NUMERICAL LITHOSPHERIC MODELLING: RHEOLOGY, STRESS AND DEFORMATION IN THE CENTRAL FENNOSCANDIAN SHIELD

KARI MOISIO

Faculty of Science,
Department of Physical Sciences,
Division of Geophysics,
University of Oulu

OULU 2005

KARI MOISIO

NUMERICAL LITHOSPHERIC MODELLING: RHEOLOGY, STRESS AND DEFORMATION IN THE CENTRAL FENNOSCANDIAN SHIELD

Academic Dissertation to be presented with the assent of the Faculty of Science, University of Oulu, for public discussion in Kajaaninsali (Auditorium L6), Linnanmaa, on December 16th, 2005, at 12 noon

OULUN YLIOPISTO, OULU 2005

Supervised by
Professor Pertti Kaikkonen

Reviewed by
Professor Sierd Cloetingh
Professor Giorgio Ranalli

ISBN 951-42-7950-6 (nid.)
ISBN 951-42-7951-4 (PDF) http://herkules.oulu.fi/isbn9514279514/

ISSN 0355-3191 http://herkules.oulu.fi/issn03553191/

OULU UNIVERSITY PRESS
OULU 2005

Moisio, Kari, Numerical lithospheric modelling: rheology, stress and deformation in the central Fennoscandian Shield
Faculty of Science, Department of Physical Sciences, Division of Geophysics, University of Oulu, P.O.Box 3000, FIN-90014 University of Oulu, Finland
2005
Oulu, Finland

Abstract

This thesis deals with the analysis of the rheological structure and tectonic modelling of the Fennoscandian Shield. First, a short introduction to the geology and geophysics of the Fennoscandian Shield is presented followed by a description of rheological concepts. Second, the applied modelling procedures, together with the sources of error are explained. Last a brief summary of each original paper including conclusions is given.

Understanding rheological conditions through the entire lithosphere and even deeper is the key for understanding the deformation of the earth's interior. Thus, investigating the rheological structure and possible consequences resulting from tectonic loading are required to some extent when interpreting geophysical data into tectonic models.

In this thesis rheological structure is obtained by calculating rheological strength in different locations of the central Fennoscandian Shield. These locations are mainly situated along different deep seismic sounding (DSS) profiles as they provide necessary geophysical information required for model construction. Modelling begins by solving the thermal structure in the lithosphere, as rheological behaviour, mainly ductile flow is strongly controlled by temperature. Results from these calculations show that the rheological structure of the lithosphere depends on the thermal conditions resulting in significant areal variations. Generally, the central Fennoscandian Shield can be considered to be rheologically rather strong. Rheologically weak layers are however usually found in the lower crust. Correlation of the rheological structure with earthquake focal depth data shows that brittle fracture is the relevant mechanism in the earthquake generation and that non-occurrence of deep earthquakes implies low stress or high strength conditions deeper in the crust.

Calculated rheological structure is furthermore used as a material parameter in the structural models which are solved next. These results suggest that it is highly unlikely that any considerable ductile deformation in the crust of the central Fennoscandian Shield exists and it seems that the present-day thermal and mechanical conditions in the investigated area do not favour such processes in significant amounts.

Keywords: finite-element method, rheological strength

Acknowledgments

Studies for this thesis were done during the years 1996-2005. In the beginning I studied under the Lithospheric Graduate School organization and later on as teaching personnel in the Division of Geophysics at the University of Oulu. I have crossed paths with several colleagues since the beginning of my studies and would not have been able to carry out all of this work alone. I would like to express my deepest gratitude to all of those who have been involved in this work.

Especially, I would like to mention the supervisor of this work, Professor Pertti Kaikkonen, who first introduced me to this field of geophysics and offered me the possibility to do a masters thesis. Without his contribution this work would not have been possible. Also, Professor Sven-Erik Hjelt should be mentioned as he was responsible for organising the Lithospheric Graduate School which gave me the opportunity to continue my studies in the field of geophysics. Colleagues in Amsterdam in the early years of my post graduate studies helped me greatly to really get this work started. I thank professors Giorgio Ranalli and Sierd Cloetingh for reviewing this thesis. Also, all personnel in the facilities of the Division of Geophysics during these several years must be mentioned as they have in many ways helped me to carry out this long-lasting work. I thank CSC, the Finnish IT center for science for providing the required software and hardware for calculations and Mr. Aaron Bergdahl for revising the language of the manuscript Special and dearest thanks to my family, Anu, Niklas and Oona who during these many years gave me something else to think about. You light up my life.

Oulu, November 2005 Kari Moisio

List of the original papers

This thesis consists of a summary and the following five original papers that are referred to in the text by their Roman numeral:

I Kaikkonen P, Moisio K & Heeremans M (2000) Thermomechanical lithospheric structure of the central Fennoscandian Shield. Physics of the Earth and Planetary Interiors 119: 209-235.

II Moisio K, Kaikkonen P & Beekman F (2000) Rheological structure and dynamical response of the DSS profile BALTIC in the SE Fennoscandian Shield. Tectonophysics 320: 175-194.

III Moisio K & Kaikkonen P (2001) Geodynamics and rheology along the DSS profile SVEKA'81 in the central Fennoscandian Shield. Tectonophysics 340: 61-77.

IV Moisio K & Kaikkonen P (2004) The present day rheology, stress field and deformation along the DSS profile FENNIA in the central Fennoscandian Shield. Journal of Geodynamics 38: 161-184.

V Moisio K & Kaikkonen P (2005) Three-dimensional numerical thermal and rheological modelling in the central Fennoscandian Shield. Submitted to Journal of Geodynamics.

Contents

1 Introduction

During the last decades numerical modelling has been a useful and a powerful tool used in geophysical investigations. In the early years numerical methods, e.g., finite element and finite difference methods were commonly used for solving partial differential equations in engineering applications. Soon after, geophysicists realised the usefulness of these methods, e.g., in solving electromagnetic problems such methods have been widely used for over 30 years. The most severe limitations in those days came mostly from computer hardware, which were primitive in the sense of application runtimes. As computer capabilities have since developed rapidly, this limitation has more or less vanished. At present it is possible to solve more and more complicated and larger numerical models in a shorter time frame. This has widened the possibilities of numerical methods to apply to all areas of geophysics. Geodynamical modelling of tectonic problems is one of the latest applications in geophysical and geological research. Thin-sheet modelling of large-scale stress fields, basin formation and subduction zone modelling are some of the tectonic cases where numerical methods have been used successfully. Rheology, describing how material deforms, is one of the most important aspects to consider when studying geodynamic processes. It was originally defined as the study of flow, but in geosciences it is used when discussing rock deformation mechanisms. Viscous flow, elasticity, fracture and creep mechanisms are the deformation modes that describe most of the rheological behaviour in the earth.

This thesis is based on this summary and five separate papers where thermal, rheological and tectonic, i.e., structural modelling has been applied. The main purpose has been to investigate thermal and rheological structure together with analysing the possible state of stress and deformation in the Fennoscandian Shield. First, the research area is reviewed with known geophysical conditions together with the geological setting. The next chapter discusses how rheology is defined, explaining the most important deformation mechanisms and commonly used rheological equations, followed by chapters describing the modelling methods and error sources. Last, a brief summary of each original paper is given together with conclusions.

2 The Fennoscandian Shield: geology and geophysics

The Fennoscandian Shield, surrounded by the Caledonides in the west and young Phanerozoic sediments in the south, can be divided into three domains: the Archaean, the Svecofennian and the Scandinavian (Gaál & Gorbatschev 1987). Our studies have mostly concentrated on the Svecofennian and the Archaean domains. The oldest Archaean domain occupies the eastern part of the Shield and was formed during the Saamian (3.1–2.9 Ga) and the Lopian (2.8–2.6 Ga) orogenic events. The youngest Scandinavian domain occupies a relatively small area in the southwest and was formed during the Gothian event (1.75–1.5 Ga). The Svecofennian domain occupies the rest of the Shield and was formed by an accretional type orogeny, called the Svecofennian, by deformation and high-grade metamorphism and by extensive crustal melting during the period of 1.9–1.55 Ga (Windley 1992). It is believed to have been a collision between the Archaean Karelian craton and a Palaeoproterozoic island arc complex which produced large amounts of juvenile mantle derived crust in subduction like processes (Huhma 1986). The Lake Ladoga-Bothnian Bay zone (LLBB) is a boundary between the Archaean and the Proterozoic areas. This boundary has always been the focus of great geological and geophysical interest due to sulphidic ores found in it.

Geophysical investigations have been carried out quite intensively not only in the vicinity of the LLBB zone but also in the whole of the Fennoscandian Shield. Seismic, electromagnetic, thermal and gravity investigations have been conducted to study the large-scale deep structures, to say nothing of the experiments done for economic purposes. Geophysical investigations are a vital source of data needed in rheological and geodynamical calculations. They also provide valuable information as reference data for comparing calculated models with in-situ data.

Seismic studies include, for example, the deep seismic sounding (DSS) profiles FENNOLORA (e.g., Guggisberg et al. 1991), BALTIC (Luosto et al. 1985, 1990), Sovetsk-Kohtla-Järve (SKJ) profile (Sharov et al. 1989), SVEKA (Grad & Luosto 1987, Luosto et al. 1994, Yliniemi et al. 1996), BABEL (BABEL Working Group 1993a, 1993b, Korja & Heikkinen 1995), POLAR (Luosto et al. 1989, Luosto 1991), FENNIA (FENNIA Working Group 1996), Pechenga-Kovdor-Kostomuksha (Azbel et al. 1989) and the quarry blasts profiles in the central part of Finland (Yliniemi 1991). Seismic tomography research for the central Fennoscandian Shield, i.e. the SVEKALAPKO project, was accomplished in the late 1990's with 143 seismic stations deployed in an array of 1000 km by 900 km. That study focused on the deep structure of the lithosphere-asthenosphere system up to the depth of 400 km (Sandoval et al. 2003, 2004).

Geothermal research has been carried out quite intensively in the Fennoscandian Shield, especially in Finland. Thermal data and numerous modelling results have been presented, e.g., by Puranen and co-authors (1968), Järvimäki and Puranen (1979), Kukkonen (1988, 1989, 1993, 1994, 1995, 1996, 1998), Pasquale and co-authors (1991), Cermák and co-authors (1993), Kukkonen and Jõeleht (1996) and Jokinen and Kukkonen (2000). Also, studies for evaluating crustal heat production in the Fennoscandian Shield have given valuable input data for thermal models (Kukkonen 1989, 1998, Jõeleht & Kukkonen 1998, Kukkonen & Lahtinen 2001).

Gravity investigations of Puranen and co-authors (1978), Lähde (1985), Elo and Korja (1993), and Elo (1997) provide important information about the density of the rocks in the Fennoscandian Shield. The deep geoelectric structure is best resolved with magnetotelluric studies. For the Fennoscandian Shield, these results are presented by Ádám and co-authors (1982), Hjelt and Korja (1993), Kaikkonen and co-authors (1983), Korja and Koivukoski (1990, 1994), Korja and Hjelt (1993), Korja and co-authors (1993) and Korja (1997).

Earthquake activity in Finland is generally quite low and the magnitudes of the earthquakes are relatively small, usually a local magnitude (M_L) of around 2. However, analyses of these events provide valuable data in the form of focal depths. These data give indications about rheological conditions in the crust and can be used for verifying rheological models (Ahjos & Uski 1991, 1992). Analyses of the earthquake source mechanism provide information about the tectonic setting in the vicinity of the event. Fault plane solutions for some of the Finnish earthquakes are presented in Uski et al. (2003). The number of these solutions is quite low due to the sparse coverage of seismic stations and the small magnitudes of earthquakes in Finland, but lately improvements have been made in the upgrading of seismic station instruments and station coverage and also fault plane solution techniques (Uski et al. 2003). Also, geodetic observations (Chen 1991, Kakkuri 1997) provide necessary information for understanding the existing stress conditions in the Fennoscandian Shield.

3 Rheology

3.1 Basic concepts

In geology and geophysics the term rheology refers to the study of mechanical properties and their role in the deformation and the flow of the materials that form the Earth. We can describe the Earth's rheology mathematically in many ways, but it is always a function of intrinsic, i.e., material parameters, and extrinsic, i.e., environmental parameters. Basic rheological equations describing deformation are called constitutive equations, which relate stress and strain. The stress is a result of the force acting on a surface surrounding or within a body, and comprises both the force and the reaction of the material on the other side of the surface (Park 1989). The state of the stress at a point can be expressed by the stress tensor, which has nine components. Stresses cause solids to deform, i.e., stresses produce changes in the distances separating neighbouring small elements of solids. The amount of deformation of a body is expressed by strain. Strain is the change in size and shape of a body resulting from the action of an applied stress field. Constitutive equations are usually derived empirically in laboratory conditions. Many earlier calculations were based on an approach where materials were assumed to behave elastically. It is still a valid assumption in many cases, e.g., in thin sheet modelling.

3.2 Deformation mechanisms

Hooke's law describes elastic behaviour, where the stress is linearly proportional to the strain. Characteristic for the elasticity is an instantaneous and total recovery of the deformation upon application and removal of a load, respectively. Quite often upper crustal rocks residing in low temperature and pressure show elastic behaviour. Also, seismic waves can be treated as elastic phenomena.

In rheology the term solid refers to a material that does not yield if the stress is below a certain threshold value whereas a material that shows the steady-state flow under a constant stress is called a fluid. If material is behaving like a fluid different constitutive equations must be used. The characteristic material parameter for a fluid is viscosity, which is a measure of the resistance of a fluid to flow. The term creep refers to a very slow flow under a constant stress, i.e., the materials that are creeping can actually be considered as fluids over a long time scale. In geodynamics, viscous behaviour is often applied for modelling the deformation of the mantle. Either the Newtonian or the non-Newtonian description can be used. Newtonian fluid has a linear relationship between the strain rate and the stress whereas the non-Newtonian fluid has a non-linear, quite often a power-law dependence, where the strain rate is proportional to the nth power of the stress. The latter is assumed to be very common at high temperatures. In addition, between elastic and viscous deformation a stage of a transient, i.e., time dependent flow, can exist. Combinations of these three phases can be modelled with analogue models like the Maxwell's viscoelastic body and the Burgers general linear body (see e.g., Mase 1970, Ranalli 1995).

Failure of material takes place when permanent deformation occurs. Material fails when deviatoric stress reaches a certain critical value called the yield strength or the yield stress. Failure can occur as discontinuous deformation, i.e., fracturing or continuous irrecoverable deformation, i.e., plastic flow (Ranalli 1995). Material that fails by fracturing is called brittle and material that fails by a plastic flow is termed ductile. Yielded flow and fluid flow can be considered similar, with the exception that the former flow occurs only when the yield stress is reached. Earthquakes, faulting and folding are thought to be examples of fracturing and plastic flow, which thus have a major role in geodynamics, especially in the upper lithosphere.

3.2.1 Brittle deformation

The upper lithosphere can be considered to have pre-existing fracture planes, faults, present. Brittle failure resulting from sliding along these fracture planes can be described using Amonton's law (Turcotte & Schubert 1982). The coefficient of friction is used as a measure of a resistance to sliding. It is expressed with shear stress and normal stress with the consequence that the shear stress required for sliding increases with increasing normal stress. Perhaps a more common mechanism of brittle failure in the upper lithosphere is shear fracture. This is often described using the Navier-Coulomb criterion (Scholz 1990, Ranalli 1995), which states that fractures occur across the plane where the shear stress first reaches a value dependent on the material parameters and on the normal stress. This failure criterion resembles Amonton's law but it refers to the formation of a new fracture plane in contrast to failure along pre-existing planes. Another difference is that the coefficient of internal friction is used as a measure for the increase in shear strength caused by the increase in normal stress. The Navier-Coulomb criterion is however also valid for sliding along pre-existing cracks as the cohesion, i.e., inherent shear strength of the rock is negligible compared to the cohesion of the solid rock. Also, if the coefficient of internal friction is replaced by sliding friction, the relation is reduced to Amonton's law (Ranalli 1995).

Faulting is the result of shear fracture and sliding under stress. Combining Amonton's law and the Navier-Coulomb criterion, we can derive a widely used relation for describing brittle failure. This is often presented by Byerlee's law, which can be written in terms of the principal stress difference, lithostatic pressure and pore fluid pressure (Sibson 1974, Ranalli 1995)

$$\sigma_1 - \sigma_3 = \alpha\rho g z(1-\lambda), \text{ where } \alpha = \begin{cases} \dfrac{\mathrm{r-1}}{\mathrm{r}}, & \text{normal fault} \\ \mathrm{r-1}, & \text{thrust fault} \\ \dfrac{\mathrm{r-1}}{1+\beta(\mathrm{r-1})}, & \text{strike slip fault} \end{cases} \tag{1}$$

$$\text{and } \mathrm{r} = \left[\left(1+\mu^2\right)^{1/2} - \mu \right]^{-2}, \tag{2}$$

σ_1 and σ_3 are the maximum and minimum principal stresses [Pa], respectively, ρ is the density [kgm^{-3}], g is the gravitational acceleration [ms^{-2}], z is the depth [m], α is a parameter depending on the faulting type, λ is the hydrostatic pore fluid factor, β is the magnitude of the intermediate stress (between 0 and 1), and μ is the sliding friction coefficient. This relation strictly applies to sliding along pre-existing faults and to the uppermost crust only. Shimada (1993) has concluded that, particularly for granites, a change in fracture mechanism from a typical Navier-Coulomb type to a high-pressure type takes place as pressure increases in the brittle regime. High pressure fracturing is not so dependent on pressure and it decreases as the temperature increases. This implies that the brittle zone should be divided into two different brittle regimes.

Brittle-ductile transition (BDT) is a regime in the lithosphere, above which rocks exhibit brittle deformation and below which they fail by plastic (or ductile) flow. This layer is probably gradual, where the deformation mechanism changes slowly. This zone is quite often correlated with the maximum depth of crustal earthquakes, as the earthquakes are assumed to be a consequence of brittle fracture. However, the real lower boundary of crustal earthquakes is connected to changes in frictional behaviour in the fault zone from stick-slip to stable sliding (from velocity weakening to velocity strengthening) (Scholz 1990). Correlation between the critical temperature and the BDT depth, together with a known distribution of earthquake focal depths, can be used for analysis and verification of the rheological models. This is why good and reliable earthquake data is valuable. The critical temperature for a frictional behaviour change for granitic composition is between 300 and 400 °C (Scholz 1990, Blanpied et al. 1991). Fig. 1 shows the BDT depth, depth of the critical temperature of 350 °C and the earthquake epicentres with focal depths based on data from Paper I. However, earthquake data in Fig.1 has been updated and it covers the time from the beginning of 1965 until the end of January 2005.

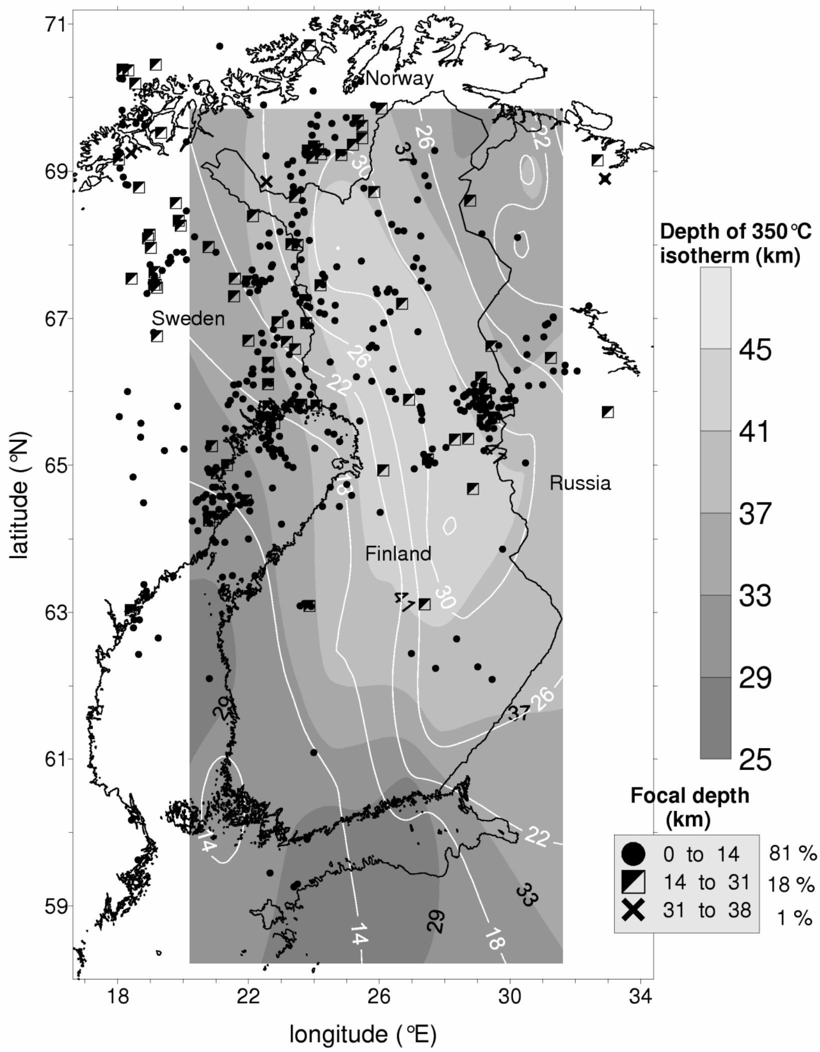

Fig. 1. The depth (km) distribution of the crustal brittle-ductile transition (BDT) (white contour lines) on the contour map of the depth (km) of the 350 ºC isotherm for the compressional wet rheology. The percentage distribution of focal depths classified into three groups is also shown in the figure. (updated from Paper I).

3.2.2 Ductile deformation

Ductile flow occurring in lithospheric rock is usually described using empirical relationships, i.e., creep equations. These empirical relations are verified to lithospheric conditions, which are usually beyond any laboratory limits, by microphysical models of creep. The two most important classes of creep mechanism are diffusion creep and dislocation creep (see e.g., Turcotte & Schubert 1982, Kirby 1983, Tsenn & Carter 1987, Ranalli 1995). Diffusion creep results from the diffusion of atoms and crystal lattice vacancies through the interior of crystal grains or along grain boundaries subjected to a deviatoric stress. These are thermally activated processes and they exhibit a linear relationship between stress and strain rate and they can be considered equivalent to Newtonian fluid flow. Typically these mechanisms are connected to low stress regimes but they are also grain size dependent. Low grain size (< 100 μm) favours diffusion creep which enables the existence of creep, for example in the shear zones.

In higher stress regimes the dislocation creep is often dominant although some amount of diffusion creep can also be influential to the overall strain rate. Dislocation creep is a result of a motion of dislocations in the crystal lattice structure. The dislocations are defined as imperfections in the crystal structure with two basic end members, edge and screw dislocations. Two principal classes are dislocation glide and dislocation climb. These mechanisms are also thermally activated. For the dislocation glide, activation can also be attained by an increase in stress. This means that there is a finite strength in the rocks deforming in such a manner. Ductile creep occurs only when this critical value is exceeded either by change in temperature and/or stress. At high temperatures creep is a result of thermal activation only and the strength of the rock has vanished. Deformation beyond this point is characterised by some viscous flow dependent on physical and the mechanical conditions. Dislocation creep has an exponential dependence on temperature and pressure with non-linear stress and strain rate relation in a power-law form. Dislocation creep can be considered as equivalent to non-Newtonian flow if the required activation level is exceeded.

Plastic flow is a result of failure in rocks. This is usually connected to the dislocation glide creep. Ductile steady-state (constant rate of flow under constant stress) properties of a material can be described by an empirical constitutive equation, i.e., the power-law creep, which is dominant in high stress and temperature conditions. The ductile flow law gives the stress difference necessary to maintain a given strain rate (Kirby 1983)

$$\sigma_1 - \sigma_3 = \left(\frac{\dot{\varepsilon}}{A_p}\right)^{\frac{1}{n}} \exp\left(\frac{E_p}{nRT}\right), \tag{3}$$

where A_p [$Pa^{-n}s^{-1}$], n and E_p [$Jmol^{-1}$] are empirically determined material parameters for the power-law, T is the temperature [K], R = 8.314 $Jmol^{-1}K^{-1}$ (universal gas constant) and $\dot{\varepsilon}$ is the strain rate [s^{-1}]. At higher stresses (>200 MPa) the power law breaks down into the other strain rate-stress relationship (Tsenn & Carter 1987). For an olivine mineralogy, which is typical for conditions in mantle, this relation is well described by the Dorn law (Goetze & Evans 1979)

$$\sigma_1 - \sigma_3 = \sigma_D \left[1 - \left(-\frac{RT}{E_D} \ln\left(\frac{\dot{\varepsilon}}{A_D} \right) \right)^{\frac{1}{2}} \right], \tag{4}$$

where A_D [s^{-1}] E_D [$Jmol^{-1}$] and σ_D [Pa] are material parameters for the Dorn law.

The yield strength and the failure of material are affected by many intrinsic and extrinsic parameters. Temperature and pressure have a major influence on how the material fails. Increasing temperature has a strong effect on reducing the ductile strength while increasing pressure has a major effect on increasing brittle strength. Also, the presence of fluids (water, melts, gaseous phases) in the pore space of rocks drastically reduces the brittle strength. Pore fluid pressures are usually assumed to be hydrostatic when the pore fluid factor, defined as a ratio of pore fluid pressure to overburden pressure, has a value of about 0.4. Higher pore fluid pressures are however possible and they reduce the fracture strength dramatically. For a dry rock the pore fluid factor is assumed to be zero.

4 Numerical modelling

4.1 Rheology

Rheological profiles, sometimes referred to as strength envelopes, are constructed by calculating the brittle and the ductile strengths for a certain faulting type and a strain rate as a function of depth. The lowest value of the brittle and the ductile strength at a current depth is referred to as rheological strength. A faulting type between compressional and extensional ones can be assumed from fault plane solutions if available. Also, whether the rock is dry or wet must be taken into account. For the brittle regime, wet rock affects the strength of the rock through the action of pore fluid pressure. The pore fluid factor is usually chosen based on earlier studies, as a determination of an in-situ value is not possible or practical. For the ductile regime there is a similar effect on creep strength and the conditions are often referred to as hydrous. The mechanism of hydrolytic weakening is assumed to be responsible for this softening. Usually the upper or even the middle crust is assumed to exhibit wet conditions but this depends greatly on the tectonic and geological environment. In Fig. 2 the effects of two different pore fluid factor values for a thrust fault regime are shown for a typical strength envelope. Also, the brittle and the ductile regions and the BDT zone are illustrated. This strength envelope has many sources of uncertainties which are reviewed in the next chapter. However, one uncertainty is worth mentioning here as it is connected to the maximum stresses, i.e., peaks shown in Fig. 2. It is generally understood that the maximum stresses in the lithosphere are unlikely to exceed values of 200 MPa (Lamontage & Ranalli 1996, Ranalli 2000) and in this sense the use of ordinary brittle failure criteria most likely results in unrealistic strength values if extrapolated to and below depths of the middle crust. This is also discussed more in Paper V.

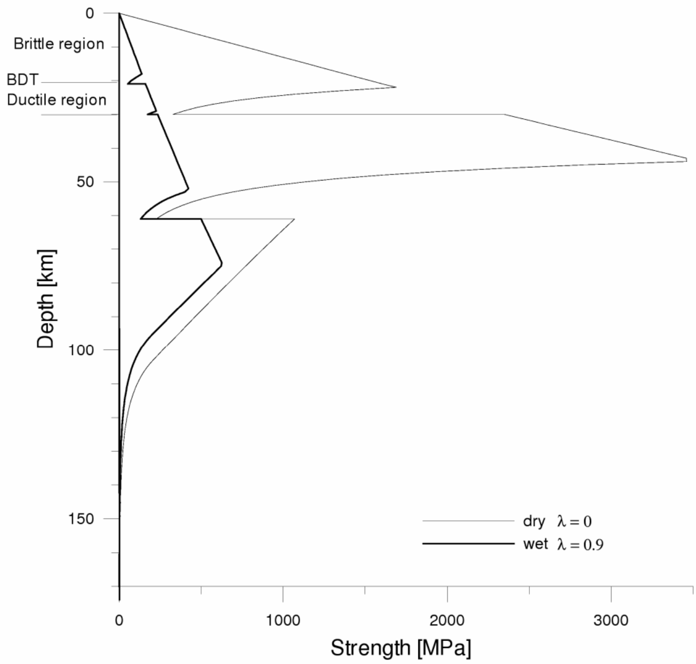

Fig. 2. Typical strength envelope calculated for thrust fault regime with pore fluid factors λ = 0 and λ = 0.9 Label 'ductile region' corresponds only to profile calculated with λ = 0.

4.2 Thermal modelling

Temperature has a major importance in rheological and the tectonic modelling as, for instance, ductile flow is dependent on temperature in a non-linear way. Thus, the temperature in the lithosphere must be known in order to resolve the rheological structure properly. Conductive heat transfer, understood to be the main mechanism of the heat transport in the lithosphere, is described with Fourier's law of heat conduction where the heat flow is directly proportional to the temperature gradient

$$q = -k\frac{dT}{dz}, \tag{5}$$

where q is the heat flow [Wm^{-2}], T is the absolute temperature [K], k is the coefficient of thermal conductivity [$Wm^{-1}K^{-1}$], and z is the depth [m]. This equation can be used to determine the heat flow density (HFD) with temperature measurements in boreholes as a function of a depth, together with the laboratory estimates of thermal conductivity.

Taking into account heat production and expanding the equation into three dimensions, we can write the heat conduction equation in a general form as

$$\rho\, c_p \frac{\partial T}{\partial t} = k\left(\frac{\partial^2 T}{\partial x^2} + \frac{\partial^2 T}{\partial y^2} + \frac{\partial^2 T}{\partial z^2}\right) + A\,, \tag{6}$$

where t is the time [s], ρ is the density [kgm^{-3}], c_p is the specific heat capacity [$Jkg^{-1}K^{-1}$] and A is the volumetric rate of the heat production [Wm^{-3}]. Heat production for the lithosphere can only be indirectly estimated. For example, measurements from outcrops and till samples provide estimates of heat production values in the uppermost part of the crust. Relationships between heat flow-heat production, seismic P-wave velocity-heat production and lithological models of the lithosphere based on the DSS data can be used to estimate proper values for the rest of the lithosphere.

If we assume steady state conditions ($\partial T/\partial t = 0$), i.e., no temperature change over time, we end up with an equation that can be used to calculate the stable geotherms of the lithosphere. This differential equation can be solved in many analytical and numerical ways (see e.g., Turcotte & Schubert 1982, Fowler 1990, Ranalli 1995). We have used the finite element method, to which a short introduction is given later. More complexity is brought to the solution if temperature dependence of the thermal conductivity is taken into account. This non-linear relationship is

$$k(T) = k_{ref}\left[\left(\frac{1}{1+bT}\right) + c(T + 273.15K)^3\right], \tag{7}$$

where T is the temperature [°C], k_{ref} is the thermal conductivity at the room temperature [$Wm^{-1}K^{-1}$] and b (0.0015 1/K) and c (1×10^{-10} $Wm^{-1}K^{-4}$) are empirical constants. Values of b and c are taken from the data by Schatz and Simmons (1972) and Zoth and Haenel (1988).

4.3 Structural modelling with the finite element method

The finite element method is a numerical approach by which partial differential equations can be solved in an approximate manner. The differential equations, which describe the physical problem considered, are assumed to be applicable over a certain region. A characteristic feature of the finite element method is that instead of seeking approximations that hold over the entire region, the region is divided into smaller parts, so called finite elements, and the approximation is then carried out over each element. By this way even non-linear behaviour can be approximated to behave linearly over single elements.

In static problems, which is often the case in tectonic studies, the static equilibrium equation can be presented in a form where displacements are the basic unknowns (in the following parentheses correspond to a matrix form of a parameter)

$$[K][u]=[f] \tag{8}$$

where K is a stiffness matrix, u is a vector of unknown displacements, f is a vector of forces.

Basic procedures in the finite element method (Zienkiewicz & Taylor 1989) are

1. The continuum is divided into finite elements for which expressions for the stiffness matrix can be easily derived. These element stiffness matrices can be assembled into the global stiffness matrix with the demand of internal continuity for the assembled structure.
2. The elements are interconnected at a discrete number of nodal points (nodes) situated on their boundaries. The displacements of these nodes will be the basic unknown parameters of the problem.
3. Interpolation (shape) functions are chosen to uniquely define the state of displacement within each element in terms of its nodal displacements.
4. The shape functions now uniquely define the state of strain within an element in terms of the nodal displacements. These strains, together with any initial strains, stresses and the constitutive properties of the material, define the state of stress throughout the element and also on its boundaries.
5. With the principal of virtual work, for example a formulation for the stiffness matrix can be derived.

In the following, a brief formulation of a finite element problem is presented.

(1)-(2) After the region is divided into **m** elements, the global stiffness matrix and the global force vector can be assembled from individual elements **e** with nodes **i**

$$K_{ij}=\sum_{e=1}^{m}K_{ij}^{e} \quad \text{ja} \quad f_i=\sum_{e=1}^{m}f_i^{e} \tag{9}$$

(3) The displacement vector can be approximated with shape functions $\mathbf{N_i}$, which describe the behaviour of the unknown parameter, in this case the displacement inside the individual element. Usually, different types of polynomials are used. Thus, the displacement vector in the element defined with **n** node points is

$$[u]\approx[\hat{u}]=\sum_{i}^{n}N_i a_i^e=[N_1,N_2,N_3,\cdots,N_n]\begin{bmatrix}a_1\\a_2\\a_3\\\vdots\\a_n\end{bmatrix}^e=[N][a]^e, \tag{10}$$

where a^e are the discrete displacements of the nodes.

(4) When the displacements are known at every point in the element we can derive the strains at any point based on the chosen relation between the displacement and the strain

$$[\varepsilon]^e = [S][u], \tag{11}$$

where the matrix S is a suitable linear operator. Using the shape function definition above, we can approximate the strain as

$$[\varepsilon]^e = [B][a]^e, \tag{12}$$

where B is a strain-displacement matrix, i.e., $[B]=[S][N]$. When the strains are known, the stresses can be derived based on the chosen constitutive equation. For elastic behaviour, i.e., Hooke's law, there is a linear relationship between the stress and the strain

$$[\sigma]=[D][\varepsilon], \tag{13}$$

where D is the constitutive matrix, i.e., in the elastic case the matrix containing the material properties as elastic constants. If we add initial strains ε_0, e.g., thermal strain and also initial stress σ_0 we get

$$[\sigma]=[D][\varepsilon-\varepsilon_0]+[\sigma_0] \tag{14}$$

(5) The last step is to derive expressions for the stiffness matrix and the force vector in order to solve the basic equation (8), for the unknown displacements.

The principle of virtual work states that a virtual (very small) change of the internal strain energy must be offset by an identical change in external work due to the applied loads

$$\delta U = \delta V , \tag{15}$$

where: U = U_1+U_2= strain energy (internal work), V = $V_1 + V_2 + V_3$ = external work and δ is a virtual operator. The virtual strain energy is defined as

$$\delta U_1 = \frac{1}{2}\int_V \varepsilon^T \sigma \, dV \tag{16}$$

if neglecting energy U_2 connected to moving surface resistances. Next, the external virtual work will be considered. Neglecting the inertia term V_1 leaves body forces F_B and surface stresses F_A

$$\begin{aligned} \delta V_2 &= -\int_V \bar{u}^T F_B \, dV \\ \delta V_3 &= -\int_A \bar{u}^T F_A \, dA , \end{aligned} \tag{17}$$

Combining the earlier Equations (10), (12) and (14) results in

$$
\begin{aligned}
\delta U_1 &= \int_V \delta\varepsilon^T \sigma \, dV = \int_V (B\delta a)^T [DBa - D\varepsilon_0 - \sigma_0] dV \\
&= \delta a^T \left[\int_V B^T DBdV\, a - \int_V B^T D\varepsilon_0 dV + \int_V B^T \sigma_0 dV \right] \\
&= \delta a^T [Ka - \Psi_0 + \Sigma_0]
\end{aligned}
\tag{18}
$$

and

$$
\begin{aligned}
\delta V_1 + \delta V_2 &= -\int_V \delta \bar{u}^T F_B \, dV - \int_A \delta \bar{u}^T F_A \, dA \\
&= -\int_V \delta a^T N^T F_B \, dV - \int_A \delta a^T N^T F_A \, dA \\
&= -\delta a^T \int_V N^T F_B \, dV - \delta a^T \int_A N^T F_A \, dA \\
&= -\delta a^T P_B - \delta a^T P_A
\end{aligned}
\tag{19}
$$

By using the relation of the principle of virtual work $\delta U = \delta V$ we obtain a linear system of Equations

$$
[K][u] = [f],
$$

$$
\text{where} \begin{cases} K = \int_V B^T DBdV \\ f = P_B + P_A + \Psi_0 + \Sigma_0 \\ = -\int_V N^T F_B \, dV - \int_A N^T F_A \, dA - \int_V B^T D\varepsilon_0 dV + \int_V B^T \sigma_0 dV. \end{cases}
\tag{20}
$$

After the nodal displacements **u** have been solved, the strains and the stresses in the element can be obtained from Equations (12) and (14), respectively.

5 Sources of error

Results of numerical rheological modelling are subjected to many uncertainties and thus a brief review of the most important error sources is given next. One of the major sources of error is connected to the uncertainties associated with lithospheric geotherms as temperature has a major influence in controlling ductile deformation. Fernandez & Ranalli (1997) have divided the uncertainties related to rheological calculations into two groups: operational and methodological errors. Operational errors include inaccurate estimation of lithospheric structure and composition, errors in temperature, improper material and rheological parameters, pore fluid pressure and friction coefficient etc. Methodological errors originate from the basic assumptions used in the rheological calculations. They include, for example, inaccurate rheological flow laws both in the brittle and ductile regimes and also assumption of the uniform strain and constant strain rate.

The measured surface HFD value is the parameter that indicates what the deep geothermal structure is like. The surface HFD is quite often used as a constraint when geotherms are derived in a numerical or analytical way. The surface HFD values are evaluated from temperature measurements taken in the boreholes and large uncertainties can be connected to these estimations. For example, groundwater flow, erosion, climatic temperature change (from annual to ice-ages) and topography bias measured values (see e.g., Kukkonen 1989, 1995). Furthermore, the sites of the HFD measurements cannot usually be chosen randomly resulting in unevenly distributed and possible large distances between the measurement sites that complicate the evaluation of the two- (2-D) or three-dimensional (3-D) geothermal structure. Quantitatively speaking variation in the surface HFD with an amount of ±10 mW/m^2 can result in a temperature change of ±100 °C at the depth of 50 km Moho depth. Thermal conductivity and heat production values are also required in lithospheric thermal modelling. These values are derived from laboratory measurements of rock samples but they usually only correspond to the conditions in the uppermost crust. For the rest of the lithosphere they are usually indirectly estimated, e.g., from relationships between the heat flow - heat production, the seismic P-wave velocity - heat production and the lithological models of the lithosphere based on the DSS data. Inaccurate values for these parameters can also result in a temperature change of ±100 °C at the 50 km depth similarly as the surface HFD variation of ±10 mW/m^2 (Jokinen & Kukkonen 1999).

From the operational errors, one significant error is related to rheological parameters themselves. Fernandez and Ranalli (1997) have compiled widely used parameters for different lithospheric layers and rather significant variations can be found among them (Kirby & Kronenberg 1987a, 1987b, Ranalli 1995). Also, the parameter values themselves seem to differ slightly from each other depending on the literature used (e.g., Cloetingh & Burov 1996). As the correct value is difficult to estimate, selecting a proper value for the rheological parameter can be of major importance as large variations in rheological strength can be generated. Other uncertainties are introduced, as the existence of possible hydrous environments is difficult to estimate. The upper crust is quite often assumed to have some amounts of water occupying pore spaces. A pore fluid pressure reduces the strength required for fracturing significantly. Hydrous environments can also occur deeper in the crust and the mechanism of hydrolytic weakening reduces the ductile strength considerably. An assumption of the proper tectonic setting, i.e., a faulting type for the brittle deformation, has to be estimated from the available earthquake data.

From the methodological errors, the inaccuracy of rheological flow laws must be kept in mind as those expressions are derived from experimental data, which are furthermore extrapolated to lithospheric conditions. The brittle failure criterion of Byerlee's law has been derived from results that are confirmed only to the depths of the upper crust. Therefore, using it to much larger depths results in strengths that are most likely somewhat unrealistic. Different brittle mechanisms are also very likely to occur in the deeper parts of the lithosphere as pressure increases (Ord & Hobbs 1989, Shimada 1993). Also, ductile flow mechanisms can differ from those commonly used, at least for some compositions, e.g., peridotite (e.g., Drury et al. 1991). Assuming a constant strain rate in the evaluation of rheological strength is not necessarily realistic. This is because in tectonically active areas different rates of deformation, together with anisotropic strain rates, are more likely to exist (e.g., Beaumont et al. 1994). Fig. 3 shows the strength envelopes calculated with four typical strain rate values for tensional and compressional faulting.

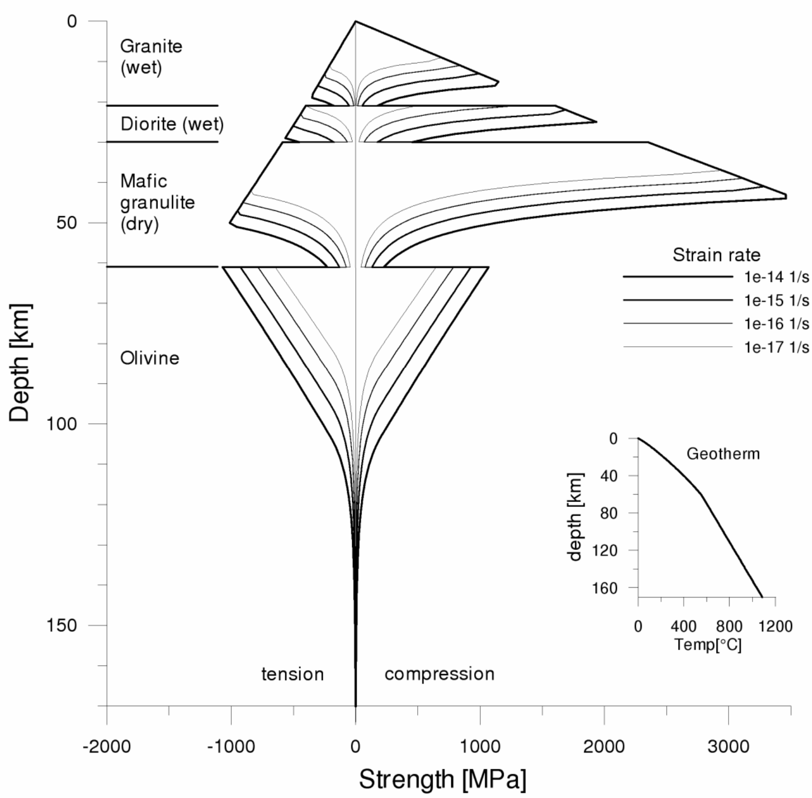

Fig. 3. Effect of the varying strain rate on the strength envelope. Used lithology and structural layers are shown next to the vertical axis and geotherm below right.

6 Summary of the original papers

In these original papers several lithospheric 2-D rheological models along the different DSS profiles together with the 3-D rheological model were calculated for the central Fennoscandian Shield. Analysis of the rheological structure and its consequences in the Shield were discussed. The rheological structure was furthermore used as a non-linear elastoplastic material property in different structural models, which were solved using the finite element method. Results give indications of the state of stress and possible deformation at the lithosphere. Next, a brief summary of each paper is presented.

6.1 Paper I

In this paper the present-day thermomechanical structure of the central Fennoscandian Shield was studied. The strength envelopes, i.e., rheological profiles, were calculated for 35 points located in different DSS profiles with the purpose of creating a map for indicating the rheological conditions in the central Fennoscandian Shield. The one-dimensional geotherms were calculated by solving the heat conduction equation for a steady state condition. Also, geotherms calculated by Jokinen and Kukkonen (1999, 2000) were used for 6 points for verification and comparison with our thermal and rheological models. The geotherms and the strength values at these common points correlated rather well. The strength envelopes were mainly calculated for a compressional, i.e., thrust fault regime, for both dry and wet conditions. A compressional condition was assumed as it is thought to be the dominating stress component in Fennoscandia (e.g., Kakkuri 1997). Results of this paper are presented with different rheological parameters describing the differences and the areal variations in the rheological structure of the central Fennoscandian Shield. Such parameters were the ICS (integrated crustal strength) and the ILS (integrated lithospheric strength), the MSC and MSL depths (depths of the mechanically strong crust and lithosphere, respectively), the RHTL (the rheological thickness of the lithosphere) and the BDT depth (brittle-ductile transition). These all show that the lithosphere in the central Fennoscandian Shield can be considered to be rather strong rheologically. For example, BDT depth for dry compressional conditions varies between 15 and 40 km and for wet compressional conditions between 11 and 32 km. The depth of the critical temperature where the frictional behaviour changes, i.e., depth of the 350 °C isotherm, varies between 25 and 44 km. Comparison of the earthquake focal depth data with the rheological parameters shows that the conventional mechanism of earthquakes, i.e., brittle fracture, is responsible for the earthquakes in the central Fennoscandian Shield.

6.2 Paper II

This paper is partly based on the previous one as the rheological profiles calculated in Paper I for the DSS profiles BALTIC and SKJ were used in a structural model of this study. Numerical modelling was applied for studying the present-day state of stress and deformation under different tectonic loading conditions at the seismic profiles in question. The finite element method was used for solving the structural problem for the 2-D model, which was constructed from existing seismic data along the profiles. One-dimensional geotherms were used in the calculation of the rheological vertical profiles which furthermore were applied to the 2-D structural model. Modelling was applied to different models with linear elastic and nonlinear elasto-plastic material properties. Calculated strength profiles were used as nonlinear material parameters in the elasto-plastic case where creep strength was translated into plastic yield stress. Different tectonic load cases were analysed with displacement, force and pressure type boundary conditions in a compressional setting.

The elastic case showed that the highest stress intensities accumulate in the middle and the lower crust. The elasto-plastic case was used to analyse if the generated stress field is high enough to overcome the yield strength. If the wet rheological conditions were applied to the entire crust, the lower crust attained a high enough stress field resulting in plastic deformation. The dry model showed different results with no significant plastic deformation in the crust. As in Paper I, we used the Monte-Carlo simulated geotherms in the structural model. This modelling gave consistent results compared to earlier ones and confirmed that our thermal models were accurate enough. The comparison between the rheological behaviour and geophysical investigations, e.g., electromagnetic, seismicity and seismic studies on the BALTIC-SKJ profile, was also considered.

6.3 Paper III

In this paper, a similar modelling procedure to that of Paper II was used. Finite element modelling was applied for the DSS profile SVEKA, for studying the structural present-day conditions in the lithosphere. Modelling basically followed the same procedure as in Paper II, with the exception that the thermal and the rheological calculations were made in two dimensions. The structure of the model was based on the seismic model of the SVEKA profile (Yliniemi et al. 1996, Kozlovskaya and Yliniemi 1999). The thermal and the structural models were both solved with the finite element method using the same mesh in both problems. Different model cases were analysed with varying properties, i.e., composition, boundary condition and thermal structure always assuming compressional condition.

Rheological calculations showed that the lower crust was characterised by a weak ductile layer. Numerical modelling revealed what kind of variations are to be expected in rheological response with varying conditions. Results also showed that the highest stress intensities were found in the lower crust and that different cases showed only minor ductile deformation in the lowermost crust in the Proterozoic part of the profile. These calculations implied that ductile lower crust is highly unlikely in the central Fennoscandian Shield.

6.4 Paper IV

In this paper a similar procedure to that described earlier was applied to the DSS profile FENNIA for studying rheological structure and modelling the present-day stress field and the state of deformation. The 2-D thermo-structural model for the FENNIA profile was constructed using the seismic P-wave velocity model as a basis for the model. This model has a higher resolution, i.e., element sizes are smaller than in the previous papers. The thermal finite element model was solved assuming 2-D conductive steady-state conditions and the solved temperature was furthermore used in the calculation of rheological strength. In contrast to the previous papers, the temperature dependence of the thermal conductivity was also taken into account in the thermal model. Strength values were included into the structural finite element model as elastoplastic non-linear material parameters. The structural model had the same geometry and element distribution as the thermal model. Compressional pressure simulating tectonic loading was applied for generating the stress field.

The 2-D rheological model showed that in the central Fennoscandian Shield rheologically weak crustal layers could be generated under suitable conditions. A very important factor is an assumption of wet crustal conditions for ductile creep properties. Results of the structural modelling implied that no significant crustal deformation is generated with the used thermal, material and boundary conditions. The crustal deformation is limited, with both the wet and the dry conditions, to very narrow zones in the upper and middle crust. Uncertainty analysis of the effects of the temperature on rheological and structural models was also done with 3 thermal model variants with large uncertainty bounds. This showed that the effect of temperature on modelling rheological and stress conditions is significant.

6.5 Paper V

In this paper the previously described modelling procedure was conducted in the three dimensions for the thermal, rheological and structural models. In this way the limitations met in the 2-D models, namely the incorrect direction of the applied boundary condition, can be avoided. The finite element mesh was oriented so that its longest axis is approximately in the same direction as the assumed maximum horizontal stress, i.e., NW-

SE and so the applied boundary conditions were optimally oriented in comparison to the in-situ stress conditions. The horizontal boundaries of the mesh were derived from several seismic measurements done it the central Fennoscandian Shield. These boundaries divide the crust into the upper, the middle and the lower crust. As rheological calculations require knowledge of thermal structure, the 3-D conductive steady-state model was first solved using the finite element method. Temperature values were then used in the calculation of rheological strength, which were subsequently used as non-linear material parameters in the structural finite element model having the same geometry and element distribution as the thermal one. This structural model was subjected to compressional boundary conditions in order to simulate the tectonic stress conditions in the area.

The calculated rheological structure in the 3-D model has similar features as seen in the earlier 2-D rheological models. A cross-section of rheological strength showed a layered structure in the crust with two individual rheologically weak layers in the granitic upper and the dioritic middle crust. A comparison of the rheological structure with earthquake focal depth data showed that the earthquake events are located in the brittle regime as most of them occurred between the depths of 3 and 10 km. Also, it was possible to conclude that as the earthquakes are infrequently below a depth of 15 km, the stress conditions at these depths were such that tectonic stress is not large enough to exceed the true strength. Resulting stress conditions, after the model was subjected to compressional boundary conditions, had values of around 50 MPa. However, in comparison with the earlier 2-D modelling, this 3-D model did not result in any significant deformation. The largest differences between the 2-D and 3-D models were in the distribution of the stress field. Although the same boundary condition was used the results were quite different. The degree of deformation was much closer to that of earlier modelling. This was however also affected by the worse resolution of the 3-D model

.

7 Conclusions

Understanding rheological conditions throughout the entire lithosphere and even deeper is the key for understanding the deformation of the lithosphere. Thus, investigating rheological structure and possible consequences resulting from tectonic loading is in some sense required when interpreting geophysical data into tectonic models. The thermal, rheological and structural models investigated in this thesis are only a scratch on the surface of tectonic modelling. However, they provide important information in describing possible areas of deformation, the distribution of the stresses, in understanding the occurrence of the earthquakes and also for developing more sophisticated and complex models. This thesis also discusses the uncertainties connected to the various parameters needed in the modelling. Thus, the results of these models should be analysed with caution as values of different parameters can generate rather large variations to them. Also, the verification of the thermal, rheological and structural models is very difficult as in-situ data about the conditions deep in the lithosphere are difficult to obtain. This, together with the error sources, is one of the major problems in these studies. Xenoliths give some estimation of the thermal conditions to a depth of a few hundred kilometres. Fault plane solutions provide information about the tectonic setting and focal depth data is very valuable as well. In addition, high quality stress-measurements would also be useful although they would be limited to very shallow depths.

This thesis has shown that rheological structure depends strongly on the thermal conditions resulting in significant areal variations. Generally, the central Fennoscandian Shield can be considered rheologically rather strong. Depending on the applied properties, rheologically weak layers are however usually found in the lower crust and also in the upper and the middle crust, although these layers are very thin in most cases.

Ranalli (1995) connects the existence of the weak lower crust to lateral extrusion, i.e., the horizontal flow of ductile lower crustal material. This process relaxes the topography of the Earth’s surface and affects the topography of the Moho. At longer wavelengths lateral extrusion also depends on the rheology of the lower crust. The observed present day features of the Moho are thus not necessarily a result of the last tectonic event, i.e., post-orogenic ductile flow should also be considered.

The correlation between rheological structure and earthquake focal depth data in the central Fennoscandian Shield showed that brittle fracture is the relevant mechanism in earthquake generation. The BDT depth in our models is usually well below the depth where most of the earthquakes occur. Also, non-occurrence of deep earthquakes implies low stress or high strength conditions deeper in the crust. Numerical structural models used to simulate the stress and strain conditions in the lithosphere showed how stresses are distributed and whether any major deformation is possible. Based on the several models, it is highly unlikely that any considerable ductile deformation in the crust exists and it seems that the present-day thermal and mechanical conditions in the central Fennoscandian Shield do not favour such processes in significant amounts.

References

Ádám A, Kaikkonen P, Hjelt S-E, Pajunpää K, Szarka L, Verö J & Wallner A (1982) Magnetotelluric and audiomagnetotelluric measurements in Finland. Tectonophysics 90: 77-90.

Ahjos T & Uski M (1991) Earthquake epicenters in northern Europe. In: Korhonen H & Lipponen A (eds) Structure and dynamics of the Fennoscandian lithosphere. Institute of Seismology, University of Helsinki, Report S-25: 81-86.

Ahjos T & Uski M (1992) Earthquakes in northern Europe in 1375-1989. Tectonophysics 207: 1-23.

Azbel IYa, Buyanov AF, Ionkis VT, Sharov NV & Sharova VP (1989) Crustal structure of the Kola Peninsula from inversion of deep seismic sounding data. In: Freeman R, von Knorring M, Korhonen H, Lund C & Mueller St (eds) The European Geotraverse, Part 5: Polar Profile. Tectonophysics 162: 1-85.

BABEL Working Group (1993a) Integrated seismic studies of the Baltic shield using data in the Gulf of Bothnia region. Geophys J Int 112: 305-324.

BABEL Working Group (1993b) Deep seismic reflection/refraction interpretation of crustal structure along BABEL profiles A and B in the southern Baltic Sea. Geophys J Int 112: 325-343.

Beaumont C, Fullsack P & Hamilton J (1994) Styles of crustal deformation in compressional orogens caused by subduction of the underlying lithosphere. Tectonophysics 232: 119-132.

Blanpied ML, Lockner DA & Byerlee JD (1991) Fault stability inferred from granite sliding experiments at hydrothermal conditions. Geophys Res Lett 18: 609-612.

Cermák V, Balling N, Kukkonen I & Zui VI (1993) Heat flow in the BALTIC Shield-results of the lithospheric geothermal modelling. Precamb Res 64: 53-65.

Chen R (1991) On horizontal crustal deformation in Finland. Rep Finn Geod Inst 91:1, Helsinki.

Cloetingh S & Burov EB (1996) Thermomechanical structure of European continental lithosphere: constraints from rheological profiles and EET estimates. Geophys J Int 124: 695-723.

Drury MR, Vissers RLM, van der Wal D & Hoogerduijn Strating EH (1991) Shear localization in upper mantle peridotites. Pure Appl Geophys 137: 439-460.

Elo S (1997) Interpretations of the Gravity Anomaly Map of Finland. Geophysica 33: 51-80.

Elo S & Korja A (1993) Geophysical interpretation of the crustal and upper mantle structure in the Wiborg rapakivi granite area, SE Finland. Precamb Res 64: 273-288.

FENNIA Working Group (1996) P- and S-velocity crustal structure of the Fennoscandian Shield beneath the FENNIA profile in southern Finland. EGS XXI General Assembly, The Hague, The Netherlands, 6-10 May, 1996. Ann Geophys Part I, C64.

Fernàndez M & Ranalli G (1997) The role of rheology in extensional basin formation modelling. Tectonophysics 282: 129-145.
Fowler CMR (1990) The Solid Earth – An Introduction to Global Geophysics. Cambridge Univ. Press, Cambridge.
Gaál G & Gorbatschev R (1987) An outline of the Precambrian evolution of BALTIC Shield. Precamb Res 35: 15-52.
Goetze C & Evans B (1979) Stress and temperature in the bending lithosphere as constrained by experimental rock mechanics. Geophys J R astr Soc 59: 463-478.
Grad M & Luosto U (1987) Seismic models of the crust of the Baltic shield along the SVEKA profile in Finland. Ann Geophys 5B: 639-650.
Guggisberg B, Kaminski W & Prodehl C (1991) Crustal structure of the Fennoscandian Shield: A traveltime interpretation of the long-range FENNOLORA seismic refraction profile. Tectonophysics 195: 105-137.
Hjelt S-E & Korja T (1993) Lithospheric and upper mantle structures, results of electromagnetic soundings in Europe. Phys Earth and Planet Inter 79: 137-177.
Huhma H (1986) Sm-Nd, U-Pb and Pb-Pb isotopic evidence for the origin of the Early Proterozoic Svecokarelian crust in Finland. Geol Surv Finland, Bull 337.
Jokinen J & Kukkonen I (1999) Random modelling of the lithospheric thermal regime: forward simulations applied in uncertainty analysis. Tectonophysics 306: 277-292.
Jokinen J & Kukkonen IT (2000) Inverse Monte Carlo simulation of the lithospheric thermal regime in the Fennoscandian Shield using xenolith-derived mantle temperatures. J Geodyn 29: 71-85.
Jõeleht A & Kukkonen IT (1998) Thermal properties of granulite facies rocks in the Precambrian basement of Estonia and Finland. Tectonophysics 291: 195-203.
Järvimäki P & Puranen M (1979) Heat flow measurements in Finland. In: Cermák V & Rybach L (eds) Terrestial Heat Flow in Europe. Springer, Berlin, 172-178.
Kaikkonen P, Vanyan LL, Hjelt S-E, Shilovsky AP, Pajunpää K & Shilovsky PP (1983) A preliminary geoelectrical model of the Karelian megablock of the Baltic Shield. Phys Earth and Planet Inter 32: 301-305.
Kakkuri J (1997) Postglacial deformation of the Fennoscandian crust. Geophysica 33: 99-109.
Kirby SH (1983) Rheology of the lithosphere. Rev Geophys 21: 1458-1487.
Kirby SH & Kronenberg AK (1987a) Rheology of the lithosphere: selected topics. Rev Geophys 25: 1219-1244.
Kirby SH & Kronenberg AK (1987b) Correction to 'Rheology of the lithosphere: selected topics'. Rev Geophys 25: 1680-1681.
Korja A, Korja T, Luosto U & Heikkinen P (1993) Seismic and geoelectric evidence for collisional and extensional events in the Fennoscandian Shield-implications for Precambrian crustal evolution. Tectonophysics 219: 129-152.
Korja A & Heikkinen PJ (1995) Proterozoic extensional tectonics of the central Fennoscandian Shield: Results from the BABEL experiment. Tectonics 14: 504-517.
Korja T & Koivukoski K (1990) Magnetotelluric soundings along the SVEKA profile in the central Fennoscandian (Baltic) Shield, Finland. Dep Geophys, Univ Oulu, Report 17.
Korja T & Hjelt S-E (1993) Electromagnetic studies in the Fennoscandian Shield-electrical conductivity of Precambrian crust. Phys Earth and Planet Inter 81: 107-138.
Korja T & Koivukoski K (1994) Crustal conductors along the SVEKA profile in the Fennoscandian (Baltic) Shield, Finland. Geophys J Int 116: 173-197.
Korja T (1997) Electrical Conductivity of the Lithosphere-Implications for the Evolution of the Fennoscandian Shield. Geophysica 33: 17-50.
Kozlovskaya E and Yliniemi J (1999) Deep structure of the Earth's crust along the SVEKA profile and its extention to north-east. Geophysica 35: 111-123.

Kukkonen IT (1988) Terrestial heat flow and groundwater circulation in the bedrock in the central Baltic Shield. Tectonophysics 156: 59-74.

Kukkonen IT (1989) Terrestrial heat flow in Finland, the central Fennoscandian Shield. Geol Surv of Finland, Nuclear Waste Disposal Research, Report YST-68.

Kukkonen IT (1993) Heat flow map of northern and central parts of the Fennoscandian Shield based on geochemical surveys of heat producing elements. Tectonophysics 225: 3-13.

Kukkonen IT (1994) Present thermal structure of the Lithosphere in the Finnish part of the Fennoscandian Shield. In: Pajunen M (ed) High temperature-low pressure metamorphism and deep crustal structures. Geol Surv of Finland, Guide 37: 69-74.

Kukkonen IT (1995) Thermal aspects of groundwater circulation in bedrock and its effect on crustal geothermal modelling in Finland, the central Fennoscandian Shield. Tectonophysics 244: 119-136.

Kukkonen IT (1996) Temperature and heat flow density in the bedrock on the GGT/SVEKA transect. In: Ekdahl E & Autio S (eds.) Global Geoscience Transect/SVEKA-Proceedings of the Kuopio Seminar, Finland 25.-26.11.1993. Geol Surv of Finland, Report of Investigation 136: 23-30.

Kukkonen IT (1998) Temperature and heat flow density in a thick cratonic lithosphere: The SVEKA transect, central Fennoscandian Shield. J Geodyn 26: 111-136.

Kukkonen IT & Jõeleht A (1996) Geothermal modelling of the lithosphere in the central BALTIC shield and its southern slope. Tectonophysics 255: 25-45.

Kukkonen IT & Lahtinen R (2001) Variation of radiogenic heat production rate in 2.8-1.8 Ga old rocks in the central Fennoscandian shield. Phys Earth and Planet Inter 126: 279-294.

Lamontagne M & Ranalli G (1996) Thermal and rheological constraints on the earthquake depth distribution in the Charlevoix, Canada, intraplate seismic zone. Tectonophysics 257: 55-69.

Luosto U (1991) Moho depth map of the Fennoscandian Shield based on seismic refraction data. In: Korhonen H & Lipponen A (eds) Structure and dynamics of the Fennoscandian lithosphere. Institute of Seismology, University of Helsinki, Report S-25: 43-49.

Luosto U, Korhonen H, Kosminskaya I, Zverev SM, Lund C-E, Sharov N, Lanne E, Tuppurainen A, Ilmola V-M & Foursov AN (1985) First results from the DSS study of the BALTIC profile in SE Finland. Institute of Seismology, University of Helsinki, Report S-11.

Luosto U, Flueh E, Lund C-E & Working Group (1989) The crustal structure along the Polar profile from seismic refraction. In: Freeman R, von Knorring M, Korhonen H, Lund C & Mueller St (eds) The European Geotraverse, Part 5: Polar Profile. Tectonophysics: 162, 51-85.

Luosto U, Tiira T, Korhonen H, Azbel I, Burmin V, Buyanov A, Kosminskaya I, Ionkis V & Sharov N (1990) Crust and upper mantle structure along the DSS BALTIC profile in SE Finland. Geophys J Int 101: 89-110.

Luosto U, Grad M, Guterch A, Heikkinen P, Janik T, Komminaho K, Lund C, Thybo H & Yliniemi J, (1994) Crustal structure along the SVEKA'91 profile in Finland. In: Makropoulos K & Suhadolc P (eds) European Seismological Commission, XXIV General Assembly, 1994, September 19-24, Athens, Greece. Proceedings and Activity Report 1992-1994 Vol II: 974-983.

Lähde S (1985) Determining and use of regional mean densities of Finnish bedrock in helping geophysical interpretation (in Finnish). MSc Eng Thesis, Helsinki University of Technology, Mining Department.

Mase GE (1970) Continuum mechanics. Schaum's outline series. McGraw-Hill Inc, New York.

Ord A & Hobbs B (1989) The strength of the continental crust detachments zones and the development of plastic instabilities. Tectonophysics 158: 269-289.

Park RG (1989) Foundations of structural geology. Blackie & Son Ltd, Glasgow.

Pasquale V, Verdoya M & Chiozzi P (1991) Lithospheric thermal structure in the Baltic shield. Geophys J Int 106: 611-620.

Puranen M, Järvimäki P, Hämäläinen U & Lehtinen S (1968) Terrestial heat flow in Finland. Geoexploration 6: 151-162.

Puranen R, Elo S & Airo M-L (1978) Areal and geological variation of rock densities, and their relation to some gravity anomalies in Finland. Geoskrifter (Aarhus University) 10: 123-164.

Ranalli G (1995) Rheology of the Earth (2nd edit). Chapman & Hall, London.

Ranalli G (2000) Rheology of the crust and its role in tectonic reactivation. J Geodyn 30: 3-15.

Sandoval S, Kissling E, Ansorge J & the SVEKALAPKO Seismic Tomography Working Group (2003) High-resolution body wave tomography beneath the SVEKALAPKO array – I. A priori three-dimensional crustal model and associated traveltime effects on teleseismic wave fronts. Geophys J Int 153: 75-87.

Sandoval S, Kissling E, Ansorge J & the SVEKALAPKO Seismic Tomography Working Group (2004) High-resolution body wave tomography beneath the SVEKALAPKO array – II. Anomalous upper mantle structure beneath the central Baltic Shield. Geophys J Int 157: 200-214.

Schatz JF & Simmons G (1972) Thermal conductivity of earth materials at high temperatures. J Geophys Res 77: 6966-6983.

Scholz CH (1990) The mechanics of earthquakes and faulting. Cambridge Univ. Press, Cambridge.

Sharov NV, Zagorodny VG, Glaznev VN & Zhamaletdinov AA (1989) Struktura litosfery I verkhney mantiy Baltiyskogo shchita. In: Sollogub VB (ed) Litosfera Tsentralnoy I Vostochnoy Evropy: Vostochno-Evropeyskaya Platforma. Naukova Dumka, Kiev, 52-77 (In Russian).

Shimada M (1993) Lithosphere strength inferred from fracture strength of rocks at high confining pressures and temperatures. Tectonophysics 217: 55-64.

Sibson RH (1974) Frictional constraints on thrust, wrench and normal faults. Nature 249: 542-544.

Tsenn MC & Carter NL (1987) Upper limits of power law creep of rocks. Tectonophysics 136: 1-26.

Turcotte DL & Schubert G (1982) Geodynamics, Applications of continuum physics to geological problems. John Wiley and Sons, New York.

Uski M, Hyvönen T, Korja A & Airo M-L (2003) Focal mechanisms of three earthquakes in Finland and their relation to surface faults. Tectonophysics 363: 141-157.

Windley B (1992) Precambrian Europe. In: Blundell D, Freeman R & Mueller St (eds) A continent revealed: The European Geotraverse. Cambridge University press, Cambridge, 139-214.

Zienkiewicz OC & Taylor RL (1989) The Finite Element Method, Volume 1. Mc-Graw Hill, New York.

Zoth G & Haenel R (1988) Thermal conductivity. In: Haenel R, Rybach L & Stegena L (eds) Handbook of Terrestrial Heat-Flow Density Determination. Kluwer, Dordrecht, 449-466.

Yliniemi J (1991) Deep seismic soundings in the University of Oulu. Institute of Seismology, University of Helsinki, Report S-25: 1-6.

Yliniemi J, Jokinen J & Luukkonen E (1996) Deep structure of the Earth’s crust along the GGT/SVEKA transect extension to northeast. In: Ekdahl E & Autio S (eds) Global Geoscience Transect/SVEKA-Proceedings of the Kuopio Seminar, Finland 25.-26.11.1993. Geol Surv of Finland, Report of Investigation 136: 56.