INSTITUTE OF SEISMOLOGY
UNIVERSITY OF HELSINKI

REPORT S-41

LITHOSPHERE 2000

A SYMPOSIUM ON THE STRUCTURE, COMPOSITION AND EVOLUTION OF THE LITHOSPHERE IN FINLAND

Geological Survey of Finland, Auditorium
Espoo, Otaniemi, October 4-5, 2000

PROGRAMME AND EXTENDED ABSTRACTS

edited by Lauri J. Pesonen, Annakaisa Korja and Sven-Erik Hjelt

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2000

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LITHOSPHERE 2000

A SYMPOSIUM ON
THE STRUCTURE, COMPOSITION AND EVOLUTION OF THE LITHOSPHERE IN FINLAND

PROGRAMME AND EXTENDED ABSTRACTS

The symposium will take place on October 4-5, 2000
in the Auditorium of the Geological Survey of Finland, Espoo, Finland.

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Finnish National Committee of the International Lithosphere Programme (ILP)
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ORGANIZING COMMITTEE AND EDITORS:

Lauri J. Pesonen	Department of Geophysics P.O. Box 4, FIN-00014 University of Helsinki Finland E-mail: lauri.pesonen@helsinki.fi http://www.geophysics.helsinki.fi
Annakaisa Korja	Institute of Seismology P.O. Box 26 FIN-00014 University of Helsinki Finland E-mail: annakaisa.korja@seismo.helsinki.fi http://www.seismo.helsinki.fi
Sven-Erik Hjelt	Department of Geophysics Institute of Geosciences P.O. Box 3000 FIN-90014 University of Oulu Finland E-mail: seh@babel.oulu.fi sven-erik.hjelt@oulu.fi http://www.gf.oulu.fi

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PROGRAMME

Wednesday, October 4, 2000, 9.00-18.00

9.00-10.00	Registration at the Geological Survey of Finland, Espoo, Otaniemi
10.00-11.00	***Opening Session*** ***Chair J. Kakkuri***
10.00-10.05	Opening of the Symposia by Chairman of the National Committee of the ILP *Juhani Kakkuri* Welcoming words by *Heikki Vartiainen (Ministy of Trade and Industry)*
10.05-10.55	The Lithosphere in Finland *(Ilmari Haapala)*
10.55-11.20	***Coffee Break***
11.20-13.00	***The Lithosphere Under the Fennoscandian Shied, Part I*** ***Chair K. Korsman***
11.20-11.45	The Structure of the Crust and Upper Mantle in Fennoscandia as Imaged by Electromagnetic Waves (*Toivo Korja and the BEAR Working Group*)
11.45-12.10	Review of Some Features of the Seismic Velocity Models in Finland *(Pekka Heikkinen and Urmas Luosto)*
12.10-12.35	Deep Seismic Tomography of the Crust and Lithosphere-Asthenosphere System in the Fennoscandian Shield Area – Pre-studies and First Results *(Tellervo Hyvönen, Elena Kozlovskaya and the SVEKALAPKO Seismic Tomography Working Group)*
12.35-13.00	Determination of Crustal Gravity Field of the Fennoscandian Shield Using Deep Seismic Sounding Interpretations *(Juhani Kakkuri and Zhitong T. Wang)*
13.00-14.00	***Lunch Break***
14.00-16.10	***The Lithosphere Under the Fennoscandian Shied , Part II*** ***Chair S.-E. Hjelt***
14.00-14.25	Gravity Anomalies in Finland Due to Crustal and Upper Mantle Sources *(Seppo Elo)*
14.25-14.50	Thermal State of the Lithosphere in the Central Part of the Fennoscandian Shield *(Ilmo T. Kukkonen and Petri Peltonen)*
14.50-15.15	Kimberlites, Carbonatites and Their Mantle Sample: Constraints for the Origin and Temporal Evolution of the Lithospheric Mantle in Fennoscandia *(Petri Peltonen, Hugh O´Brien, Juha Karhu and Ilmo T. Kukkonen)*
15.15-15.40	Rheological Structure and Numerical Geodynamical Modelling in the Fennoscandian Shield *(Kari Moisio and Pertti Kaikkonen)*
15.40-16.10	***Coffee break***

16.10-18.30 ***Poster Session***
Chair P. Heikkinen

16.10-16.20 Poster presentations à 2 min
16.20-18.00 Posters with refreshments

P1. Stanislav S. Gornostayev, Eero Hanski, Kauko V.O. Laajoki and Sergiy E. Popovchenko. Chromitite-bearing Ultramafic Rocks in the Ukrainian Shield: New Evidence for Paleoproterozoic Ophiolitic Mantle Rocks

P2. Eero Hanski, Jouni Vuollo, Jani Wennerstrand and Stanislav Gornostayev. Ultramafic Rocks in the Nunnanlahti Greenstone Belt, Eastern Finland: a Potential Example of Archaean Ophiolitic Rocks

P3. Pentti Hölttä and Markku Väisänen. Tectonic evolution of the Central Lapland Greenstone Belt –metamorphic and structural observations

P4. Annakaisa Korja and Pekka Heikkinen. The Accretionary Svecofennian Orogen - in the Light of Seismic BABEL Lines

P5. Jouni Vuollo, Heikki Salmirinne, Lauri J. Pesonen, Vladimir Stepanov, George Fedotov and Dimitri Frank-Kamenetsky. The Eastern Fennoscandian Mafic Dyke Swarms GIS-Database

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Thursday, October 5, 2000, 08.30 - 18.00

8.30-11.45 ***Crustal Studies, Part I***
Chair I.T. Kukkonen

8.30-8.55 Magnetic Signatures of the Finnish Bedrock and Their Relationship with Geological Boundaries *(Meri-Liisa Airo and Tapio Ruotoistenmäki)*

8.55-9.20 Fennoscandian Crustal Model as Reflected by Petrophysical Interpretation of Potential Field Anomalies *(Juha V. Korhonen and Heikki Säävuori)*

9.20-9.45 Granite Emplacement During the 1.83 Ga Late-orogenic Stages in Southern Finland *(Tom Stålfors and Carl Ehlers)*

9.45-10.10 Paleoproterozoic Trondhjemite Migmatites in Southern Finland *(Tapio J. Koistinen, Kalevi Korsman and Petri Virransalo)*

10.10-10.35 Extensive Zone of Mafic-Felsic Magma Interaction in the Svecofennian: the Hyvinkää-Mäntsälä Gabbroic Belt, Southern Finland *(Toni Eerola, Riku Raitala, Jyrki Bergström, Taina Eloranta, Niilo Kärkkäinen, and Ragnar Törnroos)*

10.35-11.00 ***Coffee break***

11.00-12.40 ***Crustal Studies, Part II***
Chair A. Korja

11.00-11.25 Granulite Metamorphism and Formation of the Lower Crust in Finland *(Kalevi Korsman, Pentti Hölttä and Annakaisa Korja)*

11.10-12.15 Mafic Dyke Swarms - Geological Evolution of the Palaeoproterozoic in the Fennoscandian Shield *(Jouni Vuollo, Hannu Huhma and Lauri J. Pesonen)*

12.15-12.40 Meteorite Impact Cratering - Implications for the Fennoscandian Lithosphere *(Lauri J. Pesonen, Andreas Abels, Martti Lehtinen and Jüri Plado)*

12.40-13.30 ***Lunch Break***

13.30- 15.30	***The Evolution of the Fennoscandian Shield: Observations and Models***	
	Chair H. Huhma	
13.30-13.55	Evolution of Proterozoic Surface Environments: Evidence from Carbon and Strontium Isotope Ratios in Sedimentary Carbonates *(Juha A. Karhu)*	
13.55-14.20	Crustal Boundaries of East European Craton – Keys to Proterozoic Amalgamation *(Mikko Nironen, Annakaisa Korja, Raimo Lahtinen and Pekka Tuisku)*	
14.20-14.45	Evolution of the Fennoscandian Lithosphere in the Mid-Proterozoic: the Rapakivi Magmatism *(O. Tapani Rämö, Annakaisa Korja, Ilmari Haapala, Olav Eklund, Sören Fröjdö and Matti Vaasjoki)*	
14.45-15.00	***Coffee Break***	
15.00-15.25	Fennoscandia as a Part of a Proterozoic Supercontinent *(Satu Mertanen and Lauri J. Pesonen)*	
15.25-15.50	Geological Evolution of the Central Lapland Greenstone Belt *(Eero Hanski and Hannu Huhma)*	
15.50-16.40	***The Ongoing and New Lithosphere Projects and the Future of Lithosphere Research in Finland***	
	Chair P. Nurmi	
15.50-16.00	GGT	*(K. Korsman)*
15.50-16.00	EUROPROBE/Svekalapko	*(S-E. Hjelt)*
16.00-16.10	ESF/GEODE	*(H. Papunen)*
16.10-16.20	ESF/Impact	*(L.J. Pesonen)*
16.20-16.30	New projects 2000 -	*(S.-E. Hjelt)*
16.30-16.40	***Break***	
16.40-18.00	***Final Session and General Discussion***	
	Chair L. J. Pesonen	
16.40-16.45	Summary of the Lithosphere Studies	*(I.T. Kukkonen)*
16.45-16.50	Summary of the Crustal Studies	*(R. Lahtinen)*
16.50-16.55	Summary of the Studies of the Evolution of the Fennoscandian Shield	*(O.T. Rämö)*
17.00-18.00	Final Discussion	
18.00	Concluding Remarks	*(L.J. Pesonen)*

PART I: INVITED PAPERS

The Lithosphere in Finland

Ilmari Haapala

Department of Geology, P.O. Box 11, FIN-00014 University of Helsinki, Finland
E-mail: ilmari.haapala@helsinki.fi

The knowledge about the structure and evolution of the continental crust in Finland increased drastically during the last quarter of the 20th century, and the very last years brought an insight also into the structure and composition of the upper, lithospheric mantle. This advancement is based largely on systematic long-term basic research done at various institutions (Geological Survey of Finland, Finnish Geodetic Institute, Finnish universities). New findings, such as the discovery of diamond-bearing kimberlites in eastern Finland in the 1980's, have activated the scientific research of the lithosphere. This paper gives a short review of the geological, geophysical and geochemical research in Finland on the structure, composition and evolution of the lithosphere.

Keywords: Finland, Archean, Paleoproterozoic, lithosphere, crust

1. The Research Institutions

The *Geological Survey of Finland* and its predecessors have mapped and studied the bedrock of Finland since 1865. The geological mapping in scale 1:400,000 was completed in 1980, and over 50% of the country is already covered by maps with a scale of 1:100,000. The first geological map over Finland (scale 1:2,5 million) was published by *J.J. Sederholm in 1897.* The currently published thematic maps in 1:1 million scale synthesize the lithologic, structural, geophysical and ore-geologic knowledge of the country. Airborne geophysical mapping with magnetic, electromagnetic and radiometric measurements from an altitude of 150 m with a line spacing of 400 m was started in 1951 and completed in 1972; it was followed by more detailed mapping from a height of 30-50 m with a line spacing 200 m *(Peltoniemi,1982).*

Isotopic age determination and isotopic petrogenetic studies have been carried out at the Geological Survey since the 1960's, and they have been of invaluable help in age classification of the bedrock units and in developing petrogenetic models for the crustal evolution (e.g., *Huhma, 1986*). The paleomagnetic studies made at the Geological Survey and at the University of Turku have produced abundant new information of the latitudinal drift and rotation of Fennoscandia from the Archean to the present (e.g. *Mertanen and Pesonen, 1997*). Very significant integrated geological and geophysical work has been done along the Global Geoscience Transect (GGT) SVEKA-profile, where the different Finnish institutions worked as a team and Kalevi Korsman of the Geological Survey of Finland acted as the driving force (*Korsman et al., 1999*).

Since its foundation in 1918, the *Finnish Geodetic Institute* has made the first-order triangulation over Finland. The loops between the triangulation chains have been filled in later with trilateration measurements. These measurements have formed the basis for atlas of Finland. Accurate geodetic measurements enable also sophisticated studies on the recent crustal deformation in Finland (e.g. *Kakkuri and Chen, 1992*). Since 1945, the Geodetic Institute has made systematic gravimetric measurements over the country. First gravimetric anomaly maps of Finland were published in 1962 (*Honkasalo, 1962*), more accurate ones, based on 22500 observation points (with an average station separation of 5 km) were published by *Kiviniemi (1980).* This grid provides valuable information on gravity field over Finland and delineates major crustal structures. Later, a more dense gravity data set was used for geological mapping and raw material exploration (*Elo, 1997*). Further, the Geodetic

Institute has also produced land uplift maps, based on mareographic observations on the coast and repeated precise levellings in inland areas (*Kakkuri, 1993*).

The *Institute of Seismology of the University of Helsinki*, has conducted deep seismic soundings in Finland and adjacent areas since 1960's as international collaboration projects. The data collected from the recent refraction and wide-angle reflection profiles have been used to calculate contour maps which show the thickness of the crust (the depth of the Moho) and the main crustal layers (*Luosto, 1991 and 1997; Korja et al., 1993*).

The researchers at the *Department of Geophysics of the University of Oulu*, have studied the electric conductivity of the lithosphere by applying various electromagnetic methods. Together with the airborne surveys of the Geological Survey these studies allow to focus from shield-scale structures to small local conductivity structures. The results have been summarized by *Hjelt and Korja (1993)* and *Korja (1997)*.

The geological departments of the *Universities of Helsinki, Turku, Oulu and Åbo Akademi* have all contributed to the research of the lithosphere by their research projects, dealing with the various aspects of tectonic, petrologic and geochemical evolution of the crust.

2. The Structure of the Lithosphere

The main superficial structural features of the lithosphere of Finland and surrounding Fennoscandia (Fig. 1) are known from systematic geologic mapping. The deep seismic sounding profiles and maps showing the thickness of the crust and its main crustal layers have opened new views for understanding the structure and evolution of the lithosphere in Finland. Two observations are of special interest: (i) the anomalously thick crust (55-64 km) in eastern and central Finland, and (ii) the thin (normal) crust (down to 40 km) around the Bothnian Bay and, especially, in areas of rapakivi granite batholiths (*Luosto, 1991 and 1997; Korja et al., 1993*). Most of the variations in the thickness of the crust are related to corresponding variations in the thickness of the lower crustal layer. Comparison with gravity data shows that the variations in the crustal thickness are mostly compensated by density variations within the crust (*Korsman et al., 1999*). In regions of the thinned crust, the lower crust is characterized by strong subhorizontal reflectivity and a sharp Moho boundary, whereas in regions of the thicker crust, the reflectivity decreases strongly in the lowermost crust. All this suggests that in areas of an anomalously thick crust the lowermost crustal layer consists of very high-density material.

The up-bulging transitional zone between the two Moho boundaries (M1 and M2; Fig. 2) below the rapakivi granite batholiths represent a mafic underplate (a composite mass of mantle peridotite and of intruded mafic magma) related to the mantle upwelling in an extensional environment under a large cratonized continent (*Haapala, 1989; Haapala and Rämö, 1992; Korja et al., 1993; Korja, 1995; Rämö and Haapala, 1996; Korsman et. al, 1999*). The thinning of the lower crust is caused by extensive partial melting of the deep crust and transit of the granitic magmas to higher levels, and to the extension of the ductile lower crust (Fig. 2).

The knowledge of the structure and composition of the lithospheric mantle in Finland is still fragmentary. Some information can be obtained by geophysical (seismic, geoelectromagnetic, gravimetric, geothermal) studies, from the chemical and isotopic composition of the mantle-derived dike rocks, and directly from the mantle xenoliths. From the dispersion of seismic surface waves *Calcagnile (1982)* estimated that in the central parts of the Fennoscandian Shield, including most of Finland, the thickness of the lithosphere is more than 170 km (Fig. 1). In the global geoelectromagnetic data, the asthenosphere is normally visible as a conductive zone. The available data from the Fennoscandian Shield indicate that in the central and southwestern parts of the shield, the asthenospheric layer is missing or is electrically weak, but in the peripheral areas the conducting asthenosphere layer appears more clearly and is closer to the surface, in the north at a depth of 100 to 150 km (*Hjelt and Korja, 1993; Korja, 1997*).

Obviously the most reliable information on the composition, structure and pressure – temperature conditions of the upper mantle is provided by the kimberlite-hosted mantle xenoliths found recently in eastern Finland, around the Archean – Proterozoic boundary (*Peltonen et al., 1999*). On the basis of geothermometric and geobarometric studies of the ultramafic mantle xenoliths, *Kukkonen and Peltonen (1999)* and *Peltonen (2000)* concluded that the xenoliths were trapped into the uprushing kimberlite melt at the depths of 94 to 230 km and at the temperatures of 828 to 1356°C. No textures indicating partial melting have been detected in the xenoliths. Thus the xenoliths represent petrographically solid lithospheric mantle, and suggest that the thickness of the lithosphere is in eastern Finland anomalously high, more than 230 km. The kimberlite contains also small diamond-bearing eclogitic xenoliths that represent small mafic magma bodies in the ultramafic lithospheric mantle. Isotopic studies indicate that the upper part of the lithospheric mantle below the Archean crust is about 3.2 to 2.6 Ga and below the Svecofennian crust about 1.9 Ga in age. This is in agreement with the model that subcrustal mantle was stabilized (and depleted) at the same time as the orogenic crust above it. There are indications that below a depth of about 150 km the lithospheric mantle was stabilized at about 1.8 Ga under both the Archean and the Proterozoic terranes *(Peltonen, 2000)*. When reaching the asthenosphere, the isotopic geochronometers must approach the zero line.

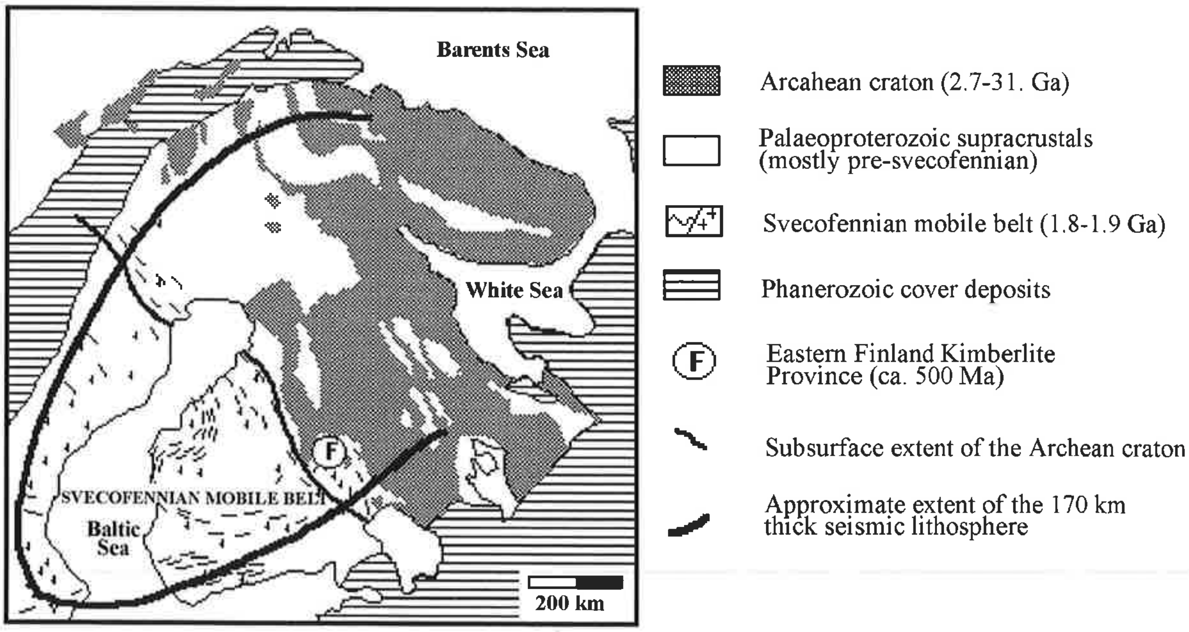

Figure 1. Main geological features of the eastern part of the Fennoscandian Shield (simplified from Peltonen et al., 1999).

3. The Svecofennian Orogeny

The Svecofennian orogeny (1.9 to 1.8 Ga) was preceded by several stages of the rifting of the Archean continent between 2.5 and 2.0 Ga and the formation of the Jormua ophiolite association at 1.95 Ga (*Kontinen, 1987; Peltonen et al., 1996; Korsman et al., 1999*). The first widely cited plate-tectonic model for the Svecofennian orogenic belt was presented by Hietanen (1975). She interpreted the belt as an island arc system accreted to the Archean craton in the east. The newest models of *Nironen (1997)* and *Korsman et al. (1999)* are based on large volumes of new geologic and geophysical data. According to their interpretations, the Svecofennian orogenic crust was formed by a complex accretion of island arcs from west and south to the Archean continent at 1.91 to 1.89. This accretion led also to tectonic thicknening of the crust and lithosphere, especially near the Archean - Paleoproterozoic boundary. Subsequent thickening of the Svecofennian crust at 1.89 to 1.87 Ga was related to mafic underplating which caused also remelting and increased metamorphism (*Korsman et al., 1999*). Anatectic melting of the thick felsic sedimentary - volcanic complex of southern Finland produced the 1.84 to 1.81 Ga migmatite-forming potassic granites. Migmatites and associated rocks are intruded by the 1.80 to 1.77 Ga post-tectonic granitoids which in some places show a bimodal magmatic association.

4. Reworking of the Crust

The isotopic methods (U-Pb, Sm-Nd, Pb-Pb, Rb-Sr, K-Ar, Re-Os) have been used widely and successfully, often together with geochemical studies, to model the petrogenesis of granitic and other igneous rocks. By isotopic methods it is possible to determine the age of the rock and to estimate the age (crustal residence time) and nature of the source material.

Kouvo et al. (1983) showed by using the U-Pb, Sm-Nd, Rb-Sr and Pb-Pb methods that the Nattanen-type 1.8 Ga granites of northern Finland, although mostly located in the Palaeoproterozoic crust, have a major Archean crustal source component. In an extensive study on the origin of the Svecofennian crust, *Huhma (1986)* showed by U-Pb and Sm-Nd studies that all of the 1.9 to 1.8 Ga granitoids of southern and central Finland consist mainly of newly mantle-derived material, with only a minor admixture of older continental crust. Whereas nearly all of the 1.9 to 1.8 Ga granitoids from northern and eastern Finland, close to the Archean craton, had a major Archean source component. The presence or absence of the Archean component in granites has subsequently been used to locate the boundary between the (deep) Archaean crust and the accreted Paleoproterozoic Svecofennian crust (e.g., *Lahtinen and Huhma, 1997*) (Fig. 2). The wide tectonothermal impact of the Svecofennian orogeny to the Late Archean basement of eastern Finland was proved with K-Ar datings of biotite and hornblende by *Kontinen et al. (1992)*.

The extensive U-Pb, Sm-Nd, and Pb-Pb studies of *Rämö (1991)* confirmed that the main source of the Finnish rapakivi granites is the Svecofennian crust. He also showed that the granites of the Salmi rapakivi batholith in Russian Karelia, located at the Archean-Proterozoic boundary, has a mixed Archean-Proterozoic source. On the basis of geochemical modeling, he concluded that the source had intermediate to felsic (granodioritic) composition.

The episodes of mafic under- and intraplating before, during and after the Svecofennian orogeny have produced bimodal mafic - felsic associations with local mingling and mixing of the melts. These mingling - mixing phenomena are especially well visible in the postorogenic granitoid complexes of southwestern Finland *(Lindberg and Eklund, 1988)*, and in the anorogenic rapakivi granite complexes *(Rämö, 1991; Eklund et al., 1994; Salonsaari, 1995)*.

5. The Paleomagnetic Studies

Paleomagnetic studies provide an important addition to the studies of the geologic history of Fennoscandia. By measuring the natural remanent magnetization of precisely dated igneous rocks, it is possible to construct the drift history of the Shield. On the basis of the studies by K.J. Neuvonen, L.J. Pesonen and S. Mertanen it is possible to calculate the movement of the Shield from the late Archean to the late Neoproterozoic (e.g., *Mertanen, 1995; Mertanen and Pesonen, 1997; Neuvonen et al., 1997*). Recently, the paleomagnetic data have been used to make reconstructions of the Fennoscandia with other shields (Laurentia, Ukraine, Amazonian

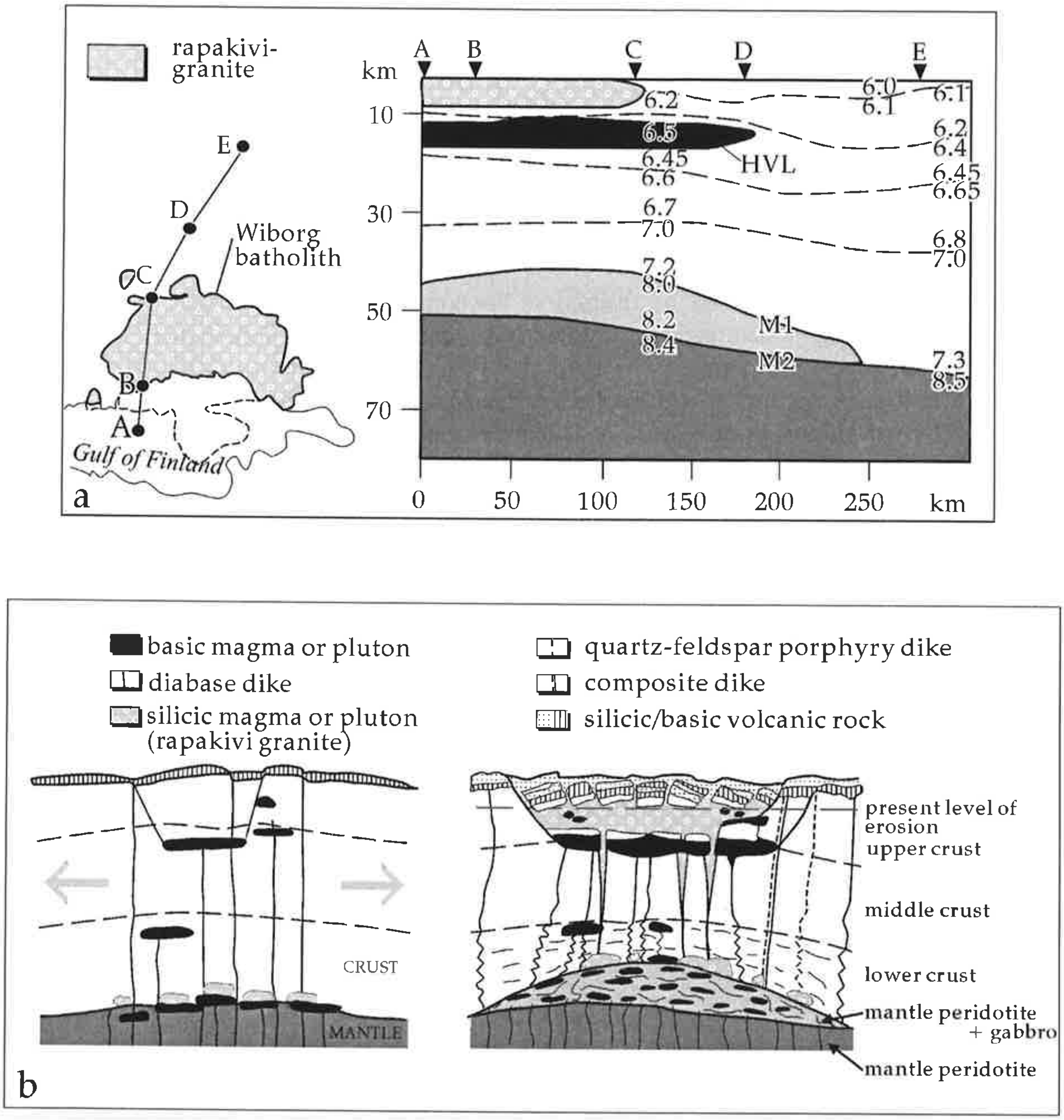

Figure 2. **a)** The structure of the lithosphere across the Wiborg rapakivi granite batholith in SE Finland (modified from Luosto, 1991, and Korja et al., 1993). **b)** a two-stage model for the origin and emplacement of the bimodal rapakivi granite complexes of Finland (Haapala, 1989; Rämö and Haapala, 1996). Mantle-derived basic magmas are intruded in part at the mantle-crust boundary producing a mafic underplate (mantle peridotite + gabbro) and in part to higher levels producing diabase dikes, gabbroic intrusions and volcanic rocks. The thermal effect of the basic magmas (about 1200°C) causes extensive melting of the intermediate - silicic part of the lower crust producing granite magmas from which the rapakivi granites, porphyry dikes and rhyolites crystallized. Mingling of the basic and silicic magmas lead to local hybridization (e.g., Salonsaari, 1995). Partial melting and the transit of the granitic magmas to higher crustal levels, coupled with extensional stretching, lead to thinning of the lower crust.

craton, Congo craton) at various times in the geological past (*Mertanen and Pesonen, 2000, this volume*).

6. Conclusions

Combined geologic (petrologic, geochemical, isotopic) and geophysical (seismic, electromagnetic, gravimetric, magnetic, paleomagnetic, geothermal) studies in the 1980's and the 1990's have produced new infórmation on the structure and evolution of the crust and the lithosphere in Finland and adjacent areas. The mantle topography is relatively well known, but only little data exists of the lithospheric mantle. Fragmentary but valuable information on the composition and age of the lithospheric mantle is obtained by studying the mantle xenoliths of the kimberlites in eastern Finland. Seismic tomographic studies are needed to obtain a general picture of the structure and thickness of the litospheric mantle. Such studies are not only of academic interest, but can also be utilized in economic geology, e.g., selecting areas for diamond prospecting. Paleomagnetic studies have enabled us to follow the drift of the Fennoscandian Shield from the Archean to the Phanerozoic, and the method is useful in reconstructing ancient continents. Geochemical and isotopic methods are in key position in modeling the reworking of the crust and in estimating the nature and role of the mantle magnetism.

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The Structure of the Crust and Upper Mantle in Fennoscandia as Imaged by Electromagnetic Waves

Toivo Korja[1] and the BEAR Working Group[2]

[1]Geological Survey of Finland, P.O. Box 96, FIN-02151 Espoo, Finland
E-mail: toivo.korja@gsf.fi

Information on electrical conductivity of the crust and upper mantle beneath the Fennoscandian are reviewed and potential implications are discussed with a special emphasis on the project Baltic Electromagnetic Array Research (BEAR) and its preliminary results.

Keywords: geoelectromagnetic arrays, electrical conductivity, lithosphere, asthenosphere, Fennoscandia

1. Introduction - Electrical Conductivity

The electric and electromagnetic soundings are the most sensitive geophysical methods for the detection of tectonically important, although volumetrically minor, constituents of the Earth's lithosphere, i.e. carbon/graphite, free saline fluids, hydrogen, and small percentage of partial melt. In tectonically active regions partial melts and free saline fluids form conducting structures that are distinct for different tectonic environments such as continental rifts, mid-ocean ridges, oceanic subduction zones and convergent continental margins.

In stable regions, ancient tectonic processes have in many places in the upper crust left electrically conducting traces, which give information on collisions of crustal blocks revealing the location of palaeosuture zones or terrane boundaries. The enhanced electrical conductivity may be due graphite- and/or sulphide-bearing metasedimentary rocks deposited in various tectonic settings of the continental margin *(Stanley, 1989; Korja and Hjelt, 1993; Boerner et al. 1996; Zhamaletdinov, 1996)*. Lower crustal conductivity is more enigmatic but the likely candidates for the enhanced electrical conductivity are carbon or saline fluids *(Hyndman and Hyndman, 1968; Haak and Hutton, 1986; Marquis and Hyndman, 1992; Jödicke, 1993; Duba et al., 1994; Korja et al., 1996)*. Yet, several crucial questions are open: Is the enhanced conductivity of ancient tectonic origin or is it a result of today's tectonic processes? What is the role of material transport from the upper crust to the lower crust in subductions/collisions? Has the lower crust been spatially adequately sampled by the electromagnetic methods? The last problem, in particular, is of importance because, after the paper of *Haak and Hutton (1986)*, it has commonly been thought that the continental lower crust is usually conductive. The results from the Fennoscandian Shield contrast this opinion and suggest that a resistive lower crust

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[2]The BEAR Working Group consists of researchers from 16 institutes from the following countries *Germany* (University of Göttingen, GeoForschungsZentrum-Potsdam, Technical University of Braunschweig), *Finland* (University of Oulu, Finnish Meteorological Institute in Helsinki and Nurmijärvi and Geological Survey of Finland in Espoo), *Russia* (Geological Institute of the Kola Science Centre in Apatity, St. Petersburg University, St. Petersburg Filial of the Institute of Terrestrial Magnetism, United Institute of the Physics of the Earth in Troitsk and P.P. Shirshov Institute of Oceanology in Moscow), *Sweden* (University of Uppsala and Geological Survey of Sweden in Uppsala), *UK* (University of Edinburgh) and *Ukraine* (National Academy of Sciences of Ukraine in Kiev and Lviv Centre of Institute of Space Research in Lviv).

may be quite common at least in Precambrian shields. This again raises a question on spatial sampling, as there is a tendency to map "anomalous" regions (e.g. old continental margins and suture zones) whereas data from "boring" homogeneous areas are fewer.

In continental areas, a comparison between field observations (magnetotelluric soundings) and laboratory measurements on candidate samples of continental upper mantle (primarily on single crystalline olivine) (see e.g. *Constable et al., 1992*) demonstrates that there must be some conducting component within the dry olivine/peridotite matrix that is not present in an oceanic mantle lithosphere. This may explain the one to two orders of magnitude higher conductivities of the continental mantle observed in the field studies (*Jones, 1999; Bahr and Duba, 2000*). Carbon and hydrogen from dissolved water has been proposed to solve the discrepancy (*Duba and Shankland, 1982*; *Karato, 1990*). Alternatively, *Bahr and Duba (2000)* explained the difference between the laboratory and field data by accommodating the upper mantle conductance into a "thin asthenospheric" layer. This would leave the upper mantle highly resistive - and compatible with laboratory results - except for the conducting asthenospheric layer where enhanced conductivity would be due to the presence of a small percentage of partial melt or some other conducting phase such as carbon or hydrogen.

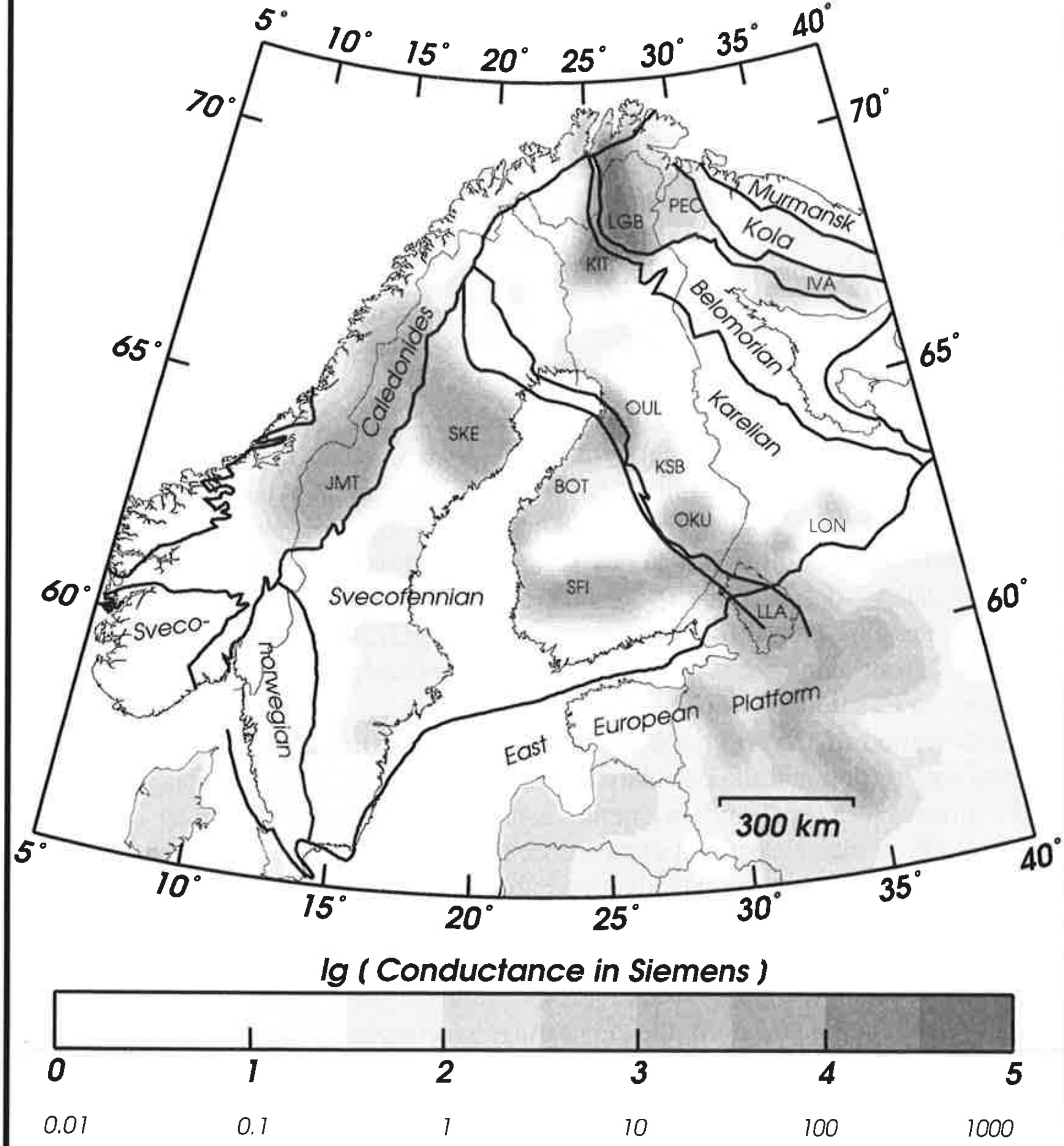
Electromagnetic methods map regions, where conditions are favourable for electric currents to flow no matter at which depths or in which tectonic environment such regions are located. Graphite or saline fluids, for example, may as has been proposed (see references above), provide such pathways in the upper crust, in the lower crust and likely also in the upper mantle. It is, however, quite necessary and useful to make a distinction between the upper crustal, lower crustal and upper mantle conductivity because the accumulation, formation and distribution of these constituents strongly depends on the tectono-geological environment as well as on the physico-chemical conditions. Yet it should be clear that the knowledge of the upper crustal structures and physical properties – easily accessible also by direct observations – will constrain the properties of the deep lithosphere provided that a correct scaling from surface conditions to those prevailing in deep lithosphere is performed (see e.g. *Bahr, 1997*).

2. Upper Crustal Conductors and Collisional Orogenic Belts

Electromagnetic studies of the Precambrian regions throughout the world have shown that elongated conductivity structures are parallel to, and approximately collocated with, collisional orogenic belts e.g., in the Fennoscandian, Canadian and Australian Shields (Fig. 1; *Korja and Hjelt, 1993, 1998; Chamalaun and Barton, 1993; Boerner et al., 1996*). The conductive bodies possess remarkable lateral continuity over hundreds of kilometres and extremely high electrical conductivity (> 1 S/m), the latter suggesting that metallic conduction mechanism (likely dominated by graphite) is required.

Figure 1. Upper crustal conductivity in Fennoscandia. Integrated conductance (conductivity multiplied by thickness) in Siemens for the uppermost 20 km is shown as a grey-shaded map. Note that the areas with conductance below 10 S are shown in white. Thick lines represent boundaries of the major crustal segments in the Fennoscandian Shield, thin lines are the national borders. The map has been produced using the conductance data compiled by *Korja et al.* (2000) as a part of the BEAR research. Note that the interpolation algorithm has smoothed the conductances and therefore the borders of the crustal conductors appear smooth although in reality they usually represent sharp lithological contrasts. The conductors labelled in the map: BOT - Bothnian, IVA - Imandra-Varzuga, JMT - Jämtland, KIT - Kittilä, KSB - Kainuu Schist Belt, LGB - Lapland Granulite Belt, LLA - Lake Ladoga, LON - Lake Onega, OKU - Outokumpu, OUL - Oulu, PEC - Pechenga, SFI - Southern Finland, SKE - Skellefteå.

Crustal conductance: 0 - 20 km
(conductance = conductivity x thickness)



*Lower row of values represent equivalent thickness in metres of conducting material having a resistivity of 0.01 ohm*m (e.g. interconnected graphite).*

In the Fennoscandian Shield, electromagnetic induction research (magnetotelluric soundings, magnetometer array studies) on the deep structure as well as airborne and other ground electromagnetic surveys on the near-surface structure have provided a globally unique electromagnetic data set to estimate crustal conductivity (*Jones, 1992; Korja and Hjelt, 1998*). The electrical structure of the shield is characterised by elongated belts of conductors that are limited to the upper half of the crust except for some occasional, albeit important, penetrations into the lower crust. In most cases conductors within these belts are inclined and reach a depth of several kilometres or even several tens of kilometres, indicating that ancient collisional tectonics were responsible for the present-day geometry of conductors because a transportation mechanism is required to carry sedimentary material to the depths of tens of kilometres.

The close spatial coincidence of the near-surface (airborne surveys) and deep conductors (magnetotelluric soundings) suggests a genetic relationship between them. The high-resolution airborne electromagnetic data reveal a very complicated internal structure of the deep conductors. The large conductance values and correlation with the surcafe geology suggest that the electronic conduction mechanism is required to explain the high conductivity values. Electronic conductor is the dominant conduction mechanism, for example, in graphite and sulphide -bearing rock types. The importance of the two rock types seems to vary from conductor to conductor. The metasedimentary rocks now responsible for enhanced conductivity seem to have been deposited in the Fennoscandian Shield during a relatively short time interval, roughly between 2.1 and 1.9 Ga ago *(Hjelt and Korja, 1998*). The actual tectonic position of the conductors is not fully clear yet but as the carbon is organic, the sedimentation must have taken place in shallow or deep-sea basins. The following interval capable to produce major amounts of conductive lithologies was the Cambrian, when highly conducting black shales (alum shales, so called due to abundance of aluminium) were deposited on top of the rifted western margin of the Baltica *(Andersson et al., 1985*). The Alum shale formation formed a major decollement along which the Caledonian nappes were transported several hundreds of kilometres in the Caledonian orogeny.

The classification of the metasedimentary rocks of North Karelia (OKU in Fig. 1) on the basis of tectonic setting and the correlation of the identified tectonic assemblages with physical properties (mainly electrical and magnetic) indicate the following (*Korja et al., 1999*): (i) graphite provides a major electrical conduction, (ii) a euxinic depositional environment can be implicated in all cases of the conductive assemblages, (iii) it is possible to identify two major depositional settings that are common to all conductors. The extensional basins that were initial rift or pull apart -basins and that developed a higher energy sediment input sometime after initial formation. Compressional basins represent foreland basin system that developed on the Karelian foreland in response to the advancing Svecofennian Arc Complex, (iv) the pre-collisional stage of the basin and margin formation (extensional basins) do not seem to produce elongated conductors with large amounts of conducting materials, whereas the post-collisional stage (compressional basins) contains the dominant proportion of conducting material. But this is contrasted by the Caledonides where the major part of the conducting lithologies (alum shales) were deposited during the pre-collisional stage.

3. Enhanced Lower Crustal Conductivity - Conducting Phase from Surface or Mantle?

The resistive regions between the conductive belts serve as transparent windows to probe deeper structures and properties of the Shield. The deep Palaeoproterozoic crust in the central part of the shield (Central Finland Granitoid Complex) is one to two orders of magnitude more conducting than the Archaean lower crust (Karelian crust in Eastern Finland and Russian Karelia) (*Korja and Koivukoski, 1994; Kovtun et al., 1989*). The lower crust in the south-western parts of the Svecofennian Orogen and in the Sveconorwegian Orogen, on the other hand, seem to be as resistive as the Archaean lower crust (*Rasmussen, 1988; Gharibi et al., 2000*). A notable exception is the transition zone between the Archaean Karelian Province and the Svecofennian Arc Complex, where lower crust is also rather conductive even though it is considered to be composed of Archaean material (Archaean cratonic rocks beneath the Palaeoproterozoic Karelian supracrustal cover).

The cause of the enhanced lower crustal conductivity is not yet clear, but spatial and temporal variations as described above imply that a cause for the enhanced lower crustal conductivity in a stable continental environment should be searched from "local geology" rather than from "universal processes" affecting pervasively the continental lower crust. Studies in the Lapland Granulite Belt (LGB in Fig.1; *Korja et al., 1996*) have shown that the lower crustal conductivity may be explained by graphite-bearing shear zones developed in ductile environment in agreement with the petrologic evidence of dry deep crust (e.g. *Bucher-Nurminen, 1990*; *Frost and Bucher*, 1994). Moreover, it has been suggested (op. cit.) that the carbon is organic. Consequently, this model implies that enhanced lower crustal conductivity is due to the transportation of carbon-bearing material from the surface to the deep crust in tectonic processes and no deep source for the carbon is required. Alternatively, the carbon could be derived from the mantle e.g. via precipitation from CO_2-bearing fluids (e.g. *Frost et al., 1989*). Other possible sources include e.g. magnetite provided that it forms interconnected pathways. This model also suggests a deep source, e.g. additions of mafic material into the lower crust, for the enhanced lower crustal conductivity.

4. Upper Mantle –Is There an Electrical Asthenosphere beneath Fennoscandia?

Electrical conduction in the asthenosphere is an outstanding indicator of the upper mantle's physical state. At temperatures over ca. 1300 °C, the mantle material becomes partially melted and conducting pathways for electrical currents are created. If no other conduction mechanism is active, electrical asthenosphere is intimately connected with the thermal state of the upper mantle. Existing electromagnetic data from the central and southern Fennoscandia do not show well-conducting asthenospheric structures, while results from the more peripheral regions indicate the presence of an upper mantle conducting layer with the depths to the upper surface of the conductor ranging from 70 km to ca. 130 km (*Krasnobaeyeva et al., 1981; Jones, 1982; Jones et al., 1983; Kaikkonen et al., 1983; Rasmussen et al., 1987; Pajunpää, 1988; Korja and Koivukoski, 1994; Zhamaletdinov, 1990*).

These studies, however, have been subjected to a number of electrical distortion effects due to the upper crustal heterogeneities (e.g. the lack of detailed analyses of galvanic and inductive distortions or the use of amplitude data only) and the lack of proper control of the source currents in the ionosphere (*Osipova et al., 1989*). Should these results be relied upon, then the deep geoelectric structure varied over short scales and furthermore the lithospheric thicknesses of only 70 km were in contradiction to the low heat flow in Fennoscandia, in particular in the N and NE parts of the Shield, where a shallow "electrical asthenosphere" was suggested. Yet at high latitudes, as in Fennoscandia, a severe limiting factor for upper mantle studies is the proximity of the ionospheric source region of the inducing EM fields. The closeness of the source region results in a non-uniform excitation model with the traditional EM data, which are not solely dependent on the Earth's conductivity structure, but also on the geometry of the source currents. Recent studies (e.g. *Viljanen et al., 1995; Viljanen, 1996; Viljakainen, 1996*) have demonstrated that the spatial variations of the source field become significant at periods exceeding first thousands of seconds (and temporarily even at much shorter periods), which are responsible for the deep lithosphere - asthenosphere penetration. Thus electromagnetic transfer functions, estimated using the magnetotelluric Tikhonov-Cagniard assumptions (plane waves), would give distorted and even false conducting structures at the asthenospheric depths. The usual way to overcome the source distortions is the careful selection of plane wave events (time intervals with plane wave assumptions fulfilled) and the use of advanced processing techniques (robust and multi-station) to obtain reliable transfer function estimates from the limited fragments of the data records. This strategy is to be based on both the external monitoring of the source structure (geometry and temporal behaviour) and on the internal time series processing criteria. Alternatively, one may attempt to incorporate the non-uniform primary field into the modelling. If the source field heterogeneities can be measured directly by a proper experimental set-up, then their effect becomes part of the modelling. A full control on the spatial behaviour of the source field can be obtained by a 2-D array with the dimensions designed for upper mantle studies and for the expected spatial variations of the inducing field. In this case transfer functions can be estimated as a function of spatial frequencies (as well as temporal frequencies) and no assumptions on the spatial wave number are required.

5. The BEAR Project

The project Baltic Electromagnetic Array Research (BEAR) uses data from a shield-wide magnetotelluric and magnetometer array to determine the electrical conductivity of the upper mantle beneath the Fennoscandian Shield (Fig. 2; *Korja and BEAR Working Group, 1998*; *BEAR Working Group, 1999*). The BEAR project completed the data acquisition on June-July 1998 by acquiring during a period of 1.5 months, magnetic and electric time series data simultaneously from 46 magnetotelluric sites. The magnetic data from 20 permanent magnetometer sites (IMAGE and SAMNET projects) extended the size of the array to 61 sites in an area of 1000 x 1400 km in central Fennoscandia (14°-33° E and 57°-71° N) and five sites in Arctic Ocean (Spitzbergen) (74°-78° N).

The main objectives of the BEAR project are: (i) to estimate a complete set of transfer functions, which contain reliable information on the Earth's interior and are free from the distortions caused by the inhomogeneous polar magnetotelluric source, (ii) to apply the best existing techniques and to develop new methods for the data interpretation, to resolve deep conductivity structures taking into account the influence of the crustal conductors, and (iii) to provide electrical conductivity models and to compare and integrate these models with other available geophysical and geological models as well as geodynamic reconstructions available from the SVEKALAPKO project for the Fennoscandian Shield in order to determine the thickness of the lithosphere, the properties of the asthenosphere, and the disposition of the major lithospheric structures.

The original time series data have now been processed (*Sokolova et al., 2000b*) and the analysis of the internal properties of the electromagnetic transfer functions is under progress (*Lahti et al., 2000*). As a result of the work of the first year, an extensive and unique electromagnetic time series database has been compiled. This database is the principal data set for the BEAR teams to investigate the electrical properties of the Fennoscandian Shield, but it will provide an opportunity to use data from such a large electromagnetic array as an important

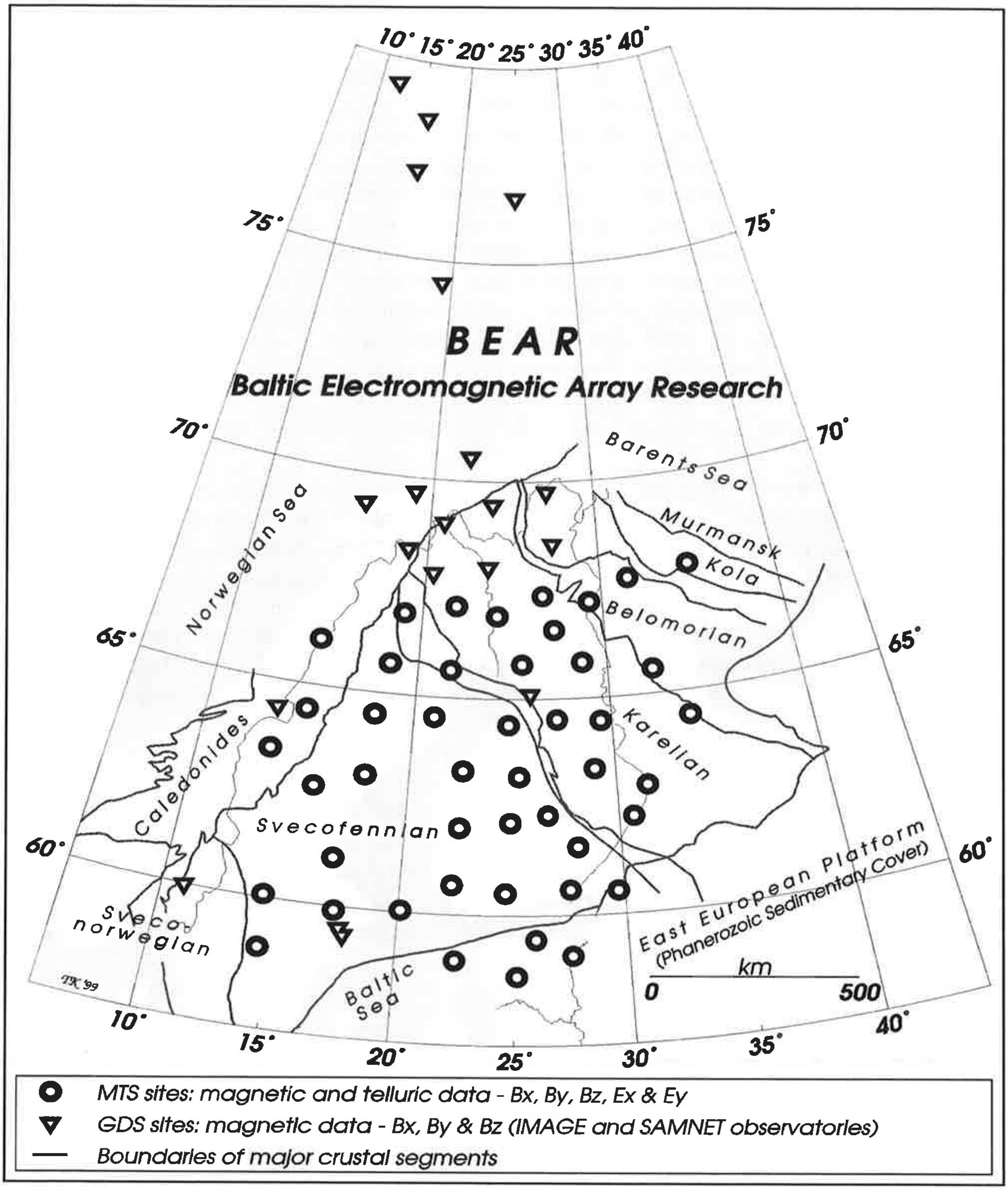


Figure 2. The BEAR array used for ultra deep electromagnetic sounding to probe electrical properties of the upper mantle beneath Fennoscandia.

reference in local studies in Fennoscandia as well as in similar large scale array studies elsewhere.

6. Results of the BEAR Research

The most important result concerning the processing of the time series data, is the compilation of the electromagnetic transfer function (TF) database (*Sokolova et al., 2000b*). The BEAR TF database includes the estimates obtained by various processing methods (single site and remote reference techniques, robust estimators and array schemes) and by several teams. The period interval of stable and reliable transfer function estimates, robustly averaged over the individual estimates, ranges from 10 s to 3 h at almost all sites and up to 6-12 or even 24 hours in many sites. The quality of estimated transfer functions and their robustness in the presence of the different distortions was permanently controlled with almost all possibilities given by the modern processing techniques and by the excessive amount of simultaneous data recordings. The processing task provided also a unique comparison of a number of different data processing techniques (applied by 5-6 teams) both on complicated real data and on synthetic time series data. The results indicate a good compatibility of the results obtained by the different approaches and suggest therefore that the final transfer functions are reliably estimated.

Perhaps the most striking result was the finding that in most cases the advanced the robust techniques are capable to practically eliminate the distortions in the transfer function estimates caused by the non-uniform source field (*Sokolova et al., 2000a*). Distortion effects are clearly seen in the data, but the data-adaptive robust techniques automatically reject the seriously distorted time domain events and further downweight the outliers in the data, and thus they eliminate distortions up to periods of 3-6 hours at properly observed sites. This fact would really extend the applicability of the standard (plane wave) magnetotelluric method to at least one decade longer periods than previously expected at high latitudes close to the polar source region.

The existence of a conductive layer in the uppermost mantle nearly everywhere beneath the Shield has been confirmed. This fulfils one of the main aims of the BEAR project and will provide, after careful analyses and modelling of the geometry and properties of the conductor, important information on the electrical conductivity of the lithosphere-asthenosphere system. More challenging - from the methodological point of view - is, however, the finding that almost all the sites show a moderate departure from a two dimensional behaviour (*Lahti et al., 2000*). Three dimensional inversions are therefore required to obtain reliable lithospheric conductivity models in Fennoscandia.

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Review of Some Features of the Seismic Velocity Models in Finland

Pekka Heikkinen and Urmas Luosto

Institute of Seismology, P.O. Box 26, FIN-00014 University of Helsinki, Finland
E-mail: pekka.heikkinen@seismo.helsinki.fi

This work is focused on features that have had less attention in the interpretation of the refraction/wide-angle data: i) structures in the upper mantle, ii) variations in the lower crust and iii) the extent of the low velocity zone in the upper crust. The preliminary results indicate that a high-velocity boundary in the mantle, observed earlier on BALTIC profile, exists also on SVEKA81 and possibly FENNIA. The lower crust seems to be divided into two layers in all the profiles in the Proterozoic Svecofennian Domain. Major exceptions are the southern end of the BALTIC profile and the northern end BABEL 1, where the high-velocity layer at the base of the crust does not exist. The low velocity zone seems to be connected with the Central Finland Granitoid Complex. It is not observed on BALTIC, disappears towards north on SVEKA81 and to south on FENNIA and diminishes in size across the BABEL line 1 to west.

Keywords: crustal structure; upper mantle; Proterozoic; Fennoscandia; deep seismic sounding

1. Introduction

Much of our knowledge of the Earth's crust and uppermost mantle is based on results of the seismic refraction and reflection surveys (*Mooney and Brocher, 1987*). In refraction surveys – also called Deep Seismic Soundings (DSS) - the spacing between receivers is typically 1-2 km, the shot point intervals are tens of kilometres and recording offsets are up to several hundreds of kilometres. In reflection surveys the receivers and source points are more densely spaced (50-200 m) and offsets much smaller, 3-15 kilometres. The two data sets provide complementary information of the Earth's structure. Near-vertical reflections map acoustic impedance changes with high lateral resolution and are capable of showing horizontal variations in the crustal structure. On the other hand, the velocity models based on the nearly horizontally travelling refracted waves and wide-angle reflections show more or less smoothly changing layered the structures. Refracted waves and wide angle reflections have low lateral resolution but they provide good control on the average velocity distribution, especially in the lower crust, where near-vertical reflections do not accurately constrain the seismic velocities (*Christensen and Mooney, 1995, Rudnick and Fountain, 1995*). Using refracted waves it is also possible to study structures with low acoustic impedance contrasts like transition zones or homogeneous batholiths.

The near-vertical reflection data are processed to be displayed as vertical cross-sections and they can be more or less directly compared with the geology *(Yilmaz, 1987*). The refraction data are usually interpreted by the method of trial and error using forward ray-tracing modelling (*Cervený and Pšencík, 1983*). Due to the method used, the interpretation of a refraction survey is not unique. The most reliable results in the models are the average velocity distributions and the depth of the Moho boundary.

Most of the seismic measurements up to the 1970's were refraction surveys (e.g. *Giese at al., 1976*). Wider use of seismic reflection profiling in deep crustal studies began around 1975 (e.g. *Oliver et al., 1976; Barazangi and Brown, 1986*). In recent years a large number of reflection

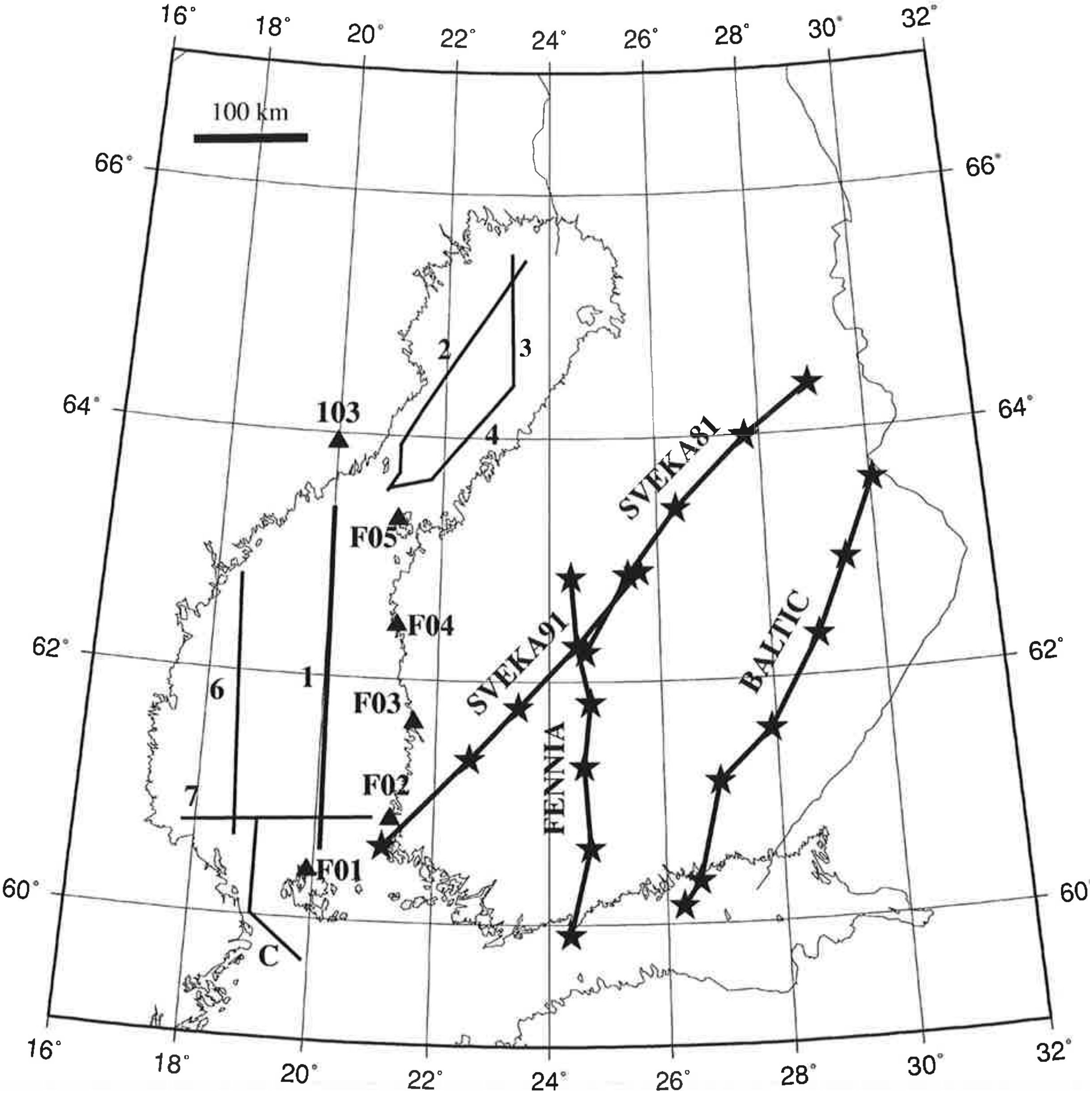


Figure 1. The seismic refraction profiles and BABEL reflection lines on the Fennoscandian Shield. The numbers 1,2,3,4,6,7 and C mark the corresponding BABEL line. The shot points of the refraction profiles are marked with stars and the locations of the land stations for BABEL lines are marked with dots.

surveys have been conducted in Precambrian regions, in Canada (*Clowes et al., 1996*) and also in Australia (*Korsch, 1998*).

In Finland, SVEKA81 in 1981 and BALTIC in 1982, were the first proper deep seismic refraction surveys (*Luosto et al., 1984; Luosto et al., 1990*). The SVEKA81 profile was continued to south-west in 1991 (SVEKA91). In 1994 the measurements were conducted along the FENNIA profile (*Luosto et al., 1994; FENNIA Working Group, 1998*), south of the SVEKA81 profile. The largest reflection survey in the Shield has been the BABEL project in 1989 during which more than 2000 km of deep reflection lines were shot in the Gulf of Bothnia and Baltic Sea (*BABEL Working Group, 1993a,b*). The air-gun signals were also recorded on land providing refraction/wide-angle reflection data (see Fig.1 for profiles).

General reviews of the results of the refraction surveys in the Fennoscandian Shield have been made earlier by Luosto (*1997, and references therein*). The BABEL near-vertical reflection data and their interpretations are discussed elsewhere in this volume (*Korja and Heikkinen, 2000*). We thus chose to concentrate here on some interesting features, to which less attention has been paid in the interpretation of the refraction/wide-angle data. The questions we would like to discuss here are: i) the possible structures in the upper mantle, ii) variations of the structure of the lower crust and iii) changes in the extent of the low velocity zone in the upper crust.

The data considered here are from the DSS profiles in the Central and Southern Finland: SVEKA81, SVEKA91, BALTIC and FENNIA. We also use wide-angle recordings of the BABEL line 1, from which new data have recently become available from the stations F01 and 103. As these stations are on the southern and northern extensions of the profile, the velocity model constructed from their data coincides with the profile itself and can be compared directly with the near-vertical reflection section. The earlier velocity model (*Heikkinen and Luosto, 1992*) based on the data from the stations F02-F05 on the mainland of Finland, images velocity distribution between the line 1 and the coast, i.e. about 40-50 km east of the BABEL 1. For short, we shall in the following distinguish these two models by calling the new velocity model BABEL1/M (Fig. 2) and the older one BABEL1/E.

2. Mantle Structures

Most of the velocity models mentioned above do not show boundaries in the upper mantle as the main focus in interpretation has been on the crustal structure. One exception is the velocity model of BALTIC, where a deeper boundary with velocity of about 8.4 km was observed at the depth of 50 km, about 10 km below Moho in the southern end of the profile. The structure is similar also in the northern end where the 8.4 km/s-boundary is at the depth of 60 km. Strong reflections from the southern shot points also indicate an existence of a deeper layer, which is interpreted to be at the depth of 75-95 km in the middle of the profile (at distances 180-340 km from the beginning of the profile), dipping gently to south.

In other profiles, which are shorter than BALTIC, we have also tried to find indications of similar layers by studying later reflections. In SVEKA81 an interesting new observation is a high velocity boundary (V_p=8.5 km/s) below Moho at the depth of 70 km with. Due to the short length of the profile, this layer can be observed only as reflections. It also seems that this boundary dips gently to south. Below this boundary there are also indications of a layer at the depth of 90 km.

The FENNIA profile, which runs to south from the profile SVEKA81, shows thinning of the crust towards south. However, instead of the gradual change in the crustal thickness observed beneath SVEKA91, a large step in the Moho depth was required to make the observed Pn-refractions possible. As a step of this amount is not very common, we have tried to find alternative models. If we introduce beneath the Moho a layer of higher velocity (8.4-8.5 km/s), like in the BALTIC profile, it is possible to make the Moho boundary more gently dipping and still find solutions that satisfy the observations. However, this is a very preliminary observation and requires further studies.

In the BABEL wide-angle data similar boundaries have not yet been found, which may be due to lower signal strength. The most surprising observation in the data from the station 103 is a strong, north-dipping reflector in the mantle in the middle of the profile (Fig. 2). In the near-vertical data comparable features are not observed.

Although most of the DSS profiles in Finland are relatively short for studying sub-Moho features, the very preliminary results seem to indicate that a high-velocity boundary exists not only beneath BALTIC, where it was earlier observed but also beneath SVEKA81 and possibly FENNIA.

3. High Velocity Layer in the Lower Crust

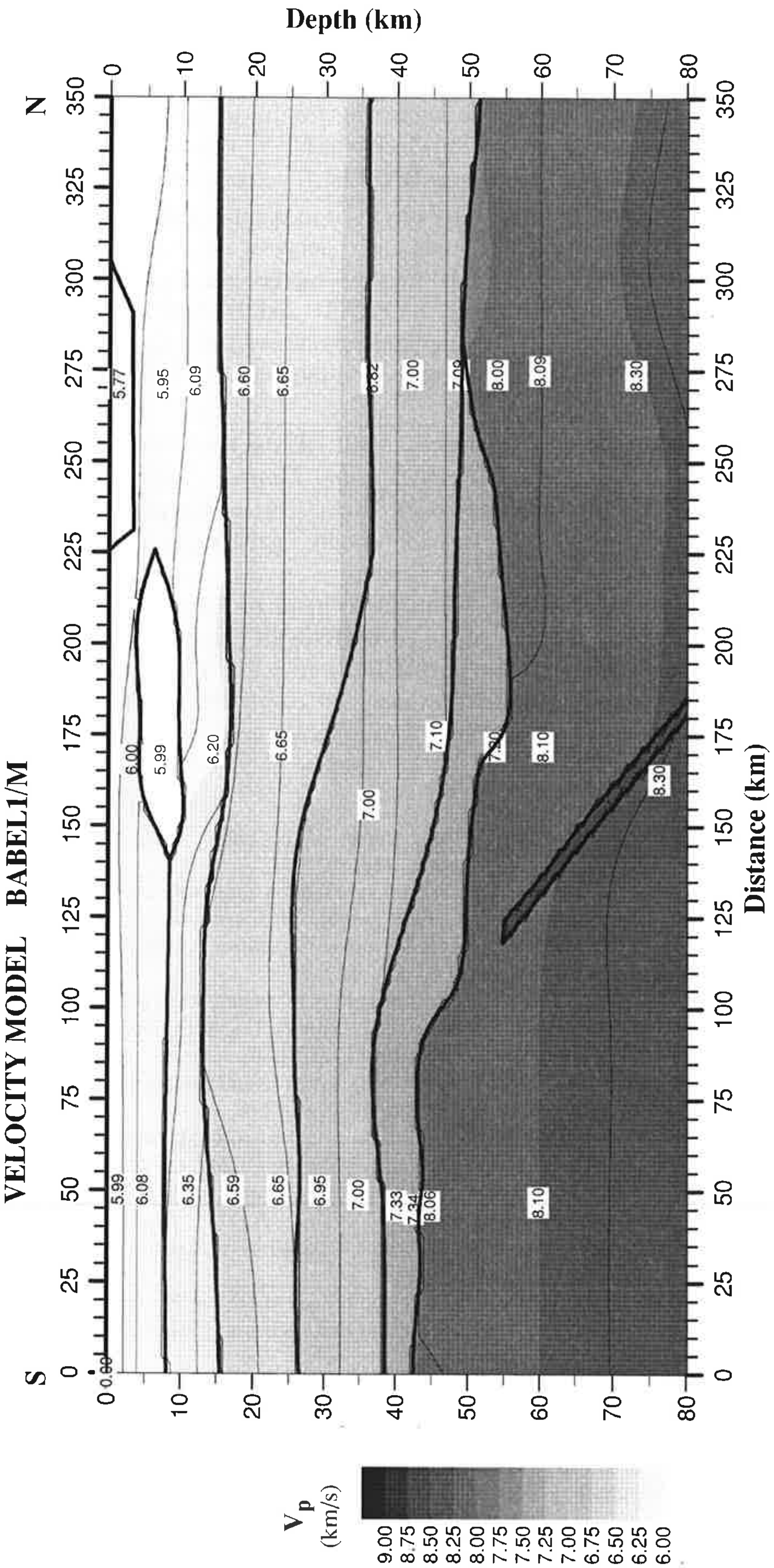
The most important and surprising result of the DSS profiles mentioned above was that the crust in the Proterozoic Svecofennian is exceptionally thick, up to 60 km. In southern Finland as well as in the Archaean in the north, the crust thins to values of 45-48 km (*Luosto 1997 and references therein; Yliniemi, 1986; Yliniemi at al., 1996*). It was also found that the Moho depth changes were mostly due to variation of the thickness of the lower crust (*Korja et al., 1993*) where Vp > 7.0 km/s. However, the structure of the lower crust can differ between profiles. In most cases it can be divided in two separate layers. In BALTIC and BABEL B (*BABEL Working Group, 1993b*) the velocity change in the lower crust is interpreted to be gradual with velocity increasing from about 7.0 to 7.3 km/s. This may be caused by the fact that this layer is too thin to be observed directly as a first refracted arrival. In SVEKA81 where this layer is thicker, about 15 km, it can be clearly seen from two opposite directions, which confirms the existence of this boundary. In BABEL 1 from the station F04, a corresponding refracted can be seen, but only in one direction. In other profiles (SVEKA91 and FENNIA) the layer is thinner and observed only as reflections, which do not constrain the velocity strongly.

Reprocessing and closer examination of the BALTIC data showed reflections from a lower crustal layer, which can be interpreted coming from a boundary where velocity increases from 7.2 km/s to 7.35 km/s. These changes improved the match between the calculated and observed traveltimes. This layer disappears at the southern end of the profile as well as beneath the Archaean like in the SVEKA81 profile.

In the velocity model BABEL1/E the high-velocity layer at the base of the crust is about 12-13 km thick in the central part of the profile, thinning strongly towards north and more gently to south. In the new velocity model BABEL1/M (Fig. 2) the high-velocity layer is thinner, disappearing in the north at about 300 km from the beginning of the profile.

The lower crust seems to be divided into two distinct layers in all profiles in the Proterozoic Svecofennian Domain. Major exceptions are the southern end of the BALTIC profile and the northern end BABEL 1, where this layer does not exist. It may be interesting to note that both of these areas were later intruded by rapakivi granites.

Figure 2. The velocity model of the BABEL line1 determined from the stations F01 and 103.


VELOCITY MODEL BABEL1/M
Depth (km)
Distance (km)
S
N
Vp (km/s)
9.00
8.75
8.50
8.25
8.00
7.75
7.50
7.25
7.00
6.75
6.50
6.25
6.00

4. Low Velocity Zone

Many of the upper crustal features find explanations from the lithological variations, which can be correlated with surface geology. One interesting feature in the upper crust not so readily explainable is the low velocity zone (LVZ), which typically is at the depth of 5-10 km. It is seen in SVEKA81, SVEKA91 as well as in the at BABEL1/E velocity models.

As the low velocity layer is not directly observed as a refracted wave, its interpretation is based on the decay of the P-wave amplitude and a delay of the P-wave, which has refracted from the layer beneath the LVZ. This kind of features can be produced also by lateral variations in the velocity structure. In two models, SVEKA81 and BABEL1/E, the LVZ is perhaps determined best. The weakening of the amplitude is observed in a similar way at stations, for which the rays have passed quite different paths and thus the low velocity zone is the simplest possible explanation.

In the southern end of the velocity model BABEL1/E the low velocity layer disappears and it is not visible in BABEL 7 at all (*Heikkinen and Luosto, 1992*). In the velocity model BABEL1/M the LVZ is observable only over a small part in the middle part of the profile, between 140-220 km. In FENNIA it was found only in the northern end of the profile. In BALTIC the LVZ is not observed.

The pronounced low velocity zone seems to be more or less connected with the Central Finland Granitoid Complex. It is not observed on BALTIC and disappears towards north in SVEKA81 and diminishes in size across BABEL 1 to west. Either the composition of the Granitoid Complex is such that under prevailing PT-conditions the P-velocity decreases as observed for some felsic rocks or the change is caused by a compositional change.

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Deep Seismic Tomography of the Crust and Lithosphere-Asthenosphere System in the Fennoscandian Shield – Pre-studies and First Results

Tellervo Hyvönen[1], Elena Kozlovskaya[2] and
the SVEKALAPKO Seismic Tomography Working Group

[1]Institute of Seismology, P.O. Box 26, FIN-00014 University of Helsinki, Finland
[2]Department of Geophysics, P.O. Box 3000, FIN-90014 University of Oulu, Finland
E-mail: lhyvonen@seismo.helsinki.fi

Two large seismic tomographic experiments aiming to study the lithosphere-asthenosphere system beneath the Fennoscandian Shield and its southern margin have been carried out recently. They are the EUROPROBE/TOR -project in northern Germany, Denmark and southern Sweden in 1996-97 and the EUROPROBE/SVEKALAPKO -project in southern Finland and northwestern Russia in 1998-99. The first results of the TOR project and preliminary studies in the SVEKALAPKO array area demonstrated that 3D seismic tomography experiments and seismic anisotropy studies are able to reveal distinct differences between deep lithosphere structures in the areas of important tectonic boundaries.

Keywords: Seismic tomography, lithosphere, asthenosphere, Fennoscandian Shield, SVEKALAPKO, TOR

1. Introduction

Two large seismic tomographic experiments aiming to study the lithosphere-asthenosphere system beneath the Fennoscandian Shield and its southern margin have been carried out recently. They are the EUROPROBE/TOR -project in northern Germany, Denmark and southern Sweden in 1996-97 and the EUROPROBE/SVEKALAPKO -project in southern Finland and northwestern Russia in 1998-99. The first results of the TOR project and preliminary studies in the SVEKALAPKO array area demonstrate that 3D seismic tomography experiments and seismic anisotropy studies are able to reveal distinct differences between deep lithosphere structures at important tectonic boundaries.

The Fennoscandian Shield is one of the best places on the Globe for studying the thermal and mechanical processes controlling the evolution of ancient lithosphere and seeking to understand the contrasted signatures of the Archaean and Proterozoic lithospheres. The lack of thick sedimentary cover and relatively low seismic noise provide favourable bedrock conditions for major geophysical experiments (particularly for deep magnetotelluric sounding and seismic tomography studies) to define the lithosphere-asthenosphere systems below the different parts (Lapland-Kola, Karelian and Svecofennian) of the Fennoscandian Shield *(Hjelt and Daly, 1996)*.

Tomography projects aimed at revealing deep Earth structures have become increasingly popular over the last decade. Passive seismic tomography experiments have been carried out within the PANCARDI and TESZ EUROPROBE projects (seismic tomography experiment in Vrancea Zone and TOR project). The main target of TOR is the boundary between the Proterozoic Fennoscandian Shield and Phanerozoic Europe. The recording seismic stations were deployed in a band from East-Central Sweden through Denmark into Northern Germany. The installation of TOR stations was finished in October 1996 and they were operated until the spring and early summer of 1997.

2. EUROPROBE/TOR -project

The TOR project is essentially very similar to the EUROPROBE / SVEKALAPKO passive seismic experiment. Thus all experiences gained from TOR data acquisition and interpretation are relevant for the SVEKALAPKO project. The TOR experiment can be considered as 2.5-D rather than 3-D due to array configuration (a 900-km long profile with a 100-km width). Nevertheless, the first results of the TOR project have proved that such an array is capable of solving many problems relevant to modern seismology (*Gregersen et al., 1999, 2000*). These include an estimation of the deep lithosphere-asthenosphere structure by tomographic inversion of travel times and wave-forms, the studies of scattering, inelastic attenuation by signal coda simulation. Furthermore, crustal and mantle discontinuities can be found by the receiver function method, tomographic imaging of seismic sources, investigations of anisotropy by polarisation of seismic body waves and by surface waves studies. The most important first results of the TOR project are a refined 3-D crustal model and P-wave velocity structure of the lithosphere-asthenosphere system across the Trans-European Suture Zone (TESZ) that allows us to distinguish three lithospheric blocks of significantly different thickness corresponding to the Phanerozoic European lithosphere, the TESZ and the Fennoscandian Shield *(Arlitt et al., 1999, 2000)* and lateral variations of seismic anisotropy of the mantle lithosphere around the TESZ *(Gossler et al., 1999; Wylegalla et al., 1999; Vecsey et al., 2000).*

3. EUROPROBE/SVEKALAPKO -project

The SVEKALAPKO deep seismic tomography experiment was conducted in Fennoscandia (in Finland and partly in Russia) from September 1998 until May 1999. The seismic array of 85 short period and 39 broad band stations was installed in a regular 50 x 50 km grid covering an area of 600 by 400 km, and was designed to sample seismic body and surface wave fields with a resolution of about 30 km at depth of 300 km. Two additional temporary small aperture seismic arrays were recording in Finland (in the centre and in the northeastern part of the array) and one small array in Russian Karelia (see Fig.1). The equipment was provided by scientific institutions from Finland, France, Germany, Poland, the Netherlands, Sweden, Switzerland and Russia. The staff at the Oulu Unit at Sodankylä Geophysical Observatory, the Department of Geophysics, University of Oulu and the Institute of Seismology, University of Helsinki carried out the main part of the fieldwork.

During the data acquisition period teleseismic and regional earthquakes, local events and quarry blasts were recorded. The final SVEKALAPKO event list includes 1356 seismic events: 701 selected teleseismic (epicentral distance > 20°) earthquakes within a magnitude range of 3.9-8.3, 75 regional earthquakes within a magnitude range of 1.2-5.2, 580 local events (explosions) and 120 strong quarry blasts. During the recording period eight local micro-earthquakes occurred in Finland, four of them inside the NW corner of the SVEKALAPKO tomography array. The preliminary analysis of the recordings has demonstrated the high quality of the SVEKALAPKO data set (Fig. 2). All of the SVEKALAPKO event data were transformed into MiniSEED format to provide a convenient data access for all members of working group.

4. Preliminary Studies within the SVEKALAPKO Array

The SVEKALAPKO deep seismic tomography project has the advantage of being able to utilise the results of earlier seismic investigations to form a reliable crust model, which is a necessary condition for the teleseismic tomography. Three-dimensional P-velocity images of the crust for the area of SVEKALAPKO seismic array has been created by linear interpolation of velocity-depth values selected from the DSS results (*Hyvönen and Malaska, 1999*) and by creating time maps of P- and PmP-arrivals (*Luosto et al., 1999*) based on the results of DSS profiles in the Fennoscandian Shield. Earlier investigations give a lithosphere thickness of 160-200 km beneath the central Fennoscandian Shield from surface wave studies (*Calcagnile*

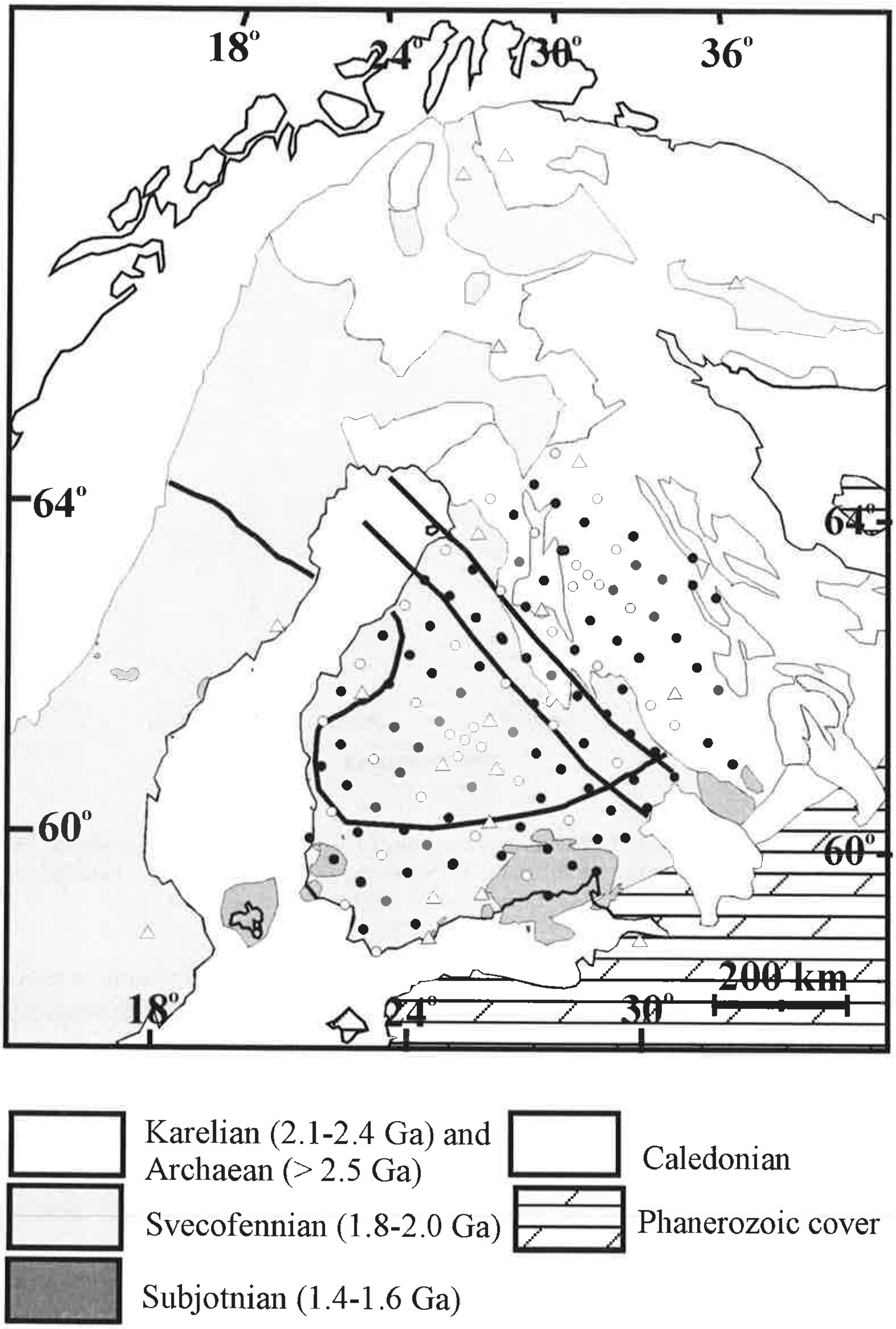

Figure 1. The SVEKALAPKO seismic tomography array on a simplified geological map (black dots - Short period stations (SP), grey dots - Braod band stations (BB), triangles - permanent stations).

et al., 1997) and from body waves (*Babuska et al., 1988*), and these results can be also used as a-priori data to constrain the tomographic models.

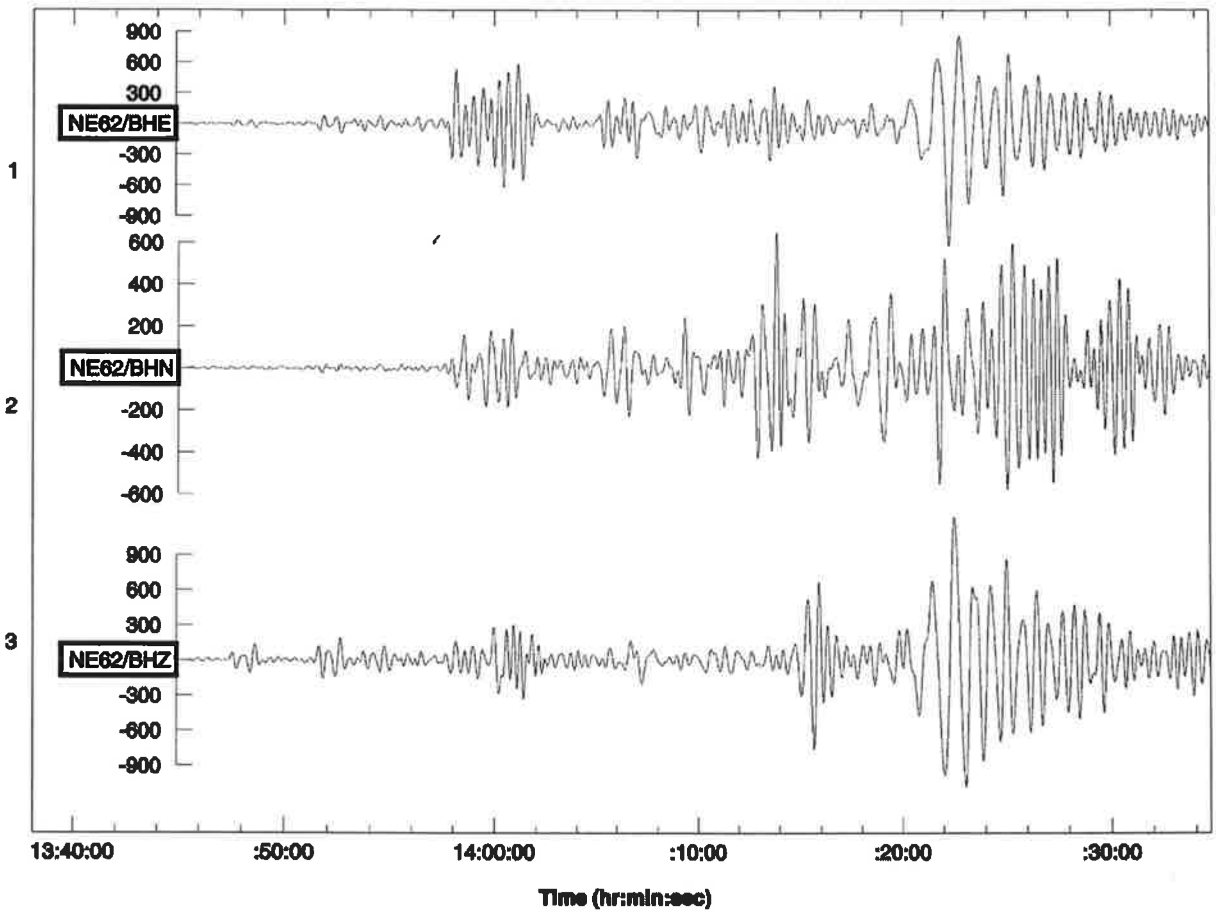

Figure 2. An example of SVEKALAPKO seismic recording of 3-component broad band station FE09 (= NE62 at KAF) in the centre of the array: Earthquake on September 28th, 1998 in Jawa, Indonesia (origin time: 133430.4 UT, location: 8.194º S, 112.413º E, focal depth: 152 km, magnitude: 6.4).

One of the most important parts of SVEKALAPKO project is a seismic anisotropy study beneath the Fennoscandian Shield. The anisotropy of the lithosphere in Fennoscandia has been determined earlier using both surface and body waves (*Babuska et al., 1998; Plomerova et al., 1999a; Silver, 1996*). *Plomerova et al.* (*1996, 1999b*) have measured teleseismic shear-wave splitting values between 0-1 s in Fennoscandia, and beneath seismic station KAF in the centre of SVEKALAPKO array the splitting was 0.9 s, which indicates an anisotropy in the deep lithosphere under this area. The main source of the observed anisotropy seems to be the dominant olivine orientation inside the sub-crustal lithosphere, the crustal anisotropy being relatively small due to the very heterogeneous structure of the crust.

Analysing P-wave travel time residuals showing azimuth-incident angle dependence of relative P- wave residuals can also monitor lateral changes of P-wave velocity anisotropy. *Plomerova et al.* (*1999b*) carried out such an analysis using P-arrival data sets of Fennoscandian seismic stations as well as of permanent seismic stations in Finland. In Fennoscandia the observed high P-velocity directions were interpreted by seismic wave propagation through large dipping anisotropic structures possibly representing stabilised remnant paleo-subduction zones. Corresponding dipping anisotropy patterns were observed beneath stations in the Archaean regions of Finland, and P-wave travel time residuals close to the deep Archaean-Proterozoic paleosuture seemed to be affected by the anisotropy structures. Instead in the stations on the Proterozoic domain of Finland the behaviour of P-residuals seemed to be more scattered, but in general the fast P-velocities propagated from the south-west. These studies show distinct difference in structural behaviour between the main tectonic units covered by the SVEKALAPKO array, i.e. the Archaean-Proterozoic border zone.

The next anisotropy investigation will be the joint interpretation of both shear wave splitting and P-wave travel time residuals with the method developed by *Sileny and Plomerova (1996)* resulting in a 3-D image of the anisotropy in the lithosphere mantle. This together with the dense station coverage in the SVEKALAPKO tomography experiment will allow us to map the anisotropy of the deep lithosphere in a more detailed way than has been possible earlier.

5. Conclusions

The first results of the TOR project show that large seismic tomography experiments are necessary when studying the structure of the deep lithosphere. Preliminary investigations on the SVEKALAPKO area indicate that there exist distinct structural differences beneath the array, and between the Archaean and Proterozoic territories in the lithosphere.

An advantage of the SVEKALAPKO seismic tomography project is that the results of other SVEKALAPKO sub-projects can be used as a-priori data to constrain the seismic models used in the interpretation of various seismic data sets. Especially important are geothermal investigations, analysis of the crust - upper mantle xenoliths and the preliminary results of the SVEKALAPKO BEAR Working Group that confirm the existence of a conductive layer in the uppermost mantle nearly everywhere beneath the Fennoscandian shield (*Bear Working Group, 1999*). The theoretical modelling of rock properties under high pressure and temperature constrained by real geothermal, petrophysical and geoelectric data can be a useful tool to establish the dependence between various rock properties and integrate the data of other SVEKALAPKO sub-projects into inversion algorithms (*Kozlovskaya and Hjelt, 2000*).

The present status of the SVEKALAPKO deep seismic tomography project leads us to conclude that it will be an important step in the development of deep seismic investigations in Finland and provide unique information about the structure of the deep lithosphere-asthenosphere system beneath the Fennoscandian Shield.

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Determination of Crustal Gravity Field of the Fennoscandian Shield Using Deep Seismic Sounding Interpretations

Juhani Kakkuri and Zhitong T. Wang

Finnish Geodetic Institute, FIN-02430 Masala, Finland
Email: Juhani.Kakkuri@fgi.fi

Deep Seismic Sounding (DSS) data have been used for determining the layered structure of the Fennoscandian crust. The 3-D seismic velocity structure model derived from the DSS-data has further been converted to a 3-D density model using the empirical relationship that holds between the seismic velocities and crustal densities. Structural effects on the gravity field of the shield area have been estimated from the 3-D density model.

The study performed shows that the mass deficiency of the Fennoscandian crust must have caused a geoidal depression twice as deep as that one determined from the gravimetric data. This proves that the crust is isostatically compensated by the upper mantle, i.e. an anomalously high-density upper mantle must exist beneath Fennoscandia.

Keywords: crustal structure, deep seismic sounding, geoid, Moho, postglacial rebound.

1. Introduction

Details of the geoid can extensively be explored by means of the deep seismic sounding (DSS). This is possible because the data obtained from the DSS can be used to construct a 3-D velocity structure model for the area to be studied, the Fennoscandian Shield area in our case. The velocity model can further be converted to a 3-D density model using the empirical relationship that holds between the seismic velocities and crustal mass densities. Undulations of the geoid can then be estimated from the 3-D density model.

The velocity at which compressional seismic P-waves travel through homogeneous materials depends on the density ρ, the bulk modulus k, and the shear modulus n as follows:

$$v_p = \sqrt{\frac{k + \frac{4}{3}n}{\rho}} \tag{1}$$

Thus, when the elastic parameters k and n are known, the density can be calculated from the observed velocity of the p-waves. Unfortunately the elastic parameters are poorly known for materials inside the Earth, and the equation shown is not applicable as such. For practical applications, it can be replaced by a linear relation known as Birch's law

$$v_p = \alpha(m) + b\rho \tag{2}$$

where α depends on the mean atomic weight m only, and b is a constant. For plutonic and metamorphic rocks, which are the main types of rocks in the shield areas, the mean atomic weight plays an insignificant role and can be safely neglected from the density–velocity relation. In the sedimentary rock areas, where the p-velocities depend on the burial and geologic age of the rock, the use of Birch's law is, however, questionable. Density data, for example, from drilling holes should be used instead of DSS-data.

According to DSS-data, a three-layer crustal structure exists in most parts of the Fennoscandian Shield, characterized by p-wave velocities of 6.0 – 6.5 (for the upper crust), 6.6 – 6.9 (mid crust), and 7.0 – 7.3 km/s (lower crust). More complicated structures exist only in quite few places, mostly in the vicinity of the transition zones from continental crust to the oceanic crust along the North Atlantic coast.

2. Mathematical modelling

Gravitational potential of a body can be written in the spherical coordinate system as follows:

$$V(r,\theta,\lambda)=G\int\frac{\rho(r',\theta',\lambda')}{\sqrt{r^2+r'^2-2rr'\cos\psi}}r'^2\sin\theta'dr'd\theta'd\lambda' \qquad (3)$$

where ψ is the angle between the vectors $\overline{OQ}$ and $\overline{OP}$ as shown in Figure 1, $\rho(r',\theta',\lambda')$ is the density of a mass element at point Q, and G is the Newtonian gravitational constant. The angle ψ can be solved from

$$\cos\psi=\cos\theta\cos\theta'+\sin\theta\sin\theta'\cos(\lambda-\lambda') \qquad (4)$$

For practical computations, it is convenient to expand Equation 3 into series of spherical harmonics. For details, see *Wang (1998)*.

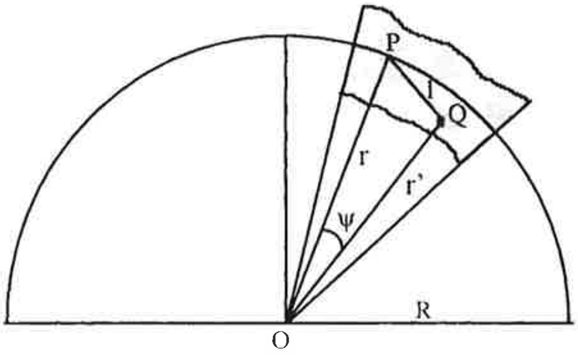

Figure 1. Arrangement of features in a spherical approximation. The shaded area represents the whole crust from the surface down to Moho.

Because of the complexity of the features, it is common practice to represent the potential field of the crust by slicing it into small spherical volumes, finite elements, that take the form of spherical prisms and are filled with homogeneous masses (Fig. 2). The potential field of the whole crust is then the summation of those of the spherical elements. In practice, the masses above the geoid (most part of the continental topographic masses) and those below the geoid (most parts of the Earth's crust) are to be treated separately (see *Wang,1998; Kakkuri and Wang,1998*).

3. Computation of the Fennoscandian crustal geoid

The geoid undulation N can be obtained from the well-known Bruns' formula

$$N=\frac{T}{\gamma} \qquad (5)$$

where T is the disturbing potential on the geoid and γ is the normal gravity. In order to evaluate the disturbing potential T, two models, namely a mass model and a normal model are needed (see Fig. 3). Here the mass model is a simplified three-layered crustal model derived from the DSS-data. In

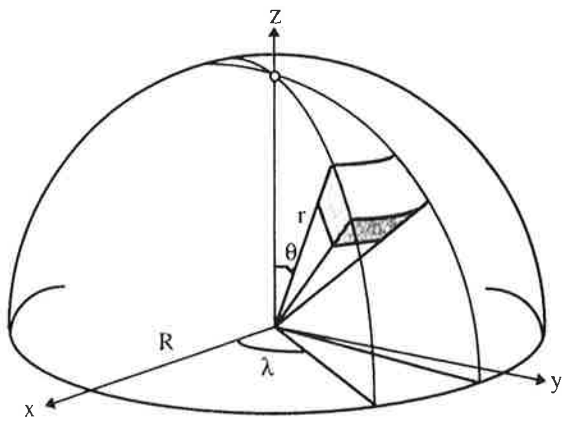

Figure 2. A finite element of a spherical prism.

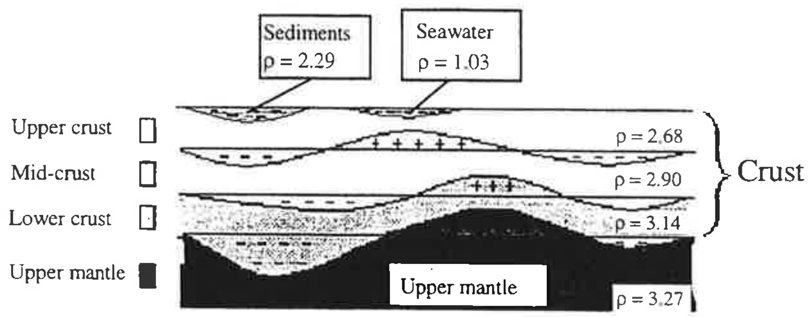

Figure 3. Mass models used for estimating the geoidal undulations from the crust. Straight lines show the boundaries of the normal model and curved lines those of the mass model derived from the DSS-data. Positive and negative signs show the areas of mass surplus and mass deficiency, respectively.

the normal model masses are assumed to be distributed in three homogeneous layers with the depth of each layer being the volume-weighted mean depth of the corresponding layer of the mass model. The density of each layer is determined by the condition that the total masses in the mass model and in the normal model are to be the same in each layer, i.e., the total of mass anomalies is to be zero. As a result, in terms of the spherical harmonics of the disturbing potential, the coefficient of the zero-degree harmonic vanishes.

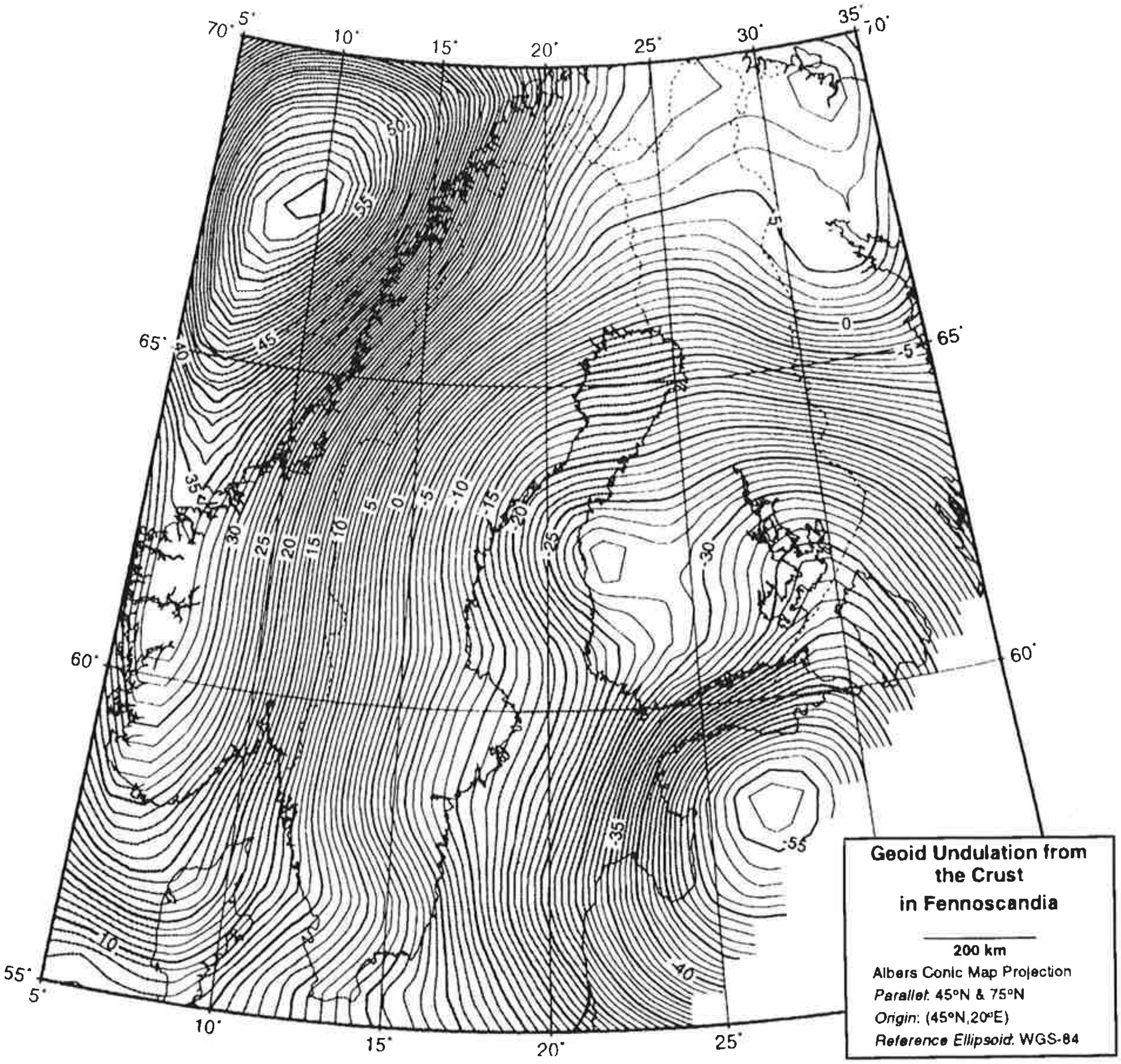

Figure 4. A Fennoscandian crustal geoid calculated from the mass models constructed from DSS-data.

4. Results and conclusions

The Fennoscandian crustal geoid computed by *Wang (1998)* from the DSS-data is shown in Figure 4. When confronting it with the Fennoscandian gravimetric geoid, a bulge approximately 18 metres tall is found. This fact can be interpreted as an indication of the existence of an anomalously dense upper mantle beneath the Fennoscandian crust.

It can also be stated that the influence of the layered structure of the crust on the geoid is mainly due to the variation of the geometric shape of the crustal layers. Variation of the density inside the layers plays a secondary role but is not insignificant. The accuracy obtained is found to be sufficient for geophysical interpretations of the undulations of the Fennoscandian geoid.

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Gravity Anomalies in Finland Due to Crustal and Upper Mantle Sources

Seppo Elo

Geological Survey of Finland, P.O.Box 96, FIN-02151 Espoo, Finland
E-mail: seppo.elo@gsf.fi

The paper briefly describes the status of gravity data in Finland, examines major gravity anomalies, discusses interpretations of gravity anomalies in terms of crustal and upper mantle structure, defines main density modes of the Precambrian crust, and introduces as an example a regional gravity anomaly map from Ylivieska-Vihanti-Kiuruvesi area, in Finland.

Keywords: gravity data, gravity anomaly, bedrock density, crustal structure, Finland

1. Introduction

Finland is exceptionally well covered by gravity measurements. The Finnish Geodetic Institute has established a national gravity net with an average station separation of 5 km (e.g. *Kääriäinen and Mäkinen, 1997*). The regional gravity data of the Geological Survey of Finland covers about 20% of the Finnish territory with an average station interval of about 0.5 km (e.g. *Elo, 1998*). The Nordkalott and Mid-Norden projects have compiled the gravity data from Norway, Sweden and Finland into Bouguer anomaly maps and databases (e.g. *Korhonen et al., 1989; Ruotoistenmäki et al., 1997*). The Fennoscandian Map project has compiled the gravity data from Norway, Sweden, Finland, Russia and Estonia into a preliminary Bouguer anomaly map and database (e.g. *Korhonen et al.*, 1999). Petrophysical data necessary in the interpretation of gravity anomalies is comprehensive (e.g. *Puranen, et al., 1978 ; Puranen, 1989; Korhonen et al., 1997; Airo, 1999*). The geological and geophysical framework including deep seismic soundings, deep electromagnetic soundings, paleomagnetic studies and geothermal research, is well developed as exemplified by this symposium.

The interpretations of the gravity anomalies have dealt mainly with crustal sources but upper mantle sources have also been suggested (e.g. *Elo and Korja, 1993; Elo, 1997; Wang, 1998*). In the following, I shall review a number of results and discuss some of the problems.

2. Gravity Anomalies

The Bouguer anomaly over the Scandinavian mountains is negative (-70 mGal at minimum), while free-air anomaly remains positive (+40 mGal at maximum) as it should in the case of nearly isostatic equilibrium (*Balling, 1980*). The negative Bouguer anomaly and the side minimum of the free-air anomaly extend to Finland (*Elo, 1997*).

Finland belongs to the Fennoscandian land uplift area. A regional Bouguer anomaly minimum of about -20 mGal, with a half-amplitude width of about 600 km along the short axis, is associated with the land uplift maximum, although gravity variations smaller in area but greater in amplitude tend to mask it (*Balling, 1980*; *Kakkuri, 1986*). According to the three-dimensional modelling (*Elo, 1997*) the depth of the source cannot lie deeper than 100 to 300 km.

Wang (*1999*) calculated a crustal geoid using a density structure of the crust derived from the results of the deep seismic soundings. According to him, the effects of the crustal structure on the geoidal depression are surprisingly large, being about twice the observed geoidal depression, which clearly indicates the existence of very dense subcrustal materials. This is in agreement with the idea that the root of the deep lithosphere in Fennoscandia is formed by a crystallization of the asthenospheric materials. Those materials may have become crystallized into semisolid or solid form during the long cooling history of the Shield.

An alternative interpretation of the discrepancy between the observed geoid and Wang's crustal geoid is that the results of the deep seismic soundings cannot reliably be transformed into a true crustal density structure. The relationship between seismic velocity and density in the continental crust is problematic. For example, according to *Barton (1986)*, for example, calculated gravity profiles for the continental crust show that, due to the range of densities possible for rocks of each seismic velocity and vice versa, the use of a seismic velocity measurement of a rock as the only indication of its density does not provide a useful constraint when attempting to reproduce observed gravity variations. Another example is provided by the deep seismic sounding profile BALTIC which traverses the Haukivesi regional gravity high of +40 mGal in amplitude, one of the largest gravity maxima in southern Finland. According to three-dimensional gravity modelling the source of this anomaly must lie in the upper crust extending to a depth of about 16 km. However, in the deep seismic model (*Luosto, 1985*) there is no indication of this large mass surplus. Especially among the granulites, the relationship between the density and the P-wave velocity seems to be ambiguous. There exist data on a group of granulites having densities between 2740 - 2920 kg/m^3 and seismic velocities between 5.4 - 5.9 km/s with no discernible correlation (*Кобранова, 1986*). Consequently, there may well be large volumes of moderately dense granulites in the upper and middle crust not indicated by seismic models.

According to deep seismic soundings (*Korja et al., 1993*), the Proterozoic crust in southern Finland is very thick, while topographic elevations are below 300 m. The Archaean crust in eastern Finland is distinctly thinner than the Proterozoic crust. The crustal thickness in southern Finland ranges from 58-62 to 44-46 km. The thickness of the lower crustal high-velocity layer accounts for most of the variation in total crustal thickness. The thick crust alone should reduce gravity by 100 to 200 mGal depending on what density contrast is assumed. Such negative gravity anomalies are not observed.

On the contrary, striking Bouguer anomaly highs, approximately 40-50 mGal in amplitude (e.g. in Haukivesi and the Gulf of Bothnia), coincide with the thickest crust in Proterozoic southern Finland. According to gravity modelling, the lower crust has a mean density somewhere between the densities of a normal lower crust and an upper mantle; a considerable mass surplus must exist in the upper and middle crust. It seems that a mass transfer from the upper mantle and the lower crust to what is now the upper and middle crust took place in association with the crustal thickening. This mass transfer was largely due to mafic magmatism and thrusting of highly metamorphosed crust towards surface.

Under the Archaean Kuhmo greenstone belt the crustal thickness changes in such a way that it is about 10 km thinner in the northeast than in the southwest. The crustal thickening to the southwest is associated with about 17 km thick high-velocity lower crust, which disappears in the northeast. The 10 km step in Moho should cause a Bouguer anomaly step of nearly 167 mGal if a density contrast of 400 kg/m^3 is assumed. The Bouguer anomaly on the eastern side of the Kuhmo belt is only about 12 mGal higher than in the western side. Based on the amplitude-gradient ratio of Bouguer anomaly, most of the observed 12 mGal step has a source in the upper crust. Obviously the lower crust again has a mean density somewhere between densities of a normal lower crust and an upper mantle. The upper part of the high-velocity lower crust in the southwest, being juxtaposed to the normal lower crust in the northeast, represents a mass surplus. The lower part of the high-velocity lower crust, being juxtaposed to the upper mantle, represents a mass deficit. Taken together they more or less preserve isostatic equilibrium and counterbalance each other's gravity effect (*Elo, 1997*).

The 3 to 6 km thick greenstone belts in Lapland cause Bouguer anomaly highs lying on a broader Bouguer anomaly low due to the crustal thickening of the order of 5 to 8 km. Here the mass deficit due to the crustal thickening dominates and is not entirely counterbalanced by the mass surplus within the crust (*Elo, 1992*).

A crustal thickness of 44-46 km is observed in southeastern Finland under the Wiborg rapakivi granite batholith. An extensive Bouguer anomaly low is associated with the batholith. A pronounced regional high surrounds this low. The range of values is nearly 60 mGal. The anomaly pattern, resembling an inverted fedora, arises when the gravity low, which is due to the relatively low-density upper and middle crust, is added onto a broader gravity high, which is due to the uplifted Moho. The model derived from geophysical data supports a Proterozoic mantle upwelling, which provided excess heat to cause uplift and partial melting of both the upper mantle and the lower crust. Partial melting of the upper mantle produced intrusions of gabbro-anorthosites and diabases, while extensive melting of the lower crust produced rapakivi magmas and caused their emplacement into the upper crust (*Elo and Korja, 1993*).

The general pattern of a gravity high near the Barents Sea coast coupled with a gravity low seawards implies a mass deficit due to relatively low-density sedimentary rocks under the sea associated with crustal thinning (*Elo et al., 1989*).

As a conclusion of the above examples, mass anomalies due to crustal thickness variations are more or less compensated for by mass distributions within the crust.

3. Densities

The distribution of rock densities within the exposed crust is not unimodal. The main component is the granodioritic one, 2685-2720 kg/m^3, consisting mainly of plutonic rocks as in Central Finland Granitoid area, or metamorphosed rocks in the schist and migmatite belts. The granitic component, 2600-2660 kg/m^3, is common being the main component in the Proterozoic rapakivi granite areas and in the Archaean gneiss-granite terraines. The high-density mode from 2740 to 2920 kg/m^3 comprises high-grade metamorphic, mafic intrusive and mafic volcanic rocks and has the largest standard deviation of the components. *Elo and Säävuori (1997)* demonstrated the connection between the Kiuruvesi gravity anomaly maximum and high-grade metamorphism and the abundance of deformed mafic intrusions. The rock samples including granodiorites to gabbros and high-grade cordierite bearing gneiss from the metamorphic core complex have a mean density of 2850 kg/m^3. Distinct Bouguer anomaly maxima in Finland are caused by greenstones and other mafic volcanic rocks, ultramafic and mafic intrusions, diabase dykes and sills, carbonatite intrusions, granulite belts and upthrust blocks from the middle crust in general, schist and migmatite belts when surrounded by a gneiss-granite basement or granites. Distinct Bouguer anomaly minima are caused by a) granite intrusions, rapakivi granites in particular and potassium rich granites of various ages in general, b) thick layers of quartzites in metamorphosed schist belts, c) granite-gneiss basement domes or tectonic blocks in metamorphosed volcano-sedimentary surroundings, d) preserved basins of Mesoproterozoic and younger sedimentary rocks, e) meteorite impact structures, f) strongly weathered bedrock and e) occasionally by abnormally thick overburden.

4. A Gravity Anomaly Map of the Ylivieska-Vihanti-Kiuruvesi Region

Figure 1 shows how information on the structure of the upper crust is revealed by the regional gravity data. For example, in the SE corner of the map lies the Kiuruvesi area, where a major regional gravity high is clearly associated with the rocks of high metamorphic grade. In the NW part, there are several gravity highs caused by large mafic intrusions. The Vihanti regional gravity high, in the upper right hand corner, is partly due to the exposed mafic intrusions, and partly to the dense subsurface rocks of the relatively high metamorphic grade. In the western part of the map, the ovoidal gravity lows surrounded by the gravity highs are due to large granitoids. A diagonal zone of faults and shear lenses runs from NW to SE. Major shear zones occur in NE-SW direction. Granitization and fracturing appear to have occurred along the major shear zones. There is evidence of more clear-cut faults, as well, for example two WSW-ENE faults in the central part of the map.

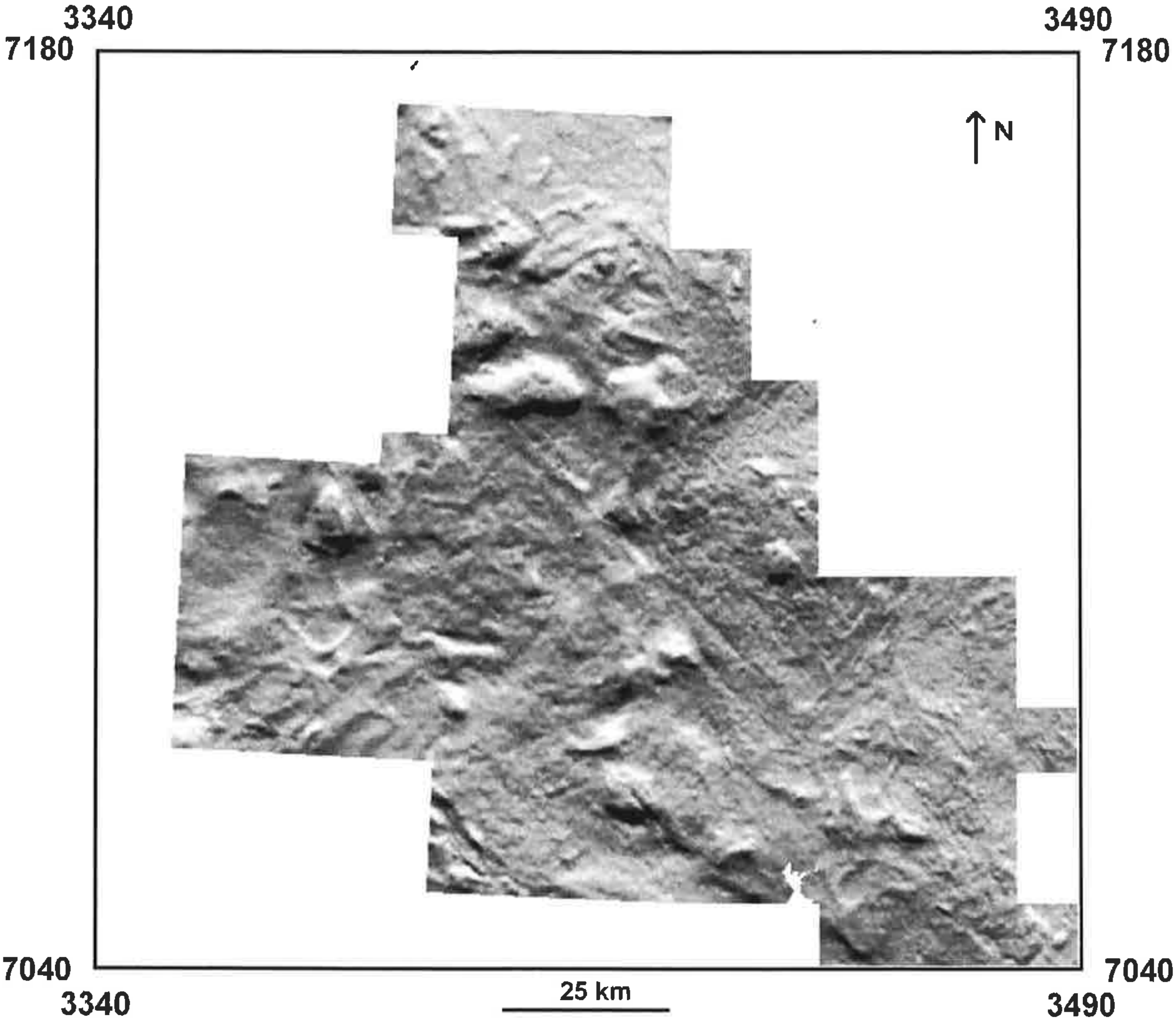

Figure 1. Ylivieska-Vihanti-Kiuruvesi Bouguer anomaly map illuminated from north. Regional gravity data from the Geological Survey of Finland and Outokumpu Mining Co. Average station interval 0.5 km.

5. Conclusions

Evidently gravity data contains a wealth of geological information. The interpretation of this information in terms of crustal and upper mantle structures requires a profound understanding of geological and geochemical processes, advanced computer systems of geophysical modelling, versatile frameworks for taking into account all the complementary information and for maintaining composite models, and even numerical modelling of geological processes themselves.

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Thermal State of the Lithosphere in the Central Part of the Fennoscandian Shield

Ilmo T. Kukkonen and Petri Peltonen

Geological Survey of Finland, P.O. Box 96, FIN-02151 Espoo, Finland
E-mail: ilmo.kukkonen@gsf.fi

The knowledge of lithospheric thermal conditions is usually based on borehole measurements and theoretical models of the heat transfer in the lithosphere. Deep temperatures and heat flow densities (HFD) can sometimes be constrained by direct petrological and geophysical temperature-dependent data. In this paper, we shortly review the HFD measurements in Finland, the central part of the Fennoscandian Shield. The results of geothermal modellings of the crustal and upper mantle temperature and the HFD supported by xenolith-derived deep temperature information are also discussed.

Keywords: Heat flow density, lithosphere, geotherms, xenoliths, Fennoscandian Shield

1. HFD in Finland

Geothermal heat flow density (usually given in mW m^{-2}) expresses the amount of energy conducted to the Earth's surface in unit time across unit area. It is the key geothermal parameter characterizing the internal thermal conditions of the lithosphere, and it is determined as the product of the geothermal gradient and thermal conductivity of the rocks. The gradient is measured in boreholes *in situ*, but thermal conductivity requires laboratory measurements of drill core samples. In Finland, boreholes are usually less than 1 km deep. Heat flow data have been reported by *Puranen et al. (1968), Järvimäki and Puranen (1979), Kukkonen and Järvimäki (1992) and Kukkonen (1988,1989a,1989b, 1993, 2000).*

The Upper crust. Measured heat flow density values in Finland in the uppermost 1 km of the bedrock range from very low values of less than 15 mW m^{-2} to 69 mW m^{-2}, whereas an average value of all 46 sites (53 boreholes) is 37 mW m^{-2} *(Kukkonen, 2000).* The very low values are apparently palaeoclimatically disturbed *(Kukkonen et al., 1998).*

The geothermal gradient in Finland is typically 8-15 K km^{-1}. Temperatures at 500 m below surface are usually between 8 and 14 EC, whereas at 1 km the temperature ranges from 14 to 22 EC. Values either extrapolated from geotherms or calculated with thermal models suggest that temperatures exceeding 40 EC should be encountered at 1-1.5 km depth. However, in order to reach 100EC, depths from 6 to 8 km would be required.

Heat production due to radioactive decay of ^{235}U, ^{238}U, ^{232}Th and ^{40}K in the lithosphere is a key factor in understanding heat flow variations in continental areas. Relationships between measured heat flow density and radiogenic heat production rate in the upper crust in Finland have been studied by *Kukkonen (1989a, 1989b, 1993).* High heat production rates are generally related to increased the HFD values. Heat production rate and HFD are lowest in the Archaean areas in eastern Finland, whereas the highest values are encountered in the Paleoproterozoic Svecofennian migmatites and the Mesoproterozoic Wiborg rapakivi batholite in southern Finland. The described areal trend in the HFD is characteristic of the Fennoscandian Shield in general. The HFD increases from the Archaean areas in the Russian Karelia and Kola Peninsula to the younger Proterozoic areas in Finland and Sweden in a NE-SW direction (Fig. 1). At least in the Finnish part of the Shield, this variation can mostly be attributed to areal variations in the upper crustal heat production rate *(Kukkonen and Jõeleht,1996; Kukkonen, 1998).*

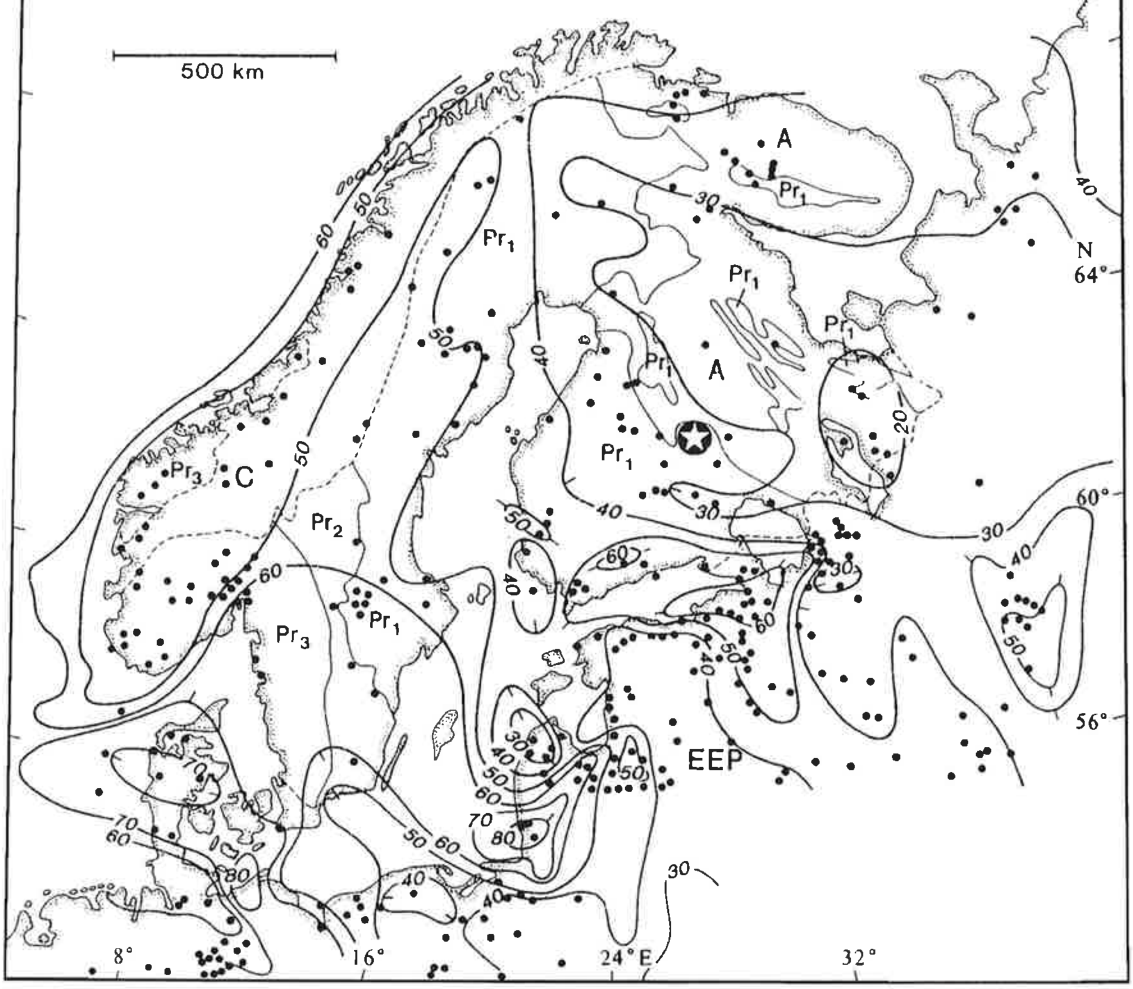

Figure 1. The heat flow density (mW m^{-2}) in the Fennoscandian Shield and adjacent areas. Adapted from *Čermák et al. (1993), Kukkonen and Jõeleht (1996) and Kukkonen et al. (1998).* A - Archaean; Pr_1 - Palaeoproterozoic (1750-2500 Ma); Pr_2 - Mesoproterozoic (1200-1750 Ma); Pr_3 - Neoproterozoic (900-1200 Ma); C -Caledonides; EEP - East European Platform. The star indicates the location of the kimberlite pipes in eastern Finland.

The Lithosphere. The lithospheric thermal regime is basically controlled by the deep HFD signal from the upper mantle added with radiogenic heat generated in the lithosphere. Due to the concentration of heat producing elements in the crust in general, and in the upper crust in particular, the HFD and geothermal gradient decrease downward in the lithosphere. Therefore, direct extrapolation of the data measured in boreholes less than 1 km deep is not sufficiently accurate for estimating the deep temperatures in the lithospheric depth range (0-250 km). Thus, the knowledge of the deep thermal conditions must be based on theoretical models of heat transfer in the lithosphere.

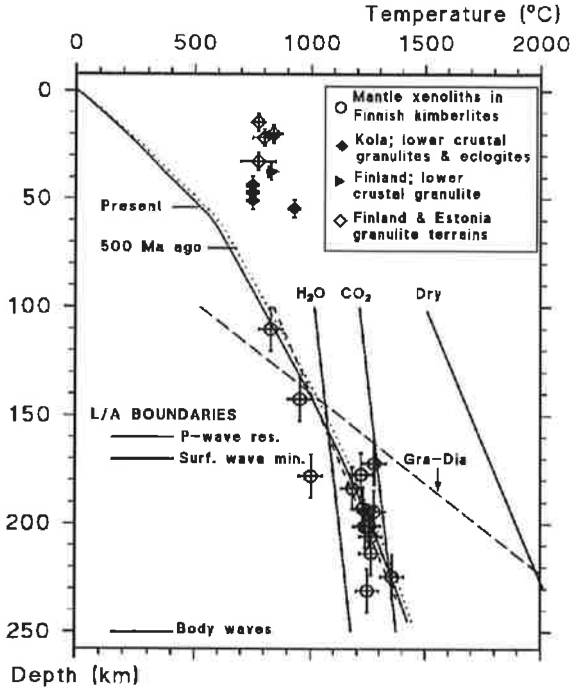

Figure 2. The xenolith-calibrated geotherm and xenolith P-T data for eastern Finland. Solid line is the modelled geotherm representing the present thermal state of the lithosphere, the dotted line corresponds to conditions 500 Ma ago (calculated assuming 10 % higher lithospheric heat production rate). The broken line indicates the average temperature gradient calculated from the xenolith P-T data. 'H_2O', 'CO_2' and 'dry' indicate the peridotite solidi with excess volatile contents or dry rock. 'Gra-Dia' indicates the phase boundary of graphite and diamond. Lithosphere thickness values from seismic P-wave residual studies, surface wave and body wave data are included. Mantle xenolith depth values were recalculated from the original data. Data for granulite terrains and granulite facies xenoliths are given for comparison with error bars based on original data. Figure adapted from *Kukkonen and Peltonen (1999).*

2. Modellings of Lithosphere Temperature and P-T Data on Mantle Xenoliths

Simulations of deep temperatures and the HFD's in the central part of the Fennoscandian Shield of the SVEKA and BALTIC deep seismic transects have been presented by *Kukkonen and Jõeleht (1996), Kukkonen (1998), Kukkonen and Peltonen (1999) and Jokinen and Kukkonen (2000).* Results of these modellings suggest that the temperature at 50 km (approximate average Moho depth in Finland) is about 400-500 EC in the Archaean areas, but it increases to about 500-600 EC in the Proterozoic areas. The mantle HFD is low, in the order of 12 ± 5 mW m^{-2}, both in the Proterozoic and the Archaean areas.

Important additional data on deep thermal conditions have been obtained from the kimberlite-hosted mantle xenoliths in Kaavi, eastern Finland *(Kukkonen and Peltonen,1999; Peltonen et al., 1999).* By combining the petrological pressure-temperature data of the xenoliths with a 2-dimensional thermal model it was possible to obtain a xenolith-calibrated geotherm. The results suggest that the P-T data of the mantle samples are in agreement with a conductive geotherm corresponding to a surface HFD value of 36 mW m^{-2} which according to borehole measurements is representative for the area. All studied mantle xenoliths are lithospheric in character, and they equilibrated at depths ranging from 110 to 230 km (Fig. 2).

Thus, the lithosphere is at least 230 km thick in eastern Finland. Temperatures in the lower lithosphere are 1200-1400EC, which is close to the melting temperature of volatile-bearing peridotite, but no partial melting textures can be seen in the xenoliths. The volatile content of the lower lithosphere is therefore presumably very low. Based on the P-T data of the xenoliths, the average temperature gradient of the upper mantle is 4.0 ± 0.7 K km^{-1} and HFD 12 ± 4 mW m^{-2}. Similar values are given by the forward and inverse models *(Kukkonen and Peltonen, 1999; Jokinen and Kukkonen, 2000).*

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Kimberlites, Carbonatites and Their Mantle Sample: Constraints for the Origin and Temporal Evolution of the Lithospheric Mantle in Fennoscandia

Petri Peltonen, Hugh O´Brien, Juha Karhu and Ilmo T. Kukkonen

Geological Survey of Finland, P.O. Box 96, FIN-02151 Espoo, Finland
E-mail: petri.peltonen@gsf.fi

This work summarises our current knowledge of the petrological nature and chemical and isotopic composition of the subcrustal lithospheric mantle in Fennoscandia. New age determination and Sm-Nd and carbon isotope results are provided for the Finnish kimberlites and carbonatites. Their lithospheric vs. asthenospheric origin is discussed. Our synthesis of the available data implies that the lithospheric mantle in Fennoscandia has had a complex history and consists of several domains and layers with distinct stabilization ages and compositional evolution through time.

Keywords: kimberlite, carbonatite, mantle xenolith, lithosphere, Fennoscandian Shield

1. Introduction

Most Archaean cratons are underlain by anomalously thick (typically ca. 200 km) cold mantle keels generally distinguished by fast and anisotropic seismic velocities relative to the underlying asthenospheric mantle *(e.g. Polet and Anderson, 1995)*. Petrological studies of orogenic lherzolite massifs (exposed lithospheric mantle section within continental shear zones) and mantle xenolith suites recovered from kimberlites and lamproites have implied that these keels consist of mantle peridotites depleted in their basaltic constituents such as Ca, Al and Fe *(Boyd and Mertzman, 1987)*. Recent studies have demonstrated a secular evolution in the composition of the subcontinental lithospheric mantle (SCLM) peridotites becoming less depleted from Archaean through Proterozoic to Phanerozoic time *(Griffin et al., 1999)*.

Radiometric dating of the rocks originating from SCLM has proven difficult because of the open-system behaviour of elements such as Nd, Sr, U and Pb at the high ambient temperatures prevailing in the SCLM. The first evidence for the antiquity of cratonic roots was provided by the Sm-Nd study of garnet inclusions in diamonds by *Richardson et al. (1984)* who demonstrated that the SCLM of the Kaapvaal craton was subjected to a geochemical enrichment in the Archaean. Their study also implied that the SCLM was at least ca. 150 km thick already in the Archaean – the thickness required to stabilize diamond. Progress in Re-Os isotope dating has documented the close correlation between the ages of the stabilization of SCLM and formation of the overlying crust. In many shield areas (Kaapvaal, Siberia, Wyoming, Tanzania) the crust and mantle parts have remained coupled for billions of years *(Pearson, 1999)* which is directly related to the lower density of the SCLM relative to the asthenospheric mantle *(Griffin et al., 1999)*.

A detailed study of the SCLM is crucial for our understanding of crustal processes as well. This is because major modifications in the SCLM by magmatic activity (e.g. thermal erosion of the base of the SCLM by plume activity) or tectonic processes (e.g. rifting) may cause uplifting, magmatic activity or formation of world-class mineral deposits in the crustal part of lithosphere. Until recently, lack of mantle samples from the Fennoscandian lithospheric mantle has prevented the study of many fundamental topics of the lithospheric evolution. The recent discovery of diamondiferous kimberlite pipes in eastern Finland (Tyni, 1997) has substantially improved the situation by providing us mantle samples (xenoliths) from depths between 100-230 km.

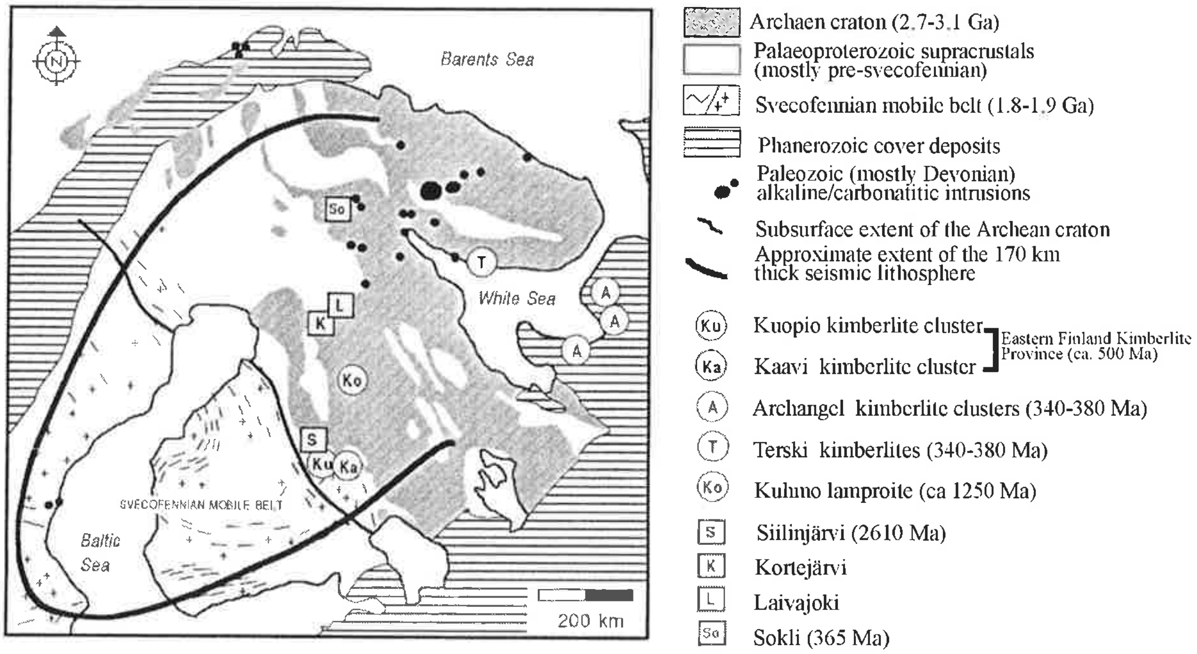

Figure 1. Generalised geology of the eastern part of the Fennoscandian Shield showing the major occurrences of Paleozoic kimberlites and alkaline-carbonatitic intrusions .

2. Kimberlites and Carbonatites of Fennoscandia – Lithosphere or Asthenosphere Melts?

Kimberlites. Kimberlite pipes form from ultramafic, volatile-charged, incompatible element-rich magmas that represent a mixture of liquid, peridotite and eclogite detritus carried from the depth and of the megacryst suite minerals like Ti-pyrope, Mg-ilmenite and Cr-diopside. There are two endmember kimberlite types based on examples from South Africa: Group I with abundant rounded grains (macrocrysts) of olivine, in a monticellite, perovskite, magnesian ulvöspinel-magnetite, Ba-rich mica, calcite and serpentine matrix, and Group II typically with abundant phlogopite ± olivine in a matrix of phlogopite, K-richterite and other diagnostic minerals. Olivine lamproites are similar to Group II kimberlites that exist, for example, in Western Australia (Argyle diamond mine), Montana and Wyoming (USA), and southern Spain but no rocks identical to Group II kimberlites have been found outside of Southern Africa.

Approximately 20 kimberlites have been discovered so far in the Eastern Finland Kimberlite Province which consists of Kuopio and Kaavi kimberlite clusters (Fig. 1). All of them have typical Group I mineralogies, major and trace element compositions and intrusion morphologies. Age determinations have been made on four of these kimberlites by employing ion microprobe analyses of the U-Pb systematics in perovskites. Ages obtained range from 589 Ma to 626 Ma, suggesting that the younger K-Ar determination (434 Ma) reported in *O'Brien and Tyni (1999)* is erroneous. Recalculations using these older ages has significantly reduced the spread in the Sr-Nd isotope field for the Kaavi-Kuopio kimberlites (Fig. 2). In terms of Nd and Sr isotopes, the Finnish kimberlites are similar to Group I kimberlites elsewhere. This worldwide uniformity *(Smith, 1983)* strongly suggests that the source of Group I kimberlite is either from well-mixed mantle (i.e., asthenosphere) or from lithospheric mantle that has been converted physically by heating and chemically by melt infiltration into asthenosphere-like mantle. Either way, Group I kimberlites do not directly provide information on the isolated aged roots of cratons they transect excluding the xenoliths and xenocrysts they contain.

Group II kimberlites and related olivine lamproites, however, have Sr-Nd isotopic compositions that reflect long-term storage of Rb- and LREE-enriched mantle rocks separate from the asthenosphere (Fig. 2). The transitional Group II kimberlite/olivine lamproite in Kuhmo, Finland, appears to be very similar in terms of composition, mineralogy and isotopic composition (Fig. 2) to one group of ca. 365 Ma old diamondiferous rocks from the Archangel area *(Mahotkin et al. 2000)*. However the Kuhmo dyke/pipe appears to be considerably older, ca. 1250 Ma (perovskite U-Pb, O'Brien, unpubl. data). The enriched isotopic signature of the Kuhmo and Archangel rocks indicate that the Karelian craton SCLM contains veins of mica and amphibole, along with trace minerals that over time have developed extreme isotopic compositions. Either sublithospheric magmas were contaminated by the material from these veins to produce the isotopic signatures (a mixing process) or these magmas actually represent direct melts from the metasomatized Karelian Craton SCLM.

Carbonatites. Most carbonatites have similar, although not exactly the same, Sr-Nd isotopic compositions as Group I kimberlites (Fig. 2), and the straightforward interpretation is that they both orginate in the well mixed asthenospheric mantle. However, experiments on melting of carbonated peridotite indicate that most carbonatites should be produced at the solidus inflection at a depth of around 100 km, well within the lithospheric mantle *(Wyllie and Lee, 1999)*. It is proposed that the asthenospheric isotopic signatures result from multiple episodes of invasion of carbonatite melts into the lithosphere mantle that freeze and remelt with successive influxes, rapidly building zones of carbonated wehrlite. Melting of these modified zones can then produce carbonatites at the depth of 100 km with asthenospheric isotopic compositions *(Harmer, 1999)*. The key point is that the process cannot take hundreds of millions of years otherwise the isotopic signatures would indicate this aged enrichment.

Due to the low carbon abundancy in the mantle, the volume of the source rock required to generate carbonatite melt may be 1000-1000 times higher than the volume of the carbonatite itself. Therefore, carbonatites (and carbonaceous kimberlites) can be expected to give a good estimate of the average carbon isotope composition of their ultimate mantle source. Kimberlites from the pipes of the Eastern Finland Kimberlite Province contain 10-15% calcite generally present as fine-grained disseminations in the groundmass. Isotopic compositions of carbon were measured on five samples from four separate pipes. The $\delta^{13}C$ values range from -2.2 to -4.6 ‰ (PDB), with an average value of -3.5 ‰. Surprisingly, carbonatite intrusions from the Karelian domain have $\delta^{13}C$ values in the same range, irrespective of the age of the intrusion. Carbonates from the late Archean Siilinjärvi carbonatite complex span from -3.1 to -4.5 ‰ (n = 8). The Paleoproterozoic Laivajoki and Kortejärvi carbonatites have $\delta^{13}C$ values in the range from -3.6 to -4.9 ‰ (n = 8; *Nykänen et al,. 1997*) and the Devonian Sokli carbonatite from -2.7 to -4.1 ‰ (n = 10). Accordingly, no significant differences in $\delta^{13}C$ signatures can be shown to exist among the various carbonatite intrusions nor between the carbonatites and the kimberlite pipes.

The total span of $\delta^{13}C$ values for kimberlites and carbonatites from the Karelian domain of the Fennoscandian Shield is from -2.2 to -4.9 ‰. Worldwide carbonatite complexes show only slightly more variation in their $\delta^{13}C$ values with 91% of all carbonatites having $\delta^{13}C$ values between -2 and -8 ‰ *(Deines, 1989)*. Two alternative explanations exist for such a small range and the apparent lack of any secular evolution in the isotope composition. One interpretation is that the SCLM of the Karelian craton was subjected to geochemical enrichment already in the Archaean, and that the average $\delta^{13}C$ value of SCLM has not changed significantly since then.

However, considering the complex magmatic and thermal evolution of the Karelian craton it is considered unlikely that the Karelian SCLM has remained intact for eons with respect to carbon isotopes. Therefore, the other alternative, where the uniform carbon isotope

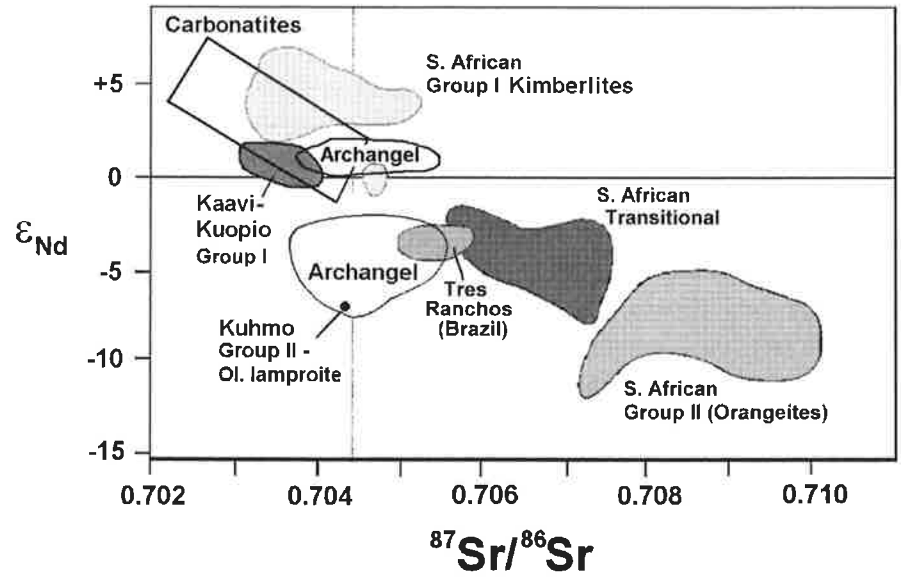

Figure 2. $^{87/86}Sr$ vs. ε_{Nd} diagram of the whole rock samples from kimberlites from Finland and other kimberlite provinces. Carbonatite box includes data from all carbonatites < 200 Ma old. Vertical and horizontal lines indicate the Bulk Earth isotopic compositions. Data from Smith (1983), Harmer (1999), and Mahotkin et al. (1999).

composition reflects the composition of the well-mixed sublithospheric convective mantle, is considered more likely. Preliminary initial Sr-Nd isotopic data on calcite samples from the Siilinjärvi and Sokli carbonatites are near those measured from the Finnish kimberlites. However, if an additional Sr-Nd substantiates a depleted signature for carbonatites ranging in age from 2.6 to 0.3 Ga, then it is more likely that the carbonatite sources (perhaps at the depth of 100 km) formed relatively close to the time of emplacement. In this case the constancy of the carbon isotopes is just a reflection of a uniform asthenospheric carbon reservoir from which the lithospheric carbonatite sources have been formed. This would imply that the carbonatite data provide very little information about the isotopic composition of the carbon stored for long periods in the Karelian SCLM, which at the correct depth, may be in the form of diamond.

3. The Mantle Xenolith Testimony: Origin of 250 km Thick Continental Roots

Three types of mantle xenoliths have been recovered from the kimberlite pipes of the Eastern Finland Kimberlite Province: garnet-spinel peridotites, garnet peridotites, and bimineralic diamondiferous eclogites. *Garnet-spinel peridotites*, most of which originate from depths of ca. 100-150 km, are all highly depleted fine grained granuloblastic and equigranular harzburgites. They have metasomatic overprint with the development of minor hydrous phases. *Garnet peridotites* are compositionally and texturally distinct from the garnet-spinel peridotites. They are of deeper origin (170-230 km), consist of coarse grained harzburgites, lherzolites and wehrlites and are all texturally similar to SCLM of other cratons such as Kaapvaal and Siberia. However, their Nd and Sr isotopic composition is not typical for Archaean SCLM, being more akin to off-cratonic lithospheric peridotites. *Eclogite xenoliths* are bimineralic eclogites. They are derived from depths of ca. 250 km which is comparable to the depth of origin of the deepest garnet peridotites *(Peltonen et al. 1999).*

Probably the most important contribution of the mantle xenolith study is that the SCLM within the central Fennoscandian Shield is compositionally and texturally heterogeneous. The sketch illustrated in Figure 3. attempts to combine the xenolith data with what is currently known from the geodynamic evolution of the craton margin (e.g. Korsman et al., 1999). The uppermost part of the SCLM of both the Karelian and Svecofennian domain most likely represents the lithospheric mantle which was isolated from the convecting mantle at the time of the formation of the overlying crust. The present geometry of the boundary between the Karelian and Svecofennian upper SCLM is determined by the initial rifting of the craton at ca. 2.0 Ga ago and by a subsequent accretion of the Svecofennian oceanic lithosphere on to the craton margin. This boundary is likely to be almost vertical because the deepest harzburgite xenoliths originate from depths of ca. 150 km (Fig. 3).

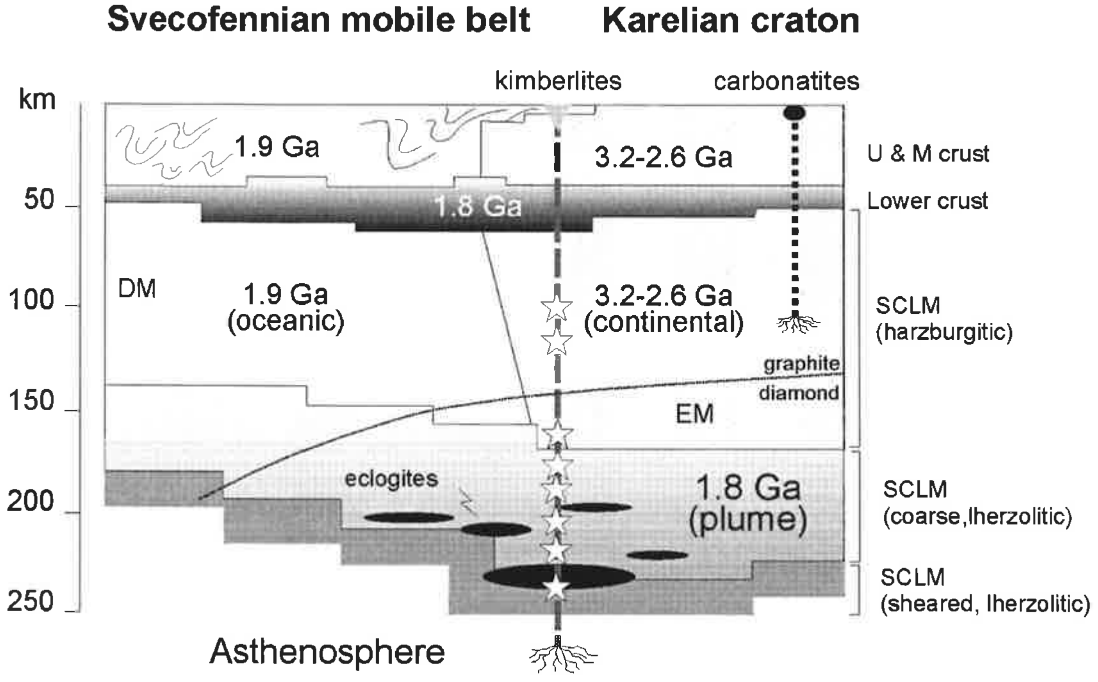

Figure 3. Simplified sketch illustrating the possible configuration of the major lithospheric domains within the central Fennoscandian Shield across the Proterozoic–Archaean transition. The stabilization age of the upper SCLM is believed to broadly coincide with the formation ages of the overlying crustal parts of the lithosphere. Lower part of the SCLM has distinct composition: it is believed to represent younger, ca. 1.8 Ga old, postorogenic addition to the base of lithosphere. This postorogenic growth of the lower SCLM is suggested to be coevel with generation of the mafic lower crust and postorogenic granites in the upper crust. Lowermost sheared lherzolite layer of the SCLM is not represented in the studied xenolith suite but its presence is based on the inference from other cratons worldwide. Mostly, this lowermost layer is likely to represent asthenospheric melt -filtrated 1.8 Ga peridotites, but even younger additions, related to the e.g. 1.6 Ga rapakivi and 0.5 Ga kimberlite magmatism remain a possibility. Location of the Eastern Finland Kimberlite Province and the approximate locus of origin of the studied mantle xenoliths (white stars) are indicated.

The lower SCLM of both the Karelian and Svecofennian domains is distinct in texture and composition from that of the upper SCLM (Fig. 3). As implied by the xenolith petrography and thermobarometry, the lower SCLM is composed of relatively fertile coarse grained garnet peridotites with some diamondiferous eclogite pods at its base. No direct age determinations are available, but U-Pb datings of kimberlitic and lower crustal zircons bear evidence that the formation of the lower lithospheric mantle and mafic lower crust and the emplacement of postorogenic 1.8 Ga granites can all be related *(Hölttä et al., 2000)*. This can be explained by a model where a plume impinged on the base of the 3.2-1.9 Ga old lithosphere at the postorogenic stage. This resulted in thermal erosion of the pre-existing lithosphere and its replacement by the new material originating from the 1.8 Ga old plume head.

The xenolith data thus implies that the SCLM beneath the Karelian craton margin is at least 230 km thick *(Kukkonen and Peltonen 1999)*. This should, however, be considered as a minimum estimate since the deepest xenoliths originating from this level are still typical coarse grained lithospheric peridotites with no evidence for the vicinity of the lithosphere–asthenosphere boundary. Although no "sheared" xenoliths are present in our sample suite it is reasonable to assume – by inference with xenolith studies from other cratons (*Boyd, 1987*) – that the coarse garnet peridotites are underlain by a layer of "sheared" peridotites with mylonitic textures. Traditionally such sheared lithospheric peridotites have been considered to be of asthenospheric origin *(Nixon and Boyd 1973)*. Recent Re-Os isotope results have, however, indicated that in most cases they give ancient formation ages and that they thus actually represent lowermost parts of the ancient cratonic roots which have been infiltrated by asthenospheric melts *(Pearson 1999)*. Therefore, we conclude that the true lithosphere – asthenosphere boundary within the Karelian craton margin lies some tens of kilometres beneath the maximun depth of 230 km indicated by the xenolith thermobarometry.

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Rheological Structure and Numerical Geodynamical Modelling in the Fennoscandian Shield

Kari Moisio and Pertti Kaikkonen

Department of Geophysics, P.O. Box 3000, FIN-90014 University of Oulu, Finland
E-mail: kari.moisio@oulu.fi

Rheological modelling can be used to estimate the state of the deformation in the lithosphere. One-dimensional models give approximations for the brittle and the ductile failure, the mechanical strength and the decoupling of the crust and mantle, for instance. Geothermal models are required and important, as ductile mechanisms are strongly dependent on the temperature.

Keywords: Fennoscandian Shield, structural analysis, finite-element analysis, thermal analysis, rheology

1. Background

Rheological properties control the deformation of the lithosphere. They are dependent on the intrinsic and extrinsic parameters, such as the temperature and the composition of the rock. Lithosphere is often divided into two mechanically different regions where main forms of deformation are the brittle and the ductile ones. In the brittle region, deformation is a result of the fault and crack movement, whereas in the ductile region, material deforms by creep i.e., slow plastic flow. These mechanisms are modelled with the known rheological laws derived from laboratory experiments. By using these laws to calculate the required stress to achieve deformation, we obtain the rheological strength, i.e., stress difference necessary for deformation to occur. This strength is defined as a difference between maximum and minimum principal stresses and it is calculated both for the brittle and the ductile mechanisms and the lowest one of these gives us the relevant (yield) strength at relevant depth. Brittle deformation is often described by the Byerlee's law *(Ranalli, 1995),* which gives the stress difference necessary for the brittle fracture to occur. Ductile steady-state (constant rate of flow under constant stress) properties of a material can be described by an empirical constitutive equation, which often has a power-law dependence between the stress and strain rate. At high stresses and temperatures this power-law creep *(Kirby, 1983; Ranalli, 1995)* is dominant and the ductile flow law gives the stress difference necessary to maintain a given strain rate.

2. Rheology of the Lithosphere in Finland

Several deep seismic sounding (DSS) profiles located in the Fennoscandian Shield were used in studying the present-day thermomechanical structure of the Fennoscandian Shield *(Kaikkonen et al., 2000)*. These profiles are located in different tectonic units, which represent different stages in Precambrian crustal and lithospheric growth. We constructed several lithospheric rheological strength profiles and analysed their implications and consequences on the Fennoscandian Shield. We first calculated the present-day one-dimensional geotherms from which the strength envelopes i.e., strength profiles for a certain faulting type and strain rate, were obtained. These profiles were mainly calculated for a compressional, i.e., thrust fault regime, and for both the dry and the wet conditions. We analysed the results with the help of different rheological parameters, such as the integrated crustal (ICS) and lithospheric strengths (ILS), the depths of the mechanically strong crust (MSC) and lithosphere (MSL), the rheological thickness of the lithosphere (RHTL) and the brittle-ductile transition depth (BDT)

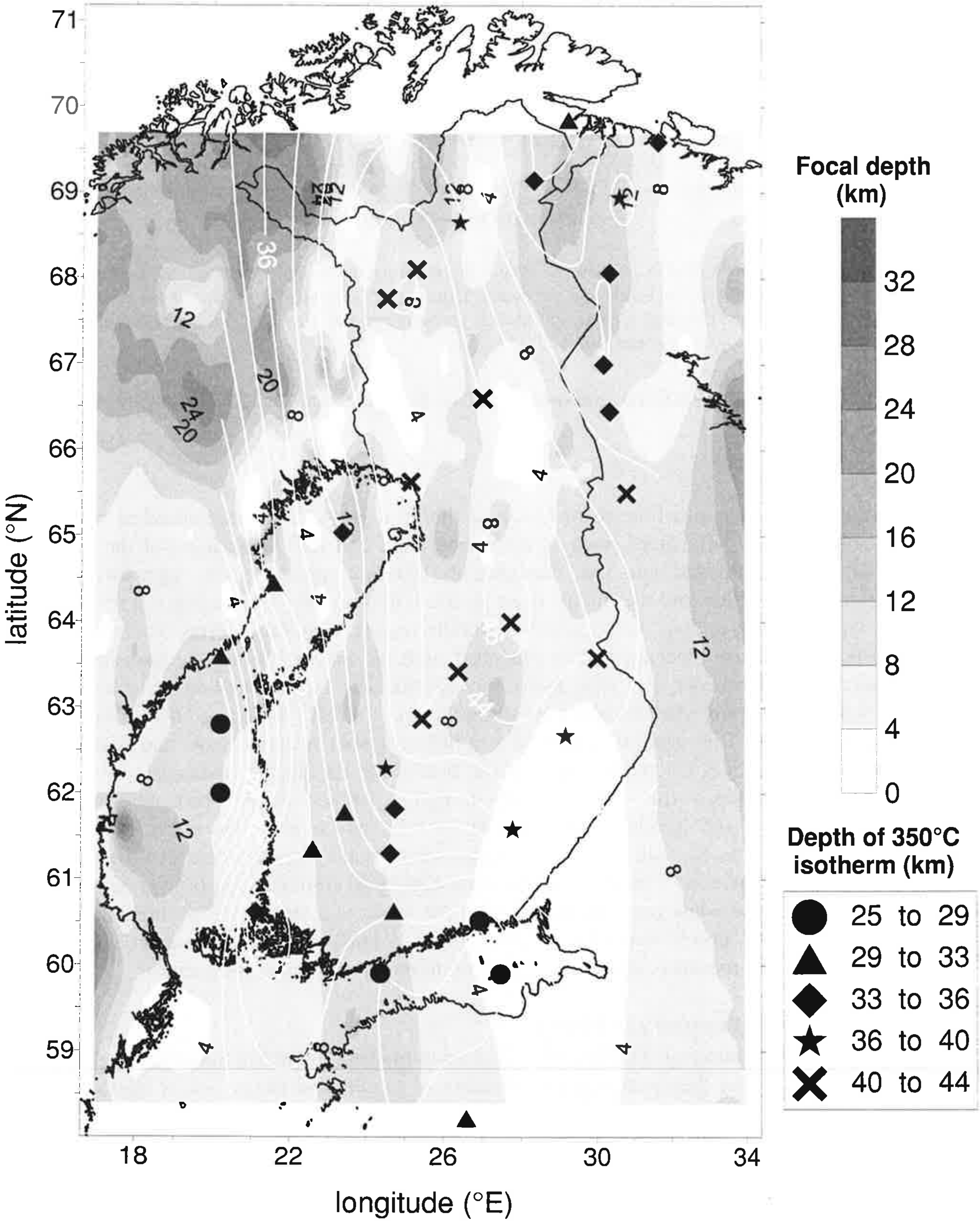

Figure 1. The grey-tone map of earthquake focal depths in kilometers, white contour lines indicate the depth of the mechanically strong crust (MSC- depth where the ductile strength is less or equal to 50 MPa) in kilometers for the compressional wet rheology and the depth of the 350ºC isotherm in kilometers shown in symbols *(taken from Kaikkonen et al., 2000).*

(for details see *Kaikkonen et al., 2000*). These imply that the lithosphere in the central Fennoscandian Shield can be considered rheologically rather strong. The results show lateral variations in the thermomechanical structure reflecting the geometry of the lithosphere and following roughly the same trend as the geochronological development of the Fennoscandian Shield, i.e., Archaean areas are stronger than the younger and warmer Proterozoic areas. The mechanical structure shows distinct decoupling of the weak lower crust and the strong upper mantle, especially in the case of a wet crustal rheology.

The rheological parameters and the earthquake focal depth data *(Ahjos and Uski, 1991; 1992)* were jointly analysed. The BDT depths for dry and wet crust are in the ranges of 15 to 40 km and 11 to 32 km, respectively. The focal depths coincide better with the wet than the dry BDT depths. The dry model yields higher BDT depth values and does not fit well with the depth limit of 31 km of the earthquakes. If we assume a frictional transition (velocity weakening/velocity strengthening) temperature of 350°C, which is located at the depths of 25-44 km with an average value of 35 km, it is in good agreement with the focal depth limit of 31 km (Fig. 1). Consequently, it seems that the velocity weakening/velocity strengthening explains best the real lower boundary of earthquake occurrence.

3. Finite-element Modelling

We have applied finite-element modelling to the two DSS profiles BALTIC and SVEKA in central Finland in order to study the state of stress and the possible deformation of the deep structure. The finite element method was used to solve the numerical problem for two-dimensional lithospheric structural models constructed from existing seismic and thermal data (*e.g. Grad and Luosto, 1987; Luosto et al., 1990; Kukkonen and Jõeleht, 1996; Yliniemi et al., 1996; Kukkonen, 1998*). Rheological strength profiles were used as non-linear material parameters in the elasto-plastic model where the creep strength was translated into a plastic yield stress. Different tectonic load cases, composition, material properties and thermal conditions were analysed. For the BALTIC profile we used a granitic composition for the upper and a dioritic composition for the lower crust. The mantle was assumed to be homogeneous and to be composed of olivine. In the SVEKA profile, wet granite and wet diorite were used to for the upper and the middle crustal lithologies, respectively, and dry diabase or dry mafic granulite for the lower crust. The mantle was assumed to be composed of an olivine.

Figure 2 shows the stress intensity and the stress ratio in the BALTIC profile. The stress intensity is a parameter defined as the largest of the absolute values of the principal stress differences and it depicts the overall state of stress in the structural model. The stress ratio is defined as a ratio between the equivalent stress and the yield stress. It has values larger or equal to one (≥ 1), when yielding is occurring, and values less than one (<1), when the stress state is elastic or near to a failure. For the BALTIC profile, the numerical modelling *(Moisio et al., 2000)* revealed that with a wet rheological model the lower crust results in plastic deformation along the entire profile. This ductile layer separates mechanically the crust and the mantle from each other. The largest stress intensities are found in the lower crust. The dry rheological model resulted in slightly different results with no significant plastic deformation. For the SVEKA profile *(Moisio and Kaikkonen, 2000),* we applied methodologically similar kind of analysis and also studied if there is any connection between the deep electrically conductive structures and the ductile lower crust. Modelling, however, showed that the present-day mechanical and thermal conditions of the central Fennoscandian Shield do not seem to favour any plastic deformation in the SVEKA profile, and there is no connection

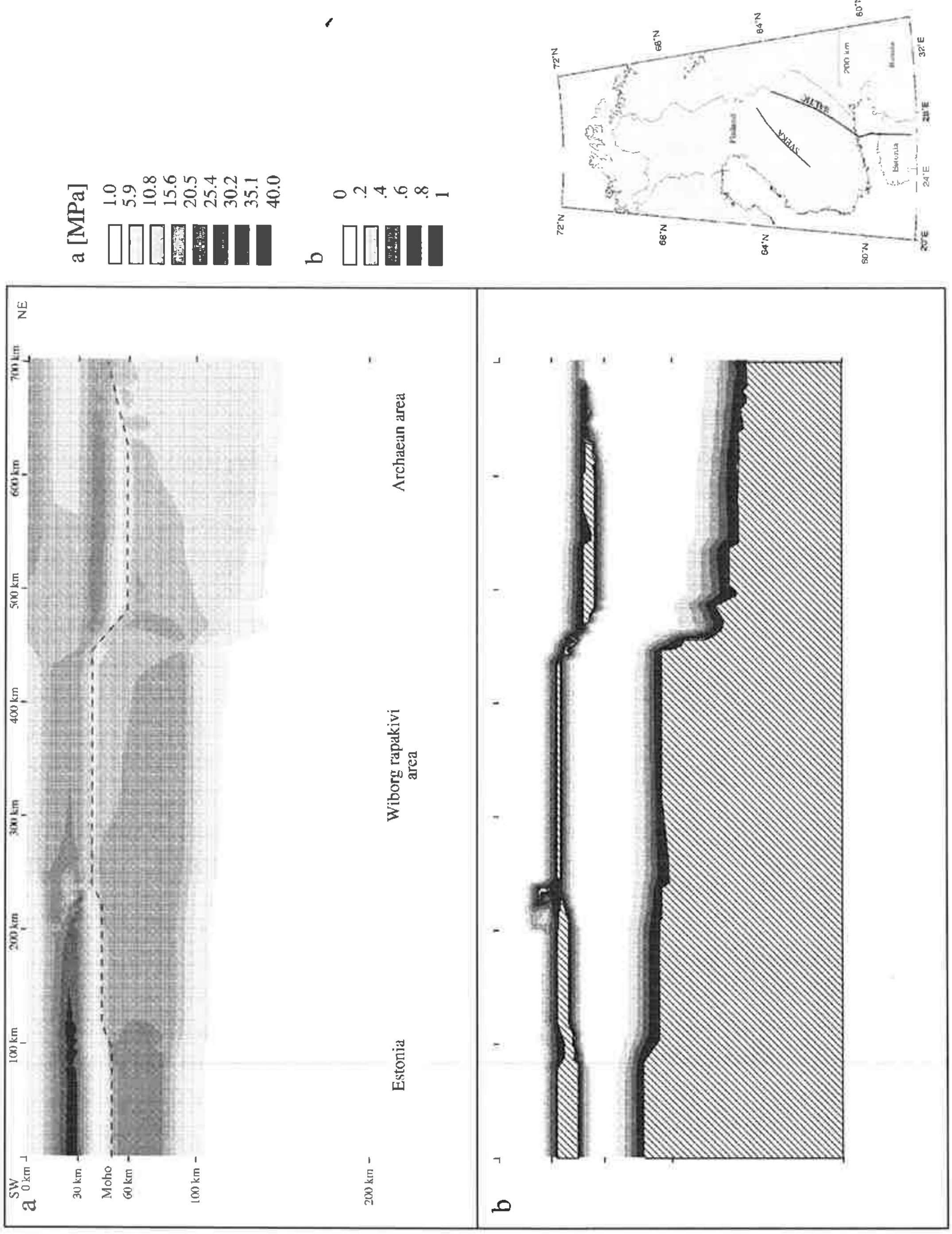

Figure 2. The stress intensity in MPa (a) and the stress ratio (b) for the elasto-plastic model of the DSS profile BALTIC with a wet crustal rheology. The stress ratio shows the yielded areas of the model with ratio higher or equal to one with rasterized areas *(taken from Moisio et al., 2000).*

between rheologically weak ductile layers and electrically conductive structures of the lower crust.

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Magnetic Signatures of the Finnish Bedrock and their Relationship with Geological Boundaries

Meri-Liisa Airo and Tapio Ruotoistenmäki

Geological Survey of Finland, P.O. Box 96, FIN-02151 Espoo, Finland
E-mail: meri-liisa.airo@gsf.fi

Structural bedrock provinces are characterized by their continuous aeromagnetic patterns which show a coherent magnetization level and follow the stratigraphical features of the structural province. The main mineral compositions within these provinces also obey typical overall rules so that the provinces can often be classified due to their distributions of petrophysical properties. The structural provinces are outlined by discordant zones that interrupt the stratigraphic markers. These zones can be interpreted to reach a great depth, and they also have a great lateral extent from several tens to hundreds of kilometers. Along these zones vertical and/or lateral movement of the bedrock provinces may also have taken place. Because of the related crustal scale shearing and/or fracturing, magnetomineralogical changes within these deformation zones are common and the zones are typified commonly by a weak magnetization, declined to different degrees. These deformation zones act as geological boundaries, which separate adjacent structural provinces and divide the bedrock into crustal blocks.

Keywords: airborne geophysics, Precambrian, deformation, magnetite, pyrrhotite

1. Airborne Magnetic Data and Maps of Finland

The airborne high altitude, low resolution magnetic data of Finland were measured by the Geological Survey of Finland (GTK) during the years 1951-1971 and 1993 (eastern border zone), with a terrain clearance of 150 m and a line spacing of 400 m. The original analog data were digitized to a 1 km by 1 km grid (see e.g. *Korhonen, 1980*). At present about 85% of onshore Finland is also covered by low-altitude (~35 m), high resolution data with a line spacing of 200 m, and observations made 4 to10 times /s. Copies of maps and grid matrices are available from the Geological Survey of Finland. A detailed up-to-date description of low-altitude flight data and configuration is available on the GTK www-page: ***http://www.gsf.fi/aerogeo/eng0.html.*** This paper summarizes shortly the magnetic characteristics which describe the Finnish bedrock in the aeromagnetic data.

2. Relationships between Magnetic Signatures and Geology

The benefit of magnetic surveys in geological studies is that they provide 3D information of the subsurface geology because magnetic anomalies arise from sources at all depths within the crust. Magnetic anomalies are produced by variable concentration of magnetic minerals in rocks, mainly magnetite and pyrrhotite in shield areas. The magnetic signatures reflect compositional and structural information about the bedrock, even in areas of weak magnetization. Coherent units of continuous magnetic patterns correspond to structural bedrock blocks. They are generally separated and outlined by weakly magnetic lineaments or zones, which mark their boundaries. Because magnetic interpretations are limited due to the non-uniqueness of potential field source distributions, the most important control on the reliability of magnetic models is information on magnetic properties, and understanding the factors that determine magnetization for the geological units.

3. Factors Influencing on Magnetic Anomaly Pattern

Magnetic properties. The induced and remanent magnetizations (J_i and J_r, respectively) and their relationships in rocks depend on the amount of magnetic material and its composition, magnetic grain size, grain shape and grain texture. The age of remanent magnetization has a minor effect on the magnetization intensity unless the amount of the ferromagnetic minerals is large (e.g., unless the volume % of magnetite is at least > 0.4). Together J_i and J_r are responsible for the anomaly amplitude, and their ratio $Q = J_r/J_i$ (Königsberger ratio) expresses the relative importance of remanent magnetization for magnetic signature by affecting the anomaly shape (*Airo, 1993; 1995*). When $Q < 1$ (the induced magnetization is dominant) the magnetic anomalies form typically smoothly continuous patterns. On the other hand, $Q > 1$ indicates that remanence dominates induced magnetization. Then the anomalies show generally irregular patterns with a chained alteration of magnetic highs and lows.

Magnetic mineralogy. The observed high remanence values are related either to pyrrhotite or to a small magnetite grain size or its lamellar structure. Also the irregular, broken shape of magnetite grains seems to increase the remanence. Pyrrhotite anomalies are distinguished from magnetite anomalies by their combination with electrically conductive horizons.

Geometrical factors, such as volume, dip and strike of magnetic bodies as well as their depth of the upper surface influence on the magnetic anomaly patterns. Deep structures produce long wavelength anomalies, while shallow ones are recognized by their short wavelength anomalies. Folding is shown as curved, banded magnetic patterns, but the long-wavelength regional folding results in symmetrical variation of magnetic field.

The effect of ***deformation*** on magnetic signatures depends on whether J_i or J_r were changed during deformational processes. This is further a function of the volume and composition as well as grain size and shape of the newly formed magnetic minerals. For example, if secondary magnetite has small grain size, like small inclusions in mafic silicate grains, then the increased remanence may affect the anomaly shape.

Metamorphism has profound effects on magnetic properties of mafic rocks. Therefore changes in metamorphic grade accross a study area should be taken into account in magnetic interpretation. The changes in magnetic properties reflect the re-distribution of ferric iron Fe^{3+} during metamorphism (see also *Airo,1999a*). Magnetite formation is favoured by high total iron contents and an intermediate redox state / oxidation ratio.

Hydrothermal alteration may greatly modify the magnetic properties, owing to the creation or destruction of ferromagnetic minerals. For example, serpentinisation of olivine-rich ultramafic rocks tends to produce abundant magnetite. Likewise carbonate alteration redistributes magnetite, without destroying it. But intense talc-carbonate alteration demagnetizes the rock with iron entering magnesite at the expense of magnetite.

Fault and fracture zones are generally expressed as weakly magnetic lineaments cross-cutting geological units. The lacking of magnetic minerals within fault zones may simply result from erosion, but in deep large-scale fracture and shear zones hydrothermal alteration results also in changes in magnetic mineralogy. Faulting and thrusting post-date the main formation stage of magnetic minerals, but shearing is generally associated with simultaneous changes in magnetic mineralogy. That is why shear zones sometimes contain magnetic banded or spotted local anomalies. Shear zones and thrust faults are typical geological boundaries, which separate regions or geological blocks of different deformation and structure. Shear lenses related to low angle thrust zones are often observed in aeromagnetic data.

4. Tools for Analysing Magnetic Signatures and Their Relations

Visual examination of coloured or grey-shaded total field magnetic data allows a rapid general means for identification, characterisation and outlining of various geological units and their boundaries. These include the arrangement of magnetic patterns and abrupt terminations or displacements of these patterns, combined occasionally with extensive linear gradients. Structural features are typically enhanced by shaded relief presentations, particularly when using appropriate light angle. Application of various filtering and transformation processes may produce secondary products with improved information content.

Vertical derivatives of magnetic field enhance high frequencies relative to low frequencies and eliminate long-wavelength regional effects. The process resolves the effects of adjacent anomalies and gives peaks over the tops of sources. Source outlines are indicated by steep gradients and inflections. First vertical derivative data is a basic necessity in magnetic interpretation. Second vertical derivative has a more resolving power, but its application demands high-quality magnetic data since 'noise' is amplified to the same degree as shallow signal. The method is also sensitive to changes of survey height.

Horizontal derivatives in the x and y directions also enhance high frequencies and thus the magnetic response of shallow structural features or narrow bodies. The process is useful for mapping the continuity of anomalies, as well as for mapping body outlines since it produces anomaly peaks approximately located over the edges of wide bodies.

3D Analytic signal is a function related to magnetic fields by the derivatives. It is independent of both the direction of magnetization and the direction of the Earth's magnetic field. This means that all bodies with the same geometry have the same analytic signal. As the peaks of analytic signal functions are symmetrical and occur directly over the edges of wide bodies and directly over the centres of narrow bodies, analytic signal maps provide simple indications of magnetic source geometry. They also define source positions regardless of any remanence in the sources.

Upward continuation simulates flying the survey at a higher altitude because of computation of the magnetic field at higher level. The process smooths out high-frequency anomalies relative to low-frequency anomalies. This means that upward continuation is useful for suppressing the effects of shallow anomalies when details on deeper anomalies are required, and when outlining large-scale geological units and regional folding.

Petrophysical properties can be utilized both locally and regionally if good coverage of representative samples is available (i.e. a great number of samples representing both anomalous signatures and the background). The rock densities and magnetic properties are useful in describing and characterizing distinctive geological units or bedrock blocks, because the densities and magnetic properties reflect compositional variation in the rocks and their magnetic mineral concentration and texture (*Airo 1999b*).

5. Geological Boundaries

The following short description of the magnetic signature of the major geological features of Finnish bedrock is based on the low-altitude magnetic data by the Geological Survey of Finland (GTK). The data grid of the aeromagnetic map of Finland in Figure 1 has been compiled by *Arkimaa et al. (1999)*. The areas still lacking low-altitude data have been completed with the existing high-altitude data.

General descriptions of the Finnish bedrock are available in e.g. *Gaál and Gorbatschev. (1987), Gaál et al. (1989) and Korsman et al. (1999)*. The plate tectonic models in the following have been adopted mainly from *Ruotoistenmäki (1996)*.

AP - Archaean-Proterozoic boundary
EA - Eastern Finland Archaean block
IA - Iisalmi Archaean block
PA - Pudasjärvi Archaean block
NA - Northern Finland Archaean block
SFS - Southern Finland Svecofennian block
KA - Kainuu schist belt

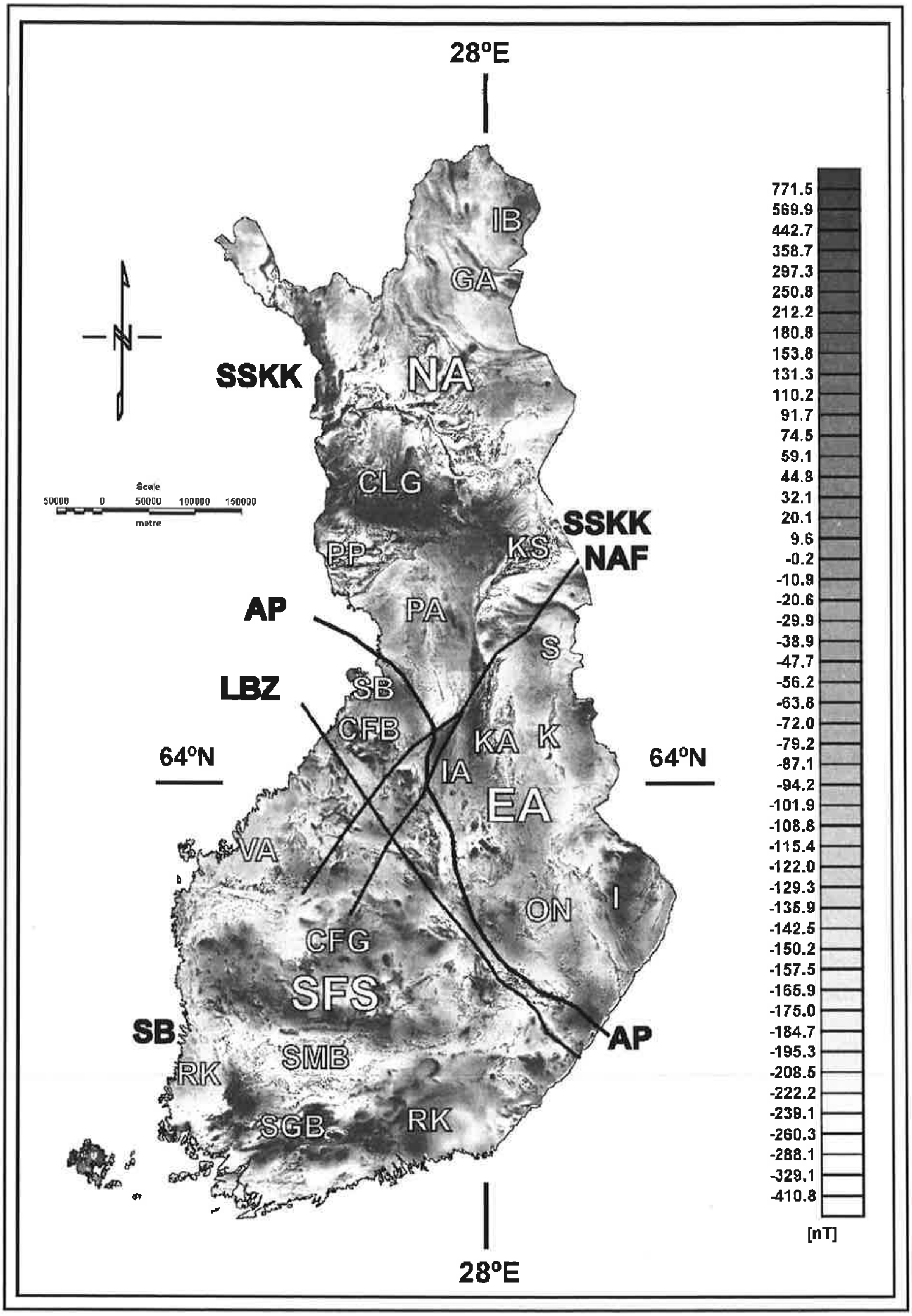

Figure 1. Aeromagnetic map of Finland. The regional geological features are considered in the text.

NAF - Proterozoic Näränkävaara - Auho fracture zone
ON - Outokumpu Nappe
GA - Granulite Arc
IB - Archaean Inari block
SSKK - Salla-Sodankylä-Kittilä-Karasjokk greenstone belts
CLG - Central Lapland granitoid complex
PP - Peräpohja supracrustal belt
KS - Kuusamo supracrustal belt
LBZ - Ladoga-Bothnian Bay fault zone
CFG - Central Finland granitoid block
CFB - Central Finland back arc basin
SMB - Southern Finland migmatite belt
SGB - Southern Finland granitoid belt
VA - Vaasa arc
RK - Rapakivi intrusions
SB - Satakunta and Muhos sedimentary basins

5.1 The Archaean-Proterozoic Boundary (~1.9 Ga)

The geological boundaries are generally visible on magnetic maps as borders of areas characterized by varying magnetic anomaly intensity and / or patterns. These areas generally correspond to well-defined lithological blocks - often of different geological nature. Typically, the block boundaries are also fractures and faults or shear zones of local or regional scale from kilometers to hundreds of kilometers. They appear on magnetic maps mostly as regions of decreased magnetization, which disrupt or outline the continuous or coherent magnetic patterns of lithological blocks. The depth of the largest faults can reach the lower crust.

The major block boundary in the Finnish bedrock is the Archaean-Proterozoic boundary (AP, in Fig. 1) separating the Archaean block(s) in the East from the Proterozoic blocks in the West. The AP boundary has been roughly adapted from the Finnish bedrock map by *Korsman et al. (1997)*. It represents tectonically the 1.9 Ga old Svecofennian collision zone between island arc provinces and Archean Karelian craton. The Svecofennian part is magnetically more variable than the Karelian craton and characterized by more "plastic" signatures. The Karelian craton reveals patterns of more "brittle" nature. The collision zone AP is visible as regional and local scale shear zones shown as magnetic minima. Active movements are still going on at some places.

5.2 Regional Block Boundaries

The major lithological blocks of Finland can be further divided into the Eastern Finland (EA), Iisalmi (IA), Pudasjärvi (PA), and Northern Finland (NA) Archaean blocks and into the Southern Finland Svecofennian block (SFS). These are considered in more detail in the following.

Eastern Finland Archaean blocks. The AP collision-zone separating the eastern Finland Archaean block (< 3.2 Ga) from the southern Finland Proterozoic rocks is covered by early Proterozoic Kainuu schist belt (KA, ~ 2.2-1.9 Ga). The schist belt is broken by numerous Svecofennian (ca. 1.9 Ga), collision related, wedge shaped, thrust fractures. These have effectively complicated the stratigraphic image of the schist belt. In magnetic data the belt is a profound magnetic maximum where a significant component is due to pyrrhotite bearing black schists (metamorphosed black shales) and magnetite rich metavolcanites and serpentinites. The schist belt is cut by the sinistral Proterozoic Näränkävaara - Auho fracture zone (NAF). The vertical component of the fracture zone was interpreted in the Näränkävaara area (in NE part of the zone) to be at least one kilometer in dimension (eastern part uplifted relative to the western side). The course of the zone is visible on magnetic maps as discontinuities in Kainuu and Näränkävaara areas.

The Archaean Ilomantsi - Kuhmo - Suomussalmi greenstone belts (I, K, S, ~ 3.0-2.7 Ga) have been squeezed into wedge shaped rift-type fractures within the surrounding granodioritic crust. The greenstone belts form distinct, narrow, high-amplitude linear anomalies, cut by numerous younger fractures. On regional scale they form a border zone where the regional magnetic and gravimetric anomaly level of the surrounding crust increase abruptly from west to east. This reflects vertical movements along the fractures which control the present picture of the greenstone belts.

The Outokumpu Nappe (ON, ~ 1.95-1.92 Ga) overlies the SE part of the Archaean EA block. It forms curvilinear and linear anomaly zones which are mainly due to magnetite bearing serpentinites and pyrrhotite bearing black schists. These zones reflect the course of the outcropping sections of the nappe complex.

Northern Finland Archaean block .The northern Finland Archaean block (< 3.2 Ga) is largely covered and cut by Proterozoic volcanic, sedimentary and intrusive rocks having generally a strong Archaean isotopic (ε_{Nd}) component. The ca. 1.9 Ga old Granulite Arc (GA), south of the Archaean Inari block (IB), is characterized by curvilinear magnetic anomalies due to magnetite containing mafic layers in granulite beds. The magnetic pattern reflects north-south trending thrust structures which uplifted the granulite belt almost simultaneously with the Svecofennian collision in the south.

The Salla-Sodankylä-Kittilä-Karasjokk greenstone belts (SSKK, ~ 2.5-2.0 Ga) form a complicated high anomaly zone north of the central Lapland granitoid complex (CLG, ca. 1.8 Ga). CLG is characterized by an exceptionally intense regional magnetic (and gravity) anomaly zone. The regional granitoid anomaly is decorated by mainly NW and NE directed, linear, 100 km long fractures. Geochemically the granites have a relatively low iron content in silicates and a high iron oxide (magnetite) content (*Puranen, 1989*). The southern part of CLG is outlined by regional synforms of the Peräpohja (PP) and Kuusamo (KS) supracrustal belts which contain early Proterozoic, highly magnetic layered intrusions (~ 2.5-2.4 Ga). PP and KS are shown as strongly folded triangular magnetic anomaly zones to the SW and SE of the CLG. Geometrically the entire structure - from Salla-Sodankylä-Kittilä-Karasjokk greenstone belts to Peräpohja-Kuusamo belt and Central Lapland granitoid complex in between - can be explained by a relatively simple interference structure, formed by large scale E-W and NE-SW trending regional folds. The fold structures possibly traverse through the whole crust. This is reflected as regional magnetic and gravity maxima corresponding to the region of CLG, surrounded by regional minima to the North and the South.

Southern Finland Svecofennian blocks (< 1.9 Ga) The Archaean-Proterozoic boundary (AP) is very sharp. On its western side no significant Archaean isotopic (ε_{Nd}) components in igneous rocks have been observed – opposite to as is common to the Proterozoic rocks on the eastern side of AP. The boundary zone is also characterized by a very thick crust (down to 60 km) which is stabilized by highly metamorphosed, high density mafic intrusions and granulites on its both sides. They have been uplifted along NW-SE wedge shaped vertical faults during the Svecofennian collision. On magnetic maps these highly metamorphosed units emerge as high-amplitude, roundish magnetic anomalies.

Roughly parallel to AP runs the dextral Ladoga-Bothnian Bay fault zone (LBZ, < 1.9 Ga). LBZ is characterized in magnetic data as numerous linear NW-SE trending brittle fractures. As interpreted from gravity data, LBZ is regionally 10-20 km wide. It is relatively ductile, formed during the early stage of the S-SW directed Svecofennian collision. The brittle faults shown on magnetic maps reflect late activity of LBZ, partly active even today.

The Svecofennian crust between the AP and LBZ zones forms a significant ore potential belt in Finland. It can be interpreted as a relict of a back arc basin (CFB, 1.9-1.87 Ga old) of the Central Finland island arc granitoid block (CFG, 1.89-1.87 Ga) on its western side. Besides the mafic and granulitic rocks, the CFB is also characterized by felsic and mafic volcanites, intrusions and black schists which give the belt a very complicated magnetic anomaly pattern (Fig. 1).

The magnetic anomaly pattern of CFG is less complicated, which is typical of blocks dominated by felsic intrusions. In the central parts of CFG, more mafic, magnetite bearing intrusions are shown as roundish magnetic highs. The southern part of CFG forms an E-W trending anomaly zone containing more complicated magnetic structures. It is also characterized by a very thick crust (ca. 60 km). This is interpreted to be due to the north trending Proterozoic collision of the southern Finland migmatite (SMB) and granitoid belts (SGB) at about 1.9-1.8 Ga ago. SMB and SGB are interpreted as a back-arc basin and island arc, respectively. The regional magnetic anomaly amplitude of the SMB is distinctively lower than that of the southern part of CFG. Geochemically the intrusions within SMB are more iron-rich than those in the southern CFG. However, in SMB the iron is concentrated into silicates while CFG contains more iron oxides, mainly magnetite. The complicated magnetic signature of SMB is largely due to pyrrhotite bearing metasedimentary rocks (mainly black schists), which have been folded during the collision(s). The southern granitoid belt is characterized by highly magnetic mafic intrusions, which are cut by sharp and distinct faults. The large amount of mafic intrusions in SMB and in SGB have also acted as a stabilizing factor in the formation of the thick crust. The SMB and SGB arc complex has collided 'vertically' against CFG from south. However, in the region of the Vaasa arc (VA), the thrusting has been from the west.

Post-orogenic , ~1.6 Ga old rapakivi intrusions (RK) are mainly observed in aeromagnetic data as homogeneous low magnetic anomaly zones with sharp boundaries. Similar signatures characterize also the Satakunta and Muhos Mesoproterozoic (1.4-1.2 Ga old) sedimentary basins (SB).

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Fennoscandian Crustal Model as Reflected by the Petrophysical Interpretation of the Potential Field Anomalies

Juha V. Korhonen and Heikki Säävuori

Geological Survey of Finland, P.O. Box 96, FIN-02151 Espoo, Finland
E-mail: juha.korhonen@gsf.fi

The geological, geophysical and geodetic institutions of Fennoscandia and nearby areas are jointly compiling for the first time digital maps in scale 1:1,000,000 on the lithology, magnetic anomalies and Bouguer anomalies. The aim is to present these data in a coherent way, with a similar information structure and with comparable processing methods. The maps and databases will be ready in 2001.

The initiative to measure and geologically interpret geophysical data of the area has continued for more than fifty years. The research volume on the correlation between geophysical anomalies, petrophysical properties and geological structures is increasing. Hence a considerable and useful investment on the geophysical information has been made. Although a lot of high quality, raw and reduced data already exist, their major usage still waits to come. A virtual crustal model has been proposed as a more effective way to organize and combine the old and new information.

Keywords: crust, magnetic anomalies, Bouguer anomalies, physical properties, Fennoscandia

1. Introduction

The requirement of systematic regional interpretation of the aerogeophysical and gravity data sets is the background reason for the Crustal Model Program for Finland. The aim of the program is to collect and homogenize reliable base material (e.g. Fig. 1.), to make it easily accessible for interpretation, to develop petrophysically based applications for potential field interpretations, especially to decrease its ambiguity, and to produce structural end evolutionary models. These and interpretation examples would be gradually joined together by the geographic information systems (GIS), and would become a virtual reality: a computer based geological model in three dimensions (3D) and time combined with an interactive display, accessible in the Internet.

The project arranges workshops that are discussion and meeting forums for the professionals working with the crustal model problems in the Fennoscandian Shield and in nearby areas. The First Workshop was held on the petrophysical interpretation of the potential fields at the Geological Survey of Finland (GTK) at Espoo in 1997 *(Korhonen, 1997)*. The theme of the Second Workshop combined seismic and potential field interpretations and it was held as a part of the EAGE meeting in Helsinki in 1999 (*Korhonen, 1999*). The Third Workshop is planned to be held in 2001 at Espoo and it is aimed to emphasize the geological interpretation of the geophysical data.

2. The Problem Setting in the Interpretation

The geological interpretation of the potential field anomalies relies on the knowledge of the composition, structure and evolution of the bedrock by using information on its physical properties. In each study area we should ask the following questions:

- What are the minerals, rock types and stratigraphic and tectonic units associated with geophysical anomalies?
- Which components of the anomalies are caused by the near surface parts of outcropping geological units and which are caused by their downward extensions? And further more, what are the contributions from the deep, unexposed sources, and what causes them? What is possible to known of their geometry?

- What are the geological and physical processes that have influenced the present day physical properties?
- What is the general spatial and temporal overview of the bedrock that can be deduced from geophysical anomalies, petrophysical properties and geological data (*Korhonen et al., 1997*)?

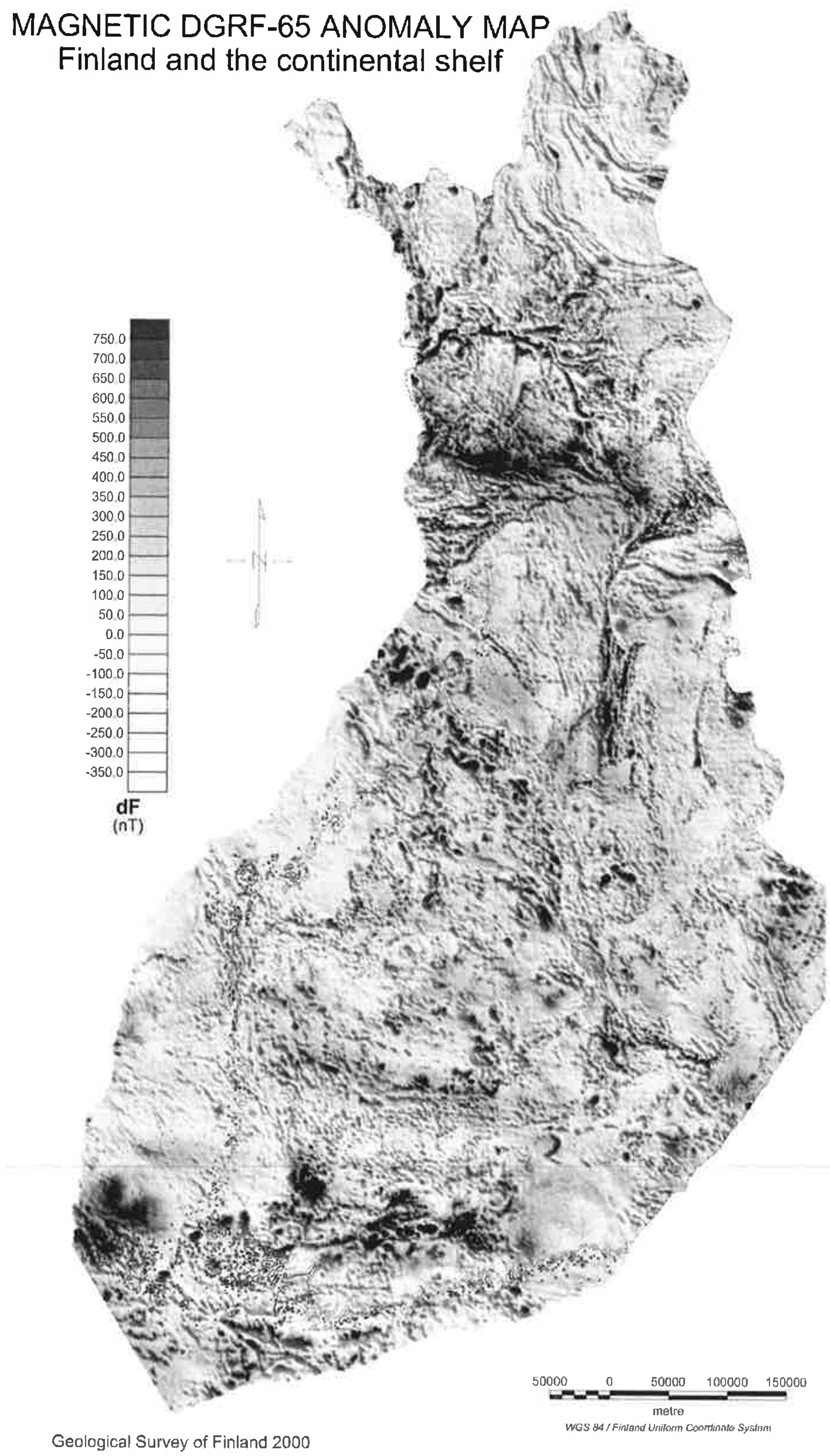
3. Finnish Petrophysical Data

Airborne geophysical surveying started in 1951 in Finland. Soon it was realized that precise data on the petrophysical properties of rocks of the survey areas are required for geological interpretation of the geophysical anomalies. Hence the measurements of the physical properties of rock samples were started in 1953, and more systematic measurement on regional sample sets and drill cores in 1963. The measured properties are density, magnetic susceptibility, and intensity of remanence plus - optionally - the direction of remanence, porosity, electrical conductivity, induced polarization effect, and seismic P-velocity. In order to have a more spatially representative data set, a petrophysical program covering all of Finland was initiated in 1980. The petrophysical laboratory was computerized in 1982 and two new laboratories began in the regional offices of the Geological Survey at Kuopio and Rovaniemi. To manage the accumulating information the first petrophysical database was established in 1973. The present relational database (RDB) consists of 133 000 records (*Puranen et al., 1968; Puranen, 1989; Korhonen, 1992; Säävuori and Hänninen, 1997*). A final report of the mapping program is in preparation. The most direct use of the petrophysical data base is to provide material constants for the interpretation of geophysical anomalies. Together with the magnetic and gravity anomaly maps, the petrophysical properties have been used to develop stratigraphic, tectonic and evolution models and to identify candidate structures of possible extraterrestrial origin (e.g. *Airo, 1999; Korhonen et al., 1997 and references therein; Wennerström and Airo,1998*).

4. Global Petrophysical Data

The magnetic and gravimetric methods are extensively used for geological studies worldwide. In all scales, the usefulness of the geophysical data is considerably increased if combined with petrophysical data. The results from a global questionnaire study among the petrophysical indicates that most of the current digital sample data is clustered in the Fennoscandian Shield where it exhibits an average data density of 30 measurements/100 km^2 of magnetic susceptibility and bulk density. Combined with the geological data it allows an estimation of the average upper crustal magnetization for interpretation of the Gulf of Finland, Kiruna and Baltic MAGSAT magnetic anomalies and the corresponding magnetic provinces in the more near surface studies. For most of the ca. sixty magnetic anomalies observed at satellite altitudes, there is only little petrophysical control on. Literature study indicates that it is likely that more petrophysical data exist at various geological institutions. Hopefully these data can be added to the databases. The Fennoscandian Shield is, however, an exceptional area in the sense that digital petrophysical data can easily be applied in the geological interpretation of the small-scale geophysical anomalies (*Korhonen and Purucker, 1999*).

Figure 1. The aeromagnetic anomaly map of Finland and adjoining continental shelf. The anomaly values are calculated by first by reducing the total field measurements to 1965.0 and then subtracting the main field as represented by the (Definitive) International Reference Field (DGRF-65) (*Korhonen et al., compilation in progress*).


MAGNETIC DGRF-65 ANOMALY MAP
Finland and the continental shelf
750.0
700.0
650.0
600.0
550.0
500.0
450.0
400.0
350.0
300.0
250.0
200.0
150.0
100.0
50.0
0.0
-50.0
-100.0
-150.0
-200.0
-250.0
-300.0
-350.0
dF
(nT)
N
50000 0 50000 100000 150000
metre
WGS 84 / Finland Uniform Coordinate System
Geological Survey of Finland 2000

5. Regional Geological Background of the Shield

The main geological entities of the Fennoscandian Shield evolved and stabilized in the Palaeo-Mesoproterozoic. Less than 20 per cent of the Shield originates from the Neoarchaean and has been affected by the Palaeoproterozoic Svecofennian Orogeny. Neoproterozoic mafic and Phanerozoic alkaline magmatism added minor masses to the uppermost part of the Shield. The Shield was at least partly covered by sediments in the Palaeozoic.

The Precambrian crystalline rocks of the Shield continue below the Neoproterozoic-Phanerozoic sediments of the East European craton, establishing the Fennosarmatian Shield. It is not known, what factors caused the Precambrian rocks of the Shield to be exposed at the upper surface of the crust while the platform area has remained covered by Phanerozoic sediments? What is the reason for the updoming of the Shield?

Another problem is the shield-wide correlation of the geological units across the areas covered by water and across the national borders. Complications arise from the variation in the geological nomenclature, the definitions of the units and the evolution theories that differ from country to country. The main emphasis on the geoscientific studies has been on the regional geology and the main mineral potential areas inside each country. Geological correlation of the whole Shield has been the responsibility of individual geologists rather than the joint task of the major organizations of the area.

6. The Fennoscandian Map Project

To see the Fennoscandian Shield geologically as one entity seven NW-European organizations started a joint project to merge the numerical geological, geophysical and petrophysical data on a scale of 1:1 million in 1997. The participants are the Geological Surveys of Finland, Norway and Sweden, the Northwest Department of Natural Resources of Russia, with its subsidiary companies, the Finnish Geodetic Institute, the National Land Survey of Sweden and the Norwegian Mapping Authority. The Geological Survey of Finland (GTK) coordinates the project.

The practical aim is to compile a general overview of the geological structure and evolution of the Shield area in a digital format. The main geophysical goals include the compilation of uniform magnetic and gravity data sets and maps, the use of petrophysical data in geological interpretation of the sources of the major geophysical anomalies, to trace them to depth, and to identify unexposed anomaly sources. All the geophysical information is to be used in the geological correlation and in establishing a geological evolution model of the area. This work has been preceded by several inter-Nordic and Finnish-Russian-(Estonian) joint projects starting from NORDKALOTT-project in 1980 (e.g. *Korhonen, 1989*).

Preliminary magnetic and gravity databases were established and preliminary maps compiled in a scale of 1:10,000,000 (*Korhonen et al., 1999*). In 1999 - 2000, the petrophysical data base was completed and the base map compiled (*Korhonen et al., 2000*). In the last year of the project, the final digital maps are to be completed, the interpretations are to be made and map explanations are to be written.

The Finnish data contribution is based on the aeromagnetic grid 1 km x 1 km (*Korhonen, 1980*), Bouguer anomaly data of the Finnish Geodetic Institute (*Kääriäinen and Mäkinen, 1997*), regional gravity surveys of the GTK (*Elo, 1997*) and the digital petrophysical database of the GTK (*Korhonen et al., 1997 and references therein*).

7. Petrophysics and Interpretation

To correlate the magnetic and gravity anomalies with geological formations, the density and magnetic characteristics of the geological formations of the Shield are calculated from petrophysical data. The available data consists of 330.000 density and magnetic susceptibility measurements, 170.000 measurements of the intensity of remanent magnetization and 7000 measurements of its direction. For the identification of the shallow and the deep seated source of the anomalies, geologically weighted density and magnetization grids with cell size 5 km x 5 km have been calculated. The depth extent of the selected major surficial, geological formations have been interpreted from the geophysical anomalies and from their petrophysical properties. The identified deep anomaly sources are delineated and geologically interpreted. This information contributes to the understanding of the 3-D structure and the evolution of the Precambrian of the area, and to the explanation of the geological map. Petrophysical data provide information to what extent the magnetic and gravity sources are located in the same geological formations. The spatial distribution and the geological correlation of crustal magnetic masses is defined by first calculating the pseudogravimetric anomaly from the magnetic total field component data and then applying similar modeling techniques as used for the gravity interpretation.

The present border of the Shield is not visible in the magnetic or gravity anomalies patterns. Instead, the average magnetic anomaly field of the central part of the Shield is low (-120 nT) and increases outwards to an average value of +120 nT at a radius of 500 km, rapidly decreasing close to zero outside. This circular symmetry is roughly concentric with the postglacial Fennoscandian landuplift and is more pronounced on the western and northern parts of the Shield. The positive magnetic anomalies are associated with Neoarchaean-Mesoproterozoic igneous rocks, which are mainly felsic in composition. Some parts of the anomalies are caused by unexposed magnetic sources. The main part of the circular structure was formed or reactivated in the Svecofennian orogeny. A substudy focuses on evolution of the structure and on its possible connections with the evolution of the Shield and with land uplift of the same area. Both endogenic and extraterrestrial origin models are being tested. The Shield outside the structure consists of the accreted Precambrian terrains of the Kola, Sveconorwegian and Lofoten areas. These terrains are characterized by their higher magnetization in central parts than in the border areas, thus they differing by their nature from the central part of the Shield (*Korhonen et al., 2000*). The final results of the project will be published in 2001.

8. Trends

A major part of the smooth magnetic and gravity anomalies are caused by sources that are not outcropping and, hence, must be interpreted by indirect information. Traditionally, the rock samples of high metamorphic grade have been used to obtain statistical information on petrophysical properties of deep-seated rocks. More specific estimates can be found if xenoliths of kimberlites are available, e.g. in the Archaean part of the Fennoscandian shield. The measurements on the thermomagnetic properties of the samples provide us with information on the magnetic mineral composition and on the values of the induced magnetization at higher temperatures. Reflection seismic studies provide data of the geometry of the hidden bodies and is therefore ultimately necessary for a reliable interpretation of the potential field anomaly sources at depth. The geothermally calculated Curie-isotherms of the magnetic minerals in the lithosphere may be used as additional boundary values for the numerical magnetic interpretations.

The petrophysical parameters can be tied with the time scale of the geological evolution with geochronological isotope studies. The isotope studies may further indicate later thermal and extraterrestrial events altering the original physical properties. Petrophysical data that is measured parallel with rock geochemical compositions provide another cross-scientific analysis.

The accumulation of the digital data, the results of the numerical interpretation plus development of GIS software and the increasing capacity of the computers gradually provide opportunities for building a virtual crustal model for the geological studies. This will be a storage, retrieval and interactive display of a digital 3D model, where the environment of evolution, the geometry of the units, and the geological attributes are traceable from the present day to the geological history.

Metadata

Metadata information on global petrophysical databases is given at web address: **http://www.gsf.fi**
Metadata on Finnish regional geophysical data sets of the GTK are found at address: **http://info.gsf.fi**

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Granite Emplacement During the 1.83 Ga Late-orogenic Stages in Southern Finland

Tom Stålfors and Carl Ehlers

Åbo Akademi University, Department of Geology and Mineralogy, FIN-20500 Turku, Finland
E-mail: tom.stalfors@abo.fi

Late-orogenic (1.83 Ga) granites and migmatites forms in southern Finland a 100 × 500 kilometer long zone which is connected to a major transpressive shear zone system. The granites were intruded along sub-vertical ductile shear zones and were emplaced subhorizontally parallel to the overturned (D_2) axial planes. They are composed of small melt batches, possibly originating from different sources and collected to form sheet-like plutons, with individual layer thicknesses at metre-scale or less.

Keywords: Granites, partial melts, melt generation, geochemistry, emplacement

1. Introduction

The country rock in southern Finland formed mainly during the Svecofennian orogeny about 1.9 Ga ago. The late Svecofennian domain in southern Finland is a large crustal segment characterised by roughly E-W trending subhorizontal migmatites and granites (Fig. 1a). Combined ductile E-W shear movements and NNW-SSE compressional movements defined a transpressional tectonic regime during the emplacement.

About 1.83 Ga ago, the middle and lower crust was partially melted as a result of crustal thickening and subsequent extension. During this event, S-type migmatites and granites were formed along a 100 × 500 kilometre long zone. This zone, known as the Late Svecofennian Granite - Migmatite zone (LSGM zone) is bordered both in the north and the south by earlier metasediments and volcanics. It was intruded by later (~1.6 Ga) rapakivi granites (Fig. 1a). Partial melts that moved upwards through the crust formed either granitic massifs in the middle and upper crust or froze as migmatites at greater depths. Major transpressive shear zones border the LSGM zone. The LSGM zone is a tectonic and metamorphic zone that cross-cuts the earlier Svecofennian granitoids and also the E-W trending threshold between the thinner and the thicker crust that underlies southern Finland (Fig. 1b).

2. The Tectonics

Three deformational phases are identified in the Svecofennian rocks: D_1, D_2, and D_3. The late Svecofennian granites have recorded only the D_3 structures, which gives us the "late-tectonic" timing of the emplacement. Due to the high metamorphic grade, the oldest deformational structures are very sparse, even if some isoclinal F_1 folds can be observed. The tight F_2 folds are overturned towards W or NW, while the F_3 folds are upright and transpose earlier structural features parallel to the E-W trending F_3 axial planes. The late-orogenic granites were emplaced during this event and were deformed during D_3. In places, microcline phenocrysts indicate the sense of movement (upper side towards W or NW) during D_3 in horizontal granite sheets. This imbrication emphasises the fact that emplacement took place both during and after the late stages of D_3 deformation. The granites were intruded along sub-vertical ductile shear zones, which define a major transpressive shear zone system, and emplaced subhorizontally parallel to the

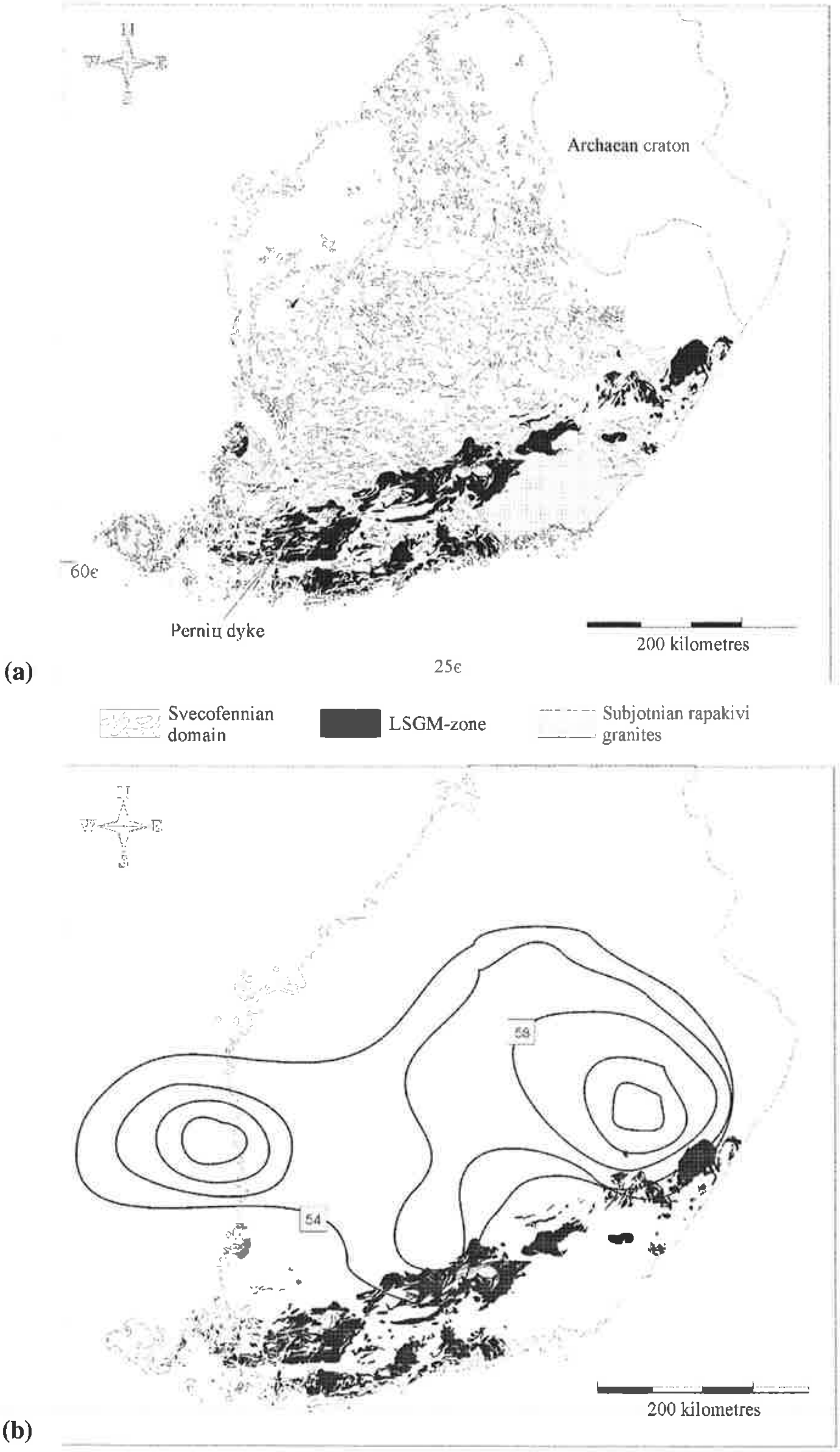

Figure 1. (a) Geological overview of southern Finland. The Svecofennian domain consists mainly of granodiorites, gneisses and gabbros. The LSGM-zone is characterised by late-orogenic migmatites and granites, which are intruded by later rapakivi granites. (b) The LSGM zone crosscuts the E-W trending threshold between the thinner and the thicker crust that underlies southern Finland. Crustal thickness (the depth to the Moho) contours are in kilometers and modified after Luosto (1997).

overturned D_2 axial planes. The driving force was strike-slip dilatancy pumping with alternating compressional and extensional areas.

3. The Emplacement of Granites

We show data that late-orogenic plutons in southern Finland are built up by accumulations of small individual batches of crustally derived melt. Some plutons are visibly composed of different layers of granites, and they show individual variations in their REE compositions. The great volumes of chemically variable late-orogenic granites and migmatites indicate that the melts were transported and emplaced as small batches over an extended time interval, possibly extracted from different protoliths. These migmatites and granites were generated and emplaced at a depth of about 15 to 20 kilometres and pressures between 4 and 6 kbars. If all the granitic material had been molten at the same time, the self-supporting palaeosome structures (mesosome) would have collapsed. Preserved pre-migmatitic structures, even in places with melt percentages > 50%, indicate several successions of melt production. The palaeosome structures will collapse when the amount of melt exceeds the critical melt fraction (30-50%). If stress is applied to the system, even lesser amounts of melt will result in the breakdown of the palaeosome. Earlier experiments *(e.g. Miller et al., 1988)* indicate that during deformation, an initial dramatic weakening will occur at melt fractions as low as 10-15%.

During plastic deformation, melt volumes above the permeability threshold (1-5 %, *Maaloe, 1982*) will be squeezed out of the matrix if the melt can migrate to areas with lesser pressures *(Sawyer, 1991)*. These low-pressure sites are generated throughout a transpressional regime, so the amount of melt volumes never need to exceed a couple of percents, even though the cumulative amount of partial melts in the end of the generation process can be over 50%.

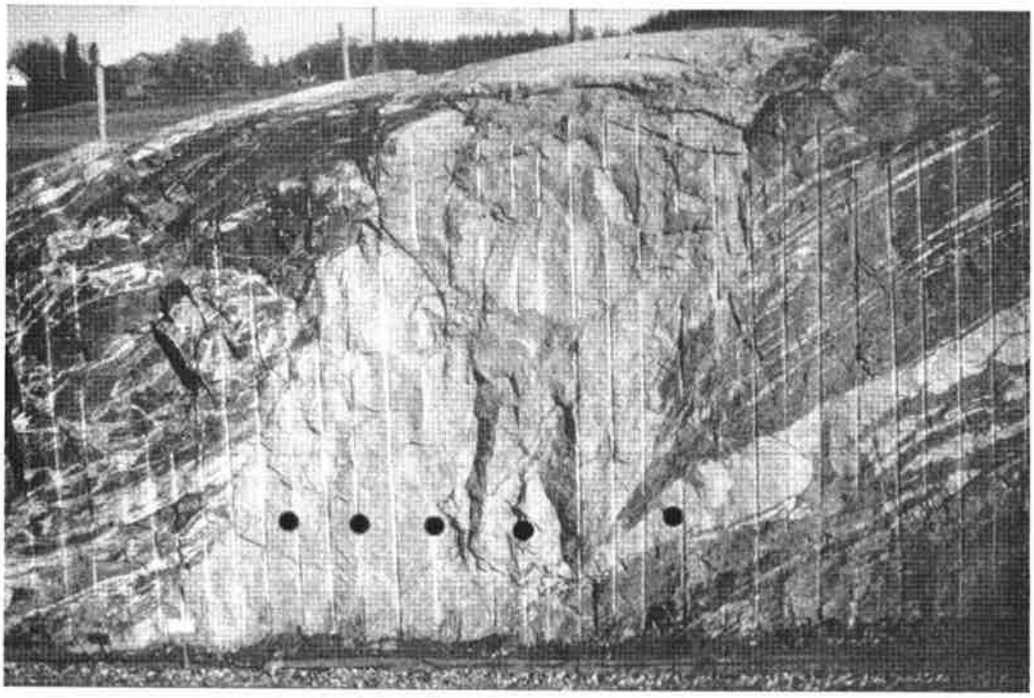

Figure 2. A granite dyke crosscutting mica- and hornblende gneisses at Perniö, SW Finland. The width of the dyke is ca. 4 metres. The circles mark the sampling locations.

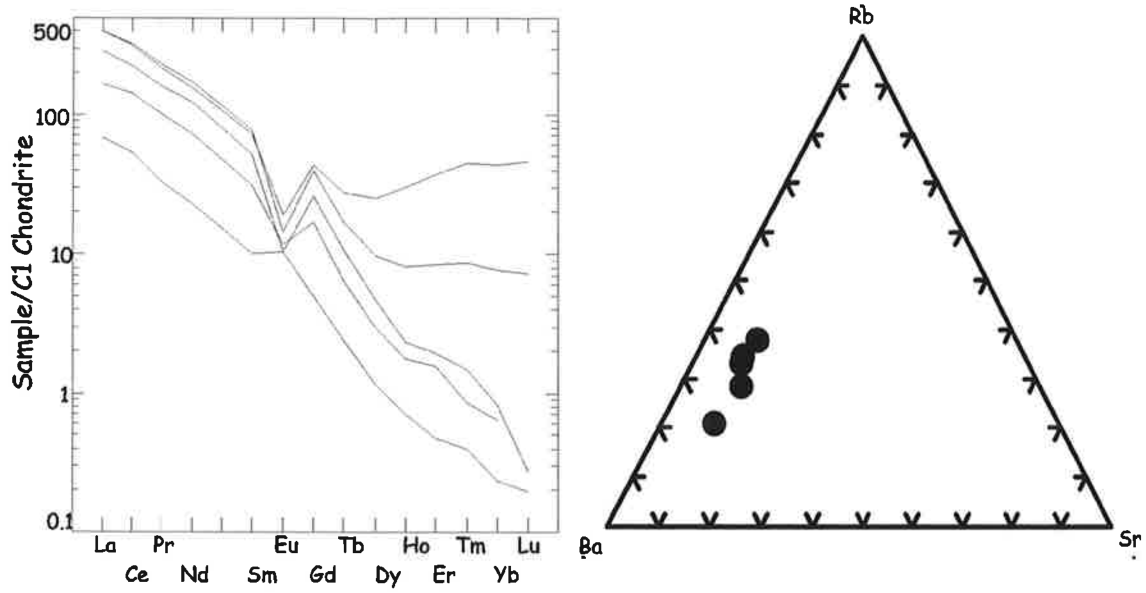

Figure 3. REE and Rb-Ba-Sr ternary diagrams from a crosscutting dyke at Perniö, SW Finland.

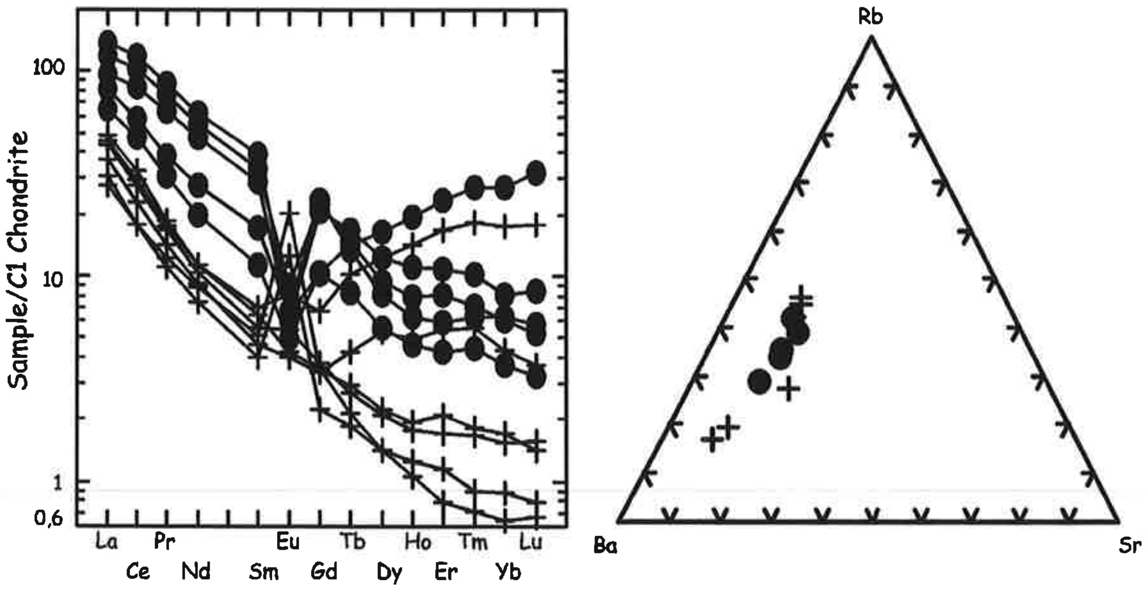

Figure 4. REE and Rb-Ba-Sr ternary diagrams from a pluton in south-central Finland. Based on geochemical evidence, the samples are divided into two groups (crosses and circles) corresponding to the different magmas, possibly derived from different sources. This division is based both on the zirconium concentrations and the Eu/Eu* ratios. The Eu/Eu* ratio for the samples marked with a circle is less than one, whilst the samples marked with a cross have in general an Eu/Eu* ratio of more than 1.

4. The Geochemistry

Geochemical analyses over a crosscutting granitic dyke in SW Finland (Perniö, Fig. 2) confirm the emplacement hypothesis. Both the REE and the Rb-Sr-Ba diagram (Fig. 3) indicate that the melt was successively differentiated during melting or transportation. These five samples were taken from the 4 metre wide crosscutting granitic dyke mentioned above (Fig. 2). Large deviations in the REE suggest that the melt was transported rather as independent batches than as one major pulse of magma. These melt batches were transported through vertical shear zones created by D_3 and were emplaced at a structural level in the crust, perhaps corresponding to the brittle-to-ductile transition zone or some other horizontal discontinuity. A granitic pluton is thus an accumulation of several individual subhorizontal layers with thickness at metre-scale or less.

Field observations and geochemical data further suggest that all the material in neither a granitic pluton nor a migmatite were molten at the same time. The failure of the palaeosome structures was prevented during the late stages of the Svecofennian orogeny by keeping the amount of melts present at any time well below the melt fraction needed for the internal structure to fail.

A temperature estimate, based on the solubility of zirconium in granitic melts *(Watson and Harrison, 1983)*, gives us a range between 850 and 700°C during emplacement. The highest temperatures are obtained for the least fractionated samples, which indicate a successive drop of the temperature during emplacement of the individual batches. We further argue that the previous

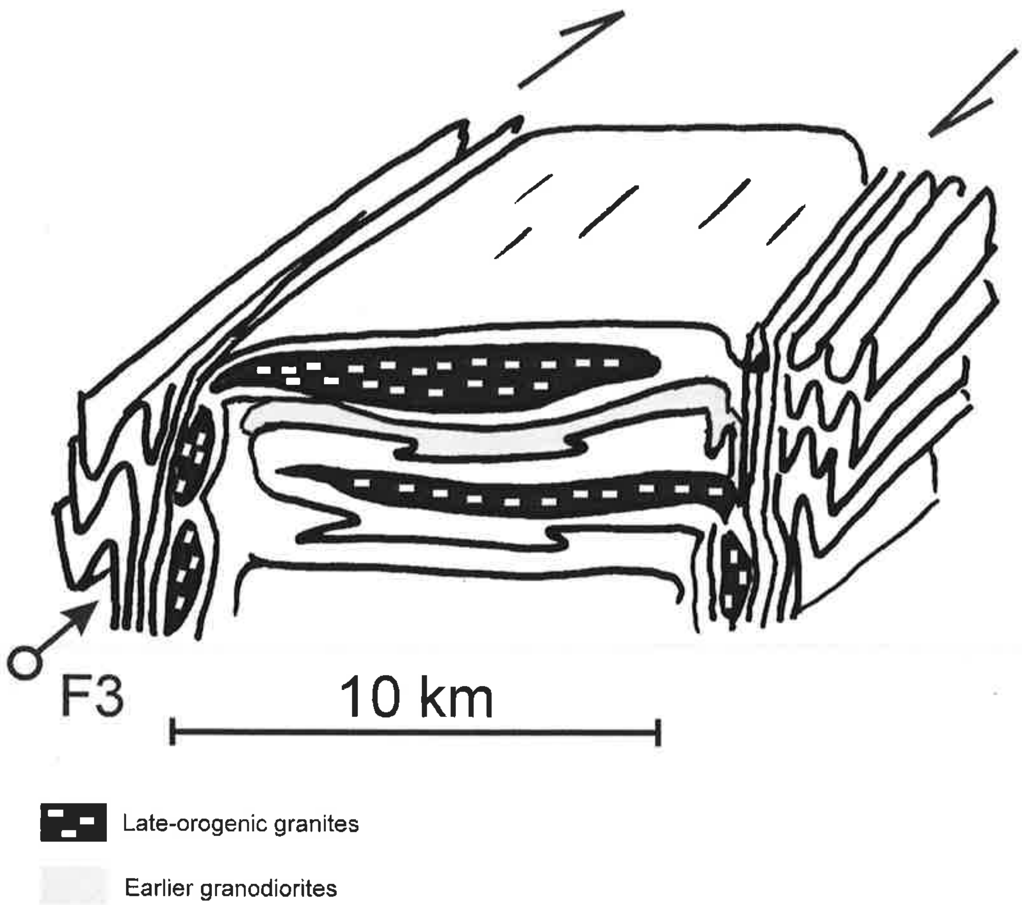

Figure 5. A proposed emplacement model for the granites in southern Finland.

melt batches were solidified before the next one was injected into the vein. This explains the internal structure of the dyke. If successive melt batches would enter a dyke containing molten material, it would result in a dyke where the central parts contained the most differentiated magmas and less differentiated earlier magma batches would be pushed aside and concentrate closer to the wall rock.

We have also sampled and analysed granite sheets from a single pluton lying close to the northern boarder of the LSGM zone in south-central Finland, about 10 kilometres south of Hämeenlinna. Eleven samples, spacing about 3 to 5 metres show large deviations both in their REE and feldspar compositions (Fig. 4). The correlation between the lanthanides and the zirconium concentration suggests that at least two different magmas are present. A comparison of the Eu/Eu* values of these samples suggest the same division into two groups.

The order of the intrusion of the individual layers is not so clear, but the nature of a differentiated and complex origin is still obvious. Our interpretation is that the pluton is built up of small layers of different granitic magmas that intruded in a random manner, thus constructing a visibly layered pluton with occasional remnants of earlier material separating some layers. The samples above indicate that this layered nature of granites is a widespread phenomenon, visible throughout the LSGM zone.

5. Conclusions

We propose the following model of the emplacement for the LSGM zone. The microcline granites (S-granites) occur as subhorizontal sheets surrounded by more intensely deformed and sheared zones. The sheets are gently folded during the D_3 phase, while the steep dykes of the granite that occur along the margins of the sheets are strongly deformed and drawn out into boudins during the same D_3 phase (Fig. 5). Sheets of earlier intruded granodiorites (1.89 Ga) are frequently exposed subhorizontally interlayered with the later S-granites.

Local concentrations of granodiorite sheets stabilized the subhorizontal D_2 structures forming areas that resisted or minimized the subsequent F_3 folding. These structures acted as repositories for granite melt batches that were squeezed from the surrounding, more intensely folded and sheared supracrustal gneisses. The stretched and deformed granite dykes in the gneiss area suggest a strike-slip dilatancy pumping as a possible mechanism for the transport of granite melts from the shear zones to the subhorizontally layered sheeted massifs.

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Paleoproterozoic Trondhjemite Migmatites in Southern Finland

Tapio J. Koistinen, Kalevi Korsman and Petri Virransalo

Geological Survey of Finland, P.O. Box 96,FIN-02151 Espoo, Finland
E-mail: petri.virransalo@gsf.fi

In southern Finland two major types of Paleoproterozoic migmatites make up broad zones dominating the image of the geological maps of the area. The southern part of Finland consist of granite migmatites and the surroundings of the Central Finland granitoid complex are trondhjemite migmatites. Migmatite types are different with respect to the composition of the original paleosome material, composition of the neosome, age of the formation and relationships with deformational phases. The trondhjemite migmatites developed together with the rapid regional temperature rise and onset of intensive deformation early in the geological evolution, while the granite migmatites have a longer history, extending to the late orogenic phases. In this paper the latter type is not dealt with in details.

Keywords: Palaeoproterozoic, migmatites, Finland, Raahe-Ladoga Zone, Central Finland granitoid complex

1. Introduction

Paleoproterozoic trondhjemite migmatites abound along the Raahe-Ladoga Zone (RLZ) near the SW margin of the Archaean craton and southwestwards in many parts of southern Finland, where some non-migmatitic belts are also met with (Fig. 1,2). They are not found northeast of the line Parikkala-Savonranta-Kaavi-Juankoski-Oulu. The distribution of the trondhjemite (-tonalite) veins broadly correlates with the distribution of the early intrusions belonging to the Central Finland granitoid complex and analogous intrusions elsewhere in the southern part of Finland. In the northeast, such early intrusions do not extend far into the Archaean area. At few localities they are observed to intersect the margin of the Archaean crust, as is the case NE from Varkaus (where unexposed Archaean crust is below the migmatites) or in western Kaavi and in Juankoski (where due to deeper section the Archaean crust is partly exposed). The northeastern boundary of these intrusions is rather abrupt in the northeast, and consequently the northeastern boundary of the trondhjemite migmatite terrain there is rather abrupt also (within a range of a few kilometers).

2. Granite migmatites

Another major type of Paleoproterozoic migmatites is the granite migmatite, frequent in the broad zone running from southwestern Finland towards east-north-east until Kitee and extending to the northwestern Lake Ladoga. Granitic veins also occur in the Oulujärvi and Oulu areas, and close to the central Lapland granite with analogies in the north. The migmatites in South Finland were made famous by J.J. Sederholm. In all these areas also large masses of potassium-rich granites occur.

The distribution of both major groups of migmatites is shown in the bedrock map of Finland *(Korsman et al., 1997)*. The two major groups of migmatites have distinctive positions in the evolution of the crust in Finland.

3. Trondhjemites

The trondhjemitic veins appeared very early during the metamorphic and igneous evolution. In the RLZ, the 1885 Ma pyroxene granites and their melted wall materials are superposed on the regionally formed trondhjemitic veined gneiss. The veins are closely associated with the progressive and rather complex formation of so called F_2 fold structures (a collective term for a coherent group of structures). The time span for the appearance of the metamorphic

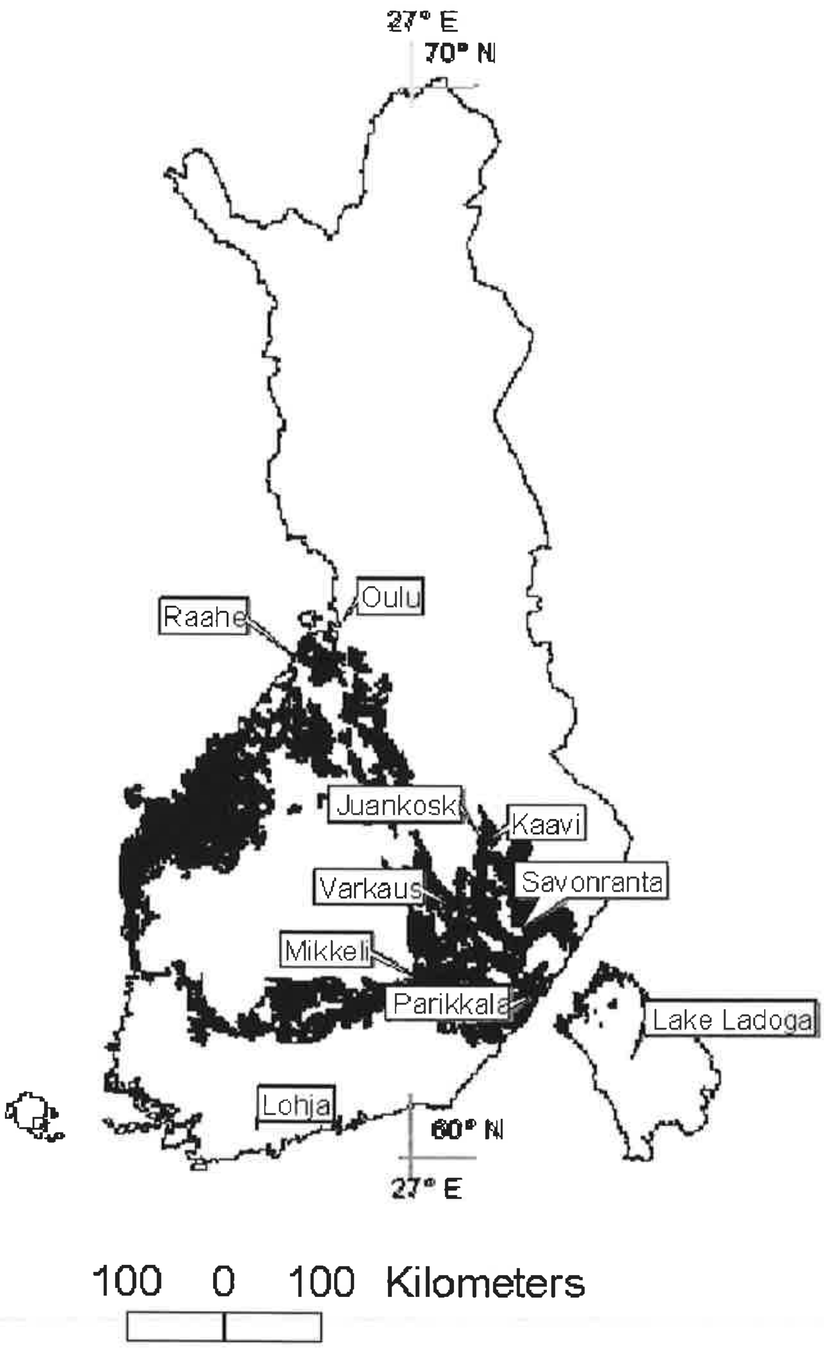

Figure 1. The trondhjemite migmatites of Finland.

trondhjemitic veins is quite narrow as the age of the volcanism is only ca. 1.9 Ga in the region. In contrast, the granite migmatites of southern/southeastern and northern Finland broadly coincide with younger, <1850 Ma igneous materials, and they have their own distribution patterns.

It has been shown that the composition of the original paleosome material as well as partial water pressure during metamorphism are important factors in determining the composition of the migmatite neosome *(Korsman et al., 1999)*. This accounts for the prevalence of the potassium granite veining in southern-southwestern Finland, with dominantly pelitic paleosome

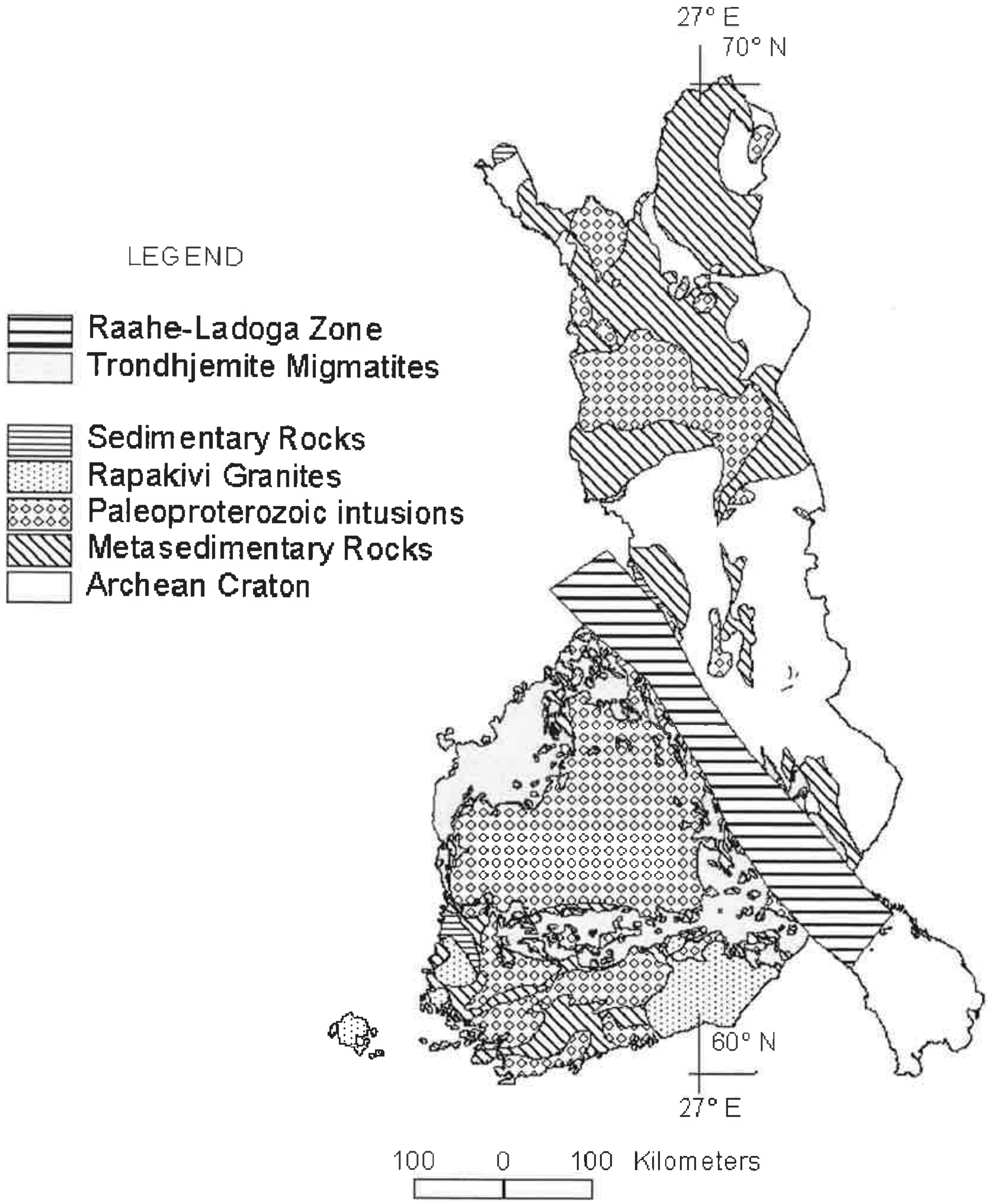

Figure 2. The Raahe-Ladoga Zone and the simplified lithological map of Finland.

material. In this region the veins associated even with the earliest structures well predating the discordant potassium granite masses are often all granitic, as for example in the localities southwest of Lohja. Where the late granite happens to extend marginally to the trondhjemite migmatite province, e.g. southwest of Mikkeli, the granitic veins are clearly seen superposed on the trondhjemitic veins.

In trondhjemite migmatites, with increasing melt component and consequently increasing veining, the end result may be a variable mass of veins with metamorphic banding. The Vaasa granite or the Vaasa migmatite complex (*M.I. Lehtonen et al.*, mapping in progress) is a good example of such large inhomogeneous mass but at the same time more limited occurrences are frequent in different parts of the trondhjemite migmatite zones. In places the melt has migrated from its host region to form breccia migmatites and separate neomagmatic intrusions nearby.

While the trondhjemitic vein evolution has temporally a close relationship with the D_2 fold and shear evolution, the earliest intrusion components of the central Finland granitoid complex also were emplaced during or slightly after D_2. These early granitoids often display magmatic flow structures with resultant alignment of porphyritic crystals if present. The magmatic flow banding has frequently been accentuated and immediately followed by metamorphic S_2 fabric before the final solidification of the crystallizing mass. The expression of the magmatic flow as well as the metamorphic overprint varies within wide limits. The variation depends on the position of the magma and rock in the D_2 stress field and on whether the crystallization coincides with D_2 folds or the later shearing aspects. Some of the intrusions remain unaffected and are homogeneous. The early mafic to ultramafic bodies are often megaboudins within the veined gneiss owing to the effects of D_2. Many later igneous members of the granitoid complex, often porphyritic, clearly postdate D_2 and where they are foliated this foliation is related to post D_2 evolution.

The rather intimate association of the trondhjemitic (-tonalitic) veins in the migmatites and the early granitoids, both within narrow time span, implies that they are related to a common crustal regime, which is characterized by a rapid temperature rise linked with the availability of granitoid magma from a deep source to an upper crustal level.

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Granulite Metamorphism and Formation of the Lower Crust in Finland

Kalevi Korsman[1], Pentti Hölttä[1] and Annakaisa Korja[2]

[1]Geological Survey of Finland, P.O. Box 96, FIN-02151 Espoo, Finland
[2]Institute of Seismology, P.O. Box 26, FIN-00014 University of Helsinki, Finland
E-mail: Kalevi.Korsman@gsf.fi

The Svecofennian Domain in central Fennoscandia is characterized by thick crust (> 50 km) with large crustal thickness variations concentrated in the high-velocity ($v_p > 7$ km/s) lower crust (depths greater than 35 km). The few exposed medium-pressure granulites (Lapland, Varpaisjärvi) are poor representatives of the present lower crust, which mainly consists of mafic granulites on the basis of geophysical and xenolith data. The Svecofennian crust was thickened tectonically by collisional events followed by mafic underplating that was the major contribution to the composition of the present lower crust.

Keywords: Lower crust, granulite, eclogite, seismic velocity, Finland

1. Introduction

The lower crust has been a subject of continuous interest for earth scientists because of its importance in the evolution of the continental crust. Precambrian shields and platforms have generally a thick crust (40-60 km) with a high-velocity layer ($v_p > 6.8$ km/s) in the lowest third of the crust. The lower crust is thought to consist of mafic granulite facies rocks, which form a dominant part of the lower crustal xenoliths. However, the presence of intermediate to felsic rocks in most xenolith suites supports the existence of these lithologies in the lower crust. Restitic, granulite facies metapelites, which have lost granitic melt fraction, also have high seismic velocities and therefore cannot be seismically distinguished from mafic granulites. The predominance of high-velocity minerals (garnet, pyroxene, olivine) in the lower crust is necessary to explain the downward increasing P-wave velocities. In many regions the lower crust is reflective, which indicates compositional variation (*Holbrook et al., 1992; Downes, 1993; Rudnick and Fountain, 1995*).

The Svecofennian Domain in central Fennoscandia is characterized by thick crust (> 50 km) with large crustal thickness variations that are concentrated in the high-velocity ($v_p > 7$ km/s) lower crust (> 35 km) (*Korja et al., 1993*). The present pressure (> 12 kbar) and temperature conditions (> 430°C) (*Kukkonen and Peltonen, 1999*) of the lower crust are compatible with eclogite facies mineralogy as established by *Ringwood and Green (1966)* and *Thompson (1990)* (Fig. 1). The geophysical signatures such as the seismic velocity, density and inferred magnetite content of the lower crust (*Henkel et al., 1991*) are, however, more compatible with the granulite facies mineralogy (*Holbrook et al., 1992*). The mafic eclogite has seismic velocities over 8 km/s and thus it must be regarded as part of the geophysical mantle.

The PT-paths of granulite terrains seem to correlate with their lithological compositions. Terrains that have been isothermally uplifted are dominated by felsic compositions, whereas those that have cooled isobarically have a significantly larger component of mafic lithologies *(Rudnick and Fountain, 1995)*. Isothermally decompressed granulite terrains may represent double-thickened crust due to collision and exhumed by erosion and/or extensional collapse of the resulting orogen. Uplift of granulites formed in this way may be unrelated to the event that caused the granulite facies metamorphism. On the other hand, granulites that show evidence of isobaric cooling are often regarded as products of magmatic underplating. Geochronological constraints, when available, support long-term residence of isobarically cooled terrains in the deep crust (*Wells, 1980; Bohlen and Mezger, 1989; Bohlen, 1991; Rudnick and Fountain, 1995*). Consequently, these probably provide better examples of lower crust than the isothermally decompressed granulites in which the high temperatures and pressures that they record are only a transient feature (*Rudnick, 1992; Rudnick and Fountain, 1995*).

2. Exposed Lower Crustal Rocks in Finland

The lower crustal rocks can be sampled from two main sources: granulite terrains and xenoliths, which have been carried to the Earth's surface by rapidly erupting alkaline and basaltic magmas. The granulite terrains represent isothermally uplifted parts of the lower crust whereas xenoliths comprise isobarically cooled samples.

In the Palaeoproterozoic Lapland Granulite Belt the metamorphic pressures at the thermal maximum have been 6-12 kbar (*Raith and Raase, 1986; Belyaev and Kozlov, 1997*) and in the Archaean Varpaisjärvi granulite complex the 'peak' metamorphism has taken place at around 9±1 kbar (*Hölttä and Paavola, 2000*). These pressures suggest metamorphic crystallization at the depths of ca. 20-35 km. In the Lapland granulite area, the thickness of the present crust is 42-46 km (*Korja et al., 1993*); therefore at least the highest-pressure granulites were metamorphosed at depths comparable to the present lower crust.

In Finland, the Varpaisjärvi area is the best example of an Archaean exposed granulite terrain, consisting of various lithologies and having a complex evolution from the Middle-Archaean to the Palaeoproterozoic (*Hölttä, 1997; Hölttä and Paavola, 2000; Hölttä et al., 2000a*). In the Varpaisjärvi granulite area, the crust is thick, close to 60 km (*Korja et al., 1993*). Here the 9 kbar granulites, metamorphosed at the depth of ca. 30 km, would represent only mid-crustal conditions. However, the thickening of the crust was evidently a Proterozoic event (*Koistinen, 1981; Pajunen and Poutiainen, 1999; Korsman et al., 1999*) and therefore the Varpaisjärvi granulites may represent the lower crust of the Karelian continent.

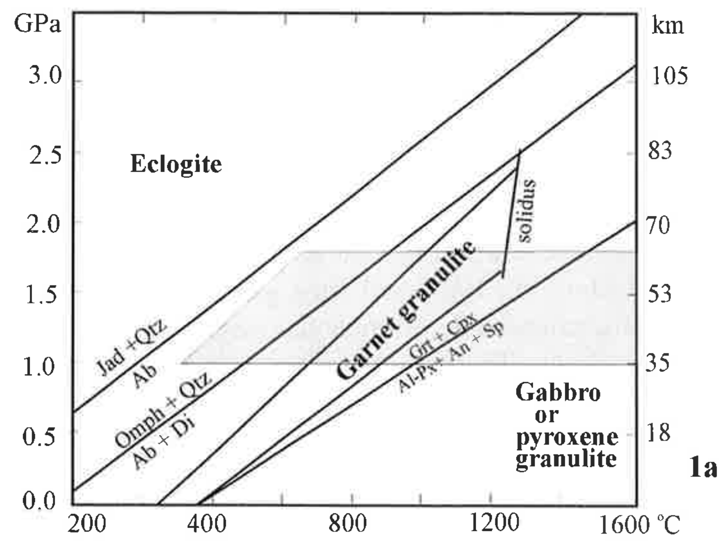
The Varpaisjärvi granulite area and the Lapland Granulite Belt are dominated by intermediate and felsic lithologies. They are isothermally uplifted (*Korja et al., 1996; Tuisku and Korja, 1996; Hölttä and Paavola, 2000*) and the high abundance of supracrustal rocks suggests that these granulites are not good representatives of the deep crust. They are more likely upper crustal rocks that have been through an orogenic cycle.

During the last decade, diamond prospecting has revealed several kimberlite pipes in eastern Finland. A kimberlite cluster is located at Kaavi, where it intrudes through Palaeoproterozoic lower crust, Archaean upper crust and overlying Palaeoproterozoic cover rocks. These kimberlites contain both mantle and crustal xenoliths. The crustal xenoliths that yield pressures of 7.5-12.5 kbar are plgioclase-bearing mafic granulites, not eclogites (*Hölttä et al., 2000b*). Because the crust in the Kaavi area is exceptionally thick, exceeding 65 km (*Korja et al., 1993*), these pressures indicate that the xenoliths originate from the middle crust and upper parts of the present lower crust, from depths of ca. 25-42 km (*Hölttä et al., 2000b*). However, samples of the lowermost crust (42-65 km) have not yet been identified in Kaavi. The xenoliths in the Kandalaksha region in the Kola peninsula are also plagioclase-bearing granulites, even the highest-pressure ones (12-15 kbar) (*Kempton et al., 1995*). In the Kandalaksha region the thickness of the crust is 42-46 km, which indicates that these xenoliths really represent the lowermost part of the crust.

3. The Formation of the Svecofennian Lower Crust

The thickest parts of the Svecofennian crust attain thickness of over 60 km. Such thickness is typical of continental crust that has experienced tectonic thickening by stacking of one continental plate over another. The mechanical and thermal properties of the thickened crust are very much dependent on the initial geotherm of the underlying plate. If the underlying plate had a high initial geotherm (80 mW), then most of the lower plate would remain in granulite facies.

But if the initial geotherm had low values (40 mW), then most of the lower plate would have to slide into eclogite facies (*Richardson and England, 1978*; Fig. 1).



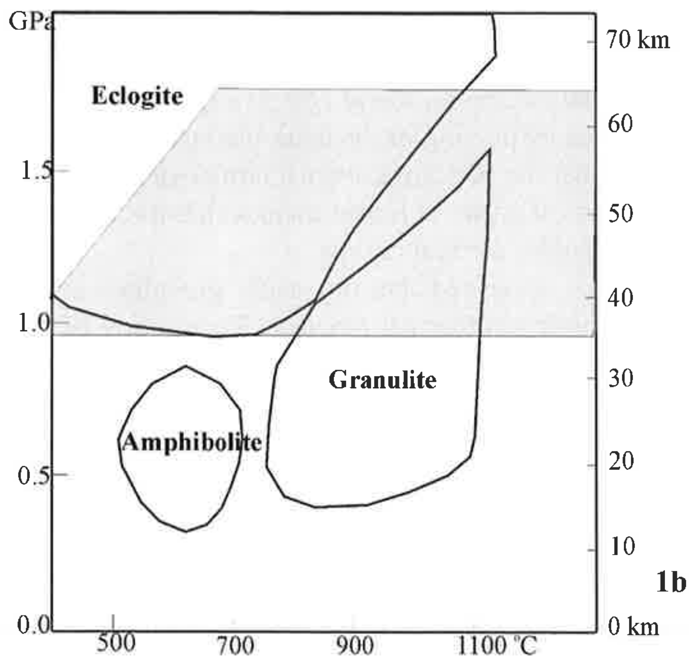


Figure 1. Pressure temperature diagrams showing the stability field of eclogite and mafic granulite.
Grey area indicates the possible pressure temperature range of the lower crust during crustal evolution. The low temperature part images the present conditions and the high temperature part images the metamorphic conditions during formation of the lower crust. Lower crustal material that follows isobaric cooling trends moves from right to left.
a) Eclogite, garnet granulite and pyroxene granulite stability fields after *Ringwood and Green (1966)*
b) Eclogite, granulite and amphibolite stability fields after *Fyfe (1978)*.

After the collision of the Archaean Karelian craton and the Central Svecofennian Arc Complex, the lithosphere suffered from delamination, that induced magmatic underplating and consequently increased the thickness of the high velocity layer in the Archaean-Svecofennian boundary zone and in the Central Svecofennian Arc Complex. The upwelling of the mantle induced partial melting of the lower crust and caused high temperature metamorphism in the middle and upper crust at around 1.885 Ga (*Korsman et al., 1999; Rämö and Nironen, 1999*). Because the crust was rather thick and warm, the gabbros within the present high velocity layer sled into the granulite facies whereas deeper parts of the underplate were metamorphosed within the eclogite field and are regarded as a part of the geophysical mantle.

In southern Finland the 1.84-1.81 Ga metamorphism resembles to some extent metamorphic evolution of the tectonically thickened crust, but it is impossible to explain the ancient high heat flux of this zone by tectonically thickened crust alone (*Korsman et al 1984*). This metamorphism seems to be partly connected with bimodal magmatism and magmatic underplating of this age (*Korsman et al., 1999*). Therefore the entire Svecofennian lower crust is composed of metamorphic gabbroic and metasedimentary rocks connected with magmatic underplating and tectonic thickening. According to some observations (e.g. *Haudenschild, 1988*), after high temperature metamorphism at 1.84-1.81 the cooling down to the closure temperature of the K-Ar system in biotite (ca. 300°C) took place rather late, at around 1.65 Ga.

The seismic crust (v_p < 7.7 km/s) is underlain by an upper mantle layer with P-wave velocities between 8.0 and 8.2 km/s at the southern and northern parts of the BALTIC profile (*Luosto et al., 1990*). *Elo and Korja (1993)* interpreted this layer to have a density contrast of about 350 kg/m^3 and to comprise mafic underplate material associated with the Subjotnian (1.65-1.556 Ga) rapakivi related magmatism. Because the velocity values and the density contrast are compatible with mafic eclogite, *Korja (1995)* suggested that the crust is underlain by mafic eclogite that originated by pooling of the underplating material at the original crust mantle-boundary. This implies that the petrological crust comprising the eclogite layer is by about 7 km thicker than the geophysical crust. This also implies that the underplating took place in the eclogite facies and under a colder thermal regime.

Ellis and Maboko (1992) suggested that dry mafic granulites at lower crustal depths are probable in areas of decreasing geothermal gradient. Their study of the retrograde granulite-eclogite transition indicates that mafic, two pyroxene granulites are metastable in the eclogite stability field and that the eclogite formation through retrograde reactions requires fluid involvement and shearing. *Korja (1995)* suggested that the high seismic reflectivity of the high velocity lower crust may be explained by high density, high-velocity eclogite shear zones (8.1-8.4 km/s) within the lower density, lower velocity granulitic lower crust (6.5-7.4 km/s) (*Austrheim, 1990; Jamtveit et al., 1990; Ellis and Maboko, 1992; Fountain et al., 1994; Korsman et al., 1999*).

4. Conclusions

The Svecofennian crust was thickened tectonically by collisional events followed by mafic underplating. At present, the mafic underplate is observed as high velocity lower crust and uppermost mantle. The crust-mantle boundary represents geophysical transition from mafic granulites to eclogites.

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Mafic Dyke Swarms - Geological Evolution of the Palaeoproterozoic in the Fennoscandian Shield

Jouni Vuollo[1], Hannu Huhma[2] and Lauri J. Pesonen[2,3]

[1]Geological Survey of Finland, P.O. Box 77, FIN-96101 Rovaniemi, Finland
[2]Geological Survey of Finland, P.O. Box 96, FIN-02151 Espoo, Finland
[3]Institute of Geophysics, P.O. Box 4, FIN-00014 University of Helsinki, Finland
E-mail: jouni.vuollo@gsf.fi

Dyke swarm studies have provided valuable information on the Palaeoproterozoic geological evolution of the Fennoscandian Shield and have enabled continental reconstructions to be made, e.g. in the North Atlantic area. Several dyke swarms comprising a voluminous number of dykes are visible in the eastern and northern parts of the Fennoscandian Shield. It has been suggested that these can be divided into at least five main groups in terms of their geochemical composition, absolute age and occurrence. The groups are dated to 2.5, 2.45, 2.33, 2.1 and 1.97 Ga. Some speculations can be made on tectonic movements between the Kuhmo, Taivalkoski and Pudasjärvi Archaean blocks during the time interval 2.45-1.98, and it can be proposed from the dyke orientations that these blocks were rotated several tens of degrees during that time. Dyke swarm investigations have also indicated some uplifted Archaean granulites in the eastern Fennoscandian Shield

Keywords: mafic dykes, swarms, geological evolution, Palaeoproterozoic, Fennoscandian Shield

1. Introduction

Mafic dyke swarms are exceptional geologic time markers that often punctuate major episodes of crustal rifting. A knowledge of the timing of a dyke emplacement is essential for the understanding the tectonic evolution of rift-related environments and for the regional correlation of igneous activity. Dyke swarms are important in continental environments as they are often the only surviving evidence of quite considerable geological events (e.g. rifting, mantle plumes, plate subductions, or crustal "break-up") and can be used to monitor the geological history of the continents over long periods of time. Dyke swarms are also ideal indicators in the reconstruction of the Precambrian crustal blocks *(e.g. Halls and Palmer, 1990; Neuvonen et al., 1997).*

2. Paleoproterozoic dyke swarms

Several dyke swarms comprising a voluminous number of dykes are visible in the eastern and northern parts of the Fennoscandian Shield, and thus Finland, Russian Karelia and the Kola Peninsula provide a good area for research in the formation of Palaeoproterozoic dykes in particular. A new version of the eastern Fennoscandian dyke swarm map is shown in Figure 1 and will also be presented as a poster in this symposium *(Vuollo et. al., 2000, this volume).*

Integrated studies of the geochronology, geochemistry and palaeomagnetism of the Palaeoproterozoic dyke swarms aim at identifying various dyking events and their relationship to economically important layered intrusions and ophiolites in the eastern Fennoscandian Shield. The studies are also aimed at establishing the earliest part of the Proterozoic apparent polar wander path for Fennoscandia. Dykes were sampled in areas belonging to the Karelian and Kola Provinces where overprinting by the ~1.88 Ga Svecofennian orogen is expected to be minimal, i.e. in the Taivalkoski and Suomussalmi areas of eastern Finland, Suoperä and Lake

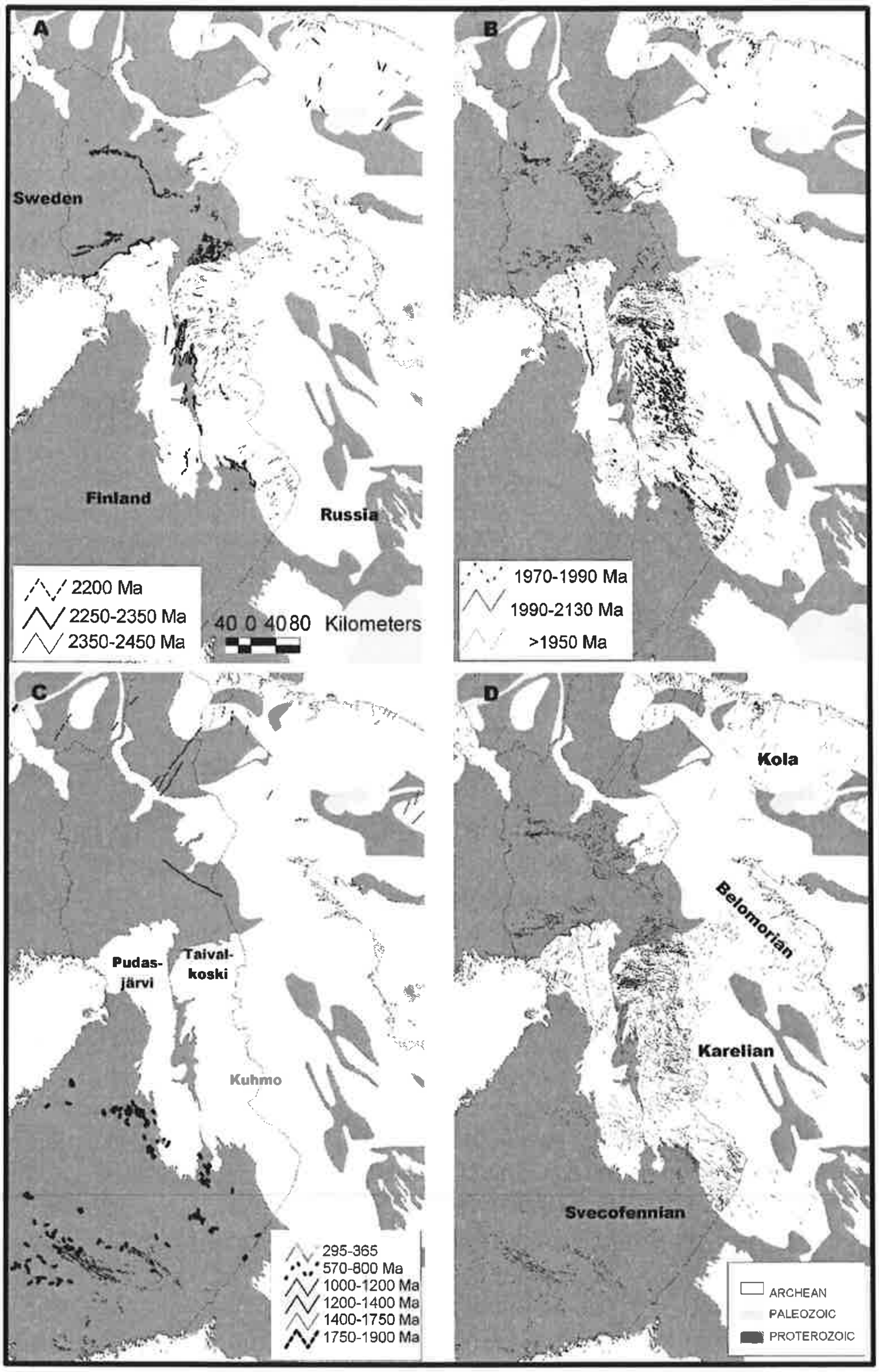

Figure 1 (A-D). The diabase dyke swarms of the Eastern Fennoscandian Shield. Dykes that can be assigned to specific swarms are shown as various line segments. Dykes that cannot presently be assigned (Fig B) to a particular Paleoproterozoic swarm (>1950 Ma) are shown as thin lines.

Pääjärvi near the Burakovsky Intrusion in Russian Karelia, and the mid-Kola Peninsula. Together with previous zircon ages (summarized in *Vuollo, 1994*), the results indicate that there were several dyke intrusion events in the eastern Fennoscandian Shield between 2.5 Ga and 1.97 Ga. It is suggested that these dykes can be divided into at least five main groups in terms of their geochemical composition, absolute age criteria and occurrence. The groups are dated to 2.5, 2.45, 2.33, 2.1 and 1.97 Ga. The 2.5 Ga age group is seen only on the Kola Peninsula, but the others are visible throughout the Karelian Province (Fig. 1 and Table 1).

Table 1. U-Pb and Sm-Nd age determinations and trends (trd) of mafic dyke swarms in the Eastern Fennoscandian Shield.

	U-Pb						**Sm-Nd**					
Block	Kuhmo	*Trd*	Taival-koski	*Trd*	Pudas-järvi	*Trd*	Kuhmo	*Trd*	Taival-koski	*Trd*	Pudas-järvi	*Trd*
Dyke swarm												
1,Bon-norite	2395	40°	-	60°	-	280°	2370$\pm$77	40°	-	60°	-	280°
Gabbro-norite	2446+5/-4	340°	-	-	-	-	2421$\pm$30	340°	-	-	-	-
Tholeiite	-	320°	-	-	2378	330°	2476$\pm$30	320°	-	-	2461$\pm$150	330°
Fe-tholeiite	-	-	-	285°	-	-	-	-	2407$\pm$35	285°	-	-
2, Tholeiite	-	270°	2331+19/-3	278°	-	-	2349$\pm$30	270°	-	-	-	-
3, Fe-tholeiite	-	280°	-	-	2114$\pm$14	330°	2054$\pm$40	280°	-	-	-	-
4, Fe-tholeiite	1981+5/-4	320°	-	-	-	350°	1992$\pm$47	320°	-	-	-	-

The 2.45 Ga dyke swarm can be divided into five subgroups in terms of their field relations, geochemistry and isotope content: 1) NE-trending boninitic-noritic dykes, 2) NW-trending gabbro norite dykes, 3) NW-trending tholeiitic dykes, 4) NW-trending Fe-tholeiitic dykes and 5) E-W-trending plagioclase porphyry dykes. Groups 1, 2 and 5 have similar geochemical trends, showing slight calc-alkaline affinity, while groups 3 and 4 have a typical tholeiitic trend. Sm-Nd isotope studies carried out on groups 1-4 show that groups 1 and 2 represent a (boninitic) magma type with negative initial ε_{Nd} values consistent with those recorded previously in layered intrusions *(Huhma et al., 1990; Turchenko et al., 1991)*. The data on groups 3 and 4 (tholeiitic magma type) show positive ε_{Nd} values and give new information on the magmatic evolution of the 2.45 Ga event. The results indicate that at least the dykes of groups 1, 2, 3 and 5 may be coeval with 2.45 Ga intrusions and thus could be candidates for the parent magma of the layered intrusions of that age. The younger dyke swarms (2.33-1.98 Ga) form a homogeneous group with the chemical composition of typical continental tholeiitic basalts, meaning that classification according to their chemistry is virtually impossible. However in the Kuhmo block, it is possible to classify the dykes of ages ~2.1 Ga and 1.98 Ga by reference to their geochemistry, age and orientation. The first group comprises dykes with a trend of 280° and the second with trend of 320° (Fig. 2 and Table 1).

3. Block movements

At this stage, few speculations can be made on tectonic movements between the Kuhmo, Taivalkoski and Pudasjärvi Archaean blocks during the time interval 2.45-1.98 Ga (Fig. 2).

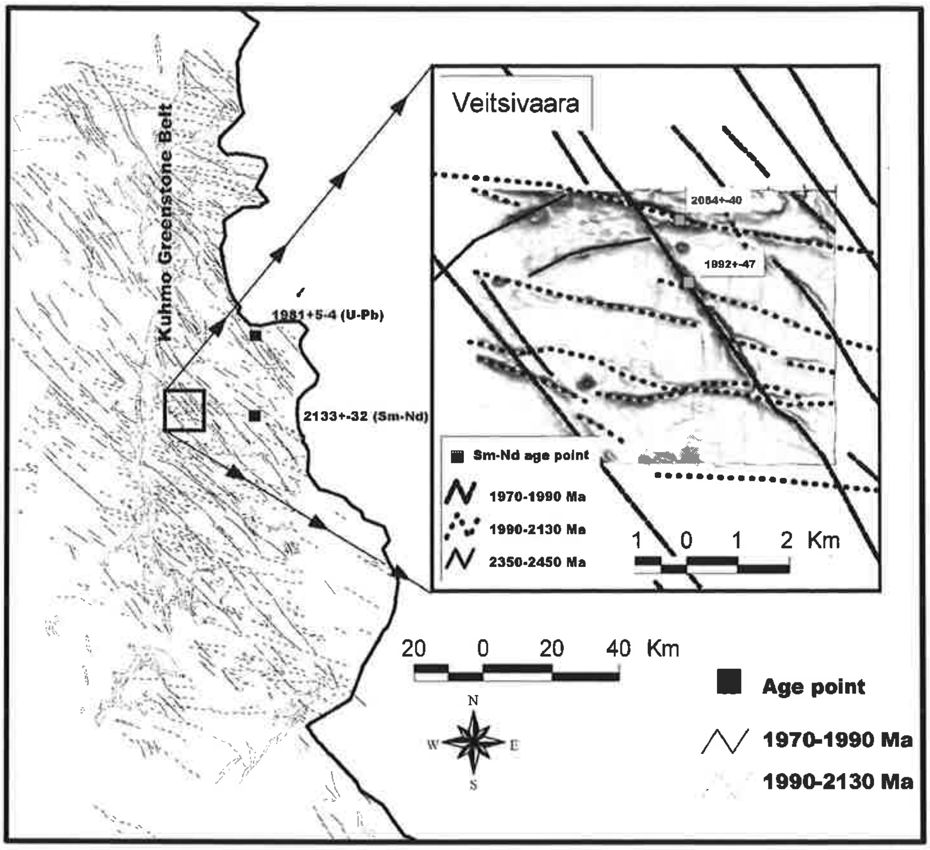

Figure 2. Distribution of ~2.1 Ga and 1.98 Ga dyke swarms in the Kuhmo block. Inset map shows detailed dyke distribution of the Veitsivaara area together with aeromagnetic map.

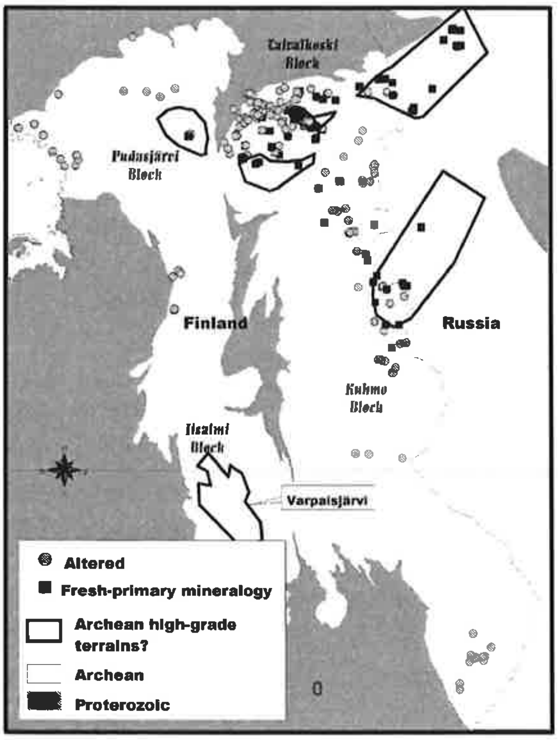

Figure 3. Uplifted Archaean granulites? are drawn according to aeromagnetic maps and diabase mineralogy in the regions of the Taivalkoski-Pudasjärvi-Kuhmo blocks and Varpaisjärvi granulites according to *Korsman et al. 1997*.

Based on geochemical analogies and assuming that the primary trends of the dyke swarms have been identical, it can be proposed that the Kuhmo block has been counter clockwiserotated by several tens of degrees relative to the Pudasjärvi block. For example, the boninitic-norite dykes with an age of 2.44 Ga show a difference of almost 50° in their orientation whereas differences of about 40° can be seen in the 2.1 Ga dykes, in the two blocks. In the case of the younger Palaeoproterozoic dykes (1.98 Ga) the differences in dyke orientations are slightly less, around 30°. A similar type of relative rotation in the Archaean granulite blocks has been presented earlier in the Varpaisjärvi terrane (*Neuvonen et al., 1997*). These results provide only some indications of the movements which may have taken place between the crustal blocks and which can be traced by studying dyke swarms. Further age determinations are needed to identify crustal events in the Taivalkoski and Pudasjärvi blocks.

Dyke swarm investigations in the regions of the Taivalkoski-Pudasjärvi-Kuhmo (Finland and Russian Karelia) and the Kola Peninsula Archaean blocks have also indicated clearly delimitable areas of high metamorphic grade (uplifted Archaean granulites?) in different parts of the Fennoscandian Shield. The clearest indication of such Archaean areas is the preservation of primary magmatic minerals between the dyke swarms. High metamorphic grade Archaean regions(?) of this kind are presented in Figure 3.

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Meteorite Impact Cratering - Implications for the Fennoscandian Lithosphere

Lauri J. Pesonen[1,2], Andreas Abels[3], Martti Lehtinen[4] and Jüri Plado[2,5]

[1]Department of Geophysics, P.O. Box 4, FIN-00014 University of Helsinki, Finland;
[2]Geological Survey of Finland, P.O. Box 96, FIN-02151 Espoo, Finland
[3]Institute of Planetology, University of Münster, 48149 Münster, Germany
[4]Finnish Museum of Natural History, FIN-00141 University of Helsinki, Finland
[5]Institute of Geology, University of Tartu, 51014 Tartu, Estonia
E-mail: Lauri.Pesonen@helsinki.fi

The database of Fennoscandia-Baltic region contains twenty-seven structures of a confirmed meteorite impact origin. The majority of the impact structures have been found in the Proterozoic crystalline rocks. The diameter of the structures ranges from ~ 0.08 to 52 km and the age from pre-historic to ca 2.4 Ga. There are no traces of Archaean impacts in Fennoscandia so far. In few cases, geochemical analyses have led to define the projectile type. Three anomalies appear. *First*, there is a relatively large number of small (D < 5 km) but old (> 200 Ma) structures. *Second*, there is a tendency of the structures to be located south of 63.5°N. *Third*, seven structures are of Cambro-Ordovician age forming a noticeable peak in the crater ages. These biases bear importance in studying the variation of the flux of projectiles colliding with Earth and/or in understanding the preservation differences of the terranes containing impact structures. Although the global database reveals ten structures with diameters > 80 km, there are no such large structures known in Fennoscandia so far. This could be simply due to the relatively small size of Fennoscandia or due to its complex geologic history. Some of the large circular structures in satellite images and geophysical maps may represent deeply eroded remnants of impacts, albeit shock-metamorphic features or impact rocks have not been found so far.

Keywords: impact structures, meteorites, cratering, evolution, Fennoscandia

1. Introduction

The study of impact structures in Fennoscandia has recently gained increased attention for three reasons. *First*, impact structures have a potential for containing economical resources, such as diamonds (Lappajärvi), metallic ores (Siljan, Paasselkä), building and jewelry materials (Sääksjärvi) and water reservoirs (Lappajärvi; see *Pesonen et al., 1999a*). *Second*, scientists have become aware of the geological and biological consequences of impact on the evolution of the Earth (*Grieve and Pesonen, 1996*). *Third*, impact craters have preserved down-dropped sediments of pre- and post-impact origin (*e.g.*, Lappajärvi, Karikkoselkä, Tvären), which serve as unique rocks (not otherwise available due to erosion) for geologists to reconstruct the past geology of the region.

The Fennoscandian Shield has a more than 3 Ga long geologic history and thus the cumulative effects of impact-cratering have potentially played a major role in the creation of tectonic patterns and in the redistribution of elements in the lithosphere, atmosphere and biosphere. It is in a unique position for impact cratering research for several reasons:

(i) the cratonic part of the shield has peneplane topography and has been stable since the last major Svecofennian orogeny at 1.9 Ga ago,

(ii) it is well-exposed,

(iii) there are presently more than seventy craterlike structures in Fennoscandia (*Henkel and Pesonen, 1992; Wickman, 1992*), of which twenty seven are now identified as being of extraterrestrial origin, and

(iv) high-quality satellite (*e.g.*, LANDSAT, NOAA, SPOT, SAR) images for the entire Fennoscandia region are available, as well as high-resolution airborne and ground geophysical and geological data, providing a remarkable database for impact cratering research (*Pesonen, 1996a*).

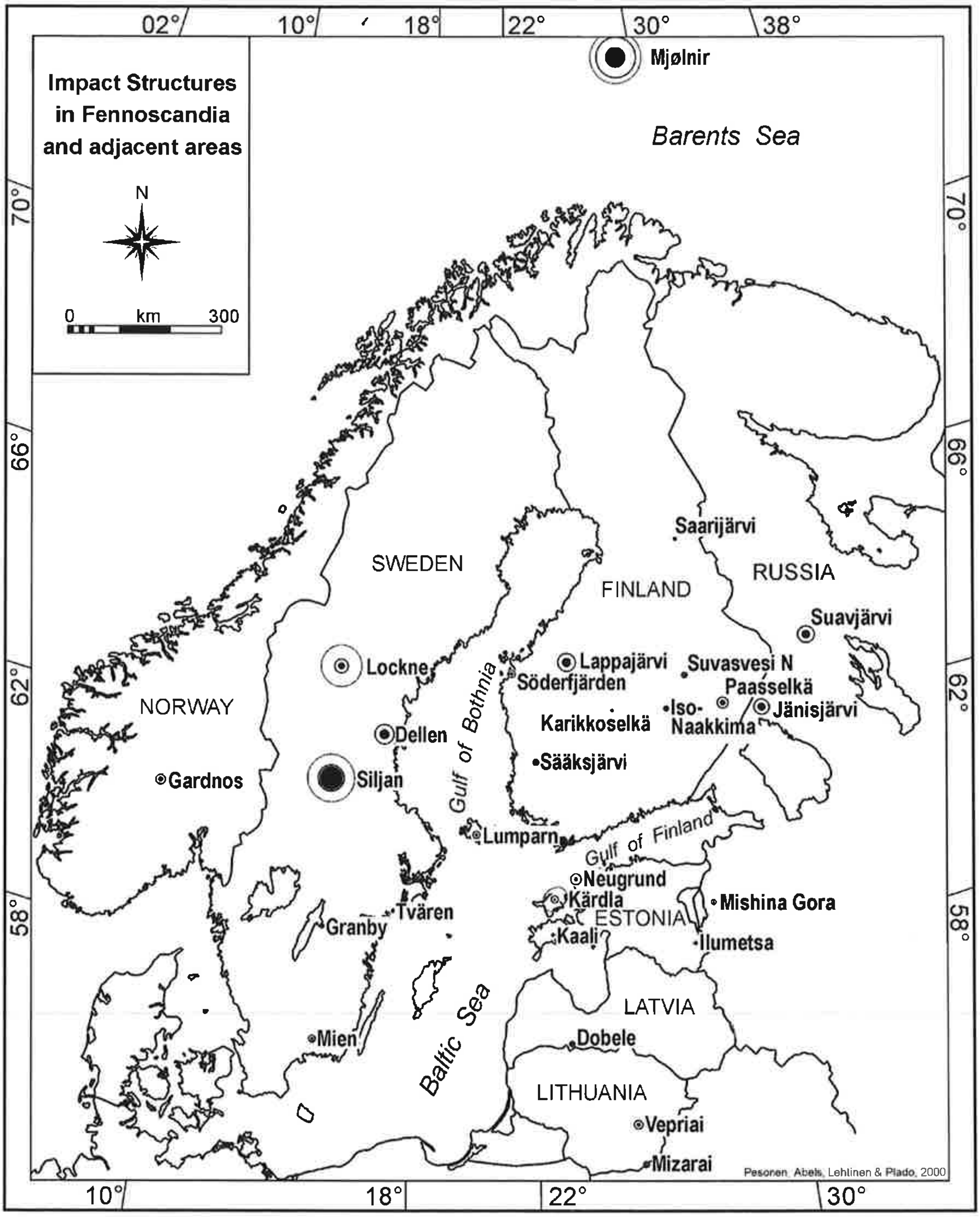

Figure 1. Location of the 27 confirmed impact structures in Fennoscandia and adjacent areas. Data are listed in Table 1. The symbols have one to one ratio, including 'fracture halos' (solid circles) and ejecta occurrences (dotted circles), if observed. The location of two Quaternary crater fields (Kaali and Ilumetsa in Estonia) is marked with crosses.

The purpose of this paper is to provide a new look at the Fennoscandian impact record and to stress the importance of these structures and events for the development of the lithosphere of Fennoscandia and to promote research in this field of Earth sciences (*Pesonen et al., 2000; Abels, 2000; Plado, 2000*).

2. Fennoscandian-Baltic Impact Structures

Table 1 and Figure 1 summarize the morphometric and other data of the twenty seven confirmed impact structures so far known in Fennoscandia and the Baltic area. Since the last catalogue of *Pesonen (1996a)*, structures like Karikkoselkä, Lumparn, Suavjärvi, Neugrund, Saarijärvi and Paasselkä, have been discovered and/or proven to be of the impact origin by proving shock metamorphic features. This high rate in finding new impact craters is due to various factors: (i) an increased research towards impacts in general, (ii) application of geophysical methods on suspected structures, (iii) deep drilling into suspected structures, (iv) comprehensive petrographic and mineralogical studies of drill cores and collected unusual rock specimens, and (v) application of advanced methods in tracing strongly eroded impact structures.

Two marine structures are known in Fennoscandia-Baltic area. The complex Mjölnir structure (D ~ 40 km) is located in the Barents Sea and the Neugrund structure (D ~ 7 km) in the Baltic Sea, close to the coast of Estonia (Fig. 1). All other structures are on present continental land area. However, many of them (*e.g.*, Lockne, Tvären, Söderfjärden, Kärdla, Karikkoselkä) have been interpreted to have formed in an ancient shallow sea as deduced from the attitudes of the allochthonous breccias and sediments in these structures. It is noticeable that four of the structures formed on sea (Lockne, Kärdla, Tvären and Granby) are of nearly similar age (450-460 Ma) and therefore the possibility that a fragmented asteroid or comet shower caused them cannot be missed (*Puura et al., 1994*). One of the structures (Suvasvesi N in Finland; *Pesonen, 1996b*) is part of a probable double structure of which the northern member has been so far proven to be an impact structure. Figure 1 shows also that all but two (Mjølnir and Saarijärvi) impact structures are so far found in south-central part of Fennoscandia. Twelve of these (Suavjärvi, Jänisjärvi, Paasselkä, Suvasvesi N, Iso-Naakkima, Karikkoselkä, Lappajärvi, Söderfärden, Lumparn, Dellen, Siljan and Gardnos) may appear in a "belt" which runs from central Karelia to central Norway. This may reflect (i) a difference in the glacial cover (moraines) between south-central and northern Fennoscandia, (ii) movements of tectonic features ("hinges") which have augmented the erosional or the preservational differences between south-central and northern Fennoscandia, (iii) paleogeographic extents of former protecting sedimentary rocks (see *Abels, 2000*), or (iv) simply a lack of activity in searching impact structures in northern Fennoscandia.

The only onshore structure so far known north of 63.5º (Saarijärvi) has been found already in the early 1930's. The impact origin was confirmed by a deep drilling in the mid-1990's (*Pesonen et al., 1997*).

Several Fennoscandian impact structures contain a well-preserved breccia lens, sometimes associated with an impact-melt layer, whose age can vary by an order of magnitude (Lappajärvi melt has an age of ~71 Ma and Jänisjärvi ~ 698 Ma, respectively; Table 1). The preservation of melt-breccia bodies in Jänisjärvi must be due to that sedimentary rocks have covered them for a considerably long time. This concept is supported by the presence of pre-impact (*e.g.*, Lappajärvi and Karikkoselkä) and post-impact (*e.g.*, Lumparn, Iso-Naakkima, Saarijärvi, Lockne) sedimentary rocks in some structures. The discoveries of these naturally trapped sediments are of paramount importance in reconstructing the geological history of Fennoscandia beyond the glaciations, since in most part of the shield (particularly in Finland) the sediments have been eroded away.

Table 1. List of confirmed impact structures in Fennoscandia and adjacent areas.

Name (Country)		Coordinates		Diameter (km)			State	Morphology	Impactites	Age (Ma , method)		Target	History		
		lat. N	long. E												
Dellen	S	61°50'	16°45'	a: 19	b: 22	c: 20	5	c, tec, l (r, u)	m, mb, lb	89.0 ± 2.7	(Rb/Sr)	cry	a: 1888	b: 1910	c: 1968
Dobele	La	56°35'	23°15'	a: 4.5	b: -	c: -	3, 4	c, bu	lb	325 - 255	(bstr)	sed (Si+De+Cb)	a: -	b: 1978	c: 1978
Gardnos	N	60°40'	09°00'	a: 4.7	b: -	c: 5	6	c, e	lb, mb	700 - 612	(str)	cry + sed (INP)	a: 1945	b: 1990	c: 1990
Granby	S	58°26'	14°56'	a: 3.2±0.3	b: -	c: (= a)	3, 4	s, bu	lb	~465	(bstr)	cry + sed (Ca+Or) + wat	a: -	b: 1976	c: 1980
Illumetsa	E	57°57'	27°24'	a. 0.08			2?, 3	s, pc, f (2, 5?)	lb, ej?	0.006	(C^{14})	sed (Q+De)	a: 1938	b: -	c: 1960
Iso-Naakkima	F	62°11'	27°09'	a: 2.5	b: -	c: -	6	s, tec?, b	lb, bd	1200±100	(pmag, str, bstr)	cry + sed?	a: 1989	b: -	c: 1993
Jänisjärvi	R	61°58'	30°55'	a: 14	b: -	c: -	5	c, l (r, u)	m, mb, lb, sc	698 ± 22	(Ar/Ar)	cry + sed (INP?)	a: 1921	b: 1971	c: 1973
Kaali	E	57°24'	22°40'	a: 0.11			2, 3	s, l + pc, f (9)	pr, ej, lb, sc	0.00395	(hist, C^{14})	sed (Q+Si)	a: 1827	b: 1919	c: 1937
Kärdla	E	58°59'	22°40'	a: 4	b: 15	c: 4	2, 3, 4	c, bu	ej, lb, mb, bd	~456	(bstr)	cry + sed (INP?+Ca+Or [~140]) + wat [<100]	a: 1967	b: -	c: 1980
Karikkoselkä	F	62°15'	25°15'	a: 1.25	b: -	c: 1.5	4	s, l (r)	lb, sc*	< 450 (bstr), 260-230 (pmag)		cry + sed (Ca+Or?) + wat	a: 1994	b: 1994	c: 1996
Lappajärvi	F	63°10'	23°40'	a: 17	b: 35	c: 19	3, 5	c, l (r, u)	m, mb, lb	77.3±0.4 (Ar/Ar), 71±5 (U/Pb)		cry + sed (IMP [~18]+Ca+Or+?)	a: 1858	b: ~1955	c: 1967
Lockne	S	63°00'	14°49'	a: 13.5	b: 24	c: -	2, 3, 4	c, e	ej, lb, bd, mf	~455	(bstr)	cry + wat [=200] + sed (Ca+Or [~81])	a: 1940	b: 1988	c: 1990
Lumparn	F	60°09'	20°08'	a: 7	b: 10	c: 8 - 9	5	c, tec, b (r)	lb, mb, bd	600 - 560	(pmag. str)	cry + (wat + sed?)	a: 1926	b: 1979	c: 1992
Mien	S	56°25'	14°52'	a: 5.5	b: 9	c: 7 - 8	4	c, l (r, u)	m, mb, lb	121 ± 2.3	(Ar/Ar)	cry	a: 1890	b: 1910	c: 1965
Mishina Gora	R	58°43'	28°03'	a: 2.5	b: 9	c: -	5	s	lb, mb, sc	350 - 250	(str)	cry + sed (INP+Ca+Or+De)	a: ~1935	b: 1974	c: 1980
Mizarai	Li	54°01'	23°54'	a: 5	b: -	c: -	4	c, bu	m, lb, bd, sc*	520 - 480	(str, bstr)	cry + sed (INP+Ca) + wat	a: 1978	b: 1978	c: 1988
Mjølnir	N	73°48'	29°40'	a: 40	b: -	c: -	2, 4	c, m	ej	149 - 141	(str)	sed (Pe+Tr+Ju) + wat [200-400]	a: 1993	b: 1993	c: 1996
Neugrund	E	59°20'	23°31'	a: 7	b: 20	c: (= a)	3, 4	c, m	lb	~540	(str)	cry + sed (INP+Ca) + wat [<100]	a: 1927	b: 1996	c: 1998
Paasselkä	F	62°09'	29°24'	a: 10.5±0.5	b: -	c: -	5	c?, l (r)	lb	Precam?	(el)	cry + sed	a: 1984	b: 1992	c: 1999
Sääksjärvi	F	61°25'	22°23'	a: 4.5	b: -	c: 5 - 6	5	s?, l (r)	m, mb	514 ± 12	(Ar /Ar)	cry	a: 1969	b: 1969	c: 1969
Saarijärvi	F	65°17'	28°23'	a: 1.5	b: -	c: -	6	c?, l (r, u?)	lb, sc*	> 600	(bstr)	cry + (wat + sed)?	a: 1937	b: 1992	c: 1997
Siljan	S	61°05'	15°00'	a: 45	b: 55	c: 52	7	c, e	bd, lb, sc*	368 ± 1 (Ar/Ar), 345 (K/Ar)		cry + sed (Or+Si+De? [~500])	a: ~1830	b: 1963	c: 1971
Söderfjärden	F	62°41'	21°35'	a: 6.6±0.4	b: 11	c: (= a)	3, 5	c, bu (r)	lb, bd	530 - 510	(bstr)	cry + (wat [<50])?	a: 1973	b: 1978	c: 1978
Suavjärvi	R	63°07'	33°23'	a: -	b: 16	c: -	6, 7	c, e	lb, mb?	~2400	(str)	cry + sed?	a: -	b: -	c: 1996
Suvasvesi-N.	F	62°41'	28°11'	a: 4.5	b: -	c: -	5	c?, tec?, l (r)	lb, mb	560-530 *or* 260-230 (pmag)		cry	a: 1986	b: 1986	c: 1995
Tvären	S	58°46'	17°25'	a: 2	b: 4	c: 3 - 4	5	s, b (r)	lb, bd, mf	~457	(bstr)	cry + sed (Ca+Or [25-80]) + wat [~100]	a: 1931	b: 1963	c: 1991
Vepriai	Li	55°05'	24°34'	a: 7±0.5	b: -	c: 8	2, 3, 4	c?, bu	lb, mb, sc	350 - 250	(bstr)	sed (INP+Ca+Or+Si+De [>525]) + wat	a: -	b: 1978	c: 1978

Impact structures listed from top to bottom in alphabethical order; numbering of impact structures continued from Henkel and Pesonen 1992, and Pesonen 1996; names of nos. 34 and 66 changed.

Country: E - Estonia, F - Finland, La - Latvia, Li - Lithuania, N - Norway, R - Russia, S =-Sweden

Diameter: a: craterform (as far as preserved rim-crest diameter; otherwise steepest gradient in morphology and/or gravity; for markedly oval structures axis length variation is indicated); b: limit of recognized impact-induced structural disturbance; c: estimation of original rim-crest diameter ; State = state of preservation: 1 - ejecta largely preserved; 2 - ejecta partly preserved; 3 - elevated rim ar least partly preserved; 4 - crater-fill products largely preserved; 5 - crater-fill products partly preserved; 6 - only remnants of crater-fill preserved, crater floor exposed; 7 - crater floor removed, substructure exposed

Morphology (type of crater, conditions of exposure): c - complex crater, s - simple crater; e - exposed, m - submarine, l - lake, b - bay (r - rim exposed, u - central uplift exposed), tec - tectonically modified, bu - buried under sedimentary rocks, pc - penetration crater, f - crater field (in brackets number of craters; only diameter of largest crater given)

Impactites = observed impact lithologies (in boulder, drill core or outcrop): ej - ejecta outside craterform, m - melt rock, mf - fragments of melt in resurge (backwash) deposit, mb - melt-bearing breccia (e.g., suevite), lb - lithic breccia, bd - breccia dyke (incl. pseudotachylites), pr - projectile fragments, sc - shatter cones (in situ *, otherwise in allochthonous breccias)

Age: str - stratigraphic relationships, bstr - biostratigraphic dating, pmag - paleomagnetic dating, hist - historically documented event, el - estimation due to apparent erosion level

Target: sed - sedimentary rocks, cry - crystalline rocks (metamorphic + magmatic), wat - water; IMP - late Mesoproterozoic, INP - late Neoproterozoic, Ca - Cambrian, Or - Ordovician, Si - Silurian, De - Devonian, Cb - Carboniferous, Pe - Permian, Ju - Jurassic, Q - Quaternary; in [square brackets] approximate thickness and water depth, respectively, in meters.

History: year of (a) discovery or first description of lithic or melt-bearing breccia, melt rock, and/or (largest) craterform, (b) first suggestion of impact-origin, (c) final confirmation of shock-features or meteoritic material.

3. The Recognition of Old Impact Craters: the Role of Geophysics

The increased coverage of high-resolution airborne geophysical and digital elevation data has increased the prospects of discovering new impact structures. This is the case particularly in the Fennoscandian Shield, where the crystalline basement has only a relatively thin cover of Quaternary sediments, if it is present at all. Because the basement is often exposed, it is easy to sample the target rocks and impactites for petrophysical analyses. This has led to a petrophysical database of impactites and fractured target rocks and by this way to a better understanding of the changes in the physical properties of rocks caused by an impact (*e.g., Plado et al., 2000; Pesonen et al., 2000; Plado 2000; Sturkell and Ormö, 1998*). These have aided the interpretation of the geophysical anomalies of the impact structures and, in many cases, allowed a unique solution for the geometry of the structures (e.g., Kärdla, Karikkoselkä and Lockne) to be achieved. Moreover, because the geophysical signatures of the impact structures differ from those of the surroundings, the geophysical data are valuable in identifying old and deeply eroded impact structures (*e.g.*, Iso-Naakkima, Saarijärvi and Paasselkä; *Elo et al., 1993; Pesonen et al., 1997; Pesonen et al., 1999b*).

In most cases, the impact structures clearly truncate the pre-existing structural trends as can be seen in geophysical (*e.g.*, magnetics, gravity and electromagnetics) and topographic maps (*e.g.*, Suvasvesi N, Karikkoselkä; *Werner, 1999; Pesonen et al., 2000*).

4. Huge Multiring Structures in Fennoscandia

Although the global database reveals more than ten impact structures with diameters > 80 km, there are no such large impact structures known in Fennoscandia so far. This could be simply due to the relatively small size of this region or due to its complex geologic history. Some of the large circular structures seen in satellite images and in geophysical maps (*e.g.*, Uppland, Marras, Nunjes, Lycksele, Harads, Valga; see *Pesonen, 1996a*) may represent deeply eroded remnants of deformed impact structures, albeit shock-metamorphic features or impact-generated rocks, have not been found so far, or are not preserved anymore. However, if these structures are not the result of an impact, we should be able to find convincing endogenic explanations for them, *e.g.*, block-faulting due to orogenic tectonics or rising magma bodies.

With the possible exception of the large Siljan and Mjølnir structures, the Fennoscandian impact events have not caused global effects in the atmosphere and the biosphere. However, they have been locally very destructive events for distances of several crater radii. Their destructive effects on the crust or the lithosphere are probably relatively minor. For structures smaller than 20 km, the central uplift -induced deformation may reach at maximum a depth of 2 km. Fracturing, however, reaches much deeper levels. The depth effects of impact structures are not yet well understood, but they bear potential importance for alteration processes taken place in lithosphere. For instance, mineral deposition could be promoted through hydrothermal systems induced by the impact or by the shielding effects of impactites. It is almost certain that some deeply eroded, very large impact structures in Fennoscandia still await discovery.

5. Crater Diameters and Crater Ages

Table 1 lists the original (*i.e.*, rim-to-rim) and present diameters of Fennoscandian impact structures as well as the outer limit of their structural disturbance. Original diameters have been collected from the literature and they range from 0.08 km (Ilumetsa) to 52 km (Siljan). With the exception of Siljan and Mjølnir, the diameters are generally less than 20 km and often less than 6 km. Many Fennoscandian impact structures are probably deeply eroded so that the present diameter is much less than the original one. Examples of deeply eroded structures are Saarijärvi and Paasselkä, where we see today only the very bottom of the autochthonous breccias. The Sääksjärvi structure (ca. 514 Ma) is probably eroded to the level of the bottom of the melt layer since the drillings (six of them) have not penetrated the melt

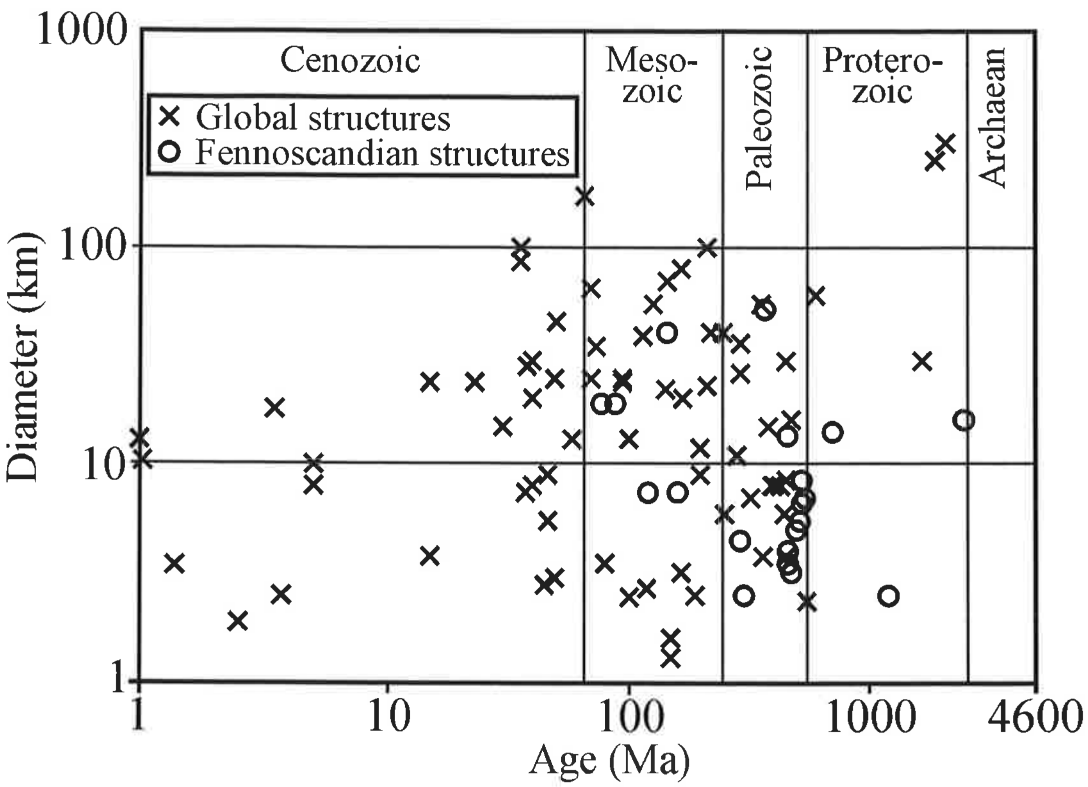

Figure 2. Age vs. diameter of impact structures in Fennoscandia (open circles) and globel (crosses). Only the structures with the present diameter ≥ 1 km and age ≥ 1 Ma are plotted.

body, although impact melt boulders are present in the moraine. This probably indicates that the melt layer was still in the breccia lens during the last glaciation, which ultimately wiped out the remains of this melt body.

Five structures containing impact melt (Lappajärvi, Dellen, Mien, Sääksjärvi and Jänisjärvi) have been dated with isotopic methods although great care must be taken when interpreting the results (*e.g.*, Dellen, see *Deutsch and Schärer, 1994*). Several structures have been dated stratigraphically (*e.g.*, Lumparn, Kärdla, Gardnos, Misarai; *Puura et al., 1994*) or biostratigraphically (*e.g.*, Tvären, Kärdla, Saarijärvi, Lockne, Iso-Naakkima, Vepriai; *Puura et al., 1994*). In these cases the age is a minimum (biostratigraphy) or a maximum (stratigraphy) age. Recently, some structures have been dated with paleomagnetic methods, *e.g.*, Karikkoselkä and Suvasvesi N, where the age uncertainty is, however, quite large (ca. ±30 Ma) and sometimes ambiguous.

The general age pattern of Fennoscandian craters does not follow the global pattern, which reveals a clear bias towards young (< 120 Ma) structures due to erosion (*Grieve and Pesonen, 1996*). The bias in Fennoscandia is towards both young (< 100 Ma) and old (> 500 Ma) impact craters. It is noteworthy that some of the old (> 200 Ma) craters have very small diameters (~5 km; e.g. Iso-Naakkima, Suvasvesi N, Gardnos) resulting from protection by sedimentary rocks and tectonic peculiarities (*e.g.*, covering by Caledonian nappes).

A characteristic feature in the ages of Fennoscandian record is that at least seven structures are of Cambro-Ordovician age and form a noticeable peak in the histogram of crater ages (Fig. 2). If this observation is real and not just a chance, it bears importance in studying the possible variation of the influx rate of projectiles colliding with our planet (see *Culler et al., 2000*) and/or in understanding the tectonic, erosional or preservation differences between the terranes containing impact structures.

7. Projectiles

Using geochemical techniques and mixing calculations it has been possible in few cases to identify the projectile type of the impactor. For example, previously the Sääksjärvi impact structure was suggested to be caused by chondritic meteorite (see *Grieve and Pesonen, 1996*) but more the recent analysis suggests that it was an iron meteorite (*Schmidt et al., 1997*). The projectile of the Lappajärvi impact, was probably a carbonaceous or H-chondritic body (*Göbel et al., 1980*).

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Crustal Boundaries of East European Craton – Keys to Proterozoic Amalgamation

Mikko Nironen[1], Annakaisa Korja[2], Raimo Lahtinen[1] and Pekka Tuisku[3]

[1] Geological Survey of Finland, P.O. Box 96, FIN-02151 Espoo, Finland
[2] Institute of Seismology, P.O. Box 26, FIN-00014 University of Helsinki, Finland
[3] Department of Geosciences, P.O. Box 3000, FIN-90014 University of Oulu, Finland
E-mail: mikko.nironen@gsf.fi

We give short descriptions of the crustal units in the western East European Craton that comprises the Fennoscandian Shield as well as the southern part of the Fennoscandian segment and the western Sarmatia, which are covered by Phanerozoic platform sediments. Based on existing geophysical and geological data, we attempt to point out the importance of the boundaries between these units for the reconstruction of the Proterozoic crust, and for the timing of final amalgamation of the entire East European Craton. Focus is on the Svecofennian Orogen - the established sutures that bound the orogen as well as the inferred sutures and terrane boundaries within the orogen. We also present problems that we consider most important for future studies on the Svecofennian Domain.

Keywords: tectonics; accretion; Proterozoic; Svecofennian orogeny; Fennoscandia

1. Introduction

Similarities between Precambrian and modern orogens had been observed previously, but *Hietanen (1975)* was the first to present a plate tectonic interpretation of the Svecofennian Orogen in central Fennoscandian Shield. The accumulations of geochronological data (mainly whole-rock U-Pb on zircon) and structural mapping from the 1970´s have allowed new plate tectonic interpretations and reconstructions to be made (*see Gaál and Gorbatchev, 1987*). The increase of isotope data (U-Pb, Pb-Pb, Sm-Nd; *Huhma, 1986; Skiöld, 1988)* enabled definition of the boundary (suture s in Fig. 1) between the Achaean cratonic crust and the Proterozoic juvenile crust. *Gaál and Gorbatchev (1987)* presented a comprehensive model for the evolution of the Fennoscandian Shield. Studies involving metamorphism and deformation (*Korsman et al., 1988; Tuisku & Laajoki, 1990*) enabled the understanding of changes in p-T conditions during crustal thickening as the result of plate collision.

The use of geophysical methods (especially electromagnetic, seismic refraction and reflection studies) in the 1980's allowed a plate tectonic interpretation to the electric conductors, seismic reflectors and crustal thickness variations in the Fennoscandian Shield *(T. Korja, 1993; A. Korja et al., 1993*). *Lahtinen (1994)* presented a model for the Svecofennian Orogen in Finland involving several island arc accretionary units. *Nironen (1997)* presented a kinematic plate tectonic model for the Svecofennian Orogen. The latest summary of the Svecofennian Orogen, based on the GGT/SVEKA Transect, is given by *Korsman et al. (1999)*. *Gorbatschev and Bogdanova (1993)* extended their tectonic model to the entire East European Craton including areas covered by Phanerozoic rocks (Fig. 1).

Here we give short descriptions of the geological units of the western East European Craton - especially the Svecofennian Orogen - based on existing literature (references are limited to latest publications and summaries). We also attempt to characterize and classify the boundaries between the units that are important in a plate tectonic framework.

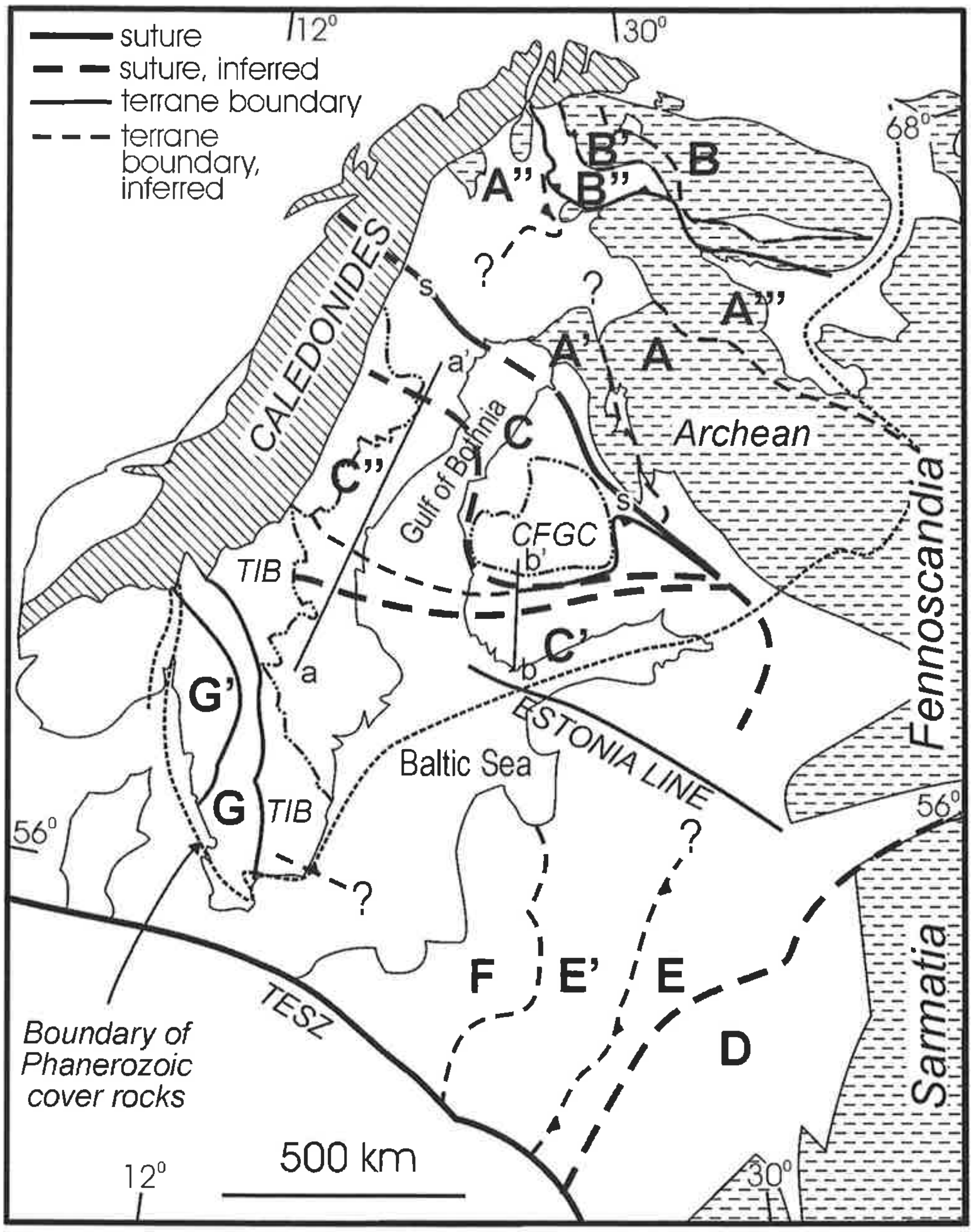

Figure 1. Main tectonic boundaries in the western East European Craton. Crustal units are marked with capital letters (A-G) and explained in text and Table 1. CFGC - Central Finland Granitoid Complex, TIB - Trans-Scandinavian Igneous Belt, TESZ - Trans-European Suture Zone.

2. Crustal Units

The East European Craton consists of the *Fennoscandian*, *Sarmatian* and *Volgo-Uralian* crustal segments *(Gorbatschev and Bogdanova, 1993)*. The main crustal units of Fennoscandia and its marginal areas are shown in Figure 1 and listed in Table 1. The southern part of Fennoscandia and western Sarmatia are covered by Phanerozoic platform sediments ranging in thickness from 1 to 5 km.

A. The *Karelian Domain* consists of Achaean granitoid gneiss complexes and supracrustal rocks (e.g. greenstones) ranging in age between 3.1-2.5 Ga, and having Paleoproterozoic cover.

The Achaean craton may be divided into at least four terranes (A: Kuhmo, A': Pudasjärvi, A'': Finnmark, and A''': Belomorian) separated either by Proterozoic supracrustal sequences containing ophiolite complexes (A – A', A – A") or by Proterozoic structural reworking and isotope resetting (A – A'''). Proterozoic rocks comprise autochthonous supracrustal rocks deposited on the Achaean basement from 2.45 Ga onwards, allochthonous younger units, and the ca. 1.95 Ga ophiolites. Large areas of the Proterozoic supracrustal rocks in Northern Sweden and Finland are intruded by Svecofennian (1.9-1.8 Ga) plutonic rocks *(Gaál and Gorbatschev, 1987; Peltonen et al., 1996; Korsman et al., 1999).*

B. The *Kola Domain* consists of a complex set of Achaean blocks separated by Proterozoic structures or supracrustal sequences. The Paleoproterozoic Pechenga-Varzuga belt separates the Murmansk and Central Kola blocks (B) from the Inari-Tersk block (B'). The geochemical characteristics of the Pechenga area may indicate an opening of a marine basin. To the south, another marine basin is manifested by the fault and thrust bound Lapland-Umba granulite belt (B"), which is characterized by high-pressure metamorphism in its southern border *(Melezhik and Sturt, 1994; Tuisku and Makkonen, 1999).*

The *Svecofennian Domain* is divided into the Central (C) and Southern (C') Svecofennian accretionary complexes, the enigmatic Bothnian Basin (C") and the Trans-Scandinavian Igneous Belt (TIB). In Finland the accretionary complexes differ from each other by the age of peak metamorphism and by the composition of sedimentary rocks and migmatite leucosome. The age of peak metamorphism appears to be the same in the entire Swedish Svecofennian.

C. The 1.92 Ga primitive island arc rocks adjacent to the Achaean craton in Finland are the oldest documented lithologies in **the *Central Svecofennian accretionary complex*** but an older protolith (ca. 2.1-2.0 Ga) has been proposed as a nucleus for the Central Finland Granitoid Complex (*Lahtinen and Huhma, 1997*). The accretionary complex is dominated by 1.90-1.87 Ga old island arc-type volcanic rocks, varying from less mature (Skellefte) to mature (Tampere and Arvidsjaur), and calc-alkaline granitoids. Extension-related 1.88-1.87 Ga plutonic rocks in central Finland and 1.80-1.77 Ga granitoids in northern Sweden form the latest major magmatic episodes. The migmatites with tonalite leucosome in Finland were formed from immature psammites 1.89-1.88 Ga ago whereas metamorphism took place ca. 1.83 Ga ago in Sweden (*Weihed et al., 1992; Korsman et al., 1999*).

C'. The *Southern Svecofennian accretionary complex* comprises remnants of several volcanic belts, e.g. the 1.90-1.89 Ga Bergslagen-Orijärvi belt formed in an intra-arc basin of a mature continental arc (older nucleus?) (*Allen et al., 1996*). Metapelite-quartzite-carbonate association characterizes the sedimentary sequences. Plutonism shows age groups of 1.89-1.85 Ga, 1.84-1.82 Ga and 1.81-1.79 Ga. The 1.84-1.82 Ga S-type granites and migmatites with granite leucosome form a belt that extends from southeastern Finland to central Sweden. The granitoids of the 1.81-1.79 Ga group are found roughly parallel to the northern margin of the inferred terrane boundary C-C' in Finland (*Korsman et al, 1999*).

C". The *Bothnian Basin* is composed of a ca. 10 km sequence of metagreywackes, accumulated from 1.95 Ga until at least 1.87 Ga and intruded by 1.88-1.87 Ga granitoids, 1.82-1.81 Ga S-type granites and subsequently by 1.80-1.77 Ga granitoids (TIB) (*Lundqvist et al., 1998)*. Shallow-dipping structures (as imaged by seismic reflectors and electromagnetic anomalies; *Korja, 1993; Korja and Heikkinen, 2000 (this volume)*) differ from the vertical structures in the east (C) and indicate less shortening. The age of peak metamorphism (ca. 1.83 Ga) is the same as in C' but younger than in C in the east (1.88 Ga).

The *Trans-Scandinavian Igneous Belt* (TIB) is a N-S trending belt that host three age groups of granites called the TIB 1 (1.85-1.77 Ga), TIB 2 (1.7 Ga) and TIB 3 (1.68-1.65 Ga). The TIB 1 granites are the most voluminous and overlap in age with late Svecofennian magmatism. TIB magmatism has been suggested to reflect Andean-type continental arc, extensional environment or extensional delamination (*Lundqvist and Persson, 1999).*

Table1. Crustal units of the western East European Craton.

Crustal unit	**Age range**	**Lithology**	**Metamorphism**	**Proterozoic suture / terrane boundaries (=> shows direction of closure)**	**Indications for a tectonic boundary**
A. Karelian Domain	Archaean (3.1-2.6 Ga), Proterozoic (2.5-1.92 Ga)	Archaean gneisses, Proterozoic continental and marine rocks	1.87 Ga, high T, low P (over-printing)	A', A''=>A (closure of intracratonic basin?) A''' undefined	Ophiolites, thrust tectonics
B. Kola Domain	Archaean (2.7 Ga), Proterozoic (2.5-1.80 Ga)	Archaean gneisses, Proterozoic greenstones and paragneisses	B'': 1.90 Ga, high T, moderate P	B, B'=>B'', B''=>A (init. max. 1.90 Ga) (closure of intracratonic basin or arc - continent collision)	Proterozoic marine and volcanic arc rocks, electric conductors, seismic reflectors
C. Central Svecofennian accretionary complex	1.91-1.87 Ga	Volcanic arc sequence (+ older crust?)	1.88 Ga (Finland), 1.83 Ga (Sweden), high T, low P	C=>A (init. max. 1.92 Ga) (arc - continent collision)	Ophiolites, thrust tectonics, electric conductors, seismic reflectors, isotopes
C'. Southern Sveco-fennian accretionary complex	1.90-1.80 Ga	Volcanic arc sequence (+ older crust?)	1.88 Ga?, 1.85-1.81 Ga, high T, low P	C'=>C (init. 1.89 Ga), (arc - arc collision) C'=>C'' (thrusting?)	MORB-type volcanics, electric conductors, seismic reflectors, lithology, isotopes
C''. Bothnian Basin	1.95-1.77 Ga	Turbiditic metagreywackes, granitoids	1.82 Ga, high T, low P	C''=>C (arc - arc collision) C'=>C'' (thrusting?)	seismic reflectors, lithology
D. Sarmatian (Archaean) Domain, Osnitsk-Mikashevichi Belt	Archaean (3.7-3.0 Ga), Proterozoic (2.0-1.95 Ga)	Archaean gneisses, calc-alkaline igneous rocks			
E. Central Belarus - Vitebsk terrane	2.0-1.9 Ga	Volcanic arc sequence	moderate T, P	E=>D (arc - continental arc collision)	Seismic reflectors
E'. Belarus-Baltic Granulite Belt, East Lithuanian Belt	1.9-1.85 Ga	Intermediate gneisses, granulites and granites	1.80 Ga, high T low P	E'=>E (arc - arc collision)	Seismic reflectors
F. West Lithuanian Granulite Belt, Latvian Domain, South Estonian Domain	1.85-1.80 Ga	Granulites, ?	high T, moderate P	F=>E'?	Seismic reflectors
G. Southwest Scandinavian Domain	1.69-65 Ga (G), 1.66-1.59 Ga (G')	Volcanic arc sequences, gneissic granites		G, G'=>C' (arc - continent collision)	TIB magmatism, transpressional zones, seismic reflectors

D. Western Sarmatia comprises both Achaean and Proterozoic terranes. The westernmost of the latter, **the *Osnitsk-Mikashevichi Belt,*** is located upon the northwestern margin of Sarmatia and consists of ca. 2.0 Ga to1.95 Ga plutonic and volcanic rocks *(Bogdanova, 1999).*

E. *The Central Belarus - Vitebsk terrane* consists of 1.98 Ga metavolcanic gneisses that have been intruded by 1.90 Ga granites. The rocks **in the *Belarus-Baltic Granulite Belt* (E')** are mostly mafic rocks of igneous origin, associated with charnockites and metapelitic granulites. These rocks as well as the rocks of the *East Lithuanian Belt* have been interpreted as island arc and back-arc sequences *(Bogdanova et al., 1994; Bogdanova, 1999).*

F. The *West Lithuanian Granulite Belt,* the *Latvian Domain* and the *SW-Estonian Domain* comprise a complex and a rather ill-defined terrane with ages around 1.85-1.80 Ga Similar ages have been obtained from southeastern Sweden (*Bogdanova, 1999*).

Offshore southeastern Sweden there is a seismic reflectivity zone that is moderately NE dipping and reaching the mantle, which has been interpreted as a terrane boundary *(Abramovitz et al., 1997).*

The *Southwestern Scandinavian Domain* was mainly formed during the Gothian orogeny 1.75-1.55 Ga ago. The sequential growth of the Fennoscandian Shield towards the west included (at least) two accretionary events. The 1.69-1.65 Ga *Ätran - Klarälven terrane* (or Eastern Segment; G) extends northwest below the Caledonides to western Norway. It contains calc-alkaline, deformed granitoids with ages equivalent to TIB 3 granites in the east. The 1.66-1.59 Ga *Idefjorden terrane* (G') consists of island arc type volcanic-sedimentary sequences intruded by 1.59 Ga calc-alkaline granitoids. Both terranes have been ascribed to the eastward subduction and accretion against the Svecofennian continental crust. The rocks in both terranes were deformed during Gothian orogeny and subsequently during Sveconorwegian orogeny (1.15-0.9 Ga) that did not produce major crustal addition to the Fennoscandian Shield. *(Åhäll and Gower, 1997).*

3. Crustal Boundaries

A boundary between two terranes can be called a suture if oceanic lithosphere can be demonstrated to have existed between the terranes. It is to be noted that a terrane boundary denotes the exposed (or upper crustal) margin and the site where the suture extends to the mantle is usually offset. The lithological and isotopic (U-Pb, Pb-Pb and Sm-Nd) differences between the Achaean units A and A' and the Proterozoic unit C as well as the geophysical data allow the boundary to be called a suture (line **s** in Fig. 1). The terrane boundary associated with this suture is located further east and northeast due to thrusting during collision.

The boundaries within the Karelian Domain (A-A''') and Kola Domain (B-B'') are based on the existence of the Proterozoic mobile and thrust belts. Especially the ophiolites in some of these belts are good evidence of marine crust, yet their tectonic setting is still controversial (narrow, Red Sea type or larger ones). The inferred suture between units C and C' and between C and C'' is problematic because isotopic data have yielded fairly similar ages for these terranes. Data for the existence of the suture are the northeast-dipping conductors and reflectors (*BABEL Working Group, 1993*; Fig. 2). Data for the terrane boundary between the units C and C' includes the lithological differences and thrust tectonics, and between C' and C'' a south-dipping reflector in the Gulf of Bothnia (*Korja and Heikkinen, 2000 (this volume)*).

The U-Pb isotopic data allows the boundary between units D and E to be called a suture although the isotopic and especially lithological data are scarce. Seismic data across the EUROBRIDGE profile suggest slightly east dipping boundaries between units D and E as well as between E and E', consistent with a suture dipping southeast beneath the craton margin and successive docking of the terranes against the craton *(EUROBRIDGE Seismic Working*

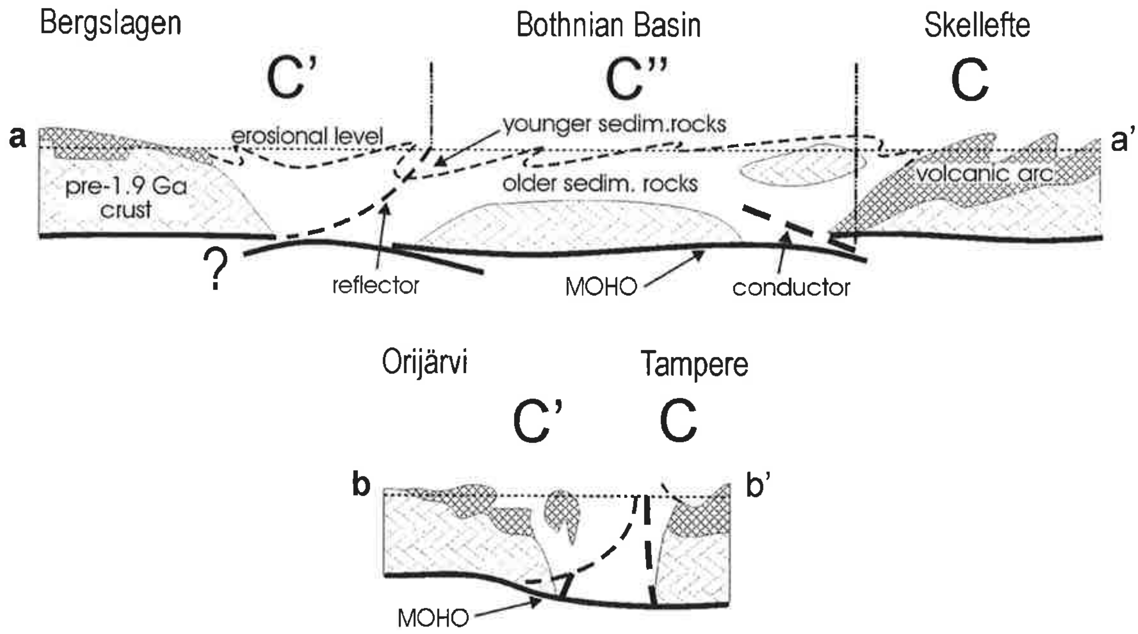

Figure 2. Schematic sections across a) central Sweden (a - a') and b) southern Finland (b - b'). See Fig. 1 for location of the cross-sections. The cross-sections generalize the models by Heikkinen and Luosto (2000, this volume) and Korja and Heikkinen (2000, this volume).

Group, 1999). The seismic data suggest a west-dipping boundary between the units E' and F but the geological character of the boundary is obscure.

The eastern part of the Southwest Scandinavian Domain and the TIB contain a complex set of (transpressional) shear zones that are mainly Sveconorwegian in age. The eastern boundary of the unit G may be regarded as the intracratonic Gothian deformation front but there is no general agreement where the boundary between the Svecofennian and the Gothian Orogens is located.

The boundary named "the Estonia Line" is a distinct shear zone in aeromagnetic maps that appears to divide mafic-dominated rocks in the south from more felsic rocks in the north. The age and geological meaning of the zone is obscure but it is located in an important area considering accretion directions of the Proterozoic arc complexes towards the two Achaean cratons.

4. Critical Aspects

We consider the following problems important with respect to the evolution of the Svecofennian Domain.

1. The pre-1.91 Ga evolution of the Svecofennian Domain. Does the geophysical data support or discard the existence of older Proterozoic nuclei within the Svecofennian?

2. Differences in the tectono-metamorphic evolution within the Svecofennian. Is the Bothnian Basin a separate terrane (Fig. 2)? Are there still unknown terrane boundaries within the Svecofennian?
3. Are the development and persistence of an anomalously thick crust, as seen in parts of the Svecofennian Domain, related to the existence of older nuclei, to continued convergence or to both in these areas?
4. Does TIB 1 represent the extension of the Svecofennian orogeny or initiation of a new Gothian orogeny? Do TIB 2 and TIB 3 belong to the Gothian orogeny? Is it reasonable to group the three magmatic events under "TIB magmatism"?
5. At what time did the Fennoscandian, Sarmatian and Volgo-Uralian crustal segments amalgamate to form the East European Craton? Paleomagnetic data suggests the amalgamation of Fennoscandia and Sarmatia by ca. 1.7-1.6 Ga (*Elming et al., 1998*) but the apparent curving of suture **s** and the boundary between the terranes F and G suggest a closer and older history for these two segments and thus contradict with the paleomagnetic data. How far to the south does the Svecofennian Domain extend?

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Evolution of the Fennoscandian Lithosphere in the Mid-Proterozoic: the Rapakivi Magmatism

O. Tapani Rämö[1], Annakaisa Korja[2], Ilmari Haapala[1], Olav Eklund[3], Sören Fröjdö[4], and Matti Vaasjoki[5]

[1]Department of Geology, P.O. Box 11, FIN-00014 University of Helsinki, Finland
[2]Institute of Seismology, P.O. Box 26, FIN-00014 University of Helsinki, Finland
[3]University of Turku, Department of Geology, FIN-20014 Turku, Finland
[4]Åbo Akademi University, Dept. Geology and Mineralogy, FIN-20500 Turku, Finland
[5]Geological Survey of Finland, P.O. Box 96, FIN-02151 Espoo, Finland
E-mail: tapani.ramo@helsinki.fi

Petrologic, geochronological, geochemical (Nd isotopes in particular), and geophysical (seismic and gravimetric) data recently acquired on the mid-Proterozoic (1.67–1.47 Ga) rapakivi granite suites of the Fennoscandian (or Baltic) Shield and its southern continuation are reviewed. Three lithospheric models put forward to explain the origin of the bimodal rapakivi granite magmatism are discussed in the light of these new data.

Keywords: crustal evolution, geophysics, isotope geology, Proterozoic, rapakivi granite

1. Introduction

The central part of the Fennoscandian Shield comprises a complex segment of Paleoproterozoic crust that was differentiated from the mantle ~2.1–1.8 Ga ago (*Huhma, 1986*). This involved accretion of several arc complexes to the Neoarchaean craton in the eastern part of the shield at ~1.91–1.89 Ga, subsequent extensional collapse, mafic underplating, and anatectic (1.84–1.82 Ga) magmatism, and emplacement of shoshonitic stocks at ~1.80 Ga (see *Korsman et al., 1999,* and references therein). After a ~0.3 Ga period of quiescence, a major tectonothermal event commenced and is recorded by the 1.67–1.47 Ga rapakivi granites and related mafic and intermediate rocks that formed as the Paleoproterozoic Svecofennian crust was reorganized, presumably in response to mantle upwelling and extension (e.g., *Haapala and Rämö, 1990*).

The significance of the Proterozoic rapakivi granites and related mafic rocks has been recognized worldwide (e.g., *Haapala and Rämö, 1999*). They are present on all continents and presumably represent the most voluminous silicic intraplate magmatism on Earth. As the magmatic association of the rapakivi granites features, almost exclusively, only silicic (mainly crust-derived) and mafic (mainly mantle-derived) rocks, rapakivi magmatism provides a handle on the evolution of both pre-existing crust and subcontinental mantle. In the classic rapakivi terrane of southeastern Fennoscandia, petrological, geochemical, geochronologic, and geophysical methods have been widely utilized to constrain magmatic evolution and tectonic setting of the rapakivi granites and mafic rocks associated with them (see *Haapala and Rämö, 1990, 1999,* and *Rämö and Haapala, 1995, 1996,* and references therein). Here we summarize the current knowledge on the evolution of the Fennoscandian lithosphere in the mid-Proterozoic (~1.7–1.5 Ga) time, focusing on, not only the classic Finnish rapakivi occurrences, but also those in the western (central Sweden) and eastern (Russian Karelia) parts of the shield as well as in its concealed southern continuation (Estonia, Latvia).

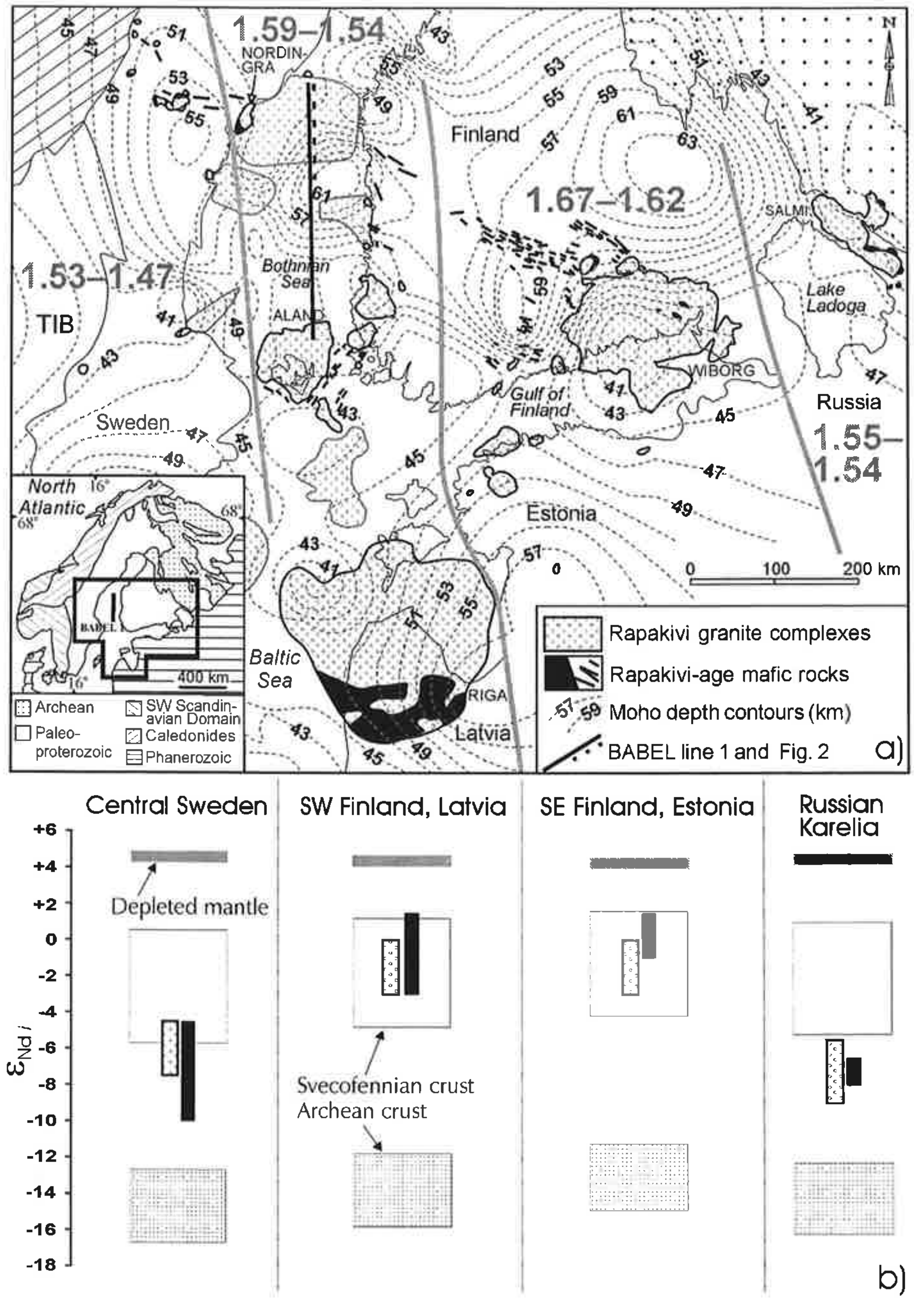

Figure 1. **(a)** A map showing the 1.67–1.47 Ga rapakivi granite complexes and the contours of crustal thickness of the south-central part of the Fennoscandian Shield. Thick subvertical lines outline four rapakivi age zones (1.53–1.47, 1.59–1.54, 1.67–1.62, and 1.55–1.54 Ga). The inset shows an area relative to the major crustal domains of the Shield. TIB is the 1.85–1.7 Ga Transscandinavian Igneous Belt. Note that the area south of the Gulf of Finland is covered by the Phanerozoic sedimentary rocks. The map is compiled mainly after *Koistinen (1994)*, *Rämö et al. (1996)*, *Andersson (1997)*, *Korja et al. (2000, submitted)*, *Korsman et al. (1999)*, and *Persson (1999)*. **(b)** A diagram showing the variation of initial Nd isotopic composition (expressed as $\varepsilon_{Nd\ i}$ -values) in the rapakivi granites (pattern with small open circles) and the related mafic rocks (black pattern) of the four age zones. The composition of depleted mantle (*DePaolo, 1981*), the ~1.9 Ga Svecofennian crust (*Huhma, 1986*; *Patchett and Kouvo, 1986*), and the Archaean crust of the Fennoscandian Shield (*Rämö et al., 1996*) are shown at the time of interest. Note that the rapakivi complexes in central Sweden and Russian Karelia show a significant Archaean source component. The data is from *Ahl et al. (1997)*, *Alviola et al. (1999)*, *Amelin et al. (1997)*, *Andersson (1997)*, *Claesson and Kresten (1997)*, *Fröjdö et al. (1996)*, *Neymark et al. (1994)*, *Persson (1999)*, *Rämö (1991)*, *Rämö et al. (1996)*, *Suominen (1991)*, and *Vaasjoki et al. (1991)*.

2. Petrology and Geochronology

The 1.67–1.47 Ga (Subjotnian) rapakivi granites are found as five large batholiths and several smaller batholiths and stocks and they are associated with quartz-feldspar porphyry dikes (and minor silicic volcanic rocks) that sharply cut the surrounding metamorphic bedrock (Fig. 1a). They are spatially and temporally associated with mafic rocks that include gabbroids, anorthosite, and diabase dikes (e.g., *Haapala and Rämö, 1990*; *Rämö, 1991*; *Eklund et al., 1996*). The gabbroic and anorthositic rocks are found as relatively small bodies (except in the Riga batholith; Fig. 1a) within or at the margins of the rapakivi granite complexes, whereas the diabase dikes form several dike swarms that transect the metamorphic bedrock surrounding the complexes, and, in some instances, cut the rapakivi plutons as well. Evidence for coexisting silicic and mafic magmas, and local mingling and mixing between these end members, has been reported from several rapakivi complexes (*Rämö, 1991*; *Eklund, 1993*; *Salonsaari, 1995*). Volumetrically minor intermediate (monzodioritic) rocks are present in at least two rapakivi plutons, Ahvenisto and Åland (*Eklund et al., 1994*; *Alviola et al., 1999*).

During the past two decades, the Fennoscandian rapakivi complexes have been the subject of extensive U-Pb chronological work (see *Haapala and Rämö, 1999*, and references therein) and fall into four age zones that delineate a gross east-west pattern (Fig. 1a). The classic Wiborg batholith and associated plutons in southeastern Finland and Estonia are 1.67–1.62 Ga old whereas those in southwestern Finland, Latvia (the Riga batholith), and west-central Sweden (Nordingrå) are slightly younger, 1.59–1.54 Ga. The rapakivi granites in central Sweden (west of Nordingrå) have been dated at 1.53–1.47 Ga and those in Russian Karelia (Salmi) at 1.55–1.54 Ga. U-Pb data on the associated mafic and intermediate rocks show that, even though they are often cut by the rapakivi granites, no substantial age difference has been observed (e.g., *Vaasjoki et al., 1991*; *Alviola et al., 1999*).

In terms of petrography, mineral chemistry, geochemistry, and magmatic association, the Fennoscandian rapakivi granites are typical subalkaline A-type granites (e.g., *Haapala and Rämö, 1992*). Their petrogenesis has often been attributed to mafic underplating and crustal anatexis in an extensional tectonic regime (e.g., *Haapala and Rämö, 1990*; *Rämö, 1991*; *Korja and Heikkinen, 1995*). Petrochemical studies suggest that rapakivi complexes in southeastern Finland (*Rämö, 1991*; *Kosunen, 1999*) and central Sweden (*Andersson, 1997*) were derived from quartzofeldspathic protoliths that underwent dehydration melting in response to thermal perturbations in the mantle.

3. Nd isotope Geochemistry

The Sm-Nd isotopic system has proven to be an invaluable indicator of the overall age and character of the protoliths involved in rapakivi granite genesis (e.g., *Rämö and Haapala, 1996*). As the rapakivi granites and associated mafic rocks contain material from the deep crust and upper mantle, respectively, they provide information from the unexposed parts of the lithosphere.

The initial Nd isotopic compositions of the Fennoscandian rapakivi granites show important variations (Fig. 1b). The Finnish, Estonian, and Latvian rapakivi granites and the Nordingrå pluton of Sweden have $\varepsilon_{Nd\,i}$-values between –3 and 0. This suggests derivation from the ~1.9 Ga Svecofennian crust (e.g., *Haapala and Rämö, 1990*) and that, in this region, the 1.9 Ga crust is quite homogeneous in terms of overall mantle separation age. The Riga batholith of Latvia has, however, a slightly higher $\varepsilon_{Nd\,i}$-value than the other plutons, which suggests that the lithosphere may be slightly more juvenile in the south (*Rämö et al., 1996*). Compared with the Finnish occurrences, the Salmi batholith of Russian Karelia is quite different in having more negative $\varepsilon_{Nd\ i}$-values (–9 to –5.5). This indicates a substantial Archaean component in the protolith of the batholith, which is not surprising as it is situated between Paleoproterozoic and Archaean domains (*Rämö, 1991*; see also *Neymark et al., 1994*; Fig. 1a). Similar, yet far more surprising, is the Nd isotopic composition registered by the rapakivi granites of central Sweden with $\varepsilon_{Nd\ i}$-values between –7.5 and –4.5 (Fig. 1b). No exposed Archaean crust is known in central Sweden but the rapakivi granites show that in the deep parts of the crust, an Archaean component is present (*Andersson, 1997*).

The mafic rocks associated with the rapakivi granites show Nd isotopic compositions that differ only from those of the granites marginally (Fig. 1b). The mafic rocks associated with the Finnish rapakivi granites show $\varepsilon_{Nd\ i}$-values that range from slightly positive (+1.5) to slightly negative (–3). This variation has been ascribed to a slightly depleted mantle protolith (ε_{Nd} +1 to +2; *Rämö, 1991*; *Fröjdö et al., 1996*) and the contamination of the mantle-derived melts by the Svecofennian crust. The mafic rocks associated with the Russian Karelian and the central Swedish rapakivi granites (ε_{Nd} –8 to –6.5 and –10 to –4.5, respectively) are indicative of an enriched mantle protolith and/or contamination by a crust with an Archaean component (see *Neymark et al., 1994*; *Andersson, 1997*). The fact that the Swedish mafic rocks have, on average, slightly more negative $\varepsilon_{Nd\ i}$-values than the associated granites (–10 to –4.5 vs. –7.5 to –4.5; *Andersson, 1997*; *Persson, 1997*) suggests that the upper mantle in this area may include a slightly larger Archaean component than the lower crust. Thus the Nd isotopic composition of some of these mafic rocks may reflect rather mantle source characteristics than crustal contamination (*Rämö and Korja, 2000*).

4. Geophysics

Extensive geophysical studies have been performed in order to reveal the deep structure of the Fennoscandian Shield (see *Korsman et al., 1999*, for review). Deep seismic soundings along several profiles (e.g., SVEKA, BALTIC) have yielded a comprehensive picture of the structure of the Svecofennian lithosphere. Figure 1a shows the variation of crustal thickness (Moho depth) across the central and southern part of the Fennoscandian Shield as well as its southern continuation.

The Svecofennian Orogen is characterized by a thick crust (at maximum, 65 km) that has particularly steep and local ovoid thinnings associated with the rapakivi granites (e.g., *Luosto, 1997*; *Korsman et al., 1999*; see Fig. 1a). The rapakivi granite batholiths are found as relatively thin (~5–10 km) sheet-like bodies in the uppermost part of the crust and are characterized by circular regional Bouguer gravity anomaly minima and smooth magnetic patterns with sharply cutting anomaly contacts (*Elo and Korja, 1993*; *Korja et al., 1993*). The associated gabbroic and anorthositic rocks form local positive anomalies. The large Bouguer gradients result from the combination of the upper crustal rapakivi granites and thinning of the lower crust beneath them. The crustal section that hosts the rapakivi granites is characterized by listric seismic reflectors that tend to flatten out at either the lower to middle crustal boundary or at the Moho boundary, whereas the rapakivi granite and related mafic intrusions appear as upper crustal (0–20 km) non-reflective bodies delineated by listric and normal shear zones (Fig. 2). The lower crust is highly reflective and tends to thin and bow up, presumably in response to extension and removal of anatectic material to higher crustal levels. In places, it is also underlain by an unreflective lowermost crust and a clear, reflective Moho. Highly reflective structures within the batholiths are related to contacts between rapakivi granite and associated mafic rocks or the younger Postjotnian diabases. The gabbroic parts have higher velocity, density, and magnetic susceptibility than the surrounding rocks (*Korja et al., 2000*).

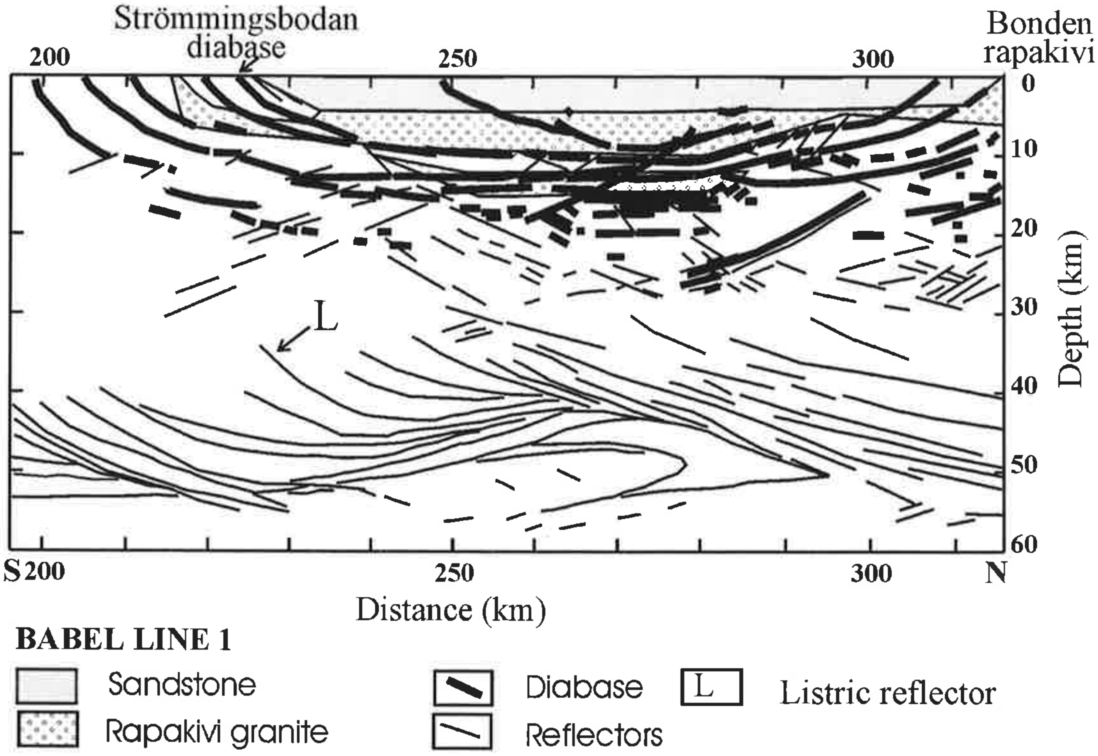

Figure 2. Geological interpretation of the northern half of the deep seismic sounding line BABEL 1, showing tectonic structures resulting from the emplacement of the rapakivi granites and Postjotnian diabase dikes in an extensional tectonic setting. A system of listric shear zones (L) and conjugate transfer faults govern the spatial and temporal relationships of the Subjotnian rapakivi granites and gabbroic and anorthositic rocks, Jotnian sandstones, and Postjotnian diabases. After *Korja and Heikkinen (1995) and Korja et al., 2000.*

The highly reflective lower crust shows upward concave structures that may result from upward movement of the lower crust as the lithosphere was in a process of overall extension. The high reflectivity can be interpreted as having been caused by mafic underplating and intraplating that acted as a thermal energy source for the lower crustal melting that generated the rapakivi granite magmas. The listric shear zones that detach either at the lower crust to middle crust boundary or the Moho boundary can be envisioned as pathways for the ascent of the rapakivi and associated mantle-derived mafic magmas. The space created by the extending upper crust was occupied by the rapakivi granite magmas as well as the uprising lower crust and mantle. Cratonic basins with continental sediments were developed due to thermal contraction of the rift zones (Fig. 2; *Korja and Heikkinen, 1999*; see also *Klein and Hsui, 1987*). In fact, the Baltic Sea has many characteristic features of paleorifts: topographic low (now under water), relatively thin crust with large crustal gradients, and voluminous bimodal magmatism (*Korja and Heikkinen, 2000*).

5. Lithospheric Models

There are three different models to explain the origin of the rapakivi granites in relation to orogenic evolution (*Haapala and Rämö, 1999*):

(1) Melting of thickened orogenic crust (e.g., *Vorma, 1976*; *Windley, 1991*);

(2) Mafic underplating and subsequent melting of the crust by mafic magmas (e.g., *Bridgwater et al., 1974*; *Anderson, 1983*; *Haapala and Rämö, 1990*); and

(3) Intracratonic magmatism related to convergent processes at craton margins (e.g., *Van Schmus, 1996*; *Bettencourt et al., 1999*).

In the light of the extensive geophysical evidence for crustal thinning associated with rapakivi magmatism, the hypothesis of orogenic crustal thickening leading to rapakivi magmatism is obsolete. Anatexis by dehydration requires high thermal input which, at least in terms of the huge volume of the Fennoscandian rapakivi granites, would probably be impossible to be accounted for by tectonic thickening only.

The extensional tectonic setting, silicic-mafic magmatism, and geochemical characteristics of the rapakivi granites are best explained by the mafic underplate model. The ultimate cause of the underplating remains largely open, however. It could have been due to active or passive rifting, extensional collapse of orogen, or deep mantle plumes, or could have been triggered by compositionally instable domains in the upper mantle (*Haapala and Rämö, 1992, 1999*). It should also be noted that the age of the rapakivi granite magmatism across southern Fennoscandia (Fig. 1a) does not fit a hot spot track but may rather reflect a spreading plume head underneath the lithosphere (cf. *Zeyen et al., 1997*; *Korja and Heikkinen, 2000*). In the plume hypothesis, lateral spreading of the thermal source and net strengthening of thinned crust could have led to cessation of extension and anatexis in any one area and spreading of deformation and magmatism to adjacent areas of weaker (and thicker) crust (*Korja and Heikkinen, 1995*).

The third model that explains the origin of the rapakivi granites as intraplate manifestations of convergent margin processes has been suggested for the rapakivi-type anorogenic granites of mid-continental U.S.A. (*Van Schmus, 1996*) and western Amazonian craton (*Bettencourt et al., 1999*). Interestingly, the Gothian (~1.5–1.65 Ga) arc-type granitoid magmatism on the southwestern flank of the Fennoscandian Shield was coeval with the emplacement of the rapakivi granites in the east-central part of the shield. Whether there was a genetic relationship (see *Åhäll and Connelly, 1998*) between these two granitoid suites is yet to be determined.

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Fennoscandia as a Part of a Proterozoic Supercontinent

Satu Mertanen [1] and Lauri J. Pesonen[1,2]

[1] Geological Survey of Finland, P.O. Box 96, FIN-02151 Espoo, Finland
[2] Department of Geophysics, P.O. Box 4, FIN-00014 University of Helsinki, Finland
E-mail: satu.mertanen@gsf.fi

Global palaeomagnetic data were updated by palaeomagnetists from different continents at a Workshop held in Aarhus, Denmark in November 1999. The aim of the workshop was to define reliability criteria for palaeomagnetic data in order to facilitate continental reconstructions between different cratons through time. By continental reconstructions it is possible to study the development of the Earth and also to reveal geological connections between stable cratons and orogenic belts and between economically important formations which were once assembled together, but are now dispersed due to later break up of supercontinents. The main purpose of the Palaeoproterozoic subgroup of the Workshop was to study Precambrian supercontinents that might have existed before the formation of the supercontinent Rodinia at ca. 1.0 Ga ago (e.g. Dalziel, 1991). Global continental reconstructions were established for six time periods where data are available from many of the continents, at 2.0, 1.83, 1.65, 1.25, 1.1 and 1.05 Ga (see Pesonen et al., 2000). In addition, well-defined data with less strict confidence criteria also exist from Archaean and Early Palaeoproterozoic formations in some of the cratons.

Keywords: Fennoscandia, reconstruction, palaeomagnetism, Archaean, Proterozoic

1. The Archaean

Archaean palaeomagnetic data have been obtained from the Fennoscandian Shield (Karelian craton), Laurentia (Superior craton), Australia (Pilbara craton) and Africa (Kaapvaal craton). Based on these data it is suggested that at 2.85 Ga the Karelian craton was at equatorial palaeolatitudes, but at 2.6 Ga it had drifted to high palaeolatitudes, indicating that considerable drift of cratons took place already during the Archaean. Palaeomagnetic data from Superior craton also indicate high palaeolatitudes at 2.7-2.6 Ga and therefore Karelia and Superior may have been together at that time. Pilbara and Kaapvaal cratons occupied high palaeolatitudes at 2.87 Ga (*Zegers et al., 1998*), in contrast to low palaeolatitudes of Karelia, suggesting that at least during the early Archaean no global supercontinent existed (see *Mertanen, 2000*).

2. The Early Palaeoproterozoic

Palaeomagnetic data from 2.45 Ga Karelian dykes of the Karelian craton and from the Matachewan dykes of the Superior craton imply that the two cratons may have been joined at 2.45 Ga, both occupying a shallow palaeolatitude near the equator (Fig. 1). The dykes are aligned, suggesting that they formed a single dyke swarm, together with the layered intrusions on both cratons at ca. 2.45 Ga ago. Palaeomagnetic data from Lapland-Kola Orogen are not so well defined, but according to recent studies (see *Mertanen et al., 1999a*) the palaeopoles of Karelia and Kola differ at 2.4 Ga, suggesting that the two cratons were ca. 1000 km apart. In addition to those of Karelia and Superior, 2.4 Ga old palaeopoles are available from the Yilgarn craton of Australia, indicating that Yilgarn was at high palaeolatitudes and not connected to Karelia and Laurentia.

At 2.0 Ga there are no firm palaeomagnetic evidences to make an assembly of continents. Palaeomagnetic data do exist from the Superior, Ukraine, West African, Kalahari and Congo/São Fransisco cratons which show that at 2.0 Ga the cratons occupied low palaeolatitudes. Well-defined poles exist especially from Laurentian 1.998-2.217 Ga dykes and sills (*Buchan et al., 2000*), but due to more conspicuous overprinting, reliable Fennoscandian data from coeval formations are still lacking. The oldest well-defined Fennoscandian data after 2.45 Ga are from the 1.93 Ga Tsuomasvarri intrusion from the Lapland-Kola Orogen. Coeval data are also available from the Amazonian craton showing that Amazonia and Fennoscandia occupied relatively similar palaeolatitudes at ca. 1.93 Ga (equatorial) and at ca. 1.8 Ga (20°) which may indicate that there was a connection between these cratons during 1.93-1.8 Ga.

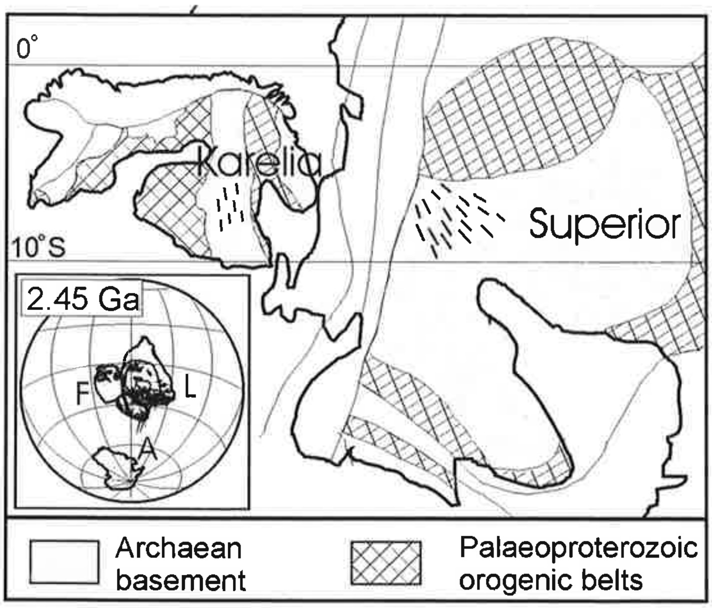

Figure 1. Continental reconstruction at 2.45 Ga showing the occurrence of coeval Matachewan and Karelian dykes (*Mertanen et al. 1999b*). Based on this reconstruction it is suggested that the Superior Craton of Laurentia and the Karelian Craton of the Fennoscandian Shield were joined at ca. 2.45 Ga ago. The inset map shows the relative locations of the Fennoscandian Shield (F), Laurentia (L) and Australia (A).

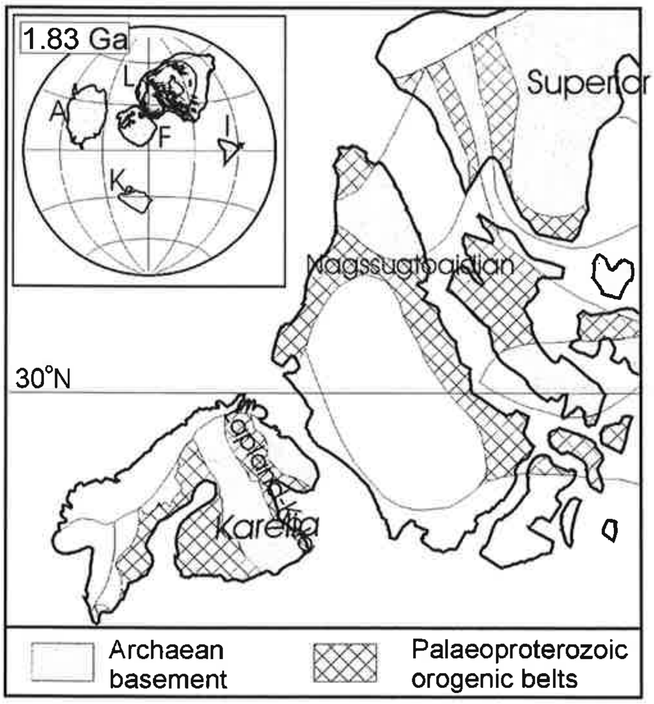

Figure 2. Continental reconstruction of the Fennoscandian Shield and Laurentia at 1.83 Ga. The inset map shows the relative locations of the Fennoscandian Shield (F), Laurentia (L), Amazonia (A), Kalahari (K) and India (I).

3. The Middle Palaeoproterozoic

In the 1.8 Ga reconstruction, both the Fennoscandian Shield and Laurentia have drifted to higher palaeolatitudes compared with those at 2.0 - 1.93 Ga. For the Fennoscandian Shield, data from the 1.83 Ga Haukivesi lamprophyric dykes were used and for Laurentia, data from the 1.827 Ga Sparrow dykes (see *Buchan et al., 2000*). This reconstruction (Fig. 2) places the Kola Peninsula adjacent to the present East Greenland so that the Lapland-Kola belt and the Nagssugtoqidian belt of Greenland form a ca. 90° angle. The relative orientation of the Fennoscandian Shield and Laurentia is about the same at 1.83 Ma and at 1.88 Ga (*Pesonen, 1995*), when less reliable data are used, but differs from that at 2.45 Ga, indicating that between 2.45 Ga and 1.88 Ga the two cratons were separated and eventually collided at a different position before 1.88 Ga. Palaeomagnetic data of the age of 1.9-1.8 Ga from the Kola and Karelia cratons are in agreement implying that the two cratons formed a unity at that time. Palaeomagnetic data also exist from the Amazonian, Kalahari and Indian cratons, but they cannot be matched with each other like the Fennoscandian Shield and Laurentia.

Reliable palaeomagnetic data from 1.65 Ga formations of different continents are still too sparse for global reconstructions. Well-defined data exist from the Fennoscandian Shield, Amazonian, Gawler (Australia) and South China cratons and they all, except for Gawler, occupy low equatorial palaeolatitudes. Compared with the palaeoposition at 1.83 Ga, both the Fennoscandian Shield and the Amazonian craton have drifted to lower palaeolatitude preserving about the same relative position as in 1.83 Ga.

4. The Mesoproterozoic

Well-defined palaeomagnetic data have been obtained both from the 1.267 Ga Mackenzie dykes in Laurentia and from the 1.26 Ga Jotnian dykes from the Fennoscandian Shield which allows a reliable reconstruction (Fig. 3). At that time the Fennoscandian Shield and Laurentia had approximately the same relative orientation as at 1.83 Ga, suggesting that between 1.83 and 1.26 the two cratons drifted together. Both continents have drifted ca. 30° towards the equator and coevally rotated counterclockwise by ca. 90°. Another noticeable feature in the configuration is the closeness of the Congo/São Fransisco craton with the southern part of the Fennoscandian Shield. From Amazonia there are no reliable palaeomagnetic data from that time. Australia has maintained the same high palaeolatitude as at 1.65 Ga and is not connected to the joined Fennoscandia/Laurentia.

According to a new palaeomagnetically derived geodynamic scenario, the Fennoscandian Shield and Laurentia started to drift apart after 1.25 Ga about 1.2 Ga ago. The Congo/São Fransisco craton collided with the southern Fennoscandian Shield causing the formation of the Sveconorwegian orogen. Likewise, the Amazonian craton collided with Laurentia resulting in the Grenvillian orogen. Consequently, a theoretical Congo Ocean existed at 1.1 Ga ago between Laurentia/Amazonia and Fennoscandia/Congo-São Fransisco (Fig. 4). From the Amazonian craton, however, no reliable palaeomagnetic data are available from the 1.1 Ga formations, but its palaeoposition is based on geological interpretations. Geological evidences indicate that both the Fennoscandian Shield and the Congo craton experienced a rifting and orogenic episode at ca. 1.25-1.0 Ga, manifested by the Sveconorwegian orogeny and dyke swarms in the Fennoscandian Shield and the continent-continent type Kibaran orogeny in the Congo craton at 1.3-1.1 Ga, thus giving support to the palaeomagnetically derived reconstruction. At 1.1 Ga the Gondwana supercontinent, comprising dominantly of Australia, East Antarctica and India already existed, locating north of Laurentia (Fig. 4).

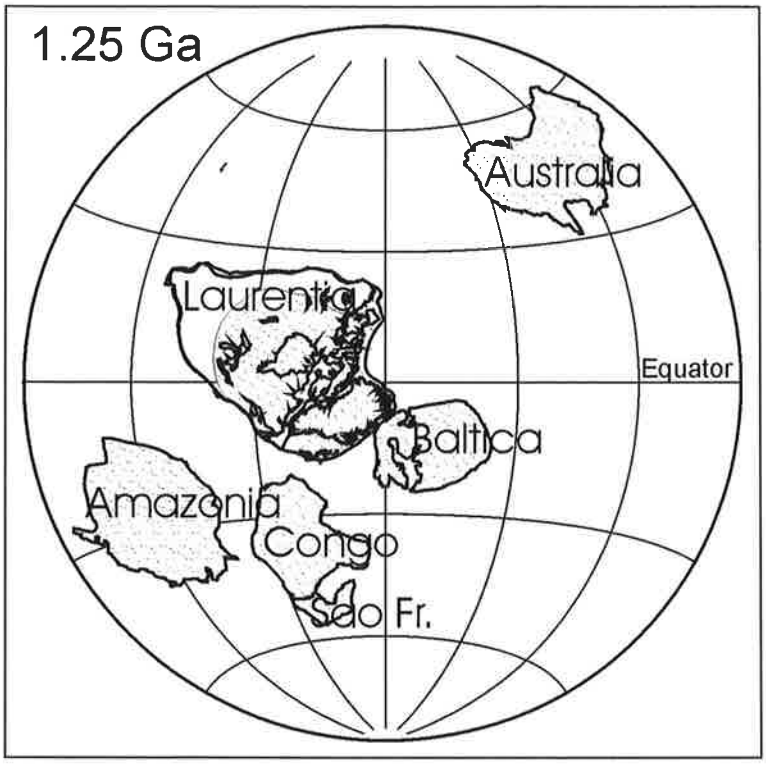

Figure 3. Continental reconstruction of the Fennoscandian Shield (Baltica), Laurentia, Amazonia, Congo/São Fransisco and Australia at 1.25 Ga.

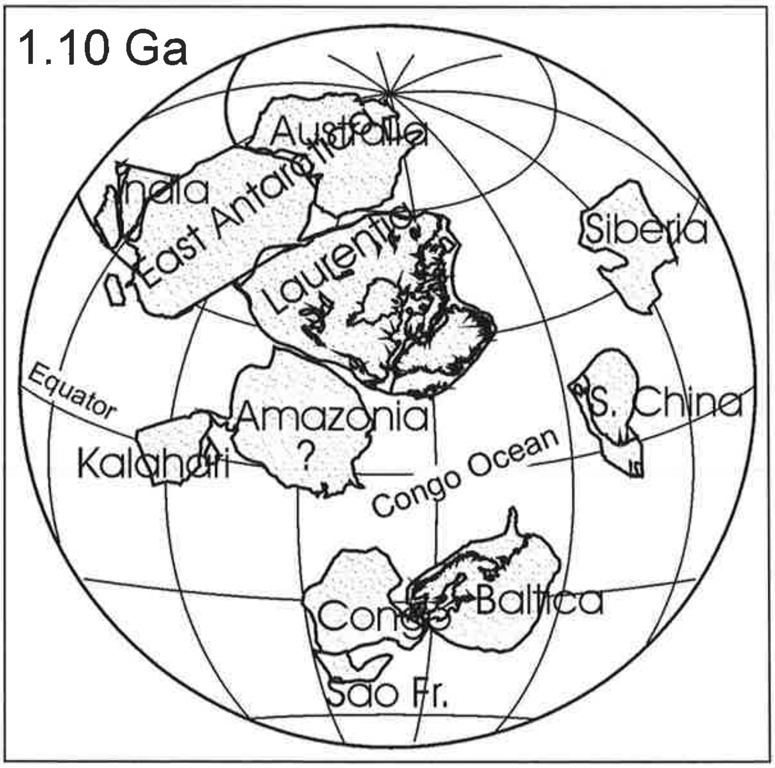

Figure 4. Continental reconstruction of the Fennoscandian Shield (Baltica), Congo/São Fransisco, Laurentia, Amazonia, Kalahari, South China, Siberia and Gondwana (Australia, East Antarctica, India) at 1.10 Ga.

5. The Rodinia Supercontinent Assembly

Continental reconstruction based on upgraded palaeomagnetic data at 1.05 Ga are in close agreement with the previously proposed reconstruction of the Rodinia supercontinent. Laurentia has drifted to lower palaeolatitudes, resulting to the closure of the Congo Sea due to the collision of the SE coast of Greenland to the northern Fennoscandia/Congo-São Fransisco craton. These cratons amalgamated with other continents, Gondwana, Amazonia, Siberia, South China and Kalahari to form the supercontinent Rodinia.

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Geological Evolution of the Central Lapland Greenstone Belt

Eero Hanski[1] and Hannu Huhma[2]

[1]Geological Survey of Finland, P.O. Box 77, FIN-96101 Rovaniemi, Finland
[2]Geological Survey of Finland, P.O. Box 96, FIN-02151 Espoo, Finland
E-mail: eero.hanski@gsf.fi

Six hundred million years of the Palaeoproterozoic depositional evolution is recorded in the Central Lapland Greenstone Belt, beginning ca. 2450 Ma ago with the eruption of the mantle plume-related, komatiitic to rhyolitic volcanics on Archaean gneisses. In the following 300-400 Ma period, deposition of a thick, transgressive succession of the quartzite-dolomite-basalt-pelite association took place, accompanied by komatiitic to picritic volcanic rocks at a late stage. The hitherto extensional geotectonic evolution was interrupted by a collisional event ca. 1920 Ma ago, which led to the thrusting of a slab of an ancient oceanic lithosphere onto older cratonic rocks. The supracrustal rock sequence was completed with the deposition of molasse-like, coarse-clastic sediments less than 1880 Ma ago.

Keywords: evolution, geochronology, greenstone belt, Palaeoproterozoic, Lapland

1. Introduction

In the northern part of the Fennoscandian Shield, Palaeoproterozoic supracrustal rocks cover the Archaean basement as an almost uninterrupted zone extending from northern Norway through Central Finnish Lapland and into Russia (Fig. 1). The Finnish part of this zone is called the Central Lapland Greenstone Belt (CLGB). Although the supracrustal rocks of the CLGB have most often been regarded as belonging to the Karelian formations, their correlation with the traditional Karelian formations, viz. the Sariolian, Jatulian and Kalevian, has been controversial for decades, partly due to the lack of reliable age determinations. Nevertheless, the discovery made already by *Hackman (1927)*, indicating the presence of two principal stratigraphic entities separated by a major unconformity, has generally been accepted. For the younger unit containing exclusively metasedimentary rocks, *Hackman (1927)* coined the name **'Kumpu quartzites'**, while *Sederholm (1932)* later introduced the frequently used the term **'Lapponian'** for the widely distributed volcano-sedimentary sequence of the lower unit. A premise for the adoption of the latter term and implicit in subsequent stratigraphic schemes (e.g., *Silvennoinen, 1985*) was the notion that the Lapponian rocks are *older* than Jatulian, i.e. exceeding 2.2 Ga. However, geological, geochemical and geochronological investigations carried out since 1984 by the Lapland Volcanite Project have revealed that both the Kumpu quartzites and a large part of Sederholm's Lapponian volcanic rocks are *younger* than the Jatulian and do not have any counterparts in terms of conventional Karelian stratigraphic units (*Lehtonen et al., 1998*). This contradiction then led to rejection of the previous stratigraphic nomenclature (*Räsänen et al., 1995*). The former Lapponian rocks are now assigned to five lithostratigraphic groups which are, from oldest to youngest, the Salla, Onkamo, Sodankylä, Savukoski, and Kittilä Groups, while the traditional Kumpu formations are divided into two units, the Lainio and Kumpu Groups (*Lehtonen et al., 1998*). Their occurrence in the central part of the belt is shown in Figure 2.

2. The Salla Group

The early Proterozoic depositional history in Finnish Lapland began in a continental environment with subaerial eruption of intermediate to acid volcanic rocks and deposition of minor arkosic sediments of the Salla Group, occurring most widely in the SE part of the

CLGB. Generation of these rocks was related to an initial stage of rifting of the Archaean craton. A U-Pb zircon age of ca. 2520 Ma has been measured for a volcanic breccia at Peurasuvanto (*Pihlaja and Manninen, 1988*), but due to slightly heterogeneous data, the precise age for this volcanic unit remains questionable. More confident dating results have been recently obtained for felsic metavolcanics from the southern shore of the Lokka reservoir and from the western side of the Koitelainen intrusion, yielding U-Pb zircon ages of 2438±14 and 2438±11 Ma, respectively.

Figure 1. Palaeoproterozoic Central Lapland Greenstone Belt (grey) in Finnish Lapland.

3. The Onkamo Group

A type section of the next unit, the Onkamo Group has been mapped around the Möykkelmä granite-gneiss basement dome (Sodankylä), comprising a subaerially erupted sequence of komatiites, tholeiitic basalts, basaltic andesites and andesites which, in common with the Salla Group rocks, display a strong signature of crustal contamination in their chemical composition (*Räsänen et al., 1989*). The Salla Group is found to be cut by ca. 2440 Ma, chromitite-bearing mafic layered intrusions at Akanvaara and Koitelainen, but such a relationship is not observed in the Onkamo Group, thus suggesting a possible coeval nature for the intrusions and the Onkamo Group metavolcanites. Tholeiitic basalts of the Möykkelmä area are exceptionally rich in chromium (700-1100 ppm). It seems that the Salla and Onkamo Groups do not deviate much in age and belong to a magmatic episode producing large volumes of igneous rocks in the form of layered intrusions, dyke swarms and lavas over large areas within a short interval of time, thus representing characteristic features of magmatism related to a starting mantle plume (cf. *Ernst and Buchan, 1997*).

4. The Sodankylä Group

The active volcanic period was later followed by a more tranquil period resulting in the deposition of widespread epiclastic sedimentary rocks of the Sodankylä Group, which are now represented mainly by quartzites and less abundant carbonate rocks with occasional stromatolitic structures and mica schists. These are accompanied by less abundant mafic metavolcanites. Mafic magmatism is also manifested by hypabyssal intrusions forming concordant, 2.2 Ga, differentiated mafic-ultramafic sills (assigned to the Haaskalehto-type intrusions in Central Lapland) in a fashion similar to the differentiated sills in eastern Finland which are observed within Jatulian quartzites and the underlying Archaean basement. The coexistence of this particular-type of sills with quartzites is not a mere coincidence with regard to the correlation of Jatulian-type quartzites in eastern and northern Finland (*Hanski, 1986*). Primary sedimentary structures indicative of tidal action suggest that at least a part of the Sodankylä Group was deposited under continental margin conditions (*Lehtonen et al., 1998*).

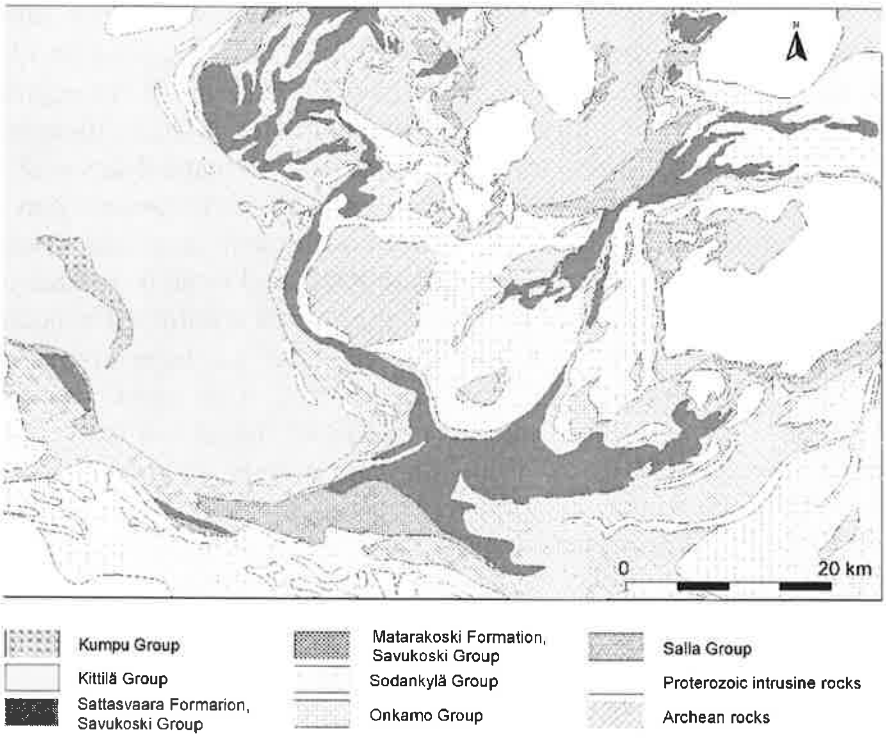

Figure 2. Stratigraphical map of the central part of the Central Lapland Greenstone Belt (simplified after *Lehtonen et al., 1998* and *Korsman et al., 1997*).

5. The Savukoski Group

A gradual deepening of the depositional basin led to the accumulation of more fine-grained sediments, phyllites and black schists and some mafic tuffites of the Savukoski Group (Matarakoski Formation in Fig. 2). The pelitic metasediments have been intruded by the Ni-Cu-sulphide-bearing Keivitsa mafic intrusion dated at c. 2050 Ga, thus providing a minimum age for these sedimentary rocks. These are overlain by primitive volcanic rocks of the komatiite-picrite association which can be followed for more than 350 km towards the north to northern Norway. They exhibit volcaniclastic and pillow structures with spectacular outcrops in the Sattasvaara area (Sattasvaara Formation in Fig. 2). Geochemically they form a diverse group of rocks varying from extremely LREE-depleted, but moderately Ti-enriched komatiites to LREE- and HSFE-enriched picrites. Special efforts have recently been paid to unravel the age of these rocks, giving the following results: The seven whole rock – clinopyroxene pairs analysed from komatiites at Jeesiörova provide an average Sm-Nd age of 2056±25 Ma (1 std; *Hanski et al., in press*). The initial ε_{Nd} values range from +2 to +4, with the most LREE-depleted komatiites having the highest ε_{Nd}. The available data also suggest that the komatiites and picrites were generated from a geochemically heterogeneous, but isotopically similarly depleted mantle source.

6. The Kittilä Group

The Kittilä Group, which forms a large volcanite-dominated complex in Central Finnish Lapland (Kittilä area), is in tectonic contact with other lithostratigraphic units described above. It comprises various kinds of submarine mafic metavolcanites and their cogenetic dykes, geochemically resembling modern E-MORB, N-MORB, OIB and IAT, associated with graphite schists, BIF and other chemical metasediments. Based on whole-rock and clinopyroxene analyses, a Sm-Nd isochron age of 1990±35 Ma with an initial ε_{Nd} value of +3.7±0.2 has been obtained for the tholeiitic lavas of the Vesmajärvi Formation. At the eastern edge of this volcanic complex, there exists a chain of serpentinites interpreted as dismembered ophiolitic mantle rocks (*Hanski, 1997*). These are cut by ultramafic to mafic dykes with boninitic, island arc tholeiitic and calc-alkaline affinities. On the other hand, in the western part of the complex, dacitic to rhyolitic dykes are intimately associated with mafic lavas and dykes demonstrating their contemporaneous emplacement. Hence these felsic rocks can be utilized to determine the age for the associated mafic rocks. Four dated porphyries record U-Pb zircon ages between 2012±5 and 2018±7 Ma and have similar positive initial ε_{Nd} values (+3.8) to those of the tholeiitic mafic metavolcanites, suggesting that no old sialic crust was involved in their generation. Another, younger felsic magmatic pulse of dykes and small plutons within the Kittilä Group has been dated to ca. 1920 Ma. The current interpretation suggests that the Kittilä Group is an allochthonous complex representing a block of ancient oceanic lithosphere, which was obducted to its present position at the time of the generation of the younger group of felsic dykes (*Hanski et al., 1998*).

7. The Kumpu and Lainio Groups

Molasse-type conglomerates and quartzites of the Lainio and Kumpu Groups lie unconformably on the folded rocks of the older stratigraphic units. These two groups were distinguished on the basis of their deformation history (*Lehtonen et al., 1998*), but it has now become evident that the deposition of both of them took place relatively late. *Hanski and others (1997)* demonstrated the existence of felsic porphyry pebbles in a Kumpu Group conglomerate, having petrological characteristics and an age (1928±6 Ma) similar to those of the younger felsic dyke type found within the Kittilä Group. Recent NORDSIM studies show that in both the Lainio and Kumpu Groups, there exists a detrital zircon population with an age of about 1880 Ma which matches well with the age of the felsic to intermediate plutons of the Haaparanta Suite occurring in western Finnish Lapland (*Hanski et al., 2000*). A Haaparanta Suite provenance for the conglomerates is further corroborated by the conventional U-Pb zircon dates of 1888±22 Ma and 1873±11 Ma obtained for granitoid and felsic porphyry pebbles in the Kellostapuli and Vesikkovaara conglomerates of the Lainio Group, respectively.

At Latvajärvi, the Lainio Group also comprises minor intermediate to felsic volcanics, which have been dated at 1880±8 Ma and can be correlated with coeval felsic porphyries occurring widely in northern Sweden.

8. Conclusions

In conclusion, the ~600 Ma evolution of the CLGB commenced with plume-related, intracratonic rift magmatism producing a variety of rock types ranging from crustally contaminated komatiites to rhyolites and large layered intrusions. These were subsequently covered by epicontinental clastic sediments and then by deeper-water pelitic sediments. The corresponding stage (Jatulian to Lower Kalevian) in central and eastern Finland has traditionally been connected with sedimentation occurring during a platform to continental margin development. Following submarine eruptions of komatiitic to picritic volcanites have been dated to ca. 2060 Ma. This magmatic event has also been related to mantle plume activity (*Puchtel et al., 1998*), but this interpretation has some problems as there is no evidence for a preceding stage of strong crustal uplift in the stratigraphy. Tectonic emplacement (obduction) of a slab of oceanic lithosphere (the Kittilä Group) took place at ca. 1920 Ma and was followed by high-pressure metamorphism of the Lapland Granulite Belt, intrusion of the 1880 Ma Haaparanta Series plutonic rocks, extrusion of minor intermediate to felsic volcanics and deposition of the Lainio and Kumpu Group coarse-clastic sediments. Still later, intrusion of granites occurred ca. 1800 Ma ago.

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PART II: POSTERS

Chromitite-bearing Ultramafic Rocks in the Ukrainian Shield: New Evidence for Paleoproterozoic Ophiolitic Mantle Rocks

Stanislav S. Gornostayev[1,3], Eero Hanski[2], Kauko V.O. Laajoki[1] and Sergiy E. Popovchenko[3]

[1]Department of Geosciences, P.O. Box 3000, FIN-90014 University of Oulu, Finland
[2]Geological Survey of Finland, P.O. Box 77, FIN-96101 Rovaniemi, Finland
[3]National Mining University of Ukraine, 19 Karl Marx Ave., Dnepropetrovsk, 49027, Ukraine
E-mail: ssg@cc.oulu.fi

The Golovanevsk suture zone, which separates the Volyno-Podolsky (Western) and Central Ukrainian geoblocks of the Ukrainian Shield, contains Paleoproterozoic fault-bounded, ultramafic massifs with characteristics comparable to those of ophiolitic complexes, including Paleoprotorezoic ophiolites of the Fennoscandian Shield. Among these Ukrainian massifs are Kapitanov and Lipovenki, hosting chromitites with platinum-group elements (PGE) characteristics typical of ophiolitic rocks. At Kapitanov, Al-rich massive and disseminated chromite ores are hosted by a layered dunite-peridotite sequence. The Lipovenki massif is composed of mantle dunite-harzburgite and hosts nodular, schlieren and sulfur-poor massive chromitites with textures and chromian spinel compositions similar to those reported from ophiolitic complexes. The platinum-group minerals (PGM) in chromitites from the Lipovenki and Kapitanov are mostly represented by sulfides, arsenides and sulfarsenides of Ir-group PGE. All these observations suggest that the Golovanevsk suture zone contains remnants of Paleoproterozoic oceanic lithosphere.

Keywords: ophiolite, chromitite, Paleoproterozoic, platinum-group elements, Ukrainian Shield

1. Introduction

Ophiolitic rocks, which represent slabs of the oceanic lithosphere tectonically emplaced to a continental environment, provide key evidence for the presence of major suture zones. The number of well-established occurrences of ophiolitic rocks in the early Precambrian terranes is still very small with the best-known examples being restricted to the Fennoscandian and Canadian Shields (*Kontinen, 1987; Scott et al., 1992; Vuollo et al., 1995; Hanski, 1997*). According to *Dagelaysky (1997)*, the Ukrainian Shield is divided into three geoblocks separated by two, N-S trending suture zones (Fig. 1). The Golovanevsk suture zone between the Volyno-Podolsky (Western) and Central Ukrainian geoblocks is known to contain ultramafic rocks and related chromite deposits in the Pobugskoe region, as revealed for the first time by N.T. Vadimov in 1952 (unpublished data) in connection with drilling for lateritic nickel deposits. The existence of ophiolitic rocks in the Ukrainian Shield was suggested by *Nalivkina (1977)* based on interpretations of regional geological and geophysical data. Later investigations by the Ministry of Geology of Ukraine have produced more detailed geological, geophysical and tectonic data which support her suggestions *(Galetsky et al., 1984)*. Our recent studies have provided more evidence for the ophiolitic nature of some of the Paleoproterozoic chromite-bearing ultramafic bodies in the Ukrainian Shield (*Gornostayev et al., 1999; Gornostayev et al., 2000*). In this paper, we present mineralogical and geochemical data on two ultramafic massifs, Kapitanov and Lipovenki, located in the Pobugskoe region within migmatites and gneisses of the Teterev-Bug series (*Ershov, 1960; Kanevskii, 1996*).

2. The Kapitanov Massif

The Kapitanov massif is a layered, fault-bounded dunite-peridotite body that is 2500 m long and 260 m wide. The age of the ultramafic rocks in the Pobugskoe region is estimated between 1.96-2.1 Ga based on K-Ar dating on amphiboles (*Kanevskii, 1981*). The predominant peridotite (harzburgite) at Kapitanov is composed of olivine, orthopyroxene, chromite, accessory green spinel and minor clinopyroxene (Table 1). The massif also includes pyroxenites, gabbros and carbonate-rich rocks.

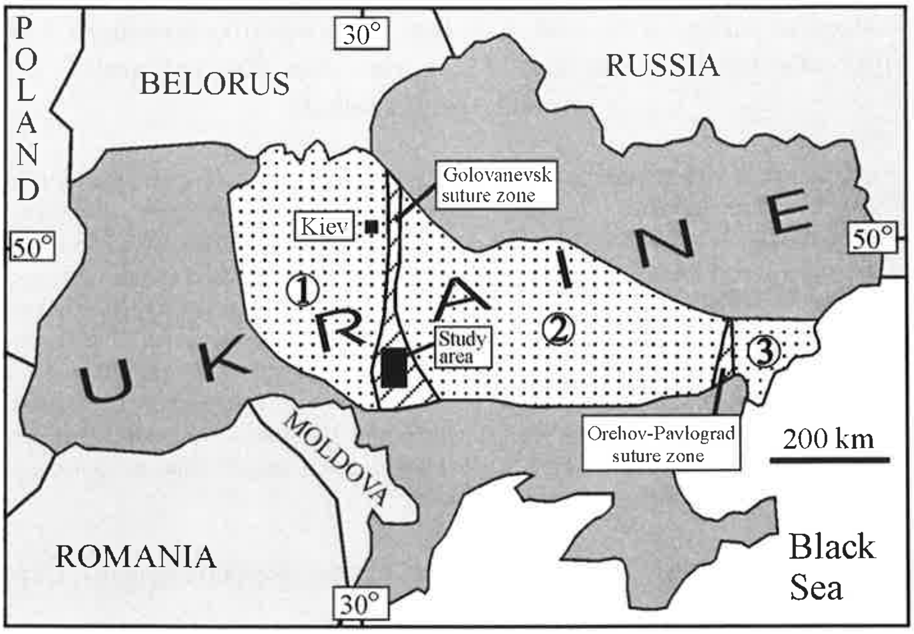

Figure 1. Location of the study area in the Golovanevsk suture zone, Ukrainian Shield. Geoblocks of the Shield (stippled): 1- Volyno-Podolsky (Western), 2 - Central, 3 - Priazovsky (Eastern) (simplified after *Dagelaysky, 1997*).

Harzburgites and, rarely, pyroxenites host massive and disseminated chromite ores, which extend up to 2-16 m in thickness and 40-250 m in length and chromian spinel enriched in Al (27.07-36.18 wt.% Al_2O_3). The chromian spinel grains in the ores range from 0.002 to 1 mm in size. Other minerals include ilmenite, ferritchromite (rims around chromian spinel), olivine (Fo_{80-90}), magnetite, serpentine, orthopyroxene and clinopyroxene. The ultramafic host rocks are often highly serpentinized. Some samples of the host rocks and ores contain abundant dolomite and calcite (Table 1), which are thought to represent metasomatic minerals (*Ershov, 1960*).

Opaque minerals in the serpentine matrix and in chromian spinel are represented by nickeline, millerite, gersdorffite, maucherite, As-bearing tucekite, galena and chalcopyrite. Among the PGM found in the ores are irarsite, laurite and ruarsite with minor amounts of anduoite, sperrylite, stibiopalladinite and unidentified PGE phases. These minerals occur as irregular grains of 1-5 μm in size attached to chromian spinel edges, within interstitial silicates or as euhedral, 1-9 μm crystals included in chromian spinel. The composition of PGM species (e.g., irarsite) varies between grains found as inclusions in chromian spinel and those enclosed in silicates. The PGM formation processes probably took place during two stages: i) high temperature exsolution from chromian spinel at subsolidus stage or precipitation of magmatic minerals on growth surfaces of chromian spinel (e.g., *Garuti et al., 1999* and references therein) and ii) subsequent low-temperature PGM formation probably during serpentinization.

Table 1. Representative microprobe analyses* of minerals from peridotites (1-9) and chromitites (10) of the Kapitanov massif. Also shown are chromite analyses from the Outokumpu (11, *Vuollo et al., 1995*) and New Caledonia (12, *Moutte, 1982*) chromitites for comparison.

	1 Spinel	2 Magnetite	3 Chromite	4 Olivine	5 Cpx	6 Calcite	7 Calcite	8 Calcite	9 Dolomite	10 Chromite	11 Chromite	12 Chromite
SiO2	0,00	0,00	0,00	38,73	52,64	–	–	–	–	0,00	–	0,16
TiO2	0,00	0,61	0,10	0,00	0,20	–	–	–	–	0,15	0,03	0,09
Al2O3	48,77	2,21	38,41	0,00	2,21	–	–	–	–	36,11	22,07	32,91
Cr2O3	11,12	8,94	22,27	0,00	0,22	–	–	–	–	26,28	48,27	34,13
V2O3	0,01	0,05	0,10	0,00	0,01	–	–	–	–	0,45	–	0,12
FeO	18,86	29,33	23,67	19,22	2,98	0,71	0,54	0,50	2,29	22,04	15,00	11,15
Fe2O3**		7,12	56,67	5,56	–	–	–	–	–	–	3,30	0,32 2,92
MnO	0,93	1,00	0,47	0,53	0,15	0,32	0,25	0,45	0,77	0,22	0,26	0,23
MgO	12,44	0,87	8,84	40,63	16,32	2,35	2,67	3,13	19,23	9,59	12,97	16,64
CaO	0,00	0,00	0,00	0,00	24,25	54,59	52,64	53,49	30,38	0,00	–	–
CoO	0,11	0,02	0,06	0,02	0,01	–	–	–	–	0,07	–	–
NiO	0,48	0,48	0,16	0,34	0,04	0,02	0,03	0,00	0,05	0,27	–	0,16
ZnO	0,32	0,00	0,18	0,00	0,00	–	–	–	–	0,17	0,06	–
BaO	–	–	–	–	–	0,00	0,00	0,04	0,00	–	–	–
SrO	–	–	–	–	–	0,00	0,00	0,00	0,05	–	–	–
SO3	–	–	–	–	–	0,00	0,06	0,10	0,06	–	–	–
CO2**	–	–	–	–	–	42,01	43,81	42,29	47,17	–	–	–
Total	100,16	100,18	99,82	99,47	99,03	100,00	100,00	100,00	100,00	98,65	98,98	98,51

"Samples: 1, 2, 8, 9 = 3541/358-1; 3-7 = 3541/405-1; 10 = 3536/195.0"
* analyses were made using a JEOL JCXA-733 microprobe at the University of Oulu.
** calculated based on stoichiometry

3. The Lipovenki Massif

The Lipovenki massif comprises three separate, closely associated ultramafic bodies occurring as subvertical and relatively small fault-bounded slices or lenses. They are known as the Western, Shkolnoe and Eastern bodies. The ultramafic bodies are mainly composed of serpentinized dunite-harzburgite, and are covered by lateritic nickeliferous rocks. The Western ultramafic slice (1.7 km long and c. 250 m wide) hosts three concordant chromitite ore bodies up to 130 m long and 20-25 m wide. The main ore body is composed of massive chromitite and nodular and irregular schlieren-type ores.

The massive chromitites consist of tabular and irregular chromite grains (0.1-3 mm) and, rarely, euhedral and rounded chromite crystals separated by serpentine. Only few grains of sulfides have been found in the ores. The massive ores contain straight and mosaic cracks similar to those found in ophiolitic complexes. The nodular and schlieren chromitites consist of euhedral to subhedral chromite grains, 0.1-1 mm in size, located in a serpentine matrix. The chromite grains in these areas are locally altered along margins and cracks, which are also filled with serpentine. Some of the grains are rimmed by magnetite. Chromite grains in nodular and schlieren ores frequently contain diverse inclusions of silicates, sulfides and sulfarsenides in contrast to sulfide-poor massive ores. All the ore types are weathered in the upper part of the main ore body close to the lateritic cover rocks.

The chromites of massive, nodular and schlieren ores from Lipovenki have compositions similar to those reported from ophiolitic complexes. The Cr/(Cr+Al) values fall in the range of 0.58-0.65, the NiO (0.08-0.33 wt.%) and TiO_2 (0.09-0.22 wt.%) contents are low and comparable with those measured for chromites from podiform chromitite deposits (*Roberts and Neary, 1993; Stowe, 1994*). The $Fe^{3+}/(Fe^{3+}+Al+Cr)$ and $Mg/(Mg+Fe^{2+})$ values for the Lipovenki chromites vary in the ranges of 0.12-0.69 and 0.39-0.57, respectively. The PGM observed within the Lipovenki massif include laurite, irarsite and unidentified Rh-bearing phases. All the PGM phases have been identified in nodular and schlieren ores and are found as inclusions both in silicates and chromite.

4. Chondrite-normalized PGE Patterns

We analyzed four whole-rock chromitite samples for PGE using ICP-MS. Two of them (3541/392, 3541/401) are massive ores from the Kapitanov deposit and the rest two represent the nodular (LO1/1) and schlieren (L-SH/02) ore types from the Lipovenki deposit. The results are listed in Table 2 and shown as chondrite-normalized plots in Figure 2 together with fields for ophiolitic chromitites after *Talkington and Watkinson (1986)* and Vasarakangas chromitites from the Outokumpu region after *Vuollo et al. (1995)*. Both Ukrainian deposits possess PGE characteristics similar to those found in chromitites from ophiolitic mantle tectonites, including high Ir-group PGE/Pd-group PGE, but they differ from each other to some extent. The Lipovenki samples show smooth, PPGE-depleted curves, whereas the Kapitanov samples display positive Ru and negative Pt anomalies due to relatively high Ru and Pd concentrations. Although the former patterns are common among the ophiolitic chromitites, the latter zigzag patterns are not unknown, as these are observed, for example, in the Kempirsai massif, Urals and the Bragança massif, Portugal (*Bridges et al., 1993; Melcher et al., 1999*).

Table 2. PGE and Au analyses (ppb) for chromitites from the Kapitanov and Lipovenki deposits compared with data from Vasarakangas chromitites, Outokumpu, after *Vuollo et al. (1995)*.

Sample	Deposit	Os	Ir	Ru	Rh	Pt	Pd	Au
3541/401	Kapitanov	36,0	28,1	174	13,1	7,12	14,5	0,99
3541/392	Kapitanov	30,8	18,7	310	12,7	11,6	19,0	1,14
LO1/1	Lipovenki	70,0	89,9	130	11,3	4,30	<1.0	<0.5
L-SH/02	Lipovenki	21,1	18,1	27,2	2,94	1,80	<1.0	<0.5
POL-4	Vasarakangas	34	20	98	8	6	2	5
POL-6	Vasarakangas	56	34	114	11	6	2	10

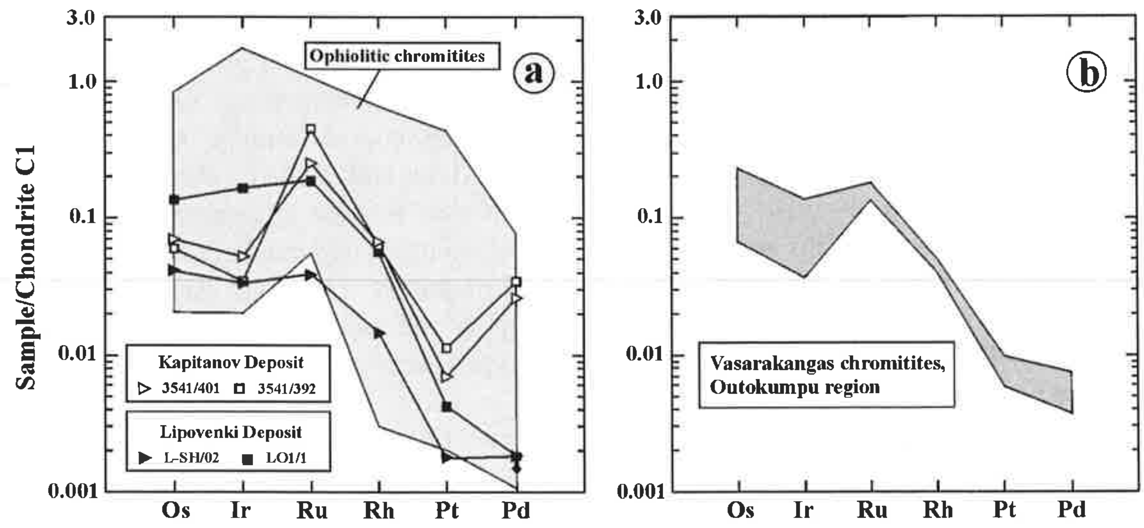

Figure 2. a) Chondrite-normalized PGE patterns for chromitites from the Kapitanov and Lipovenki deposits compared with a reference field for ophiolitic chromitites after *Talkington and Watkinson (1986)*. **b)** PGE characteristics of Vasarakangas chromitites from the Outokumpu region, Finland, after *Vuollo et al. (1995)*.

5. Discussion and Conclusions

The Kapitanov and Lipovenki Paleoproterozoic ultramafic massifs are fault-bounded slices or lens-shape bodies hosting chromitites with PGE and PGM characteristics typical of ophiolitic rocks. At Kapitanov, Al-rich massive and disseminated chromite ores are hosted by a layered dunite-peridotite sequence. The Lipovenki massif is composed of mantle dunite-harzburgite and hosts nodular, schlieren and sulfur-poor massive chromitites with textures and chromite compositions similar to those reported from ophiolitic complexes. All these observations allow us to conclude that the studied ultramafic massifs are remnants of Paleoproterozoic oceanic lithosphere and the Golovanevsk zone between the Volyno-Podolsky (Western) and Central Ukrainian geoblocks represent a major suture zone.

Paleomagnetic studies suggest that the Ukrainian and Fennoscandian Shields have drifted independently until the Paleoproterozoic and that they were part of a large continent during the Paleoproterozoic and Mesoproterozoic (*Elming et al., 1993*). This large shield was split up between ca. 1.35 and 1.2 Ga and the formation of East European Platform took place in the late Precambrian (1.07-0.57 Ga). It should be noted that the magnetization ages on the Ukrainian Shield are often questionable since the whole-rock K-Ar dating is the principal radiometric method that has been employed (*Mikhailova and Kravchenko, 1987; c.f., Elming et al., 1993*). Nevertheless, the available paleomagnetic and geochronogical data are compatible with the suggestion that the emplacements of the Paleoproterozoic Ukrainian and Fennoscandian ophiolitic rocks were related to the same collisional event. More reliable age determinations of the ophiolitic rocks would be crucial for supporting this hypothesis. On the other hand, the recognition of suture zones within the Ukrainian Shield is significant for the interpretation of paleomagnetic data.

Acknowledgements

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Ultramafic Rocks in the Nunnanlahti Greenstone Belt, Eastern Finland: A Potential Example of Archaean Ophiolitic Rocks

Eero Hanski[1], Jouni Vuollo[1], Jani Wennerstrand[2] and Stanislav Gornostayev[3]

[1]Geological Survey of Finland, P.O. Box 77, FIN-96101 Rovaniemi, Finland
[2]Department of Geology, P.O. Box 11, FIN-00014 University of Helsinki, Finland
[3]Department of Geosciences, P.O. Box 3000, FIN-90014 University of Oulu, Finland
E-mail: eero.hanski@gsf.fi

The Nunnanlahti greenstone belt in eastern Finland contains large lenses of serpentinites and soapstones which are suggested to be potential examples of Archaean ophiolitic rocks. This view is supported by several geological, geochemical and mineralogical observations. These include the tectonic contacts of the lenses with the surrounding rocks, the highly refractory major element geochemistry of the ultramafic rocks, the high forsterite content in olivine relicts, and the presence of chromitite pods and bands as well as abundant mafic to ultramafic dyke rocks within serpentinites and soapstones.

Keywords: ophiolite, greenstone belt, Archaean, Nunnanlahti, Finland

1. Introduction

Discoveries of Palaeoproterozoic ophiolitic rocks in Finland, Canada and the U.S.A. have resulted in general acceptance of the operation of plate tectonic processes at least c. 2.0 Ga ago. In contrast, Archaean plate tectonics have remained as a controversial issue because of the lack of universally agreed examples of ophiolitic complexes of that age. Our report deals with an occurrence of ultramafic bodies in the Archaean Nunnanlahti greenstone belt in eastern Finland. These occurrences are well-known due to the presence of exploitable soapstone deposits (*Rossi, 1997*) and possess features, which make them potential candidates for Archaean ophiolitic rocks. The Nunnanlahti greenstone belt (2x15 km) is situated in eastern Finland (Fig. 1) and is composed mainly of felsic and mafic metavolcanites (tholeiites and Fe-tholeiites) with the latter exhibiting pillow structures in places *(Kohonen et al., 1989)*.

2. Ultramafic Rocks, Dykes and Chromitites

All the ultramafic bodies are situated on the hanging-wall side of the major Nunnanlahti Shear Zone. The ultramafic rocks, which are now represented by serpentinites and soapstones, occur as lens-like, massive bodies (up to 700mx1400 m in size) having tectonic contacts with their metavolcanic country rocks. In soapstone quarries, chlorite-rich lenses and bands are observed which probably were originally mafic to ultramafic dyke rocks. These rocks have undergone pervasive metamorphic and metasomatic alteration together with their host rocks and possess LREE-depleted chondrite-normalized REE patterns as opposed to LREE-enriched patterns obtained for Fe-tholeiitic Proterozoic dykes observed in the region. The chemical composition of the serpentinites is highly refractory with the Al_2O_3 and TiO_2 contents ranging up to 0.9 wt.% and 0.06 wt.%, respectively, which indicates that the rocks were originally mainly dunites and depleted harzburgites. On a Cr/Al vs. Ni/Al diagram (Fig. 2), the Nunnanlahti serpentinites plot along the trend of ophiolitic ultramafic rocks and mostly within or close to the field for the Nuttio serpentinites from Central Lapland (*Hanski, 1997*). They differ clearly from Archaean komatiitic rocks on this diagram.

Figure 1. Location of Nunnanlahti serpentinites and Palaeoproterozoic ophiolitic ultramafic rocks (Outokumpu, Jormua, Nuttio) in Finland.

Relict olivines in serpentinites have a highly magnesian composition with Fo up to 97 %, one of the features shared with ophiolitic serpentinites in Central Finnish Lapland (*Hanski, 1997*). These olivines also possess extremely low contents of CaO and Cr_2O_3 varying in the ranges of 0.00-0.01 wt % and 0.00-0.03 wt%, respectively. These values contrast with the reported CaO and Cr_2O_3 contents for high-Fo (>90%) olivines from komatiites, which normally fall in the ranges of 0.05-0.35 wt % and 0.10-0.30 wt %, respectively. On the other hand, MnO contents (0.08-0.15 wt %) are low compared with those of olivines formed during regional metamorphism in ultramafic rocks (e.g., *Blais and Auvray, 1987*).

The Nunnanlahti serpentinites and soapstones are uniformly rich in accessory chromite resulting in high Cr_2O_3 contents commonly falling between 0.5 to 1.0 wt%. In several places, we have observed lens-like chromite segregations and their more altered magnetite-rich derivatives reaching 20-30 cm in size (Fig. 3) as well as a few cm thick, locally brecciated or folded metachromitite bands or layers. The chromite segregations resemble small podiform chromitite lenses observed in younger ophiolitic ultramafic rocks. The abundance of chromite does not fit well with a komatiitic parental magma as this mineral appears relatively late as a liquidus phase in komatiitic liquids (*Murck and Campbell, 1986*). We have started to search for mineral inclusions in chromitite and have so far identified nickel and iron sulfides, galena and sulfarsenides and one grain of tin.

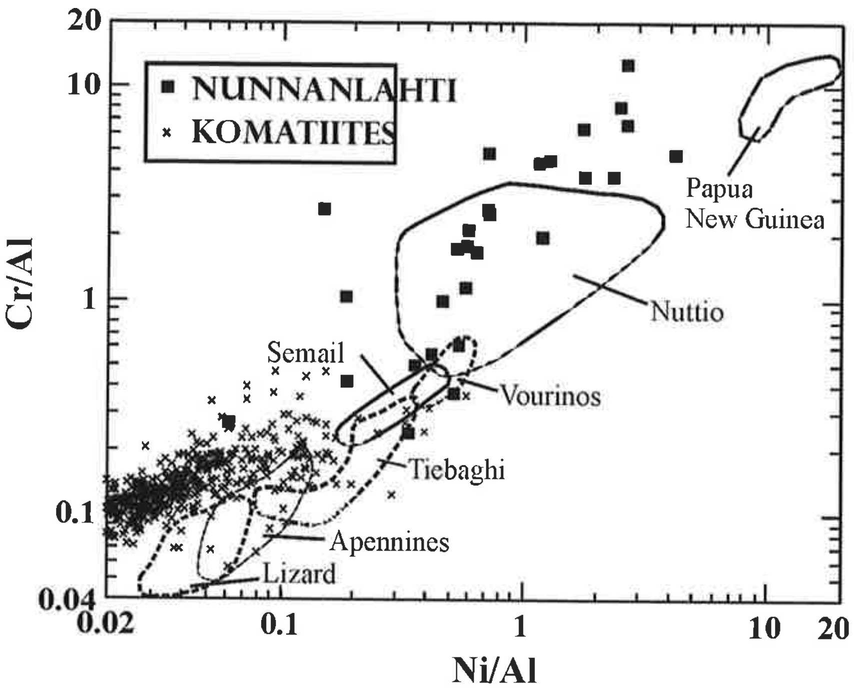

Figure 2. A Cr/Al vs. Ni/Al plot for ophiolitic mantle rocks, komatiitic rocks and Nunnanlahti serpentinites and soapstones. The reference data for the ophiolitic rocks is taken from *Roberts and Neary (1993)*, *Jaques and Chappel (1980)*, *Rampone et al. (1995)* and *Hanski (1997)* and for the komatiitic rocks from various literature sources.

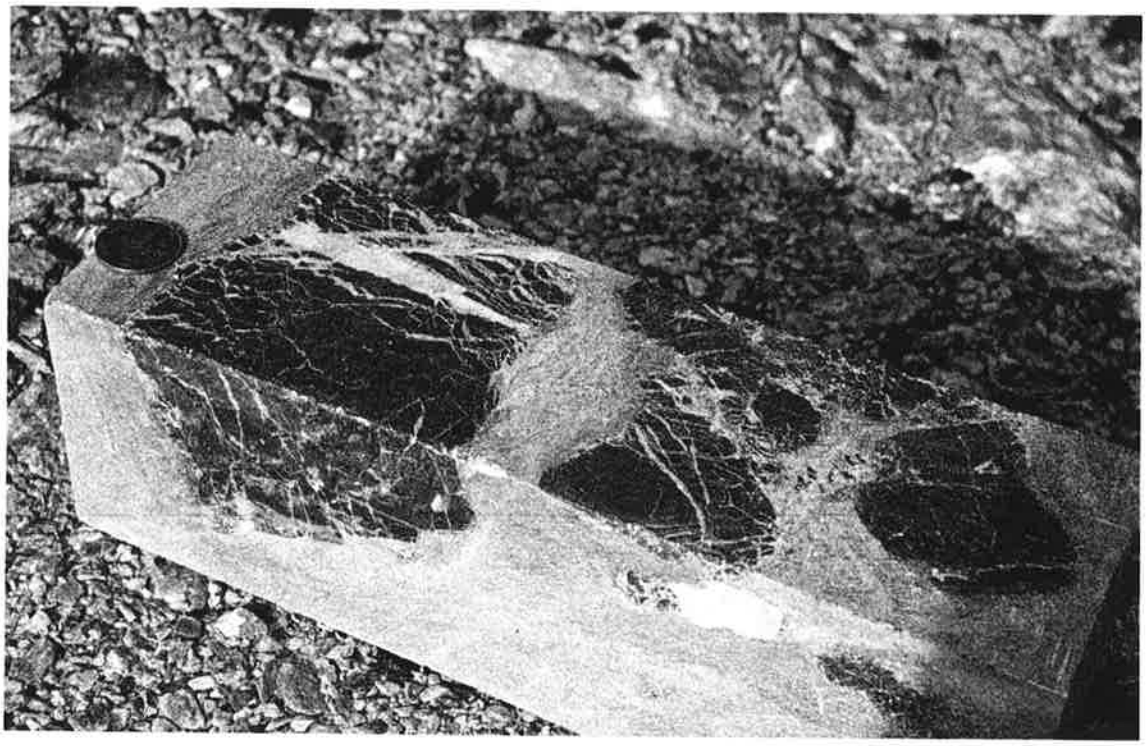

Figure 3. Metachromitite lenses (dark) with carbonate veinlets (light) in a soapstone block from Nunnanlahti. The diameter (2.5 cm) of the coin servs as a scale on the left edge of the block.

3. Conclusions and Discussion

All the above listed characteristics are compatible with the suggestion that the Nunnanlahti serpentinites and soapstones were originally plutonic rocks and represent fragments of ancient lower oceanic lithosphere rather than olivine cumulates of komatiitic lavas. One option is that they are tectonic slices belonging to the Palaeoproterozoic Outokumpu Ophiolite Complex (see Fig. 1), but at present we prefer their Archaean origin. So far, the supposed Archaean age for these rocks is based on the observation that they are cut by mafic dykes correlated with the widely occurring Palaeoproterozoic dyke swarms in the Archaean basement (*Sorjonen-Ward, 1997*). It would be important to get the age problem of the Nunnanlahti ultramafic rocks resolved, potentially using the Os or Pb isotopic method.

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Tectonic Evolution of the Central Lapland Greenstone Belt – Metamorphic and Structural Observations

Pentti Hölttä and Markku Väisänen

Geological Survey of Finland, P.O. Box 96, FIN-02151 Espoo, Finland
E-mail:pentti.holtta@gsf.fi

The Palaeoproterozoic Central Lapland Greenstone Belt was formed by the volcanic activity and sedimentation related to the prolonged rifting between ca. 2.5-1.9 Ga. After that the belt was affected by the compressional Svecofennian tectonism. The early deformation created the main foliation in the area associated with recumbent folds during a northward transport episode. The late deformation produced the overprinting folds of differing orientations, possibly caused by the continued Svecofennian movement and the simultaneous opposing movement of the Lapland granulite belt. Peak metamorphism, as indicated by the timing of porphyroblast growth, was reached during the late stage of deformation. The changes in metamorphic grade are mostly connected with the shear zones, except for in the vicinity of granitoids that provided the heat for the highest grade assemblages.

Keywords: Palaeoproterozoic, Fold and Thrust Belt, Shear Zones, Oceanic Crust, Greenschist Facies

1. Introduction

The early evolution of the Central Lapland Greenstone Belt (CLGB) can be described in the simplest form as a prolonged rifting starting at around 2.5 Ga and continuing until the Svecofennian orogeny at ca. 1.9 Ga. After that, compressional tectonics have prevailed. Early rift-related volcanism is contaminated by the underlying older crust, but the latest phases of the mantle derived volcanism, the Kittilä formation, bear no evidence of crustal contamination and is regarded as an allochthonous oceanic unit separated by thrusts *(Hanski, 1997)*. Younger quartzites and conglomerates unconformally cap all the previous rocks.

2. Structures

Previous structural investigations have suggested either a complex polyphase folding from various directions *(Lehtonen et al., 1998)* or a progressive development of a fold and thrust belt with opposing transport directions from both south and north simultaneously with a dextral rotation of structures *(Ward et al., 1989)*. In the present stage of our investigations, we simply refer to the structural evolution as the early and late stages of deformation.

The early deformation (D1) produced tight to isoclinal recumbent folds with a penetrative axial plane foliation, S1, the main foliation in the area. Microscopically, however, it is a composite foliation and an earlier bedding-parallel foliation is crenulated in fold hinges. A NNE-SSW mineral lineation on the foliation plane possibly correlates with the tectonic translation direction. Kinematic indicators are few, but some asymmetric recumbent folds, sigmoidal inclusion trails in garnets, asymmetric stair-stepping structures, S-C structures and shear bands suggest a south to north transport direction. Subhorizontal D1 structures are deformed by later sets of folds (D2) that in the southern part of the study area are E-W trending, overturned N-vergent folds, whereas in the northern part, folds with a similar style are mostly upright and N-S trending. In the NE part of the area, W-vergent folds were observed. A new S2 foliation was formed only in tight folds, but most porphyroblasts (e.g. andalusite, kyanite, staurolite) grew during D2 as a folded S1 is preserved as inclusion trails in porphyroblasts. The lack of unambiguous overprinting relationships makes it difficult to know if late folds of different orientations are of the same or of different generations.

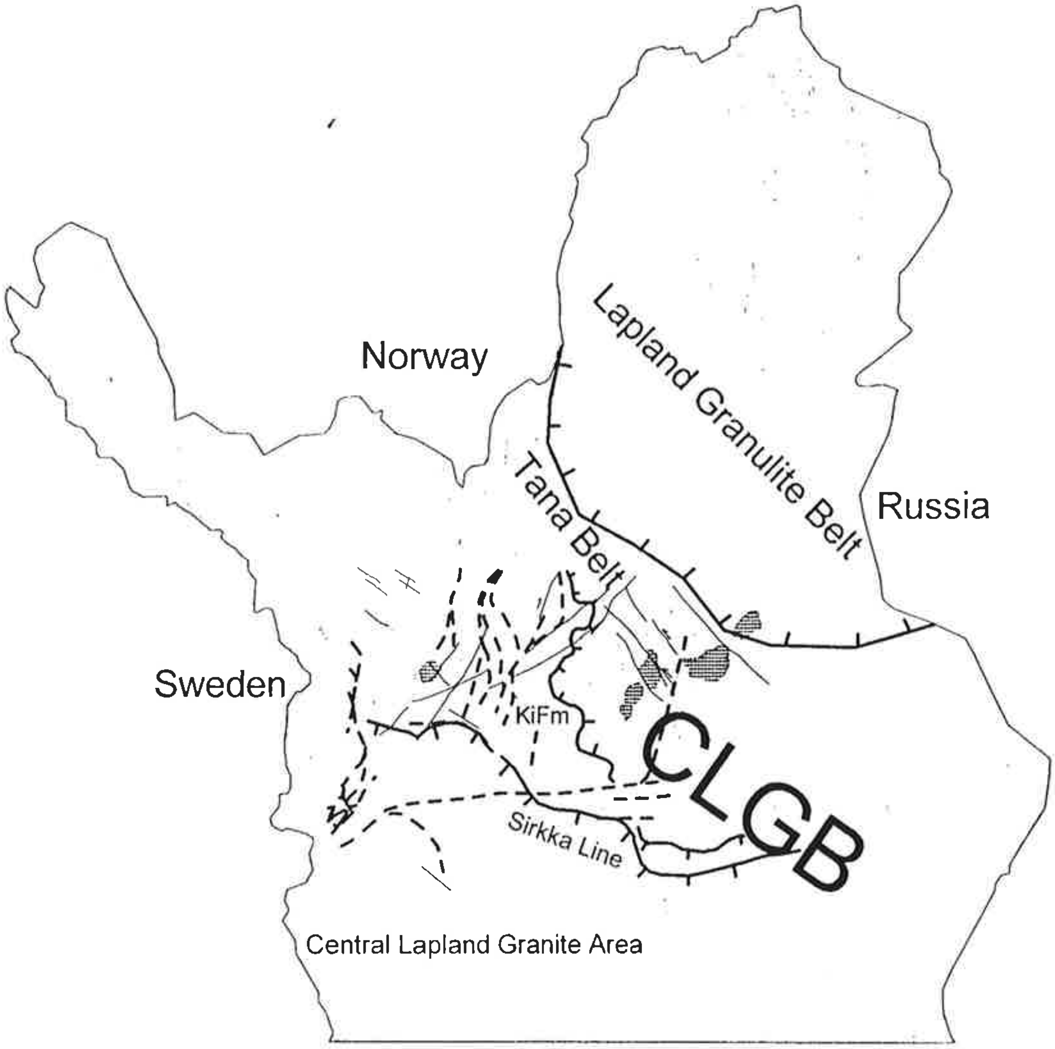

Figure 1. Major tectonic features of Lapland. CLGB - Central Lapland Greenstone Belt, KiFm - Kittilä Formation, solid thick lines - thrusts, solid dashed lines - shear zones, solid thin lines - late faults, grey patche - 1.77 Ga granites.

There are faults of many generations and orientations in the CLGB. The most prominent of them is the Sirkka Line, a WNW-ESE trending major crustal lineament that separates the allochthonous oceanic unit in the N from older formations. *Berthelsen and Marker (1986)* and *Gaál et al. (1989)* interpreted the Sirkka Line as a southward dipping thrust zone. Our field observations support this interpretation as two generations of overturned or recumbent folds were identified in the vicinity of the zone. Gently dipping high strain zones were also found in the west close to the major N-S steep fault zones and in the NE along the eastern border of the oceanic formation, consistent with the idea of fold and thrust belt *(Ward et al., 1989)*. The steep shear zones have two main orientations, E-W and N-S. In the west along the Swedish border, a steep zone of several N-S trending faults bounding the gently dipping domains show both northward and eastward movement directions suggesting a transpressional tectonism. Further east, the N-S oriented steep shear zones affect mafic-ultramafic rocks, but so far no kinematic indicators were found. E-W trending steep zones apparently seem to be associated with the Sirkka Line. Occasionally, dextral shear senses could be identified. The CLGB was still later affected by conjugate sets of faults that even cut the 1.77 Ga granites (Fig. 1).

3. Metamorphism

The area between the Lapland granulite belt and the Central Lapland granite area can be divided into several metamorphic zones which seem to differ in P-T development. The metamorphic grade varies from greenschist facies in the Kittilä greenstone belt to transitional granulites in the Tana belt. The changes in metamorphic grade are mostly connected with the shear zones, which separate allochthonous blocks. There is an abrupt increase in metamorphic grade across an E-W trending shear zone when coming from the Kittilä greenstone belt to the Central Lapland granite area, where metapelites are medium-grained, locally garnet-bearing gneisses which are strongly retrogressed close to the shear zone. In few exposures these contain all three aluminium silicates: andalusite, sillimanite and kyanite, sillimanite replacing kyanite. The higher grade is possibly caused by the heat flow from the granitoids; this is seen well in the western side of the Central Lapland granite area where the metamorphic grade increases from fine grained schists to andalusite gneisses and then to sillimanite gneisses towards the granite contact.

In the poster presentation we will show the structural and metamorphic data and discuss their significance to the evolution of the Svecofennian Orogen in northern Finland.

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The Accretionary Svecofennian Orogen-
in the Light of Seismic BABEL Lines

Annakaisa Korja and Pekka Heikkinen

Institute of Seismology, P.O. Box 26, FIN-00014 University of Helsinki, Finland
E-mail: Annakaisa.Korja@seismo.helsinki.fi

The BABEL profiles B, C, 1, 4, 4A and 3 form a 1200 km long nearly continuous cross-section through the Fennoscandian Shield. The near-vertical marine reflection profiles display a wide range of crustal structures that can be associated with both the accretionary Svecofennian orogeny (1.9-1.8 Ga) and the following Subjotnian and Jotnian rift-stages (1.65-1.11 Ga). The Svecofennian accretionary orogeny took place, when a number of micro-plates with island arc affinities and surface expression of a large archipelago accreted to the continental Karelian plate. Some of the accreting blocks seem to have had older cores that have acted as crustal indentors during the collision.

Keywords: reflection seismics, tectonics, accretion, Palaeoproterozoic, Svecofennian, Fennoscandia

1. Introduction

The BABEL lines were situated in the Baltic Sea and its northern extension the Gulf of Bothnia. The deep marine seismic reflection BABEL lines B, C, 1, 4, 4A and 3 (Figure 1) that were gathered as an international co-operation in 1989, were designed to study the Proterozoic structures and the evolution of the Fennoscandian Shield. The compilation of the first results was published in 1993 *(BABEL Working Group, 1993a,b)*. As a lot of new geophysical and geological data have been published since the first interpretations and because the data has been reprocessed at the University of Helsinki, we have reinterpreted the BABEL data as a one continuos data set. The interpretation focuses on the regional scale reflection structures and on their plate tectonic implications for the evolution of the Svecofennian orogeny.

2. BABEL Line B

In the southernmost part of BABEL profile B (Fig. 2a; > 400 km), the crust is relatively thin (40 km) but it grows stepwise in thickness along the profile and reaches its maximum thickness of 58 km in the middle part of the profile (175 km). Where the crust is thickest the reflectivity is obscure and the Moho boundary is not very distinctive. The crustal structure of the southern part (400-250 km) is dominated by apparently NE-dipping reflectors that begin in the upper crust and flatten out at the lower crust or at the crust-mantle boundary. Offshore Småland (410-340 km), the crustal reflectors are cut by an unreflective, mushroom-shaped body in the lower and middle-crust. At distances between 350 km and 250 km, the upper crust above the dominantly NE-dipping reflectors has both NE- and SW-dipping reflectors that resemble open folds.

Based on correlation of reflectivity patterns of Åland (profiles C, 1, 7) and Bothnian (profiles 1 and 6) rapakivi granite batholiths, the Landsortens Deep rapakivi granite is identified in the middle part of the profile (190-140 km), where Jotnian sandstone and Subjotnian Baltic Sea porphyries have been found at the sea bottom *(Ahlberg, 1986)* and where *Puura and Flodén (1999)* have suggested a rapakivi granite batholith based on Bouguer anomaly correlation.

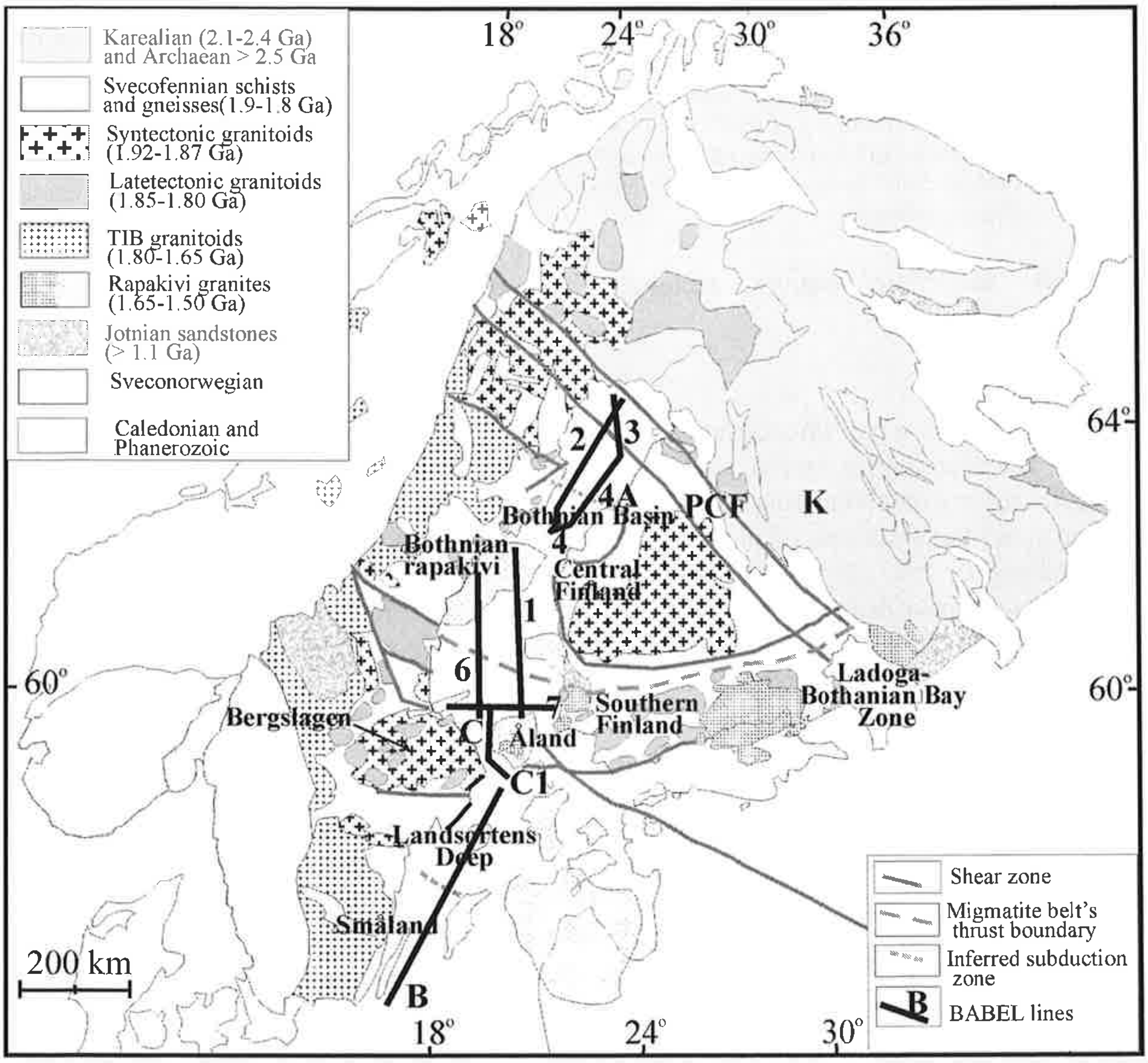


Figure 1. Geological map of the Fennoscandian Shield with major tectonic boundaries and BABEL lines B, C, 1, 2, 3, 4, 6, and 7.

3. BABEL Line C

Although line C has a turning point (Fig. 1), it is shown here as a single profile (Fig. 2b). The turning point at 82 km is marked on the line drawn profile with a straight line.

Along profile C, the Moho is approximately at 50 km of depth. The northern part of the profile (0-82 km) shows prominent, apparently S-dipping reflectors that crosscut weaker reflectivity patterns. The SE-NW-striking part of the profile shows flat or moderately dipping reflectors in the lower crust, unreflective middle crust and highly reflective upper crust that hosts mainly SE-dipping listric reflectors that flatten out at the depth of 22 km. The uppermost central part of the profile hosts mainly NW to N dipping and minor S-dipping reflectors that start from within stratified reflectivity associated with Jotnian Åland sedimentary basin. The basin structure is crosscut by an unreflective near vertical-structure where the Märket Postjotnian diabase dyke is outcropping.

4. BABEL Line 1

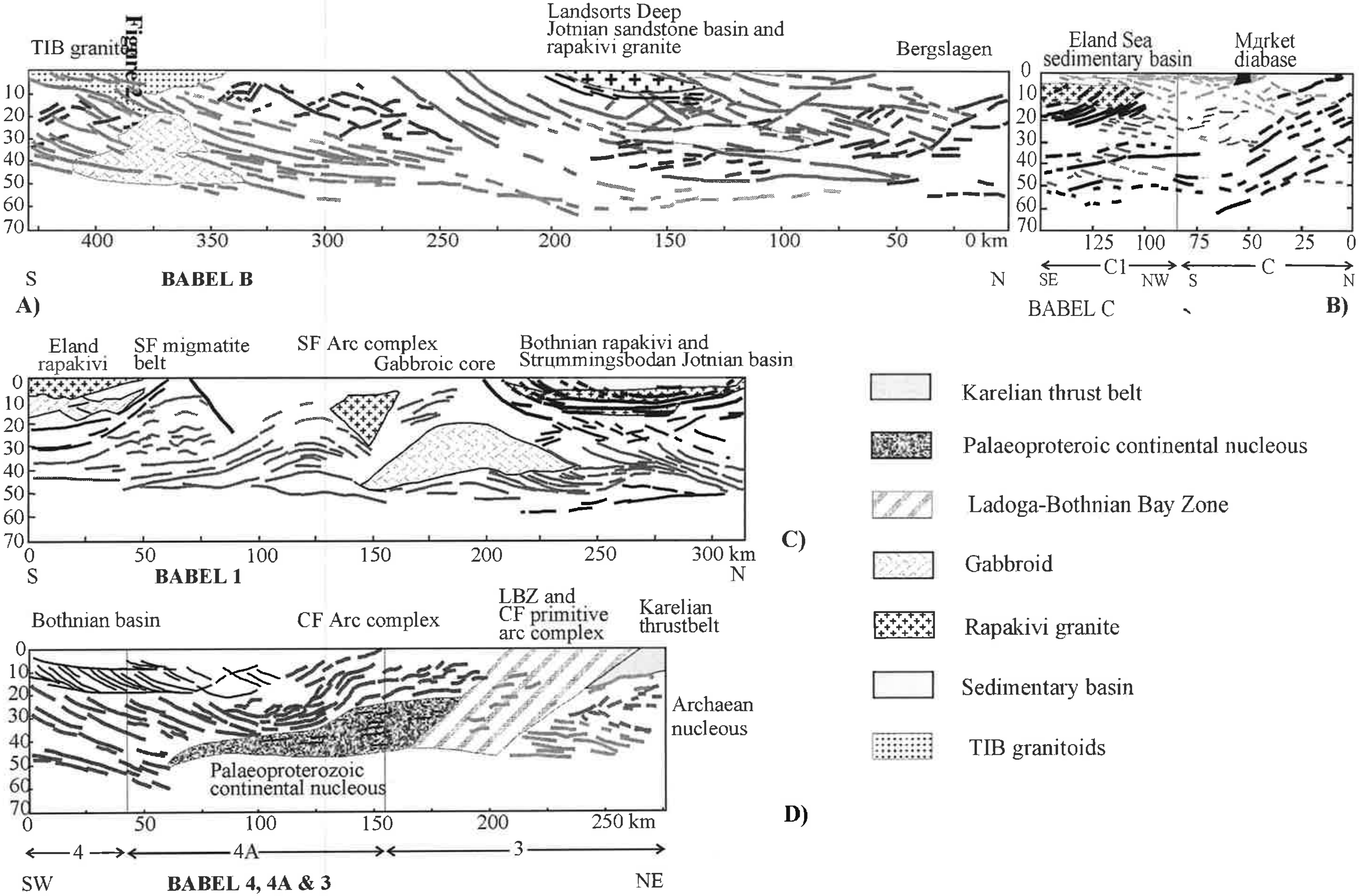
The most prominent feature on BABEL profile 1 (Fig. 2c) is the highly reflective lower crust. The crustal thickness increases stepwise from the south (42 km) towards the north and is at the largest (56 km) in the middle of the profile (180-220 km) where the base of the reflectivity is rather obscure but traceable. The thickness of the crust decreases again towards the northern end of the profile, where it reaches 52 km *(Heikkinen and Luosto, 1992)*. The profile hosts large regional differences in both the intensity and style of reflectivity.

In the southernmost part (0-60 km) of the BABEL profile 1, the upper crust is relatively unreflective, but hosts isolated reflectors mimicking graben and horst -structure. This has been explained to arise from the contact between the lower velocity granite (6.15 km/s) and the higher velocity gabbro-anorthosite (6.50 km/s) *(Malaska and Heikkinen, 1997; Korja et al., 2000)*. The unreflective upper crustal area is delineated from the neighbouring highly reflective area to the north by SE-dipping reflectors that tend to flatten in the middle to lower crust *(Korja and Heikkinen, 1995)*.

Small-scale reflectors dipping both to the south and to the north are characteristic for the highly reflective middle crust in the southern part (30-85 km). The reflection pattern mimics small-scale open folding. The block is terminated in the north by a single 60-degrees dipping reflector reaching the surface at the distance of 67 km. The lower crust of the area is characterized by dipping reflectors that tend to flatten out in the lower crust or to offset the Moho boundary (30 km, 75 km). These reflectors are continuos and as they rise to the adjacent middle to lower crust they resemble large-scale open-folds (90-150 km). North of the fold pattern is an unreflective body that based on gravity data has been interpreted to be composed of gabbroic intrusions *(Pedersen et al, 1992; Korja et al., 2000)*.

The bright saucer-shaped reflectors interpreted as Postjotnian diabase sills *(BABEL Working Group, 1993)* dominate the otherwise unreflective upper crust of the northern part of the profile (220-320 km). The middle crust is less reflective but it hosts distinctive dipping events that tend to join the concave band of lower crustal reflectors between 35 km and 40 km of depth and distances 240 to 300 km. The reflective band is underlain by a 5 km thick, unreflective layer, whose bottom is marked by a clear reflective event (265- 275 km) at the crust-mantle boundary at the depth of 52 km *(Heikkinen and Luosto, 1992)*.

Figure 2.


TIB granite
Landsorts Deep
Jotnian sandstone basin and
rapakivi granite
Bergslagen
S
BABEL B
N
A)
Eland Sea
sedimentary basin
Mдrket
diabase
SE
C1
NW
S
C
N
BABEL C
B)
Eland
rapakivi
SF migmatite
belt
SF Arc complex
Gabbroic core
Bothnian rapakivi and
Strummingsbodan Jotnian basin
S
BABEL 1
N
300 km
C)
Bothnian basin
CF Arc complex
LBZ and
CF primitive
arc complex
Karelian
thrustbelt
Archaean
nucleous
Palaeoproterozoic
continental nucleous
250 km
4
4A
3
SW
BABEL 4, 4A & 3
NE
D)
Karelian thrust belt
Palaeoproteroic continental nucleous
Ladoga-Bothnian Bay Zone
Gabbroid
Rapakivi granite
Sedimentary basin
TIB granitoids

5. BABEL Lines 4, 4A and 3

Lines 4, 4A and 3 are shown as one profile but the turning points at 41 km and at 156 km are marked with vertical lines on the line-drawing (Figure 2d). Line 2, where crustal geometry is rather similar, is not displayed in this paper.

Profile 4 and southern parts of profile 4A, host a set of steeply dipping reflectors that are cut by E-dipping low-angle reflectors at the depths of 5 km and 15 km. The middle crust is weakly reflective whereas the lower crust hosts prominent, steep NE-dipping reflectors that form the crust-mantle boundary. They also continue into the mantle at distance 35 km to 60 km. The dip of the reflectors has been determined from three-dimensional block diagrams.

In the north-eastern part of profile 4 (100-130 km) the middle crustal reflectors change dip to mainly SW. Further NE (130-200 km), at the end of profile 4 and at the beginning of profile 3, the reflectivity pattern changes to a complex of SW and NE dipping reflectors that together mimic open -fold structures. The lower crust underneath (80-170 km) is void of prominent reflectors and shows no distinct direction of reflectivity. However, the Moho-boundary is sharp and slightly undulating between 42 and 45 km in depth.

On profile 3, Between 200 km and 250 km in distance, a 30 to 50 km wide, subvertical unreflective block cuts through the crust. The Moho boundary is subhorizontal but less clear than in its surroundings. Within this block minor SW-dipping-reflectors exist. In the northernmost part of profile 3, the reflectors dip to the SW in the upper crust and to the NE in the lower crust. The middle crust is void of prominent reflectors and shows no distinct direction of reflectivity.

6. A Preliminary Tectonic Interpretation

The BABEL profiles B, C, 1, 4, 4A and 3 form a 1200 km long nearly continuous cross-section through the Fennoscandian Shield. In the following we present a preliminary tectonic model for the formation of the accretionary Svecofennian orogen based on the BABEL profiles. The interpretation will start from the oldest Karelian part in the north (line 4) and proceed through the Svecofennian (lines 3,1,C, B) will finish at the youngest TIB granites at the Svecofennian Gothian boundary in the south (line B). The offshore reflection structures have been correlated with the closest onshore geological and geophysical data, i.e. the northern profiles 1,3 and 4 have been correlated with Finnish bedrock units and the profile B with Swedish units.

BABEL profiles 4, 4A and 3 (Fig. 2d) images a series of collisional terrain boundaries between the Southern-Finland Arc complex, Central-Finland Arc-complex, Central Finland primitive arc complex and the Karelian continent. The Bothnian Basin area is underlain by a set of upper crustal E-dipping low-angle reflectors which in turn are underlain by prominent, NE-dipping lower to middle crustal, which have been associated with a subduction zone and accompanying accretion prism (BABEL Working group, 1993). Together the NE dipping crustal units of the BABEL profile 1, 2 and 4, are interpreted to image an accretionary prism. To the north of the proposed subduction zone (dipping mantle reflector, *BABEL Working Group, 1993*) the reflectors dip SW and display open fold structures. The reflective package has been thrust over a less reflective lower crustal indenture

Figure 2. Schematic geological interpretation of the BABEL profiles B, C, 1, 4, 4A, and 3

a) BABEL B images the architecture of the Central and Southern Swedish Svecofennides.

b) BABEL C images mostly extensional features related to the Subjotnian and Jotnian rifting event.

b) BABEL 1 images the internal architecture of the Southern Finland Arc Complex.

c) BABEL 4, 4A & 3 images a series of collisional terrain boundaries between the Southern-Finland Arc complex, Central-Finland Arc-complex, Central Finland primitive arc complex and the Karelian continent.

(30-50 km), which may be part of an older Proterozoic nucleus that has been proposed by *Lahtinen (1994)* and *Lahtinen and Huhma (1997)* to exist beneath the Central Finland Granitoid complex. The narrow Central Finland Primitive Arc Complex at the craton margin has weak SW-dipping structures, which indicate thrusting towards the NE. These structures, however, have been overprinted by a 30 km wide, subvertical unreflective band cutting through the crust, which is interpreted as the offshore continuation of the Raahe-Ladoga transform fault zone *(Koistinen and Saltykova, 1999)*. The Karelian cratonic crust is imaged as a less reflective, wedge-shaped block that has been both over- and underthrust in the Svecofennian Orogeny.

BABEL profile 1 (Fig. 2c) images the internal architecture of the Southern Finland Arc Complex. The unusually thick crust (55-60 km) host unreflective, high density, gabbroic intrusions (200km x 50km x 20km) *(Korja et al., 2000)*, which are here interpreted as a magmatic core of an island arc. To the south a highly reflective ramp anticline structure developed against this structure. Further south, the crust was thickened via thrust-slices comprising the Southern Svecofennian migmatite, schist and volcanic belts. North of the magmatic core, the lower and middle crustal listric crustal reflectors dip NE and detach at an upward concave band of reflectors interpreted as an old Moho boundary. The reflectors are interpreted to image Svecofennian deformation zones, within the abovementioned accretionary prism, that have been inverted to listric shear zones during the Subjotnian extension. In the northern and southern ends of BABEL profile 1, Bothnian (submarine) and Åland rapakivi granite batholiths, respectively, are imaged as unreflective, upper-crustal bodies delineated by reflectors interpreted as listric and normal faults *(Korja and Heikkinen, 1995; Korja et al., 2000)*.

Profile C images mostly extensional features related to the Subjotnian and Jotnian rifting event. In the south the listric reflectors flattening out at depth of 22 km are interpreted to image the bottom of the subcrustal continuation of the Åland rapakivi granite. The apparently S-dipping reflectors in the north (0-60 km) may trace a relatively unknown Bouguer anomaly lineament that runs from northern Uppland to the Åland Island.

Profile B images the architecture of the Central and Southern Swedish Svecofennides. The southern and central parts of the profile are characterised by NE dipping crustal reflectors and Moho offsets as well as stepwise increasing thickness of the crust. As a whole the profile resembles an accretionary continental margin with retreating subduction zone to the southwest. Remnants of dipping mantle reflectors interpreted as subduction zones have been interpreted from wide-angle data by *Abramovitz and others (1997)* beneath central part of profile B and recently by Balling (2000) beneath the northern part of profile A.

In the southern part of profile B, a mushroom-shaped body intrudes from the mantle through the NE-dipping, collisional structures and is thus younger in age. The upper part of the intrusive body meets the apparently NE-dipping band of reflectors, interpreted as a shear zone. The shear zones are located offshore Småland, where TIB granites outcrop on land (Fig. 1) and where geophysical anomalies *(Wonik and Hahn, 1992)* suggest that the granites continue offshore as well. It is thus interpreted that the TIB-granite magmas were emplaced as shallow level plutons that rose along shear zones from the middle to lower-crustal mantle-originated magma chambers. As the age of the reflection pattern is not known, we cannot determine which of the three TIB events *(Larsson and Berglund, 1992)* is imaged in the seismic section.

In the northernmost parts of the profile, B (0-75 km), offshore Bergslagen, the geometries change, so that reflectivity dips southwards or south-eastwards and the geometry shows similarities with the Karelian continental boundary, where Proterozoic cover rocks have been thrust NE on the cratonic crust (BABEL Working Group, 1993b). The data seems to support *Allen's et al. (1996)* suggestion of an older core underlying the Bergslagen area.

7. Conclusions

The near-vertical marine reflection profiles display a wide range of crustal structures that can be associated with both the accretionary Svecofennian orogeny (1.9-1.8 Ga) and the following Subjotnian and Jotnian rift-stages (1.65-1.11 Ga). The Svecofennian accretionary orogeny took place, when a number of micro-plates with island arc affinities and surface expression of a large archipelago accreted to the continental Karelian plate. Some of the accreting blocks seem to have had older cores that have acted as crustal indentors during the collision. Magmatic and extensional processes modified the crustal and lithospheric structures during Subjotnian and Jotnian events (1650-1100 Ga).

The results show that the seismic reflection surveys are of great importance when crustal architecture and tectonic evolution are studied.

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The Eastern Fennoscandian Mafic Dyke Swarms GIS-Database

Jouni Vuollo[1], Heikki Salmirinne[1], Lauri J. Pesonen[2,3], Vladimir Stepanov[4], George Fedotov[5] and Dimitri Frank-Kamenetsky[6]

[1]Geological Survey of Finland, P.O. Box 77, FIN-96101 Rovaniemi, Finland
[2]Geological Survey of Finland, P.O. Box 96, FIN-02151 Espoo, Finland
[3] Institute of Geophysics, P.O. Box 4, FIN-00014 University of Helsinki, Finland
[4]Geological Institute, Karelian Research Center, RAS, Petrozavodsk, Russia
[5]Geological Institute, Kola Science Centre, RAS, Apatity, Russia
[6]State Company "Mineral", St. Petersburg, Russia
E-mail: jouni.vuollo@gsf.fi

A new mafic dyke swarm GIS-database has been crated at the Institute of Geosciences, University of Oulu, and the Geological Survey of Finland together with Russian Organizations. All the available data on dyke swarms are gathered into a GIS system (ArcInfo-ArcView) to help to combine and correlate the information. We will present an updated global dyke map of the eastern Fennoscandian Shield produced from this database.

Keywords: Fennoscandian Shield, mafic dyke swarms, GIS-database

1. Introduction

Research on the Palaeoproterozoic basic igneous activity in the eastern Fennoscandian Shield and correlation with the North Atlantic Area is taking place at the Institute of Geosciences, University of Oulu, and the Geological Survey of Finland, Rovaniemi. The project funded by the Academy of Finland. The aim is to obtain a clear picture of spatial and temporal distribution of mafic dyke swarms, their geochemical characteristics, geophysical properties, geochronology and relationship to ore critical magmatic events.

The role of GIS (Geological Information System) has been emphasized during the last few years, and geoscientists all over the world are increasingly relying on this method in creating maps and so on. By putting spatial data in an integrated system, where it can be organized, analyzed and mapped, it is possible to find patterns and relationships that were previously unrecognized. Integrated GIS data on various geological and geophysical phenomena and their characteristics will make it possible, for instance, to gain a better understanding of the geological history of the Fennoscandian Shield.

2. Eastern Fennoscandian Mafic Dyke Swarms GIS Databases

The first Global Mafic Dyke GIS Database was produced in Ottawa and funded by LITHOPROBE Canada. It included information on more than 300 dyke swarms, ~30 mantle plumes and isotope ages. In 1997, our project started to build up the "Eastern Fennoscandian Mafic Dyke swarms GIS Databases" in order to collect all the available geoscientific data for GIS purposes. Last year "State Company Mineral" started to collect Russian Dyke Swarm GIS Databases and we will present a database combined from these results.

Large amounts of valuable data, which can tell us about the geological evolution of the Fennoscandian Shield (ancient continental rift systems, subductions, mantle plumes etc.), are now available. All the data are gathered into a GIS system (ArcInfo-ArcView) to help to combine and correlate the information. The databases include both vector and raster spatial data as well as a large

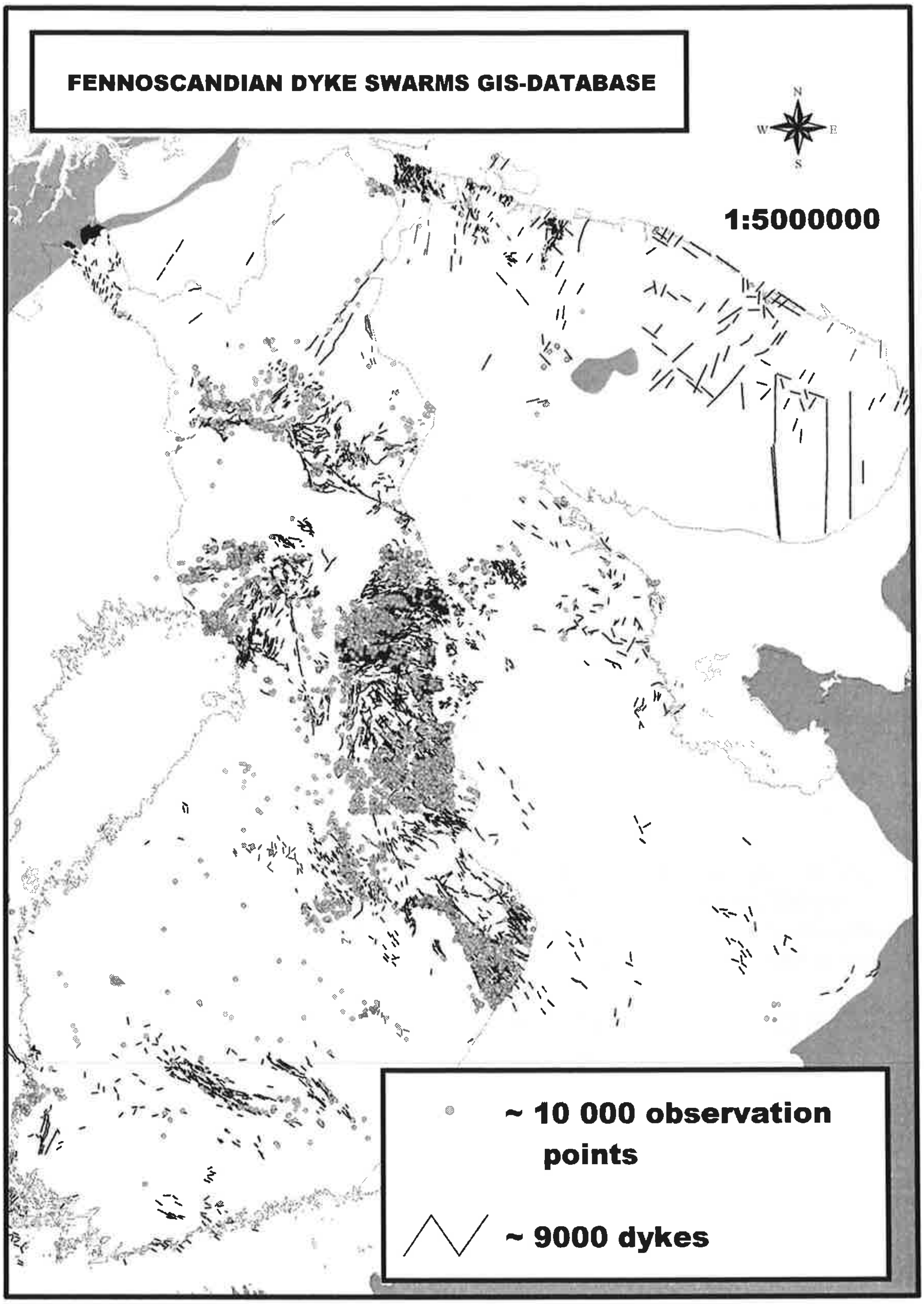

Figure1. Fennoscandian Dyke Swarms GIS-Database – observation points and dykes.

amount of attribute data. Dykes from all 52 of the 1:100 000-scale bedrock maps of eastern Finland and from many other detailed maps have been digitized. We have collected all the aeromagnetic maps of our study area in the form of TIFF pictures (pixel size 50m). The database includes more than 10 000 observation points in both Finland and Russia. The petrophysical database (density, susceptibility, remanence and Q-value) extracted from the national database of the Geological Survey of Finland (GTK) consists of 3700 samples. There is also a geochronological database that includes 60 U-Pb and 20 Sm-Nd age determinations on diabases of different types and ages all over the study area. The geochemical database consists of more than 1500 whole-rock analyses and numerous REE and PGE analyses, and the mineral analyses database comprises hundreds of silicate and oxide microprobe determinations. A totally updated, digital Eastern Fennoscandian dyke swarm map has been compiled in the scale of 1:50 000 with the aid of all these databases.

The poster will present an updated global dyke map for the eastern Fennoscandian Shield (Finland and some areas in Russian Karelia and the Kola Peninsula). The maps presented here have been produced from this database. Some detailed regional maps for the eastern part of Finland, Russian Karelia and the Kola Peninsula will be presented, showing the distribution and characteristics of the dyke swarms.

In the future, the final and most notable aim of the project will be to make a reconstruction of the crustal blocks of the eastern part of the Fennoscandian Shield by means of dated dyke swarms and to correlate the dyke swarms on an intercontinental basis in order to enable models to be developed for continental reconstructions (Fennoscandian Shield, Scotland, Greenland, Canadian Shield). Integrated GIS data on various geological and geophysical phenomena and their characteristics will make it possible to gain a better understanding of the geological history of the Fennoscandian Shield. Some preliminary crustal block reconstructions are presented in this volume (*Vuollo et al. 2000*), and the first palaeomagnetic evidence for a 2.44 Ga continental reconstruction of the North Atlantic region has been published by *Mertanen et al. (1999)*.

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LIST OF AUTHORS

Andreas Abels
Institute of Planetology, University of Münster, 48149 Münster, Germany
Meri-Liisa Airo
Geological Survey of Finland, P.O. Box 96, FIN-02151 Espoo, Finland
E-mail: meri-liisa.airo@gsf.fi
Jyrki Bergström
Department of Geology, P.O. Box 11, FIN-00014 Helsinki, Finland
Toni Eerola
Department of Geology, P.O. Box 11, FIN-00014 Helsinki, Finland
E-mail: tonieve.eerola@kolumbus.fi
Carl Ehlers
Åbo Akademi University, Department of Geology and Mineralogy,
FIN-20500 Turku, Finland
Olav Eklund
University of Turku, Department of Geology, FIN-20014 Turku, Finland
Seppo Elo
Geological Survey of Finland, P.O.Box 96, FIN-02151 Espoo, Finland
E-mail: seppo.elo@gsf.fi
Taina Eloranta
Department of Geology, P.O. Box 11, FIN-00014 Helsinki, Finland
George Fedotov
Geological Institute, Kola Science Centre, RAS, Apatity, Russia
Dimitri Frank-Kamenetsky
State Company "Mineral", St. Petersburg, Russia
Sören Fröjdö
Åbo Akademi University, Dept. Geology and Mineralogy, FIN-20500 Turku, Finland
Stanislav S. Gornostayev
Department of Geosciences, P.O. Box 3000, FIN-90014 University of Oulu, Finland
E-mail: ssg@cc.oulu.fi
Ilmari Haapala
Department of Geology, P.O. Box 11, FIN-00014 University of Helsinki
E-mail: ilmari.haapala@helsinki.fi
Eero Hanski
Geological Survey of Finland, P.O. Box 77, FIN-96101 Rovaniemi, Finland
E-mail: eero.hanski@gsf.fi
Pekka Heikkinen
Institute of Seismology, P.O. Box 26, FIN-00014 University of Helsinki, Finland
E-mail: pekka.heikkinen@seismo.helsinki.fi
Hannu Huhma
Geological Survey of Finland, P.O. Box 96, FIN-02151 Espoo, Finland
hannu.Huhma@gsf.fi
Tellervo Hyvönen
Institute of Seismology, P.O. Box 26, FIN-00014 University of Helsinki, Finland
E-mail: lhyvonen@seismo.helsinki.fi
Pentti Hölttä
Geological Survey of Finland, P.O. Box 96, FIN-02151 Espoo, Finland
E-mail: pentti.holtta@gsf.fi
Pertti Kaikkonen
Department of Geophysics, P.O. Box 3000, FIN-90014 University of Oulu, Finland

Juhani Kakkuri
Finnish Geodetic Institute, FIN-02430 Masala, Finland
E-mail: Juhani.Kakkuri@fgi.fi
Juha A. Karhu
Geological Survey of Finland, P.O. Box 96, FIN-02151 Espoo, Finland
E-mail: juha.karhu@gsf.fi
Tapio J. Koistinen
Geological Survey of Finland, P.O. Box 96,FIN-02151 Espoo, Finland
Juha V. Korhonen
Geological Survey of Finland, P.O. Box 96, FIN-02151 Espoo, Finland
E-mail: juha.korhonen@gsf.fi
Annakaisa Korja
Institute of Seismology, P.O. Box 26, FIN-00014 University of Helsinki, Finland
E-mail: annakaisa.korja@seismo.helsinki.fi
Toivo Korja
Geological Survey of Finland, P.O. Box 96, FIN-02151 Espoo, Finland
E-mail: toivo.korja@gsf.fi
Kalevi Korsman
Geological Survey of Finland, P.O. Box 96, FIN-02151 Espoo, Finland
E-mail: kalevi.korsman@gsf.fi
Elena Kozlovskaya
Department of Geophysics, P.O. Box 3000, FIN-90014 University of Oulu, Finland
Ilmo T. Kukkonen
Geological Survey of Finland, P.O. Box 96, FIN-02151 Espoo, Finland
E-mail: ilmo.kukkonen@gsf.fi
Niilo Kärkkäinen
Geological Survey of Finland, P.O. Box 96, FIN-02151 Espoo, Finland
Email: niilo.karkkainen@gsf.fi
Kauko V.O. Laajoki
Department of Geosciences, P.O. Box 3000, FIN-90014 University of Oulu, Finland
Raimo Lahtinen
Geological Survey of Finland, P.O. Box 96, FIN-02151 Espoo, Finland
E-mail: raimo.Lahtinen@gsf.fi
Martti Lehtinen
Finnish Museum of Natural History, FIN-00141 University of Helsinki, Finland
E-mail: martti.lehtinen@helsinki.fi
Urmas Luosto
Institute of Seismology, P.O. Box 26, FIN-00014 University of Helsinki, Finland
E-mail: luosto@seismo.helsinki.fi
Satu Mertanen
Geological Survey of Finland, P.O. Box 96, FIN-02151 Espoo, Finland
E-mail: satu.mertanen@gsf.fi
Kari Moisio
Department of Geophysics, P.O. Box 3000, FIN-90014 University of Oulu, Finland
E-mail: kari.moisio@oulu.fi
Mikko Nironen
Geological Survey of Finland, P.O. Box 96, FIN-02151 Espoo, Finland
E-mail: mikko.nironen@gsf.fi
Hugh O´Brien
Geological Survey of Finland, P.O. Box 96, FIN-02151 Espoo, Finland
Petri Peltonen
Geological Survey of Finland, P.O. Box 96, FIN-02151 Espoo, Finland
E-mail: petri.peltonen@gsf.fi

Lauri J. Pesonen
Department of Geophysics, P.O. Box 4, FIN-00014 University of Helsinki, Finland
E-mail: lauri.pesonen@helsinki.fi
Jüri Plado
Institute of Geology, University of Tartu, 51014 Tartu, Estonia
Sergiy E. Popovchenko
National Mining University of Ukraine, 19 Karl Marx Ave., Dnepropetrovsk, 49027, Ukraine
Riku Raitala
Department of Geology, P.O. Box 11, FIN-00014 Helsinki, Finland
Tapio Ruotoistenmäki
Geological Survey of Finland, P.O. Box 96, FIN-02151 Espoo, Finland
E-mail: tapio.ruotoistenmaki@gsf.fi
O. Tapani Rämö
Department of Geology, P.O. Box 11, FIN-00014 University of Helsinki, Finland
E-mail: tapani.ramo@helsinki.fi
Heikki Salmirinne
Geological Survey of Finland, P.O. Box 77, FIN-96101 Rovaniemi, Finland
Vladimir Stepanov
Geological Institute, Karelian Research Center, RAS, Petrozavodsk, Russia
Tom Stålfors
Åbo Akademi University, Department of Geology and Mineralogy,
FIN-20500 Turku, Finland
E-mail: tom.stalfors@abo.fi
Heikki Säävuori
Geological Survey of Finland, P.O. Box 96, FIN-02151 Espoo, Finland
Email: heikki.saavuori@gsf.fi
Pekka Tuisku
Department of Geosciences, P.O. Box 3000, FIN-90014 University of Oulu, Finland
E-mail: pekka.tuisku@oulu.fi
Ragnar Törnroos
Department of Geology, P.O. Box 11, FIN-00014 Helsinki, Finland
E-mail: ragnar.tornroos@helsinki.fi
Matti Vaasjoki
Geological Survey of Finland, P.O. Box 96, FIN-02151 Espoo, Finland
E-mail: matti.vaasjoki@gsf.fi
Petri Virransalo
Geological Survey of Finland, P.O. Box 96,FIN-02151 Espoo, Finland
E-mail: petri.virransalo@gsf.fi
Jouni Vuollo
Geological Survey of Finland, P.O. Box 77, FIN-96101 Rovaniemi, Finland
E-mail: jouni.vuollo@gsf.fi
Markku Väisänen
Geological Survey of Finland, P.O. Box 96, FIN-02151 Espoo, Finland
E-mail: markku.vaisanen@gsf.fi
Jani Wennerstrand
Department of Geology, P.O. Box 11, FIN-00014 University of Helsinki, Finland
Zhitong T. Wang
Finnish Geodetic Institute, FIN-02430 Masala, Finland

PART III: INVITED PAPERS

CONTINUED

Evolution of Proterozoic Surface Environments: Evidence from Carbon and Strontium Isotope Ratios in Sedimentary Carbonates

Juha A. Karhu

Geological Survey of Finland, P.O. Box 96, FIN-02151 Espoo, Finland
E-mail: Juha.Karhu@gsf.fi

Palaeoproterozoic surface environments of the Earth were subjected to many inter-related tectonic, climatic and biogeochemical events, which can be studied using variations in the isotopic composition of sedimentary carbonates. The earliest known continental glaciations occurred between about 2450 and 2200 Ma. Glaciations were followed by a major perturbation in the biogeochemical cycle of carbon, evidenced by a strong enrichment in ^{13}C of marine sedimentary carbonates between about 2200 and 2100 Ma. In contrast to the Palaeoproterozoic, the Mesoproterozoic Era was characterized by a prolonged period of environmental and tectonic stability, but the succeeding Neoproterozoic Era was again a time of great geologic, environmental and biologic revolutions. Major changes affected the Neoproterozoic carbon cycle. Periods with positive $\delta^{13}C$ values are punctuated with sharp negative excursions, associated with evidence of continental glaciations. The $^{87}Sr/^{86}Sr$ record of Proterozoic seawater shows a generally increasing trend from the Palaeoproterozoic to the Cambrian. This may be interpreted as resulting from a growing flux of radiogenic strontium supplied from weathering of the continental crust, whereas hydrothermal sources became less important.

Keywords: Carbonate sediments, carbon, C-13/C-12, strontium, Sr-87/Sr-86, Proterozoic

1. Introduction

Many questions related to the evolution of the surface environments of the Earth can be resolved by studying the variations of the isotopic composition of seawater through time. Isotope ratios of dissolved inorganic carbon and strontium in ancient seawater are preserved in sedimentary carbonate rocks. These isotope curves show secular variations due to changes in the biogeochemical and geochemical carbon and strontium cycles, and they may be used to set constraints for models simulating the evolution and the interactions of the surface environments on the Earth. In addition, the isotope curves may be utilized as a correlation tool for widely separated sedimentary sections. This is possible, because the oceans, at any time, are homogeneous with respect to the isotope ratios of carbon and strontium, due to long residence times of these elements in the oceans (*Holland, 1978*). Chemostratigraphic correlation is especially useful in the Proterozoic where biostratigraphic data is lacking.

2. Palaeoproterozoic

The surface environments of the Palaeoproterozoic Earth were subjected to many inter-related tectonic, climatic and biogeochemical events. Climatic events included the earliest known continental glaciations in North America, Fennoscandia, Australia and South Africa. Glaciogenic deposits in North America and Fennoscandia are constrained to have deposited between 2450 and 2200 Ma.

The Palaeoproterozoic glaciations were followed by a major positive excursion in the carbon isotope values of sedimentary carbonates between 2200 and 2100 Ma (Fig. 1; *Karhu and Holland, 1996*). The excursion reflects a similar change in the isotopic composition of the oceans, and it demonstrates that a major perturbation affected the Palaeoproterozoic carbon cycle.

The best evidence for the timing of the Paleoproterozoic excursion comes from the Fennoscandian Shield, where nearly all the Jatulian and equivalent dolomite formations have unusually high $\delta^{13}C$ values (*Karhu, 1993; Melezhik and Fallick, 1996*). The deposition of many of these carbonate units has been constrained using U-Pb data on zircons from overlying or underlying volcanic beds or from cross-cutting intrusions. Outside Fennoscandia, similarly ^{13}C-enriched Palaeoproterozoic carbonates have been found in North America, South America, India and in several locations of central and southern Africa. All these findings confirm that the excursion was global in character.

Glaciogenic sediments are not known to be in immediate contact with the ^{13}C-enriched carbonates. For example, on the Fennoscandian Shield the Sariolian glaciogenic rocks are separated from the ^{13}C-enriched Jatulian dolomites by a weathering interval and an unconformity (*Marmo and Ojakangas, 1984*). The Palaeoproterozoic Duitschland Formation in South Africa appears to be an exception, as ^{13}C-enriched carbonates have there been deposited prior to the Palaeoproterozoic glaciation (*Buick et al., 1998*). Sedimentary carbonates of the Duitschland Formation may, however, represent a separate carbon isotope excursion, predating the major anomaly between 2200 and 2100 Ma (*Buick et al., 1998*).

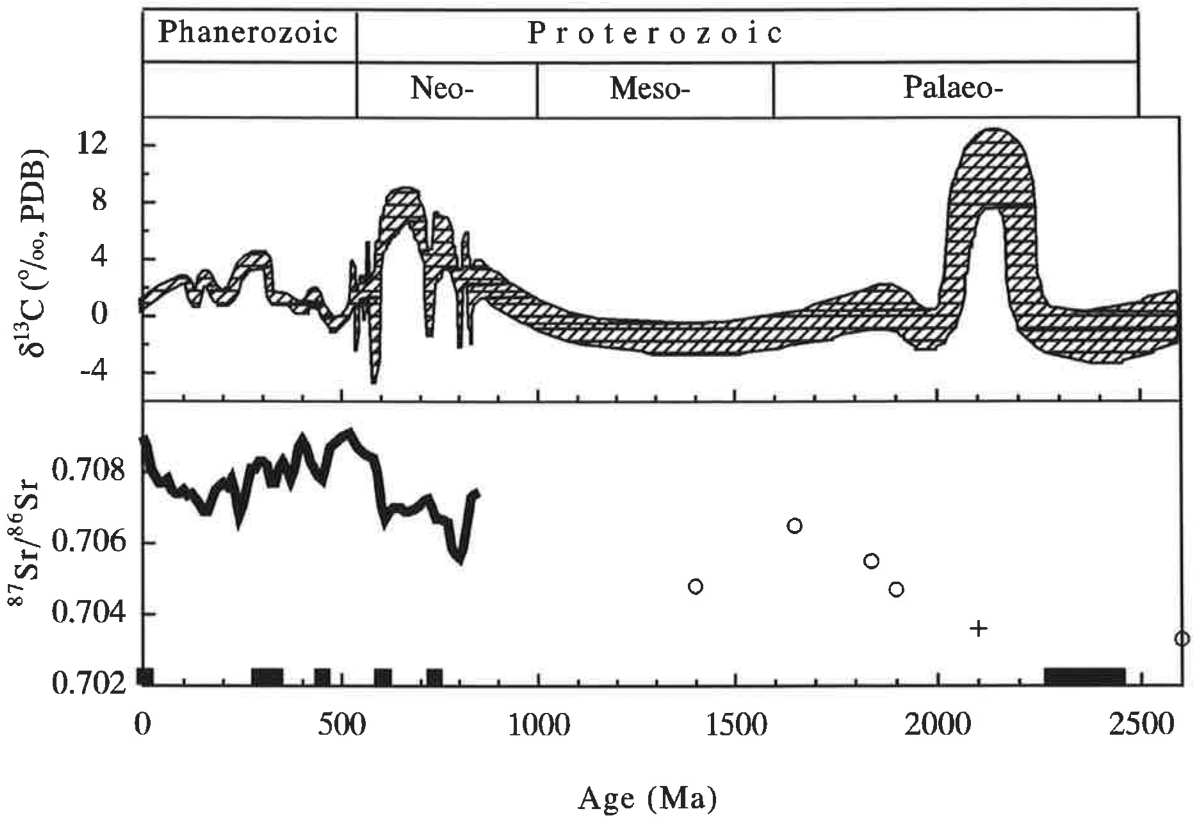

Figure 1. Summary of secular variations of the carbon and strontium isotope ratios in seawater. Sources for carbon isotope data are *Holser et al. (1988), Kaufman and Knoll (1995), Karhu and Holland (1996)* and *Brasier and Lindsay (1998)*. Sources for strontium isotope data are *Burke et al. (1982)* for the Phanerozoic and *Kaufman and Knoll (1995)* for the Neoproterozoic. Open dots refer to $^{87}Sr/^{86}Sr$ data from *Veizer et al. (1992)* for the Palaeoproterozoic and from *Hall and Veizer (1992)* for the Mesoproterozoic. The cross marks the lowest $^{87}Sr/^{86}Sr$ ratio measured from Jatulian dolomites (ca. 2100 Ma) of the Fennoscandian Shield (unpublished data, J. Karhu). Glacial periods are indicated as solid bars.

3. Mesoproterozoic

The Palaeoproterozoic time of climatic and biogeochemical perturbations was followed by a prolonged period of environmental and tectonic stability (*Brasier and Lindsay, 1998*). The time of stability lasted for about 1000 Ma and included the whole Mesoproterozoic Era and parts of the Paleoproterozoic and Neoproterozoic Eras. The period is characterized by invariable $\delta^{13}C$ values around 0 ‰ (Fig. 1), suggesting that the surficial redox conditions remained stable. In addition, evidence for climatic extremes, such as continental glaciations, is absent.

4. Neoproterozoic

The succeeding Neoproterozoic Era was, in strong contrast to the preceeding period, a time of great geologic, environmental and biologic revolutions. Major changes affected the carbon cycle, as is evidenced by thick sedimentary sections with ^{13}C-enriched carbonates having highly positive $\delta^{13}C$ values. The periods with positive $\delta^{13}C$ values are punctuated with sharp negative excursions (Fig. 1), which are invariably associated stratigraphically with evidence of continental glaciations. At least two (*Kennedy et al., 1988*), but possibly as many as four discrete glaciations (*Jacobsen and Kaufman, 1999*) have been recognised. The two broad levels with glacial sediments have been identified as the Sturtian (about 700-760 Ma) and Varangian (about 600 Ma) ice ages (*Jacobsen and Kaufman, 1999*). Some of these glaciers seem to have flowed into ice-covered tropical seas, possibly during a period of almost complete global glaciation as a "snowball Earth" (*Kirschvink, 1992; Hoffman et al., 1998*).

5. Secular Variation of Strontium Isotopes

For the analysis of strontium isotope ratios, sedimentary carbonate samples are screened carefully for alteration, because the strontium isotope signatures are very sensitive to diagenetic and metamorphic reactions (*Kaufman et al., 1993*). Due to alteration effects, the $^{87}Sr/^{86}Sr$ ratios tend to increase, and therefore the lowest $^{87}Sr/^{86}Sr$ for a given time interval typically provide the most reliable estimates of coeval seawater values.

The Palaeoproterozoic and Mesoproterozoic evolution of the marine $^{87}Sr/^{86}Sr$ values is poorly known, but more data exist for the latter part of the Neoproterozoic Era (Fig. 1). The generally increasing $^{87}Sr/^{86}Sr$ trend from the Paleoproterozoic to the Cambrian may be interpreted as resulting from a growing flux of radiogenic Sr supplied from weathering of the continental crust, while the hydrothermal sources became less important (*Veizer, 1989*). The secular variations of the strontium isotope ratios in the Neoproterozoic and Phanerozoic are ascribed to the same mechanism. Low $^{87}Sr/^{86}Sr$ values are suggestive of a higher hydrothermal flux due to active plate motions, whereas high values testify for mountain building and increased erosion rates. The steep increases in marine $^{87}Sr/^{86}Sr$ values during the Vendian and Cenozoic (Fig. 1) have been ascribed to Pan-African and Himalayan continental collisions, respectively.

6. Conclusion

The evolution of the Proterozoic surface environments during the Palaeoproterozoic and Neoproterozoic Eras were characterized by major tectonic, climatic and biogeochemical revolutions. These two periods were separated by a prolonged, 1000 Ma interval of environmental stability in crustal dynamics, redox conditions and global climate. The $^{87}Sr/^{86}Sr$ ratios of Proterozoic seawater show an increasing trend from low, mantle-like values at the Archaean-Proterozoic boundary to highest values in the Cambrian. The seawater carbon and strontium isotope curves help us to understand the evolution of the surficial system of the

Earth, but in addition, these isotope curves are emerging as an important tool for stratigraphic correlation in the Proterozoic, where a biostratigraphic framework is lacking.

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Extensive Zone of Mafic-Felsic Magma Interaction in the Svecofennian: the Hyvinkää-Mäntsälä Gabbroic Belt, Southern Finland

Toni Eerola[1], Riku Raitala[1], Jyrki Bergström[1], Taina Eloranta[1], Niilo Kärkkäinen[2] and Ragnar Törnroos[1]

[1] Department of Geology, P.O. Box 11, FIN-00014 Helsinki, Finland
[2] Geological Survey of Finland, P.O. Box 96, FIN-02151 Espoo, Finland
E-mail: tonieve.eerola@kolumbus.fi

The occurrence of magma mingling and mixing has been recently observed in all investigated mafic-ultramafic intrusions of the Lower Proterozoic Hyvinkää-Mäntsälä Gabbroic Belt in southern Finland. Structures and textures indicating magma mingling include mafic magmatic enclaves (MME), double enclaves and disrupted synplutonic mafic dykes with crenulated and chilled margins and quartz ocelli, corroded K-feldspar xenocrysts with or without plagioclase mantles, acicular apatite and uralitized pyroxene phenocrysts. Observation of such features in a wide area suggests that coeval mafic-felsic magma interaction took place in an extensive zone during the Svecofennian orogeny.

Keywords: Magma mingling, layered intrusions, Hyvinkää-Mäntsälä Gabbroic Belt, Svecofennian orogeny, Lower Proterozoic, southern Finland.

1. Introduction

The intrusion of basaltic magma into the crust is often intimately associated with the generation of silicic magmas by crustal melting *(Huppert & Sparks, 1988)*. During this process, the magmas could act together and eventually be mixed and mingled. This interaction between mafic and felsic magmas is a common phenomenon but a complex petrologic process which occurs in all tectonic environments (*Didier and Barbarin, 1991; Wiebe, 1993; Neves and Vauchez 1995; Salonsaari, 1995)*. Structures and textures indicating magma mingling include mafic magmatic enclaves (MME), double enclaves and disrupted synplutonic mafic dykes with crenulated and chilled margins and quartz ocelli, corroded K-feldspar xenocrysts with or without plagioclase mantles, acicular apatite and uralitized pyroxene phenocrysts (*e.g. Vernon 1983; Didier and Barbarin, 1991; Sylvester 1998)*.

In Finland, magma mingling and mixing have been observed in syn- and post-orogenic stages of the 1.95-1.80 Ga Svecofennian orogeny and in the Middle Proterozoic anorogenic rapakivi granite complexes (*e.g. Nironen and Bateman, 1989; Salonsaari, 1995; Eklund et al., 1998; Rämö et al., 1999*).

This paper describes features suggesting the extensive occurrence of mingling and mixing of mafic and felsic magmas in the Svecofennian Hyvinkää-Mäntsälä Gabbroic Belt (HMGB) in southern Finland (Fig. 1). Some related geological and economic implications are also suggested. The emphasis is on the Soukkio Complex in Mäntsälä, earlier thought as an example of the granitisation of gabbro by potassic metasomatism (*Härme, 1958; 1978; e.g.* Protected Educational Site 108 in Fig. 2; Kananoja & Grönholm 1993).

2. The Hyvinkää-Mäntsälä Gabbroic Belt

The HMGB forms a 150 km long and 20 km wide E-W trending zone of several 1.88-1.87 Ga mafic-ultramafic intrusions, occurring from Somero to the Wiborg Rapakivi Granite Complex (Fig. 1). These intrusions crosscut the surrounding 2.0-1.9 Ga metavolcano-sedimentary island arc sequence of Häme Belt (*Kähkönen et al., 1994*) and are easily distinguishable in the aeromagnetic maps as high anomalies. All the mafic-ultramafic intrusions show well-developed dynamic and non-dynamic layering, strong crystal fractionation, and magmatic differentiation *(Raitala, 1997; Bergström, 2000; Eerola, 2000; Eloranta, 2000)*. The Vähävesi, Karkkila and Hyvinkää (1.88 Ga; *Patchett and Kouvo, 1986*) gabbros are the major intrusions and several minor mafic-ultramafic bodies occur in the Somero, Mäntsälä and Pukkila areas. The Mäntsälä intrusions (Fig. 2) comprise hornblende-gabbros, hornblendites and anorthosites, the most important of which are the Hirvihaara and Soukkio Gabbros (1.87 Ga; *Huhma, 1986*).

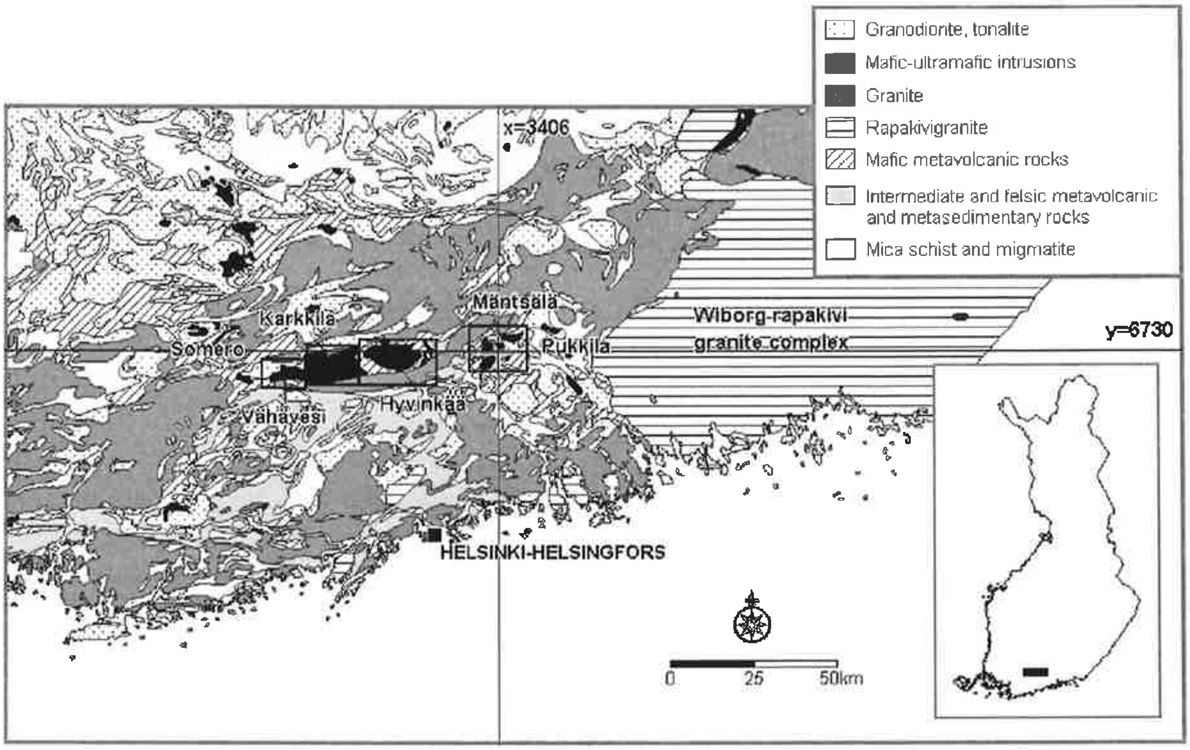


Figure 1. Precambrian basement of Southern Finland (after *Koistinen, 1994*) showing the main mafic-ultramafic intrusions of the Hyvinkää-Mäntsälä Gabbroic Belt. Investigated intrusions are framed.

3. Magma Mingling and Mixing Features

The Mafic-Silicic Layered Complex of Soukkio in Mäntsälä. The Soukkio Complex consists of layered tholeiitic gabbro (Site 1 in Fig. 2) and peraluminous calc-alkaline porphyritic monzonite-granite *(Eerola, 2000)*. The complex is clearly distinguished in aeromagnetic and gravity maps by high anomalies. The complex displays features related to mingling in a 1-2 km wide zone surrounding the gabbro at Lake Kilpijärvi. In the western side of the complex, hybridized felsic ameboidal layers occur with fine grained mafic and ultramafic layers with chilled margins against the felsic ones, cut by felsic dykes and pipes (Site 2 in Fig. 2). These features are similar to those described by *Wiebe, (1993)* in the mafic-silicic layered intrusions in Coastal Maine, USA (e.g. Pleasant Bay gabbro-diorite), where basic magma was injected into and trapped in a silicic magma chamber, forming layering of felsic and mafic magmas. In the central part of the Soukkio Complex (Site 3 in Fig. 2) there are abundant MMEs and mafic pillows hosted by monzonite (*Eerola, 2000*). MMEs are globules of hot mafic magma quenched against the granitic host in which they were injected *(e.g. Vernon, 1984; Neves and*

Vauchez 1995; Sylvester, 1998). The mafic pillows represent disruption of mafic magma by granitic veins when both were molten *(e.g. Wiebe, 1993)*. In the eastern side of the complex, synplutonic mafic, felsic and intermediate hybrid dykes intruded crystallizing granite (Figs. 3a, 3b) and were disrupted during a magmatic flow of the host *(e.g. Neves and Vauchez 1995; Salonsaari, 1995; Sylvester 1998; Mellqvist, 1999)*(Site 4, Fig. 2).

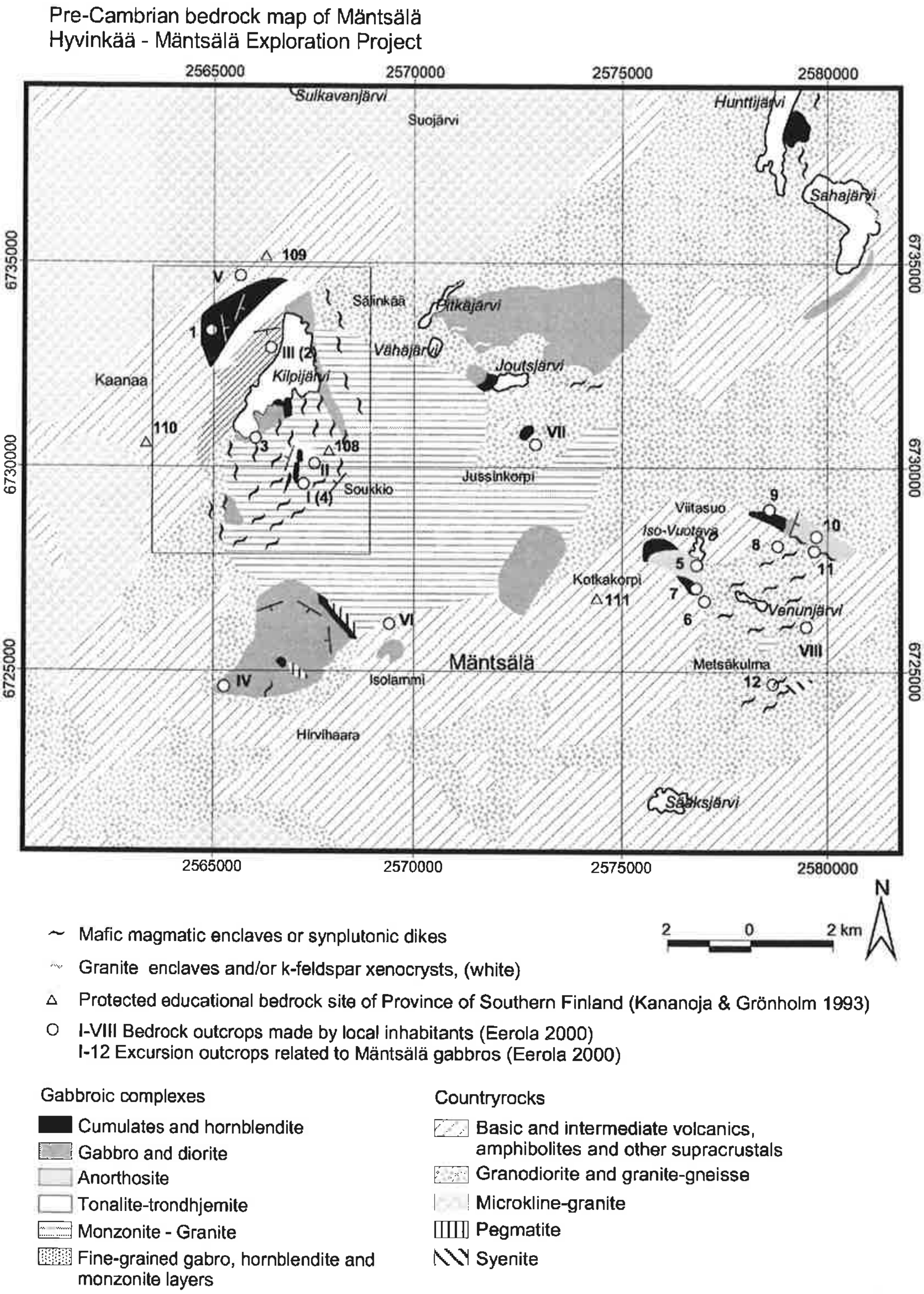
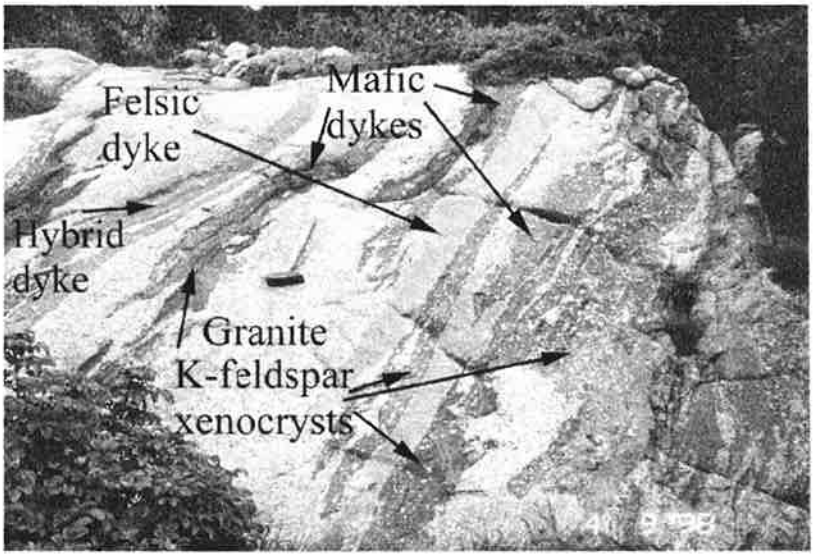


Figure 2. Precambrian basement map of the Mäntsälä region (after *Kaitaro, 1956*). The Soukkio Complex is framed.

In all these features are found rapakivi-textured K-feldspar ovoids, quartz-ocelli, acicular apatite, and double enclaves, typical textures and structures indicating thermal disequilibrium and mechanical mixing of contrastant magmas, suggesting magma mingling *(cf. Vernon, 1983; Didier and Barbarin, 1991; Neves and Vauchez 1995; Salonsaari, 1995; Sylvester, 1998; Mellqvist, 1999)*. Tonalites (Fig. 2) were probably formed by magma mixing.



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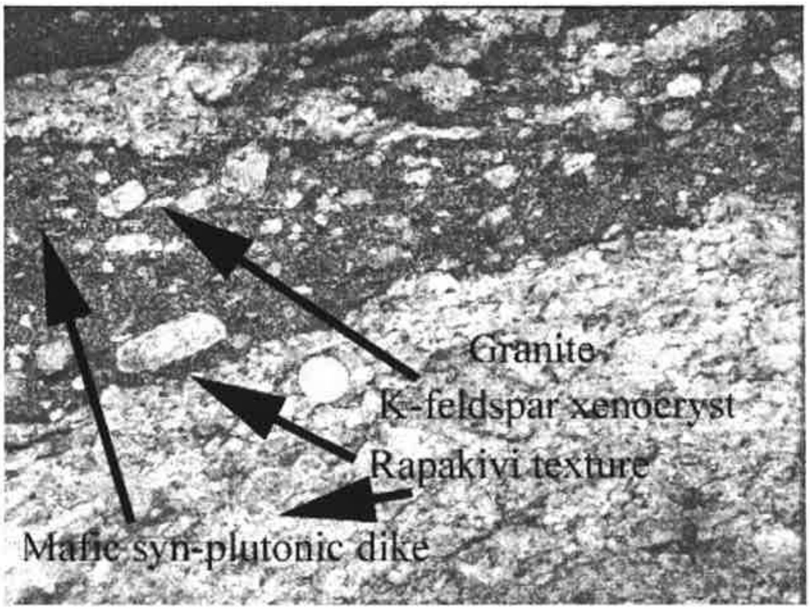

3b

Figure 3. Synplutonic dykes at Soukkio Complex, Mäntsälä.
3a. Mafic, felsic and hybrid synplutonic dykes hosted by granite, Soukkio Complex, Mäntsälä. Hammer has length of 0.5 m. Photo by T. Eerola.
3b. Detail of an synplutonic mafic dyke (but not of 3a) with rapakivi textured K-feldspar xenocrysts and megacrysts and crenulated margins against the granite, Soukkio Complex, Mäntsälä. Note the rapakivi-textured K-feldspar ovoids also in the granite. The coin is 2 cm wide. Photo by T. Eerola.

The preliminary geochemical data (*Eerola, 2000*) shows that MMEs and synplutonic mafic dykes have compositions intermediate between those of the granite and gabbro (Fig. 4a), evidencing mixing of the enclaves with the host. The granite hosted enclaves and coexisting mafic magma have very similar REE compositions (Fig. 4b), in accordance with observations of *Salonsaari, (1995)*. REE data for the granite is not available.

As the late-orogenic microcline-granite intruded the crystallized Soukkio granite *(Härme, 1978)* and gabbro, it does not caused any kind of interaction or effect on the gabbro. This and evidences above rules out the K-metasomatic granitisation of gabbro by microcline-granite advocated by *Härme, (1958; 1978)*. Based on modern concepts and literature, *Eerola, (2000)* suggested that presented features were produced by magma mingling.

Textures and structures suggesting magma mingling are also found elsewhere in the Mäntsälä region (Hirvihaara, Pitkäjärvi, Joutsjärvi and Jussinkorpi intrusions and in the Viitasuo-Metsäkulma area, e.g. Sites 8 and 12; Fig. 2)(*Eerola, 2000*).

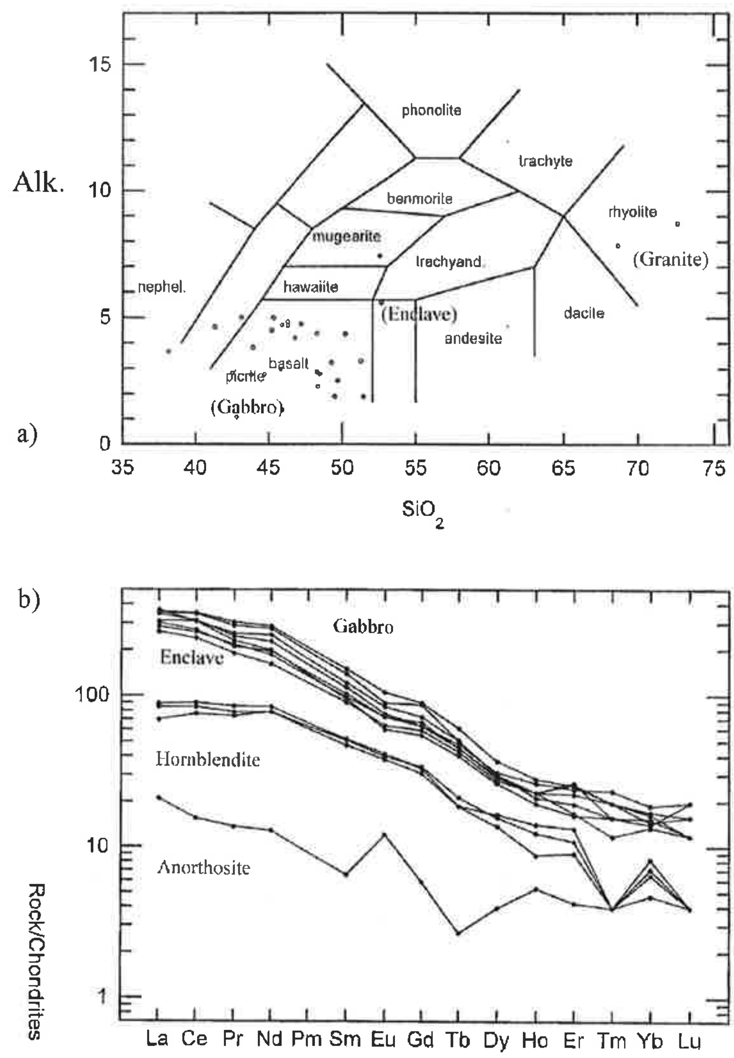

Figure 4. Rock classification diagrams.
4a. Rock classification diagram of *Cox et al., (1979)*, with whole rock analyses of gabbros, enclave and granites of the Soukkio Complex.
4b. Chondrite-normalized REE diagram (*Nakamura, 1974*) with whole rock analyses of gabbros, hornblendites, anorthosites and enclave from the Soukkio Complex, Mäntsälä.

The Hyvinkää Basic Layered Intrusion. The Hyvinkää intrusion is the largest of the synorogenic, layered gabbroic intrusions of the HMGB. It shows five major lithologic units in stratigraphic order: layered ultramafic cumulates, plagioclase-pyroxene-olivine cumulates, layered hypidiomorphic gabbro-norites, homogenous gabbro-diorites and granophyre (*Raitala, 1997*). The granophyre is largely product of crustal anatexis. The contact of the uppermost diorites and the lowermost granophyre shows classic features of interaction between mafic and felsic magmas. The melanogranophyric sequence, product of magma mixing, is a relatively thin local (0-50 m) zone between diorite and felsic granophyre. The felsic granophyre shows S-type geochemical characteristics, is extremely heterogeneous, and forms the major part of the granophyre unit. The lower part of the felsic granophyre is characterized by mafic volcanogenic xenoliths. Features suggesting magma mingling are found also in a smaller mafic-felsic body to the north of the Hyvinkää intrusion.

The Karkkila Basic Layered Intrusive Complex. The Karkkila Complex shows features of mixing and mingling in large area in the form of hybrid rocks carrying MMEs (Fig. 5) with quartz-ocelli and rapakivi textured K-feldspar xenocrysts (*Bergström, 2000*). Mafic and felsic magmas were mingled and mixed in a separate magma chamber, generating hybrid magma which invaded the basic layered body.

The Vähävesi Basic Layered Intrusion. Interaction of contrasting magmas occured in the southern margin of the Vähävesi intrusion, where there is a hybrid zone of quartz-diorite and granodiorite with sub-rounded MMEs with quartz-ocelli and K-feldspar xenocrysts (*Eloranta, 2000*). The felsic magma, intruding the gabbro, formed a large zone in the roof area and was possibly produced by the melting of crustal rocks (*Eloranta, 2000*).

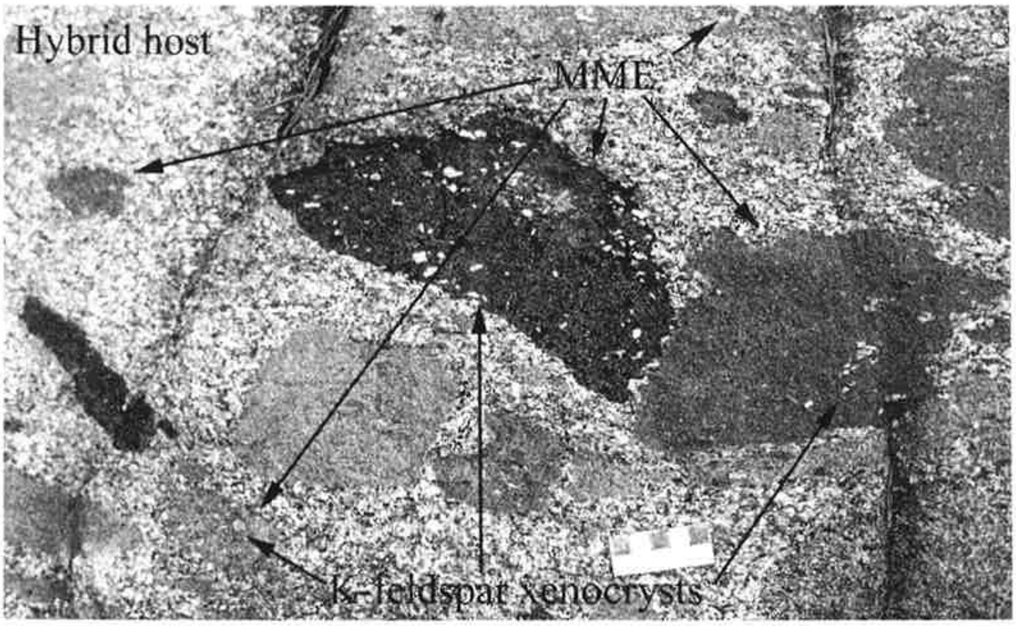

Figure 5. A swarm of different types of MMEs with K-feldspar xenocrysts, hosted by felsic hybrid rock in the Karkkila complex. The scale is centimetric. Photo by J. Bergström.

4. Geological and Economical Implications

Textures and structures related to the interaction of coeval contrasting magmas have been observed in all investigated intrusions of the HMGB. This reveals a widespread zone of coeval contrasting magmatism related to interaction between mantle magmas and crustal rocks during the Svecofennian orogeny in southern Finland. The intrusion of several mafic bodies of the HMGB produced widespread crustal melting, generation of granitic magma and subsequent mingling and mixing.

The correlation of increased TiO2 and P2O5 contents in mafic intrusive rocks and the occurrence of coeval felsic and basic magmas are regarded as positive indicators for Fe-Ti-P ore potential in mafic-ultramafic intrusions, especially in the Middle Proterozoic anorthosite-rapakivi-granite associations (*Emslie, 1978*). The high Ti and P in the parental melt of the Kauhajärvi gabbro, western Finland, rich in apatite, ilmenite and magnetite, are suggested to be a result of an interaction of tholeiitic and granitic magmas in a late- or post-orogenic bimodal magmatic system (*Kärkkäinen and Appelqvist, 1999*).

Many mafic intrusions in the HMGB are enriched in TiO2 (2-5%) and P2O5 (2-3%), and there are indications of Fe-Ti-P mineralizations in this region. The most promising prospects are an apatite-ilmenite mineralization with 11.3% TiO2 and 2.8% P2O5 at the small Mustalampi Gabbro (probably a satellite of the Vähävesi intrusion, *Eloranta 2000*), near Karkkila *(Härme, 1955; Kärkkäinen and Appelqvist, 1999)* and a nelsonite boulder (with 11.2% TiO2, 5.4% P2O5) found at Onkimaanjärvi to the SE from Mäntsälä.

The observations presented here may well be found in many other Svecofennian complexes and the phenomenon of mafic-felsic magma interaction is probably much more common than has previously been believed.

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