Three-dimensional seismic velocity model of the West Bohemia/Vogtland seismoactive region

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SUMMARY

In this paper, we present a smooth 3-D seismic model \( WB2012 \) for the West Bohemia/Vogtland earthquake swarm region defined by means of seismic tomography. Inverted data were represented by a set of 2920 \( P \)-wave traveltimes from controlled shots fired in a framework of different experiments and a set of 11339 \( P \)- and \( S \)-wave arrival times from 661 local earthquakes between 1991 December and 2010 March. We used a standard tomographic approach for independent calculation of \( P \)- and \( S \)-wave velocity fields in a rectangular grid whose size was 1 km in all coordinates. The traveltimes and rays were calculated by a numerical solution of the eikonal equation. While locating seismic events, our new \( WB2012 \) model yielded arrival time residuals on average by 13 per cent lower and hypocentre depths by 0.95 km shallower compared to the locations of the foci in the standard 1-D vertically inhomogeneous isotropic velocity model of the West Bohemia swarm region \( WB2003 \). Further, we converted the \( P \)- and \( S \)-wave velocities to the bulk modulus \( K \) and Poisson’s ratio \( \nu \). The bulk modulus (\( \approx 40 \)–70 GPa) correlates acceptably with the tectonic and geological structure of the area. The anomalously low values of the Poisson’s ratio (\( \approx 0.15 \)) are typical for the most active focal zones of Nový Kostel and Lazy in West Bohemia.

Key words: Controlled source seismology; Body waves; Seismic tomography; Crustal structure; Europe.

1 INTRODUCTION

The area of West Bohemia (Czech Republic) and Vogtland (Saxony, Germany), latitude \( \approx 49.8^{\circ} \)–\( 50.7^{\circ} \) N, longitude \( \approx 12^{\circ} \)–\( 13^{\circ} \) E, is one of the most active intra-plate earthquake swarm areas in Europe.

This region is situated in the western part of the Bohemian Massif where three principal tectonic units (Saxothuringian, Moldanubian and the Teplá-Barrandean) approach each other. It is intersected by an ENE–WSW trending neotectonic structure, the Eger rift and by the Mariánské Lázně fault (MLF) striking in the NNW–SSW direction. Quaternary volcanism was active there until the Holocene (Gögen & Wagner 2000); two extinct volcanoes, Komorní Hůrka (the age of about 0.3 Ma) and Železná Hůrka in West Bohemia, are located only 15 and 25 km apart from the main epicentral zone (Fig. 1). Most of the seismic energy has been released there in the form of quite frequently reoccurring earthquake swarms which comprise a series of \( 10^3 \)–\( 10^4 \) individual events, mostly with magnitudes of \( M_L < 4 \). Typically, the events are clustered in several focal zones (Nový Kostel, Kraslice-Klingenthal, Plesná, Lazy, etc.; for details see, e.g. Horálek & Fischer 2010), the depths of foci in the whole region varying between 5 and 20 km. The most of recent activity (more than 90 per cent of the total seismic moment released) took place near the village of Nový Kostel, where the hypocentres are located to a depth of 7–10 km. The largest instrumentally recorded swarms so far were those of 1908 and 1985/86, which reached maximum magnitudes of \( M_L \approx 5.0 \) and 4.6, respectively; the seismic potential of the region is assessed to correspond roughly to a single event of local magnitude \( M_L \approx 5.5 \) (Fischer et al. 2010). Most of the available seismic data come from the local network, WEBNET, consisting of 13 networked and 10 autonomous three-component seismic stations, which cover an area of about 900 km\(^2\) (for details, see www2012a).

Another notable feature of geodynamic activity in Western Bohemia and Vogtland is a high flux of mantle-derived CO\(_2\) being manifested by several moffetes and numerous carbonized mineral springs and CO\(_2\) vents (e.g. Weinlich et al. 1998). It is generally assumed that the earthquake swarm activity is closely related to crustal fluid movements (Horálek & Fischer 2008). Hence, the region in question has been regarded as a unique intraplate geodynamics laboratory, which has been investigated together with regions characterized by the occurrence of earthquake swarms, for example, southern Iceland, or by swarm-like seismicity induced by injected fluids into deep boreholes at the HDR site Soultz-sous-Forêts in France (Horálek et al. 2010).

Localization of hypocentres (Fischer & Horálek 2000, 2003; Fischer 2003; Horálek & Fischer 2010), magnitude estimation (Horálek et al. 2009) and determination of focal mechanism of stronger events (Horálek et al. 2002; Vavryčuk 2011; Horálek & Šilený 2013) represent basic interpretation procedures. All these tasks depend on the \( P \)- and \( S \)-wave velocity model. Therefore, the...
Figure 1. The Western Bohemia/Vogtlând region (after Horálek & Fischer 2008) is depicted with the earthquake epicentres, WEBNET seismic stations and principal faults (EG, Eger rift; MLF, Mariánské Lázně fault). The main Nový Kostel focal zone is marked by dashed rectangle, the hidden main fault plane NK (strike 170°, dip 80°), on which all earthquake swarms and majority of microswarms occurred during last 25 years, is denoted by a grey rectangle. A dashed ellipse marks a contiguous focal zone Kraslice–Klingenthal. The maximum compression striking 160° (Svancara et al. 2008) is indicated in the lower right corner. The solid line indicates the Czech–German border, squares mark towns in West Bohemia, SE Saxony and NE Bavaria. The German KTB superdeep borehole is situated in the lower left corner.

determination of the optimum velocity model was naturally subject of great concern in the past. Novotný (1996) summarized previously published 1-D velocity models and proposed a new velocity model, WB95. Málek et al. (2000, 2004) published a number of 1-D velocity models based on both earthquake and shot data, which were inverted using different approaches. A 1-D weak anisotropic model and its isotropic simplification developed by Málek et al. (2005) have been used in various studies as standard velocity models. Much effort was committed to studying the importance of anisotropy using moment tensor solutions (Rössler et al. 2004, 2007). However, the impact of the 3-D velocity variations was never surveyed.

Therefore, our goals in this work are: (i) to derive a 3-D smooth velocity model of the focal region of the West Bohemia seismic swarm area and of nearby surroundings by means of seismic tomography; (ii) to collect all relevant data to solve the problem and to enable the simple addition of new data in the future; (iii) to perform all calculations as simply and transparently as possible and to ensure easy verification of results; (iv) to access the reliability of the resulting model and (v) to enable the applicability of the new velocity model in different problems.

A new 3-D velocity model can be applied in diverse seismological/geological problems, for example, localization of the earthquake hypocentres; inversion of the amplitudes (or waveforms) to seismic moment tensor; application of static corrections and migration of seismic reflection data; solving joint seismic and gravity inversion; formulating new hypotheses regarding deep geological structure, etc.

2 DATA

We limited the construction of the tomographic velocity model to the rectangular region 11.73°E–13.07°E and 49.74°N–50.71°N, covering an area of 95 km × 107 km and hence 10 165 km² (Fig. 2). Traveltimes of P waves generated by shots and arrival times of P and S waves generated by earthquakes represent data, which we used for seismic tomography. For the active seismic surveys, we were able to fix both source and receiver locations. However, for local earthquakes only the receiver locations were known exactly. Hypocentre coordinates of earthquakes were updated in each iteration.

Data of controlled shots come from measurements performed on the target territory within the framework of the German experiment, DECORP91 (www2012b) and from international refraction experiments, CELEBRATION 2000 (Guterch et al. 2003a,b) and SUDETES 2003 (Grad et al. 2003). Although the latter two provided besides the Pg also the Pn and PmP phases, we reduced the data set to Pg phases only because of easy and consistent implementation in seismic tomography. Unfortunately, no S phases were available for these kinds of measurements. Most fieldwork was performed using single vertical geophones and S waves could not be picked reliably.
Figure 2. A map showing the location of the target region (right) and a detailed view (left) on the distribution of earthquakes epicentres (light blue circles), controlled shots (red stars), geophone positions (black crosses) and permanent seismological stations (green triangles) with topographic map in the background. The rectangle in the left map specifies the exact extent of the tomographic model presented later on in Figs 4 and 5.

Table 1. Data summary.

<table>
<thead>
<tr>
<th>Data summary</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Number of shots</td>
<td>52</td>
</tr>
<tr>
<td>Number of earthquakes</td>
<td>661</td>
</tr>
<tr>
<td>Number of P-wave onsets from explosions</td>
<td>2920</td>
</tr>
<tr>
<td>Number of P-wave arrivals form earthquakes</td>
<td>5562</td>
</tr>
<tr>
<td>Number of S-wave arrivals from earthquakes</td>
<td>5777</td>
</tr>
</tbody>
</table>

All earthquake data come from the WEBNET database. The complete earthquake catalogue and bulletin of arrivals was excessively large from the perspective of seismic tomography and also highly redundant. Therefore, we reduced original earthquake data so that the density of initial hypocentres in the $1 \times 1 \times 1$ km boxes did not exceed 85 events per box. In this way, the final earthquake catalogue was reduced to 661 events.

Basic statistical features of our data sets are depicted in Fig. S1. Most of the rays have epicentral distances of less than 40 km, the most frequent value is 12 km and the longest ray is 75 km long. The azimuthal distribution of rays is relatively regular except for the azimuth interval $-25^\circ$ to $+25^\circ$. (Table 1 also has a summary of the data.) The digital version of data tables is accessible on the Internet (www2012d), so that in the future both revision of our results, alternative processing applying other methods and supplementing input data by new measurements are possible.

3 METHOD

We solved our problem in the local Cartesian left-oriented system $O(x, y, z) \equiv (12.25E, 50.25N, 0)$, where the altitude 0 corresponds to the true sea level. Lambert’s projection is used to transform Cartesian and geographic coordinates (for parameters of the projection, see the HDR files at www2012d). The horizontal axes $x$ and $y$ are oriented in W–E and S–N directions, respectively; the vertical axis $z$ is oriented positively downwards which results in negative $z$-coordinates of points located along the surface. All shots and stations/geophones are considered with the true heights and no static corrections or station residuals are necessary for compensating the surface undulations. The velocity model is defined independently for $P$ and $S$ waves. The grid size is $nx = 96$, $ny = 108$, $nz = 34$, the grid spacing is 1 km $\times$ 1 km $\times$ 1 km. The total number of unknown parameters both for $P$ waves $n_P$ and for $S$ waves $n_S$ is $n_P = n_S = nx \times ny \times nz = 352 512$, that is, much higher than the number of measurements (the total number of data is 14 259; see also Table 1). The used discretization is consistent with the forward calculation code and velocity models can be immediately included in forward modelling.

The forward problem was evaluated by using the software package NonLinLoc (www2012c). The time fields corresponding to the first arrivals of both $P$ and $S$ waves in the tested velocity model were calculated with the program Vel2Time. The sources were placed sequentially at all shot points and to all seismic stations used in the next step of hypocentres’ localization. The time fields were computed utilizing a numeric solution of the eiconal equation by means of a finite difference algorithm (Vidale 1988; Podvin & Lecomte 1991). Next, all hypocentres were relocated by the NLLoc program using the first-arrival time fields corresponding to the given velocity model. The relocation involves both spatial coordinates and origin times and provides traveltime residuals, which are model-dependent and used later (together with residuals corresponding to shots) for velocity model optimization. In the last step of the forward problem, the traveltime residuals from shots were calculated and rays from shots and earthquakes were traced. This part was solved using a slightly modified program, Time2Eq. The rays were approximated by a continuous sequence of short linear segments, the direction of which being parallel with the local gradient of the time field in a given point.

We solved the inverse problem by using a classical ‘box-like’ method of seismic tomography. The real medium is approximated by a set of homogeneous cubes, which are centred on the nodes of the velocity grid. Representation of velocity models is therefore of two types: grid representation for forward calculations and cell representation for inverse calculations. Such duality is necessary due to the particularity of the forward and inverse problems. No difficulties can occur for smooth models without discontinuities.
Optimization of both $P$ and $S$ velocity models was solved separately using the same algorithm. The discrete 3-D velocity grid of $P$ waves $\alpha_{ijk}$ or $S$ waves $\beta_{ijk}$ is arranged into a slowness vector $s$:

$$s_{li} = \frac{1}{\alpha_{ijk}} \text{ (for P waves)} \quad \text{or} \quad s_{li} = \frac{1}{\beta_{ijk}} \text{ (for S waves)}$$

$I = k + (j - 1) \ast nx + (i - 1) \ast nx \ast ny$.

$$i = [1...nx], \quad j = [1...ny], \quad k = [1...nx].$$

(1)

The measured traveltimes along all rays are arranged into a vector $t_{obs}$ and synthetic traveltimes predicted in the currently tested velocity model into a vector $t_{calc}$. Let matrix $A = [\alpha_{ij}]$ be composed of the lengths of segments of the $ith$ ray crossing the $jth$ cell. The subject of the solution is a system of linear equations:

$$A s = t_{calc}, \quad A(s + \Delta s) \approx t_{obs} \rightarrow A \Delta s \approx t_{obs} - t_{calc} = \Delta t.$$

(2)

where the slowness model correction $\Delta s$ is searched using traveltime residuals $\Delta t$. Matrix $A$ is extremely sparse and the problem can be simplified by removing all zero columns of matrix $A$ and corresponding components of the vector $\Delta s$ and, thus, solving a reduced system of linear equations. The reduced system $A' \Delta s' = \Delta t$ and the original system $A \Delta s = \Delta t$ can be transformed to each other by using a contraction matrix $C$:

$$A \Delta s = A' \Delta s' = \Delta t, \quad A' = AC, \quad \Delta s' = C^T \Delta s, \quad CC^T = I.$$

(3)

The contraction matrix is constructed starting with identity matrix $C = I_{mn \times m}$ ($m$ is the number of columns in $A$) and then removing all columns from $C$ that correspond to zero columns in $A$, so matrix $C$ becomes a rectangular matrix having $m$ rows and $q$ columns, $q < m$. The reduced system (3) is much simpler compared to the original system (2), but it is still ill-posed and a possible way how to overcome this problem is to use a suitable regularization. We used a second-order Tikhonov regularization, which forces minimum curvature of the $\Delta s'$ solution:

$$A' \Delta s' = \Delta t \rightarrow \left[ A' \lambda L_2 \right] \Delta s' = \begin{bmatrix} \Delta t \ 0 \end{bmatrix}.$$

(4)

The operator $L_2$ in (4) generates second differences of the slowness-correction field ($d^2 \Delta s' / \Delta x^2 \sim \Delta s_{i+1,j,k} - 2 \Delta s_{i,j,k} + \Delta s_{i-1,j,k}$, $d^2 \Delta s' / \Delta y^2 \sim \Delta s_{i,j+1,k} - 2 \Delta s_{i,j,k} + \Delta s_{i,j-1,k}$, $d^2 \Delta s' / \Delta z^2 \sim \Delta s_{i,j,k+1} - 2 \Delta s_{i,j,k} + \Delta s_{i,j,k-1}$, etc.). The value of the damping factor $\lambda$ was adjusted such that the Euclidean norm calculated from traveltime residuals in the corrected model did not decrease by more than 50 per cent to allow only small steps between iterations and to keep the iteration process stable:

$$\Delta s'(\lambda) = (A^T A + \lambda L_2^T L_2)^{-1} (A^T \Delta t).$$

(5)

implying that the optimum value of $\lambda$ must be calculated repeatedly in each iteration. After adjusting the value of $\lambda$ the reduced vector of slowness corrections $\Delta s'$ is expanded into the original vector $\Delta s$:

$$\Delta s = C \Delta s'.$$

(6)

and, using inverse mapping (eq. 1) the slowness model is updated in the true 3-D geometric configuration. Gridpoints of the slowness model corresponding to zero rows of matrix $C$ are not modified by this procedure since they correspond to cells which are not crossed by any ray. In order to preserve the smoothness of the corrected model, the Delaunay triangulation is applied to all unconstrained cells and the slowness model is smoothly interpolated in the whole volume.

4 INVERSION

Seismic tomography is non-linear and, usually underdetermined (due to sparse model sampling with respect to data amount, as in our case) and non-unique inverse problem that must be solved iteratively. Uniqueness is achieved by adding artificial regularizing terms but in this way some arbitrariness is included in calculations and the final solution can be biased. The overall structure of our approach is as follows:

(1) to set-up initial velocity/slowness models for $P$ and $S$ waves;
(2) to relocate all earthquakes;
(3) to trace rays both for earthquakes ($P + S$) and for shots ($P$) and calculate the traveltime residuals;
(4) to set-up all relevant equations and solve them for the optimum slowness corrections;
(5) to apply calculated slowness corrections and interpolate unconstrained slowness gridpoints and transform the complete slowness fields into the velocity fields;
(6) if the process stopped to converge (indicated by no further misfit reduction) or computing time has been exhausted, output results and exit;
(7) go to 2.

Items (2) and (3) were solved using the software package NonLinLoc, all other items were solved using MATLAB.

To obtain a reliable velocity model we repeated the inversion procedure 100 times using different starting models and averaged these particular results into one final representative model. Next the covariance matrix from all 100 inverted velocity models was evaluated and its diagonal elements were used for estimating the errors of velocities in individual cells. Starting models were generated applying random additive perturbation to the 1-D isotropic model by Málek et al. (2005):

$$v(x,y,z) = v(r) = v^c(z) \left( 1 + \sum_{i=1}^{20} \phi_i (r) \right), \quad r = [x,y,z]^T,$$

(7)

$$\phi_i (r) = A_i e^{-\frac{(r-r_c)^T C_i (r-r_c)}}.$$

that is by adding 20 random 3-D Gaussian perturbation terms to the 1-D background. In (7), $v^c$ is the propagation velocity (either of $P$ or $S$ waves) in the 1-D model by Málek et al. (2005), $A_i$ is randomly selected relative amplitude of the $ith$ perturbation term ($-0.1 < A_i < +0.1$), $r_c$ is the central point of the velocity perturbation lying randomly inside the model and $C_i$ is symmetric positive definite matrix controlling scaling and orientation of the velocity perturbation in such a way so that most of the perturbation pattern does not exceed the margins of the model. Each perturbation term in (7) clearly represents a Gaussian-like body randomly shifted, scaled and rotated whose isosurfaces are 3-D ellipsoids. Thus, each starting model is a 3-D model and quite different from the 1-D background model. The perturbations of the $P$ and $S$ model are independent, so it is generally possible that non-physical $v_p$-to-$v_s$ relations are generated, but this is of no importance for the solved problem. An example of a starting 3-D model is given in Fig. S2.

The initial and final rms’ for inversions using 100 different starting models are shown in Fig. 3. As it is obvious from this figure, a significant improvement of the data fit was achieved in each inversion. The average rms of all initial models was 0.492 s, while that
of final models was around 0.073 s (average improvement ratio is 1:0.148). The dispersion of individual final rms values is substantially smaller than that of initial rms'. Only in two cases (inversions with sequence numbers 81 and 82 in Fig. 3) the final rms' were comparable with starting rms' of other models which is evidently due to complex starting velocity models initialized too far from the solution. All this indicates that the inversion is formally consistent.

The results of the $q$-th inversion ($q = 1, 2, \ldots, 100$) are velocity models defined in gridpoints $(x_i, y_j, z_k)$ for $P$ waves $\gamma(i,j,k)$ and for $S$ waves $\beta(i,j,k)$. Using mapping from eq. (1), such 3-D velocity grids can be transformed into vectors $\nu$:

$$
\nu_i^q(l) = \alpha(i,j,k), \quad \nu_j^q(l) = \beta(i,j,k), \quad l = k + (j - 1) \ast nx + (i - 1) \ast nx \ast ny, \quad q = 1, 2, \ldots, 100.
$$

Having a set of the particular vectors the average vectors $\nu^\alpha$ and $\nu^\beta$ and corresponding covariance matrices $C^\alpha$ and $C^\beta$ can be calculated:

$$
\nu^\alpha = \langle \nu_i^q \rangle_q, \quad C^\alpha = \left( \langle (\nu_i^q - \nu^\alpha) \ast (\nu_i^q - \nu^\alpha) \rangle_q \right)_{ij},
$$

$$
\nu^\beta = \langle \nu_j^q \rangle_q, \quad C^\beta = \left( \langle (\nu_j^q - \nu^\beta) \ast (\nu_j^q - \nu^\beta) \rangle_q \right)_{ij}.\tag{9}
$$

The symbols $\langle \rangle_q$ in eq. (9) are used for ensemble average, that is, for averaging 100 different final models in our case. The diagonal elements of the covariance matrices are variances of individual elements of the vector $\nu$. Standard deviations of individual components of the vector $\nu$ are determined as follows:

$$
\nu^\alpha \rightarrow \nu^\alpha \pm \sigma^\alpha, \quad \sigma^\alpha = \sqrt{\text{diag}(C^\alpha)},
$$

$$
\nu^\beta \rightarrow \nu^\beta \pm \sigma^\beta, \quad \sigma^\beta = \sqrt{\text{diag}(C^\beta)}.\tag{10}
$$

The velocity vectors $\nu^\alpha$ and $\nu^\beta$ and their errors $\sigma^\alpha$ and $\sigma^\beta$ are transformed back to 3-D grids. These four grids are final results of our tomographic inversion. The errors of the velocity fields calculated via eq. (10) reflect data uncertainties and forward mismodelling in case of unique inverse problem. If the inverse problem is not unique, velocity parameters are coupled and their different combinations fit the measured data equally well. In such a case, the covariance matrix $C$ reflects the above-mentioned factors as well but the controlling factor becomes the problem of non-uniqueness. Inverting a sequence of different starting models and presenting the average model together with its error is how to deal with ill-posed inverse problems like seismic tomography.

### 5 P- AND S-WAVE VELOCITY MODELS

The representative model (i.e. that obtained by averaging 100 different final models via eq. 9) both for $P$ and $S$ waves is shown in Fig. 4. Only selected horizontal and vertical planar cross-sections are presented here. The colour scale is adjusted to show only those parts of the model which were crossed by rays and the velocities were calculated by using real data rather then by interpolation. Since the final model was acquired as the average of 100 partial inversion runs, only cells crossed by rays in at least 50 partial models are displayed, otherwise the cells are masked. Corresponding errors of the velocity maps in Fig. 4 are drawn in Fig. 5. Accordingly, only parts of the model constrained by real data are shown. Relative errors of the velocity model are rather high, ranging from 5 to 15 per cent. The highest errors are, as expected, at a greater depth of the model, while the shallow parts of the model are more correctly determined. Presentation of 3-D spatial data in printed documents is rather demanding, however, computerized graphics provide much better facilities. A user-friendly presentation of velocity models can be found in the web page (www2012d), from where the velocity models can also be downloaded in a numerical form. To give at least a basic idea about the spatial distribution of the $P$- and $S$-wave velocities, four depth-velocity cross-sections along parallel profiles C–C′, D–D′, E–E′ and F–F′ (for their locations see Fig. 4) are shown in Fig. 6. These profiles are oriented perpendicularly to the MLF with 10 km spacing. It is generally assumed that the majority of local seismicity in the western part of the Bohemian Massif is related to the NW end of this fault (perhaps fault system). Focal mechanisms of the 2000 and 2008 swarms as well as the spatial distribution of the foci in the Novy Kostel focal zone signify that the main fault plane is fairly steep (dip ~ 80°) and deepening towards SSW (Vavryčuk 2011; Horálek & Šilený 2013). The MLF is manifested on the velocity maps by decreased velocities, particularly those of $P$ waves. The inclination of the velocity anomalies (velocity isolines), clearly visible, especially on the profile D–D′, corresponds well to the expected inclination of the MLF.

Although as our new 3-D $P$- and $S$-wave velocity models now become available, an application of a proper 1-D model is still beneficial in many seismological tasks (e.g. hypocentre localization or source mechanisms retrieval) owing to its simplicity. Hitherto, a commonly used reference 1-D isotropic model of the upper crust (Vlcek 2005) is composed of piecewise constant gradient layers, and was also derived using $P$-wave traveltimes of shots and $P$- and $S$-wave arrival times of local earthquakes. To simply compare our new 3-D velocity model (hereafter WB2012)...
with the reference 1-D isotropic model \textit{WB2005}, we constructed a 1-D approximation of \textit{WB2012}. This approximation consisted of averaging the velocity field along selected depth horizons and by estimating the standard deviation as functions of depth (both the \textit{WB2005} and averaged \textit{WB2012} models are depicted in Fig. 7). One can see that lateral variations of the \textit{WB2012} model are centred around the standard model \textit{WB2005} in case of \textit{P} waves. The lateral variability decreases with the depth. But with \textit{S} waves the situation is different; besides lateral variations, the new model provides systematically lower velocities, especially for the depth range of 0–4 km. The differences between the models \textit{WB2005} and \textit{WB2012} can be mapped in space as well. The appropriate relative velocity differences between these two models are demonstrated in Fig. S3 (an analogy of Fig. 4). The relative differences are in the range

Figure 4. Horizontal cross-sections of the \textit{P}-wave (a) and \textit{S}-wave (b) velocity fields at a depth of 3 km; Vertical cross-sections along profiles A–A’ (c, d) and B–B’ (e, f). The profiles intersect at the station NKC (indicated by a white triangle). Approximate course of the Mariánské Lázně fault (MLF) is shown by a dashed white line. Parallel profiles C–C’, D–D’, E–E’ and F–F’ cross the MLF perpendicularly, depth-velocity cross-sections along these profiles are displayed in Fig. 6.
Three-dimensional seismic velocity model

Figure 5. Relative accuracy of the velocity maps from Fig. 4. The projection is the same as in Fig. 4 with a different colour scheme.

of ±15 per cent. The notable feature is lowered values (negative differences) in the central part of the model below the station NKC, which typically can be attributed to the existence of a fractured volume as a consequence of the earthquake activity.

6 MODEL RESOLUTION

An important feature of any tomographic velocity model is its resolution, that is, specification of the size of the smallest anomaly, which can be interpreted. The simplest qualitative characteristics on how well various parts of the model are constrained are hit count maps giving the number of rays crossing individual cells. Hit counts averaged along horizontal layers are presented in Fig. 8 (cells not crossed by rays are not included in the averaging). If only depths ranging between 0 and 10 km are considered (i.e. the depths in which we present the velocity models), the average hit count is ~15 for P waves and ~20 for S waves. Note that, the number of P-wave rays is higher than the S-wave rays—nearly every cell, which is crossed by a particular S-wave ray is also crossed by the
Figure 6. Depth-velocity cross-sections along the profiles C–C′, D–D′, E–E′ and F–F′ crossing perpendicularly the seismoactive Mariánské Lázně fault (MLF). For the geographical location of the profiles refer to Fig. 4. Low-velocity anomalies clearly correspond to the MLF fault zone. The profile coordinate of the MLF at the surface is 22.8 km.

corresponding P-wave ray. But there are also many cells, which are crossed only by a few P-wave rays related to the controlled shots. This is why the average hit for the P waves is lower than that for the S waves.

Vertical cross-sections showing spatial distribution of ray density are in the upper part of Fig. 9. Good ray coverage indicates a good resolution. The ray density is relatively homogeneous except for computation volume boundaries where the ray density is lower.

A comparison of Figs 6 and 9 shows that model details are practically independent of the ray density and do not correlate with each other. Note that, the hit count parameter is merely an indicator of how the solution of the problem is resolved. We have therefore introduced another indicator, which quantifies the diversity of local ray direction. We calculated the directional vectors of ray segments crossing each cell. Next the components of these vectors were changed according to their signs to values ±1 and in this way
The 1-D WB2012 model was averaged along horizontal planes together with an estimation of the standard deviation (black error bars). The lateral variability of this model decreases with the depth. The S-wave velocities in the WB2012 are systematically lower compared to the WB2005 model, especially in the upper part.

Figure 7. Comparison of WB2005 (red) and 1-D representation of WB2012 (black) velocity models. The 1-D WB2012 model was averaged along horizontal planes together with an estimation of the standard deviation (black error bars). The lateral variability of this model decreases with the depth. The S-wave velocities in the WB2012 are systematically lower compared to the WB2005 model, especially in the upper part.

The values of corr, can be in the interval <−1; +1>. Perfect resolution corresponds to corr, = +1, nevertheless, we postulated all cases when corr, > 0.7 as acceptably resolved. We made the resolution analysis many times using various values of the checkerboard pattern size and shift. The 2-D functions showing the dependence of the corr, on the horizontal (L_x = L_y = L_z) and on the vertical (L_z) checkerboard sizes can be seen in Fig. 10. We used the same size in the x and y coordinate directions and averaged the results for all shifts (i_{off}, j_{off}, k_{off}) of the pattern. Cross-sections of a selected recovered pattern are shown as an example in Fig. S4. A set of examples more appropriately documenting the resolution analysis can be found at the web page www2012d.

As expected, the resolution of S waves is slightly worse than that of P waves. Only structures whose sizes are larger than ~15 km horizontally and ~6 km vertically can be considered as real, smaller objects must not be interpreted at all. Therefore, it is necessary for any geological interpretation of our WB2012 velocity model to take the model resolution into account. Let us emphasize that calculated part of the model. Consequently, indications provided by the hit count cross-sections can be considered as relevant.

To determine the true resolution of the WB2012 model, we applied the checkerboard test to our data. The slowness update is calculated in each iteration cycle via eqs (5) and (6). In the test we consider known perturbation pattern \( \Delta x^b \) and related time residual \( \Delta t^b \):

\[
\Delta x^b(l) = \sin \left( \frac{i + i_{off}}{L_x} \pi \right) \sin \left( \frac{j + j_{off}}{L_y} \pi \right) \sin \left( \frac{k + k_{off}}{L_z} \pi \right) \\
\Delta t^b = A \Delta x^b,
\]

where \( L_x, L_y, L_z \) are checkerboard pattern sizes along coordinate axes and \( i_{off}, j_{off}, k_{off} \) are pattern offsets, expressed in grid coordinates. Obviously, the slowness perturbation can reach only values in the range of \(<−1; +1>\). Next, the perturbation pattern \( \Delta x^b \) is transformed into \( \Delta x^b\omega \) using the contraction matrix \( C \) so as to exclude parameters corresponding to cells without rays; the time residual \( \Delta t^b \) is used for getting the appropriate solution:

\[
\Delta x^b\omega = C^T \Delta x^b \\
\Delta x^i = (A')^T A' + \lambda L_1^2 L_2^{-1} (A')^T \Delta t^b \\
= ((A')^T A' + \lambda L_1^2 L_2^{-1})^{-1} (A')^T A' \Delta x^b\omega = R \Delta x^b\omega \\
R = ((A')^T A' + \lambda L_1^2 L_2^{-1})^{-1} (A')^T A'.
\]

The matrix \( R \) in eq. (12) is the resolution matrix corresponding to the linearized system of equations including the regularization term. We have applied eq. (12) separately to both \( P \) and \( S \) waves using the same values of \( \lambda \) as in the last iteration of the inversion via eq. (5). In an ideal case the recovered pattern \( \Delta x^i \) should be equal to the contracted input pattern \( \Delta x^b\). In fact, some discrepancy is observed which follows from the non-uniqueness of the inverse problem; specifically, it depends strongly on the size of the checkerboard pattern \( (L_x, L_y, L_z) \) and weakly on its shift \( (i_{off}, j_{off}, k_{off}) \). The discrepancy between input and recovered checkerboard patterns can be quantified, for example, by the correlation:

\[
corr = \frac{\langle (\Delta x^i)^2 \Delta x^b\omega \rangle}{\sqrt{\langle (\Delta x^i)^2 \rangle \langle \Delta x^b\omega \rangle}} \\
= \frac{\langle (\Delta x^i)^2 \Delta x^b\omega \rangle}{\sqrt{\langle (\Delta x^i)^2 \rangle} \sqrt{\langle \Delta x^b\omega \rangle}} \\
= \frac{\langle (\Delta x^i)^2 \Delta x^b\omega \rangle}{\sqrt{\langle (\Delta x^i)^2 \rangle} \sqrt{\langle \Delta x^b\omega \rangle}}.
\]
the presented resolution analysis is based on the correlation, hence it expresses a measure of similarity but not equality. Such an approach, that is, classification based on similarity/dissimilarity, is fully consistent with the search for boundaries between different geological/petrological units. Absolute velocities (see previous section) must be interpreted together with their accuracy so as to avoid potential blunders.

7 PERFORMANCE OF THE 1-D WB2005 AND 3-D WB2012 VELOCITY MODELS

We tried to compare location results using the reference model WB2005 and the new model WB2012. In the first test, we localized a reduced set of earthquakes used in our tomographic computations with epicentres less than 20 km from the central station NKC (Nový Kostel). Localization of events from the margins or outside the seismic network are potentially unstable, therefore, these events were excluded. The results are summarized in Table 2 and presented as histograms in Fig. S5. Both models (WB2005 and WB2012) yield nearly the same horizontal coordinates of the hypocentres (the differences are of the order $\sim 10^2$ m, see Table 2), however, the differences in the vertical coordinates are more significant. Model WB2012 provides shallower hypocentres compared to WB2005; the difference is on average 0.826 km. The average depth of hypocentres is 8 km. So, the relative difference of depth due to alternate velocity models reaches 10 per cent, origin times in the model WB2012 are on average 0.111 s sooner. Obviously, we are treating the well-known trade-off between the hypocentre depth and origin time. The average rms of arrival times for both phases $P$ and $S$ reached 0.084 s with the WB2005 model and dropped down to 0.067 s when using the WB2012 model. So, the relative improvement is 20 per cent in favour of the WB2012 model but this test utilized a part of the data that were used previously for optimizing only the WB2012 model.

The next location test was based on the relocation of a voluminous set of 4,671 events, which occurred in the time period from 2008 August 4 to 2009 February 24, during the 2008 West-Bohemia earthquake swarm. This swarm has been analysed by several authors (Horálek et al. 2009; Fischer et al. 2010; Hiemer et al. 2012) who showed that hypocentres tend to cluster in a relatively small volume along the active fault plane. The amount of clustering along planar structures is usually understood as a measure of the localization quality in such cases. We localized the whole set of events in both WB2005 (Fig. 11a) and WB2012 (Fig. 11b) velocity models and calculated centroid coordinates for the whole set (see Table 3). It is...
Three-dimensional seismic velocity model

Figure 10. Two maps showing correlation between input and recovered checkerboard patterns for (a) $P$ waves and (b) $S$ waves. Horizontal axes, size of the pattern in horizontal directions (the same in $x$ and $y$ coordinates); vertical axes, size of the pattern in a vertical direction. White circles at [20 km, 8 km] indicate the position of the example given in Fig. S4. Recovered structures correlating with the true structures by $corr > 0.7$ are considered as well resolved.

Table 2. Localization differences due to velocity models.

<table>
<thead>
<tr>
<th></th>
<th>Mean difference</th>
<th>WB2012</th>
<th>WB2012A</th>
<th>WB2012B</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\langle x_w - x_r \rangle$ [km]</td>
<td></td>
<td>+0.275</td>
<td>−0.012</td>
<td>+0.062</td>
</tr>
<tr>
<td>$\langle y_w - y_r \rangle$ [km]</td>
<td></td>
<td>−0.143</td>
<td>−0.108</td>
<td>−0.278</td>
</tr>
<tr>
<td>$\langle z_w - z_r \rangle$ [km]</td>
<td></td>
<td>−0.826</td>
<td>−0.356</td>
<td>−0.445</td>
</tr>
<tr>
<td>$\langle T_w - T_r \rangle$ [s]</td>
<td></td>
<td>+0.111</td>
<td>+0.214</td>
<td>+0.119</td>
</tr>
</tbody>
</table>

It is apparent that an influence of the 3-D model to horizontal coordinates is negligible, while hypocentre depths in the WB2012 are on average by 0.663 km shallower compared to those in the WB2005 model, similarly as in the case of relocating the set of events included in the tomographic calculations. To quantify the effect of the velocity model on the hypocentre clustering, we calculated basic parameters of the best-fitting planes, which correspond to hypocentres localized in models WB2005 and WB2012. Both planes are nearly parallel, the angle between their normals is only 4°. The mean distance between individual hypocentres and the best-fitting plane is 0.3237 km and 0.3058 km regarding WB2005 and WB2012 models, respectively. Thus, from the perspective of event clustering the WB2012 model is also better. Finally, we have calculated traveltime residuals in both models, histograms of which are indicated in Figs S6a and S6b. The WB2005 model yields average rms 0.134 s, while WB2012 gives 0.116 s, that is, in value by 13 per cent lower. Again the WB2012 model is better. This localization test utilizes independent data, not used for optimizing both WB2005 and WB2012 models.

Figure 11. Hypocentres of the 2008 earthquake swarm localized using the WB2005 (left) and WB2012 (right) velocity models. The same projection is used in both cases. The application of the WB2012 velocity model results in enhanced clustering of hypocentres and upwards shift of hypocentres by 0.663 km.
Systematic differences in location results due to the application of different models (WB2005 or WB2012) are noteworthy; apparently, only two factors play an important role in the construction of these velocity models: amount of inverted data and geometry of the considered problem. For a deeper insight into the problem we performed several variants of the calculation and comment only on the two most important ones (WB2012A and WB2012B). We tried to eliminate the asymmetry between the amount of P- and S-wave data in these variants.

First, we have excluded all data of shots and inverted only P and S arrival times from earthquakes. As a result, the data set was reduced to 5929 P- and 6221 S-wave arrivals. The resulting model WB2012A is similar to the WB2012 model and also includes similar velocity anomalies. Distribution of the S-wave velocities is nearly identical to those in the WB2012 model; the P-wave velocities in the upper part of the model (depth down to ~4 km) are around 5 per cent lower, and the P-wave velocity field is slightly smoother compared to the WB2012 model. The location differences relative to the model WB2005 are again small as regards the horizontal coordinates but the depth of foci is systematically about 0.356 km shallower and origin time is systematically about 0.214 s later (see Table 2).

In the second variant (WB2012B), we extended the data set by fictitious S-wave onsets from shots, which cannot be reliably picked in real seismograms. We generated artificial readings \( t_{\text{sh}} = \sqrt{3} t_{\text{p}} \) for each P-wave onset from each measured shot. We balanced the contribution of the P and S waves but introduced an artificial constraint \( v_p/v_s \approx \sqrt{3} \) for the upper part of the model crossed by quasihorizontal rays from shots. This extended data set contained 8845 and 9138 readings of the P- and S-wave onsets, respectively. Adding artificial data may result in a non-physical velocity model from the inversion, however, a similar constraint was also used by Mälke et al. (2005) who calculated the model WB2005 that we used for comparison. The P-wave velocity fields in models WB2012 and WB2012B are almost identical. On the contrary, the S-wave velocities in the WB2012B and WB2012 differ: they are systematically higher in the WB2012B model but only in its uppermost part, that is, down to a depth of about 1 km. The location differences against the model WB2005 are summarized in Table 2. It is evident that the greatest difference is again in the depth coordinate and again the hypocentres are shallower. We have presented all three variants of the 3-D model on the web page www2012d: (1) WB2012—all available data used (both local earthquakes and shots); (2) WB2012A—only local earthquake data used, all shots excluded; (3) all available data used and artificial S-wave readings from shots added with the artificial constraint \( v_p/v_s \approx \sqrt{3} \). For most of the problems the selection of the model can be unimportant; in other cases, the suitability of a particular model should be checked individually. From the perspective of a deeper geological structure (\( z > \sim 2 \) km), all the 3-D velocity models are practically equivalent. The depth of earthquake foci may be slightly shallower than have been assumed until now. Shallower than standard hypocentres were suggested, for example, by Rössler et al. (2007) based on the moment tensor inversion.

In the two variants of computation we just described, we only modified P- or only S-wave data. Even if the P- and S-wave tomographic inversions (eqs 2–6) are mutually independent in one iteration cycle, the resulting models are in fact coupled. Changing only one part of the data (reducing the number of P-wave picks of explosions or adding artificial S-wave readings) results in changes of hypocentral coordinates of earthquakes due to relocation (including origin times), which automatically causes changes in both (i.e. P and S) models.

Different depths of hypocentres using models WB2005 and WB2012 may have further consequences. For instance, the rays and Green’s functions needed for the estimation of the moment tensor are different. It has been proved by means of synthetic data (e.g. Sileny 2004, 2009) that the velocity model influences moment tensor, especially its non-double-couple components. In reference to West Bohemia/Vogtland seismicity, the non-double-couple components are one of the key factors in a concept of the fluid-driven triggering of earthquakes. It is, therefore, evident that a comparative calculation of moment tensors using both models WB2005 and WB2012 is unavoidable. However, solving this problem is beyond the scope of this paper.

Finally, we tried to ‘localize’ the shots used in model calculations so as to compare the WB2005 and WB2012 models by checking the differences between true and calculated shot coordinates. Unfortunately, shot data are not always suitable for hypocentre localization due to bad shot-station configurations. Nevertheless, we performed the localization test for the whole set of 51 shots which provided at least five P-wave onsets regardless of the measurement geometry since a poor azimuthal distribution of stations was the same if either the WB2005 or WB2012 model was used.

Epicentre mislocation achieved by using the newer WB2012 model is displayed as a histogram in Fig. S7a. Most epicentres are localized with accuracy below 2 km; mean deviation is 1.128 km. The goal of this test was to compare localization errors of the WB2012 model relative to the WB2005 model. For each shot we have evaluated the relative error \( \Delta r \) of epicentre localization in WB2012 versus WB2005 models. It can be seen in Fig. S7b that \( \Delta r > 1 \) holds in 35 per cent of the cases (a smaller localization error if WB2005 used), while \( \Delta r < 1 \) holds in 64 per cent cases (smaller localization error if WB2012 is used). On the basis of this analysis the model WB2012 is favoured, but this test concerns P waves only.

**8 BULK MODULUS AND POISSON’S RATIO**

A straightforward way to compare the seismic model and geological structure is to examine the P- and S-wave propagation velocities. This technique is not usually the best because \( v_p \) and \( v_s \) as a rule, are strongly correlated. Much better information is provided, for example, in the doublet of \( v_p \) and \( v_s/v_p \). Nevertheless, parameters directly controlling the deformation response to a stress impulse give the best quantitative characteristic of the geological medium. Such parameters, which can be derived from seismic measurements and
which define the mechanical properties in a framework of isotropic models, are the bulk modulus \( K \) and the Poisson’s ratio \( \nu \):

\[
K = \rho \left( a^2 - \frac{4}{3} \beta^2 \right), \quad \nu = \frac{\lambda}{2(\lambda + \mu)} = \frac{\frac{\nu^2}{2} - 1}{\frac{\nu^2}{2}}.
\] (14)

Bulk modulus is a measure of rock resistance to being compressed and is defined as volumetric stress over volumetric strain. Poisson’s ratio is defined as the ratio of transverse contraction strain to longitudinal extension strain in the direction of the applied load, thus in some sense a measure of liquidity. An ideal liquid is characterized by \( \nu = 0 \), an ideal elastic body that resists the normal stress with no shear deformation is characterized by \( \nu = 0.5 \). The bulk modulus is calculated for the density \( \rho = 2700 \text{ kg m}^{-3} \). In what follows, only Poisson’s ratio and bulk modulus are discussed. Another popular form of the model characterization is the \( \nu/\nu_{WB2012} \) ratio available online as cross-sections at www2012d.

The horizontal cross-sections \( K = K(x, y) \) and \( \nu = \nu(x, y) \) for the depth level of \( z = 3 \text{ km} \) are shown in Fig. 12 (maps a and b). This depth provides most of the information regarding seismic wave velocities, from which \( K \) and \( \nu \) are calculated. Only parts of cross-sections constrained by data are drawn, similar to the velocity maps in Fig. 4. The slope of the bulk modulus surface \( \text{i.e. } \text{grad}(K) = (\text{d}K/\text{d}x) + (\text{d}K/\text{d}y) \) is constructed by means of the edge detection tool in the Grass GIS software (www2012e), is depicted in Fig. 12(c). A simplified geological map with basic tectonic lines is presented for comparison in Fig. 12(d).

It can be recognized from Fig. 12(c) that boundaries between regions with a different value of bulk modulus are to a great extent tectonically conditioned. This is clearly visible by comparing the \( \text{grad}(K) \) with the faults. Dark colours correspond to high values of the bulk modulus slope and thus can be indicative of edges of the bulk modulus surface. Note that, the directions of ‘edges’ correspond quite well to northern part of the MLF system (MLF, azimuth \( \sim -30^\circ \) NW–SSE) and the Eger rift system (EGR, azimuth \( \sim +60^\circ \) EEN–WWS). Some of dark linear segments can be identified with particular faults. However, it should be emphasized that geological map and faults reflect only information along the Earth’s surface.

On the other hand, the map of Poisson’s ratio (Fig. 12b) correlates much less with the geological/tectonic setting but shows two major negative anomalies with \( \nu \sim 0.14 \) (the standard value of \( \nu = 0.25 \) corresponds to the standard \( \nu/\nu_{WB2012} = \sqrt{3} \) ratio). Both these anomalies lie in seismically active regions: \( \sim[12.50E, 50.25N] = \text{Novy Kostel focal area}, \sim[12.60E, 50.10N] = \text{Lazy focal area} \). The lower values of \( \nu \) coincide with the increased ability of rock to a brittle failure (the higher the value of \( \nu \) the closer the medium is to a fluid material, which cannot generate seismic events). It may imply that abrupt changes of the bulk modulus are linked to tectonic lines and the low Poisson’s ratio is a characteristic feature of the seismogenic volume.

9 CONCLUSIONS

The \( W2012 \) is the first 3-D \( P \)- and \( S \)-wave velocity model of the upper crust in the West Bohemia/Vogtland earthquake swarm region. This model was derived by means of the tomographic inversion of 11 339 \( P \)- and \( S \)-wave arrival times from local earthquakes and of 2920 \( P \)-wave traveltimes from controlled shots. Since 1-D models are beneficial in many seismological applications, we also constructed a 1-D simplification of the \( W2012 \) model by averaging the velocity field along horizontal planes. The \( W2012 \) model was properly tested for resolution by applying the checkerboard test and for the localization ability by localizing local earthquakes (including the 2008 swarm) and controlled shots; resultant locations were compared with those obtained using the 1-D velocity model \( WB2005 \) by Mälek et al. (2005), which hitherto had been used as a standard velocity model by the WEBNET group and also by other researchers using WEBNET data. Thanks to independent \( P \)- and \( S \)-wave velocities in the \( W2012 \) model, we were able to calculate bulk modulus \( K \) and Poisson’s ratio \( \nu \) and in this way get an essential idea of mechanical properties of the medium in the upper crust in the region in question.

Basic parameters of the \( W2012 \) model and the most important outcome of this paper can be summarized as follows:

1. The model includes most of the earthquake swarm area of West Bohemia, above all the main Nový Kostel focal zone and is relevant down to 10 km.
2. The \( P \)-wave velocities in the \( WB2005 \) model and in a 1-D approximation of the \( W2012 \) model are nearly the same while \( S \)-wave velocities in the \( W2012 \) are systematically lower, mainly in its upper part, in depths between 0 and 4 km. Relative differences between \( P \)- and \( S \)-wave velocities in the 1-D model \( WB2005 \) and 3-D model \( W2012 \) vary in the range of \( \pm 15 \) per cent. The notable feature is lowered values (negative differences) in the central part of the model, that is, beneath the main Nový Kostel epicentral area. This feature may be due to a fractured rock in the earthquake swarm volume.
3. Localizations of individual earthquakes using the \( W2012 \) model yield lower arrival time residuals compared to those using the \( WB2005 \) model (by 13 per cent on average); horizontal coordinates of the hypocentres in both \( WB2005 \) and \( WB2012 \) models are nearly the same, however, differences in the vertical coordinates are significant: hypocentres in \( WB2012 \) are by 600–800 m shallower compared to the hypocentre depths in the \( WB2005 \) model (which corresponds to \( \sim 10 \) per cent of focal depths of local swarm events). Localization ability of surface events (shots) is similar in both models.
4. The checkerboard test indicates the resolution of the tomographic \( W2012 \) model. The model is of a significant importance for geophysical purposes but any interpretations of crustal structures should be undertaken with caution. Only structures whose sizes are larger than \( \sim 15 \text{ km} \) and \( \sim 6 \text{ km} \) in horizontal and vertical directions, respectively, can be considered as real; smaller objects may not be interpreted. Nevertheless, the basic structures, particularly the MLF, are very well recognizable in the \( W2012 \) model.
5. The bulk modulus \( K \approx 4070 \text{ GPa} \) correlates rather well with the tectonic and geological structure of the region in question; boundaries between areas with different \( K \) are to a great extent tectonically conditioned; directions of edges of the \( K \)-field correspond well to the MLF and to the Eger rift system. Poisson’s ratio \( \nu \approx 0.15 \) correlates much less with the geological/tectonic setting but two major negative anomalies with \( \nu \approx 0.14 \) clearly correspond to most active focal zones, Nový Kostel and Lazy, which may indicate increased ability of rock to brittle failure.

The primary aim of this paper is to provide the new \( W2012 \) model to the broader scientific community and to initiate its verification from various geophysical/geological aspects, in this way accomplishing its improvement. The model in the numerical form can be found on the web page (www2012d).
Figure 12. The sequence of maps displaying (a) bulk modulus in MPa, (b) Poisson’s ratio and (c) slope of the bulk modulus map highlighting the edges. A simplified geological map is given in (d). All the maps are drawn in the same geographical scale and include significant faults (thick black lines) and focal zones as follows: MLF, Mariánské Lázně fault; EGR, Eger rift; NKC, Nový Kostel focal zone; LAC, Lazy focal zone. The Czech–German border is marked by a light blue line. Cross-sections in maps (a)–(c) refer to the depth level of 3 km. Geological data in map (d) were downloaded from www2012f.
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REFERENCES


SUPPORTING INFORMATION

Additional Supporting Information may be found in the online version of this article:

Figure S1. A statistical summary of data in graphical form, (a) histogram of epicentral distances; (b) histogram of ray azimuths. Most rays are ∼12 km long and they have a nearly uniform azimuthal distribution.

Figure S2. 3-D perspective views of one of the starting models. Colour scales for P and S waves are different, units are [km s⁻¹].

Figure S3. Relative differences of the WB2012 model related to the standard WB2005 velocity model. Arrangement of the maps is the same as that in Fig. 4.

Figure S4. An example of a reconstructed checkerboard pattern the size being 20 × 20 × 8 km (for its position refer to Fig. 10). The projections are the same as in Figs 4 and 5, etc. Here only typical cross-sections are shown, but the appropriate correlations.
were calculated inside the whole 3-D volume crossed by rays. The geometry of the true checkerboard pattern is indicated by dashed lines.

**Figure S5.** Histograms of differences between hypocentral coordinates obtained by localization of a subset of earthquakes close to the NKC station in 1-D *WB2005* and 3-D *WB2012* velocity models. All histograms are normalized, thus the vertical scales are relative only. Horizontal scales are in [km] in all cases except for (d) which is in [s]. (a) *x*-coordinate, mean difference 47 m; (b) *y*-coordinate, mean difference 77 m; (c) depth, mean difference 951 m; (d) origin time, mean difference 0.077 s.

**Figure S6.** Normalized histograms of traveltime residuals, from localizing an independent set of 4672 earthquakes occurring during the time period 2008 August 4 to 2009 February 24, (a) model *WB2005*—mean residual 0.134 s; (b) model *WB2012*—mean residual 0.116 s. Application of the *WB2012* model results in lowering the average rms by 13 per cent.

**Figure S7.** The results of trial localizations of shots. (a) Histogram of the epicentre mislocations if the model *WB2012* was used. (b) Comparison of the shot-localization accuracy if using either *WB2005* or *WB2012* velocity models. Shots better localized in model *WB2005* correspond to symbols above 1; those better localized in model *WB2012* correspond to symbols below 1 (http://gji.oxfordjournals.org/lookup/suppl/doi:10.1093/gji/ggt295/-/DC1).

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