

Moment Tensor Inversion Based on the Principal Component Analysis of Waveforms: Method and Application to Microearthquakes in West Bohemia, Czech Republic

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ABSTRACT

We develop and test a new hybrid approach of the amplitude and waveform moment tensor inversions, which utilizes the principal component analysis of seismograms. The proposed inversion is less sensitive to noise in data, being thus more accurate and more robust than the amplitude inversion. It also suppresses other unmodeled phenomena, like a directivity of the source, errors caused by local site effects at individual stations, and by time shifts in arrivals of observed and synthetic signals due to an inaccurate velocity model. This inversion is computationally less demanding than the full waveform inversion and thus applicable to large sets of earthquakes. The approach is numerically tested on synthetic data with various levels of noise. The applicability of the inversion is demonstrated on inverting more than 800 microearthquakes that occurred during the 2014 activity in West Bohemia, Czech Republic. The analysis revealed several distinct clusters of moment tensors. Focal mechanisms corresponding to moment tensors of three clusters are left-lateral strike slips associated with the most active fault in the focal zone. Another cluster is characterized by right-lateral strike slips associated with the fault conjugate to the main fault. Finally, we identified a cluster with pure reverse focal mechanisms that are anomalous and not expected to occur in the region. These mechanisms were not detected in previous seismic activity, and they have an unfavorable orientation with respect to regional tectonic stress. This might indicate a presence of local stress heterogeneities caused, for example, by an interaction of faults or fault segments in the focal zone.

INTRODUCTION

The seismic moment tensor (MT) describes equivalent body forces acting at a seismic point source. It comprises not only a double-couple (DC) component representing shear faulting

on a planar fault in isotropic media but also non-double-couple (non-DC) components produced, for example, by shear faulting on nonplanar faults, by cavity collapses in mines, by tensile faulting induced by fluid injection in geothermal or volcanic areas, or by seismic anisotropy in the focal zone (Julian *et al.*, 1998; Miller *et al.*, 1998; Nakamichi *et al.*, 2003; Templeton and Dreger, 2006; Ford *et al.*, 2008; Minson and Dreger, 2008; Šílený and Milev, 2008; Vavryčuk *et al.*, 2008; Guilhem *et al.*, 2014). Hence, MTs carry information about the orientation of activated fractures, fracturing mode, and elastic properties of the material in the focal zone (Vavryčuk, 2006; Vavryčuk and Hrubcová, 2017).

The MT inversion is a delicate procedure that requires an accurate location of the source, a detailed velocity model, good azimuthal coverage of stations on the focal sphere equipped with sensors of rather broad and accurately defined frequency response, and low noise in data (Šílený, 2009; Ford *et al.*, 2010; Stierle, Bohnhoff, *et al.*, 2014; Stierle, Vavryčuk, *et al.*, 2014). We can invert amplitudes of individual seismic phases (Vavryčuk *et al.*, 2008; Fojtíková *et al.*, 2010; Kwiatek *et al.*, 2016), amplitude ratios (Miller *et al.*, 1998; Hardebeck and Shearer, 2003; Jechumtálová and Šílený, 2005), or full waveforms (Dziewonski *et al.*, 1981; Kikuchi and Kanamori, 1991; Šílený *et al.*, 1992). The full waveforms are often inverted, assuming a point-source approximation and a time-independent focal mechanism. The inversion is performed in the time domain (Dreger and Woods, 2002; Sokos and Zahradník, 2008; Zahradník *et al.*, 2008; Adamová *et al.*, 2009), in the frequency domain using amplitude spectra (Cesca *et al.*, 2006) or complex spectra (Cesca and Dahm, 2008), or using a combined time-frequency approach (Vavryčuk and Kühn, 2012). Each inversion method is suitable for a different range of epicentral distances and frequencies of analyzed waves. The waveform inversion is preferable when applied to broadband data of moderate or large earthquakes recorded at regional or global seismic networks. By contrast, the amplitude inversion is advantageous when applied to short-period data of small earthquakes or microearthquakes recorded at local networks. To extend the applicability of the MT inversion or to get more robust results, some hybrid approaches propose various combinations of joint

inversion of polarities, amplitudes, waveforms, or amplitude ratios (Li *et al.*, 2011a,b; Fojtíková and Zahradník, 2014; Vavryčuk and Kim, 2014).

The analysis of MTs of microearthquakes is challenging for several reasons. First, the waveforms are complex and of high frequency, reflecting small-scale inhomogeneities in the Earth's crust. Second, the waveforms are often noisy because of low magnitude of events. Finally, the microseismic data comprise usually a large number of earthquakes that cannot be processed manually and need an automatic or semiautomatic processing. All these facts restrict the applicability of the waveform inversion and favor a much simpler amplitude inversion. The waveform inversion is rather laborious and time-consuming, and needs a detailed velocity model for computing accurate Green's functions. In comparison, the amplitude inversion is simpler and less computationally demanding than the waveform inversion. However, it is usually less accurate; it needs observations at more stations, and it is more sensitive to noise in data than the waveform inversion.

In this article, we describe a new inversion technique that improves the performance of the amplitude MT inversion by incorporating some aspects of the waveform inversion. The idea is based on computing the source time function and effective amplitudes of direct waves needed in the inversion by applying the principal component analysis (PCA) to seismograms. The proposed hybrid approach of the amplitude and waveform inversions is less sensitive to noise, being thus more accurate and more robust than the amplitude inversion. Still, the inversion is simple and computationally less demanding than the waveform inversion and thus applicable to large sets of earthquakes. The approach is numerically tested on synthetic data with various levels of noise. The applicability of the inversion is demonstrated by inverting more than 800 microearthquakes that occurred during the 2014 activity in West Bohemia, Czech Republic.

METHOD

Principal Component Analysis

The PCA is a statistical method that uses an orthogonal decomposition to transform possibly correlated observations into a set of linearly uncorrelated variables called the principal components. The principal components are ordered in such a way that the first component has the highest possible variance retaining as much as possible of the variation present in the data. Each component has the highest variance under the constraint of its orthogonality to the preceding components. The resulting vectors form an uncorrelated orthogonal basis (Jackson, 1991; Jolliffe, 2002). The goal of the PCA is to condense the maximum amount of variance present in the data into the fewest number of principal components. The PCA is commonly used in natural as well as social sciences, typically, in problems related to analysis of large data sets and to reducing their multidimensionality. The PCA is widely used in geophysics and seismology (e.g., Hagen, 1982; Dong *et al.*, 2006; Ocana *et al.*, 2008; Kositsky and Avouac, 2010).

Earthquake-related applications of the PCA include the analysis of ionospheric and geomagnetic anomalies associated with earthquakes (Hattori *et al.*, 2006; Lin, 2012a,b), spatio-temporal analysis of crustal deformations produced by large earthquakes (Kositsky and Avouac, 2010; Gualandi *et al.*, 2014); and signal detection, filtering, and polarization analysis of seismic data (Bataille and Chiu, 1991; Wagner and Owens, 1996; Muti and Bourennane, 2007). In MT inversions, the PCA was first applied by Vasco (1989, 1990) for finding the common source time function from six MT rate functions. This approach appeared useful and effective, being widely applied by other researches (Cesca and Dahm, 2008; Davi *et al.*, 2010; Eyre *et al.*, 2013; Zecevic *et al.*, 2013).

PCA Moment Tensor Inversion

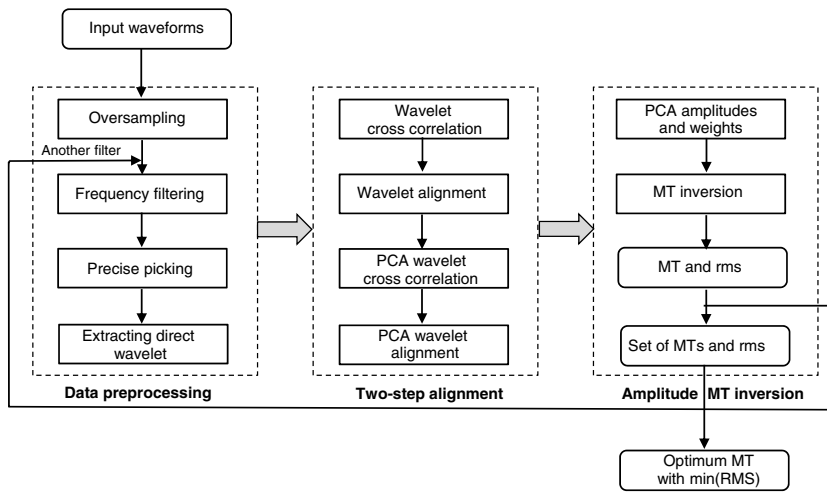
In the approach of Vasco (1989, 1990), the waveforms are first inverted using the generalized linear inversion for six MT rate functions. If the point-source approximation with a time-independent focal mechanism is assumed, the MT rate functions should be identical functions of time, except for scaling. However, the MT rate functions are frequently different because of noise in data, inaccurate Green's functions, and/or numerical errors. For this reason, the PCA is applied to extract the common source time function from the six MT rate functions. Alternatively, the source time function can be approximated by averaging the MT rate functions (Ruff and Tichelaar, 1990) or by minimization techniques (Šílený, 1998; Wéber, 2009).

The presented approach applies the PCA to inverting MTs in a different way. If we assume a point-source approximation and invert for MTs using direct waves, the waveforms recorded at stations should be identical functions of time. The exceptions are waves distorted by overcritical reflections that might introduce phase shifts. If we exclude stations with such anomalous records, we can apply the PCA to waveforms recorded at stations for finding the common wavelet radiated by the source. Once we find the source time function, we can invert amplitudes represented by the calculated PCA coefficients at each station for a time-independent MT. Such an approach is advantageous for several reasons: (1) we eliminate or suppress all station-dependent effects in input data like locally generated noise, site effects, or source directivity; (2) the inversion is computationally more efficient because the waveform inversion is substituted by the amplitude inversion; and (3) the inversion is more stable and robust because the sensitivity of the inversion to noise and other unmodeled effects is minimized.

Mathematical Description

Assuming $\mathbf{X}$ to be the data matrix composed of seismic waves recorded at the network of stations (i.e., each column of matrix $\mathbf{X}$ contains a single station component) and preprocessed to have a zero mean, we compute the eigenvalues and eigenvectors of the covariance matrix $\mathbf{XX}^T$. Because the covariance matrix is symmetric, it can be diagonalized in the following way (Jolliffe, 2002):

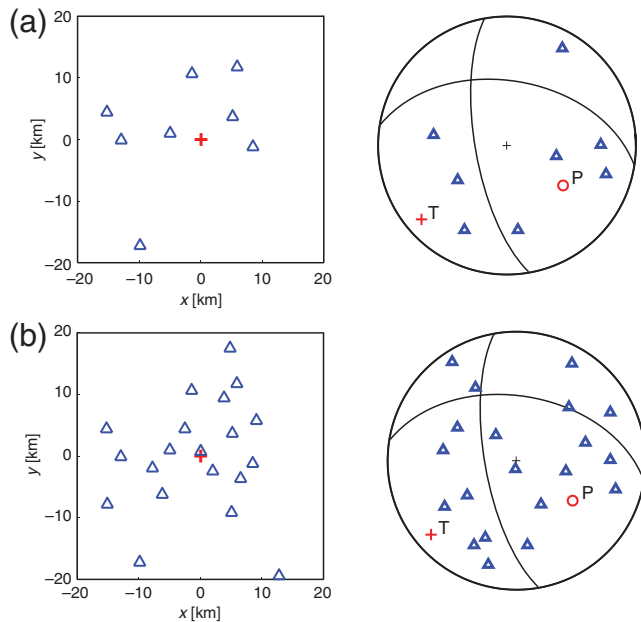
$$\mathbf{XX}^T = \mathbf{WDW}^T \quad (1)$$



▲ **Figure 1.** Flowchart of the principal component analysis (PCA) moment tensor (MT) inversion.

in which $\mathbf{W}$ is the matrix of orthonormal eigenvectors corresponding to eigenvalues ranked in a descending order which form diagonal matrix $\mathbf{D}$. The eigenvectors are the sought-after principal components. The PCA is closely related to the singular value decomposition, because the square roots of the eigenvalues of $\mathbf{X}\mathbf{X}^T$ are the singular values of $\mathbf{X}$. Subsequently, the PCA decomposition of $\mathbf{X}$ reads

$$\mathbf{X}_{\text{PCA}} = \mathbf{X}\mathbf{W} \quad (2)$$



▲ **Figure 2.** Station configurations used in the numerical tests. (a) 8 stations and (b) 20 stations. (Left) Map view and (right) focal spheres. Blue triangles mark positions of the stations. The red cross in the left plots shows the epicenter. The red cross and open circle in the right plots mark the T and P axes of the shear focal mechanism used in the tests. The strike, dip, and rake angles of the focal mechanism are 170° , 70° , and -45° , respectively.

and

$$\mathbf{X} = \mathbf{X}_{\text{PCA}}\mathbf{W}^T. \quad (3)$$

Hence, data matrix $\mathbf{X}$ can be reconstructed by multiplying the matrix $\mathbf{W}^T$ of the principal components by matrix $\mathbf{X}_{\text{PCA}}$ of the corresponding PCA coefficients.

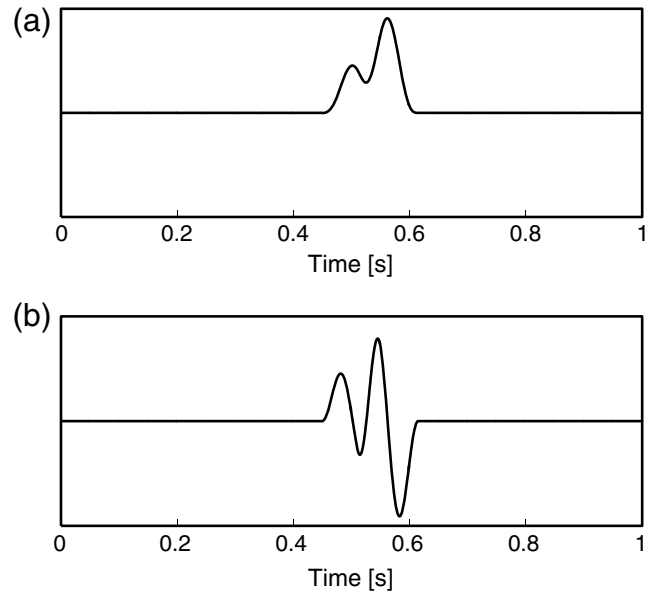
If matrix $\mathbf{X}$ is formed by a set of similar traces, the first principal component is dominant and corresponds to the common wavelet. Therefore, the coefficients of the PCA decomposition of the individual seismic traces, which are multiplied by the first principal component in equation (3), are the sought-after effective wave amplitudes (the PCA amplitudes). In other words, the PCA amplitudes are the multiplication factors of the common wavelet that reproduce most accurately the individual traces.

Having computed the PCA amplitudes, we proceed similarly as in the standard MT inversion of amplitudes directly measured from waveforms. The inversion solves a set of linear equations expressed in the matrix form as follows:

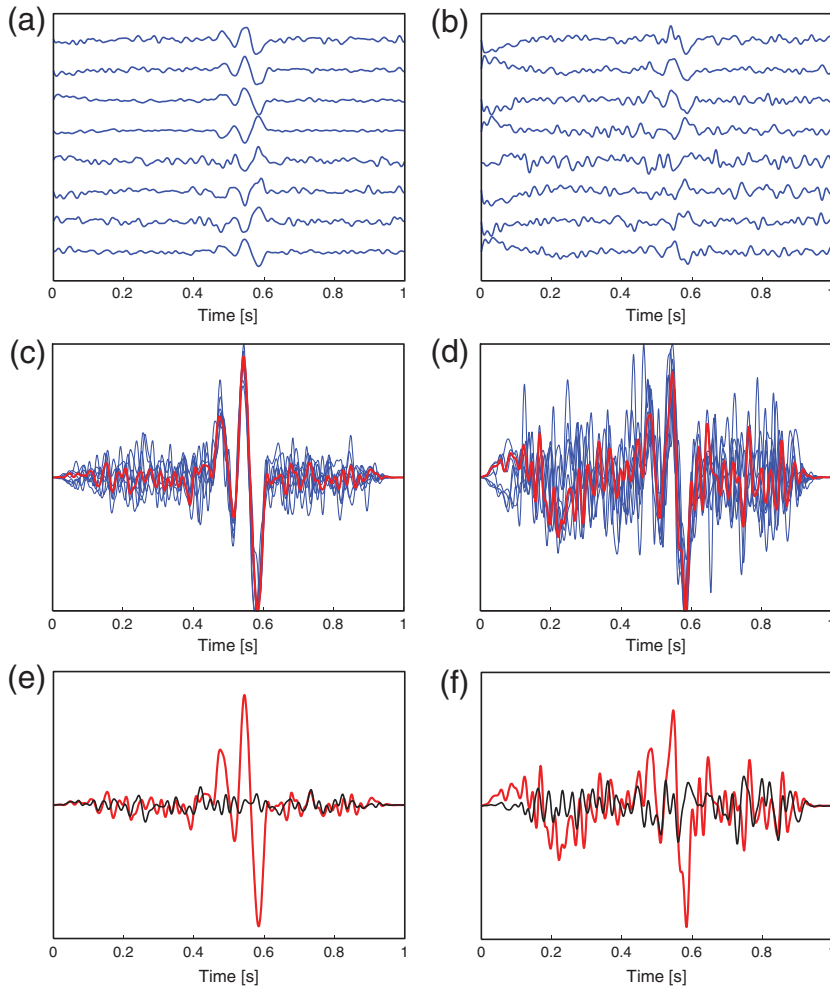
$$\mathbf{G}\mathbf{m} = \mathbf{u}, \quad (4)$$

in which $\mathbf{G}$ is the $K \times 6$ matrix of the spatial derivatives of the Green's function

$$\mathbf{G} = \begin{bmatrix} G_{11} & G_{12} & G_{13} & G_{14} & G_{15} & G_{16} \\ G_{21} & G_{22} & G_{23} & G_{24} & G_{25} & G_{26} \\ \vdots & \vdots & \vdots & \vdots & \vdots & \vdots \\ G_{K1} & G_{K2} & G_{K3} & G_{K4} & G_{K5} & G_{K6} \end{bmatrix}, \quad (5)$$



▲ **Figure 3.** The (a) displacement and (b) velocity records of the P wavelet used in the numerical tests.



▲ **Figure 4.** (a, b) Noisy synthetic records of the *P* waves for the station configuration with eight stations, (c, d) the *P*-wave records (blue) together with the first principal component (red), and (e, f) the first (red) and second (black) principal components. Maximum noise level is (left) 50% and (right) 150%. Some of the traces in (c) and (d) are displayed with a flipped polarity to be consistent with the polarity of the common wavelet.

$\mathbf{m}$ is the vector of six components of relative MT $\mathbf{M}$

$$\mathbf{m} = [M_{11} \ M_{22} \ M_{33} \ M_{23} \ M_{13} \ M_{12}]^T, \quad (6)$$

and $\mathbf{u}$ is the vector of wave amplitudes at K one-component sensors. Quantities G_{ki} are the components of the Green's function matrix for the k th sensor:

$$\begin{aligned} G_{k1} &= G_{k1,1}, G_{k2} = G_{k2,2}, G_{k3} = G_{k3,3}, G_{k4} \\ &= G_{k2,3} + G_{k3,2}, G_{k5} = G_{k1,3} + G_{k3,1}, G_{k6} \\ &= G_{k1,2} + G_{k2,1}, \end{aligned} \quad (7)$$

in which $G_{kl,m} = \partial G_{kl} / \partial \xi_m$ is the spatial derivative of the Green's function defined as the amplitude produced by the force couple at the source acting along the l axis with its arm in the m axis and recorded at the k th sensor along the sensor's direction.

Because the MT inversion is linear, it is fast and computationally undemanding. In addition, the MT inversion produces the source time function, which is a time integral of the common wavelet (i.e., the first principal component function). The source time function can further be used for calculating the scalar moment and subsequently for proper scaling of the relative MT in equation (6).

Inversion Steps and Code Structure

The MT inversion of a single earthquake consists of data preprocessing and picking, alignment of traces, calculation of the PCA amplitudes, and the MT inversion using the PCA amplitudes (see Fig. 1). Individual steps of the inversion can be summarized as follows:

1. Data preprocessing

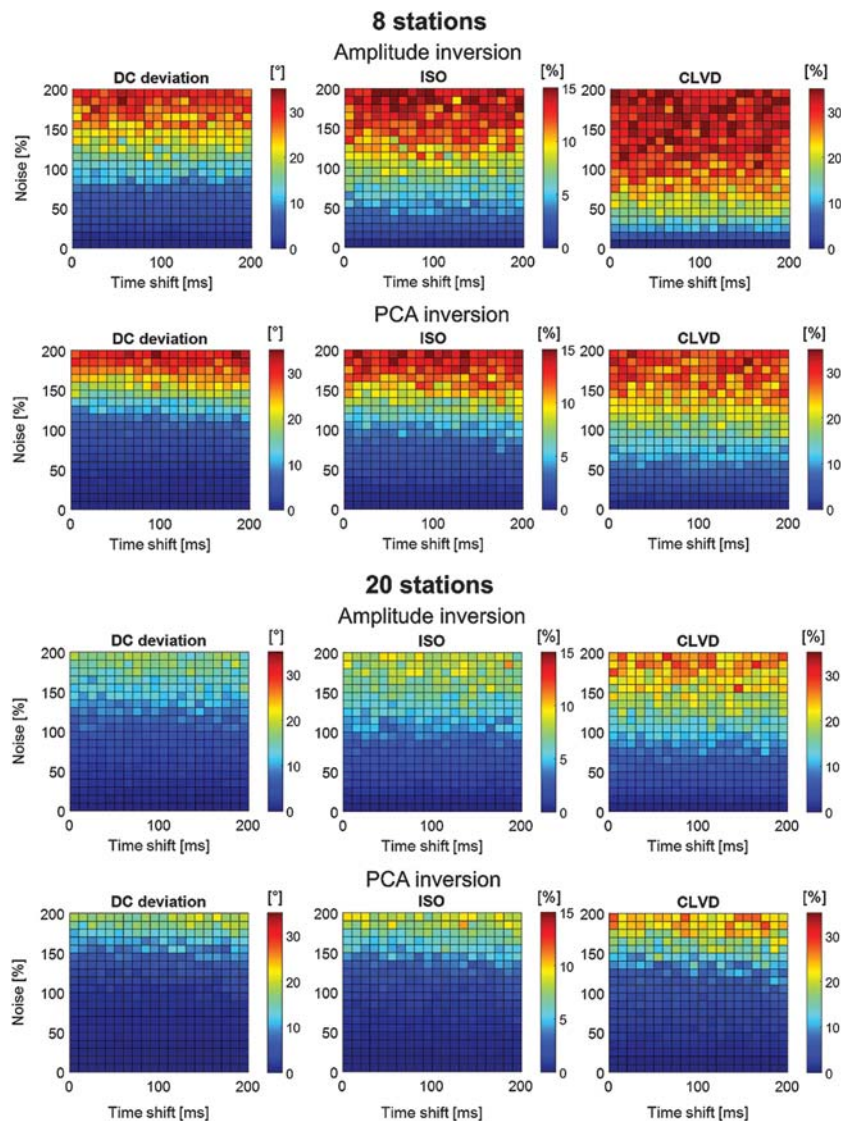
- First, we oversample data to be able to perform an accurate alignment of waveforms by tiny shifts in the next processing of waveforms.
- The oversampled data are band-pass filtered to enhance the signal-to-noise ratio (SNR).
- A rough alignment of waveforms is done using manual picking, if available, or using an automatic picking algorithm. In the automatic processing, we use the Suspension Bridge Picking (SBPx) algorithm of Meier (see [Data and Resources](#)), based on optimizing a ratio of integrated weighted amplitudes from before and after the candidate pick found by the standard short-term average/long-term average algorithm.

2. Two-step accurate alignment of waveforms

- The waveforms at all stations are shifted in time to maximize the cross correlation with the waveform of the highest SNR. Such aligned data are used for computing the principal components.
- The waveforms at all stations are aligned using the cross correlation with the computed first principal component.
- The aligned data are again decomposed using the PCA, and the principal components are refined.

3. Calculation of the PCA amplitudes and the MT inversion

- We compute the ratio between the maximum of the first and second refined principal components to check whether the method is applicable. Because the waveforms at various stations should be similar (except for amplitudes), the first principal component must be dominant. If not, the point-source approximation assumed in the inversion is not valid, and the inversion might fail.



▲ **Figure 5.** Tests of sensitivity of the amplitude and PCA inversions to errors in time arrivals and noise in data. The station configuration is with (upper six panels) 8 stations and (lower six panels) 20 stations. The color-coded double-couple (DC) deviation is calculated as the average of deviations of the true P and T axes from the retrieved P and T axes (see equation 8). The color-coded absolute values of isotropic (ISO) and compensated linear vector dipole (CLVD) components indicate the errors of the inversions, because the true ISO and CLVD are zero.

- The PCA coefficients of the first principal component serve as the effective amplitudes (including their polarity) used for the standard-amplitude MT inversion.
- The correlation coefficients between individual traces and the first principal component serve as the weights in the linear MT inversion scheme. If a wavelet at a station is significantly different from the common wavelet found by the PCA, its correlation coefficient is low, and its weight is suppressed in the inversion.
- The inversion produces a set of MTs because it is run repeatedly for several alternative band-pass filters. The

optimum MT is that which produces the minimum root mean square (rms) of differences between synthetic and observed amplitudes. The optimum MT is relative because it is calculated from amplitudes but not from the low-frequency asymptote of displacement records.

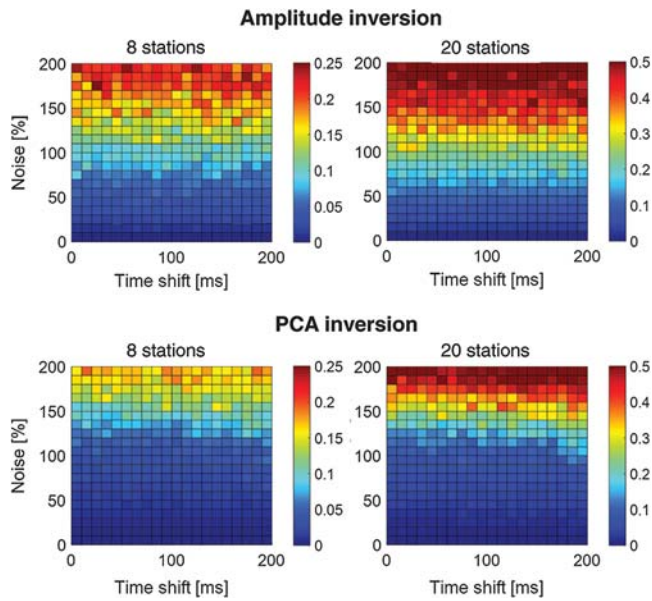
- The scalar moment of the optimum MT is calculated by integrating the common (displacement) wavelet.

A crucial part of the PCA MT inversion, which calculates the PCA wavelet and the PCA amplitudes from input traces, is coded in MATLAB and freely available on the web (code PCA-DECOMPOSITION, see [Data and Resources](#)).

NUMERICAL TESTS

In this section, we perform numerical tests illustrating the effectiveness and robustness of the proposed inversion scheme. The velocity model, station configurations, event location, and focal mechanism are used to mimic observations of microearthquakes in the West Bohemia region used as an example in the [Application to the 2014 Earthquake Sequence in West Bohemia](#) section. We use synthetic data of one event recorded by two station configurations consisted of 8 and 20 stations, respectively. The input data are contaminated by various levels of noise. The depth of the event is 9 km, and the epicentral distance of stations is less than 20 km. We assume a shear earthquake with the focal mechanism defined by strike, dip, and rake of 170°, 70°, and −45°, respectively (see Fig. 2). This focal mechanism is typical for the activity in West Bohemia and represents the principal focal mechanism in the region ([Vavříček, 2011](#)). The Green's function needed for calculating synthetic amplitudes, and subsequently for inverting for the MT, was calculated for a 1D gradient isotropic velocity model using the ray method ([Červený, 2001](#)). The arrival

times of the direct *P* waves are calculated by the kinematic two-point ray tracing. Geometrical spreading along the ray is obtained from the width of an elementary ray tube constructed from neighboring rays. The ray amplitudes include the conversion coefficients at the free surface. The source time rate function is formed by two one-sided pulses of different amplitudes (see Fig. 3). The total width of the pulse is about 0.18 s. The prevailing frequencies range from 10 to 16 Hz, corresponding to magnitudes of 2.5–1.5. The data are contaminated by random white noise with a uniform distribution. The noise level ranges from 0% to 200% of the maximum am-



▲ **Figure 6.** The root-mean-square (rms) residuals of the amplitude and PCA inversions as a function of errors in time arrivals and noise in data. (Upper row) The amplitude inversion and (lower row) the PCA inversion. The rms is calculated using equation (9).

plitude of the noise-free wavelet. The noisy waveforms are band-pass filtered in the 2–40 Hz frequency range (see Fig. 4). In addition, the waveforms are randomly shifted in time to mimic the errors in arrival times of the Green's functions, due to an inaccurate velocity model. The time shifts vary from 0 to 0.2 s. The sampling frequency is 250 Hz.

The vertical synthetic noisy P waveforms were generated 50 times for each combination of random noise in amplitudes and in time shifts and inverted repeatedly for the MT to obtain statistically significant results. We applied two MT inversions: (1) the standard amplitude inversion and (2) the PCA-based inversion, and compared their accuracy. The inversions differ in the way the P -wave amplitudes are measured at synthetic waveforms. In the amplitude inversion, the maximum amplitudes of the P waves are measured and inverted for the MT. In the PCA inversion, the effective amplitudes of the P waves are computed according to the [Inversion Steps and Code Structure](#) section and inverted for the MT.

The errors produced by the amplitude and PCA inversions are quantified by calculating the deviation of the DC component of the retrieved MTs from that of the synthetic MT. The DC deviation δ is computed as the average of deviations of the P/T axes of the retrieved focal mechanisms (defined by direction vectors $\mathbf{p}$ and $\mathbf{t}$) from the P/T axes of the true focal mechanism (defined by direction vectors $\mathbf{p}_{\text{true}}$ and $\mathbf{t}_{\text{true}}$):

$$\delta = \frac{1}{2} (\cos(|\mathbf{p} \cdot \mathbf{p}_{\text{true}}|) + \cos(|\mathbf{t} \cdot \mathbf{t}_{\text{true}}|)), \quad (8)$$

in which the dot means the scalar product. The deviation δ calculated for individual retrieved MTs is then averaged over

50 realizations of random noise. Another option for quantifying the DC deviation could be, for example, calculating the Kagan angle (Kagan, 1991); see also a discussion in Tape and Tape (2012).

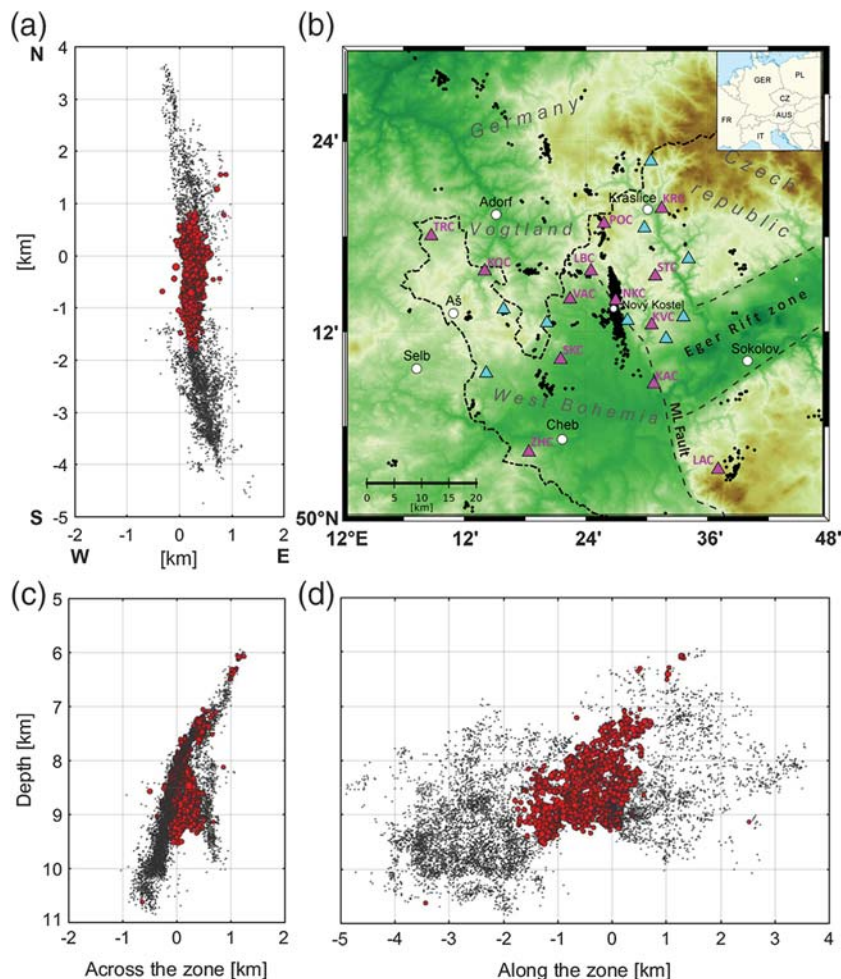
In addition, we quantify the inversion errors by calculating the non-DC components of the MTs, which are decomposed into the isotropic (ISO) and compensated linear vector dipole (CLVD) components. Advantages of this MT decomposition and properties of other possible MT decompositions are discussed in detail in Vavryčuk (2015). Because the true MT is pure DC, the true values of the ISO and CLVD components are zero. Consequently, the non-DC components of the retrieved MTs are spurious and directly measure the errors of the inversion. The percentages of the ISO and CLVD components were calculated using equations (6)–(10) of Vavryčuk (2015). The percentages are averaged over 50 realizations of random noise similarly as for the DC deviation.

Figure 5 indicates that both inversions are equally insensitive to errors in arrival times produced usually using a simplified velocity model in the Green's function calculations. The insensitivity to inaccurate arrival times of waves is a great advantage of the amplitude inversion compared to the inversion applied to waveforms with no time-shift corrections. The PCA inversion also works well because the P waveforms at all stations are aligned by the two-step cross-correlation procedure; see Figure 1 and the [Inversion Steps and Code Structure](#) section. However, as regard the sensitivity of the MT inversions to noise in amplitudes, the performance of both approaches is different. Figure 5 demonstrates a clear superiority of the PCA inversion over the amplitude inversion. The PCA inversion produces lower errors in the orientation of the DC component as well as lower non-DC components of the MT. This applies to both station configurations, with 8 stations (upper panels) and 20 stations (lower panels). Because the true MT is pure DC, the true values of the ISO and CLVD are zero (blue color in panels in the middle and right columns of Fig. 5). In addition, we observe that the spurious CLVD is about three times higher than the spurious ISO, indicating that the CLVD component is significantly more sensitive to numerical errors of the inversion than the ISO component.

A better performance of the PCA inversion compared to the amplitude inversion is also documented in Figure 6, showing the rms of the MT solutions defined as the normalized differences between the synthetic and observed amplitudes in the MT inversion:

$$\text{rms} = \frac{\sqrt{\sum (A_i^{\text{synth}} - A_i^{\text{obs}})^2}}{\sqrt{\sum (A_i^{\text{synth}})^2}}, \quad (9)$$

in which i defines the station component, and the summation is over all stations and components which recorded the analyzed event. Obviously, the inversion of noise-free data produces the correct MT and the zero rms. When inverting noisy data, the rms is nonzero and increases with increasing noise. This increase is, however, different being lower for the PCA



▲ **Figure 7.** Locations of microearthquakes in the West Bohemia region. The locations are shown in (a, b) the map view and in (c) vertical sections across the fault zone and (d) along the fault zone. The triangles in (b) show the positions of the West Bohemia Network (WEBNET) stations: magenta triangles, online stations; blue triangles, offline stations. The dots in (a, c, and d) mark the locations of microearthquakes that occurred in the period of 1995–2015 (in black) and in the 2014 earthquake sequence (in red). The hypoDD method was used to compute the earthquake locations.

inversion than for the amplitude inversion. This indicates that calculating the common wavelet by applying the PCA to waveforms before the inversion reduces noise in the input data efficiently.

APPLICATION TO THE 2014 EARTHQUAKE SEQUENCE IN WEST BOHEMIA

The West Bohemia Swarm Area

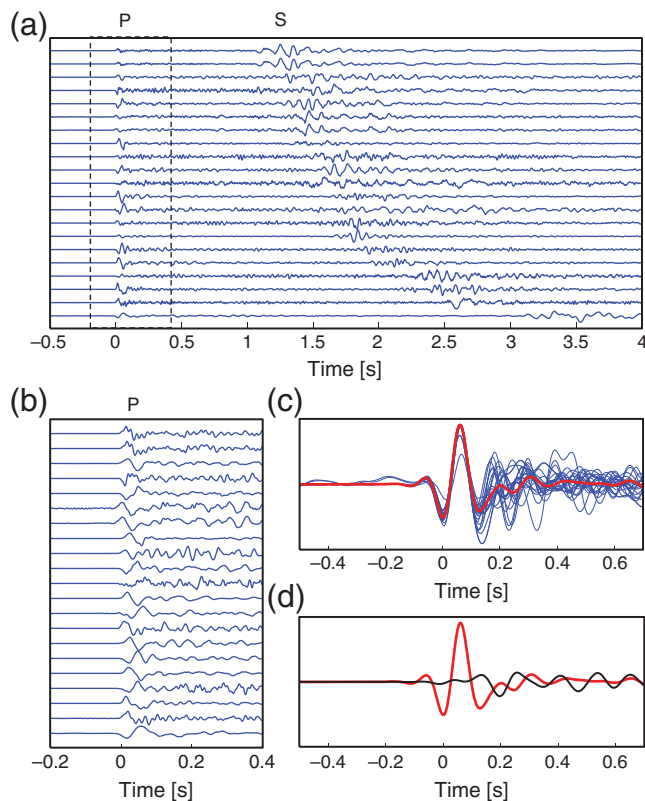
The West Bohemia region is situated in the western part of the Bohemian Massif at the contact of three Variscan structural units: (1) Saxothuringian, (2) Teplá-Barrandian, and (3) Moldanubian (Babuška *et al.*, 2007). The triple junction of the units is characterized by thinning of the crust with depth of Moho ranging from 28 to 32 km (Hrubcová *et al.*, 2013).

The tectonic structure of the area is characterized by two main fault systems: (a) the Mariánské Lázně northwest–southeast fault system and (b) the Ore-Mountain west-southwest–east-northeast fault system (Fig. 7b). However, the recently active faults in the area are the left-lateral strike-slip fault in the north–south direction with a strike of N169° E and the right-lateral strike-slip fault in the northwest–southeast direction with a strike of N304°E. The seismically active faults were identified at depth by foci clustering and by focal mechanisms (Vavryčuk *et al.*, 2013), but they have also some geological evidence on the surface (Bankwitz *et al.*, 2003). The maximum compressive stress determined from focal mechanisms has an azimuth of N146°E (Vavryčuk, 2011). The stress analysis indicates that the two active faults in the region with strikes of N169°E and of N304°E are optimally oriented for shearing with respect to the tectonic stress, thus forming a pair of conjugate principal faults in the area. The deviation of the principal faults from the σ_1 axis is about 32° and corresponds to fault friction of 0.5.

Data

The West Bohemia swarm area is characterized by persistent seismic activity with a periodic occurrence of earthquake swarms. Prominent earthquake swarms occurred in 1985/1986, 1994, 1997, 2000, 2008, 2011, and 2014 at the same epicentral area (Fig. 7), called the Nový Kostel seismic zone (Fischer *et al.*, 2014; Čermáková and Horálek, 2015; Hainzl *et al.*, 2016). The swarm duration was typically from two weeks to two months, and the activity was focused at depths of 6–11 km (Fig. 7c,d). The strongest instrumentally recorded earthquake was the M_L 4.6 earthquake on 21 December 1985.

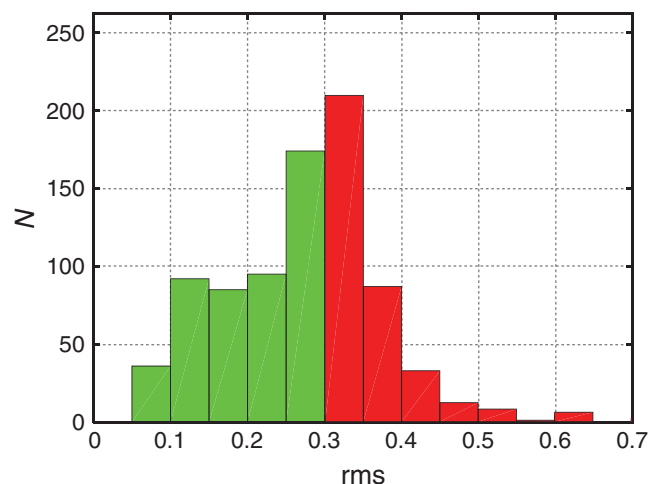
The most recent major seismic activity occurred in 2014 (Hainzl *et al.*, 2016). It covered a period of almost three months and consisted of three distinct phases, with the strongest events on 24 May (M_L 3.5), 31 May (M_L 4.2), and 3 August (M_L 3.6). The seismic activity was monitored by 22 short-period three-component seismic stations of the West Bohemia Network (WEBNET) consisting of 13 online and 9 offline stations. The stations have epicentral distances less than 25 km and cover the area uniformly with no major azimuthal gaps (Fig. 7b). Most of the stations are short period with a sampling frequency of 250 Hz and a flat frequency response between 1 and 60 Hz. In addition, the station with the nearest epicentral distance (NKC) was equipped by a broadband STS-2 seismometer. In total, about 5600 events with the local mag-



▲ **Figure 8.** Application of the PCA to records of the microearthquake of 24 May 2014 at 16:14:30 with magnitude of M_L 2.1. (a) The velocity records of the vertical component at the WEBNET stations aligned according to the P -wave arrivals. (b) A detailed time window marked by dashed lines in (a). (c) The aligned P waveforms at all stations (blue lines) together with the first principal component (red line). (d) A comparison of the first (red) and second (black) principal components. Polarities of some traces in (c) are flipped according to the common wavelet.

nitide M_L larger than -0.5 were detected during the 2014 activity.

We analyzed 833 microearthquakes with magnitude M_L larger than 0.5 recorded at least at 16 WEBNET stations. The data set covers the whole period of the activity, including the three strongest events. The P and S arrival times were picked manually for the online stations and automatically using the SBPx algorithm of Meier (2015) for the offline stations. The locations of the events were calculated using the NonLin-Loc code (Lomax *et al.*, 2000). The 1D-layered velocity model proposed by Málek *et al.* (2005) was used for locating the events. The Green's functions were computed by the ray method (Červený, 2001) using a 1D-gradient velocity model obtained by smoothing the layered model of Málek *et al.* (2005). The velocity records were oversampled from 250 Hz to 1 kHz and band-pass filtered (see Fig. 8). We applied the two-sided Butterworth filters with the low corner frequency of 1 Hz and with the high corner frequency alternatively of 6, 8, 10, and 12 Hz. The PCA analysis was run on data filtered in several frequency bands because analyzed earthquakes cover a

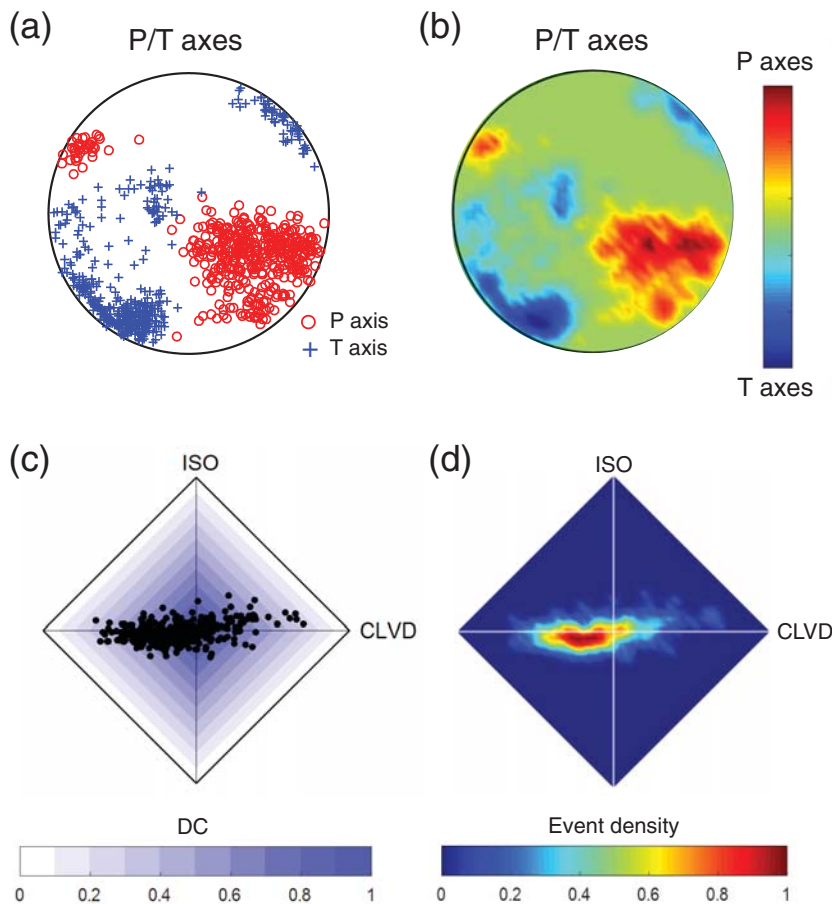


▲ **Figure 9.** Histogram of the rms residuals of the 833 MTs retrieved by the PCA inversion. Green/red indicates reliable/unreliable MTs defined by the rms lower/greater than 0.3. The rms is calculated using equation (9).

wide range of magnitudes from 0.5 up to 4.2, so the predominant frequency of the P wave can vary. Based on the P -wave arrival times, we cut the window containing the direct P wave (see Fig. 8b). Before calculating the effective P -wave amplitudes using the PCA, we applied the two-step alignment, as described in the [Inversion Steps and Code Structure](#) section and computed the common wavelet (see Fig. 8c,d). The computed PCA amplitudes corresponding to the individual frequency bands are used in the MT inversion. The optimum frequency band is that which produces an MT with the lowest rms.

Retrieved Moment Tensors

The retrieved MTs of the 833 analyzed earthquakes are differently accurate. This is indicated by a large scatter of the rms of the MT solutions (see Fig. 9). Therefore, we eliminated unreliable solutions and selected the 440 most accurate MTs by applying the two quality criteria: (1) the rms lower than 0.3 (see Fig. 9) and (2) the number of stations recording the earthquakes higher or equal to 18. As seen from Figure 10, the clusters of the P and T axes of the reliable solutions do not overlap and are well separated. However, both P and T clusters are rather large with a varying density (Fig. 10b) and can be divided into several smaller subclusters corresponding to several distinct types of MTs. For example, the T axis is nearly horizontal for a majority of earthquakes, but some earthquakes are anomalous, having the T axis nearly vertical (blue plus signs close to the center of the focal sphere in Figure 10a). By contrast, the non-DC components behave differently. They form just one cluster elongated along the CLVD axis (Fig. 10c,d). As observed in the synthetic tests, the higher variation of the CLVD can partly be caused by a higher sensitivity of the CLVD to numerical errors of the inversion. The CLVD and ISO components are negative for a majority of MTs.



▲ **Figure 10.** The DC and non-double-couple (non-DC) components of the 440 reliably determined MTs displayed in (a, b) the focal sphere and in (c, d) the CLVD–ISO diamond plot. (a) The P/T axes of individual microearthquakes. (b) The density plot of the P/T axes. (c) The non-DC components of individual microearthquakes. (d) The density plot of the non-DC components. The color scale of the density plots is normalized to its maximum value. For a detailed description of the CLVD–ISO diamond plot, see Vavryčuk (2015).

Classification of the Moment Tensors

The existence of several typical MTs (and their DC components) is confirmed by the cluster analysis (Hartigan, 1975). To identify typical MTs, we applied the k -means clustering (Jain, 2010). In this method, each cluster is defined by its centroid, and the positions of the centroids are found by minimizing the sum of distances of all classified objects. The only parameter controlling clustering is the number of clusters. However, the crucial role in clustering also plays the definition of distance measured between two MTs, $\mathbf{M}^{(1)}$ and $\mathbf{M}^{(2)}$. Here, we apply the definition proposed by Willemann (1993):

$$D = \cos^{-1} \left(\frac{1}{2} \frac{M_{ij}^{(1)} M_{ij}^{(2)}}{M^{(1)} M^{(2)}} \right), \quad (10)$$

in which $M^{(1)}$ and $M^{(2)}$ are the scalar moments defined by their Euclidean norm:

$$M^{(1)} = \sqrt{\frac{1}{2} M_{ij}^{(1)} M_{ij}^{(1)}} \\ \text{and } M^{(2)} = \sqrt{\frac{1}{2} M_{ij}^{(2)} M_{ij}^{(2)}} \quad (11)$$

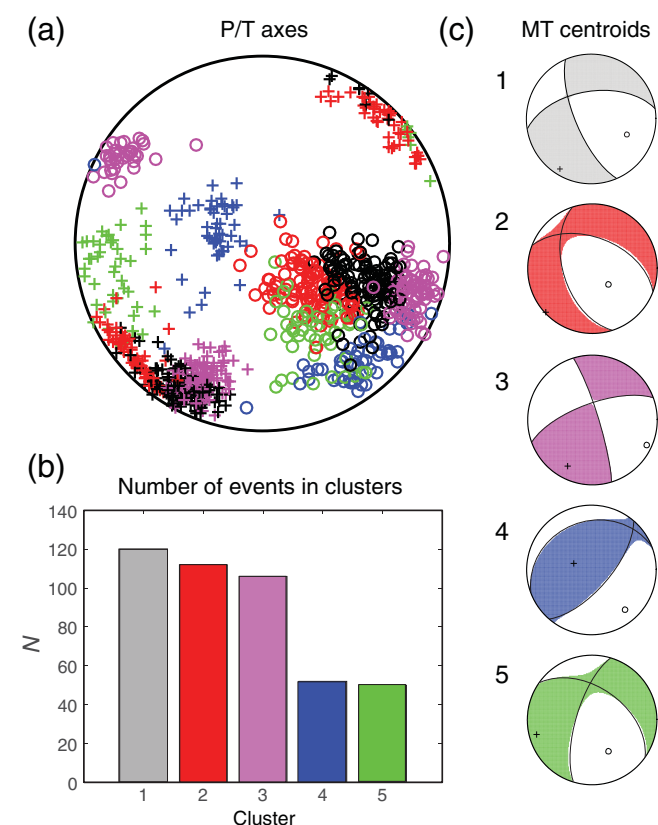
(Silver and Jordan, 1982). The cosine distance in equation (10) measures the angle between two MTs. This angle is, however, different from the frequently used Kagan angle (Kagan, 1991). We do not use the Kagan angle because it can be applied to DC MTs only, and its physical interpretation is controversial (Willemann, 1993). A distance measure of MTs similar to equation (10) is also applied by other authors (Tape and Tape, 2012; Cesca *et al.*, 2014).

Figure 11 and Tables 1 and 2 show the results of the classification of the MTs into five clusters. They present the number of earthquakes in individual clusters, the centroid moment tensors (CMTs) and their DC and non-DC components, and the mean rms of earthquakes in the clusters. The MTs are most frequently grouped in clusters 1 and 2, representing the left-lateral strike slips with a normal component, and in cluster 3 representing the left-lateral strike slips with a weak reverse component. The less frequent MTs are those in cluster 4, which are almost purely reverse, and those in cluster 5, which are the right-lateral strike slips with a normal component. The left-lateral strike slips (clusters 1–3) are associated with the most active fault in the focal zone having the strike between 155° and 170° . This fault was active during all prominent earthquake swarms in the last 40 years. The right-lateral strike slips (cluster 5) are associated with a less active fault

with strike of 300° – 310° . These mechanisms appeared, for example, in the 2008 swarm activity. Both faults are symmetrically oriented with respect to the maximum compression of the regional tectonic stress and are close to the principal faults in the region (Vavryčuk, 2011).

Interestingly, the almost pure reverse focal mechanism of cluster 4 is anomalous. It is not well oriented for shearing under the regional tectonic stress, and it has not been detected in the previous seismic activity in the region. Because the three strongest microearthquakes of the 2014 activity display this type of the focal mechanism, the 2014 activity seems to be exceptional and distinctly different from the previous seismic activities. This observation is also supported by anomalies in the magnitude–frequency relations found by Hainzl *et al.* (2016).

The CMTs differ not only in the orientation of the DC component but also in the percentage of their DC and non-DC components. Clusters 1 and 3 are formed by almost pure DC earthquakes (DC > 93%), with the mean rms of 0.22 and 0.27, respectively. By contrast, cluster 2 consists of earthquakes



▲ **Figure 11.** Classification of the 440 reliably determined MTs into five clusters. (a) The focal sphere with the P/T axes. (b) Histogram of the number of events in individual clusters. (c) Focal mechanism plots corresponding to the centroid moment tensors of individual clusters. The individual MT clusters are color coded. The black open circles and plus signs in (c) mark the P and T axes, respectively.

with the lowest DC percentage (DC $\sim 75\%$), with the mean rms of 0.13. No positive correlation between the percentage of the non-DC components and the rms indicates that the non-DC components are not errors produced by the inversion but that they are of physical origin. Except for cluster 3, which has negligible non-DC components, the ISO and CLVD are

negative. The negative values of the ISO and CLVD have also been observed in other earthquake swarms in the studied area being interpreted in terms of fault weakening due to erosion by fluids manifested by shear-compressive faulting (Vavryčuk and Hrubcová, 2017).

DISCUSSION AND CONCLUSIONS

The numerical tests demonstrated that the PCA-based MT inversion is an efficient and robust tool for determining MTs of microearthquakes. The method is less sensitive to seismic noise in data than the amplitude inversion. It also suppresses other unmodeled phenomena, such as a directivity of the source, errors caused by local site effects at individual stations, and time shifts in arrivals of observed and synthetic signals due to an inaccurate velocity model. The method is fast and less computationally demanding than the waveform inversion. It can work in automatic or semiautomatic regime, and hence it is suitable for processing large sets of earthquakes.

The analysis of the 2014 earthquake activity in West Bohemia using the PCA-based MT inversion revealed several distinct clusters of the MTs. The CMTs differ by the orientation of their DC component as well as by the amount of the non-DC components. Three CMTs are left-lateral strike slips associated with the most active fault in the focal zone. Interestingly, two of them are almost shear, with the DC higher than 94%, and the third type is slightly non-DC (DC = 75%), with a negative CLVD and ISO. We identified two additional clusters: (1) one cluster with right-lateral strike slips and (2) one cluster with reverse focal mechanisms. Both CMTs have DC about 80%, with a negative CLVD and ISO. The right-lateral strike slips are associated with another fault previously active in the focal zone. By contrast, the reverse focal mechanisms in the area are anomalous and unexpected. They have not been detected in the previous seismic activity in the region, and they also have an unfavorable orientation with respect to regional tectonic stress. This might indicate a presence of local stress heterogeneities caused, for example, by an interaction of faults or fault segments in the focal zone.

Table 1
Centroid Moment Tensors (CMT) of Five Clusters

Cluster	Number of Events	Mean rms	CMTs					
			M11	M22	M33	M23	M13	M12
1	120	0.221	0.4100	-0.1750	-0.2747	-0.3915	0.0445	0.4557
2	112	0.130	0.2224	0.2572	-0.5911	-0.2583	0.2314	0.3836
3	106	0.274	0.3688	-0.4224	0.0661	-0.2056	-0.1597	0.5223
4	52	0.220	-0.3519	-0.1596	0.4556	-0.3868	0.2784	0.3073
5	50	0.184	-0.2355	0.4940	-0.3239	-0.3125	0.3250	0.3073

Number of events is the number of earthquakes in individual clusters; mean rms is the root mean square averaged over earthquakes in individual clusters. Components of the CMTs are expressed in the Cartesian coordinate system with x_1 -N, x_2 -E, x_3 -down.

Table 2
Double-Couple (DC) and Non-Double-Couple (non-DC) Components of the CMTs

Cluster	Strike1 (°)	Dip1 (°)	Rake1 (°)	Strike2 (°)	Dip2 (°)	Rake2 (°)	DC (%)	CLVD (%)	ISO (%)
1	156.7	70.8	−37.5	260.9	54.9	−156.3	93.7	−4.5	−1.8
2	162.0	52.9	−53.2	290.9	50.3	−128.4	74.5	−20.8	−4.8
3	342.9	82.3	20.8	249.9	69.4	171.7	97.6	1.8	0.6
4	40.8	66.8	78.1	248.9	25.9	115.7	82.6	−15.0	−2.5
5	305.0	49.1	−153.1	196.7	70.0	−44.1	78.9	−18.2	−2.9

The DC, isotropic (ISO), and compensated linear vector dipole (CLVD) percentages were calculated using equations (6)–(10) of Vavryčuk (2015).

DATA AND RESOURCES

The data used in this study are available from the authors upon request. The MATLAB codes for the principal component analysis (PCA) analysis of waveforms are available at <http://www.ig.cas.cz/pca-decomposition> (last accessed June 2017). The MT decomposition and visualization (Fig. 10) were performed using the open-access MATLAB code MT-Decomposition at <http://www.ig.cas.cz/mt-decomposition> (last accessed June 2017). The Suspension Bridge Picking (SBPx) algorithm of Meier available at <https://www.mathworks.com/matlabcentral/fileexchange/51996-suspension-bridge-picking-algorithm-sbpx> (last accessed June 2017) was used for the automatic picking of data. ☒

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REFERENCES

- Adamová, P., E. Sokos, and J. Zahradník (2009). Problematic non-double-couple mechanism of the 2002 Amfilochia M_w 5 earthquake, Western Greece, *J. Seismol.* **13**, 1–12.
- Babuška, V., J. Plomerová, and T. Fischer (2007). Intraplate seismicity in the western Bohemian Massif (central Europe): A possible correlation with a paleoplate junction, *J. Geodyn.* **44**, 149–159, doi: [10.1016/j.jog.2007.02.004](https://doi.org/10.1016/j.jog.2007.02.004).
- Bankwitz, P., G. Schneider, H. Kämpf, and E. Bankwitz (2003). Structural characteristics of epicentral areas in central Europe: Study case Cheb basin (Czech Republic), *J. Geodyn.* **35**, 5–32.
- Bataille, K., and J. M. Chiu (1991). Polarization analysis of high-frequency, three-component seismic data, *Bull. Seismol. Soc. Am.* **81**, no. 2, 622–642.
- Čermáková, H., and J. Horálek (2015). The 2011 West Bohemia (central Europe) earthquake swarm compared with the previous swarms of 2000 and 2008, *J. Seismol.* **19**, 899–913.
- Červený, V. (2001). *Seismic Ray Theory*, Cambridge University Press, Cambridge, United Kingdom.
- Cesca, S., and T. Dahm (2008). Frequency domain inversion code to retrieve time-dependent parameters of very long period volcanic sources, *Comput. Geosci.* **34**, no. 3, 235–246.
- Cesca, S., E. Bufo, and T. Dahm (2006). Amplitude spectra moment tensor inversion of shallow earthquakes in Spain, *Geophys. J. Int.* **166**, 839–854.
- Cesca, S., A. T. Sen, and T. Dahm (2014). Seismicity monitoring by cluster analysis of moment tensors, *Geophys. J. Int.* **196**, no. 3, 1813–1826.
- Davi, R., G. S. O'Brien, I. Lokmer, C. J. Bean, P. Lesage, and M. M. Mora (2010). Moment tensor inversion of explosive long period events recorded on Arenal volcano, Costa Rica, constrained by synthetic tests, *J. Volcanol. Geoth. Res.* **194**, no. 4, 189–200.
- Dong, D., P. Fang, Y. Bock, F. Webb, L. Prawirodirdjo, S. Kedar, and P. Jamason (2006). Spatiotemporal filtering using principal component analysis and Karhunen-Loeve expansion approaches for regional GPS network analysis, *J. Geophys. Res.* **111**, no. B03405, doi: [10.1029/2005JB003806](https://doi.org/10.1029/2005JB003806).
- Dreger, D., and B. Woods (2002). Regional distance seismic moment tensors of nuclear explosions, *Tectonophysics* **356**, 139–156.
- Dziewonski, A. M., T. A. Chou, and J. H. Woodhouse (1981). Determination of earthquake source parameters from waveform data for studies of regional and global seismicity, *J. Geophys. Res.* **86**, 2825–2852.
- Eyre, T. S., C. J. Bean, L. De Barros, G. S. O'Brien, F. Martini, I. Lokmer, M. M. Mora, J. F. Pacheco, and G. J. Soto (2013). Moment tensor inversion for the source location and mechanism of long period (LP) seismic events from 2009 at Turrialba volcano, Costa Rica, *J. Volcanol. Geoth. Res.* **258**, 215–223.
- Fischer, T., J. Horálek, P. Hrubcová, V. Vavryčuk, K. Bräuer, and H. Kämpf (2014). Intra-continental earthquake swarms in West-Bohemia and Vogtland: A review, *Tectonophysics* **611**, 1–27, doi: [10.1016/j.tecto.2013.11.001](https://doi.org/10.1016/j.tecto.2013.11.001).
- Fojtíková, L., and J. Zahradník (2014). A new strategy for weak events in sparse networks: The first-motion polarity solutions constrained by single-station waveform inversion, *Seismol. Res. Lett.* **85**, no. 6, 1265–1274, doi: [10.1785/0220140072](https://doi.org/10.1785/0220140072).
- Fojtíková, L., V. Vavryčuk, A. Cipciar, and J. Madarás (2010). Focal mechanisms of micro-earthquakes in the Dobrá Voda seismoactive area in the Malé Karpaty Mts. (Little Carpathians), Slovakia, *Tectonophysics* **492**, 213–229, doi: [10.1016/j.tecto.2010.06.007](https://doi.org/10.1016/j.tecto.2010.06.007).
- Ford, S. R., D. S. Dreger, and W. R. Walter (2008). Source characterization of the 6 August 2007 Crandall Canyon seismic event in central Utah, *Seismol. Res. Lett.* **79**, no. 5, 637–644.
- Ford, S. R., D. S. Dreger, and W. R. Walter (2010). Network sensitivity solutions for regional moment-tensor inversions, *Bull. Seismol. Soc. Am.* **100**, no. 5A, 1962–1970.
- Gualandi, A., E. Serpelloni, and M. E. Belardinelli (2014). Space-time evolution of crustal deformation related to the M_w 6.3, 2009 L'Aquila earthquake (central Italy) from principal component analysis inversion of GPS position time-series, *Geophys. J. Int.* **197**, no. 1, 174–191.
- Guilhem, A., L. Hutchings, D. S. Dreger, and L. R. Johnson (2014). Moment tensor inversions of $M \sim 3$ earthquakes in the Geysers geothermal fields, California, *J. Geophys. Res.* **119**, 2121–2137.

- Hagen, D. C. (1982). The application of principal components analysis to seismic data sets, *Geoexploration* **20**, nos. 1/2, 93–111.
- Hainzl, S., T. Fischer, H. Čermáková, M. Bachura, and J. Vlíček (2016). Aftershocks triggered by fluid intrusion: Evidence for the aftershock sequence occurred 2014 in West Bohemia/Vogtland, *J. Geophys. Res.* **121**, 2575–2590, doi: [10.1002/2015JB012582](https://doi.org/10.1002/2015JB012582).
- Hardebeck, J. L., and P. M. Shearer (2003). Using *S/P* amplitude ratios to constrain the focal mechanisms of small earthquakes, *Bull. Seismol. Soc. Am.* **93**, 2432–2444.
- Hartigan, J. A. (1975). *Clustering Algorithms*, John Wiley & Sons, New York, New York.
- Hattori, K., A. Serita, C. Yoshino, M. Hayakawa, and N. Isezaki (2006). Singular spectral analysis and principal component analysis for signal discrimination of ULF geomagnetic data associated with 2000 Izu Island earthquake swarm, *Phys. Chem. Earth* **31**, 281–291.
- Hrubcová, P., V. Vavryčuk, A. Boušková, and J. Horálek (2013). Moho depth determination from waveforms of microearthquakes in the West Bohemia/Vogtland swarm area, *J. Geophys. Res.* **118**, 120–137, doi: [10.1029/2012JB009360](https://doi.org/10.1029/2012JB009360).
- Jackson, J. E. (1991). *A User's Guide to Principal Components*, John Wiley & Sons, New York, New York.
- Jain, A. K. (2010). Data clustering: 50 years beyond *k*-means, *Pattern Recogn. Lett.* **31**, no. 8, 651–666.
- Jechumtálová, Z., and J. Šílený (2005). Amplitude ratios for complete moment tensor retrieval, *Geophys. Res. Lett.* **32**, L22303, doi: [10.1029/2005GL023967](https://doi.org/10.1029/2005GL023967).
- Jolliffe, I. (2002). *Principal Component Analysis*, Wiley Online Library.
- Julian, B. R., A. D. Miller, and G. R. Foulger (1998). Non-double-couple earthquakes 1: Theory, *Rev. Geophys.* **36**, 525–549.
- Kagan, Y. Y. (1991). 3-D rotation of double-couple earthquake sources, *Geophys. J. Int.* **106**, 709–716.
- Kikuchi, M., and H. Kanamori (1991). Inversion of complex body waves—III, *Bull. Seismol. Soc. Am.* **81**, 2335–2350.
- Kositsky, A. P., and J.-P. Avouac (2010). Inverting geodetic time series with a principal component analysis-based inversion method, *J. Geophys. Res.* **115**, no. B03401, doi: [10.1029/2009JB006535](https://doi.org/10.1029/2009JB006535).
- Kwiątek, G., P. Martínez-Garzón, and M. Bohnhoff (2016). HybridMT: A MATLAB/Shell environment package for seismic moment tensor inversion and refinement, *Seismol. Res. Lett.* **87**, no. 4, 964–976.
- Li, J., H. S. Kuleli, H. Zhang, and M. N. Toksöz (2011a). Focal mechanism determination of induced microearthquakes in an oil field using full waveforms from shallow and deep seismic networks, *Geophysics* **76**, no. 6, WC87–WC101.
- Li, J., H. S. Kuleli, H. Zhang, and M. N. Toksöz (2011b). Focal mechanism determination using high-frequency waveform matching and its application to small magnitude induced earthquakes, *Geophys. J. Int.* **184**, no. 3, 1261–1274.
- Lin, J.-W. (2012a). Nonlinear principal component analysis in the detection of ionospheric electron content anomalies related to a deep earthquake (> 300 km, *M* 7.0) on 1 January 2012, Izu Islands, Japan, *J. Geophys. Res.* **117**, no. A06314, doi: [10.1029/2012JA017614](https://doi.org/10.1029/2012JA017614).
- Lin, J.-W. (2012b). Ionospheric total electron content seismo-perturbation after Japan's March 11, 2011, *M* = 9.0 Tohoku earthquake under a geomagnetic storm; a nonlinear principal component analysis, *Astrophys. Space Sci.* **341**, 251–258.
- Lomax, A., J. Virieux, P. Volant, and C. Berge (2000). Probabilistic earthquake location in 3D and layered models: Introduction of a Metropolis-Gibbs method and comparison with linear locations, in *Advances in Seismic Event Location* Thurber, C. H. Thurber and N. Rabinowitz (Editors), Kluwer, Amsterdam, The Netherlands, 101–134.
- Málek, J., J. Horálek, and J. Janský (2005). One-dimensional *qP*-wave velocity model of the upper crust for the West Bohemia/Vogtland earthquake swarm region, *Stud. Geophys. Geod.* **49**, no. 4, 501–524.
- Miller, A. D., G. R. Foulger, and B. R. Julian (1998). Non-double-couple earthquakes 2: Observations, *Rev. Geophys.* **36**, 551–568.
- Minson, S. E., and D. S. Dreger (2008). Stable inversions for complete moment tensors, *Geophys. J. Int.* **174**, no. 2, 585–592.
- Muti, D., and S. Bourennane (2007). Survey on tensor signal algebraic filtering, *Signal Process.* **87**, no. 2, 237–249.
- Nakamichi, H., H. Hamaguchi, S. Tanaka, S. Ueki, T. Nishimura, and A. Hasegawa (2003). Source mechanisms of deep and intermediate-depth low-frequency earthquakes beneath Iwate volcano, northeastern Japan, *Geophys. J. Int.* **154**, no. 3, 811–828.
- Ocana, E., D. Stich, E. Carmona, F. Vidal, M. Breton, M. Navarro, and A. Garcia-Jerez (2008). Spatial analysis of the La Paca, SE Spain, 2005 seismic series through the relative location of multiplets and principal component analysis, *Phys. Earth Planet. In.* **166**, nos. 3/4, 117–127.
- Ruff, L. J., and B. W. Tichelaar (1990). Moment tensor rate functions for the 1989 Loma Prieta earthquake, *Geophys. Res. Lett.* **17**, 1187–1190.
- Šílený, J. (1998). Earthquake source parameters and their confidence regions by a genetic algorithm with a 'memory', *Geophys. J. Int.* **134**, 228–242.
- Šílený, J. (2009). Resolution of non-double-couple mechanisms: Simulation of hypocenter mislocation and velocity structure mismodeling, *Bull. Seismol. Soc. Am.* **99**, no. 4, 2265–2272.
- Šílený, J., and A. Milev (2008). Source mechanism of mining induced seismic events—Resolution of double couple and non double couple models, *Tectonophysics* **456**, nos. 1/2, 3–15.
- Šílený, J., G. F. Panza, and P. Campus (1992). Waveform inversion for point source moment tensor retrieval with variable hypocentral depth and structural model, *Geophys. J. Int.* **109**, 259–274.
- Silver, P. G., and T. H. Jordan (1982). Optimal estimation of the scalar seismic moment, *Geophys. J. Roy. Astron. Soc.* **70**, 755–787.
- Sokos, E., and J. Zahradník (2008). ISOLA a Fortran code and Matlab GUI to perform multiple point source inversion of seismic data, *Comput. Geosci.* **34**, 967–977.
- Stierle, E., M. Bohnhoff, and V. Vavryčuk (2014). Resolution of non-double-couple components in the seismic moment tensor using regional networks: 2. Application to aftershocks of the 1999 *M_w* 7.4 Izmit earthquake, *Geophys. J. Int.* **196**, no. 3, 1878–1888, doi: [10.1093/gji/ggt503](https://doi.org/10.1093/gji/ggt503).
- Stierle, E., V. Vavryčuk, J. Šílený, and M. Bohnhoff (2014). Resolution of non-double-couple components in the seismic moment tensor using regional networks: 1. A synthetic case study, *Geophys. J. Int.* **196**, no. 3, 1869–1877, doi: [10.1093/gji/ggt502](https://doi.org/10.1093/gji/ggt502).
- Tape, W., and C. Tape (2012). Angle between principal axis triples, *Geophys. J. Int.* **191**, 813–831.
- Templeton, D. C., and D. S. Dreger (2006). Non-double-couple earthquakes in the Long Valley volcanic region, *Bull. Seismol. Soc. Am.* **96**, no. 1, 69–79.
- Vasco, D. W. (1989). Deriving source-time function using principal component analysis, *Bull. Seismol. Soc. Am.* **79**, no. 3, 711–730.
- Vasco, D. W. (1990). Moment-tensor invariants: Searching for non-double-couple earthquakes, *Bull. Seismol. Soc. Am.* **80**, no. 2, 354–371.
- Vavryčuk, V. (2006). Spatially dependent seismic anisotropy in the Tonga subduction zone: A possible contributor to the complexity of deep earthquakes, *Phys. Earth Planet. In.* **155**, 63–72, doi: [10.1016/j.pepi.2005.10.005](https://doi.org/10.1016/j.pepi.2005.10.005).
- Vavryčuk, V. (2011). Principal earthquakes: Theory and observations from the 2008 West Bohemia swarm, *Earth Planet. Sci. Lett.* **305**, 290–296, doi: [10.1016/j.epsl.2011.03.002](https://doi.org/10.1016/j.epsl.2011.03.002).
- Vavryčuk, V. (2015). Moment tensor decompositions revisited, *J. Seismol.* **19**, no. 1, 231–252, doi: [10.1007/s10950-014-9463-y](https://doi.org/10.1007/s10950-014-9463-y).
- Vavryčuk, V., and P. Hrubcová (2017). Seismological evidence of fault weakening due to erosion by fluids from observations of intraplate earthquake swarms, *J. Geophys. Res.* **122**, doi: [10.1002/2017JB013958](https://doi.org/10.1002/2017JB013958).
- Vavryčuk, V., and S. G. Kim (2014). Nonisotropic radiation of the 2013 North Korean nuclear explosion, *Geophys. Res. Lett.* **41**, doi: [10.1002/2014GL061265](https://doi.org/10.1002/2014GL061265).
- Vavryčuk, V., and D. Kühn (2012). Moment tensor inversion of waveforms: A two-step time-frequency approach, *Geophys. J. Int.* **190**, 1761–1776, doi: [10.1111/j.1365-246X.2012.05592.x](https://doi.org/10.1111/j.1365-246X.2012.05592.x).

- Vavryčuk, V., M. Bohnhoff, Z. Jechumtálová, P. Kolář, and J. Šílený (2008). Non-double-couple mechanisms of micro-earthquakes induced during the 2000 injection experiment at the KTB site, Germany: A result of tensile faulting or anisotropy of a rock? *Tectonophysics* **456**, 74–93, doi: [10.1016/j.tecto.2007.08.019](https://doi.org/10.1016/j.tecto.2007.08.019).
- Vavryčuk, V., F. Bouchaala, and T. Fischer (2013). High-resolution fault image from accurate locations and focal mechanisms of the 2008 swarm earthquakes in West Bohemia, Czech Republic, *Tectonophysics* **590**, 189–195, doi: [10.1016/j.tecto.2013.01.025](https://doi.org/10.1016/j.tecto.2013.01.025).
- Wagner, G. S., and T. J. Owens (1996). Signal detection using multi-channel seismic data, *Bull. Seismol. Soc. Am.* **86**, no. 1A, 221–231.
- Wéber, Z. (2009). Estimating source time function and moment tensor from moment tensor rate functions by constrained L_1 norm minimization, *Geophys. J. Int.* **178**, 889–900.
- Willemann, R. J. (1993). Cluster analysis of seismic moment tensor orientations, *Geophys. J. Int.* **115**, 617–634.
- Zahradník, J., J. Janský, and V. Plicka (2008). Detailed waveform inversion for moment tensors of M 4 events: Examples from the Corinth Gulf, Greece, *Bull. Seismol. Soc. Am.* **98**, 2756–2771.
- Zecevic, M., L. De Barros, C. J. Bean, G. S. O'Brien, and F. Brenguier (2013). Investigating the source characteristics of long-period (LP) seismic events recorded on Piton de la Fournaise volcano, La Reunion, *J. Volcanol. Geoth. Res.* **258**, 1–11.

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