7 GEOTHERMAL OBSERVATIONS

For the purposes of the investigation of Swedish seismotectonics, observations of crustal temperature are relevant, chiefly for predicting the thermal state of the crust and the control that this may have over the mechanical (and hence seismogenic) properties of the crystalline rocks at depth. The chief purpose of this section therefore is to review the current state of knowledge of the geotherm beneath Sweden, and its regional variations.

Direct observations of the temperature gradient measured in boreholes can also be employed to show the existence of present-day or past perturbations in the geotherm of various origins, including the flow of fluids into or out of the upper crust.

7.1 BACKGROUND

The earliest studies on the temperature gradient were undertaken in mines and, as a result of the significant thermal anomalies found associated with the highly conductive metalliferous ore bodies, this procedure was extended in the hope that it could become a prospecting tool within mineral exploration boreholes (Parasnis, 1975; 1982). In the 1970s downhole thermal measurement was carried out within boreholes drilled in the search for hydrocarbons in the Phanerozoic sediments around Skane. Geothermal studies in the search for suitable hot-dry rock geothermal projects, also targeted a number of granites, in particular along the western coast of Sweden near Gotheborg (Landstrom et al, 1980; Eliasson et al, 1988). Through the 1980s a considerable additional body of data has been collected as part of research undertaken in the pursuit of a suitable radioactive waste repository. In the deep borehole drilled at Gravberg, Dalarna, in the search for Mantle derived methane, for the first time in Sweden, the geotherm was followed for more than 5km depth (Malmquist, 1987).

A map of near-surface heatflows for the whole of Fennoscandia is provided by Cermak (1990). The results of the European Geotraverse project that spans almost the whole of Sweden, include attempts to extrapolate from the near-surface thermal observations, to model the geothermal gradient of the lower crust and upper mantle (Balling, 1990; Pasquale et al, 1990, 1991).

7.2 OBSERVATIONS

The original observations on which all geothermal gradients are constructed are obtained with the use of highly sensitive thermistor probes or microcircuits. Observations are almost all collected from within small diameter boreholes too narrow to permit the convective flow of the water-column. It is important that the water in the borehole has equilibrated to the temperature of the wallrock, requiring boreholes to be left undisturbed for a period, ideally of at least a year, prior to observation. An absolute accuracy of 0.1 °C with a resolution of 0.01 °C was quoted by Kukkonen (1988) for observations from Finland.

Temperature measurements have to be corrected for the thermal conductivity of the rock in which they are measured to derive heatflows. The available set of Fennoscandian borehole heatflow observations (from Cermak, 1990) are illustrated in Figure 7-1.

7.3 PERTURBATIONS TO THE GEOTHERM

The geotherm represents the flow of heat from heat-sources in the crust and mantle, to the Earth's surface. Over very long time-periods and in the absence of any changes in either heat-input or output, this gradient will equilibrate. However while the principal heat-sources associated with radioactivity show very long-term thermal decay times, the heat-sink of the atmosphere and hydrosphere at the Earth's surface shows considerable variation in temperature. While annual temperature cycles extend to only the top few metres, long-term changes reflecting climatic variation, most importantly that associated with the last Ice Age, can extend to many tens or even hundreds of metres. As the period of the Ice Age lasted longer than the period since deglaciation, the Holocene warming induces a shallow temperature

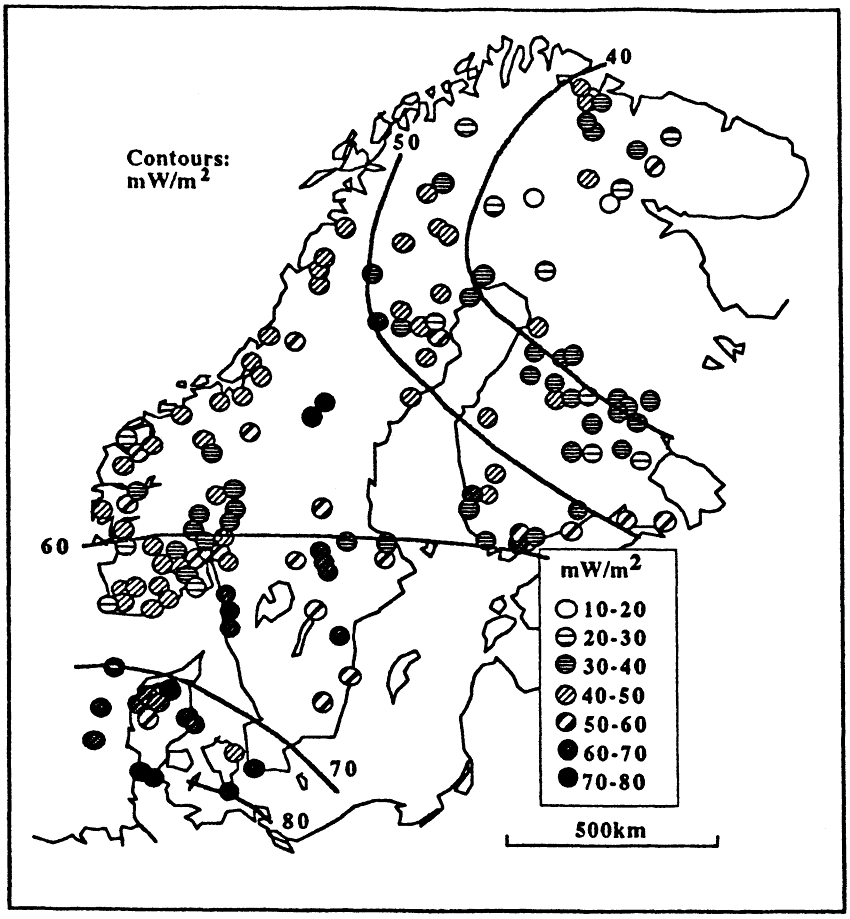


Figure 7-1: Surface heatflows in Fennoscandia.

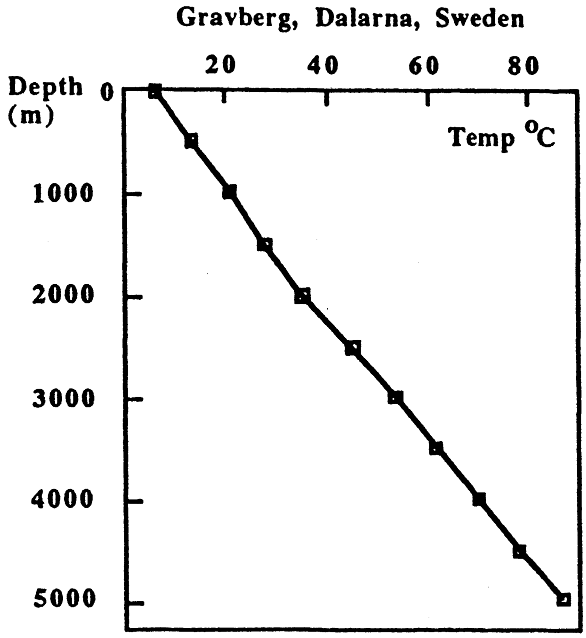


Figure 7-2: Gravberg borehole geotherm.

gradient down to some depth (typically of a few hundred metres) below which the steeper gradient is that of the Ice Age. As a generality, every perturbation in surface temperature has a chance to be recorded if it endured for a period longer than any subsequent variation. A number of graphical methods have been employed to compensate for the effect of the lower temperatures of the last Ice Age on the near surface temperature gradient (see for example Balling, 1990).

The geothermal gradient becomes refracted through materials of differing conductivity. Variations in conductivity provide some significant contrasts in the relationship between heatflows and geothermal gradients for the crystalline rocks typical of much of the Baltic Shield, and become of particular importance when boreholes pass through sedimentary formations, as exist in the Skane region.

A further important source of perturbations to the conductive temperature gradient comes from fluid circulation. Where there is a vertical component in groundwater flow, the conductive temperature gradient will be perturbed. The magnitude of this perturbation will be determined by the vertical component of the flow-rate of the water. A detailed study of the impact of fluid flow on temperature gradients reported from boreholes in Finland has been undertaken by Kukkonen (1988, 1992), but no comparable study has been undertaken with regard to boreholes in Sweden.

7.3.1 **The Record of the Vertical Component of Fluid Flow**

The overall vertical extent of an anomaly resulting from fluid flow will reflect the time-period over which the flow has been sustained. A recent change in the flow-regime, as results when drilling fluid passes into a fracture-zone, produces a temperature spike. Long-term this spike relaxes into a change in the temperature gradient with depth.

Of seventeen drillholes, between 270 and 1080 m depth in Finland, Kukkonen (1988) found sharp vertical changes in heat flow density and temperature gradient in eight, with groundwater flow considered to be responsible for the temperature variations in a further five holes. No similar comparative study of the impact of groundwater flow on borehole temperatures exists for Sweden. From results collected by Parasnis (1975) in the Aitik region it is clear that water was draining into the hole passing into fractures at around 250 m and 500 m depth, encountered in a number of boreholes, significantly affecting the measured temperature profiles. Temperature profiles were similarly affected in the borehole KLJ 1 at Lansjarv, indicating fluid flow through the borehole, as they were in borehole KF111 at the Finnsjon site, where flow from the surface was passing into the upper high conductivity edge of fracture zone 2, that also marked a change in temperature gradient (Tiren, 1989).

It might be anticipated that the equation of input and output required by simple fluid convection would mean that those boreholes in which the temperature is depressed by downward circulation are compensated by those in which temperatures are raised by upward moving water. An asymmetry is however noteworthy, that in borehole observations from Sweden and Finland perturbations to the measured temperature gradient associated with downward circulating fluid appear more common and larger than those attributable to upward circulation. This effect may chiefly be the result of bore-hole siting on hills (K Ahlbom personal communication) rather than a response to increasing crustal porosity (see Section 8.5.2).

7.4 EXTRAPOLATING THE GEOTHERM TO DEPTH

The extrapolation of near-surface thermal observations to depth demands assumptions about the concentration of heat-producing radiogenic elements in the crustal rocks and also their conductivities. It also requires adjusting the near-surface gradient to take into account the residual effects of the colder surface temperatures of the last glaciation. The uncertainty in this extrapolation can be expressed in terms of the likely error on the estimation of either the range of temperatures that might exist at some depth or the range of depths at which a particular temperature is surpassed. The seismotectonic significance of this depth range concerns the brittle-ductile behaviour of the rockmass and hence is itself dependent on the mineralogy of the lower crust. With such inherent uncertainties any depth is likely to have errors of at least 10-20%.

7.4.1 Deep Boreholes

The best correlation of this extrapolated geotherm comes from deep boreholes. The deepest borehole in the Baltic Shield, that of the superdeep borehole on the Kola Peninsula (USSR) reaching to a depth in excess of 11,500 m, revealed some important thermal anomalies that also could be correlated with observations of hydraulic conductivity and salinity (Kremenetsky et al, 1989 and Kukkonen, 1989). The heatflow showed steps at a number of depths between 1300 and 5000 m, increasing from about 29 to 50 mW/m^2, probably reflecting downward circulating groundwater. At 4500-5000 m the heatflow density reached a value of 70 mW/m^2 correlating with the highest values of dissolved solids found in the borehole, indicating upward circulation. Active water influx zones were encountered down to 11,000 m.

The significance of observations from this single superdeep borehole may not be typical. The temperature profile of the Gravberg borehole suggests a fairly regular temperature gradient with depth down to about 7000 m (see Figure 7-2). However Kukkonen (1991) has calculated that once bulk hydraulic conductivities rise higher than 10^{-8} to 10^{-7} m/s, heat transfer by forced and free convection can begin to dominate over conduction, and that due to the preponderance of downward moving water encountered in boreholes (see above) the heat flow density in central Fennoscandia may be underestimated.

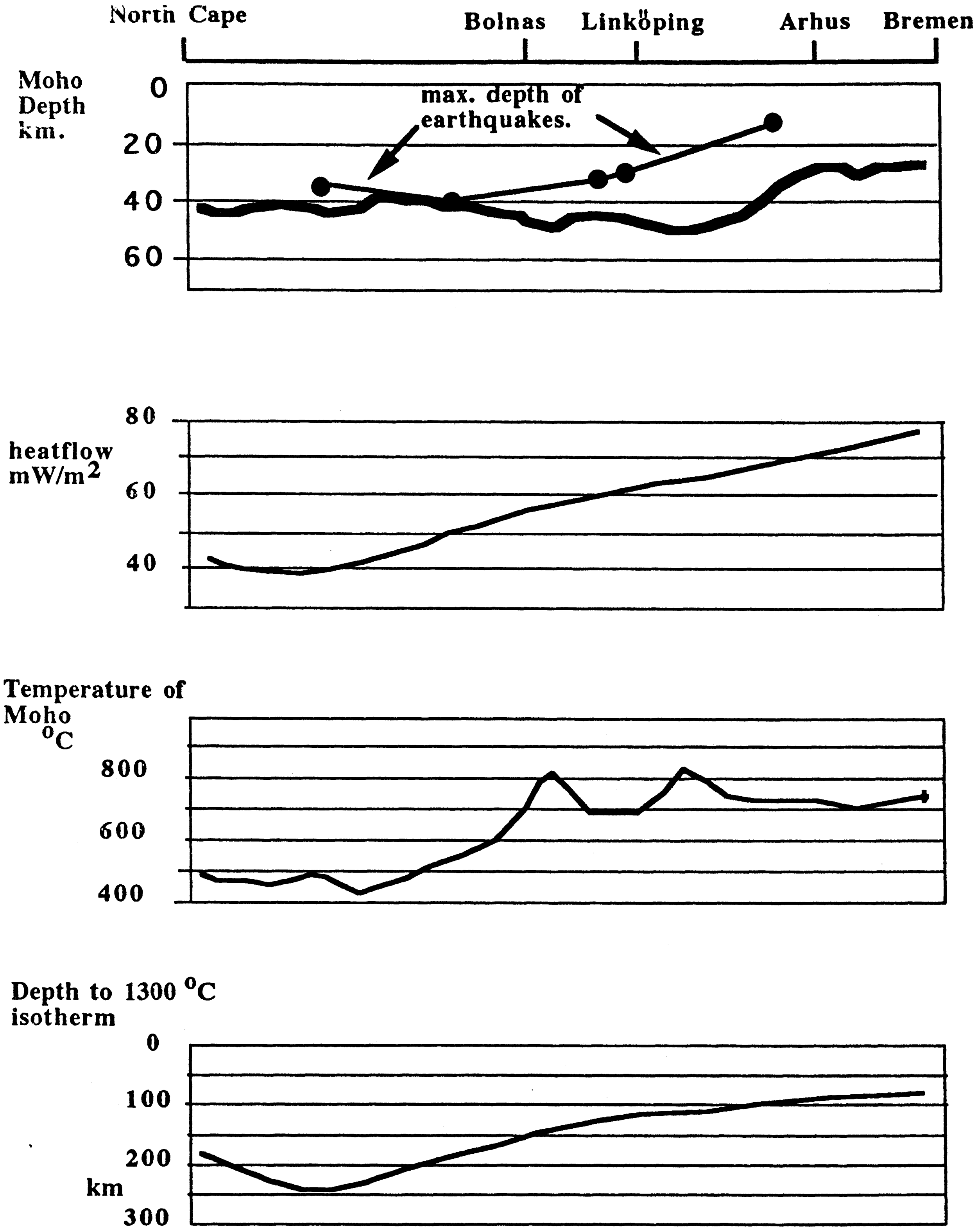


Figure 7-3: Projection of surface geotherm to depth along the line of the northern segment of the European Geotraverse.

7.4.2 Radiogenic Heat Production

Regional variations in the radio-isotope concentration of the crustal rocks, as between granites and potassium-depleted high grade metamorphics, cause significant variation in internal heat generation that have to be accommodated when extrapolating temperatures deep into the crust or underlying mantle. Numerous attempts have been made to refine these latter extrapolations from chemical and conductivity studies of rocks considered typical of the middle and lower crust and attempts have been made to find a correlation between the seismic velocities of crustal rocks and their radiogenic heat generation (see for example Rybach and Buntebath, 1984). The recent thermal models of Pasquale et al, (1991) are considered to provide the best available model of crustal and upper mantle heat-flows and temperatures beneath Sweden.

7.4.3 The Swedish Geotherm

Heatflows adjusted for the effects of the Ice Age, vary from above 50 mW/m^2 in southern Sweden falling to 35 mW/m^2 in central Sweden and rising towards the north reaching 45 mW/m^2 in the Caledonides of the North Cape, Norway. The modelled profile along the line of the geotraverse is illustrated in Figure 7-3. Significant variations in the temperature regime exist beneath Sweden. The 400 °C isotherm sinks from 28 km in southern Sweden to almost 40 km in central Sweden. Temperatures at the Moho vary from 350 up to 900 °C in southern Sweden. These projected geotherms reveal significant variations in the heatflow out from the Mantle below Sweden indicating differences in Mantle rheology and origin.

These profound variations in geothermal gradient constrain the rheological behaviour of the crust and upper mantle underlying Sweden, affecting the controls of earthquake generation and lithospheric deformation. The implications of these thermal models are discussed in Section 8 of this report.

8 CURRENT SEISMOTECTONICS

Seismotectonics concerns the geological controls of seismicity. A seismotectonic understanding demands the integration of observations of crustal deformation, crustal structure and properties with information on earthquake locations, sizes, styles, distributions etc.

8.1 TECTONICS OR REBOUND?

The spatial and temporal correlation of the uplift of Fennoscandia with the melting of the Fennoscandian ice-sheet has provided convincing evidence of the isostatic recovery of the region in response to the removal of the ice-load. Yet despite the universal acceptance of this model, somewhat surprisingly the cause of Fennoscandian seismicity is still much debated as to whether the earthquakes are a result of postglacial rebound or tectonics.

To summarise a long and complex history: in the 18th Century observers outside Scandinavia would often express incredulity at the stories of rising land because there were insufficient earthquakes (as for example the Italian mathematician Paolo Frisio, quoted in Ekman, 1991a). Towards the end of the 19th Century once land uplift had been generally accepted to be the result of glacial unloading seismic activity was considered the result of crustal rebound. However this view changed in the 1960s following increasing knowledge on the role of Tertiary uplift in the evolution of the Norwegian landscape (Kvale, 1960) and a tectonic explanation became increasingly predominant in the 1970s and 1980s following the investigation of earthquake activity away from plate boundaries (intraplate seismicity).

As intraplate seismicity could occur within the continental interiors in the absence of glacial unloading it appeared likely that seismicity in Fennoscandia had a common cause, somehow related to slow intraplate tectonic deformation. The discovery of apparent consistency among Fennoscandian stress orientations was employed as support for this view, and the evidence of significant Tertiary intraplate deformation within seismic profiles across the Norwegian continental margin through long periods of the Tertiary appeared to suggest some long-term geological evidence for such processes.

The chief elements of data introduced into a discussion as to whether tectonics or deglaciation was the principal control of seismicity has been that relating to the stress field (an earlier attempt to compare coseismic strain energy with the strain energy predicted from crustal flexure was somewhat inconclusive, Bath, 1972). Missing from this debate was any consideration of the strainfield. This is principally because of the difficulty of obtaining an accurate empirical overview of strain, as discussed in Section 4.8 above. However unlike stress that can record events in the past, the strain field and seismicity should both record only what is happening in the crust today.

8.1.1 **The Tectonic Strainfield**

Although the crystalline rocks of the Fennoscandian shield are particularly intractable for revealing tectonic activity (see Section 2.4 above), the Baltic Shield is almost surrounded by sedimentary basins (see Figure 2-3), from within which it is possible to obtain some upper bound estimates of deformation rates. Detailed studies of the geological deformation along the Norwegian continental shelf through the past 50 million years also provide some important constraints on the maximum level of tectonic strain that might be anticipated across Fennoscandia in the absence of any glaciations.

Important manifestations of Neogene compressional deformation are seen around the continental shelf margin in the Norwegian Sea. These compressional structures have amplitudes of up to 1 km. However most of the mid-Norway structures can be seen to have ceased uplift in the mid-Pliocene - the subsequent sedimentation being undeformed (Mutter, 1984). There is today rapid, presumed thermal, subsidence of the central Norwegian margin around the More Basin (Hamar and Hjelle, 1984). Despite an intensive examination of commercial seismic reflection profiles very few indications of faults cutting or deforming Quaternary sediments along the margin are known (see Muir Wood and Forsberg, 1987). It is therefore only possible to place some kind of upper bound on the amount of deformation likely to have been overlooked as distributed across a broad range of faults along the margin. It is highly unlikely that more than 100 m of deformation has been distributed over the 200 km wide mid-Norway continental margin in the past 2 million years (an annual upper bound strain of 2.5 x 10^{-10}). Only along the Lofoten margin is it possible that compressional deformation may continue (Mutter, 1984); although even along this margin it seems unlikely that the Quaternary horizontal shortening adds up to more than 200-400 m, a horizontal strain lower than 10^{-9}.

This continental margin tectonics demonstrates that the sedimentary basins along the shelf have acted as a buffer zone of thin warm lithosphere, between the stronger lithosphere of the oceanic crust and that of the old, cold, relatively rigid Baltic Shield. Within the shield strain rates are likely to have been significantly lower than along the margin. The Baltic Basin to the southeast of Sweden overlies the shield and in the extreme south comprises sediments as young as Neogene age (Ulmishek, 1991). The geological evidence of deformation within the Baltic Basin, over the whole of the past 500 million years, provides some major constraints on deformation rates. Across much of the basin there is no evidence for fault movements within the past 300 million years. Through the whole of the Tertiary (65My) it would be difficult to miss more than 100 m of deformation across the whole 500 km wide basin: equivalent to a strain rate of 4 x 10^{-12} per year, two orders of magnitude lower than that of the continental margin.

Hence estimates of the *upper bound* tectonic strain rate may be 10^{-9} to 10^{-10} across the Norwegian continental margin but within the Baltic Shield itself probably two orders of magnitude lower: 10^{-11} to 10^{-12}.

8.1.2 The Rebound Strainfield

As discussed in Section 4.1.6 the horizontal strain-field across Fennoscandia has been explored directly within geodetic studies in Finland, although in many regions the inherent errors are as large as the claimed signal. Strain-rates as high as a few parts of 10^{-7} were claimed (Chen, 1991), with both dilatational and compressive strain found in different regions. Estimates of the flexure implied by changes in the rebound gradient in Fennoscandia were shown to imply strains higher than 10^{-9} averaged over distances of 100 km (see also Section 4.1.6).

Gasperini et al (1991) have modelled the horizontal strain accompanying Fennoscandian rebound in a finite element code cast in axisymmetric formulation for a viscoelastic half-space, employing a high viscosity lithosphere, 120 km thick, underlain by a more fluid mantle (viscosity = 10^{21} Pa.s). At 13,000 years after the deglaciation of a 2500 m thick, 800 km radius ice-cap located on a shield region, horizontal strain-rates across the uplifting region are found to be around 2 x 10^{-9}/yr (see Figure 8-1). James and Morgan (1990) modelled the Laurentian ice-cap and found (for a uniformly stratified lithosphere and mantle 13,000 years after deglaciation) horizontal velocities of up to 3 mm/yr, that averaged across the 2000 km radius (Laurentian) rebound dome were also equivalent to a strain rate of ca. 10^{-9}/yr.

All such models have employed a laterally homogeneous lithosphere beneath the rebound dome itself and assume no sharp variations in ice-load, as are associated with a topographic front or continental shelf margin. The levels of horizontal strain measured in Finland suggest that these lateral variations may locally raise the strain-rate perhaps one order of magnitude higher than the predictions of the uniform models of the cratonic lithosphere employed by Gasperini et al, and James and Morgan.

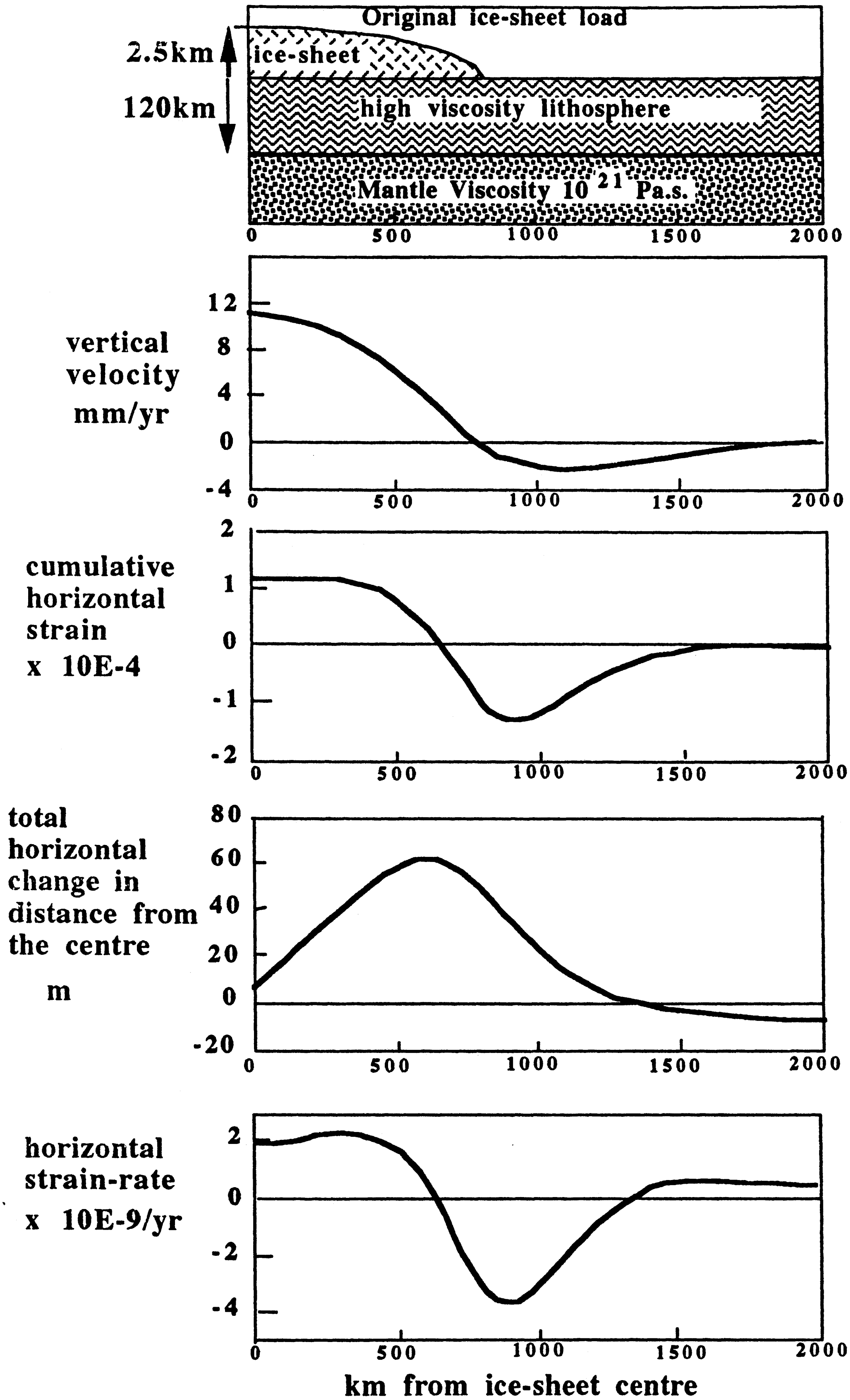


Figure 8-1 The horizontal strain-field of rebound, modelled for 13,000 years following (instantaneous) unloading (from Gasperini et al., 1991).

8.1.3 The Competing Strainfields of Rebound and Tectonics

From the relative magnitude of the modelled and measured rebound strainfield as compared with the upper bound estimates of the tectonic strain field, rebound should dominate the strainfield of the Baltic Shield by around two to three orders of magnitude. Along the Norwegian continental shelf, although the values are more comparable, rebound may also dominate the upper bound tectonic strain rate. This result is not surprising; the rates of crustal uplift over the past few thousand years are as high as those interseismic deformations found at the fastest moving plate boundaries.

The failure to answer the rebound versus tectonics problem in the past appears to have been an assumption that uplift was proceeding without any accompanying horizontal deformation. The observations and models of the horizontal strain field now suggest that rebound may dominate over any other cause of crustal deformation, not only across the rebound dome, but also over the surrounding collapsing forebulge to great distances into the Norwegian Sea and Central Europe.

8.1.4 A Tectonic Component in Addition to Rebound?

A number of authors have proposed that the current uplift of Fennoscandia comprises a subsidiary tectonic component of deformation in addition to rebound (eg. Mörner, 1977). Uplift, in response to the arrival of a lighter body of underlying mantle material, has been proceeding in western Scandinavia at various periods of the Tertiary: long-term uplift rates are likely to have been around 100 m per million years (0.1mm/yr), although lower rates of uplift over much of central Sweden imply rates 10-20% of this. Hence any non-rebound component of uplift is likely to be no more than 1%, and more likely 0.1%, of the maximum measured uplift of 10mm/yr. Such a tectonic component is effectively unobservable within the present geodetic and tide-gauge data. As long as rebound associated with the inward streaming of mantle material continues, it becomes difficult to conceive physically how any tectonic process associated with mantle motions can 'communicate' with Fennoscandia.

8.1.5 Reconciling Stress with Strain

The strainfield evidence provides support for a model in which the current deformation state of Fennoscandia is dominated by rebound. It therefore becomes necessary to enquire why the stress-field data has continued to be seen to support a tectonic argument (see for example Gregersen et al, 1991). The tectonic argument should first be stated: it is that the regional consistency of the stress-field reveals the influence of plate boundary forces (generally referred to as ridge push) and that if rebound was the dominant seismotectonic influence then the stress-field should show a radial or tangential influence.

That ridge push is not the simple control of seismicity is suggested by the significant asymmetry that exists between the opposing coastlines of the North Atlantic (Sykes, 1965), and the almost complete absence of seismicity along the continental margin to the western of the British Isles. The relatively low seismicity of the coast of Greenland in comparison with the relatively seismic continental margin of Norway clearly lends more support to the role of deglaciation than to 'ridge push' as the control of earthquake generation.

Arguments against the control of postglacial rebound that employ the evidence of the stress-field have considered that post-glacial uplift is a process in its own right: in effect that if rebound is the dominant control, that the stress-state should be a response to doming. However the process of rebound that is at present underway began around 50,000 years ago with glacial loading. The uplift should not be considered to be that of a dome but rather a subsidence bowl that continues to shallow until it eventually disappears. Hence the crust is returning to its original pre-glacial state of stress, not developing a new state characteristic of rebound.

Before glacial loading the crust was subject to a tectonic strain field. Although this cannot be known in detail there is widespread evidence from north west Europe to suggest that this comprised a NW-SE orientation of the principal horizontal stress direction. The orientation of the minimum stress direction is likely to have been either vertical, consistent with a compressional tectonic regime, or horizontal, consistent with a strike-slip regime (see Section 6). During the accumulation of an overlying ice-cap this stress-state was altered, first as a result of the increase in vertical load, but subsequently there would have been an increase of stress radial to the centre of loading, as a result of sub-crustal flow out from beneath the ice-load. The cumulative strainfield of loading is expected to be equivalent but of opposite sign to the strainfield of deglaciation, as discussed in Section 8.1.2 and illustrated in Figure 8-1.

Deglaciation removes the first of these stress changes as a simple response to unloading, but the second cause of stress modification is relieved far more slowly during rebound, as a result of crustal expansion that accompanies the return of sub-crustal mantle flow. As the rebound is today considered to be about 90-95% complete (Ekman, 1991b), so changes in the stress-state that accompanied isostatic depression should be 90-95% relieved. Ideally once rebound has ceased, the crust should return to its pristine stress-state. Hence the effect of rebound is to return the crust to its pre-glaciation stress-state rather than to impose a new stress-field. In such circumstances it becomes difficult to define exactly what stress-field should be considered 'evidence' of the influence of rebound.

8.1.6 **Modifications of the Stress State from Strain Release**

The chief cause of disturbance in this system will come if stresses are relieved through brittle fault rupture at some point in the glacial loading/unloading cycle. In such circumstances the stress-state will not return to its original state because there will have been permanent strain. That there is ongoing brittle fault rupture occurring within Fennoscandia is demonstrated by the occurrence of earthquakes. Currently the earthquakes are individually too small and the level of seismicity is too low to have a significant regional impact on crustal stresses. However if seismicity was higher and earthquakes much larger at some other part of the glacial cycle it is likely that the stress-field could have been significantly altered over large areas. Such areas would be unable thereafter to return to their pristine stress-state.

Hence it is likely that the current stress-state of Fennoscandia should be fairly close to the stress-state that existed prior to glaciation, except for regions in which there has been large scale stress-release in fault displacement and consequently permanent strain. It may be possible to identify such regions from stress-field anomalies, as that known from around the Lansjarv Fault. For regions unaffected by permanent strain continuing rebound is taking the stress-field back to its pristine state, not generating a stress-field 'characteristic' of rebound. The disparity between the pre-glacial tectonic stress-state and that encountered today (taking into regard the fact that rebound is not yet quite complete) could provide a measure of the degree to which there has been local strain release within the glaciation/deglaciation cycle.

8.1.7 **Where is permanent strain most likely to occur?**

The areas most likely to have suffered permanent strain, can be predicted to show some azimuthal variation relative to the centre of the ice-load. For a circular ice-cap located on a flat laterally homogeneous crust, the flanks of the downwarped crust will have suffered a radial increase in stress during loading. The impact of this increase will depend on the pre-existing stress-state that for Fennoscandia, as discussed in Section 8.1.1 above, is known to have a maximum principal stress oriented NW-SE. On those flanks of the downwarped bowl facing the principal horizontal stress direction, radial strain raises sigma 1 during downwarping and on glacial unloading this will tend to take the rock into failure (see Figure 8-2). In

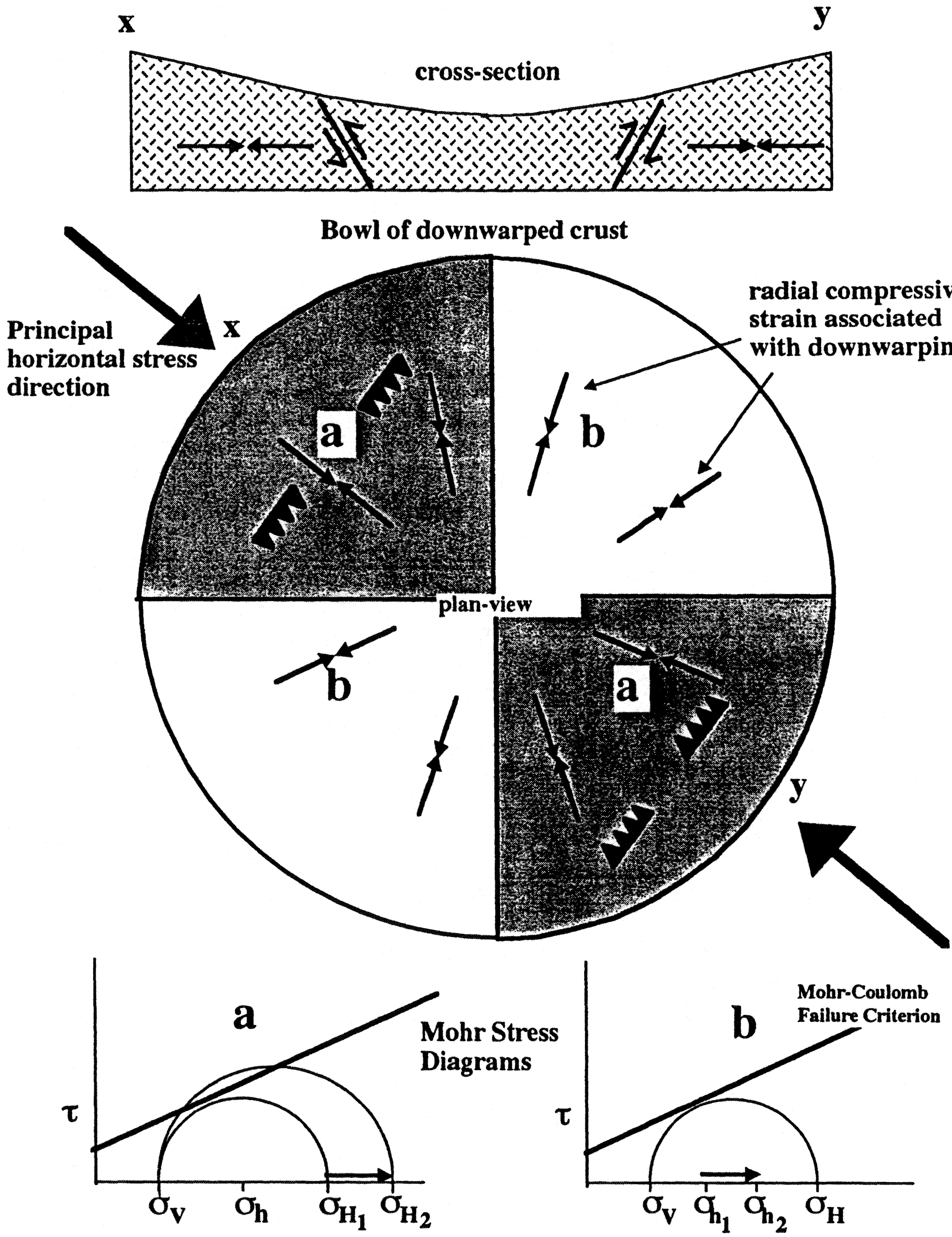


Fig 8.2: Model for the concentration of major strain release immediately following rapid glacial unloading (for a compressional tectonic stress-state).

contrast on those flanks of the downwarped crust lying orthogonal to this orientation it is the least principal horizontal stress that will be raised during crustal downwarping, leaving the failure conditions unaltered (for a compressional stress-state) or even reduced (for a strike-slip stress state).

For Fennoscandia one might therefore predict that stresses on both the north-east and south-west flanks of the rebound dome are likely to be returning to the pristine tectonic state as the minor horizontal stress continues to decline, while regions to the north-west and south-east of the rebound dome are more likely to have been modified as a result of permanent strain, in the immediate aftermath of glacial unloading. From Figure 8-3 it can be seen that many of the high quality in situ and focal mechanism stress observations from the region to the north-west of the rebound centre, in northern Sweden and northern Norway, do appear to give anomalous or rotated orientations of the principal horizontal compressive stress direction, suggesting that these regions may have had their stress-state modified as a result of permanent strain. In contrast, to the south-west of the rebound centre, in southern Sweden (outside Skane), the principal horizontal stress direction is far more consistently oriented NW-SE. From the stress observations there is less evidence to support permanent strain changes during deglaciation to the south-east of the rebound dome in Finland.

8.1.8 Tectonic Strain During Glacial Loading

During the existence of the ice-sheet the original horizontal tectonic stress may have continued to increase slowly as a result of the externally applied plate boundary forces (and intraplate strain) that prevailed prior to glaciation. (However it is also possible that the increase in vertical load will tend to force the horizontal release of strain to the periphery of the ice-sheet.) Once the load is removed this additional tectonic stress (stored as long as the load remains) will be relieved. This process is not expected to show any azimuthal variation around the rebound dome and, while capable of generating a

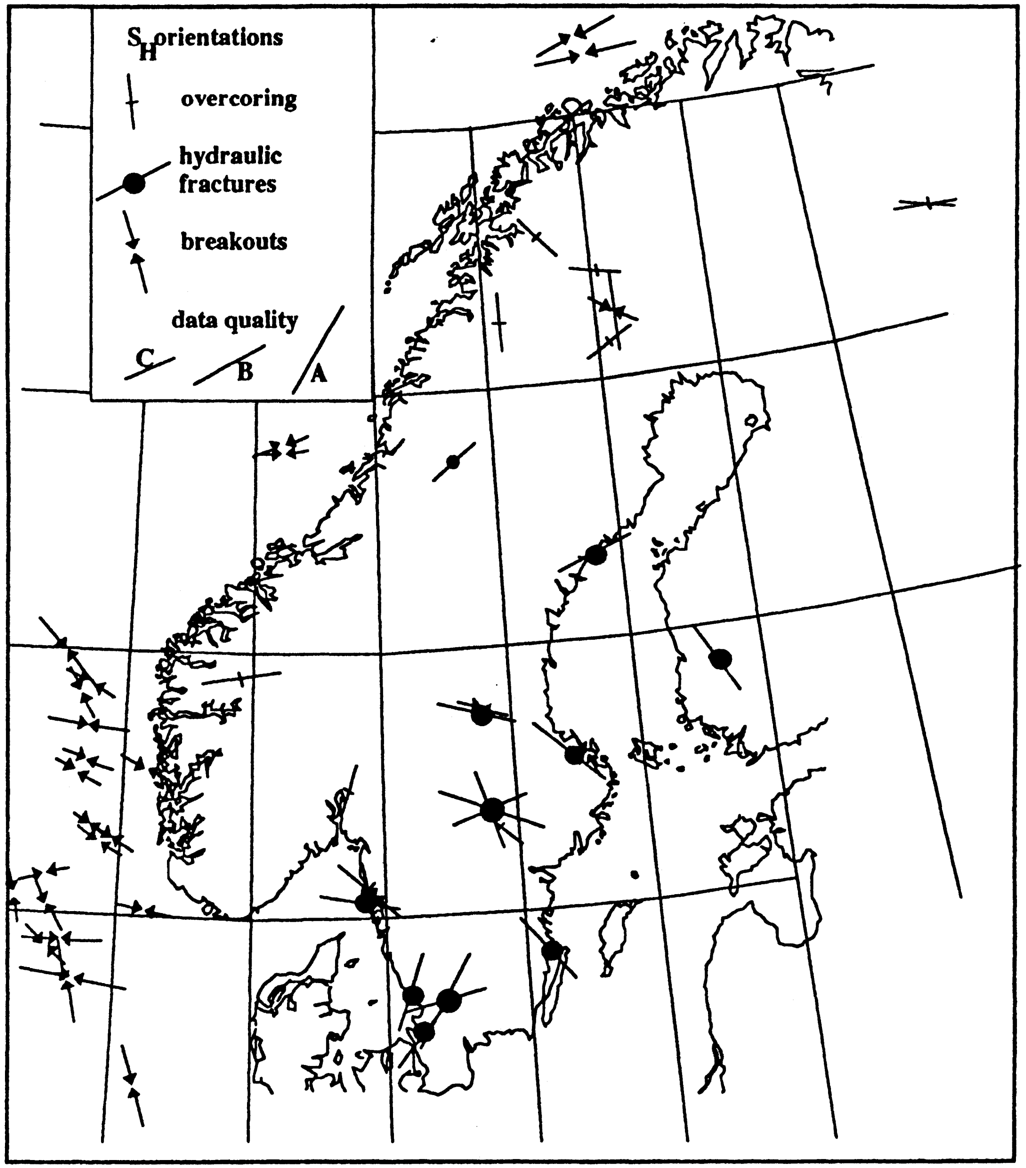


Figure 8-3: Highest quality (Grade A, B, and C) in situ stress observations for Fennoscandia.

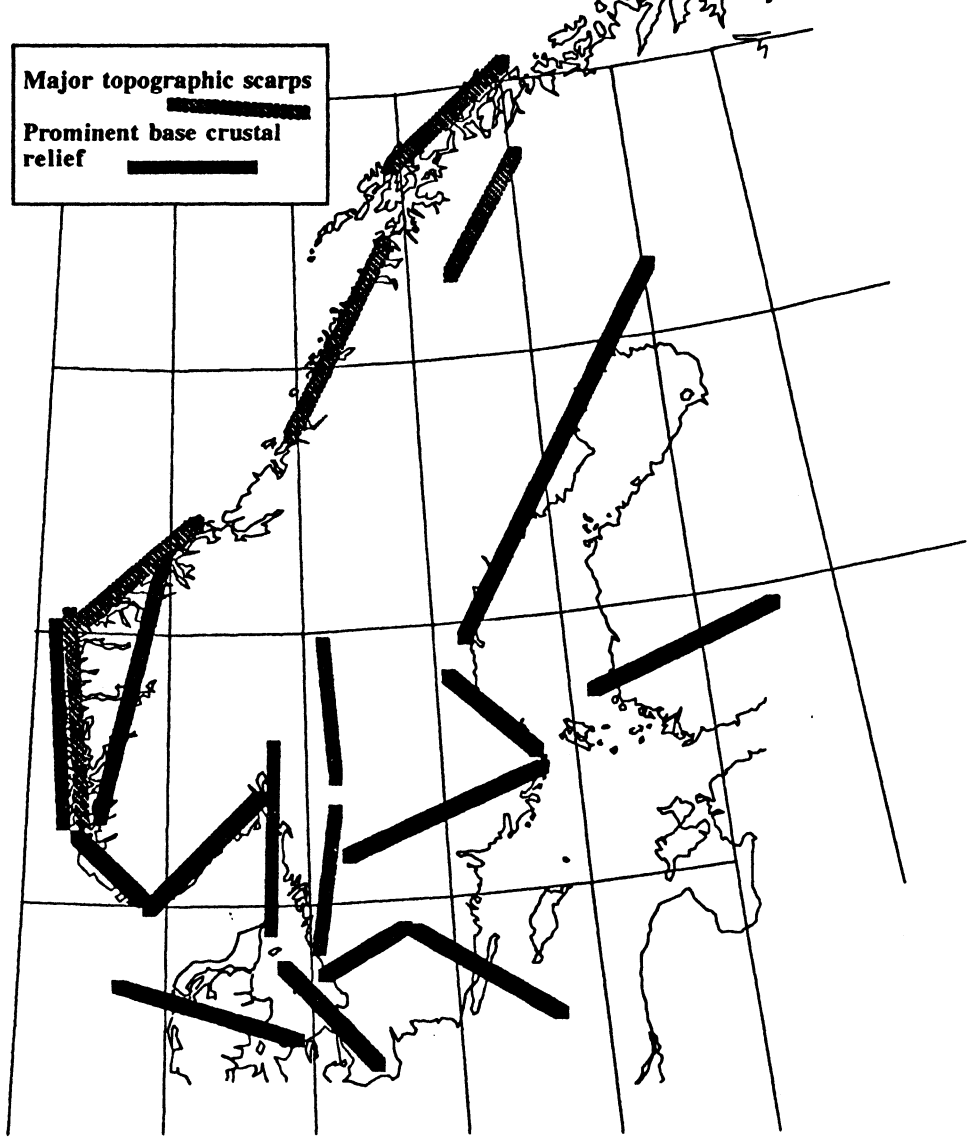


Figure 8-4: Principal linear topographic and Moho relief scarps in Fennoscandia.

burst of seismicity following glacial unloading, should not be the cause of any regional alteration in the stress-field.

8.2 REBOUND SEISMOTECTONICS: THE REAL WORLD

In a perfectly uniform lithosphere, without topography, on which a circular disc-shaped ice-load accumulated and subsequently melted, in the absence of any anisotropy in the underlying strain (and hence stress-field), the deformation associated with rebound would be evenly distributed around the rebound dome. Such a simple model is very far removed from the situation in Fennoscandia, in which the plan of the ice-cap was far from circular, the topographic relief is as large as the ice-thickness, both crustal thickness and heatflow show a twofold variation, and there is a marked asymmetry in the distribution of ice-load, topography, and lithospheric rheology, all within an anisotropic strain and stressfield.

In many publications the Fennoscandian rebound surface continues to be modelled as a simple smooth dome. In part, as discussed in Section 4.4 this is because the level of error inherent in the geodetic data and the wide-separation of survey lines and tide-gauges prevents anything other than a smooth surface being interpolated. However perhaps the fundamental reason is because these models are employed by researchers only concerned with the longest wavelength component of rebound, in order to gain one of the most important insights into the rheology of the Upper Mantle. For such Mantle modellers shorter wavelength variations in rebound are simply noise. However it is this noise that is of principal interest to those exploring how postglacial rebound affects crustal deformation.

8.2.1 Rebound Deformation

The sharpest changes in gradient of the rebound strain field are likely to be related to locations where there was a pronounced lateral variation in load, as across topographic scarps bounding areas of high and low relief, or where the underlying crust and lithosphere show variations in rheology.

(i) Topographic controls

Where an ice-cap overlies a marked topographic scarp between a highland and a lowland region there will be differential unloading during deglaciation and hence differential rebound. For Fennoscandia the most prominent major topographic scarps lie along the western coastline of western and northern Norway. One of the few major inland topographic scarps is the eastern boundary of the Caledonian mountains in northern Sweden (see Figure 2-3).

(ii) Rheology

The rheology of the lithosphere is principally controlled by two factors: mineralogy and the geotherm. Rocks whose mineralogy is dominated by quartz/feldspar (those of the crust) are weaker than rocks whose mineralogy is dominated by olivine (those of the mantle). For quartz subjected to a differential stress of 50 MPa, a temperature of about 300 °C would produce a strain rate of 10^{-15} sec^{-1}, whereas for dry olivine, a temperature of about 800 °C would be required (Kusznir and Park, 1987). Hence for the same geotherm, the thickness of the crust becomes of great importance: a lithosphere with a thick crust tending to deform more readily. It is for this reason that deformation becomes concentrated within the thicker continental crust along the ocean margins.

However temperature is also of great significance to viscosities and strain-rates. A temperature rise of 100 °C lowers viscosity by around one order of magnitude (Gasperini et al, 1991). For the lowest geotherms, characteristic of the shields (<45 Wm^{-2} see Figure 7-1), any applied force is carried to great depths typically in excess of 80 km (Kusznir and Park, 1987). As the heatflow increases so the boundary of ductile deformation rises up to 40 km for around 60 Wm^{-2} and by 80 Wm^{-2} is located at a depth of about 10 km. Hence for the same set of boundary forces crustal stresses rise with increasing heatflow, as does the tendency for faulting in the crust.

Extrapolations of temperature gradients from near-surface geothermal measurements can be checked against observations of the depth distribution of earthquakes, which provide an indication of the depth to the brittle-ductile transition zone (Chen and Molnar, 1984). The maximum depth of the small number of accurately observed earthquake hypocentres in Fennoscandia shows some pronounced regional variations, that are broadly consistent with the range of near-surface geothermal gradients (see Figure 7-4). The deepest accurately observed hypocentres are down to 40 km, close to the base of the crust in the centre of the shield, while depths have not been recorded below 20 km close to the Fennoscandian Border Zone.

The best determinant of lithospheric rheology across Fennoscandia is likely to be the thickness of the crust allied to the temperature gradient (see Figure 7-4). There are very significant variations in both these properties. Towards the centre of the Baltic Shield, the heatflow is generally low, and hence it is crustal thickness that may be the dominating influence. In particular the thinner crust of northern Sweden may be expected to make the lithosphere more rigid than the thicker crust of Finland. A zone of thicker (>50 km) and hence weaker crust is also located under central eastern Sweden (see Figure 2-2). However in passing out from the centre of the shield the heatflow variations are so pronounced that it is likely that these, rather than crustal thickness, will be the predominant control. In particular the high heat flows and raised geotherm of south-western Sweden will make the lithosphere far less rigid than that in the centre of the shield (see also Figure 2-1).

8.2.2 Rheology and Rebound

Figure 8.4 highlights some of the most important topographic and Moho relief boundaries around Fennoscandia. As topography is relatively subdued across most of Sweden and Finland, variations in rebound gradients are most likely to reflect changes in underlying rheology. Where the rheological properties of the crust or lithosphere vary from one province to its neighbour, deformation may become concentrated along the weaker margins of the more rigid province.

Comparison of Figure 8.4 and Figures 4.6 and 4.7 illustrates the degree to which crustal thickness appears to influence rebound gradients. It can be seen that a number of the most significant changes in gradient are found across zones where the crust shows a significant thickness variation. In central northern Sweden where the shield crust is relatively thin, the geodetic profiles reveal almost no internal deformation. However where the crust increases in thickness towards the south, the geodetic profiles reveal gradient changes.

The Oslo Graben is the most prominent indent of thinner lithosphere into the Fennoscandian Shield and is likely to concentrate deformation along the southern flank of the rebound dome. The western coast of southern Sweden also marks a very pronounced change in crustal thickness and an increase in geothermal gradient that again is liable to concentrate deformation. To the west of the rebound centre the form of the rebound surface is steeper (and along the coast of Norway perturbed by topographic effects and changes in lithospheric rheology) far more than across the cratonic region to the south-east that passes across Finland into Russia.

8.3 REBOUND DEFORMATION AND SEISMICITY

Changes in the gradient of rebound imply concentrations of crustal deformation leading to brittle coseismic fault rupture, as modelled by Stein et al (1979). In Sweden there are three prominent boundaries between thinner and thicker crust: (i) first along the crest of the rebound dome on the western coastline of the Gulf of Bothnia; (ii) along the western coast of Southern Sweden, and (iii) along the northern edge of the thicker crust of eastern central Sweden. From the distribution of seismicity it appears that deformation is principally distributed along and around these zones.

As discussed in Section 4, geodetic observations have many problems of data-collection, inherent errors, time and length scales for revealing deformation. If it is accepted that all earthquakes within the Baltic shield are a result of rebound it becomes possible to employ the evidence of seismicity for mapping this deformation far more effectively than would be possible from any geodetic survey. In turn the distribution of seismicity may be seen to give primary information on prominent variations in lithospheric rheology.

8.3.1 Depth Dependent Stress Release

Focal mechanisms of earthquakes show some pronounced regional variations, that may give insights to crustal flexure within the rebound dome (see Figures 8-5a and 8-5b). A component of extension characterises earthquakes in the Oslo Graben, but compression is found from focal mechanisms along its north-east margin. Extensional events are also located to the north-west of reverse fault earthquakes on both the east and west coasts of southern Sweden. An extensional component of fault-displacement also dominates across southern Norway. All these variations may represent ridges (of extension) and saddles (of compression) in the rebound surface. Employing only focal mechanisms of earthquakes for which accurate hypocentral depths are available, in Norway extension is found to be consistently distributed at shallower depths than compression, while in Sweden the shallowest earthquakes have given compressional focal mechanisms, although extension dominates at mid-crustal depths (see Figure 8-6). The areas subject to shallow earthquake generation appear to be relatively restricted (see Figure 8-7) implying that significant crustal flexure is itself localised along rheological (as in Sweden) or topographic (as in Norway) boundaries. Depth dependent stress release is itself consistent with earthquake generation through crustal flexure.

8.3.2 Deformation Around The Baltic Shield

As noted in Section 5, before proceeding to compare levels of seismicity or seismic energy release across Fennoscandia and the western continental shelf it is important only to employ sets of data complete above some threshold. These data-sets can be compared in a variety of ways. In Figure 8-8a comparison is made between the seismicity of the

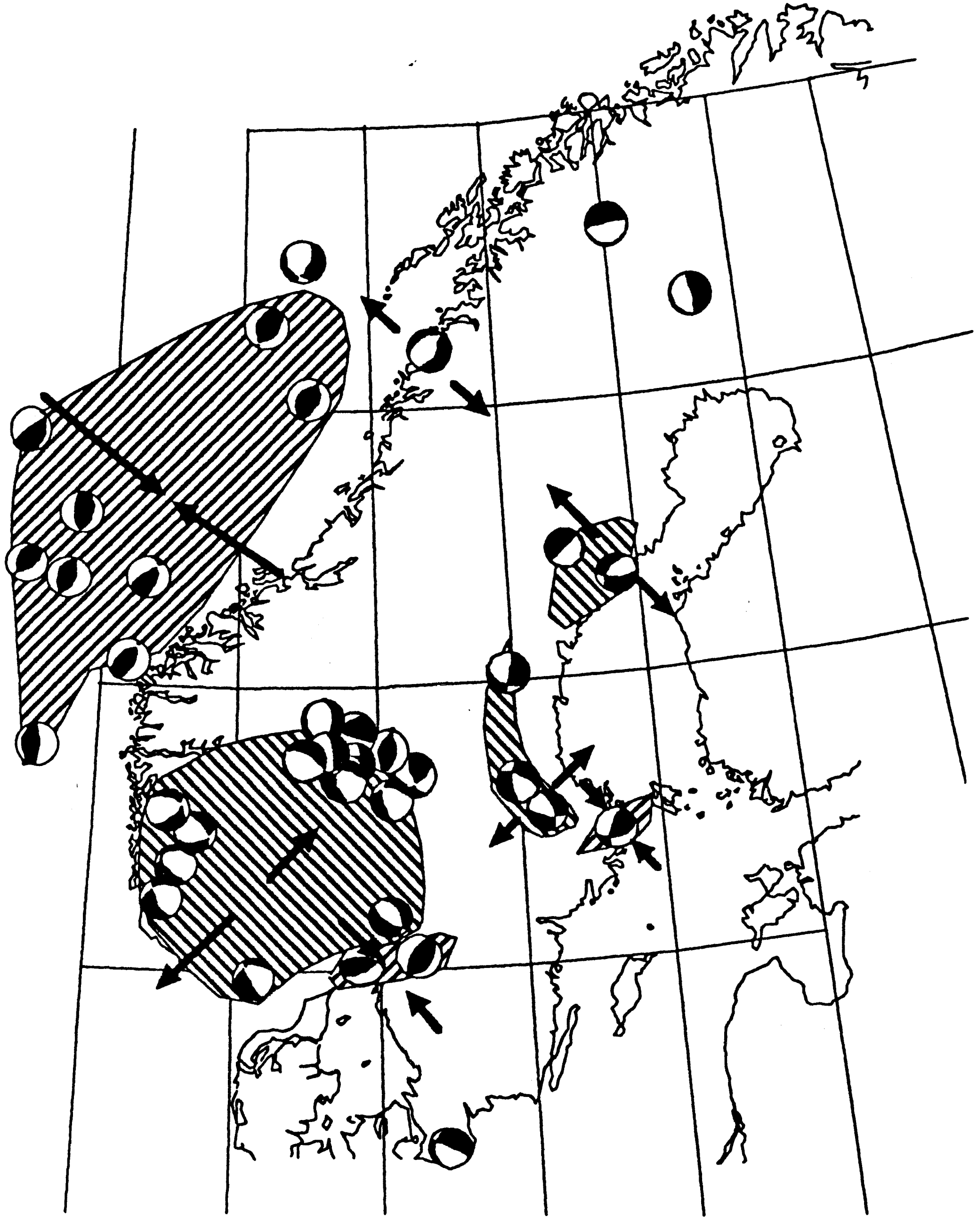


Figure 8-5a High quality dip-slip focal mechanisms in Sweden and Norway.

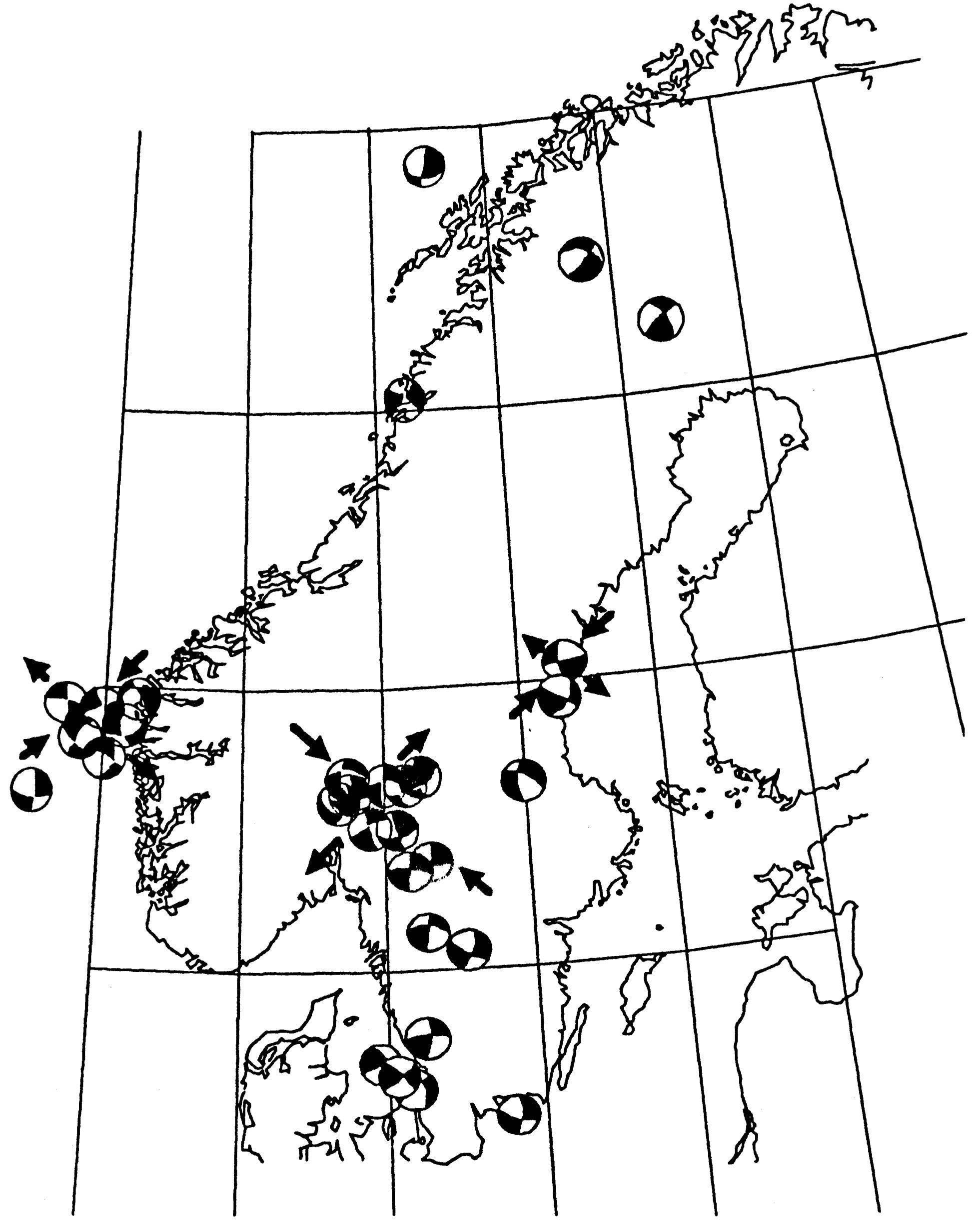


Figure 8-5b High quality strike-slip focal mechanisms in Sweden and Norway.

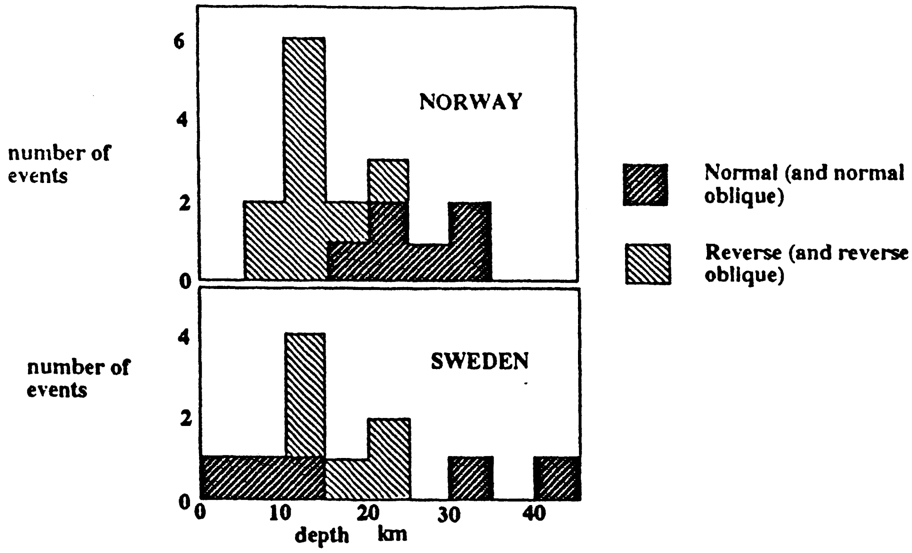


Figure 8-6: Variation of fault style with depth Norway vs. Sweden.

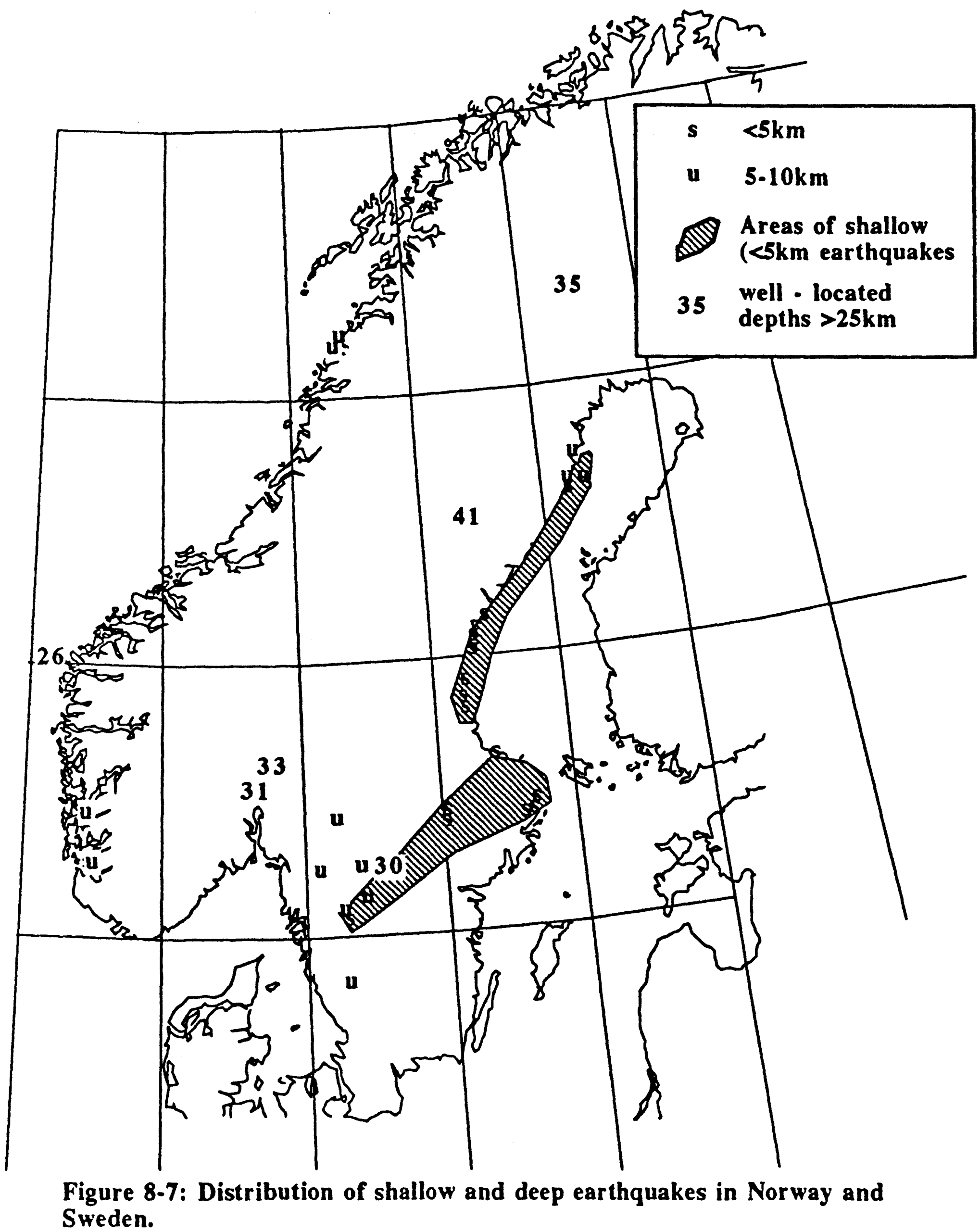


Figure 8-7: Distribution of shallow and deep earthquakes in Norway and Sweden.

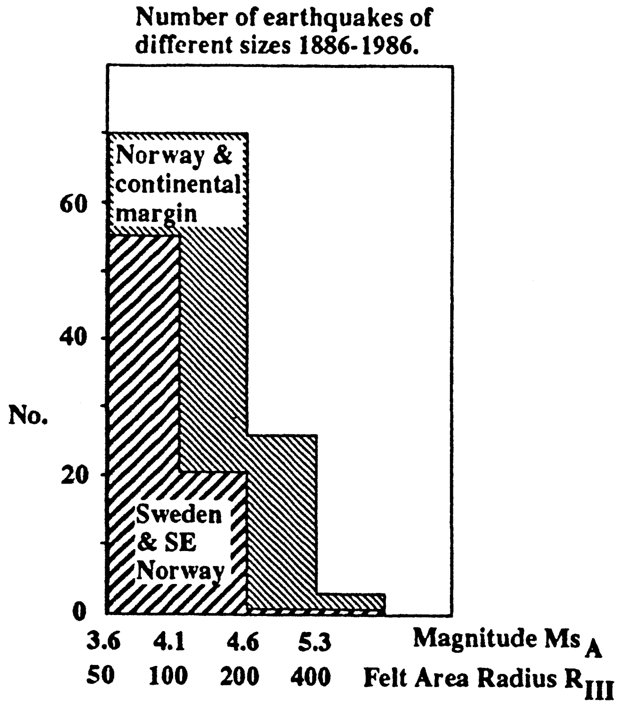


Figure 8-8: Comparative seismicity of Norway and Sweden.

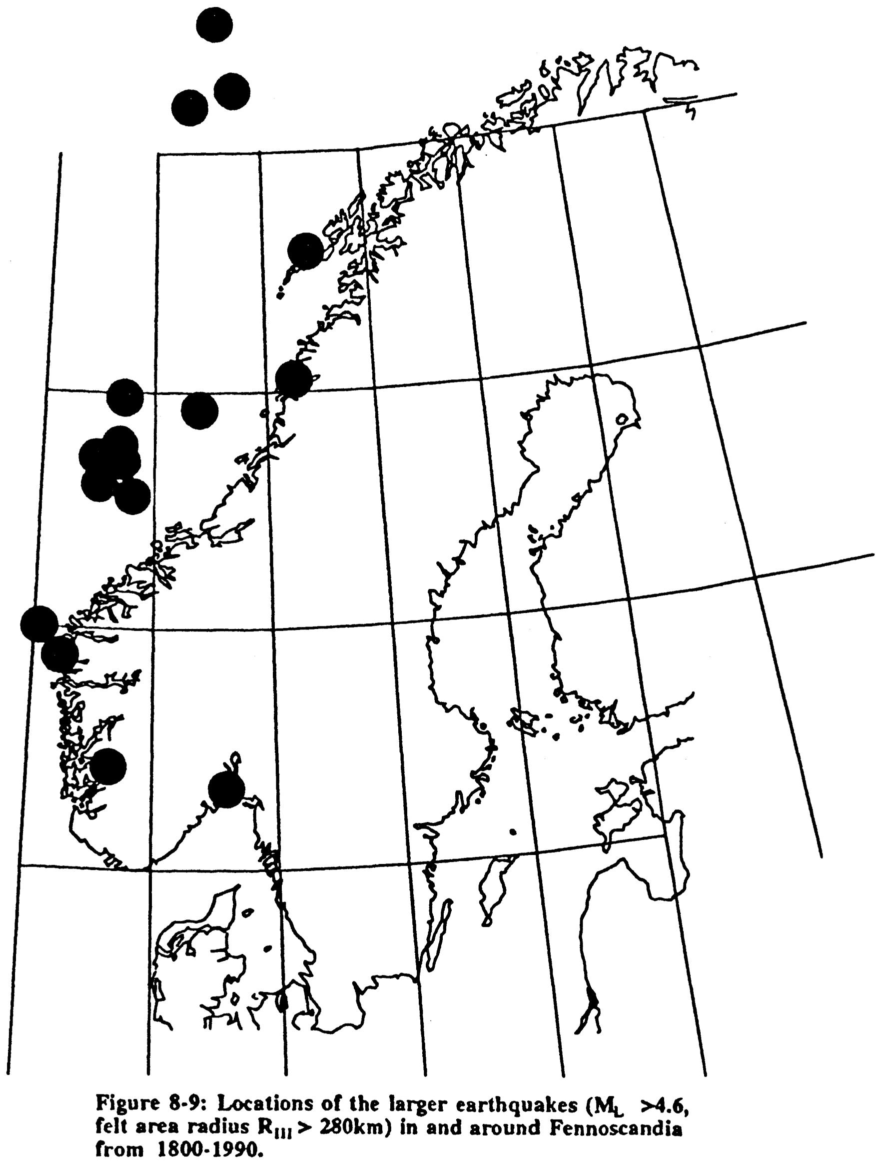


Figure 8-9: Locations of the larger earthquakes (M_L >4.6, felt area radius R_{III} > 280km) in and around Fennoscandia from 1800-1990.

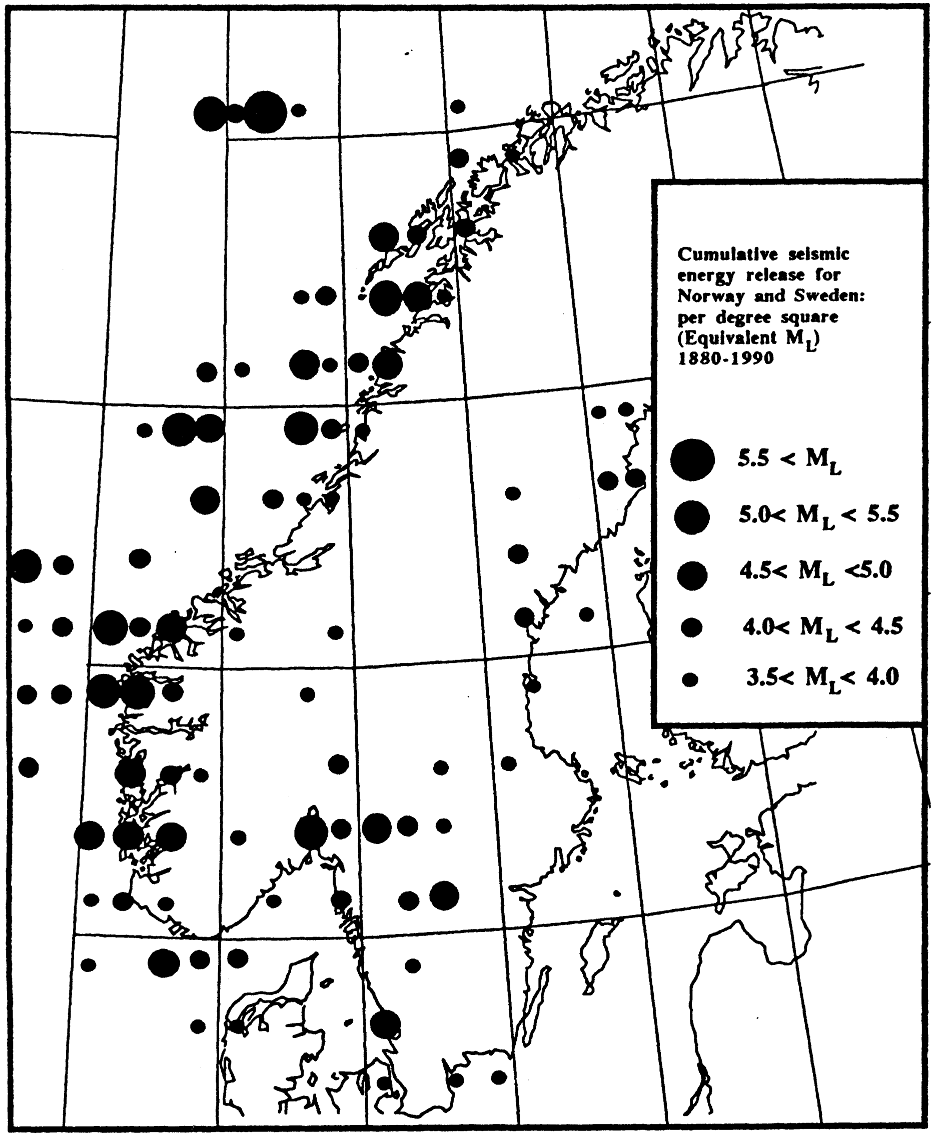


Figure 8-10: Seismic energy release for Norway and Sweden (in equivalent M_L), summed per degree(lat.) x degree (long.): 1880-1990.

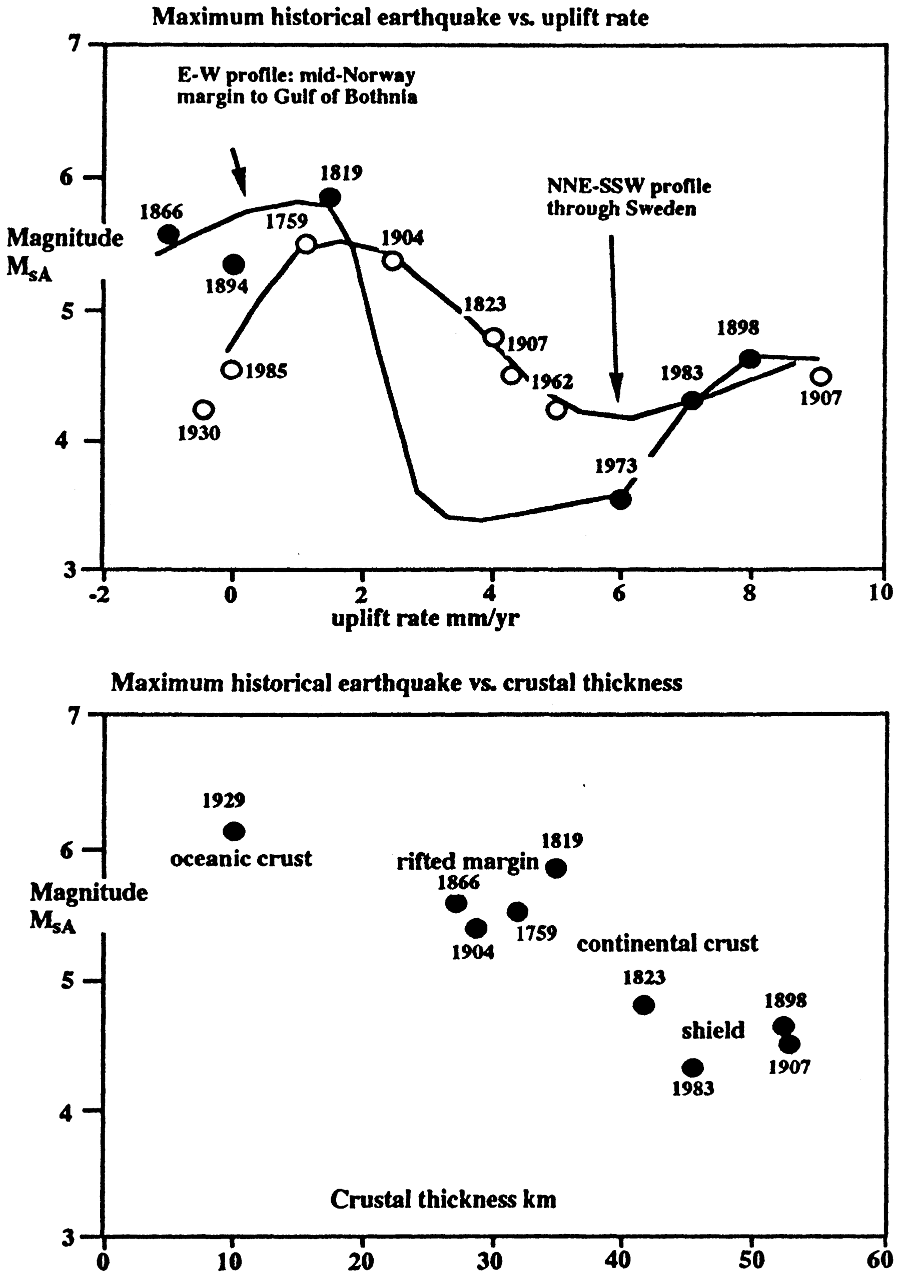


Figure 8.11: Maximum magnitudes for Norway and Sweden vs. uplift rate and crustal thickness.

approximately equivalent areas of Sweden (including south-east Norway) with that of western and northern Norway including the adjacent continental shelf. In Figure 8-9 the distribution of the larger magnitude earthquakes since 1800 is shown, while in Figure 8-10 the cumulative seismic energy release for the period 1880-1990 (in the form of an equivalent magnitude M_L) has been summed for each degree square of Sweden, Norway, Denmark and the adjacent continental margins. As is apparent from Figures 8-8, 8-9 and 8-10 the seismicity is more than one order of magnitude higher in Norway relative to Sweden.

In Figure 8-11 the maximum known historical earthquake magnitudes are plotted against both uplift rate and crustal thickness. It will be seen that the largest earthquakes are currently found around the margins of the rebound dome, although magnitudes appear to rise slightly once again towards the centre of rebound (caution should be used in attempting any statistics from these plots as the areas sampled decrease with increasing uplift rate). Again there is an interesting comparison to be made with the profiles of uplift shown in Figure 4-4. In passing from the mid-Norway continental shelf towards the centre of rebound, the highest magnitudes are found around the marked changes in uplift gradient located around the coast, as well as within the forebulge. The highest magnitudes known from central northern Sweden may be less than M3. Earthquake size increases again around the Gulf of Bothnia. In contrast, in passing into the rebound dome from south-west Sweden, there appear to be mild geodetic gradient changes and moderate magnitude earthquakes distributed along much of the length of the country.

Along the Norwegian continental shelf concentrations of seismicity mostly lie beyond the rebound dome within the marginal forebulge area that is sinking as a result of the streaming of the underlying mantle material back towards the centre of the shield. The collapse of this forebulge induces a radial increase in stress. Those areas along the Norwegian continental margin that lie facing the North Atlantic spreading ridge, are expected to suffer continuing increases in the maximum horizontal stresses, inducing reverse fault earthquakes late into the forebulge collapse, and it is in this offshore region of the Norwegian continental margin that many of the largest earthquakes have been located within the past century (see Figures 8-9 and 8-10), that from the recent observations of focal mechanisms (see Figure 8-5a) are consistently associated with almost pure reverse fault solutions.

If there was a corresponding forbulge to the south-east of the rebound dome (around the south-east border of Finland), one could anticipate similar coseismic deformation. The absence of such activity might in turn reveal asymmetries in sub-lithospheric mantle flow, or resultant crustal deformation.

Across much of the Baltic shield, as horizontal stresses continue to be relieved radial to the centre of uplift, observations of stress (see Section 3), strain localisation (see Section 4.1.5) and distributed seismicity confirm that the low rates of deformation are accommodated by the jostling of crustal blocks, with very small movements being concentrated along pre-existing major faults and fracture-zones.

8.4 SPATIAL AND TEMPORAL VARIATIONS IN SEISMICITY

8.4.1 Temporal Variations in Seismicity

In Figure 8-12 decadal average water-level changes at the longest running tide-gauges around the Swedish coast (from Sjöberg and Fan, 1986) have been plotted alongside annual seismic energy release (from Meyer and Ahjos, 1985). There appears to be some consistency between the fall-off in water-level changes in the 1940s with a decline in seismic energy release. The water-level observations might reflect either or both fluctuations in uplift rate, or some oceanographic effects. A correlation with a greater number of peaks and troughs would be preferred before an inter-relation between changes in apparent uplift rate and seismic energy release could be proved statistically.

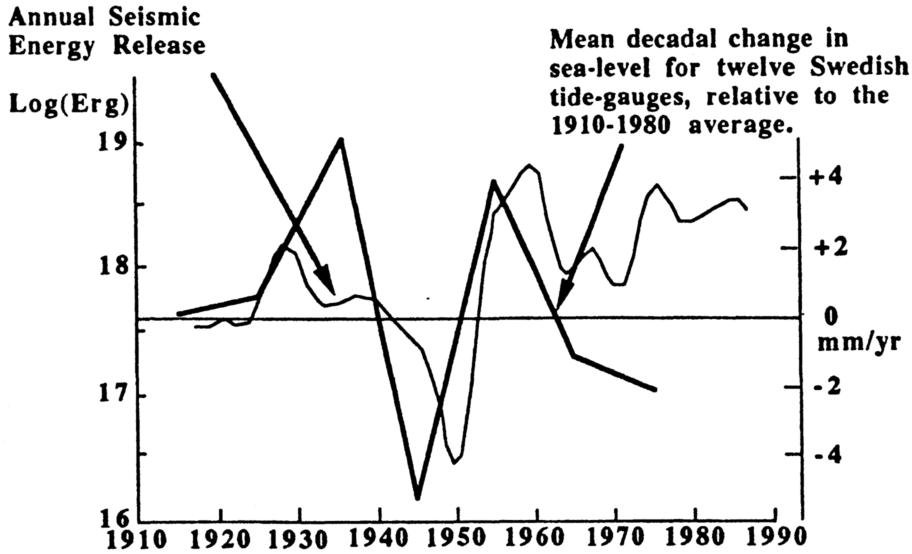


Figure 8-12: Annual Fennoscandian seismic energy release compared with decadal average land-level changes from Swedish tide gauges.

A correlation between seismic energy release in Fennoscandia with that of mid-Atlantic spreading ridge was proposed by Meyer and Scherman (1988) and Skordas et al, (1991). The correlation is chiefly founded on an apparent dramatic slump in seismic energy release both on the North Atlantic spreading ridge and in Fennoscandia in the 1940s, unfortunately coinciding with the war-affected slump in observational seismology.

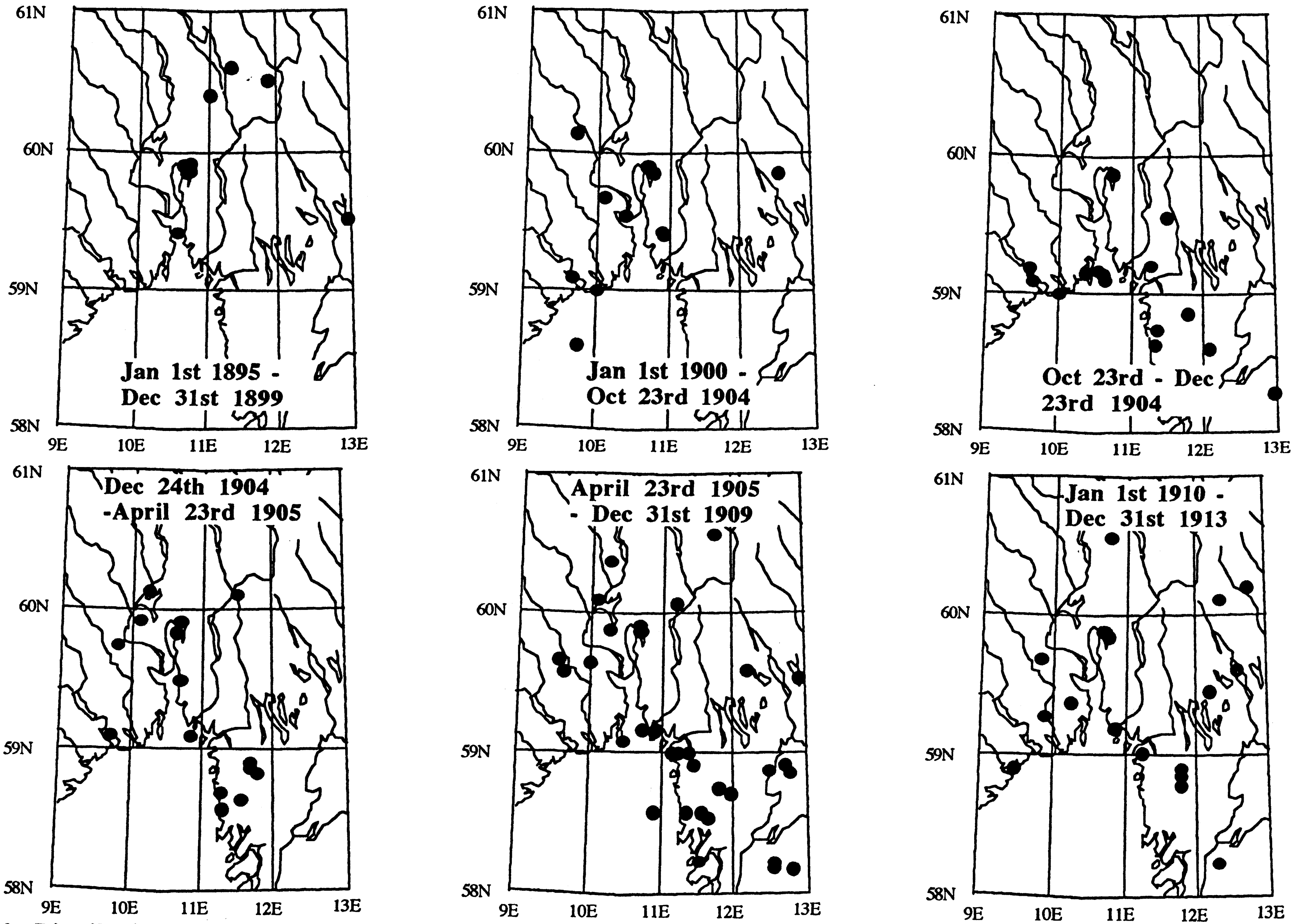
A connection between Fennoscandia and the adjacent spreading ridge, if demonstrated, is perhaps not surprising as it would indicate that strain in the lithosphere (and pressure in the asthenosphere) are interconnected. Such regional scale inter-communication has been identified in other regions of the world. Such a correlation is not, as some have claimed, a demonstration that the spreading ridge is driving all seismicity in Fennoscandia. It is equally probable that variations in the rate of rebound (as are suggested by the tide-gauge data for the 1940s, see Figure 8-12) are affecting seismicity at the spreading ridge. From the model of Gasperini et al, (1991) the cumulative horizontal strain accompanying post-glacial rebound (since 13,000 BP) has been up to 120m across the region, while over the same time-period the mid-

Atlantic spreading ridge should have widened by an estimated 260m. The underlying asthenosphere has had to accommodate two parallel zones of upwelling. Perhaps some of the dilation anticipated at the spreading ridge has instead been occurring beneath Fennoscandia?

In terms of mass movement of the asthenosphere the uplift of Fennoscandia is probably the dominant process. The volume of asthenospheric material emplaced during spreading (taken as an equivalent dyke widening by 20mm per year extending down to a notional depth of 100km) reflects an areal expansion of 2,000m2/yr, whereas the uplift of Fennoscandia across a NW-SE profile (1000km across and with a central uplift of 10mm/yr) reflects an areal expansion of ca. 5,000 m2/yr. Hence rebound may continue to have a more profound and widespread impact on surrounding lithospheric strains and the asthenospheric flow regime, than sea-floor spreading. During deglaciation the asthenospheric flux beneath Fennoscandia must have been more than an order of magnitude larger than beneath the adjacent spreading ridge.

8.4.2 Spatio-Temporal Clustering

On a number of occasions there appear to have been regional changes in seismicity in some part of Fennoscandia that cannot be considered to reflect a simple mainshock-aftershock sequence. Following the Oslofjord earthquake of 1904, there was a significant regional increase in seismicity over an area of more than 10,000 km^2, that was far too extensive to reflect aftershock activity in the vicinity of the original mainshock. In Figure 8-13 the reported seismicity of this region is illustrated for a series of time windows, for five year intervals in the decade before and after the earthquake, as well as shorter periods immediately following the earthquake on October 23rd 1904. Prior to the event, activity was located to the north of the Oslo Graben, but following the earthquake extended over a larger region to the south-east. It is difficult to know what the baseline of seismicity was in this region, prior to the 1890s as macroseismic data only began to be collected systematically around this time.



8.13: Distribution of (felt) earthquakes before and after the October 23rd 1904 Oslofjord Earthquake.

It is possible that both the earthquake and the regional increase in seismicity reflect some underlying process, perhaps some aseismic regional strain release or strain wave, in the lower crust or underlying mantle. There is unfortunately no suitable tide-gauge data for this region for this period to test whether a geodetic signature of such a process can be located.

Another spatio-temporal concentration of seismicity was noted by Porkka and Korhonen (1977) in the mid 1970s along a WNW-EWSE trending zone passing from the Norwegian coast through to northern Finland. This may also have a cause in some strain release beneath the brittle crust. Such clusters of earthquakes may reveal that the crustal deformation that accompanies rebound is, at least on occasion, regionally discontinuous. As with all other aspects of rebound seismicity, the localisation of earthquakes does not appear to be controlled by the existence of any major crustal fault, but rather by broader processes perhaps responding to variations in lithospheric rheology.

8.4.3 **Seismic Swarms**

True seismic swarms (concentrations of seismicity at a single location that may continue for several years) have been reported at a number of locations in Fennoscandia, in particular at Meloy, south of Bodo, Norway in 1978 (Bungum et al, 1979) and close to nearby Luroy where swarm activity continued for more than a decade following the 1819 August 31st earthquake (Muir Wood, 1989b). Similar swarm activity may explain the concentration of reported earthquakes around Hernosand, on the Swedish coast of the Gulf of Bothnia, in the mid 18th Century (Kjellen, 1910). All these areas can be predicted to be in zones of shallow extension (Hernosand on the crest of the rebound dome, and Luroy and Meloy, within the zone of steeper rebound gradients inferred to pass along the Norwegian coast). Prominent seismic swarms, and other spatio-temporal event clusters are shown in Figure 8-14. Such earthquake swarms are also found in central Scotland and parts of Arctic Canada (Basham et al, 1977) and appear to be characteristic of areas subject to glacial rebound, potentially reflecting some combination of crustal extension, crack propagation and fluid flow.

8.5 CRUSTAL STRAIN AND GROUNDWATER

8.5.1 Hydrogeological Consequences of Coseismic Strain

Significant increases in the surface discharge of groundwater have been observed following a number of major earthquakes and both the magnitude and geographical extent of groundwater discharge have been found to show a strong correlation with the modelled strainfield of elastic rebound surrounding a fault rupture (Muir Wood and King, 1993). For a normal fault earthquake, elastic rebound following fault rupture is compressional; for a reverse fault earthquake, extensional. Compressional strain is found to correlate closely with the volume of water released, indicating that to a depth of around 5-10 km, strain is accommodated by changes in the apertures of water-filled cracks, and hence strain changes lead to an alteration in crustal porosity. Typically, a Magnitude 7 normal fault earthquake is found to release around 0.5 km^3 of water over a region of 10,000 km^2 and a period of 8 months.

8.5.2 Hydrogeological Consequences of Rebound Strain

The crust of Fennoscandia has been subject to widespread horizontal strain as a result of post-glacial rebound. It is to be anticipated that this strain has been accommodated by changes in the aperture of cracks. As the rebound dome predominantly involves crustal extension, an increase of crustal porosity can be predicted across Sweden over the past 10,000 years. This in turn should have drawn surface water into the crust. In a crust with a dynamic porosity of 0.1%, a cumulative post-glacial strain of 10^{-4} since deglaciation (consistent with the models of Gasperini et al, 1991, see Figure 8-1), distributed down to a depth of 10 km would draw surface groundwater to an average depth of 1000 m. This depth is of course very sensitive to the dynamic porosity of the rockmass.

It has been widely noted that water with a very low dissolved ion content is found down to several hundred metres in boreholes drilled in many parts of the Baltic Shield (see Lahermo and Lampen, 1987). Saline groundwaters are generally first encountered at depths typically of between 200 and 1000m, although in parts of the region are found closer to the ground surface. A plot of available wells for both Sweden and Finland is shown in Figure 8-15, for which data has been taken from published well-logs.

There are several species of saline and brackish groundwater that can be discriminated. Much of the saline water is pre-Holocene in age. In some areas the chemistry is typical of saline waters found in crystalline shields, although never reaching the brine concentrations encountered in the Canadian shield. Other waters may have formed as a result of chemical fractionation during the formation of a deep glacial permafrost. In lowland areas, below the highest shoreline of the 7,000 BP Litorina Sea, surface recharge from Baltic sea water is likely. Below this shoreline the lower the elevation the longer that a given region has spent submerged during the Holocene: the Litorina shoreline rises from about 30 m at Helsinki up to 100 m close to the modern centre of the rebound dome. However even below the Litorina shoreline no simple relationship is found between depth to the saline water interface and elevation (Lahermo and Lampen, 1987), and the fresh water layer varies from almost zero to several hundred metres, the average thickness being 20-60 m.

In Finland about half the deep boreholes for which detailed water sampling was undertaken are below the Litorina shoreline although this proportion is higher along the eastern coast of Sweden (see Figure 8-15). However, in both Sweden and Finland there is no simple relationship between the elevation of the land above present sea-level and the depth to saline water. In particular around the coast of the Gulf of Bothnia freshwater is commonly encountered to a depth of several hundred metres. Neglecting the lowland sites, submerged at some period since 7,000 BP, freshwater is typically found to depths in excess of 500 m. The chief exceptions to this are boreholes at Urjala in

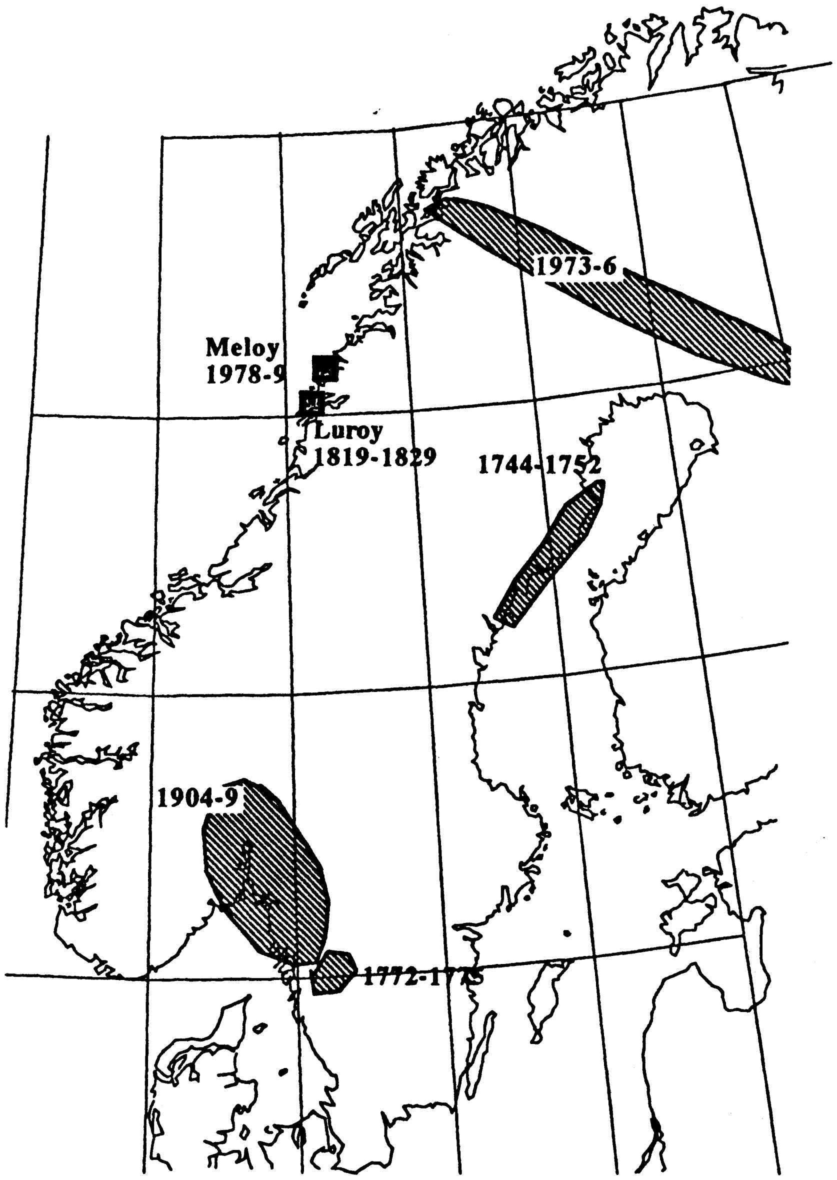


Figure 8-14: Seismic swarms (squares) and other regional spatio-temporal earthquake clusters.

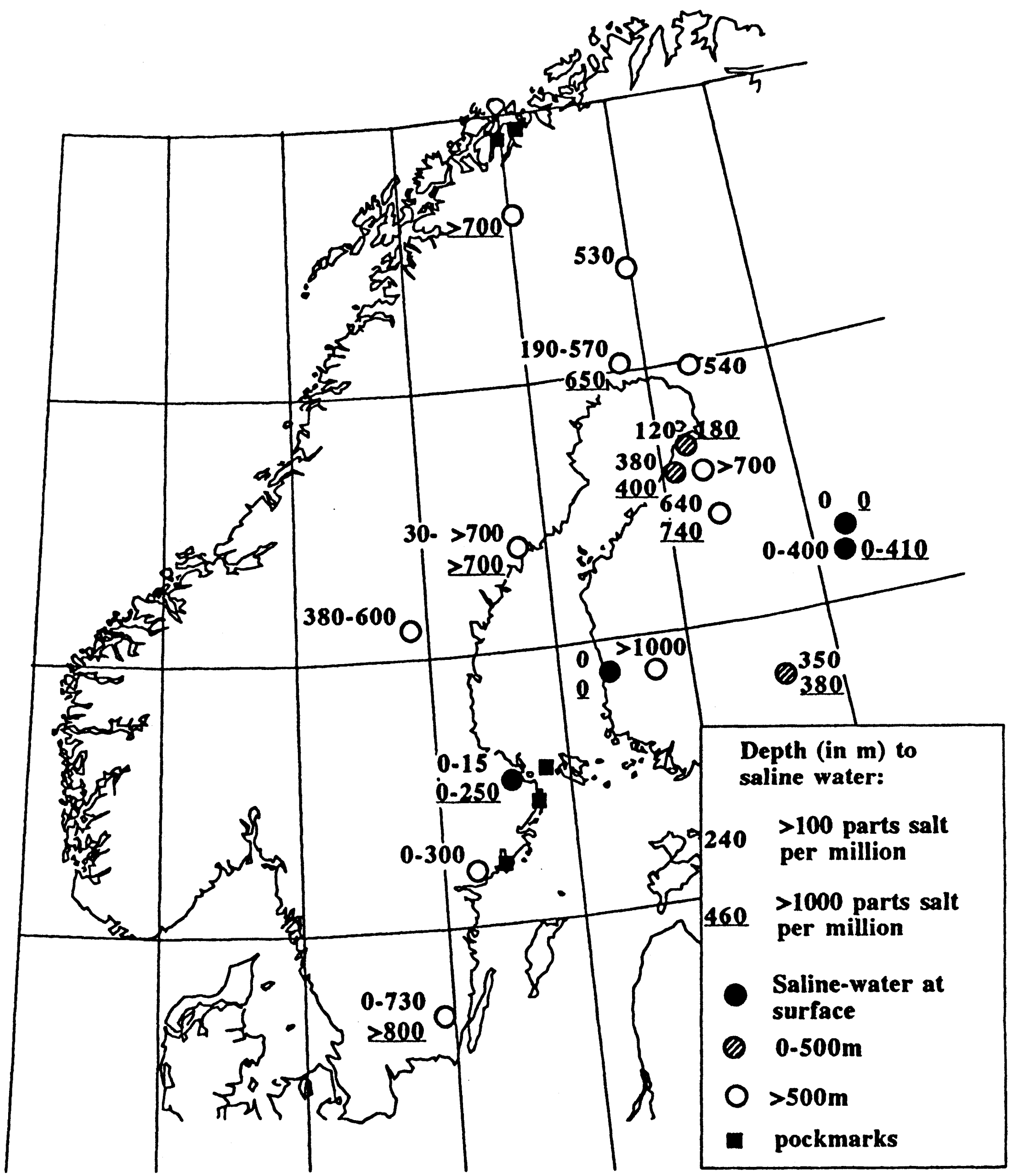


Figure 8.15 Depth to saline groundwater and location of pockmarks within the Fennoscandian rebound dome.

southern Finland and a number of sites in the Outukumpu sulphide ore belt in central eastern Finland where saline water has been found up to the surface.

If the depth to the interface with saline groundwater does show any relationship with the strainfield then it is to be anticipated that those areas subject to compressional strain, as a result of flexure within the rebound dome, should have a boundary between fresh and saline groundwater at far shallower levels than those areas known to be in extension. In Finland the geodetic studies of horizontal strain (see Figure 4-9) can be compared with these depths. It is to be noted that some correspondence appears between those areas where saline water is found at the surface (generally in southern Finland) with areas identified from the geodetic evidence as subject to compressional strain. All areas in which the boundary with saline water is located at depths of several hundred metres (chiefly in north-western Finland) appear geodetically to be in extension. The average depth of the freshwater in these latter regions is within the same order as the simple strain-estimate calculation presented above.

In Sweden there are no direct observations of the horizontal strainfield to be compared with the well data. Saline water has only been found up to the surface along the eastern coast of central and southern Sweden (in an area below the Litorina sea-level). However among these wells there are several in which, as at Finnsjon, water at relatively shallow depths is of higher salinity than the adjacent sea. Although the picture is confused as a result of submergence for some part of the Holocene, by analogy with Finland this area might be inferred as one suffering compressional strain.

8.5.3 Horizontal Strain and Pockmarks

Pockmarks are considered to reflect the expulsion of gas or water through superficial sediments (Hovland and Judd, 1988). If the crust is undergoing compressional strain, the reduction in the aperture of cracks is likely to expel water, and where these cracks are covered by sediments of relatively low permeability, raised fluid pressures may tend to be relieved through surface vents. Most of the pockmarks known from north-west Europe have developed in the Holocene in the forebulge regions of the Fennoscandian and Scottish ice-sheets, areas which are likely to have been subject to compressional strain during forebulge collapse. Such compressional strain may have squeezed out fluid, just as compressional strain in the aftermath of an extensional fault movement squeezed out groundwater (Muir Wood and King, 1993).

However pockmarks are also known from along the eastern coast of central Sweden within the area subject to rebound, and hence generally expected to be in extensional strain. The pockmarks so far identified are located in the Aland Sea, and from two coastal zones: about 50km to the north-east of Stockholm at Wettershaga and 50 km to the south of Stockholm at Sundsbadarna (Floden and Soderberg, 1988). Recurrent, sometimes explosive, eruptions of gas are claimed by Floden and Soderberg to continue in the bay close to Laparo Island, in the vicinity of the Wettershaga pockmarks.

Shallow reverse fault earthquakes also occur in this region of the central east coast of Sweden (see Wahlström, 1980 and Figure 8-7) including the shallow reverse fault earthquake at Bergshamra on 11th November 1979 (Wahlström, 1980). Hence pockmarks, continuing gas-bursts, surface saline water and shallow reverse fault earthquakes all occur in the same locality. It is possible that they are all a manifestation of compressional strain as a result of the existence of a local saddle in the rebound surface.

8.5.4 Compressional Strainfields within the Rebound Dome

There are a number of characteristics that can be predicted of regions at present subject to shallow compressional deformation as a result of the existence of saddles in the rebound surface. Shallow compressional stress relief phenomena ('pop-ups') are well known from around the border of eastern Canada with USA (Adams, 1989). One

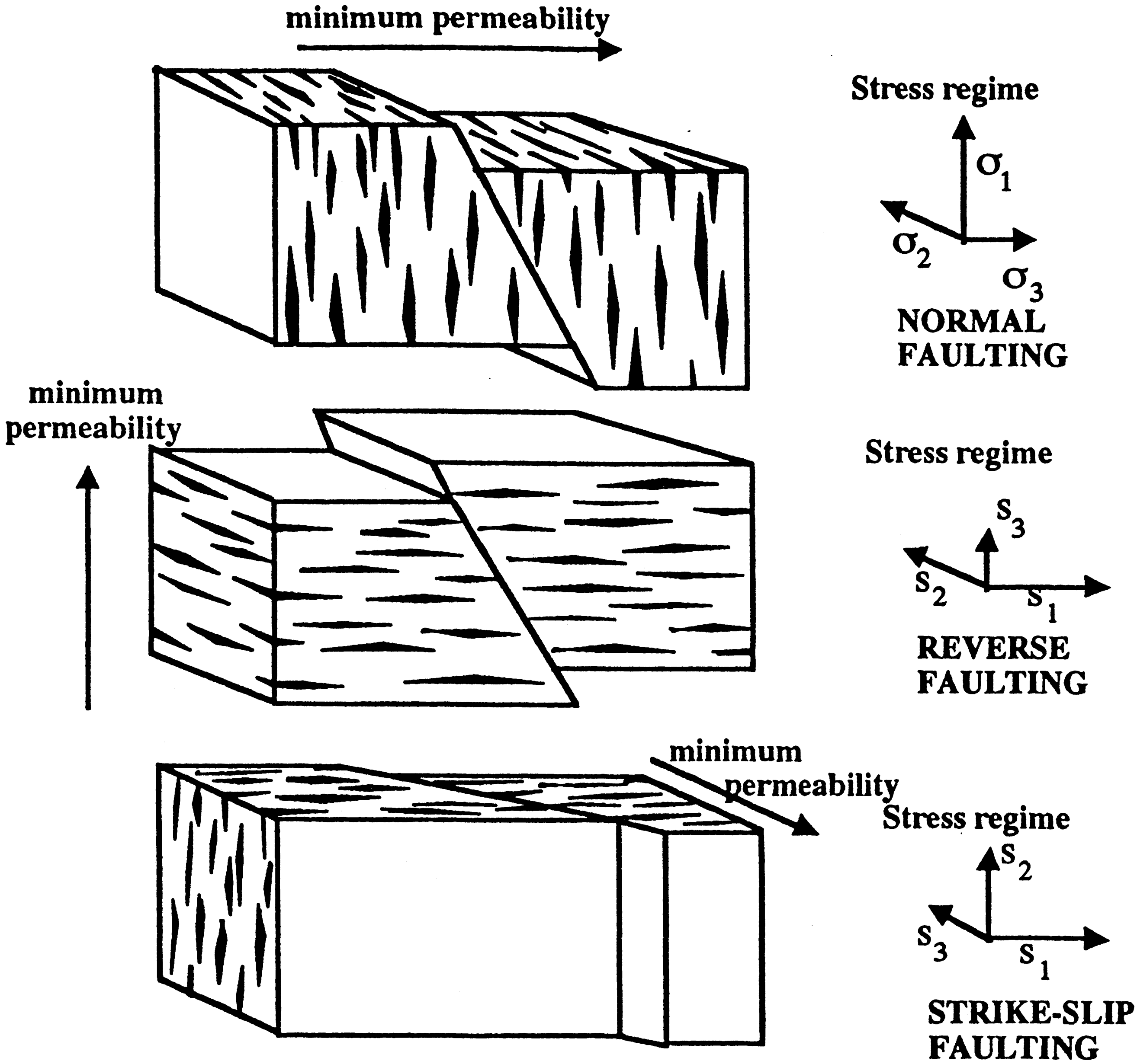


Figure 8-16 Varying orientations of major open fractures under different stress regimes, and their impact on permeability anisotropy.

manifestation of this stress-release is the reverse displacement of boreholes drilled through dip-slip faults in the excavation of road cuttings. The last of these shallow stress-release phenomena have only been identified in Fennoscandia in Lebesby, Norwegian Finnmark.

Permeability anisotropy within the shallow crust may also be an indicator of the stress-regime. Fractures are most likely to dilate perpendicular to the minimum principal stress direction (see Figure 8-16). In areas subject to active crustal compression dilatant fractures will tend to lie close to horizontal. It is to be noted that the Finnsjon site, located in the midst of the zone of shallow compressional deformation in central eastern Sweden, appears to be characterised by flow along low-angle fractures (Tiren, 1989). This contrasts with other regions of Sweden (in a strike-slip stress regime) in which flow is concentrated on high angle fractures.

Changes in crustal deformation that accompany flexures in the rebound surface will continue to affect the hydrogeology. As the rates of rebound change, so the stress regime will alter and consequently the aperture and even the orientation of fractures along which fluid flow is channelled. These local effects are superimposed on the major crustal strain changes that accompany loading and unloading which are likely to flush water into and out of the crust through the glaciation-deglaciation cycle, raising and lowering the saline-freshwater interface. The hydrogeology, just as the seismotectonics of Fennoscandia, has been profoundly dynamic through the Quaternary.

9 SEISMOTECTONICS OF THE GLACIATION / DEGLACIATION CYCLE

The seismotectonic character of Sweden is likely to have changed markedly through the glaciation/deglaciation cycle as a result of major changes in vertical load, fluid pressures and crustal strain. The discovery in northern Sweden and Finnish Lapland of a remarkable province of reverse faulting dating from the very end of deglaciation (Lagerbäck, 1979, Kujansuu, 1964, see Section 3.3) has revealed the degree to which there has been pronounced variation in regional seismic energy release and seismotectonic state through the glaciation/deglaciation cycle. However, the concentration of the mapped surface fault rupture in this area also suggests that this form of deglaciation seismotectonics was not uniformly distributed across Sweden. The question remains what kind of variation in the seismotectonic state and seismic energy release existed across other parts of the country and at other periods of the cycle.

9.1 ICE LOADING

The Weichselian ice-sheet is known to have been initiated from mountain glaciers, probably starting from around 80 ka (Lundquist, 1987). However after an Early Weichselian advance the ice-sheet retreated, and it was not until about 50 Ka that the main Fennoscandian ice-sheet began to accumulate. In northern Sweden a prolonged period of extreme cold and permafrost conditions preceded the accumulation of the latest Weichselian ice-sheet (Lagerbäck, 1988b), which appears to have been frozen to the rock throughout its existence over a large part of northeastern Norbotten. The formation of this thick permafrost prior to the arrival of an ice-sheet may have prevented hydraulic conductivity between fractures in the sub-surface rockmass and surface groundwater.

The ice-divide continued to migrate towards the east as the crest of the ice-sheet shifted from the mountains of Norway towards the trough of the Gulf of Bothnia. Southern Sweden is considered to have remained ice-free as late as 24,000 BP (Lundquist, 1983), but the ice subsequently advanced rapidly reaching the outskirts of Hamburg at 20,000 BP (Ehlers, 1981).

9.1.1 **Stress-State During Ice Loading**

The stress-state in Fennoscandia prior to the accumulation of the ice-sheet can only be inferred to be one in which the maximum horizontal stress was NW-SE and the minimum principal stress vertical (see Section 8.1.6). The imposition of an additional vertical load would have served to increase vertical stresses, far more than the horizontal stresses (that are likely to have shown a corresponding increase by about one third) reducing the deviatoric stress and hence taking the rock away from failure conditions (see Figure 9-1). Flow of mantle material from beneath the ice-load to its periphery would have caused an increase in stresses radial to the downwarped crust, as discussed in Section 8.1.6. However for the duration of the ice-sheet it is likely that the increase in vertical stress significantly exceeded any increase in horizontal stress and as a result the region covered by the ice-sheet was almost aseismic, consistent with the crust beneath the Greenland and Antarctic ice-sheets today (Johnston, 1987, 1989).

9.2 DEGLACIATION REBOUND

The glacial maximum was at around 20,000 BP. Deglaciation began in the Barents Sea, where rapid unloading associated with the floating off of the ice-cap, may have led to some resultant crustal deformation. The ice-sheet began to retreat from southern Sweden at around 13,000 yr BP. This retreat became interrupted during the Younger Dryas period from 11,000 to 10,500 yr BP when the ice-sheet underwent a readvance to the west. It is likely that there was some increase in the thickness of the western ice-sheet at this period (Anundsen, 1985). Within the overall course of deglaciation this readvance appears to have affected the pattern of rebound.

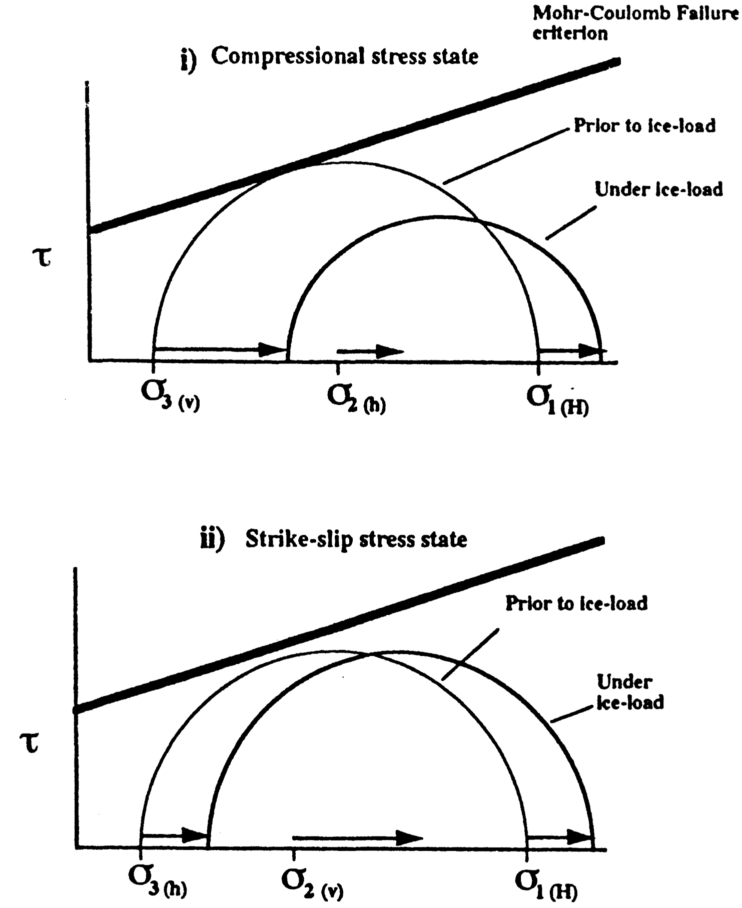


Figure 9.1: Mohr stress diagrams illustrating the impact of ice-loading on a crust i) in a compressional stress state, and ii) in a strike-slip stress state.

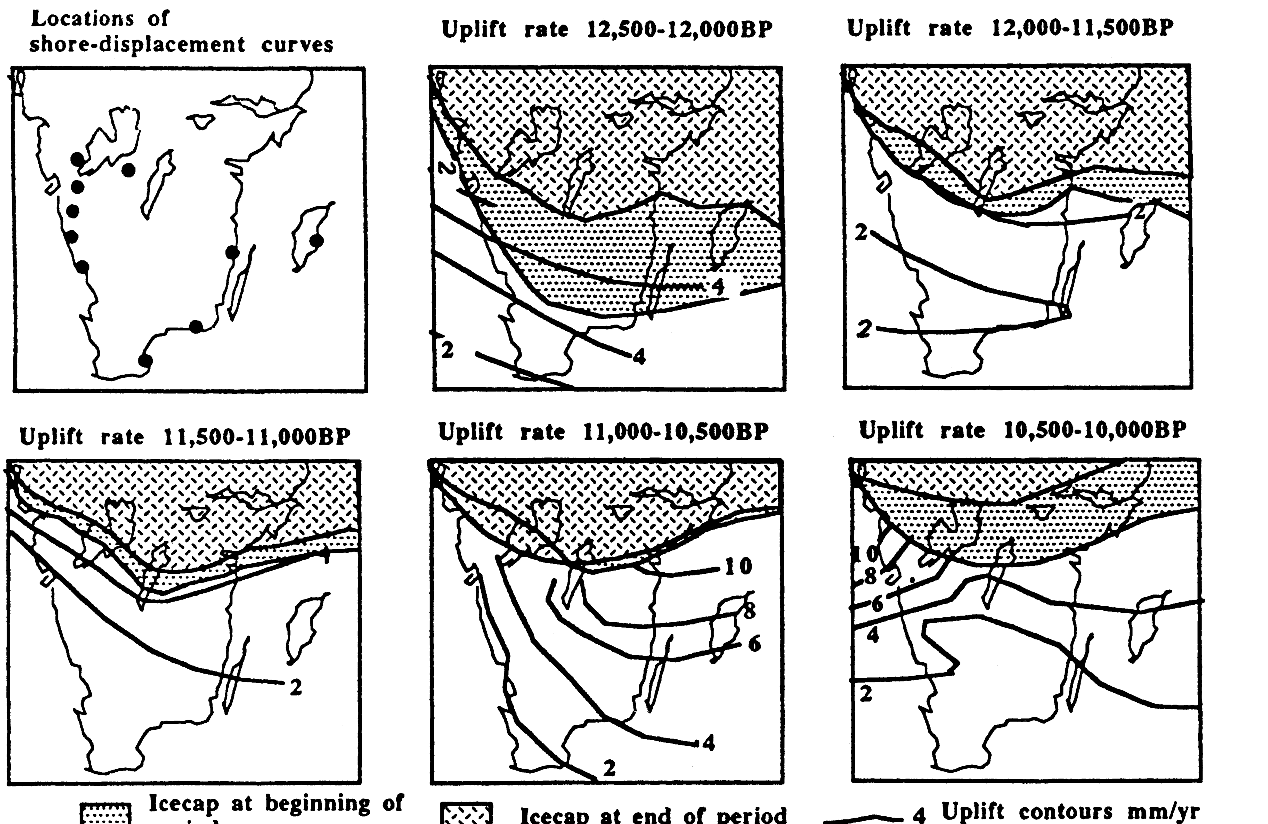


Figure 9.2: Averaged 'absolute' uplift rates in southern Sweden (in mm/yr) derived from shore displacement curves and compensated for relative eustatic sea-level, for 500 year time-periods during deglaciation.

9.2.1 Deglaciation Rebound in Southern Fennoscandia

Lake isolation studies around southern Sweden provide evidence for the change of relative sea-level through this period (Bjorck, 1989) (see Section 4.3). The shore displacement curves obtained from these studies have been compensated for the contemporary eustatic changes in sea-level from which it is possible to gain evidence of the rate of land elevation. This has been sampled over a series of 500 year time-periods in Figure 9-2 to show the average rate of land-uplift in mm/yr from 12,500 to 10,000 yr BP. These results indicate a change in the direction of tilt through time, in particular during and immediately following the Younger Dryas Readvance, as can be seen by a comparison of average uplift in the period 11,000-10,500 with that between 10,500-10,000 BP. There is also evidence for a rim of rebound around the margins of the ice-sheet for the period from 12,500-11,500 when the ice-sheet was undergoing rapid marginal retreat.

In a study of the elevation of the 10,300 BP Baltic Ice Lake shoreline around the southern Baltic, Svensson (1991) compared the elevation of the shoreline with the present uplift rate at the same location (see Section 4.3 and Figure 4.10). This elevation reflects the average rate of uplift through the period from 10,300 BP to the present. Svensson fitted a best-fit line through the data relating present elevation to current uplift-rate (see Figure 9-3), and found regionally consistent anomalies that were larger than any potential errors (see Figure 9-4). A NE-SW oriented zone passing from eastern Finland to Estonia showed a present-day uplift rate up to 1 mm/yr higher than that predicted from the best-fit line. This contrasts with the Lake Vattern area of southern Sweden where the present day uplift is more than 1 mm/yr lower than average (the maximum differences in elevation for any single uplift rate exceed 50 m, or around 20% of the absolute change in elevation since 10,300 BP).

These observations demonstrate that since 10,300 BP the Lake Vattern area has undergone more rapid average rebound relative to areas in eastern Finland. The simplest explanation for these

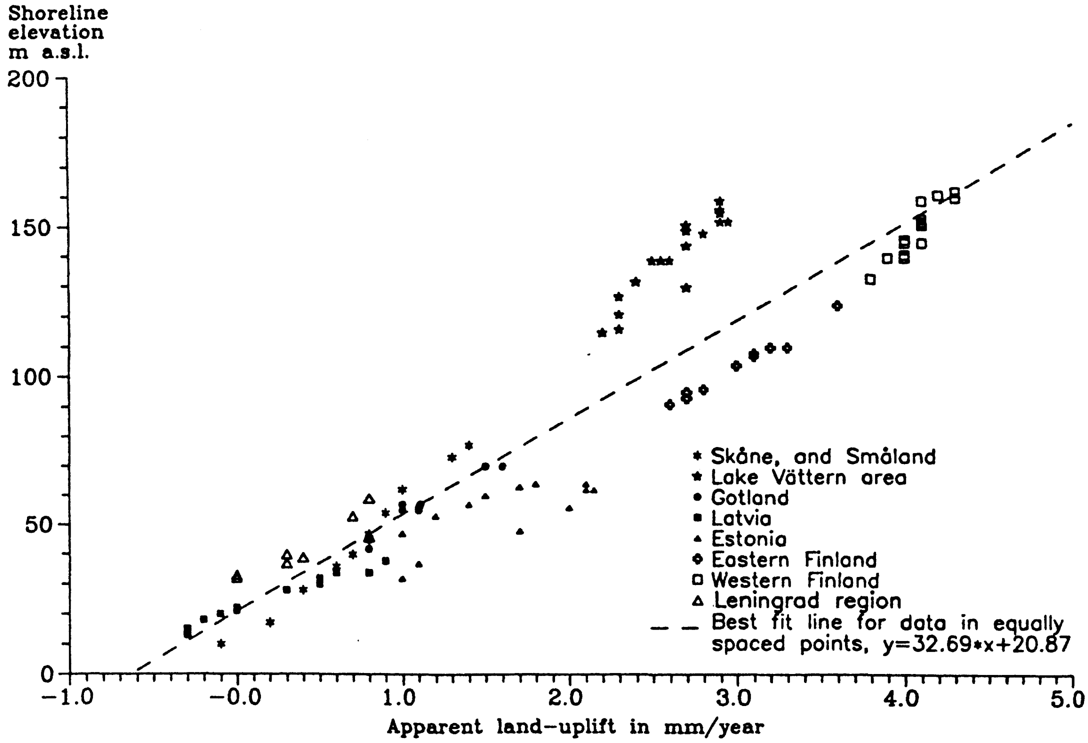


Figure 9-3 Rates of present-day uplift versus present elevation for the Ancylus Lake shoreline, southern Fennoscandia (Svensson, 1991).

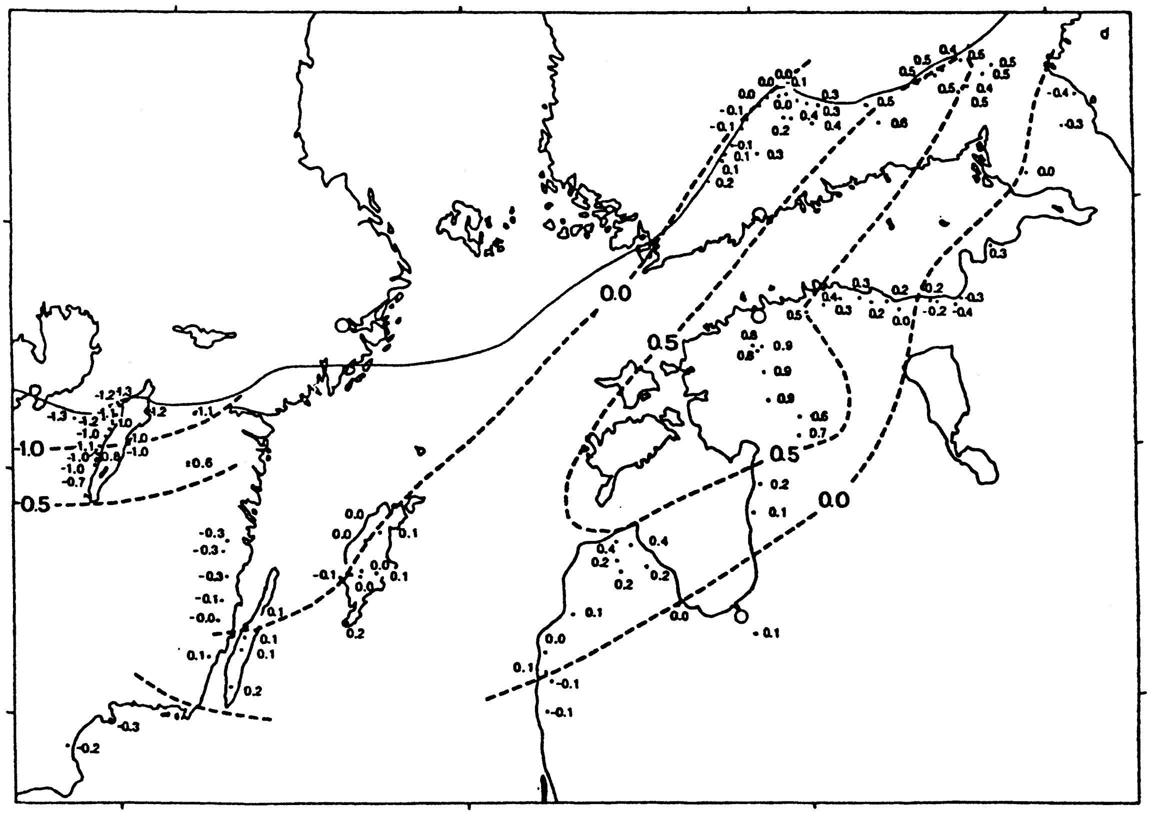


Figure 9-4 Regional anomalies of present uplift rates relative to the elevation of the Ancylus Lake shoreline (from Svensson, 1991).

regional variations is that more of the total rebound had already occurred prior to 10,300 BP in Estonia and Eastern Finland than had occurred in southern Sweden. Although there may be some association with lithospheric properties (the thicker >50 km crust of the Lake Vattern area having undergone a slower uplift prior to 10,300 BP, than that of the <40 km thick crust of Estonia, see Figure 2.2), the delayed rebound in southern Sweden probably reflects the more rapid unloading of the ice-cap to the east than to the west in particular during the Younger Dryas period (see above). Hence in the period before 10,300 BP it appears that an outer rim of rebound, oriented NE-SW, extended across Estonia into southern Finland. The relative uplift (in comparison with the areas to the northwest and southeast) within this rim would have to be about 2 mm/yr over 1000 years. The form and amplitude of this rim may have been a northeasterly continuation of the rim observed across southern Sweden between 12,500 and 11,500 BP (see Figure 9-2 above).

9.2.2 Deglaciation Seismicity in Southern Fennoscandia

A number of claims have been made as to the level of seismicity and faulting associated with deglaciation in southern Sweden (see for example Mörner 1978). From knowledge of the rates and gradients of the rebound surface it is possible to provide important constraints as to the form and magnitude of the crustal strain-field at this period and hence the likely levels of deformation and seismicity.

There are three distinct factors that contribute to seismic energy release on glacial unloading: (a) the release of tectonic strain stored during the existence of the ice-sheet; (b) the release of strain as a result of the outward flow of the upper mantle beneath the ice-sheet, and (c) strain reflecting crustal deformation in response to unloading.

(a) In southern Sweden any tectonic strain accumulated only over a period of about 10,000 years, rather than the 40,000 years over which the ice-sheet remained in place in northern Sweden. Employing the upper bound cratonic strain-rates discussed in Section 8.1.1, an annual strain of 10^{-11} for 10,000 years across a region 400 km across (NW-SE) is only 40 mm. This could be released by a series of earthquakes no larger than magnitude 5.

(b) As discussed in Section 8.1.6 and illustrated in Figure 8-2, as a result of the disposition of southern Sweden relative to the centre of the rebound dome, the effect of the outward movement of mantle material beneath the ice-sheet will have been to raise the minimum horizontal stress. This in turn will either have left the deviatoric stress unaffected (for a compressional tectonic stress-state) or reduced the deviatoric stress (for a strike-slip tectonic stress state).

(c) Rapid or differential unloading may have produced crustal deformation, and of all three factors involved in deglaciation this is the only process that appears likely to significantly raise seismicity levels in southern Sweden. At such time particular orientations of faults may have been reactivated according to the local strain field of rebound. For example when the western coast of southern Sweden appears to have undergone rapid rebound between 10,500 and 10,000 BP, a saddle zone of compression is found to the south of this area.

In Figure 9-5 some of the rebound profiles from southern Sweden, reconstructed from the lake isolation studies (see Figure 9-2), are compared with modern uplift profiles along the same traverse. From these it can be seen that uplift gradients were up to five times steeper than those prevailing today in the same region (but only about 1.5 times steeper than those found close to the coast of western Norway today, see Figure 4-6). The rate of angular strain across the rim of rebound, identified between 12,000 and 11,500 BP, was (averaged over 500 years) two to three times that found across the centre of the modern rebound dome (compare with Figure 4-5).

These comparisons could help inform an understanding of the level of seismicity consequent on this deformation, as well as the style and orientation of faults likely to have been reactivated at this period. From a preliminary inspection of the information on deformation rates it appears that seismicity levels may have been one or perhaps two orders of magnitude higher than today, but not the five or six orders of magnitude higher than today implied by the evidence for deglaciation seismicity in northern Sweden.

During the phase of rapid ice retreat, following the Younger Dryas, the Lake Vanern region became rapidly unloaded. Lake isolation studies to the north of Lake Vanern (Risberg and Sandgren. 1993, Axen, 1993), have indicated some possible complexity in the pattern of uplift immediately following deglaciation (ca 9500 BP). Lake isolation has not proved easy to date, as a result of low levels of organic content, and at a number of lakes younger dates are found in the lacustrine sediments than in the underlying marine deposits. Hence it is as yet not possible to resolve the significance of these apparent variations in relative elevation of isolation, with regard to any simple tectonic model, although as a result of the rapid unloading this region is an important one for indicating the potential for deglaciation tectonics in southern Sweden.

9.2.3 Deglaciation Rebound in Northern Fennoscandia

Lake isolation studies or shorelevel studies have not been undertaken in sufficient detail around the central Finland and Gulf of Bothnia region to determine the regional distribution of land-level changes and uplift rates from the period from 10,000 to 9,000 years BP. However it is known that there was a remarkable retreat of the ice-cap across Finland at this period, Lundquist, 1987 (see Figure 9-6). The thickness of the ice-sheet in northern Sweden is believed to have been at least

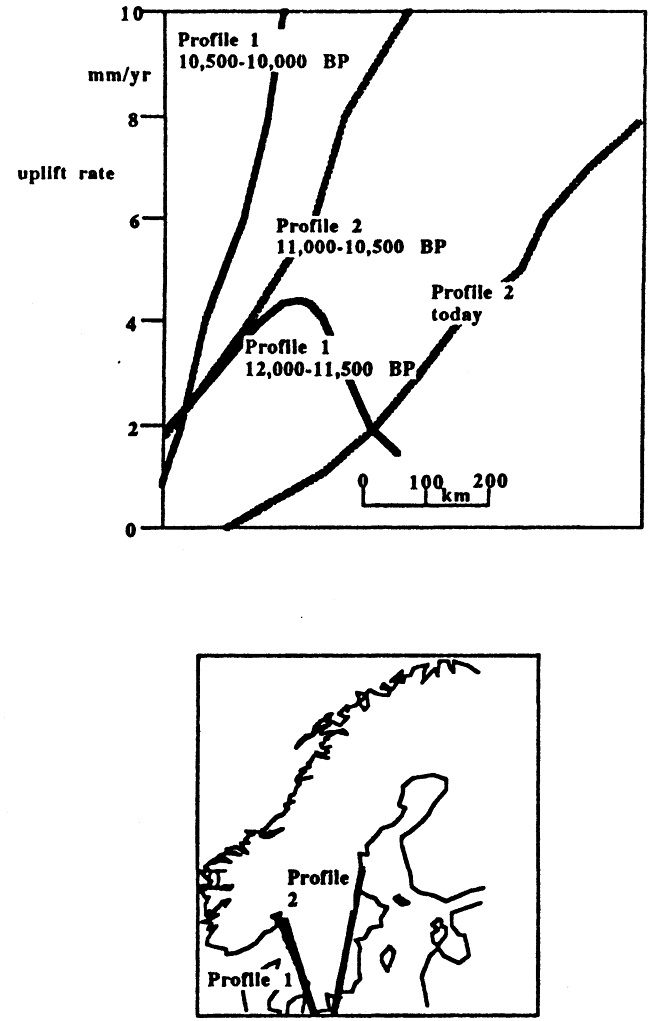


Figure 9-5: Comparison of deglaciation rebound profiles (adjusted for eustatic sea-level changes) in South Sweden with those prevailing today.

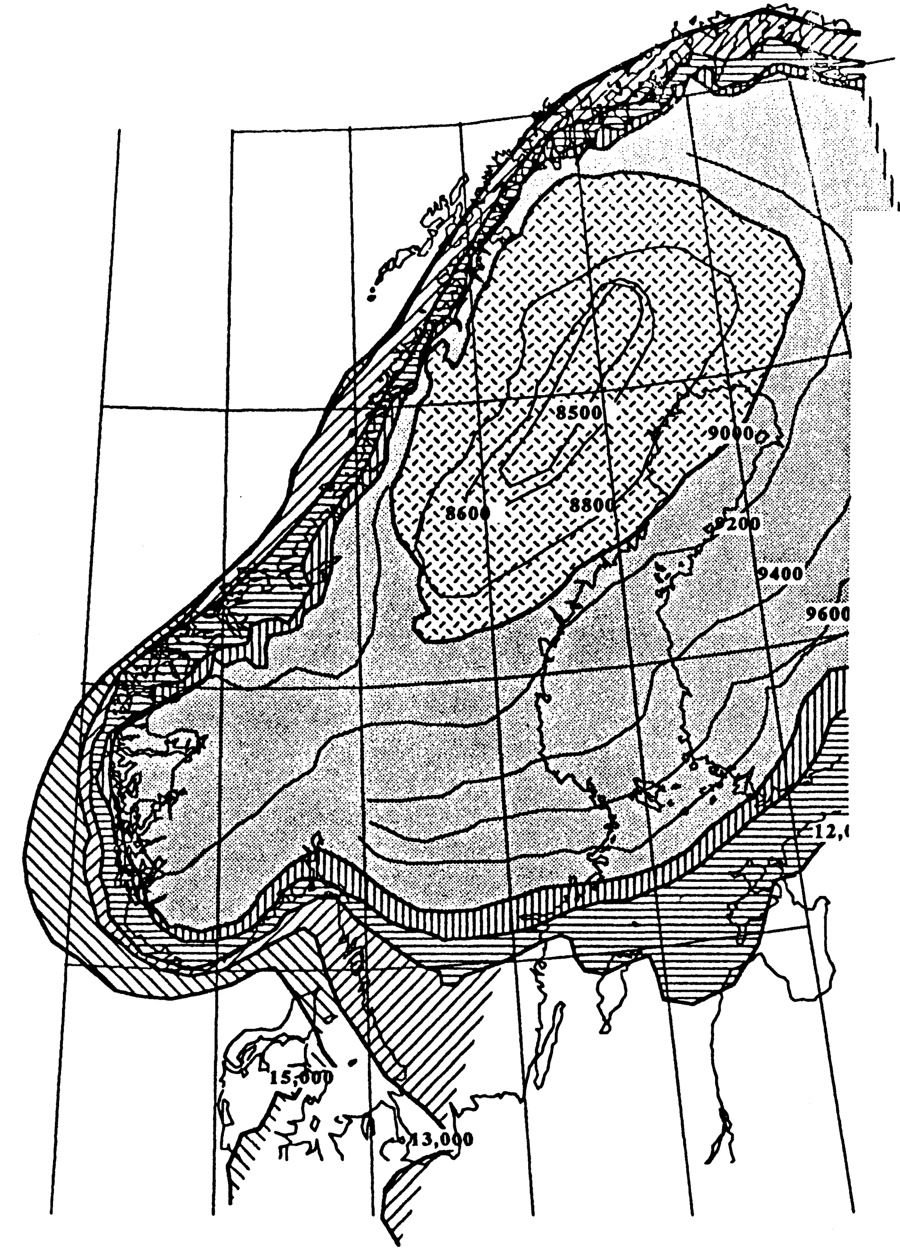


Figure 9-6: Deglaciation ice-front positions in years BP.

500-1000 m at the time that the southeasterly edge of the ice-sheet retreated across the Gulf of Bothnia.

As can be seen from Figure 9-6, during the course of the retreat across Finland between 10,000 and 9,000 BP, and in particular between 9,500 and 9,000 BP, the south-east ice-sheet margin retreated 300 km (at rates approaching 1 km/yr) while over the same period the retreat was only about 20 km to the northwest (Lundquist, 1987). The ice-cap margin in much of western Finland was consistently below the level of the contemporary sea. In the direction of retreat across Finland the topography fell towards the Gulf of Bothnia. Within the gulf the water depth at this time was in excess of 200 m over large areas, capable of floating off a thick ice-shelf. Rebound rates are known to have been very high immediately following ice unloading overtaking any eustatic rise in sea-level. Hence the marine limit in the northern Gulf of Bothnia consistently records deglaciation.

Through this period unloading became increasingly asymmetric. The pattern of modern rebound reveals that the centre of the ice-load was approximately distributed along the western margin of the Gulf of Bothnia. By 9,000 BP all the remaining ice-load was distributed to the northwest of this axis. Hence the crust of Finland was rising rapidly relative to the crust of northern Sweden.

9.2.4 Deglaciation Tectonics in Northern Fennoscandia

The rate of ice-cap retreat into northern Sweden must have slowed once the ice-cap margin was above the contemporary sea-level. Uplift was now occurring faster to the east than to the west, where the ice-cap still remained. It was exactly at this time, that the first of the faults of the Lapland Post-Glacial Fault Province ruptured.

As discussed in Section 3.3.2 these fault-ruptures can be divided into three groups. The eastern group of lowland fault-scarps, all lie around the 200 m contour, close to the marine limit (which falls from more than 200 m in southern Norbotten to around 170 m around the Finnish Border with Sweden, Fromm, 1965), and includes the two faults in Finland that lie orthogonal to the main NNE-SSW strike (see Figure 3-3). Excavations across and in the vicinity of the Lanjsarv Fault have shown that the fault moved approximately at the time the marine limit was being formed (Lagerbäck, 1990), probably within decades or at most a century of local ice-melting. A number of the landslides in the vicinity of the Lansjarv Fault moved over or alongside stagnant ice (Kujansuu, 1972, Lagerbäck, 1990).

All these faults broke in response to flexure exceeding the shear-strength of the faulted crust. This flexure was the result of differential rebound. The rim of ice-margin rebound seen across Estonia and eastern Finland before 10,300 BP (Figure 9-4) may have continued to migrate towards the north-west across Finland during the ice-cap retreat, before crossing the centre of ice-loading at around 9,000 BP. However the concentration of flexure at the marine limit reflects rapid glacial unloading followed by rapid uplift of the lowlands to the east. The crust thins from Finland into northern Sweden (see Figure 2-2), making the lithosphere stronger and more prone to brittle rupture in place of flexure.

The exact faults that moved, their length etc was determined by available fractures in the crust. In some cases movement ruptured a long fault, in others a series of individual shorter fault segments (high displacement to length ratios, see Figure 3-8, could suggest that a few of the faults were either very strong or newly created at depth). All these lowland faults probably moved within a few decades of deglaciation. The movement on one fault would have concentrated strain at the end of the fault so encouraging the rupture on a neighbouring fault. The whole set of these lowland faults may have been triggered within a century.

Although significantly higher than the marine limit, the Lainio Suijavaara Fault is located in a region where the marine limit passes deep inland along the valleys of the Tornea and Muinio rivers, and also where the lowland fault group is itself relatively incoherent. This central fault group may be an inland infilling of the dis-placement represented by the lowland fault group.

Above the marine limit the ice front is likely to have continued to retreat towards the north west, although the exact pattern of deglaciation in this region is largely inferred. The advancing rim of rebound followed the retreating ice-sheet and the faults are likely to have continued to move in ascending order of elevation. In each case (with the exception of the two orthogonal faults at the northern end of the lowland fault group) crustal flexure involved the southeasterly lowland side rising more rapidly than the northwest upland side, reflecting ice-retreat. For the Highland Fault Group an additional topographic factor also came into play. Along the eastern edge of the Caledonides in northern Sweden the average elevation falls by more than 500 m over a distance of 50 km (Muir Wood, 1989). Across this topographic front the variation in the original ice-load also contributed to differential rebound during glacial unloading (see Figure 9-7).

It is not impossible that more than one movement occurred on the individual faults at this period, as continued flexural strain across the region of the fault might continue to be relieved on the same zone of weakness. One major movement might have been followed by a series of relatively minor movements on different sections of each fault system. However no conclusive evidence has been found to indicate more than a single episode of displacement. This flexure was itself migrating at a speed comparable to that of the retreating ice-cap, and once the fault no longer lay within the flexure it is likely to have ceased activity.

9.2.5 Deglaciation Seismicity in Northern Fennoscandia

The size of the earthquakes associated with these fault ruptures has been estimated from the dimensions and displacements of the surface faults (see Section 3.3.9). The cumulative seismic energy release of all these faults is estimated to be about 5×10^{21} Nm (equivalent to one earthquake M8.4). The seismicity that accompanied deglaciation is likely to have lasted at least several centuries, but probably less than 1000 years.

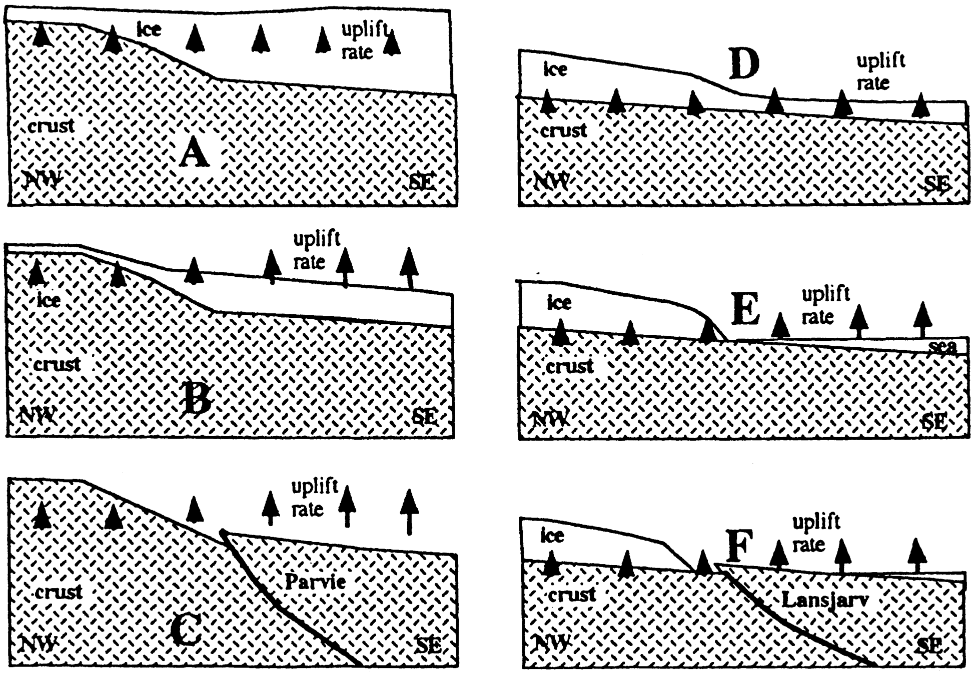


Figure 9-7: Models for the generation of the Parvie and Lansjarv Fault displacements.

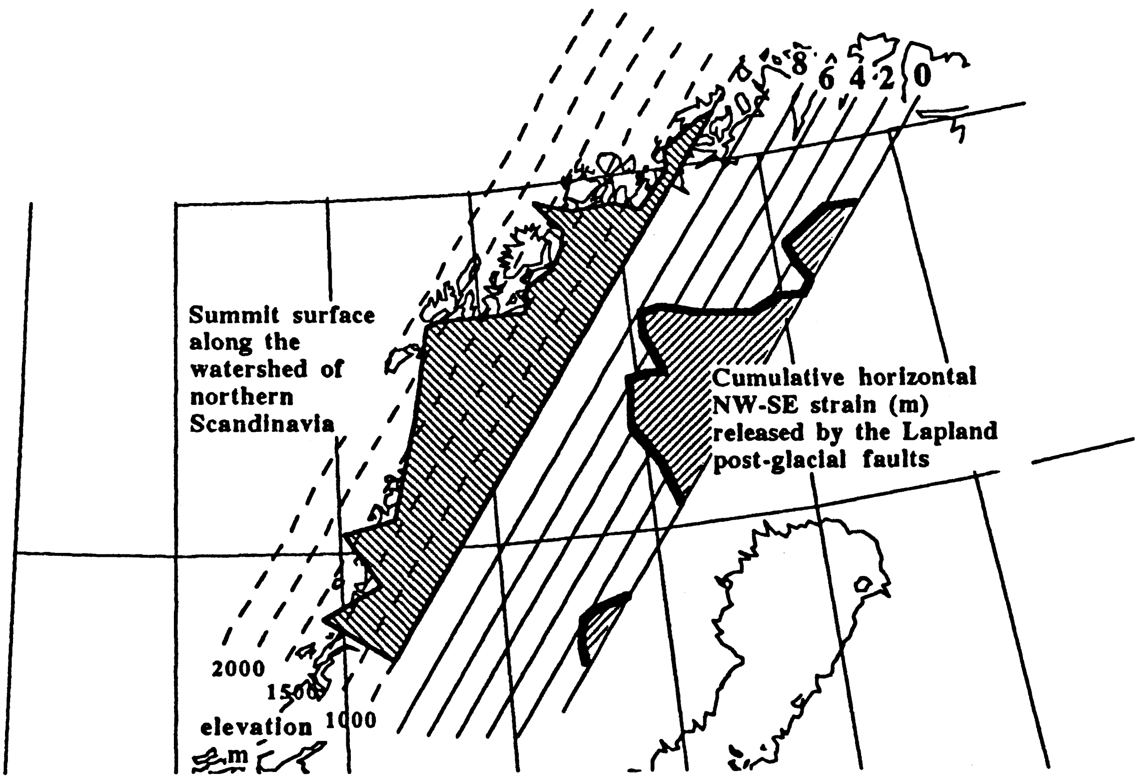


Figure 9-8: Comparison between cumulative horizontal NW-SE strain represented by the Lapland post-glacial reverse fault scarps, with the elevation of the summit surface along the watershed of northern Scandinavia.

Taking 500 years as a representative time-period indicates an annual seismic energy release of 10^{19} Nm. This is 100,000 to 1,000,000 times higher than this same region today (employing local magnitude moment conversions derived by Kim et al, 1989) and 1,000 to 10,000 times higher than for an equivalent area of the currently active Norwegian continental shelf. The Lapland post-glacial fault province covers an area of approximately 100,000 km^2, within which the areally adjusted seismic energy release was 10^{14} Nm per year per km^2. This is a level consistent with a very active continental plate boundary, higher than currently prevails in Iran and almost comparable with that of California.

9.2.6 **Differential Vertical Uplift**

The elastic rebound model of fault rupture indicates that the crust was being flexured prior to faulting. This differential rebound is known to be critically related to the period when the ice-cap margin was retreating across the region. The vertical displacement of the Lansjarv faults is typically 4-8 m. If the fault concentrated the flexure over a width of 50 km, this would be equivalent to a rise of 20 mm/yr in the uplift rate over this distance of 50 km for a duration of 200-400 years. This would imply a rebound gradient almost twenty times the highest gradients known today (around the Norwegian coast). It would be of great interest to see if evidence could be found in support of such pronounced differential rebound at this period.

9.2.7 **The Horizontal Component of Deformation**

The displacement of the deglaciation faults reflects a significant horizontal component of crustal shortening. By making some assumptions about fault-dip (modelled as 60 degrees) and average displacement along the individual faults it is possible to find the overall horizontal shortening. This is a maximum of about 10 m, more typically 6 m (see Figure 9-8). In effect Lulea became 10 m closer to Narvik as a result of this displacement (although subsequent horizontal extension is likely to have increased this separation once again). Along a 300km NW-SE traverse through the centre of the province this reflects a strain of 3×10^{-5}. Where did this strain come from?

There are two potential causes of this horizontal shortening: either tectonics or glacial loading.

(a) Tectonics

The tectonic cause (generally ascribed to ridge push - see Section 8.1.1) would be some compressional strain accumulated during the whole duration of the existence of the ice-cap and subsequently relieved at the time of deglaciation (see Muir Wood, 1989). However the accumulation of 10 m of crustal shortening in 40,000 years, would produce almost 400 m of horizontal shortening through the Quaternary, and imply a remarkable tectonic strain rate of around 10^{-8} per year, three orders of magnitude higher than the upper bound estimate for the tectonic strain of this craton discussed in Section 8.1.1. Such strain levels would be manifest in a series of major reverse faults, and fault bounded topography. A phase of similar reverse faulting would be expected at the end of each glaciation, and it is likely that the same faults would be repeatedly reactivated, as in any other area of active tectonics. Hence the post-glacial fault-scarps would be expected to have formed on the site of multiple phases of Quaternary reverse displacement. There is however little evidence for any earlier Quaternary displacement on these structures, and no evidence to support high levels of Quaternary tectonic strain. While some tectonic strain across this region during the duration of the ice-sheet is unquestioned this is expected to produce more nearly 100mm of crustal shortening; only 1% of the 10 m estimated.

The chief argument that has been used in support of a tectonic explanation in the past comes from the regionally consistent orientation of most of the faults, approximately orthogonal to the spreading direction in the Norwegian Sea. However the connection may not be so simple. As a result of the opening of the Norwegian Sea the continental margins became uplifted, roughly parallel with the spreading axis. The topography of these margins controlled the build up of the ice-cap and also controlled the subsequent deglaciation. The horizontal crustal shortening indicated by the faults shows some approximate similarity with the summit surface along the watershed (see Figure 9-8).

(b) Loading strain

As discussed in Section 8.1.7 strain-release following deglaciation will tend to be concentrated on the flanks of the rebound dome facing the principal horizontal compressive direction, and can be predicted to involve horizontal shortening along structures orthogonal with this direction. The movement of mantle material out from under the load towards its margins creates a radial compressional strainfield across the bowl, that for the Fennoscandian ice-sheet has a magnitude modelled as 10^{-4} (see Figure 8-1; Gasperini et al, 1991). If the load is removed more rapidly than the mantle material can stream back beneath the downwarped bowl, the reduction in vertical stress should allow some component of this horizontal strain to be relieved through faulting.

The importance of the strain-field is demonstrated by the existence of two faults orthogonal to the general trend, at the northern end of the group of lowland faults (see Figure 3-3, Kujansuu, 1964). The orientation of these high angle reverse faults is inconsistent with a motivation from ridge push. However if radial strain is sufficiently high the minimum and maximum horizontal stress orientations may switch. In common with all the other Lowland Faults, these faults involved the uplift of the submerged lowland relative to the inland areas. However in contrast to the other faults of this group on the northern fringes of the expanded Gulf of Bothnia the marine limit ran more nearly E-W than N-S. The flexure that preceded faulting therefore involved the relative uplift of the southern block.

While it is impossible to assign a fault displacement to a single cause it appears that conceivably 99% of the crustal shortening represented by the Lapland postglacial faults reflects strain associated with crustal downwarping, rather than tectonics.

9.3 POST-DEGLACIATION SEISMOTECTONICS

Subsequent to the final ice-cap melting, at least for the Lowland Group of PGFs, the cause of compressional flexure had entirely disappeared, and subsequently this region towards the centre of the rebound dome is likely to have been in extension. Hence in any region in which there was stress-release as a result of permanent strain accompanying reverse fault displacement the NW-SE horizontal stress would not have accumulated again. This is confirmed by observations of the stress-field around the Lansjarv Fault that show a NE-SW orientation of the principal horizontal compressive stress and very low horizontal stress magnitudes (see Section 6.3.3). Studies of shear-wave splitting seen in recordings of local earthquakes at Lansjarv indicate that the maximum horizontal stress at this locality is oriented NE-SW. It would be of interest to know to what distance from the fault stress-release (and consequently coseismic strain) could still be detected.

However the influence of topography on glacial unloading that was important for the development of the highland group of the Lapland postglacial faults could potentially have continued to have an influence on the local pattern of rebound for some time thereafter, with a rapid decrease in rebound rates in passing to the west. This would have led to further minor reverse fault displacements on the faults within the Highland Group.

Following the final melting of the ice-cap between 8500 and 8,000 BP it is likely that northern Sweden rose very rapidly. This may have created a saddle in the rebound surface with the remainder of southern Fennoscandia rising more slowly. Such a saddle would have created conditions of shallow crustal compression as are found on the area around the Canada - New York State Border today (Adams, 1989). A saddle in the rebound surface, leading to shallow compressional stress relief could conceivably have some bearing on the generation of boulder caves at around 8,000 BP (see Section 5.4.4).

Elsewhere in Sweden for the past 8,000 years rates of crustal deformation associated with rebound (and by inference seismicity levels) have probably been close to those prevailing today.

9.4 PREVIOUS GLACIATIONS

The record of glaciation and deglaciation is inevitably far less well established for earlier glacial episodes. The Saalian ice advance is however known to have penetrated deeper into Holland, Germany and Poland than the Weichselian (see Ehlers, 1990). Three separate ice advances have been recognised in the Saalian, the ice-maximum remaining in place for a long duration estimated as from 160 to 130 ka (Forsstrom et al, 1988).

The decay of the Saalian ice-sheet was also more rapid, without the delays and minor readvance that characterised the Weichselian deglaciation. Evidence from deep sea cores indicates that the switch from the Saalian glacial to the Eemian interglacial was very rapid and led to the worldwide melting of glaciers within a period of about 2,000 years, although the sea-level continued to rise at a slower pace for another 4,500 years. One result of this rapid warming may have been melting of the ice-sheet to the west in south Norway and SW Sweden, and the stagnation of ice-masses in other areas (Ehlers et al, 1984). This would have had important implications for the pattern of crustal deformation and sea-level rise that accompanied deglaciation.

At a few sites around the Baltic, marine sediments have been found deposited during the early stages of the Eemian inter-glacial. These deposits are found at elevations up to 115 m in western Finland and in southeastern Finland the Eemian seas reached higher levels than those found during the Flandrian, establishing a sea interconnection between the Baltic and the White Sea via the Lake Onega basin (Forsstrom et al, 1987). However in Estonia the Eemian shallow marine interglacial sequence is found at a depth of 61-75 m below sea-level.

As the ice melted more rapidly than at the end of the Weichselian, there was a different pattern of resulting uplift during deglaciation, with a higher marine limit around the Gulf of Bothnia, and a less orderly ice-sheet retreat. The larger Saalian ice-sheet had a centre of mass located further to the south-east than that of the Weichselian. Hence NNE-SSW trending reverse faulting might be predicted to have formed on the western flank of the downwarped crust, perhaps in eastern Finland. The seismotectonic consequences of differences in deglaciation of the Saalian and earlier Quaternary glaciations remain to be explored.

10 CONCLUSIONS

10.1 CURRENT SEISMOTECTONICS

The directly observed and modelled strainfield of the Fennoscandian rebound dome and the diversity of focal mechanisms strongly suggests that all current seismicity in Sweden is a response to deglaciation.

No evidence has been found for any late Neogene tectonic activity in mainland Sweden in the absence of the Ice Ages. There may however be some active tectonics along the Fennoscandian Border Zone and along the Lofoten and More continental margins offshore Norway. However the levels of strain associated with this tectonic activity are one or more orders of magnitude lower than the strainfield associated with postglacial rebound.

As revealed by the detailed geodetic information on land-uplift and the focal mechanisms of earthquakes a signficant proportion of the Fennoscandian rebound dome is undergoing mild extension, relieving the high horizontal stresses that prevailed as a result of crustal downwarping beneath the ice-sheet. However there may also be 'saddles' of shallow compression within the rebound dome. Seismological, geodetic and hydrogeological information suggest that one such saddle area exists along the central coast of eastern Sweden.

Within the rebound dome there is some consistency between those areas of flexure identified from geodetic data and the concentrations of seismicity. Flexures probably tend to be localised where there is a transition in rheology (most critically crustal thickness) but also may reflect topography (that affected the original ice-load), see Figure 10-1.

Shallow compressional stress-relief phenomena (pop-ups, borehole displacements across reverse faults exposed in road cuttings) are unknown from Scandinavia except in Finnmark. These phenomena are very common along the border area of southern Canada and northern New England, USA. This reflects the fact the in contrast with Canada the rebound profile for Fennoscandia is almost everywhere extensional (areas of increasing

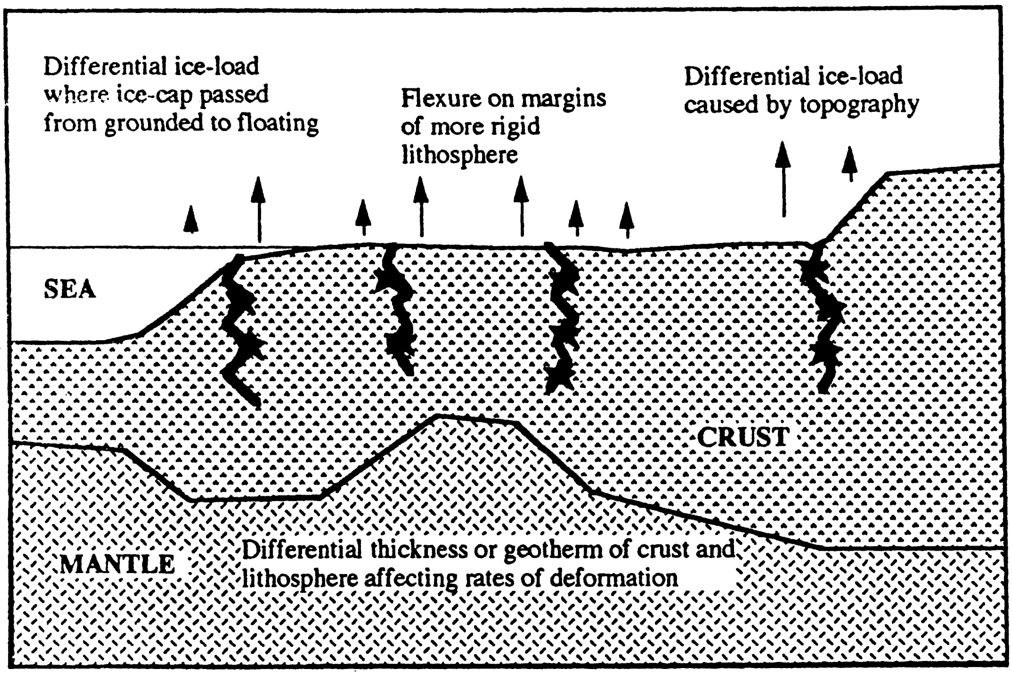


Figure 10-1: The impact of topography and lithospheric rheology on the localisation of deformation during postglacial rebound.

compressional strain are chiefly located offshore to the west of Scandinavia). However it is possible that shallow compressional stress relief phenomena were prevalent in central Sweden as a result of compressional strain reflecting a concave rebound profile soon after 9000 BP when the ice had finally left northern Sweden.

Some geodetic observations suggest locations in Scandinavia where strain appears to be localised along faults or some other narrow zones. The movements along these features appear to be episodic, have maximum averaged rates of around 1 mm/yr and discontinuous motions. Faults may even move in different directions at different periods. These faults appear to accommodate some of the overall strain, move aseismically and reflect the fact that mild extension is destressing faults of the appropriate orientation.

Some of the largest earthquakes occur within the former forebulge offshore to the west of Scandinavia. Within this region horizontal stresses continue to be raised as a result of forebulge collapse, in contrast to the majority of the rebound dome in which stresses are radially relieved.

Maximum earthquake depths in Sweden show a good correlation with the temperature gradient. In areas of low thermal gradient as in the centre of the Baltic Shield earthquakes occur down to the base of the crust.

The association of seismicity with rebound suggests that within the rebound dome the present level of seismicity will continue to slowly decay (over the next few thousand years) as the rate of rebound declines. Maximum magnitudes of historical earthquakes show some general correlation with the form of deformation across the rebound dome. In particular the largest earthquakes continue to occur in thinner warmer crust around the margins of the dome.

10.2 DEGLACIATION SEISMOTECTONICS

The current stress-field is returning to the state that prevailed prior to ice-loading everywhere except in regions in which permanent strain has occurred during the glaciation/deglaciation cycle. As stress-changes that accompany glacial loading and unloading are radial to the centre of the ice-sheet it is only on the northwest and southeast flanks of the downwarped crustal that these affect the NW-SE maximum horizontal stress state. On glacial unloading it is in these areas that permanent strain, through major faulting, is predicted. The current stress-state should provide a good indicator as to those regions that have been subject to permanent strain.

Southern Sweden lies on the south-west flank of the centre of the ice-load in which radial strain affected the minimum rather than the maximum horizontal stress. Hence there is no reason for any significant permanent strain as a result of major fault movements during deglaciation. This is confirmed both by the absence of surface fault scarps of the type identified in northern Sweden and also by the stress-field that suggests little variation relative to the tectonic stress-field that prevailed prior to glacial loading.

In southern Sweden the form of the strainfield accompanying deglaciation can be mapped in some detail from dating studies of relative sea-level changes during lake isolation. In southernmost Sweden this strainfield may have been no more than an order of magnitude more intense than that prevailing today.

Although currently the largest earthquakes are restricted to the margins of the shield this was not the case during deglaciation, when the largest earthquakes occurred towards the centre of the rebound dome, in the cold thick shield crust.

The Lapland post-glacial faults were triggered by rapid differential asymmetric unloading at the centre of the downwarped crust as a result of the melting of the ice-cap across Finland between 10,000 and 9,000 BP. These faults may have ruptured down to the base of the crust. The largest of the accompanying earthquakes probably had a magnitude somewhat in excess of M8.

The approximately 10m horizontal shortening released by these faults dominantly reflects the strain of downwarping imposed on the crust. These faults are located in a region in which the crust is now in mild extension and hence have no potential for renewed reverse displacement.

10.3 HYDROGEOLOGICAL IMPACT

The extensional strain that accompanied the rebound of Fennoscandia causes an increase in crustal porosity. Above the level of the sea this will have drawn freshwater into the crust to depths typically of several hundred metres. The greatest porosity increases are likely to be found where the highest rates of extension have occurred in the aftermath of a reverse fault earthquake in which the surrounding crust relaxes in extension.

Where there is a saddle in the rebound surface, the crust is subject to shallow compression leading to a decrease in crustal porosity with the principal control of fluid movement along shallow-dipping fractures. In such areas pre-Holocene saline groundwater is likely to be squeezed close to the surface. These conditions appear to prevail along the central eastern coast of Sweden.

During the course of glacial loading, the water absorbed into the crust during rebound extension will be squeezed out again. This flushing process in the upper crust has a tidal character, ultimately controlled by the Milankovich cyclicity, that determines ice ages.

Hydrogeological changes both as a result of gradual and sudden (earthquake related) strain may affect Swedish bedrock radioactive waste repositories. In particular:

(a) outbursts of groundwater that accompany certain styles of earthquakes;

(b) upwelling of ground water from shallow compression; and

(c) changes in the overall location of the boundary between pristine and recent recharge water in the crust, as a result of the crustal strain.

11 REFERENCES

Adams J. (1989) Postglacial faulting in eastern Canada: nature, origin and seismic hazard implications, Tectonophys., 163, pp 323-331.

Adams J., Wetmiller E.J., Hasegawa H.S. and Drysdale J. (1991) The first surface faulting from a historical earthquake in North America, Nature, 352, pp 617-619.

Agrell H. (1981) Sillberga gryt - en sentida sprickgrotta i Uppland, Grottan 4, pp 28-29.

Ahjos T., Saari J., Penttila E. and Korhonen H. (1984) Earthquakes and seismic hazard in Finland, Engineering Geology 20, pp 1-12.

Ahjos T. and Uski M. (1991) Earthquakes in northern Europe in 1375-1989. Submitted to Tectonophysics.

Ahlin S. (1982) NNE-SSW striking Precambrian faults of south-western Sweden. G.F.F., 104, pp 182-185.

Ahlin S. (1987) Phanerozoic faults in the Vastergotlanbd basin area, SW Sweden, G.F.F., 109, pp 221-227.

Aki K. (1965) Maximum likelihood estimate of b in the formula log N = a - bM and its confidence limits. Bull Earthq Res Inst, 43, 237-239.

Aki K. and Richards P. (1980) Quantitative Seismology, Volume I. Freeman, San Francisco, 557 pp.

Alsaker A. Kvamme L.B., Hansen R. A., Dahle A. and Bungum H. (1991) The M_L scale in Norway. Bull Seismol Soc Am, 81, 379-398.

Ambraseys N.N. (1985a) Magnitude assessment of northwestern European earthquakes, J. Earthq. Eng. Struct. Dyn., 13, pp 307-320.

Ambraseys N.N. (1985b) The seismicity of Western Scandinavia, J. Earthq. Eng. Struct. Dyn., 13, pp 361-399.

Andersen O. B., Kejlso E., and Remmer O. (1974) Secular movements within Jutland as determined from repeated precise levellings 1885-1894 and 1943-53, Geodaetisk Instituts Skrifter 3, 3rd Series, Vol 40.

Anundsen K. (1982) Undersokelser Vedrorende Mulige Kvartaere forkastninger i Sorvest-Norge.

Anundsen K. (1985) Changes in shore-level and ice-front position iin Late Weichsel and Holocene, southern Norway. Norsk Geografisk Tidsskrift 349., pp 205-225.

Anundsen K. (1989) Late Weichselian relative sea-levels in southwest Norway: observed strandline tilts and neotectonic activity, G.F.F., 111, pp 288-292

Arvidsson R. (1991) Seismodynamics of Sweden deduced from earthquake-source and seismic-wave characteristics, Doctoral dissertation, Uppsala University.

Arvidsson R. and Kulhanek O. (1990) Major earthquakes in Sweden, their generation and focal mechanisms. XXII General Assembly ESC Barcelona, 17-22 September 1990, Programme and Abstracts, 171 (abstract).

Arvidsson R., Gregersen S., Kulhanek O. and Wahlström R. (1991) Recent Kattegat earthquakes - evidence of active intraplate tectonics in southern Scandinavia. Phys Earth Planet Int, 67, 275-287.

Arvidsson R., Wahlström R. and Kulhanek O. (1992) Deep-crustal earthquakes in the southern Baltic Shield. Geophys. J Int, 108, pp767-777.

Axberg S. (1980) Seismic stratigraphy and bedrock geology of the Bothnian Sea. Stockholm Contributions to Geology.

Axen E. (1993) Stratigraphical investigation of two small lakes in south-western Varmland, western Sweden, indicating early Holocene irregular isostatic uplift, Swedish Nuclear Power Inspectorate (SKI/SKB) Technical Report, 24p.

Babuska V., Plomerova J. and Sileny J. (1987) Structural model of the subcrustal lithosphere in central Europe. Composition, Structure and Dynamics of the Lithosphere. Asthenosphere System, edited by K. Fuchs and C. Froidevaux, Geodynamics Series, Vol 16.

Babuska V., Plomerova J. and Pajdusak P. (1988) Seismologically determined deep lithosphere structure in Fennoscandia, GFF 110, pp 380-382.

Bakkelid C. (1986) The determination of rates of land uplift in Norway, Tectonophys. 130, pp 307-326.

Bakkelid S. and Skjothaug P. (1985) Aktive geologiske forkastininge i Eigersund - forlopige resultatater (unpublished report).

Balling N. (1990) Heat flow and lithospheric temperature along the northern segment of the European Geotraverse, in Proceedings of the Sixth Workshop on the European Geotraverse (EGT) Project: data compilations and synoptic interpretation, 1989, Einsielden, pp 405-416 eds. Freeman R. and Mueller St., European Science Foundation, Strasbourg.

Bannister S.C., Ruud B.O. and Husebye E.S. (1991) Tomographic estimates of sub-Moho seismic velocities in Fennoscandia and structural implications, Tectonophys., 189, pp 37-53.

Basham P.W., Forsyth D.A. and Wetmiller R.J. (1977) The seismicity of northern Canada, Can. J. Earth Sci., 14, pp 1646-1667.

Båth M. (1956) An earthquake catalogue for Fenoscandia for the years 1891-1950, Sveriges Geol. Unders., C545, pp 1-52.

Båth M. (1972) Zum Studium der Seismizitat von Fennoskandia, Gerlands Beitr. Geophys. 81, pp213-226.

Båth M. (1978) Energy and tectonics of Fennoscandian earthquakes, Tectonophys. 50, ppT9-T17.

Båth M., Kulhanek O., van Eck T. and Wahlström R. (1976) Engineering analysis of ground motion in Sweden. Seismological Institute, Uppsala University, Report 5-76, 59 pp.

Becker J-M, Lithen T. and Nordquist A. (1988) Experience of motorized trigonometric levelling (MTL) - comparison with other techniques, National Land Survey, Professional Papers, 1988:23, Gavle.

Berberian M. (1979) Earthquake faulting and bedding thrusts associated with the Tabas-e-Golshan (Iran) earthquake of September 16, 1978, Bull. Seism. Soc. Am., 69, pp 1861-1887.

Bergsten K.E. (1943) En senglacial forkastning i norra Ostergotland. Sven Geogr. Arsb., 1943: pp 1-16.

Berthelsen A. (1987) A tectonic model for the crustal evolution of the Baltic Shield, pp 31-57 in J-P Schaer and J.R . Rodgers (eds) The Anatomy of Mountain Ranges, Princeton Univ. Press, Princeton. N.J.

Berthelsen A. and Marker M. (1986) 1.9-1.8 Ga strike-slip megashears in the Baltic Shield and their tectonic implications, Tectonophys. 128, pp 163-181.

Bjarnason B. and Stephansson O. (1987) Non-linear and dis- continuous stress variation with depth in the upper crust of the Baltic Shield. Proc. 6th Congress Int. Soc. for Rock Mech., Montreal, Canada, pp 809-812.

Bjarnason B., Ljunggren C. and Stephansson O. (1989a) New developments in hydrofracturing stress measurements at Lulea University of Technology. Int. J. Rock Mech. Min. Sci. & Geomech. Abstr., 26, 579-586.

Bjarnason B., Zellman O. and Wikberg P. (1989b) Drilling and borehole description. In Interdisciplinary study of post-glacial faulting in the Lansjarv area Northern Sweden (Edited by G. Backblom and R. Stanfors) Technical Report 89-31, pp 7:1-7:14. Swedish Nuclear Fuel and Waste Management Co., Stockholm.

Bjorck S. (1989) The geodynamic and geologic value of a regional network of shore displacement curves: examples from sdouthern Sweden, G.F.F., 111, pp 292-295.

Bjorck S. and Digerfeldt G. (1982) Late Weichselian shore displacement at Hunneberg, southern Sweden, indicating complex uplict, G.F.F., 104, pp 132-155.

Bjorkman H. and Tragardh J. (1982) Differential uplift in Blekinge indicating late-glacial neotectonics, G.F.F., 104, pp 75-79.

Blomquist R.G. (1990) Deep groundwaters in the crystalline basement of Finland, with implications for nuclear waste disposal studies, G.F.F., 112, pp 369-374.

Boe R., Sorensen S. and Hovland M. (1992) The Karmsundet Basin, SW Norway: stratigraphy. structure and neotectonic activity, Norsk Geol. Tidds., 72, pp281-283.

Bonilla M.G., Mark R.K. and Lienkaemper J.J. (1984) Statistical relations among earthquake magnitude, surface rupture length and surface fault displacement. Bull. Seismo. Soc. Am., 74, pp 2379-2411.

Brown E.T. and Windsor C.R. (1990) Near surface in situ stresses in Australia and their influence on underground construction. The Institution of Engineers Australia, Tunnelling Conference, Sydney 11-13 Sept. 1990, pp 18-48.

Brudy M. (1990) Modellierung des Europaeischen Spannungsfelds unter besonderer berucksichtigung neuer daten aus Nordwesteuropa. Diploma thesis, Univ. Karlsruhe.

Bungum H. and Husebye E.S. (1974) Analysis of the operational capabilities for detection and location of seismic events at Norsar. Bull Seismol Soc Am, 64, 637-656.

Bungum H., Hokland B.K., Husebye E.S. and Ringdal F. (1979) An exceptional intraplate earthquake sequence in Meloy, northern Norway, Nature 280, pp 32-35.

Bungum H., Havskov J., Hokland B.Kr. and Newmark R. (1986) Contemporary seismicity of northwest Europe. Annales Geophys, 4, B, 5, 567-576.

Bungum H., Alsaker A., Kvamme L.B. and Hansen R.A. (1991) Seismicity and seismotectonics of Norway and nearby continental shelf areas, J. Geophys. Res., 96, pp 2249-2266.

Cermak V. (1989) Crustal heat production and mantle heat flow in Central and Eastern Europe, pp 195-215 in Uyeda S., Chapman D. and Zwart H.J. eds., Thermal Aspects of Tectonics, Magmatism and Metamorphism, Tectonophys., 159, pp 195-215.

Cermak V. (reporter) (1990) Heat flow along the EGT: an activity report. In 'The European Geotraverse: Integrative Studies', Results from the Fifth Study Centre, Eds. Freeman R., Giese P. and Mueller St., European Science Foundation, Strasbourg, pp 387-399.

Chen R. (1991) On the horizontal crustal deformation in Finland, Reports of the Finnish Geodetic Institute 91.1. 98pp.

Chen W-P. and Molnar P. (1984) Focal depths of intracontinental and intraplate earthquakes and their implications for the thermal and mechanical properties of the lithosphere, J. Geophys. Res., 88, pp 4183-4214.

Clauss B., Marquart G. and Fuchs K. (1989) Stress orientations in the North Sea and Fennoscandia, a comparison to the western European stress field. In Earthquakes at North-Atlantic Passive Margins: Neotectonics and Postglacial Rebound, edited by S. Gregersen and P. Basham, Nato ASI Series C, Kluwer Academic Publishers, pp 277-289.

Davenport C.A., Ringrose P.S., Becker A., Hancock P and Fenton C. (1989) Geological investigations of late and post glacial earthquake activity in Scotland, pp 175-194 in S. Gregersen and P. W. Basham eds., Earthquakes at North-Atlantic Passive Margins: Neotectonics and Postglacial Rebound, Kluwer, Dordrecht.

De Geer G. (1938) Jordbavningar i Bromma. Sven Turistforen., 1938, 24pp.

De Geer G. (1940) Geologia Suecia Principes. K. Sven Vetenskapsakad Handl., 3, pp 1-360.

Denham D. (1988) Australian seismicity - the puzzle of the not-so-stable continent, Seism. Res. Lett., 59, pp 235-240.

Doe T.W., Ingervald K., Strindell L., Leijon B., Hustrulid W., Mayer E. and Carlsson H. (1983) In situ stress measurements at Stripa mine, Sweden. Report LBL-15009, SAC44. Lawrence Berkeley Laboratory, Berkeley.

Doe T.W. and Korbin G.E. (1987) A comparison of hydraulic fracturing and hydraulic jacking stress measurements. Proc. 28th U.S. Symp. on Rock Mechanics, Tuscon, 29 June-1 July.

Donner J. (1980) The determination and dating of synchronous late Quaternary shorelines in Fennoscandia, pp 285-293 in Mörner N-A. ed. Earth Rheology, Isostasy and Eustasy, Wiley, Chichester.

Drury M.J. (1984) Perturbations to temperature gradients by water flow in crystalline rock formations, Tectonophys., 103., pp 19-32.

Eddington P.K., Smith R.B. and Renggli C. (1987) Kinematics of Basin and Range intraplate extension, pp 371-392 in Coward M.P., Dewey J.F. and Hancock P.L. eds., Continental Extensional Tectonics, Geol. Soc. Lond, Special Publications, 28.

Ehlers J. (1981) Some aspects of glacial erosion and deposition in North Germany, Annals of Glaciology, 2, pp 143-6.

Ehlers J. (1990) Reconstructing the dynamics of the North-West European Pleistocene ice sheets, Quat. Sci. Rev., 9, pp 71-83.

Ehlers J., Meyer K-D, and Stephan H-J. (1984) The Pre-Weichselian glaciations of North-West Europe, Quat. Sci. Rev., 3., pp 1-40.

Ekman M. (1976) Undersokning av Lutningsandringar i Jordskorpan vid Forkastningslinjer i Sodra Sverige, Statens landmateriverk, Gavle.

Ekman M. (1977) Berakning av geoidens rorelse i samband med den postglaciala landhojningen i Fennoscandia, Profess. Papers 1977/5, National Land Survey, Gavle.

Ekman M. (1986a) Apparent land uplift at 20 sea level stations in Sweden 1895-1984, Lantmatereriet, Tekniska skrifter, 1986:6, Gavle.

Ekman M. (1986b) A reinvestigation of the world's second longest series of sea level observations: Stockholm 1774-1984, National Land Survey, Professional Papers, 1986:4, 7pp Gavle

Ekman M. (1991a) A concise history of postglacial land uplift research (from its beginning to 1950), TERRA Nova, 3, pp 358-365.

Ekman (1991b) Gravity change, geoid change and remaining postglacial uplift of Fennscandia, TERRA Nova, 3, pp 390-392.

Ekman M., Eliassson L., Pettersson L. and Sjöberg L.E. (1982) Bestamning av landhojningen i Sverige med geodetiska metoder, National Land Survey, Professional Papers 1982:13, Gavle.

Eliasson T., Lindblom U., Slunga R., Sundquist U. and Wallroth T. (1988) The Swedish Hot-Dry-Rock Project; some preliminary achievements, in Boden A and Eriksson K.G. eds., Exploration of Deep Continental Crust, Vol2, Deep Drilling in Crystalline Bedrock, Springer-Verlag, Berlin.

Emery K.O. and Aubrey D.G. (1991) Sea Levels, Land levels, and Tide Gauges, Springer Verlag, New York 237p.

Engell-Sörensen L. and Havskov J. (1987) Recent North Sea seismicity studies. Phys Earth Planet Inter, 45, 37-44.

Fernlund J.M.R. (1988) The deglaciation sea level and the marine limit in the Lahom-Halmstad area, Sweden, G.F.F., 110, pp 207-220.

Feyling-Hanssen R.W. (1966) Geologiske observasjoner i Sandnes-ọmradet, Norges Geol. Unders. 242, pp 26-43.

Floden T. (1980) Seismic stratigraphy and bedrock geology of the central Baltic, Stockholkm. Contrib. Geol. 35., pp 1-240.

Floden T. and Soderberg P. (1988) Pockmarks and related seabed structures in some areas of Precambrian bedrock in Sweden, Geol. Surv. of Finland, Special Paper 6, pp 163-169.

Forsstrom L., Aalto M., Eronen M. and Gronlund R. (1988) Stratigraphic evidence for Eemian crustal movements and relative sea-level changes in Eastern Fennoscandia, Palaeogeog. Palaeoclim. Palaeoecol. 68, pp 317-335.

Fromm E. (1965) Beskrivning till jordartskarta over Norbotten lan nedanfor Lappmarksgransen. Sveriges. Geologiska Undersokning Ca 39, pp 1-236.

Gasperini P., Sabadini R. and Yuen D.A. (1991) Deep continental roots: the effcts of lateral variations of viscosity on post-glacial rebound, pp 21-32 in R. Sabadini er al., (eds). Glacial Isostasy,. Sea-Level and Mantle Rheology, Kluwer, Dordrecht.

Gorbatschev R. (1980) The Precambrian development of southern Sweden. G.F.F., 102, pp 129-136.

Gorbatschev R. and Gaal .G. (1987) The Precambrian history of the Baltic Shield. American Geophysical Union Geodynamics Series, Vol. 17. pp. 149-159.

Grunthal G. and Stromeyer D. (1992) The recent crustal stress field in central Europe - trajectories and finite-element modeling. J. Geophys. Res. (in press).

Gregersen S. and Basham P.W. (1989) Earthquakes at North Atlantic Passive Margins: Neotectonics and Postglacial Rebound, 1-6.

Gregersen S., Korhonen H. and Husebye E.S. (1991) Fennoscandian dynamics: Present-day earthquake activity, Tectonphys. 189, pp 333-344.

Gronlie O.T. (1922) Strandlinjer, moraener og skjaelforekomster i den sydlige del av Troms fylke, Norge Geol. Unders., 94, pp 1-39.

Gumaelius O. (1883) Samling af underrattelser om jordstotar i Sverige, G.F.F., 3, pp 341-6.

Gutenberg B. and Richter C.F. (1942) Earthquake magnitude, intensity, energy and acceleration, Bull. Seism. Soc. Am., 32, pp 163-191

Hallbjorn L. (1986) Rock stress measurements performed by Swedish State Power Board. In Proc. Int. Symp. on Rock Stress and Rock Stess Measurements, Stockholm (Edited by O. Stephansson), pp 197-206, Centek Publishers, Lulea.

Hamar G.P. and Hjelle K. (1984) Tectonic framework of the More basin and the northern North Sea, pp349-358 in Proceedings of the North European Margin Symposium, Trondheim 1983, (eds. Spencer A.M. et al.,) Graham and Trotman, London.

Hansen J.M. (1980) Laesos postglacial udvikling i relation til den Fennoskandiske Randzone, Dansk geol. Foren. Arsskrift for 1979, pp 23-30.

Harper J.F. (1990) Plate dynamics: Carribean map corrections and hotspot push. Geophys J. Int., 100, pp 423-431.

Hast N. (1958) The measurement of rock pressure in mines. Swedish Geological Survey, Ser. C. No. 560, 183 p.

Hiltscher R. (1969) Beitrag zur Gebirgsspannungsmessung nach dem Bohrlochboden-Entspannungsverfahren. Proc. Symposium on Determination of Stresses in Rock Masses, Lisboa, pp 245-264.

Hiltscher R., Martna J. and Stindell L. (1979) The measurement of triaxial rock stresses in deep boreholes. Proc. 4th Int. Congr. on Rock Mechanics, Montreux, Vol. 2, pp 227-234.

Hovland M. and Judd A.G. (1988) Seabed Pockmarks and Seepages: Impact on Geology, Biology and the Marine Environment, Graham, and Trotman, London 293pp.

Hubbert M.K. and Willis D.G. (1957) Mechanics of hydraulic fracturing. AIME Trans., Vol 210, pp 153-168.

James T.S. and Morgan W.J. (1990) Horizontal motions due to post-glacial rebound, Geophys. Res. Lett., 17, pp 957-960.

Johnsen E. (1981) Kvartaere trekk fra Beiarns geologiske historie, pp93-106 in Arbok for Beiarn, Utgitt av Beiarn historielag, publ. Egil Trohaugs Forlag A.S., Bodo.

Johnston A.C. (1987) Suppression of earthquakes by large continental ice sheets, Nature 330, pp 467-9.

Johnston A. C. (1989) The effect of large ice sheets on earthquake genesis, pp581-599 in S. Gregersen and P. W. Basham eds., Earthquakes at North-Atlantic Passive Margins: Neotectonics and Postglacial Rebound, Kluwer, Dordrecht.

Kaarianen E. (1975) Land uplift in Finland on the basis of sea level recordings, reports of the Finnish Geodetic Insititute 75:5, 14pp.

Kakkuri J. and Vermeer M. (1985) The study of land uplift using the third precise levelling of Finland. Rep. Finn. Geod. Inst. No 85:1, Helsinki.

Kasahara K. (1981) Earthquake Mechanics, Cambridge University Press, Cambridge.

Kim W-Y, Kulhanek O., van Eck T. and Wahlström R. (1985) The Solberg, Sweden, earthquake of September 29, 1983. Seismological Department, Uppsala University, Report 1-85, 49 pp.

Kim W-Y, Skordas E., Zhou Y-P and Kulhanek O. (1988) Source parameters of major earthquakes near Kiruna, northern Sweden, deduced from synthetic seismogram computation. SKB TR 88-23, 45 pp, Stockholm.

Kim W-Y., Wahlström R. and Uski M. (1989) Regional spectral scaling relations of source parameters for earthquakes in the Baltic Shield, Tectonophys., 166, pp 151-161.

Kinck J.J., Husebye E.S. and Lund C-E. (1991) The South Scandinavian crust: structural complexities from seismic reflection and refraction profiling, Tectonophys, 189, pp 117-133.

Kjellen R. (1910) Sveriges jordskalf. Forsok till en seismisk landsgeografi. Goteborgs Hogskolas Arsskrift 15:2, pp 1-211.

Klasson H. and Leijon B. (1990) Rock stress measurements in the deep boreholes at Kuhmo, Hyrynsalmi, Sievi, Eurajoki and Konginkangas. Technical Report YJT-90-18, Nuclear Waste Commision of Finnish Power Companies. 122 pp.

Kremenetsky A.A., Milanovsky S. Yu. and Ovchinnikov L.N.N. (1989) A heat generation model for continental crust based on deep drilling in the Baltic Shield, Tectonophys., 159., pp 231-246.

Kuivamaki A. (1986) Observations of the Venjarvi and Ruostejarvi postglacial faults, Geol. Surv. Finland Nuclear Waste Disposal Research, Report YST-52, 20p.

Kuivamaki A., Paananen M. and Kukkonon I. (1988) The Pasmajarvi Fault - A reactivated pre-existing fracture zone, Abstract in 18th Nordic Geological Winter Meeting, Copenhagen, Jan. 1988.

Kujansuu R. (1964) Nuorista siiroksista Lapisaa (Recent Faults in Finland) Geologi, 16, pp 30-36.

Kujansuu R. (1972) On landslides in Finnish Lapland, Geological Survey of Finland, Bulletin 256, pp 1-22.

Kukkonen I.T. (1988) Terrestial heat flow and groundwater circulation in the bedrock in the central Baltic shield, Tectonophys., 156, pp 59-74.

Kukkonen I.T. (1989) terrestial heat flow and radiogenic heat production in Finland, the central Baltic Shield., Tectonophys., 164., pp 219-230.

Kukkonen I.T. (1992) Aspects of deep geothermal modelling in the central Fennoscandian shield, Tectonophys., in press.

Kukkonen I and Kuivamaki A. (1985) Geological and geophysical observations of the Pasm_ajarvi and Suasselka postglacial faults, Geol. Surv. Finland, Nuclear Waste Disposal Research, Report YST-46, 14p.

Kulhanek O. and Wahlström R. (1981) Macroseismic observations in Sweden 1977-1979. Swedish Geological Survey, Ser C, 776, 13 pp.

Kulhanek O. and Wahlström R. (1985) Macroseismic observations in Sweden 1980-1983. Swedish Geological Survey, Ser C, 808, 28 pp.

Kulhanek O. and Wahlström R. (1992) Macroseismic observations in Sweden 1984-1990, Swedish Geological Survey, Ser C, 825 38p.

Kulhanek O., van Eck T., John N., Meyer K. and Wahlström R. (1981) Source parameters and seismotectonic implications of the recent Bergshamra, Sweden, earthquake. Earth Evolution Sciences, 1, 66-70.

Kulhanek O., van Eck T., John N., Meyer K. and Wahlström R. (1983) Spectra of the earthquake sequence February-March, 1981, in south-central Sweden, Tectonophys, 93, 337-350.

Kumpas M.G. (1984) Seismic interpretation of the Tornquist Zone in Denmark and Sweden, G.F.F., 106, pp 388-389.

Kusznir N.J. and Park R.G. (1987) The extensional strength of the continental lithosphere: its dependence on geothermal gradient, and crustal composition and thickness. Editors Coward M.P., Dewey J.F. and Hancock P.L., Continental Extensional Tectonics, Geological Society Special Publication No. 28, pp 35-52.

Kvale A. (1960) Norwegian earthquakes in relation to tectonics, Universitete i Bergen Arbok, Matematisk-naturvitenskapelig serie, 10, 17p.

Kvamme L.B. and Hansen R.A. (1989) The seismicity in the continental margin areas of northern Norway. In: Gregersen S and Basham P (eds), Earthquakes at North-Atlantic Passive Margins: Neotectonics and Postglacial Rebound, Kluwer, Dordrect, Ser C, 429-440.

Lagerbäck R., (1979) Neotectonic structures in northern Sweden, G.F.F., 100, pp 263-269.

Lagerbäck R. (1988a) Postglacial faulting and paleoseismicity in the Lansjarv area, northern Sweden. SKB Technical Report 88-25. Swedish Nuclear Fuel and Waste Management Co. Stockholm, 37pp.

Lagerbäck R. (1988b) Periglacial phenomena in the wooded areas of Northern Sweden - relicts from the Tarendo Interstadial, Boreas, 17, pp 487-499.

Lagerbäck R. (1990) Late Quaternary faulting and palaeoseismicity with particular reference to the Lansjarv area, northern Sweden, G.F.F., 112, pp 333-354.

Lagerbäck R. and Witschard F. (1983) Neotectonic in northern Sweden - geological investigations. SKBF/KBS Technical Report 83-58. Swedish Nuclear Fuel and Waste Management Co. Stockholm 58pp.

Lagerlund E. (1977) Till studies and neotectonics in northwest Skane, south Sweden, Boreas, 6, pp 159-166.

Lahermo P.W. and Lampen P.H. (1987) Brackish and saline groundwaters in Finland, pp103-109 in Fritz P. and Frape S.K. Saline water and gases in crystalline rocks, Geol. Ass. Canada Spec. Paper 33.

Landstrom O., Larson S. A., Lind G. and Malmquist D. (1980) Geothermal investigations in the Bohus Granite area in South-Western Sweden, Tectonophys., 64., pp 131-162.

Leijon B. (1986) Application of the LuT triaxial overcoring technique in Swedish mines. Proc. Int. Symp. on Rock Stress and Rock Stress Measurements, Stockholm, (Edited by O. Stephansson), pp 569-579. Centek Publishers, Lulea.

Leijon B.A. (1989) Relevance of pointwise rock stress measurements - an analysis of overcoring data. Int. J. Rock Mech. Min. Sci. & Geomech. Abstr. Vol. 26, No. 1, pp 61-68.

Li B. (1967) Rock stress in mines. Dr. Ing. Thesis, Norwegian University of Technology, Trondheim.

Lindquist J.G. (1984) Heat flow density measurements in the sediments of three lakes in Northern Sweden, in Terrestial Heat Flow Studies and the Structure of the Lithosphere eds. Cermak V. Ryback L. and Chapman D.S., Tectonophys., 103, pp 121-140.

Ljunggren C. and Raillard G. (1986) In-situ stress determination by hydraulic tests on preexisting fractures at Gide test site, Sweden. Research Report TULEA 1986:22, Lulea University of Technology, Lulea, Sweden, 77 p.

Ljunggren C. and Raillard G. (1987) Rock stress measurements by means of hydraulic tests on pre-existing fractures at Gide test site, Sweden. Int. J. Rock Mech. Min. Sci & Geomech. Abstr., Vol. 24, No. 6, pp 339-345.

Ljunggren C. (1990) Hydraulic fracturing and hydraulic tests on preexisting fractures in a foliated rock - a comparison of results and techniques. Rock Mechanics Contributions and Challenges: Proc. 31st U.S. Symp. on Rock Mechanics, Golden, Colorado, (Edited by W.A. Hustrulid and G.A. Johnson), pp 1027-1034. Balkema, Rotterdam.

Ljunggren C. and Bjarnason B. (1990) A comparison of hydraulic measuring techniques for borehole rock stress measurements. Report BeFo 211:1/90, Swedish Rock Engineering Research Foundation, Stockholm, 54 p.

Lundquist J. (1961) Beskrivning till karta over landisens avsmaltning och hogsta kustlinjen i Sverige, Sveriges Geol. Undersokn., Ba 18, pp 1-148.

Lundquist J. (1983) The glacial history of Sweden, in Ehlers J. (ed) Glacial Depoists in North-West Europe, pp 77-82. Balkema, Rotterdam.

Lundquist J. (1987) Late Weichselian glaciation and deglaciation in Scandinavia, pp 269-292, in Quaternary Glaciations in the Northern Hemisphere, Report on the International Geological Correlation Programme, Project 24, eds. V. Sibrava, D.Q. Bowen and G.M. Richmond, Pergamon Press, Oxford.

Lundquist J. and Lagerbäck R. (1976) The Parve Fault: A late-glacial fault in the Precambrian of Swedish Lapland. G.F.F. 98, pp 45-51.

Lykke-Andersen H. (1981) Indications of neotectonic features in Denmark, Z. Geomorph. 40, pp 43-54.

Lykke-Andersen H. (1987) Thickness of Quaternary deposits and their relation to the pre-Quaternary in the Fennoscandian boder zone in Kattegat and Vendsyssel, Boreas, 16, pp 369-371.

Matikainen R. (1988) Measurements of the state of stress in Finland. Research Report A64. Finnish Association of Mining and Metallurgy Engineering (in Finnish).

Matsuda T., Yamazaki H., Nakata T. and Imaizumi T. (1980) The surface faults associated with the Rikuu Earthquake of 1896, Tokyo Univ. Earthquake Res. Inst. Bull., 55, pp 795-855.

Meissner E. (1986) The Continental Crust - A Geophysical Approach, Academic Press, Orland, 426pp.

Meissner R., Wever Th., and Fluh E.R. (1987) The moho in Europe - implications for crustal development. Ann. Geophys., 5B, pp 357-364.

Meyer K. and Ahjos T. (1985) Temporal variations of energy release by earthquakes in the Baltic Shield, Geophysica 21., 51-64.

Meyer K.O and Scherman S. (1988) Stress migration in the North Atlantic and intraplate seismicity in Scandinavia - a proposal, Tectonophysics 156, 175-8.

Mörner N-A. (1969) The Late Quaternary history of the Kattegatt Sea and the Swedish West Coast: deglaciation, shorelevel displacement. chronology, isostasy and eustasy, Sveriges Geol. Undersokn., C-140, pp 1-487.

Mörner N-A. (1977) Past and present uplift in Sweden: glacial isostasy, tectonism and bedrock influence, G.F.F., 99, pp 48-54.

Mörner N-A., (1978) Faulting fracturing and seismic activity as a function of glacial-isostasy in Fennoscandia, Geology, 6, pp 41-45.

Mörner N-A., (1985) Palaeoseismicity and geodynamics in Sweden, Tectonophys., 117, pp 139-153.

Mörner N-A, Somi E., Zuchiewicz, (1989) Neotectonics and palaeoseismiocity within the Stockholm intracratonal region in Sweden, Tectonophys., 163, pp 289-303.

Mueller St. and Panza G.F. (1984) The lithosphere-asthenosphere system in Europe. In Proc. of "First EGT Workshop: The Northern Segment", edited by D.A. Galson and St. Mtller, Copenhagen, 28-30 October 1983.

Mueller B., Fuchs K., Mastin L., Gregersen S., Pavoni N., Stephansson O., Ljunggren C. and Zoback M.L. (1992) Regional patterns of stress in Europe. J. Geophys. Res.

Muir Wood R. (1989a) Extraordinary deglaciation reverse faulting in northern Fennoscandia pp141-173 in S. Gregersen and P. W. Basham eds., Earthquakes at North-Atlantic Passive Margins: Neotectonics and Postglacial Rebound, Kluwer, Dordrecht.

Muir Wood R. (1989b) The Scandinavian earthquakes of 22 December 1759 and 31 August 1819. Disasters, 12, pp 223-236.

Muir Wood R. and Forsberg C.F. (1987) Regional crustal movements on the Norwegian continental shelf, ELOCS (Earthquake Loading on the Norwegian Continental, Shelf) Rep.1-3, 148pp, Norwegian Geotechnical Institute, Oslo, NTNF/NORSAR, Kjeller and Principia Mechanica Ltd, London.

Muir Wood R. and King G.C.P. (1993) Hydrological Signatures of Earthquake strain, (J. Geophys. Res., submitted).

Muir Wood R. and Steedman R.S. (1992) Earthquake induced groundwater changes and liquefaction. Proc 10th World Conf Earthquake Eng, Madrid, July, pp1337-1341.

Muir Wood R. and Woo G. (1987) The historical seismicity of the Norwegian continental shelf, ELOCS (Earthquake Loading on the Norwegian Continental, Shelf) Rep.2-1, 124pp, Norwegian Geotechnical Institute, Oslo, NTNF/NORSAR, Kjeller and Principia Mechanica Ltd, London.

Mutter J.C. (1984) Cenozoic and late Mesozoic stratigraphy and subsidence history of the Norwegian margin, Geol. Soc., Am. Bull., 95, pp 1135-1149.

Myrvang A. and Grimstad E. (1981) Rock stress measurements as a practical tool in Norwegian mining engineering practice. Proc. AIME Annual Meeting, Chicago.

Nesje A. and Dahl S.O. (1990) Authochthonous block fields in southern Norway: implications for the geometry, thickness and isostatic loading of the Late Weichselian Scandinavian ice sheet., J. Quat. Sci., 5, pp 225-234.

Niini H. (1987) Bedrock Fractures affecting land uplift in Finland, Geol. Surv Finland, Special Paper 2, pp 51-64.

NOAA, World Data Center A: World Stress Map. 315 Marine Dr., Boulder, CO. 80220, USA (data available on floppy disk).

Norling E. (1984) Kimmerian tectonics, stratigraphy and palaeogeography of Skane, G.F.F., 106, pp 392-4.

Olesen O. (1988) The Stuoragurra fault, evidence of neotectonics in the Precambrian of Finnmark, northern Norway. Norsk. Geol. Tidsskrift., 68, pp 107-118.

Olesen O., Henkel H., Lile O.B. Muring E. and Steinar J. (1991) Detailed geophysical investigations of the Stuoragurra postglacial fault, Finnmark, northern Norway, NGU Report 90.160, 36pp.

Paananen M. (1987) Geophysical studies of the Venejarvi, Ruostejarvi, Suasselka and Pasmajarvi potglacial faults in northern Finland, Geol. Survey of Finland, Nuclear Waste Disposal Research, Report YST-59, 97 poages.

Panza G.F., Duda S.J., Cernobori L. and Herak M. (1989) Gutenberg's surface-wave magnitude calibrating function: theoretical basis from synthetic seismograms. Tectonophysics, 166, 35-43.

Parasnis D.S. (1975) Temperature phenomena and heat flow estimates in two Precambrian ore-bearing areas in North Sweden, Geophys. J. R. Astron. Soc., 43, pp 531-554.

Parasnis D.S. (1982) Geothermal flow and phenomena in two Swedish localities north of the Arctic Circle, Geophys. J. R. Astron. Soc., 71, pp 545-554.

Pasquale V., Cabella C. and Verdoya M. (1990) Deep temperatures and lithospheric thickness along the European Geotraverse, Tectonophys., 176, pp 1-11.

Pasquale V., Verdoya M. and Chiozzi (1991) Lithospheric thermal structure in the Baltic Shield, Geophys J. Intl., Vol 106, pp 611-620.

Porkka M.T. and Korhonen H. (1977) Earthquakes in the northern parts of the Baltic Shield, pp 97-104 in C. Kortman ed., Fault tectonics in the eastern part of the Baltic Shield, The Committee for Scientific and Technical Co-operation between Finland and the Soviet Union, Helsinki

Ramberg I.B. and Spjeldnaes N. (1978) The tectonic history of the Oslo region, pp 167-194 in I.B. Ramberg and E.R. Neumann eds., Tectonics and Geophysics of Continental Rifts, Reidel, Dordrecht.

Risberg J. and Sandgren P. (1993) Shore displacement and evidence of irregular isostatic uplift during early Holocene in south-western Varmland, western Sweden. NSK Report, 30p.

Roberts D. (1991) A contemporary small-scale thrust-fault near Lebesby, Finnmark, Norsk. Geol. Tidsskr., 71, pp 117-120.

Rokoengen K. and Ronningsland T.M. (1983) Shallow bedrock geology and Quaternary thickness in the Norwegian sector of the North Sea between 60°30'N and 62°N, Norsk Geol. Tiddskr. 63, pp83-102.

Rokoengen K. and Saettem J. (1983) Shallow bedrock geology and Quaternary thickness off northern Helgeland, Vestfjorden and Lofoten, IKU Report P-155.

Rummel F. (1986) Stresses and tectonics of the upper continental crust - a review. In Proc. Int. Symp. on Rock Stress and Rock Stress Measurements, Stockholm (Edited by O. Stephansson), pp 177-186. Centek Publishers, Lulea.

Rybach L. and Buntebarth G. (1984) The variation of heat generation, density and seismic velocity with rock type in the continental lithosphere, in Terrestial Heat Flow Studies and the Structure of the Lithosphere eds., Cermak V., Rybach L. and Chapman D.S., Tectonophys., 103., pp 335-344.

Saari J. (1990) Microearthquakes in the Baltic Shield: Seismotectonic analysis for nuclear power plant in the Loviisa area, south-eastern Finland. Nuclear waste commission of Finnish power companies, Report YJT-90-16, 31 pp.

Scholz C.H. (1990), The mechanics of earthquakes and faulting. CUP, Cambridge.

Shapira A., Kulhanek O. and Wahlström R. (1979) Detectability of regional events by means of the Swedish seismograph station network. Seismological Department, Uppsala University, Report 7-79, 17 pp.

Sivhed U. (1991) A pre-Quaternary, post-Palaeozoic erosional channel deformed by strike-slip faulting, Scania, southern Sweden, G.F.F. 113, pp 139-143.

Sjöberg L.E. (1983) Land uplift and its implications on the geoid in Fennoscandia, Tectonophys. 97, pp 97-101.

Sjöberg L.E. (1986) Determination of the land uplift from old water marks and tide gauge data at Ratan and Lovgrundet/Bjorn, Sweden, Neotectonics, 1, pp 51-59.

Sjöberg L.E. (1987) Comparsion of some methods of determining land uplift rates from tide gauge data, Zeit f. Vermessungswesen. 2, pp 69-73.

Sjöberg L.E. and Fan H. (1986) Studies on secular land uplift and long periodic variations of sea level around the coasts of Sweden, Dept. of Geodesy, Royal Institute of Technology, Stockholm., Report 3. 59pp.

Sjöberg R. (1987) Caves as indicators of neotectonics in Sweden. Z. Geomorph. N.F. Suppl. Bd., 6: pp 141-148.

Skordas E. and Kulhanek O. (1991) Spatial and temporal variations of b-values in Fennoscandia. Unpublished manuscript.

Skordas E., Meyer K., Olsson R. and Kulhanek O., (1991) Causality betwen interplate (North Atlantic) and intraplate (Fennoscandia) seismicities, Tectonophys., 185, pp 295-307.

Slunga R. (1981) Earthquake source mechanism determination by use of body-wave amplitudes - an application to Swedish earthquakes. Bull Seismol Soc Am, 73, 1499-1511.

Slunga R. (1985) The seismicity of Southern Sweden, 1979-1984, final report. FOA Report C 20572-T1, 45 pp.

Slunga R. (1989a) Focal mechanisms and crustal stresses in the Baltic Shield: In Gregersen, S and Basham P (eds), Earthquakes at North Atlantic Passive Margins: Neotectonics and Postglacial Rebound, Kluwer, Dordrecht, Ser C, 261-276.

Slunga R. (1989b) Earthquake mechanisms in Northern Sweden Oct 1987-Apr 1988. SKB TR 89-28, 165 pp, Stockholm.

Slunga R. and Nordgren L. (1987) Earthquake measurements in Southern Sweden Oct 1 1986 - March 31 1987. SKB TR 87-27, 60 pp, Stockholm.

Slunga R. and Nordgren L. (1990) Earthquake measurements in Southern Sweden Apr 1 1987 - Nov 30 1988. SKB, Arbetsrapport 90-19, 90 pp, Stockholm.

Slunga R., Norrman P. and Glans A-C (1984) Seismicity of Southern Sweden. FOA Report C 20543-T1, 106 pp.

Sorensen R., Bakkelid S., and Torp B. (1987) Nasjonalatlas for Norge, Hovedtema 2: Landformer, berggrun og losmasser, Kartblad 2.3.3.

Stein S., Sleep N., Geller R., Wang S. and Kroeger G. (1979) Earthquakes along the passive margin of eastern Canada, Geophys. Res. Lett., 5, pp 537-540.

Stephansson O. (1988) Ridge push and glacial rebound as rock stress generators in Fennoscandia. In Geological Kinematics and Dynamics: From Molecules to the Mantle (Edited by C. Talbot). Bull. Geol. Instns of Uppsala, 14, 39-48.

Stephansson O. (1989) Stress measurements and modelling of crustal rock mechanics in Fennoscandia. In Earthquakes at North-Atlantic Passive Margins: Neotectonics and Postglacial

Rebound (Edited by S. Gregersen and P.W. Basham), pp 213-229. Kluwer Academic Publishers, Dordrecht.

Stephansson O. and Angman P. (1986) Hydraulic fracturing stress measurements at Forsmark and Stidsvig, Sweden. Bull. Geol. Soc. Finland, 58, 307-333.

Stephansson O., Dahlstrom L-O., Bergstrom K., Sarkka P., Vtinen A., Myrvang A., Fjeld O. and Hansen T.H. (1987) Fennoscandian Rock Stress Data Base - FRSDB. Research Report TULEA 1987:06, Lulea University of Technology, Lulea.

Stephansson O. and Ljunggren C. (1988) Swedish stress data for the World Stress Map Project and geological implications of the data in FRSDB. In Proc. of the "Workshop on Nordic Rock Stress Data, Trondheim", SINTEF Report, October.

Stephansson O., Sarkka P. and Myrvang A. (1986) State of stress in Fennoscandia. Proc. Int. Symp. on Rock Stress and Rock Stress Measurements, Stockholm (Edited by O. Stephansson), pp 21-32. Centek Publishers, Lulea.

Stephansson O., Savilahti T. and Bjarnason B. (1989) Rock mechanics at the deep borehole at Gravberg, Sweden. In Proc. Rock at Great Depth, Pau, vol. 2, pp 863-870.

Stephansson O., Savilahti T. and Bjarnason B. (1990) Rock mechanics of the Gravberg well no 1. Technical Report U(G) 1990/46, Swedish State Power Board, 58 p.

Stephansson O. and Shen B. (1991) Modelling of faulted rock mass response to glaciation, thermal loading and seismicity. Engineering Geology, June 1991.

Strand T and Kulling O. (1972) Scandinavian Caledonides, Wiley Interscience, London.

Suutarinen O. (1983) Recomputation of land uplift values in Finland, Reports of the Finnish Geodetic Institute, 83:1, 16p.

Svensson N-O. (1991) Postglacial land uplift patterns of south Sweden and the Baltic Sea region, TERRA Nova, 3, pp 369-378.

Sykes L.R. (1965) The seismicity of the Arctic, Bull. Seism. Soc. Am., 55, pp 519-536.

Talbot C.J. (1986) A preliminary structural analysis of the pattern of postglacial faults in northern Sweden, SKB Technical Report 86-20, Swedish Nuclear Fuel and Waste Management Co. Stockholm 36pp.

Talbot C.J. (1990a) GPS networks to measure active strains in Sweden, G.F.F. Abstract, Lund Geologiske Foreningens Oct 4th.

Talbot C.J. (1990b) Problems posed to a bedrock radwaste repository by genetly dipping fracture zones. G.F.F. 112, pp 355-359.

Talbot C.J. and Slunga R. (1989) Patterns of active shear in Fennoscandia. pp441-466 in Gregseren D. and Basham PO.W. eds., earthquakes at North Atlantic Passive Margins: Neotectonics and postglacial rebound, Kluwer, Dordrecht.

Tanner V. (1930) Studier over Kvartarsystemet i Fenoskandias Nordliga Delar: IV , Bull. Geol. Komm. Finland, No. 88.

Tiren S.A. (1989) Geological setting and deformation history of a low angle fracture zone at Finnsjon, Sweden, part 2. in Characterization of Fracture zone 2, Finnsjon study-site, SKB Technical Report 89-19.

Tynni R. (1982) The reflection of geological evolution in Tertiary and interglacial diatoms and silicoflagellates in Finnish Lapland, Geol. Survey of Finland Bull., 320, pp 1-40.

Ulmishek G. (1991) Geologic Evolution and Petroleum Resources of the Baltic Basin, pp 603-632, in Interior Cratonic Basins, AAPG Memoir 51.

Ussiso I. (1977) Computation of land uplift and mean sealevel in Sweden, G.F.F., 99, pp 42-48.

Vanek J., Zatopek A., Karnik V., Kondorskaya N.V., Riznichenko Yu V., Savarensky E.F., Solov'ev S.L. and Shebalin N.V. (1962) Standardisation of magnitude scales. Izv, Geophys Ser, 153-158.

Verio A. (1987) On nonlinearities in the Fennoscandian land uplift within the main sulphide ore belt, Journal of Geodynamics, 8, pp 151-161.

Verio A. (1990) Vertical movements at some faults in Finland, Geological Survey of Finland, Nuclear Waste Disposal Research, Working Report 4-90, (presented at 11th Meeting of the Nordic Geodetic Commission, Copenhagen).

Vidal G. (1984) Lake Vattern, G.F.F., 106, p 397.

Vuorela P., Kuivamaki A. and Paananen M. (1987) Neotectonic bedrock movements. A preliminary survey of the Pasmajarvi fault. Geol. Surv of Finland, Nuclear Waste Disposal research, Report 57, 15pp.

Wahlström R. (1978) Magnitude-scaling of earthquakes in Fennoscandia. Seismological Institute, Uppsala University, Report 3-78, 44 pp.

Wahlström R. (1980) Near-surface earthquakes in the Baltic Shield, notably Sweden. Tectonophys, 68, T31-T35.

Wahlström R. (1989) Seismodynamics and postglacial faulting in the Baltic Shield. In: S Gregersen and P Basham (Editors), Earthquakes at North-Atlantic Passive Margins: Neotectonics and Postglacial Rebound, Kluwer, Dordrect, Ser C, 467-482.

Wahlström R. (1990) A catalogue of earthquakes in Sweden from 1375-1890, G.F.F. 112, pp 215-225.

Wahlström R. and Ahjos T. (1984) Magnitude determination of earthquakes in the Baltic Shield. Annales Geophys, 2, 553-558.

Wahlström R., Linder S-O and Holmqvist C. (1987) Near distance seismological monitoring of the Lansjärv neotectonic fault region. SKB TR 88-12, 27 pp, Stockholm.

Wahlström R., Linder S-O, Holmqvist C. and Mårtensson H-E (1989). Near-distance seismological monitoring of the Lansjärv neotectonic fault region, Part II. SKB TR 89-01, 17 pp, Stockholm.

Wahlström R. and Ahjos T. (1984) Magnitude determination of earthquakes in the Baltic Shield, Annales Geophysicae 2, pp 553-557.

Wannas K. (1989) Seismic stratigraphy and tectonic pattern of the Bothnian Bay. Stockholm Contributions to Geology.

Yielding G., Ouyed M., King G.C.P. and Hatzfeld D. (1989) Active tectonics of the Algerian Atlas Mountains from aftershocks of the 1980 El Asnam earthquake, Geophys. J. Intl., 99, pp 761-788.

Zoback M.L. and Zoback M. (1980) State of stress in the conterminous United States. J. Geophys. Res., 85, 6113-6156.

Zoback M.L., Zoback M.D., Adams J., Assumpcao M., Bell S., Bergman E.A., Blumling P., Brereton N.R., Denham D., Ding J., Fuchs K., Gay N., Gregersen S., Gupta H.K., Gvishiani A., Jacob K., Klein R., Knoll P., Magee M., Mercier J.L, Mueller B.C., Paquin C., Rajendran K., Stephansson O., Suarez G., Suter M., Udias A., Xu Z.H. and Zhizhin M. (1989) Global patterns of tectonic stress. Nature 341, No. 6240, 291-298.

List of SKB reports

Annual Reports

1977-78
TR 121
KBS Technical Reports 1 – 120
Summaries
Stockholm, May 1979

1979
TR 79-28
The KBS Annual Report 1979
KBS Technical Reports 79-01 – 79-27
Summaries
Stockholm, March 1980

1980
TR 80-26
The KBS Annual Report 1980
KBS Technical Reports 80-01 – 80-25
Summaries
Stockholm, March 1981

1981
TR 81-17
The KBS Annual Report 1981
KBS Technical Reports 81-01 – 81-16
Summaries
Stockholm, April 1982

1982
TR 82-28
The KBS Annual Report 1982
KBS Technical Reports 82-01 – 82-27
Summaries
Stockholm, July 1983

1983
TR 83-77
The KBS Annual Report 1983
KBS Technical Reports 83-01 – 83-76
Summaries
Stockholm, June 1984

1984
TR 85-01
Annual Research and Development Report 1984
Including Summaries of Technical Reports Issued during 1984. (Technical Reports 84-01 – 84-19)
Stockholm, June 1985

1985
TR 85-20
Annual Research and Development Report 1985
Including Summaries of Technical Reports Issued during 1985. (Technical Reports 85-01 – 85-19)
Stockholm, May 1986

1986
TR 86-31
SKB Annual Report 1986
Including Summaries of Technical Reports Issued during 1986
Stockholm, May 1987

1987
TR 87-33
SKB Annual Report 1987
Including Summaries of Technical Reports Issued during 1987
Stockholm, May 1988

1988
TR 88-32
SKB Annual Report 1988
Including Summaries of Technical Reports Issued during 1988
Stockholm, May 1989

1989
TR 89-40
SKB Annual Report 1989
Including Summaries of Technical Reports Issued during 1989
Stockholm, May 1990

1990
TR 90-46
SKB Annual Report 1990
Including Summaries of Technical Reports Issued during 1990
Stockholm, May 1991

1991
TR 91-64
SKB Annual Report 1991
Including Summaries of Technical Reports Issued during 1991
Stockholm, April 1992

1992
TR 92-46
SKB Annual Report 1992
Including Summaries of Technical Reports Issued during 1992
Stockholm, May 1993

Technical Reports
List of SKB Technical Reports 1993

TR 93-01
Stress redistribution and void growth in butt-welded canisters for spent nuclear fuel
B L Josefson[1], L Karlsson[2], H-Å Häggblad[2]
[1] Division of Solid Mechanics, Chalmers University of Technology, Göteborg, Sweden
[2] Division of Computer Aided Design, Luleå University of Technology, Luleå, Sweden
February 1993

TR 93-02
Hydrothermal field test with French candidate clay embedding steel heater in the Stripa mine
R Pusch[1], O Karnland[1], A Lajudie[2], J Lechelle[2], A Bouchet[3]
[1] Clay Technology AB, Sweden
[2] CEA, France
[3] Etude Recherche Materiaux (ERM), France
December 1992

TR 93-03
MX 80 clay exposed to high temperatures and gamma radiation
R Pusch[1], O Karnland[1], A Lajudie[2], A Decarreau[3],
[1] Clay Technology AB, Sweden
[2] CEA, France
[3] Univ. de Poitiers, France
December 1992

TR 93-04
Project on Alternative Systems Study (PASS). Final report
October 1992

TR 93-05
Studies of natural analogues and geological systems. Their importance to performance assessment.
Fredrik Brandberg[1], Bertil Grundfelt[1], Lars Olof Höglund[1], Fred Karlsson[2], Kristina Skagius[1], John Smellie[3]
[1] KEMAKTA Konsult AB
[2] SKB
[3] Conterra AB
April 1993

TR 93-06
Mineralogy, geochemistry and petrophysics of red coloured granite adjacent to fractures
Thomas Eliasson
Chalmers University of Technology and University of Göteborg, Department of Geology, Göteborg, Sweden
March 1993

TR 93-07
Modelling the redox front movement in a KBS-3 nuclear waste repository
L Romero, L Moreno, I Neretnieks
Department of Chemical Engineering, Royal Institute of Technology, Stockholm, Sweden
May 1993

TR 93-08
Äspö Hard Rock Laboratory Annual Report 1992
SKB
April 1993

TR 93-09
Verification of the geostatistical inference code INFERENS, Version 1.1, and demonstration using data from Finnsjön
Joel Geier
Golder Geosystem AB, Uppsala
June 1993

TR 93-10
Mechanisms and consequences of creep in the nearfield rock of a KBS3 repository
Roland Pusch, Harald Hökmark
Clay Technology AB, Lund, Sweden
December 1992

TR 93-11
Post-glacial faulting in the Lansjärv area, Northern Sweden. Comments from the expert group on a field visit at the Molberget post-glacial fault area, 1991
Roy Stanfors (ed.)[1], Lars O Ericsson (ed.)[2]
[1] R S Consulting AB
[2] SKB
May 1993

TR 93-12
Possible strategies for geoscientific classification for high-level waste repository site selection
Lars Rosén, Gunnar Gustafson
Department of Geology, Chalmers University of Technology and University of Göteborg
June 1993

ISSN 0284-3757

CM Tryck AB, Bromma 1992