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A workshop dealing with the Finnish contribution to the International Global Geoscience Transect (GGT) Program was held at the Geological Survey of Finland Regional Office in Kuopio from 25-26/11/1993. The primary goal of the Program is to compile geological and geophysical data along representative profiles across orogenic belts, in order to facilitate tectonic interpretations and enhance understanding of lithospheric structure and evolution. The articles in this volume deal almost exclusively with the Finnish GGT transect, which includes a number if major tectonic units between Kustavi, on the southwest coast, and Kuhmo in the northeast of the country, and covers an area 160 km in width and 770 km in length, parallel to the Svekokarelian (SVEKA) Deep Seismic Sounding (DSS) profile.

According to K-Ar studies the Archaean bedrock was strongly affected by Palaeoproterozoic orogenic processes. The high temperatures, large volumes of fluids and shear type of deformation associated with the Svecofennian reactivation of the Archaean basement of eastern Finland imply the possibility that gold might also have been remobilized during this process. New studies concerning the mantle section of the Jormua Ophiolite indicate continental break-up slightly before the formation of the new oceanic basin in the west.

Five principal epochs of mineralization and metallogenic provinces can be recognized in the development of the collisional Raahe-Ladoga Zone between the Archaean continent and Palaeoproterozoic lithosphere. Economically most important are the early orogenic evolution of the Kainuu-Outokumpu rift and the Pyhäsalmi island arc as well as the evolution of synorogenic Kotalahti Ni-belt.

The geological background for the GGT/ SVEKA transect is also described by the geology of the Tampere Schist Belt (TSB) and Vammala Migmatite Belt (VMB). The TSB was formed in an evolved arc setting and is separated by a fault from the VMB which likely represents tectonised forearc to trench or marginal basin deposits. It is also possible that the VMB contains Palaeoproterozoic crust older than ca 1.905 Ga i.e. equivalents of the sources of the 1.91 - 2.0 Ga zircons so common in the TSB graywackes.

On of most distinctive features of this transect is the presence of unusually thick Proterozoic crust, which increases in thickness from 50 km to around 60 km in the central part of the profile, near the Raahe-Ladoga Zone. The northeastern part of the profile traverses Archaean terrain which has a more characteristic crustal thickness of around 40 km. The three dimensional crustal structure of the Bothnian Sea has been deduced from the nearvertical reflection data of the Baltic and Bothnian Echoes from the Lithoshere (BABEL) experiments.The studies of temperature and heat flow density in the bedrock on the GGT/ SVEKA transect indicate that the observed surface heat flow density variations can mostly be explained by upper crustal heat sources. The middle and lower crustal, heat production seems to be lower in the Archaean than in the Proterozoic domain.

Key words (GeoRef Thesaurus, AGI): crust, bedrock, deep-seated structures, tectonics, geophysical surveys, deep seismic sounding, geotraverses, SVEKA, Precambrian, symposia, Finland

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PREFACE

The articles of this report represent the presentations as given at GGT/ SVEKA workshop in Kuopio 25 - 26.11.1993. The papers review geological and geophysical investigations concerning the structure of the deep crust and the possible surface expression of deep features. One of the papers 'Deformation, metamorphism and the deep structure of the crust in the Turku area, southwestern Finland' by Markku Väisänen, Pentti Hölttä, Jyri Rastas, Annakaisa Korja and Pekka Heikkinen has been published previosly in Geological Survey of Finland, Guide 37. The results of continuing research activity will be presented in future workshops, and it is anticipated that the final report will be published during 1996.

Elias Ekdahl

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SVECOKARELIAN (PALAEOPROTEROZOIC) TECTONO-THERMAL EFFECT ON THE ARCHAEAN BEDROCK OF NORTH KARELIA, EASTERN FINLAND

by

Asko Kontinen and Jorma Paavola

Kontinen, A. & Paavola, J. 1996. Svecokarelian (Palaeoproterozoic) tectono-thermal effect on the Archaean bedrock of North Karelia, eastern Finland. In: Ekdahl, E. and Autio, S. (eds.) Global Geoscience Transect/ SVEKA - Proceedings of the Kuopio Seminar, Finland 25. - 26.11.1993. Geologian tutkimuskeskus, Tutkimusraportti – *Geological Survey of Finland, Report of Investigation* 136, 7 - 8.

Key words (GeoRef Thesaurus, AGI): bedrock, basement, thermal history, recrystallization, Svecokarelian Orogeny, absolute age, Proterozoic, Archean, Ilomantsi, Lieksa, Finland

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Introduction

A regional K-Ar dating project of the Archaean basement in eastern Finland was carried out by Geological Survey of Finland (GSF, Kuopio) from 1984 - 1992 (Kontinen et al. 1992). Widespread resetting of biotite and hornblende ages as a response to the Svecokarelian orogeny was documented as a result of this extensive project, totalling 61 hornblende and 68 biotite K-Ar dates. Accordingly, large parts of the Archaean basement were heated 1900 - 1850 Ma ago to temperatures in excess of 500 °C. Based on petrographic aspects of the dated minerals in thin sections, it was concluded that the resetting of K-Ar age mineral system was not controlled primarily by volume diffusion but rather by metamorphic mineral recrystallization. Archaean mineral dates were observed mainly from high grade metamorphic blocks and are attributed to the rigidity and dry nature of the high grade rocks making them relatively more resistant to the resetting processes (recrystallization).

To provide a more precise understanding of the regional heating pattern, thermal history and recrystallization processes in the basement, we have obtained a further 20 hornblende and biotite K-Ar dates, nine hornblende Ar-Ar dates and several zircon and titanite U-Pb age determinations, in combination with field and petrographic work, particularly in the North Karelian region. This study has been done in co-operation with isotope geologists from GSF (Espoo) and the University of Leeds.

The Ilomantsi area

Our earlier conclusion concerning the relatively modest Svecokarelian heating in the area S-SE from Heinävaara-Ilomantsi-Hattuvaara line (Ilomantsi domain) has been confirmed by the new data. The generally older hornblendes (K-Ar ages up to 2689 Ma) in the Ilomantsi domain mean that Svecokarelian temperatures generally did not exceed 500 °C (which is the commonly accepted hornblende blocking or resetting temperature; e.g. Harrison 1981), in contrast to the area to the north of the domain, where hornblende is generally reset to show Svecokarelian ages. U-Pb titanite ages from the Ilomantsi region are all Archaean but, vary sympathetically with the hornblende K-Ar ages so that slightly lower titanite ages come from

the samples with the lowest hornblende K-Ar ages.

The Koitere area

The area to the north of the Heinävaara-Ilomantsi-Hattuvaara line, from Koitere up to Vieki and Nurmijärvi (Koitere domain), has proved to be a very interesting study area with respect to the nature of Svecokarelian reactivation and related deformation and metamorphism (Sorjonen-Ward 1993). This is because the primary upper amphibolite facies Archaean metamorphism of the rocks in the Koitere domain permits typical Presvecokarelian mineral parageneses (typically clinopyroxene + green hornblende + brown biotite + fresh plagioclase + undulose quartz + magnetite) and rock textures (igneous, gneissic) to be readily distinguished from the greenschist to epidote amphibolite facies Svecokarelidic parageneses (typically bluegreen hornblende + green biotite + heavily saussuritized plagioglase + annealed, non-undulous quartz ± carbonate ± iron sulphides) and textures (protomylonite).

The Svecokarelian reactivation is best ascribed as to regional recrystallization, with localized zones of protomylonitization. Most of the Koitere domain is strongly affected by the protomylonitization and related metamorphic rehydration processes. Rocks retaining Archaean upper amphibolite facies minerals and textures have survived only in scattered fault-bounded blocks that have to varying degrees escaped the Proterozoic deformation and metamorphic retrogression; these blocks are seen on areal aeromagnetic maps as domains of high magnetic response. The Proterozoic age of the protomylonitization is indicated by the Proterozoic hornblende and biotite ages obtained from the protomylonitic rocks and the deformation and metamorphism of Proterozoic diabase dykes within the protomylonitic domains and relative lack of evidence for deformation in dykes of the same swarms within the relict upper amphibolite facies blocks. In conclusion, the new information from the Koitere domain strongly supports our earlier view of the major impact of the Svecokarelian orogeny on the mineralogy and texture and structure of the basement rocks.

Implication for the gold problem

The high temperatures, large volumes of fluids and shear type of deformation associated with the Svecokarelian reactivation of the Archaean basement of eastern Finland imply the possibility that gold mineralizations in the basement may be, if not Svekokarelian in origin, at least significantly remobilized during the Svecokarelian reactivation. This possibility gets support from the fact that many of the metadiabases in the basement area are remarkably rich in tourmaline. The Svecokarelian metamorphic fluids transported not only water and carbon dioxide but also boron (O'Brien et al. 1993; Bornhorst et al. 1993) and hence possibly also gold, which tends to accompany boron in hydrothermal fluids.

Influence on rock magnetic properties

Because of the alteration of magnetite to nonmagnetic phases in association with the Svecokarelian reactivation and related metamorphic processes, the magnetic properties of the Archaean rocks have also been extensively modified. As a consequence, in many areas the intensity patterns on the regional aeromagnetic maps reflect rather the effects of the Svecokarelian reactivation than pristine Archaean bedrock structures.

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STRATIGRAPHY AND EVOLUTION OF THE PALAEOPROTEROZOIC SUPRACRUSTAL ROCKS IN THE TAMPERE AREA, SOUTHERN FINLAND: A PRELIMINARY REVIEW

by

Yrjö Kähkönen

Kähkönen, Y. 1996. Stratigraphy and evolution of the Palaeoproterozoic supracrustal rocks in the Tampere area, southern Finland: a preliminary review. In: Ekdahl, E. and Autio, S. (eds.) Global Geoscience Transect/ SVEKA - Proceedings of the Kuopio Seminar, Finland 25. - 26.11.1993. Geologian tutkimuskeskus, Tutkimusraportti – *Geological Survey of Finland, Report of Investigation 136*, 9 - 12.

Key words (GeoRef Thesaurus, AGI): metasedimentary rocks, metavolcanic rocks, stratigraphy, evolution, Proterozoic, Tampere, Finland

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The Tampere Schist Belt (TSB) is a discontinuous Palaeoproterozoic volcanic-sedimentary belt which is about 200 km long in E-W direction and about 20 km across at its widest. It is separated by a fault from the Vammala Migmatite Belt (VMB) to the south, while to the north it is bounded by the Central Finland Granitoid Complex. The TSB is composed mainly of metasediments, and arc-type basaltic, intermediate and rhyolitic metavolcanics, with intermediate compositions predominating (Ojakangas 1986, Kähkönen 1989). The TSB metavolcanics have U-Pb zircon ages of ca. 1904 - 1890 Ma (Kähkönen et al. 1989) while zircons in the graywackes are mostly 1.91 - 2.0 Ga in age, with an additional Archaean component (Huhma et al. 1991, Claesson et al. 1993). In its central parts near Tampere, the TSB comprises a major E-W striking F1-syncline with subvertical axial planes and subhorizontal F1-axes (Nironen 1989a). Stratigraphic columns vary from place to place but pyroclastic rocks dominate among the metavolcanics and the metasedimentary rocks are characterized by turbidites. This preliminary paper discusses selected parts of the supracrustal successions in the Tampere area.

Table 1. A tentative generalized stratigraphic scheme of the central TSB. Zircon ages from Kähkönen et al. (1989), Nironen (1989b), Huhma et al. (1991) and Claesson et al. (1993).

	zircon age
Takamaa formation mafic volcanics	1.889 Ga
Arc-type volcanics and related sedimentary rocks including the Veittijärvi-type conglomerates	1.905 - 1.89 Ga pebbles 1.89 - 1.884 Ga
Myllyniemi turbidites	clastic zircons ca. 1.91 - 2.0 Ga and Archean
Haveri formation extensional basalts	

In the central TSB near Tampere, around Lake Näsijärvi (1 : 100 000 map sheets 2123, 2124 and 2142), the succession exposed in the southern limb of the major syncline is up to 7 - 8 km thick (Kähkönen & Leveinen 1994). Its lower parts are dominated by turbiditic metasedimentary units of which the Myllyniemi formation is the lowermost (Table 1). Metavolcanic rocks are found in the southernmost part of the belt at Tohloppi but, contrary to the original suggestion of the author, they are younger than the Myllyniemi metaturbidites (H. Huhma, personal communication, see also Kähkönen 1994). Consequently, the existence of

early thrusts is considered likely. A part of the graywackes and conglomerates of the Myllyniemi formation are rich in quartz but the bulk of the metaturbidites were derived from volcanic sources. The upper part of the succession is dominated by metavolcanic rocks. Near Lake Pulesjärvi, 10 - 15 km east of Lake Näsijärvi, the related metasedimentary rocks comprise fluvial strata deposited on subaerial volcanic apron(s) (Leveinen, 1990). A stratovolcano or a complex of stratovolcanoes probably emerged above sea level in this area. In contrast, the Veittijärvi conglomerate at Ylöjärvi, 5 - 10 km east of Lake Näsijärvi, represents deposition in submarine channels (Rautio 1987). In the core of the major syncline the conglomerate is overlain by the 1.889 Ga (Kähkönen et al. 1989) basalt-rich Takamaa formation (called the Upper Volcanic Unit at Ylöjärvi by Kähkönen 1989). The northern limb at Ylöjärvi is dominated by metavolcanics.

The succession at Viljakkala, 30 - 35 km NW of Tampere, includes the Haveri formation (Mäkelä 1980, Kähkönen & Nironen 1994) which contains pillow-basalts and is considered to be the lowermost unit in the northern limb of the major syncline. The Haveri basalts are exceptional in the TSB since they show extensional, not arc-type, affinities. This formation is overlain by the turbiditic Osara formation, and the latter seems to be overlain by the Harhala formation, composed of arc-type andesites, dacites and rhyolites.

At Orivesi, 30 - 40 km NE of Tampere, the northern limb of the major syncline is dominated by the andesitic-dacitic-rhyolitic Koskuenjärvi formation (called the Intermediate Unit at Orivesi by Kähkönen 1989). Metaturbidites are absent in this 1 km thick unit and the metavolcanic rocks were evidently deposited in subaerial environments. The Koskuenjärvi formation has a zircon age of 1904 Ma (Kähkönen et al. 1989). The southern limb of the major syncline at Orivesi is dominated by metaturbidites and subaqueous pyroclastic rocks.

The differences between the successions of the southern and northern limb are attributed to two factors. Rapid facies changes are common in environments characterized by stratovolcanoes and small sedimentary basins. In addition, the E-W striking Paarlahti fault or shear zone tectonically excised part of the succession from the northern limb.

The Mauri "arkose", 25 - 40 km W of Tampere, comprises a unit of fluvial-deltaic sandstones up to 2 km thick (Matisto 1968). Clasts of potassium feldspar are common but since the bulk of the clasts are actually fine-grained felsic volcanics the rock should rather be classified as a litharenite (with predominantly volcanic provenance) than as an arkose. The Mauri arenites occur on the southern limb of an anticline but the precise stratigraphic position is not quite clear. The ca. 1.9 Ga clastic zircon (Matisto 1968) is commonly thought to suggest that the Mauri arenites are relatively high in the succession of the TSB.

In the Suodenniemi area, about 50 km west of Tampere, there seems to be a large syncline with a subvertical NW-SE striking axial plane but in the southwest the structure is more complicated (Perttula 1982, field observations of the author). The Suodenniemi succession is rather poorly exposed but it is interpreted as being ca. 5 km thick in the northeast. In the southwest, structural complexities make the stratigraphic interpretation problematic. The Suodenniemi succession is dominated by turbiditic metasedimentary rocks. Two major units with abundant mafic metavolcanics have been identified in the NE limb. The Hoivasvuori metavolcanics (ca. 0.4 km thick) are intercalated among the northeasternmost metaturbidites and mainly consist of coarse mafic pyroclastic rocks although uralite-porphyritic lava locally displaying pillow structures is also present (observations by the author). The pyroclastic mafic to intermediate metavolcanics at Punapäät (ca. 0.5 km, or more, thick) occur high in the sequence, on both sides of the hinge zone. They grade upward into horizons dominated by metaturbidites with volcanic provenance. A massive plagioclase porphyry up to 0.5 km thick at Suodenniemi is considered by the author to be subvolcanic rather than crystal tuff in origin (cf. Perttula 1982). The metasedimentary rocks at Suodenniemi appear to be predominantly mudstones and fine-grained graywackes but cobble-pebble conglomerates, rare quartz-rich arenites and conglomerates, coarse-grained graywackes and arkoses are present as well. Some of the turbidite-dominated units are rich in intermediate to felsic tuffs and redeposited tuffs. The precise stratigraphic position of the Kari arkoses, 1 - 2 km SW of the hinge zone, is not known, due to structural complexities but they are probably high in the succession.

The supracrustal rocks of the Vammala Migmatite Belt (VMB) are dominated by metasediments (e.g., Matisto 1971). These consist mostly of turbiditic graywackes and mudstones, although black schists and graphite schists are more common than in the TSB. In addition, the VMB contains sedimentary carbonates and arenites. The Rämsöö conglomerate at Vesilahti, about 30 km SW of Tampere, represents feeder-channel deposits of a submarine fan but clast compositions differ from typical TSB conglomerates. In contrast, the Naistenmatka conglomerate at Pirkkala (about 5 km SW of Tampere),

resembles TSB conglomerates since it is characterized by volcanic clasts. However, it differs from the Veittijärvi and Tohloppi conglomerates due to the absence of pink feldspar porphyry clasts. Most of the turbidite gneisses in the northern VMB resemble the turbiditic TSB metasediments in chemical composition (Lahtinen 1994).

Metavolcanic rocks are not common in the VMB but mafic metavolcanics with MORB-like affinities do occur in the southern part of the belt (Lahtinen 1994). The basaltic to andesitic amphibolites at Kiimajärvi, about 50 km WSW of Tampere, have mixed tectonomagmatic affinities according to the Ti vs. Zr diagram (unpublished data of the author) but similar characteristics are found in certain metavolcanics of the TSB as well.

In general, the TSB was formed in an evolved arc setting (Kähkönen 1989). The Haveri metabasalts are older than the bulk of the 1.905 - 1.89 Ga arc-type TSB metavolcanics and represent extension of pre-1.91 Ga Palaeoproterozoic crust. The 1.905 - 1.89 Ga arc-type volcanism was also preceded by deposition of the bulk of the TSB metaturbidites (e.g., of the Myllyniemi formation) since the youngest clastic zircons in the latter are slightly older than 1.905 Ga. Some of the metaturbidites, however, probably have a significant component derived from the 1.905 - 1.89 Ga TSB volcanics.

The ca. 1.904 Ga Koskuenjärvi formation at Orivesi is the oldest unit in the TSB dated so far by U-Pb zircon methods. Based on Yli-Kyyny (1990) and his personal communication, the 2 - 3 km thick succession of metavolcanics at Kiialanniemi, Ikaalinen (50 - 60 km NW of Tampere), resembles the Koskuenjärvi formation since both units are dominated by intermediate to felsic subaerial rocks and both seem to occur relatively low in their respective local stratigraphic successions. In contrast, the metavolcanics interbedded with metaturbidites on the southern limb of the major syncline near Tampere are subaqueous pyroclastic rocks. The bulk of the volcanic succession at Suodenniemi also represents submarine sequences.

The supracrustal rocks of the VMB differ from those of the TSB in many respects and may largely represent tectonised forearc to trench or marginal basin deposits incorporated within an accretionary prism complex or complexes. Such complexes would consist mainly of submarine fan and abyssal turbidites but some of the black schists might represent pelagic sediments. Local emergence of the prisms would result in occasional shallow-water or subaerial depositional environments. The variation in the tectonomagmatic affinities of the VMB metavolcanics is thought to reflect the occurrence of both extensional and arc-type settings in and near the accreted basin(s). It is also possible that the VMB contains Palaeoproterozoic crust older than ca. 1.905 Ga, i.e., equivalents of the sources of the 1.91 - 2.0 Ga zircons so common in the TSB graywackes.

The differences in the character of the supracrustal rocks of the TSB and VMB may be explained by differences in the character of crustal components of convergent systems. Although the two belts are separated by a major fault, they do not necessarily represent separate suspect terranes, i.e., they do not need to be mutually exotic.

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THE EVOLUTION AND METALLOGENESIS OF THE RAAHE-LADOGA ZONE

by

Elias Ekdahl

Ekdahl, E. 1996. The evolution and metallogenesis of the Raahe - Ladoga Zone. In: Ekdahl, E. & Autio, S. (eds.) Global Geoscience Transect/ SVEKA - Proceedings of the Kuopio Seminar, Finland 25. - 26.11.1993. Geologian tutkimuskeskus, Tutkimusraportti — *Geological Survey of Finland, Report of Investigation 136, 13 -17.*

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Introduction

The bedrock of central Finland is divided into the Archean Karelian and Proterozoic Svecofennian Domains. These two domains are separated by the Raahe-Ladoga Zone, at which thus defines the margin of the Archean craton. This major geological and tectonic unit contains highly metamorphosed volcanics and sediments and records several epochs of mineralization. The Svecfennian Domain in southwestern Finland lacks evidence for an underlying Archean basement. Since Hietanen (1975) first presented a plate tectonic interpretation for Finland, with the oceanic Svecofennian plate subducted beneath the continental Karelian block around 1.9 Ga, a number of authors have pursued this theme. The Raahe - Ladoga Zone has an integral place within such interpretations but in addition, records events that took place prior to the postulated collision.

Tectonic evolution of the Raahe - Ladoga Zone

The plate tectonic model presented by Park et al. (1984) involved progressive southwards migration of arcs including the Skellefte (2000 - 1950 Ma) and Tampere (1950 - 1900 Ma) volcanic arcs. Continental basic igneous activity (2.3 - 2.05 Ma) related to a phase of tensional activity preceded the break-up stage, which was represented by the genesis of the Outokumpu assemblage, attributed to the influence of a mantle diapir. Obduction of the Outokumpu back arc basin took place during the later orogenic phase. The model of Gaál (1986), involving a marginal basin and magmatic arc system on the active continental margin, related to eastwards subduction was subsequently modified (Gaál 1990), involving a reversal of subduction after collision. Thinning of continental crust led to marginal basin formation (Outokumpu rift zone). Westwards subduction of oceanic crust was associated with the formation of oceanic island arcs, intruded by tonalites at 1930 - 1900 Ma, along with the obduction of Outokumpu association back onto the foreland. Subsequent reversal of subduction resulted in the development of a continental margin magmatic arc between 1900 - 1870 Ma, with associated massive sulphide deposits.

An alternative interpretation involves prolonged, possibly multiple phases of subduction in a N-NE direction towards the continental margin, with back arc spreading. On a large scale the evolution of the Raahe-Ladoga Zone can be compared to a Wilson cycle, as is indicated by the chronogram based on 70 zircon ages determined by the GSF

(Fig. 1). The chronogram indicates that the total duration of the orogenic processes was at least 150 Ma. Orogenesis can be considered to have commenced at about 1.97 Ga, as mentioned by Koistinen (1981) in connection with the initiation of tectonism in the Outokumpu region. Bimodal arc volcanism represents the early orogenic stage, whereas the synorogenic stage is characterized by tholeiitic to calc-alkaline magmatism close to the active continental margin between 1.90 - 1.87 Ga. The I-type characterisitics of the granitoids can be accounted for by appealing to Andino type plate tectonics (Nurmi and Haapala 1986). The highly metamorphosed Savo schists on the suture zone represent a transitional zone between the Svecofennides and Karelides. Studies by the author in the Pielavesi district, support the notion that their development was, at least to some extent, coeval.

Tectonically the Raahe - Ladoga Zone (RLZ) is at least 150 km wide and can be considered to extend as far as the Kainuu Schist Belt in the northeast. The most significant mineralization epochs appear to have been associated with magmatism immediately prior to and during orogenesis.

Although the RLZ is primarily a synorogenic feature, it also contains various elements recording earlier events, including extension, rifting and break-up. The RLZ is characterized by dextral faults and shear zones that were generated by prolonged plate convergence along a S-SW vector (Fig. 2). The effects of this deformation are apparent for a considerable distance into the foreland. Dextral NW-trending wrench faults with some antithetic conjugate NE-trending faults and dextral N-trending zones dominate the structural pattern of the Archean craton over an area at least 150 km wide.

Conjugate fracture systems have also controlled volcanism in the island arc zone, with volcanic zones and complexes defining an en echélon chain (Fig. 3). Subduction resulted in D_1 recumbent folds and N-NE directed thrusting which were progressively steepened during D_2. This progressive deformational event was associated with amphibolite and granulite facies metamorphism in the marginal zone, and climaxed prior to the emplacement of the pyroxene granites about 1880 - 1890 Ma. Continued convergence resulted in the formation of dextral shear zones along the continental margin with thrusting of the tightly folded trench sequence towards the NW and NNW (Fig. 3 A). This deformation resulted in the formation of the prominent S_3 transverse foliation and SE-plunging L_3 mineral elongation lineation in the marginal zone. The orientation of ore deposits throughout the marginal zone was largely determined by this D_3 phase of deformation, which must have been a prolonged event, continuing until the intrusion of the late orogenic granites. D_3 deformation also influenced the geometry of metamorphic domains in the Kiuruvesi-Haukivesi complex (Korsman et al. 1988). The NW- NE- and N-trending fault zones that were

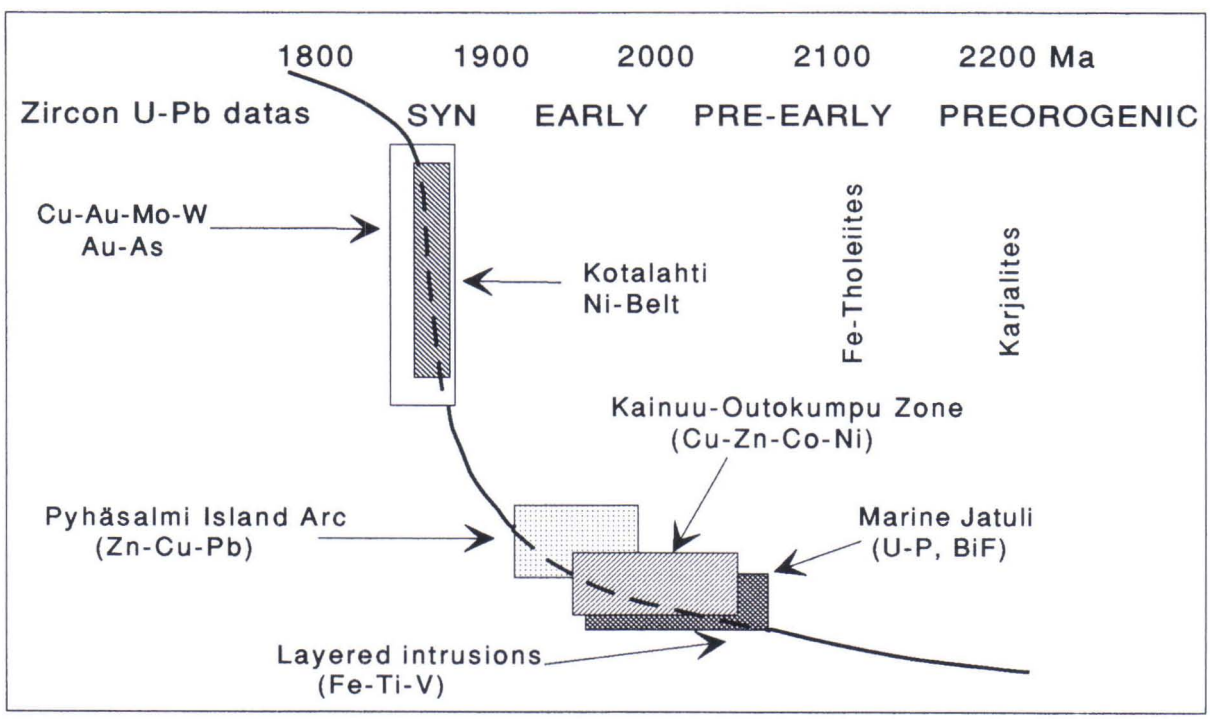

Fig 1. Descriptive chronogram of the Raahe-Ladoga Zone; compiled from about 70 U-Pb zircon data (y-axis).

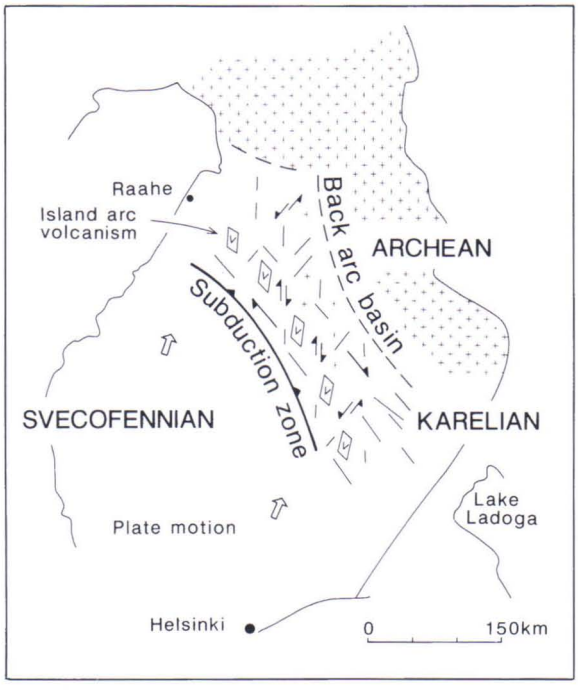

Fig 2. Plate tectonic interpretation of the evolution of the Raahe-Ladoga Zone.

initiated in the crust during subduction were frequently reactivated, resulting in vertical differential block movements during late stages of deformation. Late stage movement along these structures is demonstrated by brittle displacements.

Probably as a consequence of prolonged subduction, the Kainuu-Outokumpu back arc was emplaced onto the continental foreland some time after 1.97 Ga, with the vergence of the nappes coinciding with the direction of subduction (Ward 1987). The strongly developed SW-plunging lineation associated with this thrusting event in both Archean basement and Proterozoic cover has been described by numerous authors (Wegmann 1928; Väyrynen 1933, 1954; Koistinen 1981; Paavola 1984, 1988; Park 1988). The syngenetic ore deposits of the Outokumpu region were also polydeformed and are tectonically aligned within the NE-SW thrusting direction (Fig. 3 B). Thrusts are best developed to the N of the Suvasvesi Shear Zone, although deformed remnants of the thrusting and its associated lineation are found further southwest near Juva and Leppävirta. Similarly, N-NE directed thrusting in Kainuu is recorded by thrust-nappes within the Kainuu Schist Belt itself.

The D_3 deformation along the Haukivesi and Suvasvesi shear zones took place within the D_2-D_3 interval defined by Koistinen (1981) for the Outokumpu district. Deformation throughout the Raahe-Ladoga Zone has been a progressive, continuous process resulting from protracted compression. Volcanism, ore formation and later reorientation and remobilization of ore deposits has been controlled principally by N-NW-trending synthetic and N-NE-trending antithetic shear and fault zones. Brittle deformation during the final stages of orogeny was also controlled by reactivation of the same structures.

Metallogenesis

The geological history and metallogenic evolution of the Raahe - Ladoga Zone is illustrated on the chronogram in Figure 1.

Five principal epochs of mineralization and

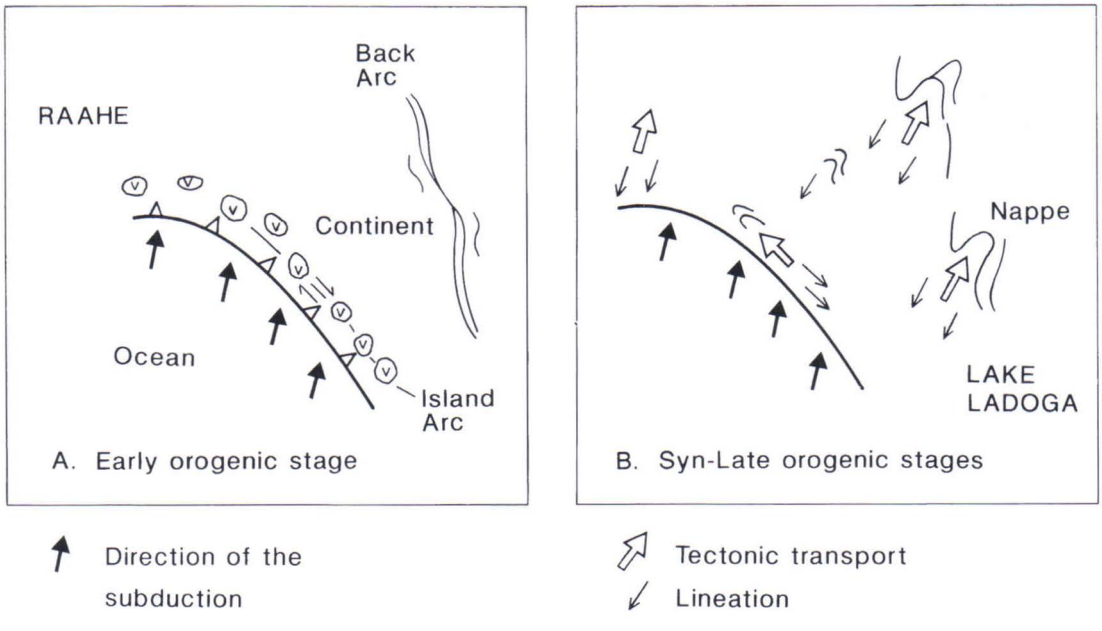

Fig 3. Schematic illustration of tectomic events in the Raahe-Ladoga Zone.

metallogenic provinces can be recognized in the development of the Raahe-Ladoga Zone.

1) Preorogenic stage
- Low-Al tholeiitic layered intrusions (2.2 Ga)
- Fe-tholeiites (2.1 Ga)

2) Pre- to early orogenic stage
- Stratiform U-P occurrences and iron formations within the Marine Jatulian association (ca. 2.08 - 1.97 Ga)
- Layered intrusions containing Ti-V-Fe deposits (ca. 2.06 Ga)

3) Early orogenic stage
- Kainuu-Outokumpu zone with Cu-Zn-Co-Ni deposits (≥1.96 Ga)
- Pyhäsalmi Island Arc with Kuroko-type deposits (≥1.92 Ga)

4) Synorogenic stage
- Kotalahti Ni-Belt (ca. 1.89 - 1.88 Ga)
- Porphyry type Cu±Mo±Au±W occurrences (≤1.88 Ga)

5) Syn-late orogenic stage
- Epigenetic Au-As occurrences (ca. 1.88 - 1.85 Ga)

Extensional magmatism is recorded troughout the whole of the Karelian Domain, represented by 2.2 Ga Al-poor tholeiitic sills (karjalites) and 2.1 Ga Fe-tholeiitic dykes (Vuollo et al. 1992). Some weak U and Cu mineralization has been found at the margins of these intrusions.

Prior to the early orogenic stage, the clastic and chemical marine Jatuli sediments with minor volcanics were deposited, including limited developments of iron formatious and U-P deposits.

The Kainuu Outokumpu ophiolite zone is almost 300 km in length and in addition to a number of exhausted deposits contains some tens of prospects. The massive Outokumpu-type deposits have been described as Cyprus-type (Koistinen 1981).

The Vihanti-Pyhäsalmi zone is interpreted as an island arc. The volcanic sequence consists of basaltic and andesitic lavas and in the uppermost parts, of pyroclastic rhyodacites and dacites. Kuroko-type Zn-Cu-Pb deposits are located within the hydrothermally altered upper parts of the pile.

Synorogenic ore deposits are represented by Ni-Cu mine-ralization associated with mafic and ultramafic intrusions of the Kotalahti belt. Porphyritic granitoids containing Cu, Au, Mo and W occurrences are characteristic of Cordilleran-type orogeny and are also closely associated with the Raahe - Ladoga Zone. The subduction zone setting was also conductive to the formation of epigenetic Au-As mesothermal deposits.

The Svecofennian Orogen is considered here to comprise three successive orogenic events, each of which contains the same broad subduction-related style of metallogenesis. The early orogenic stage is characterized by volcanic-hosted stratabound Zn-Cu-Pb deposits within the Pyhäsalmi Island Arc (1.92 Ga), the Skellefte-Tampere Arc (1.90 - 1.89 Ga) and the Bergslagen-Orijärvi Arc (1.90 - 1.86 Ga). The succeeding synorogenic stages are characterized by mafic to ultramafic magmatism with Ni-deposits and the late phases by calc-alkaline I-type granitoids with porphyry-type Cu±Au±Mo±W mineralization and epigenetic Au-As deposits. The Skellefte-Tampere Arc also contains post-orogenic (1.80 - 1.77 Ga) Li-Be-Sn-Nb-Ta pegmatites.

The evolution of lead isotope characteristics from the Outokumpu-type pure mantle composition to the more mature orogenic lead of the sulphide deposits which young to the southwest is an important feature in characterization of metallogenetic provinces (Vaasjoki 1981; Vaasjoki and Sakko 1988). Deep seismic reflection profiling (BABEL Working Group 1990) and magnetotelluric investigations (Korja 1990) also support the modelling presented above.

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EXTENSIONAL STRUCTURES OF SOUTHWESTERN FINLAND BASED ON THE BABEL REFLECTION RESULTS

by

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Korja, A. & Heikkinen, P. J. 1996. Extensional structures of southwestern Finland based on the BABEL reflection results. In: Ekdahl, E. and Autio, S. (eds.) Global Geoscience Transect/ SVEKA - Proceedings of the Kuopio Seminar, Finland 25. - 26.11.1993. Geologian tutkimuskeskus, Tutkimusraportti – *Geological Survey of Finland, Report of Investigation* 136, 19 - 22.

Key words (GeoRef Thesaurus, AGI): crust, extension tectonics, shear zones, faults, reflection methods, BABEL, Proterozoic, Southwestern Finland

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Introduction

Orogenic processes include both compression associated with collision and extension associated with destruction and denudation of the orogen. Details may differ but in all orogenies pre-, syn- and postorogenic periods and structures can be found. Research on Precambrian orogenies is focused on the pre- and synorogenic stages, which dominate surface lithologies and tectonic patterns on both local and regional scales. Only minor attention is paid to structures associated with the destruction of the orogen although those structures are most likely to be best preserved and thus most easily imaged as continuous seismic reflections. Surface lithology and deep crustal structures image complimentary features of the tectonic history. Observations from both sources are required when the whole of the orogenic evolution is studied. This has been well demonstrated in the Basin and Range Province of North America, where the destruction of thick collisional Laramide crust has been explained by post-collisional extension (Wernicke 1985).

The Svecofennian crust (2.0 - 1.75 Ga) consists of granitoid batholiths surrounded by polydeformed schist belts metamorphosed to medium or high grade (Fig. 1). It has been intruded by Subjotnian rapakivi batholiths and associated mafic intrusions (1.52 - 1.67 Ga). Sedimentary basins developed in the area in Riphean, Jotnian and Vendian times (1700 - 570 Ma) and the older basins were intruded by Postjotnian diabase dykes (1.2 Ga). The bedrock of the southern Fennoscandian Shield is highly fractured and consists of a mosaic of faulted blocks, also revealed by metamorphic studies (Korsman et al. 1988, Hölttä 1986) and lineament interpretations (e.g. Parkkinen and Huomo 1978, Kuivamäki and Tuominen, 1985). The fracture zones are observed on aeromagnetic and gravimetric maps as abrupt changes in magnetic and Bouguer gradient patterns.

The purpose of the paper is to describe and interpret the postcollisional crustal deformation of the Svecofennian seen on the seismic reflection sections. The crustal deformation is compared to surface geology and an extensional evolution model for the crust is proposed.

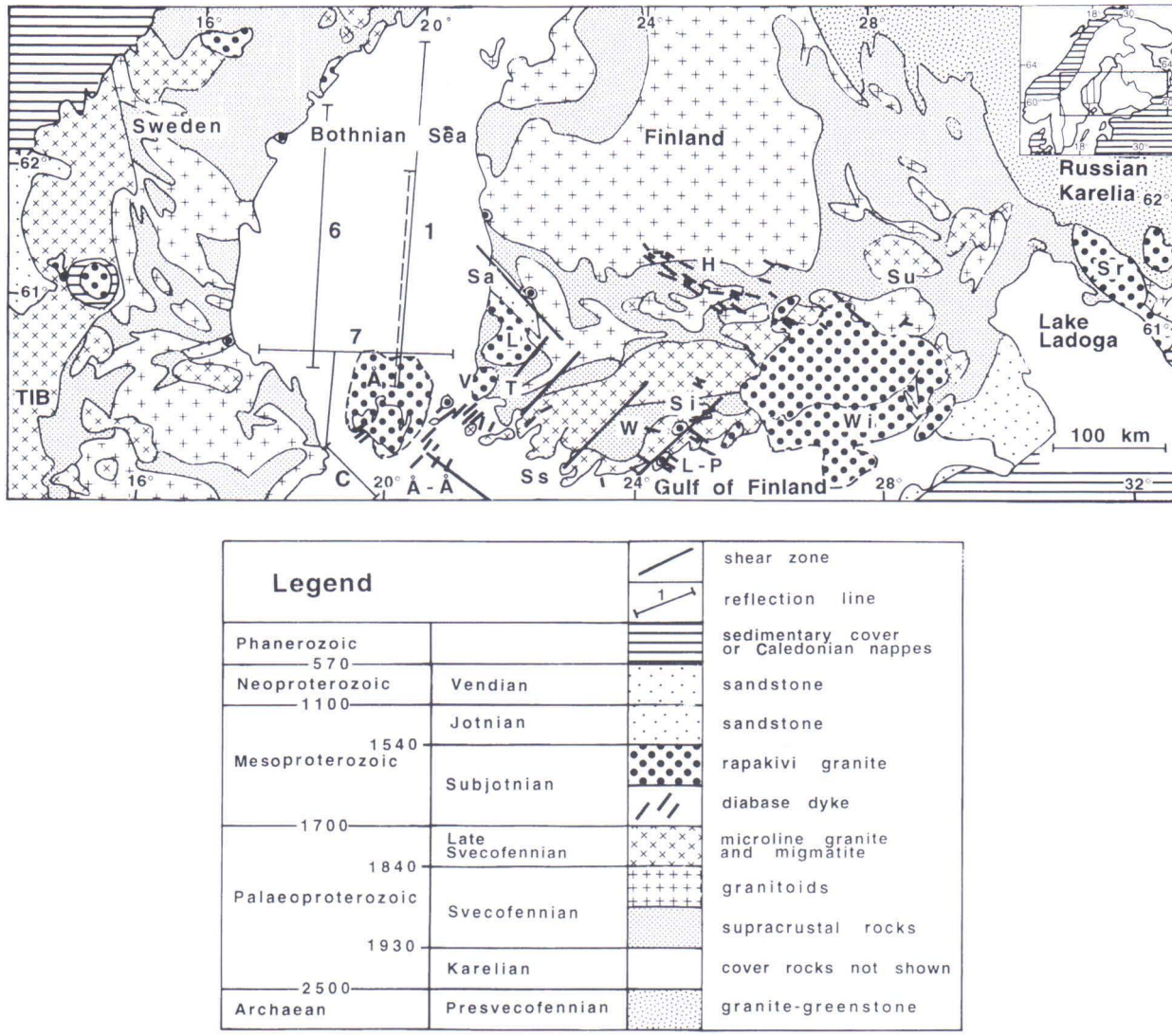

Fig. 1. Geological map of the Central-Fennoscandian Shield showing the BABEL reflection lines 1,6,7, and C. (Korja and Heikkinen 1995). L; Laitila, Å; Åland, V; Vehmaa, Wi; Wiborg rapakivi batholith, L-P; Lahti-Porkkala lineament, Ss; Suomusjärvi shear zone, Sa; Satakunta graben formation, H; Häme diabase dike swarm, Å-Å; Åland-Åboland dike swarm, T; Turku area.

Seismic data and structural interpretations

The three-dimensional crustal structure of the Bothnian Sea has been deduced from the near-vertical reflection data of the lines 1, 6, 7, and C of the BABEL (Baltic and Bothnian Echoes from the Lithosphere) experiment (BABEL Working Group 1993) (Fig. 1). For this study the CDP-stacked sections have been migrated with a velocity of 6.5 km/s and displayed as stacked envelope sections. The display enhances large scale crustal structures.

The reflection sections show large scale listric shear zones dipping towards the SE with apparent strikes between 030° and 040° (Fig. 2). The listric shear zones flatten at major detachment zones at 35 - 40 km and 48 km depth. The detachment zones are associated with changes in P-wave velocity from 6.8 km/s to 7.0 km/s and from 7.4 km/s to 8.2 km/s corresponding to the middle and lower crustal boundary and the Moho boundary. The detachment zone geometry and cross-cutting relationship are well developed on line 7. The listric normal faults related to the Moho detachment zone are younger and in the SE actually appear to cut the upper detachment zone. The southern end of line 6 is dominated by the older set of listric faults and the southern end of line 1 is dominated by the younger set of listric faults and consequently it is not easy to correlate these two lines.

The BABEL lines 1 and 7 cross-cut the Åland rapakivi granite batholith. The crust beneath the batholith is thinner and the lower crust has higher velocity and higher reflectivity than in the surrounding areas (BABEL Working Group 1993). The

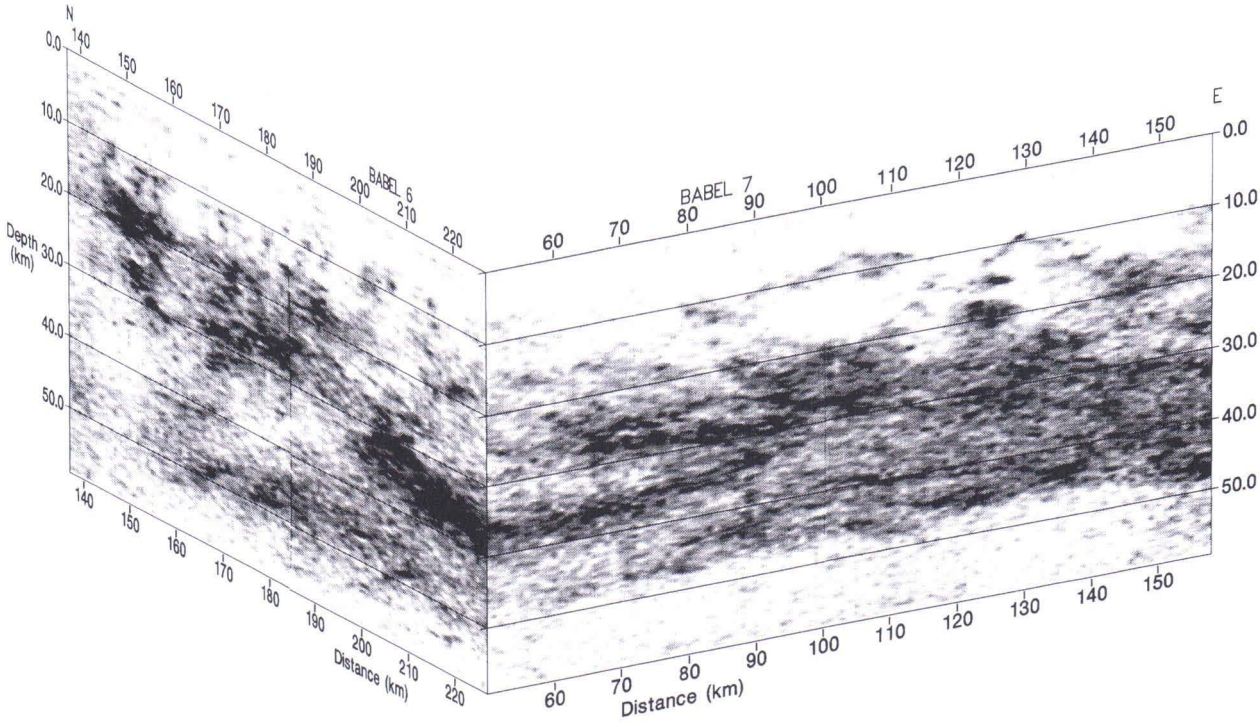

Fig. 2. Three-dimensional cross section of the BABEL lines 1 and 7 displayed as stacked envelope section (Korja and Heikkinen 1995). View is to the NW. Note the highly reflective listric shear zone dipping toward the viewer.

batholith itself is distinguished by a rather transparent area within which good reflections mimic graben and horst structures in the upper crust. The reflection image is interpreted as mafic Subjotnian intrusions and dykes occupying faulted and tilted block surfaces within the rapakivi granite. Although weaker, horst and graben structures are also observed the western part of line 7, where the Bothnian Sea Basin occupies the upper most part of the crust. Weak faults extend upwards to the floor of the Basin and consequently the Bothnian Sea Basin can be interpreted as a thinskinned sedimentary basin.

The Subjotnian to Jotnian Bothnian Sea Basin is offset to the NW of the major crustal thinning zone situated along the Gulf of Finland (Korja et al. 1993). The rapakivi granitoids related to mafic underplating of the crust, lie either within this major crustal thinning zone or to the SE of it. The geometry and spatial distribution of the listric faults, the thinskinned sedimentary basins, the major crustal thinning zone and anorogenic rapakivi batholiths can all be explained by a simple shear extensional delamination model of the crust (Wernicke 1985, Lister et al. 1986).

Discussion

The thinning of the upper crust via extension leads to a mosaic-like upper crust, where high-grade rocks outcrop next to low grade rocks (Fountain 1989). The simultaneous upward bowing of the lower crust and mantle in the central thinning zone causes the old metamorphic PT-trends of the crust to bow upwards. This results in mosaic crustal units within which the average metamorphic grade increases towards the Moho uplift (Wernicke 1985). In Finland, metamorphic grade generally increases towards the south, where S-type potassium granites and migmatites dominate. Metamorphic grade locally reaches granulite facies (Korsman et al. 1988, Hölttä 1986, Ploegsma and Westra 1991) that abruptly changes to amphibolite facies. The metamorphic block structure is thus in accordance with the extensional model.

In the Turku area, the peak metamorphic conditions (800° C, 5.5 kb) of low-pressure - high-temperature granulites were attained by 1840 Ma. By 1718 Ma the granulites had attained depths corresponding to 2 kb (Hölttä 1986). Approximately the same depth range is suggested for intrusion and crystallization of the rapakivi granites as well as for the contact metamorphic reactions in pyroxene-hornfels facies aureoles (Vorma 1972). This means that the high-grade rocks experienced about 10 km of uplift prior to 1718 Ma and consequently relatively little uplift/erosion is related to the rapakivi emplacement. Based on the surface observa-

tions of the uplift time frame we are inclined to relate the extensional deformation to postcollisional extension (Dewey 1988) of the Svecofennian crust and isochronous exhumation of the deeper levels of the crust.

This conclusion is supported by the late ductile shear deformation described from the nearby Åland archipelago by Brannigan (1987). According to him the postorogenic Seglinge granitoid batholith (1815 Ma) has been deformed by a late deformational shearing event (De+4). Later brittle phases of deformation control the emplacement of the Åland -Åboland dike swarm (1520 - 1600 Ma). The dominant direction of the sinistral shearing is 130° and the conjugate dextral shearing direction is 030°. The conjugate direction 030° agrees well with the shear zone directions of 035 - 040° observed in the seismic sections.

The apparent strike of the observed listric faults (035 - 040° and 040 - 045°) coincides rather well with the strike of the Subjotnian Åland-Åboland dike swarm, the magnetic lineament joining the rapakivi batholiths of Kökarsfjärden, Fjälskär, Peipohja, Vehmaa and Laitila (Suominen 1991), the Lahti-Porkkala lineament joining the Obnäs and Bodom rapakivi batholiths and the Suomusjärvi shear zone (Ploegsma, 1991), and also the lineaments (035/305°, 040°) related to the Wiborg rapakivi batholith (Parkkinen and Huomo 1978, Kuivamäki and Tuominen 1985). The conjugate direction agrees well with the axis of the Satakunta graben formation, the Häme diabase dike swarm, and Brewen and Hällefors dike swarms in Sweden. The conjugate direction is interpreted as a transfer fault system transferring shear strain between listric faults nucleating at different locations and levels of the extending crust (Lister et al. 1986).

The higher reflectivity and higher velocity of the lower crust beneath the Åland rapakivi granitoid together with the extensional listric shear zones point to mafic underplating of the crust in an extensional tectonic setting. This agrees well with the proposal of Haapala and Rämö (1991) that rapakivi granites are partial melts of the crust for which mafic underplating of the crust provided a heat source in an extensional tectonic setting.

Conclusions

Reflection sections image extensional listric shear zone structures formed in the post-Svecofennian time. They can be interpreted as imaging the post-collisional collapse of the Svecofennian orogen. The intrusion of the rapakivi granitoid batholiths was controlled by the listric faults in an extensional regime.

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TEMPERATURE AND HEAT FLOW DENSITY IN THE BEDROCK ON THE GGT/ SVEKA TRANSECT

by

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Introduction

The GGT/ SVEKA transect is characterized by heat flow density (HFD) and radiogenic surface heat production values that decrease from the southwest towards the northeast. On the Bothnian Bay coast HFD reaches 50 mWm^{-2} but at the NE end of the transect the corresponding value is only about 30 mWm^{-2}. These variations correlate with the age and geotectonic setting of the rocks on the profile (Kukkonen 1989a, 1992, 1993).

The present study discusses the present thermal regime of the lithosphere on GGT/ SVEKA with the aid of two-dimensional numerical forward models using finite difference methods. The simulation code SHEMAT (Clauser 1988) was used. Coupled heat and fluid transfer simulations were calculated for the crust.

The major question tested with the present simulations is: Can the surface HFD variation along the GGT/ SVEKA transect be attributed to upper crustal heat sources, or do we need a lateral variation in the Moho heat flow density or in middle and lower crustal heat generation values? Furthermore, does advective heat transfer in the upper crust produce significant deviations from the conductive heat flow values?

Compilation of model parameters

On a crustal scale, available geothermal measurements are limited to the very surface: heat flow measurements are made in drill holes usually shallower than 1 km, and rock thermal properties are measured from outcrop samples or from drill cores. These measurements cannot be extrapolated linearly into deeper levels of the lithosphere. Knowledge of the structure and composition of the deep lithosphere is therefore essential.

In order to calculate the thermal and hydraulic regime of the subsurface a wide spectrum of parameters is needed. In the present study forward thermal modelling is performed, ie. starting from given parameter values the subsurface temperature and heat HFD are calculated using certain boundary conditions. The necessary parameters are the vertical and horizontal variations in thermal conductivity, radiogenic heat production rate and hydraulic permeability. Further data are also needed concerning the horizontal variations in the hydraulic heads that provide the driving force for groundwater flow, and which is used as a constraint upon the hydraulic flow field employed in the model.

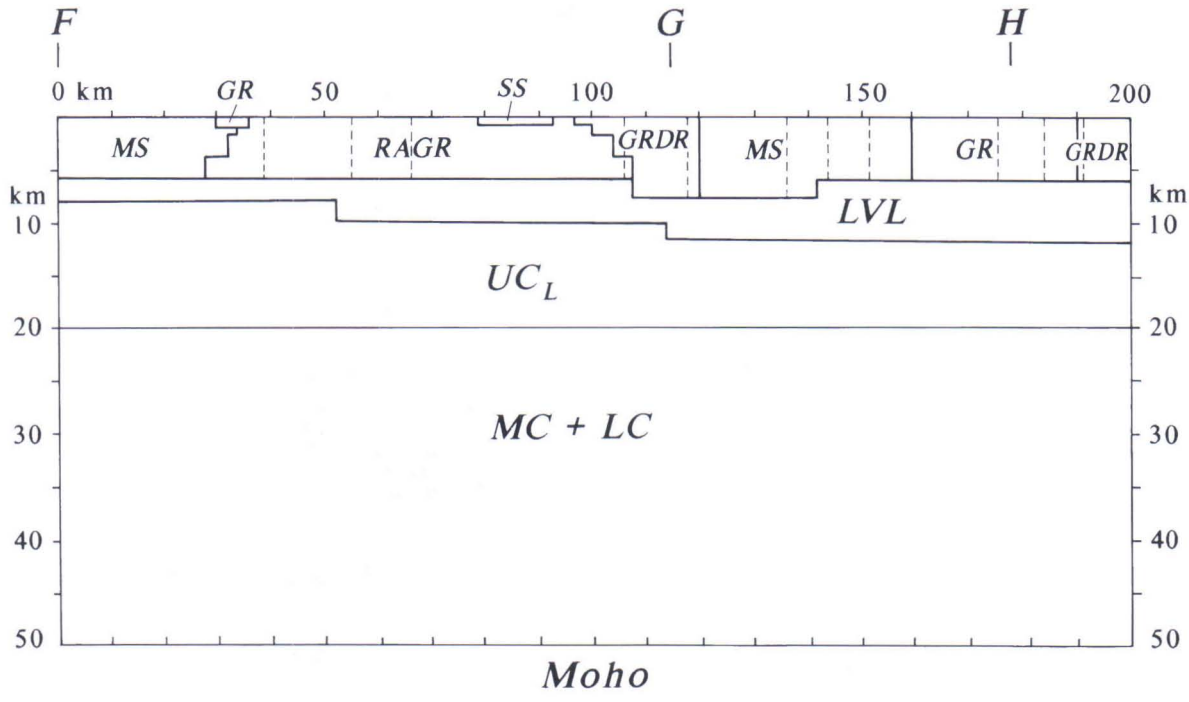
Subsurface structures were simplified from deep seismic sounding (DSS) data (Grad and Luosto

1987; U. Luosto 1993, written communication) and gravity field modellings where available (Elo 1982). Thermal properties (conductivity and heat production rate) were derived from previous laboratory measurements of different rock types, published data on surface geology (lithological maps), and geochemical surface data, all of which were supported by DSS data. Both lithogeochemical and glacial till geochemical data were used.

A fundamental problem in thermal modelling pertains to crustal heat production. Much lithospheric thermal modelling has been based on heat production estimated with the aid of the heat production-P-wave velocity relationship (Rybach and Buntebarth 1984; Čermák and Bodri 1993). Although it is a very elegant technique for estimating heat production values at deep crustal levels, it is not utilized in this study; critical discussions on the relationship and its implications can be found in the literature (Fountain 1986; 1987; Rybach and Buntebarth 1987). Instead, the DSS data was utilized by adopting the most plausible lithological interpretations for the V_P and Poisson ratio data. The modelled DSS data on GGT/ SVEKA by Grad and Luosto (1987) show considerable lateral variation in V_P and Poisson ratio values in the uppermost 10 km (above the low velocity layer, LVL). When compared to surface geology it is apparent that this variation can be attributed to the surface lithological variations extending to depths of several kilometres. Below the LVL the lateral variation practically disappears and the crust seems to be only vertically stratified.

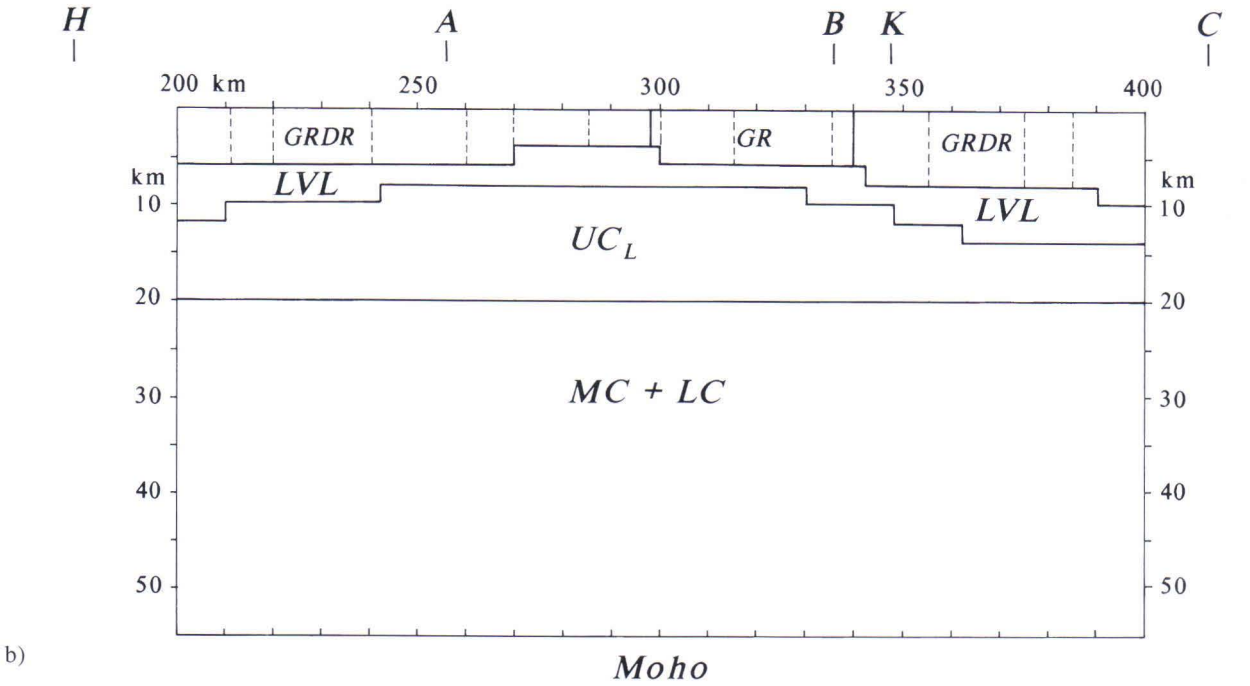
Holbrook et al. (1992) have compiled seismic P- and S-wave velocities and Poisson ratios for different rock types measured in the laboratory at pressures and temperatures corresponding to middle and lower crustal conditions. On a V_P-Poisson ratio diagram different lithological types can be distinguished, although considerable overlap naturally exists. Plotting the GGT/ SVEKA data on these graphs yields plausible estimates of the subsurface rock compositions. The following very preliminary estimates can be given: Upper crust below the LVL (appr. 10 - 20 km) - granite, granodiorite, and felsic amphibolite facies gneiss; middle crust (20 - 40 km) - intermediate granulite and metapelite; and lower crust (40 - 55 km) - metapelitic granulite, pyroxenite and amphibolite. Below the Moho the GGT/ SVEKA data is approximate to pyroxenite, eclogite and dunite.

Heat production values for the uppermost 10 km were taken from several published lithogeochemical studies of particular targets. The values ranged



a)

Fig. 1. Distribution of lithological domains in the thermal model. a) southwestern, b) central and c) northwestern part of the GGT/ SVEKA transect. The horizontal dimension measures distances along the GGT/ SVEKA transect from shotpoint F. Locations of shotpoints are indicated above the model. Vertical dashed lines indicate the locations of the vertical fracture zones included in model variant 3 (see text). Abbreviations: MS = metasediments and other supracrustal rocks, GR = granite, SS = sandstone, RAGR = rapakivi granite, GRDR = granodiorite, GRGN = granite gneiss, HVB = high velocity body, LVL = low velocity layer, UC_L = upper crust (lower part), MC = middle crust, LC = lower crust.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200 km
250
300
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MC + LC
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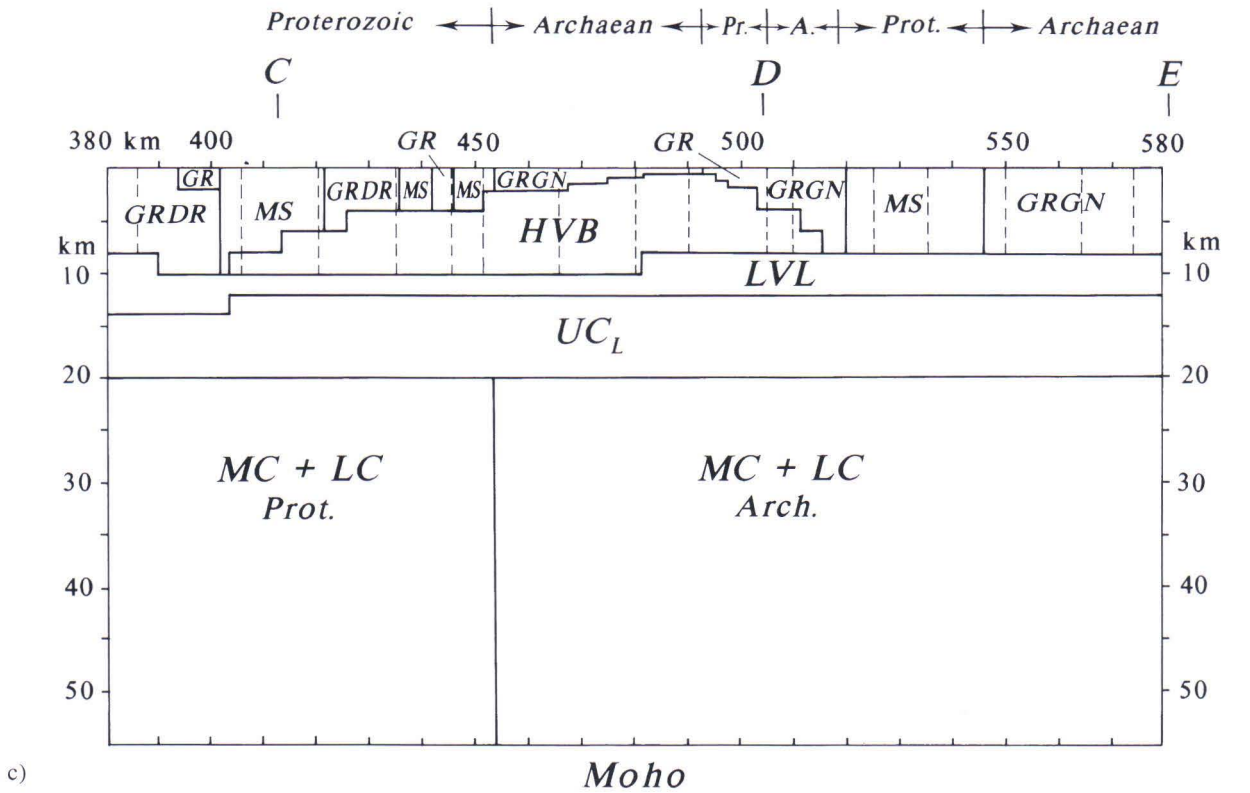
50
Moho

b)


Proterozoic
Archaean
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380 km
400
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GRGN
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km
LVL
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MC + LC
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MC + LC
Arch.
30
40
50
Moho

c)

Fig. 1.

from 0.8 to 3.1 μW m^{-3}. The middle and lower crustal heat production values were estimated with the aid of published values for middle and lower crustal rocks from other areas brought to the surface either as xenoliths or in tectonically emplaced or exhumed mafic/intermediate granulite complexes (Rudnick 1992; Percival et al. 1992). The applied values ranged from 0.3 to 0.1 μW m^{-3}.

Hydraulic permeability values were generalized from published in situ borehole measurements in Finland (Teollisuuden Voima Oy 1992; Ahonen and Blomqvist 1994) and the vertical variation of permeability was assumed to follow qualitatively the seismically interpreted fracture frequency variation with depth (Grad and Luosto 1992). The seismic surface wave data on the GGT/ SVEKA profile suggested rapidly decreasing fracture frequency with depth in the uppermost 4 km. The applied permeability values in the whole crustal model range from $1 \cdot 10^{-15}$ to $1 \cdot 10^{-21}$ m^2. The lowest values were assigned to the middle and lower crust, thus making them practically impermeable.

The DSS results (Grad and Luosto 1987) revealed a conspicious low velocity layer (LVL) 2 - 6 km thick in the upper crust between 5 - 15 km depths. In the present study the LVL is assumed to be a fractured, fluid-filled zone with a homogeneous 'lithology' (in terms of heat production and thermal conductivity) resembling the upper 10 km of the upper crust. In advective simulations the LVL is assumed to have a moderate permeability ($1 \cdot 10^{-16}$ m^2).

The hydraulic heads used for constraining advective heat transfer were estimated from topographic elevation data assuming that the groundwater table is identical with the ground surface, which is an acceptable assumption since the mean annual precipitation in Finland is 600 - 700 mm. In reality the groundwater table is usually a few metres below the ground surface, but this error is negligible for the purposes of the present model.

Considering the large degree of uncertainty in the parameter values the model was kept as simple as possible. Most of the variation in parameter values is concentrated in the upper crust where the information density is highest, whereas the middle and lower crust are considered to be much more homogenous units.

Model discretization and boundary conditions

The present model is 580 km long, beginning starting from shotpoint F and is divided into three segments, each 200 km long and partly overlapping with the others. Boundaries between segments were chosen to coincide with local water divides determined from the topographic elevation data. In the vertical direction the model extends to the Moho, which was taken to be at 50 km depth in the southwestern part and at 55 km in the central and northeastern parts of the model (Fig. 1).

The compiled 2-dimensional model was discretized with a constant cell size (2 km) in the horizontal but with a varying cell size (0.5 km - 10 km) in the vertical direction. This vertical variation in the cell size is mainly due to the requirement for controlling the permeability structure, especially in the uppermost 10 km.

The applied thermal boundary conditions were as follows. The upper boundary was kept at a constant temperature (5°C), no heat flow was permitted across the lateral boundaries and a constant heat flow density was prescribed at the lower boundary. However, between simulations the basal HFD was varied to find the best estimate for HFD at the Moho.

The hydraulic boundary conditions were as follows. At the upper surface the groundwater table was kept constant (recharge and discharge across the boundary permitted). No water was allowed to flow across the lateral or lower boundaries.

Model variants

Three main versions were calculated with the present crustal model: (1) a variant with only conductive heat transfer, (2) a variant with both conductive and advective heat transfer (mean permeability values were applied for different domains and depth intervals of the model), and (3) a variant in which both conductive and advective heat transfer were allowed and there were several additional vertical high permeability (10^{-15} m^2) zones in the upper crust extending down to the LVL (Fig. 1). The locations of these zones were determined using the strong horizontal gradient zones in the elevation and gravity data and were intended to represent deep reaching fracture zones. As a result a vertical 'fracture zone' was present about every 10 km in the model, which is in agreement with the lineament density studies of Vuorela (1982).

All the simulations were calculated for steady-state conditions and no time-dependent processes were included

The simulated HFD values at 0.5 km depth were compared with the measured drill hole data along the profile and the model parameters were adjusted until a reasonable agreement was obtained between the model and measurements. In practice, the Moho heat flow and the middle and lower crustal heat

production in the Archaean domain were found to be the only parameters that needed any fine calibration.

Results

An example of the crustal temperature and HFD calculated with the model is presented in Figure 2 (Archaean-Proterozoic boundary, model variant 1, conductive heat transfer). Horizontal heat flow density profiles sampled from the model (at 0.5 km depth) are given for all model variants in Figure 3. Conductive crustal geotherms and HFD-depth cur-

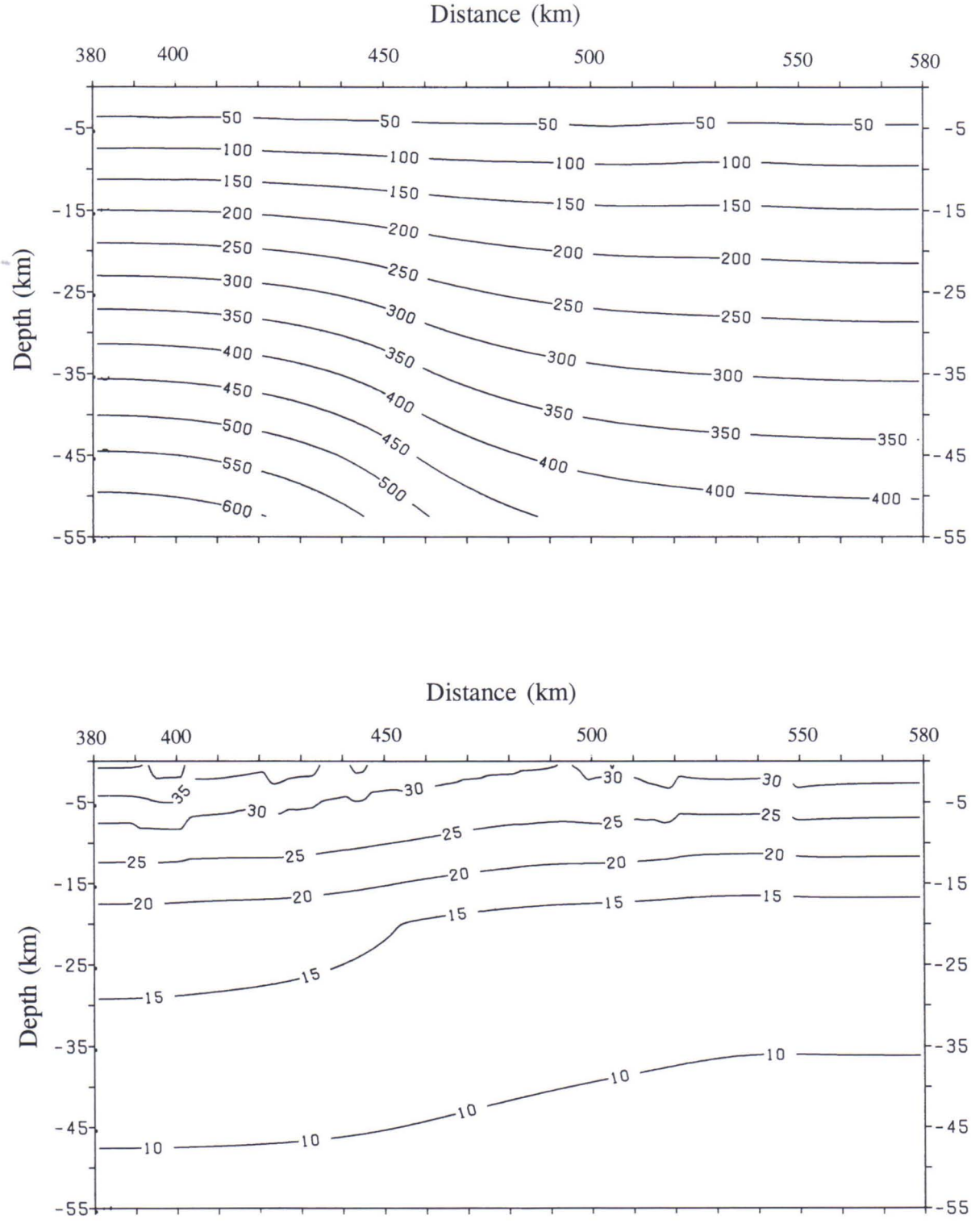

Fig. 2. Results of conductive heat transfer modelling in the northeastern part of the model (Fig. 1c). Temperatures in °C (above) and HFD in mWm^{-2} (below).

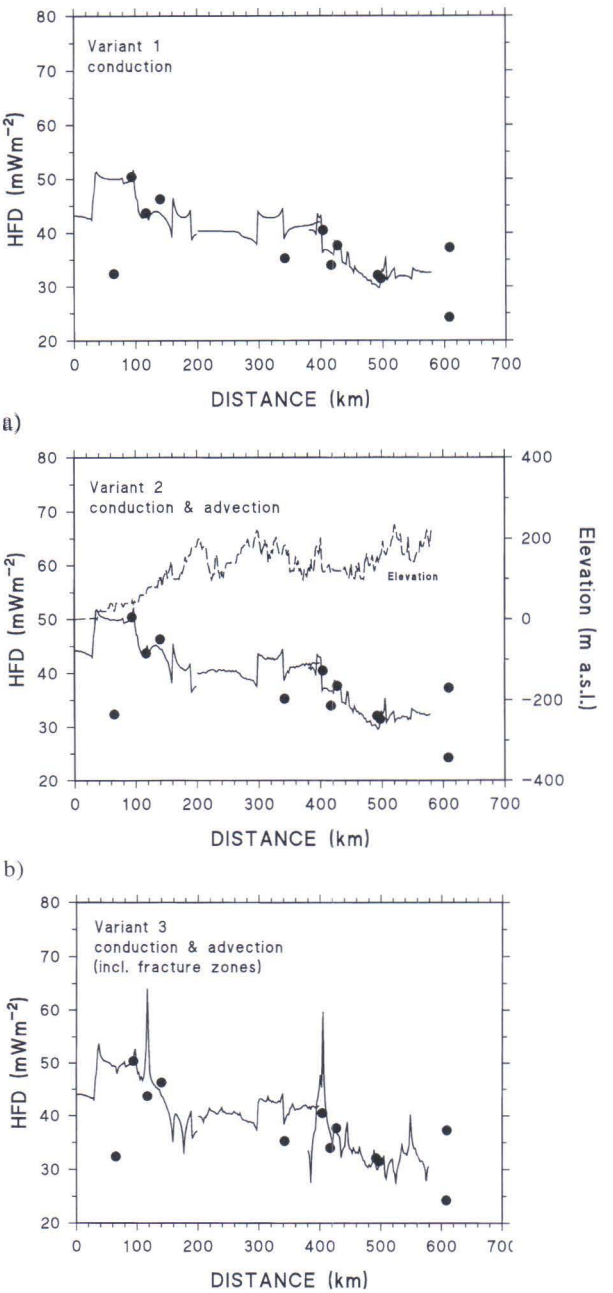

Fig. 3. Horizontal HFD profiles sampled from the model at 0.5 km depth. a) model variant 1, conductive heat transfer, b) model variant 2, conductive and advective heat transfer, c) model variant 3, conductive and advective heat transfer with additional high permeability fracture zones in the upper crust (cf. Fig. 1). The dots indicate HFD values measured in drill holes in the vicinity of the GGT/ SVEKA transect and projected onto it. Topographic elevation of the ground surface (hydraulic head) is shown with model variant 2.

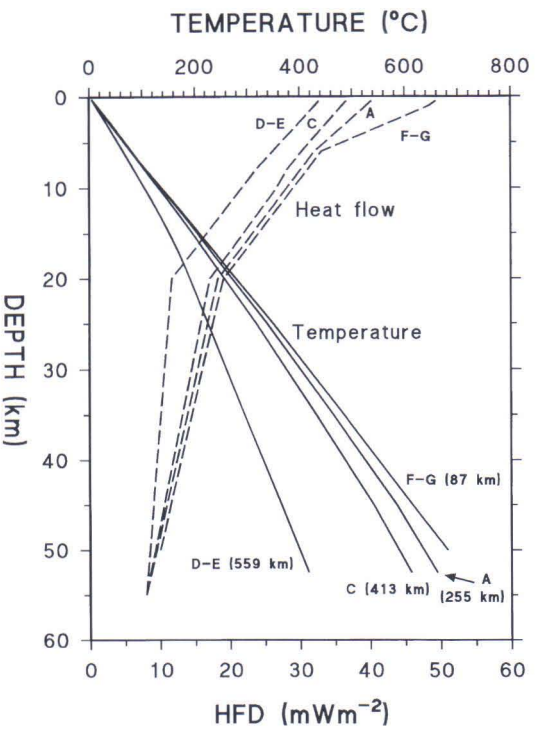

Fig. 4. Selected temperature-depth and HFD-depth profiles from the conductive simulation (model variant 1). The curves correspond to the rapakivi-sandstone area in the southwest (between shotpoints F-G), Central Finland Granitoid Complex (A), Lake Ladoga-Bothnian Bay Zone (C) and Archaean granite gneiss (D-E) in the northeast.

ves from different locations are presented in Figure 4.

Variations in upper crustal heat production as deduced from lithogeochemical data and applied in the model seem to agree reasonably well with the measured heat flow values (Fig. 3). There was no need to adjust the middle and lower crustal heat production (0.3 μWm^{-3}) in the Proterozoic part of the model, but in the Archaean part this parameter had to be decreased to 0.1 μWm^{-3}. The Central Finland area remains somewhat problematic and the simulated heat flow is about 7 mWm^{-2} higher than the measured drill hole value at Konginkangas (Fig. 3, at 342 km). This may be an indication that the middle and lower crust is depleted in heat producing elements, or only local heat transfer effects on the particular site.

Some of the HFD values projected onto the transect deviate significantly from the simulated results (Fig. 3). These discrepancies are mainly due to the fact that the particular geological structures are not represented in the 2-dimensional model. This is for instance the case for the Eurajoki (Fig. 3, at 65 km, 32 mWm^{-2}) and Hyrynsalmi (at 609 km, 37 mWm^{-2}) sites.

The simulations suggest relatively small variations in the lower crustal temperatures of the Proterozoic part of the GGT/ SVEKA transect (Fig. 4). According to the conductive simulations the temperature at 50 km depth decreases gradually in the Proterozoic area from about 680°C on the Bothnian Bay Coast through about 640°C in the Central Finland Granitoid Complex and further to about 580°C in the Lake Ladoga-Bothnian Bay zone. In the Archaean part of the model the temperature at 50 km quickly decreases to about 400°C.

HFD at the Moho (the mantle HFD) is relatively small. In the southwestern part of the model a value of 10 mWm^{-2} was found to be satisfactory and in the central and northeastern parts it is 8 mWm^{-2}.

The difference between the horizontal HFD profiles of the pure conductive simulation (model variant 1) and the conductive and advective simulation (model variant 2) is very small (< 3 mWm^{-2}). This is for two reasons. First, the applied permeability values are not very high, and second, the hydraulic gradients are not very high either. Groundwater recharge and discharge takes place over relatively short distances (about 10 km) in the model, and that is not capable of generating substantial advective heat transfer effects.

Adding vertical fracture zones to the model (variant 3) produces sharp HFD minima and maxima where local head variations are large. These spikes are not apparent in the measured HFD values (Fig. 3), and we may conclude that in this respect the model variant 3 may not be fully representative of the real situation.

The present simulations indicate fluid flow rates (darcy velocities) of the order of 10^{-12} - 10^{-13} m/s in the upper crust. The Peclet numbers were usually smaller than 0.01 which means negligible advective disturbances to the thermal field (model variant 2).

Discussion and conclusions

The simulated temperatures are somewhat higher and Moho HFDs lower than those suggested by earlier results (Kukkonen 1992) based on 1-dimensional models and heat production determined from V_p-A relationships. The present results indicate that the observed surface heat flow density variations can mostly be explained by upper crustal heat sources (lithological variation). Further, the middle and lower crustal heat production seems to be lower in the Archaean than in the Proterozoic domain.

Circulation of groundwater does not seem to produce significant deviations to the thermal field and it seems that conductive heat transfer could be taken as a working hypothesis for further studies. However, it must be remembered that only a few sites have been measured in situ and most of these represent bedrock blocks that were chosen because of their small fracture frequency and presumably low hydraulic permeability. In general, the hydraulic permeablity of the crust is very poorly constrained and increasing its value in the model by one order of magnitude from those applied in the present simulations would affect the thermal field considerably. On the other hand, the available HFD measurements do not seem to support large regional advective disturbances in the thermal field, although local flow anomalies have been indicated in earlier studies (Kukkonen 1988).

The present model is a steady-state model and the palaeoclimatic disturbances created by the Weichselian glaciation and Holocene climatic changes in ground surface temperature have not been taken into account. Forward modellings with half-space models have indicated that if the heat transfer is purely conductive the palaeoclimatic effects should have decreased the present HFD by only a few mWm^{-2} (Kukkonen 1987, 1989b). If included in the model this effect could be easily compensated for by increasing the Moho HFD's correspondingly.

The present model results should be regarded as preliminary only. Further work is needed, for instance, to obtain a more detailed picture of upper crustal heat production values, particularly in the Central Finland Granitoid area. At the moment the model has not been quantitatively analysed for its sensitivity to changes in the parameter values; this remains to be tested by future simulations.

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GEOELECTRIC STUDIES IN THE FENNOSCANDIAN SHIELD AND ALONG THE GGT/ SVEKA TRANSECT

by

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Korja, T.,S.-E. Hjelt, S.-E., Kaikkonen, P., Pernu, T., Salmirinne, H. & Tiikkainen, J. 1996. Geoelectric studies in the Fennoscandian Shield and along the GGT/ SVEKA transect. In: Ekdahl, E. and Autio, S. (eds.) Global Geoscience Transect/ SVEKA - Proceedings of the Kuopio Seminar, Finland 25. - 26.11.1993. Geologian tutkimuskeskus, Tutkimusraportti – *Geological Survey of Finland, Report of Investigation 136*, 31 - 47.

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Methods for determining electrical conductivity

Information concerning electrical conductivity of the subsurface at a crustal scale is inferred mainly from magnetotelluric soundings (MT) and magnetometer array studies (MV). These techniques utilize the simultaneous registration of several spatial components of the Earth's electromagnetic field, thus providing directional information about the subsurface conductivity (see e.g. Fig. 3). Locally controlled source data (e.g., data from direct current and VLF resistivity surveys, frequency soundings, and airborne electromagnetic mapping) provide information concerning near-surface structures, e.g. the possible surface expressions of deeper conductors.

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, 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 us to investigate in detail both the position of conductors with respect to geological and tectonic structures and the internal structure of the conductors themselves. The principal survey characteristics of the geoelectric methods used in the Fennoscandian Shield are given in Table 1 (Korja and Hjelt 1993).

Geoelectric studies along the GGT/ SVEKA transect include crustal scale investigations with magnetometer arrays (Pajunpää 1987) and magnetotelluric soundings (Korja and Koivukoski 1994). The magnetotelluric GGT/ SVEKA profile consists primarily of 82 sites between Kustavi, near the city of Uusikaupunki in SW Finland, and Malahvianvaara, which is close to the national border between Finland and Russia. From Malahvianvaara the profile continues across into Russian Karelia where seven MT soundings were carried out in 1993 (Salmirinne et al. 1993). Approximately 100 additional MT soundings are located within the transect research corridor (Fig. 1) and include three NS-directed profiles trending perpendicular to the boundary between the Southern Finland Schist Belt area and the Central Finland Granitoid Complex (old SVEKA (SVE), TAMPERE II (TRE), and SATA (SA) profiles; see Fig. 9). Other clusters of sites are within the Central

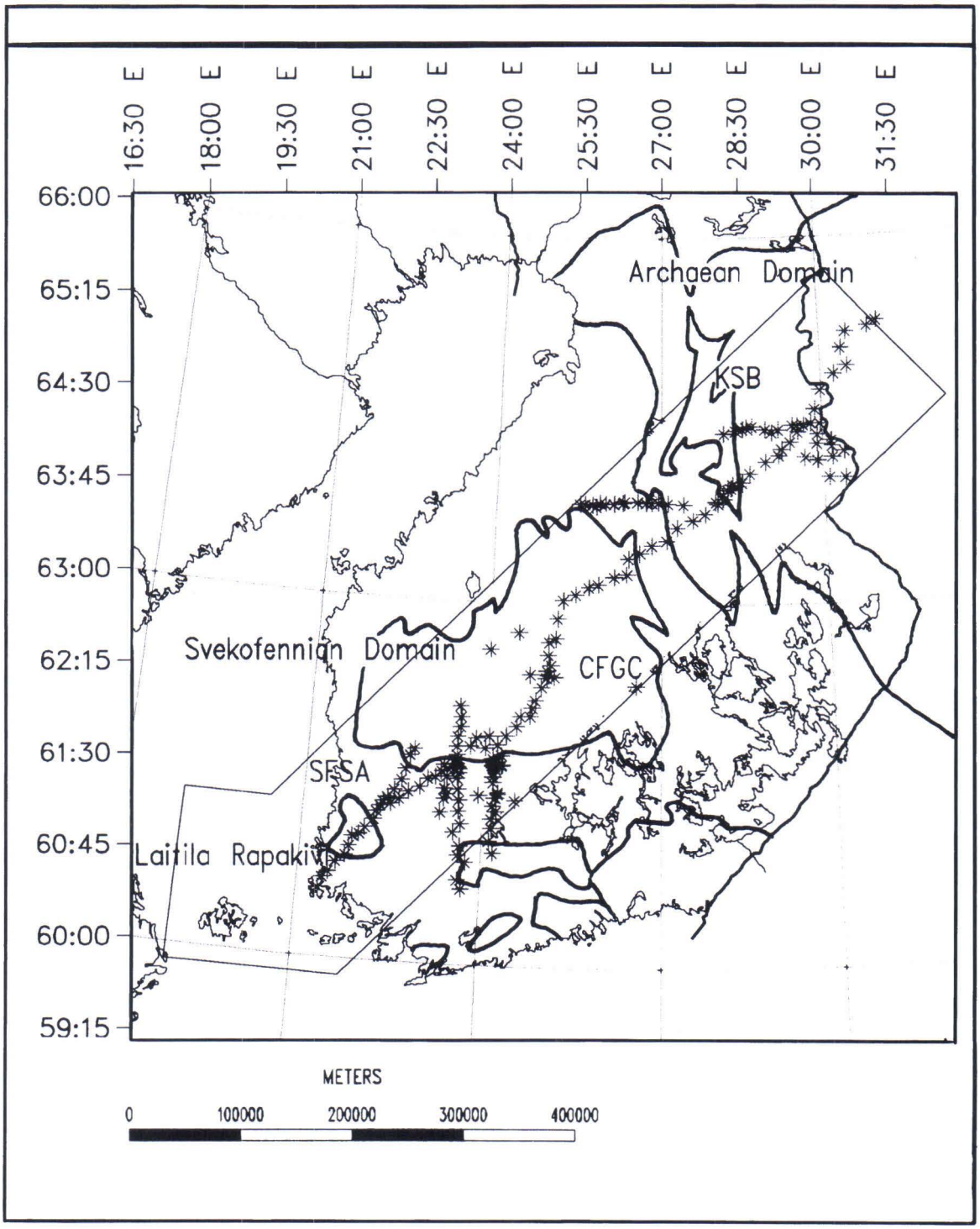

Fig. 1. Magnetotelluric (MT) sites within the GGT/ SVEKA transect research corridor. Thick black lines denote the boundaries of major geological units. CFGC-Central Finland Granitoid Complex; KSB-Kainuu Schist Belt; SFSA-Southern Finland Schist Belt Area.

Finland Granitoid Complex and the Archaean Kuhmo block. These sites yield a larger regional coverage, enabling improved estimates of the average electrical properties of the lithosphere beneath these areas. The MT sites in the eastern part of the OULU IV profile (Kokkola-Vieremä; Vaaraniemi 1989) also fall within the GGT/ SVEKA transect.

Local DC and VLF resistivity profilings and audiomagnetotelluric soundings have also produced data for use in investigating near-surface structures (e.g Kaikkonen and Pajunpää 1984; Pernu et al. 1989; Hjelt and Fokin 1990; Korja and Koivukoski 1994). Almost the entire transect research corridor is covered by low-altitude airborne electromagnetic

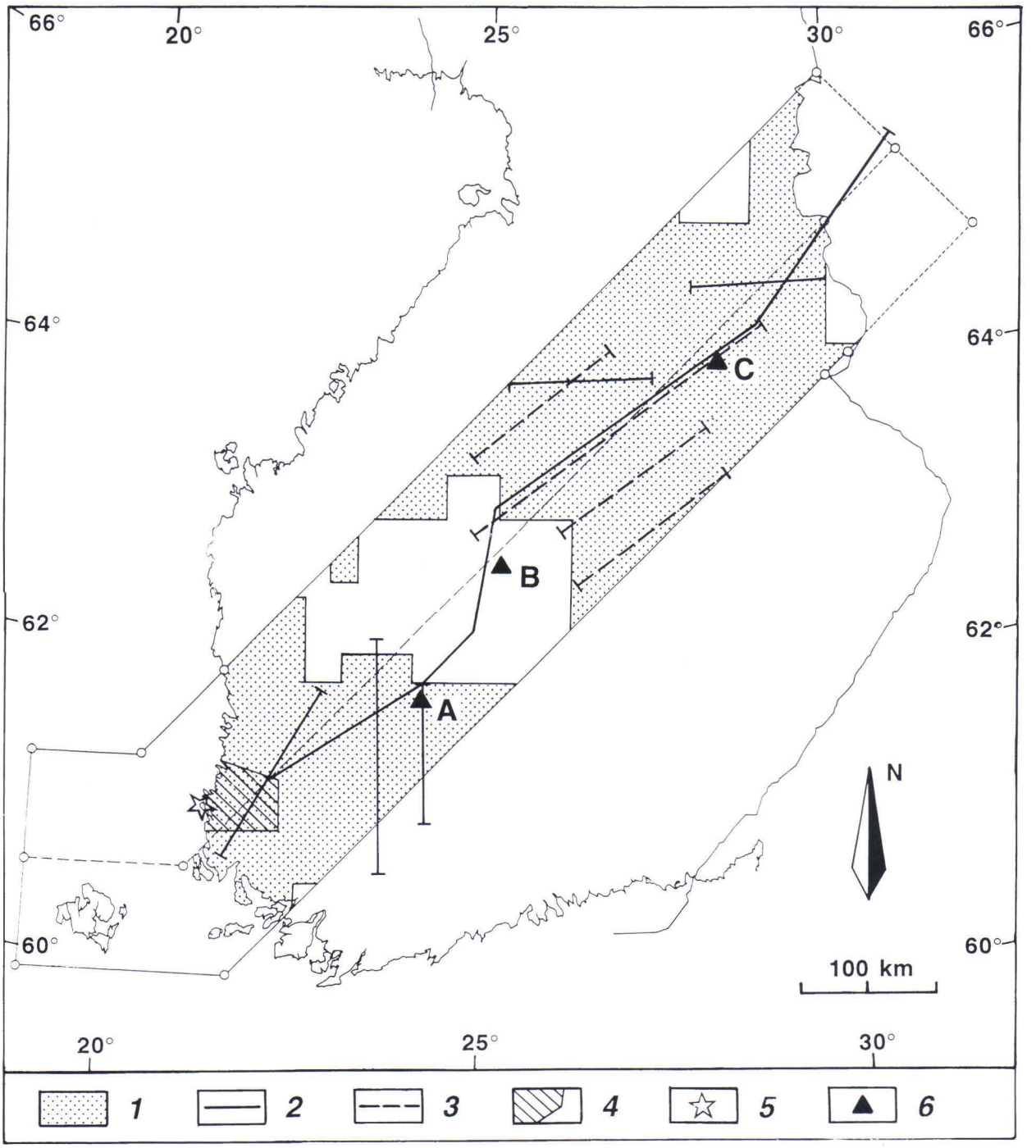

Fig. 2. Geoelectric data along the GGT/ SVEKA transect. 1-coverage of airborne low-altitude (40 m) electromagnetic data produced by the Geological Survey of Finland; 2-magnetotelluric profiles; 3-audiomagnetotelluric profiles, 4-approximate region mapped by the signal from the Fennoskan Link source; 5-cathode of the Fennoskan Link; 6-audiomagnetotelluric, DC resistivity, and VLF resistivity profiling surveys (A-Tampere Schist Belt; B-Central Finland Granitoid Complex; C-Talvivaara).

surveys carried out by the Geological Survey of Finland (Fig. 2).

Data that can be obtainable by observing the signal from the Fennoskan DC Power Link between Finland and Sweden have recently been used to estimate the electrical properties of the crust around the cathode of the power line (close to Pyhäranta near the city of Rauma) (Fig. 2) (Kaikkonen et al. 1993).

Understanding MT models

Geoelectric data are usually transformed firstly into layered (1D) Earth models. Results of most EM surveys today are presented as 2D block models from forward modelling or smooth 2D inversi-

on models whereas 3D conductivity models are still rare. At present 3D modelling is mainly used in estimating 3D effects of conducting structures and not for inverting experimental data into Earth models. All these rather crude descriptions of real Earth structures have then to be further translated into the language of geology and tectonics.

A schematic example of the interpretation of electromagnetic data and conductivity models is given in Figure 4. The figure shows a generalized model for the structure of the "collisional" conductors (see below) found in the Fennoscandian Shield, as well as the effect of sampling distance on various modelling conductors. By increasing the lateral

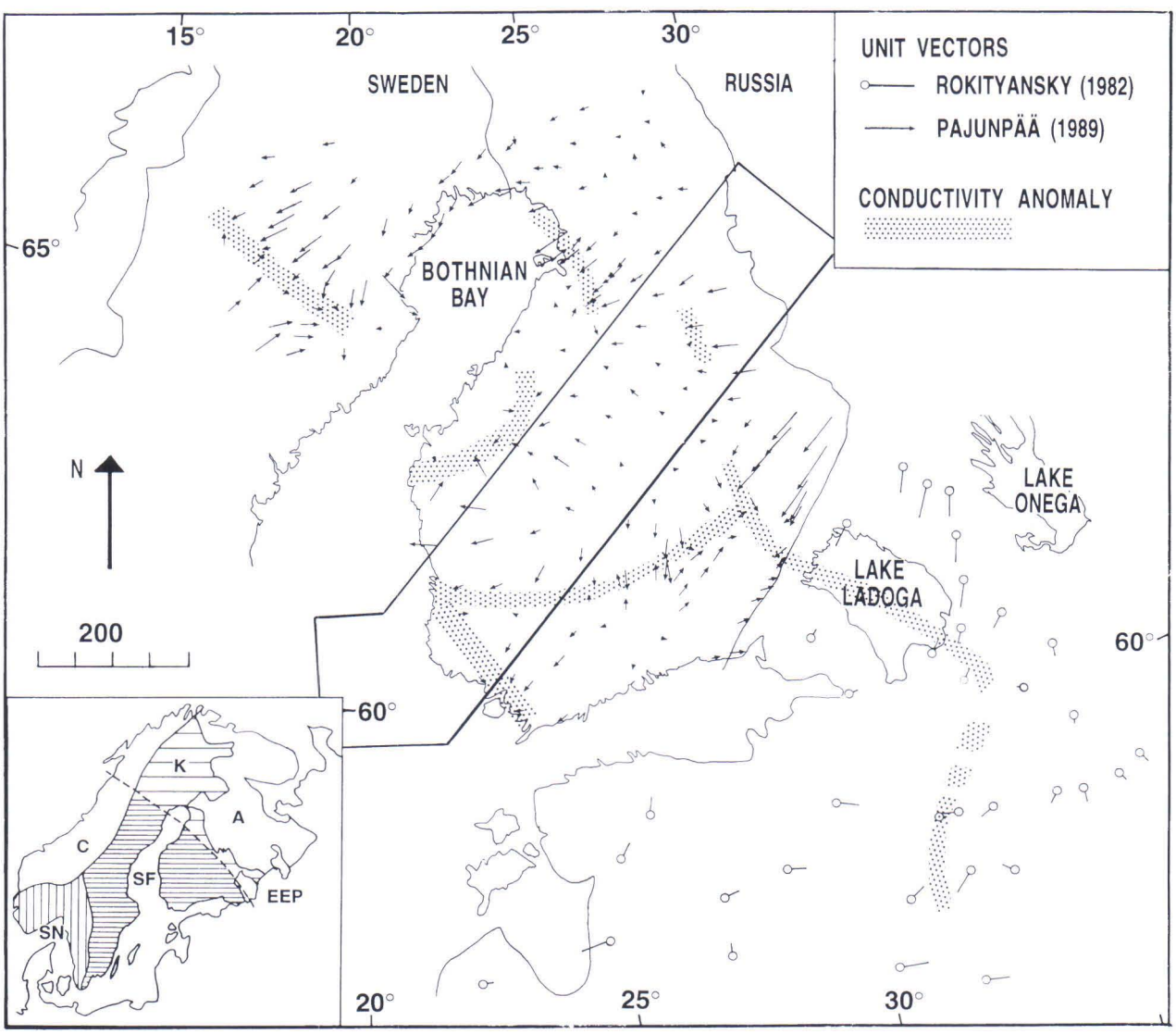

Fig. 3. Conductivity anomalies and reversed real induction vectors in the Fennoscandian Shield showing an example of directional information produced by EM methods. Vectors that point towards better conducting regions were taken from various sources (see original references in Korja and Hjelt, 1993). Vectors are shown at a period of 322 s for Finland and Sweden and at periods of 600 s to 1800 s for Russia. Stippled zones denote electrical conductivity anomalies inferred from the behaviour of the induction vectors. Inset: A-Archaean, K-Karelian, SF-Svecofennian, SN-Sveconorwegian, C-Caledonian, EEP-East European Platform.

sampling density, smaller and smaller structures can be identified and consequently the internal structure of a large scale block conductor can be resolved. The magnetovariational method allows the lateral position of the conductor (Fig. 3) and some estimation of the depth of the conductor to be determined (uppermost panel in Fig. 4). 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 be determined (second panel in Fig. 4). 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 (cf. Table 1).

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. 4) embedded within more resistive host rocks (white areas of several thousands of Ωm in Fig. 4). 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 Figure 4 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 lowermost panel in Figure 4 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. Note that the models in Figure 4 are conceptual and do not directly describe any specific real structure identified in the Fennoscandian Shield.

Electrical conductivity of the Fennoscandian Shield

We shall next briefly outline the main results of the electromagnetic investigations undertaken in the Fennoscandian Shield in order to relate the structures found within the GGT/ SVEKA transect to their wider context. More complete reviews of electromagnetic studies carried out in the Fennoscandian Shield can be found in Korja and Hjelt (1993), Hjelt and Korja (1993) and Korja (1994). The main features of the results are shown in three figures: The location and geometry of the main belts of conductors are shown together with the major tectonic units of the shield in Fig. 5 (see also Fig. 3, where regions of strong induced crustal electrical currents are delineated with the help of real reversed induction vectors obtained by the MV method). The two-dimensional models with seismic information are shown in Figure 6 and the lower crustal conductivity variations within the shield in Figure 7.

Table 1. Geoelectromagnetic methods used for studies of lithospheric electrical conductivity in the Fennoscandian Shield and the principal survey characteristics.

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	1-250 km
Audiomagnetotellurics (AMT)	horizontal/vertical	0.1-2 km	0.5-5 km
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
Magnetohydrodynamic (MHD)3)	horizontal/vertical	0.5-25 km	-15 km
Fennoskan 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 direct current power transmission link between Finland and Sweden across the Gulf of Bothnia.

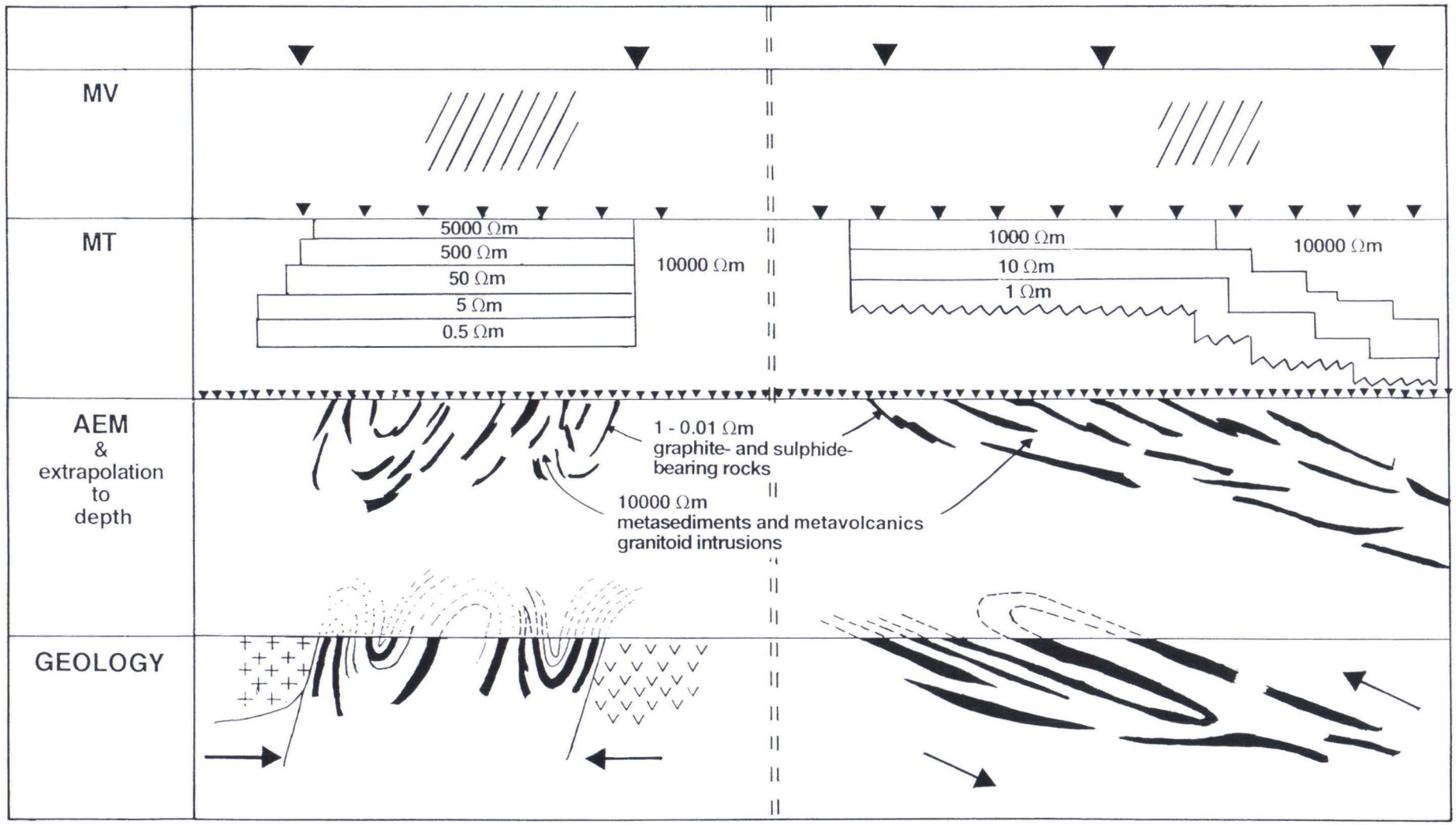

Fig. 4. Mapping and zooming in EM investigations. The diagram shows the effect of decreased sampling distance: when the lateral sampling distance is progressively decreased additional details of the internal structures are revealed. The models shown here are conceptual ones and do not directly represent any particular example. See other details in the text. Figure is after Korja and Hjelt (1993).

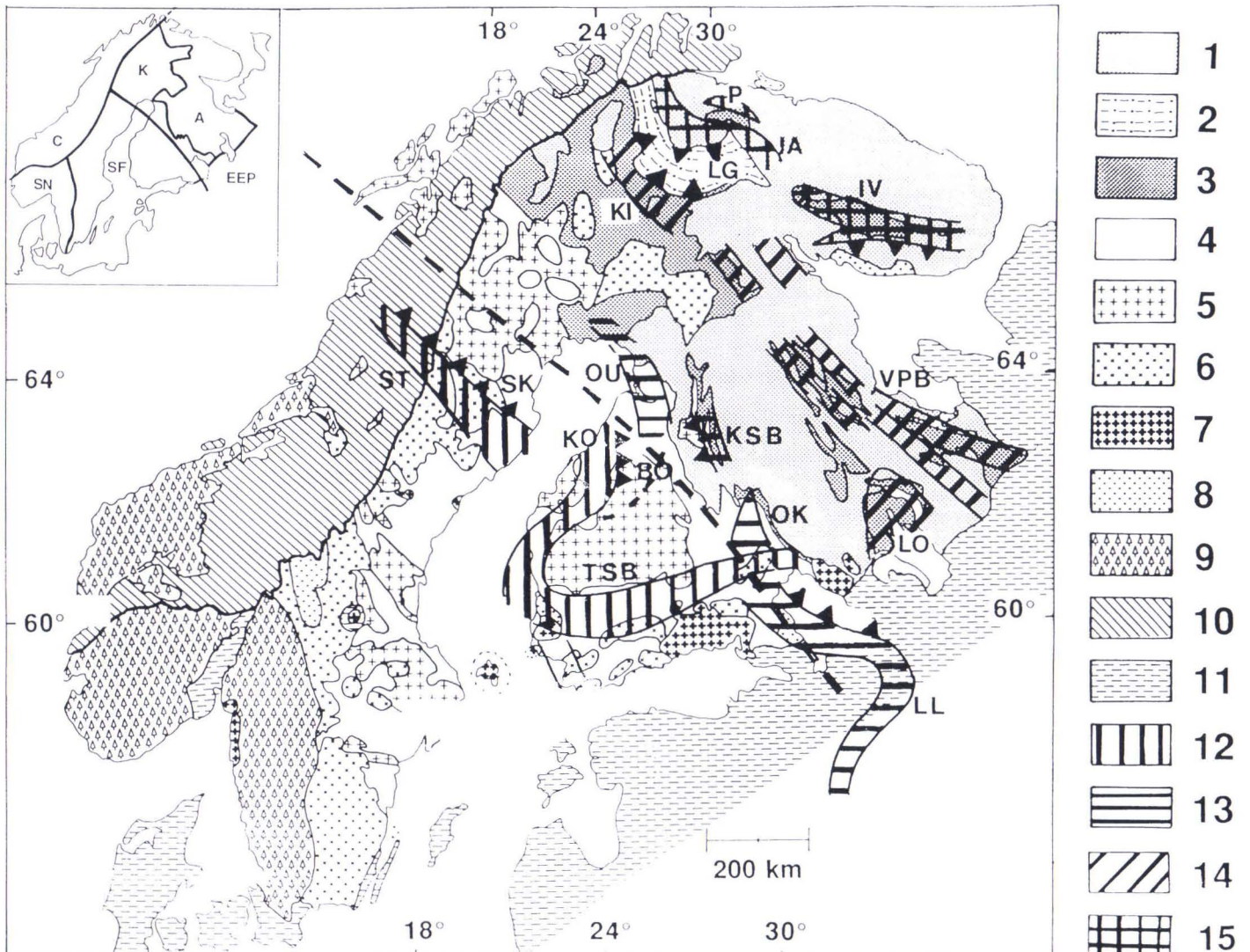

Fig. 5. Belts of crustal electrical conductors in the central and eastern Fennoscandian Shield. The geometry of the conductors is based on MV, MT, and MHD data. Structures with related tectono-geological background are shown by equivalent patterns. More detailed descriptions of the relations between individual conductors are given in the text and in Korja (1990, 1994) and Korja et al. (1993). Abbreviations used for the conductors: BO - Bothnian; IA - Inari-Allarechen; IV - Imandra-Varzuga; KI - Kittilä; KO - Kokkola; KSB - Kainuu Schist Belt; LG - Lapland Granulite; LL - Lake Ladoga; LO - Lake Onega; OK - Outokumpu; OU - Oulu; P - Pechenga; SK - Skellefteå; ST - Storavan; TSB - Tampere Schist Belt / SFCA; VPB - Vetrenny-Poyas Belt. Figure is after Korja et al. (1993). Inset as in Figure 3.

Results have revealed a complex electrical structure for the crust of the Fennoscandian Shield, with resistivity variations from 106 Ωm to below 10 - 1 Ωm. 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 (Fig. 5). The internal structure of the conductors may be very complicated as is evident from the airborne electromagnetic data (see e.g. Fig. 9). The resistive blocks are usually relatively homogeneous laterally but have vertical variations in conductivity. These blocks serve as transparent geoelectric windows for probing the electrical properties of the lower crust and upper mantle. It should be noted, however, that a careful 3-D analysis should be carried out in order to investigate possible inductive effects on measured responses due to surrounding conductive structures.

Conductive belts

The four main belts of elongated conductors in the central and eastern parts of the shield are as follows (Fig. 5; Korja et al. 1993):

(i) The NW-SE trending zone of conductors from the Kittilä Greenstone Belt (KI) to the Vetreny-Poyas Greenstone Belt (VPB).

(ii) A set of NW-SE trending, southwestward dipping, highly conducting zones in the Kola Peninsula and in northern Finland including the Inari-Allarechen (IA), Pechenga (P) and Imandra-Varzuga (IV) conductors.

(iii) A belt of discontinuous conductors coincident with the Archaean and Svecofennian boundary including the Lake Ladoga (LL), Outokumpu (OK), Kainuu Schist Belt (KSB), and Oulu conductors (OU).

The Lake Ladoga (LL) conductor is a northeastward dipping conductor at a depth of about 10 km and may indicate conducting metasedimentary rocks that were deeply buried in the collision between the Svecofennian island arc complex and the Archaean craton. The LL conductor also coincides with a region containing some of the thickest crust in the shield. Beneath Lake Ladoga the northeastward dipping conductor turns westwards which may indicate the continuation of the Archaean-Proterozoic boundary underneath the Phanerozoic cover. Further to the south the LL can no longer be resolved due to the screening effect of the Phanerozoic sedimentary cover.

In the Kainuu Schist Belt (KSB) and Outokumpu (OK) regions the near-surface conductors mapped by the AEM surveys are composed of conducting graphite- and sulphide-bearing black schist layers which were thrust onto the Archaean craton during the collision. A deeper conductor beneath the Outokumpu region (MV data) most likely represents deeply buried metasedimentary rocks (cf. Lake Ladoga conductor above). The KSB and OK both coincide with thick crust.

(iv) A 500 km long conductor crosses southern Finland in an E-W direction and marks the collision of crustal blocks within the Svecofennian Domain. It is not clear however, how and if the conductor continues into Sweden. Airborne magnetic and other geophysical data indicate that the southern Finland conductor may continue towards the Skellefteå region (towards the Skellefteå (SK) and Storavan (ST) conductors, Fig. 5) via the Kokkola conductor because all conductors are associated with a regional magnetic minimum several tens of kilometres in width that contains a number of elongated narrow local magnetic anomalies.

The main features of these conductors are listed below:

(i) Conductors are highly conducting with 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 include graphite, sulfides and some oxides.

(ii) Conductors are usually caused by extremely conducting (> 1 S/m) graphite- and sulphide-bearing metasedimentary rock layers several tens of metres in width; "host" rocks to these conducting metasedimentary rock layers are usually very resistive (few thousands of Ωm); metasedimentary rock layers are usually interconnected in a very complicated manner, forming a network of conductors (see e.g. Fig. 9 below).

(iii) The belts of thin conducting metasedimentary rock layers (i.e. large scale crustal block conductors detected by MV and MT) have a variety of geometrical orientations: they may be nearly horizontal (Kittilä Greenstone Belt conductors, Outokumpu nappe -conductor), they may be inclined (Granulite Belt, Skellefteå, Kokkola, Kainuu Schist Belt and Lake Ladoga conductors), or they may be nearly vertical (Southern Finland and 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

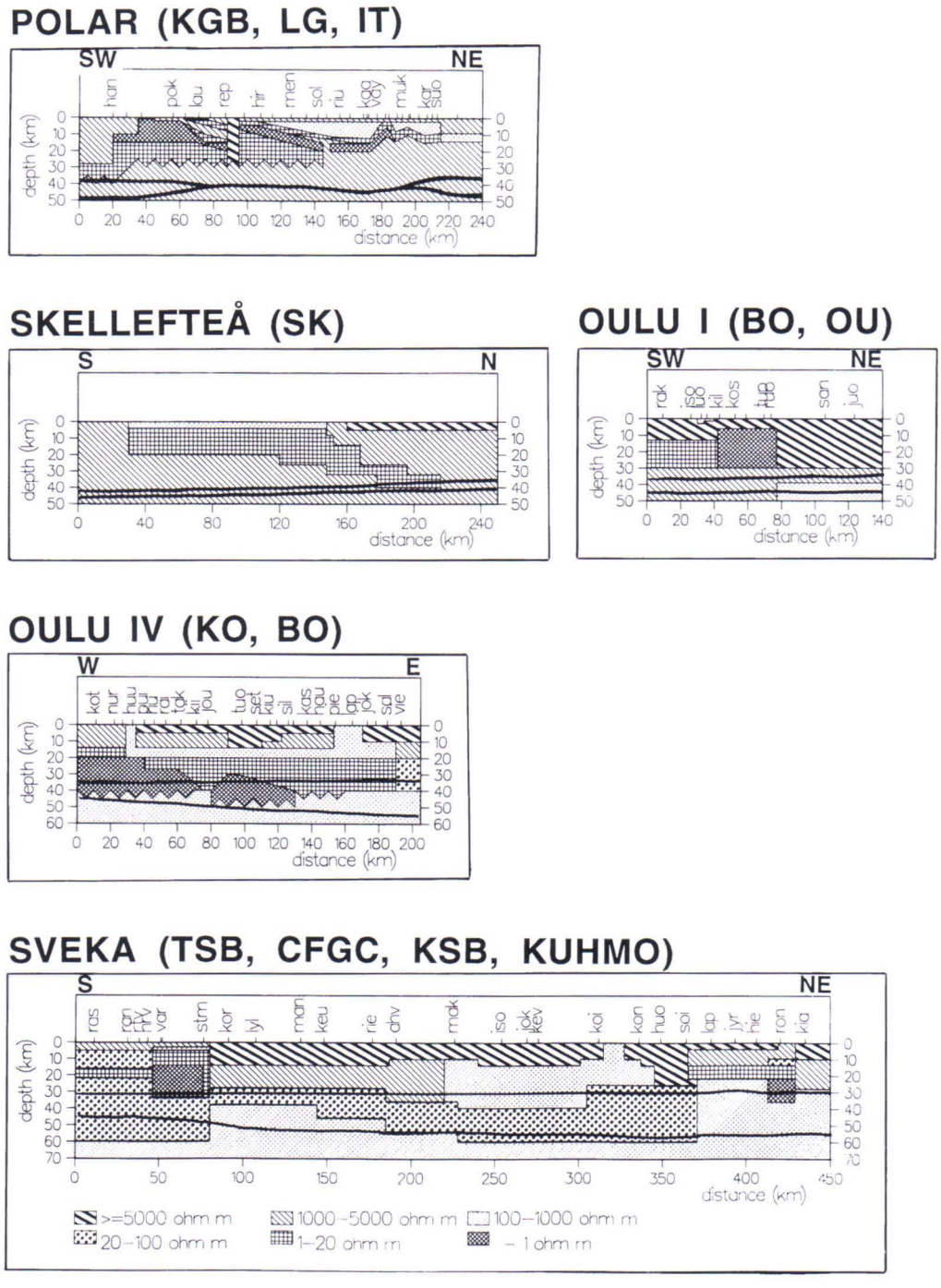

Fig. 6. 2D geoelectric models across the belts of crustal conductors. The resistivity scales of all models are the same and are shown in detail below the SVEKA model. The profiles have no vertical exaggeration. The lower and upper thick black lines indicate the Moho boundary and the boundary between the upper and lower crust, respectively (Korja et al. 1993). Figure is after Korja and Hjelt (1993).

processes formed both.

(v) The 2D-models of the conductors (Fig. 6) show that most of these conductors are located in the seismically defined upper and middle crust, i.e. conductors do not penetrate into the lower crustal layer as defined by high P-wave velocities. In those places where conductors do extend into the lower crust (e.g. Skellefteå conductor) the high velocity layer is either absent or very thin.

(vi) All conductors have been attributed to the presence of conducting metasedimentary rocks. The sediments were deposited in a variety of environments (extensional rift basins, continental margins, shallow and deep sea oceanic basins). Subduction

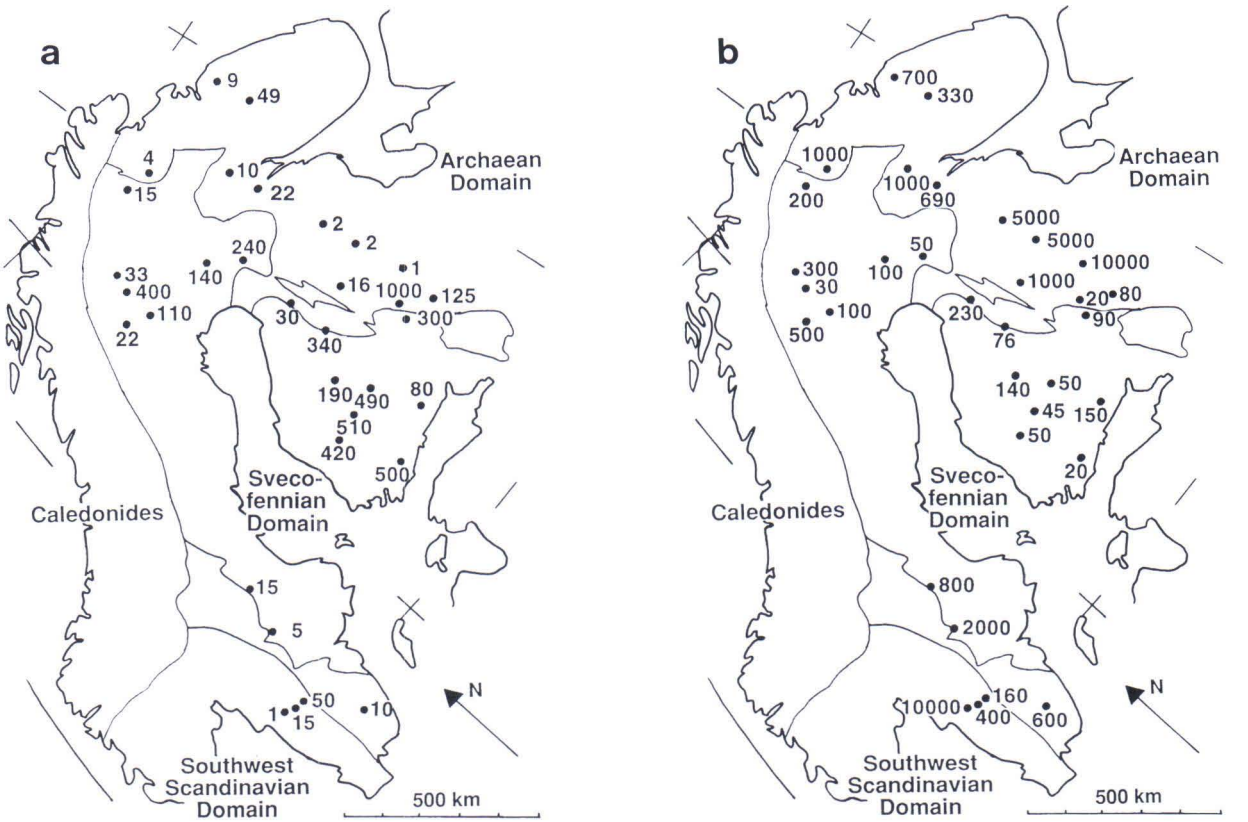

Fig. 7. Lower crustal conductivity in the Fennoscandian Shield. a) Estimates of the total conductance (in Siemens) of the lower crust determined from MV and MT data. The thickness of the lower crust is from seismic velocity data (Korja et al. 1993) and the resistivities of this part of the crust are from electrical models. See other details in text. Thin solid lines represent boundaries between the main tectonic domains of the shield. b) Average electrical resistivity of the lower crust recalculated from the total conductance estimates of the lower crust shown in Figure 7a (average resistivity [Ωm] = thickness of the lower crust [m] / total lower crustal conductance[S]). Figure is after Korja and Hjelt (1993).

and collision deformed sedimentary sequences between crustal masses and underthrust them deep into crust or emplaced them closer to the surface. Consequently these conductors can be considered as tectonic markers or "scars" of old collisional zones ("collisional conductors").

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 (Korja and Hjelt 1993). The main problem when estimating lower crustal conductivity is how to define the lower crust. If there were a distinct lower crustal conducting layer (i.e. if there were a clear decrease in resistivity at a particular depth, followed by a systematic increase beneath the conductor) we could define the conducting lower crust to coincide with the layer and then calculate the conductance of this layer.

In our approach (see alternative definitions for the lower crust in 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 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.

Estimates of lower crustal conductance and the resistivity of the lower crust estimated in the manner described above are given in Figure 7a. The data show rather large lateral variations in conductance within the shield and although these are partly due to variations in the thickness of the high velocity lower crustal layer it seems that the Archaean lower crust is more resistive than the Palaeoproterozoic Svecofennian crust. This feature is also evident in the resistivity map (Fig. 7b).

The lower crust beneath the Karelian Domain between the Archaean and Svecofennian Domains seems also to be also rather conductive even though it is considered to be composed of Archaean mate-

rial (Archaean cratonic rocks beneath the Palaeoproterozoic Karelian supracrustal cover). This might therefore indicate that the lower part of the Archaean crust in the Karelian Domain was also affected by the Svecofennian orogeny, perhaps by the additions of Palaeoproterozoic mafic material.

The electrical asthenosphere

Data for the Fennoscandian Shield were compiled from several sources, and a complete reference list is given by Korja (1990). The major problem in Fennoscandia and in its northern part, in particular, is the proximity of the source of the magnetotelluric field. The closeness of the source may produce evidence for a conducting layer in sounding responses even though no such layer exists in reality.

The results from the more southerly parts of the shield nevertheless indicate the presence of conducting asthenosphere in peripheral regions of the shield. It is apparent, however, that in the central and southwestern parts of the shield the asthenospheric layer is absent or is electrically weak. 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).

Electrical conductivity along the GGT/SVEKA transect

Main features

The old SVEKA MT profile, for which the 2D geoelectric model is available (Korja and Koivukoski 1994), coincides with the GGT/ SVEKA MT profile between Keuruu in the CFGC and the Kainuu Schist Belt. From Keuruu the old SVEKA MT profile turns to the south, cross perpendicular to the Tampere Schist Belt near Orivesi, and terminates near Hämeenlinna (Figs. 1 and 2). Consequently we are able to give only a qualitative description of conductivity structures along the SW and NE parts of the GGT/ SVEKA transect (i.e. from Kustavi to Keuruu and from KSB to Kalevala in Russian Karelia) and a quantitative description only along the central part of the GGT/ SVEKA transect (i.e. from the Southern Finland Schist Belt area to KSB). The 2D geoelectric model for the old SVEKA MT profile is given in Figure 8. A detailed description of the model can be found in Korja and Koivukoski (1994).

The main electrical conductivity features within the GGT/ SVEKA transect are as follows:

(1) Three zones of enhanced electrical conductivity have been detected along the GGT/ SVEKA transect: (i) a NW-SE trending zone beneath the Laitila rapakivi area, (ii) an E-W trending zone beneath the Southern Finland Schist Belt area, and (iii) N-S trending zone beneath the Kainuu Schist Belt (Fig. 3).

The Satakunta conductor is observed in both the magnetometer array data and the magnetotelluric data, as well as in recent Fennoskan Link data (Kaikkonen et al. 1993). The conducting zone seems to be about 20 km wide beneath the Laitila rapakivi area but neither quantitative estimates of the actual resistivities nor the depth and thickness of the conductor are available. It seems, however, that the Satakunta conductor is a shallow feature.

In the Southern Finland Schist area and Kainuu Schist Belt the conductors exceeding 15 000 S extend from the upper crust to the middle crust. Resistivity decreases gradually with depth from 100 Ωm to below 1 Ωm (Fig. 8). The conductor beneath the Southern Finland Schist area is part of a long conductivity anomaly crossing southern Finland and may be interpreted as a terrane boundary within the Svecofennian Domain. The conductor beneath the KSB forms, together with the Lake Ladoga, Outokumpu, Kainuu Schist Belt and Oulu conductors, a discontinuous conductive belt along the Archaean-Proterozoic boundary. High conductivity beneath both belts is most likely caused by inclined graphite- and sulphide-bearing metasedimentary rock layers. The SW-dipping set of conductors west of the Kainuu Schist Belt (Fig. 8) may represent a zone of sheared rocks between the Svecofennian and Archaean crust. The LBBZ, which is defined as a major shear zone between the Archaean and Svecofennian Domains, cannot be distinguished clearly from its surroundings with respect to electrical conductivity in the middle and lower crust.

On the basis of AMT data (140 soundings), however, it is possible to distinguish three electrically different upper crustal (< 5 km) units in the central part of the GGT/ SVEKA (Kaikkonen and Pajunpää 1984), with the central block, which coincides with the LBBZ, being the most conductive and the easternmost block the most resistive. 2D modelling of the AMT data gives resistivity values of 500 Ωm to 1000 Ωm for the upper part of the crust in the LBBZ and resistivity values varying from 1000 Ωm up to 200 000 Ωm on both sides of

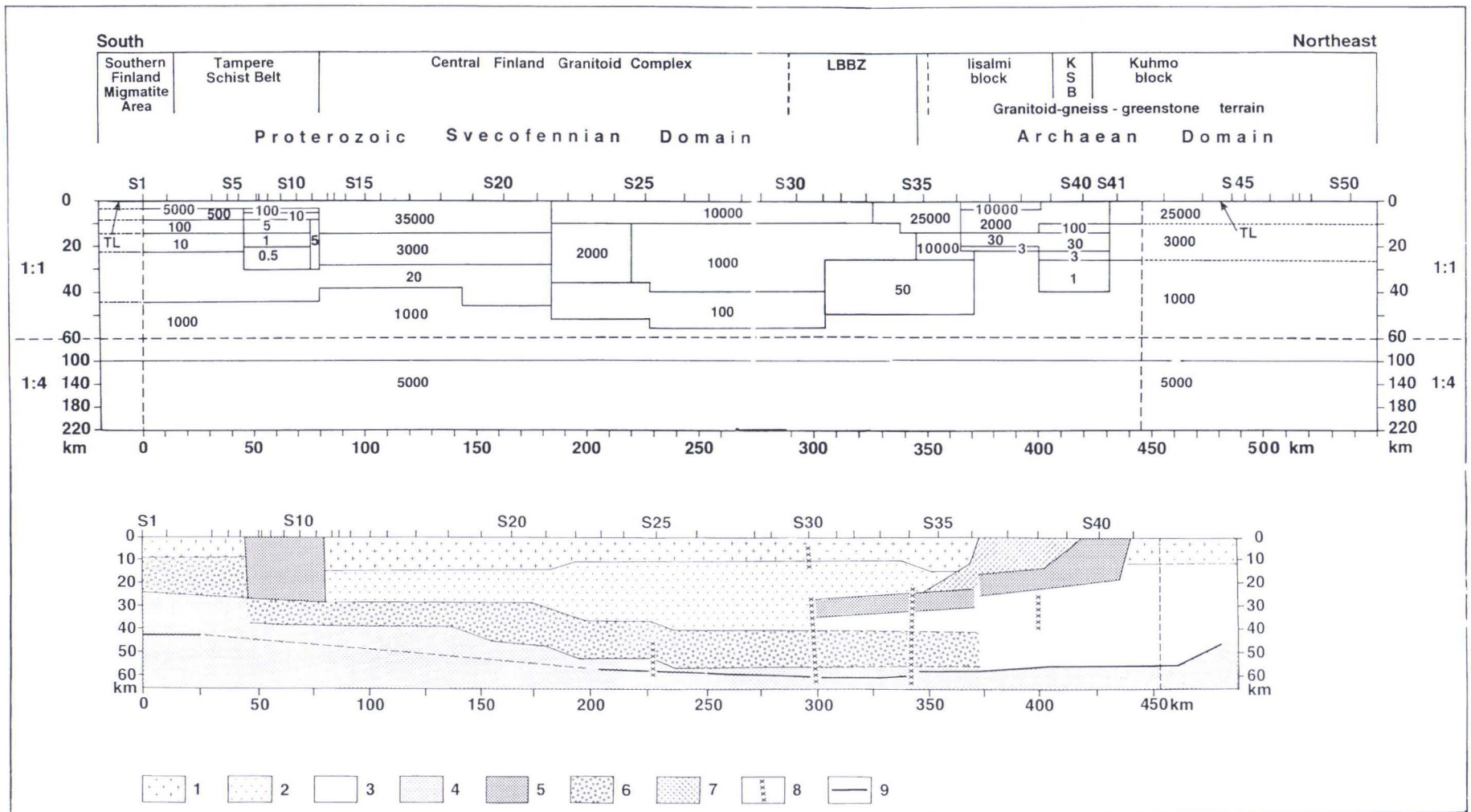

Fig. 8. 2D geoelectric model of the old SVEKA profile (upper panel) and its interpretation (lower panel). Numbers in the model denote resistivity values (Ωm). Scale is 1 : 1 (no vertical exaggeration) for upper 60 km and 1 : 4 (vertical reduction) for deeper parts. TL, which extends over the whole model, denotes a thin layer included in the upper part of the model to simulate a conductive overburden (conductance 0.1 S). LBBZ = Lake Ladoga - Bothnian Bay Zone, KSB = Kainuu Schist Belt, KGB = Kuhmo Greenstone Belt. Note that the geoelectric model east of the KSB is based on a 1D model of an average response from 13 sites within the Kuhmo block. Geoelectric interpretation model: 1, 2, and 3-upper, middle, and lower crustal resistive rocks; 4-resistive lower lithosphere; 5-upper crustal (collisional) conductors; 6-well-conducting middle/lower crustal layers; 7-resistive upper crustal rocks beneath the Varpaisjärvi-Iisalmi block; 8-fracture zones from seismic data; 9-refraction seismic Moho.

the LBBZ. The airborne electromagnetic data also show that the uppermost crust contains electrical conductors although they are not so abundant as e.g. in the Bothnian region.

(2) The upper crust in both the Palaeoproterozoic Central Finland Granitoid Complex and the Archaean Domain is highly resistive, with model resistivities ranging from 10 000 to 35 000 Ωm (Fig. 8). Magnetotelluric data also indicate that in south-western Finland the upper crust is resistive except for the conductive Satakunta conductor.

(3) A well-developed lower crustal conducting layer of 170 to 900 S is found beneath the Proterozoic Central Finland Granitoid Complex (CFGC) (see below). This conducting layer terminates at the Archaean-Proterozoic boundary. Magnetotelluric data from the SW-part of the GGT/ SVEKA indicate that the middle to lower crust is also conductive in southwestern Finland, to the southwest of the Southern Finland Schist area. The Archaean lower crust is typically resistive, with a conductance less than 30 S. There is, however, evidence in the magnetotelluric data for the existence of a conducting layer beneath the Moho, i.e. in the upper mantle, in the Archaean Domain. Whether this is a real conducting layer or just a side effect related to a remote crustal conductor (e.g. the conductor beneath the KSB) is not yet clear. 2D modelling or inversion of magnetotelluric data is required before the real nature of this layer can be assessed.

(4) In general the Proterozoic crust appears to be more conductive than the Archaean crust.

Southern Finland Conductivity Anomaly

The ability to resolve smaller and smaller structures by decreasing the lateral sampling distance is demonstrated by studies of the southern Finland conductivity anomaly. Magnetometer array studies with a sampling distance of about 30 - 40 km firstly detected a zone of strong current concentration. This is shown in Figure 9 as a dashed line crossing southern Finland. Magnetotelluric studies with much smaller lateral sampling distances of 5 - 10 km or even less, enabled the positions of the boundaries of the conductor to be located more precisely with respect to the northern and southern geological units (Central Finland Granitoid Complex and Southern Finland Migmatite Area, respectively). Magnetotelluric studies along the old SVEKA profile (SVE in Fig. 9) also indicated that the resistivity of the conductor decreases with depth, from about 100 Ωm at a depth of 3 km to 0.1 Ωm at mid-crustal depths. The uppermost layer has a resistivity of about 5000 Ωm.

The results of the geoelectric cross-section of the Tampere II profile (TRE in Fig. 9) obtained by 1D inversion (Pernu et al. 1989) indicates that the geoelectric structures are continuous between the two profiles. 2D modelling of the magnetovariational data from the Mikkeli profile (MIK in Fig. 9) crossing the Southern Finland conductivity anomaly about 140 km east of the old SVEKA profile has revealed a conductor with a resistivity of 0.5 Ωm at a depth of 12 km (Pajunpää 1986). Taking all the data into account, it is fairly clear that the Southern Finland conductivity anomaly is caused by a good conductor, which is more or less continuous over its entire length of 400 to 500 km and is located within the upper and middle crust.

Airborne electromagnetic surveys showed that the entire region is characterized by a number of long, thin conductors (thin dashed and solid lines in Fig. 9) forming a very complex conductivity pattern. Other electrical data (e.g. AMT, DC, and VLF-R soundings and profilings) and other geophysical studies (e.g. magnetic and gravimetric) showed that structures are nearly vertical and may extend in depth to several kilometres (Pernu et al. 1989). It thus seems that the gradually increasing conductivity beneath the TSB is caused by a network of thin, long, almost vertical, several kilometres deep and extremely conducting graphite- and sulphide-bearing metasedimentary rock layers that were tilted and deeply buried during convergence of two crustal terranes. The MT method images thin vertical conductors as one 40 to 80 km wide conductor with a conductivity that increases gradually with depth. The magnetotelluric block model apparently indicates a depth of 3 km for the conductor, but AEM data suggest that the deep conductor is exposed in the form of a complex network of thin elongated conductors. Protoliths to the graphite-bearing rocks would have been carbon-bearing sedimentary rocks deposited in a closed ocean basin between the colliding crustal terranes. In addition to the graphite- and sulfide-bearing metasedimentary rocks, shear zones and fractures filled with saline water (0.25 Ωm) may partly enhance the conductivity.

Central Finland Granitoid Complex

The 2D geoelectric model of the Central Finland Granitoid Complex (CFGC) was constructed from measured magnetotelluric data by forward modelling using a trial-and-error approach (Korja and Koivukoski 1990, 1994). It was assumed after MV- and MT data and other geophysical and geological evidence that the geoelectric strike direction is E-W in the southern part of the profile and approximately N-S (-5° from N to W) in the northeastern part of the

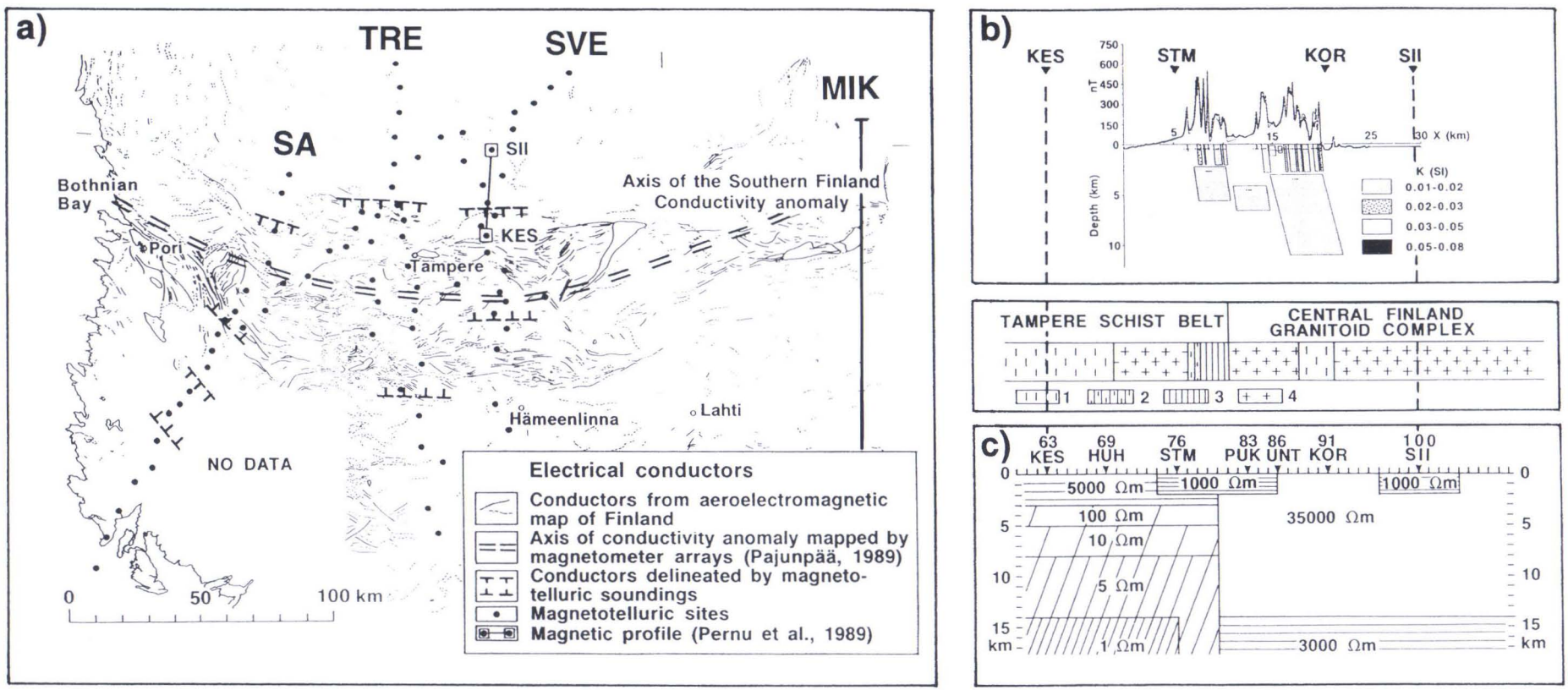

Fig. 9. Electrical conductivity in southern and southwestern Finland. a) Electrical conductors from airborne electromagnetic maps (Peltoniemi et al. 1992), magnetometer arrays (Pajunpää 1987) and MT sounding profiles (Pernu et al. 1989, Korja and Koivukoski 1990, 1994). b) A magnetic model (Pernu et al. 1989) based on airborne magnetic data with the baseline Btot = 50.670 nT. c) The two-dimensional geoelectric model on the TSB part of the old SVEKA profile (Korja and Koivukoski, 1994). Figure is after Korja (1994).

profile. Measured responses (impedance tensors or apparent resistivity and corresponding phase curves) were first rotated to coincide with the direction of geoelectric strike (parallel/ E-polarization responses) and the direction perpendicular to the strike (perpendicular/ H-polarization responses). Consequently the following data was used in modelling the old magnetotelluric SVEKA profile (Hämeelinna - Keuruu - Kainuu Schist Belt; Korja and Koivukoski 1994):

southern part of the profile:

H-pol / perpendicular to strike	NS-responses(no rotation)
E-pol / parallel to strike	EW-responses (no rotation)

northern part of the profile:

H-pol / perpendicular to strike	-5° from E to N rotated EW-responses
E-pol / parallel to strike	-5° from N to W rotated NS-responses

Modelling of these data led to the model shown in Figure 8 and the model is valid beneath both the TSB and KSB. Beneath the CFGC the model requires some modifications due to subsequent analysis of magnetotelluric impedance tensor data.

Recently a new impedance tensor decomposition method (Groom and Bailey 1989) was applied to the data of the CFGC to remove distortions produced by local near-surface 3-D inhomogeneities (i.e. to remove departures from 2D responses) and to recover regional 2D responses (with the underlying assumption that larger scale regional structure is 2D or 1D and not 3-D). This new analysis of the data (decomposition) does not change the main conclusions made previously (Fig. 8) except for one important aspect, namely that the lower crust and upper mantle seem to be electrically anisotropic. That is, although the conclusions concerning the upper crust beneath the CFGC remain valid, and the lower crust is still electrically more conductive, the conductivity is by nature anisotropic (see below) rather than isotropic, as was implied by the previous analysis (Korja and Koivukoski 1994).

The previous analysis of the magnetotelluric data (Korja and Koivukoski 1990, 1994) led to inconsistent geoelectric strike directions from site to site within the CFGC when the traditional Swift-rotation (i.e. minimization of the off-diagonal elements of the impedance tensor) was applied to obtain the geoelectric strike direction. Because no stable strike direction within the CFGC was found the strike direction of 90° (E-W) for the southern part of the profile and -5° (from N to W) for the northwestern part of the profile was selected according to other available evidence as described above. The tensor decomposition, however, recovered very consistent 2D regional geoelectric strike directions within the CFGC, in particular for long periods.

The 2D regional geoelectric strike directions obtained from the Groom-Bailey tensor decomposition are shown in Figure 10 at four different period intervals; short periods represent upper crustal depths and longer periods lower crustal and upper mantle depths. The results show that at upper crustal depths (0.1 to 10 s) the geoelectric strike is not very stable and hence no dominant strike exists. The longer period data, however, reveal a very stable strike direction of about 60° from N to E, implying that deeper in the crust and in the upper mantle electric currents flow approximately in that direction. The figure contains more than 1200 strike estimates (42 sites x about 30 periods/site).

Rotation of the decomposed impedance tensor data (i.e. the data from which assumed distortions of near-surface 3-D inhomogeneities have been removed) into the direction of 60° from N to E reveals that at long periods the parallel (E-polarization) and

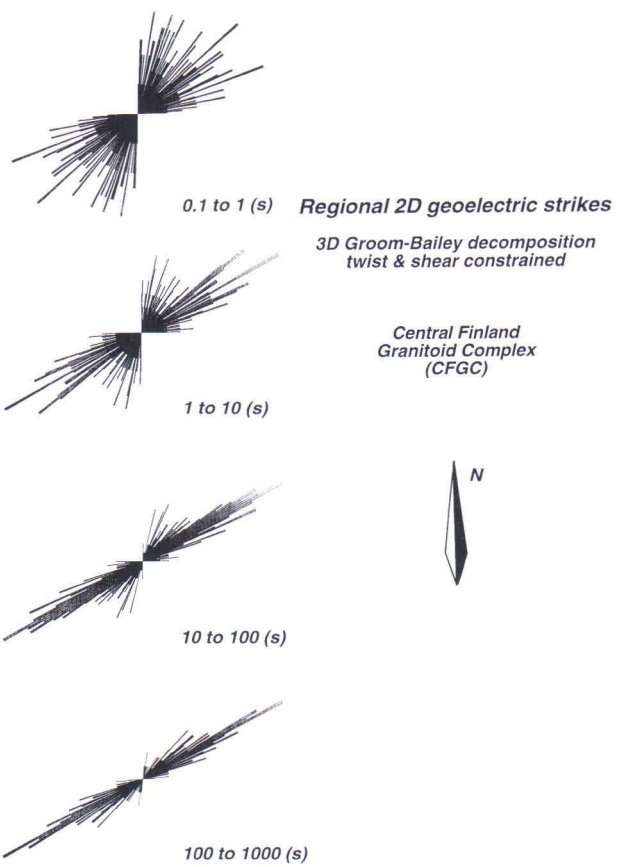

Fig. 10. Distribution of regional 2D geoelectric strike direction estimates at four period intervals for the magnetotelluric data of the CFGC. Estimates obtained by 3D Groom-Bailey decomposition of magnetotelluric impedance tensor data with constrained shear and twist. Geomagnetic north is in the top of the figure. Maximum of strike direction estimates at the period interval from 100 to 1000 s is obtained in the direction of 60 degrees.

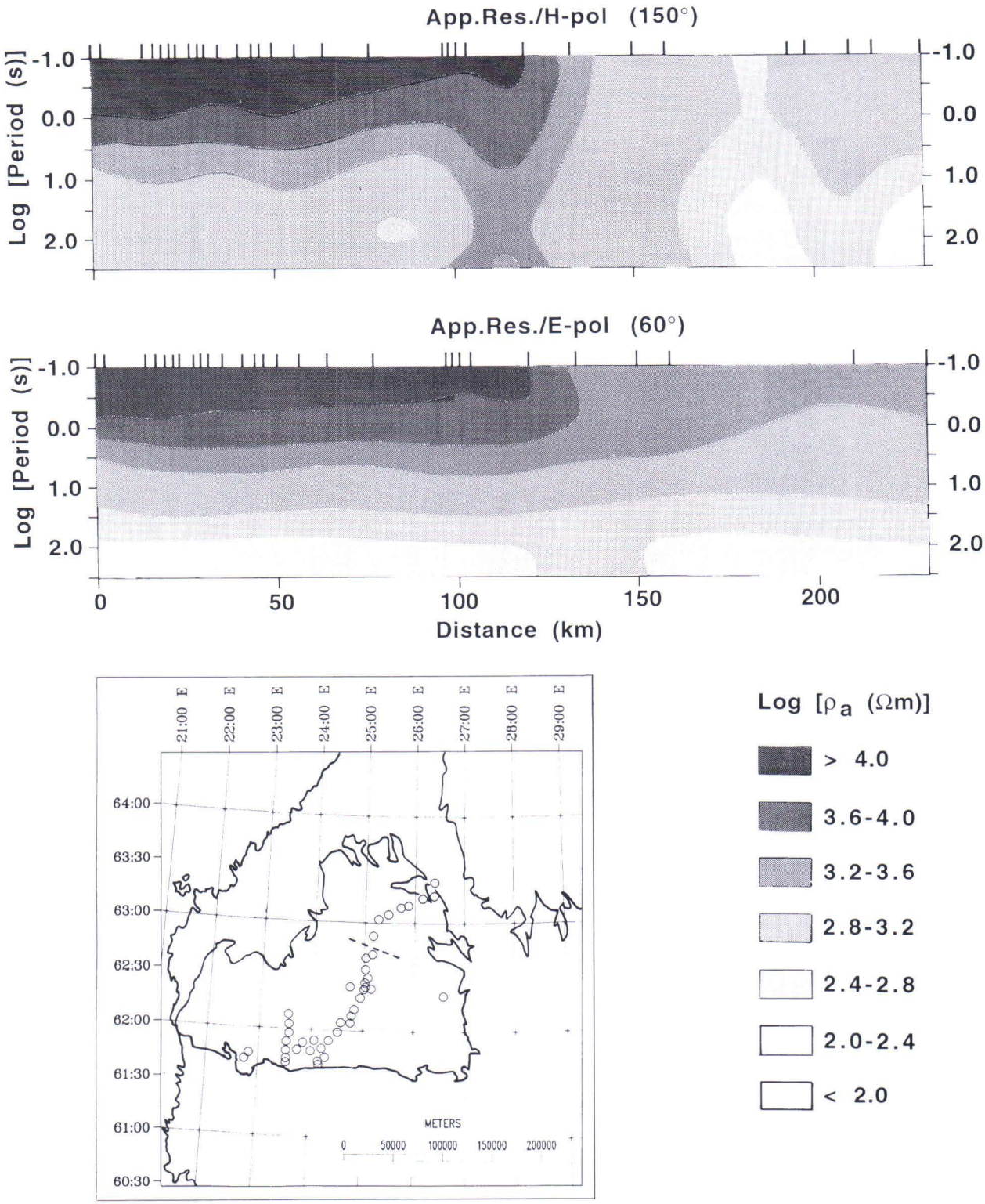

Fig. 11. Apparent resistivity pseudosections of regional 2D responses of magnetotelluric data from the CFGC. H-polarization data (upper panel) represent a case where the telluric line direction (150°) is perpendicular to the regional geoelectric strike and E-polarization data (lower panel) a case where the telluric line direction (60°) is parallel to the regional geoelectric strike. Vertical axis show the logarithm (lg) of the sounding period and the horizontal axis the distance from the southern end of the profile. Resistivity scale is logarithmic (lg). Vertical bars above the pseudosections denote the positions of sounding sites (open circles in the site map) projected onto the profile. Note the division of the isotropic upper crust into two distinct units and anisotropic nature of the lower crust and upper mantle. The location of the upper crustal conductivity boundary is shown by the dashed line in the site map.

perpendicular (H-polarization) responses 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 rotated apparent resistivities (H-pol/150° and E-pol/ 60°) along the NE-SW directed profile in the CFGC are shown as pseudosections in Figure 11.

The data show that the upper crust is isotropic and highly resistive in the CFGC, the first 10 to 14 km having resistivities of 35 000 to 10 000 Ωm. Resistivities then decrease dramatically from 3000 to 1000 Ωm in the middle crust (taken from the old 2D model shown in Fig. 8). Furthermore, the CFGC can be divided geoelectrically into two units, the southern unit having a more resistive upper crust than the northern one. The boundary between the two units is slightly south of the Saarijärvi-Karstula line in the middle of the CFGC (Fig. 11).

In contrast to the isotropic upper crust, the lower crust and upper mantle beneath the CFGC are electrically anisotropic: electrical resistivity is more than an order of magnitude lower in the direction of 60° than in the direction of 150°. The anisotropic splitting of phases begins approximately at periods of 5 s, which according to preliminary modelling corresponds to depths of about 30 km. No estimates of the thickness of the anisotropic part of the lithosphere are yet available.

There are several possible causes for anisotropy: the lower lithosphere may be genuinely anisotropic due to microscopic properties (electrical conductivity of minerals) or macroscopic properties (conductive and resistive "layers" or "dykes") or it may be attributed to side effects from surrounding conducting structures (structural anisotropy). 3-D modelling, however, is the only way to determine that the anisotropy is not due to inductive effects of conductors surrounding the CFGC.

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THE DRIFT OF FENNOSCANDIA IN THE LIGHT OF PALAEOMAGNETISM WITH EXAMPLES FROM THE FINNISH GGT/ SVEKA TRANSECT

by

L. J. Pesonen and S. Mertanen

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Key words (GeoRef Thesaurus, AGI): paleomagnetism, pole positions, SVEKA, block structures, dikes, continental drift, paleolatitude, Precambrian, Fennoscandia, Finland

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The main goal of the Finnish GGT-project is to develop an integrated model of the structure and evolution of the lithosphere in Finland (Kukkonen and Pesonen 1991). The GGT-profile (F-E, Fig. 1) crosses several larger intrusions and dyke swarms with ages ranging from Archaean to Jotnian, that are suitable for paleomagnetic studies. Figure 1 shows the paleomagnetic study targets in the Finnish GGT. One of the aims of these studies is to search for primary magnetizations and to detect magnetic overprints of various ages. Examples of post-Jotnian dykes magnetically overprinting the Jotnian sandstone in Satakunta were described in our previous GGT-article (Pesonen et al. 1992). Widespread Svecofennian overprints in Early Proterozoic and Archaean rocks have been described by Mertanen et al. (1989). Another aim of these studies is to measure past horizontal or vertical movements of the Fennoscandian shield or of the various units that constitute the shield. In this paper we show some examples of detecting local block movements within the Finnish GGT-profile. We also present a new paleolatitude curve for Fennoscandia and compare magnetic latitude data with paleoclimatic evidence.

Paleomagnetism and tectonic models

In order to detect relative movements between

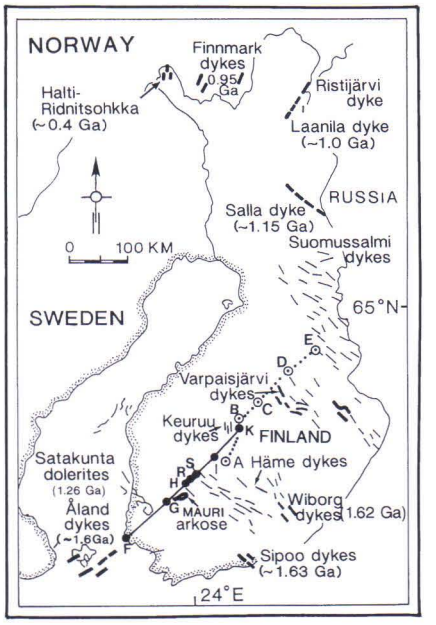

Fig. 1. The Finnish GGT/SVEKA transect (profile F to E) showing palaeomagnetic study sites.

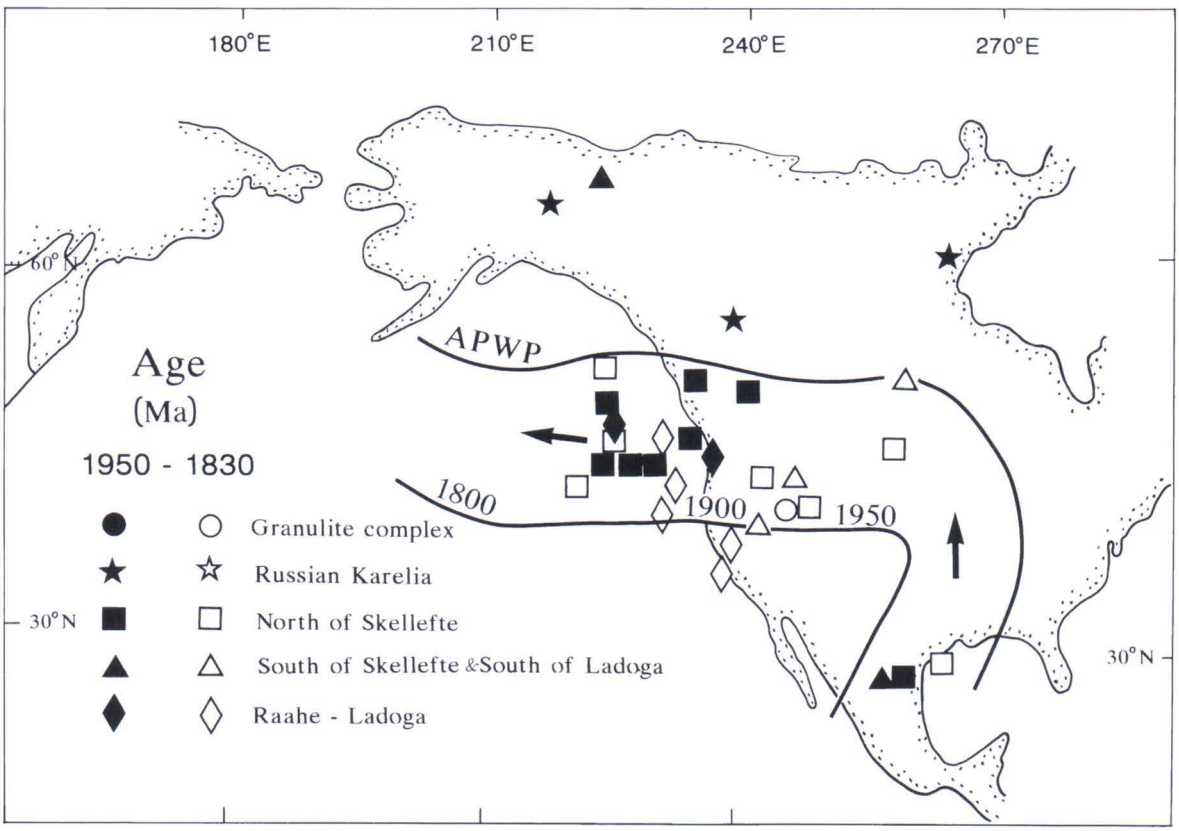

Fig. 2. Svecofennian APWP of Fennoscandia where good quality poles are plotted according to tectonic blocks as shown in the index. Data from Pesonen et al. (1991).

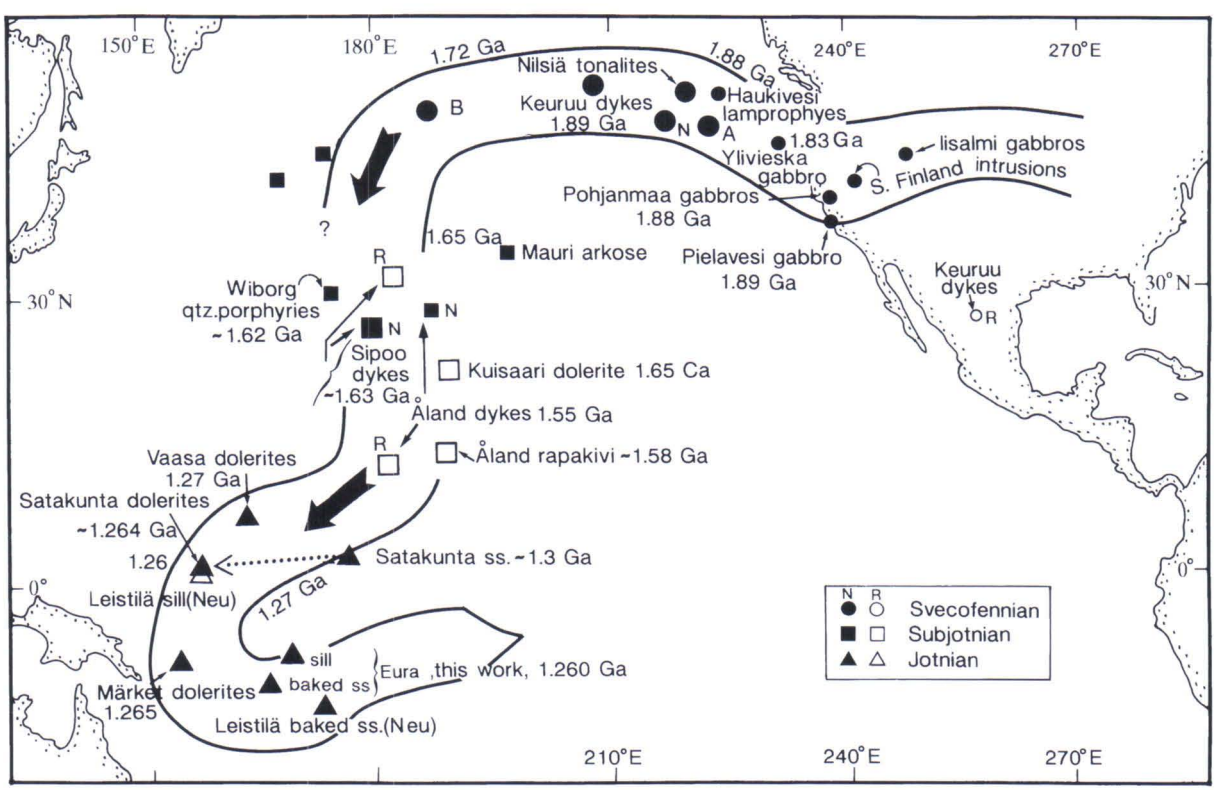

Fig. 3. Paleomagnetic poles from the GGT-transect and some other relevant poles from Fennoscandia plotted on the Fennoscandian APWP for the period 1.9 - 1.25 Ga. Closed (open) symbol denotes normal (reversed) polarity. New data are labelled. (Neu) = Neuvonen (1965, 1973), A (B), Varpaisjärvi component A (component B) from Figure 4.

tectonic blocks, high resolution paleomagnetic data from blocks with coeval magnetization ages are required. Figure 2 shows good quality paleopoles from Svecofennian units in Fennoscandia that have estimated magnetization ages of ~1.95 - 1.83 Ga (Pesonen et al. 1991).

We can see that there is a continuous apparent polar wander path (APWP) during 1.95 - 1.83 Ga involving poles from widely distant Svecofennian blocks and showing only a few deviating poles. These poles (shown as stars in Fig. 2) were obtained from sedimentary formations in Russian Karelia, and lie slightly to the north of the main Svecofennian APW track. These deviating poles may be due either to small, less than 2000 km, block movements between the southeastern and western Svecofennian domains or due to age difference of the poles. However, the poles may even be erroneous since they lack structural control and paleomagnetic tests. More paleomagnetic determinations with tilt and baked contact tests, coupled with precise datings of these Karelian formations, are needed to determine the tectonic origin of these aberrant poles.

Results from the GGT/SVEKA transect

Figure 3 shows the paleomagnetic poles from the GGT/SVEKA transect. We note here in passing that the poles for the Åland, Sipoo and Ahvenisto rapakivi massifs and associated dyke swarms (ages ca. 1.65 - 1.55 Ga) form part of the puzzling Subjotnian loop (see Pesonen et al. 1989) and that the poles from the Paleoproterozoic Keuruu dykes (~1.89 Ga), while fitting the APWP, still need to be investigated with respect to the polarity asymmetries, since the reversed polarity pole does not coincide with the normal pole (see Figure 3; Pesonen, 1987).

Jotnian

New paleomagnetic poles from the Satakunta Eura dolerites and baked sandstones of Jotnian age (~1.265 Ga) are plotted in Figure 3. A positive baked contact test has been obtained for the Eura sill (Kiperinjärvi) since the poles (Fig. 3) for this sill and for baked sandstones agree well within error limits, but disagree with the slightly older Satakunta sandstone pole cut by the sills (age ~1.3 Ga; Neuvonen 1973). The new results are based on NRM multicomponent analysis of these samples (see Pesonen et al. 1992), and observe that the new poles depart slightly towards the east (= towards younger ages) compared to the previous Satakunta dolerite pole (Fig. 3; Neuvonen 1965), for which no multicomponent analysis was done. This could be due to better resolution of the characteristic remanence with the multicomponent analysis, or due to the scarcity of the new determinations compared to previous ones. Another possibility is, as already inferred from Neuvonen´s previous poles (Neuvonen 1965; 1973), that the Eura sills, for which the baked contact test was performed (Leistilä baked sandstone), record a younger magnetization than the other Satakunta intrusions. U-Pb (Zr) ages from the Eura sill (Ämmänpelto, Kauttua) yield an age of 1.260 Ga which is slightly younger than most of the ages obtained from other Jotnian intrusions in Satakunta (1.258 - 1.264 Ga), Vaasa (1.268 - 1.273 Ga), and Åland (1.265 Ga) (Suominen 1991).

Mauri arkose

Twenty eight samples were measured from the Svecofennian Mauri arkose in the Tampere block (Fig. 1). The mean pole lies some 20° to the south of the Fennoscandian APWP (Fig. 3). If the remanence is primary, this could be attributed to structural tilting since the pole is uncorrected for tilt. However, if the magnetization turns out to be post-tilting (i.e., secondary), the departure suggests a late Svecofennian remagnetization age (~1.7 Ga).

Varpaisjärvi dykes

Samples were collected from Jatulian dykes intruding in the tectonically distinct Pällikäs and Jonsa blocks at Varpaisjärvi within the Archean Iisalmi block (Fig. 1) to test the block rotation hypothesis of Hölttä et al. (1992), involving a 20° anticlockwise rotation of the Jonsa block relative to the Pällikäs block (Fig. 4). This hypothesis is supported by two observations: (i) Archaean structural trends show an anticlockwise change of ~20° from Pällikäs to Jonsa, and (ii) the strikes of the dykes show a similar change in trend (Fig. 5). Previously, both Neuvonen et al. (1981) and Pesonen (1987) have tested the rotation hypothesis using paleomagnetic data of the Varpaisjärvi dykes which show that the NRM declinations of the two blocks do indeed depart by ~25°, suggesting that relative block rotation has taken place. However, the *sense* of rotation suggested by paleomagnetic declinations is clockwise, not anticlockwise!

Figure 4b shows new preliminary palaeomagnetic data for the dykes from the Jonsa and Pällikäs blocks based on multicomponent analysis. For comparison, the results by Neuvonen (1992) on the same dykes without multicomponent analysis are shown in Figure 4a. The main conclusions are as

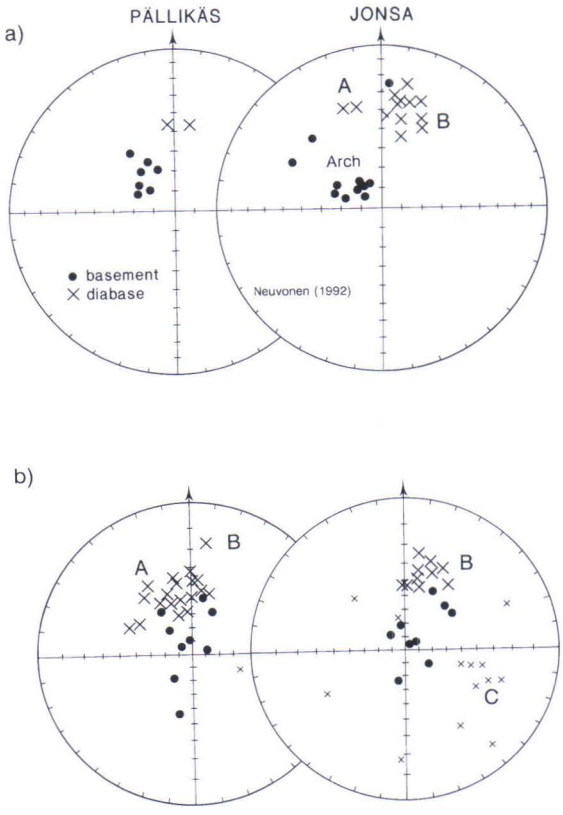

Fig. 4. Paleomagnetic data for the Varpaisjärvi dykes and Archaean basement rocks from the Pällikäs (left) and Jonsa (right) blocks. a) Data of Neuvonen (1992) without, and b) preliminary data of this work with multicomponent analysis. Components A (= Svecofennian), B (= Jatulian and/or late Svecofennian) and C are explained in text.

follows. Paleomagnetic directions from both studies show at least three main groups: (i) the Archean basement directions, (ii) the Pällikäs dyke directions (= component A), and (iii) the Jonsa dyke directions (= component B). The multicomponent analysis reveals more magnetization components (see for example component C, small crosses), but the origin of these less well defined components is still unclear. It is evident from Figure 4 that the directions of the two groups (A and B) partly overlap: some of the Jonsa (B) directions appear to plot onto the Pällikäs (A) group and vice versa. The difference between the two populations is nevertheless significant.

Based on the interpretation that the Pällikäs A direction is a typical Svecofennian direction (~1.9 Ga; see Fig. 3; e.g. Pesonen 1987; Mertanen et al. 1989) and on the observation that the Pällikäs dykes are more altered than the Jonsa dykes (Paavola, pers.comm. 1993), we propose a new model (Fig. 5) which can explain both the palaeomagnetic directions and the block rotation:

Fig. 5 (on the right). The rotation model for the Varpaisjärvi Jonsa and Pällikäs blocks. (a) At ~2.7 Ga structural trends in both the Jonsa and Pällikäs blocks are roughly NNE-SSW. Both blocks are affected by tensional fields which produce the NW-SE tensional fractures acting as precursors for Jatulian dykes. The basement blocks will be magnetized coherently by the ~2.7 Ga remagnetization (uplift and slow cooling). (b) The Jonsa block will be rotated by ~20°anticlockwise at ~2.68 Ga producing a corresponding deviation between the Archean structural trends and magnetizations as well as the tensional cracks. (c) The Jatulian (~2.2 - 2.0 Ga) dykes intrude both blocks but now with a ~20° difference in their strikes. They will be coherently magnetized with a NNE declination. (d) The dykes in the Pällikäs block will be later (~1.9 Ga) remagnetized (component A) by the Svecofennian orogeny, leaving only a few Jatulian relict magnetizations, while the Jonsa block, representing dry enderbitic basement, will escape more easily the Svecofennian overprint and preserves the Jatulian magnetization direction (B).

Model

a) During the Archaean the blocks were aligned in a NW-SE trending belt (Fig. 5a). The late Archean deformation and orogeny at ~2.7 Ga produced the NNE structural trends in and around the two blocks and created the NW-SE trending tension cracks, which will later (at ~2.2 Ga) become the conduits for the dolerite magma.

b) During post-orogenic cooling the cratons were egionally remagnetized at about 2.68 Ga ago (Fig. 5a). Slightly later, but still during the Late Archean, as a response to a change in regional stress, an anticlockwise rotation of ~20° of the Jonsa block with respect to Pällikäs block takes place (Fig. 5b). Unfortunately, this late Archaean rotation of 20° cannot be easily detected palaeomagnetically due to the steep inclinations (about 75 - 85°) in the basement rocks, which cause declinations to be widely scattered in both the Pällikäs and Jonsa blocks (Fig. 4). A paucity of tectonic activity then ensues until Jatulian time (2.2 Ga).

c) At ~2.2 Ga, during the Jatulian, the dykes intruded both blocks (Fig. 5c). The magma will follow the *pre-existing* fractures, resulting in the Jonsa dykes having E-W trends and the Pällikäs dykes having NW trends, respectively. The dykes will be magnetized at 2.2 Ga with a NNE declination and moderate steep inclination (direction B), as is still seen in the Jonsa block today (Fig. 4 and 5d).

d) The Pällikäs dykes also had this NNE-trending B magnetization originally, but they were almost totally overprinted by the Svecofennian orogeny (~1.9 Ga), which produced the NNW magnetization direction (component A) (Fig. 3 and 4). Some dykes in the Pällikäs block have survived the overprint and reveal the primary Jatulian remanence as relict, and conversely, a few of the Jonsa dykes have been remagnetized during the Svecofennian orogeny (Fig. 5d). The model thus predicts that the Pällikäs dykes, which are of coeval age and geochemically similar

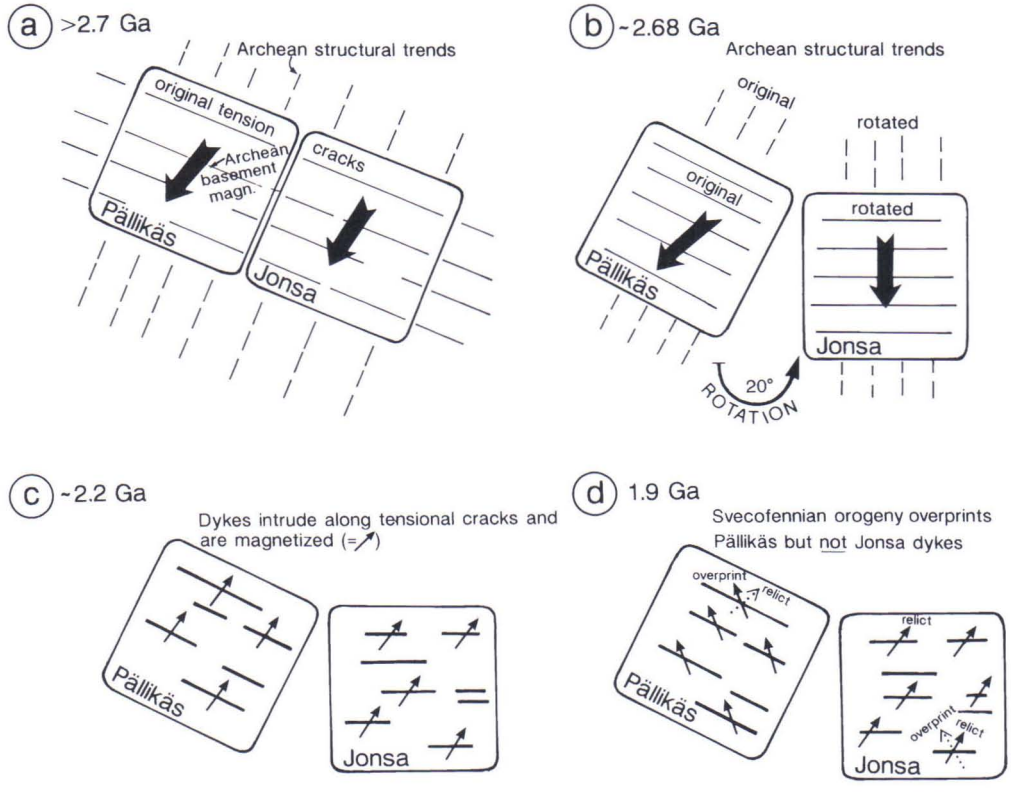

Fig. 5.

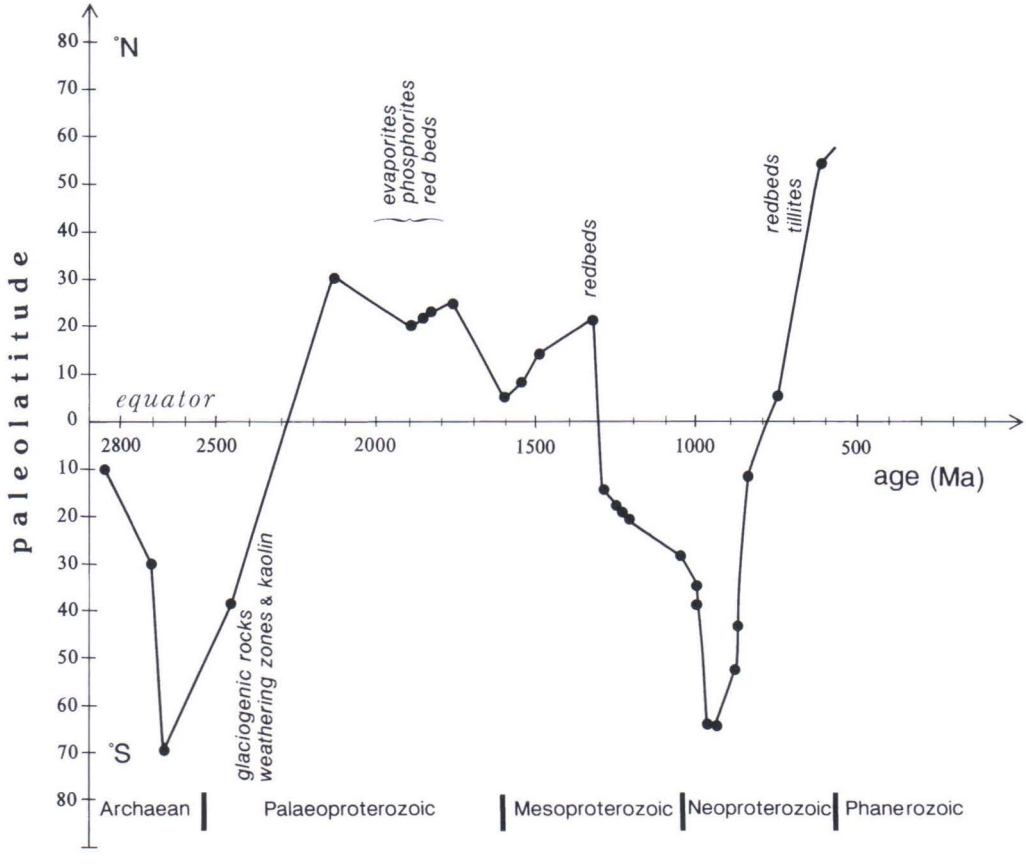

Fig. 6. Paleolatitude curve for Fennoscandia during 2.8 - 0.6 Ga ago, where the polarity adoption is such that Fennoscandia was in southern latitudes during the Archean-Early Proterozoic. Some palaeoclimatic indicators are also shown (Pesonen et al. 1989; Elming et al. 1993).

to the Jonsa dykes, should be more altered than the Jonsa dykes as evidenced by a thorough Svecofennian overprint direction. The model also predicts a ~20° difference in remanence direction for the Archean basement of the Jonsa and Pällikäs blocks, which could be discernible after more detailed sampling and multicomponent analysis of magnetization data. In testing this possibility, one has to remember that the Jonsa block represents a granulite facies area which may have been uplifted relatively more rapidly than the amphibolite grade Pällikäs block (see also Neuvonen 1992).

Fennoscandian drift

In the last example we have calculated the paleolatitude curve for whole Fennoscandia based on the best quality poles from the updated paleomagnetic database (Fig. 6; Pesonen et al. 1991).

Included in Figure 6 are some paleoclimatic indicators of latitude, such as evaporites, phosphorites, red beds and weathering zones which suggest low latitudes, and glaciogenic rocks (dropstones, tillites) which suggest more higher latitudes (Elming et al. 1993). Generally, palaeomagnetic data are consistent with paleoclimatic indicators but more data from precisely dated rocks are required to prove the palaeomagnetic latitude curve. A noticeable feature in the latitude curve is that during most of Precambrian time, Fennoscandia was apparently located at low to moderate latitudes (Pesonen et al. 1989).

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THE GGT/ SVEKA TRANSECT DIGITIZATION AND THE FORMAT FOR DIGITIZED INFORMATION

by

Petri Virransalo

Virransalo, P. 1996. The GGT/ SVEKA transect digitization and the format for digitized information. In: Ekdahl, E. and Autio, S. (eds.) Global Geoscience Transect/ SVEKA - Proceedings of the Kuopio Seminar, Finland 25. - 26.11.1993. Geologian tutkimuskeskus, Tutkimusraportti — *Geological Survey of Finland, Report of Investigation* 136, p. 55.

Key words (GeoRef Thesaurus, AGI): geotraverses, SVEKA, digitization, digital data, maps

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The publication formats of the GGT/ SVEKA transect are a program package DIGIMAP published by American Geophysical Union and a printed poster map. At present (15.11.1993) these programs are not available in Finland, but digitization guidelines and data file formats are described in Götze and Williams (1993). Digital data for transect consists of original geologic and geophysical data that is used to construct the maps and cross sections of the transect and graphical images of the maps and cross sections themselves. Data transfers to DIGIMAP are tested from ARC/ INFO and PC-ARC/ INFO (ESRI), AutoCAD (Autodesc), Designer (Micrografix) and Grid Master (Numonics) (op cit).

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DEEP STRUCTURE OF THE EARTH CRUST ALONG THE GGT/ SVEKA TRANSECT EXTENSION TO NORTHEAST

by

Jukka Yliniemi[1)], Jarkko Jokinen[2)] and Erkki Luukkonen[2)]

Yliniemi, J., Jokinen, J. & Luukkonen, E. 1996. Deep structure of the earth crust along the GGT/ SVEKA transect extension to northeast. In: Ekdahl, E. and Autio, S. (eds.) Global Geoscience Transect/ SVEKA - Proceedings of the Kuopio Seminar, Finland 25. - 26.11.1993. Geologian tutkimuskeskus, Tutkimusraportti — *Geological Survey of Finland, Report of Investigation* 136, p. 56.

Key words (GeoRef Thesaurus, AGI): crust, deep-seated structures, deep seismic sounding, geotraverses, SVEKA, Archean, Kuhmo, Finland

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The structure of the late Archaean domain, including the Kuhmo greenstone belt (KGB), the surrounding migmatites and granitoids in eastern Finland have been investigated by deep seismic sounding, in which mine tremors have been used (Lahnaslampi/ Finland and Kostamus/ Russia) as energy sources. This profile extends the SVEKA-transect by about 100 km further to the northeast. In the interpretation given here data obtained in 1981 from shot points along the SVEKA-transect have also been used. The measurements along the northeastern extension were carried out during the years 1984, 1985 and 1992.

The two-dimensional crustal model was calculated using raytracing techniques. It is immediately apparent from the reverse record sections that the structure of the crust to the northeast from the KGB is different from that to the southeast of the greenstone belt. The record section from the Kostamus shotpoint shows distinct reflected waves from the upper, midle and lower crust and the PmP-waves are very strong. In contrast the profile section to the southwest of the KGB shows clear reflected waves from the upper crust only.

There is also a positive Bouguer anomaly, with a notable increase over the KGB when we examine GGT-profile to northeast. According to petrophysical density measurements we know that the bedrock to the east of the KGB (2700 kg/m^3) has a density about 100 kg/m^3 higher than that of the bedrock to the west (2600 kg/m^3). The petrophysical density difference in the uppermost part of the crust can explain the Bouguer anomaly, but the overall increase might be attributed to the thickness variations within the late Archaean crust, as interpreted from the seismic data. Because either of these models fulfil the present isostatic condition or the seismic model (the gravity model was constructed using Birch's equation), the structure of the crust must therefore be more complicated.

The generalized seismic model, when coupled with the results of earlier structural and gravity studies from the area, have following features: (1) the KGB is 1.0 - 4.5 km thick; (2) the thickness of the crust decreases from 46 km beneath the KGB to 43 km at Kostamus, and (3) the thickness increases rapidly to 57 km to the southwest of the KGB and the data indicate the existence of a deep crustal reverse fault with a magnitude of displacement of about 10 km. The reverse is also perpendicular to the late Archaean fault movements in the KGB.

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RESULTS OF PRELIMINARY STUDIES OF THE CENTRAL FINLAND GRANITOID COMPLEX

by

R. Lahtinen, M. Nironen and P. Virransalo

Lahtinen, R. , Nironen, M. & Virransalo, P. 1996. Results of preliminary studies of the Central Finland Granitoid Complex. In: Ekdahl, E. and Autio, S. (eds.) Global Geoscience Transect/ SVEKA - Proceedings of the Kuopio Seminar, Finland 25. - 26.11.1993. Geologian tutkimuskeskus, Tutkimusraportti — *Geological Survey of Finland, Report of Investigation* 136, p. 57.

Key words (GeoRef Thesaurus, AGI): upper crust, granites, geophysical surveys, geochemical surveys, SVEKA, Proterozoic, Central Finland

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The Central Finland Granitoid Complex (CFGC) is a vast area consisting mainly of granites and granodiorites. The Sveka transect in the CFGC covers an area of about 20 000 km^2. About 40% of the transect area is covered by 1 : 100 000 bedrock maps and the rest of the area by 1 : 400 000 maps. High altitude airborne magnetic maps and gravity maps are also available. Some detailed studies on geology and economical potential have been done. This area was also sampled by the Rock Geochemistry Research Project (RGRP) with variable sampling density (1 sample/ 40 - 80 km^2).

The main objective of the first stage is to form a geological outline of the upper crust in the transect area by using geophysical, geological and geochemical data. The area for which only 1 : 400 000 bedrock maps were available were mapped at a scale of 1 : 50 000 with emphasis on areas showing magnetic and gravity anomalies. Some linear structures were studied in detail. Sampling for petrophysical measurements covers the whole study area. Samples for the geochemical purposes (128) were taken to supplement the RGRP sampling, especially from rocks responsible for geophysical anomalies. Field observations are stored in the KALPEA data base.

The field work shows that magnetic highs are mainly caused by mafic intrusives, subvolcanic rocks and some granites with abundant magnetite. Regional magnetic lows seem to coincide with granite areas. The mafic rocks formerly considered as diorites were found to be rich in quartz and K-feldspar, i.e. monzonitic. At least some linear belts formerly interpreted as quartz-feldspar rocks are mylonitized granites and the belts thus represent large fault structures.

A compilation map has been prepared from the field and geophysical data using also geochemical classifications. At the second stage the knowledge of the upper crust will be integrated with deep seismic, gravity and magnetotelluric studies to model the structure of lithosphere.

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PETROLOGY OF THE MANTLE SECTION OF THE JORMUA OPHIOLITE, NE FINLAND

by

Asko Kontinen[1)] and Petri Peltonen[2)]

Kontinen, A. & Peltonen, P. 1996. Petrology of the mantle section of the Jormua Ophiolite, NE Finland. In: Ekdahl, E. and Autio, S. (eds.) Global Geoscience Transect/ SVEKA - Proceedings of the Kuopio Seminar, Finland 25. - 26.11.1993. Geologian tutkimuskeskus, Tutkimusraportti – *Geological Survey of Finland, Report of Investigation* 136, 58 - 59.

Key words (GeoRef Thesaurus, AGI): ophiolite, upper mantle, peridotites, gabbros, clinopyroxenite, dikes, cumulates, Proterozoic, Jormua, Finland

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Introduction

Ophiolites, in addition to mantle xenoliths and "orogenic lherzolites", permit the direct study of the petrology, geochemistry, and physical properties of the upper mantle. On-land fragments of ophiolitic mantle are invariably serpentinised but, if carefully screened for alteration, may provide more representative mantle samples than xenoliths.

At present, only three early Proterozoic or older rock complexes are known that comprehensively satisfy the definition of ophiolite, and only one of these - the 1.95 Ga old Jormua Ophiolite Complex (JOC) in the Kainuu Schist belt - exposes its mantle section as well (Kontinen 1987).

Structure and composition of the mantle section

The mantle section of the Jormua Ophiolite is the largest (30 km^2) and best exposed component of this ophiolite, being preserved as several large fault-bounded blocks that are spatially associated with metagabbros, a sheeted mafic dyke complex, and basaltic metalavas (Fig. 1). The mantle section is composed of serpentinised and regionally metamorphosed peridotites (depleted lherzolites-harzburgites, some dunitic cumulates) cross-cut by various ultramafic to mafic intrusive phases. No primary minerals, except for some chromite relicts are preserved in the serpentinites. However, traces of mantle tectonic foliations and lineations are still visible, being defined by seams and streaks of altered chromite. Stable metamorphic mineral parageneses of the serpentinites formed during the peak regional metamorphism include antigorite-olivine and antigorite-olivine-tremolite, which yield metamorphic peak temperature estimates of 480°C and 530°C with pressures of 2 kb and 5 kb respectively.

The mantle lherzolites of the Jormua Ophiolite have relatively uniform compatible element but variable incompatible element abundances. Such heterogeneity is only partly explained by their variable depletion. In addition, chemical data indicates that Jormua lherzolites have been infiltrated and metasomatised by basanitic-nephelinitic melt, which is also represented by rare OIB-like dykes enclosed by serpentinites.

Considerable variation occurs in the composition and morphology of the intrusive phases crosscutting the mantle peridotites between each of the various fault-bonded blocks of the mantle section. The Lehmivaara block is particularly rich in metabasaltic dykes, the Antinmäki and the Kannas blocks in irregular-shaped gabbro lenses and pods, and the Hannusranta block in clinopyroxenitic dykes and

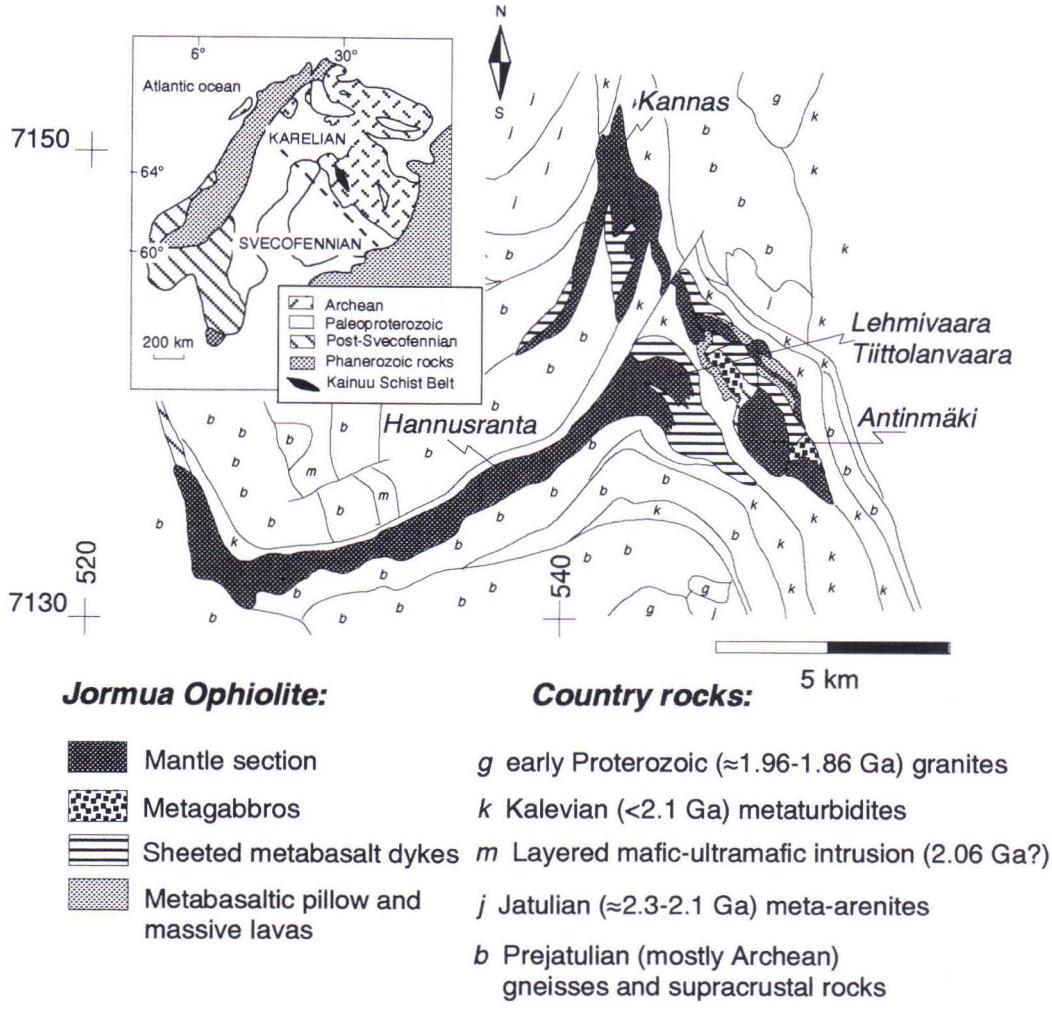

Fig.1. Generalized geological map of the Jormua Ophiolite (modified after Kontinen, 1987). Major geological features of the Fennoscandian Shield are shown in the inset map.

pods (Fig. 1). Whole rock composition and mineral textures of the clinopyroxenites and gabbros imply that they cannot represent melt compositions but are crystal cumulates ("dyke cumulates" of Harte et al. 1993). Gabbroic and clinopyroxenitic dyke cumulates are expressions of the upward transportation of MORB- and OIB-like melts, respectively, from the mantle. The gabbros resemble Jormua dykes and lavas. Within some of the gabbroic pods differentiation proceeded to extreme iron enrichment as expressed by voluminous ilmenite precipitation. Such pods represent magmas which evolved from primary MORB-like melts within the uppermost mantle but failed to reach the petrological Moho discontinuity (Peltonen et al. 1995). Clinopyroxenitic dyke cumulates in turn show evidence of emplacement into peridotites undergoing deformation, and crystallization at relatively high confining pressure. Extrusive rocks, coeval with these cumulates, are not present within the crustal unit of the Jormua Ophiolite. We believe that the clinopyroxenitic dyke cumulates intruding the mantle peridotites within the Hannusranta block are expressions of continental break-up related magmatism that slightly predated the formation of the new oceanic basin.

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KYANITE-CHLORITE-CORDIERITE ROCK - A PALAEOPROTEROZOIC OVERPRINT ON ARCHAEAN TONALITIC GNEISSES

by

Matti Pajunen

Pajunen, M. 1996. Kyanite-chlorite-cordierite rock - a Palaeoproterozoic overprint on Archaean tonalitic gneisses. In: Ekdahl, E. and Autio, S. (eds.) Global Geoscience Transect/ SVEKA - Proceedings of the Kuopio Seminar, Finland 25. - 26.11.1993. Geologian tutkimuskeskus, Tutkimusraportti — *Geological Survey of Finland, Report of Investigation* 136, p. 60.

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Several occurrences of distinctive coarse-grained kyanite-chlorite-cordierite rock have been found in the Archaean Nurmes-Sotkamo area of eastern Finland. These rocks occur in NW-SE- and NE-SW-trending fractures, faults and shear zones that cut across Archaean structures and metamorphic assemblages. In these zones intense metasomatism has altered the primary tonalitic gneisses to heterogeneous kyanite-chlorite-cordierite rocks. Metasomatic zonation was generated by concurrent leaching and accumulation of rock-forming components in different parts of the zones. Metasomatism occurred under prograde metamorphism as shown by mineral reactions proceeding from chlorite-muscovite to kyanite-staurolite and finally to cordierite-sillimanite stability. According to these mineral parageneses the metamorphic grade increases southwards, reflected in more widespread stabilization of sllimanite, whereas in the northern ocurrences, kyanite and/or andalusite crystallized instead. In the vicinity of the Kainuu schist belt the foliation is more pronounced than in the highly metamorphosed southern targets, where inoriented crystals reflect static crystallization. In south the metamorphic peak was attained at about 600 - 620°C/4 - 5kbar. The peak of cordierite-sillimanite metamorphism and metasomatism is dated from cogenetic xenotime as 1852±2 Ma old. The kyanite-chlorite-cordierite rocks provide evidence for the presence of fluids that were capable of changing rock compositions and for prograde amphibolite facies conditions in the zones of most intensive fluid flow. The metamorphic-metasomatic succession records the P-T evolution of the Archaean crust during Palaeoproterozoic era and indicates that the present erosion surface was deeply buried beneath Archaean±Palaeoproterozoic crust at about 1850 Ma.

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SEISMIC STUDIES ON LITHOSPHERIC STRUCTURE ALONG THE GGT/ SVEKA TRANSECT - A REVIEW

by

Urmas Luosto

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Key words (GeoRef Thesaurus, AGI): crust, deep seismic sounding, velocity structure, geotraverses, SVEKA, Precambrian, Finland

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The first deep seismic sounding measurements along the GGT/SVEKA transect were made in 1981 between Keuruu and Kuhmo. Interpretation of this data set showed that the crust in the central part of the profile is unexpectedly thick, about 60 km, with an average velocity down to the Moho of about 6.6 km/s (Luosto et al. 1984, Grad & Luosto 1987). In the upper crust a low velocity layer was found and in the NE part of the profile a high velocity body. Fracturing of the uppermost crust was evaluated by a detailed study of P- and S-wave velocities (Grad & Luosto, 1993). The uppermost 200 meters is the most fractured part of the crust and shows no significant differences between the southern and northern parts of the profile. On the other hand it was found that at depths of 0.2 - 1 km in the Archaean part of the profile the fracturing is twice as dense as at equivalent levels in the Proterozoic crust. The results suggest an isolated saturation of cracks in the uppermost crust beneath this part of the GGT/SVEKA transect.

In 1991 the next international deep seismic sounding experiment in Finland was carried out on a line from Kustavi to Kannonkoski, southwest of the "original" seismic SVEKA profile. According to a preliminary interpretation the crust in the SW part is nearly 50 km thick but increases steadily towards the northeast, being about 60 km in the NE part of the profile (Luosto et al., 1996). A low velocity layer was found between 6 and 13 km. Intensive reflections were observed from inclined reflectors in the depth range 10 - 20 km, roughly beneath the Satakunta sandstone formation, Tampere schist belt and Keuruu gabbro intrusions.

The extensive marine seismic reflection survey BABEL (Baltic and Bothnian Echoes from the Lithosphere) was carried out in 1989 (BABEL Working Group 1993). The profile line 7 from Uusikaupunki to Gävle also connects the GGT/ SVEKA transect to Sweden. In addition to reflection recordings on the ship, wide angle reflections were recorded on land, too. This made it possible to integrate interpretations of both reflection and refraction data. The crustal thickness varies between 45 and 50 km, the smallest values being found below the Ahvenanmaa rapakivi intrusion. According to the wide angle data there is no low velocity layer in the upper crust. Reflection data show several intensive reflectors dipping eastwards, coinciding with refraction boundaries in the deeper more horizontal parts of the reflectors. These phenomena were interpeted to indicate an episode of crustal extension (Korja & Heikkinen 1994).

Crustal structure in the northern part of the

geotraverse was studied using recordings from the NE part of GGT/ SVEKA transect together with recordings of quarry blasts at Kostamuksha (Yliniemi et al. 1993). According to these investigations the Moho boundary rises steeply to the northeast roughly below the Kainuu and Kuhmo schist belts where the crust decreases in thickness by about 10 km.

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