Seismodynamics of Sweden deduced from earthquake focal mechanisms

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SUMMARY
Focal mechanisms from 18 major Swedish earthquakes, $M_L(UPP) \geq 3$, reveal that:
1. several of the studied earthquakes have focal mechanisms with significant extensional stresses which fit well to the idea of postglacial rebound as an important stress generator,
2. in areas like south-western Sweden, the north-westerly trending compressional axes indicate that ridge push from the North Atlantic Ridge is also a considerable stress contributor;
3. some local tectonic features, like the Skälderviken depression in Kattegat, are momentous for seismotectonic interpretations. In addition to first $P$ polarities, full waveform modelling for frequencies up to 3 Hz and epicentral distances up to 225 km provides rather robust focal-mechanism determinations. Focal depths of seven events, ranging from 11 to 37 km, are well resolved within a few km by waveform modelling.

Key words: earthquakes, focal depth, seismotectonics, Sweden.

1 INTRODUCTION
Sweden is a part of Fennoscandia, an intraplate region, which has been subjected to deglaciation and subsequent land uplift during Quaternary time in a way similar to that of north-eastern Canada. It is typical for intraplate areas to have few and geographically scattered earthquakes, often with thrust-fault mechanisms indicating compressive acting stresses associated with global tectonic forces (e.g. see Sykes 1978). Focal mechanisms in south-eastern Canada, representing an old shield area, fit well into this pattern (e.g. see Wahlström 1987). However, deviations from this trend are seen in north-eastern Canada where focal mechanisms show normal-faulting style on land and thrust-faulting style beneath the sea (Stein et al. 1989). A possible explanation of the deviation from the general intraplate stress pattern in north-eastern Canada could be that besides the global plate tectonic forces, the postglacial rebound also acts as a significant stress contributor (Stein et al. 1979; Hasegawa & Adams 1990; Stein et al. 1989; Spada et al. 1991). When compared with non-glaciated areas along the Atlantic seaboard, the rate of earthquake occurrence appears to be higher along the glaciated margins (Stein et al. 1989). Results from recent modelling of lithospheric stresses indicate that except for the plate tectonic concept favouring e.g. ridge-push forces, also postglacial rebound, spreading stresses due to density contrasts in the lithosphere and sedimentation, should be considered in earthquake-generating processes in intraplate deglaciated regions (e.g. see Stein et al. 1989). This is also underlined by the fact that the interior of the recent glaciated areas of Greenland and Antarctica are almost void of earthquakes, which may probably be due to a stress-suppressing mechanism caused by the ice sheets. Upon removal of the ice, stresses are released (Johnston 1989). The same mechanism may explain the large postglacial faults observed in northern Sweden (Lagerbäck 1979, 1990; Muir-Wood 1989).

In Fennoscandia, during recent years, the ridge push has been emphasized as a major cause for earthquakes (Husebye et al. 1978; Slunga 1989, 1991; Gregersen, Korhonen & Husebye 1991; Skordas et al. 1991). Nevertheless, some recent works also reveal other plausible causes. For example, a study on Norwegian earthquakes suggests that for different regions, ridge push, postglacial rebound, density contrasts and sedimentation, play a crucial role in generating seismic activity (Bungum et al. 1991). The latter view is also supported by focal-mechanism styles obtained from the four largest Swedish earthquakes (Arvidsson, Wahlström & Kulhanek 1992).

The aim of the present paper is to study the earthquake-generating processes in Sweden and to some extent in the rest of Fennoscandia, by making use of available new focal mechanisms for major regional earthquakes [$M_L(UPP) \geq 3.0$]. In doing this, new focal mechanisms are derived for seven Swedish earthquakes and compiled together with 11 previously constructed mechanisms. From this set of 18 mechanisms we attempt to present a plausible seismotectonic interpretation.
2 SEISMOTECTONIC SETTINGS

Generally speaking, Fennoscandia consists of the Precambrian Baltic Shield in the central and eastern part and of the younger Caledonides in the west (Fig. 1). The Baltic Shield in its turn is divided up into three major domains, the Archean Domain in the north-east, the Fennoscandian Domain in the central and south-eastern parts, and the South-west Scandinavian Domain in the south-west (Gaal & Gorbatschev 1987). The continental margin, offshore Norway, is described as a passive margin (e.g. see Bungum et al. 1991) even though some authors have emphasized a possible coupling between earthquake occurrence offshore of western Norway and extensions of the Senja and Jan Mayen fracture zones (Husebye et al. 1978; Sykes 1978).

The largest known neotectonic movement in Fennoscandia is the postglacial rebound with a maximum coastline rise of approximately 280 m, at about 64°N (Mörner 1979). Several neotectonic faults in Sweden have been surveyed (Fig. 1). The most pronounced are the postglacial faults in Lapland, the northernmost part of the country (Lagerbäck 1979), and the Skålderviken depression, off-coast of southern Sweden just north of the Tornquist zone (Lykke-Andersen 1987; Lind & Lykke-Andersen 1990).

Compared with other parts of the world, Fennoscandian seismicity is low to moderate with most of the larger earthquakes occurring offshore of Norway (e.g. see Bungum et al. 1991). Only a few major events took place in Norway, Sweden, Finland and north-western USSR (Fig. 2). The three largest earthquakes in the region, all with magnitudes about 5.5 or larger, are: the 1759 event in Kattegat; the 1819 earthquake near Bodø in northern Norway; and the 1904 earthquake south of Oslo (e.g. see Muir-Wood & Woo 1987). It has been pointed out that the largest earthquakes with $M > 5$ occur in areas where the crust is thin (Kinck, Husebye & Lund 1991). As the record of instrumentally located earthquakes reveals, the Tornquist zone is almost aseismic in its eastern part (Arvidsson et al. 1991). However, west of Jutland, major earthquakes are occurring along the Tornquist zone (Gregersen et al. 1991). As far as the seismicity of Sweden is concerned, the quakes are rather scattered, although several areas of seismicity concentra-

![Figure 1. Map of Fennoscandia with the major geological boundaries and tectonic lineaments. A: Archean Domain; SF: Svecofennian Domain; SW: Southwestern Scandinavian Domain; C: Caledonides; TZ: Tornquist zone; SD: Skålderviken depression; M: Mylonite zone; PZ: Protogine zone; O: Oslo graben; SK: Skellefteå proterozoic subduction zone; P: Postglacial faults; LB: Ladoga–Bothnian Bay lineament.](https://academic.oup.com/gji/article-abstract/116/2/377/654315)
Seismodynamics of Sweden

Figure 2. Seismicity of Fennoscandia 1963–1989. Only epicentres for earthquakes (circles) with magnitude $M_L(\text{UPP}) \geq 3$ are plotted (99 per cent completeness level for Sweden and Finland, Arvidsson & Wahlström 1992). The size of the circles corresponds to the magnitude, $M_L$, within the range 3.0 to 5.2.

lations are obvious. Lagerbäck (1979) associates some of the Lapland seismicity with postglacial faults. This is further supported by the fact that in Lapland, two distinct alignments of epicentres appear on the seismicity map for earthquakes with magnitude three and larger (Fig. 2). These alignments coincide with the two largest postglacial faults in this part of the country. Further south in the Gulf of Bothnia, a clustering of earthquakes coincides with the site of a major suture zone, the Skellefteå proterozoic suture zone (Fig. 1), which is observed both in magnetotelluric data (Rasmussen, Roberts & Pedersen 1987) and seismic profiling where a clear step in the Moho depth is defined (BABEL Working Group 1990). Earthquakes in this region could be explained by the local basement inhomogeneity concept (Hinze et al. 1988) which assumes concentration of stresses at inhomogeneities in the basement. One seismic zone in south-central Sweden appears to be interconnected with the Oslo-graben area in south-eastern Norway (Husebye et al. 1978; Arvidsson et al. 1992). An earlier study shows that a high density of earthquakes coincides with a high density of large-scale lineaments (order of 10 km and greater) whereas where lineaments were absent, earthquakes were also absent (Arvidsson, Wahlström & Kulhanek 1987). A pocket of seismicity, containing two of the four largest Swedish events in the last 40 years, is associated with the Skålderviken depression, offshore of southern Sweden. During the time of modern instrumentation, say, since the late 1950s, a certain temporal variation in seismic energy release has been observed in Sweden. Note, e.g. that the four largest earthquakes occurred within a three-year period from 1983 to 1986 (Arvidsson et al. 1992). At least in one case, major Swedish earthquakes could be associated with a known neotectonic fault. Observe that the three earthquakes occurring in Kattegat in 1985, 1986 and 1990, correlate well with the neotectonic faulting of the Skålderviken depression (Arvidsson et al. 1991).

In conclusion, the seismicity in Sweden appears to be related to older deformation zones and inhomogeneities in the basement. One peculiarity of Swedish earthquakes is that the deduced focal depth places some of the events in the lower crust, sometimes close to Moho (Kim et al. 1985; Arvidsson et al. 1992). A similar phenomenon is observed also in Norway (Bungum et al. 1991).

3 HYPOCENTRE AND FOCAL-MECHANISM DETERMINATION

The present study makes use of data from 18 major ($M_L(\text{UPP}) \geq 3.0$) earthquakes in Sweden. Basic source
parameters of the events are listed in Table 1. For seven shocks (Nos 7, 11, 13, 14, 15, 16 and 17, see Table 1) we present new results, whereas for the remainder we compile the corresponding parameters taken from other sources.

An iterative approach has been applied to derive the epicentre, focal depth and focal mechanism. A hypocentre was first determined by means of available arrival times. Thereafter, depth phases on the observed seismograms were identified with the aid of synthetic seismogram modelling. Then the depth was quantified by computing the rms error between the computed and observed traveltimes of the depth phases. The procedure is further described

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<th>Location Lon.</th>
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<th>Focal P</th>
<th>Focal T</th>
<th>Focal t</th>
<th>Focal p</th>
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P: compressional axis; T: tensional axis; t: trend; p: plunge; all in degrees clockwise from north.

help of synthetic modelling. Derived location parameters are summarized in Table 1.

3.2 P-polarity mechanisms

To determine the focal mechanisms, we first utilized the \( P \) polarities and \( SV/P \) ratios as input to the FOCMEC program (Snoke et al. 1984). The current version of the program allows the user to weight the \( P \)-polarity data in two quality classes (Wahlström 1987). To ensure homogeneity, available seismograms from stations in Fennoscandia, within reasonable epicentral distances, were checked by the present authors. Records within the cross-over distances of 130–170 km were discarded due to the ambiguity of \( Pg \) and \( Pn \) readings. \( SV/P \) ratios were used only for distances less than 130 km. For individual events, 10 to 14 polarities and up to three \( SV/P \) ratios were used.

In three of the cases (Nos 11, 13, 16, Table 1) the solutions derived from \( P \)-polarity readings showed only one family of mechanisms. For event 14, two families of solutions were obtained, though the orientation of stresses in the horizontal plane was quite stable for the different solutions. Solutions for events 7, 15 and 17, had to be further constrained by making use of corresponding \( SV/P \) ratios (Table 1; Fig. 3). We shall continue our discussion on focal mechanisms in the following paragraph on synthetic modelling.

![Figure 3](https://academic.oup.com/gji/article-abstract/116/2/377/654315)
3.3 Seismogram modelling

Methods for calculating seismograms at local regional distances can be divided into three main classes: The generalized ray method (e.g. see Wallace & Helmberger 1982), the reflectivity method (Kind 1979a; Kennett, 1980) and the wavenumber integration method (Wang & Herrmann 1980; Bouchon 1981). So far, synthetic waveform modelling has not been applied at local and regional distances in such a number as at teleseismic distances. From long-period data recorded at regional distances several earthquake focal mechanisms have been successfully constrained and more recently, with the use of broad-band seismometers, the frequency band has been extended from low frequencies to frequencies up to 3 or 4 Hz (Fan & Wallace 1991; Zhao & Helmberger 1991). Works using only short-period data have been rather few but some attempts with promising results have also been made in the Scandinavian region (Kim 1987; Kim et al. 1988; Bungum et al. 1991; Arvidsson et al. 1992). The relative scarcity of studies using short-period seismometers is due to several factors. Methods for calculating full waveform seismograms at short distances were not fully developed before the early 1980s. Another factor is that the calculation of high-frequency seismograms at local and regional distances are comparatively computer intensive and high-speed machines have made the required computations easier. Further, many major earthquakes produce saturated seismograms which are hardly (in this context) of any use. However, there is a special class of data, like that considered in this work, which needs special attention. We mean high-frequency seismograms, namely from small earthquakes in areas of low seismicity where these events provide the only key to determine faulting parameters. These earthquakes are usually too small to provide enough information (above the level of microseisms) at long periods but are well recorded in the relatively high-frequency band.

In the present work a forward modelling procedure has been applied and synthetic waveforms were constructed by making use of a propagator matrix-wavenumber integration method (Wang & Herrmann 1980; Herrmann & Wang 1985), which allows for the construction of the full waveform at high frequencies and short epicentral distances. The source–time function was approximated by a triangular pulse of 0.2 s duration. All seismograms, both synthetic and observed, were low-pass filtered using cut-off frequencies between 2 and 3 Hz, to eliminate the high-frequency background noise. Varying cut-off frequencies were applied in order to ensure a high signal-to-noise ratio which varied both with respect to the size of the studied event and with the background microseismic noise level. Appropriate instrument responses were added to the synthetic seismograms through frequency-domain multiplication.

Basically, forward modelling of waveforms at local and regional distances is a two-fold task with respect to the crustal structure and the seismic source. The structural problem is often solved by a priori information from, e.g. seismic profile measurements. When only long-period $P_m$ waves are used, a simple one-layered crust is sufficient to describe the generated seismic waves for different tectonic regimes (e.g. see Wallace & Helmberger 1982; Fan & Wallace 1991). The difficulty of modelling increases with increasing frequency so that the resolution of structural complexities become more important for higher frequencies. Up to a few Hz, which is the frequency band considered in this study, the waveforms can be modelled with plane-layered structures at local and regional distances (Fan & Wallace 1991; Zhao & Helmberger 1991). One trade-off with different models is that the focal depth may vary with crustal structure within say two-to-four km (e.g. see Fan & Wallace 1991). However, this variation still provides a much better focal-depth constraint when compared with that derived from arrival-time locations and readings from sparsely distributed seismograph stations, which is the case in Fennoscandia.

The starting structural model (Earth model 1, Table 2) is that used in Arvidsson et al. (1992) which is based on a model from the EUGENO-S experiment (Green et al. 1988). This model, which has a Moho depth of about 41 km, is characterized by a mid-crustal reflector at 23 km, similar to that observed in data from the Fennolora long-range seismic profile (Galsou, Mueller & Munch 1984; Galsou & Mueller 1986; Guggisberg 1986). The mid-crustal reflector is here interpreted as the Conrad discontinuity. However, to obtain a better fit between observed and synthetic data, in some cases, this model was altered and smoothed (Earth model 2, Table 2). At short distances, model changes imposed usually only small differences, whereas at larger distances, effects of model selection are more significant. One potential problem, in the present synthetic modelling, is the variation of the crustal thickness from about 30 to 50 km beneath Sweden which has been found both from the Fennolora profile (Galsou et al. 1984; Galsou & Mueller 1986; Guggisberg 1986) and from inversions of teleseismic waves (Bungum, Pirhonen & Husebye 1980). Beneath the epicentral areas studied here, the crust varies, according to seismic investigations (Knick et al. 1991), between 38 and 50 km and due to teleseismic inversions between 40 and 46 km (Bungum et al. 1980). Thus, in order to examine the impact of the varying Moho depth, modelling was performed not only with the original structural models (Table 2) but also with models where the Moho depth was adjusted so that it approximated the average Moho depth between considered station-earthquake pairs. Results obtained from the two models with varying Moho depth did not alter significantly. This may be due to the fact that, as observed in earlier works (Helmberger & Johnson 1977), the seismogram wave train is dominated by upgoing waves and thus is more sensitive to the shallow structure. Another explanation is that a higher velocity in the lower layers may compensate for the additional crustal thickness. However, in cases where lower crustal reflections such as $PmP$ and $SmS$ are observed at short distances, as in Fig. 4(a), this is not true since the crustal thickness versus the recorded distance is rather large.

Several previous studies show that the focal depth of earthquakes recorded at regional distances can be determined by identifying depth-dependent phases in true records with the aid of synthetic seismograms computed for different focal depths (Kind 1979b; Kim et al. 1985, 1988; Zhao & Helmberger 1991; Arvidsson et al. 1992). By careful examination of seismograms from the events studied here, we found depth phases such as $P^*P$, $Pn$, $sPn$ and $S$ to $P$ converted (and critically reflected) head wave at the free
Table 2. Earth model 1.

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<td>41-</td>
<td>7.90</td>
<td>4.51</td>
<td>3300</td>
</tr>
</tbody>
</table>

Surface, for all events, even though not at all stations. The Moho reflected phases \( PmP \) and \( SmS \) were also used for one event (Event 15; Table 1). Initially, the focal depth was determined by aligning the observed seismograms with the synthetics computed for the whole range of the crustal thickness (Fig. 4). By this comparison, an initial check on identification of depth phases observed as well as on initial focal depth was made. Thereafter, the focal depth was quantified by computing the root-mean-square error, \( \text{RMS}_r \), of the observed and theoretical traveltime difference between the considered depth phase and the first \( P \). The error is defined as:

\[
\text{RMS}_r = \left[ \frac{1}{N} \sum (\text{obs}_i - \text{comp}_i)^2 \right]^{1/2},
\]

where \( \text{obs}_i \) and \( \text{comp}_i \) is the difference in traveltime between the first \( P \) and the depth phase in the observed seismogram and the theoretically computed difference, respectively. \( N \) is number of observations, and the summation is performed over all available stations. The

![Figure 4](https://academic.oup.com/gji/article-abstract/116/2/377/654315)

(a) Synthetic seismograms (event 15, Table 1) demonstrating the effect of variation of depth from surface to Moho on calculated synthetic seismograms. 33, 41 and 50 indicate the crustal thickness and OBS the observed seismogram made at station VNY at epicentral distance of 70 km (vertical component). (b) Depth determination for event 15 (Table 1), recorded at station VNY (vertical component), by using the technique of synthetic-seismogram modelling. Numbers to the left indicate focal depth used in the calculations and OBS indicates the true, i.e. observed, seismogram. Dashed lines show arrivals of phases in the observed trace and in corresponding synthetics. Identified phases are labelled according to conventional nomenclature (see e.g., Kulhanek 1990).
RMS$_T$ was computed for theoretical values of comp, covering the whole crust thickness. For all studied events, clear minima defined the 'true' depth (Fig. 5). Even with records from only one or two stations, the focal depth can be reasonably well estimated. As already mentioned, arrival-time readings provided stable results in only two cases (events 11 and 16; Table 1) which agree to within 2 km near the depth determined through RMS$_T$. Since we determine the depth from the time difference between the first P-type wave arrival and the considered depth phase, the impact on focal depth of possible location error for different stations, here estimated to be less than 5 km, is rather small, say about 1 km. Indeed, refracted phases such as Pn and Sn are indifferent to location errors. In this context, it is interesting to note that some of the considered events are located in the lower crust. This is in good agreement with earlier results which indicate the existence of a seismogenic lower crust beneath Fennoscandia (Kim et al. 1985; Arvidsson et al. 1992; Bungum et al. 1991).

Events 7, 11, 15, 16 and 17 (Table 1) were used for seismogram modelling. For events 13 and 14 (Table 1), the small amount of data precludes the use of synthetics for well-constrained focal-mechanism determinations. The best-fit mechanism was derived by using the 'average' P-polarity focal mechanism as a starting trial. This solution was then perturbed by varying one of the parameters strike, dip and rake, while keeping the remaining two fixed (Fig. 6). The fit between the observed and synthetic seismograms was based on several criteria. A reasonable correspondence between the relative amplitudes of the major arrivals such as Pg, Sg, Pn, Sn, pP^*P, PmP, SmS, Pn, sPn, a minimum phase shift between the observed and synthetic seismograms for direct P and S waves and correspondence in polarities. It was not possible to model the coda waves in full detail, but as follows from our experience, a rough fit of the envelope and of main arrivals were within reach. The final mechanism from the modelling procedure should not deviate significantly from observed P polarities used in the P-polarity mechanism. An acceptably good correspondence was achieved between the observed and synthetic seismograms for events 7, 11, 16 and 17. In the example shown in Fig. 7, the fit satisfies the above criteria, except perhaps for the radial component made at UPP. The deviation between the observed and synthetic seismograms can be due to a low signal-to-noise ratio of the true seismogram and/or to structural complexities. The slight misfit in arrival times for different phases reflects most likely the error in location. The best fit is generally found for records from epicentral distances of up to, say, 120 km, but reasonable results were obtained for distances up to 225 km. Event 15 (Table 1) showed rather poor fit except for one station, despite the short epicentral distances used. The reasons are probably several. From the polarity mechanism, it is observed that all stations applied in the modelling procedure of this particular event were close to nodal planes and hence even small inaccuracies in the mechanism may be crucial. Further, available seismograms indicate more complex structure than that assumed in this study (Earth models 1 and 2, Table 2). Despite these pitfalls, it seems that the focal depth of event 15 is quite well determined to be about 20 km and that the modelled seismogram (Fig. 4b) to some extent supports the FOCMEC mechanism.

When modelling ground motions recorded at larger distances, one initial problem was that some later phases (PmP, SmS, etc.) appeared systematically stronger on synthetic seismograms than on corresponding true records. A similar phenomenon was also observed in a previous study considering several events in south-central Sweden (Arvidsson et al. 1992). In this context, it is worthwhile to mention that by imposing a lower crust velocity gradient in the crustal model, which may also be physically more plausible, significant changes in the calculated wave trains are obtained (Cormier, Mandal & Harvey 1991). By introducing a velocity gradient in a part of the lower crust model (Earth model 2, Table 1), a satisfactory fit could also be obtained for some records from stations at larger distances (events 7; Fig. 7). Likewise, relative amplitudes of the reflected phases reduced to corresponding proportions. As follows from Fig. 3, for events 7, 11, 16 and 17 (Table 1) the achieved best fits of the synthetics are associated with mechanisms that are quite close to the first-polarity mechanisms, thus verifying the FOCMEC results. For event...
Figure 5. Depth determination. The ordinate shows the RMS error of theoretical and observed difference between depth phases and first P. The abscissa gives the focal depth in km.
14, the polarity mechanisms were rather ambiguous. Even though we had for this event records from only two stations (one three component) the modelling results indicate that the strike-slip style is to be preferred before the normal faulting.

The fault planes deduced in this work are shown in Fig. 8. The final mechanisms are chosen either as the synthetic mechanism whenever available (events 7, 11, 16, 17), or the mechanism with the average strike from the family of solutions given by FOCMEC (events 13, 14, 15).

3.4 Grouping of focal mechanisms

We tried to assemble the seven mechanisms derived in the present work together with 11 additional mechanisms available for major Swedish earthquakes with magnitudes above 3 in one map (Fig. 8; Table 1). At the first glance, it seems from Fig. 8, that the mechanism styles of the 18 events studied here are quite different. This is in line with observations from in situ measurements also indicating a rather scattered picture of observed stresses (Stephansson et al. 1987; Clauss, Marquardt & Fuchs 1989). However, after a more thorough examination, we feel that some categorization with respect to the geographical distribution, faulting styles and orientation of stress axes, can be made even with the limited number of mechanisms to hand. At least four seismic zones become more distinct:

1. Kattegat zone offcoast of southern Sweden, with mainly strike-slip faulting and north–south directed compressional axes (events 6, 8, 18).
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Pn pPn sPn Sn sSn

AZ = 167°, Δ=224 KM

UPP Z

UPP T

UPP R

HRN Z

Pn pPn Sn

AZ = 187°, Δ=177 KM

KTB Z

P S

AZ = 200°, Δ=90 KM

10. s

Figure 7. Synthetic seismograms of event 7 (Table 1) for stations UPP (3 component), HRN (Z) and KTB (Z). Observed (upper traces) and synthetic seismograms (lower traces) are displayed. Epicentral distances, Δ, and source-to-station azimuths, AZ, are also given for each station.

(2) South-central Sweden zone, showing mixed faulting styles with the largest and best constrained events having strike-slip to normal faulting. The compressional axes are trending north-westerly (events 4, 9, 10, 11, 12, 15).

(3) Eastern Sweden-Gulf of Bothnia zone, displaying a strong component of normal faulting (events 3, 5, 7, 14, 17). Directions of principal stresses are rotated with respect to south-western Sweden (Fig. 9). Most of the landward earthquakes here lie within areas which were under the sea-level directly after the deglaciation some 8000 yr BP.

(4) Northern Lapland zone, with mixed types of mechanisms both with compressional as well as extensional characteristics (events 1, 2, 13, 16). The deviatoric stresses are rotated with respect to Gulf of Bothnia earthquakes. The directions of compressive axes range from west-south-west to north-west with a dominant east-west trend (Fig. 9).
Figure 8. Compilation of focal mechanisms for 18 major Swedish earthquakes from the period 1967–1990. Events 7, 11, 13, 14, 15, 16 and 17, are those derived in the present study. The remaining mechanisms are taken from other sources (see Table 1). Beach-ball illustrations are used to indicate compressional (shaded) and extensional (white) quadrants. The black and white dots exhibit the position of compressional and extensional axes, respectively. Lower hemisphere projections are used. Numbers indicate the event number used in Table 1.
4 DISCUSSION

4.1 Seismodynamics of Sweden

Seismodynamics of the Kattegat zone are anomalous with respect to the hypothesis of a ridge push along the North Atlantic Ridge. In this context, we also wish to emphasize that the distance from the North Atlantic Ridge, which is often solely discussed in the plate-tectonic concept, is about the same as the distance to southern Italy, i.e. to the Mediterranean collision zone (Gregersen 1992; Muller et al. 1992). Focal mechanisms of the events from the Skälderviken depression (events 6, 8 and 18, Table 1, Fig. 8) correlate well with neotectonic faulting at the north-eastern flank of the depression. The north-south oriented compressional axes (Fig. 9) and the extensional features of the depression contradict the hypothesis of ridge push from the North Atlantic Ridge. The cause of this depression can be sedimentation in Kattegat. However, at present we cannot disregard spreading stresses due to density contrasts (edge of the Baltic Shield) or other factors discussed above (Arvidsson et al. 1991).

South-central Sweden and south-eastern Norway have recently been treated in Arvidsson et al. (1992) and in Bungum et al. (1991), respectively. On the basis of seismicity patterns (Husebye et al. 1978; Talbot & Slunga 1989) and focal-mechanism solutions (Arvidsson et al. 1992) it is reasonable to assume that the two areas constitute a continuous seismotectonic province. There are indications, such as the direction of compression or the significant component of normal faulting, that several stress generators, e.g. ridge push and postglacial uplift, are acting in the area.

In the eastern Sweden–Gulf of Bothnia zone, earthquake mechanisms deviate more from the ridge-push concept when compared e.g. with south-western Sweden. In spite of some scatter, the direction of compression has rotated from a north-westerly trend to north-south (Fig. 9). The significant extensional component with mechanisms ranging from strike-slip to normal faulting close to the centre of uplift, supports postglacial rebound stresses. One additional factor which increases the differential stress due to the rebound is the unloading of water due to the decreasing water depth caused by the postglacial rebound (Anderson 1980). The concentration of earthquakes along the Swedish coast of the Gulf of Bothnia may be associated (Stephansson 1979) with stress release along the dominant zone of weakness running along the Swedish shore line in the Gulf of Bothnia. Some of the release here may be ascribed to sedimentation since a number of near-surface events have been identified from the appearance of rather distinct short-period $R_g$ waves on
corresponding seismograms (Wahlström 1980). Modelling indicates that sedimentation may produce significant stresses (Stein et al. 1989). However, in the Barents Sea a low seismicity is observed despite a high sedimentary loading rate and a close proximity to the North Atlantic Ridge (Husebye et al. 1975).

The few and rather varying mechanisms available for the Lapland region make an interpretation difficult. It is likely that both the postglacial uplift and ridge-push mechanisms are acting in the area as earthquake generators. It is interesting to note that events 1, 2, 13 and 16 all are located on, or in the vicinity, of distinct postglacial faults (Fig. 1). Events 2, 13 and 16, with their NE to NS striking planes could possibly be associated with the NNE strikes of the present faults.

4.2 Implications for Fennoscandia

For Fennoscandia and neighbouring regions, a certain large-scale seismotectonic pattern is also apparent. First, the extensional axis appears to rotate around the centre of uplift. It is north-easterly–south-westerly oriented in south-central Sweden while in Kattegat and the Gulf of Bothnia it has rotated and trends more or less east–west. Further north, in Lapland, it turns to north–south and north-west–south-east. The compressional axis also changes between these regions. It is trending north-west–south-east in southern Sweden, north–south in the Gulf of Bothnia and west-south-west–east-north-east to north-west–south-east in Lapland (Fig. 9). Secondly, in the Gulf of Bothnia, as well as in south-central Sweden and southern Norway, there is a significant component of extension. From Fig. 9 it follows that principal stresses close to the centre of the rebound are, in general, oriented along the gradient of the postglacial uplift. Focal mechanisms display thrust faulting at the outer rim of the uplift, in both the Kola Peninsula (Assinovskaya 1986) and the North Sea (Bungum et al. 1991). Directions of compressional axes are north-east–south-west in the Kola peninsula, whereas in the Norwegian Sea they are east–west to north-west–south-east. Lastly, as follows from the present study and from Bungum et al. (1991), in Norway and the western part of Sweden, the directions of compressional axes are, in general, north-west–south-east.

The distribution of stresses described above agrees well with the modelled behaviour of stresses in a rebound area where extension is dominating in the centre of the uplift and compression at the brim and where principal stresses are oriented along the gradient of uplift (e.g. see Stein et al. 1989; Spada et al. 1991). Skordas & Kulhanek (1992) observed a clear correlation between the change of b values and the rate of postglacial rebound. It also follows from their work that the postglacial rebound should be considered as one of the major stress generators in the region. It is worth mentioning that in the Baffin bay area, which has a similar tectonic development to the Baltic Shield, a similar pattern of focal mechanisms was also observed (Stein et al. 1989; Hasegawa & Adams 1990). On the other hand, the rather homogeneous direction of compressional axis in the western part of Fennoscandia agrees well with the direction postulated from the ridge-push concept (e.g. Slunga 1989, 1991; Skordas et al. 1991; Bungum et al. 1991).

5 CONCLUSION

The results of the present study can be summarized as follows:

1. Focal mechanisms of several major Swedish earthquakes show a significant component of extension which contradicts the hypothesis of a ridge-push generating concept. It is likely that the postglacial rebound also represents a significant earthquake generator in the area.

2. In some parts of Fennoscandia, in the south-west in particular, the NW-trending compressional axes indicate a significant stress contribution from the ridge push.

3. Local effects, such as the Skålderviken depression, are of importance for faulting styles of individual earthquakes in the area.

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