Originally published as:

Bock, G., Grünthal, G., Wylegalla, K. (1996): The 1985/1986 western Bohemia earthquakes: modelling source parameters with synthetic seismograms. - Tectonophysics, 261, 1-3, 139-146

ELSEVIER

Tectonophysics 261 (1996) 139–146

TECTONOPHYSICS

The 1985/86 Western Bohemia earthquakes: Modelling source parameters with synthetic seismograms

Günter Bock*, Gottfried Grünthal, Kurt Wylegalla

GeoForschungsZentrum, Telegrafenberg A17, 14473 Potsdam, Germany

Received 5 April 1995; revised version received 22 August 1995

Abstract

A major swarm of microearthquakes (main shock: $M_L = 4.6$) occurred in December 1985 and January 1986 in the western Bohemia region (50.2°N, 12.4°E). Here, we revisit the 1985/86 series by modelling some pronounced features of the observed P waveforms frequently seen in the records of the CLL, BRG and the GRF seismograph stations. Travel times and waveforms can be explained by a simple two-layer crustal model consisting of an upper crust with $V_P = 5.9$ km/s and thickness $t = 18$ km, and a lower crust with $V_P = 6.8$ km/s and $t = 11$ km. The crust–mantle boundary is at 29 km depth. Waveforms recorded at distances between about 100 and 120 km consist of a low-amplitude first P wave which has travelled through the lower crust. This phase is followed about 0.5 to 1 s later by the large-amplitude main event. About 3 s after the main event another phase stands out with similar amplitude. It probably represents the depth phase sP_mP which leaves the focus upwards as an S wave, is converted at the surface to P, and then follows the path of P_mP , the reflection from the crust–mantle boundary, to the receiver. We used this depth phase to constrain the focal depth of the main shock to 9 km. Depths of the stronger events ($M_L > 3$) of the swarm are within the depth range from 7 to 10 km. Inversion of body wave amplitudes and P wave polarities for the focal mechanism of the main event reveals a strike–slip mechanism which agrees well with previously published results.

1. Introduction

Earthquake swarms have repeatedly occurred in the Czech–German border region of western Bohemia and Vogtland. About 8000 events were instrumentally recorded in the 1985/86 swarm which was thoroughly studied by many researchers (Bormann, 1989; Procházková, 1987). A dense network of local and regional seismic stations provided tight constraints on the locations and focal mechanisms of the stronger events. The events of the 1985/86 series are located in western Bohemia at the northern tip of

the tertiary Cheb basin, close to the northern end of the surface expression of the Mariánské Lázně fault zone (Grünthal et al., 1990; Zahradník et al., 1990). The epicentral area is depicted in Fig. 1 together with the major fault zones in the area.

A characteristic feature of seismogram recordings of earthquakes in western Bohemia is the presence of a prominent phase in the P wave coda arriving about 3 s after P_g (Fig. 2). The secondary phase is clearly seen at stations located in different azimuths and at epicentral distances greater than 100 km. At closer distances, a corresponding phase is not seen as evidenced by the Moxa (MOX) recording in Fig. 2 and other examples from the literature

* Corresponding author.

Epicentre map for the Vogtland region

1962 - 1991

without foreshocks and aftershocks

Fig. 1. Epicentre map for western Bohemia and southern Saxony for the period 1962–1991 without foreshocks and aftershocks that occurred during swarm activity. The surface expressions of major fault zones are marked by heavy lines, those of secondary faults by thinner lines, following Dudek (1987), Šantrauček (1986), Šantrauček and Kolářová (1962), Zeman (1983) and the Geological Map of Saxony (1994). The line marked MLF represents the Mariánské Lázně fault zone. The arrow at the NNW end of the MLF indicates the location of the 1985/86 swarm.

(e.g. Plešinger et al., 1987). Seismograms showing a similar phase were published by several investigators (e.g. Procházková et al., 1987; Strauch, 1989b) but to the best of our knowledge, a satisfactory explanation for its presence has not yet been provided.

The absence of the later phase at close distances rules out continuing focal activity as a viable explanation. Using travel time and synthetic seismogram modelling we show in this paper that the phase is best explained as a depth phase and that it can be used as a sensitive indicator for variations in focal depth. We also show that by incorporating measured amplitudes of P and S phases together with those of the depth phase, focal mechanisms can be derived using data from relatively few stations.

2. The depth phase sP_mP

The main event of the 1985/86 series was recorded by many stations well distributed in azimuth (Zahradnik et al., 1990). The epicentre was located at 50.22°N, 12.46°E, focal depth is given as 10 km, and the origin time was 10:16:20.5 UTC (Strauch, 1989a). In our study we used data of the former Potsdam seismograph network (Strauch, 1989b) and the Gräfenberg (GRF) array (Harjes and Seidl, 1978). The stations Berggiesshübel (BRG), Collm (CLL), Moxa (MOX), and Tiefenort (TIE) are located to the northeast and northwest, and the stations of the GRF array to the southwest of the epicentral area.

First arrival times of P were used to derive a laterally homogeneous crustal model with the intercept time method (e.g. Telford et al., 1976). The intercept time method is based on the assumption of a layered model with the seismic velocities in each layer being constant. Observations are interpreted by approximating the first arrival times of P by straight line segments in the time–distance curve. The slopes of the straight line segments and their intercepts with the time axis allow velocities and layer thicknesses of the model to be determined.

Adopting 10:16:20.5 UTC as origin time, and a focal depth of 9 km for the Dec. 23, 1985, event, travel times are best explained by a two-layered 29 km thick crust (Table 1) and an upper mantle P wave velocity of 7.9 km/s. These values are well within the range reported for Moho depths and upper mantle velocities in central Europe (Prodehl and Giese, 1990). The observed picks of the first P wave arrivals are shown plotted in Fig. 3 together with the arrival times predicted by the model. The misfit between observed and theoretical arrival times does not exceed 0.5 s over a large distance range. A lower crustal layer with an apparent P wave velocity $V_p = 6.8$ km/s is inferred from first arrivals at distances between 100 and 120 km. The travel time branch associated with these arrivals is termed P_b (Fig. 3, also Fig. 4); they are P waves that are refracted from the lower crust. P_b becomes a first arrival only because the source was located relatively deep. In explosion seismic studies where the sources are located close to the surface, P_b could not be observed as first arrival for the model of Table 1 because the lower

Fig. 2. Trace-normalized and bandpass-filtered (0.5 to 3 Hz) displacement seismograms of the main event on December 21, 1985, 10:16. The unlabelled seismograms are recordings of the Gräfenberg array. Superimposed travel times are for the crustal model of Table 1.

crustal layer is too thin. The error of the intercept time estimates for the straight line segments of the time–distance curve associated with P_b and P_n is about 0.5 s resulting in approximately 3 km for the uncertainty of the crustal interface depths given in Table 1. The upper crustal layer is not well resolved by the earthquake data because data from only one station were available at close epicentral distances. The value of 5.9 km/s chosen for the P velocity in the upper crustal layer is an average which is consistent with results of deep seismic sounding studies

(Schulze and Bormann, 1990; DEKORP Research Group, 1994), but it does of course not account for the inferred lateral variability of upper crustal structure in the area. We also note that the inferred transition depth from upper to lower crust at 18 km is well in line with results of the deep seismic sounding studies just cited.

Synthetic seismograms for the model of Table 1 are shown in Fig. 4. They were calculated with the reflectivity method (Kind, 1979a; Müller, 1985). The Q_P and Q_S in Table 1 denote the quality factors for P and S waves respectively that were employed in the calculations. Comparison of synthetics with the observed seismograms and travel times (Figs. 2 and 4) suggests that the observed later phase between 3 and 5 s reduced time is the depth phase sP_mP . It leaves the source upwards as an S wave, is converted at the free surface to a P wave and then follows

Fig. 3. Reduced travel times (reduction velocity = 6 km/s) as a function of epicentral distance. Squares denote the observed first arrival times of P. The first P arrivals predicted by the crustal model in Table 1 are shown as line segments for a source depth of 9 km. The lower crustal phase P_b becomes a first arrival between about 100 and 120 km distance.

Table 1
Gradient model

Depth (km)	P-wave (km/s)	S-wave (km/s)	Density (g/ml)	Q_P	Q_S
0	5.90	3.41	2.67	1350	600
14	5.90	3.41	2.67	1350	600
22	6.80	3.93	2.80	1350	600
29	6.80	3.93	2.80	1350	600
29	7.90	4.56	3.3	1350	600
40	8.1	4.68	3.32	1350	600

Fig. 4. Synthetic seismogram section (bottom) and travel times (top) for the model of Table 1 and a focal depth $h = 9$ km. The reflection from the interface between upper and lower crust is labelled P_bP , its associated depth phase sP_bP .

the P_mP path (i.e., reflection from the crust–mantle boundary) to the receiver.

Travel times of sP_mP are sensitive to focal depth. This is illustrated in Fig. 5 which compares the GRF-B2 seismogram of the main event with synthetics calculated for a range of focal depths. It is obvious that a depth near 9 km matches the observation best. This absolute depth estimate is model-dependent. Systematic errors may arise from errors in the Moho depth and errors in crustal velocities. Based on the uncertainties of the model parameters previously discussed and on numerical modelling, we estimate that the absolute depth value derived from sP_mP is accurate to within about ± 2 km. As shown in the following section, this is of the same order as the likely depth range of the major events of the 1985/86 earthquake swarm. Estimates of depth variations relative to a reference event are not subject

Fig. 5. Vertical-component seismogram of the main event recorded at station B2 of the Gräfenberg array (bottom) and synthetics for focal depths ranging from 3 km to 13 km. Amplitudes are proportional to ground velocity.

to these large uncertainties and are therefore given in the following.

3. Constraints on focal depths

We have used the phase sP_mP to investigate variations in focal depth of 24 events ($M_L > 3$) that occurred between December 6, 1985, and January 23, 1986. The depth phase was well recorded at station B2 of the GRF array. Amplitude-normalized displacement seismograms are shown in Fig. 6. They have been aligned along the first minimum of the phase labelled P_g . The onset times of sP_mP are marked by vertical bars. The sensitivity of the sP_mP travel time to variations in focal depth is about 0.26 s/km. The sensitivity of travel time differences to focal depth is approximately 0.24 s/km for $sP_mP - P_g$, and 0.35 s/km for $sP_mP - P_mP$. Variations of epicentral distance on differential travel times are small (~ 0.02 s/km) and were neglected in working out the variations of focal depth indicated in Fig. 6.

The correct identification of the reference phase would be difficult on the basis of travel times alone. The reference phase chosen in Fig. 6 is labelled P_g . The Moho reflection P_mP however arrives almost at the same time as P_g (Figs. 2 and 4). Based on the synthetics of Fig. 4, we believe, however,

Fig. 6. Trace-normalized displacement seismograms, bandpass-filtered between 0.5 and 3 Hz, of station B2 of the Gräfenberg array for 24 events occurring in December/January 1985/86 in western Bohemia. The origin times are displayed to the left of the traces. The numbers to the right give depth perturbations (in km) relative to the main event.

that the main contribution to the reference phase along which the seismograms in Fig. 6 were aligned probably comes from P_g . We have therefore used the value of 0.24 s/km which is associated with P_g as reference phase to convert the variations in the arrival times of sP_mP to variations in focal depth. Using a higher value would reduce the depth perturbations derived under the assumption of P_g as reference phase. Therefore, the depth perturbations relative to the depth of the main event, that are shown to the right of the traces in Fig. 6, must be regarded as maximum possible values. Adopting a focal depth of 9 km for the main event, it follows that the stronger events ($M_L > 3$) of the 1985/86 series in western Bohemia were constrained to the depth range between about 2 km above, and 1 km below the main event. Adopting 9 km as the depth of the main event, absolute depths would range between 7 and 10 km. There is no evidence for any of the events investigated lying outside this depth interval.

4. Focal mechanism

A well-constrained fault-plane solution for the main event was published by Zahradník et al. (1990). The solution is based on P wave polarities at many

stations well distributed in azimuth. The mechanism corresponds to a strike-slip motion with a small normal-faulting component along two possible fault planes striking approximately N–S and E–W and dipping steeply to the west and north, respectively.

The fact that the fault-plane solution of the main event is well constrained by polarity data provides a good opportunity to investigate the question of whether data from relatively few stations are sufficient to derive a focal mechanism. In this study we re-investigated the main event of the 1985/86 series using measured amplitudes of P_n , P_g , sP_mP , and S_g , and P wave polarities to derive a fault-plane solution. The method is similar to the one introduced by Kisslinger (1980) and Kisslinger et al. (1981) for local earthquakes; an application employing synthetic seismograms calculated with the reflectivity method was described by Bock (1993a). Following Bock (1993a), the method is briefly described here.

The amplitudes of various seismic phases that can be reliably identified in the seismograms are compared with theoretical amplitudes calculated with the reflectivity method for a point shear dislocation in the layered model of Table 1. Differences between observed and theoretical amplitudes are minimized as a function of fault strike Φ , fault dip δ , direction of

the slip vector λ and scalar seismic moment M_0 . The nomenclature for the definition of fault-orientation parameters is that used by Aki and Richards (1980). A logarithmic l_1 norm is used to judge the goodness of fit. The quantity

$$M = \sum_{i=1}^n |\log_{10} a_o - \log_{10} a_c|_i \quad (1)$$

is minimized as function of Θ , δ , and λ where the a_o and a_c denote observed and calculated amplitudes, and n is the number of all phases used in the inversion.

Confidence limits are provided following Gephart and Forsyth (1984). The value of M from Eq. (1) at the 68% confidence limit is given by

$$M_{68} = \frac{\sqrt{n(\pi/2 - 1) + n}}{n - 3} M_{\min}, \quad (2)$$

where M_{68} is the sum of the absolute values of the misfit at the 68% confidence limit, and $M_{\min}$ is the value of M obtained from Eq. (1) for the best fitting model.

For the Dec. 23, 1985, 10:16, event, the best fitting parameters are given in the following table (all values in degrees):

	Strike Φ	Dip δ	Rake λ
First nodal plane	181	61	–19
	(169, 191)	(49, 85)	(–48, –8)
Second nodal plane	280	74	–150
	(264, 286)	(42, 84)	(–173, –139)

The values in parentheses give the 68% confidence intervals calculated after (2). For the scalar seismic moment we obtain $M_0 = 5 \times 10^{15}$ N m.

The range of possible fault plane solutions is shown in Fig. 7. The focal mechanism published by Zahradnik et al. (1990) clearly falls within the 68% confidence limits of our solution. It was shown, e.g. by Horálek et al. (1987) and Strauch (1989a), that the epicentres of the 1985/96 swarm are distributed along a N–S line. This favours the N–S striking plane as the rupture plane. The focal mechanism supports the seismotectonic model of Grünthal et al. (1990) which states that seismic activity occurred on fault elements whose N–S trend is different from the 153° strike direction of the Mariánské Lázně fault zone.

A comparison of synthetic and observed seismograms is made in Fig. 8 for the stations B2 and C1

Fig. 7. Fault plane solution for the main event (December 21, 1985, 10:16) of the 1985/86 earthquake swarm in western Bohemia. The P and T axes shown are for the best-fitting solution obtained in this study with a limited dataset. The corresponding nodal planes are drawn as solid lines and labelled with their strike and dip values (in degrees). The 68% confidence range of the solution is indicated by the pairs of nodal planes labelled 1 and 2; the P and T axes associated with them are denoted by P1, T1 and P2, T2 respectively. The dashed lines indicate the fault plane solution published by Zahradnik et al. (1990) which was based on P wave polarities observed at many stations.

of the GRF array. Three-component recordings are available for C1 at an epicentral distance $x = 153$ km while at B2 ($x = 120$ km) only the vertical component is recorded. Although the match of waveforms is not perfect, amplitudes of the major phases are reasonably well matched by the synthetics.

5. Conclusions

Using the arrival times of P and S alone, focal depths of local earthquakes can only be resolved if data from stations located very close to the epicentre are available. If this condition is not satisfied focal depth usually becomes less well constrained. Depth-sensitive phases recorded at distant (i.e., distances several times greater than focal depth) stations are occasionally used in detailed studies of selected source regions (e.g. Barbano et al., 1985; Bock,

Fig. 8. Observed (solid lines) and synthetic (dashed lines) seismograms for the stations B2 and C1 of the GRF array. Phases used in the inversion are labelled.

1993b; Kind, 1979b; Langston, 1987; Zonno and Kind, 1984) to obtain a better-than-normal depth constraint.

Seismogram recordings of earthquakes occurring during 1985/86 in western Bohemia show a prominent depth phase that was identified as sP_mP . The depth phase consistently shows up on many seismograms and could be used to constrain the focal area of the major swarm events to the depth range $7 \text{ km} < h < 10 \text{ km}$. The inferred depth for the main event which occurred on Dec. 21, 10:16 UTC, was about 9 km. A focal mechanism could be derived for the main event using amplitude and polarity data from relatively few stations. The method may provide a promising tool to investigate the mechanism of many more events of the western Bohemia swarm, however, this was beyond the scope of the present study.

Acknowledgements

Data of the Gräfenberg array were made available by the Seismologische Zentralobservatorium Erlangen.

References

- Aki, K. and Richards, P.G., 1980. *Quantitative Seismology, Theory and Methods*, Vol. I. W.H. Freeman and Co., San Francisco.
- Barbano, M.S., Kind, R. and Zonno, G., 1985. Focal parameters of some Friuli earthquakes (1976–1979) using complete theoretical seismograms. *J. Geophys.*, 58: 175–182.
- Bock, G., 1993a. The Woods Reef (New South Wales) earthquake of 14 November 1990: Focal mechanism derived from amplitude ratios and synthetic seismograms. *Aust. J. Earth Sci.*, 40: 369–376.
- Bock, G., 1993b. Depth phases from local earthquakes. *BMR J. Aust. Geol. Geophys.*, 13: 275–279.
- Bormann, P. (Editor), 1989. *Monitoring and analysis of the earthquake swarm 1985/86 in the region Vogtland/Western Bohemia*. Veröffentlichung 110, Zentralinstitut für Physik der Erde, Potsdam, 282 pp.
- DEKORP Research Group, 1994. Crustal structure of the Saxothuringian Zone: Results of the deep seismic profile MVE-90 (East). *Z. Geol. Wiss.*, 22: 647–769.
- Dudek, A., 1987. Geology and tectonic pattern of the western Bohemia seismic area. In: D. Procházková (Editor), *Earthquake Swarm in Western Bohemia*. Geophysical Institute, Czech. Acad. Sci., Praha, pp. 34–37.
- Geological Map of Saxony, 1994. *Geologische Übersichtskarte des Freistaates Sachsen 1:400,000*. Sächsisches Landesamt für Umwelt und Geologie, Freiberg.
- Gephart, J.W. and Forsyth, D.W., 1984. An improved method for determining the regional stress tensor using earthquake focal mechanism data: Application to the San Fernando earthquake sequence. *J. Geophys. Res.*, 89: 9305–9320.
- Grünthal, G., Schenk, V., Zeman, A. and Schenková, Z., 1990. Seismotectonic model for the earthquake swarm of 1985–1986 in the Vogtland/West Bohemia focal area. *Tectonophysics*, 174: 369–383.
- Harjes, H.-P. and Seidl, D., 1978. Digital recording and analysis of broadband seismic data of the Gräfenberg (GRF) Array. *J. Geophys.*, 44: 511–523.
- Horálek, J., Vavryčuk, V., Plešinger, A., Pšenčík, I., Jedlička, P. and Soukup, J., 1987. Refined localization of selected Jan. 5–Feb. 6, 1986 events of the west-Bohemian earthquake swarm. In: D. Procházková (Editor), *Earthquake Swarm in Western Bohemia*. Geophysical Institute, Czech. Acad. Sci., Praha, pp. 226–235.
- Kind, R., 1979a. Extensions of the reflectivity method. *J. Geophys.*, 45: 373–380.
- Kind, R., 1979b. Observations of sP_n from Swabian Alb earthquakes at the GRF array. *J. Geophys.*, 45: 337–340.
- Kisslinger, C., 1980. Evaluation of S to P amplitude ratios for determining focal mechanisms from regional network observations. *Bull. Seismol. Soc. Am.*, 70: 999–1014.
- Kisslinger, C., Bowman, J.R. and Koch, K., 1981. Procedures for computing focal mechanisms from local (SV/P)_z data. *Bull. Seismol. Soc. Am.*, 71: 1719–1729.
- Langston, C.A., 1987. Depth of faulting during the 1968 Meckering, Australia, earthquake sequence determined from wave-

- form analysis of local seismograms. *J. Geophys. Res.*, 92: 11 561–11 574.
- Müller, G., 1985. The reflectivity method: a tutorial. *J. Geophys.*, 58: 153–174.
- Plešinger, A., Horálek, J., Vavryčuk, V., Pšenčík, I. and Zedník, J., 1987. Complex analysis of local digital observations of the 1985/86 west-Bohemian earthquake swarm: preliminary results. In: D. Procházková (Editor), *Earthquake Swarm in Western Bohemia*. Geophysical Institute, Czech. Acad. Sci., Praha, pp. 343–356.
- Procházková, D. ed., 1987. *Earthquake swarm 1985/86 in Western Bohemia*. Geophysical Institute, Czech. Acad. Sci., Praha, 415 pp.
- Procházková, D., Tobyaš, V., Knaislová, D., Soukup, J. and Michek, R., 1987. Earthquake swarm in Western Bohemia 1985/86-data. In: D. Procházková (Editor), *Earthquake Swarm 1985/86 in Western Bohemia*. Geophysical Institute, Czech. Acad. Sci., Praha, pp. 299–306.
- Prodehl, C. and Giese, P., 1990. Seismic investigations around the EGT in Central Europe. In: R. Freeman, P. Giese and St. Mueller (Editors), *The European Geotraverse: Integrative Studies*. European Science Foundation, Strasbourg, pp. 77–98.
- Šantrauček, P., 1986. Main structural features of the tertiary Cheb region (in Czech). In: K. Mrázová and L. Ruprechtová (Editors), *Počítačové spracovani údajov československej seizmickej siete*. Geophys. Inst. Slovak Acad. Sci., Bratislava, pp. 22–30.
- Šantrauček, P. and Kolářová, M., 1962. *Nerostné suroviny chebské panse s vypoctem zasob hnědého uhlí v kategorii C2*. Geofond Praha, unpublished manuscript.
- Schulze, A. and Bormann, P., 1990. Deep seismic sounding in eastern Germany. In: R. Freeman, P. Giese and St. Mueller (Editors), *The European Geotraverse: Integrative Studies*. European Science Foundation, Strasbourg, pp. 109–114.
- Strauch, W., 1989a. Refined localization of Vogtland/Western Bohemia events by means of digital recordings. In: P. Bormann (Editor), *Monitoring and Analysis of the Earthquake Swarm 1985/86 in the Region Vogtland/Western Bohemia*, Veröffentlichung 110, Zentralinstitut für Physik der Erde, Potsdam, pp. 115–123.
- Strauch, W., 1989b. Data evaluation at the Potsdam centre-seismogram examples. In: P. Bormann (Editor), *Monitoring and Analysis of the Earthquake Swarm 1985/86 in the Region Vogtland/Western Bohemia*, Veröffentlichung 110, Zentralinstitut für Physik der Erde, Potsdam, pp. 97–101.
- Telford, W.M., Geldart, L.P., Sheriff, R.E. and Keys, D.A., 1976. *Applied Geophysics*. Cambridge University Press, 860 pp.
- Zahradník, J., Antonini, M., Grünthal, G., Janský, J., Procházková, D., Schmedes, E., Špičák, A. and Zedník, J., 1990. A combined study of macroseismic data and focal mechanisms applied to the West-Bohemian earthquake, Czechoslovakia. *Pure Appl. Geophys.*, 133: 53–71.
- Zeman, A. (Editor), 1983. *Tektonické a blokové poměry Českého masivu z hlediska zájmových oblastí jaderných elektráren* (souhrnná studie). Archiv. Ústř. Úst. Geol. Praha.
- Zonno, G. and Kind, R., 1984. Depth determination of North Italian earthquakes using Gräfenberg data. *Bull. Seismol. Soc. Am.*, 74: 1645–1659.