Ridge push and glacial rebound as rock stress generators in Fennoscandia

OVE STEPHANSSON

Stephansson, O. 1988 12 30: Ridge push and glacial rebound as rock stress generators in Fennoscanida. *Bulletin of the Geological Institutions on the University of Uppsala,* N.S. Vol. 14, pp. 39–48. Uppsala ISSN 0302-2749.

This contribution summarizes and interprets some of the output from the established Fennoscandian Rock Stress Data Base (FRSDB). It is suggested that three stress-generating mechanisms are responsible for the virgin stress state in the Fennoscandian Shield. Assuming a remaining 140 m of uplift from the latest glaciation, the remaining maximum horizontal stress due to subsidence is estimated to be only 2–3 MPa. The locked–in stresses due to rock creep from loading of a 3 km thick ice sheet provides the excess horizontal stresses, but the stress gradient with depth is insufficient. To explain the present stress state in Fennoscandia, ridge push is introduced as an active stress system. Based on the results from finite element modelling of viscoelastic relaxation in the mantle and stress migration to the crust, a composite stress diagram for the upper and lower crust in Fennoscandia has been constructed.

At present, stresses generated by ridge push cannot be distinguished from glacial rebound. The scatter in the recorded stress data is discussed in connection with the frictional strength of faults and surface topography.

Ove Stephansson, Division of Rock Mechanics, Luleå University of Technology, S-951 87 Luleå, Sweden. *Received 30th June 1987; Revision received 21st December 1987.*

Introduction

This paper is concerned with the inherent difficulties in the interpretation of in-situ stress determinations in general, and stresses for the Fennoscandian Shield in particular. Data from rock stress measurements in Fennoscandia have now been collected and stored in a data base (Stephansson et al., 1986). This allows a thorough analysis of the magnitude and orientation of crustal stresses. The data will also throw new light on the origin of crustal stresses.

The first rock stress measurements in Fennoscandia by Hast (1958) indicated that horizontal stresses exceed the vertical stress. Energy release from the late glaciation was assumed to be responsible for this stress state. This paper analyses the possible contribution to the stress state of the isostatic response of the crust to an ice sheet loading. The theory proposed by Walcott (1970) is applied to an ice sheet of thickness 3 km, and the contribution of stresses from a remaining uplift of about 140 m is estimated.

Glacially induced stress changes can be generated by the creep of rock material. Loading from an ice sheet will cause creep in crustal rock. Deglaciation will then unload the rocks and the induced stresses will relax. However, the viscoelastic-plastic nature of the crust will lock in some of the induced stresses as demonstrated for the general case by Voight (1966) and with special application to Fennoscandia by Bergman (1977). In this paper, the "locked" stresses from the loading of an ice sheet 3 km in thickness will be analysed and an attempt will be made to fit the result obtained to the stress state compiled from the Fennoscandian Rock Stress Data Base.

Fennoscandia is known to be a part of a broad midplate compressive stress province of the Eurasian plate, where the excess horizontal stresses suggest a far-field source. Since the western part of the plate has no associated subducting slab, the most likely plate driving force is ridge push (Bott and Kusznir, 1984). Thus, a ridge spreading stress at the Mid Atlantic Ridge is assumed to be one of the contributors to the stress field in Fennoscandia (Klein and Barr, 1986). The applied push force at the ridge will be concentrated in the upper lithosphere as a result of creep and stress decay in the lower lithosphere. This effect of stress amplification in the crust was studied by Bott and Kusznir (1984) and later by Hasegawa et al. (1985), and is here ap-

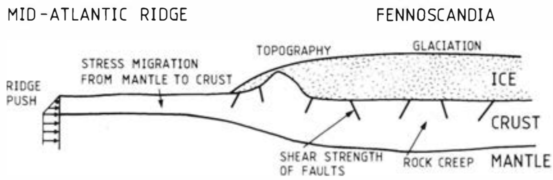


Fig. 1. Schematic stress-generating mechanisms in Fennoscandia.

plied to the stress state in the Fennoscandian Shield. Based on existing data in the Fennoscandian Rock Stress Data Base, and taking account of the stress amplification in the crust from ridge push, a possible stress gradient for the upper and lower Fennoscandian crust is formulated.

The following stress-generating mechanisms are considered in this paper: (i) isostatic response and creep of crustal rocks from ice loading, (ii) ridge push, and (iii) stress migration from the earth's mantle to the crust (Figure 1). The effects of topography and the shear strength of faults and joints are also discussed.

The stress state in Fennoscandia

Many compilations of in-situ stress measurements, from various parts of the world, have now been published. The most recent is presented in the proceedings of the International Symposium on Rock Stress and Rock Stress Measurements, Stockholm. The contributions by Stephansson et al. (1986), Klein and Barr (1986), and Herget (1986) are of special interest to this study.

Changes of stresses with depth

Stephansson et al. (1986) presented some preliminary results from the Fennoscandian Rock Stress Data Base (FRSDB) which, in 1986, contains almost 500 entries from 102 sites in Finland, Norway and Sweden. Regression analyses of maximum and minimum horizontal stresses versus depth were presented for four different stress measurement methods. The following conclusions can be drawn concerning the variation of stress with depth:

- There is a large horizontal stress component in the uppermost 1 000 m of bedrock.
- Discrepancies in the variation of stress magnitude with depth have been obtained wherever different rock stress measurement methods are used. Of all the methods used, the overcoring method of Hast (1958) gives by far the largest stress gradients and intercepts of stress at the bedrock ground surface, cf. figure 4 in Stephansson et al. (1986).
- The maximum and minimum horizontal stresses exceed the vertical stress as estimated from the weight of the overburden.
- Stress measurements from the Leeman, Leeman-NTH, Leeman-LuT overcoring methods have revealed minor differences in the magnitudes of the minimum horizontal stress, σ_{HMIN}, and the vertical stress σ_V.
- Regression analyses of the principal stresses versus depth for the overcoring rock stress measurement methods give the following results:

$$\begin{aligned} \sigma_1 &= 0.050\,z + 7.9\ ,\ r = 0.61 \\ \sigma_2 &= 0.32\,z + 4.2\ ,\ r = 0.60 \\ \sigma_3 &= 0.019\,z + 0.6\ ,\ r = 0.56 \end{aligned} \qquad (1)$$

where z is the depth in metres and the stresses are expressed in MPa, and r is the regression coefficient.
- The ratio of the average horizontal stress to the vertical stress versus depth is represented in Figure 2. It can be seen that there is considerable

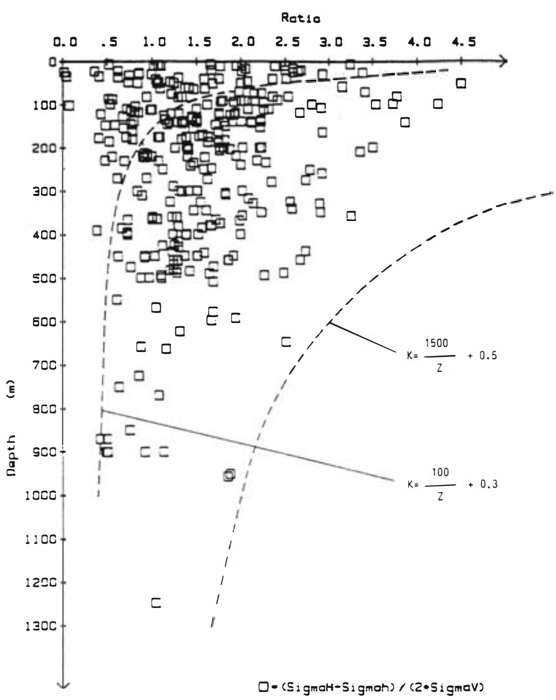


Fig. 2. Ratio of average horizontal stress/vertical stress as a function of depth for rock stress measurements in Fennoscandia. The dashed curves are from Brown and Hoek (1978). Open squares are data points.

scatter in the data points for the uppermost 200–300 m, but this ratio tends to a limiting value of unity for depths exceeding 1 000 m. The "trend lines" between measured in-situ stresses and depth, as suggested by Brown and Hoek (1978), are shown in Figure 2.

In the discussion of stress gradients in the upper crust, it should be noted that a major discontinuity in the stress field with depth has been reported from several test sites. Three such stress discontinuities are reported for areas in the Fennoscandian Shield. Martna et al. (1983) and later Stephansson and Ångman (1986) reported stress jumps of the order of 20 MPa for the maximum horizontal stress across a major subhorizontal fracture zone at 320 m depth for a vertical borehole at Forsmark, central Sweden. A similar result was obtained from the stress measurements at Lavia, central Finland (Bjarnason and Stephansson, 1987) where a stress jump of about 20 MPa was inferred from measurements below a major fracture zone at 420 m depth in Proterozoic granodiorite rocks. Pronounced changes in the stress fields have been reported in the nappes around the headrace tunnels of the Vietas power plant in the Swedish Caledonides by Martna and Hansen (1986).

Current measurements indicate that stress discontinuities can exist across both fracture zones and lithological boundaries. The stresses have different orientations and magnitudes across such discontinuities. The stress changes are of the order of a few tens of megapascals and are of the same order of magnitude as the stress drops generated by intraplate earthquakes.

Orientation of stresses

In 1958, Hast published a compilation of in-situ rock stress measurements made in boreholes within the Fennoscandian Shield. Most of these measurements were conducted in mines and for several measuring sites in the vicinity of mining operations.

In a later compilation which included Hast's data, Ranalli and Chandler (1975) concluded that within the southern part of the Fennoscandian Shield, the maximum horizontal stress is directed approximately E-W, while in the northern part of the Shield the trend is more N-S. This early picture of the stress direction in Fennoscandia was upheld until Slunga et al. (1984) published data from earthquake fault plane solutions of the southern parts of the Fennoscandian Shield. Their result may be summarized as follows: The seismicity of the southern part of the Baltic Shield was monitored for four years by a digital regional seismic network, with a station spacing of 100 km. Some 160 earthquakes were analysed for location, focal depth, seismic moment, fault plane solution and static stress drop. More than 90 % of the earthquakes occured at

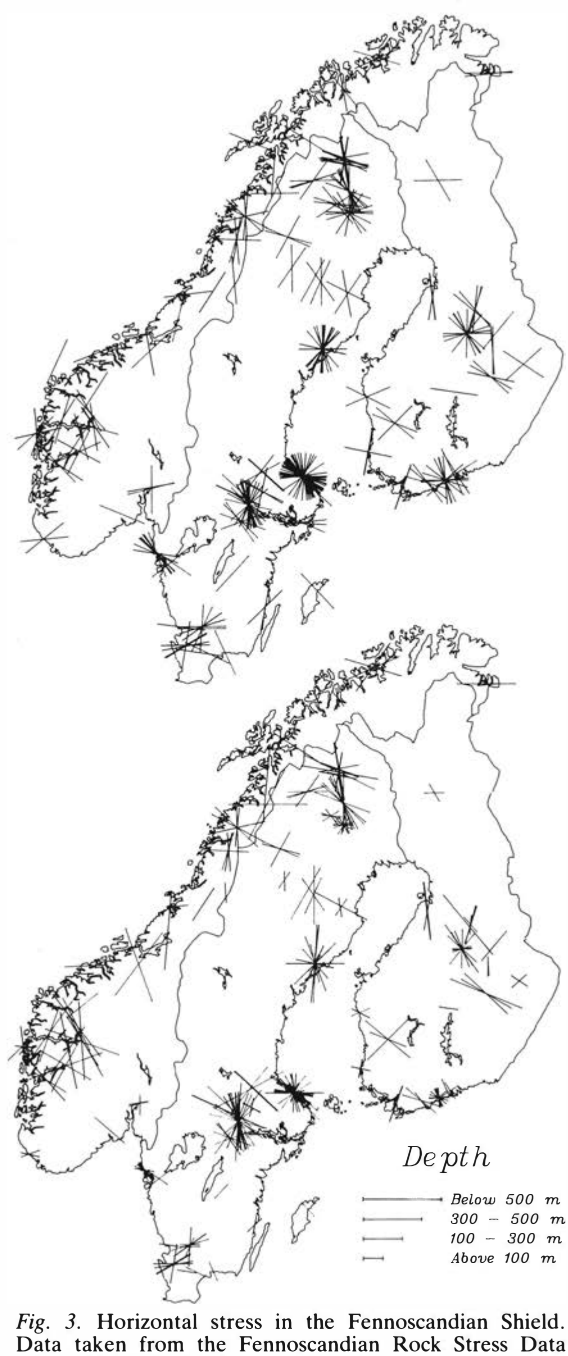


Fig. 3. Horizontal stress in the Fennoscandian Shield. Data taken from the Fennoscandian Rock Stress Data Base (FRSDB), Stephansson et al. (1987). A) Direction of maximum horizontal stress for all data points and sites at all depths. B) Direction of maximum horizontal stress at different depths.

depths of less than 19 km. The orientation of the horizontal stresses relaxed by the events is very consistent, the principal compression was characteristically NW-SE. The geographical distributions of the events are related to the regional geology.

In a recent paper, Klein and Barr (1986) presented a compilation of previously published western European in-situ stress data, together with results of wellbore breakout analyses of wells drilled in the North Sea, the Atlantic Ocean and onshore Britain. They found that within central and northern Europe, the North Sea, the Atlantic Ocean, the British Isles and northern Scandinavia the regional direction of maximum horizontal stress is aligned approximately NW-SE. In southern Scandinavia they claim that the maximum stress direction is approximately E-W. However, the data they present are based on old measurements, and their statement is not valid. Klein and Barr (1986) concluded that the consistent NW-SE orientation of the maximum horizontal in-situ stress direction over western Europe is dominated largely by plate tectonic boundary forces acting on the Eurasian tectonic plate. These forces include plate edge forces (for example compressive ridge push forces perpendicular to the Eurasian/ African continental collision zone as expressed by the Alps).

A compilation of the magnitude and direction of the major horizontal stress at each of the measuring sites for the virgin stress state in Fennoscandia is shown in Figure 3A. The variations in the direction of the horizontal stress with depth at each site is apparent.

The direction of maximum horizontal stress at different depths in the Fennoscandian Shield is shown in Figure 3B. The large number of measurements and the scatter in the direction for individual sites is demonstrated for Forsmark and Stripa in central Sweden, Kiruna and Malmberget mines in northern Sweden and Helsinki in southern Finland. Furthermore, there is a slight tendency for the maximum horizontal stress to align with the axis of the Caledonides. At depths below 300 m, the maximum horizontal stress tends to act in the NW-SE direction in central Sweden and Finland, cf. Figure 3B.

Glacially induced stress changes

Isostatic response of an elastic lithosphere

By studying the nature of the deformation or the distribution of the compensation produced by an ice sheet, the isostatic response of the crust to loading can be assessed. Walcott (1970) studied the isostatic response to ice sheet loading of the earth's crust in Canada, whereby the lithosphere was treated as an elastic sheet overlying a fluid substratum. By applying the well-known theory of elastic bending of a thin plate in two dimensions, the lifting force caused by the elastic bending was determined by Walcott (1970).

The stress state can be determined from the second derivative of deflection, and the maximum stress difference will occur at the base of the lithosphere and at about 200–300 km from the edge of the ice sheet (Walcott, 1970). The magnitude is approximately 20 MPa per kilometre of elevation of the ice sheet. The maximum horizontal stress at the upper surface of the upper crust is estimated to be 8.5 MPa per kilometre of elevation of ice. Given an ice sheet of 2 km thickness, the horizontal stress due to subsidence is 17 MPa and compressive at the upper surface of the crust (Walcott, 1970).

The total depression of the Fennoscandian region by the latest glaciation is estimated to be 900 m (Kukkari, 1986). Assuming 140 m of uplift remains, as stated by the same author, the residual maximum horizontal stress due to subsidence will be of the order of 2–3 MPa. This horizonal stress is much lower than the measured excess of horizontal stress in Fennoscandia.

The storage of stress during rock creep

Voight (1966), and later Bergman (1977), studied the influence of creep on a rock mass due to ice loading and melting. The stresses generated can be attributed to three mechanisms:

a) elastic deformation from ice load
b) creep from ice load
c) residual stress after ice melting.

Is it possible to explain the excess of horizontal stress in Fennoscandia by any of these mechanisms?

Let us determine the state of stress for a point in the earth crust at depth H and with volume weight γ. Assume the rock to be covered with an ice sheet of thickness H_1 and volume weight γ_1. The vertical stress now becomes

$$\sigma_V = \gamma H + \gamma_1 H_1 \tag{2}$$

and for an elastic rock mass the horizontal stress is

$$\sigma_H = K_o \sigma_V \tag{3}$$

where $K_o = \dfrac{\nu}{1 - \nu}$

For Poisson's ratio $\nu - 1/6$ we have $K_o = 0.2$.

Due to longterm loading under the weight of the ice sheet, the rock mass starts to creep. The stresses tend to approach a hydrostatic state and the ratio K_c of horizontal to vertical stress has the following bounds

$$K_o < K_c < 1$$

As the ice is melting, the vertical stress is reduced by an amount equal to the wieght of the ice sheet. The horizontal stresses are being locked-in due to irreversible rock creep. The resulting ratio K of the horizontal to the vertical stress is given by

$$K = K_c + \frac{H_1 \gamma (K_c - K_o)}{\gamma H} \qquad (4)$$

Let us assume "transverse isotropy" in the plane of horizontal stress, so that

$$(\sigma_1 + \sigma_2)_H = 2\sigma_H = 2K\sigma_V = 2K\gamma H \qquad (5)$$

Equations (4) and (5) then give

$$(\sigma_1 + \sigma_2)_H = 2[K_c (\gamma H + \gamma_1 H_1) - K_o\gamma_1 H_1] \quad (6)$$

Using this model, Figure 4 shows the sum of horizontal stresses $(\sigma_1 + \sigma_2)_H$ versus depth for an ice sheet with $H_1 = 3\ 000$ m and $\gamma_1 = 9.5$ kN/m^3, and the rock properties $\gamma = 27$ kN/m^3 and $v = 1/6$ and $K_c = 0.3$, 0.6 and 1.0. The same figure shows the increase in vertical stress with depth for the elastic case $K_o = 0.2$ and the average horizontal stress of the Fennoscandian Shield from the Fennoscandian Rock Stress Data Base (FRSDB). The measured stress at the ground surface is consistent with a creeping rock mass and residual stress for $K_c = 0.3$. The average measured stress gradient versus depth is much steeper and not in accordance with the theory of creep induced stresses from ice loading. At this stage, an additional stress generating mechanism must be sought to explain the steep stress gradient and provide the additional stresses between the line for $K_c = 0.3$ and the measured horizontal stress in the Fennoscandian Shield. One possible mechanism is ridge push.

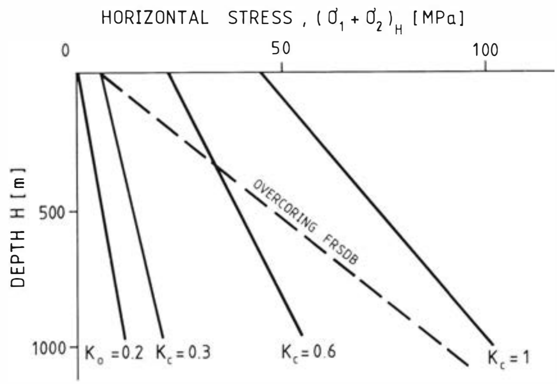


Fig. 4. Sum of horizontal stresses versus depth generated by assuming a model of an ice sheet of thickness 3 km and volume weight 9.5 kN/m^3. K_o and K_c are the resultant ratios of horizontal to vertical stress. Dashed line indicates natural stress gradient deduced from FRSDB by regression analysis. This gradient is steep and suggests an additional stress mechanism from a ridge push.

Stress systems generated by plate tectonics

Stress systems affecting the lithosphere can be divided into two main categories which will be refered to as "renewable" and "non-renewable" types (Bott and Kusznir, 1984).

Renewable stress systems are those which persist as a result of boundary or body forces, even though the strain energy is progressively dissipated. Stresses arising from plate boundary forces and from isostatically compensated surface loads are the two main contributors to stress in the Fennoscandian Shield. The persistant plate boundary forces cause the plates to move relative to each other. Geological evidence, thermodynamic considerations and comparison of observed stresses with predicted values all indicate that the plates are driven by boundary forces rather than by mantle drag through the bottom boundary. This means that the plate moves faster, in the order of a few centimetres per year for the Eurasian plate, than the underlying mantle. Mantle drag therefore tends to oppose plate motion rather than drive it. Bott and Kusznir (1984) distinguish three principal types of driving forces:
- ridge push (compression, 20–30 MPa)
- slab pull (tension, 0–50 MPa (?))
- trench suction (tension, 0–30 MPa(?))

Examples of simple stress states within lithospheric plates caused by plate boundary forces are shown in Figure 5. The most relevant model for the Fennoscandian continental crust of the Eurasian plate is given by example (b) in the figure. Ridge compressive forces at the Mid Atlantic Ridge, will change to trench tensile forces near the Mediterranean trench. Later in this section we will analyse a recent model to determine the stress distribution in the lithosphere.

For renewable stress systems, the lithosphere acts as a "strain energy reservoir" fed by the action of tractions and body forces, and is relieved at ap-

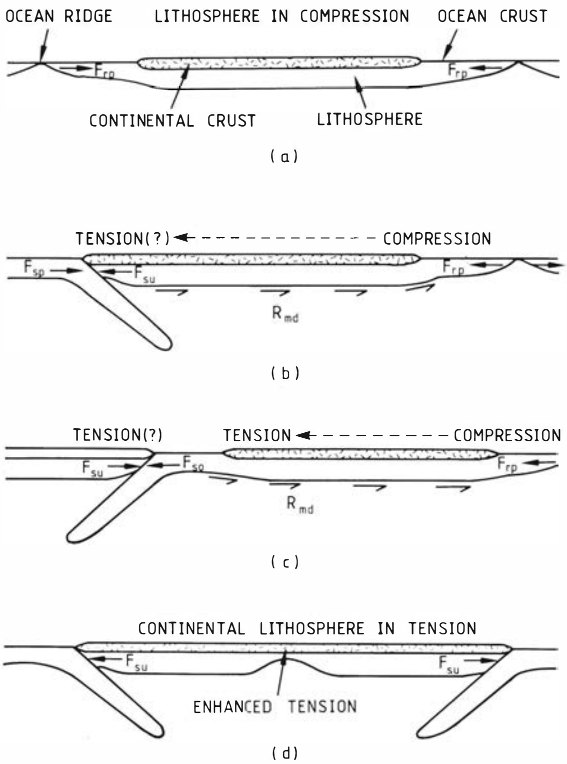


Fig. 5. Examples of simple stress systems within lithospheric plates caused by plate boundary forces (after Bott and Kusznir, 1984). A) Ridge push force developed at ocean ridges on opposite sides of a plate. This compresses the entire plate; example: present African plate. B) Ridge push force on one side of a plate and trench suction force on opposite side, causing stress system grading from compressive at the ridge to possible tensile at the trench (if local overriding plate resistance is high, compression may occur throughout the plate); example present South American plate and Fennoscandia. C) Ridge push force on one side of a plate and slab pull on opposite side, stress as in B); example Carboniferous basin formation in Great Britain. D) Trench suction on opposite sides of an entirely continental plate producing tension throughout; example Pangea just before its break-up.
F_{rp} = ridge push; F_{sp} = slab pull; F_{su} = trench suction; F_{md} = mantle drag.

proximately the same rate by tectonic activity. Loading of the lithosphere by surface topography, lateral density variation or ice sheets generates local stress fields.

Non-renewable stress systems are those which may be completely dissipated by release of the strain energy initially present. Bott and Kusznir (1984) list the following significant sources for non-renewable stress systems:

- bending stresses at subduction zones
- membrane stresses caused by changes in the radii of curvature of a plate as it migrates in latitude
- thermal stresses due to temperature changes in the lithosphere
- other mechanisms, e.g. phase transition, tidal (10^{-3} MPa).

Of these stress generating systems, membrane stresses are thought to be the only stresses of relevance to crustal rock mechanics or neotectonics in the Fennoscandian Shield.

Models of global stress

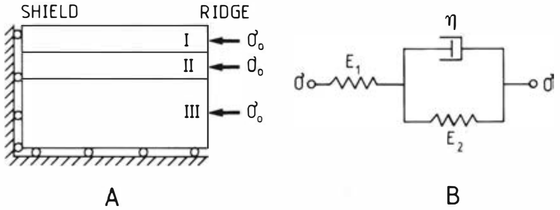
Global stress models have been calculated for a variety of possible driving forces. One of the first examples of intraplate stress modelling using a finite element technique was presented by Stephansson and Berner (1971). Based on the gravity model by Talwani, they modelled a section of the crust east of the Mid Atlantic Ridge. It was found here that most of the stresses were transmitted in the crust, and also that very low deviatoric stresses appeared in the mantle. The ridge compressive force was created by the light material and the elevation at the Mid Atlantic Ridge.

Salomon et al. (1980) presented global intraplate stress models by using a finite element method in which the effects of a wide variety of possible driving force combinations could be simulated. The most realistic global stress models include ridge pushing forces as an essential element. As in the modelling conducted by Stephansson and Berner (1971), the horizontal stress at the ridge was generated by the ridge elevation. In one possible global model that provides a reasonably good fit with most of the intraplate stress orientation data, the following forces are included:

(1) symmetric pushing force at the ridge, of 10 MPa across a plate 100 km in thickness
(2) a pulling force at the trenches of magnitude 10 MPa
(3) a resistive force of magnitude 10 MPa at the continental collision zone
(4) a drag stress proportional to the plate velocity.

The magnitude and orientation of the principal horizontal deviatoric stress for that particular model of plate driving forces gave a pronounced NW-SE direction of the principal horizontal deviatoric stress in the lithosphere of Fennoscandia.

The facts that viscosity approaches infinity in the brittle and elastic upper lithosphere, and that viscosity is finite and decreases with depth in the lower lithosphere, means that applied push forces at the



LAYER	THICKNESS H (km)	YOUNG'S MODULUS $E_1 \cdot 10^{11}$ (Pa)	YOUNG'S MODULUS $E_2 \cdot 10^{8}$ (Pa)	POISSON'S RATIO ν	VISCOSITY η (PaS)
I	20	1.053	1.0	0.219	10^{60}
II	20	1.284	1.0	0.197	10^{26}
III	60	1.802	1.0	0.263	10^{23}

C

Fig. 6. Linear viscoelastic modelling of lithosphere, redrawn from Hasegawa et al. (1985). A) Finite element model of: I, upper crust, II, lower crust, and III, upper mantle subjected to initial horizontal plate tectonic stress σ_0. B) Viscoelastic model. C) Parameters for three-layered crust-upper mantle model of Canadian Shield.

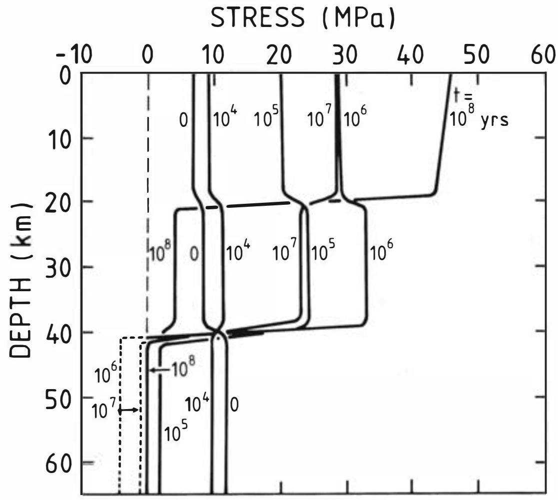


Fig. 7. Temporal pattern of horizontal deviatoric stress component for model in which 10 MPa horizontal stress is applied to boundary on right to depth of 100 km. After Hasegawa et al. (1985).

ridges will be concentrated in the upper lithosphere as a result of creep and stress decay in the lower lithosphere. This effect of stress amplification in the upper lithosphere of a shield area was studied by Bott and Kusznir (1984). After an initial application of a uniform compressive stress of 10 MPa across a 150 km thick lithosphere, and assuming a power law creep model, they obtained an interesting stress distribution. Here, stress relaxation by creep in the lower lithosphere resulted in progressive amplification of the upper lithospheric stress and gave rise to stress differences of the order of 20–25 MPa over a time span of 1–100 Ma.

The most recent and also very attractive model of upper crustal stresses and vertical stress migration of a shield type lithosphere has been conducted by Hasegawa et al. (1985). Based on existing knowledge of the rheology of the crust and the upper mantle they studied a three-layered model of the Canadian Shield. Since the overall geological evolution of Canada and Fennoscandia is very similar, the model is most applicable to the crustal rock mechanics of the present study. Figure 6A shows one of the three-layered plane strain models selected for finite element calculations. The boundary conditions are kept as simple as possible, with a free surface and fixed boundaries at the shield and bottom of the upper mantle, respectively. A spreading ridge stress of $\sigma_0 = 10$ MPa is applied to the continental lithosphere; the vertical extent of this applied stress corresponds to the thickness of the oceanic lithosphere. The effect of gravity is omitted in the calculations.

The linear rheological model selected here is kept as simple as possible (figure 6B). Figure 7 illustrates the temporal and spatial variation of deviatoric horizontal stress in the three-layered model from time t = 0 to t = 10^8 years. After t = 10^5 years the stress in the upper mantle has relaxed entirely. Because of mechanical coupling of the layers, the relaxed stress in the upper mantle is now "shouldered" by the crust, and at t = 10^8 years induced stress in the lower crust "migrates" to the upper crust. Thus, the model illustrates the tectonic process whereby the maximum horizontal stress migrates upward, resulting in a stress amplification of 40–50 MPa in the upper crust after a time span of about $2 \cdot 10^8$ years starting from the most recent opening of the Atlantic ocean.

Based on the results from the finite element modelling of the elastic-viscoelastic relaxation and contributions from other stress generation mechanisms, Hasegawa et al. (1985) constructed a composite stress diagram for the upper (0–20 km) and the lower (20–40 km) crust in eastern Canada. As stated earlier, there is a strong possibility that the stress state in the eastern Canadian Shield is similar to that of Fennoscandia, and the relaxation model of Hasegawa et al. (1985) has therefore been applied to this study.

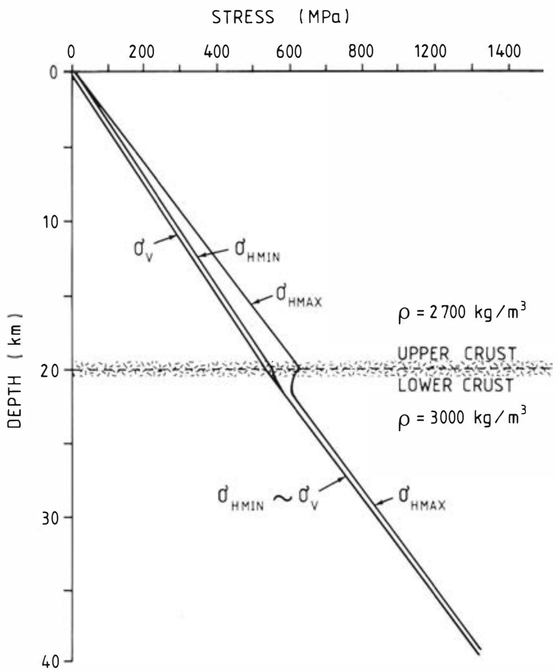


Fig. 8. Composite stress diagram for upper and lower crust in Fennoscandia. Stress gradients are consistent with shallow stress measurements and fault plane solutions which are predominantly of thrust fault type. Absolute stress levels are obtained from the stress diagram of Hasegawa et al. (1985).

A composite stress diagram for the upper and lower crust in Fennoscandia is shown in Figure 8. The modelled variation of σ_{HMAX}, σ_{HMIN} and σ_V with depth is consistent with the measured stresses in the Fennoscandian Shield (Stephansson et al., 1987). Earthquake fault plane solutions by Slunga et al. (1984) provide additional support for this variation of stresses with depth. Many earthquakes are associated with a dominant strike-slip component that would imply that either σ_{HMIN} is close to σ_V, or is locally less than σ_V. This is the principal reason for drawing σ_{HMIN} close to σ_V in the composite stress diagram. For Fennoscandia, the following stress variations with depth are suggested:

Upper crust, z = 0–20 km

$$\begin{aligned} \sigma_{HMAX} &= 5 + 32\,z \\ \sigma_{HMIN} &= 2 + 28\,z \\ \sigma_V &= 27\,z \end{aligned} \tag{7}$$

Lower crust, z = 20–40 km

$$\begin{aligned} \sigma_{HMAX} &= 560 + 30\,(z-20) \\ \sigma_{HMIN} &= 540 + 30\,(z-20) \\ \sigma_V &= 540 + 30\,(z-20) \end{aligned} \tag{8}$$

where the depth, z, is in kilometres and the stresses are expressed in megapascals.

The vertical stress is taken to be equal to the overburden stress, $\sigma_V = z \cdot \varrho \cdot g$, and the horizontal differential stresses are measured from this datum. The combined contribution of the spreading ridge stress and viscoelastic relaxation is uncertain to within a factor of 3. Residual stress due to incomplete post-glacial rebound is of the order of a few megapascals, for an estimated remaining uplift of about 150 m. The contribution from the non-renewable membrane stress is of the order of 10 MPa for a viscoelastic membrane. A differential stress that increases linearly with depth and is observed in many regions is supposed to be caused by basal drag (Figure 1). The composite stress field indicates a differential stress that varies from a few tens of megapascals at 5 km to about 100 MPa at a depth of 20 km, i.e. at the base of the upper crust. Below the upper and lower crustal interface, the horizontal differential stress is governed solely by the spreading ridge stress of 10–30 MPa.

Discussion

Crustal stresses in the Fennoscandian Shield have been assessed in this paper. It has been shown that the excess horizontal stress in the upper crust cannot be explained solely on the basis of glacially induced stresses. The remaining isostatic rebound from the Weichselian glaciation causes an additional horizontal stress of the order of a few megapascals. The locked-in stresses from rock creep after deglaciation, can, to some extent, explain the presence of excess of horizontal stress in the upper part of the crust, but the much steeper gradient of the measured maximum horizontal stress must also be explained. By superposition of the two stress generating mechanisms due to glaciation we get closer to the natural situation, but still the magnitudes of the stresses generated are insufficcient to explain the magnitudes of stresses measured in boreholes. It is therefore suggested that the remaining excess horizontal stress is caused by the ridge push applied by the Mid Atlantic Ridge. At this stage of the analysis of virgin stresses in the Fennoscandian Shield, there is no method by which the three stress-generating mechanisms may be distinguished. More data are needed concerning the stresses applied at the spreading ridge.

A survey of the orientation and magnitude of the stresses in Fennoscandia reveals that the magnitude of horizontal stresses increases from the Caledonides in the west to the Archean terrain in the east (Fig. 3A). One must bear in mind that the stresses and the orientations have been obtained from differnet depths and also measured in different rock types. However, a stress increase would be expected towards the west, since stress magnitudes in mountainous areas are enhanced by the topography (Swolfs and Savage, 1986). Amplified stress magnitudes would also explain the recurrent faulting and seismicity along the axial protion of the Caledonian mountain range. Although the seismicity is enhanced along the Norwegian coast (Gregersen, 1986) there is no indication of any excess of horizontal stresses, cf. Figure 3A. The most plausible explanation for the amplification in stress magnitudes eastwards is the much longer time for vertical migration of stress from the mantle to the upper crust in the old Archean (> 2.5 Ga) and Svecokarelian (1.9 – 1.7 Ga) rock to the east, as compared with that required in the much younger Caledonian (~ 0.4 Ga) rocks in the west. The enhancement of the stresses in the old rocks can also be explained to some extent by their higher stiffness and hence their increased ability to store the migrated stresses from the mantle, cf. Figure 3A.

Savage et al. (1986) have shown that the stress field induced in a anisotropic rock mass under gravity and vanishing horizontal displacements depends on the type and magnitude of the rock mass anisotropy and the orientation of the rock fabric with respect to the ground surface. The gravity-induced stress distributions predicted by their models are very similar to stress-ratio-with-depth plots published in the literature. These models are valid for stratified rock masses or regularly jointed rock masses in which joints are horizontal or vertical. For rock masses with vertical rock fabrics the authors were able to show that the induced stress field is triaxial. Although the magnitude of the stress field induced under gravity is strongly affected by the rock mass structure, the irregularly faulted and jointed Precambrian rock in Fennoscandia would reduce the possibility of generating a positive anomaly of horizontal stress. If the assumption of lateral restraint is fulfilled, a certain portion of the horizontal stress might originate from the influence of rock fabrics on gravity-induced stresses. At this stage of knowledge about the state of stress in the earth is crust, we cannot assess the importance of rock fabric and are unable to determine this contribution relative to other stress generating mechanisms.

Although the maximum horizontal stress in Fennoscandia exceeds the vertical stress, the very large scatter in the data obtained in the uppermost part of the crust must still be explained (Fig. 2). There are several possible explanations for this, one being the probable irregularity of natural fracture patterns and its influence on gravitational loading. Another possibility is the change in magnitude and direction of stresses near faults. Yet another explanation could be the large scatter in the frictional strength of discontinuities in the uppermost part of the earth's crust. Frictional sliding can be expected to occur at any time on a fault plane if the magnitude of shear stress along the fault plane, τ, is greater than or equal to the frictional resistance to sliding, $\mu\ \sigma_n$, where μ is the frictional strength and σ_n is the normal stress acting on the fault plane. A comprehensive summary of laboratory data on friction by Byerlee (1978), for a wide variety of rock types, indicates that values of μ for most rocks range between 0.6 and 1.0. This range of values for the coefficient of friction is valid for normal stresses up to 100 MPa. In-situ stress measurements made at depth in areas of active faulting have confirmed these results (Zoback and Healy, 1984). This variation of μ alone can cause considerable differences in the stress state in Fennoscandia and elsewhere.

In reviewing the results of laboratory-scale testing of rock and rock joints, Barton (1977) found that rock joints exhibit a wide range of shear strengths under low effective normal stress levels. Conversely, under high effective normal stresses shear strength for joints varies little, despite large variations in the triaxial compressive strength of rocks at the time of fracture. This supports the general trend obtained for the variation in the ratio of average horizontal to vertical stress with depth shown in Figure 2. Close to the bedrock surface, where the normal stress on the fracture plane is low, there is a large scatter of the values obtained for the stress ratio of horizontal versus vertical stress (Fig. 2). The ability of faults to transmit shear stresses depends strongly on the roughness of the fault surface. At greater depth, stress transmission across faults is determined by the shear strength of the asperities on the fault surface.

Acknowledgements. – The author wishes to thank Kjell Bergström and L-O Dahlström for their help in extracting information from the Fennoscandian Rock Stress Data Base. Göran Olofsson typed the manuscript, Monica Leijon prepared the drawings and Peter Digby corrected and improved the English of the manuscript. Christopher Talbot, Peter Cobbold, and an unidentified referee made valuable comments. I am thankful for their work. Discussions with members of the project group on Rock Mass Stability at Swedish Nuclear Fuel and Waste Management Co. have been most valuable to this work.

The work presented in this paper was funded jointly by the National Swedish Board for Technical Development, Grant No. 85 – 4791, and the Swedish Natural Science Council, Grant no. 3447 – 136.

REFERENCES

Barton, N.R. 1977: The shear strength of rock and rock joints. *Int. J. Rock Mech. Min. Sci. & Geomech. Abstr. 13,* 255–279.

Bergman, S.G.A. 1977: Rock stress measurement in the bedrock of Scandinavia, conditions, results and interpretation. *SKB Report 64.* Swedish Nuclear Waste Company, Stockholm, 24 p. (In Swedish)

Bjarnason, B. & Stephansson, O. 1987: Non-linear and discontinuous stress variation with depth in the upper crust of the Baltic Shield. *Proc. 6th Int. Cong. Rock Mech.,* Montreal, Canada.

Bott, M.H.P. & Kusznir, N.J. 1984: The origin of tectonic stress in the lithospere. *Tectonophysics 105,* 1–13.

Brown, E.T. & Hoek, E. 1978: Trends in relationships between measured in-situ stresses and depth. *Int. J. Rock Mech. Min. Sci. & Geomech. Abstr. 15,* 211–215.

Byerlee, J.D. 1978: Friction of rock. *Pure application. Geophys. 116,* 615–626.

Gregersen, S. 1986: Crustal inhomogeneities and seismicity near the margins of the North Atlantic Ocean. *J. Geodyn. 6,* 5–12.

Hasegawa, H.S., Adams, J. & Yamazaki, K. 1985: Upper crustal stresses and vertical stress migration in eastern Canada. *J. Geophys. Res. 90,* 3637–3648.

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

Herget, G. 1986: Changes of ground stresses with depth in the Canadian Shield. In: O. Stephansson (Ed) Proc. Int. Symp. on rock Stress and Rock Stress Measurements, Stockholm, 1–3 Sept., 1986. Centek Publ., Luleå, 61–68.

Klein, R.J. & Barr, M.V. 1986: Regional state of stress in western Europe. In: O. Stephansson (Ed) Proc. Int. Symp. on Rock Stress and Rock Stress Measurements, Stockholm, 1–3 Sept., 1986. Centek Publ., Luleå, 31–44.

Kukkari, J. 1986: Newest results obtained in studying the Fennoscandian land uplift phenomenon. *Tectonophysics, 130,* 327, 331.

Martna, J., Hiltscher, R. & Ingevald, K. 1983: Geology and rock stresses in deep boreholes at Forsmark in Sweden. *Proc. 5th Int. Cong. Rock Mech.,* Melbourne, Australia, F111–F116.

Martna, J. & Hansen, L. 1986: Initial rock stresses around the Vietas headrace tunnels Nos. 2 and 3, Sweden. In: O. Stephansson (Ed) Proc. Int. Symp. on Rock Stress and Rock Stress Measurements, Stockholm, 1–3 Sept., 1986. Centek Publ., Luleå, 605–613.

Ranalli, G. & Chandler, T.E. 1975: The stress field in the upper crust as determined from in-situ measurements. *Geol. Rundschau 64,* 653–674.

Salomon, S.C., Richardson, R.M: & Bergman, E.A. 1980: Tectonic stress: Models and magnitudes. *J. Geophys. Res. 85,* 6086–6092.

Savage, W.Z., Amadei, B.P. & Swolfs, H.S. 1986: Influence of rock fabric on gravity-induced stresses. In: O. Stephansson (Ed) Proc. Int. Symp. on Rock Stress and Rock Stress Measurements, Stockholm, 1–3 Sept., 1986. Centek Publ., Luleå, 99–110.

Slunga, R., Norrman, P. & Glans, A–C. 1984: Baltic Shield seismicity, the result of a regional network. *Geophys. Res. Letters 11,* No. 12, 1247–1250.

Stephansson, O. & Berner, H. 1971: The finite element method in tectonic processes. *Phys. Earth Planet. Interiors 4,* 301–321.

Stephansson, O. & Ångman, P. 1986: Hydraulic fracturing stress measurements at Forsmark and Stidsvig, Sweden. *Bull. Geo. Soc. Finland* 58, Part 1, 307–333.

Stephansson, O., Särkkä, P. & Myrvang, A. 1986: State of stress in Fennoscandia. In: O. Stephansson (Ed) Proc. Int. Symp. on Rock Stress and Rock Stress Measurements, Stockholm, 1–3 Sept., 1986. Centek Publ., Luleå, 21–32.

Stephansson, O., Dahlström, L-O., Bergström, K., Särkkä, P., Väätäinen, A., Myrvang, A., Fjeld, O. & Hansen, T.H. 1987: Fennoscandian rock stress data base – FRSDB. *Research Report LULEA* 1987:06, Luleå University, Luleå.

Swolfs, H.S. & Savage, W.Z. 1986: Topographic modification of in-situ stress in extensional and compressional tectonic environments. In: O. Stephansson (Ed) Proc. Int. Symp. on Rock Stress and Rock Stress Measurements, Stockholm, 1–3 Sept., 1986. Centek Publ., Luleå, 89–98.

Voight, B. 1966: Restspannungen in Gestein. *Proc. 1st Int. Cong. Rock Mech.,* Lissboa, Portugal, II, 45–50.

Walcott, R.I. 1970: Isostatic response to loading of the crust in *Canada. Can. J. Earth Sci. 7,* 716–727.

Zoback, M.D. & Healy, J.H. 1984: Friction, faulting and in-situ stress. *Annales Geophysicae 2,* 689–698.

Printed by Ekenäs Tryckeri Ab, Ekenäs, Finland December 30th, 1988