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Electrical Conductivity of the Lithosphere - Implications for the Evolution of the Fennoscandian Shield

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Abstract

The electromagnetic induction research on the deep structure in the Fennoscandian Shield and the airborne surveys in Finland have provided a globally unique electromagnetic data set to estimate lithospheric conductivity. The data allow to focus from shield-scale structures to small local structures, and enables to interpret the conductivity structures in terms ranging from tectonic processes affecting the entire shield to small processes responsible for enhanced electrical conductivity within shear zones.

The Shield is characterized by elongated belts of conductors that are limited to the upper half of the crust except for some occasional penetrations into the lower crust. Due to the large conductances involved and the correlation with the surface geology the primary conductivity mechanism invoked to explain this is electronic as found in graphite- and sulphide-bearing rocks. The conductive belts surround more resistive blocks and mark boundaries between distinct crustal units.

The resistive regions serve as transparent windows to probe deeper properties of the shield. The deep Palaeoproterozoic crust seems to be more conductive, in particular in the central part of the shield, than the Archaean lower crust. There are also an increasing amount of evidence that the lower crust and even the upper mantle may be electrically anisotropic. Studies in the Lapland Granulite Belt have shown that the enhanced lower crustal conductivity may be explained by graphite-bearing shear zones in agreement with the petrological evidence of dry deep crust.

Currently available information on the upper mantle conductivity is sparse. An additional methodological difficulty at high latitudes is imposed by the proximity of the auroral zone. The results nevertheless indicate that in the central and southwestern parts of the shield the asthenospheric layer is absent or is electrically weak whereas in peripheral regions the upper mantle conductor associated with the asthenosphere might be closer to the surface and more conducting.

Key words: electrical conductivity, lithosphere, lower crust, terrane boundaries, Precambrian, Fennoscandia

1. Introduction

This paper is aimed at: (i) giving a brief introduction to geoelectromagnetic methods and the main factors controlling the electrical conductivity in the continental lithosphere and upper mantle, (ii) summarizing the geoelectromagnetic studies and their results used to investigate the electrical properties of the lithosphere mainly in the central and northeastern part of the Fennoscandian Shield, (iii) showing how the combination of large scale regional and small scale detailed local studies provides the best models for the tectono-geological interpretation of conducting structures, and (iv) demonstrating that geoelectromagnetic studies can provide valuable information on the geological and tectonic structures of old Precambrian cratons and their present physical state.

Geoelectromagnetic soundings are the most sensitive geophysical methods for the detection of volumetrically minor but tectonically important constituents, i.e. carbon/graphite, free saline fluids, and small percentage of partial melt. Hence electromagnetic techniques are complementary to many other geophysical techniques which also are sensitive to the bulk parameters of the medium. The electromagnetic methods provide independent and complementary data and allow to constrain the Earth models obtained from seismic data, in particular. Rapid improvements in geoelectromagnetic methodology and an increase in the number of electromagnetic studies covering various tectonic environments have considerably improved in last few years our knowledge on the electrical properties of the Earthis crust and upper mantle (e.g. *Hjelt*, 1988; *Jödicke*, 1992; *Hjelt* and *Korja*, 1993; *Brown*, 1994 for upper and middle crust; *Jödicke*, 1992; *Jones*, 1992 for lower crust; *Schultz*, 1990 for upper mantle). The improved knowledge on the electrical properties of subsurface as well as the availability of spatially coinciding geophysical data sets (e.g., magnetotelluric and seismic data) has made it possible to interpret geoelectrical models in much more specific tectono-geological terms ranging from models for stable shields (e.g. *Korja, A. et al*., 1993; *Boerner et al*., 1995; *Korja et al*., 1996a) to models for ongoing continent-continent collisions (e.g. *Pous et al*., 1995).

In stable regions old tectonic processes have left in many places electrically conducting traces in the upper and middle crust. These conductors, composed primarily of graphite- and/or sulphide-bearing rocks of sedimentary sequences, can give information e.g. on collisions of crustal blocks revealing the location of possible palaeosuture zones or terrane boundaries (Fig 1. where the map of electrically conducting zones of Europe is shown; *Hjelt* and *Korja*, 1993).

Studies on deep continental crust (*Jones*, 1992) have led to a general conclusion that the electrical conductivity of the continental lower crust, estimated from surface measurements, is higher than estimated from the electrical conductivity determinations of appropriate rocks at relevant pressure and temperature conditions. Recent studies (e.g. *Rasmussen*, 1988; *Kellet et al*., 1992) indicate also that the deep lithosphere is likely to be electrically anisotropic. Results from the Canadian and Fennoscandian Shields (*Mareschal et al*., 1995; *Korja et al*., 1996b,c) suggest that the anisotropy is inherited from old tectonic processes and therefore bear knowledge on the evolution of the deep lithosphere. In tectonically stable areas, free saline H_2O-bearing fluids and graphite forming an interconnected and conductive phase within otherwise resistive rock matrix have been the main proposals for enhanced lower crustal conductivity.

Shields with exposed Precambrian crust provide excellent windows to probe the deep structure of continental lithosphere and to investigate its evolution through the Palaeoproterozoic and Archaean times as well as to study the present day dynamics of continental lithosphere. Of special interest is the relationship between the thick, floating (?) craton and the convecting mantle. The Fennoscandian (Baltic) Shield (Fig. 1) is the largest exposed part of the Precambrian East European Craton (*Gorbatschev* and *Bogdanova*, 1993). It has practically no sedimentary rocks on land areas except for a very thin overburden of quaternary sediments. Therefore the electrical screening effect of the upper crustal conducting sediments in particular is minimal. The conductors within the surficial parts of the Fennoscandian bedrock are reasonably well known because of systematic airborne electromagnetic mapping. Consequently it is possible to infer the geoelectrical structure of the crust and upper mantle with a reasonable resolution from electromagnetic data.

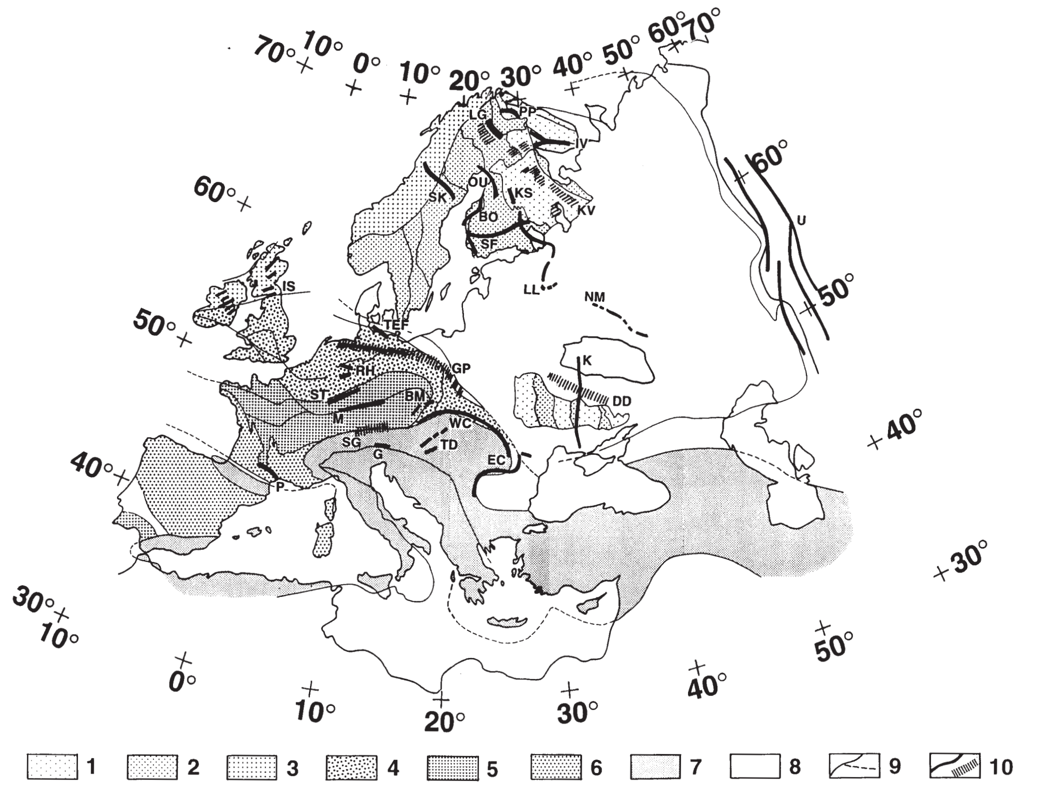

Fig. 1. Electrically conducting zones in Europe. Lines indicate only the location of conducting zones and not their actual width and represent only the surface projections of the nearest-to-the-surface parts of the conductors. In many cases the conductors are inclined and the zones show the uppermost parts of the dipping conductive bodies. The geological map show the major terranes in western Europe and the limits of the East European platform. Figure is from *Hjelt* and *Korja* (1993) where references for original studies and the names of conductors are given. 1-Archaean terranes, 2-Proterozoic terranes, 3-Caledonides (Scandinavia, Scotland, and Ireland), 4-Eastern Avalon terranes, 5-Variscan suspect terranes, 6-Variscan Gondwana, 7-Alpine orogen, 8-East European Platform, 9-boundaries of geological units, 10-elongated conductivity structures (solid lines - deep crustal conductors, ruled lines - shallow conductors caused e.g. by sedimentary and metasedimentary rocks).

2. Electrical conductivity of the rocks in the continental crust and upper mantle

Electrical conductivity of the Earth varies in the widest range (10^2 to 10^{-6} S/m) of any of the physical parameters that can be sensed from the Earth's surface and is the most sensitive physical rock parameter to small changes in various, volumetrically minor but tectonically important, constituents such as aqueous fluids, conductive minerals, and partial melt (*Haak* and *Hutton*, 1986; *Jones*, 1992). As most rocks are resistive, the conductivity of a rock volume depends primarily on the degree of interconnection and the amount of conductive phase. Water-bearing fluids in pores or cracks occupying only a few percent, or even less than 1%, of a total rock volume can decrease the bulk resistivity of the rock volume by one or two orders of magnitude. The effect is even more pronounced with much more conducting graphite, sulphides or magnetite as a conductive phase. The highly resistive rock matrix is nearly "transparent" in the electromagnetic sense and the electromagnetic methods detect mainly the conductive phase. Hence electromagnetic techniques are complementary to many other geophysical techniques, which also are sensitive to the bulk parameters of the medium, and has an advantage to be sensitive to very small amounts of e.g. saline H_2O-bearing fluids or graphite. This, however, does not imply that the electromagnetic methods could resolve the spatial distribution of these material at great depths.

Increasing temperature increases the conductivity as more and more electrons or ions become free to move whereas increasing pressure decreases the conductivity. The latter effect is mostly related to the closure of pores and cracks and occurs usually at depths comparable to lithostatic pressures well below 5 kbar (e.g. *Zhdanov* and *Keller*, 1994, p. 95). At higher pressures an increase in pressure has a very small effect in the electrical conductivity. For molten rocks, however, the pressure effect is stronger. The combined effect of increasing pressure and temperature is a decrease in resistivity. Therefore a descending resistivity curve does not necessarily indicate a presence of an anomalous conductor at depth. The discussion below is primarily targeted to indicate the possible causes for the anomalous behaviour of electrical conductivity i.e. departures from normal temperature-pressure controlled conductivity of dry rocks thought to compose the crust and upper mantle.

In upper and middle crust, especially in shield areas, the electrical conductivity is bimodal in large scale. Most rocks and formations are very resistive whereas a volumetrically minor fraction of rocks have very high conductivities, of which predominantly carbon- and sulphide-bearing ones are the most important sources of increased conductivity. *Stanley* (1989) has outlined the occurrence of thin films of carbon or sulfide minerals as a major conduction mechanism in metamorphosed shales in the Alaska Range, and proposed it as the primary cause for deep conductive anomalies in suture zones. In the Fennoscandian Shield the most conductors represent carbon- and sulphide-bearing metasedimentary rock layers but there are also several examples where shear zones containing carbon and/or sulfides are very conducting. In most cases the carbon is organic. One notable exception in the Fennoscandian Shield is the Lapland

Granulite Belt where the lowermost basal unit seems to contain also inorganic carbon that precipitated into shear zones during the upthrusting of granulites (*Korja et al*., 1996a).

The most commonly proposed causes for the enhanced electrical conductivity in the deep crust include (1) H_2O-bearing fluids trapped at the top of the lower crust (at the brittle/ductile transition) (*Gough*, 1986; *Hyndman* and *Shearer*, 1989; *Jones*, 1992) or being distributed over the entire lower crust in lamellae structures (*Sanders*, 1991; *Merzer* and *Klemperer*, 1992); (2) crystalline grain boundary phases such as carbon, sulphides and ilmenite (*Frost et al*., 1989; *Mareschal et al*., 1992; *Jödicke*, 1992; *Glover* and *Vine*, 1992; *Duba et al*., 1994), (3) underplated and intruded mafic magmas (*Warner*, 1990) or (4) lower crustal ductile shear zones (*Haak* and *Hutton*, 1986; *Korja et al*., 1996a for graphite-bearing shear zones or *Sanders*, 1991 for water-bearing shear zones). (5) In tectonically active regions partial melts are also a plausible source for enhanced conductivity (*Waff*, 1974).

Free aqueous fluids in stable continental environment contradicts the petrological arguments indicating that the deep crust is dry and that the fluid flow is highly episodic and restricted to periods of active metamorphism and tectonism (*Frost* and *Bucher*, 1994). The laminated reflectivity of the deep crust has been considered as evidence for the presence of aqueous fluids (*Hyndman* and *Shearer*, 1989), but many recent studies (*Mooney* and *Meissner*, 1992) propose lithological variations as a possible cause for the impedance contrasts. Moreover, if H_2O-bearing fluids were present during a high-temperature stage, these would disappear rapidly. It also seems difficult to transfer fluids later from the mantle into the cooler stable lower crust (*Bucher-Nurminen*, 1990). Therefore other mechanisms have been introduced based on conduction along a solid and interconnected phase. These include thin graphitic films precipitated onto grain boundaries from CO_2-rich fluids during cooling after metamorphism (*Frost et al*.,1989; *Mareschal et al*., 1992), oxide minerals (*Duba et al*., 1994) and graphite-bearing ductile shear zones (*Korja et al*., 1996a).

In extensional environments high conductivities associated to upper mantle have been found at depths from 50 to 200 km (e.g. Pannonian basin; see Fig. 2) indicating the presence of either partial melt or hydrous fluids (*Shankland* and *Waff*, 1977). Solid conductors (amorphous or graphitic carbon) (*Duba* and *Shankland*, 1982) as well as hydrogen (*Karato*, 1990) have also been suggested as explanations. Similarly, the high conductivities in the upper mantle in stable regions have been interpreted as an asthenospheric conducting layer caused by the presence of partial melt or solid conductors. The comparisons of the depths to the electrically conducting layer and the seismic low velocity zone at upper mantle depths indicate that the “electrical” asthenosphere is slightly deeper than the “seismic” asthenosphere (e.g. *Praus et al*., 1990; *Hjelt* and *Korja*, 1993) possible because the electromagnetic methods detect the depth of a warmer isotherm (*Bahr*, 1995). The correlation of the depths of various "geophysical" lithospheric thicknesses should, however, be treated with caution since each method observes different physical parameter.

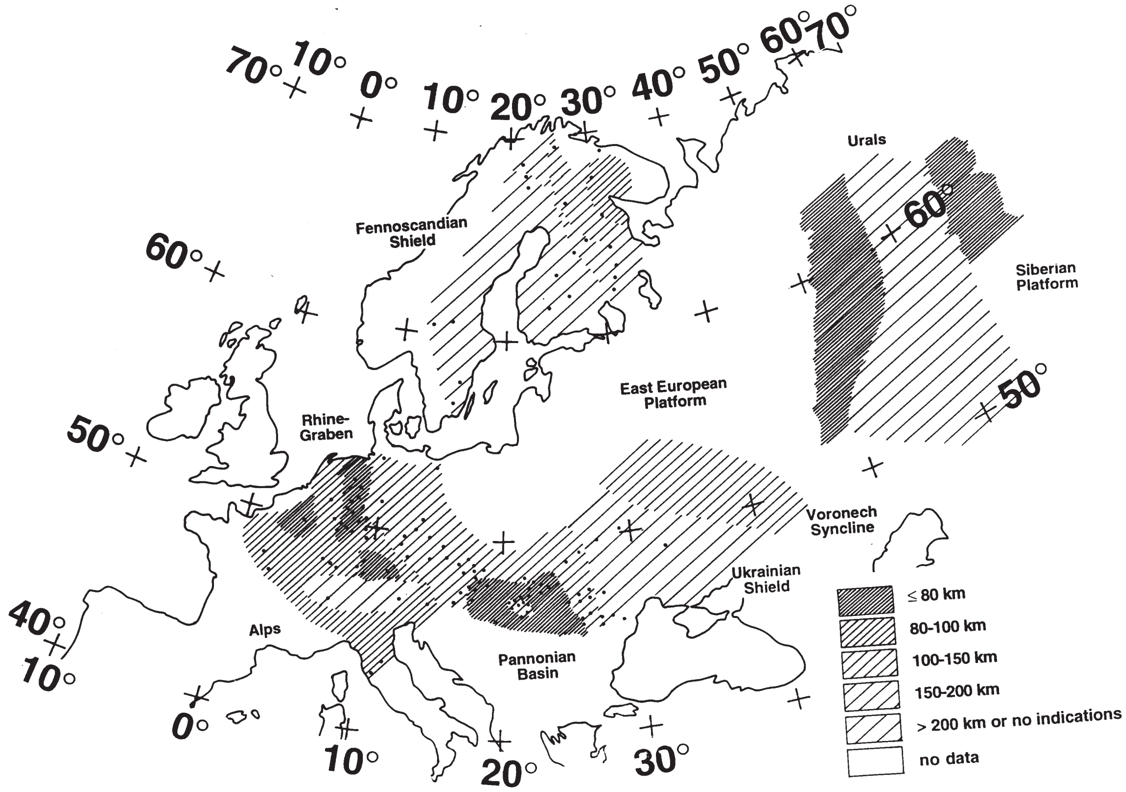

Fig. 2. A schematic map of the depths to the top of the electrical conductors coinciding approximately with the seismically defined lithosphere-asthenosphere boundary (LAB) in Europe (i.e. basically the first conducting layer in the upper mantle beneath the Moho which seems to coincide with the LAB although the electrical LAB is slightly deeper than the seismic LAB). Figure is from *Hjelt* and *Korja* (1993).

3. *Methods for determining electrical conductivity*

The subsurface conductivity can be studied by methods utilizing the time variations of the natural or artificially produced electromagnetic field and by observing the potential changes caused by d.c. electric currents injected into the Earth. The natural source methods are currently the only techniques with which the penetration to the deep crust and upper mantle can be achieved. The disadvantage is that several assumptions about the nature of the source and the primary electromagnetic field are made which may not be valid all the time or everywhere in the Earth. In practise, however, the assumptions are valid for those periods that are relevant for crustal studies almost everywhere in the Fennoscandia excluding the northernmost parts (*Viljanen*, 1996; *Viljakainen*, 1996). In the controlled source methods the properties of the source or the primary electromagnetic field are known but the sources are not powerful enough to allow the investigations of the deep crust. Therefore the use of the controlled source techniques is limited to the studies of the upper crust (0-20 km) (Table 1).

Table 1. Geoelectromagnetic methods used for studies of lithospheric electrical conductivity in the Fennoscandian Shield and the principal survey characteristics (modified from *Korja* and *Hjelt*, 1993).

Methods	**Survey characteristics**		
	Variation in conductivity	**Sampling distance**	**Depth range**
Natural source fields			
Magnetovariational (MV, GDS)[1),2)]	Horizontal	grid: 10-50 km	10-1000 km
Magnetotellurics (MT)	Horizontal/vertical	1-15 km	1-300 km
Audiomagnetotellurics (AMT)	Horizontal/vertical	0.1-2 km	0.5-20 km
Self potential (SP)[5)]	Horizontal	1-20 m	0-
Controlled sources			
VLF-resistivity (VLF-R)	Horizontal/vertical	10-50 m	0-0.5 km
Direct current (DC)	Horizontal/vertical	10-50 m	0-5 km
Airborne electromagnetic (AEM)[1)]	Horizontal	grid: 12 x 200 m	0-150 m
Controlled source tensor (AMT)[6)]	Horizontal/vertical	1-10 km	0-20 km
Magnetohydrodynamic (MHD)[3)]	Horizontal/vertical	0.5-25 km	-15 km
Fenno-Skan HVDC link[4)]	Horizontal/vertical	0.5 km -	0-

1) Magnetovariational (MV; or geomagnetic depth sounding; GDS) and airborne electromagnetic (AEM) ("wing-tip" Slingram) data also provide qualitative information on the variation of conductivity with depth.

2) Horizontal spatial gradient method (HSG) applied to magnetovariational data also gives quantitative information on the variation of conductivity with depth.

3) Direct current soundings and profilings (DCS, DCP) and frequency soundings (FS) utilizing signals from the magnetohydrodynamic (MHD) generator "Khibiny" located on Fisher Island (peninsula) of the Kola Peninsula.

4) DCS and FS utilizing signals from the high voltage direct current power transmission link between Finland (Pyhäranta) and Sweden (Forsmark) across the Gulf of Bothnia.

5) Not strictly a geoelectromagnetic method; utilizes self (spontantaneous) potentials caused by electrochemical reactions between an electronic conductor and surrounding electrolyte.

6) Depth range depends largely on how powerful sources can be used. The most powerful portable source available in Fennoscandia is a 26 kWh generator from Apatity.

Information on the entire lithosphere is inferred mainly from magnetotelluric soundings (MT) and magnetometer array studies (MV) in which the time variations of several spatial components of the Earth's electromagnetic field are recorded simultaneously. In MV studies (or in geomagnetic depth soundings; GDS) three orthogonal magnetic components are used whereas in magnetotelluric soundings two horizontal magnetic and electric components are used. The advantage of the simultaneous multi-component recording is that it provides directional information on the subsurface conductivity. Hence, even a single sounding allows to derive information on (i) the dimensionality of the conductivity structure (1D, 2D or 3D), (ii) the directionality properties of the conductivity structure (location of the conductor, geoelectric strike), and (iii) the actual resistivity variations of the subsurface beneath the observation site.

In an ideal case geoelectromagnetic studies should start with an areal mapping using a relatively large arrays with large spatial sampling distances. Anomalous conducting structures can then be studied and regional average values of conductivity be obtained at selected sites or along selected profiles with much smaller spatial sampling distance. In previous years it was common that areal sampling was carried out using large magnetometer arrays and the anomalous structures were studied by magnetotelluric sounding profiles. Modern geoelectromagnetic instruments, that are designed to record simultaneously the time variations of the five electromagnetic field components over a very wide frequency/period band (10000 Hz - 10000 s), combine these two methods. Hence one can envisage an ideal lithospheric study where a 2D array of five component instruments is employed simultaneously in order to get information about the conductivity structure of the target area in three dimensions.

Investigations of the internal "microstructure" of conductors revealed by large scale 2D arrays and/or profiles require, however, methods with denser lateral sampling and a frequency range compatible with near-surface and upper crustal depths. These include e.g. plane wave methods such as VLF resistivity (VLF-R) profilings and natural source audiomagnetotelluric (AMT) soundings, magnetic dipole type measurements such as airborne electromagnetic surveys (AEM) and controlled source tensor audiomagnetotelluric (AMT) soundings as well as direct current (d.c.) soundings and profilings using electric dipoles as a source (transmitter).

Finally, the information from boreholes and from petrophysical analysis of relevant rock samples at appropriate temperature, pressure and fluid conditions provide valuable information about the conduction mechanisms and form the link between the geophysical and geological properties.

A variety of electromagnetic methods enable us to identify structures ranging from regional scale (100 km) down to local features with dimensions of several metres or even to structures having dimensions of only a fraction of millimetres, thereby making the appropriate combination of large scale and small scale methods a powerful tool for interpreting conducting structures within the Earth. The various techniques allow to investigate in detail both the position of conductors with respect to geological and tectonic structures and the internal structure of the conductors themselves as well as to understand the conduction mechanism taking place in rocks.

4. The conductivity models: resolution and interpretation

Resolution of conductive structures depends on the spacing of sounding sites as well as the range and density of signal frequencies used for obtaining depth variation of conductivity. The ultimate limit for the resolution is, however, set by the diffusive nature of the propagating electromagnetic waves. For the vertical resolution it is safe to say that a conductivity increasing towards depth is easier to resolve than a decreasing conductivity and that due to screening the conductance (thickness x conductivity) of the overlying strata has to be at least twice smaller than the conductance of any deeper layer to be resolved by EM techniques. The depth to the conducting layer and its conductance are the best resolved parameters. Complementary information from other geophysical methods is the best way to improve the resolution. In 2D and 3D Earth the structures with dimensions smaller than the lateral spatial sampling distance and the depth of investigation for a given period are poorly resolved by electromagnetic methods. If lateral sampling is dense enough (denser than the depth of investigation) then the lateral resolution for the 2D Earth is roughly equal to lateral sampling distance for the responses perpendicular to the strike but to the depth of investigation for the responses along the strike.

Especially in shield areas electrically resistive blocks are often surrounded by narrow, highly conducting bandlike formations with a high conductance (even 20.000 Siemens). Conductive structures, which continue over considerable distances along strike are described with two-dimensional (2D) models. The 2D Earth then gives a directional total electromagnetic response i.e. responses along the strike and perpendicular to it differ from each other. This is sometimes referred to as geoelectric anisotropy. Strictly speaking this anisotropy is structural by nature. The Earth around the conductor is isotropic and the responses have an apparent anisotropic behaviour due to remote conductor. The structural ("apparent") anisotropy can be distinguished from truly anisotropic case if the measuring profile (data) extends over the conducting body or the data have a regular trend along the data profile. If the Earth hosts several 2D conductors which are so deep or so close to each other that the electromagnetic methods cannot resolve them separately then, again, the Earth gives a directional response but due to limits in resolution it is not possible to distinguish this 2D model from models, where the crust is layered (1D) and genuinely anisotropic. *Jones* (1992) calls these two cases macro- and intrinsic (micro-) anisotropy, respectively. It should be noted that magnetotelluric method can detect only horizontal (azimuthal) anisotropy and cannot distinguish between macro- and microanisotropy. A magnetic tipper vector (induction vector) can be used to distinguish between them as in an 1-D Earth the vertical magnetic field vanishes (*Zhang et al*., 1993).

A schematic example of the interpretation of electromagnetic data and conductivity models is given in Fig. 3. The figure shows a generalized model for the structure of the upper to middle crustal conductors found in the Fennoscandian Shield, as well as the effect of sampling distance on the models. By increasing the lateral sampling density, smaller and smaller structures can be identified and consequently the internal structure of a large scale block conductor can be resolved. It should, however, be kept in mind that in general the lateral resolution decreases rapidly with increasing depth and the implications for deep structure derived from the detailed near-surface structure are always by nature extrapolations.

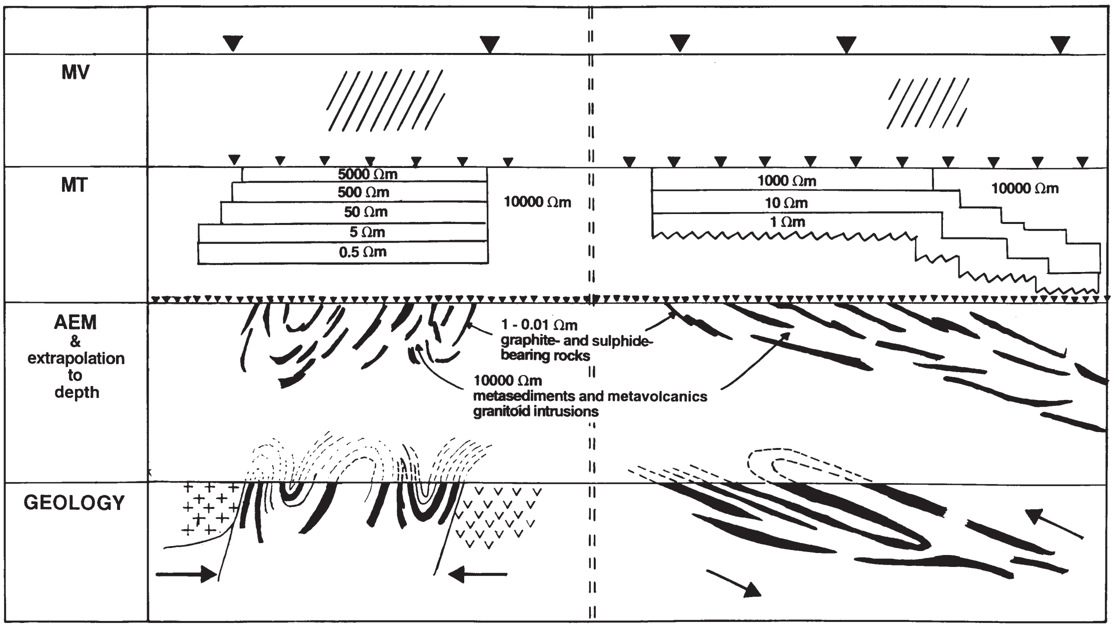

Fig. 3. A cartoon showing the effect of decreased sampling distance: When the lateral sampling distance (space between the black triangles) is gradually decreasing additional details of the internal structures are revealed. In magnetovariational (MV) and magnetotelluric (MT) studies the depth of investigation may be in favourable conditions few hundreds of kilometres whereas in airborne electromagnetic (AEM) mappings it is few tens of metres and the interpretation that the deep conductors have the properties that we image on the surface is always an extrapolation based on other available information. Figure is from *Korja* and *Hjelt* (1993).

The magnetovariational method allows the lateral position of the conductor and some estimation of the depth of the conductor to be determined (uppermost panel in Fig. 3). The magnetotelluric method, with a smaller distance between the measuring sites, allows the location and geometry of the conductor and the internal average resistivity structure to be determined (second panel in Fig. 3). The near-surface structure of a conductor, if it is also expressed at the surface, can be constructed from airborne electromagnetic data with a very dense lateral sampling (third panel in Fig. 3).

The results from different parts of the Fennoscandian Shield have shown that in most cases the large scale conductors detected by MV and MT methods are composed of thin, extremely conducting layers (black zones of 0.1 Ωm or less in Fig. 3) embedded within more resistive host rocks (white areas of several thousands of Ωm in Fig. 3). It should be noted that airborne electromagnetic data carry information only on a very thin (at a crustal scale) near-surface section of the Earth’s crust. The structure of the deeper parts of the crust in the third panel of Fig. 3 is only an extrapolation from airborne electromagnetic results in conjunction with magnetovariational and magnetotelluric data that provide information on the deeper parts of the crust. The information about the properties of deeper conductors is ultimately always an extrapolation of the truly resolved near-surface structures since it is impossible to resolve details in the scale of hundreds of metres at the depths of several kilometres or tens of kilometres.

The lowermost panel in Fig. 3 shows the first order geological interpretation of structures; the structures to the left have been interpreted (in this hypothetical case) to be remnants of rock layers that were complexly folded due to the collision of two crustal blocks, while the structures to the right are attributed to a collision between two crustal units with one block being thrust beneath the other.

5. Geoelectric data sets in the Fennoscandian Shield

During the last two decades an extensive program of geoelectromagnetic studies was carried out in Fennoscandia to develop geoelectric models for the Fennoscandian Shield. These include work done by the EM Induction Groups in Oulu and Uppsala as well as the work of several groups in former Soviet Union. The methods used for these studies have been described in Table 1. The available data sets have been reviewed by *Korja* and *Hjelt* (1993) from where a comprehensive list of references for original studies can be found.

The unique airborne electromagnetic (AEM) data collected by the Geological Survey of Finland over the last decades have been recently compiled into a surface conductance map of Finland (*Peltoniemi et al*., 1992). The airborne data, in particular, in the form of 1:100000 profile maps of real and imaginary components have been used to delineate near-surface conducting structures in order to study the internal structure of the possible surface expressions of deeper conductors (e.g., *Korja* and *Koivukoski*, 1994; *Korja et al*., 1996a). Recently the airborne electromagnetic data have been used in the form of apparent conductivity maps (*Korja et al*., 1996b) that were obtained by transforming the real- and imaginary data into apparent conductivity and depth using the program of *Pirttijärvi* (1996).

An extensive magnetometer array work started in early eighties (e.g. *Pajunpää*, 1987, 1989). The MV array data cover Finland between latitudes 60 and 66.5°N and northern Sweden between 64 and 66.5°N and have provided a general horizontal distribution of electrical conductivity in the central parts of the Fennoscandian Shield (Fig. 4a). Central and southern Sweden have been covered by similar magnetometer arrays but the data have not been fully processed and analysed yet. *Jones* (1981) and *Jones* et al. (1983) used magnetometer data of the IMS project (1976-1979) for lithospheric studies. MV data exist also from the Karelia in the former Soviet Union (e.g. *Rokityansky*, 1983).

The MT data (Fig. 4b) include soundings carried out in Russian Karelia by several groups (e.g. *Kovtun*, 1976; *Golod* and *Klabukov*, 1989; *Kovtun et al*., 1992), whereas the Finnish and Swedish data come from some early work of *Adám et al*. (1982) and *Jones* et al. (1983) and mainly from measurements made by the Uppsala and Oulu groups (e.g. *Rasmussen et al*., 1987; *Rasmussen*, 1988; *Zhang et al*., 1988; *Pedersen et al*., 1992 and *Kaikkonen et al*., 1983; *Korja et al*., 1986; *Korja et al*., 1989; *Pernu et al*.

1989; *Vaaraniemi*, 1989; *Korja*, 1990; *Korja* and *Koivukoski*, 1994; *Korja*, 1993). MT soundings have been carried out also in Denmark along the EGT transect by Uppsala and Aarhus groups (*Rasmussen et al.*, 1992; *Hjelt*, 1992). The groups in Finland, Sweden and Denmark have utilized similar equipment all through the 80's and early 90's having produced some 700 soundings. The MV and MT data form thus a comprehensive and unified data set, unique in the whole world.

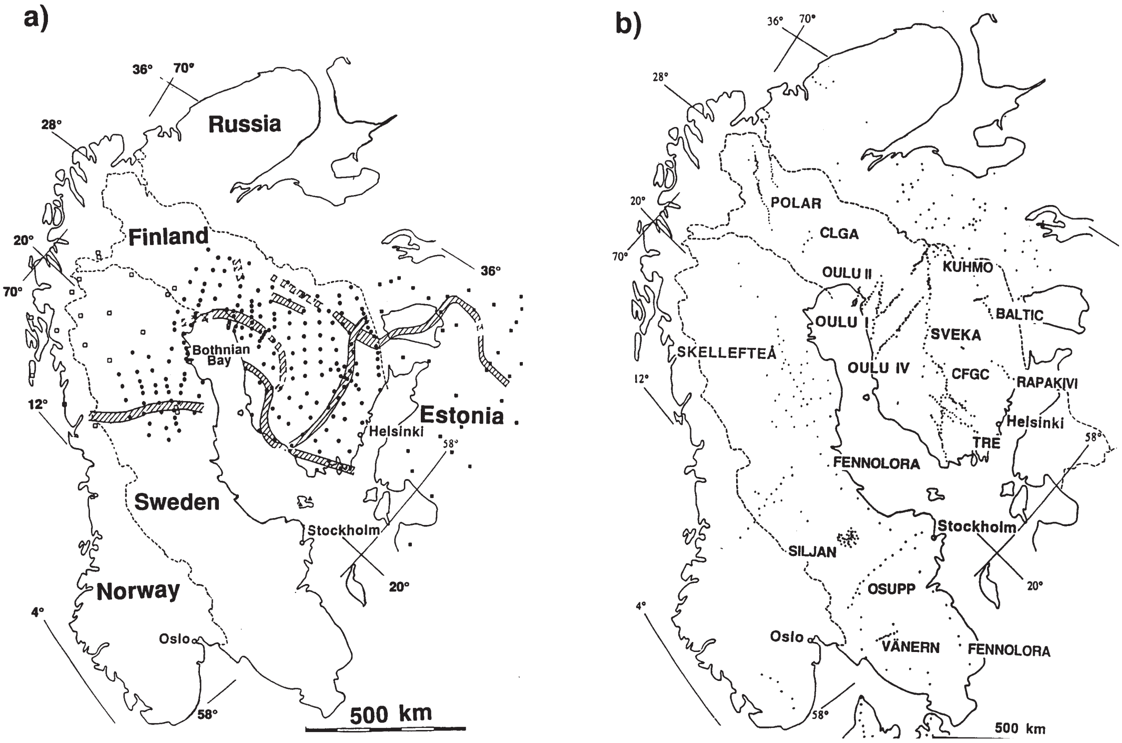

Fig. 4. Deep geoelectromagnetic studies in the Fennoscandian Shield. (a) Magnetometer sites and conductivity anomalies revealed by MV data. Central and southern Sweden have been covered by magnetometer arrays but the data have not been analysed yet. Ruled line: conductivity anomaly; dashed ruled line: conductivity boundary. (b) Magnetotelluric sites in the Fennoscandian Shield and the names of some profiles. Figure is from *Korja* and *Hjelt* (1993).

In the northeastern part of the shield frequency soundings (FS) exploiting energy from a magnetohydrodynamic (MHD) generator located at the Fischer Island of the Kola Peninsula have been used for geoelectric investigations. Data have provided both average crustal resistivity variations as well as detailed profile results (*Heikka et al.*, 1984; *Velikhov et al.*, 1987; *Vanyan et al.*, 1989; *Zhamaletdinov* 1990, *Zhamaletdinov et al.*, 1993). Recently the data that can be obtained observing the signal from the Fennoskan DC Power Link between Finland and Sweden have been used to estimate the electrical properties of the crust around the cathode of the power line (close to Pyhäranta near Rauma) (*Kaikkonen et al.*, 1996).

Locally, in various parts of the shield, audiomagnetotelluric soundings (e.g. *Kaikkonen* and *Pajunpää*, 1984, *Korja et al.*, 1989), VLF-R (*Pernu et al.*, 1989; *Korja et al.*, 1996a), d.c. dipole-dipole profilings (*Pernu et al.*, 1989; *Pernu*, 1991) and self potential measurements (*Korja et al.*, 1996a) have produced data for crustal studies.

6. *Electrical conductivity of the Fennoscandian Shield*

The results of the various geoelectromagnetic studies have recently been summarised by *Korja* and *Hjelt* (1993, 1996) including comprehensive references for original studies. *Korja, A. et al.* (1993) combined the results of electromagnetic studies and the results of an extensive deep seismic sounding program carried out in Fennoscandia (e.g. Korhonen et al., 1990, *Luosto*, 1991). Based on the geoelectromagnetic and deep seismic sounding data they showed evidence for Palaeoproterozoic collisional and extensional tectonics and proposed a model for the evolution of the shield.

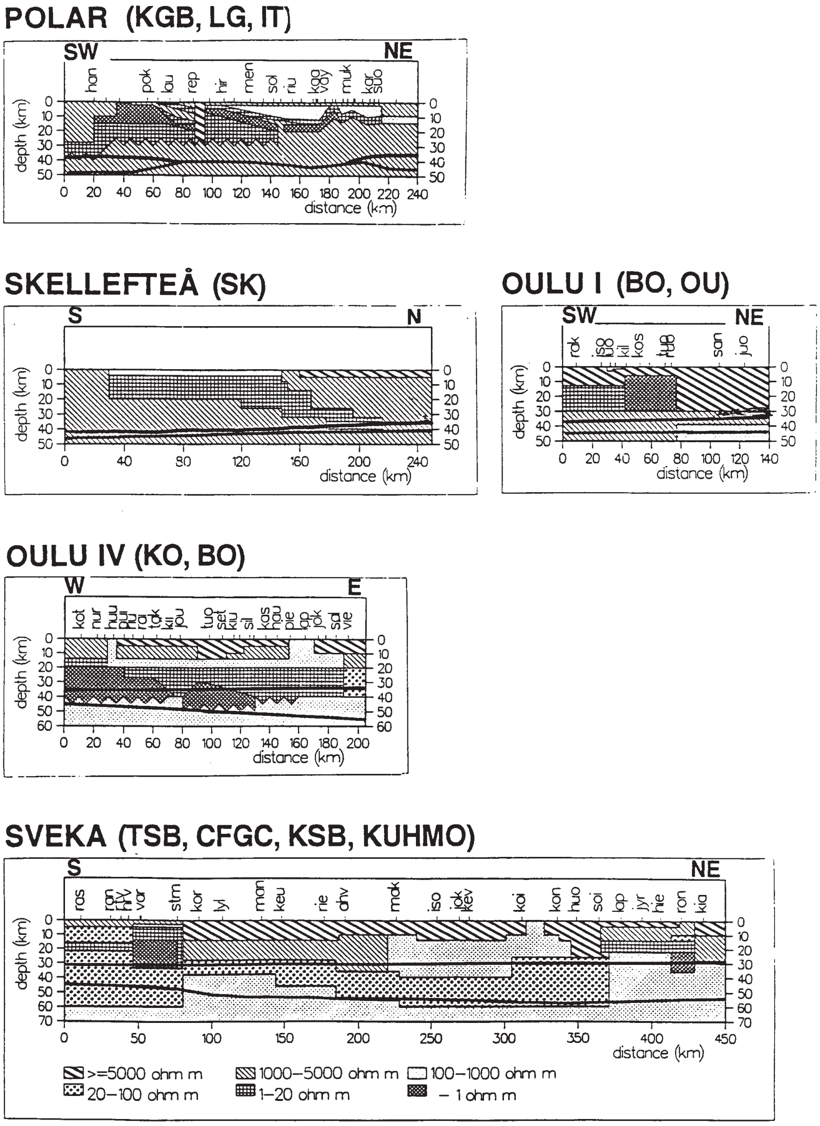

Fig. 5 Two-dimensional models of the POLAR, SKELLEFTEÅ, OULU I, OULU IV and SVEKA profiles. The resistivity scales of all models are the same and is shown in detail below the SVEKA model. The profiles have no vertical exaggeration. The thick black lines indicate the boundary between the upper and lower crust and the Moho boundary. Figure is from *Korja* and *Hjelt* (1993) where references for original studies are given.

We shall next briefly outline the main results of the electromagnetic investigations completed in the Fennoscandian Shield and then discuss on their tectono-geological interpretation in later section. Main features are shown in three figures: The two-dimensional resistivity models obtained from magnetotelluric data are shown in Fig. 5. The distribution of near-surface conductors mapped by the airborne electromagnetic surveys are shown in Fig. 6. The map is modified from the map of *Peltoniemi et al.* (1992) by using new low altitude data in Bothnian Kainuu, and Lapland regions. Airborne EM methods image very shallow structures (0-150 m) and information on the continuation of conductors deeper into the crust requires additional information from e.g. magnetotelluric soundings. The main belts of conductors compiled from all available geoelectromagnetic data (mainly MV, MT and AEM) are shown together with the major tectonic units of the shield in Fig. 7 (see also Fig. 4a for the location of the conductivity anomalies mapped by the MV method).

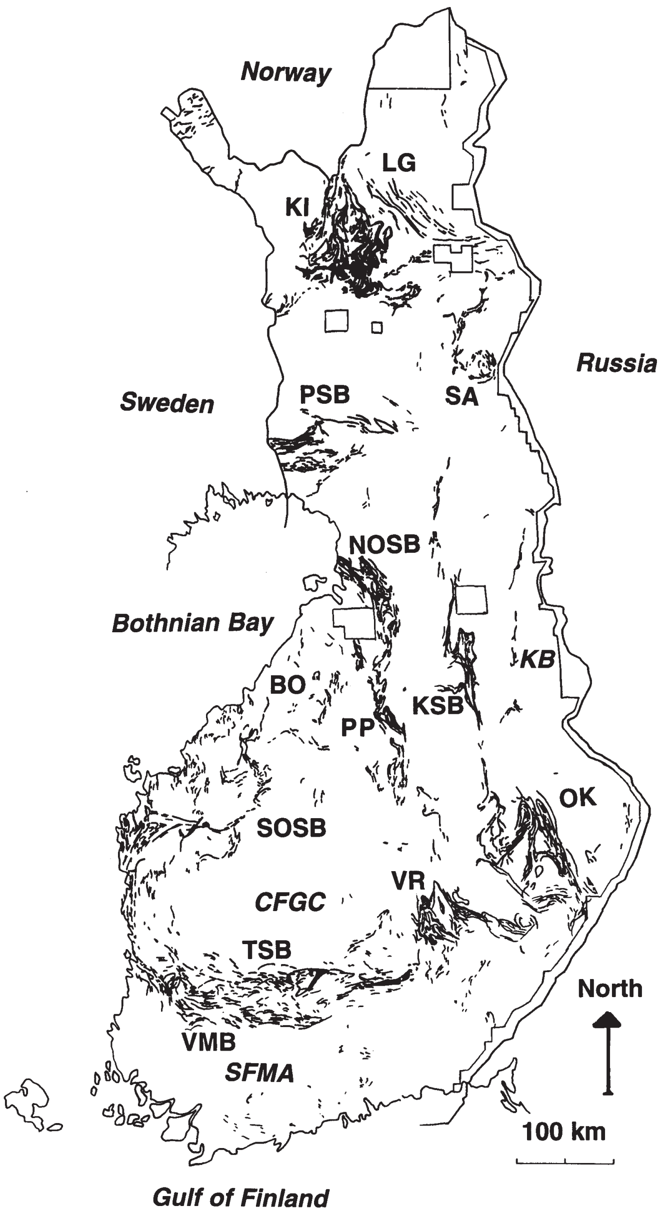

Fig. 6. Near-surface conductors (black) observed by high (150 m) and low (30 m) altitude airborne electromagnetic surveying. Map is modified from *Peltoniemi et al.* (1992). No data from squared regions. Conductors: BO=Bothnian Schist Belt, KI=Kittilä region, KSB=Kainuu Schist Belt, LG= Lapland Granulite Belt, NOSB= Northern Ostrobothnian Schist Belt, OK=Outokumpu region, PP=Pielavesi-Pyhäjärvi region, PSB=Peräpohja Schist Belt, SA=Salla Schist Belt, SOSB=Southern Ostrobothnian Schist Belt, TSB=Tampere Schist Belt, VMB=Vammala Migmatite Belt, VR=Virtasalmi-Rantasalmi region. Resistive regions: CFGC= Central Finland Granitoid Complex, KB=Archaean Kuhmo Complex, SFMA=Southern Finland Migmatite Area. Figure is from *Korja* and *Hjelt* (1996).

6.1 Upper and middle crustal conductivity: Conductive belts

Results have shown that the crustal conductivity variations in the Fennoscandia Shield are large and range from 10^6 Ωm to below 10^{-1} Ωm that are extreme values compared with those normally observed in crustal rocks (*Haak* and *Hutton*, 1986; *Jones*, 1992). Several elongated conductors which are either many hundred kilometres long or which form belts of discontinuous conductors, delineate electrically more resistive and laterally more homogeneous blocks within the shield. The main conductors, or conductive regions, in the central, northern and eastern parts of the shield are, from NE to SW, as follows:

(i) A set of NW-SE trending, southwestward dipping conducting zones in the Kola Peninsula and in northern Norway including the Polmak-Pechenga (P) and Imandra-Varzuga (IV) conductors. The Pechenga conductor reach a depth of 7-10 km (*Zhamaletdinov*, 1990) but further to northeast along the Polmak-Pechenga Schist Belt in Norway the conductor rapidly becomes very thin and shallow reaching the depth of about 1.5 km (*Zhamaletdinov et al*., 1993). The conductors in the Pechenga area can be associated with the metasedimentary rocks of the Pilgujärvi Suite of the Pechenga Complex (the graphite- and sulphide bearing rocks of the so-called “productive pile”) (*Hanski*, 1992).

(ii) A NW-SE trending, southwestward dipping Inari-Allarechen conductor (IA) along the northeastern boundary of the Lapland Granulite Belt. In Finland the conductor is nearly vertical and reaches the depth of about 10-15 km (*Korja et al*., 1989).

(iii) A northeastward dipping conductor (LG) at the lowermost basal part of the Lapland Granulite Belt and beneath it (*Korja et al*., 1989). The conductor is exposed at the SW margin of the LGB and reaches the depth of about 15 km further to the NE. The rocks of the granulite belt above the basal conductor have an average resistivity of 200-1000 Ωm which is much lower than the typical upper and middle crustal resistivities found elsewhere in the Fennoscandian Shield (see below).

(iv) The NW-SE trending zone of “sub-horizontal” conductors from the Kittilä Greenstone Belt (KI) in northern Finland to the Vetrenny-Poyas Greenstone Belt (VPB) and Lake Onega (LO) in Russian Karelia in SE. In the Kittilä Greenstone Belt the conductive Palaeoproterozoic supracrustal rocks above the Archaean craton have a thickness of about 10 km and seem to dip beneath the Lapland Granulite Belt (POLAR profile, Fig. 5) (*Korja et al*., 1989). The enhanced conductivity in the Kittilä Greenstone Belt is caused by few tens of metres wide graphite and sulphide schist layers and subordinate iron formations (*Lehtonen et al*., 1985) within otherwise resistive bedrock consisting mainly of metavolcanic rocks. The conductors around the Lake Onega (LO) can be associated with shungites (*Golod* and *Klabukov*, 1989) and reach a maximum thickness of about 2 km.

(v) Near-surface conductors (PSB) mapped by the AEM method and associated to the Palaeoproterozoic Peräpohja Schist Belt of the Karelian Province (Fig. 3). No information on the depth of these conductors is available.

(vi) A belt of discontinuous conductors coincident with the boundary zone between the Archaean Karelian and Palaeoproterozoic Svecofennian including the Lake Ladoga (LL), Outokumpu (OK), Kainuu Schist Belt (KSB), Pielavesi-Pyhäsalmi (PP), Oulu (OU) and Northern Ostrobothnian Schist Belt (NOSB) conductors. In Lake Ladoga region the deep conductor (LL part) is located at a depth of about 10 to 15 km and more to the south it turns westwards and finally becomes unresolvable due to screening of the Phanerozoic sedimentary cover (*Rokityansky*, 1983). In the Outokumpu region (OK part) the deep conductor is located in the upper to middle crust (10-15 km; *Pajunpää*, 1989) and is situated slightly to the west of the shallow conductors of the Outokumpu formation proper. In Kainuu and North-Savo the two zones of near-surface conductors have been detected by airborne and audiomagnetotelluric surveys (e.g. *Kaikkonen* and *Pajunpää*, 1984). The PP conductor along the lithological Archaean-Proterozoic (Karelian-Svecofennian) boundary is associated with the Palaeoproterozoic rocks of the northern part of the Savo Schist Belt (Pielavesi-Pyhäjärvi region) and the KSB conductor with Palaeoproterozoic Kainuu Schist Belt. The deep conductor beneath the Iisalmi block (roughly between the KSB and PP) dips southwestward (SVEKA profile; Fig. 5) and reaches a depth of 30 km close to the Karelian-Svecofennian boundary where it terminates. The correlation between the deeper conductor and the near-surface conductors of the Kainuu Schist Belt is not straightforward since part of the supracrustal rock sequences of the KSB may represent a nappe that was overthrust from SW to its present position. Around the Bothnian Bay a set of conductors have been identified of which the nearly vertical Oulu-conductor (OU part) is located at the Archaean-Proterozoic boundary (OULU I profile; Fig. 5) (*Korja et al.*, 1986). The OU and NOSB conductors may represent a single conductor that dips southwestward.

(vii) A sub-horizontal Bothnian conductor (BO) on the Finnish side of the Bothnian Bay (Oulu I profile; Fig. 5). The application of the horizontal spatial gradient (HSG) method to magnetovariational data yielded a depth of 10-15 km and a thickness of about 25 km for the Bothnian conductor (*Pajunpää*, 1988) which are comparable to the values obtained from magnetotelluric data (OULU I and IV profiles; Fig. 5) (*Korja et al.*, 1986; *Vaaraniemi*, 1989).

(viii) A NW-SE trending and northeastward dipping Skellefteå-Storavan conductor (SK, ST) in Sweden (*Jones*, 1981; *Rasmussen et al.*, 1987). The conductor plunges deeper into the crust south of the Skellefteå Belt and extends to the north and northeast about 80 km, where the upper surface of the conductor reaches the depth of 35 km (SKELLEFTEÅ profile; Fig. 5). The conductor terminates close to the SW limit of the Archaean crust inferred from the Sm-Nd isotope systematics (*Claesson et al.*, 1993). To the south of the Skellefteå Belt the conductor is located close to the surface and has an extension of at least 200 km.

(ix) An eastward dipping Kokkola (KO) conductor in the Finnish side of the Bothnian Bay OULU IV profile; Fig. 5) at middle to lower crustal depths. Magnetotelluric data from the OULU IV profile (*Vaaraniemi*, 1989; *Korja*, 1993) indicate that the most conducting rocks beneath the MV conductivity anomaly are observed at the depth of about 20 km and below. The conductor dips eastward and terminates at about 150 km to the east of the MV conductivity anomaly where the conductor reaches a depth of 40 km. This location coincides well with the conductivity boundary delineated by the MV method (dashed line in Fig. 4a) (*Pajunpää et al.*, 1983). If we projected the KO conductor to the west using the dip observed along the OULU IV profile then the conductor would reach the surface somewhere in the middle of the Bothnian Bay. There are no information on the crustal electrical properties from the sea region and it is unknown whether the conductor is exposed and what is an exact relationship between the Skellefteå and Kokkola conductors.

(x) A 500 km long, E-W trending and nearly vertical conductor (TSB&VMB) in southern Finland (*Pajunpää*, 1989; *Pernu et al*., 1989; *Korja* and *Koivukoski*, 1994). The conductor extends from the Outokumpu region in eastern Finland to the western coast of Finland. MV data indicate that in the sea region it may turn to the north. In the west the conductor is several tens of kilometres wide. Airborne survey has imaged several elongated near-surface conductors on top of the deep conductor indicating that the deep conductor is exposed in the western part. In the easternmost part the 5-10 km deep conductor becomes narrower (*Pajunpää*, 1986) and has no surface expression according to airborne data. To the north of the deep conductor, however, a region of near-surface conductors (VR) is observed (Figs. 6 and 7).

6.2 Interpretation of upper and middle crustal conductors: conductive terrane boundaries?

The crustal conductors described in the previous section share many common features and can be grouped into large scale belts. The common features are:

(i) Conductors have a conductance of several thousands of Siemens. This excludes fluids from being the primary cause for enhanced conductivity and indicates instead that an electronic conducting material is present. In crustal scale most likely candidates are graphite, sulfides and some oxides.

(ii) Most of the conductors are located in the seismically defined upper and middle crust and conductors do not penetrate into the lower crustal layer as defined by high P-wave velocities. In those places where conductors extend into the lower crust (e.g. Skellefteå conductor) the high velocity layer is either absent or very thin.

(iii) Large scale crustal block conductors have a variety of geometrical orientations: they may be subhorizontal (Kittilä Greenstone Belt), inclined (Granulite Belt, Skellefteå), nearly vertical (Southern Finland, Oulu conductor).

(iv) In those regions where magnetotelluric and seismic reflection profiles coincide (e.g. the POLAR profile and to some extent also the Bothnian Bay region) it is evident that inclined good conductors and inclined bands of reflectors coincide spatially. It is, however, not necessary that enhanced reflectivity and enhanced conductivity have a common origin. The reflectors and conductors coincide spatially because the same large scale tectonic processes formed both.

(v) Conductors are usually caused by extremely conducting (> 1 S/m) graphite- and sulphide-bearing metasedimentary rock layers hosted by rocks having a resistivity of several thousands or tens of thousands of Ωm. The sediments that are now responsible for enhanced conductivity were deposited ca. 2.1 Ga ago or later (*Korja* and *Hjelt*, 1996) in a variety of tectonic environments favourable for carbon and sulphur to accumulate. Subduction and collision deformed sedimentary sequences between crustal masses and underthrust them deep into crust or emplaced them closer to the surface. As a consequence the conductive metasedimentary rock layers are interconnected in a very complicated manner, forming a network of conductors. It should be noted, however, that graphite- and sulphide-bearing metasedimentary rocks layers are the primary but not the only source for enhanced conductivity. There are several examples where graphite- and sulphide-bearing shear zones may have very high conductivities. The saline brines in the upper crust may also provide an electrical connection between the metasedimentary rock layers and therefore further enhance the conductivity.

The upper and middle crustal conductors (i - x) described in the previous chapter can be grouped into four large scale belts: (I) the conductors within the Archaean Karelian Province, (II) the conductors within the Lapland-Kola orogen in the northeastern part of the shield, (III) the conductors at the Archaean-Proterozoic margin, and (IV) the conductors within the Svecofennian Domain (Fig. 7).

A brief synthesis, following *Korja* and *Hjelt* (1996), of the development of the conductive formations is outlined in steps related to the four stages in the Proterozoic evolution of the Fennoscandian Shield. These stages are (1) extension and rifting of the Archaean craton and deposition of sediments; (2) the Lapland-Kola orogeny and the development of conductive terrane boundaries and foreland conductive belts in the Lapland-Kola orogen and Karelian Province; (3) the first collision of an island arc complex in the Svecofennian orogeny and the development of the Archaean-Proterozoic boundary (Karelian Province - Svecofennian Domain) conductor; (4) the second collision and the development of the Southern Finland - Kokkola - Skellefteå conductor (internal Svecofennian terrane boundary).

After the cratonization at about 2.5 Ga ago and before the final break-up at about 2.0 Ga ago the Archaean craton is thought to have experienced a long (500 Ma) period of several extensional events and rifting during which several cycles of sedimentation and volcanism occurred (e.g. *Hanski*, 1992; *Gorbatschev* and *Bogdanova*, 1993). Several intracratonic extensional basins were developed and some of them ultimately were opened to small and/or large oceanic basins (e.g. Pechenga basin; *Hanski*, 1992, and the ocean southwest of the Karelian Province) where the sedimentation continued along the new continental margins. Voluminous accumulation of sedimentary and volcanogenic material filled the basins. At first stages the conditions were arid but later the sedimentation and volcanism took place in shallow and in some places in deep sea environment (e.g. *Hanski*, 1992; *Lehtonen et al.*, 1992). Only the last phase of the extensional period between 2.1 and 2.0 Ga ago seems to have been capable to produce conductive sedimentary sequences. For example, in the Pechenga Complex only the metasedimentary rocks of the uppermost volcano-sedimentary sequence, i.e. the Pilgujärvi Suite, are conductive. Similarly, in the Kittilä Greenstone Belt only the Porkonen and Kautoselka Formations of the Upper Lapponi Group contain conductive material and the sedimentation of conductive formations in the Lake Onega region took place between 2.115 and 1.98 Ga (*Karhu*, 1993).

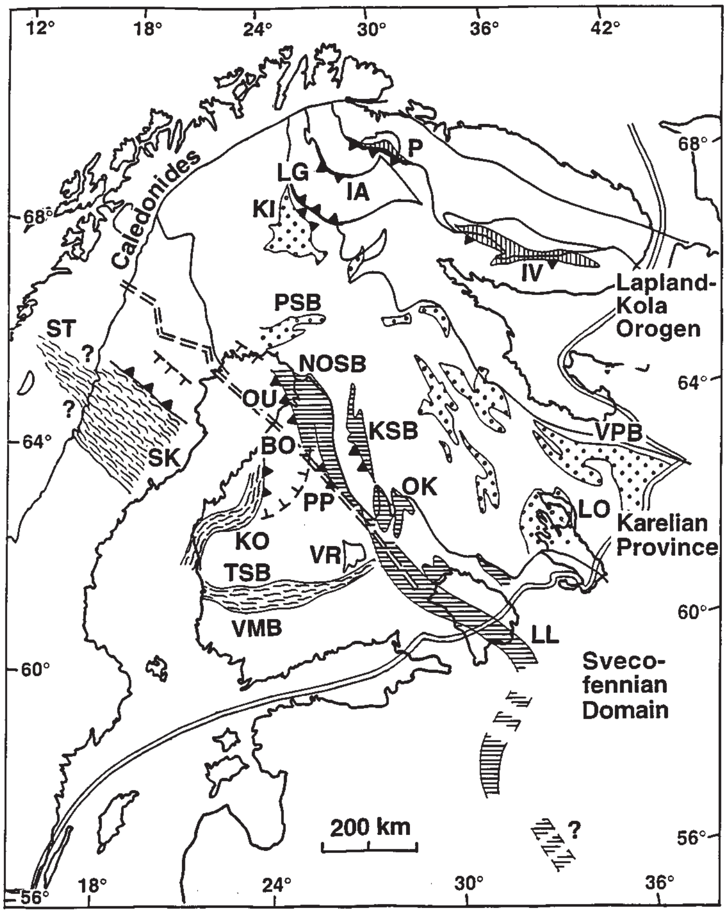

Fig. 7. Upper and middle crustal conductors in the central and northeastern Fennoscandian Shield based on MV, MT, MHD and AEM data. For inclined conductors the shaded belts show the surface projection of the uppermost part of the dipping conductive body. The dip directions are shown in places where the conductors plunge deeper into the crust. The termination of a dipping conductor is shown by a dashed line with barbs pointed towards the conductor. Geological boundaries are from *Gorbatschev* and *Bogdanova* (1993). The SW extent of the Archaean crust beneath the Palaeoproterozoic strata is shown by a double dashed line and is based on isotope studies by *Huhma* (1987) and *Öhlander et al.* (1993). Conductors: 1-Lapland-Kola region (vertically ruled): IA=Inari-Allarechen, IV=Imandra-Varzuga, LG=Lapland Granulite Belt, P=Pechenga; 2-Karelian province (circles): KI=Kittilä (Central Lapland Schist Area), LO=Lake Onega, PSP=Peräpohja Schist Belt, VPB=Vetrenny-Poyas Belt; 3-Archaean Karelian -Palaeoproterozoic Svecofennian (horizontally ruled): KSB=Kainuu Schist Belt, LL=Lake Ladoga, NOSB=Northern Ostrobothnian Schist Belt, OU=Oulu, OK=Outokumpu, PP=Pyhäsalmi-Pielavesi; 4-Svecofennian (waved): BO=Bothnian, KO=Kokkola, SK=Skellefteå, ST=Storavan, TSB/VMB/=Tampere Schist Belt/Vammala Migmatite Belt, VR=Virtasalmi-Rantasalmi. Figure is from *Korja* and *Hjelt* (1996).

It is not clear what kind of basins was developed in the Lapland-Kola region. Some models suggests that the ocean was opened between the Central Kola terrane and the Belomorian Belt (Pechenga- Imandra-Varzuga zone) or at the site of the Lapland Granulite Belt and its counterparts in the Kola Peninsula (Umba Granulite Belt and Kolvitsa Schist Belt) or that both the Lapland Granulite Belt and the Pechenga Belt have been formed in a single depositional basin (e.g. *Gaál*, 1990; *Gorbatschev* and *Bogdanova*, 1993 and references therein). Conductivity models may not alone solve the problem but they may indicate sites that separate distinct crustal masses (terranes) and provide information on deeper parts. The Polmak- Pasvik belt seems to be minor compared to the Lapland Granulite Belt: the conductors at the basal part of the Lapland Granulite Belt extend to the depth of about 15 km whereas the conductors in the Polmak-Pechenga belt flatten to very shallow structure (1.5 km) beyond the Pechenga Complex, proper (7 km). This suggests that the Lapland Granulite Belt, instead of the Polmak-Pechenga Belt, may represent a site of a hidden suture.

The second stage corresponds with the time roughly after the peak of the Lapland-Kola orogeny and the break-up of the Archaean craton in SW at about 1.95 Ga ago. The convergence in the Lapland-Kola have closed the basins and the conductive sedimentary sequences were deformed and emplaced to their current positions. The southwestward dipping Polmak-Pechenga and Imandra-Varzuga conductors were produced when the supracrustal sequences were upthrust northeastward onto the Central Kola terrane. The present day geometry of the LGB and the southwestward dipping Inari-Allarechen conductor were formed when the rocks of the LGB were upthrust southwestward onto to Karelian Province.

Within the Karelian Province the supracrustal sequences also were deformed and metamorphosed during the Lapland-Kola orogeny and most likely also later during the Svecofennian orogeny at least in the northern part of the shield. The remnants of this huge volcano-sedimentary belt extending from the Caledonian front in NW to the East European Platform in SW form now sporadically distributed schist belts including the Kittilä, Lake Onega and Vetrenny Poyas Belt conductors. The identification of ophiolites in the Kittilä Greenstone Belt (*Hanski*, 1995) combined with similar results from the southern part of this belt suggests that the belt contains rocks formed in oceanic environment. This in turn may indicate that there was an ocean east of the Karelian Province between the Karelian and the Belomorian Belt at the Palaeoproterozoic time. Consequently the conductive sedimentary rocks, that deposited during the last stages of the rifting period, may have deposited at either one or both of the newly developed continental margins. The Karelian Province may have been a foreland during the collision of the Belomorian Belt against the Karelian Province and this belt of conductors may also be considered as a “collisional conductor”, an image of a wide foreland of a collision rather than representing a terrane boundary.

The third stage represents the first Svecofennian collision. The conductors along the boundary between the Karelian Province and the Svecofennian Domain were emplaced at their current positions during the collisions of crustal masses of a Palaeoproterozoic island arc complex against the Archaean craton. The first terrane collision started soon after the final break-up, first in the southwest (Lake Ladoga region) at about 1.91 Ga ago and later in the northwest (Kainuu and Bothnian regions) at about 1.90 Ga ago (*Lahtinen*, 1994). The collision squeezed the extended cratonic margin and emplaced the sedimentary material and slices of the rifted Archaean crust onto the Karelian province to NE. In the Outokumpu region the conductors observed by the AEM method represent overthrust sedimentary material whereas the deeper conductor (at 10 - 15 km) revealed by the MV technique west of the AEM conductors may represent similar sedimentary material at deep crustal levels. The PP and KSB conductors further to the north may represent material from the same basin (margin) if the younger part of the KSB was thrust over the Archaean Iisalmi blocks or from different basins if a small basin was opened between the Karelian Province, proper, and the Iisalmi block.

The last stage represents the second Svecofennian collision, again producing crustal conductive belts. A new northward subduction south of the recently accreted terrane consumed the oceanic crust and finally closed a basin between the first and second terranes at about 1.88 Ga ago resulting in a collisional accretion of the second terrane (*Lahtinen*, 1994). Because of the geometry of the terranes the crustal shortening may have been much more intensive in southern Finland than in central Sweden. This may explain why the Skellefteå conductor (SK) has much larger dimensions than the Southern Finland Conductor (TSB). The SK extends almost 300 km (200 km of the sub-horizontal near-surface part and about 80 km of dipping part) whereas the TSB has been squeezed to its present width of 40-80 km. Moreover, the TSB is almost vertical and geological evidence indicate a northward upthrust of sedimentary sequences over the volcanic Tampere arc (*Nironen*, 1989) whereas SK was underthrust northward beneath the Skellefteå volcanic arc.

6.3 Lower crustal conductivity

The resistive blocks between the conducting belts serve as transparent windows enabling the electrical structure of the lower crust and upper mantle to be determined. Lower crustal conductance estimated by *Korja* and *Hjelt* (1993) is given in Fig. 8. Lateral conductivity variations within the shield are rather large and indicate that the Archaean lower crust is more resistive than the Palaeoproterozoic Svecofennian crust. The lower crust beneath the transition zone between the Archaean and Svecofennian Domains seems to be also rather conductive even though it is considered to be composed of Archaean material (Archaean cratonic rocks beneath the Palaeoproterozoic Karelian supracrustal cover). This might therefore indicate that the lower part of the crust in the transition zone was also affected by the Svecofennian orogeny, perhaps by the additions of Palaeoproterozoic mafic material (*Korja, A. et al.*, 1993; *Lahtinen*, 1994).

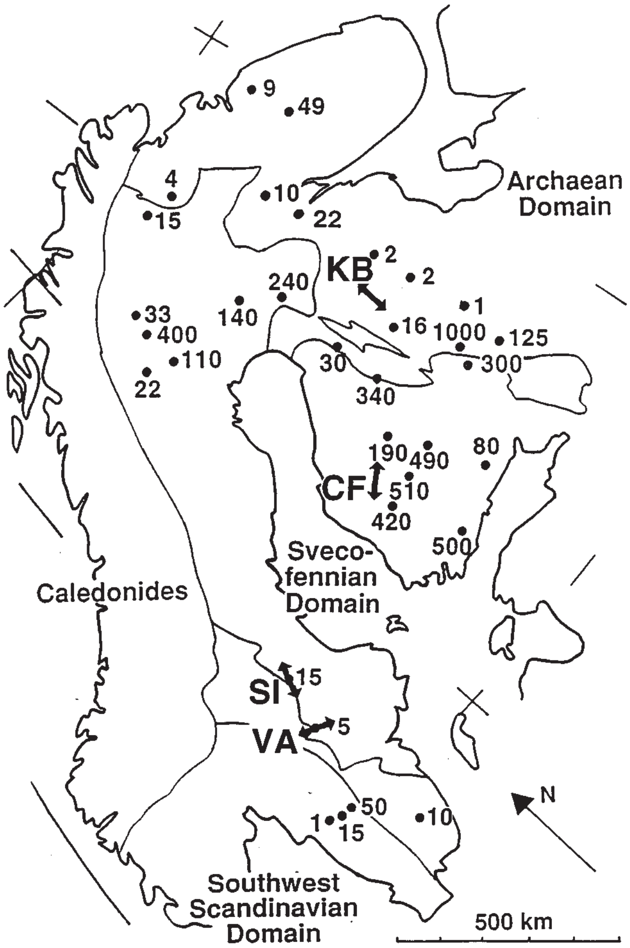

Fig. 8. Lateral variations in lower crustal conductivity. Solid arrows show the azimuth of the anisotropy (the most conductive direction) in those regions where magnetotelluric response functions representing deep crustal conductivity exhibit anisotropic features. VA=south-central Sweden; S=Siljan; CF=Central Finland Granitoid Complex and KB=Archaean Kuhmo block in eastern Finland. Figure is from *Korja* and *Hjelt* (1996).

In the approach used by *Korja* and *Hjelt* (1993) the depth and thickness of the lower crust is defined by seismic velocity data (lower crust: region where 7.0 km/s < Vp < 7.8 km/s; *Korja, A. et al.*, 1993) and the total conductance of this portion of the crust is calculated using resistivity values obtained from MT models. This approach means, naturally, that the estimates of the lower crustal conductance reflects both electrical and seismic properties of the crust, i.e. the thickness of the lower crustal high P-wave velocity layer and the average resistivity within this layer. Another objection against this approach is that there is no necessary reasons why conductivity structures should follow seismic velocity structures.

Recent results suggests that the lower crustal conductivity may be genuinely anisotropic in various areas in the Fennoscandian Shield e.g. in Värmland in southcentral Sweden (VA) (*Rasmussen*, 1988), in the Siljan area in Sweden (SI) (*Pedersen et al.*, 1989) and in central and eastern Finland (CF and KB) (*Korja et al.*, 1996b,c). The direction of the maximum conductivity (azimuth of the anisotropy) for each area have been shown in Fig. 8. The most conductive direction in the Värmland area is N75W degrees which is in conflict with the azimuth of N15E obtained in Siljan area although the two areas are only 150 km apart. The anisotropic region (layer) in Värmland is at the depths between 10 and 30 km whereas in Siljan at upper crustal depths (between ca. 10 and 20 km) and depths below 30 km. In VA and SI the anisotropic regions seem to be at least partly at different crustal (lithospheric) levels since crustal thicknesses are about 40 km and 44 km, respectively, with a rapid thickening to over 50 km towards east in the latter area (*Luosto*, 1991). Also, the Siljan region is an impact site of about 360 Ma old.

The previous analysis of the magnetotelluric data from the SVEKA profile (*Korja* and *Koivukoski*, 1994) led to inconsistent geoelectric strike directions from site to site within the CFGC and the data was modelled using isotropic models. The tensor decomposition using a new technique (*Groom* and *Bailey*, 1989), however, recovered very consistent 2D regional geoelectric strike direction of N60E degrees within the CFGC for long periods (> 5 s) (*Korja et al.*, 1996b). Rotation of the decomposed impedance tensor data into the direction of 60° from N to E reveals that at long periods the responses along and perpendicular at the strike differ from each other, by more than an order of magnitude for apparent resistivities and by about 40 degrees for phases at the longest periods available (1000 s). The results indicate that upper crust beneath the CFGC is isotropic and highly resistive, in agreement to previous results (*Korja* and *Koivukoski*, 1994), but the lower crust and upper mantle electrically anisotropic with the azimuth of 60 degrees from N to E. The onset of anisotropic behaviour at periods of 5 s corresponds to the depths of about 30 km. Hence both the lower crust (crustal thickness is about 55 km beneath the CFGC; *Luosto*, 1991) and upper mantle seem to be anisotropic.

A recent magnetotelluric study in the Kuhmo area in the Archaean Karelian Province (*Korja et al*., 1996c) indicate that also the Archaean deep lithosphere is anisotropic. On the contrary to the CFGC, the onset of the anisotropic behaviour of responses at 1 s corresponds to a depth of about 40 km according to previous 1D inversions (*Korja* and *Koivukoski*, 1994) and is compatible with the thickness of the Archaean crust (*Luosto*, 1991). Thus the entire Archaean crust seems to be isotropic whereas the upper mantle is anisotropic. Moreover, the most conductive direction (azimuth) is NS and record a 60 degree difference in comparison to the azimuth in CFGC.

No clear explanation for deep lithospheric anisotropy exists yet. The magnetotelluric data cannot resolve between micro- and macro-anisotropy: 1-D and anisotropic or small 2-D structures beyond the resolution of electromagnetic methods. The magnetovariational data both from CFGC and KB indicate a departure from 1-D structure as vertical magnetic field is not zero and induction vectors have a length of about 0.2. Therefore it seems that the anisotropy is not intrinsic caused by the electrical anisotropy of minerals but is caused by some macroscopic conductive structures. The conductive structures are, however, so small and so deep that electromagnetic methods cannot discriminate them and the Earth produces an anisotropic response. One possible cause might be near-vertical shear zones filled with conductive phase, e.g. carbon/graphite or water-bearing fluids. The difference in the azimuth between CFGC and KB suggests that the anisotropy is not related to present day tectonic stress regime but is “frozen-in” and inherited from some old processes.

The final comment concerning the electrical anisotropy of the crust and upper mantle is to warn that a special care should be taken to exclude the possibility that unknown conductors outside the research area cause the split of responses that erroneously might be interpreted as an indication of the electrical anisotropy. The entire southern Sweden has not been mapped by EM methods and therefore, for example, a conductor south or north of the Värmland and Siljan area or even a conductor between them could perhaps produce the observed split of responses in two orthogonal direction. The conductors east and/or west of the Värmland profile are not likely to cause the observed split because the profile is long and therefore one should observe an gradual change in responses (and split) eastwards or westwards along the profile. Similarly, if the split of responses in CFGC and KB were caused by conductors outside the research area, there should be a gradual lateral change in the responses (and split) along the profiles. The profiles are over 200 and 150 km long, respectively, and the responses (and their split) are stable over the entire profile. Preliminary 2-D model calculations (*Korja et al.*, 1996b,c) indicate that it is not possible to produce the observed split of responses with such 2-D models where conductors are located outside the profiles. The effect of 3-D conductivity structures surrounding, in particular, the CFGC is open. Preliminary thin sheet modellings (*Vanyan*, pers. comm; *Korja et al.*, 1996c) as well as 3-D modellings based on integral equation approach (Oksama, pers. comm.) indicate, however, that neither the surrounding (and known) conductors are capable to produce the split observed in CFGC.

Because the present day deep crust is inaccessible for direct observations, the exposed sections of deep continental crust (granulite belts) are of paramount importance for studies on the properties and processes of the deep crust. The Fennoscandian Shield hosts several such sections of which the Palaeoproterozoic Lapland Granulite Belt in the northern part of the shield is the largest.

Recent integrated study on the structure and properties of the Lapland Granulite Belt by *Korja et al*. (1996a), based locally on very dense geological and geophysical sampling, allowed them to propose a model for enhanced deep crustal conductivity and reflectivity. The main conclusion concerning the electrical conductivity was that the enhanced conductivity in the deep crust may be caused by an electron conducting mechanism acting in ductile shear zones, that contain locally-derived graphite, and to a lesser extent in those containing sulphides. The rocks in the deep crust may generally be highly resistive and the observed increase in conductivity may be a combined response from several thin, but highly conducting layers distributed throughout the deep crust. There is no need for continuous graphitic films along grain boundaries that requires a not

-so-likely pervasive flow of fluids through the deep crust. Neither is there any any need for H_2O-rich saline fluids in stable deep crust that contradicts petrological arguments on the dry deep crust. The model may also explain observed temporal variations (*Jones*, 1992) in deep crustal conductivity (Fig. 9): In active regions enhanced conductivity is caused by continuous graphite-bearing channels, shear zones and intraplated magmas (sills). In stable regions, where tectonic activity has ceased, the fluid channels tend to close and graphite will crystallize as larger grains. This would result in decreased conductivity, even though lithological layering and hence seismic reflectivity remains constant. In shear zones, however, the massive graphite will remain interconnected and will retain its conductivity even after exhumation as in the Lapland Granulite Belt.

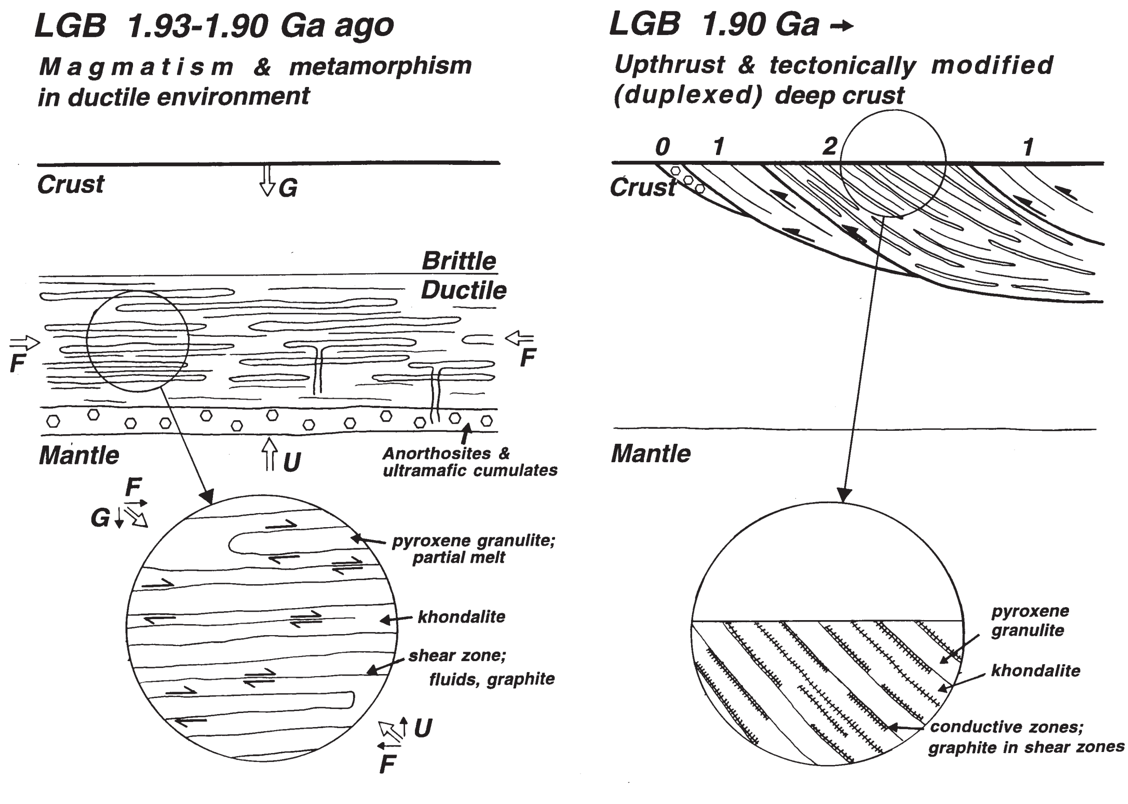

Fig. 9. Schematic models for the rocks of the Lapland Granulite Belt at two evolutionary stages and the existence of conductive structures. a) Granulite facies metamorphism and intrusion of enderbites (pyroxene granulites) took place in ductile environment in deep crust at ca. 1.93-1.90 Ga ago. Deep crust was conductive due to partial melt (magmas), fluids and graphite in shear zones. b) Deep crustal rocks were upthrust and duplexed ca. 1.90 Ga ago. Old shear zones can still be imaged by electromagnetic methods. Some of the shear zones, especially at the basal part of the belt, were formed during the upthrusting. Mantle derived fluids were flowing along the decollement as indicated by anomalous carbon isotope values. Models are based on *Korja et al.* (1996a).

6.4 *The electrical asthenosphere*

Figure 2 includes rough estimates of the depth to the upper mantle (asthenospheric) conducting layer in the Fennoscandian Shield (*Hjelt* and *Korja*, 1993). Results from the northern part of the shield indicate a presence of a conducting layer at depths ranging from 100 km to 150 km. Results from more southerly parts indicate the presence of a conducting asthenosphere in peripheral regions. It is evident, however, that in the central and southwestern parts of the shield the asthenospheric layer is absent or is electrically weak i.e. contain a small amount of conductive phase. Although crustal conductors tend to screen deeper information, in the more favourable locations, such as the Archaean in Finland, the Central Finland Granitoid Complex, and the Transcandinavian Igneous Belt (TIB) in Sweden, the total crustal and upper mantle conductance to the asthenospheric depths is only a few tens of Siemens and hence the conducting asthenospheric layer, which has total conductance of several hundred Siemens, should be detectable; 100 S corresponds for example, to layers of 50 km / 500 Ωm or 20 km / 200 Ωm.

The major problem in deep electromagnetic soundings in Fennoscandia and in its northern part, in particular, is the proximity of the source of the primary electromagnetic field. The closeness of the source may produce evidence for a conducting layer in sounding responses even though no such layer exists in reality. Recent analysis of the long period magnetotelluric data from the central parts of the Shield (*Viljakainen*, 1996) indicate that the magnetotelluric method is suitable for as long periods as 8000 s if a careful selection of plane wave events (homogeneous primary electromagnetic field) is made. These periods correspond to the depths of approximately 200 km beneath the Kuhmo region. Thus for central and southern parts of the shield the results may be valid whereas for more northerly parts a careful data evaluation (validity of the plane wave assumption) should be carried out in order to provide reliable estimates on the upper mantle conductivity. Alternatively one could map the spatial behaviour of the electromagnetic field and use the results to obtain an unbiased estimate on upper mantle conductivity. That would require an employment of a large 2-D array of simultaneously recording magnetotelluric instruments. Such an experiment is planned to be a part of the EUROPROBE/SVEKALAPKO studies.

7 Concluding remarks

The deep research in the Fennoscandian Shield as well as the airborne surveys in Finland have provided a globally unique electromagnetic data set to estimate the crustal and upper mantle conductivity. The data allow to focus from large, shield-scale structures having dimensions of the order of 100 km and more to small-scale local structures with a size of a few metres or less, and enables to interpret the conductivity structures in tectono-geological terms ranging from tectonic processes affecting the entire shield to small processes responsible for enhanced electrical conductivity within shear zones.

The electrical structure of the Shield is characterized by elongated belts of conductors that are limited to the upper half of the crust except for some occasional penetrations into the lower crust. The internal structure of conductors may be very complicated as is evident from the airborne electromagnetic data. Due to the large conductances involved and the correlation with the surface geology the conductivity mechanism invoked to explain this is electronic as found in graphite- and sulphide-bearing metasedimentary rocks. The conductors surround more resistive blocks and mark boundaries between distinct crustal units. The sedimentary rocks now responsible for enhanced conductivity seem to have been deposited during a relatively short time interval roughly between 2.1 and 1.9 Ga ago. The actual tectonic position of the conductors is not fully clear yet but as the carbon is organic the sedimentation took place in shallow or deep sea basins. The subsequent tectonic processes closed the basins, subjected the sedimentary material to temperatures exceeding 400 degrees, and emplaced them to current positions. If, for example, the deep parts of the Skellefteå conductor are caused by metasedimentary rocks, it implies that the supracrustal rocks were transported to the very deepest roots of the orogen.

The resistive regions between the conductive belts, on the other hand, serve as transparent windows to probe deeper structures and properties of the shield. The deep Palaeoproterozoic crust seems to be more conductive, in particular in the central part of the shield, than the Archaean lower crust. A notable exception is the transition zone between the Archaean Karelian Province and the Svecofennian Domain. There are also an increasing amount of evidence that the lower crust and even the upper mantle may be electrically anisotropic. Dedicated 2D and 3D modellings, however, are required to fully address this question. Studies in the Lapland Granulite Belt, that represents an exposed section of deep crust, have shown that the lower crustal conductivity may be explained by graphite-bearing shear zones in agreement with the petrological evidence of dry deep crust.

Currently available information on the upper mantle conductivity is sparse. An additional methodological difficulty is imposed by the proximity of the iono- and magnetospheric current systems. The results nevertheless indicate that in the central and southwestern parts of the shield the asthenospheric layer is absent or is electrically weak whereas in peripheral regions the upper mantle conductor associated with the asthenosphere might be closer to the surface and more conducting.

More and reliable data are clearly required on the upper mantle conductivity in order to constrain the models proposed for the structure and evolution of the deep lithosphere. An important question concerning the deep lithosphere, inaccessible for direct observations, is whether the observed (anomalous) electrical conductivity distribution is a result of currently acting processes, a result of several causally and temporally distinct processes or a combination of all. The above applies for most of the geophysical studies that observe the present day physical state of the lithosphere.

For upper and middle crustal conductivity structures a detailed evaluation on their tectonic position is required in order to understand the development of the crustal scale conductivity structures and to use them in constraining the evolutionary models of the continental crust and the Fennoscandian Shield, in particular.

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