

Attenuation tomography in West Bohemia/Vogtland

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ABSTRACT

We present a three-dimensional (3-D) *P*-wave attenuation (*Q_p*) model for the geodynamically active swarm earthquake area of West Bohemia/Vogtland in the Czech/German border region. **Path-averaged attenuation *t*^{*} is calculated from amplitude spectra of time windows around the *P*-wave arrivals of local earthquakes.** Average *t*^{*} value or *Q_p* for stations close to Nový Kostel are very low (<150) compared to that of stations located further away from the focal zone (increases up to 500 within 80 km distance). The SIMUL2000 tomography scheme is used to invert the *t*^{*} for *P*-wave attenuation perturbation. Analysis of resolution shows that our model is well-resolved in the vicinity of earthquake swarm hypocenters. The prominent features of the model are located around Nový Kostel focal zone and its northern vicinity. Beneath Nový Kostel a vertically stretched (down to depth of 11 km) and a highly attenuating body is observed. We believe that this is due to fracturing and high density of cracks inside the weak earthquake swarm zone in conjunction with presence of free gas/fluid. Further north of Nový Kostel two highly attenuating bodies are imaged which could represent fluid channels toward the surface. The eastern anomaly shows a good correlation with the fluid accumulation area which was suggested in 9HR seismic profile.

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1. Introduction

The West Bohemia/Vogtland region is located at the border of Czech Republic and Germany. It is of great interest as it exhibits ongoing geodynamical activities. Main activities in West Bohemia are related to diffuse CO₂ degassing in the form of CO₂-rich mineral waters or wet and dry mofettes (e.g., Bräuer et al., 2003; Geissler et al., 2005), Quaternary volcanoes and earthquake swarms (Fischer et al., 2014). Although fluids rising from deep crustal root zones are considered as the main reason for inducing earthquake swarm activities (e.g., Špičák and Horálek, 2001; Weinlich et al., 1998; Heinicke and Koch, 2000; Weise et al., 2001; Bräuer et al., 2005, 2009; Kämpf et al., 2013), the relation between earthquake swarms and CO₂ degassing is still a matter of debate (Babuška et al., 2016). The main tectonic features of West Bohemia including Cheb basin, Eger rift and Mariánské Lázně fault zone (MLF) as well as main mofettes (Soos, Bublák and Hartoušov), gas vents and Quaternary volcanoes (Komorní Hůrka, Železná Hůrka and Mýtina) are shown on the map in Fig. 1.

This study focuses on seismic attenuation in the West Bohemian massif and thus complements previous geophysical investigations. Seismic attenuation is the loss of energy of seismic waves passing through a

medium (e.g., Aki and Richards, 1980) and is the inverse of the quality factor *Q*. Many studies have shown that seismic attenuation is a parameter sensitive to fluid content and some significant rock properties which do not have large impact on seismic velocities (e.g., Karato, 1993; Jackson et al., 2002; Faul and Jackson, 2005; Aizawa et al., 2008), thus it sheds new lights on our existing models of *P*- and *S*-wave velocities (Mousavi et al., 2015). Seismic waves are attenuated due to anelastic and elastic processes. Geometrical spreading, multipathing and scattering are elastic processes in which energy is attenuated along the propagation path. In contrast, during anelastic processes the energy of a propagating wave converts into heat (e.g., Stein and Wysession, 2003) or permanent deformation. Theoretical and experimental studies have shown that a combination of both, intrinsic (anelastic) and scattering (elastic) attenuation is responsible for the total seismic attenuation but it is often difficult to identify the dominant mechanism (Arevalo et al., 2003).

Intrinsic attenuation is considered to be caused by a variety of mechanisms. It is mainly a result of movement along minerals, frictional sliding on grain boundaries in a fluid-saturated elastic solid, vibration of dislocations, and viscous damping from local pore fluid motion (Berryman and Wang, 2000; Winkler and Murphy, 1995). The attenuation effect of these factors is different. Intrinsic attenuation of minerals has a relatively small effect whereas viscous dissipation (Mavko et al., 1979; Walsh, 1995), the presence of water, temperature and frictional

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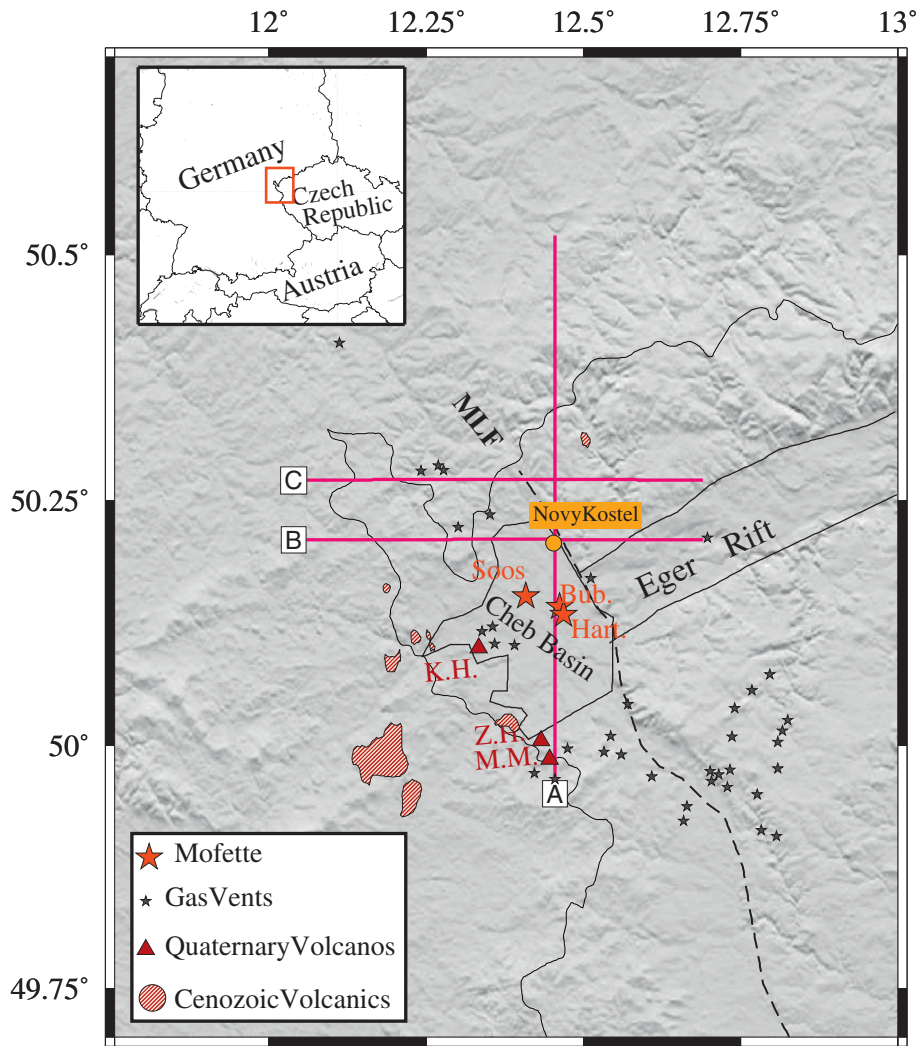


Fig. 1. Map of the study area showing the W-Bohemia/Vogtland region in the western part of Czech Republic, close to the German border. Grey background shading represents topography. Eger Rift, Cheb basin, Márianské Lázně Fault (MLF), mofettes, gas vents, Cenozoic volcanic zones, Quaternary volcanoes including Komorní Hůrka (KH), Železná Hůrka (ZH) and Mýtina (MM) and Main mofette fields inside Cheb basin including Hartoušov mofette (Hart.) and Bublák mofette (Bub.) and Soos (marked by red stars) are shown in the map. Location of three vertical tomographic sections A, B and C are shown with pink lines.

sliding between grains, are considered to have higher impact on attenuation (e.g., O'Connell and Budiansky, 1974; Johnston et al., 1979; Sato et al., 1989; Peacock et al., 1994; Winkler and Murphy, 1995). Thus, the intrinsic attenuation is strongly dependent on the permeability (Biot, 1956a,b). Intrinsic attenuation of seismic waves could even be caused by a very small amount of fluid (Mavko and Nur, 1979) which does not show up in high electrical conductivity. Intrinsic attenuation is often directly related to crack density and the presence of a small percentage of free gas in the medium (e.g., Trehu and Flueh, 2001).

In a recent study, Gaebler et al. (2015a) estimated the quality factor for elastic S-wave attenuation in West Bohemia around 700. Later, Gaebler et al. (2015b) determined scattering and absorption parameters for West Bohemia by inversion of envelopes with radiative transfer theory. They found that intrinsic attenuation dominates over scattering attenuation in this region for frequency bands from 3 to 24 Hz. Bachura and Fischer (2016) determined Q_c (coda attenuation) and measured the amount of intrinsic and scattering attenuation. They also concluded that intrinsic attenuation is dominant in the frequency range 1–18 Hz, and that intrinsic attenuation in West Bohemia is higher than that of neighbouring areas in Germany. In this paper, we present the first detailed 3-D P-wave attenuation model around the swarm quake region

in West Bohemia. Attenuation tomography provides independent constraints on derived anomalies in seismic velocities and V_p/V_s ratios (Mousavi et al., 2015). As such, using seismic attenuation as an imaging tool identifies spatial variations in fluid behaviour in Nový Kostel and provides further insight into the earthquake swarm focal zone.

2. Method

The source and attenuation parameters for an individual ray path from earthquake swarm to receiver can be determined by relating the observed spectrum of each station to its attenuation, source parameters, geometrical spreading, and site response (e.g., Sanders, 1993) as:

$$A_{ij}(f) = S_j(f) I_i(f) R_i(f) G_{ij} \exp(-\pi f t_{ij}^*) \quad (1)$$

where A_{ij} is the observed body wave spectrum of the j^{th} earthquake recorded at the i^{th} station, f is frequency, S_j is the source spectrum, I_i is the instrument response, R_i is the receiver site effect, G_{ij} is the geometrical spreading, t_{ij}^* is the path-averaged attenuation value. Geometrical spreading is considered to be independent from frequency and only depends on the distance between the source and the receiver.

The ω^2 source spectrum $S_j(f)$ can be described as follows (Scherbaum, 1990):

$$S_j(f) = \frac{\Omega_{0j}}{1 + (f^\gamma/f_{cj}^\gamma)} \quad (2)$$

where f_{cj} is the corner frequency, Ω_{0j} is the spectral moment, and γ is the source spectral fall-off (Lees and Lindley, 1994).

Inserting Eq. (2) into Eq. (1) yields:

$$A_{i,j}(f) = \frac{\Omega_{0'i,j} \exp(-\pi f t_{i,j}^*)}{[1 + (f/f_{cj})^\gamma]} \quad (3)$$

where $\Omega_{0'i,j}$ is the signal moment including all frequency-independent amplification factors (the spectral moment, geometrical spreading, instrument gain and first order – i.e. frequency independent – site amplification) (e.g., Haberland and Rietbrock, 2001). Assuming a Brune source model with $\gamma = 2$ (Brune, 1970, 1971; Hanks and Wyss 1972), we solve Eq. (3) for the free parameters f_{cj} , $\Omega_{0'i,j}$ and $t_{i,j}^*$. This is performed by iterative damped least square inversion jointly for all observations of an earthquake and involving a grid search over a range of f_c values by minimizing the differences between observed and calculated spectra.

Once $t_{i,j}^*$ values are determined, they can be inverted for the Q_p structure. The $t_{i,j}^*$ is related to $V_p(x,y,z)$ (P -wave velocity) and $Q_p(x,y,z)$ along the ray path (Hough et al., 1988). The local site effect can be described by a station constant t_0^* that was included in the inversion (Anderson and Hough, 1984).

$$t_{i,j}^* = t_0^* + \int \frac{1}{Q_p(x,y,z)V_p(x,y,z)} dr \quad (4)$$

where dr is an element along the raypath.

3. Data analyses and t^* inversion

Several temporary and permanent seismic networks have been deployed in West Bohemia to study crustal structures and geodynamic processes. For this study we analyse local earthquake data collected during 2000–2010 from permanent networks of Czech Republic (WEBNET and KRASNET) and Germany (BAYERNNETZ, SXNET and Thuringian network), temporary local network PASSEQ (Wilde-Piórko et al., 2008) and temporary stations of GFZ and University of Potsdam.

For the attenuation study we selected stations with a sampling frequency of 100 samples per second (sps) or higher. Data with a sampling frequency higher than 100 sps were down-sampled to 100 sps. Stations used in this study have a flat response in the examined frequency band, 1–30 Hz, hence we did not correct for instrument response.

The data were then fed into a spectral fitting routine following Haberland and Rietbrock (2001). The robustness of this method has been shown in a number of studies (e.g., Eberhart-Phillips and Chadwick, 2002; Haberland and Rietbrock, 2001; Rietbrock, 2001; Schurr et al., 2003; Muksin et al., 2013; Bohm et al., 2013). Furthermore, Eberhart-Phillips et al. (2008) showed that t^* values of locally recorded, high-frequency earthquakes with and without the frequency-independent assumption, are similar. On each waveform, we first selected a 2.56 s-window (256 samples) around P -wave arrivals to calculate the signal spectrum using the multitaper approach (Park et al., 1987). The same was performed for a time window preceding the selected P -wave window to obtain the noise spectrum (Fig. 3).

The selection of an event requires two criteria: 1) S-P travel-time difference larger than 1.28 s to avoid contamination of the P spectrum with S-wave signals, and 2) a minimum allowable signal to noise ratio of 2 over a continuous frequency range from 1 to 30 Hz. Considering these restrictions we have 168 events recorded at 63 stations providing

2061 calculated t^* values in an area within longitude 11° to 13.5° E and latitude 49° to 51.05° N. Due to lack of intersecting rays within the area (lots of parallel rays) we decided to exclude far stations to avoid extra smearing and projection of anomalies from broader areas into the focal zone for the purpose of the tomographic inversion. In this way, we come up with 163 events recorded at 45 stations resulting in 1602 calculated spectra. Fig. 2 gives the location of events and stations used for attenuation tomography.

To include the quality of each individual calculated spectra, the inversion method deals with relative weighting of the data and not with the absolute values of the data errors. The weights are calculated from the RMS-values of the spectral fits and classified into 5 classes (0 for full weight and 4 for zero weight).

Site effects usually show up as characteristic distortions of the spectra (e.g., spectral exaggerations or holes). Therefore, it will in every case yield a poor spectral fit and therefore higher weights in the inversion. Fig. 3. shows an example of fitting t^* for one event where the recording station MROB is affected by site effect.

The t^* or path-averaged Q_p values of an event that occurred during the swarm 2008 recorded by each station is shown in Fig. 4. A systematic trend can be observed; high path-averaged attenuation ($Q_p < 150$) at stations close to the epicenter and less path-averaged attenuation (Q_p larger than 500) further away. This points toward increased attenuation in the immediate neighbourhood of the earthquake swarm area, whereas the broader region is known to have high Q_p . This observation agrees with the results of Bachura and Fischer (2016).

The uncertainty of t^* values can be estimated by analysing neighbouring ray paths for closely located events (possibly with different source parameters, i.e., different f_c) recorded at the same station (for which we expect similar values of t^*). To derive t^* uncertainties, we divide the upper crust into $500 \times 500 \times 200$ m³ cells. The variation of t^* values generated from earthquakes within each cell (only those which contain more than 2 events) and recorded at a common station are calculated relative to the mean value (Fig. 5a). The t^* values show more or less consistent deviation relative to their averages. Histogram of t^* value variation along common ray paths relative to their averages reveals a standard deviation of ~ 0.005 s (Fig. 5b). We assume that this value is representative for the uncertainty of our t^* data set. The similar t^* values for the neighbouring rays which originated from closely located but different sources, indicate that the spectral inversion method has been able to successfully separate the source and the path effects.

4. Attenuation tomography

The determined t^* values, can be used to estimate the 3-D Q_p structure. For this purpose, the modified version of SIMUL2000 for Q_p inversion (Thurber, 1983; Eberhart-Phillips and Michael, 1998; Rietbrock, 2001) is used. SIMUL2000 uses approximate ray-tracing and pseudo-bending to perform ray path calculation and solving the forward problem (Thurber, 1983). The inverse problem is solved by an iterative, damped least square method for model perturbation to minimize the residuals ($t_{\text{obs}}^* - t_{\text{cal}}^*$), where t_{cal}^* is determined from the initial model or the model from the proceeding iteration respectively. Velocity and hypocenters are held fixed and t^* values are inverted for Q_p . According to Eq. (4), the studied area is parametrized by Q_p and V_p values in a 3-D grid node volume where we take the P -wave velocities from the recent tomography study by Mousavi et al. (2015). Q_p varies continuously in all directions with cubic B-spline interpolation between the surrounding nodes. The distance between grid nodes should be selected to yield a fair compromise between robustness of the inversion and reasonable resolution. After intensive tests for different grid sizes, considering stations, events and ray paths, location of grid nodes are set to be identical to that of the traveltime tomography (Mousavi et al., 2015). The difference here is that the outermost nodes have larger spacing to cover the distant earthquakes and stations (Fig. 2). Inversion tests with slightly different grid spacing showed more or less similar results.

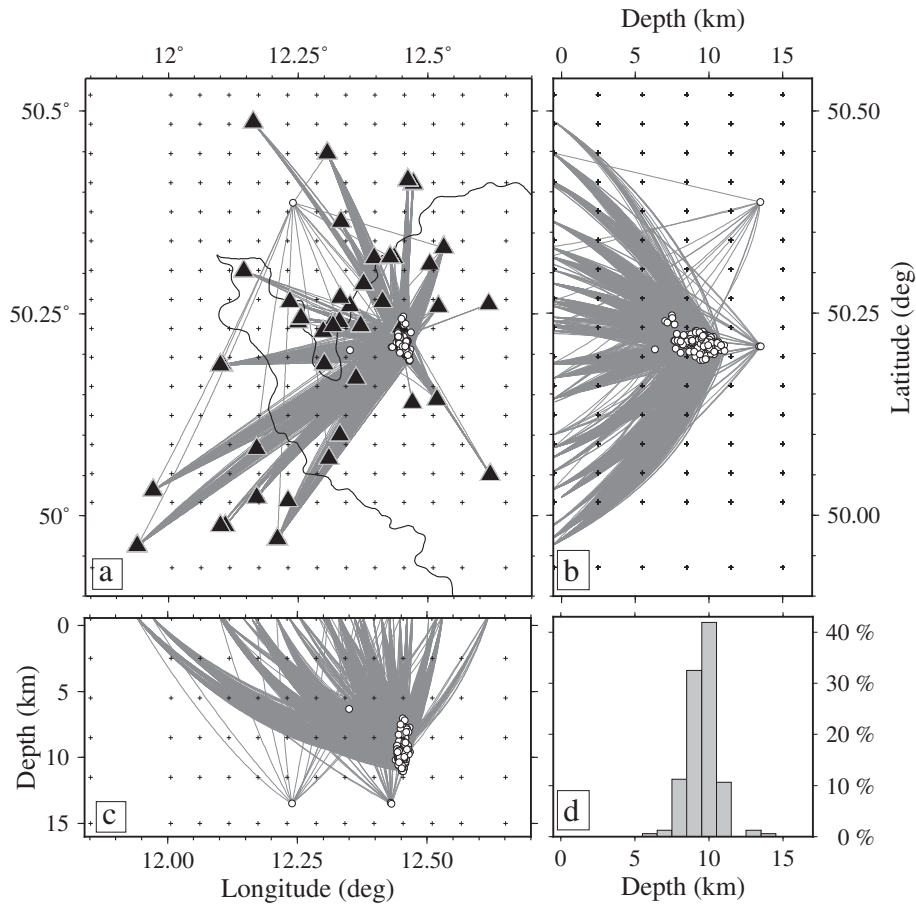


Fig. 2. Map view and cross sections of distribution of stations (black triangles) and events used in attenuation tomography (white circles). Ray coverage are shown by grey lines. The model grid nodes are indicated by black crosses. Histogram plot shows the depth distribution of earthquakes.

There is a trade-off between the uncertainty and model resolution (Menke, 1989). Obtaining a better resolution by lowering the damping values causes higher uncertainties in the results and vice versa. Therefore, a proper choice of the damping parameter is needed to stabilize the inversion process. The optimal damping for Q_p is selected by running a series of inversions with a large range of damping values and plotting the data misfit versus model variance known as the L-curve (Fig. 6) (Eberhart-Phillips, 1986). The damping value of 0.003 is a good compromise between data misfit and model variance and is used for the final attenuation inversion. Different damping values in the inversion yield very similar pattern of anomalies and it only changes the amplitude of anomalies.

It is well known that tomography models are highly dependent on the initial model (Kissling et al. 1994). We search for an optimum Q_p value in a uniform half space model within a wide range of Q_p between 100 and 1000. The resulting 3-D tomography models generally show similar pattern of anomalies, independent from the initial constant Q . To obtain a reliable Q_p model, we chose the average value of all t/t^* ($Q_p = 230$) as the initial model while this average value best explains the data.

5. Model resolution tests

A reliable interpretation of a 3-D velocity model is only possible when an estimation of resolution power is available. Resolution estimation can be divided into two categories; mathematical assessment and synthetic tests. Considering both measurements enables us to distinguish artefacts from well-resolved and robust features of the tomographic model. In this study, we examine the model resolution matrix

and two synthetic recovery tests; (1) checker-board test and (2) characteristic model. Regarding that no single resolution test can completely assess the differences between well-resolved and poorly resolved regions of a model, a good combination of several tests specifies which parts of the model are robust and believable.

5.1. Synthetic tests

The most commonly used synthetic test is “checker-board test” where an alternating pattern of high- and low-perturbations of a background model forms the synthetic model.

The advantage of this set up is that it provides information on spatial smearing and resolution for each grid nodes, since intensity and size of anomalies are equal, just with opposite sign (Eberhart-Phillips and Michael, 1998). However, it has been shown that this test is unable to retrieve complex geometries. Furthermore, in this test, the good recovery of small structures does not mean that larger structures similarly have to be well resolved (Leveque et al., 1993). Hence, it has become a common practice to try out “characteristic model” tests in which a synthetic model is constructed similar to size and shape of the anomalies obtained with “real” data (e.g., Husen et al., 2000). The purpose of a characteristic model is to test whether features comparable to those observed in the real model are recoverable.

For the checker-board tests, the background model $Q_p = 230$ (according to the initial model for the real inversion) is perturbed by $\pm 60\%$ (similar to that of recovered from real data), for the low and high Q_p anomalies across 3×3 grid nodes laterally and alternating at 2 grid nodes vertically. Synthetic t^* values are computed through the checker-board model for the same distributions of hypocentres, station

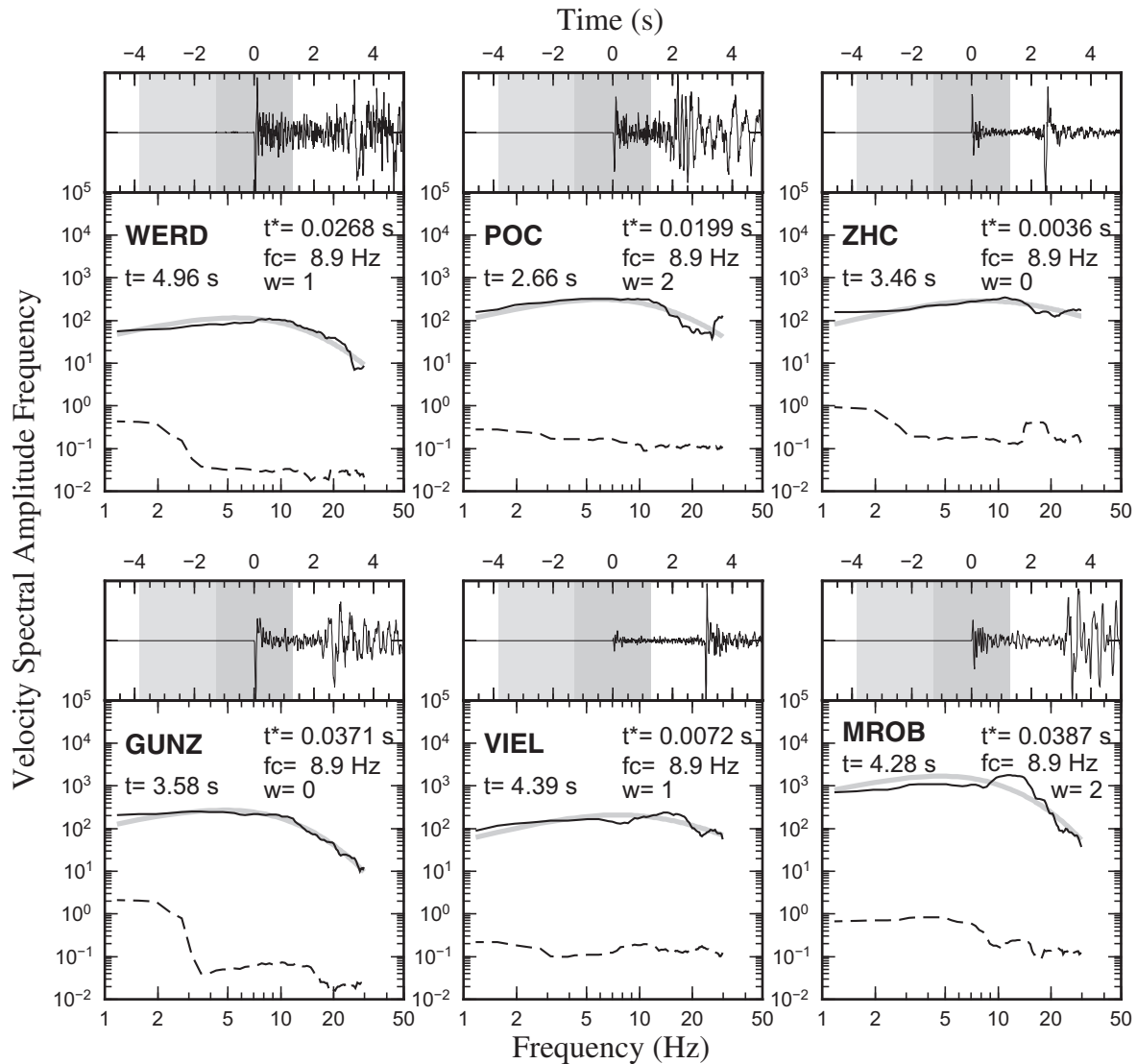


Fig. 3. The calculation of t^* values using velocity spectral analysis of six records (stations WERD, POC, ZHC, GUNZ, VIEL and MROB) of an event located in Nový Kostel occurred on 2008-10-12T07:44:56. t is traveltime, w is the weight representing the quality of the waveforms. f_c is the source corner frequency which in this event has value of 8.9. The top panels show the normalised time-series. The signal is within the darker shading windows following the noise in the lighter shading windows. The bottom panels show the noise spectrum (dashed lines), signal spectrum (black lines) and estimated spectrum resulting from spectral inversion (light-bold lines). Stations location can be found in Fig. 4.

locations, and t^* ray paths as the real data. In order to simulate similar conditions as in the data, Gaussian noise with a standard deviation of 0.005 s equivalent to the estimated average uncertainty of the real t^* values (Fig. 5b) is added to the synthetic t^* values. Map views and cross sections of the inverted Q_p models are shown in Figs. 7 and 8. The checker-board blocks located in the centre are recovered and are considered as 'well resolved'. The best recovery is found at Nový Kostel owing to the intense seismic activity. The western edges of the study area are poorly resolved due to the lack of crossing rays. The resolution at very shallow and deep layers is limited. Cross sections (Fig. 8) along profiles B and C are also provided to show vertical smearing. In profiles B and C at longitude range 12.2° to 12.4° and in profile A from latitude 50.17° to 50.30° the model is generally well resolved.

The characteristic model test works with similar principles. Inspired by the tomography of the real data, our characteristic model includes three patches of highly attenuated (low Q) anomalies in Nový Kostel and its northern neighbourhood down to depth of 5.5 km, which then continues down to depth of 8.5 km (concentrated within the swarm focal zone) while the size gradually decreases. The recovered model at 2.5 km and 5.5 km depth (Fig. 7, middle panel) shows some strong

patches of smearing (marked by black arrows). Recovery in cross-sections is fairly good although the effect of smearing is still visible. We carefully consider these recovered tomograms when it comes to interpretation of real data.

To test if there is any bias toward the recovered anomalies, e.g., due to ray geometry we performed two extra synthetic tests, one using a homogeneous model as input and the other same as the illustrated characteristic test in Fig. 7 but with opposite polarity. The output of the first test recovers a homogeneous model at depths deeper than 2.5 km. The result is a bit patchy at depth 2.5 and shallower which was already seen in the previous characteristic test. The second test, with opposite polarity confirms that our characteristic test is not biased toward low Q_p .

5.2. Spread function

The spread function introduced by Toomey and Foulger (1989) is a realistic and efficient way for representing model resolution. The Resolution matrix is an m by m matrix (where m is the number of model parameters) which describes how well each node is resolved relative to

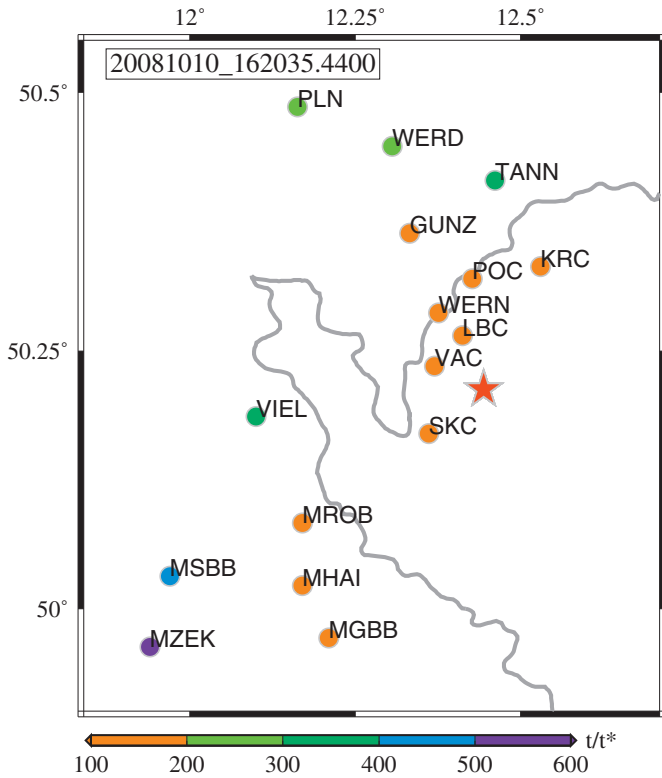


Fig. 4. Displaying t/t^* or Q values of all observing stations of an event (marked by star) occurred in swarm 2008 in Nový Kostel. Date and time of event is indicated on the top-left side of the panel.

the others. In an ideal situation this matrix would be an identity matrix, indicating that the results at each node is resolved independently from neighbouring nodes. The spread function converts the information contained in each row of the resolution matrix into a single value which describes the resolution for that node. A large spread value means that the result is smeared over a number of nodes and therefore the resolution is low. Low spread values are caused by larger diagonal elements and consequently a peak resolution. Due to the dependency of spread function to the chosen damping value and grid node spacing, there is no universally acceptable value for it.

Spread values of the real data for three depth slices together with the indicated cross-sections are shown in Figs. 7 and 8. The cross sections clearly show that spread values in the swarm region and neighbouring areas are low indicating a good resolution. In shallow layers, where rays diverge toward the location of stations (subparallel ray paths), spread values increase.

Putting all resolution tests together and comparing well resolved parts of the checker-board test with the characteristic test and spread function values helps us to define areas of low resolution and detect possible artefacts. In this study, as the conditions for the tomography are not ideal (source/receiver geometry in conjunction with the presumed structure/anomalies), we decided to be conservative in choosing the acceptable value of spread function. Areas with spread function values lower than 2 are considered as well-resolved and values between 2 and 3 are considered to be moderately resolved with smearing. For real data presentation (Section 6), the moderately resolved parts will be shown with faded intensity and non-resolved parts are suppressed.

6. Results

The results of 3-D tomographic inversion for Q_p including three map views at grid nodes 2.5, 5.5 and 8.5, and three cross-sections (A, B and C)

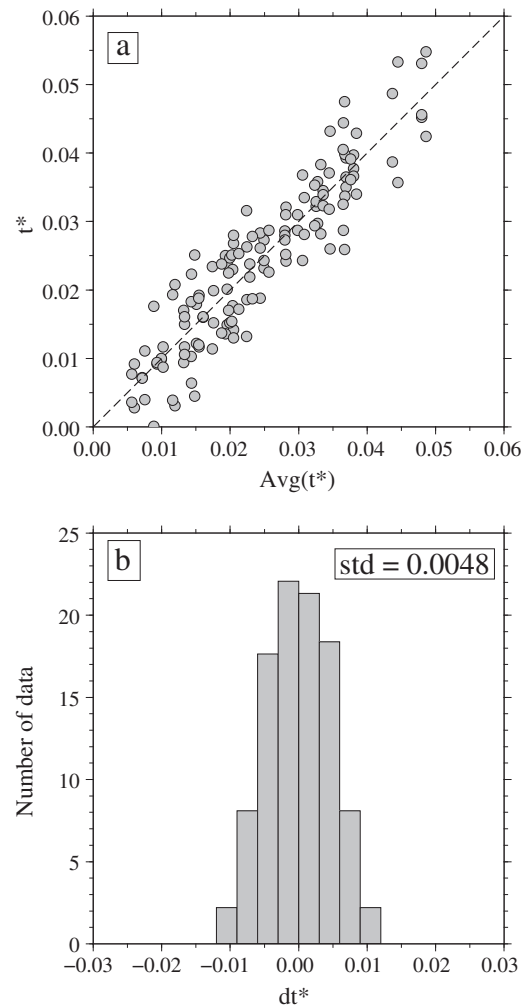


Fig. 5. Estimating the uncertainty of the derived t^* values for closely located events observed at a particular station. (a) Comparison between t^* values and the average of t^* of each cell. (b) The standard deviation is 0.005 s, which we assume to be the average uncertainty of the t^* value.

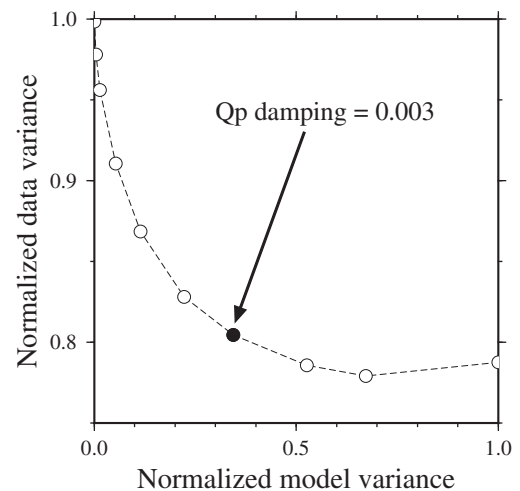


Fig. 6. Trade-off curve between data misfit and model variance for attenuation inversion. The optimal damping parameter for Q_p inversion is 0.003.

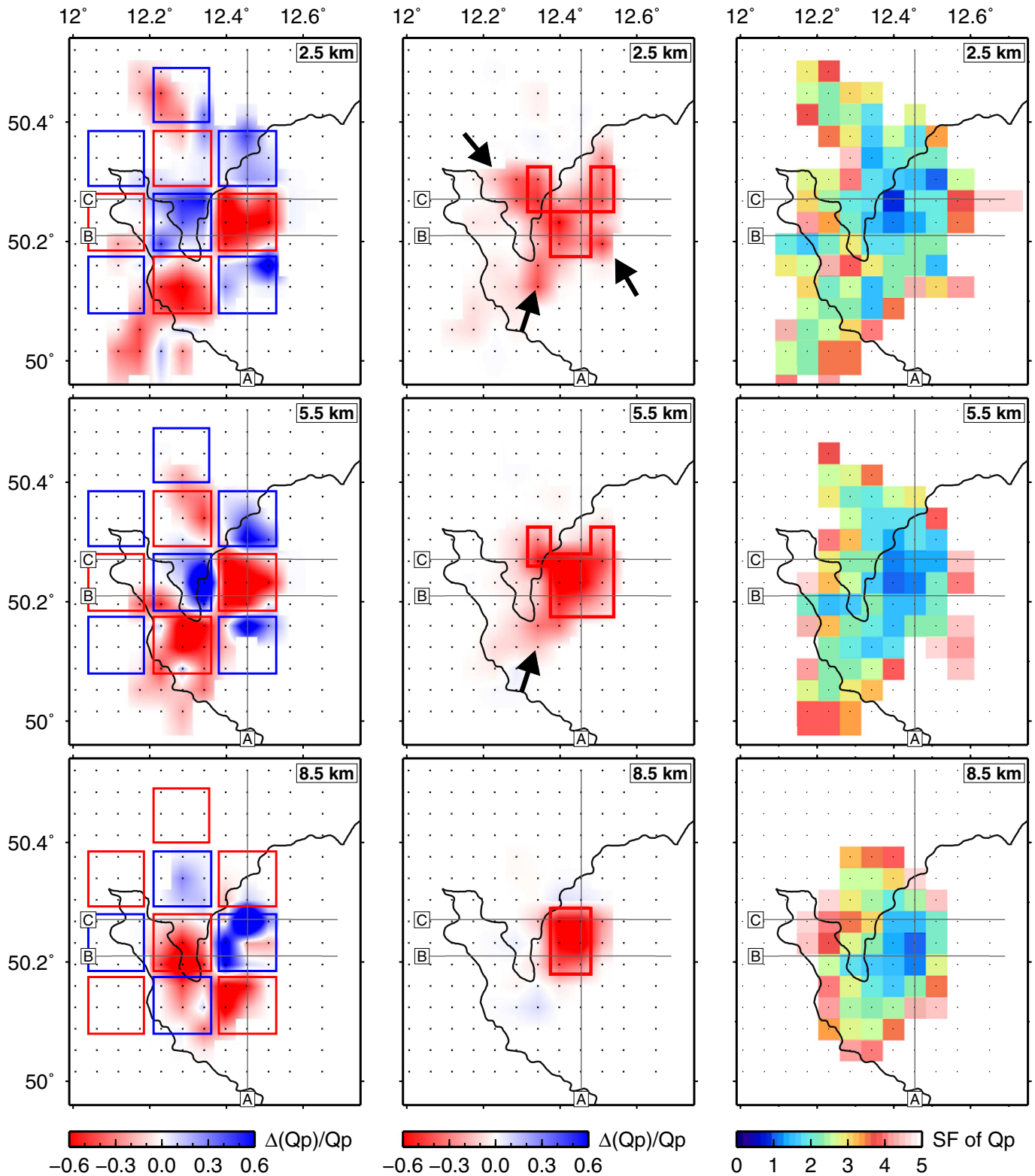


Fig. 7. The checker-board models, characteristic model and spread function (SF) at depth of 2.5, 5.5 and 8.5 km used to analyse the quality of the results. The blue and red boxes in the left indicate the high and low Qp values in checker board tests, similarly red boxes in middle panel are indication of low Qp in characteristic tests. The right panel shows the SF values. Arrows show the major smearing outside of anomaly outlines. Red and blue boxes represent the initial velocity model and the colours show their recovery.

are presented in Figs. 9. and 10 respectively. The location of these profiles can be found within the horizontal slices. To facilitate the evaluation and comparison of the results, the Vp and Vp/Vs from Mousavi et al. (2015) were added to Figs. 9 and 10. To present seismicity we

included the surface projection of the seismic events used in each tomography study onto the map views (Fig. 9). For vertical sections, seismic events within 1 km distance around the section are projected onto each profile (Fig. 10).

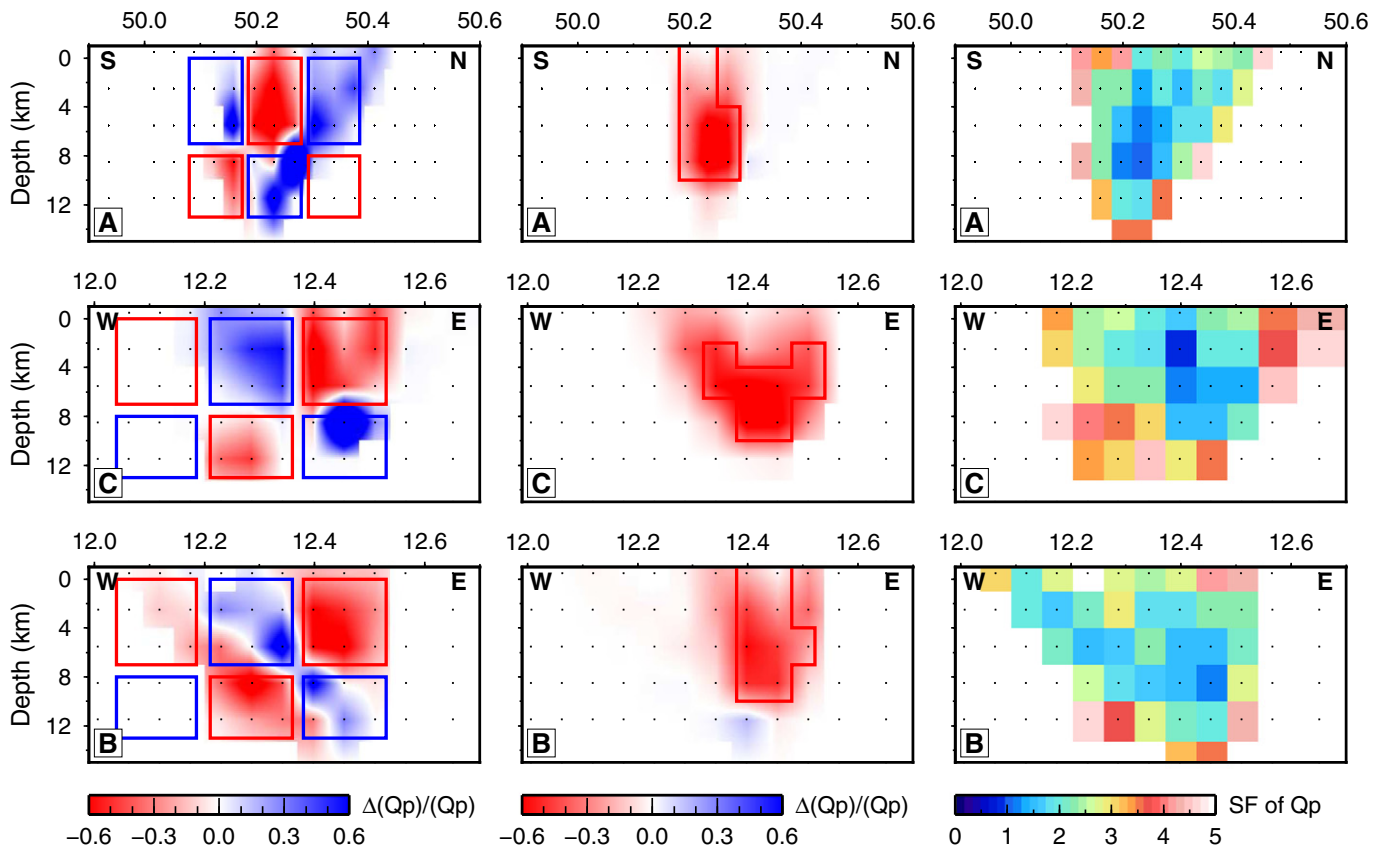


Fig. 8. Same as Fig. 7 checker-board test, characteristic test and spread function (SF) values for cross sections A, B and C. Profiles location could be found in Fig. 7. Red and blue boxes show the outline of input velocity model.

Horizontal slices reveal highly attenuating ($Q_p < 150$) features (Figs. 9 and 10) at shallow depths, ~2.5 km, located within and north of Nový Kostel area which then merge into one body at greater depth, surrounding the swarm hypocenters. From the characteristic tests we know that shallow depths are affected by smearing. Hence, some smearing can be expected for the shallow low Q_p bodies in the Nový Kostel area (see the black arrows in Fig. 7).

Another pronounced, attenuated upper crustal anomaly which also shows high V_p/V_s is imaged at German-Czech border close to the Quaternary volcano KH. The tomographic images appear blurry around the volcano due to insufficient crossing rays, indicating that the location at depth is most likely due to smearing rather than a real deep-stretched feature. Despite the lack of resolution at depth, the relatively higher attenuation beneath this area has some degrees of certainty.

Since rays reach our stations at near vertical angle, the existence of very shallow anomalies might affect the final outcome. By allowing some degree of freedom to be absorbed by a 'delay' at each station, we can estimate such scenario. All stations showed large station delays (t_i^*) except for those located at western margin. Large station delays in Nový Kostel and its vicinity suggest the continuation of the highly attenuating anomalies up to the surface.

7. Discussion

Our attenuation study revealed a number of prominent anomalies of low Q_p (< 150), which can be related to well-known features of West Bohemia. These anomalies are located at (1) Nový Kostel, covering the earthquake swarms hypocenters, (2) the northern vicinity of Nový Kostel, and (3) the Czech-German border close to the KH volcano.

7.1. Swarm area

The earthquake swarm region is probably the most extensively studied area in West Bohemia. Earthquake swarms often suggest fluid movement in critically loaded fault zones (Hainzl and Fischer, 2002). While most fluids at shallow depth in West Bohemia may be of meteoric origin, the high ratio of He and C isotopes connected with swarms suggests that some fluid originate from the mantle (Bräuer et al., 2005). The increasing V_p/V_s values observed within the Nový Kostel seismicity cluster at profile A (Fig. 10) were interpreted as an indication for a fluid passage toward the surface (Mousavi et al., 2015). The low Q_p values in the location of Nový Kostel on profiles A and B, extends deep down to ~11 km (Fig. 10, left panels) and encompasses all hypocenters. Although the resolution of the Q_p model is not comparable to our velocity tomography (Mousavi et al., 2015), the low Q_p anomaly within Nový Kostel is resolved reasonably well and suggests the presence of high fluid pore pressures in the weak and damaged zone of earthquakes.

The processes of swarm earthquakes in Nový Kostel seems to be similar to earthquake swarms in the active continental rift in New Zealand (Reyners et al., 2007) where fluids play an important role in triggering lower-crustal earthquake swarms. Here we see this in the upper-crust as well. Another similar condition is observed in KTB by Müller and Shapiro (2001) that a fluid filled crack system acts as a possible source of strong scattering attenuation. Same study observed a significant intrinsic attenuation for the frequencies < 30 Hz in this region. According to Mullick et al. (2015) fluids from a semi permeable zone located below earthquake focus intrude and produce the earthquake swarms. We speculate that the low Q_p bodies could be an indication of uprising fluids in permeable channels. Therefore, fluids pressure seems like a plausible explanation of the seismicity distribution in Nový Kostel.

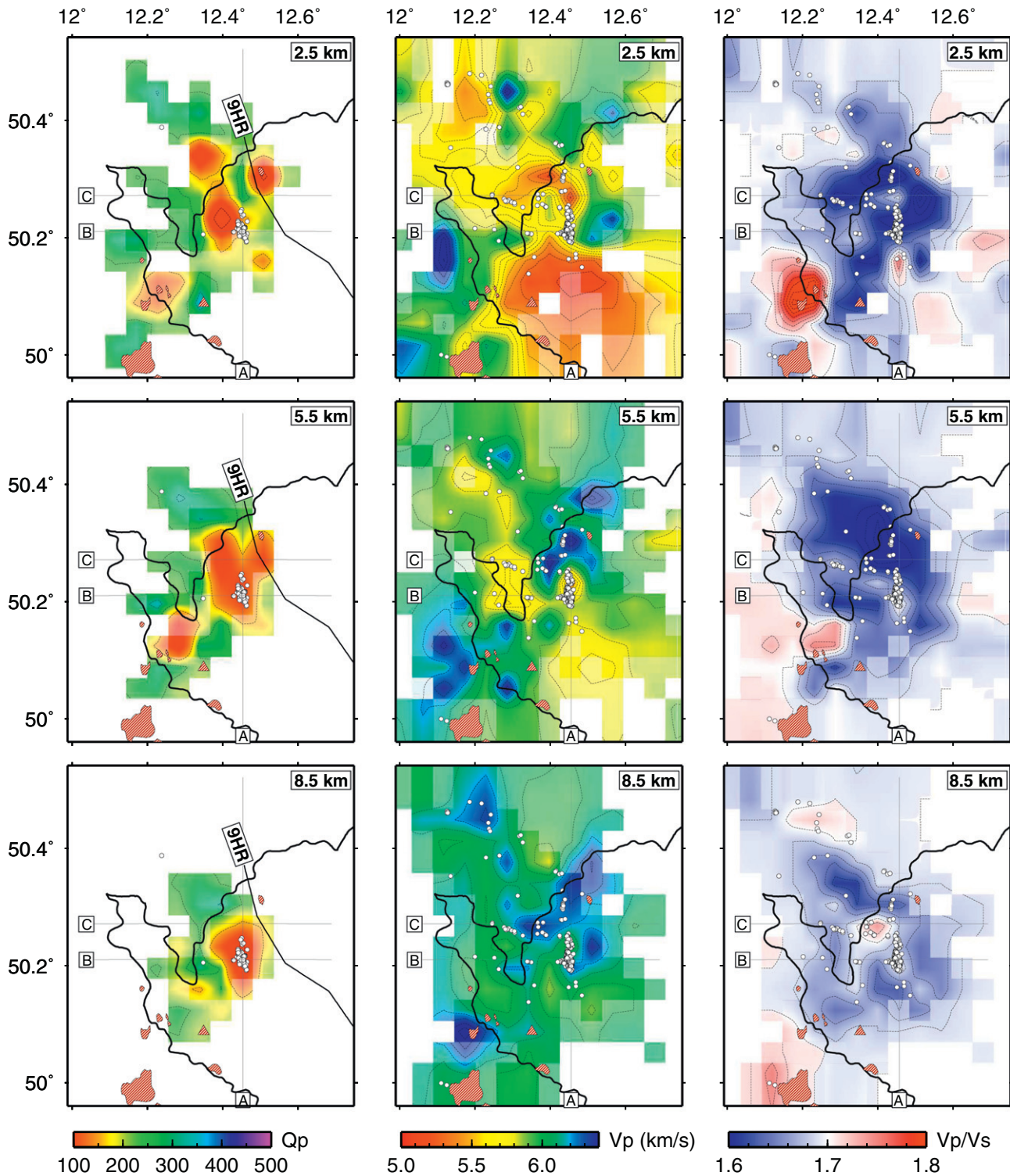


Fig. 9. The map view of the attenuation structure in this study (left panel), the Vp structure (middle panel) and the Vp/Vs structure from Mousavi et al. (2015) (right panel) at depths of 2.5, 5.5 and 8.5 km. Epicenter location of earthquakes used in each study are plotted as grey circles. Cenozoic volcanic rocks and KH volcano are shown as hatched area and triangle respectively. Location of 9HR profile is shown on the top of Qp tomograms. The Vp and Vp/Vs structure are obtained from Mousavi et al. (2015).

7.2. Low Q anomalies north of Nový Kostel

We also observed two highly attenuating anomalies located at north-east and north-west of Nový Kostel with Qp lower than 150 at 2.5 km depth (Fig. 9, left panels). These features correlate well with

the estimated low values of Poisson's ratio in the north of Nový Kostel by Růžek and Horálek (2013). The low-Qp anomalies in Nový Kostel and its northern vicinity are imaged as three patches in shallow depth (~2 km) but at depths >5.5 km, they merge into a single anomaly. It seems that they are constructing different branches of highly

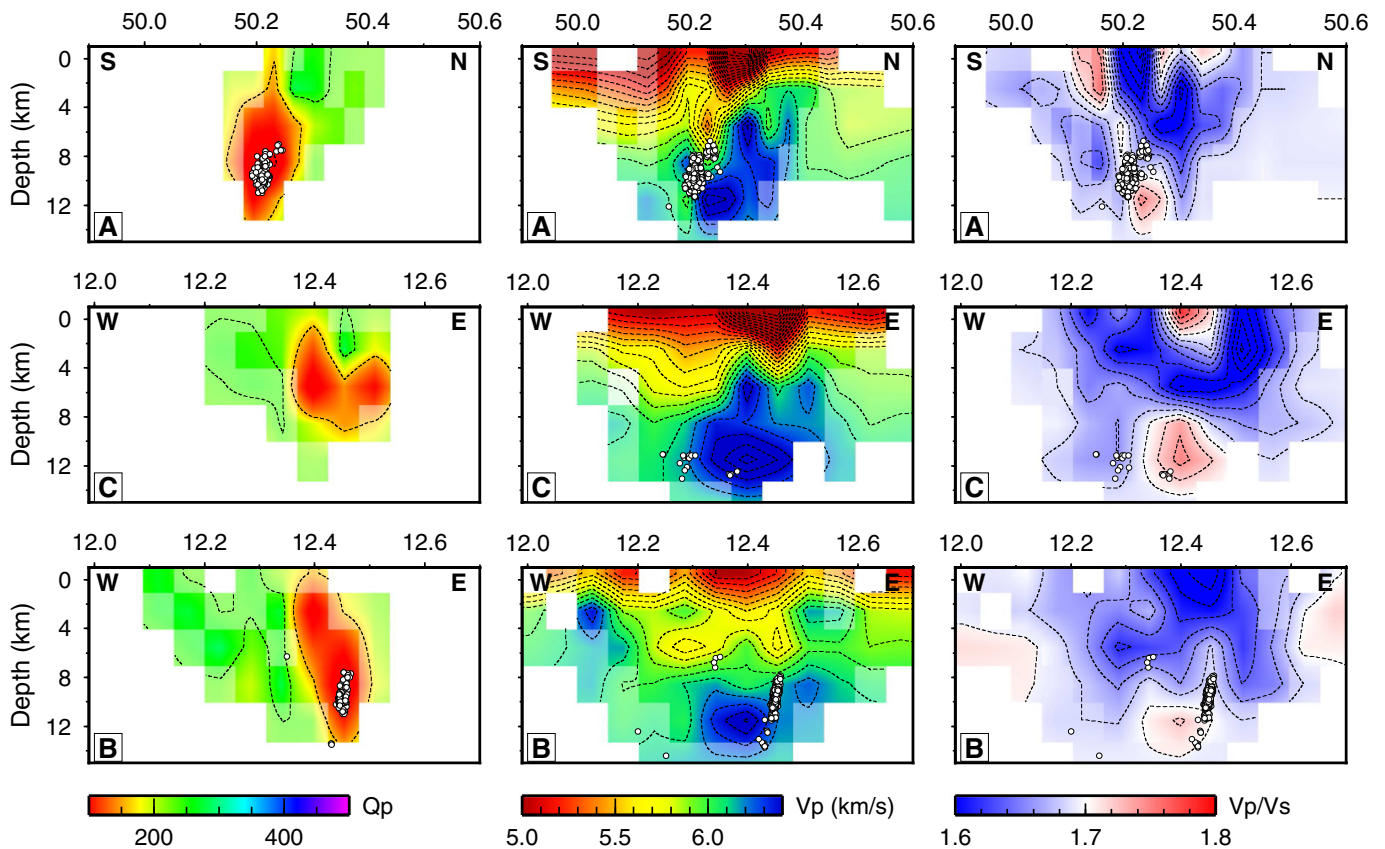


Fig. 10. The vertical cross-section along the three profiles (A–C indicated in Fig. 9) of: the Qp structure (left panels), the Vp structure (middle panels) and the Vp/Vs structure (right panels). The Vp and Vp/Vs structures are obtained from Mousavi et al. (2015).

attenuating channels which originated from the swarm focal zone (Figs. 9 and 10, left panel). We interpret these highly attenuating tunnel-like anomalies as the ascent of fluids from a common root via number of paths to the surface. The magmatic intrusive body which is imaged in the velocity tomography (Fig. 10, middle and right panel) might be the major degassing field of fluids which forms over-pressured fluid zones. This idea can be confirmed by the Vp/Vs tomography (profile A of Fig. 10, right panels) where the fluids originating from the assumed intrusive body are moving along a passage toward the surface. Helium and Carbon isotopic investigation in the north of Nový Kostel by Bräuer et al. (2003) confirmed that fluids rise up in different channels toward the surface. However, in their investigation they found no trace of CO₂-exhalation in the northern vicinity of Nový Kostel and northward along the Mariánské Lázně fault. They proposed the existence of permeability barrier traps at top layers (near the surface) which keep mantle derived fluids from ascending vertically and consequently cause a lateral distribution. We speculate that the over-pressured fluids rise along major fluid passages (our low Qp vertical anomalies) until they reach fluid caps at the shallow layers, then they find their way in minor channels to the surface at the observed fluid emission locations. Moreover, the large values of station delays might be due to the existence of fluids in shallower parts of the crust.

An interesting observation in this study is the low Qp anomaly at the north-east of Nový Kostel where the seismic reflection profile '9HR' (Tomek et al., 1997) passes through it. The location of this anomaly coincides well with a sub-horizontal reflector at 10 km distance from the north-western end of 9HR profile at depths of between 5 and 8 km (see reflector A in Mullick et al., 2015). This reflector was interpreted as a highly fragmented crust where fluids from lower crust can rise up through it and accumulate there. A similar discontinuity was observed by Hrubcová et al. (2016) who detected an interface at 4–6 km depth

was detected from the S-wave to P-wave conversions of teleseismic waves at the same location.

The low Vp and high Vp/Vs of this low Qp anomaly can be caused by an increase in fractures in saturated rocks (Sanders et al., 1995). This led us to speculate that fluid and gas accumulated adjacent to the swarm area which indicates fluid flux below but not directly above it. Unfortunately, due to the lack of crossing rays, the resolution in the peripheral edges of the model is limited. Hence, we are only given some hint of these fluid transport features.

Altogether, we think that the low Qp bodies in Nový Kostel and its vicinity as indicated by fluid studies (e.g., Weinlich et al., 1999) may represent parts of fluid channels which allow magmatic fluids originating in the upper mantle to rise through the crust and cause the observed fluid emanations and earthquake activity (e.g., Fischer et al., 2014).

7.3. Low-Q anomaly around the border of Bohemia/Bavaria

At the southwestern well-resolved part of the tomographic model, around the border region between Bohemia and Bavaria, a clear anomaly is observed particularly in Qp and Vp/Vs at the 2.5 km and 5.5 km depths slices (Fig. 9). According to the resolution tests this anomaly is located in the moderately resolved areas. Moreover, it appears roughly with the same shape at the location of the high Vp and high Vp/Vs body at the south west of the study area. A comparison with geological features mapped in the study area indicates that this anomaly coincides with the distribution of some Cenozoic volcanic rocks (compare Figs. 1 and 9). These young volcanic rocks were emplaced during multi-stage phases of volcanic activity accompanying the formation of the Eger rift (Ulrych et al., 2000; Babuška et al., 2010). In general, volcanic rocks are plausible candidates to explain low-Qp observations. The increased Vp/Vs ratios could be related with a predominantly basaltic composition of these rocks (Christensen, 1996). The anomaly is well-pronounced

and shows strong effects particularly in Vp/Vs compared to other features (e.g., 2.5 km depth slice in Fig. 9). This could be an indication for very young volcanism in comparison to other volcanic features along the Eger rift in the study region. Alternatively, the presence of fluids of a distinct chemical composition of the volcanic material could be responsible for the high Vp/Vs values of this anomaly.

8. Conclusions

In this paper, we obtained the first 3-D seismic *P*-wave attenuation tomography for West Bohemia/Vogtland. The attenuation tomography is based on path-average t^* values estimated for each earthquake-station pair. The derived Q_p values for an event at all recording stations showed that attenuation for rays travelling through the focal zone is much higher than those rays which go outside the focal zone. **The average Q_p value of ~230 in this region is much lower than in the rest of central Europe.** However, **path-average Q values to stations at larger distances (representative for Q values in the broader region) are larger than 500 which is in good agreement with previous studies**

The most important result of our tomographic study is the evidence of an anomalous volume with a very low Q_p value located at the Nový Kostel and its northern vicinity.

The tomography results reveal highly attenuating features beneath Nový Kostel and its northern neighbourhood which construct branches of anomalies from the swarm focal zone to the surface. We interpreted these anomalies as fragmented zones through which magmatic fluids can be transported through them from a common root below the swarm area, toward the surface. Such fluids may generally be an important driver of crustal seismicity in Nový Kostel. The other interesting result of this study is to demonstrate and confirm the presence of a reflector in the north-east of Nový Kostel as an accumulating zone and entrapment area of highly pressurized fluids. Furthermore, as indicated by large station delays, fluids could also exist in shallower parts of the crust.

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