Arvidsson, R., Wahlström, R., Kulhánek, O. (1992): Deep-crustal earthquakes in the southern Baltic shield. - *Geophysical Journal International*, *108*, 3, pp. 767—777.

DOI: https://doi.org/10.1111/j.1365-246X.1992.tb03468.x

Geophys. J. Int. (1992) **108**, 767–777

Deep-crustal earthquakes in the southern Baltic Shield

Ronald Arvidsson, Rutger Wahlström and Ota Kulhánek
Seismological Department, Uppsala University, Box 2101, S-750 02 Uppsala, Sweden

Accepted 1991 September 5. Received 1991 September 5; in original form 1991 July 12

SUMMARY

On 1986 July 14, one of the largest earthquakes in the Baltic Shield during this century occurred near Skövde in the province of Västergötland, Sweden, with a magnitude of $M_L(\mathrm{UPP}) = 4.5$. It was followed by the so far largest number of recorded aftershocks, more than 20, from any Swedish earthquake. The strongest aftershock, with $M_L(\mathrm{UPP}) = 3.4$, occurred about one hour after the main shock. A few months later, on November 2, an $M_L(\mathrm{UPP}) = 3.6$ event took place near Mariestad some 30 km northwest of the Skövde series. All these shocks were located in the lower crust with foci at depths between 20 and 35 km indicating active movements in the shield at depths where mainly ductile deformation is usually assumed. Waveform modelling seems to be an excellent tool to obtain information about the focal depth which is otherwise hard to determine. The calculated seismic moment and stress drop for the Skövde main shock are 5.9×10^{14} N m and 2.8 MPa, respectively, and for the Mariestad earthquake 2.3×10^{14} N m and 7.3 MPa, respectively. Focal mechanisms for the Skövde main shock and largest aftershock have been obtained from *P*-wave polarities and synthetic modelling, and for the Mariestad earthquake from *P*-wave polarities. The synthetics show good agreement with short-period seismograph records for frequencies up to at least 2 Hz and distances up to at least 200 km. A comparison of the mechanisms indicates that the faulting of the area is complex, with styles ranging from strike–slip to normal, suggesting that other factors than the push from the North Atlantic Ridge, e.g. post-glacial rebound, also contribute to the lithospheric stress pattern. This explanation is corroborated with findings from nearby southern Norway.

Key words: Baltic Shield, focal mechanism, lower crust foci, regional distance waveform modelling, seismotectonics, source parameters.

1 INTRODUCTION

Focal mechanisms from intraplate earthquakes frequently show horizontal compressional stress (Sykes 1978). Many focal mechanisms, but not all, from earthquakes in the southeastern Canadian Shield have thrust faulting with *P*-axis orientations that may be related to stress propagation from ridge push at the North Atlantic Ridge (e.g. Wahlström 1987). However, regions deviating from this pattern have also been reported, e.g., the Baffin Bay area in northeastern Canada which is affected by a post-glacial rebound of similar intensity to the one observed in the Baltic Shield. This area has normal faulting earthquakes on the landward side and thrust faulting ones on the seaside, which can partly be explained by post-glacial rebound (Stein *et al.* 1989). Reports from the Baltic Shield show a mixed picture: a recent study of Norwegian earthquakes indicates a variety of faulting styles, and different driving mechanisms are suggested (Bungum *et al.* 1991), whereas microearthquakes in southern Sweden seem to fit the ridge push hypothesis (Slunga, Norrman & Glans 1984a).

The Lake Vänern region in southwestern Sweden is one of the seismically most active parts of the Baltic Shield (e.g., see Wahlström 1988). An earthquake on 1986 July 14, near Skövde, $M_L(\mathrm{UPP}) = 4.5$, is the second largest event that has occurred in Sweden since the deployment of modern seismographs in the 1950's. The quake was followed by more than 20 recorded aftershocks, the largest number from any Swedish earthquake. The largest aftershock, magnitude 3.4, was recorded 55 minutes after the main shock. All of the focal depths in the series that have been determined with some accuracy are about 30 km. Most of the aftershocks were only recorded by a mobile seismograph network operated in the epicentral area. A few months later, on 1986 November 2, an earthquake with magnitude 3.6 and focal depth about 20 km took place near Mariestad about 30 km northwest of the Skövde earthquake series.

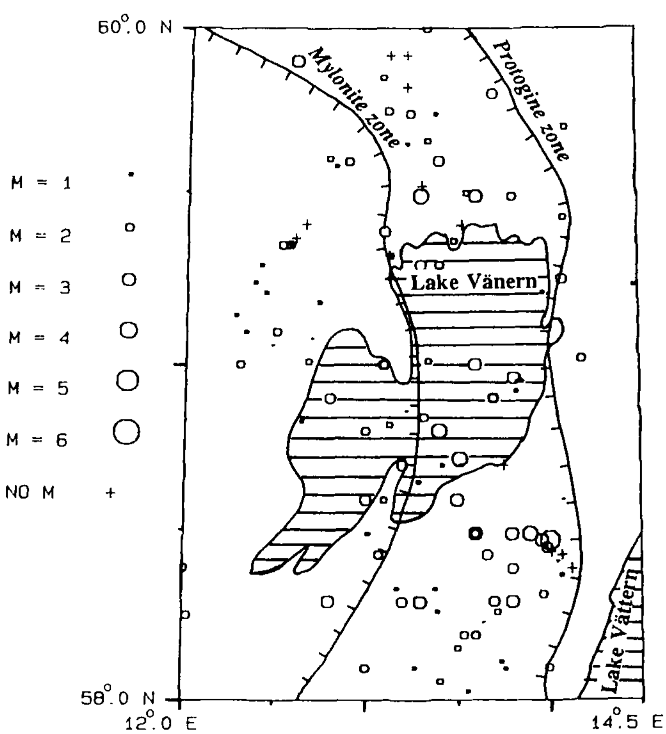
The main objective of the present study is to analyse the Skövde and Mariestad earthquakes with respect to focal depth, source mechanism, dynamic source parameters and seismotectonic implications. Being the largest recorded earthquakes in the Lake Vänern region, the events provide important information about neotectonic movements and earthquake-generating mechanisms in the southern Baltic Shield.

2 SEISMOTECTONIC SETTING

The geology of the area is characterized by stable Precambrian basement intersected by two major shear zones, roughly trending north–south, the Mylonite zone and the Protogine zone (Fig. 1). The shear zones are located west (Mylonite) and east (Protogine) of the studied events. The Protogine zone has been proposed to be an old suture zone developed under continental collision in Precambrian time (e.g. see Zeck & Malling 1976). Also intraplate faulting has been suggested as a generating mechanism (e.g. see Gorbatschev 1980). The Protogine zone constitutes the border between the older (about 1800 Ma) Svecokarelian deformation province to the east and the younger (about 1000 Ma) Sveconorwegian deformation to the west. Steep faults, striking in a north-northeasterly direction, younger and of a more brittle type than the Protogine zone, have recently been observed north-northeast of the epicentres of the Skövde earthquakes (M. Stephens and C.-H. Wahlgren 1991, personal communication). Interpretations of the EUGENO-S profile indicate that both the Mylonite and the Protogine zones are dipping to the west and transect a large portion of the crust (EUGENO-S working group 1988). The Lake Vättern graben (Collini 1951; Lind 1972) is located east of the Protogine zone and also strikes in the north–south direction. As can be seen in Fig. 1, a section of the Protogine zone coincides with the eastern shore of Lake Vänern. The lake is a depression of extensional tectonics (Ahlin 1987). This structure has been suggested to have been affected by faulting during Holocene as observed from displacements of Quaternary deposits (e.g., Björck & Digerfeldt 1982). Indications of neotectonic faulting have also been observed south of Lake Vänern (Ahlin 1983).

As for a typical intraplate region, the seismicity of the Baltic Shield is rather diffuse with few and scattered earthquakes. However, there are areas with higher seismic energy release, such as the region at and around Lake Vänern, where the earthquakes of the present study are located (Fig. 1). The Protogine zone seems to be a border between a seismically active region in the younger deformation area to the west and a virtually aseismic region in the older deformation area to the east (Fig. 1).

A certain temporal variation of the seismicity is indicated by the fact that the four largest earthquakes since 1950 in or near Sweden, M_L(UPP) > 4.0, occurred between 1983 and 1986. The two largest events were the magnitude 4.6, 1985



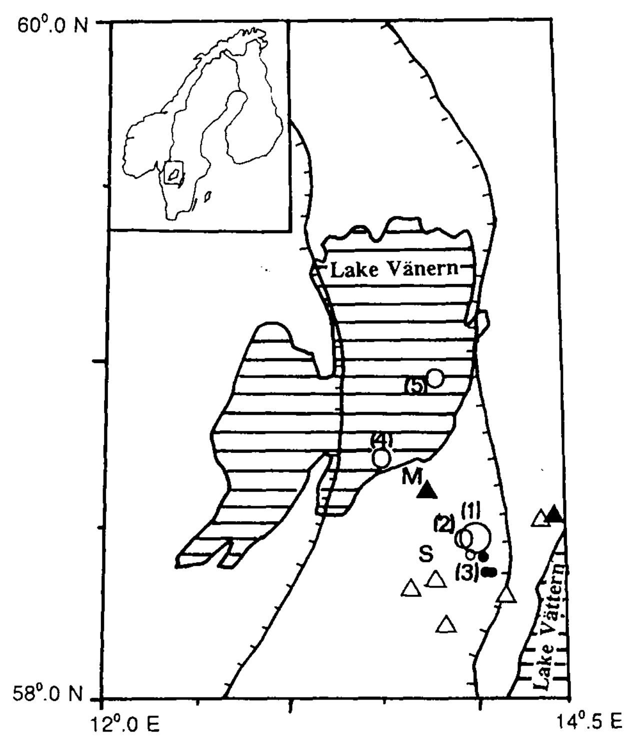


Figure 1. Left: epicentres of earthquakes in the period 1950–1989 and major tectonic structures in the Lake Vänern region. Right: open circles show locations of (1) Skövde 1986 main shock, (2) Skövde aftershock on July 14 at 14:45, (3) Skövde aftershock on July 14 at 15:28, (4) Mariestad 1986 earthquake and (5) Otterbäcken 1981 earthquake. Solid circles show locations of Skövde aftershocks recorded by field stations between July 18 and 26. Open triangles show locations of analogue stations and solid triangles show locations of digital stations of the field network. M and S denote the towns of Mariestad and Skövde.

June 15, Kattegat earthquake (Arvidsson *et al.* 1991) and the presently studied magnitude 4.5 Skövde earthquake. Also the annual number of earthquakes in the magnitude interval 3–4 was anomalously high during this four-year period.

3 INSTRUMENTAL DATA AND EARTHQUAKE LOCATIONS

Data from Danish, Finnish, Norwegian and Swedish networks with analogue and digital recordings have been used. Seismic stations used in the final steps of the analysis of locations, fault-plane solutions and dynamic source parameters are shown in Fig. 2.

For locations, the HYPOINVERSE program (Klein 1978) was used together with a two-layered crustal model (Table 1). The Skövde main shock and two aftershocks recorded the same day were located from permanent station readings (Table 2; Fig. 1). The hypocentre locations are not particularly sensitive to using other realistic models and the changes in any direction are of the order of the location error of a few kilometres. An aftershock study was undertaken by the Seismological Department, Uppsala University, with up to seven field stations deployed in the epicentral area from July 16 until September 9 (Fig. 1). Altogether 20 aftershocks were recorded and identified with the mobile network and four of these could be located with greater accuracy (Table 2; Fig. 1). The magnitude range of events recorded by the field stations is from −0.7 to 2.8 (Holmqvist & Wahlström 1987). The focal depth for the Skövde series was about 30 km (Table 2). The smaller depth estimated for the aftershock of July 14 at 15:28 (18 km) is uncertain due to the limited number of observations. The Mariestad earthquake (Table 2; Fig. 1) has a calculated depth of 21 km.

Table 1. Velocity model used for the locations (Båth 1979).

Depth	Wave velocities	
	P	S
km	km/s	km/s
0-19	6.22	3.58
19-38	6.64	3.69
38-	7.84	4.55

4 MACROSEISMIC INVESTIGATIONS

A macroseismic study of the Skövde main shock shows that it was felt over an area of about 60 000 km^2 in south-central Sweden (Fig. 3a). Information was collected from several hundreds of letters of inquiry to post officials, interviews with residents in the area, newspaper articles, etc. Available data delineate reasonably well the outer limit of the area over which the earthquake was generally perceptible. Slight damage, such as cracks in walls and roofs, were reported from several sites. A number of reports indicate intensity of

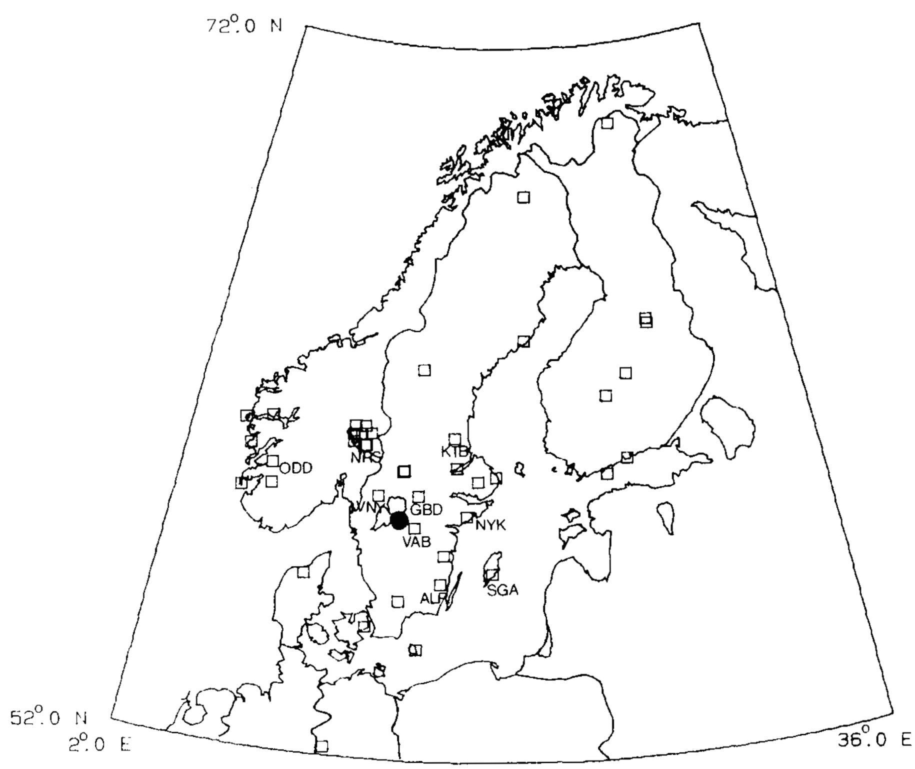


Figure 2. Locations of permanent and semipermanent stations (squares) used in the analysis. The code is given for stations mentioned in the text. The solid circle shows the area of the studied earthquakes.

Table 2. Locations of earthquakes (*cf.* Fig. 1).

Event	Date	Origin time	Epicentral location Lat.	Lon.	Focal depth	Magnitude
	y m d	GMT	°N	°E	km	M_L(UPP)
Skövde	1986 07 14	13 50 37	58.48	14.01	29+)	4.5
- " -	1986 07 14	14 45 33	58.48	13.96	29+)	3.4
- " -	1986 07 14	15 28 36	58.43	13.99	18§)	2.0
- " -	*) 1986 07 18	02 30 31	58.38	14.07	30	1.3
- " -	*) 1986 07 18	05 24 51	58.42	14.05	30	0.7
- " -	*) 1986 07 20	04 22 13	58.38	14.10	30	-0.3
- " -	*) 1986 07 26	10 50 11	58.42	14.05	30	-0.1
Mariestad	1986 11 02	07 48 00	58.71	13.51	21	3.6

*) Located from field seismograph station readings, with M_L(UPP) taken from Holmqvist and Wahlström (1987).

+) 30-35 km from synthetics.

§) Uncertain due to poor data.

shaking V (MSK scale), and at three localities the reported intensity was at least V+ (Fig. 3a). We consider the maximum intensity to be V+. By using the Blake–Shebalin formula with attenuation values for the shield from Kulhánek & Wahlström (1985) we obtain a focal depth of about 25 km. Making use of the formula of Wahlström & Ahjos (1984) we determine the macroseismic magnitude, M_M(UPP) = 4.1, i.e., somewhat less than the instrumental magnitude. The proximity in time of the large aftershock, less than one hour after the main shock, makes an independent macroseismic evaluation unreliable.

The earthquake near Mariestad was felt over an area of approximately 18 000 km^2 (Fig. 3b). The available macroseismic information consists of about 100 reports. The maximum intensity, V, was observed some 20 km east-northeast from the instrumental epicentre (Fig. 3b). The focal depth deduced from macroseismic data is about 20 km and the macroseismic magnitude, M_M(UPP) = 3.7. Both values are in good agreement with corresponding instrumental determinations (*cf.* Table 2).

5 FOCAL MECHANISMS

5.1 Polarity mechanisms

Focal mechanisms have been obtained from polarities of first *P*-wave arrivals (*Pn* and *Pg* phases) using seismograms from stations in Denmark, Finland, Norway and Sweden. Data from stations at epicentral distances between 130 and 170 km were omitted to avoid confusion between *Pg* and *Pn* phases. To determine the focal mechanism, a modified version of the program FOCMEC (Snoke *et al.* 1984; Wahlström 1987) was applied. In the algorithm of FOCMEC the focal sphere is searched for solutions satisfying the input polarities and/or *SV*/*P* amplitude ratios with a minimum number of errors. The modified version allows polarity data to be assigned different weights, full or half.

For the Skövde main shock, a large amount of data was retrieved, in total 30 azimuthally well-distributed polarities (Fig. 4). Two possible types of focal mechanism result: strike–slip faulting in which case the *P*-axis has a southeasterly trend, and normal faulting (Fig. 4).

For the largest aftershock in the Skövde series (July 14, 14:45 GMT), the derived solution differs from that of the main shock (Fig. 4). To some extent this could be an effect of the smaller data set, with only eight polarities (Fig. 4) of the aftershock, which was more than one magnitude unit smaller than the main shock. At some stations, such as NRS and SGA (see Fig. 3), the first part of the *P* coda shows large resemblance for the two earthquakes. On the other hand, at station NYK there is a difference in the polarity indicating at least a small shift of the fault plane.

For the Mariestad earthquake, totally 20 polarities with a good azimuthal coverage were obtained (Fig. 4). Four of the polarities were taken from Slunga & Nordgren (1987). The FOCMEC solution shows a clear normal faulting style with steep *P*-axis (Table 3; Fig. 4).

5.2 Synthetic seismogram modelling

The polarity mechanisms for the Skövde 1986 main shock and largest aftershock (July 14, 14:45 GMT) derived in the previous section were used as starting points for synthetic seismogram modelling in order to further constrain the mechanisms. The applied computer program uses a propagator matrix–wavenumber integration giving the full waveform (Herrmann & Wang 1985). Both observed and synthetic data were low-pass filtered at 2 Hz cut-off frequency in order to reduce scattering and other model problems inherent at higher frequencies. The source time function applied was a triangle with a duration of 0.2 s.

To find an appropriate earth model, we first checked a few different structures: an average traveltime model for Sweden (Båth 1979), a broadband receiver function for Uppsala (Q. Liu 1991, personal communication) and explosion models for the Lake Vänern region (Green *et al.* 1988). The local explosion model, profile 6E (Green *et al.* 1988), was found to give the best waveform fit. The next step was to find the appropriate Moho depth by checking arrival times for later phases of the *Lg* wavetrain at the stations NYK at 185 km and ALR at 200 km distance. The Moho depth was calculated to 41 km, i.e., in good correspondence with results from explosion models of the area indicating Moho at depths of about 41 to 43 km (Green *et al.* 1988). Once the Moho depth was established, the model was slightly perturbed in order to see if improvements were possible. *Q* values were taken from the broadband receiver function model mentioned above. The final model is given in Table 4.

Synthetics were computed for different assumed focal depths and the best fit is obtained for depths in the range 30 to 35 km for both events (e.g., see Fig. 5), i.e., in fair agreement with the depth of 29 km obtained from the location procedure (Table 2) and 25 km deduced from the macroseismic data (see above).

For the Skövde main shock, digital records of good

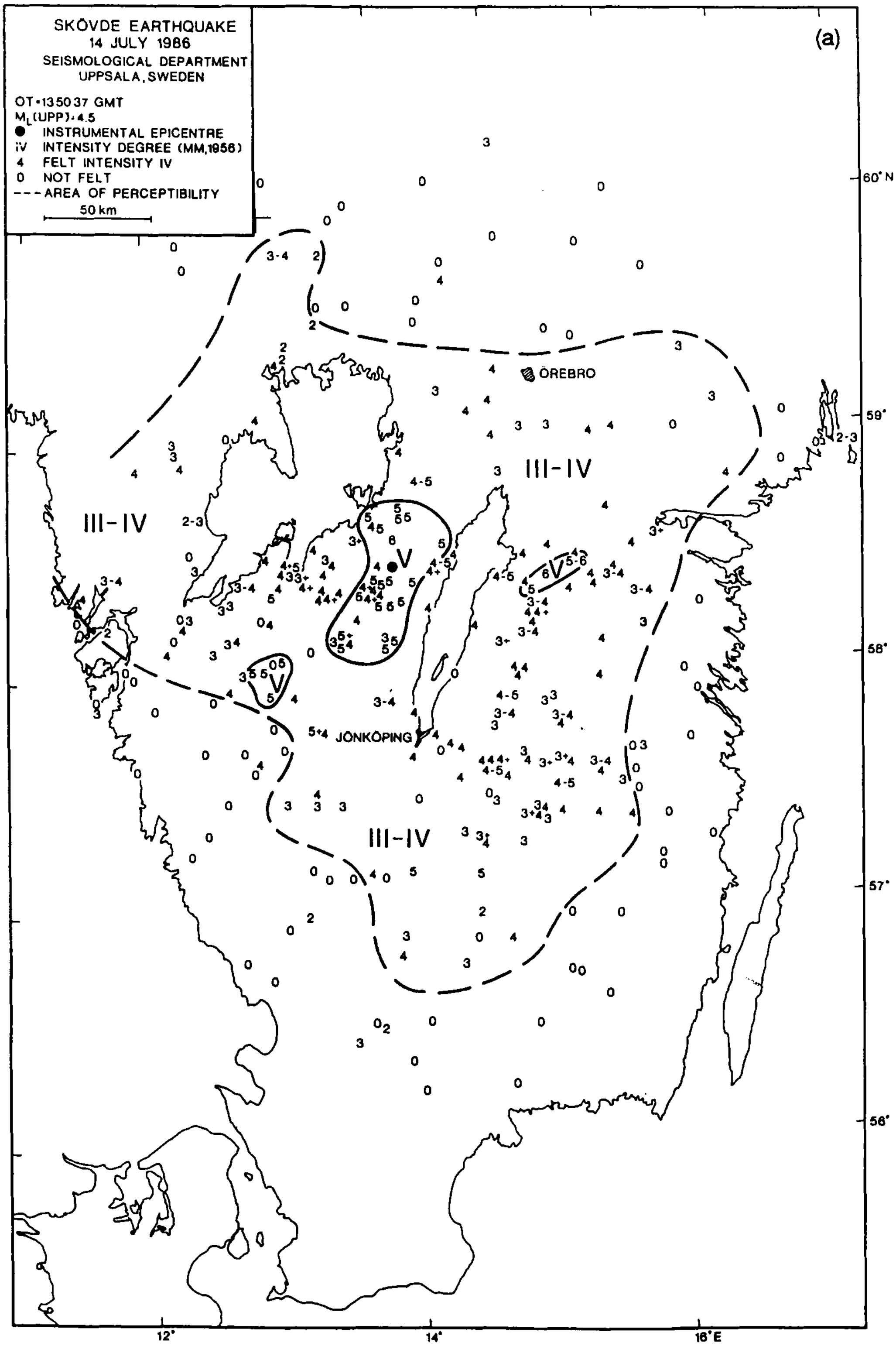


Figure 3. Isoseismal maps of the (a) 1986 July 14 Skövde main shock and (b) 1986 November 2 Mariestad earthquake.

quality are available from four stations in the epicentral distance range 113 to 200 km. Convergence of the synthetics with respect to observations was found for a mechanism with strike = 180°, dip = 90° and rake = −45° with the focal depth at 35 km (Figs 4 and 6). This mechanism is close to the two different families of solutions given by the FOCMEC program. The stations ALR, GBD and NYK all show a good agreement between observed and synthetic data (Fig. 6). For slightly different mechanisms a better fit may be found for one or two of these stations, but the overall best correspondence with respect to both the relative amplitudes and the phase polarities is for this solution. For the station

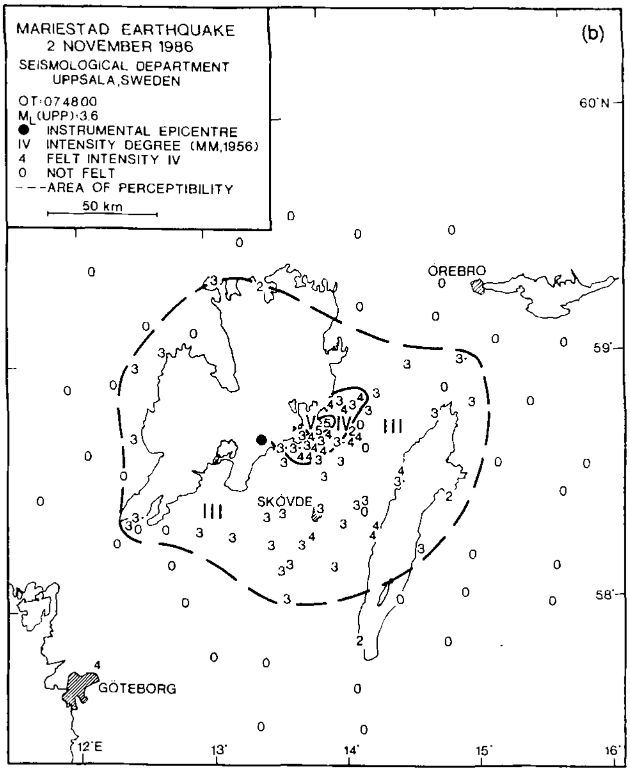


Figure 3. *(continued)*

VNY the correspondence between the observed and modelled seismograms seems to be poorer.

As can be seen in Fig. 6, there is a discrepancy between the observed and synthetic seismograms at NYK and VNY for the section between the arrivals of first P and S phases. This section is believed to contain mainly surface-reflected phases. We are not too concerned with the discrepancy since such phases are sensitive to structural details, which are not well known.

Even though the derived polarity mechanism for the largest aftershock differs from the main shock (Fig. 4), the waveforms at corresponding stations show a reasonable resemblance (Fig. 7). Thus, it seems likely that the two earthquakes had similar mechanisms. A good fit at the stations NYK and GBD is found by changing the rake for the mechanism of the main shock: strike = 180°, dip = 90° and rake = −20° (Fig. 4).

5.3 Preferred fault planes and stress axes

It is difficult to determine which of the nodal planes is the actual faulting plane for events with no observed surface manifestations. The large focal depth relative to the fault length (less than 1 km) makes the association with observed geological structures even more uncertain. The locations suggest that the earthquakes line up in a roughly northwesterly direction. However, this spatial pattern of the aftershocks is no conclusive discriminator as to the mechanisms (*cf.* Figs 2 and 4). Interpretation of data from recent seismic refraction profiles across the nearby Protogine zone shows an area of faulting dipping to the southwest (EUGENO-S working group 1988). The north–south striking planes of the Skövde earthquakes align well with the Protogine zone and may indicate movement along the zone.

For the Mariestad earthquake it is not possible to discriminate between the two nodal planes. The lake bottom bathymetry (Håkansson *et al.* 1978) shows several prevalent directions of lineaments.

The stress orientation indicated from the presently studied earthquakes is maximum compression in the approximate northwest direction. The P-axis plunge is, however, different for different events (*cf.* Fig. 4): the Skövde main shock, the largest aftershock and the Mariestad earthquake have an intermediate, shallow to intermediate and an almost vertical P-axis, respectively (Table 3; Fig. 4).

6 DYNAMIC SOURCE PARAMETERS

Dynamic source parameters were obtained using displacement spectra of Lg-waves recorded at various stations, analogue and digital, in Sweden, Norway and Finland. Only a few analogue stations could be used due to saturation of the instruments. Sampling rates of the digital stations were 40 Hz for the NORESS array (NRS), 50 Hz for the western Norway network, 60 Hz for the SKI network and 190 Hz for the field station VAB.

The seismic moment, M_0, is determined from the low-frequency spectral level, Ω_0, through the relationship (Street, Herrmann & Nuttli 1975; Herrmann & Kijko 1983)

$$M_0 = \begin{cases} 4\pi\rho\beta^3 R_0 (R/R_0)\Omega_0 & R < R_0, \quad (1a) \\ 4\pi\rho\beta^3 R_0 (R/R_0)^{1/2}\Omega_0 & R \geq R_0, \quad (1b) \end{cases}$$

where ρ is the density (2700 kg m^{-3}), β is the shear wave velocity (3.6 km s^{-1}), R_0 is the reference distance (100 km) and R is the epicentral distance. Ω_0 is obtained as the average of the low-frequency level measured by visual inspection of available spectra. The assigned value of R_0 reflects the distance where Lg-waves are believed to change from a predominantly body wave type (spherical spreading) at shorter distances to a mainly surface wave type (cylindrical spreading) at longer distances (Street *et al.* 1975).

The source radius, r, average dislocation, D_0, and stress drop, $\Delta\sigma$, were computed from the formulae (Aki 1966; Brune 1970, 1971)

$$r = 2.34\beta/(2\pi f_0), \quad (2)$$

$$D_0 = M_0/(\mu\pi r^2), \quad (3)$$

$$\Delta\sigma = M_0(7/16)(1/r^3), \quad (4)$$

where f_0 is corner frequency and μ is the shear modulus (3.3×10^{10} N m^{-1}). f_0 is computed as the average of corner frequencies measured by visual inspection of available spectra and adjusting for distance dependence (see Kim, Wahlström & Uski 1989). For more information on the applied methods, see Kim *et al.* (1989) and Arvidsson *et al.* (1991). An example of trace and ground amplitude spectra, depicting how Ω_0 and f_0 are determined (by visual inspection), is shown in Fig. 8.

Parameters for the three shocks studied are listed in Table 3. In terms of seismic moment, the Skövde main shock with 5.9×10^{14} N m is the largest Swedish earthquake during

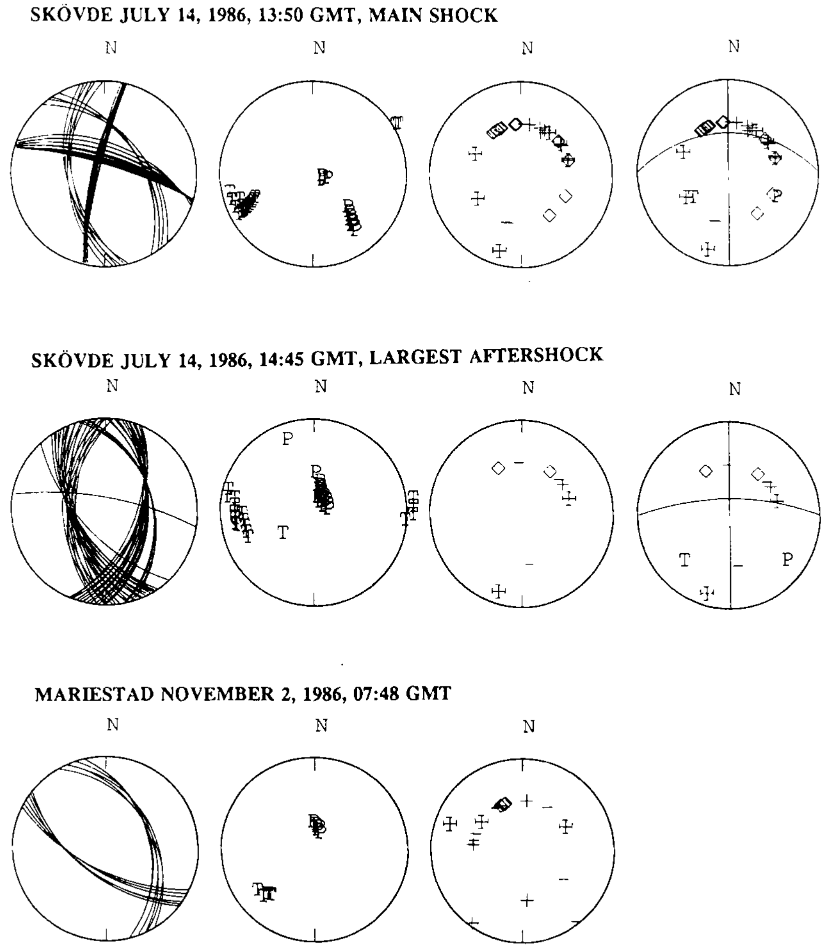


Figure 4. Lower hemisphere projections of the focal sphere for the three earthquakes for which focal mechanisms have been derived in this study: first column—distribution of possible nodal planes from FOCMEC solution; second column—corresponding maximum (P) and minimum (T) deviatoric stress axes; third column—input polarities, ✢ and ◇ are full-weighted compression and dilatation, respectively, and + and − are corresponding half-weighted polarities; fourth column—planes and stress axes obtained from synthetic seismogram modelling (Skövde earthquakes only). Resulting mechanism types are described in the text.

several decades. For comparison, the 1985 Kattegat earthquake, with magnitude M_L(UPP) = 4.6, had a calculated moment of 3.6×10^{14} N m (Arvidsson *et al.* 1991). The stress drop and dislocation of the Skövde main shock and the Mariestad earthquake are also unusually large for Baltic Shield events (*cf.* Kim *et al.* 1989).

Although reliable digital data are available for the Mariestad earthquake, significantly different corner frequencies are observed at different stations. Thus, the spectrum determined from the VAB signal shows a corner frequency of 10 Hz, whereas stations at an azimuth nearly 180° different from VAB show a corner frequency between 3.5 and 4 Hz (e.g. ODD). When the distance correction (from Kim *et al.* 1989) was taken into account, the difference was still about a factor of 2. The difference is not likely to be caused by uncertain seismograph characteristics, which are stable and known within this frequency range. However, for specific types of breaks, e.g., a unilaterally propagating rupture, the corner frequency can change considerably at sensors at the front and at the back of the rupture (e.g. see Bullen & Bolt 1985).

7 DISCUSSION

For rather high frequencies of up to at least 2 Hz and distances of up to at least 200 km, synthetic seismogram modelling shows satisfactory results stabilizing the mechanism determination. This is probably in part due to the relatively simple and homogeneous structure of the crust and upper mantle of the Baltic Shield. The success of short-distance synthetic modelling of Fennoscandian earthquakes has earlier been demonstrated by Kim *et al.* (1985) and Bungum *et al.* (1991).

The fault-plane solutions for the Skövde main shock and its largest aftershock indicate the possibility of movement along north–south striking faults related to the Protogine zone.

Several generating mechanisms have been proposed for earthquakes in Fennoscandia. The prevailing hypotheses are (a) ridge push from the North Atlantic Ridge suggested on the basis of the orientation of the P-axis of many microearthquakes (Slunga *et al.* 1984a; Bungum *et al.* 1991) and of statistical correlation (Skordas *et al.* 1991), (b)

Table 3. Source parameters.

Parameter	Skövde July 14, 1986 13:50 GMT	Skövde July 14, 1986 14:45 GMT	Mariestad November 2, 1986 07:48 GMT
Nodal plane 1: dip (°)	90°	90°	59°
strike (°)	180°	180°	122°
rake (°)	-45°	-20°	-111°
Nodal plane 2: dip (°)	45°	70°	37°
strike (°)	270°	270°	338°
rake (°)	-180°	-180°	-60°
P-axes: trend (°)	125°	133°	348°
plunge (°)	30°	14°	72°
T-axes: trend (°)	235°	227°	227°
plunge (°)	30°	14°	11°
Seismic moment (Nm)	5.9×10^{14}	2.1×10^{13}*)	2.3×10^{14}
Corner frequency (Hz)	3.0	3.7*)	5.7
Source radius (m)	450	360*)	240
Stress drop (MPa)	2.8	0.2*)	7.3
Average dislocation (mm)	28	2*)	38

*) From or derived from Kim et al. (1989).

post-glacial rebound suggested on the basis of faulting style (Bungum *et al.* 1991) and (c) lithospheric loading effects (Bungum *et al.* 1991).

The plate tectonic framework postulates horizontal principal compressive stress in a northwesterly direction with thrust faulting for earthquakes in the Baltic Shield. The mechanisms for our studied earthquakes are not clearly conclusive with this hypothesis. Also the other three earthquakes in Sweden with a magnitude above 4.0 during the time period of modern seismograph recording, Solberg 1983 (Kim *et al.* 1985), and Kattegat 1985 and 1986 (Arvidsson *et al.* 1991), have mechanisms difficult to explain from ridge push. It can be argued that anomalous behaviour of a few earthquakes can be the product of sliding guided by older faults (McKenzie 1969), but it is hard to accept that this is the case for all these large earthquakes with well-constrained mechanisms.

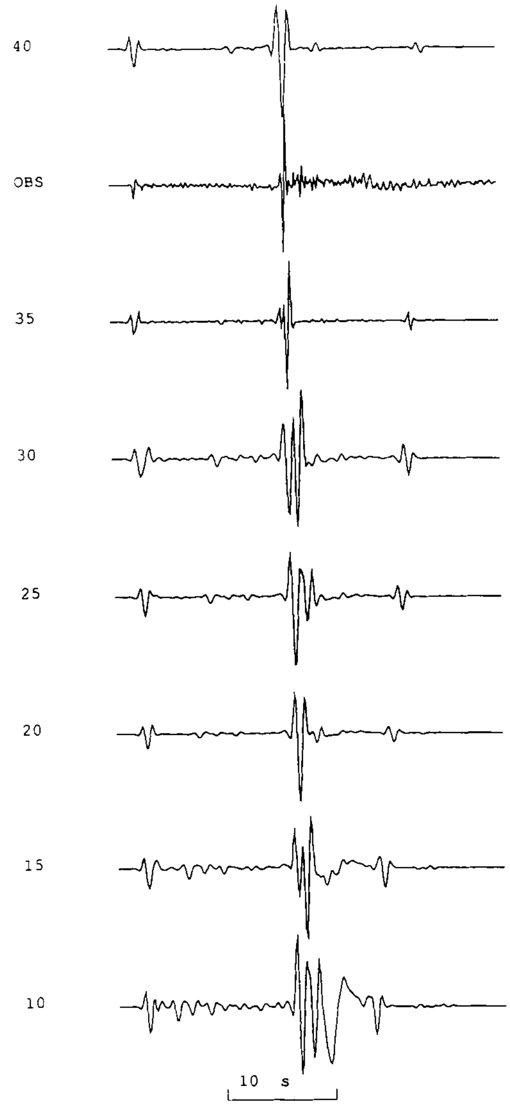


Figure 5. Observed (OBS) and synthetic seismograms for varying focal depths (10 to 40 km) at station GBD for the Skövde 1986 main shock. The best fit is observed for 35 km depth.

Table 4. Synthetic seismogram model.

Depth km	Wave velocities P km/s	S km/s	Density kg/m³	Q_P	Q_S
0-10	6.20	3.54	2700	1000	500
10-15	6.33	3.61	2900	1000	500
15-23	6.37	3.64	2900	1000	500
23-27	6.60	3.77	2900	1000	500
27-41	7.00	4.00	2900	1500	1000
41-	7.90	4.51	3300	1500	1000

Q_P: P-wave quality factor.

Q_S: S-wave quality factor.

A previously studied earthquake in the Lake Vänern region took place near Otterbäcken on 1981 February 13, magnitude 3.3, focal depth 9 km (Fig. 1). Although the mechanism derived by Kulhánek *et al.* (1983) is different from that by Slunga, Norrman & Glans (1984b), both show a predominantly horizontal component and a northwesterly trend of the *P*-axis, i.e., it is more easily associable with plate tectonic generated stresses propagated from the North Atlantic Ridge. Possibly the focal depth is a relevant parameter for the connection to plate boundary originated stresses—the 1983 Solberg earthquake (see above) with its steep *P*-axis also occurred in the deep crust.

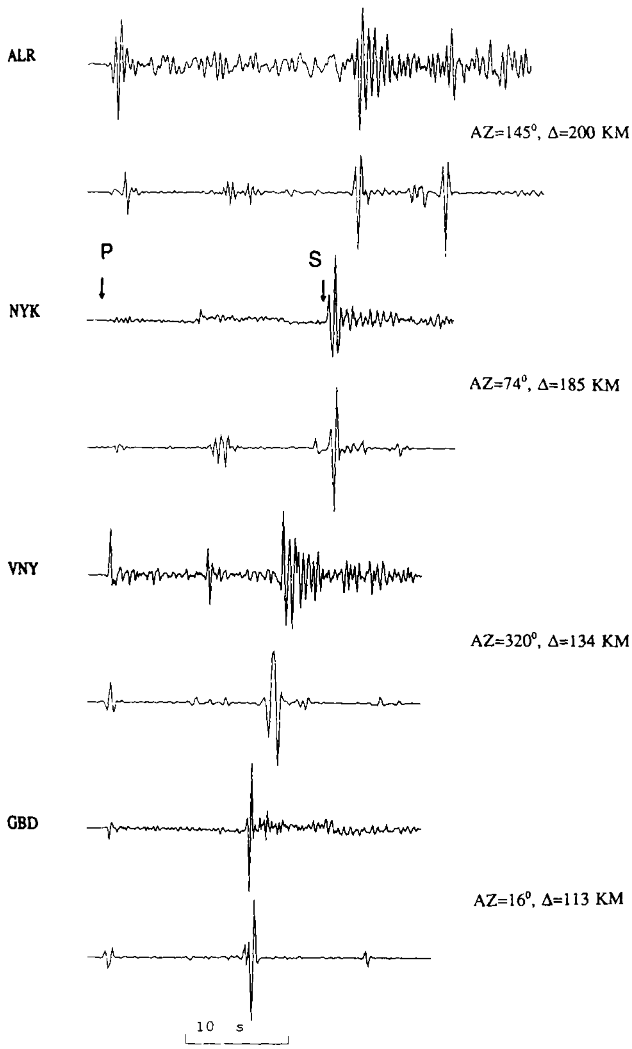


Figure 6. Synthetic (upper trace) and observed (lower trace) seismograms for the Skövde 1986 main shock at stations ALR, NYK, VNY and GBD. A focal depth of 35 km is assumed for the synthetics.

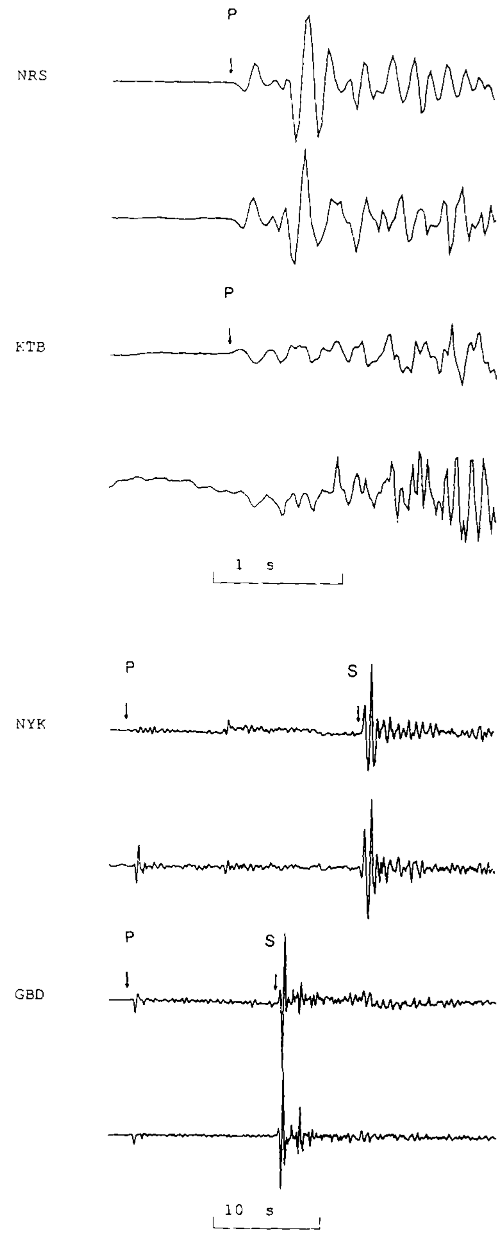


Figure 7. Comparison of observed seismograms at stations NRS, KTB, NYK and GBD for the Skövde main shock (upper trace) and largest aftershock (lower trace).

Although deviatoric stresses may to a certain extent deviate from the general stress field (McKenzie 1969), it is reasonable to assume that the relatively large earthquakes investigated in this study reflect the stress field in the Lake Vänern region. There are then similarities between this region and nearby southeastern Norway. On the basis of the seismicity pattern, the two regions have been suggested to constitute one seismotectonic unit (Husebye *et al.* 1978). The direction of the horizontal projection of the compressional axes agrees with a northwesterly direction of principal compressive stress as postulated from ridge push, but the faulting style, normal to strike–slip with an intermediate to steep *P*-axis, is hard to associate with a horizontal principal compressive stress. As a possible explanation to the normal fault mechanisms in southeastern Norway, the influence from post-glacial land uplift has been suggested in addition to the contribution from ridge push (Bungum *et al.* 1991). However, it is obvious that the mechanism(s) behind the earthquakes in the Baltic Shield is (are) more complex than straightforward stress propagation due to interplate activity at the North Atlantic Ridge.

The focal depths of the Skövde series and the Mariestad earthquake indicate that the events were located in the lower crust. The lower crust, beneath the brittle–ductile transition zone, is considered by many to be aseismic (Brace & Kohlstedt 1980; Meissner & Strehlau 1982; Chen & Molnar 1983). However, in old cratons earthquakes have been reported from deeper parts of the crust (Chen & Molnar 1983) and recently also some other lower crust earthquakes have occurred in Fennoscandia (Kim *et al.* 1985; Bungum *et al.* 1991). The focal depth of 40 km of the Solberg 1983 earthquake (Kim *et al.* 1985) has been suggested to correlate with the low Fennoscandian heat flow (Chen 1988). This seems reasonable considering the additional information presented here about other lower

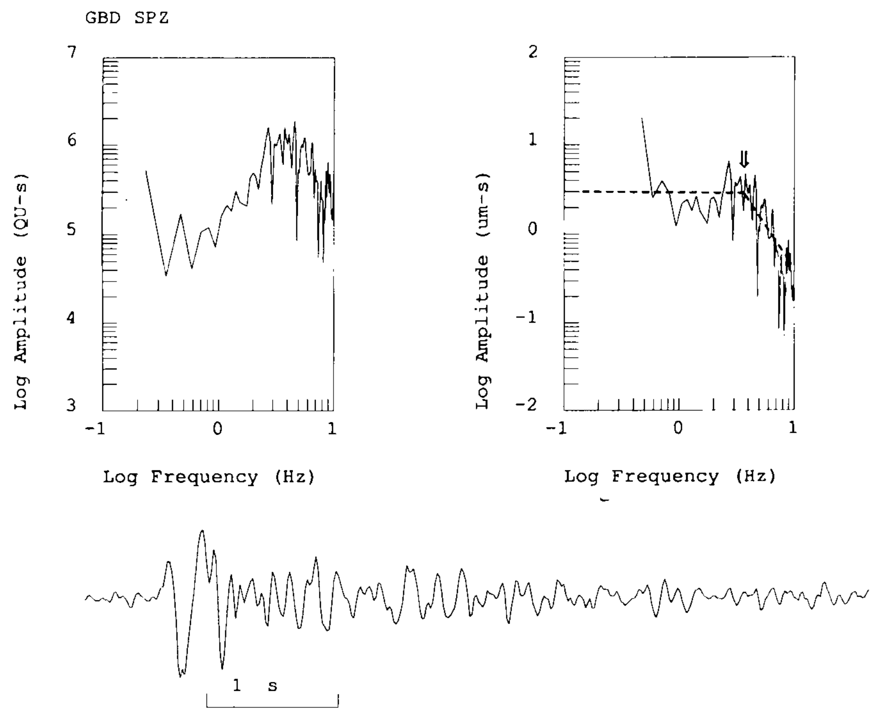


Figure 8. Trace amplitude spectrum (top left), ground displacement amplitude spectrum (top right) and section of the seismogram used for the spectral analysis (bottom), for the Skövde main shock recorded at station GBD. Linear approximations of the low- and high-frequency portions are indicated in the ground spectrum. The intersection of the two lines determines the corner frequency (arrow).

crust earthquakes in Fennoscandia. Perhaps the crust in Fennoscandia is brittle down to depths close to Moho.

Waveform modelling seems to be an excellent tool to obtain information about the focal depth, a parameter otherwise hard to determine.

ACKNOWLEDGMENTS

We are thankful to agencies which provided seismograms for this study. We are very grateful to R. B. Herrmann, who provided us with the source code for the synthetic seismogram modelling, and to H. Bungum and A. C. Johnston for valuable comments on the text. R. Wahlström was supported by the Swedish Natural Science Research Council, Contract G-GU 4102-303.

REFERENCES

Ahlin, S., 1983. Voluminous talus—a product of neotectonism?, *Geologiska Föreningens i Stockholm Förhandlingar (GFF)*, **105,** 16.

Ahlin, S., 1987. Phanerozoic faulting in the Västergötland basin area, SW Sweden, *Geologiska Föreningens i Stockholm Förhandlingar (GFF)*, **109,** 221–227.

Aki, K., 1966. Scaling law of seismic spectrum, *J. geophys. Res.*, **72,** 1217–1231.

Arvidsson, R., Gregersen, S., Kulhánek, O. & Wahlström, R., 1991. Recent Kattegat earthquakes—evidence of active intraplate tectonics in southern Scandinavia, *Phys. Earth planet. Inter.*, **67,** 275–287.

Båth, M., 1979. Earthquakes in Sweden 1979–1976, *Sveriges Geologiska Undersökning*, **C750,** 1–79.

Björck, S. & Digerfeldt, G., 1982. Late Weichselian shore displacements at Hunneberg, southern Sweden, indicating complex uplift, *Geologiska Föreningens i Stockholm Förhandlingar (GFF)*, **104,** 132–155.

Brace, W. F. & Kohlstedt, D. L., 1980. Limits on lithospheric stress imposed by laboratory experiments, *J. geophys. Res.*, **85,** 6248–6252.

Brune, J. N., 1970. Tectonic stress and the spectra of seismic shear waves from earthquakes, *J. geophys. Res.*, **75,** 4997–5509.

Brune, J. N., 1971. Correction to tectonic stress and the spectra of seismic shear waves from earthquakes, *J. geophys. Res.*, **76,** 5002.

Bullen, K. E. & Bolt, B. A., 1985. *An Introduction to Seismology,* Cambridge University Press, Cambridge, UK.

Bungum, H., Alsaker, A., Kvamme, L. B. & Hansen, R. A., 1991. Seismicity and seismotectonics of Norway and nearby continental shelf areas, *J. geophys. Res.*, **96,** 2249–2265.

Chen, W.-P., 1988. A brief update on the focal depths of intracontinental earthquakes and their correlations with heat flow and tectonic age, *Seism. Res. Lett.*, **59,** 263–272.

Chen, W.-P. & Molnar, P., 1983. Focal depths of intracontinental and intraplate earthquakes and their implications for the thermal and mechanical properties of the lithosphere, *J. geophys. Res.*, **88,** 4183–4214.

Collini, B., 1951. Beskrivning till kartbladet Gränna, *Sveriges Geologiska Undersökning,* **Aa 193,** 1–100.

EUGENO-S working group, 1988. Crustal structure and tectonic evolution of the transition between the Baltic Shield and the North German Caledonides (the EUGENO-S project), *Tectonophysics*, **150,** 253–348.

Green, C. M., Stuart, G. W., Lund, C.-E. & Roberts, R. G., 1988. P-wave crustal structure of the Lake Vänern area, Sweden: EUGENO-S Profile 6, *Tectonophysics*, **150,** 349–361.

Gorbatschev, R., 1980. The Precambrian development of southern Sweden, *Geologiska Föreningens i Stockholm Förhandlingar (GFF)*, **102,** 129–136.

Håkansson, L., Freden, C., Lindh, A., Ronge, B. & Stigh, J., 1978. *Vänerns Morfometri och Morfologi, Statens Naturvårdsverk,* Liber Förlag, Stockholm.

Herrmann, R. B. & Kijko, A., 1983. Modeling some empirical vertical Lg relations, *Bull. seism. Soc. Am.,* **73,** 157–171.

Herrmann, R. B. & Wang, C. Y., 1985. A comparison of synthetic seismograms, *Bull. seism. Soc. Am.,* **75,** 41–56.

Holmqvist, C. & Wahlström, R., 1987. The July-September 1986 Skövde aftershock sequence, *Technical Report,* Seismological Department, Uppsala University.

Husebye, E., Bungum, H., Fyen, J. & Gjöystdal, H., 1978. Earthquake activity in Fennoscandia between 1497 and 1975 and intraplate tectonics, *Norsk Geologisk tidsskrift,* **58,** 51–68.

Kim, W.-Y., Wahlström, R. & Uski, M., 1989. Regional spectral scaling relations of source parameters for earthquakes in the Baltic Shield, *Tectonophysics,* **166,** 151–161.

Kim, W.-Y., Kulhánek, O., van Eck, T. & Wahlström, R., 1985. The Solberg, Sweden, earthquake of September 29, 1983, *Report 1-85,* Seismological Department, Uppsala University.

Klein, R. W., 1978. Hypocenter location program: HYPOINVERSE, part 1: Users guide, *Open file report* 78-694, US Geological Survey, Menlo Park, CA.

Kulhánek, O. & Wahlström, R., 1985. Macroseismic observations in Sweden 1980–1983, *Sveriges Geologiska Undersökning,* **C776,** 1–13.

Kulhánek, O., van Eck, T., John, N., Meyer, K. & Wahlström, R., 1983. Spectra of the earthquake sequence February-March, 1981, in south-central Sweden, *Tectonophysics,* **93,** 337–350.

Lind, G., 1972. The gravity and geology of the Vättern area, southern Sweden, *Geologiska Föreningens Förhandlingar i Stockholm* (*GFF*), **94,** 245–257.

McKenzie, D., 1969. The relation between fault plane solutions and the direction of principal stresses, *Bull. seism. Soc. Am.,* **59,** 591–601.

Meissner, R. & Strehlau, J., 1982. Limits of lithospheric stresses in continental crusts and their relation to the depth-frequency distribution of shallow earthquakes, *Tectonics,* **1,** 73–89.

Skordas, E., Meyer, K., Olsson, R. & Kulhánek, O., 1991. Causality between interplate (North Atlantic) and intraplate (Fennoscandia) seismicities, *Tectonophysics,* **185,** 295–307.

Slunga, R. & Nordgren, N., 1987. Earthquake measurements in southern Sweden Oct. 1, Stockholm. 1986–Mar 31, 1987, *SKB Technical Report 87-27,* Swedish Nuclear Fuel and Waste Management.

Slunga, R., Norrman, P. & Glans, A.-C., 1984a. Baltic Shield seismicity, the result of a regional network, *Geophys. Res. Lett.,* **11,** 1247–1250.

Slunga, R., Norrman, P. & Glans, A.-C., 1984b. Seismicity of southern Sweden, *FOA Report C 20543-T1,* Swedish National Defence Research Institute, Stockholm.

Snoke, J. A., Munsey, J. W., Teague, A. G. & Bollinger, G. A., 1984. A program for focal mechanism determination by combined use of polarity and SV-P amplitude ratio data, *Earthquake Notes,* **55-3,** 15.

Stein, S., Cloething, S., Sleep, N. H. & Wortel, R., 1989. Passive margin earthquakes, stresses and reology, in *Earthquakes at North-Atlantic Passive Margins: Neotectonics and Postglacial Rebound,* pp. 231–259, eds Gregersen, S. & Basham, P., Kluwer, Dordrecht.

Street, R. L., Herrmann, R. B. & Nuttli, O. W., 1975. Spectral characteristics of the Lg wave generated by central United States earthquakes, *Geophys. J. R. astr. Soc.,* **41,** 51–63.

Sykes, L. R., 1978. Intraplate seismicity, reactivation of preexisting zones of weakness, alkaline magmatism, and other tectonism postdating continental fragmentation, *Rev. Geophys. Space Phys.,* **16,** 621–688.

Wahlström, R., 1987. Focal mechanisms of earthquakes in southern Quebec, southeastern Ontario, and northeastern New York with implications for regional seismotectonics and stress field characteristics, *Bull. seism. Soc. Am.,* **77,** 891–924.

Wahlström, R., 1988. Earthquake dynamics and seismotectonics in the Baltic Shield, *Geologiska Föreningens i Stockholm Förhandlingar* (*GFF*), **110,** 422–425.

Wahlström, R. & Ahjos, T., 1984. Magnitude determination of earthquakes in the Baltic Shield, *Annales Geophysicae,* **2,** 553–558.

Zeck, H. P. & Malling, S., 1976. A major global suture in the Precambrian basement of SW Sweden, *Tectonophysics,* **31,** T35–T40.