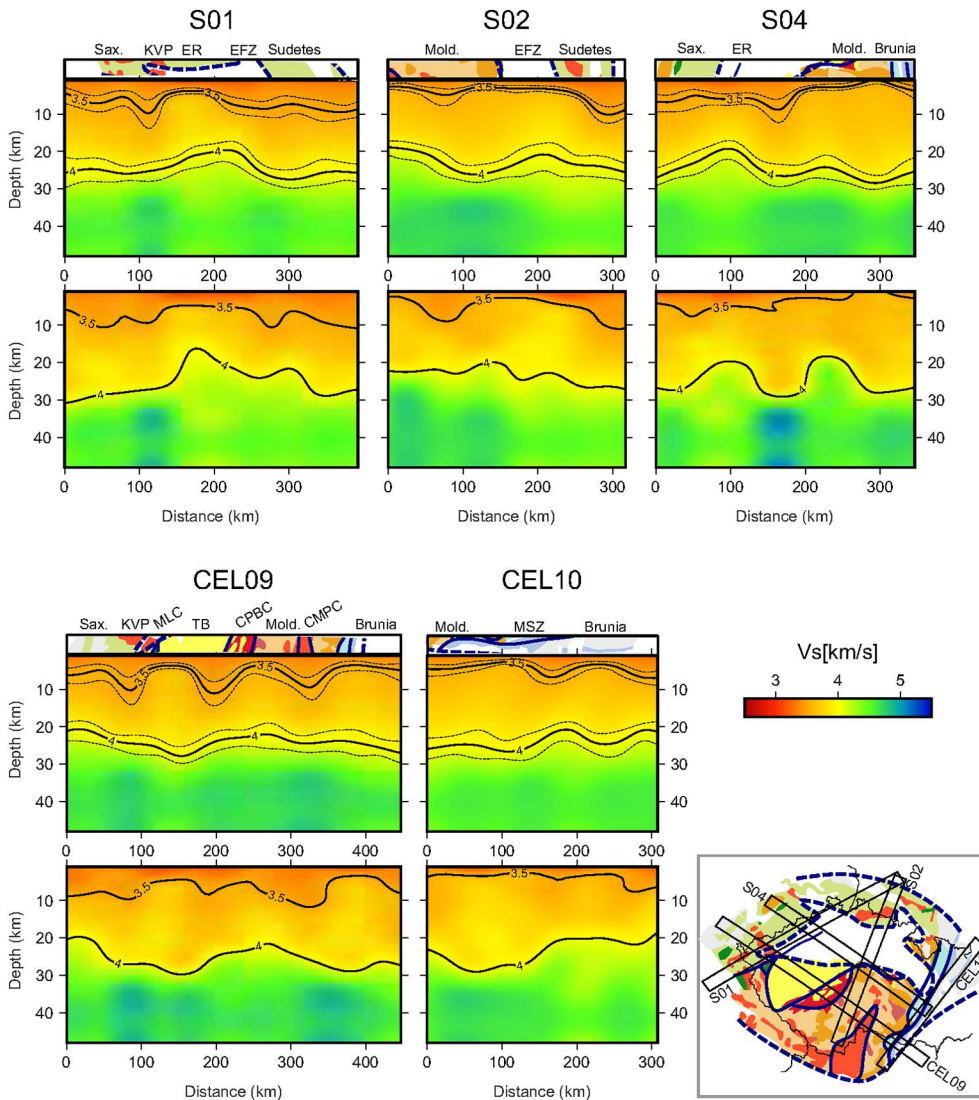


Fig. 9. The mean (top) and best (bottom) S-wave velocity model in selected profiles across the studied domain. The isoline depths of S-wave velocity values 3.5 and 4.0 km/s are shown by black curves. In the mean model, the standard deviation around these values is shown by dashed curves. The main geologic structures are denoted on the top of each profile as indicated in the map in the inset panel. MLC – Mariánské Lázně Complex, CBPC – Central Bohemian Plutonic Complex, CMPC – Central Moldanubian Plutonic Complex, ER – Eger Rift, KVP – Karlovy Vary Pluton, EFZ – Elbe Fault Zone, Mold – Moldanubian, Sax – Saxothuringian, TB – Teplá-Barrandian, MSZ – Moravo-Silesian Zone.

- The Moldanubian domain is characterized by a high-velocity anomaly, as it represents middle/lower crust denuded by erosion. The low-velocity uppermost layers are usually not present there (see also models of Růžek et al., 2007; Hrubcová et al., 2005; Majdański et al., 2006; Novotný, 2012; Růžek et al., 2016). Conversely, exceptionally high-velocity structures near the surface (see Fig. 11a and profile S04 in Figs. 9, 10) were detected in the same area also by Růžek et al. (2007) and Hrubcová and Šroda (2015). Some of these anomalies spatially coincide with exposed high-grade rocks (see Fig. 1a). In the middle of the high velocity Moldanubian domain, a low-velocity body is present in our model (see also profiles CEL09 – distance 300 km, S04 – distance 300 km and S02 – distance 100 km in Figs. 9 and 10). Similar low-velocity zone down to 15 km was found on CEL09 profile by Novotný (2012), who interpreted it as an accretionary wedge on the contact of the Moldanubian and Moravian (i.e., Brunia) domains.
- The Teplá-Barrandian domain (partially hidden under the Bohemian Cretaceous Basin) is on the contrary to the Moldanubian characterized by a low-velocity anomaly corresponding to its composition of less consolidated upper crustal rocks. An exception is a high velocity anomaly extending northward from the Moldanubian domain in the middle of the Bohemian Cretaceous Basin. The structure is also displayed on profile S02 at the distance of 200 km (Figs. 9 and 10) and evokes a promontory of the Moldanubian into the Teplá-

Barrandian under the Bohemian Cretaceous Basin (see Fig. 11b). However, this is not confirmed by known geology (Uličný et al., 2009).

- The Saxothuringian domain and the Sudetes show mainly low-velocity anomalies (see also profiles S01, S02 and S04 in Figs. 9 and 10) which reflect their similar composition and origin. Lower velocities on S02 were also obtained by Majdański et al. (2006).
- The Brunia domain in the north-eastern part shows a high velocity anomaly representing a well consolidated crystalline basement. Higher S-wave velocity values in this domain were also found by Růžek et al. (2016).

6.2.3. Plutonic bodies

In the Bohemian Massif, numerous granitic intrusions are present. Our inversion is not able to distinguish small-scale structures and thus we focus on large plutons and plutonic complexes only. Among the large plutons, the low-velocity anomaly corresponding to the Karlovy Vary Pluton may be found in our model in Fig. 11b and on profiles CEL09 and S01 in Fig. 9. The low-velocity anomaly corresponding to the Karlovy Vary Pluton is also present in models of Hrubcová et al. (2005) and Novotný et al. (2009). Moreover, Málek et al. (2004) derived 1D models for different units of the West-Bohemia region obtaining lower P-wave velocities for the plutons compared to the crystalline units. Such results are also in agreement with laboratory

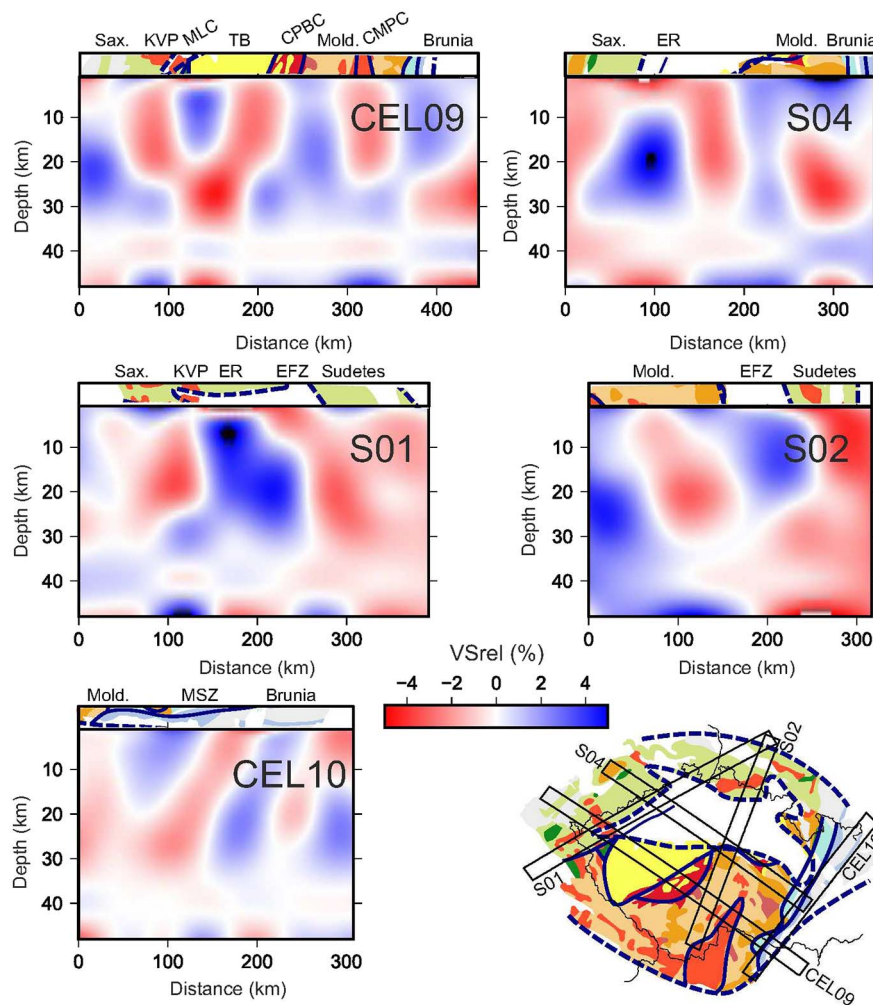


Fig. 10. The S-wave velocity perturbations with respect to average model for each depth calculated from the mean model along the studied profiles. The main geologic structures are denoted on the top of each profile as indicated in the map in the inset panel. MLC – Mariánské Lázně Complex, CBPC – Central Bohemian Plutonic Complex, CMPC – Central Moldanubian Plutonic Complex, ER – Eger Rift, KVP – Karlovy Vary Pluton, EFZ – Elbe Fault Zone, Mold – Moldanubian, Sax – Saxothuringian, TB – Teplá-Barrandian, MSZ – Moravo-Silesian Zone.

measurements for granitic rocks in this area (see also Pros et al., 1998).

Another large plutonic bodies that might be inferred by our inversion are the Central Bohemian Plutonic Complex and the Central Moldanubian Plutonic Complex. In the vicinity of the Central Bohemian Pluton a low-velocity anomaly is observed. This anomaly may,

however, belong to the adjacent Teplá-Barrandian domain, where the lower velocities are typical. In the case of the Central Moldanubian Pluton, a low-velocity anomaly is present just below its northern tip. However, it does not continue below the southern part of the Pluton, which makes interpretation of this anomaly debatable.

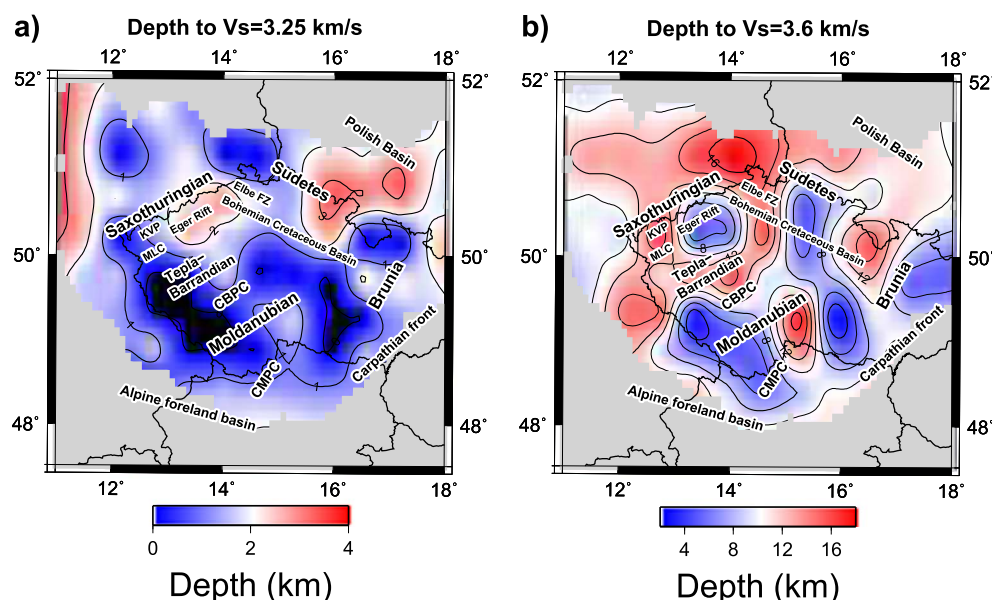


Fig. 11. Topography of the iso-S-wave velocity surface for value 3.25 km/s and 3.6 km/s of the mean model. The main geologic structures are shown (MLC – Mariánské Lázně Complex, CBPC – Central Bohemian Plutonic Complex, CMPC – Central Moldanubian Plutonic Complex, KVP – Karlovy Vary Pluton, FZ – Fault Zone).

All these three anomalies are clearly visible as low-velocity perturbations in Fig. 10 – profiles CEL09 and S01, where they extend down to ~20 km depth. For plutonic bodies, such a large depth extent is unlikely as confirmed by other studies: i) gravity modeling by Guy et al. (2011) indicated that the Central Bohemian Plutonic Complex and Central Moldanubian Plutonic Complex reach to 5 and 10 km, respectively, and ii) the Karlovy Vary Pluton was identified as a low velocity anomaly extending down to 10 km on the CEL09 profile by Hrubcová et al. (2005) and Novotný (2012). The supposed overestimation of the depth extent of these anomalies in our model may be ascribed to the vertical smoothing effect of the surface waves.

6.2.4. Eger Rift

The most distinct structure in our model is in the area of the Eger Rift: the Saxothuringian – Teplá-Barrandian boundary fault active even nowadays. Under the sediments, the Eger Rift is characterized by a very strong high-velocity anomaly extending deep in the lower crust (see Fig. 11b; profiles S01 and S04 and the southern rim of the Eger Rift anomaly is also seen at distances 100–150 km of CEL09, Figs. 9 and 10). In this area, higher S-wave velocities down to lower crust were already found by Kolínský et al. (2011). High P-wave velocities were found by Alexandrakakis et al. (2014), Mousavi et al. (2015). The S01 profile extends along the Eger Rift where the high velocity bodies were already found by Růžek et al. (2007) and by Grad et al. (2008). Grad et al. (2008) interpreted their two separate high-velocity bodies originating in the lower crust and continuing to the shallower depths as a result of the Saxothuringian subduction. The Moho updoming in this area was found by Heuer et al. (2006) and in the S04 model of Hrubcová and Šroda (2015). However, our model, despite apparent reversal of the anomaly at greater depths (see bottommost map in Figs. 5 and 6), is less sensitive to the deep structures and is unable to contradict or confirm the continuation of the high-velocity anomaly to greater depths.

7. Conclusions

We have inverted ambient noise inter-station dispersion curves into the 3D S-wave velocity model of the Bohemian Massif. The traditional two-step approach is modified by employing the finite-frequency adjoint method in the first step, and in the second step by performing Bayesian inversion by the parallel tempering algorithm for all model parameters simultaneously.

For the interpretation, we have considered the mean velocity model and its uncertainty which were calculated from all resulting 3D models sampling the posterior PDF (more than 1 million models). The mean model reveals the structures that are stable in the inversion. The estimated uncertainty helps us to assess reliability of the individual model features. As the main drawback of our model one may consider its lack of the more detailed structures for the geological interpretations. In addition, the surface geology is not perfectly related to the deep crustal structure that is imaged by surface waves inherently smoothing both laterally and vertically the true velocity structure.

The inferred 3D S-wave velocity model shows good correlation with main geologic domains of the Bohemian Massif. The Moldanubian domain is characterized by high S-wave velocities representing exposed middle/lower crustal material, except for a distinct anomaly located in its central part. Similarly to the Moldanubian, Brunia shows higher S-wave velocity anomalies. In contrast, the Teplá-Barrandian, Saxothuringian and Sudetes domains show lower S-wave velocities. In the Teplá-Barrandian domain, we have recovered a high velocity region with unclear geological interpretation resembling a promontory of the Moldanubian domain.

The most prominent high-velocity anomaly in the model is found beneath the Eger Rift. Moreover, some of the low S-wave velocity anomalies present in our model may correspond to large plutonic bodies, and the topmost low velocity structures correlate well with the sedimentary cover of the Bohemian Massif.

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