INSTITUTE OF SEISMOLOGY
UNIVERSITY OF HELSINKI

REPORT S-42

LITHOSPHERE 2002

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

Geological Survey of Finland, Auditorium
Espoo, Otaniemi, November 12-13, 2002

PROGRAMME AND EXTENDED ABSTRACTS

edited by

Raimo Lahtinen, Annakaisa Korja, Katriina Arhe, Olav Eklund, Sven-Erik Hjelt, Lauri J. Pesonen

Helsinki 2002

INSTITUTE OF SEISMOLOGY
UNIVERSITY OF HELSINKI
REPORT S-42

LITHOSPHERE 2002

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

PROGRAMME AND EXTENDED ABSTRACTS

edited by Raimo Lahtinen, Annakaisa Korja, Katriina Arhe,
Olav Eklund, Sven-Erik Hjelt, Lauri Pesonen

HELSINGIN YLIOPISTO OULUN YLIOPISTO GTK

Geological Survey of Finland, Auditorium
Espoo, Otaniemi, November 12-13, 2002

Helsinki 2002

TABLE OF CONTENTS

PREFACE

A profound knowledge of the composition, formation and evolution of the lithosphere is the key to the understanding of the geological and geophysical processes affecting human life. Two years ago the Finnish National Committee of the International Lithosphere Programme (ILP) organized a symposium, called the *"LITHOSPHERE 2000"*, to promote discussion on the latest results of lithosphere research carried out in Finland. Altogether 90 scientists, of whom 30 were students, attended the symposium where 27 papers were presented either as papers or posters. The lithosphere research has continued to be very active. The ESF and ILP based EUROPROBE subproject, SVEKALAPKO, which focuses on the structure, composition and evolution of the Precambrian Europe, is in its synthesis stage. Shield-scale mapping results on geology and geophysics have enhanced our knowledge on the lithosphere of the Fennoscandian Shield and new evolution models have been presented. A major new contribution to the future Finnish lithosphere research is the large-scale reflection seismic project FIRE with its over 1700 km of crustal-scale reflection profiles.

Encouraged by the success of the first symposium and the overwhelming amount of new data the ILP decided to organize a biennial symposium, ***LITHOSPHERE 2002***. Besides a follow-up of the *LITHOSPHERE 2000* studies, an effort has been made to pursue a wider continental-scale approach to the evolution of the Fennoscandian Shield. There is also a growing interest in the Mesoproterozoic to Phanerozoic evolution of the Fennoscandian Shield, and its effect on the Finnish bedrock. The ILP committee wishes that this symposium gives a state of art summary of the lithosphere research in Finland and inspires other geoscientists as well as the national funding agencies to promote further studies of the lithosphere.

The two-day symposium will take place in the Auditorium of the Geological Survey of Finland (GTK), in Otaniemi, Espoo, in November 12-13, 2002. The symposium will be hosted by the ILP and supported by the Universities of Helsinki, Oulu, Turku and the Geological Survey of Finland. Both oral and poster papers will be presented. Posters prepared by graduate- or post-graduate students will be evaluated and the best one will be awarded.

The symposium has the following themes:

Theme 1. **Neotectonics in the Fennoscandian Shield**
Theme 2. **Evolution of the East European Craton – Effect on the Crust and Lithosphere in the Fennoscandian Shield**
Theme 3. **The Structure and Composition of the Lithosphere in the Fennoscandian Shield**
Theme 4. **The ongoing and new lithosphere projects in Finland and Europe**
Theme 5. **General Discussion and the Poster Award**

This special volume "***LITHOSPHERE 2002***" contains the programme and extended abstracts of the symposium "Lithosphere 2002". The abstracts of both oral and poster presentations are listed in alphabetical order.

Helsinki, October 10, 2002

Raimo Lahtinen, Annakaisa Korja, Olav Eklund, Sven-Erik Hjelt, Lauri Pesonen
Organizing Committee

PROGRAMME

Tuesday, November 12, 2002, 9.00-20.00

09.00-10.00	Registration at the Geological Survey of Finland, Espoo
10.00-11.00	***Opening Session*** ***Chair S.-E. Hjelt***
10.00-10.05	Opening of the Symposia by the Chairman of the National Committee of the ILP *(S.-E. Hjelt)*
10.05-11.00	Key Note Lecture: Crustal Growth and Recycling: An Overview of LITHOPROBE Results *(Clowes, R.M.)*
11.00-12.00	***Neotectonics in the Fennoscandian Shield*** ***Chair P. Heikkinen***
11.00-11.25	Postglacial Rebound and Crustal Deformation in Finland: Recent Geodetic Results *(Mäkinen, J., Ahola, J., Jokela, J., Koivula, H., Ollikainen, M., Poutanen, M., Saaranen, V., Takalo, M. and Vermeer, M.)*
11.25-11.50	Earthquakes and Seismotectonics in Finland *(Saari, J. and Uski, M.)*
11.50-12.00	Discussions
12.00-13.00	***Lunch Break***
13.00-15.30	***Evolution of the East European Craton –*** ***Effect on the Crust and Lithosphere in Finland*** ***Chair P. Nurmi***
13.00-13.25	Palaeomagnetic Configuration of Continents During the Proterozoic *(Pesonen, L.J. and Mertanen, S.)*
13.25-13.50	SVEKALAPKO Project: A Contribution to an Improved Understanding of the Evolution of the Fennoscandian Shield *(Hjelt, S.-E. and Daly, J.S.)*
13.50-14.15	Comparison of Crustal Models of the Sarmatia and Fennoscandian Shields *(Yliniemi, J., Tiira, T. and Komminaho, K.)*
14.15-14.25	***Break***
14.25-14.50	Paleoproterozoic Tectonic Evolution of the Fennoscandian Shield – A Comparison to Modern Analogues *(Nironen, M., Lahtinen, R. and Korja, A.)*
14.50-15.15	Sedimentary Record and Magmatic Episodes Reflecting Mesoproterozoic-Phanerozoic Evolution of the Fennoscandian Shield *(Kohonen, J. and Rämö, O.T.)*
15.15-15.30	Discussions
15.30-16.00	***Coffee Break***

16.00-17.30	***The Structure and Composition of the Lithosphere in the Fennoscandian Shield, Part I*** ***Chair O. Eklund***
16.00-16.25	The Age of the Lithosphere in the Fennoscandian Shield – Sm-Nd Isotopic Evidence *(Huhma, H.)*
16.25-16.50	Crust and Upper Mantle Beneath Fennoscandia as Imaged by the Baltic Electromagnetic Array Research (BEAR) *(Korja, T., Hjelt, S.-E., Kaikkonen, P., Kozlovskaya, E., Lahti, I., Pajunpää, K., Pulkkinen, A., Viljanen, A. and BEAR Working Group)*
16.50-17.15	Lithosphere Structure Beneath Southern Finland Derived by SVEKALAPKO Seismic Array Research *(Kozlovskaya, E., Yliniemi, J., Hjelt, S.-E., Komminaho, K. and SVEKALAPKO Seismic Tomography Working Group)*
17.15-17.30	Discussions

17.30-20.00	***Posters and Refreshments*** ***Chair I.T. Kukkonen***
P1.	Crustal Conductivity Map of the Fennoscandian Shield *(Korja, T., Engels, M., Zhamaletdiniv, A.A., Kovtun, A.A., Palshin, N.A., Smirnov, M.Yu., Tokarev, A.D., Asming, V.E., Vanyan, L.L., Vardaniants, I.L. and BEAR Working Group)*
P2.	Lithospheric Conductivity along GGT/SVEKA Transect: Implications from the 2-D Inversion of Magnetotelluric Data *(Lahti, I., Korja, T., Pedersen, L.B., and the BEAR Working Group)*
P3.	The Research Project: 3-D Crustal Model of the Finnish Part of SVEKALAPKO Research Area *(Kozlovskaya, E., Pirttijärvi, M., Hjelt, S.-E., Korhonen, J.V., Elo, S. and Yliniemi, J.)*
P4.	Petrophysical Characterisics of Intrusive Rocks in Finland Compared to Intrusives in Subareas of the Seismic Reflection Profile FIRE-1 *(Ruotoistenmäki, T.)*
P5.	Geochemistry and Metamorphic Petrology of the Archaean Pudasjärvi Granulite Area, Finland *(Lalli, K., Juopperi, H. and Tuisku, P.)*
P6.	2.44 Ga Bimodal Magmatism in Koillismaa and Adjacent Russia: the Nd Isotopic Story *(Lauri, L.S., Rämö, O.T., Karinen, T. and Huhma, H.)*
P7.	The South Finland Shear Zone - Ductile Shearing of the Paleoproterozoic Crust in SW Finland *(Torvela, T. and Ehlers, C.)*
P8.	Paleoproterozoic Remobilization and Enrichment of Proterozoic Uranium Mineralisation in the East-Uusimaa Area, Finland *(Vaasjoki, M., Appelqvist, H. and Kinnunen, K.A.)*
P9.	The Bodom and Obbnäs Rapakivi Granites, Southern Finland: Distinct Compositions Imply a Paleoproterozoic Terrane Boundary *(Kosunen, P., Rämö, O.T. and Vaasjoki, M.)*
P10.	The Mineralogy of Pegmatitic Cavities in a Fractionated Rapakivi Granite - the Influence of Fluids in Elemental Mobility in a Crystallizing Front, Vehmaa Rapakivi Batholith, SW Finland *(Toropainen, V.)*
P11.	Magnetic Overprints in the Silurian Dolomites, Central Estonia *(Plado, J., Pesonen, L.J., Kirsimäe, K., Puura, V., Pani, T. and Elbra, T.)*
P12.	A Geophysical Study of the Northern Baltic Sea *(Kuusisto, M., Heikkinen, P. and Korja, A.)*
P13.	Earth Dynamics and Environmental Change Are the Headlines in the IODP, the Successor to the Successful ODP *(Strand, K., Ehlers, K. and Tiensuu, K.)*

Wednesday, November 13, 2002, 9.00-17.00

9.00- 12.00	***The Structure and Composition of the Lithosphere in the Fennoscandian Shield, Part II*** ***Chair L.J. Pesonen***
9.00-9.25	Project FIRE: Deep Seismic Reflection Sounding in Finland 2001-2005. *(Kukkonen, I., Heikkinen, P., Ekdahl, E., Korja, A., Hjelt, S.-E., Yliniemi, J., Berzin, R. and FIRE Working Group)*
9.25-9.50	Archean Terranes and Boundaries in the Fennoscandian Shield – Implications for the Archean Evolution *(Hölttä, P. and Archean Working Group)*
9.50-10.15	Geophysical Maps and Crustal Model Interpretations of the Fennoscandian Shield *(Korhonen, J.V. and Elo, S.)*
10.15-10.45	***Coffee Break***
10.45-11.10	Evolution of the Lithospheric Mantle in Eastern Finland at 3.0–1.95 Ga Based on NORDSIM Dating of the Jormua Ophiolite Complex *(Peltonen, P., Mänttäri, I. and Kontinen, A.)*
11.10-11.35	A Slab Breakoff Model for the Differentiation of the Svecofennian Crust in Southern Finland *(Eklund, O. and Shebanov, A)*
11.35-12.00	Crustal-Scale Shear Zones in the Fennoscandia Shield *(Ehlers, C.)*
12.00-13.00	***Lunch Break***
13.00- 13.50	***The Structure and Composition of the Lithosphere in the Fennoscandian Shield, Part III*** ***Chair A. Korja***
13.00-13.25	Ore Lead Isotope Systematics During the Svecofennian Orogeny *(Vaasjoki, M. and Sundblad, K.)*
13.25-13.50	Radiogenic Leads from Fissures and Sandstones within the Fennoscandian Shield: Indications of Paleozoic Weathering-Related Enrichment of Base Metals *(Sundblad, K. Vaasjoki, M. and Alm, E.)*
13.50-14.00	Discussions
14.00-15.00	***Open Forum,*** ***The Ongoing and New Lithosphere Projects in Finland and Europe*** ***Chair K. Pajunpää***

14.00-14.10	GEODE	*(Eilu, P.)*
14.10-14.20	ODP and IODP	*(Strand, K.)*
14.20-14.30	MELODY	*(Kukkonen, I.T.)*
14.30-14.40	ESF-Impact, ICDP&GISP, IGCP 440	*(Pesonen, L.J.)*
14.40-14.50	IGCP 426	*(Rämö, O.T.)*
14.50-15.00	NorFa Network	*(Eklund, O.)*

15.00-15.30 ***Coffee Break***

15.30-16.30	***Closing Session***	
	Chair R. Lahtinen	
15.30-15.40	Poster Award	
15.40-16.10	Short Session Summaries by the Chairs	
16.10-16.30	General Discussion	
16.30	Concluding Remarks	*(Lahtinen, R.)*

EXTENDED ABSTRACTS

Crustal Growth and Recycling: An Overview of LITHOPROBE Results

Ron M. Clowes, with contributions from many other LITHOPROBE scientists

LITHOPROBE and Department of Earth & Ocean Sciences,
The University of British Columbia,
6339 Stores Road, Vancouver, BC, V6T 1Z4 Canada
E-mail: clowes@lithoprobe.ubc.ca

Keywords: crustal growth, LITHOPROBE, seismic reflection, accreted terranes, collisions, underplating

1. Introduction

The North American continent has grown through the progressive accretion, dispersal and re-aggregation of continental lithosphere in tectonic cycles spanning Archean to present times. The preserved crustal record in Canada ranges from the 4020 Ma Acasta gneisses in the northwestern Slave Province of the Northwest Territories to present oceanic crustal formation at the Juan de Fuca ridge off the west coast. The basic building blocks of the continent include: significant Middle to Late Archean crust preserved in supracrustal belts and plutonic rocks of the core of the continent (the Superior, Nain, Slave, Rae, Hearne and Wyoming cratons); the Paleoproterozoic orogenic belts that permanently welded the Archean cratons together (Trans-Hudson, New Quebec, Torngat, Wopmay, Thelon-Taltson, and Penokean); and the Mesoproterozoic Grenville, Paleozoic Applachian and Mesozoic-Cenozoic Cordilleran orogenies that border the Paleoproterozoic nucleus of the continent (Figure 1). In the southern part of the amalgamated continent, Laurentia, the Keweenawan rift system almost split the continent about 1100 Ma. LITHOPROBE transects have focused scientific studies on almost all of these primary features of Canada's landmass and offshore continental margins (Figure 1). From the new geophysical, geological and geochemical data generated by these studies, enhanced understanding of these features has been achieved.

By contrasting and comparing results among the many transects, new understanding of the processes associated with crustal growth, recycling and preservation also is being derived. Within this context of a pan-LITHOPROBE synthesis *(Percival et al., 1999)*, three principal questions are being addressed: (i) How much new (i.e., juvenile) crust was extracted from the mantle, accreted to the continents and preserved through geologic time, and how does the composition of that crust vary with depth and geologic time? (ii) What determines the probability of preservation of continental crust and lithosphere, and does it vary with geologic time? (iii) What processes are responsible for the recycling of crust and lithosphere into the convecting mantle? Here I summarize some of the extensive LITHOPROBE results that relate to these issues.

2. Crustal growth

2.1 Accreted oceanic terranes

Accreted oceanic terranes or collages of such terranes, comprising ocean floor basalts, oceanic arcs and back arc assemblages, are fundamental building blocks in the development and growth of evolving orogens. Throughout the LITHOPROBE studies, examples abound among transects and regions ranging from Archean to Cenozoic.

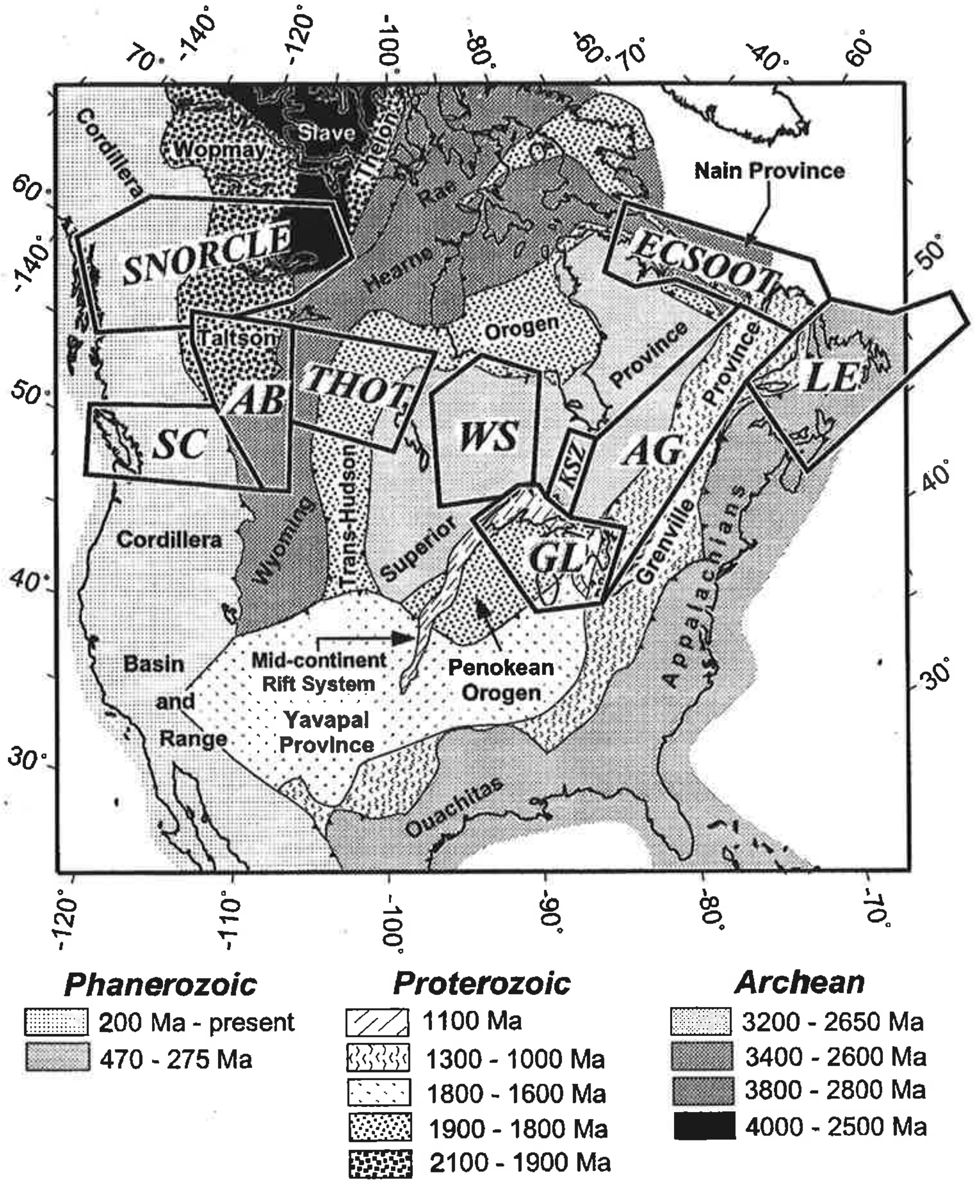


Figure 1. Location of LITHOPROBE transects (study areas) on a simplified tectonic element map of North America. Note how the Archean elements are generally stitched together, or added to, by Proterozoic elements, this amalgamation representing Precambrian North America to which was added the outlying Phanerozoic elements. The transects are: SC – Southern Cordillera; AB – Alberta Basement; SNORCLE – Slave – Northern Cordillera Lithospheric Evolution; THOT – Trans-Hudson Orogen; WS – Western Superior; KSZ – Kapuskasing Structural Zone; GL – Great Lakes International Multidisciplinary Program on Crustal Evolution (GLIMPCE); AG – Abitibi-Grenville; LE – Lithoprobe East; and ECSOOT – Eastern Canadian Shield Onshore-Offshore.

☺ In the Abitibi subprovince (eastern Ontario and Quebec) of the Archean Superior Province, results reveal the processes of accretion of primitive crust during northwest-directed subduction under the protocraton followed by additional accretion/collision to the south, the entire process occurring between about 2720 and 2680 Ma *(Calvert and Ludden, 1999)*.

☺ In the western Superior Province (western Ontario and eastern Manitoba), the type area for proposed Archean plate tectonics, recent reflection and refraction/wide-angle reflection data provide direct evidence for assembly of the craton by terrane accretion and for a large slab of remnant oceanic crust preserved within a paleo-subduction zone at the base of the crust *(White et al., 2002)*, results very similar to those from Abitibi.

☺ In the Paleoproterozoic Wopmay orogen west of the Archean Slave Province (northwestern Canada), accretion of the Hottah terrane to the Slave during the period 1900-1880 Ma took place as a result of west-dipping subduction along the rifted western margin of proto-North America *(Cook et al., 1999)*.

☺ In the Paleoproterozoic Trans-Hudson Orogen (Saskatchewan and Manitoba), all of the pre-accretion assemblages and most of the successor arc and basin deposits show a primitive character over a period of aggregation from 1910 to 1800 Ma; only the post-collisional intrusions (ca. 1790-1770 Ma) show evidence of derivation from remelting of ancient Archean crust *(Lucas et al., 1996)*. Results from the Trans-Hudson identified a previously unknown Archean microcontinent, the Sask craton, below the western half of the internal Reindeer zone, indicating the need for caution in relating subsurface volumes to surface exposures *(Lewry et al., 1994)*.

☺ In the Paleozoic Newfoundland Appalachians, closure of the ancient Iapetus ocean trapped a variety of oceanic and island-arc terranes, the Dunnage tectonostratigraphic domain, between two continental edges. However, its depth extent appears limited to the upper crust *(Hall et al., 1998)*.

☺ The Mesozoic-Cenozoic Cordillera of British Columbia and Alaska has been the cradle for many new concepts in global tectonics, including the concept of accreted terranes. In the Canadian Cordillera, about 500 km of westward growth has occurred. The Quesnel, Stikine, Cache Creek and Bridge River terranes in British Columbia represent components of this growth, although much of the surface exposures represent only thin flakes having limited depth extents.

2.2 Subduction underplating

Subduction zones can be responsible for crustal growth through subduction underplating. Two prominent examples of such underplating have been demonstrated in western North America.

☺ A pre-Tertiary and Tertiary underplate, located above the presently subducting plate, are interpreted from LITHOPROBE studies across the Cascadia subduction zone in the region of Vancouver Island (e.g., *Clowes et al., 1999)*.

☺ In south-central Alaska, seismic reflection, refraction and geological studies associated with the Trans-Alaska Crustal Transect have demonstrated that more than one third of the North American plate in southern Alaska consists of tectonically underplated oceanic lithosphere *(Fuis and Plafker, 1991)*.

2.3 Collisional processes

Collisional processes associated with orogenic development contribute to crustal growth, but the direct evidence for the sutures resulting from the collisions is often very cryptic, as they have been metamorphosed, reworked by later deformation and intruded by marginal (Andean-type) arc magmas. These magmas, derived from subducting slabs and the mantle wedges above them, form the impressive and extensive granitoid batholiths occupying the hinterlands of orogens. LITHOPROBE studies encompass a number of these.

☺ The Great Bear magmatic arc of the Wopmay orogen in the Northwest Territories is an extensive feature based on geological mapping. However, in the region of SNORCLE Transect Corridor 1, reflection data indicate that the arc extends only to depths of about 5 km *(Cook et al., 1999).*

☺ The Wathaman-Chipewyan batholith is a major component of the Trans-Hudson orogen in northern Saskatchewan and Manitoba with an exposed strike length of 900 km. This large feature formed between 1865 and 1850 Ma. Where crossed by LITHOPROBE reflection lines, the batholith is very narrow compared with its lateral extent elsewhere; interpretations indicate that it only extends to about 10 km depth.

☺ The De Pas batholith is located in the hinterland of the New Quebec orogen, northern Quebec and forms a 500-600 km long arcuate intrusion extending from the Grenville Front in southern Quebec and Labrador to the south shores of Ungava Bay in the north. Broad-based geological studies suggest that the batholith was generated in a continental magmatic arc setting between about 1840 and 1810 Ma. The tectonic scenario involves eastward-dipping subduction along the western margin of the Core Zone, a large area of mainly Archean rocks reworked in the Paleoproterozoic and located between the Archean Superior and Nain (or North Atlantic) cratons *(Dunphy and Skulski, 1996).*

☺ The Coast Mountain batholith, or Coast Plutonic Complex, in British Columbia and the Yukon extends for 1800 km along the suture zone between the accretionary Insular and Intermontane superterranes that form the western half of the Canadian Cordillera. LITHOPROBE and related studies show considerable variation along strike, with interpretations of the depth extent of the complex varying from about 10 km to almost the entire crust.

2.4 Magmatic underplating

Magmatic underplating is a known process for continental growth, but one that often is difficult to identify and establish, as the effects of the process are usually manifest in the lower crust or upper mantle. High seismic velocities, as inferred from refraction experiments and generally poor seismic reflectivity, can indicate such underplating.

☺ The Alberta Basement Transect Deep Probe refraction experiment across the Archean Hearne Province, Medicine Hat block and Wyoming Province of Alberta, Montana and Wyoming delineates a massive underplated zone, up to 30 km thick, that was emplaced in Proterozoic time after the various Archean blocks had collided and amalgamated *(Gorman et al., 2002).*

☺ Results from the Abitibi-Grenville Transect indicate a 5-8 km thick zone in the lower crust characterized by low reflectivity and high seismic velocities *(Calvert and Ludden, 1999).*

☺ In the LITHOPROBE East Transect, high velocity lower crust up to about 10 km thick, attributed to underplating, is found below the former Laurentian continental margin in western Newfoundland and below the Magdalen basin northeast of New Brunswick *(Hall et al., 1998).*

3. Crustal recycling and reorganization

Lithospheric delamination and transfer of lower crust and uppermost mantle to deeper parts of the mantle is one primary process by which crustal material is recycled into the mantle. Vast volumes of crustal material are recycled and reorganized through erosion and subsequent sedimentary deposition. LITHOPROBE has some outstanding examples of these processes.

☺ Results along Corridor 1 (Slave Province and Wopmay orogen) of the SNORCLE Transect provide the most direct evidence to date for the process of lithospheric delamination *(Cook et al., 1999)*. A 725-km-long deep reflection profile, which shows reflections dipping from within the crust to 80-100 km depth in the mantle, provides clear evidence for tectonic wedging and whole lithosphere delamination. Furthermore, the data provide the first observational evidence for subcrustal imbricate structures, as proposed for the Kaapvaal craton of South Africa on the basis of mantle xenolith studies *(Helmstaedt and Schulze, 1986)*.
☺ Off Canada's southwestern coast, plate tectonic studies indicate that more than 13,000 linear kilometers of oceanic crust have been consumed during the past 180 Ma as various plates interacted with westward moving North America *(Engebretson et al., 1992)*. However, only 500 linear kilometers of new continental crust have formed and even this crust is only partially composed of the accreted material. This requires an immense volume of recycling of oceanic crustal material into the mantle.
☺ Deep crustal seismic reflection data acquired along ~1900 km of profiles in 2000-01 in the northern Canadian Cordillera (SNORCLE Transect) reveal a tapering wedge of layered rocks that is interpreted to be an ancient sedimentary wedge *(Snyder et al., 2002)*. The wedge represents a vast volume of Proterozoic strata of largely North American affinity along the ancient margin. It compares with large sediment fans along continental margins at the mouths of rivers such as the Indus, Ganges and Amazon. The Proterozoic strata, deposited in at least three distinct periods between 1850 and 540 Ma, form a reflective tectono-depositional prism or wedge that has a volume greater than a million cubic kilometers, extends over 1000 km in length, and makes up most of the crust of the northern Canadian Cordillera.
☺ The 1100 Ma Keweenawan rift is an arcuate structure that extends from Kansas through Lake Superior to central Michigan. Studies from the LITHOPROBE GLIMPCE Transect show that this rift almost split Laurentia. In so doing, large volumes of mafic rocks, intercalated with sediments eroded from the surrounding geological features, were extruded; the package is up to 20 km thick *(Cannon et al., 1989)*. Subsequent to the end of rifting, another 10 km of post-rift sediments were deposited.

Internet contact

LITHOPROBE Secretariat e-mail: litho@lithoprobe.ubc.ca
LITHOPROBE Secretariat web site: www.geop.ubc.ca/Lithoprobe/
LITHOPROBE Seismic Processing Facility web site: www.litho.ucalgary.ca/

References

Calvert, A.J. and Ludden, J.N., 1999. Archean continental assembly in the southeastern Superior Province of Canada, *Tectonics*, 18, 412-429.

Cannon, W.F. and 11 others, 1989. The Midcontinent rift beneath Lake Superior from GLIMPCE seismic reflection profiling, *Tectonics*, 8, 305-332.

Clowes, R.M., Cook, F.A., Hajnal, Z., Hall, J., Lewry, J., Lucas, S. and Wardle, R., 1999. Canada's LITHOPROBE project (Collaborative, multidisciplinary geoscience research leads to new understanding on continental evolution), *Episodes*, 22, 3-20.

Cook, F.A., van der Velden, A.J., Hall, K.W. and Roberts, B.J., 1999. Frozen subduction in Canada's Northwest Territories: LITHOPROBE deep lithospheric reflection profiling of the western Canadian Shield. *Tectonics*, 18, 1-24.

Dunphy, J. and Skulski, T., 1996. Petrological zonation across the de Pas batholith: a tilted section through a continental arc? In ECSOOT Workshop 1996, *The University of British Columbia, LITHOPROBE Secretariat, Report 57*, p 44-58.

Engebretson, D.C., Kelley, K.P., Cashman, H.J. and Richards, M.A., 1992. 180 million years of subduction, *GSA Today*, 2, 94-95 and 100.

Fuis, G.S. and Plafker, G., 1991. Evolution of deep structure along the Trans-Alaska Crustal Transect, Chugach Mountains and Copper River basin, southern Alaska. *Journal of Geophysical Research*, 96, 4229-4253.

Gorman, A.R., and 12 others, 2002. Deep Probe: Imaging the roots of western North America. *Canadian Journal of Earth Sciences*, 39, 375-398.

Hall, J., Marillier, F. and Dehler, S., 1998. Geophysical studies of the structure of the Appalachian orogen in the Atlantic borderlands of Canada. *Canadian Journal of Earth Sciences*, 35, 1205-1221.

Helmstaedt, H.H. and Schulze, D., 1986. Southern Africa kimberlites and their mantle sample: Implications for Archean tectonics and lithosphere evolution, *in* Ross, J., ed., Kimberlites and related rocks, *Geological Society of Australia, Special Publication 14*, 358-368.

Lewry, J.F., Hajnal, Z., Green, A., Lucas, S.B., White, D., Stauffer, M.R., Ashton, K.E., Weber, W. and Clowes, R., 1994. Structure of a paleoproterozoic continent-continent collision zone: a LITHOPROBE seismic reflection profile across the Trans-Hudson Orogen, Canada. *Tectonophysics*, 232, 143-160.

Lucas, S.B., Stern, R.A., Syme, E.C., Reilly, B.A. and Thomas, D.J., 1996. Intraoceanic tectonics and the development of continental crust: 1.92 - 1.84 Ga evolution of the Flin Flon Belt, Canada. *Geological Society of America Bulletin*, 108, 602-629.

Percival, J.A., Lucas, S.B., Jones, A.G., Beaumont, C., Eaton, D. and Rivers, T., 1999. How will LITHOPROBE's final chapter be written? *Geoscience Canada*, 26, 17-26.

Snyder, D.B., Clowes, R.M., Cook, F.A., Erdmer, P., Evenchick, C.A., van der Velden, A.J. and Hall, K.W., 2002. Proterozoic prism arrests suspect terranes: Insights into the ancient Cordilleran margin from seismic reflection data. *GSA Today*, 12 (no. 10), 4-10.

White, D.J., Musacchio, G., Helmstaedt, H.H., Harrap, R.M., Thurston, P.C., van der Velden, A. and Hall, K., 2002. Images of a lower-crustal oceanic slab: Direct evidence for tectonic accretion in the Archean Western Superior Province, *Geology*, in review.

Crustal Scale Shear Zones in the Fennoscandian shield

Carl Ehlers

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

The Fennoscandian shield is composed of a number of distinct blocks separated by ductile high strain zones. Strain localization results in ductile shear zones bounded by deformation gradients. The character of the shear zone is dependent upon the nature of the protolith rocks and the temperature during deformation. Most major shear zones show signs of a prolonged activity starting with a wide ductile phase with a tight gneissose structure, with later strain partitioning into narrow mylonites and later pseudotachylites. The walls of the early shear zones are not rigid but represent domains of lower strain that can be e.g. folded, and the shear episodes could be correlated to coeval episodes of folding and intrusion. The high strain zones often wrap around lower strain domains in an anastomosing structure.

Many of the major shears in central Finland and Sweden seem to be active (or activated) around 1.8 Ga when the formation of the Svecofennian crust was more or less completed and the style of deformation shifted from horizontal thrusting and imbrication to a system of large-scale steep shears.

A Slab Breakoff Model for the Differentiation of the Svecofennian Crust in Southern Finland

Olav Eklund and Alexey Shebanov

Department of Geology, FIN-20014 University of Turku, Finland
E-mail: olav.eklund@utu.fi

The accretionary arc complex of southern Finland has been object for several differentiation events during the post-collisional stage of the crustal evolution. One major question about the evolution of the Svecofennian crust in southern Finland is whether the heat that generated the amphibolite – granulite facies rocks with temporally intruded crustal melts (the 1.83 Ga microcline granites) was caused by crustal stacking or by mafic underplating. Another question is whether the A-type granitic melts were produced solely during the anorogenic event, or do they represent a prolonged post-collisional period? We suggest that a slab breakoff event starting approximately 1830 Ma triggered the hot asthenosphere to rise. This asthenospheric rise caused partial melting of the lithospheric mantle and granulite metamorphism in the crust. The melting of the lithosphere and invasion of shoshonitic melts into the crust took place not only during the slab breakoff event, but rather as a prolonged post-collisional event that has been traced from the Åva intrusion in SW Finland and from mineral inclusions in ovoids in the anorogenic rapakivi granites.

Keywords: Granulite metamorphism, shoshonites, lithosphere, asthenosphere, slab breakoff, Fennoscandia

1. Introduction

The Late Svecofennian Granite – Migmatite Zone (LSGM) in the accretionary arc complex of southern Finland extends from Lake Ladoga to the Åland Islands. The zone is metamorphosed principally in upper amphibolite to high-T low-P granulite facies. The granulite metamorphism reached 800°C and 4-6 kbar *(Väisänen and Hölttä, 1999; Korsman et al., 1984).* According to *Väisänen et al., (2002)* the granulite facies metamorphism in SW Finland took place 1824 Ma ago, and crustal melting continued until approx. 1.81 Ga. In southeastern Finland, the granulite metamorphism was dated to 1830-1810 Ma ago *(Korsman et al., 1984).*

The LSGM was punctuated by at least 14 shoshonitic intrusions between 1815 and 1770 Ma *(Eklund et al., 1998; Väisänen et al., 2000).* This intrusive event centre close to 1800 Ma. The shoshonitic intrusions comprise ultramafic, calc-alkaline, apatite-rich potassium lamprophyres to peraluminous HiBaSr granites with $K_2O+Na_2O>5\%$, $K_2O/Na_2O>0.5$, $Al_2O_3>9$ over a wide spectrum of SiO_2 (32-78%). Characteristic for the shoshonitic rocks are their enrichment P_2O_5, F, LILE and LREE *(Eklund et al., 1998).* The εNd and iSr values for the shoshonites are similar for the mafic and the felsic end-members (SW Finland εNd = - 0.6 to +0.8, iSr = 0.7024 to 0.7032, Russian Karelia εNd= -0.8 to –0.3, iSr=0.7033 to 0.7035) *(Eklund et al., in prep.).*

In SW Finland, the early Turku shoshonites were emplaced at 4.5 kbar 1815 Ma ago *(Väisänen et al., 2000),* while the later shoshonites (like the Åva ring intrusion) was emplaced in the upper crust (less than 2 kbar) 1800 Ma ago. In SE Finland the granulite metamorphism in the Sulkava thermal dome culminated at 1830-1810 Ma ago. The minimum regional pressure determined from that area was measured to 4.5 kbar *(Korsman et al., 1984).* At 1795 Ma, the area was cut by a granite dyke that caused a contact aureole indicating a pressure of 2.5 kbar *(Niiranen, 2000).*

It is therefore concluded that the later shoshonites intruded the upper crust (roughly 2-2.5 kbar) some 30 Ma (maximum) after the metamorphic culmination (roughly 4.5-6 kbar) in the LSGM, i.e. the Svecofennian crust was exhumed some 5.6 km (minimum) between 1830 and 1800 Ma.

The situation of high-T, low-P granulite metamorphism preceding a shoshonitic magmatic event is somewhat incongruous. The P-T conditions for granulite facies are only rarely achieved by thickening following collision. Several authors have suggested that to realise the required temperatures (700 to >1000°C), especially at shallow crustal levels, either massive mantle magmas invasion into the lower continental crust is required, or the lithospheric mantle must delaminate to be replaced by hot asthenosphere. In short, granulite facies metamorphism of continental crust requires the involvement of hot asthenospheric mantle *(Thompson, 1997).* The geochemistry of the mafic shoshonites evidence that they stem from an enriched lithospheric mantle source *(Turner et al. 1996; Eklund et al 1998; Väisänen et al. 2000, Mahéo et al. 2002)*, most probably from metasomatic layers at debts around 80 km. Such sources have a solidus temperature considerably lower (900°C) compared to the peridotite dry solidus (1300°C) (McKenzie 1989). The contradiction between granulite metamorphism and the invasion of shoshonitic magmas is that the crust must have been heated by the asthenosphere prior or semi-simultaneously with the intrusion of melts from an enriched lithosphere. We thus need a model where the asthenosphere first is close to the crust over a long but narrow area and is able to cause high-T low-P metamorphism, and subsequently generate enriched low-solidus melts in a subduction enriched lithospheric mantle.

Concerning the post-collisional evolution of the accretionary arc complex of southern Finland, *Korja et al., (1993)* suggested that mafic underplating was the additional heat source required to cause the granulite facies metamorphism after delamination of the overthickened lithosphere. *Nironen, (1997)* suggested that the migmatization and emplacement of the anatectic granites in southern Finland was the result of an extensional collapse. *Väisänen et al., (2000)* correlated the 1815 Ma shoshonites from the Turku area with the regional high grade metamorphism such that the convective hot asthenospheric mantle removed parts of the lithospheric mantle and triggered partial melting of enriched parts in it. The uprising mafic melts caused the granulite metamorphism and generation of crustal melts in the already hot middle crust. *Eklund et al., (1998)* stated that the volume of shoshonitic material is difficult to estimate, but its influx into the crust can be traced over extensive areas in the shield. *Shebanov et al., (2000)* reported that micas typical for shoshonitic rocks were found in the core zones of ovoids in the anorogenic (1573 Ma) Vehmaa rapakivi batholith.

2. The slab breakoff model

When the previously mentioned models are reviewed with new geological data, we suggest a model that may explain the heat source problem for the post-collisional evolution, a slab breakoff. Our model is based on geological features recognized in southern Finland and Russian Karelia. These features are compared with those features that *Davies and von Blanckenburg (1995)* and *von Blanckenburg and Davies (1996)* suggested as indicative for a slab breakoff situation:

Features indicative for slab breakoff:

1) Coeval basaltic (lamprophyric) and granitic magmatism
2) Pronounced age maxima
3) Linear belt of small volume melts forming separate intrusions
4) Exhumation of high-P continental slices
5) Thermal metamorphism
6) Crustal extension
7) Basin with erosion products
8) Late high level shoshonites

Features in southern Finland and Russian Karelia:

1) Early Shoshonites mingled with crustal granites at 1815 Ma *(Väisänen et al., 2000)*

2) 1830-1810 Ma seems to be the age of the LSGM
3) The LSGM is a narrow thermal zone and the shoshonitic intrusions are limited to this zone
4) No high-P rocks are found in southern Finland
5) Upper amphibolite to granulite facies metamorphism in narrow areas.
6) The crust was invaded by granitic, lamprophyric and pegmatite dykes at 1800
7) Uplift noted from SW Finland between 1815 and 1800 Ma, and from Sulkava area in SE Finland between 1830 and 1800 Ma
8) High-level shoshonites crosscutting the amphibolite-granulite area is found from Ladoga Lake to the Åland islands.

Explanation of the features by the slab breakoff concept:

1) Partial melting of lithospheric mantle and crust over hot asthenosphere
2) Singular breakoff event, contrasts with a continuous subduction event
3) Heat source follows track of breakoff, and is limited in space and intensity
4) Release of buoyant crustal slivers at mantle depths
5) Heat advection on (1) rapid exhumation following release of deep mantle root, (2) by melts and conduction on mantle heat
6) Mechanical and isostatic re-equilibration (collapse) of thickened crustal edifice on release of deep mantle roof.
7) Rapid uplift on loss of mantle root leads to high erosion rates.
8) Continued heating of the mantle lithosphere after the uplift

Figure 1 represent a rough cartoon how this slab breakoff may have taken place. Note that the cartoon is out of scale, and works rather as a source for inspiration than the absolute truth.

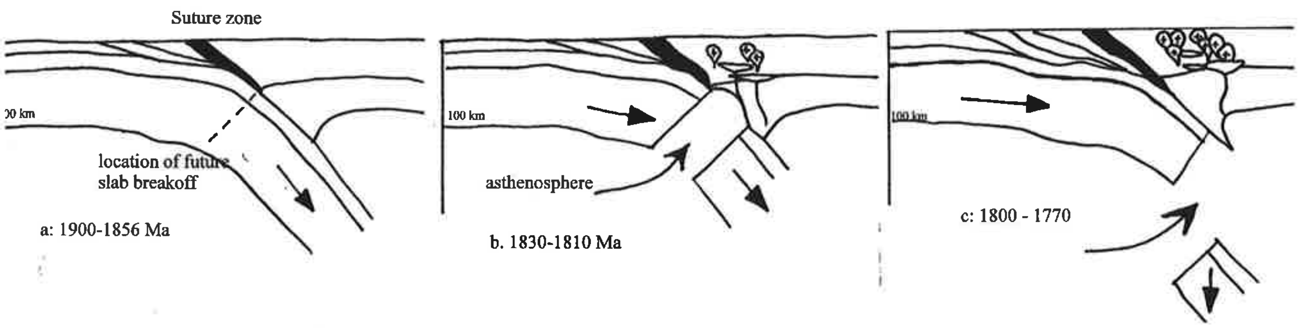


Figure 1a represents the syn-collisional stage of the Svecofennian orogeny in southern Finland. Syn-collisional granitoids intruded as late as 1.86 Ga *(Väisänen et al., 2002).* Fig. 1b. The asthenosphere rise in the gap formed by slab breakoff and melts parts of the lithospheric mantle and the crust. Fig. 1c As the detached lithosphere continues to descend, the uprising hot asthenosphere triggers melting of the metasomatized mantle forming late shoshonitic melts that intruded the upper crust. Before lithostatic equilibrium was reached, several melting events followed in the enriched lithospheric mantle. This may be the reason why the crust was invaded by shoshonitic melts over a long period. Modified after *Mahéo et al (2002).*

3. The Åva and Järppilä case studies

The bimodal shoshonitic lamprophyric-HiBaSr granite Åva intrusion in SW Finland evidence a complex history based on detailed thermobarometric studies combined with SIMS-data. The granite was formed in a deep-seated magma chamber (5-7 kbar) at 1.8 Ga. This magma chamber was reactivated, mingled and brought to the upper crust by mafic shoshonites at 1.77 Ga. The whole area was crosscut by lamprophyres most probably to 1660-1680 Ma.

Järppilä is an ovoid-bearing dyke representing a late intrusion phase in the Vehmaa rapakivi batholith. Detailed studies of the core zones of the ovoids evidence that they are significantly older than the intrusion event (1620 Ma and 1573 Ma respectively). The inherited mineral assemblage of the ovoids comprise Ba-Ti micas with high Mg#, indicative for post-orogenic granites and not for anorogenic *(Bonin, 1986)*. The iSr for the inherited mineral assemblage is 0.7018 and εNd +2, which indicate a juvenile, initially mantle derived component encapsulated in the ovoids *(Shebanov et al., 2000)*.

The data from Åva and Järppilä indicate that the post-collisional period may represent a continuous post-collisional collapse suggested by *Korja and Heikkinen (1995)* where the post-collisional extension caused thinning of the upper crust with simultaneous bow up of the lower crust and mantle upwelling.

4. Conclusions

Slab breakoff is suggested as the motor for the high-T low-P metamorphism, the rapid uplift between 1815 and 1800 Ma and the coeval and subsequent intrusion of early and late shoshonitic magmas in the accretionary arc complex of southern Finland. The asthenospheric upraise was focused along the narrow east-west trending LSGM, where the granulites as well as the high-level shoshonitic intrusions are exposed. However it seems that the post-collisional thermal event continued for a long period after the slab breakoff when the hot asthenosphere heated the lithosphere, since shoshonites intruded the upper crust probably as late as 1.68 Ga, and typical shoshonitic mineral assemblages has been found from the core zones of anorogenic rapakivi ovoids.

To verify this model, more data has to be collected from the western continuation of LSGM in Sweden.

References

Bonin, B. 1990. From orogenic to anorogenic settings: evolution of granitoid suites after a major orogenesis. *Geological Journal* ***25****, 261-270.*

Davies, J.H. and von Blanckenburg, F. 1995. Slab breakoff: A model of lithosphere detachment and its test in the magmatism and deformation of collisional orogens. *Earth and Planetary Science Letters* ***129****, 85-102.*

Eklund, O., Konopelko, D., Rutanen, H., Fröjdö, S. and Shebanov, A.D. (1998) 1.8 Ga Svecofennian Postorogenic Shoshonitic Magmatism in the Fennoscandian Shield. *Lithos* ***45****, 87-108.*

McKenzie, D.P., 1989. Some remarks on the movement of small melt fractions in the mantle. *Earth and Planetary Science Letter* ***95****, 53-72.*

Mahéo, G., Guillot, S., Blichert-Toft, J., Rolland, Y. and Pêcher, A. 2002. A slab breakoff model for the Neogene thermal evolution of South Karakorum and south Tibet. *Earth and Planetary Science Letters* ***195****, 45-58.*

Niiranen, T. 2000. Svekofennisen orogenian jälkeinen ekshumaatio ja isostaattinen tasapainottuminen kaakkois-Suomessa. *M.Sc. Thesis Dept. Geology, University of Turku. 70 pp.*

Nironen, M. 1997. The Svecofennian orogen: a tectonic model. *Precambrian Research* ***86****, 21-44.*

Korja, A. and Heikkinen, P. 1995. Proterozoic extensional tectonics of the central Fennoscandian shield: results from the Baltic and Bothnian echoes from the lithospheric experiment. *Tectonics* ***14****, 504-517.*

Korja, A., Korja, T., Luosto, U. and Heikkinen, P. 1993. Seismic and geoelectric evidence for collisional and extensional events in the Fennoscandian Shield – Implications for Precambrian crustal evolution, *Tectonophysics* ***219****, 129-152.*

Korsman, K., Hölttä, P., Hautala, T. and Wasenius, P. 1984. Metamorphism as an indicator of evolution and Structure of the crust in Eastern Finland. *Geological Survey of Finland, Bulletin* ***328****, 40 pp.*

Shebanov, A., Savatenkov, V. Eklund, O. Andersson, U.B., Annersten, H. and Claesson, S. (2000). Regional mineralogical correlation linking post- and anorogenic magmatic events from unusual barian biotites in the Järppilä rapakivi porphyries, Vehmaa batholith (S. Finland). Meeting: *Advances on Micas Rome 2-3.11 2000. Universitá degli Studi Roma Tre MURST – Progetto Fillosilicati: aspetti cristallochimici strutturali e petrologici. 183 – 186.*

Thompson, A.B. (1997) Granulite Facies Processes and Protoliths. In*: A.K. Gupta & R. Kerrich (eds.) The Dynamic Geosphere, Prof. William S. Fyfe, Felicitation Volume, Allied Publishers Ltd., New Delhi 2-12.*

Turner, S., Arnaud, N., Liu, J., Rogers, N., Hawkesworth, C., Harris, N., Kelley, S., van Calsteren, P. and Deng, W., 1996. Post-collision, Shoshonitic Volcanism on the Tibetan, Plateau: Implications for convective thinning of the lithosphere and source of ocean island basalts. *Journal of Petrology* ***37**, 45-71*

Väisänen, M. and Hölttä, P., 1999. Structural and metamorphic evolution of the Turku migmatite complex, southwestern Finland. *Bulletin of the Geological Society of Finland, **71**, 177-218.*

Väisänen, M., Mänttäri, I., Kriegsman, L.M., Hölttö, P. (2000) Tectonic setting of post-collisional magmatism in the paleoproterozoic Svecofennian orogen, SW Finland. *Lithos **54**, 63-81.*

Väisänen, M., Mänttäri, I. and Hölttä, P. 2002. Svecofennian magmatic and metamorphic evolution in southwestern Finland as revealed by U-Pb zircon SIMS geochronology. *Precambrian Research **116**, 111-127.*

von Blanckenburg, F. and Davies, J.H. (1996) Feasibility of double slab breakoff (Cretaceous and Tertiary) during Alpine convergence. *Eclogae Geol. Helv. **89**, 111-127.*

SVEKALAPKO Project: A Contribution to an Improved Understanding of the Evolution of the Fennoscandian Shield

Sven-Erik Hjelt[1] and J. Stephen Daly[2]

[1]Department of Geophysics, P.O. Box 3000, FIN-90014 University of Oulu, Finland
[2]Department of Geology, University College Dublin, Dublin, Ireland
E-mail: sven-erik.hjelt@oulu.fi

SVEKALAPKO-project, one of the five key projects of the ESF-supported EUROPROBE programme, has been operative during the years 1995 - 2001. Much of the research is still continuing.

Scientists from about 70 institutes in 19 European countries have been working together in order to gain new understanding of the Fennoscandian lithosphere, its thickness, structure and development. Some of the highlights of the 11 subprojects of SVEKALAPKO will be described in this presentation. Many of the subprojects are continuing their research after the formal end of the project. Especially integrated models will require many person-years of further research.

Keywords: Fennoscandian Shield, tectonics, crust, upper mantle, joint inversion, seismic tomography, electromagnetic deep sounding, geothermal research, potential fields, geochemistry, age determination

1. Introduction

The key issue of the SVEKALAPKO-project has been to study the nature of a lithosphere with 60km thick crust. Three key experiments were planned (*Hjelt and Daly, 1996*), a seismic tomography array study, the deep electromagnetic BEAR array study and reflection profiling. The field campaigns of the two first ones were successfully completed in 1998-99 and the analysis of the results is still going on. One line (4B) of the planned reflection programme was realized, but as a spin-off of the SVEKALAPKO work, the Russian-Finnish FIRE-project started in 2001 and will cover about 1700 km of lines across key geological structures in the Finnish part of the Fennoscandian Shield. The seismic tomography, BEAR and FIRE projects will be discussed more in detail in other papers of the LITO2002 symposium (*Kozlovskaja et al, 2002; T Korja et al, 2002 and Kukkonen et al, 2002*).

Some of the 11 subprojects of SVEKALAPKO were introduced to allow for interchange of ideas and results with other, simultaneous and related projects. Of greatest importance were the GGT-SVEKA project (*Korsman et al, 1999*) and projects dealing with metallogenic problems and production of joint geoscientific maps. [see the project website at http://babel.oulu.fi/Svekalap.html].

The annual level of external funding for SVEKALAPKO subprojects varied between 350 and 450 kiloeuros with one top year (1999) with up to 624 kiloeuros. The level of budgetary contributions to the whole projects were difficult to estimate, but it ranged from almost 1 Million euros during the first years to around 550 kiloeuros during the later years (*Daly and Hjelt, reports to EMC*).

The participants disseminated their results regularly at scientific meetings, like EUG, (EUG 10, Strasbourg, April 1999; EUG XI, April 2001, "Major boundaries in the lithosphere" symposium), EGS, IGC, numerous national and the ESC GAC-MAC and Goldschmidt meetings. The project arranged annual project workshops, where 50 - 80 papers were presented each year. (*Daly and Bogdanova, 2001; Hjelt, 1996, 1997, 1999, 2000, 2001, Philippov, 1998*).

The number of publications reached and exceeded 50 during the later years of the project. The number of abstracts were in excess of 150 annually. One special volume, "Alkaline Magmatism and Xenoliths of the Baltic Shield" Lithos Special Issue was published 2000.

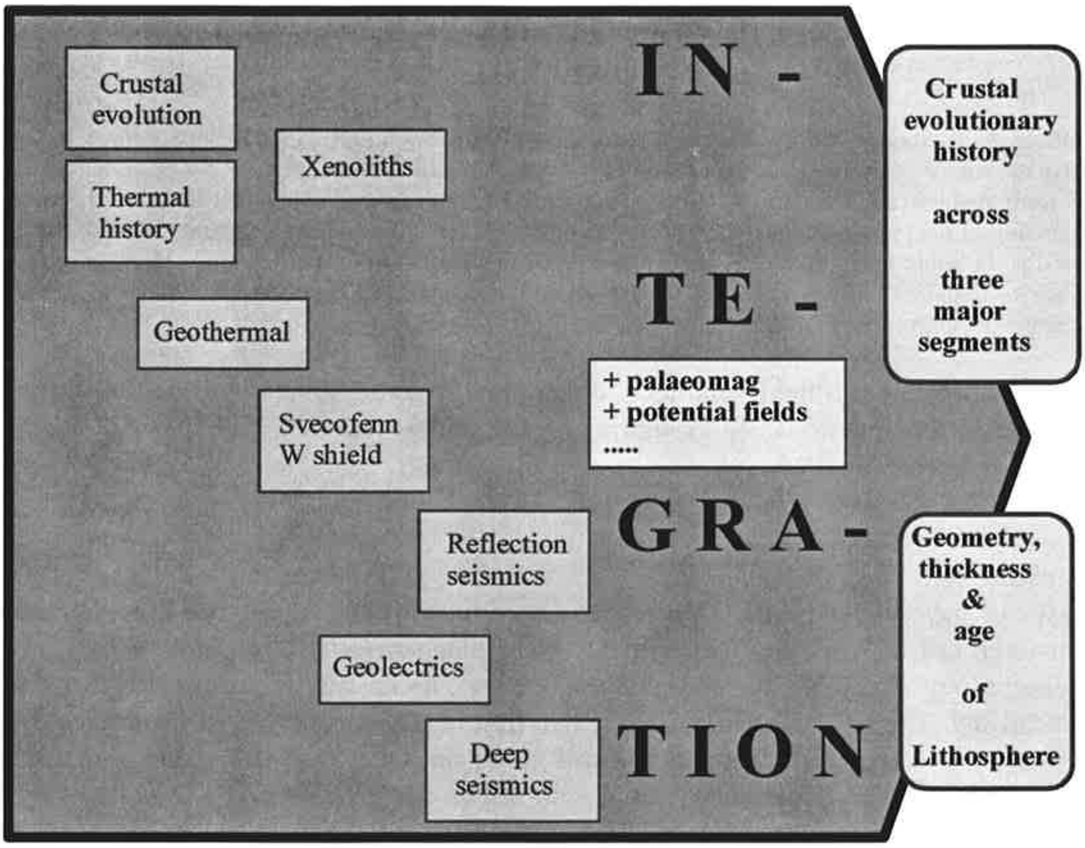


Figure 1. Major contents and aims of SVEKALAPKO research.

2. The seismic tomography experiment

The SVEKALAPKO deep seismic tomography experiment, where the field work was realized in 1998-1999, produced a large amount of detailed information about the deep structure of the crust and upper mantle, the processing and the analysis of which is still in progress.

The first preliminary results seem to indicate differences in the subcrustal anisotropic properties, a depth extension of a high-velocity keel below the central Fennoscandian Shield and increasing difference between the 410 and 660 km regions from TOR to SVEKALAPKO areas. The upper mantle P-wave velocities differ a few % from the standard reference model IASP91. Crustal velocity models are continuously improved in order to increase the accuracy of upper mantle models below the SVEKALAPKO seismic tomography array. (*Alinaghi et al, 2002; Bock et al., 2001; Hyvönen and Malaska, 2000; Kozlovskaya et al., 2002a, b; Plomerova et al, 2002; Sandoval et al., 2002).*

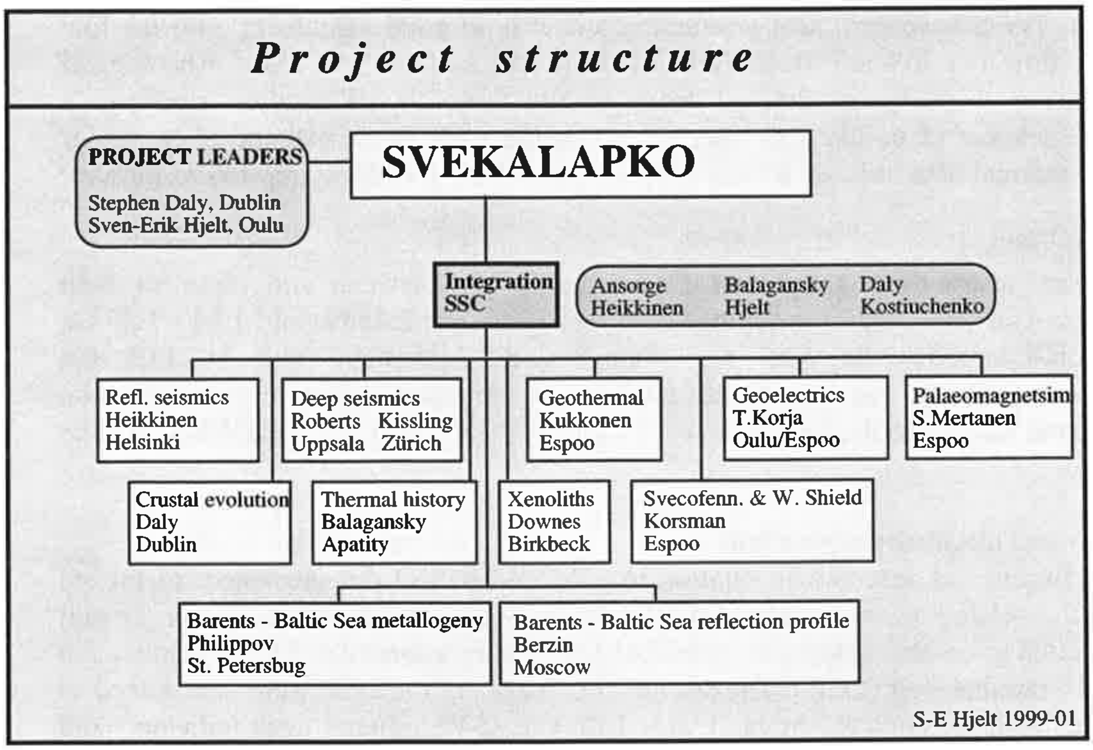


Figure 2. The organization chart of SVEKALAPKO and its subprojects.

3. The BEAR electromagnetic experiment

The Electromagnetic Array Research (BEAR) project has investigated the deep and subcrustal structure of the whole Fennoscandian Shield using the largest ever long period electromagnetic array. The preliminary results confirm the existence of an upper mantle conductor at the depth of 70 - 120 km in the Karelian part of the Shield. A deeper conducting layer at 250 - 350 km in mantle exists throughout the Shield. Below 400 km the conductivity approaches the "global curve". (*T Korja et al, 2002*)

4. Reflection seismic studies

The E-W running CDP-profile Uchta - Kem (White Sea) [SVEKALAPKO line 4B] comprises several inclined boundaries from the surface down to 25-30 km. They correlate with fault zones between the Belomorian Mobile Belt and the Karelian craton. Near-horizontal crustal boundaries and the Moho are interpreted as transitional high velocity gradients, the upper mantle being dominantly transparent (*Berzin et al, 2002*). Reflection geometry of line 4B supports geochemical and structural evidence for Neoarchaean collision of an oceanic plateau with the continental margin. Additional seismic interpretation reveals Archaean thrust-nappe structures in the Central Kola Terrane.

5. Geothermal studies

New measurements, reinterpretation of old borehole heat flow (HF) data revealed major palaeoclimatic disturbances. Thermal effects of Pleistocene glaciation can be seen down to 2 km depth from Scandinavia to the Urals.

The corrected data indicate a very low overall heat flow density: 15-20 mWm^{-2} and a low thermal gradient in the Central Kola Peninsula - 5° C/km. Heat production data for the

Shield were compiled and evaluated with respect to rock age, composition and petrophysics. The lithospheric heat production model is in good agreement with the low mantle heat flow (11 mWm^{-2} from xenolith data) and surface heat flow. Rheological modelling using a geotherm derived from mantle xenolith PT studies confirms a lithosphere thickness of c. 250 km. There is no partial melt in the asthenosphere under Finland; the thermal data indicate a 'wet' mantle. (from *Daly and Hjelt, reports to EMC*).

6. Palaeomagnetism

Continental reconstructions show that Fennoscandia lay between and close to both Laurentia and Amazonia, major groupings being recognized at 2.45 Ga and 1.88 - 1.25 Ga (Fennoscandia-Laurentia), at 1.65 Ga (Fennoscandia-Amazonia) and at 1.05 Ga (formation of Rodinia). No significant rotation or latitudinal plate motion of Kola, Belomorian and Karelia took place between 2.4 and 2.3 Ga. (from *Daly and Hjelt, reports to EMC*).

7. Xenoliths and alkaline magmatism

Among the findings of research in relation to SVEKALAPKO the subproject identified new xenoliths related to an unexposed carbonatite complex at depth, lower crustal xenoliths related to basaltic underplating c. 2.45 Ga, metamorphosed c. 1.8 Ga eclogites in the Lapland Granulite Belt (LGB). The detrital zircon ages of the LGB were determined as 2.8 - 2.0 Ga, with metamorphism ca. 1.91 - 1.87 Ga. U-Pb mineral ages indicate rapid exhumation of deep crust.

Carbonatite magmatism of the Kola Peninsula dates at 378 Ma (U-Pb). "Devonian" magmatism starts in Silurian, coeval with Caledonide compression. Volcanism is identified 20-30 Ma before intrusions. The Devonian mantle comprises of two-components, a deeper, enriched one and a shallower depleted lithosphere. (from *Daly and Hjelt, reports to EMC*).

8. Metallogenesis

The subproject has produced among others new genetic 3D models for the Pechenga and Monchegorsk ore regions, studied the correlation and geodynamic interpretation of magmatic complexes, produced a Kola-Karelia 1:1.000.000 digital metallogenic map and made Au and PGE investigations of Kola Peninsula carbonatites, U-Pb dating and geochemistry in North Ladoga region as well as sulphur isotopic studies in Lapland-Kola rift rocks revealing "black smoker" type exhalative centres. (from *Daly and Hjelt, reports to EMC*).

9. Crustal evolution and tectonics

In the SVEKALAPKO project geological field-work, geochemistry and geochronological studies were more or less focussed on the Kola Peninsula and the Karelian part of the Fennoscandian Shield. The recently discovered Neoarchaean Iringora ophiolite in the North Karelian greenstone belt (NKGB) and investigations in Khizovaara suggest that subduction-related processes have not changed substantially over the past 2.8 Ga. A modern subduction-style tectonics has operated - at least locally - in the Neoarchaean.

Detailed studies on the NKGB have revealed at least two major periods of crustal stacking and thrusting. A Neoarchaean thrust/nappe tectonics episode about 2.8 Ga ago was followed by accretion leading to Barrovian metamorphism (at 650 °C, 7 kbar). The assemblages were overprinted during the Svecofennian orogeny at 1.9 - 1.75 Ga (at 600 °C, 6 kbar).

High-quality U-Pb age determinations (both using conventional and ion microprobe techniques) have revealed two Archaean granulite-facies metamorphism episodes within the Iisalmi region: at 3.2 and 2.7 Ga. Early magmatism in the NKGB has been dated at 2.88 - 2.83 Ga.

New good quality field, geochemical and geochronological data from the Lapland-Kola orogen allow a re-evaluation of the Baltica-Laurentia correlation. Between 2.5 Ga and ca 1.8 Ga the orogenic evolution is consistent with that of correlative regions in Scotland, Greenland and North Atlantic. Rifting in all North Atlantic areas occurred in the time interval 2.5 - 2.0 Ga. This was followed by oceanic separation for not longer than about 100 Ma. The collisional event peaked at ca. 1.9 Ga and the post-collisional events continued until ca. 1.7 Ga.

Several geochemical, geochronological and petrological studies, aimed originally at investigating earlier stages of the crustal history, have lead to unexpected outcomes. There is evidence of large-scale late- or post-tectonic fluid flow in the middle crust in the Lapland-Kola orogen. In the Keivy supracrustal belt cooling to 450 °C at 1927 Ma (rutile U-Pb age) followed regional. Metamorphic rims on zircon grew at 1.72 - 1.70 Ga (ion microprobe U-Pb age). The rim is U-rich and Th-poor indicating simultaneous regional scale low temperature flow. The flow may have continued until the formation of amazonite pegmatites at 1665 - 1682 Ma (zircon U-Pb age).

The muscovites and biotites in the northern part of the Lapland-Kola orogen are too young [1.75 - 1.70 Ga (Ar ages)] to be explained simply by cooling after the 1.9 Ga collision. The possibility of infiltrating fluid of mantle origin is discussed based on recent He isotope studies. (from *Daly and Hjelt, reports to EMC*).

Selected references

Alinaghi, A., Bock, G., Kind, R., Hanka, W., Wylegalla, K, TOR and SVEKALAPKO Working Groups, 2002. Receiver function analysis of the crust and upper mantle from the North German Basin to the Archaean Baltic Shield. *Geophysical Journal International, submitted.*

Berzin, R, Yurov, YuG, and Pavlenkova, N, 2002. CDP and DSS data along the Uchta-Kem profile (the Baltic Shield). *Tectonophysics* **355**, 187 - 200.

Bock, G. (Ed.) and SVEKALAPKO Seismic Tomography Working Group, 2001. Seismic probing of Archaean and Proterozoic Lithosphere in Fennoscandia. *EOS Transactions AGU*, **82**, 50, 621, 628-629.

Daly, JS and Bogdanova, S, 2001. EUROPROBE Precambrian Time-slice symposium. *Abstracts of a joint meeting of the EUROBRIDGE and SVEKALAPKO projects.* VSEGEI, ST.Petersburg, Nov 1-3, 2001.

Daly, JS and Hjelt, S-E. *EUROPROBE - SVEKALAPKO annual reports to the EMC* (unpublished).

Hjelt, S-E, Daly, JS and SVEKALAPKO colleagues, 1996. SVEKALAPKO.Evolution of Palaeo-proterozoic and Archaean Lithosphere. pp. 56- 67 in Gee, DG and Zeyen, HJ (editors): *EUROPROBE 1996 - Lithospheric dynamics: Origin and evolution of continents.* Publ. by the EUROPROBE Secretariate, Uppsala University, 138 p.

Hjelt, S-E. (compiler), 1996. Abstracts for the SVEKALAPKO 1st Workshop, Lammi, Finland, 28.-30.11.1996. Univ of Oulu, 60 pp.

Hjelt, S-E. (compiler), 1997. Abstracts for the SVEKALAPKO 2nd Workshop, Lammi, Finland, 27.-30.11.1997. *Dept. Geophysics, Univ. Oulu, Report* **21**. 77 pp.

Hjelt, S-E (ed), 1999. SVEKALAPKO 4th Workshop, Lammi, Finland, 18.-21.11.1999, *Department of Geophysics, University of Oulu, Report* **No 22**.

Hjelt, S-E (ed): SVEKALAPKO 5th Workshop, Lammi, Finland, 2.-5.11.2000, *Department of Geophysics, University of Oulu, Report* **No 23**.

Hjelt, S-E (ed): SVEKALAPKO 6th Workshop, Lammi, Finland, 29.11.-2.12.2001, *Department of Geophysics, University of Oulu, Report* **No 24**.

Hyvönen, T., Malaska, J. 2000. Velocity modeling of the lithosphere beneath South Finland. *Physics of the Earth and Planetary Interiors*, **122,** 103-114.

Korja, T., Hjelt, S-E., Kaikkonen, P., Kozlovskaja, E., Lahti, I., Pajunpä ä , K., Pulkkinen, A., Viljanen, A. and BEAR Working Group, 2002. Crust and upper mantle beneath Fennoscandia as imaged by the Baltic Electromagnetic Array Research (BEAR). (*this volume*)

Korsman, K, Korja, T, Pajunen, M, Virransalo, P and GGT/SVEKA Working Group, 1999. The GGT/SVEKA Transect: Structure and evolution of the continental crust in the Paleoproterozoic Svecofennian orogen in Finland. *Int. Geol. Review,* **41**, 287 - 333.

Kozlovskaya, E., Yliniemi, J, Hjelt, S.-E., Komminaho, K. and SVEKALAPKO Seismic Tomography Working Group, 2002a. Lithosphere structure beneath southern Finland derived by the SVEKALAPKO seismic array research (*this volume*).

Kozlovskaya, E., Pirttijä rvi, M., Hjelt, S-E, Korhonen, J., Elo, S. and Yliniemi, J., 2002b. The research project: 3-D crustal model of the Finnish part of the SVEKALAPKO research area (*this volume).*

Kukkonen, I, Heikkinen, P. and the FIRE Working Group 2002. FIRE project - first results (*this volume*).

Philippov, N (ed), 1998. *Abstracts. Svekalapko Europrobe Project 3rd Workshop*, 26-29.11.1998, Repino, Russia, Ministry of Natural Resources of Russian Federation.

Plomerova, J., Hyvönen, T., Sandoval, S., Babushka,V and SSTWG, 2002. Structure of Fennoscandian mantle lithosphere from seismic velocity anisotropy. *XXVII General Assembly of EGS, Geophysical Research Abstracts*, 4, EGS02-A-02173.

Sandoval, S., Kissling, E., Ansorge, J. and the SVEKALAPKO STWG, 2002. Three- Dimensional Seismic Crustal Model and Associated Effects on Teleseismic Tomography Beneath Central Fennoscandia (Baltic Shield). *Submitted to J. Geoph. Res.*

The Age of the Lithosphere in the Fennoscandian Shield – Sm-Nd Isotopic Evidence

Hannu Huhma

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

Sm-Nd isotopic method has been used to constrain the crustal residence ages and in particular to study the origin of the 1.9 Ga crust in Finland. The presentation will focus on selected key issues.

Keywords: Sm-Nd isotopes, Fennoscandian Shield

1. Introduction

The formation of continental crust involves major fractionation of elements. Consequently, the Sm-Nd analyses on LREE enriched whole rocks have been successfully used in evaluating the average crustal residence ages (T_{DM}). The general approach assumes that the continental lithosphere ultimately formed from the major convective reservoir, i.e. the depleted mantle (DM), which thus provides the reference. If subsequent metamorphic Sm/Nd fractionation has taken place, the model ages T_{DM} may be strongly biased, and the use of initial ε_{Nd}-values is often more justified.

2. Mafic rocks – mantle evolution

The Sm-Nd isotopic analyses on mafic rocks provide a tool to evaluate the isotopic evolution in the mantle. Archaean mafic rocks have often seriously suffered from the Proterozoic metamorphic effects, and the determination of initial ratios is difficult. Nevertheless, the data available show that time-integrated depleted mantle reservoirs existed during Archaean times. The data on Proterozoic mafic rocks provide a wide range of initial ε-values. Mafic rocks with positive ε-values close to the model depleted mantle composition include e.g. Jouttiaapa basalts in Peräpohja, $\varepsilon(2090) \approx +4.2$ (*Huhma et al, 1990*), Jeesiörova komatiites, $\varepsilon(2060) \approx +4.0$ (*Hanski et al, 2001a*), Vesmajärvi tholeiites in Kittilä, $\varepsilon(2015) \approx +3.8$ (*Hanski et al, 1998; Hanski & Huhma, in review*) and the Svecofennian picrites, $\varepsilon(1900) \approx +3.5$. In contrast, most mafic rocks analysed in Finland provide initial values well below coeval depleted mantle, which is due to the involvement of old LREE enriched lithosphere in their genesis. These include the large 2.44 Ga old layered intrusions, $\varepsilon(2440) \approx -2$ (*Huhma et al, 1990, Hanski et al, 2001b*), the 2.05 Ga old Keivitsa intrusion, $\varepsilon(2050) \approx -3.5$ (*Hanski et al, 1997*), and the Svecofennian Ni-bearing gabbros, $\varepsilon(1900) \approx 0$. The lithospheric contribution can be either from old enriched mantle or crustal contamination.

3. Archaean crust

Sm-Nd crustal residence ages for Archaean granitoids ($^{147}Sm/^{144}Nd < 0.13$) range from 2.73 to 3.6 Ga. Typical late Archaean granitoids give T_{DM} of 2.8-3.0 Ga, and do not show any major contribution of much older crustal material in their genesis. The bulk of the analysed Archaean metasediments also suggest relatively short crustal prehistory. Granitoids with older model ages often yield indications of zircon in excess of 3 Ga.

The Nd isotopic composition of the Archaean crust at 1.9 Ga is of importance when studying the origin of the Proterozoic crust. Typical late Archaean granitoids provide ε(1900)-values close to –10, but values down to –20 have been obtained. Lower crust with more mafic Archaean lithologies should have higher ε.

4. Proterozoic crust

The crustal residence ages for Proterozoic granitoids range from 1.9 to 2.6 Ga. Granitoids in the Karelian domain tend to contain large component of Archaean crustal material (T_{DM} 2.3-2.6 Ga, $\varepsilon(1900) = -1 \ldots -10$), whereas the Svecofennian rocks represent more juvenile 1.9 Ga additions to the crust (T_{DM} 1.9-2.2 Ga, $\varepsilon(1900) = +3 \ldots -1$; e.g. *Huhma, 1986; Patchett & Kouvo, 1986; Rämö et al 2001*). However, the formation of large volumes of Svecofennian 1.9 Ga felsic crust with ε (1900) of about zero (T_{DM} 2.1 Ga) call for significant contribution from older LREE enriched lithosphere in their genesis. The most positive initial ε-values of ca. +3 have been obtained from the 1.92 Ga granitoids and volcanics in the Raahe-Ladoga zone, Central Finland, suggesting short crustal prehistory (*Lahtinen & Huhma, 1997*). In Lapland, similar juvenile ca. 1.9 Ga felsic crust exists in the Utsjoki area. The other geologically important occurrence of juvenile felsic rocks in Lapland is the volumetrically minor 2.015 Ga porphyries in the Vesmajärvi Formation, Kittilä Group. They have similar positive initial $\varepsilon(2015)$ values (+3.8) to those of the tholeiitic mafic metavolcanic rocks, suggesting that no old sialic crust was involved in their generation. 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 ca. 1.92 Ga ago (*Hanski et al., 1998*).

5. Conclusion

The Sm-Nd crustal residence ages for felsic rocks in Finland range from 1.9 to 3.6 Ga.

References

Hanski, E. J.; Huhma, H.; Lehtonen, M. I.; Rastas, P. 1998. 2.0 Ga old oceanic crust in northern Finland. *In: Hanski, E. & Vuollo, J. (eds.) International ophiolite symposium and field excursion : generation and emplacement of ophiolites through time, August 10-15, 1998, Oulu, Finland. Abstracts, excursion guide. Geological Survey of Finland. Special Paper 26, 24.*

Hanski, E. J.; Huhma, H.; Suominen, I. M.; Walker, R. J. 1997. Geochemical and isotopic (Os, Nd) study of the early Proterozoic Keivitsa intrusion and its Cu-Ni deposit, northern Finland. *In: Papunen, H. (ed.) Mineral deposits : research and exploration - where do they meet? Proceedings of the Fourth Biennial SGA Meeting, Turku/Finland/11-13 August 1997. Rotterdam: A. A. Balkema, 435-438.*

Hanski, E., Huhma, H., Rastas, P. & Kamenetsky, V.S., 2001a. The Palaeoproterozoic Komatiite-Picrite Association of Finnish Lapland. *Journal of Petrology, 42, 855-876.*

Hanski, E., Walker, R.J., Huhma, H. and Suominen I., 2001b. The Os and Nd isotopic systematics of c. 2.44 Ga Akanvaara and Koitelainen mafic layered intrusions in northern Finland. *Precambrian Research, 109, 73-102.*

Huhma, Hannu 1986. Sm-Nd, U-Pb and Pb-Pb isotopic evidence for the origin of the Early Proterozoic Svecokarelian crust in Finland. *Geological Survey of Finland. Bulletin 337. 48 p. + 2 app.*

Huhma, Hannu; Cliff, Robert A.; Perttunen, Vesa; Sakko, Matti 1990. Sm-Nd and Pb isotopic study of mafic rocks associated with early Proterozoic continental rifting : the Peräpohja schist belt in northern Finland. *Contributions to Mineralogy and Petrology 104 (3), 367-379.*

Lahtinen, Raimo; Huhma, Hannu 1997. Isotopic and geochemical constraints on the evolution of the 1.93-1.79 Ga Svecofennian crust and mantle in Finland. *Precambrian Research 82 (1-2), 13-34.*

Patchett, J. and Kouvo, O. (1986) Origin of continental crust of 1.9-1.7 Ga age: Nd isotopes and U-Pb zircon ages in the Svecokarelian terrain of south Finland. *Contributions to Mineralogy and Petrology 92, 1-12.*

Rämö, O. T.; Vaasjoki, M.; Mänttäri, I.; Elliott, B. A.; Nironen, M. 2001. Petrogenesis of the post-kinematic magmatism of the Central Finland Granitoid Complex I; Radiogenic isotope constraints and implications for crustal evolution. *Journal of Petrology, 42, 1971-1993.*

Sedimentary Record and Magmatic Episodes Reflecting Mesoproterozoic-Phanerozoic Evolution of the Fennoscandian Shield

Jarmo Kohonen[1] and O. Tapani Rämö[2]

[1]Geological Survey of Finland, P.O. Box 96, FIN-02151 Espoo, Finland
[2]Department of Geology, P.O. Box 64, FIN-00014 University of Helsinki, Finland
E-mail: jarmo.kohonen@gsf.fi, tapani.ramo@helsinki.fi

This contribution reviews the state-of-the-art knowledge regarding the Mesoproterozoic to Phanerozoic geologic record of Finland and surrounding areas. A (speculative) tectonic model for the central part of the Fennoscandian shield is presented, focusing on (1) preserved sedimentary record and other indications of ancient land surface; (2) nature of the igneous rocks; and (3) published isotope and fission track studies. Special emphasis is on the burial-exhumation history of the shield.

Keywords: sedimentary basins, intraplate magmatism, Proterozoic, Phanerozoic, Fennoscandian shield

1. Introduction

Meso- to Neoproterozoic and Paleozoic unmetamorphic (or very low grade) supracrustal rocks are found as local remnants only on the Archean and Paleoproterozoic crystalline basement of the Fennoscandian shield. Igneous events at this time are represented by the ~early Mesoproterozoic rapakivi association, Mesoproterozoic tholeiitic (or transitional) subvolcanic mafic intrusions and dikes, Phanerozoic mafic magmatism associated with Caledonian tectonism, and ~Devonian alkaline magmatism related to the Kola alkaline province. Emplacement of the 1.65–1.54 Ga rapakivi granites and associated mafic intrusions was the latest major crustal increment to the Fennoscandian shield in Finland. Mesoproterozoic (*Jotnian*) sedimentary sequences (arkose, siltstone, shale, conglomerate) are known in some ten localities within the shield. All these occupy shallow basins, tectonic depressions or grabens bordered by fractures or fault zones most of which are NW-SE oriented. Here we will describe the geologic setting of these rocks first and then present a tectonic model concerning the tectonic evolution of the shield from mid-Proterozoic to Present.

2. Supracrustal record

Major Mesoproterozoic sedimentary formations in Finland are **Satakunta** and **Muhos**. Both have marine extensions in the Gulf of Bothnia (*Winterhalter et al, 1981*; Fig. 1). A separate marine occurrence of sandstone is also known in the Åland Sea southwest of the Åland Island. In central Sweden, the continental **Dala** (Dalarna) sandstone is preserved in the core of an open N – S oriented syncline. Other Swedish occurrences include **Nordingrå** and **Gävle** on the western shore of the Bothnian Sea and **Svartälven, Mälaren,** and **Almesåkra** south of Dalarna. In NW Russia, the Mesoproterozoic supracrustal units cluster in the **Lake Ladoga – Salmi** and **White Sea – Tersk** regions and in **northern Kola Peninsula**. Many of these sedimentary formations show a spatial correlation to the rapakivi granites and Mesoproterozoic mafic dikes. In Finland, the 1265 Ma dikes of Satakunta, Åland, and Merenkurkku (*Suominen, 1991*) give the minimum depositional age for these sequences. In the Lake Ladoga region, a sill in the upper part of the basin sequence (*Amantov et al., 1996*) has a U-Pb age of ~1460 Ma (*Rämö et al., 2001*) and indicates that, at least in this region, basin formation started relatively rapidly after the emplacement of the rapakivi plutons (Salmi complex; cf. *Amelin et al., 1997*).

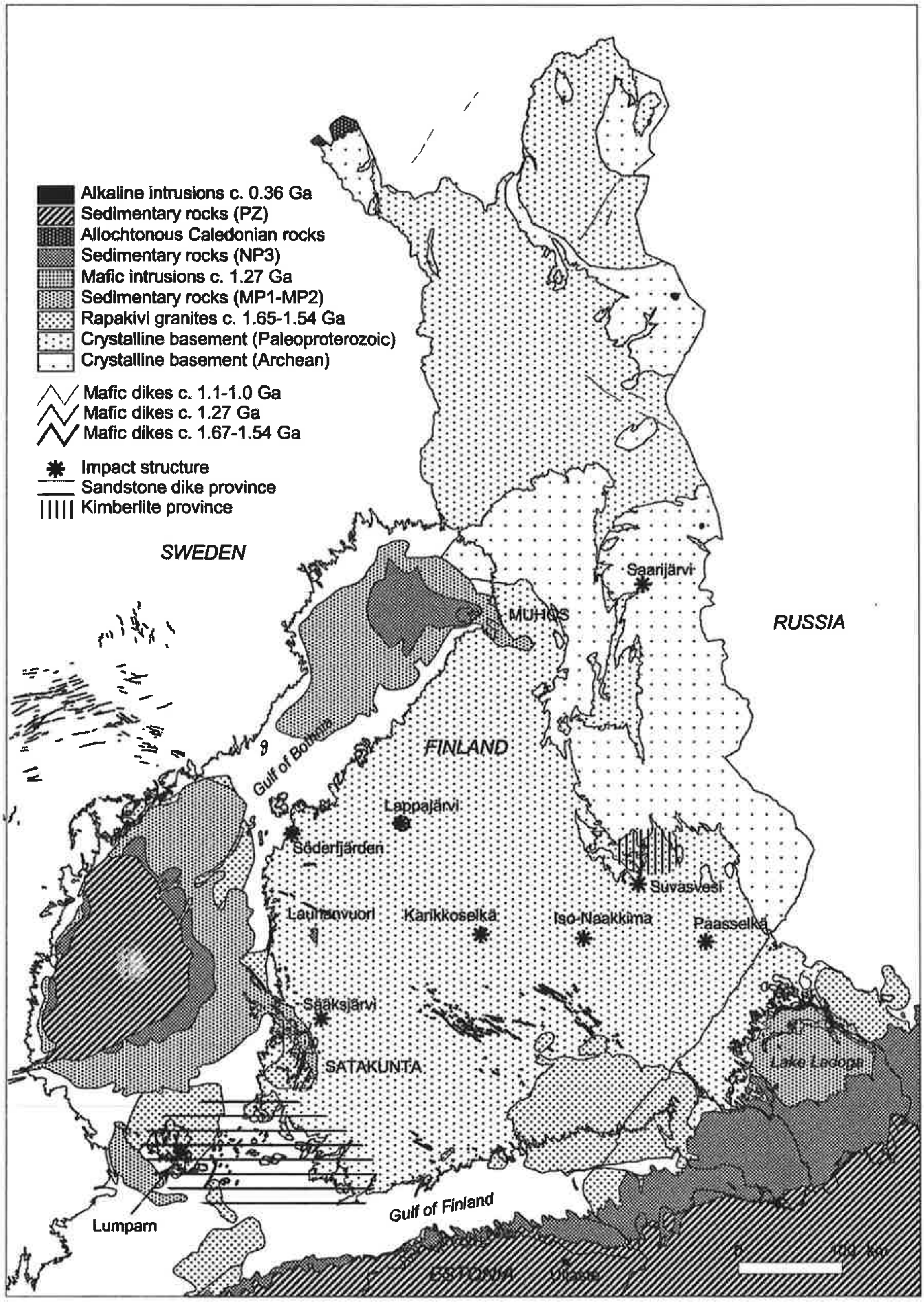


Figure 1. Mesoproterozoic to Phanerozoic geology of Finland and adjacent areas.

Towards the late Mesoproterozoic, Neoproterozoic, and Paleozoic Eras, igneous rocks become sparse. In northern Finland, scattered 1100–1000 Ma mafic dikes are found, ~370 Ma alkaline silicic rocks and carbonatite (*Kramm et al., 1993*) and kimberlites are present in east-central and northern Finland. In the far northwest part of the country, allochtonous units of the Finnish Caledonides include ~435 Ma gabbroic and ultramafic rocks (*Vaasjoki and Sipilä, 2001*).

In **Hailuoto,** the Mesoproterozoic Muhos formation is conformably overlain by the Neoproterozoic (NP3) Hailuoto formation that consists of interbedded sandstone, siltstone, and mudstone (*Veltheim, 1969*; *Tynni and Donner, 1980*). According to seismic data (*Winterhalter et al., 1981*; *Lundqvist et al. 1996*; *Koistinen et al., 2001*), sedimentary deposits comparable to the Muhos and Hailuoto formations cover a large part of the **Bothnian Bay** area. The **Lauhanvuori** sandstone covers an area of ~15 by 5 km (*Simonen and Kouvo, 1955*) with some outliers towards the north. The Lauhanvuori formation has been regarded as Neoproterozoic (NP3) (*Koistinen et al., 1997*), but Cambrian or even younger depositional age can not be excluded. A remnant of Neoproterozoic (NP3) to Cambrian sedimentary cover has preserved in **Kilpisjärvi** just underneath the basal thrust front of the Finnish Caledonides. Based on fossil studies, the major part of the Dividal group is considered to be Cambrian, but the deposition of the lower parts may have taken place during the latest Precambrian (*Lehtovaara, 1988*).

Several late Precambrian–Phanerozoic impact structures are found in southern and central Finland and bear evidence for a sedimentary record. That of **Iso Naakkima** (for location; see Fig. 1) has preserved a 100-m-thick sequence of Neoproterozoic sedimentary rocks (*Elo et al., 1993*).The impact structure of **Saarijärvi** also includes remnants of a Neoproterozoic (NP3) to Cambrian sedimentary succession (*Tynni and Uutela, 1985*; *Öhman et al., 2000*). The circular depression of **Söderfjärden** is currently considered as an impact structure (*Lehtovaara 1984, 1992*; *Abels et al., 2000*) and is filled by a ~240-m-thick sequence of Cambrian sedimentary rocks (*Lehtovaara, 1982*; *Tynni, 1982*). The largely submerged carbonate rock sequence at the Bay of **Lumparn** in Åland has maximum thickness of ~120 m (*Lehtovaara, 1982*), and is preserved in a fault bounded depression (see *Winterhalter, 1982*). The sequence consists of Lower Cambrian siltstone and sandstone, Lower Ordovician glauconitic limestone, and Middle Ordovician limestone (*Tynni, 1982*; *Lehtovaara, 1982*, and references therein). The impact origin of the down-faulted block of Lumparn has been confirmed by the recognition of shock features (*Svensson, 1993*; *Abels et al., 2000*). In the impact crater of **Karikkoselkä**, sedimentary rocks are found as blocks and fragments within the allochtonous breccia. The major part of the observed siltstones, sandstones, and mudstones are interpreted as Cambrian or Precambrian, but Ordovician microfossils have also been reported (*Uutela, 2001*).

In southwestern Finland, **clastic sandstone dikes** have been observed in many localities (see *Bergman, 1982*). In these dikes, sandstone is found as fissure fillings in the cracks of the unweathered crystalline basement. According to microfossil studies (*Tynni, 1982*), most of the dikes are Lower Cambrian, some Lower Ordovician.

3. Tectonic model

Traditionally, the Neoproterozoic and Phanerozoic evolution of the Finnish part of the Fennoscandian shield has been considered tectonically silent and depositionally retarded (e.g., *Nikishin et al., 1996*; *Puura et al., 1996*). These ideas have presumably risen from the lack of preserved sedimentary record. It is evident, however, that the western part of the craton underwent major tectonic reworking at about 1100–900 Ma (Sveconorwegian orogeny) and again about 450–350 Ma (Caledonian orogeny). The latter was preceded by opening of the Iapetus Ocean at ~600 Ma and was succeeded by the opening of the North Atlantic in the Mesozoic and Tertiary.

The key question is whether the sedimentary rocks record (1) original basin configuration or (2) selective preservation of originally more extensive cover sequences. This can be assessed, at least to some extent, by analyzing thickness variations, sedimentary facies, degree of folding, and nature of bordering faults. Reliably dated impact structures, like Lappajärvi and Sääksjärvi, are good piercing points for the ancient earth surface and valuable source of additional information. Fission track studies (e.g., *Zeck et al., 1988*; *Tullborg et al., 1996*; *Larson et al., 1999*; *Murrell and Andriessen, 2000*) have also raised questions concerning the role of the Sveconorwegian and Caledonian orogenies and, especially, the original extent of the associated foreland basins. It has been suggested (*Larson et al., 1999*), that the Caledonian foreland sediments covered large parts of Finland during Late Paleozoic and Mesozoic. *Murrell and Andriessen (2000)* suggested that final exhumation of the shield occurred not earlier than in the Tertiary. These models are in obvious contrast to the traditional ideas of minor Phanerozoic deposition and absence of substantial sedimentary cover. In the following, the tectonic evolution of the Fennoscandian shield is divided into five stages:

- Intracratonic rift basin stage (~1600–1300 Ma)
- Crustal extension episodes and Sveconorwegian orogeny (~1300–900 Ma)
- Neoproterozoic exhumation stage (~900–600 Ma)
- Platform (passive margin to early foredeep) sedimentation stage (~600–420 Ma)
- Caledonian foreland stage (~420–350 Ma), final exhumation, and birth of the shield

Intracratonic rift basin stage (~1600–1300 Ma)

Extensive geophysical evidence for crustal thinning and magmatic underplating, long duration of magmatism, and substantial volume of crust-derived magma produced require a major thermal source in the subcontinental mantle as the rapakivi association was formed. Such a thermal head is compatible with a mantle plume model (cf. *Haapala and Rämö, 1992*; *Rämö and Korja, 2000*). This model has been challenged by *Åhäll et al. (2000)* who related the Fennoscandian rapakivi magmatism to intermittent subduction events on the SW flank of the shield. This "inboard model" can not, however, explain the non-linear age distribution of the rapakivi granite batholiths (cf. *Rämö et al., 2000*). Whatever the ultimate tectonic explanation of the Fennoscandian rapakivi magmatism, the following features (see also *Rämö and Korja, 2000*) are charateristic to this stage:

- major tectonothermal activity with mantle upwelling, magmatic underplating, and emplacement of rapakivi granites
- crustal thinning and formation of intracratonic basins

Deposition of alluvial arkosic sandstones (e.g., Satakunta, Lake Ladoga) is in accord with an overall intracratonic setting. A minimum age (1265 Ma) is well constrained for the Satakunta sandstone, and the recent results from the Lake Ladoga area (northeastern Russia; see Fig.1 for location) indicate that graben formation and sedimentation of the arkosic sandstones occurred here shortly after or concurrently with the emplacement of the 1560–1530 Salmi rapakivi granite complex (*Amelin et al., 1997*; *Rämö et al., 2001*). In general, however, the depositional age of Mesoproterozoic sequences is poorly constrained. It is possible that the preserved strata represent multiple depositional stages and the basin histories may thus be longer and more complex than heretofore assumed.

The original extent of the Mesoproterozoic intracratonic basin system (MP1 to MP2) and the primary thickness of the sediments are hard to estimate reliably. However, comparisons to the assumed scale of Mesoproterozoic (*Riphean*) basins of the Russian platform (e.g., *Bogdanova et al., 1996*; *Nikishin et al., 1996*) suggest that deposition was more widespread than the preserved basins indicate.

Crustal extension episodes and Sveconorwegian orogeny (~1300–900 Ma)

The ~1265 Ma dike swarms and solitary dikes of Salla (~1100 Ma) and Laanila (~1000 Ma) plausibly reflect short-lived extensional episodes in Finland. The time span approximately corresponds to that of the Sveconorwegian orogeny (~1.2–0.9 Ga; *Johansson et al., 1991*). The 1265 Ma dikes could record the onset of the Sveconorwegial-Grenvillian orogeny (e.g., *Gorbatchev et al., 1987*; *Rämö, 1990*) or, from the supercontinent point of view, break-up of Baltica and Laurentia (*Elming and Mattson, 2001*).

The ~1265 Ma dikes are – both in Finland and in Sweden – spatially connected to the older Mesoproterozoic (MP1 or *'Subjotnian'*) intrusions and to continental sedimentary deposits (MP1/MP2). The emplacement plausibly occurred along pre-existing (MP1) crustal fractures and it is possible that initial subsidence related to this stage triggered the deposition represented by the upper parts of the Satakunta and Muhos formations. Down-faulting related to this extensional stage was apparently the main factor controlling the preservation of the Satakunta and Muhos formations. Interestingly, the ~1180 Ma kaolinic Virtasalmi saprolite (*Sarapää, 1996*) manifests the absence of sedimentary cover and attests to conditions of long-lasting continental weathering in central Finland.

The 1100–1000 Ma basaltic dikes in northern Finland, Norway, and Russia and the ~1000 Ma dikes in central Sweden roughly strike in SW-NE and NW-SE directions (Fig. 1) and could record zones of transcurrent faulting in the Sveconorwegian foreland. The effect of the Sveconorwegian orogeny to the foreland in the northeast is quite speculative, however. Nevertheless, it has been suggested that a km-scale pile of foreland sediments was deposited in proximity of the orogenic front in southwestern Sweden at ~950 Ma (*Tullborg et al., 1996*; *Larson et al., 1999*). In Finland, the Iso Naakkima sedimentary sequence overlies the Virtasalmi saprolite. The depositional age is poorly constrained at ~1000–650 Ma and it is thus impossible to speculate with connections to the Sveconorwegian foreland deposition. We estimate that, at the end of the Mesoproterozoic Era, the most of Finland was covered by continental sedimentary rocks.

Neoproterozoic exhumation stage (~900–600 Ma)

With a possible exception of the Iso Naakkima sequence (see above), observations concerning Neoproterozoic (NP1 and NP2) geological evolution are virtually absent in Finland, and thus the early Neoproterozoic history is also poorly known. It is tentatively suggested that gradual uplift and erosion characterized most of the Neoproterozoic Era in Finland. Some regional subsidence and deposition, succeeding the long period of erosion and subaereal weathering, occurred towards the end of this period.

Development of an ancient shield area with no substantial sedimentary cover in northern Europe during the Neoproterozoic (NP2/NP3 or *Pre-Vendian*) seems evident. This is manifested by erosion of pre-existing cover, continental conditions, and subaerial weathering profiles in many parts of the East European craton (e.g., *Puura et al., 1996*). The preserved *Vendian* successions in the North Baltic and Lake Ladoga - St Petersburg area show no decrease in thickness towards the north, which suggests that these deposits once were present also in southern Finland (*Puura et al., 1996*). This depositional phase appears to have been a prologue to the Cambrian basin formation.

Platform stage (~600–420 Ma)

Observations such as Cambrian clastic sandstone dikes and the ~560 Ma Sääksjärvi impact structure *(Pihlaja and Kujala, 2000)* southern Finland indicate that the current erosional level is very near to the major unconformity between the crystalline basement and

NP3/Cambrian strata. During the latest Precambrian and Cambrian, fluvial and shallow marine deposition prevailed in the slowly drowning northwestern part of the East European craton. The basin formation was contemporaneous with the break-up of the Mesoproterozoic Rodinia supercontinent and opening of the Iapetus Ocean (at ~600 Ma) in the west.

The lowermost Cambrian (mudstone and siltstone) formations of Estonia show increasing thickness towards the northeast and east and, later, towards the west and southwest. Towards the end of Cambrian, shallow marine sandstones and siltstones became dominant. The basin was apparently deepening towards the newly-formed continental margin in the west. (e.g., *Puura et al., 1996*; *Mens and Pirrus, 1997*; *Artyushkov et al., 2000*). Based on comparisons to Estonia, we suggest that, at the end of Cambrian, southern Finland was covered by a 100–350-m-thick pile of fine clastic sediments.

Towards the Ordovician and Silurian, an extensive carbonate platform developed at the craton margin. Carbonate rocks and marly clastic sediments (200 to 400 m in thickness) represent this stage in the Baltic countries (e.g., *Puura and Vaher, 1997*). Before the onset of the Caledonian collisional tectonism, the axis of maximum thickness of sediments appears to have been situated along a line from the present west coast of Estonia to the Polish coast. The basin was plausibly open towards Iapetus in west. These deposits probably covered large parts of the present day Fennoscandian shield. In central Finland, this view is supported by recent observations of Ordovician rocks in the Karikkoselkä impact structure (see above). Furthermore, the Ordovician in Estonia shows neither systematic decrease of bed thickness nor change of sedimentary facies towards the north. As the thickness of Cambrian and Ordovician rocks in the Bothnian Sea corresponds to that of Baltic countries, the cover in the Bothnian sea rather represents a tectonically preserved remnant of an originally extensive cover sequence than a separate Paleozoic basin.

The development of early Paleozoic marine basins may have been connected to development of passive margin-type conditions, followed by initial foreland subsidence and early foredeep-type basin.

Caledonian foreland stage (~420–350 Ma), final exhumation, and birth of the shield

The formation of a thrust and fold belt in Scandinavia at ~400 Ma and related crustal thickening caused a foreland basin on the eastern side of the orogen. As the crustal level in Scandinavia today represents a deep section through the Scandinavian Caledonides, vast amounts of sediments must have been deposited during the Devonian. The resultant heating of the basement due to this burial has been recorded and modeled (e.g., *Zeck et al., 1988*; *Tullborg et al., 1996*; *Samuelsson and Middleton, 1998*; *Larson et al., 1999*).

The Silurian–Devonian depositional hiatus plausibly records a rapidly changing base level configuration and base set-up related to foreland basin initiation. The role of Caledonides in Paleozoic evolution of Baltic region is reflected in basin shape. The basin axis of both the early Paleozoic carbonate depository and the Devonian sandstone dominated basin apparently follows the overall NE–SW trend of the Scandinavian Caledonides. We estimate, that at end of the Silurian, most of the present Finland was covered by 200 to 500 m of Late Neoproterozoic and Paleozoic platform sediments. The destruction of the cover occurred either in relation to the uplift of a foreland bulge in the Devonian or as a consequence of the uplift following the denudation of the Caledonides in the Late Paleozoic and Mesozoic.

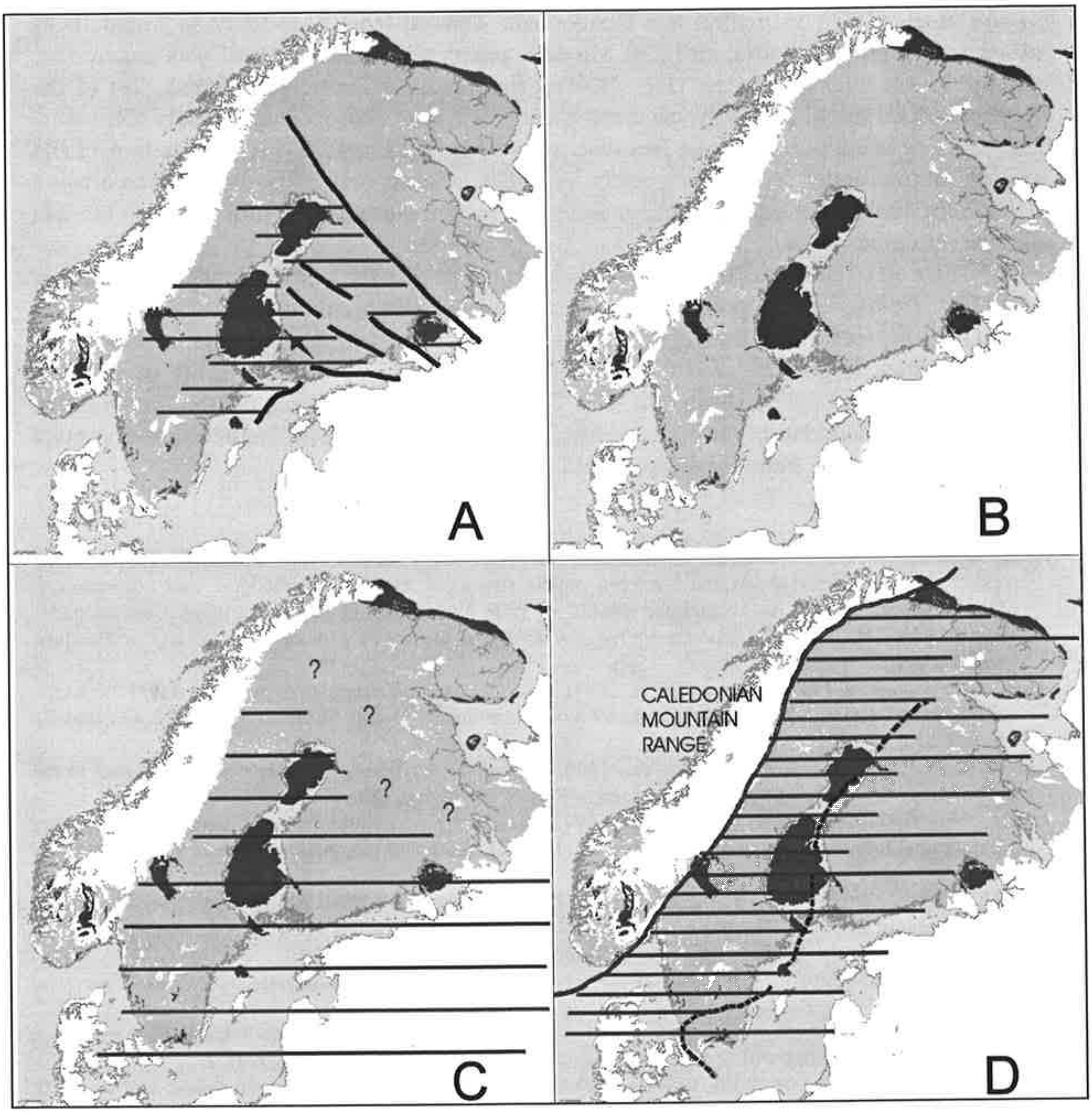


Figure 2. Schematic paleogeographic snapshots of the Fennoscandian shield: (A) at 1300 Ma (MP2), alluvial-fluvial-eolian deposition in intracratonic rift basins; (B) at 700 Ma (NP2), erosion in continental conditions, most of the Mesoproterozoic sediments have been erased; (C) at 500 Ma (Cambrian), fluvial to shallow marine deposition in passive margin setting; (D) at 400 Ma (Devonian), alluvial-fluvial deposition in a foreland basin.

Larson et al. (1999) estimated that the deposits derived from Scandinavian Caledonides buried most of Finland and, still 250 Ma ago, practically whole Finland was covered by 0.5- to 1.5-km-thick sediments (Fig. 2). The final exhumation of the Finnish part of the Fennoscandian shield probably occurred during the Late Mesozoic and Early Cenozoic. This estimate is supported by the fact that, at ~75 Ma, the Lappajärvi meteorite impact site was still buried under sedimentary rocks. A rather large amount of post-Devonian erosion in northern Finland is undisputably manifested by the current exposure of the ~365 Ma alkaline intrusions (Iivaara, Sokli).

Denudation of the Caledonides had reached a stage of hilly landscape at Late Mesozoic time (*Riis, 1996*). As a response to the opening of the North Atlantic, substantial tectonic uplift (1-2 km) took place during the Cenozoic in northwestern Scandinavia. According to some estimates (e.g., *Riis, 1996*; *Stuevold and Eldholm, 1996*), the uplift in northern Finland was around 500 m during the Late Cretaceous and Paleogene, continuing up to the Neogene. Slow denudation during Late Mesozoic and Tertiary uplift in the northwest put a finishing touch on the Fennoscandian shield as we know it today.

References

Abels, A., Bergman, L., Lehtinen, M. and Pesonen, L.J. 2000. Structural constraints and interpretations on the formation of the Söderfjärden and Lumparn impact structures, Finland. In: Plado, J. and Pesonen, L.J. (eds.) Meteorite impacts in Precambrian shields. 4th ESF Workshop, Lappajärvi-Karikkoselkä-Sääksjärvi, Finland, May 24-28, 2000. *Espoo : Helsinki: Geological Survey of Finland : University of Helsinki. Programme and abstracts, p. 26.*

Åhäll, K.-I., Connelly, J.N. and Brewer, T.S., 2000. Episodic rapakivi magmatism due to distal orogenesis?: Correlation of 1.69-1.50 Ga orogenic and inboard "anorogenic" events in the Baltic Shield. *Geology 28 (9), 823-826.*

Amantov, A., Laitakari, I. And Poroshin, Ye., 1996. Jotnian and Postjotnian: sandstones and diabases in the surroundings of the Gulf of Finland. *Geological Survey of Finland, Special Paper 21, 99-113.*

Amelin, Yu., Larin, A.M. and Tucker, R.D., 1997. Chronology of multiphase emplacement of the Salmi rapakivi granite-anorthosite complex, Baltic Shield: implications for magmatic evolution. *Contributions to Mineralogy and Petrology **127**, 353-368.*

Artyushkov, E.A., Artyushkov, Y.A., Lindström, M. and Popov, L.E., 2000. Relative sea-level changes in Baltoscandia in the Cambrian and Early Ordovician; the predominance of tectonic factors and the absence of large scale eustatic fluctuations. *Tectonophysics **320 (3-4)**, 375-407.*

Bergman, L. 1982. Paleozoic sediments in the rapakivi area of the Åland Islands. *Geological Survey of Finland Bulletin **317**, 7-27.*

Bogdanova, S.V., Pashkevich, I.K. and Gorbatschev, R., 1994. Late-Precambrian to Triassic history of East European craton: dynamics of sedimentary basin evolution. *Tectonophysics **268 (1-4)**, 1-22.*

Elming, S.-Å. and Mattson, H., 2001. Post Jotnian basic intrusions in the Fennoscandian Shield, and the break up of Baltica from Laurentia: a palaeomagnetic and AMS study. *Precambrian Research **108**, 215-236.*

Elo, S., Kuivasaari, T., Lehtinen, M., Sarapää, O. and Uutela, A. 1993. Iso-Naakkima, a circular structure filled with Neoproterozoic sediments, Pieksämäki, southeastern Finland. *Bulletin of the Geological Society of Finland **65 (1)**, 3-30.*

Gorbatchev, R., Lindh, A., Solylom, Z., Laitakari, I., Aro, K., Lobach-Zhuchenko, S.B., Markov, M.S., Ivliev, A.I. and Bryhni. I., 1987. Mafic dyke swarms of the Baltic shield. In: Halls, H.C. and Fahrig, W.F. (eds.) Mafic dyke swarms. *Geological Association of Canada Special Paper **34**, 361-372.*

Haapala, I. and Rämö, O.T., 1992. Tectonic setting and origin of the Proterozoic rapakivi granites of southern Fennoscandia. *Transactions of the Royal Society of Edinburgh: Earth Sciences **83**, 165-171.*

Johansson, L., Lindh, A. and Möller, C., 1991. Late Sveconorwegian (Grenville) high-pressure granulite facies metamorphism in southwest Sweden. *Journal of Metamorphic Geology **9**, 283-292.*

Koistinen, T. (ed.), 1994. Precambrian basement of the Gulf of Finland and surrounding area, 1:1 mill. *Geological Survey of Finland, Espoo.*

Koistinen, T., Stephens, M.B., Bogatchev, V., Nordgulen, Ø., Wennerström, M., Korhonen, J. (comp.) 2001. Geological map of the Fennoscandian Shield, scale 1:2 000 000. *Espoo : Trondheim : Uppsala : Moscow: Geological Survey of Finland : Geological Survey of Norway : Geological Survey of Sweden : Ministry of Natural Resources of Russia.*

Kramm, U., Kogarko, L.N., Kononova, V.A. and Vartiainen, H., 1993. The Kola alkaline province of the CIS and Finland: Precise Rb-Sr ages define 380-360 Na age range for all magmatism. *Lithos **30**, 33-44.*

Larson, S.Å., Tullborg, E.-L., Cederbom, C. and Stiberg, J.-P., 1999. Sveconorwegian and Caledonian foreland basins in the Baltic Shield revealed by fission-track thermochronology. *Terra Nova **11** (5), 210-215.*

Lehtovaara, J.J., 1982a. Palaeozoic sedimentary rocks in Finland. *Annales Academiae Scientiarum Fennicae. Series A. III. Geologica - Geographica **133**, 35 p.*

Lehtovaara, J.J., 1982b. Stratigraphical section through Lower Cambrian at Söderfjärden, Vaasa, western Finland. *Bulletin of the Geological Society of Finland **54** (1-2), 35-43.*

Lehtovaara, J.J., 1984. Söderfjärden, Vaasa, western Finland: a crater covered with sedimentary material since its formation in the Cambrian (in Finnish with an English summary). *Terra **96** (1), 23-33.*

Lehtovaara, J.J., 1988. The palaeosedimentology of the autochthon of the Finnish Caledonides. *Geological Survey of Finland Special Paper **5**, 255-264.*

Lehtovaara, J.J., 1992. Söderfjärden: a Cambrian impact crater in western Finland. In: Pesonen, L.J. and Henkel, H. (eds.) Terrestrial impact craters and craterform structures with a special focus on Fennoscandia: papers from a symposium held in Espoo and Lappajärvi, Finland, May 29-31, 1990. *Tectonophysics **216** (1-2), 157-161.*

Lundqvist, T., Bøe, R., Kousa, J., Lukkarinen, H., Lutro, O., Roberts, D., Solli, A., Stephens, M. and Weihed, P., 1996. Bedrock map of Central Fennoscandia. Scale 1 : 1 000 000. *Espoo : Trondheim : Uppsala: Geological Survey of Finland : Geological Survey of Norway : Geological Survey of Sweden.*

Mens, K. and Pirrus, E., 1997. Sedimentary cover; Cambrian. In: Raukas, A. and Teedumae, A. (eds.) Geology and mineral resources of Estonia. *Tallinn: Estonian Academy Publishers, p. 39-51.*

Murrel, G.R. and Andriessen, P.A.M., 2000. Longterm morphotectonic evolution in Finland; constraints on quantification and timing of vertical motions from fission track thermochronology. Fission track 2000; 9th international conference on fission track dating and thermochronology. Abstracts. *Geological Society of Australia **58**, 233-234.*

Nikishin, A.M., Ziegler, P.A., Stephenson, R.A., Cloetingh, S.A.P.L., Furne, A.V., Fokin, A., Ershov, A.V., Bolotov, S.N., Korotaev, M.V., Alekseev, A.S., Gorbachev, V.I., Shipilov, E.V., Lankreijer, A., Bembinova, E.Y. and Shalimov, I.V., 1996. Late Precambrian to Triassic history of the East European Craton: dynamics of sedimentary basin evolution. *Tectonophysics **268**, 23-63.*

Öhman, T., Pesonen, Lauri J., Raitala, J., Uutela, A. and Tuisku, P., 2000. The Saarijarvi Crater; older and larger than assumed? In: Plado, J. and Pesonen, L.J. (eds.) Meteorite impacts in Precambrian shields. 4th ESF Workshop, Lappajärvi-Karikkoselkä-Sääksjärvi, Finland, May 24-28, 2000. *Espoo : Helsinki: Geological Survey of Finland : University of Helsinki. Programme and abstracts, p. 82.*

Pihlaja, P. and Kujala, H., 2000. Some geochemical features of impact melts, suevites, and target rocks at Sääksjärvi impact structure, Finland. In: Plado, J. and Pesonen, L.J. (eds.) Meteorite impacts in Precambrian shields. 4th ESF Workshop, Lappajärvi-Karikkoselkä-Sääksjärvi, Finland, May 24-28, 2000. *Espoo : Helsinki: Geological Survey of Finland : University of Helsinki. Programme and abstracts, p. 82.*

Puura, V. and Vaher, R., 1997. Tectonics; Cover structure In: Raukas, A. and Teedumae, A. (eds.) Geology and mineral resources of Estonia. *Tallinn: Estonian Academy Publishers, p. 167-176.*

Puura, V., Amantov, A., Tikhomirov, S. and Laitakari, I., 1996. Latest events affecting the Precambrian basement, Gulf of Finland and surrounding areas. In: Koistinen, T. (ed.) Explanation to the Map of Precambrian Basement of the Gulf of Finland and Surrounding Areas, 1:1 mill. Espoo. *Geological Survey of Finland Special Paper **21**, 115-125.*

Rämö, O.T., 1990. Diabase dyke swarms and silicic magmatism - evidence from the Proterozoic of Finland. In: Parker, A.J., Rickwood, P.C. and Tucker, D.H. (eds.) Mafic dykes and emplacement mechanics : proceedings of the Second International Dyke Conference, Adelaide, South Australia, 12-16 September 1990. *Rotterdam, A. A. Balkema, 185-199.*

Rämö, O.T. and Korja, A., 2000. Mid-Proterozoic evolution of the Fennoscandian Shield, Extended abstract, 31th IGC, Rio de Janeiro.

Rämö, O.T., Korja, A., Haapala, I., Eklund, O., Fröjdö, S. and Vaasjoki, M., 2000. Evolution of the Fennoscandian lithosphere in the Mid-Proterozoic : the rapakivi magmatism. In: Pesonen, L.J., Korja, A. and Hjelt, S.-E. (eds.) Lithosphere 2000 : a symposium on the structure, composition and evolution of the lithosphere in Finland, Espoo, Otaniemi, October 4-5, 2000 : programme and extended abstracts. *Institute of Seismology. University of Helsinki, Report **S-41**, 129-136.*

Rämö, O.T., Mänttäri, I., Vaasjoki, M., Upton, B.G.J., Sviridenko, L., 2001. Age and significance of Mesoproterozoic CFB magmatism, Lake Ladoga region, NW Russia. *Geological Society of America Abstracts with Programs **33** (6), A-139.*

Riis, F., 1996. Quantification of Cenozoic vertical movements of Scandinavia by correlation of morphological surfaces with offshore data. *Global and Planetary Change **12**, 331-357.*

Samuelsson, J. and Middleton, M.F.,1998. The Caledonian foreland basin in Scandinavia: constrained by thermal maturation of the Alum Shale. *GFF **120**, 307-314.*

Sarapää, O., 1996. Proterozoic primary kaolin deposits at Virtasalmi, southeastern Finland. *Espoo: Geological Survey of Finland, 152 p.*

Simonen, A. and Kouvo, O., 1955. Sandstones in Finland. *Comptes Rendues de la Societe géologique de*

Finlande ***28***, *57-87.*

Stuevold, L.-M. and Eldholm, O., 1996. Cenozoic uplift of Fennoscandia inferred from a study of the mid-Norwegian margin. *Global and Planetary Change* ***12 (1-4)***, *359-386.*

Suominen, V., 1991. The chronostratigraphy of SW Finland with special reference to the Postjotnian and Subjotnian diabases. *Geological Survey of Finland, Bulletin* ***356***, *100 p.*

Svensson, N.B., 1993. Lumparn Bay: a meteorite impact crater in the Åland archipelago, southwest. *Meteoritics* ***28 (3)***, *445.*

Tullborg, E-L., Larson S.Å. and Stiberg, J-P., 1996. Subsidence and uplift of the present surface in the southeastern part of the Fennoscandian Shield. *GFF* ***118***, *26-128.*

Tynni, R., 1982a. New results of studies on the fossils in the Lower Cambrian sediment deposits of the Söderfjärden basin. *Bulletin of the Geological Society of Finland* ***54 (1-2)***, *57-68.*

Tynni, R., 1982b. On Paleozoic microfossils in clastic dykes in the Åland Islands and in the core samples of Lumparn. *Geological Survey of Finland Bulletin* ***317***, *35-114.*

Tynni, R. and Donner, J., 1980. A microfossil and sedimentation study of the Late Precambrian formation of Hailuoto, Finland. *Geological Survey of Finland Bulletin* ***311***, *27 p.*

Tynni, R. and Uutela, A., 1985. Late Precambrian shale formation of Taivalkoski in northern Finland (in Finnish with an English summary). *Geologi* ***37 (4-5)***, *61-65.*

Uutela, A., 2001. Proterozoic and early Palaeozoic microfossils in the Karikkoselkä impact crater, central Finland. *Bulletin of the Geological Society of Finland* ***73 (1-2)***, *75-85.*

Vaasjoki, M. and Sipilä, P., 2001. U-Pb isotopic determinations of baddeleyite and zircon from the Halti-Ridnitsohkka intrusion in Finnish Lapland: a further constraint on Caledonide evolution. *Geological Survey of Finland, Special Paper* ***33***, *247-253.*

Veltheim, V., 1969. On the pre-Quaternary geology of the Bothnian Bay area in the Baltic Sea. *Bulletin Commission géologique de Finlande* ***239***, *56 p.*

Winterhalter, B., 1982. The bedrock geology of Lumparn Bay, Åland. *Geological Survey of Finland Bulletin* ***317***, *115-130.*

Winterhalter B., Flodén T., Ignatius H., Axberg S. and Niemistö L., 1981. Geology of the Baltic Sea. In: Voipio, A.(ed.) The Baltic Sea. *Amsterdam: Elsevier, p. 1-121.*

Zeck, H.P., Andriessen, P.A.M., Hansen, K., Jensen, P.K. and Rasmussen, B.L., 1988. Paleozoic paleo-cover of the southern part of the Fennoscandian Shield; fission track constraints. *Tectonophysics* ***149 (1-2)***, *61-66.*

Geophysical Maps and Crustal Model Interpretations of the Fennoscandian Shield

Juha V. Korhonen and Seppo Elo

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

Digital and printed Geological, Magnetic Anomaly and Bouguer Anomaly maps of the Fennoscandian Shield have been compiled on a scale of 1:2,000,000 as a joint venture between the Geological Surveys of Finland, Norway and Sweden, the Finnish Geodetic Institute and the Ministry of Natural Resources of Russia.

Magnetic total intensity anomalies (DGRF-65), Bouguer anomalies (d=2670 kg/m3) and geological units from 3400 Ma to present of the Fennoscandian Shield have been digitally compiled as 1:2 000 000 maps. By means of 1: 15,000,000 insert maps facilitate comparisons of various anomaly components pseudogravimetric anomaly versus Bouguer anomaly, DGRF-65 anomaly versus pseudomagnetic anomaly, magnetic vertical derivative versus second derivative of Bouguer anomaly. Data on bulk-density, total magnetisation and lithology of samples are presented as scatter diagrams and distribution maps of the average petrophysical properties in space and time.

In the sample level, the bulk density correlates with the lithology and, together with magnetisation, establishes four principal populations of petrophysical properties. The average properties, calculated for 5 km x 5 km cells, correlate only weakly with average Bouguer and magnetic anomaly, revealing major deep- seated sources of anomalies. Pseudogravimetric and Bouguer anomalies correlate only locally with each other. The correlation is negative in the area of felsic Palaeoproterozoic rocks in W- and NW-parts of the Shield.

In 2D models the sources of gravity anomalies are explained by lateral variation of density in upper and lower crust. Smoothly varying regional components are explained by boundaries of the lower crust, the upper mantle and the asthenosphere. Magnetic anomalies are explained by lateral variation of magnetisation in the upper crust. Regional components are due to the lateral variation of magnetisation in the lower crust and the boundaries of lower crust and mantle and the Curie isotherm of magnetite.

References

Korhonen, J.V., Aaro, S., All, T., Skilbrei, J.R., Säävuori, H., Vaher, R., Zhdanova, L. and Koistinen, T., 2002. Magnetic Anomaly Map of the Fennoscandian Shield 1:2,000,000. Geological Surveys of Finland, Norway and Sweden, Ministry of Natural Resources of Russian Federation. Espoo, in print.

Korhonen, J.V., Aaro, S., All, T., Elo, S., Haller, L.Å., Kulinich, A., Kääriäinen, J., Solheim, D., Skilbrei, J.R., Säävuori, H., Vaher, R., Zhdanova, L. and Koistinen, T., 2002. Bouguer Anomaly Map of the Fennoscandian Shield 1:2,000,000. Geological Surveys of Finland, Norway and Sweden, Ministry of Natural Resources of Russian Federation. Espoo, in print.

Crustal Conductivity Map of the Fennoscandian Shield

Toivo Korja[1], Martin Engels[2], Abdoulkhay A. Zhamaletdinov[3], Aida A. Kovtun[4], Nikolai A. Palshin[5], Maxim Yu. Smirnov[4], Aleksander D. Tokarev[3], Vladimir E. Asming[3], Leonid L. Vanyan[†6], Isabella L. Vardaniants[4], and the BEAR Working Group[5]

[1]Institute of Geosciences, Department of Geophysics,
P.O. Box 3000, FIN-90014 University of Oulu, Finland
[2] Uppsala University, Department of Earth Sciences, Geophysics, Uppsala, Sweden
[3] Russian Academy of Sciences, Kola Science Centre, Geological Institute, Apatity, Russia
[4] Institute of Physics, St. Petersburg University, St. Petersburg, Russia
[6] Russian Academy of Sciences, Shirsov Institute of Oceanology, Moscow, Russia
E-mail: toivo.korja@oulu.fi

We present a three-dimensional crustal conductivity model for Fennoscandia. Such a model is needed in many applications ranging from investigations on upper mantle conductivity (BEAR project) to studies on induced currents in technological systems. Besides methodological advances made using the crustal conductivity model, data on large-scale subsurface conductivity structures and their EM signatures are also useful for the exploration of the nature and tectono-geological significance of major geoelectric units in the Fennoscandian Shield. Model is based on a compilation of all available information on crustal conductivity from a large number of geoelectromagnetic investigations made in Fennoscandia during the last two to three decades. In the resulting crustal model, lateral variations of the subsurface conductivity structure are expressed as an integrated conductance of eight separate layers. The first three layers contain the conductance of seawater, sediments (marine sediments, continental sedimentary cover and post-glacial overburden) and bedrock in the uppermost 10 km. Other layers, each 10 km thick, cover the conductance of the bedrock below the depth of 10 km. The compiled model shows the heterogeneous nature of crustal conductivity. Yet it indicates that crustal conductivity is characterized by the presence of a few elongated belts of conductors intervening more homogeneous resistive segments.

Keywords: electrical conductivity, conductance, crust, Precambrian, Fennoscandia

1. Introduction

A priori knowledge on large-scale subsurface conductivity structure is required in many applications of geophysical EM methods investigating electrical properties of the crust and upper mantle. In upper mantle studies, a priori knowledge on the crustal conductance structure has a great value, because long period electromagnetic (EM) methods are sensitive to crustal conductors even though the structure cannot be resolved due to sparse spatial sampling. In crustal studies, the effects of complicated sedimentary cover and oceans are traditionally examined in terms of its conductance, i.e. vertically integrated conductivity. Data on the large-scale subsurface conductivity structures and their EM signatures are also useful for the exploration of the nature and tectono-geological significance of the major geoelectric units, and for the understanding of their distorting influence in the results of local EM studies.

5 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 and Finnish Meteorological Institute in Helsinki and Nurmijärvi), *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, Lviv Centre of Institutes of Space Research and Carpathian Branch of Subbotin Institute of Geophysics in Lviv).

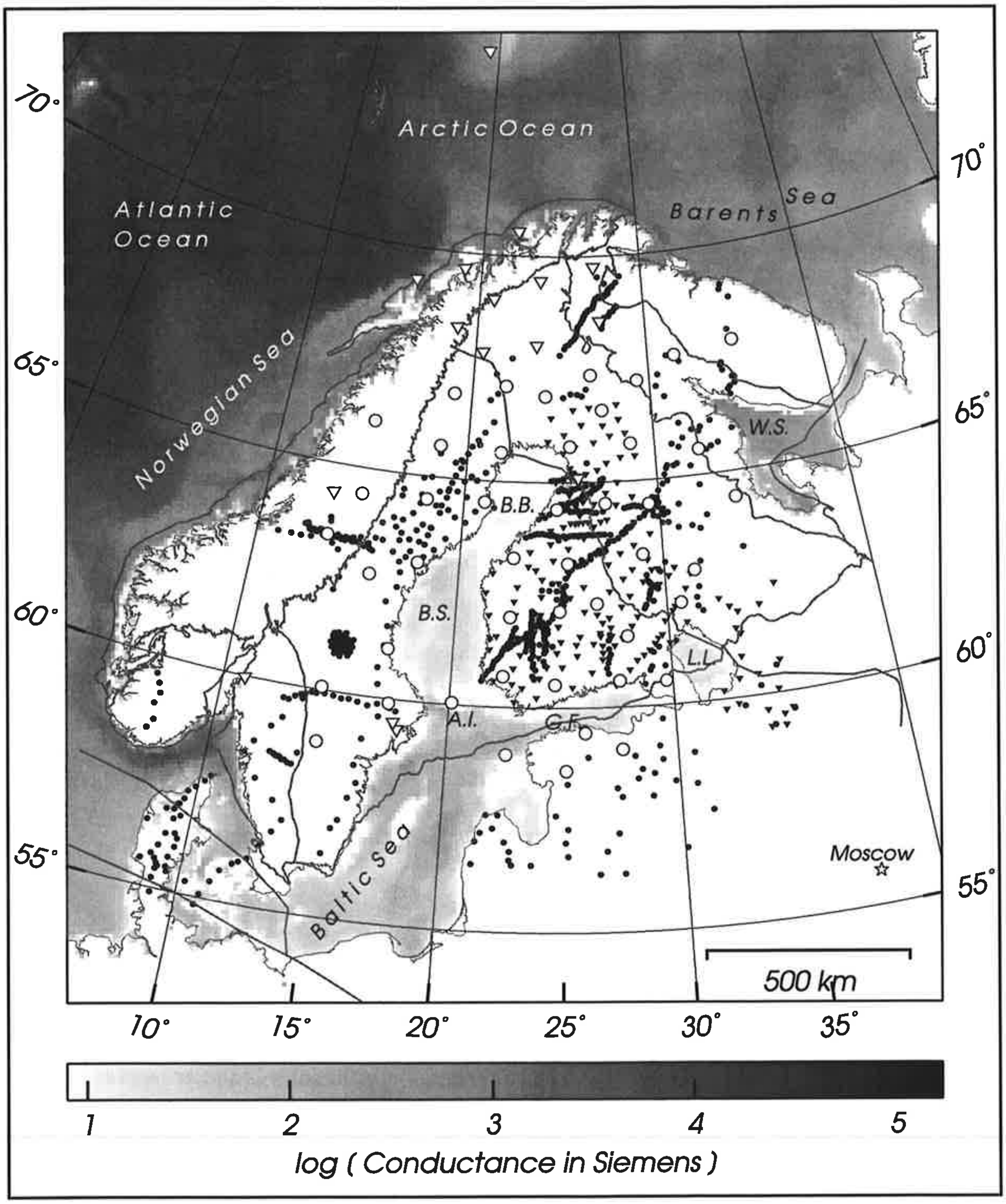


Figure 1. Crustal electromagnetic surveys in Fennoscandia. Solid inverted triangles denote the magnetometer sites of array studies and dots the sites of magnetotelluric soundings. Circles and open inverted triangles represent the magnetotelluric and magnetometer sites of the BEAR array, respectively (BEAR data not used for compilation; sites shown only for reference). In sea areas, the conductance of seawater is shown as a background. Thick grey lines denote the boundaries of the major crustal segments. A.I. = Åland islands, B.B. = Bothnian Bay, B.S. = Bothnian Sea, G.F. = Gulf of Finland, L.L. = Lake Ladoga, W.S. = White Sea. Figure is modified from Korja et al., 2002.

2. Compilation: data and elements of the database

The final model consists of water layer, sediment layer and six bedrock layers. Water conductances are obtained by converting bathymetric data (NOAA ETOPO database for North Europe; *NOAA, 1988*) to conductances for the entire database area using different conversions rates (4 S/m – 0.1 S/m) for different sea areas. At sea areas, the thickness of the marine sediments was obtained from a global sediment map of *Laske and Masters (1997)*. The sediment thicknesses were converted to conductances using the approach of *Flosadóttir et al. (1997)*, which takes into account the relation between the porosity and electrical conductivity, water conductivity and compaction of sedimentary rocks with depth. In land areas close to the Fennoscandian Shield, the conductance of the sedimentary cover is estimated primarily from MT data. It should be noted that it is not necessary to have very accurate estimates in peripheral areas because our primary target area is the Fennoscandian Shield and its vicinities. The initial data for the crustal (bedrock) conductances come from 1D and 2D resistivity models obtained through modelling/inversion of magnetotelluric and magnetometer array data (Fig. 1). Separate geoelectric units were delineated using data from magnetometer arrays and airborne electromagnetic surveys which provide information on the lateral extension of the conductivity anomalies. Finally, a resistivity model from a priori 1D or 2D models was assigned for each geoelectric unit. In peripheral areas, outside Fennoscandia, an average 1D background model for Fennoscandia (see *Korja et al., 2002a*) was used for the bedrock layers.

In the final model, lateral variations of the subsurface conductivity structure are expressed as an integrated conductance of eight separate layers. The first three layers contain the conductance of seawater, sediments and bedrock in the uppermost 10 km. Other layers cover the conductance of the bedrock below the depth of 10 km. The entire database covers an area from 0°E to 50°E and 50°N to 85°N. The size of each cell is 5′ (NS) x 5′ (EW) x 10 km (vertical), which represents a cell of c. 9300 m (NS) x 3900 m (EW) x 10000 m (vertical) at the latitude of 65°N. The cell size in the east-west direction depends on the latitude, varying from 4700 to 3200 m at 60°N and 70°N, respectively.

3. Crustal conductivity model

The crustal conductivity structure appears to be very heterogeneous in the Fennoscandian Shield. Upper crust, in particular (Fig. 2), seems to have a very complex structure evidently reflecting a complex geological history of the shield. Lower crust, on the other hand, seems to be slightly more homogeneous although large contrasts are found also in the lower crust. Although the local conductivity structure is highly complex, the following major features characterize general structure: (i) The presence of a few highly conducting conductors (S > 1000 S) that form long but rather narrow zones especially in the upper crust, (ii) the presence of large and relatively resistive areas between the conducting belts and (iii) very large variations in conductance ranging from a few S to tens of thousands of S. The conductance variations are extreme in a sense that both the very resistive (S < 10 S) and very conductive (S > 10000 S) sides are represented, and the conductance contrasts even above 100000 are met. Variations are also spatially rapid reflecting the discrete nature of crustal conductance. Yet the variations are in nature even more pronounces because the filtering applied in the final interpolation stages have smoothed the conductances.

The most dominant system of elongated conductors traverses the entire Shield from northwest to southeast in the central part of the Shield along the border between the Archaean Karelian Province in the northeast and the Palaeoproterozoic Svecofennian Domain in the southwest. A more detailed examination reveals several separate conductors including Lake Ladoga conductors in southeast in Russia (LLA in Fig. 2), southern Finland (SFI),

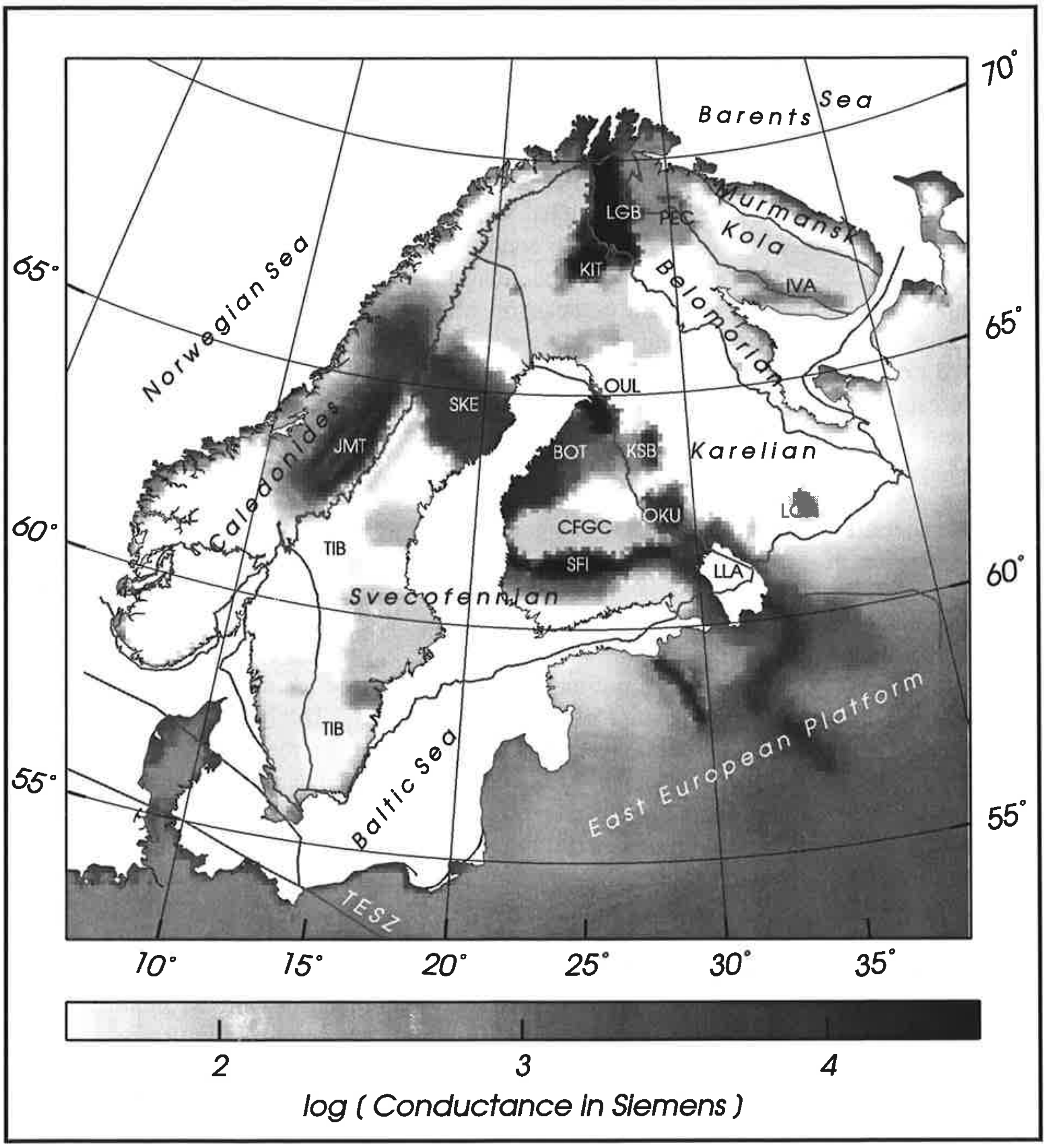


Figure 2. Crustal conductivity in Fennoscandia. Integrated conductance (conductivity multiplied by thickness) in Siemens for the depth interval of 0-20 km (sediments and bedrock) is shown as a grey shade map. Thick lines represent boundaries of major crustal segments in the Fennoscandian Shield. Map has been produced using the conductance data compiled by Korja et al. (2002). Note that the interpolation algorithm has smoothed conductances and therefore borders of crustal conductors appear smooth although in reality they usually represent sharp lithological contrasts. The conductors labelled in map: BOT = Bothnian, IVA = Imandra-Varzuga, JMT = Jämtland, KIT = Kittilä, KSB = Kainuu Belt, LGB = Lapland Granulite Belt, LLA = Lake Ladoga, LON = Lake Onega, OKU = Outokumpu, OUL = Oulu, PEC = Pechenga, SFI = Southern Finland, SKE = Skellefteå. TIB = highly resistive granites of the Trans-Scandinavian Igneous Belt.

Outokumpu (OKU), Kainuu Belt (KSB), Oulu (OUL) and Bothnian (BOT) conductors in Finland and Skellefteå (SKE) conductor in northwest in Sweden. At both ends, a single conductor is found (Lake Ladoga in Russia and Storavan-Skellefteå in Sweden; LL and STO/SKE, respectively in Fig 2) whereas in Finland the conductor is split into two branches that surround the Central Finland Granitoid Complex (CFGC in Fig 2). In the southeast, the conducting belt extends beneath the Phanerozoic sedimentary cover indicating possibly the approximate location of the Archaean-Proterozoic boundary beneath the sediments.

The second set of conductors is found in the Lapland-Kola Domain in the northeast and east around the Lapland Granulite Belt and in the Imandra-Varzuga Belt (LGB and IVA, respectively, in Fig 2) representing Palaeoproterozoic supracrustal units between the Archaean crustal segments. Minor conductors are also found in the Archaean Karelian Province in Russia, but these appear to be shallow with a small total conductance of a few tens of Siemens. The conductors around the Lapland Granulite Belt, on the other hand, have conductances of several thousands of or even tens of thousands of Siemens.

The Svecofennian-Caledonian border zone and the Caledonides seem to host another set of major conductors in Fennoscandia (JMT in Fig 2). Unfortunately the lack of data hinders to estimate the lateral extent (N-S) of the conductor.

The resistive regions between the belts of conductors provide areas where lower crustal conductivity can be observed mode accurately. Regions where upper crustal conductors extend to the lower crustal (e.g. the Bothnian Bay region and the southernmost part of the Central Finland Granitoid Complex) and where, therefore, the lower crustal conductance may reach a value of several thousands Siemens, can be considered anomalous. Excluding these areas, the lower crustal conductance varies from a few Siemens in southwestern Sweden and much of the Archaean Karelian Province to a few hundreds of Siemens in Central Finland and northwestern parts of the shield. It is interesting to note that the lower crust beneath the Archaean Belomorian Province and the southeastern part of the Archaean Karelian Province is highly resistive. On the contrary, lower crust in the northwestern part of the Karelian Province is more than one decade more conducting even though the lower crust is assumed to be Archaean. Similarly, resistive and conductive lower crust is found in the Proterozoic part of the shield. Lower crust beneath the Central Finland Granitoid Complex (northern part) has a conductance of 200-400 S whereas lower crust in western and southwestern Sweden has conductance below 10 S. This suggests that local geological processes determine the lower crustal conductivity.

4. Concluding remarks

The conductance model is a compilation of a priori local conductivity models obtained primarily from magnetotelluric data. Local 1D and 2D models are further extended (extrapolated) by the use of magnetometer array and other laterally extensive datasets (e.g. airborne electromagnetic surveys). The model is NOT a result from three-dimensional modelling of all existing data and it should be treated as a unified a priori 3D model. New studies (e.g. *Pajunpää et al., 2002; Sokolova et al., 2001*) will definitely change the model and improve it as well as fill "white" regions.

Besides better understanding of the geometry and nature of crustal conductivity variations and, consequently, better understanding of the significance of crustal conductivity in terms of tectono-geological structure and evolution of the crust, the crustal conductivity model (S-map) has shown its power in many other studies such as

- 3D modellings to investigate spatial morphology of model transfer functions and to compare modelling results with BEAR observational data including estimations of EM responses due to the asthenospheric conductivity layer beneath the heterogeneous crust (*Engels et al., 2002; Varentsov et al., 2002*)

- 3D modelling code comparison to examine the accuracy and limitations of several modelling codes with an extremely complex but realistic earth model (*Varentsov et al., 2002*)
- resolution and stability analysis to find limitations and give hints to 1D/2D interpretations (*Smirnov et al., 2002; Varentsov et al., 2002*)
- studies on upper mantle anisotropy (*Lahti et al., 2001; Korja et al., 2002b*)
- source field studies (e.g. *Engels et al., 2002*) using realistic quasi-3D Earth model in multisheet modellings
- geomagnetically induced currents (GIC) in technological systems such as power transmission lines and pipes (*Pulkkinen et al., 2000*)

References

Engels, M., Korja, T., Zhamaletdinov, A.A., Kovtun, A.A., Palshin, N.A., Smirnov, M.Yu., Tokarev, A.D., Asming, V.E., Vanyan, L.L., Vardaniants, I.L., and the BEAR Working Group, 2002. Multisheet modeling of the electrical conductivity structure in the Fennoscandian Shield. *Earth Planets Space*, 54, 559-573.

Flosadóttir A.H., Larsen J.C., Smith J.T., 1997. Motional induction in North Atlantic circulation models. *J. Geophys. Res.,* 102, 10353-10372.

Korja T., Engels M., Zhamaletdinov A.A., Kovtun A.A., Palshin N.A., Smirnov M.Yu., Tokarev A., Asming V.E., Vanyan L.L., Vardaniants I.L., and the BEAR Working Group, 2002a. Crustal conductivity in Fennoscandia - a compilation of a database on crustal conductance in the Fennoscandian Shield. *Earth Planets Space*, 54, 535-558.

Korja, T., Engels, E., Lahti, I., Kobzova, V., Ladanivskyy, B., and the BEAR Working Group, 2002b. Is Lithosphere Anisotropic in the Central Fennoscandian Shield? The book of abstracts in the 16th Workshop on Electromagnetic Induction in the Earth, Santa Fe, New Mexico, USA, 16.-22.06.2002.

Laske G., Masters G., 1997. A global digital map of sediment thickness. *EOS Transactions,* 78, F483.

Lahti I., Korja T., the BEAR Working Group, 2001. Analysis and 1D inversion of the BEAR impedance tensor data. Pp. 95-100 in M.-L. Airo and S. Mertanen (Eds.), *Proceedings of XX National Conference on Geophysics*, Helsinki, Finland. (in Finnish).

NOAA, 1988. Data Announcement 88-MGG-02. Digital relief of the surface of the Earth. NOAA, National Geophysical Data Centre, Boulder, Colorado, (http://www.ngdc.noaa.gov/mgg/global/etopo5.html).

Pajunpää, K., Olafsdottir, B., Lahti, I., Korja, T. and the BEAR Working Group, 2002. Crustal conductivity anomalies in central Sweden and Southwestern Finland. *Geophys. J. Int.*, 150, 695-705.

Pulkkinen, A., Viljanen, A., Pirjola, R., and the BEAR Working Group, 2000. Large geomagnetically induced currents in the Finnish high-voltage power system, *Finnish Meteorological Institute, Report 2000:2*, 99 pp.

Smirnov, M., Engels, M., and the BEAR Working Group, 2002. 2D Inversion Studies in the Fennoscandian Shield: Can Electrical Asthenosphere Be Resolved? The book of abstracts in the 16th Workshop on Electromagnetic Induction in the Earth, Santa Fe, New Mexico, USA, 16.-22.06.2002.

Sokolova, E.Yu., Varentsov, Iv.M., and the BEAR Working Group, 2001. The tracing of the eastern slope of the Ladoga conductivity anomaly on the 1-EV geotraverse. In: The *proceedings of the 6th SVEKALAPKO Workshop, University of Oulu, Department of Geophysics, Report 24,* 10-13..

Varentsov, Iv.M., Engels, M., Korja, T., Smirnov, M.Yu., and the BEAR Working Group, 2002. The generalised geoelectric model of Fennoscandia: a challenging database for long period 3D modelling studies within the Baltic Electromagnetic Array Research (BEAR). *Fizika Zemli.* 9, (accepted, 44 pages).

Crust and Upper Mantle Beneath Fennoscandia as Imaged by the Baltic Electromagnetic Array Research (BEAR)

Toivo Korja[1], Sven-Erik Hjelt[1], Pertti Kaikkonen[1], Elena Kozlovskaya[1], Ilkka Lahti[1], Kari Pajunpää[2], Antti Pulkkinen[3], Ari Viljanen[3], and the BEAR Working Group [4]

[1]Institute of Geosciences, Department of Geophysics, P.O. Box 3000, FIN-90014 University of Oulu, Finland
[2]Finnish Meteorological Institute, Nurmijärvi Geophysical Observatory, Nurmijärvi
[3]Finnish Meteorological Institute, Department of Geophysics, Helsinki
E-mail: toivo.korja@oulu.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). Results indicate the presence of a mantle lithospheric conductor at the depths of 80-120 km and a region of enhanced conductivity below 170-250 km. Below 400 km the mantle conductivity approaches that of global estimates.

Keywords: magnetotellurics, geomagnetic depth soundings, geoelectromagnetic arrays, electrical conductivity, lithosphere, asthenosphere, upper mantle, Fennoscandian Shield

1. Background: the BEAR project

The Fennoscandian Shield is an attractive but challenging environment to investigate the evolution and formation of the continental upper mantle beneath a stable craton and to examine its response to current plate movements. The lithosphere of the Shield evolved in a number of processes from Mesoarchaean to Palaeozoic (3.1 – 0.34 Ga) producing a complex crustal and presumably a complex lithospheric structure. Multiple collisions and rifting events of the lithospheric plates may have left observable traces into the continental lithosphere including e.g. subduction zones, imbrication of lithospheric plates and possible mantle plumes. Geologically and geophysically the crust of the Shield is fairly well studied thus giving solid basis for upper mantle studies. The recently compiled crustal 3D conductivity model over the entire Fennoscandia (*Korja et al., 2002; see also Korja et al. in this volume*) provided an efficient tool for modelling studies. The model has allowed us to produce synthetic 3D electromagnetic data from an extremely complex and demanding but realistic earth model in order to perform various numerical tests on the resolution and stability of modelling and inversion methods and to examine the effects of crustal distortions in long period magnetotelluric responses pertinent to upper mantle conductivity.

The Baltic Electromagnetic Array Research (BEAR) project focuses on determining the electrical conductivity of the upper mantle beneath the Fennoscandian Shield (*Korja and the BEAR Working Group, 2000*). The primary data to probe upper mantle electrical conductivity beneath Fennoscandia are from a shield-wide electromagnetic (EM) array that operated during a two-month-long field campaign on summer 1998 at 46 magnetotelluric and 20 magnetometer sites (Fig. 1). These recordings constitute the first real deep MT array study

---

[4] 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 and Finnish Meteorological Institute in Helsinki and Nurmijärvi), *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, Lviv Centre of Institutes of Space Research and Carpathian Branch of Subbotin Institute of Geophysics in Lviv).

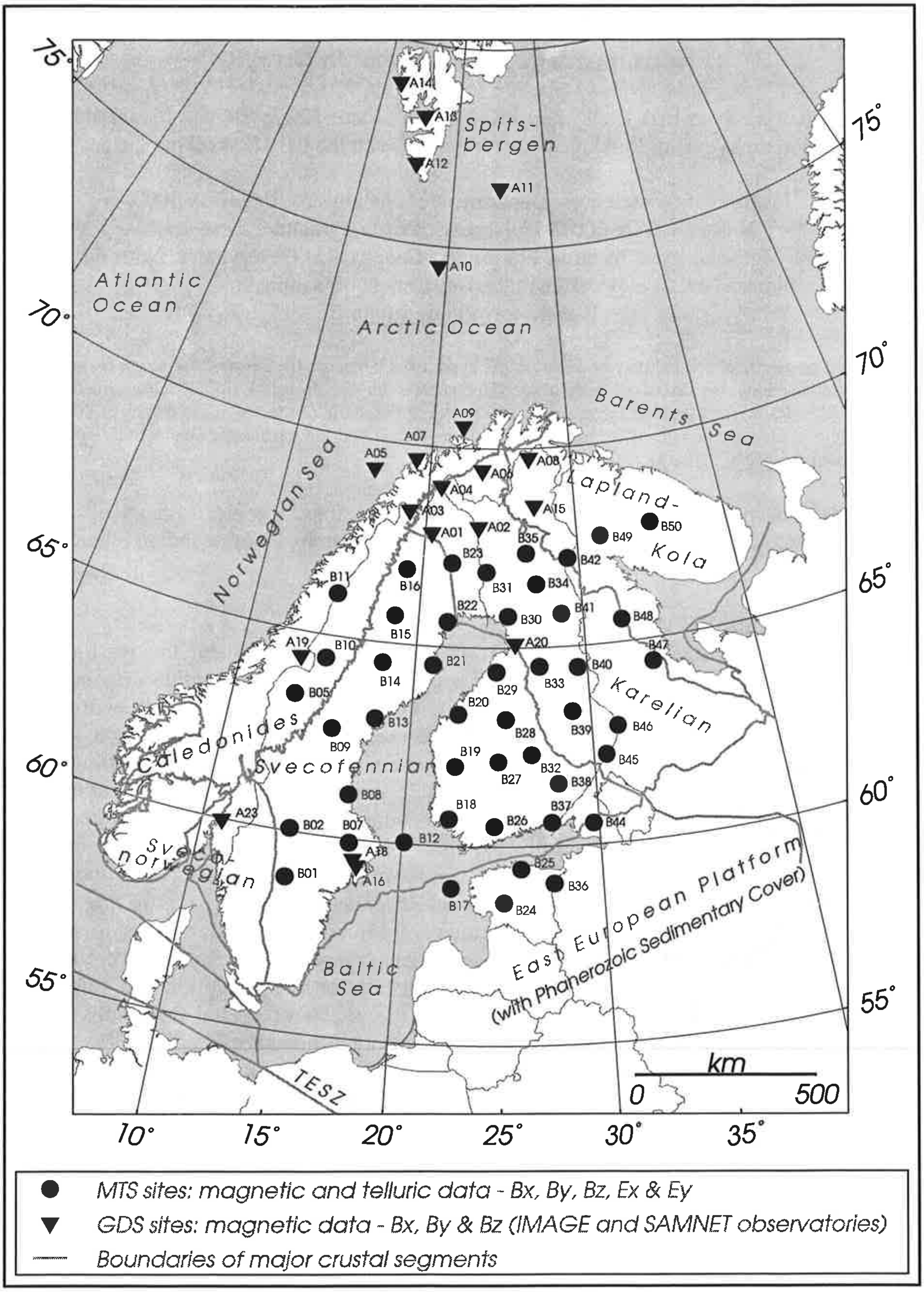


Figure 1. The BEAR array used for ultra deep electromagnetic sounding to probe electrical properties of the upper mantle beneath Fennoscandia. Magnetotelluric and magnetometer measurements have produced data at the period interval of 10 – 20000 s, which allow estimating the upper mantle conductivity to the depth of c. 400-600 km.

ever carried out and as such marks a milestone in the application of modern digital five-component and broad band (period range 1 s –100000 s) magnetotelluric instruments in simultaneous recordings over a very large area of 1000 km x 1400 km.

The acquisition of the array data in subcontinental scale in sub-polar region predetermined the need to develop new approaches and methods and has led to several important achievements with respect to processing and analysis of the BEAR data as well the efforts to analyse distortions caused by the proximity of the source region. Among the most important are

- processing of the BEAR time series data including e.g. the development of a multi-remote-references array method and efforts to determine long period responses and eliminate source effects (*Varentsov et al., 2002b; Sokolova, 2002*)
- complete distortion analysis using different decomposition techniques to determine geoelectric dimensionality of the data, regional geoelectric strike pattern as well distortions caused by crustal heterogeneities (*Lahti et al., 2001; Lahti, 2002*)
- studies on the nature and effects of the non-uniform source with multisheet modelling (e.g. *Engels et al., 2002a,b*) using a realistic quasi-3D Earth model and a realistic 3D source model from equivalent ionospheric currents (*Pulkkinen et al., 2002*)
- studies to understand why and how the source field, which varies spatially and temporally very fast, is smoothed over a large array and long recoding window into a smooth, quasi plane-wave field (*Vanyan et al., 2002*)
- 3D modellings to investigate spatial morphology of model transfer functions and to compare with BEAR observational data including estimations of EM responses due to the asthenospheric conductivity layer beneath the heterogeneous crust (*Engels et al., 2002a,b; Varentsov et al., 2002*)
- 3D modelling code comparison to examine the accuracy and limitations of several modelling codes S-map as a realistic earth model (*Varentsov et al., 2002*)
- resolution and stability analysis to find limitations and give hints to 1D/2D interpretation including analysis of the most sensitive and least distorted magnetotelluric parameters (*Smirnov et al., 2002; Varentsov et al., 2002).*

Although these results are very important steps on our (electromagnetic) stairways to upper mantle, they are more of internal interest. Therefore we concentrate below on the upper mantle conductivity structure beneath Fennoscandia as imaged by the BEAR array.

2. Upper mantle conductivity according to the BEAR research

In the following we discuss separately four major features of the upper mantle conductivity beneath Fennoscandia viz. possible upper mantle electrical anisotropy ("anisotropy"), evidence for a conducting layer in mantle lithosphere ("100 km"), enhanced conductivity at the depths corresponding to a possible asthenospheric channel ("asthenosphere"), and mantle conductivity below the lithosphere-asthenosphere system (tectosphere) at depths greater than 400 km ("below 400 km").

Upper mantle anisotropy. Electrical anisotropy can be assumed if regional geoelectrical strikes are stable over a large area and if two orthogonal phases show simultaneously a stable phase split (phase difference). Magnetotelluric data from pre-BEAR surveys (0.1-500 s; *Korja & Koivukoski, 1994*) and from the BEAR array (10 -10000 s; *Lahti et al., 2001; Lahti, 2002*) exhibit strong and stable anisotropic features in the central part of the Fennoscandian Shield including, in particular, the Central Finland Granitoid Complex (CFGC; size c. 200 km x 200 km) in the Palaeoproterozoic Svecofennian Orogen and the Archaean Karelian Province (KP). The two regional phases from decomposition analysis (McNiece-Jones extension of Groom-Bailey decomposition and Bahr's decomposition; *Groom and Bailey, 1989; McNiece and Jones, 2001; Bahr, 1991*) begin to depart at the periods of 1-5 s and a maximum split of c. 40-50 degrees is reached at the period interval of 200-600 s. At longer periods, the split

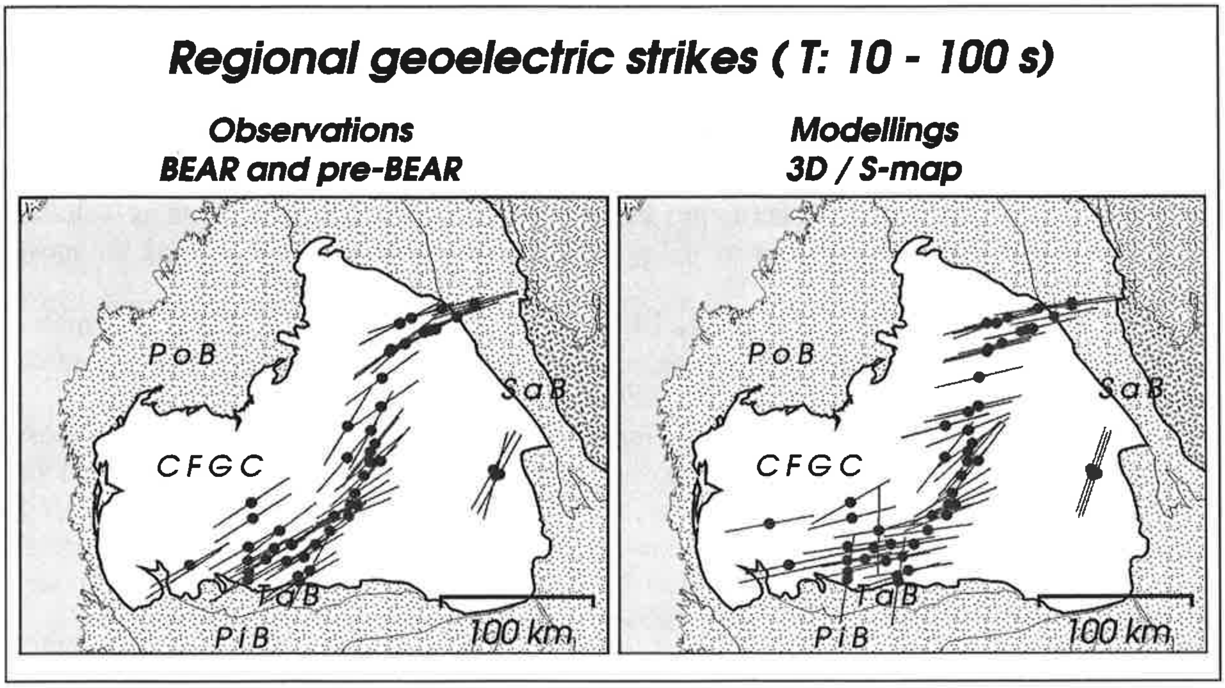


Figure 2. Upper mantle anisotropy. Electrical anisotropy can be assumed if regional geoelectrical strikes are stable over a large area and if two orthogonal phases show simultaneously a stable phase split (phase difference). Here we show observed (left) and modelled (right) regional geoelectric strikes in central Finland. Regional strikes in both cases are obtained using extended Groom-Bailey decomposition technique (Groom and Bailey, 1989; McNiece and Jones, 1996) at the period interval of 10-100 s corresponding roughly with the depths of 50-100 km. Observational data include pre-BEAR and BEAR data. Model responses are obtained from 3D forward modelling using isotropic 3D crustal model (S-map) and layered upper mantle. Nearly identical spatial behaviour of regional strikes suggests that stable strikes (indicative of anisotropy if associated with phase split) are due to isotropic 3D crust rather than genuinely anisotropic upper mantle.

slightly decreases but remains as high as c. 20 degrees at the periods of 8000-10000 s. Decomposition analysis yields also very stable regional strikes of c. N55E (+- 90 degrees) in CFGC and even in the highly resistive KP more to the northeast. N55E is the direction of the maximum phase (minimum impedance) and would therefore indicate the azimuth of anisotropy in the case of anisotropic deep lithosphere. At the same time, real induction arrows are rather short (< 0.2) though not negligible at periods shorter than 1000 s. At longer periods, the length of arrows increase reaching an average length of 0.4 in CFGC and 0.5-0.7 in KP at the periods of 6000-8000 s.

To distinguish between the two alternative models, viz. a model that contains 3D *isotropic* lithosphere or a model that requires the lithosphere (or a part of it) be *genuinely anisotropic*, numerical 3D modellings (*Korja et al., 2002*) and analogue scale modelling experiment (*Kobzova et al., 2002)* using 3D crustal model and an average 1D upper mantle reference model for Fennoscandia (*Korja et al., 2002*) were carried out. The model data were analysed in a same manner as the observed data, e.g. regional strikes and phase split were determined. Initial results suggest that (i) the stable regional strike of N55E (+-90 degrees) is primarily due to the effects of crustal conductors surrounding the Central Finland Granitoid Complex and bordering the Archaean Karelian Province as shown in Fig. 2 and (ii) a major part of the large phase split is due to known conductors although a minor part cannot be explained solely by the known crustal conductors leaving a possibility to anisotropic lithosphere. The BEAR data, however, indicates the presence of conductivity heterogeneities in upper mantle that may also contribute to observed anisotropic signature of magnetotelluric data. Therefore it appears that strong anisotropic features in data are due to isotropic 3D crust rather than genuinely anisotropic upper mantle.

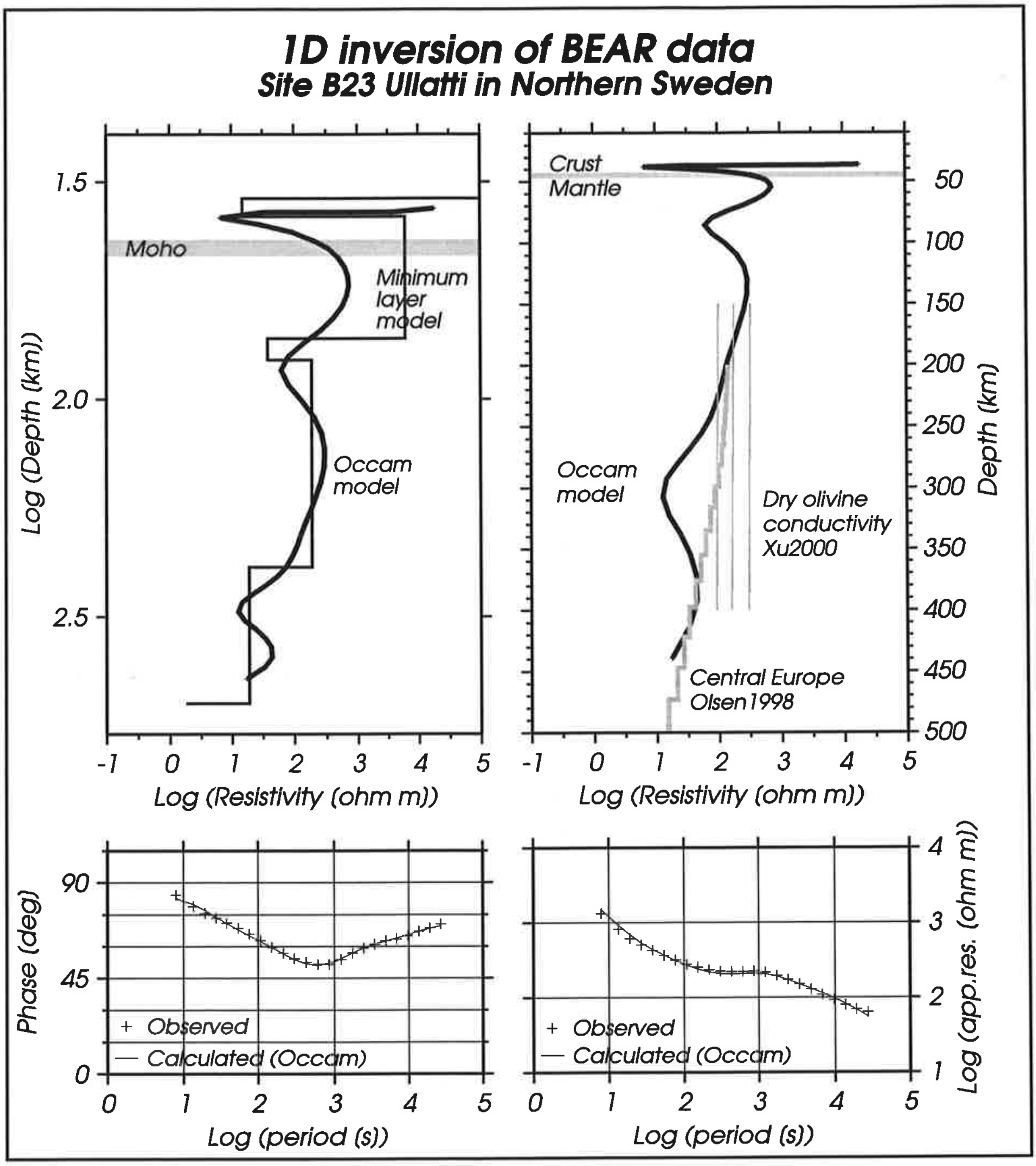


Figure 3. Main features of the upper mantle conductivity in Fennoscandian according to BEAR research. Upper panels show 1D inversion results at site B23 (Ullatti) in northern Sweden using the determinant of the impedance tensor as the inverted data. Results are representative for most of the BEAR sites. Lower panels show the quality (fit) of the inversion for phases (lower left) and apparent resistivities (lower right). Observed responses are measured responses from site B23 and calculated responses are model responses from 1D Occam inversion. Upper left panel shows, in a logarithmic scale, a smooth resistivity model from Occam inversion (thick black line) and a rough minimum layer model (grey line). Both models yield similar fit to the data. Minimum layer model indicates that a seven-layer-model (three conductors) is required to fit the data. Upper right panel shows, in a linear scale, Occam model together with (i) the observed upper mantle conductivity model for Central Europe (Olsen, 1998) based on the inversion of magnetic observatory data and (ii) upper mantle petrophysical model based on laboratory data (Xu et al, 2000) assuming dry olivine mineralogy at appropriate continental mantle geotherm +-100 C-degrees. Two main features are (a) conducting layer in mantle lithosphere at the depth of c. 90 km and (b) enhanced conductivity deeper in the upper mantle (250-4350 km), which cannot be explained by laboratory data. It is also noteworthy that deeper than 400 km, the BEAR-model coincides with the model of Central Europe.

Mantle lithospheric conductor at the depth of 100 km. 1D inversions of the determinant data from site B23 (Fig. 3) and from most of other BEAR sites (*Lahti et al., 2001*) resolve a conducting layer at the depths ranging from 70 km to 120 km. The Occam (*Constable et al., 1987*) model given in Fig. 3 is the smoothest model (a minimum structure model) that fits the data within an expected tolerance (fit of the model responses to observed responses is given in the lower panels of Fig. 3) whereas the "minimum layer model" (*Pirttijärvi et al., 1998*) has the least number of layers that also fit the data within an expected tolerance. Both models fit the data nearly identically (only a fit of Occam model is shown in Fig. 3). Both models reveal similar conductivity structure i.e. data requires the presence of three conducting layers in the middle/lower crust (S = c. 250 S for the site B23), in the mantle lithosphere (S = c. 200 S) and deeper in the upper mantle (S = c. 4000 S). The mantle lithospheric conductors at the depth interval from 80 km to 120 km are observed in areas where crustal conductivity is small. Due to relatively low conductance of the mantle lithospheric conductor (around 200 S) it is not possible to detect the mantle conductor if the conductance of crust exceeds 200 S. Further examination suggests that the depth to the top of the conductor is greater in the western and eastern parts of the array whereas shallower in the central part of the array along a NS-trending zone. The latter observation and, in general, better determination of the geometry of the mantle lithospheric conductor requires more detailed inversions as well as additional MT measurements between the BEAR sites in order to increase spatial sampling.

Asthenosphere or another upper mantle conductor deeper than 170 km. One-dimensional inversion models for site B23 indicate that electrical conductivity is enhanced considerably below the depth of c. 240 km (Fig. 3). Occam inversion resolves a layer i.e. resistivity is increasing below the conductor whereas the other approach resolves only the top of the conductor because only seven layers were allowed in inversion. The conductance of the deeper layer is c. 4000 S as estimated from the Occam model. Similar results are obtained by 1D inversions of the determinant data from other sites (*Lahti et al., 2001*) with an exception that at some sites only the top of the conducting region is resolved because the longest periods used in inversion (c. 20000 s) cannot penetrate through the conductor. Similarly, a comparison of BEAR data with the models data from multisheet and 3D volume modellings (e.g. *Engels, 2002a,b; Varentsov et al., 2002a,b*) indicate that upper mantle must contain an excess of up to 5000 S of conductive material with respect to an a priori average 1D reference model of Fennoscandia (*Korja et al., 2002*). 3D modellings indicate that the upper mantle should be less conductive in the northern part of the Shield whereas more conductive in central and southern parts. Topography of the top of the conductor or the region of enhanced conductivity is variable but according to 1D inversions the top of the conductor seem to be deeper in the central part of the Shield and shallower in western and eastern parts of the array.

Comparison of the Occam model for site B23 with (i) the observed upper mantle conductivity model for Central Europe (Olsen, 1998) and (ii) upper mantle petrophysical model based on laboratory data (Xu et al, 2000) is given in Fig. 3 (upper right panel). Olsen's model is based on the inversion of magnetic observatory data and hence represents and average upper mantle conductivity beneath Central Europe. In petrophysical models dry olivine mineralogy at appropriate continental mantle geotherm with +-100 C-degrees variations is used. Comparison with Olsen's model show that upper mantle at the depth interval of 250-350 km clearly differs from that in Central Europe – upper mantle in Fennoscandia contains conducting material c. 5000 S more than beneath Central Europe. The enhanced conductivity cannot be explained olivine conductivity as shown by the comparison of BEAR models with the petrophysical model. Partial melting and carbon films are also unlikely explanations for enhanced conductivity and thus the most likely candidate is the enhanced content of volatiles (H_2O and CO_2).

Upper mantle below 400 km. Long period (T > 4h) impedance phases derived from the BEAR data set coincides with the phases obtained earlier for Central Europe and with the global estimates (e.g. *Olsen, 1998; Semenov, 1999*) supporting the proposed spherical symmetry of conductivity structure at depths below c. 400 km. This is clearly seen in Fig. 3, where the BEAR-model from site B23 coincides with the model of Central Europe.

References

Bahr, K., 1991. Geological noise in Magnetotelluric data: A classification of distortion types. *Phys. Earth Planet. Int.* 66, 24-38.

BEAR Working Group and Varentsov, Iv.M., 2001. On the construction and verification of the volume geoelectrical model of Fennoscandia in the BEAR project. Pp. 10-13 in the proceedings of the 6th SVEKALAPKO Workshop, University of Oulu, Department of Geophysics, Oulu, Finland, Report 24.

Constable, S.C., Parker, R.L., Constable, C.G., 1987. Occam's inversion: a practical algorithm for generating smooth models from electromagnetic sounding data. *Geophysics* 52, 289-300.

Engels, M., Korja, T., Zhamaletdinov, A.A., Kovtun, A.A., Palshin, N.A., Smirnov, M.Yu., Tokarev, A.D., Asming, V.E., Vanyan, L.L., Vardaniants, I.L., and the BEAR Working Group, 2002a. Multisheet modelling of the electrical conductivity structure in the Fennoscandian Shield. Earth Planets Space, 54, 559-573.

Engels, M. and the BEAR Working Group, 2002b. On multisheet and volume 3D modelling of the Baltic Shield. In A. Hoerdt and J.B. Stoll (Eds.), Protokoll Koll. Elektromagnetische Tiefenforschung Burg Ludwigstein, September, 2001, University of Bonn, Grmany, 20 p.

Groom, R.W. and Bailey, R.C., 1989. Decomposition of magnetotelluric impedance tensor in the presence of local three-dimensional galvanic distortion. J. Geophys. Res., 94, 1913-1925.

Kobzova, V.M., Ladanivsky, B., Korepanov, V., Korja, T., and the BEAR Working Group, 2002. Physical modelling of MT field in Central Finland. An abstract in the 27th EGS General Assembly (CD-ROM), Nice, France.

Korja, T. and Koivukoski, K., 1994. Magnetotelluric investigations along the SVEKA profile in central Fennoscandian Shield, Finland. *Geophys. J. Int.,* 116, 173-197

Korja, T. and the BEAR Working Group, 2000. The structure of the crust and upper mantle in Fennoscandia as imaged by electromagnetic waves. Pp. 25-34 in L. Pesonen, A. Korja and S.-E. Hjelt (Eds.), Lithosphere 2000. Program and extend abstracts, Institute of Seismology, University of Helsinki, Helsinki, Finland, Report S-41, 179 pp.

Korja T., Engels M., Zhamaletdinov A.A., Kovtun A.A., Palshin N.A, Smirnov M.Yu., Tokarev A., Asming V.E., Vanyan L.L., Vardaniants I.L., and the BEAR Working Group, 2002. Crustal conductivity in Fennoscandia - a compilation of a database on crustal conductance in the Fennoscandian Shield. Earth Planets Space, 54, 535-558.

Korja, T., Engels, E., Lahti, I., Kobzova, V., Ladanivskyy, B., and the BEAR Working Group, 2002. Is Lithosphere Anisotropic in the Central Fennoscandian Shield? The book of abstracts in the 16th Workshop on Electromagnetic Induction in the Earth, Santa Fe, New Mexico, USA, 16.-22.06.2002.

Lahti I., Korja T., the BEAR Working Group, 2001. Analysis and 1D inversion of the BEAR impedance tensor data. Pp. 95-100. In M.-L. Airo and S. Mertanen (Eds.), Proceedings of XX National Conference on Geophysics, Helsinki, Finland. (in Finnish).

Lahti, I., 2002. Analysis and decomposition of the magnetotelluric transfer functions from the Baltic Electromagnetic Array Research (BEAR). Phil. Lic. Thesis, Department of Geopysics, University of Oulu, 81 p. (in English)

McNiece, G and Jones, A.G., 2001. Multi-site, multi-frequency tensor decomposition of magnetotelluric data. *Geophysics*, 66, 158-173.

Olsen N., 1998. The electrical conductivity of the mantle beneath Europe derived from C-responses from 3 to 720 hr . *Geophys. J. Int.*,133, 298-308.

Pirttijärvi, M., Verma, S.K. and Hjelt, S.-E.,1998. Inversion of transient electromagnetic profile data using conductive finite plate model. *Applied Geophysics*, 38, 181-194.

Pulkkinen, A., Amm., O., Viljanen, A., and the BAR Working group., 2002. Ionospheric equivalent current distribution determined with the method of elementary current systems. J. Geophys. Res. (accepted)

Semenov V.Yu., 1998. Regional conductivity structures of the Earth's mantle. *Publ. Inst. Geoph. Pol. Acad.*, C-65 (302), 122p.

Smirnov, M., Engels, M., and the BEAR Working Group, 2002. 2D Inversion Studies in the Fennoscandian Shield: Can Electrical Asthenosphere Be Resolved? The book of abstracts in the 16th Workshop on Electromagnetic Induction in the Earth, Santa Fe, New Mexico, USA, 16.-22.06.2002.

Sokolova, E.Yu., 2002. Development of new methods for the analysis of deep EM sounding data and their application in regions with complicated geoelectric structure. Cand. Sc. Dissertation. M.: OIFZ, Moscow University, Moscow, Russia,. 205 p. (in Russian).

Vanyan, L.L., Palshin, N.A., Levitin, A., and the BEAR Working Group, 2002. Two type of the ionospheric source in high latitudes. The book of abstracts in the 16th Workshop on Electromagnetic Induction in the Earth, Santa Fe, New Mexico, USA, 16.-22.06.2002.

Varentsov Iv.M., Sokolova E.Yu, Martanus E.R. et al., 2002a. Methods of the construction of a system of transfer operators of natural EM field for the high-latitude simultaneous sounding array (the BEAR experiment). *Fizika Zemli, 9.* (accepted, in Russian).

Varentsov, Iv.M., Engels, M., Korja,T., Smirnov, M.Yu., and the BEAR Working Group, 2002b. The generalised geoelectric model of Fennoscandia: a challenging database for long period 3D modelling studies within the Baltic Electromagnetic Array Research (BEAR). *Fizika Zemli. 9,* (accepted, 44 pages).

The Bodom and Obbnäs Rapakivi Granites, Southern Finland: Distinct Composition Imply a Paleoproterozoic Terrane Boundary

Paula Kosunen[1], O. Tapani Rämö[1] and Matti Vaasjoki[2]

[1]Department of geology, P.O. Box 64, FIN-00014 University of Helsinki, Finland
[2]Geological Survey of Finland, P.O. Box 96, FIN-02151 Espoo, Finland
E-mail: paula.kosunen@helsinki.fi

Two small, coeval rapakivi granite plutons west of Helsinki, Bodom and Obbnäs, display important compositional and petrographic differences that relate to different sources and crystallization histories. They also imply the presence of an east-west trending Paleoproterozoic terrane boundary.

Keywords: rapakivi granite, Paleoproterozoic, geochemistry, isotope geology

1. Introduction

The Bodom and Obbnäs rapakivi granite plutons were emplaced along the Porkkala-Mäntsälä shear zone ~1.64 Ga ago *(Vaasjoki, 1977; Kosunen, 1999)*. They cut sharply the surrounding Paleoproterozoic (~1.9 Ga) Svecofennian metamorphic bedrock that belongs to the accretionary arc complex of southern Finland with variably migmatized volcanic and sedimentary rocks, syn-orogenic ~1.89 – 1.88 Ga granitoids, and late-orogenic ~1.84 – 1.82 Ga granites *(Korsman et al., 1999)*.

The Bodom pluton consists of four main granite types (Fig. 1). A sequence of porphyritic granites (hornblende granite, hornblende-biotite granite, biotite granite) with occasional plagioclase-mantled alkali feldspar megacrysts forms the central and northern parts of the pluton, while the southwestern part is occupied by a medium-grained, mostly even-grained biotite-hornblende granite. The Obbnäs pluton is lithologically more monotonous with the bulk of the intrusion consisting of coarse-grained porphyritic hornblende-biotite granite, which shows a tendency of becoming slightly more mafic toward the southwest. In the south-westernmost part of the pluton, a more fine-grained biotite-hornblende granodiorite is present and forms variable mingling structures with the main porphyritic hornblende-biotite granite. The Obbnäs pluton shows a strong fabric with both magmatic and brittle components *(Kosunen, 1999)*.

Tholeiitic diabase dikes are found adjacent to the Obbnäs and Bodom plutons. The dikes are vertical or subvertical, generally $\leq$ 1 to 15 m wide, and can be traced in outcrops and on aeromagnetic maps for several kilometers (sometimes for several tens of kilometers; *Laitakari et al., 1996*). The dikes have not been found to cut the granites of the Bodom and Obbnäs plutons and they may thus be somewhat older than the granites.

2. Geochronology

Our new zircon U-Pb isotopic data *(Kosunen et al., in prep.)* support the previous age results *(Vaasjoki, 1977)* and suggest an emplacement age of 1640 Ma for the Obbnäs granite. A similar age of accessory titanite indicates that no significant thermal event has affected the Obbnäs granite since then. The Bodom granite contains two generations of zircon with different U-Pb ages, which is rarely observed in the Finnish rapakivi granites. Either the conditions in the magma chamber were not sufficiently severe in order to reset inherited zircons, or the Bodom granite crystallized in several stages. In the latter case, the older, dark zircon variety (~1650 Ma) may record a time at which the rapakivi melt, still accumulating at a lower crustal level, became temporarily over saturated in zirconium and the younger, light brown variety may have formed at the onset of the final solidification of the rapakivi magma

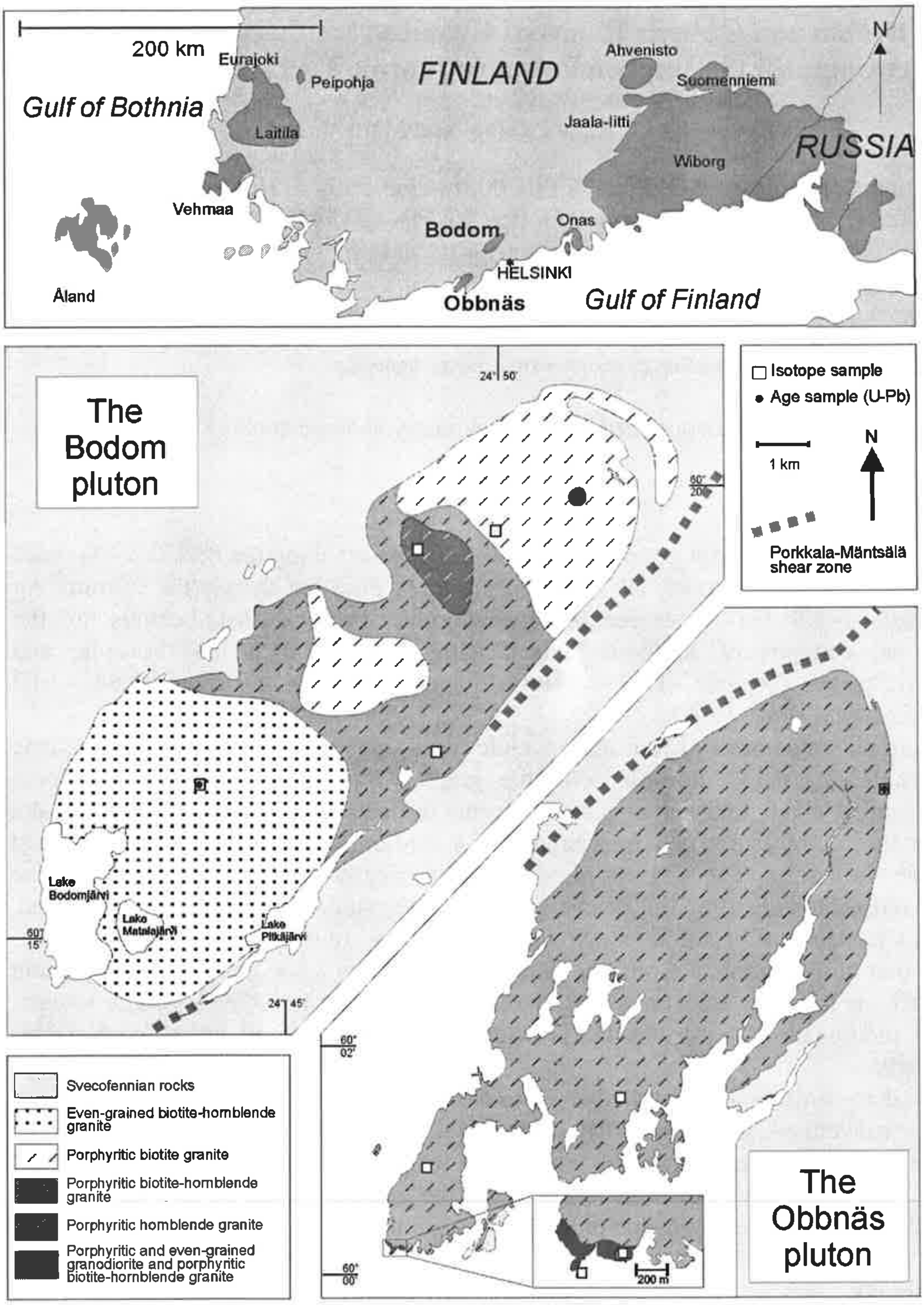


Figure 1. Location and lithology of the Bodom and Obbnäs rapakivi granite plutons.

after its intrusion into cooler country rocks 1637.5 ± 2.2 Ma ago. Thus the age difference of c. 15 Ma observed in the two zircon varieties may convey a rough idea of the time span required for the formation of a rapakivi melt in the lower crust and its subsequent intrusion to higher crustal levels. This will be examined further using the SIMS technique.

3. Geochemistry

The Bodom and Obbnäs granites are peraluminous to metaluminous and subalkaline, with average A/CNK values of 1.00 and 1.05, and K_2O/Na_2O of 2.22 and 2.12, respectively, and they show the geochemical characteristics of A-type granites and a tectonomagmatic affinity of within-plate granites. The Obbnäs pluton is, however, less evolved in character (generally lower content of K_2O, higher values of TiO_2, CaO, MgO, and P_2O_5) than Bodom, and even though the rocks of both plutons show the high overall Fe/Mg typical of the rapakivi granites, there is a marked difference, with an average FeO*/(FeO*+MgO) of 0.94 for Bodom and 0.87 for Obbnäs. The Bodom granites contain, in general, higher amounts of fluorine and trace and rare earth elements than the porphyritic biotite-hornblende granite of Obbnäs. Two significant exceptions are Ba and Sr, the amounts of which are relatively high in the Obbnäs granite. In consequence, the Rb/Ba and Rb/Sr of the Obbnäs granite are considerably lower than the respective ratios of the Bodom granites (and of the Finnish rapakivi granites in general) (Table 1).

4. Mineral chemistry

The alkali feldspar of the Bodom and Obbnäs granites is crosshatched microcline with a rather uniform composition (Or_{87} to Or_{97}), but the alkali feldspar from the Obbnäs pluton contain higher amounts of barium (Table 1). Composition of plagioclase varies according to the rock type, but the plagioclase from Obbnäs (An_{23-37}) are, in general, more calcic than the plagioclase from Bodom (An_{10-30}). In addition to the feldspars and quartz, the granites of the two plutons contain biotite and/or amphibole as their major constituents. The biotites are iron-rich, but reflect the overall difference in the Fe/Mg budget of the two plutons: the Fe/(Fe+Mg) of the biotites of the porphyritic hornblende-biotite granite of Obbnäs varies from 0.75 to 0.80, while in the biotites from the Bodom granites the values range from 0.86 to 0.98. Amphibole composition of the Bodom and Obbnäs plutons ranges from hastingsitic hornblende to hastingsite, which also are rich in iron, with Fe/(Fe+Mg) ranging from 0.96 to 0.99 in Bodom and from 0.83 to 0.89 in Obbnäs. The granites of both plutons contain accessory apatite, zircon, fluorite, and ilmenite; metamict allanite is also present (except in the porphyritic hornblende granite of Bodom). In addition, the porphyritic biotite-hornblende granite of Obbnäs contains titanite and minor ilmenomagnetite.

Zircon saturation modeling *(Watson and Harrison, 1983)* yields temperatures ranging from ~ 950°C to ~ 830°C for the Bodom and Obbnäs plutons (Table 1), while the biotite-ilmenite stability curves *(Wones and Eugster, 1965)* record temperatures from ~ 750°C to ~ 820°C for the porphyritic hornblende-biotite granite of Bodom, slightly higher temperatures from ~ 860°C to ~ 920°C for the porphyritic hornblende-biotite granite of Obbnäs, and temperatures of over 900°C for the Obbnäs granodiorite. Oxygen fugacities calculated from biotite-ilmenite stability curves for the porphyritic hornblende-biotite granites of Bodom and Obbnäs and the Obbnäs granodiorite range, in general, from –16.3 to -12.5 log fO_2. The fugacities calculated for the porphyritic hornblende-biotite granite of Bodom are clearly lower (from –16.3 to –15.2 log fO_2) than the fugacities for the porphyritic hornblende-biotite granite of Obbnäs (from –14.2 to –13.4 log fO_2). These low overall oxygen fugacities, and the more reduced nature of the Bodom pluton, are also compatible with amphibole compositions. Average pressures of about 6 kbar and 7 kbar for the Bodom and Obbnäs plutons, respectively, are suggested by the aluminum-in-hornblende barometer *(Hammerstrom and Zen, 1986)*; these pressures are, however, probably overestimates, due to the high Fe/(Fe+Mg) of the amphiboles.

Table 1. Summary of the differences between of the Bodom and Obbnäs plutons *(Kosunen et al., in prep.)*.

Pluton	**Bodom**		**Obbnäs**	
Lithology	porphyritic hbl granite porphyritic bt-hbl granite porphyritic bt granite even-grained granite		porphyritic bt-hbl granite porphyritic granodiorite even-grained granodiorite	
Whole-rock chemistry (average values or range)				
Granites	syenogranites		syeno/monzogranite	
FeOT/(FeOT+MgO)	0.94		0.87	
Rb/Ba	0.44		0.12	
Rb/Sr	3.14		0.83	
F (ppm)	3400		1200	
Ga/Al	3.8 - 5.3		3.0 - 4.2	
$(La/Yb)_N$	19.0 (porphyritic granites)		20.8	
Eu/Eu*	0.36 (porphyritic granites)		0.53	
Mineralogy and mineral chemistry of the granites (average values or range)				
Fe-Ti oxides	ilmenite		ilmenite + ilmenomagnetite, titanite	
Alkali feldspar				
BaO (wt%)	0.29		0.55	
Plagioclase				
An	10 - 30 %		23 - 37 %	
Biotite				
Fe/(Fe+Mg)	0.86 - 0.98		0.75 - 0.80	
Amphibole				
Fe/(Fe+Mg)	0.96 - 0.99		0.83 - 0.89	
Intensive parameters				
Temperature (°C)				
Zr- saturation	porphyritic hbl granite	914	881	porphyritic bt-hbl granite
	porphyritic bt-hbl granite	871	890	porphyritic granodiorite
	porphyritic bt granite	836	900	even-grained granodiorite
	even-grained granite	885		
Biotite-ilmenite stability	750 - 820 (bt-hbl granite)		860 - 920 (bt-hbl granite)	
Pressure (kbar)				
Aluminum in hornblende	6		7	
Oxygen fugacity ($\log fO_2$)				
Biotite-ilmenite stability	-16.3 to -15.2		-14.2 to -13.4	
Isotope geochemistry				
Nd				
ε_{Nd}	-1.0 to -0.8, avrg -0.8 ± 0.1 (n = 4)		-2.0 to -1.2, avrg -1.7 ± 0.3 (n = 6)	
T_{DM}	avrg 2.02 ± 0.04 Ga		avrg 2.07 ± 0.04 Ga	
Sr				
$^{87}Rb/^{86}Sr$	2.6 - 12.7		0.6 - 5.1	
Sr_I	0.7040 ± 0.0047		0.70469 ± 0.00004	
Errorchron age	1601 ± 40 Ma		1555 ± 31 Ma	

5. Isotope geochemistry

The Nd and Sr isotopic systematics also imply differences between the Bodom and Obbnäs plutons (*Kosunen et al., in prep.*; Table 1). The initial ε_{Nd} (at 1640 Ma) values of Obbnäs (–2.2 to –1.2, mean –1.7 ± 0.3 [1S.D.], n = 6) are slightly, but probably significantly, lower than those of Bodom (–1.0 to –0.8, mean –0.8 ± 0.1, n = 4). This difference is consistent with the east-west oriented Nd isotopic discontinuity *(see Fig. 7 in Rämö et al., 2001)* that presumably separates a slightly more juvenile terrain in the north from a less radiogenic terrain in the south. Our data suggest that the boundary runs between the Obbnäs and Bodom plutons. There may also be a difference in the initial Sr isotopic composition of the two plutons, although the highly-radiogenic character of the Bodom pluton ($^{87}Rb/^{86}Sr$ 2.6-12.7, as opposed to 0.6-5.1 in the Obbnäs pluton) precludes detailed comparison (Table 1). However, the relatively well-defined initial ratio of the Obbnäs pluton, 0.70469 ± 0.00004, is presumably lower than that of the Suomenniemi rapakivi pluton (initial ratio 0.7066 ± 0.0023; *Rämö,* *1999*). The Suomenniemi pluton is located north of the Nd isotopic discontinuity and is geochemically akin to the Bodom pluton.

References

Hammerstrom, J.M. and Zen, E., 1986. Aluminum in hornblende: an empirical igneous geobarometer. *American Mineralogist* ***71****: 1297-1313.*

Korsman, K., Korja, T., Pajunen, M. and Virransalo, P., 1999. The GGT/SVEKA transect: structure and evolution of the continental crust in the Paleoproterozoic Svecofennian orogen in Finland. *International Geology Review* ***41*** *(4), 287-333.*

Kosunen, P., 1999. The rapakivi granite plutons of Bodom and Obbnäs, southern Finland: petrography and geochemistry. *Bulletin of the Geological Society of Finland* ***71****, 275-304.*

Laitakari, I., Rämö, T., Suominen, V., Niin, M., Stepanov, K., and Amantov, A., 1996. Subjotnian: Rapakivi granites and related rocks in the surroundings of the Gulf of Finland. Expl. to the Map of Precambrian basement of the Gulf of Finland and surrounding area 1:1 mill (Tapio Koistinen ed.). *Geological Survey of Finland, Special Paper* ***21****: 59-97.*

Rämö, T., 1999. Sr isotopic composition of Finnish rapakivi granites: the Suomenniemi batholith. *Bulletin of the Geological Society of Finland* ***71*** *(2), 339-345.*

Rämö, O. T., Vaasjoki, M., Mänttäri, I., Elliott, B.A., and Nironen, M., 2001. Petrogenesis of the post-kinematic magmatism of the Central Finland Granitoid Complex I: radiogenic isotope constraints and implications for crustal evolution. *Journal of Petrology* ***42*** *(11), 1971-1993.*

Vaasjoki, M., 1977. Rapakivi granites and other postorogenic rocks in Finland: their age and the lead isotopic composition of certain associated galena mineralizations. *Geological Survey of Finland Bulletin* ***294****, 64 p.*

Watson, E.B., and Harrison, T.M., 1983. Zircon saturation revisited: temperature and composition effects in a variety of crustal magma types. *Earth and Planetary Science Letters* ***64****: 295-304.*

Wones, D.R. and Eugster, H.P., 1965. Stability of biotite: experiment, theory, and application. *American Mineralogist* ***50****: 1228-1272.*

The Research Project: 3-D Crustal Model of the Finnish Part of SVEKALAPKO Research Area

Elena Kozlovskaya [1], Markku Pirttijärvi [1], Sven-Erik Hjelt [1], Juha Korhonen [2], Seppo Elo [2] and Jukka Yliniemi [3]

[1] Department of Geophysics, P.O. Box 3000, FIN-90014 University of Oulu, Finland
[2] Geological Survey of Finland, P.O. Box 96, FIN-02151 Espoo, Finland
[3] Sodankylä Geophysical Observatory, Oulu Unit, P.O. Box 3000, FIN-90014 University of Oulu, Finland
E-mail: markku.pirttijarvi@oulu.fi

The objective of this research project is to construct a three-dimensional geophysical model of the crust beneath the SVEKALAPKO research area. The model will be based on the joint use of the seismic data (P- and S-waves) from the SVEKALAPKO tomography experiment, and the potential field data (gravity and magnetics), digital geological maps and the petrophysical database of the Geological Survey of Finland.

Keywords: joint inversion, potential fields, seismic tomography, crust, Fennoscandian Shield

1. Introduction

SVEKALAPKO (*Hjelt and Daly, 1996, Bock et al., 2001*) is one of the key projects of the EUROPROBE programme supported by the European Science Foundation. The project studied the Archaean and Palaeoproterozoic evolution of the Fennoscandian (Baltic) Shield. The SVEKALAPKO study area and the main geological units are illustrated in Figure 1. The SVEKALAPKO deep seismic tomography experiment, which was realized in 1998-1999, produced a large amount of detailed information about the deep structure of the crust and upper mantle (*Hyvönen and Malaska, 2000, Sandoval et al., 2002, Kozlovskaya et al., this volume*), the processing and the analysis of which is still unfinished. The question is: "How can the 3-D crustal P- and S-wave velocity models of the SVEKALAPKO area provide the necessary constraints for the potential field interpretation?"

The main target of the Crustal Model Program of the Geological Survey of Finland (GSF) is to answer questions concerning the geometrical, mineralogical and geological nature of the potential field anomaly sources of the Finnish lithosphere (*Korhonen and Säävuori, 2000*). This problem is difficult to solve without seismic data due to the principal non-uniqueness of potential field source models. From another point of view the ability of seismic methods to detect the geometry of the deep structures including sources of potential field anomalies is limited. The principal reason for this is that seismic methods, especially the reflection technique, are sensitive mainly to the horizontal and sub-horizontal boundaries and cannot reveal vertical and steeply dipping interfaces. On the contrary, the potential field data are more sensitive to vertical and sub-vertical boundaries and lateral variations of physical properties. The combination of reflection seismics with wide-angle reflection and refraction data and with tomography studies can reveal the geometry of deep structures in the crust. Integration with potential field data can in turn provide the necessary constraints for correct interpretation of seismic data.

The main condition for the joint interpretation of experimental seismic and gravity data is the established dependence between rock density and seismic velocity. Investigation of this relationship is one of the fundamental tasks in crustal studies (*Elo, 2000*). Usually this relationship is compiled from measurements of rock density and seismic velocity made under laboratory conditions. One of the main difficulties in using such relationships is a significant scattering around the mean value for all types of lithospheric rocks. Because of the scattering the density-velocity relation has to be regarded as a statistical dependence rather than as a functional relationship. Ultimately, the relationship may have to be solved separately for various geological provinces. The joint interpretation of seismic and potential field data is a new and promising approach in trying to solve this important problem.

Integration of seismic data with potential field data and results of petrophysical studies was successfully used in the EUROPROBE/EUROBRIDGE project. Integrated velocity-density models along EUROBRIDGE wide-angle reflection and refraction seismic profiles were obtained by joint inversion of seismic and gravity data constrained by a nonlinear density-velocity relationship based on gravity data inversion (*Kozlovskaya and Yliniemi, 1999, Kozlovskaya et al., 2002*).

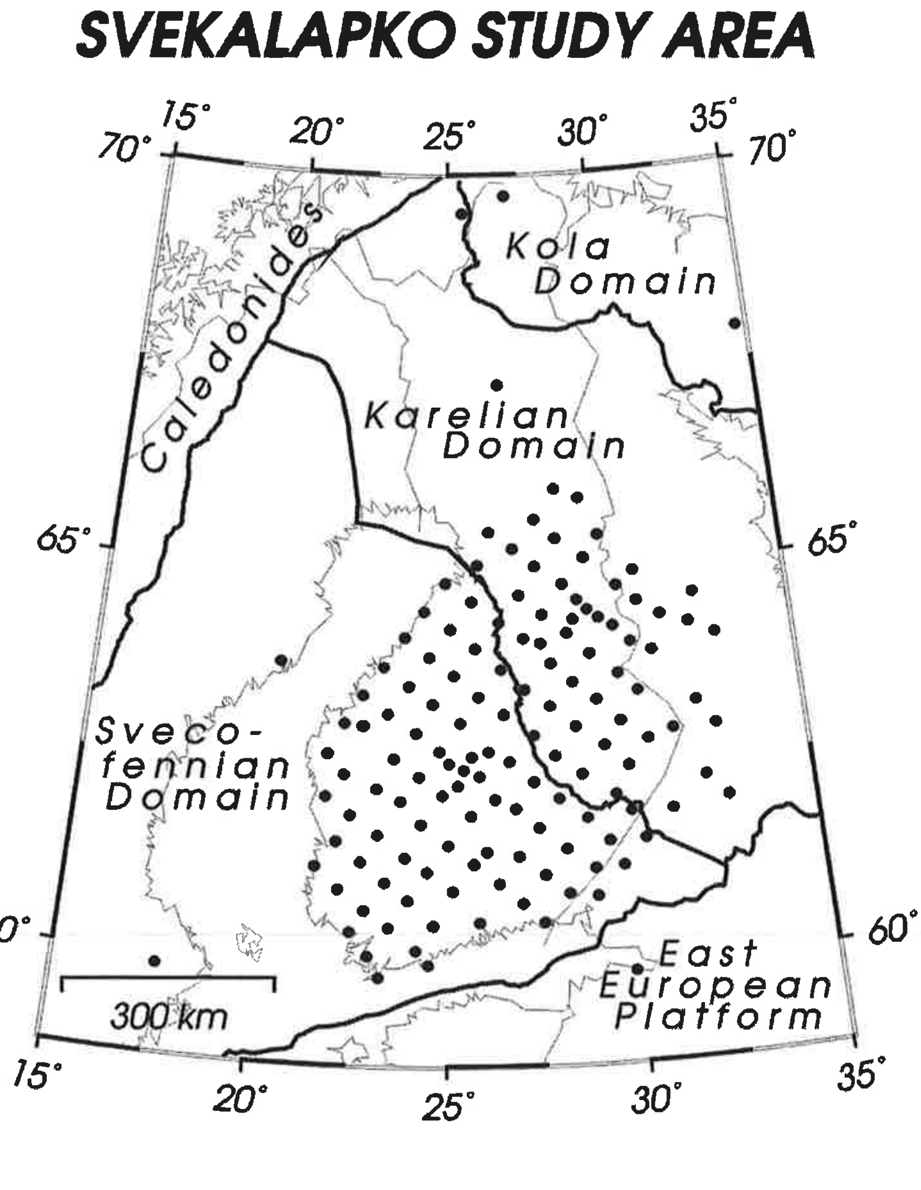


Figure 1. Geographical map of the study area and the main crustal segments (separated by thick lines). The dots show the seismic stations used in the SVEKALAPKO tomographic experiment.

2. The 3-D crustal model project

The three-dimensional crustal model (3-DCM) research project is a joint effort between scientists of the University of Oulu and the Geological Survey of Finland working in SVEKALAPKO and the Crustal Model Program. The three-year project started in November 2001 and is funded by the Academy of Finland. The main objectives are:

1. To create a 3-D crustal geophysical model that includes distribution of seismic P-and S-wave velocities, density and magnetization within the crust beneath the Finnish part of the SVEKALAPKO research area. The model will be used in petrophysical and geological interpretation.
2. To investigate the connection between deep velocity inhomogeneities in the crust and upper mantle and the regional anomalies of potential fields in southern Finland.
3. To separate the effect of the deep regional scale structures from the effect of shallow crustal structures on potential field data.
4. To continue investigations of relationship between various rock properties.

The data consists of: a) P- and S-wave arrival times of local seismic events registered during the SVEKALAPKO seismic tomography experiment, b) potential field data (Bouguer anomaly and magnetic field grids, digital maps and their transformations covering the SVEKALAPKO research area) collected and processed by the GSF and the Finnish Geodetic Institute (*Elo, 1997, Kääriäinen and Mäkinen, 1997*), and c) petrophysical laboratory data (density, magnetic susceptibility, remanence intensity and their spatial orientation, seismic velocity and other data available) on main rock types of Finland, collected and processed by the GSF (*Korhonen et al., 1997*).

Both traditional techniques of seismic data and potential field interpretation and new, recently developed joint techniques will be used. The main methods are:

a) Tomographic inversion of P- and S-wave travel times of local events complemented by forward ray-trace modeling and synthetic seismogram analysis. The result is the velocity structure beneath the SVEKALAPKO research area.
b) Joint interpretation of seismic and potential fields data using non-linear velocity-density relationships. The aim is to obtain a regional scale 3-D density distribution within the crust in the SVEKALAPKO research area.
c) The separate modeling of the residual of the observed Bouguer anomaly and gravity effect caused by the regional density distribution beneath the SVEKALAPKO research area. A more detailed structure of the uppermost crust is aimed at.
d) The separate modeling of the gravity and magnetic field data from the Geological Survey of Finland and the Finnish Geodetic Institute along selected key sections in the uppermost crust, such as major magmatic and metamorphic structures, using seismic, petrophysical and geological data as constraints.
e) Integrated interpretation of seismic and magnetic field data.

3. Results

The 3-DCM research project is still at its beginning. The research so far has concentrated on continuing the previous work of individual scientists and the preparation of data and computer software. An important part of the work is the development of forward and inverse modeling software that can handle the large size of the 3-D mesh. The following list describes some first results:

1. Compilation of a Bouguer anomaly maps at the surface and at the altitude of 10 km.
2. Preliminary 3-D P-wave velocity model on a 2×2×2 km grid.

3. Software for 3-D joint interpretation of seismic and gravity data.
4. Software for 3-D block model visualization and maintenance as well as computation of regional magnetic field anomalies.

Despite the demanding objectives of the 3D project, important contributions are expected towards better understanding the Finnish lithosphere and its processes.

References

Bock, G. (Ed.) and SVEKALAPKO Seismic Tomography Working Group, 2001. Seismic probing of Archaean and Proterozoic Lithosphere in Fennoscandia. *EOS Transactions AGU*, 82, 50, 621, 628-629.

Elo, S., 1997. Interpretations of the gravity anomaly map of Finland. *Geophysica*, 33, 51-80.

Elo, S., 2000. Gravity anomalies in Finland due to crustal and upper mantle sources. In: Pesonen, L.J., Korja, A. and Hjelt, S.-E., 2000 (Eds.). Lithosphere 2000, a Symposium on the Structure, Composition and Evolution of the Lithosphere in Finland. Programme and Extended Abstacts, Espoo, Finland, October 4-5, 2000. *Inst. Seismology, Univ. Helsinki, Report S-41*, 53-57.

Hjelt, S.-E., Daly, S. and SVEKALAPKO colleagues, 1996. SVEKALAPKO. Evolution of Palaeoproterozoic and Archaean Lithosphere. In: Gee, D.G. and Zeyen, H.J. (Eds.): *EUROPROBE 1996 - Lithospheric dynamics: Origin and evòlution of continents.* The EUROPROBE Secretariate, Uppsala University, 56-67.

Hyvönen, T. and Malaska, J., 2000. Velocity modelling of the lithosphere beneath South Finland. *Physics of the Earth and Planetary Interiors*, 122, 103-114.

Korhonen, J., Säävuori, H., Kivekäs, L., 1997. Petrophysics in the crustal model program of Finland. In: Autio, S. (Ed.), Geological Survey of Finland, Current Research 1995-1996, *GSF, Special Paper*, 23, 157-173.

Korhonen, J and Säävuori, H., 2000. Fennoscandian crustal model as reflected by the petrophysical interpretation of the potential field anomalies. In: Pesonen, L.J., Korja, A. and Hjelt, S.-E., 2000 (Eds.). Lithosphere 2000, a Symposium on the Structure, Composition and Evolution of the Lithosphere in Finland. Programme and Extended Abstacts, Espoo, Finland, October 4-5, 2000. *Inst. Seismology, Univ. Helsinki, Report S-41*, 85-90.

Kozlovskaya and E., Yliniemi, J., 1999. Deep structure of the Earth's crust along the SVEKA profile and its extention to the north-east. *Geophysica*, 1-2, 111-123.

Kozlovskaya, E., Yliniemi, J., Hjelt, S.-E., Komminaho, K. and SVEKALAPKO Seismic Tomography Working Group, 2002. Lithosphere structure beneath southern Finland derived by the SVEKALAPKO seismic array research (*this volume*).

Kozlovskaya, E., Taran, L., Yliniemi, J., Giese, R., Karatayev G., 2002. Deep structure of the crust along the Fennoscandia – Sarmatia Junction zone (central Belarus): results from a geophysical-geological integration. *Tectonophysics*, in press.

Kääriäinen, J. and Mäkinen, J., 1997. The 1979-1996 gravity survey and results of the gravity survey of Finland. *Publications of the Finnish Geodetic Institute*, 125, 24 p.

Sandoval, S., Kissling, E., Ansorge, J. and the SVEKALAPKO STWG, 2002. Three- Dimensional Seismic Crustal Model and Associated Effects on Teleseismic Tomography Beneath Central Fennoscandia (Baltic Shield). Submitted to *J. Geoph. Res.*

Lithosphere Structure Beneath Southern Finland Derived by SVEKALAPKO Seismic Array Research

Elena Kozlovskaya[1], Jukka Yliniemi[2], Sven-Erik Hjelt[1], Kari Komminaho[1]
and SVEKALAPKO Seismic Tomography Working Group

[1]Department of Geophysics, P.O. Box 3000, FIN-90014 University of Oulu, Finland
[2]Sodankylä Geophysical Observatory, Oulu Unit, P.O. Box 3000,
FIN-90014 University of Oulu, Finland
E-mail: elena.kozlovskaya@oulu.fi

The first results of the SVEKALAPKO seismic array research in Finland indicating the inhomogeneous upper mantle structure below southern Finland are discussed.

Keywords: seismic tomography, lithosphere, asthenosphere, Fennoscandia

1. Introduction

The international SVEKALAPKO seismic array research was carried out in September, 1998 - May, 1999. The study aimed at studying the lithosphere- asthenosphere system below the different parts (Lapland-Kola, Karelian and Svecofennian) of the Fennoscandian Shield (Hjelt and Daly, 1996, Bock et al., 2001). The detailed description of the field experiment and data processing procedures can be found in Raita (2001). In autumn of 2000, the ETH Zürich finished the compilation of the SVEKALAPKO event data set that includes 1356 teleseismic, regional and local seismic events. During 2000-2001 the event data set was verified and corrected at the Department of Geophysics of Oulu University, Institute of Seismology of Helsinki University and other participating institutes. The first results of SVEKALAPKO seismic array research were presented at the EUROPROBE Precambrian Time- Slice Symposium in St. Petersburg, Russia (November, 2001), at the 6th SVEKALAPKO workshop in Lammi, Finland (November, 2001) and at the XXVII General Assembly of the EGS in Nice, France (April, 2002). The different research groups are now continuing with the modeling.

2. Recent results obtained by the SVEKALAPKO Seismic Tomography Working Group from studies based on teleseismic events

Alinaghi et al. (2002) presented results of receiver function analysis of TOR and SVEKALAPKO data. They demonstrated differences in Moho depth and topography of global 410 km and 660 km seismic discontinuities between TOR and SVEKALAPKO areas. The map of the averaged crustal Vp/Vs ratio for the SVEKALAPKO area shows minor lateral variations of the ratio beneath the array. The Moho depth obtained from receiver function analysis differs slightly from that earlier compiled by Luosto (1991), i.e. the deepest Moho was detected somewhere beneath the Central Finland Granitoid Complex.

The results of teleseismic body wave tomography (Sandoval et al, 2002) revealed a number of low- and high velocity anomalies (±2% and ±3% relative to the standard IASP91 velocity model below and above 100 km depth, respectively). The tomography results also indicated very high P-wave velocity (relative to the IASP91 model) beneath the Central Finland down to a depth of ~400 km. At the moment there exists no explanation for this phenomenon.

The first results of surface wave tomography in the SVEKALAPKO area were presented by Bruneton et al. (2002). At this stage only 1-D S-wave velocity models of the

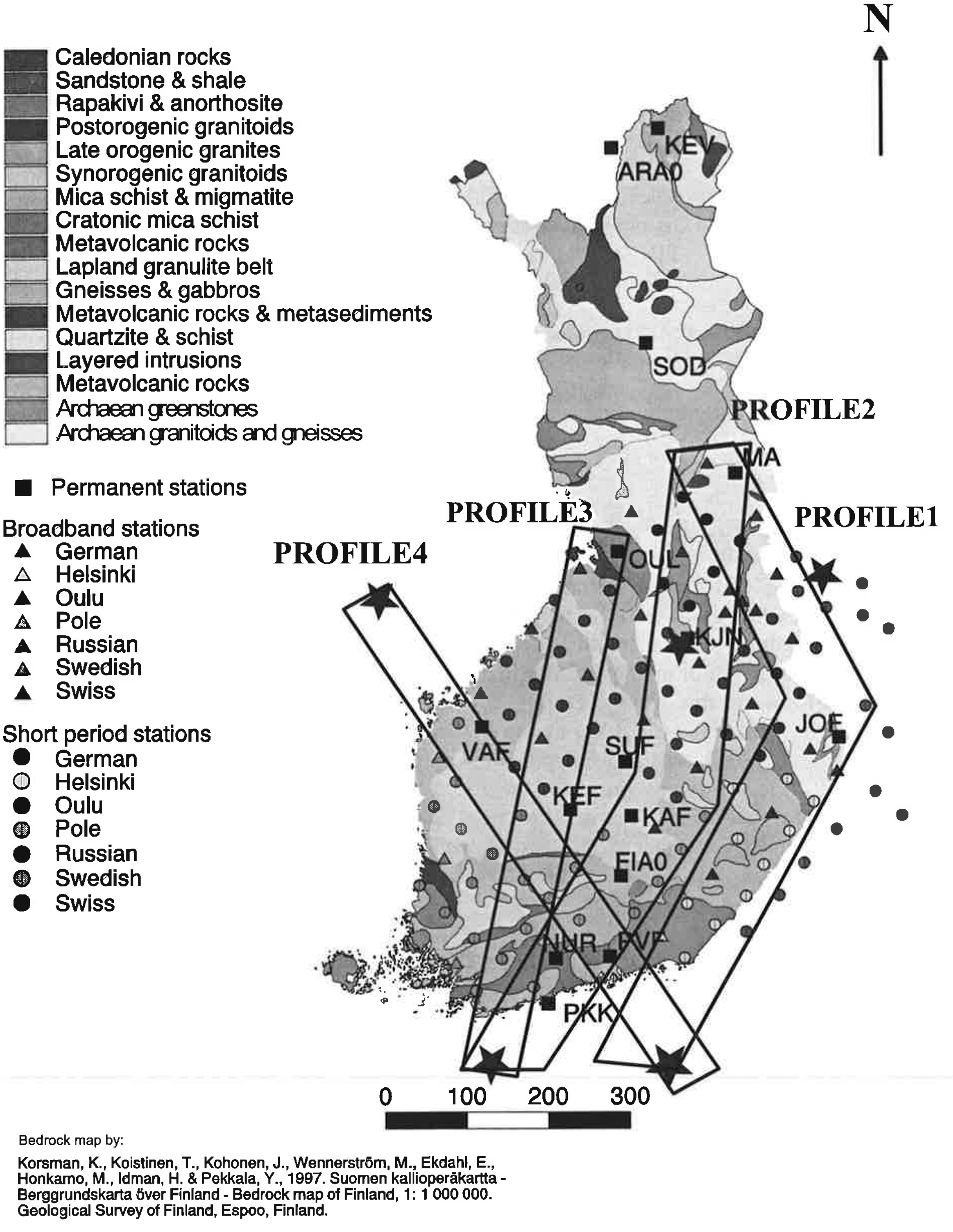


Figure 1. Location of the SVEKALAPKO array on the simplified geological map of the Fennoscandian schield. Black dots denote short period (SP) stations, white dots denote broad band (BB) stations, triangles denote permantent stations. Location of corridors along with the 2D forward raytrace modeling was performed is indicated by rectangles. Epicenters of events used for modeling of upper mantle reflectors are shown by stars.

upper mantle were obtained. The models demonstrate lateral S-wave velocity variations that can be attributed to several lithospheric blocks beneath the SVEKALAPKO research area. The result does not indicate a "seismic asthenosphere", however.

Funke et al. (2002) obtained a 1D shear wave velocity model down to a depth of 400 km with fast upper mantle (sub-Moho velocity of 4.7 km/s) from inversion of Rayleigh wave dispersion curve. The model has no low velocity zone in the upper mantle that can be associated with the "seismic asthenosphere". The dispersion curve can be explained by a nearly straight velocity profile from the Moho to the 410 km discontinuity.

Plomerova et al. (2002) presented results of seismic anisotropy investigations by joint interpretation of P-wave residual spheres and S-wave splitting parameters. They demonstrated the differences in lithosphere thickness beneath the TOR and SVEKALAPKO arrays. The result also indicated that each of the Archaean and Proterozoic domains beneath the SVEKALAPKO area is probably composed of several smaller lithospheric blocks with different orientation of the fast velocity. The thickness of anisotropic blocks was estimated to be no more than 200 km.

3. Local event studies

New information about the upper mantle structure below the central Finland was obtained from local event studies carried out at the Department of Geophysics of Oulu University. It is known that the seismic methods based upon the body waves of teleseismic events with the dominating frequency less than 1 Hz (receiver functions, teleseismic tomography, surface wave analysis, anisotropy studies based on teleseismic S-wave splitting) are able to reveal only large-scale discontinuities and inhomogeneities in the upper mantle. The more detailed information about the upper mantle structure can be obtained from reflection experiments, in which the P-waves reflected from the upper mantle have frequency in the range of 20-50 Hz. However, the ability of the reflection methods to reveal the upper mantle features is usually limited by the maximum depth of investigation of reflection experiments (~100 km). That is why recordings of local events can be considered as a very important type of seismic signal: first, they are of relatively high frequency (1-20 Hz), i.e. are able to reveal such inhomogeneities in the upper mantle that cannot be revealed by teleseismic methods. Secondly, they can be recorded at offsets of hundreds and thousands of kilometers, which means that they can penetrate to a depth of several hundreds kilometers.

The recordings of local events registered during the SVEKALAPKO seismic experiment contain very good quality reflections from the Moho boundary and a number of clear reflection events originating from the upper mantle below the SVEKALAPKO area. The results of 2D forward ray-trace modeling of reflected and refracted upper mantle P-wave phases along several corridors shown in Fig. 1 demonstrated (Yliniemi et al., 2002) that the mantle reflections originate from two different groups of boundaries: one group of phases arrive from sub-horizontal and gently dipping reflectors below the Moho boundary at a depth of 70-100 km, while the other group are phases originating from a depth of 120-150 km (Fig. 2,3). The reflections from the first group of interfaces in the upper mantle beneath southern Finland were reported earlier also by Heikkinen and Luosto (2000).

Kozlovskaya et al. (2002) presented also a refined P-wave crustal velocity model obtained as a result of joint interpretation of P-waves of local events and gravity field data. Joint interpretation of SVEKALAPKO local events and potential fields is a part of the project "Crustal model of the Finnish part of the SVEKALAPKO area" carried out by the Department of Geophysics of Oulu University in collaboration with the Geological Survey of Finland (see Kozlovskaya et al., this volume, for the project description). The model revealed high velocity in the upper mantle below southern Finland and a crustal sub-vertical high velocity anomaly, which is spatially coincident with the area of thickest Moho.

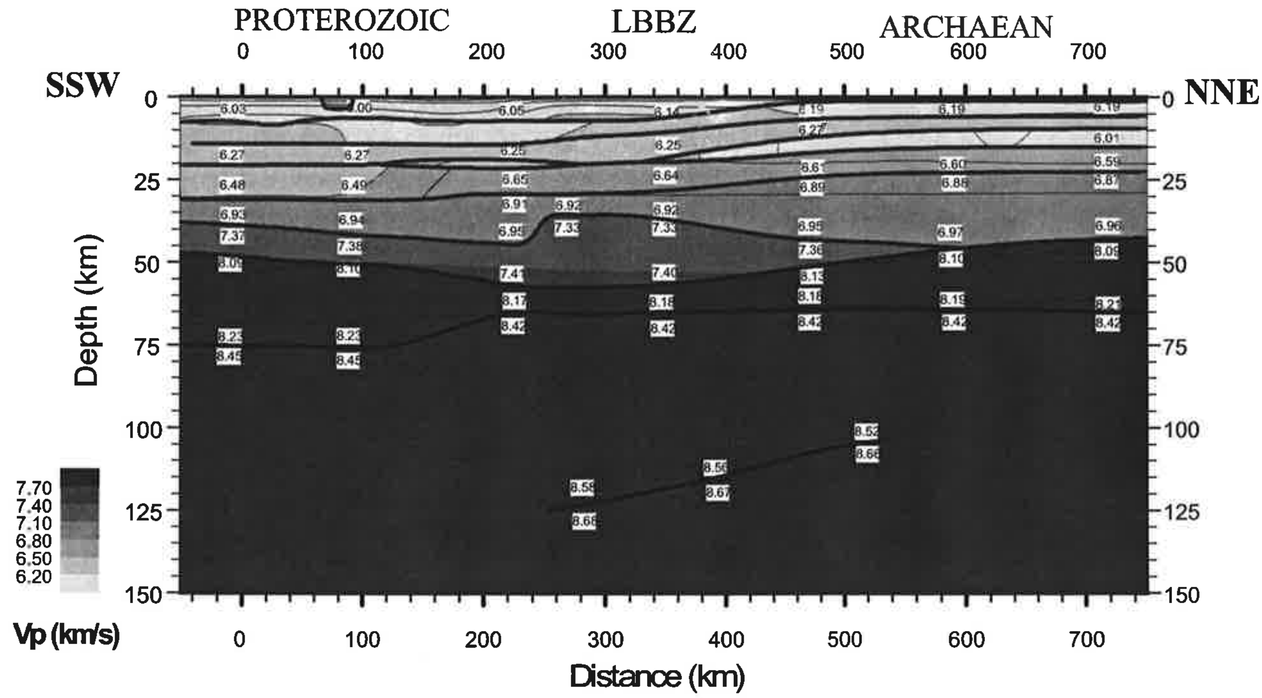


Figure 2. P-wave model along Line 2. Thin lines represent velocity isolines. Thick lines indicate major velocity discontinuities. Velocity values in km/s are shown in white rectangles. Gray colour intensity demonstrates the P-wave velocity distribution. Location of main tectonic units is indicated.

4. Summary of the first results by the SVEKALAPKO

The first results of the SVEKALAPKO STWG can be summarized as follows:

1. No low velocity layer that can be associated with the top of the asthenosphere
2. Laterally heterogeneous structure of the lithospheric mantle beneath the SVEKALAPKO area
3. Almost all the results make it possible to distinguish several big lithospheric blocks with different seismic properties beneath the SVEKALAPKO area.

Moho boundary revealed by receiver function methods do not differ significantly from that revealed earlier by the CSS (control source seismic) experiments.

4. Very high P-wave velocity in the upper mantle beneath the Central Finland
5. No clear correlation of upper mantle heterogeneities with the surface manifestation of the Ladoga- Bothnian Bay zone
6. Two groups of upper mantle reflectors (70-100 km and 120-150km) revealed by local event studies
7. High velocity area in the crust revealed by the inversion of P-waves of local events.

Proper explanation of these results in terms of petrophysics, rheology and pressure-temperature conditions in the upper mantle requires further modeling.

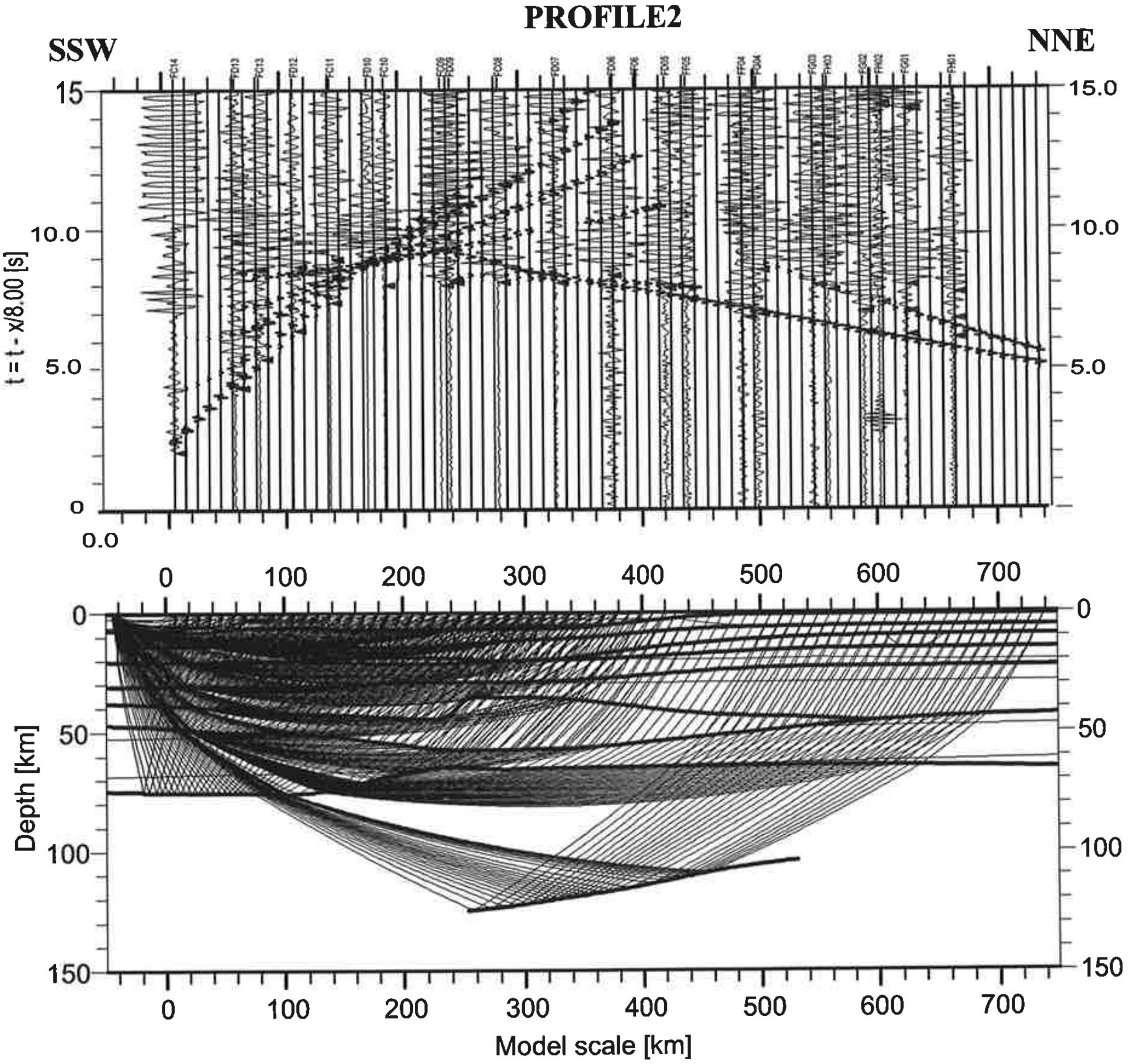


Figure 3. 2-D raytrace modeling results for local explosion on 05.11.98 in Estonia. The upper diagram shows the amplitude-normalized seismic section superimposed by synthetic seismogram. Small black triangles indicate picked arrival times of P-waves. The data are filtered by a Butterworth bandpass filter of 2-10 Hz and displayed using a reduction velocity of 8.0 km/s. The lower diagram shows the calculated rays from the explosion through the model shown in Fig. 2.

5. Origin of upper mantle reflectivity beneath southern Finland

The nature of upper mantle reflectivity in a depth range of 70- 200 km revealed by local and regional events analysis and long-range wide-angle reflection and refraction profiles in many areas of continental lithosphere is a controversial topic in present day lithospheric studies, because several competing explanations of this phenomenon exist at the moment.

One group of theories treats the upper mantle reflections as originating from global discontinuities produced by phase transition, compositional variations or change in rheological properties that can result also in seismic anisotropy (Levin and Park, 2000). These global discontinuities are denoted in literature as Lehmann discontinuity L (Lehmann, 1955), Hales discontinuity H (Hales, 1969), N discontinuity (Pavlenkova, 1996), 8^{o} discontinuity (Thybo and Perchuc, 1997). These explanations are usually based on teleseismic studies and long-range refraction profiles with relatively big distance between recording stations. Thus, basing on the results of so-called "Peaceful Nuclear Explosion" or PNE long- range profiles

Pavlenkova (1996), Tittgemeyer et al. (1996) suggested that all mantle reflections originating from this depth range could be explained by randomly alternating layers of low and high velocity material.

The other group of theories considers upper mantle reflectors as relicts of ancient subduction and collision processes. It is based mainly on high-resolution seismic reflection profiling, which provides clear evidence of merging of dipping mantle interfaces into lower continental crust in some cases (see Balling, 2000, for review). This explanation is supported also by recent results of wide-angle reflection and refraction profiling in Europe (EUROBRIDGE, POLONAISE, CELEBRATION2000) with dense sampling of recording instruments (EUROBRIDGE Seismic Working Group, 2000, Guterch et al., 1999, CELEBRATION Working Group, 2002). These surveys revealed multiple gently dipping reflectors in the upper mantle correlating with important tectonic boundaries in the crust in some cases. An attempt to explain the apparent disagreement between these two groups of theories was proposed recently by Bostock (1998, 1999), who discovered that reflectors detected by reflection surveys can merge with H and L discontinuities detected by teleseismic receiver functions method.

As for SVEKALAPKO upper mantle reflectors, they can hardly be attributed to some kind of global discontinuities; otherwise they would be detected also by the receiver function method (Alinaghi et al., 2002). More probably, they could be the relicts of ancient subduction and collision processes. A certain correlation of upper mantle reflectors with sub-vertical areas of low P-wave velocity revealed by teleseismic tomography (Sandoval et al, 2002) is also in favor of this explanation. However, further studies including more detailed petrophysical modeling, xenolith data analysis and joint interpretation with other methods data are required to explain their nature.

Appendix

The SVEKALAPKO Seismic Tomography Working Group consists of:
U. Achauer, A. Alinaghi, J. Ansorge,G. Bock, M. Bruneton, W. Friederich, M. Grad, A. Guterch, P. Heikkinen, S.-E-. Hjelt, T. Hyvönen, J.-P. Ikonen, E. Isanina, E. Kissling, K.Komminaho, A. Korja, E. Kozlovskaya, M.V. Nevsky, N.Pavlenkova, H. Pedersen, J. Plomerova, T. Raita, O. Riznichenko, R.G. Roberts, S. Sandoval, I.A. Sanina, N.V. Sharov, J. Tiikkainen, S.G. Volosov, E. Wieland, K. Wylegalla, J. Yliniemi, Y. Yurov.

References

Alinaghi, A., Bock, G., Kind, R., Hanka, W., Wylegalla, K, TOR and SVEKALAPKO Working Groups, 2002. Receiver function analysis of the crust and upper mantle from the North German Basin to the Archaean Baltic Shield. Geophysical Journal International, submitted.

Balling, N. 2000. Deep seismic reflection evidence for ancient subduction and collision zones within the continental lithosphere of northwestern Europe. Tectonophysics, 329, 269-300.

Bock, G. (ed) and SVEKALAPKO Seismic Tomography Working Group, 2001.Seismic probing of Archaean and Proterozoic Lithosphere in Fennoscandia. EOS Transactions AGU, 82, 50, 621, 628-629.

Bostock, M.G. 1998. Mantle stratigraphy and evolution of the Slave province. J.Geoph.Res., 103, 21183-21200.

Bostock, M.G. 1999. Seismic imaging of lithospheric discontinuities and continental evolution. Lithos, 48, 1-16.

Bruneton, M., Pedersen, H.A., Farra, V. and SSTW. 2002. Surface-wave tomography in Fennoscandia. XXVII General Assembly of EGS, Geophysical Research Abstracts, 4, EGS02-A-02733.

CELEBRATION Working Group, Grad, M.,2002. CELEBRATION 2000: Trans-Carpathian seismic profile CEL05 from Precambrian platform to Pannonian basin. . XXVII General Assembly of EGS, Geophysical Research Abstracts, 4, EGS02-A-03679.

EUROBRIDGE Seismic Working Group, 2000. EUROBRIDGE'97: Modelling of S-waves field on the EB'97 seismic profile. Geophysical Journal, 22,4, 87-88.

Guterch, A., Grad, M., Thybo, H., Keller, G.R., POLONAISE Working Group, 1999. POLONAISE'97 – an international sseismic experiment between Precambrian and Variscan Europe in Poland. Tectonophysics, 314, 101-121.

Hales, A.L., 1969. A seismic discontinuity in the lithosphere. Earth Planet. Sci. Lett., 7, 44-46.

Heikkinen, P., Luosto, U. 2000. Review of some features of the seismic velocity models in Finland. In: Litosphere 2000. A symposium on the structure, composition and evolution of the lithosphere in Finland.

GSF, Espoo, October 4-5, 2000. Institute of Seismology, University of Helsinki, Report S-41, 43-47 (extended abstract).
Hjelt, S. -E. and Daly, S., 1996. SVEKALAPKO Evolution of Paleoproterozoic and Archaean Lithosphere. In: Gee, D.G. and Zeyen, H.J. (Editors), EUROPROBE-1996 Lithosphere Dynamics: Origin and Evolution of Continents, 57-67.
Funke, S., Friedrich, W. and SSTWG, 2002. Rayleigh wave dispersion and a 1D model of Fennoscandian mantle. XXVII General Assembly of EGS, Geophysical Research Abstracts, 4, EGS02-A-04285
Lehmann, I., 1955. The times of P and S in northeastern America. Ann. Geofis., 8, 351-370.
Levin, V., Park, J. 2000. Shear zones in the Proterozoic lithosphere of the Arabian Schield and the nature of the Hales discontinuity. Tectonophysics, 323, 131-148.
Luosto, U, 1991. Moho depth map of the Fennoscandian Shield basedon seismic refraction data. Pp. 43 - 49 in Korhonen, H and A Lipponen (Eds): Structure and dynamics of the Fenno-scandian lithosphere,Inst. Seismol., Univ. Helsinki, Report S-25.
Kozlovskaya, E., Hjelt, S.-E., Yliniemi, j., Korhonen, J., Elo, S. and SVEKALAPKO STWG. Problem-dependent 3-D inversion of P-wave arrivals from local events recorded during the SVEKALAPKO deep seismic experiment in Finland. XXVII General Assembly of EGS, Geophysical Research Abstracts, 4, EGS02-A-00938.
Kozlovskaya, E., Pirttijärvi ,M., Hjelt, S.-E., Korhonen, J., Elo, S., Yliniemi, J. (this volume). The research project: 3-D crustal model of the Finnish part of the SVEKALAPKO research area.
Plomerova, J., Hyvönen, T., Sandoval, S., Babushka,V and SSTWG, 2002. Structure of Fennoscandian mantle lithosphere from seismic velocity anisotropy. XXVII General Assembly of EGS, Geophysical Research Abstracts, 4, EGS02-A-02173.
Pavlenkova, N.I. 1996. General features of the uppermost mantle stratification from long-range seismic profiles. Tectonophysics, 264, 261-278.
Raita, T. 2001. The seismic tomography experiment of the SVEKALAPKO project. M.Sci. Thesis. Department of Geophysics, Oulu University.
Sandoval, S., Kissling, E., Ansorge, J. and the SVEKALAPKO STWG, 2002. Three- Dimensional Seismic Crustal Model and Associated Effects onTeleseismic Tomography Beneath Central Fennoscandia (Baltic Shield). Submitted to J. Geoph. Res.
Thybo, H., Perchuc, E., 1997. The seismic 8^0 discontinuity and partial melting in continental mantle. Science, 1626-1629.
Tittgemeyer, M., Wenzel, F., Fuchs, K., Ryberg, T., 1996. Wave propagetion in a multiple-scattering upper mantle- observation and modelling. Geophys. J. Int., 127, 492-502.
Yliniemi, J., Kozlovskaya, E., Komminaho, K., Hjelt, S.-E., and SSTWG. 2002. Upper mantle structure beneath the SVEKALAPKO temporary seismic array in Finland derived from recordings of local events. XXVII General Assembly of EGS, Geophysical Research Abstracts, 4, EGS02-A-00826.

Project FIRE: Deep Seismic Reflection Sounding in Finland 2001-2005

Ilmo Kukkonen[1], Pekka Heikkinen[2], Elias Ekdahl[1], Annakaisa Korja,[2], Sven-Erik Hjelt [3], Jukka Yliniemi [4], Robert Berzin[5] and FIRE Working Group

[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]Institute of Geosciences, P.O. Box 3000, FIN-90014 University of Oulu, Finland
[4]Sodankylä Geophysical Observatory, P.O. Box 3000, FIN-90014 University of Oulu, Finland
[5]Spetsgeofizika, Moscow, Russia
E-mail: ilmo.kukkonen@gsf.fi

The project ***FIRE*** (**Finnish Reflection Experiment**) is a large-scale reflection seismic project currently being accomplished in Finland. **FIRE consortium** comprises Geological Survey of Finland, Institute of Seismology at the University of Helsinki, Institute of Geosciences at the University of Oulu and the Sodankylä Geophysical Observatory of the University of Oulu. A Russian state-owned company, Spetsgeofizika, has been hired as the seismic contractor in the project. Altogether about 1700 km of crustal scale CMP profiles will be measured in 2001-2003. The interpretation and reporting and will continue until the end of 2005. The major aim of FIRE is to compile structural and evolutionary models of the major geotectonic units in the Finnish part of the Fennoscandian Shield. This presentation provides an overview of the current situation of the project with selected examples of the first results.

Keywords: Reflection seismics, Fennoscandian Shield, crust, lithosphere

1. Introduction

The project **FIRE** (**F**innish **R**eflection **E**xperiment) is a large-scale project of solid earth geophysical and geological research in Finland. FIRE is run by a national consortium, whose partners are: Geological Survey of Finland, Institute of Seismology at the University of Helsinki, Institute of Geosciences at the University of Oulu and Sodankylä Geophysical Observatory of the University of Oulu. The seismic contractor in the project is a Russian geophysical governmentally owned company Spetsgeofizika. The funding of the contractor is based on the partial compensation of the debt of Russian Federation to Finland.

The aim of the project **FIRE** is to investigate the structure of the crust in Finland on several transects using reflection seismics during 2001-2005. The project is participated by the Geological Survey of Finland, and the universities of Helsinki and Oulu. The project is participated by about 20 scientists from the consortium institutes.

FIRE transects cross the major geotectonic units of the Precambrian in Finland, including several areas of metallogenic significance.

2. The acquisition

The seismic measurements are carried out using the Vibroseis method. Explosion sources are not used. The seismic signal is produced with five hydraulic vibrator trucks weighing each 15.4 tons. The applied signal is a linear sweep ranging from 12 to 80 Hz, and the maximum applied force is up to 60 % of the vibrator weight. The sweep duration and the correlated record length are 30 s, which corresponds to about 90 km in depth. Eight sweeps are stacked together at each shot point. The applied sounding geometry is the common depth point method. Using 360 active channels at 50 m intervals and shot points at 100 m intervals, the resulting fold is as high as 90. Thus each reflector will be recorded 90 times, which is very useful in improving the final signal to noise ratio. The recording line is 18.4 km long and the geometry is split-spread.

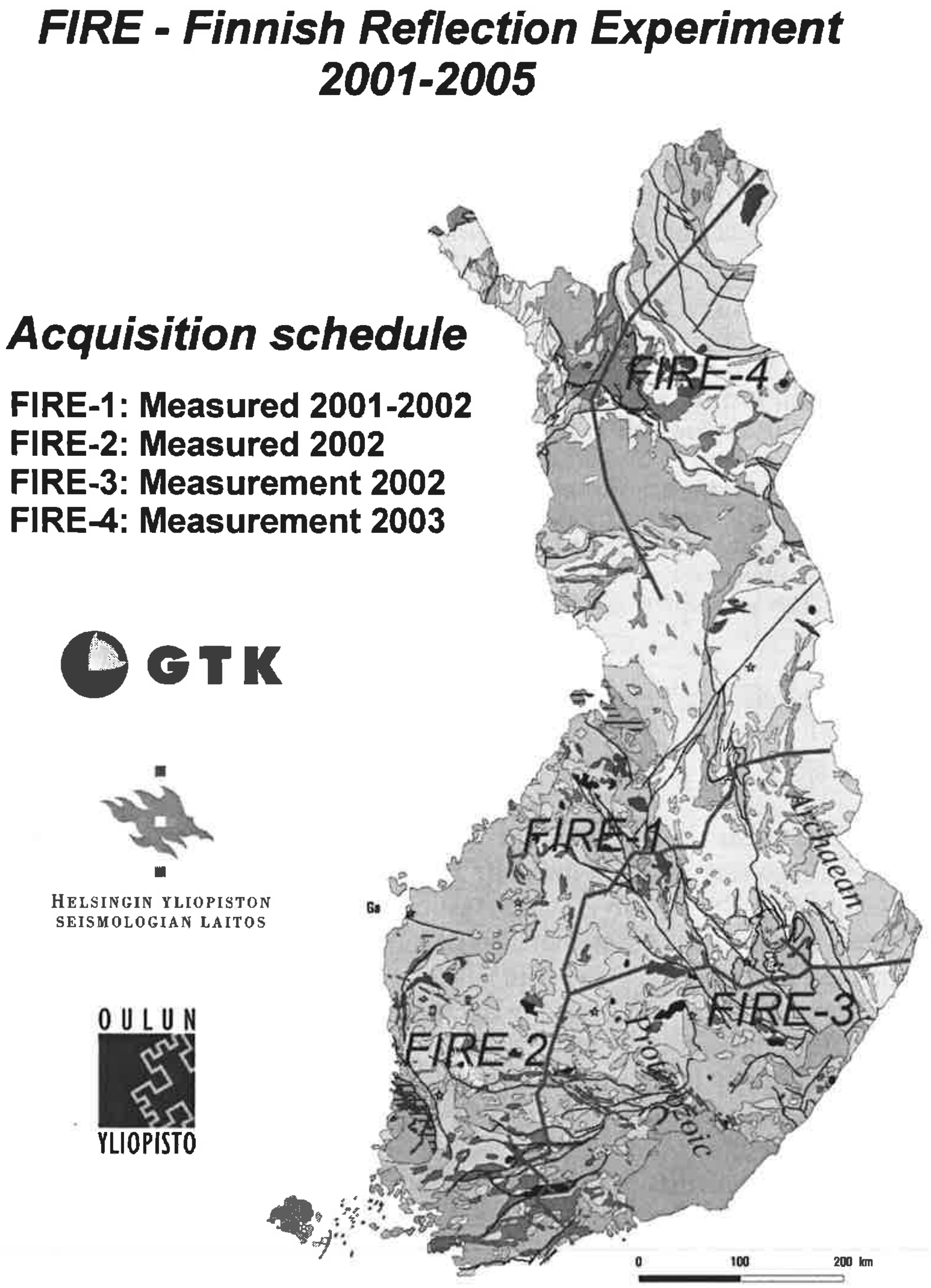


Figure 1. The location of FIRE transects plotted on a lithological map of Finland (Geological Survey of Finland).

Acquisition is done on public roads, and the geophones are installed on the road bank, typically at a distance of about 1-2 m from the edge of the road hard surface (typically asphalt or oil gravel). Each geophone channel consists of 12 vertical geophones connected in series. The signals are recorded digitally using the *INPUT/OUTPUT* Telemetric system. The sampling interval is 2 ms, which allows recording of high frequencies and results in a good resolution. Field processing is done with the *Seismic Viewer* software and final processing with *PROMAX* in the main office of Spetsgeofizika in Povarovka, as well as in the consortium institutes.

3. FIRE transects

The locations of FIRE profiles are shown in Fig. 1. The profile locations were decided using three major arguments: (1) Geological and geophysical relevance, (2) previously existing seismic wide-angle data (absolute velocity information required), and (3) road network in the transect areas. All these factors could not be ideally satisfied at the same time, but pragmatic compromises have been found. The acquisition of FIRE-1 took place in 2001-2002, FIRE-2 and FIRE-3 in 2002, whereas and FIRE-4 is scheduled for 2003.

4. The aims of FIRE

The project FIRE is expected to provide new structural information on several problems on the geology of the central part of the Fennoscandian Shield. Combining the seismic information with previous geological and geophysical data and models is expected to yield revised insights to the lithospheric structure and shield evolution. Some of the topics to be investigated in the FIRE project in Eastern and Central Finland can be shortly listed (profiles FIRE-1 and FIRE-3):

(1) Crustal reflective structures in general and their correlation with surface geology and geophysics;
(2) The subduction and collision structures between the Archaean craton and the primitive island arc system of the Pyhäsalmi area;
(3) General relations between the Archaean and Proterozoic crust, particularly in the area of thick crust;
(4) Shear and fracture systems at the Archaean-Proterozoic boundary;
(5) Properties and structure of the Archaean craton in the area of the Proterozoic overprint and outside of it;
(6) The middle and lower crustal characteristics of the Central Finland Arc Complex (the central Finland Granitoid complex);
(7) Relations between the Central Finland Arc Complex and the Tampere Schist belt;
(8) Structure and tectonic setting of the high-grade Archaean and Proterozoic blocks outcropping in central-eastern Finland;
(9) Structure and depth extent of the Proterozoic metasedimentary units and ophiolite complexes close to the border of the Archaean craton (including the Outokumpu nappe);
(10) Possible reflections related to kimberlite and alkaline magmatism;
(11) Relations between possible upper mantle reflectors and lower crustal reflectors;
(12) Correlation of seismic reflections with deep electromagnetic conductors and thermal regime of the lithosphere;
(13) Correlation of reflectors recorded on FIRE transects on continent and BABEL transects in the Bothnian Bay;
(14) Deep structure of ore bearing areas (e.g., Kuhmo and Ilomantsi greenstone belts, Pyhäsalmi area, Outokumpu, Kotalahti nickel belt).

Respectively, thematic topics to be investigated in southern Finland can be listed as follows (FIRE 2):

(1) Correlation of crustal reflectors with surface geology and geophysics;
(2) Deep structure of the southern Finland volcanic-sedimentary complex;
(3) Structure and relations of the Tampere Svecofennian collisional zone with the Central Finland Granitoid complex, as well as with the southern Finland volcanic-sedimentary complex;

(4) Structure and tectonic characteristics of the Hyvinkää-Mäntsälä layered intrusion;
(5) Deep structure of the potassium-rich granitoid belt in southern Finland;
(6) Lower crustal properties in the area of thick crust and 'normal' crust;
(7) Correlation of reflectors recorded in FIRE and BABEL projects;
(8) Deep structure of ore bearing areas (e.g., eastern part of the Vammala Ni belt, Hyvinkää-Mäntsälä layered intrusion, southern Finland leptite belt).

In northern Finland, the preliminary list of questions to be solved with the aid of FIRE data is as follows:
(1) Correlation of crustal reflectors with surface geology and geophysics;
(2) Are the Lapland Granulite Belt and the Tanaelv Belt blocks thrust from the NE, or only rootless nappes?
(3) Depth extent of the Lapland Granulite belt and its contacts with the Archaean basement;
(4) Factors contributing to the higher crustal thickness under the Central Lapland Granitoid Complex in comparison to the Greenstone belt;
(5) Internal structure of the Central Lapland Greenstone belt and continuation of the metamorphic zones in depth;
(6) Structures of the Archaean basement below the Central Lapland Greenstone Belt;
(7) Correlation of deep crustal structures with the seismically active zones and young (postglacial) faulting in central Lapland;
(8) The deep structure of the Central Lapland Granitoid Complex, its relations with the Greenstone Belt, and the origin of the migmatitic granitoid complex itself;
(9) Deep structure of ore bearing areas (e.g. polymetallic zones of Kittilä and Kolari).

5. The current situation of FIRE

At the moment of writing, in September 2002, the field-work has proceeded very efficiently, and transects FIRE-1 and FIRE-2 have been completed. The Eastern part of FIRE-3 from Ilomantsi to Leppävirta has been measured including the "side-track" over the Outokumpu nappe. In Outokumpu, a high-scale, high-resolution test was carried out on the major ore-bearing belt to obtain experience on applying reflection seismic methods to structural investigations in an ore-potential formation. At the moment, the final processing and migration of FIRE-1 has been finished, and the data from FIRE-2 and FIRE-3 are being processed.

6. The first results of FIRE

The migrated sections obtained so far have already revealed numerous previously unknown structures in the Finnish Precambrian. Among the many findings we can select the following points from FIRE-1, which extends from the Archaean craton to the Central Finland Granitoid Complex (Fig. 1).

(1) The upper on FIRE-1 crust is generally very reflective. Very bright reflectors are seen both in the Archaean as well as on the Proterozoic part of the transect. (2) Collisional events between the Svecofennian island arcs and the Archaean continent have left significant tracks in the upper and middle crustal structures. In the westernmost parts of the Archaean craton there are very sharp and bright reflectors, which can be followed from the surface to the base of the upper crust. (3) The Raahe-Ladoga belt, a major Proterozoic shear belt and sulphide-ore province is characterized by diffuse east-dipping gently sloping reflectors in the middle and lower crust. (4) The crust-mantle transition is relatively diffuse in reflection images, and only revealed by the ending of lower crustal reflectivity at depths slightly over 50 km. The depth of the reflection Moho agrees well with the refraction Moho previously determined on the Sveka DSS transect.

The Geophysical Study of the Northern Baltic Sea

Minna Kuusisto, Pekka Heikkinen and Annakaisa Korja

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

In the year 2000, the University of Hamburg organized a marine geophysical student cruise in the northern Baltic Sea. A total of 1300km of shallow reflection seismic, gravity and magnetic survey profiles were recorded at the cruise. Profiles were planned to give information about geophysics and geology of the poorly known areas. The poster give overview of the research area, presents methods used in the survey, show a preliminary results from gravity measurements and show some seismic data processing steps.

Keywords: Geophysical survey, reflection seismics, seismic data processing, Baltic Sea

1. Introduction

In the year 2000, the University of Hamburg organized a marine geophysical student cruise in the northern Baltic Sea (Fig.1). University of Helsinki and University of Tarto took part organizing the cruise and land stations. A total of 16 profiles were measured between 24.8. - 7.9.2000 by research vessel M/s Heincke. Measurements consisted of 1300 km of shallow reflection seismic, gravity and magnetic surveys and 3 temporary land stations recording seismic signal and magnetic field. Profiles are in figure 1. Profiles were planned to give information on geophysics and geology of the poorly known northern Baltic Sea. Near the coastal area of southern Finland profiles 2 and 3 crossed Åland-Paldinski-Pskov-shear zone, which is still seismically active (*Saari, 1998*). In the center of the research area the profiles crossed the Baltic Sea rapakivi area. The offshore Åland Islands the profiles reached the sedimentary basin of Åland Sea. Profile 6 crossed the deep seismic reflection profile BABEL C (BABEL Working Group, 1993) enabling comparison of dataset in the future. The margin geometry of East-European platform was also documented several times.

2. Equipments and data

The Seismic measurements carried out with the speed of 4 knots with Air Gun Harmonic Mode (25/25 in^2) as seismic source. During the measurements the air gun was at a 1 m depth and 10 m behind the ship. The high resolution streamer had 24 channels and consisted of a 120 m of tow-lead, two 10 m passive stretch sections and three 50 m active sections. Each active section had 8 hydrophone groups of 4,9 m in length. The distance between the groups was 6.25 m and the number of hydrophones per group was 16. The first active section started at the distance of 5-20 m from air gun. Shots were generated every 5 s and the shot point interval was ca.10 m.

The survey used gravity meter KSS 30, which consisted of GSS 30 gravity sensor, KT 30 stabilization and data handling subsystems. Data were recorded every 20 s and the measuring interval was ca. 40 m.

The magnetic measurements were carried out by a gradiometer (Geometrics G811). The system consisted of two proton magnetometers. During the measurements sensors were at the distance of 250 m behind the ship. The measuring interval was 20 s or 40 m. The marine magnetic data were tied to the general reference system by 3 onshore field stations (Fig.1).

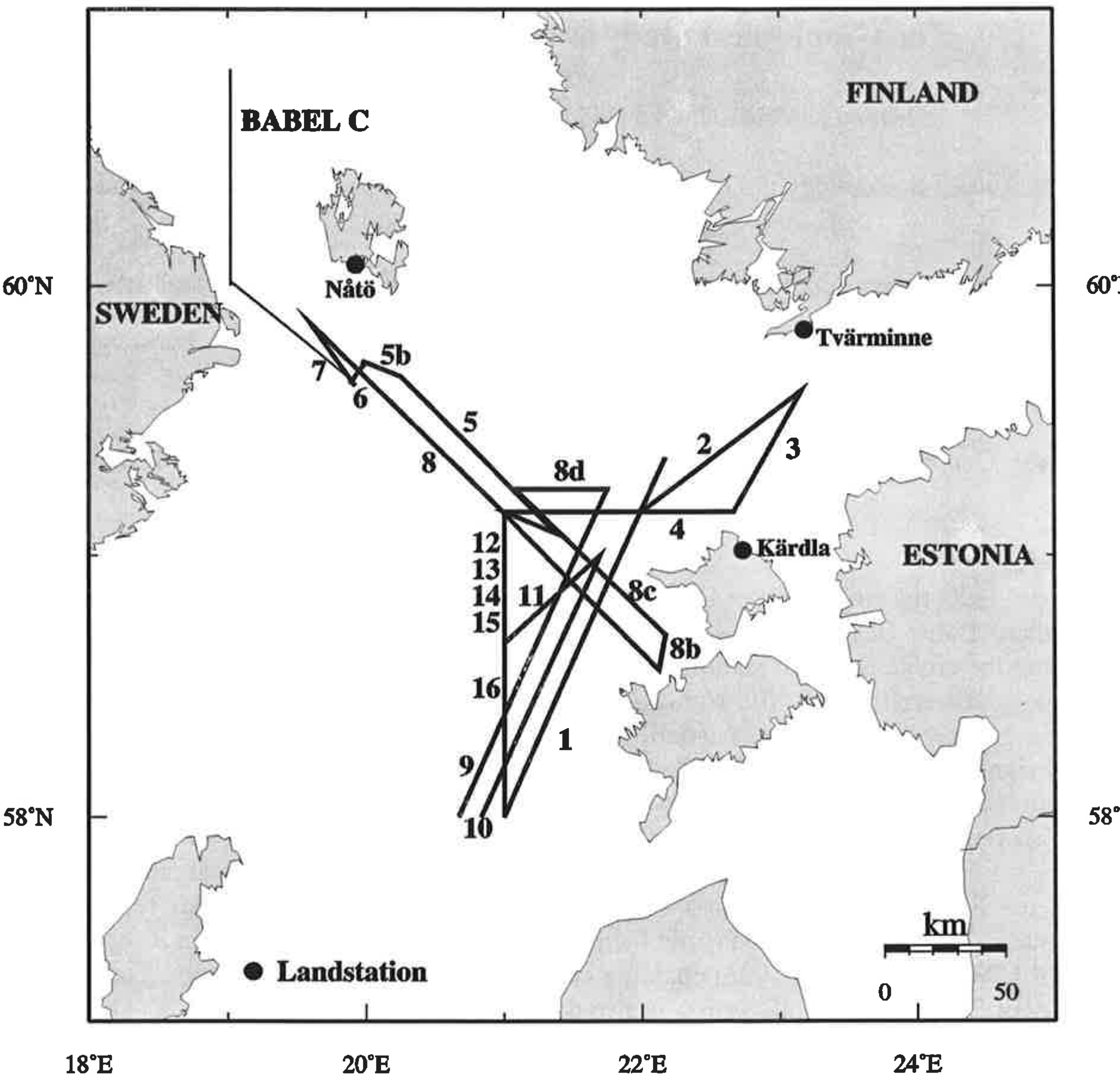


Figure 1: Geophysical survey lines in the northern Baltic Sea. Seismic survey profiles were 1-9 and 12-15, magnetic measurements carried out on profiles 1-12 and gravity measurements on all profiles. Land stations recorded continuous magnetic and seismic signal.

3. Preliminary processing of the seismic data

The reflection seismic data (Fig. 2a) was processed by using standard data processing tools included in the Seismic Unix software package (*Cohen and Stock well, 2000).*

The processing sequence was the following:
- -Editing
- -Spherical divergence correction
- -NMO-correction with water velocity (1500m/s)
- -Band pass filter (20-350 Hz)
- -Spectral balancing
- -Wavelet shaping filter
- -Predictive deconvolution
- -Muting

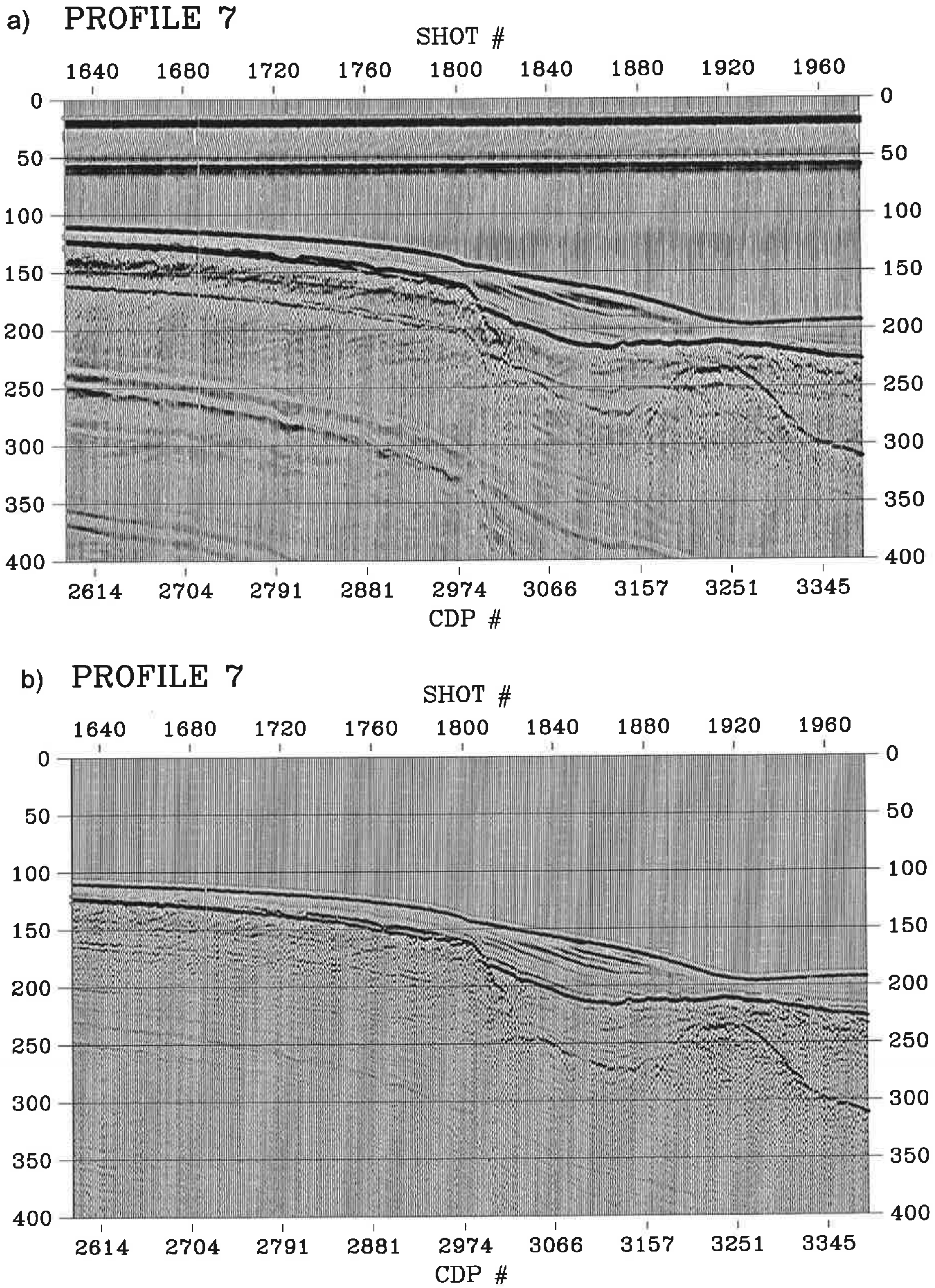


Figure 2. An example of the seismic data from the profile 7. Vertical axis is the two way travel time in milliseconds. Shot point interval is 10 m.

a) Near trace gather before band pass filtering.

b) Near trace gather after band pass filtering, spectral balancing, wavelet shaping filtering, predictive deconvolution and muting.

The wavelet from the single air gun used in this survey contains a ghost about 50 ms after the main pulse. It was found out that this ghost could be most efficiently removed by wavelet shaping filter.

The multiple reflections from the sea bottom are often the major problem in marine seismic surveys. Usually they are eliminated by predictive deconvolution. In this survey the water depth along the profiles varies approximately from 20 to 200 m, causing the corresponding variation in the time delays between the primary signal and the multiple reflections.

The best results in removing the multiple reflections were obtained when the deconvolution parametres (prediction distance and filter length) were allowed to vary with the water depth.

References

BABEL Working Group, 1993. Integrated seismic studies of the Baltic Shield using data in the Gulf of Bothnia region. Geophys. J. Int., 112, 305-324.

Cohen, J. K. and Stockwell, Jr. J. W., 2000. CWP/SU: Seismic Unix Release 34: a free Package for seismic research and processing, Center for Wave Phenomena, *Colorado School of Mines*.

Koistinen, T. (Editor), 1994. Precambrian basement of the Gulf of Finland and surrounding area, 1:1mill. *Geological Survey of Finland,* Espoo, Finland.

Saari, J., 1998. Regional and local seismotectonic characteristics of the area surrounding the Loviisa nuclea power plant in SE Finland. *Institute of Seismology*, University of Helsinki, Rep. S-36.

Lithospheric Conductivity Along GGT/SVEKA Transect: Implications from the 2-D Inversion of Magnetotelluric data

Ilkka Lahti[1], Toivo Korja[1], Laust Pedersen[2] and BEAR Working Group

[1]Department of Geophysics, P.O.Box 3000, FIN-90014 University of Oulu, Finland
[2]Department of Earth Sciences, University of Uppsala, Sweden
E-mail: ilkka.lahti@oulu.fi

GGT/SVEKA transect traverses the main tectonic units in the central part of the Fennoscandian Shield in NE-SW direction. These units are the Archaean Karelian Province in the northeast and several Palaeoproterozoic arc complexes in the Svecofennian Domain in the southwest. Since 1985, over 150 magnetotelluric (MT) soundings of which 140 are short period and 10 long period soundings have been made in the survey area. We have performed several 2-D Occam inversions of the MT data using the REBOCC code (*Siripunvarapom and Egbert, 2000*) to generate smooth conductivity models for the survey area. The best fitting model with the RMS error below 3.0% is obtained by using the determinant of impedance tensor as the inverted parameter.

Highly conductive dipping conductors at both sides of the boundary zone between the arc complexes in southern and central Finland are seen in the final model (profile 1). Both conductors represent borders of major crustal segments possibly indicating two subduction zones in the area. In contrast, only minor conductivity variations are seen at the lithological boundary between the Karelian and Svecofennian domains in central Finland whereas a southwestward dipping conductor beneath the Palaeoproterozoic Kainuu Belt is revealed. The conductor suggests the presence of Palaeoproterozoic graphite bearing sedimentary rocks beneath the Archaean rocks of the Iisalmi complex. Lower crustal conductor is absent to NE from the Kainuu Belt while the conductor is present in the Palaeoproterozoic Svecofennian Domain to southwest from the Kainuu Belt. Thus, the main conductivity boundary between the Archaean and Palaeoproterozoic lithosphere is located beneath the Kainuu Belt rather than beneath the Ladoga-Bothnian Bay Zone.

Keywords: Magnetotellurics, 2-D inversion, crustal conductivity, Fennoscandian Shield, Svecofennian

1. Introduction

GGT/SVEKA transect traverses the main tectonic units in the central part of the Fennoscandian Shield. The profile 1 (Fig. 1) is NE-SW directed and has the length of 720 km. The profile begins from the western part of the Archaean Karelian Province in the northeast in Russia, crosses the boundary zone between the Karelian Province and the Palaeoproterozoic Svecofennian arc complexes, traverses across the main tectonic units of the northern part of the Svecofennian Domain and ends in the Mesoproterozoic rapakivi area in the southwest close to the coast of the Bothnian Sea (*Korsman et al., 1999*). The profile 2 (Fig. 1) is N-S directed and has the length of 250 km. The profile 2 crosses the Häme (HaB), Pirkkala (PiB) and Tampere Belts (TaB) and ends at the Central Finland Granitoid Complex (CFGC).

Magnetotelluric data acquired along profiles consist of short period pre-BEAR data (period range 0.1 – 1000 s; *Korja and Koivukoski, 1994*) from c. 110 sites and long period BEAR data (10 – 10000 s) from c. 10 sites. Average site distances are 7.5 km and 100 km for mid period and long period data, respectively.

2. Inversion procedure

The dimensionality and strike analysis from the survey area indicates local distortions and varying regional strike azimuths while the regional dimensionality is approximately 2-D in most parts of the transect. Thus, the conventional 2-D inversion procedure of fitting both

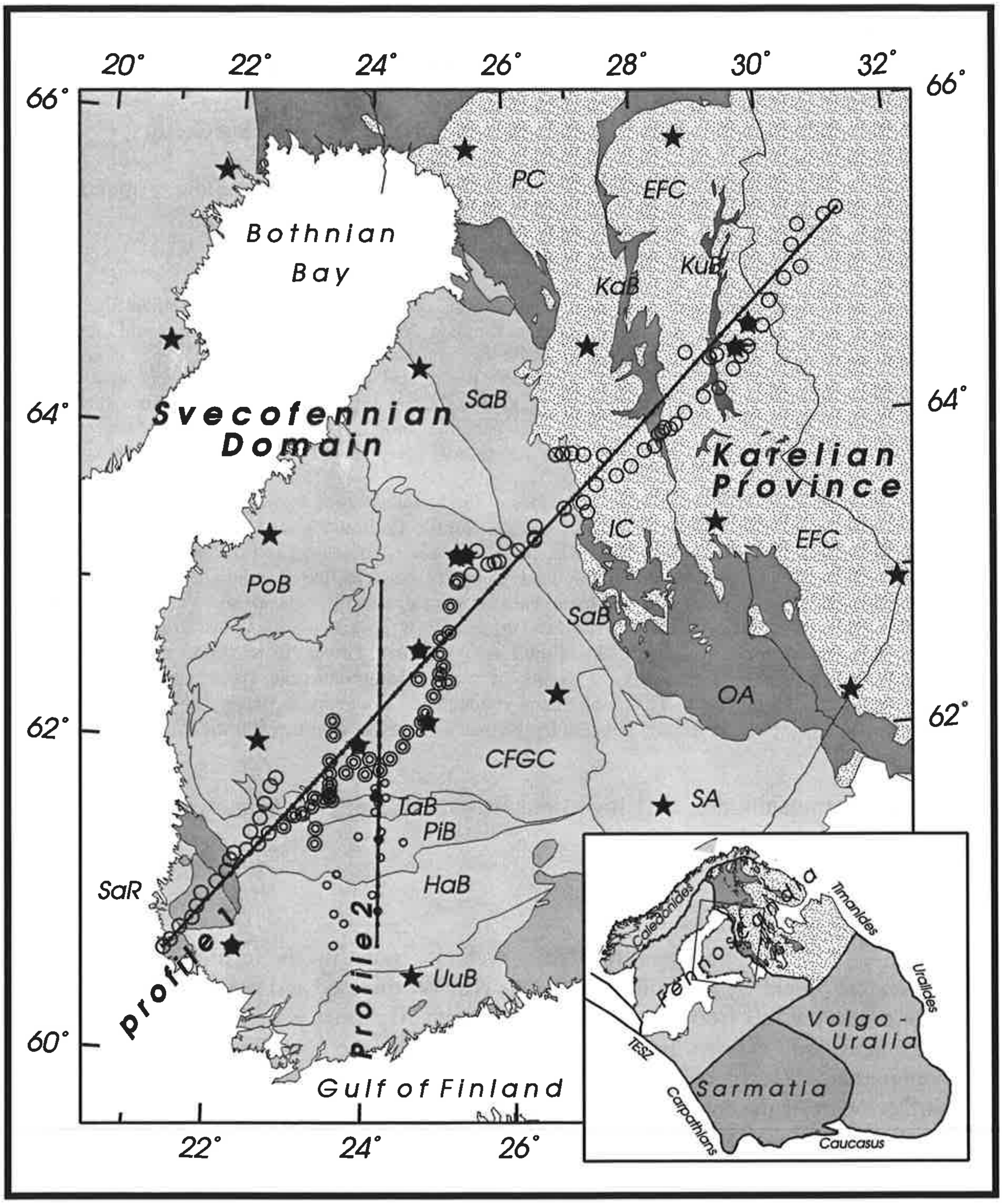


Figure 1. Magnetotelluric GGT/SVEKA profiles. Circles indicate mid-period (0.1–1000 s) pre-BEAR magnetotelluric sites that were used in the inversion. Large circles and small circles belong to profile 1 and profile 2, respectively. Stars show the location of long period (10 – 10000 s) sites including BEAR sites and four earlier long period sites. Lines indicate the location of the 2D models of the profiles 1 and 2. Geology is simplified primarily from *Korsman et al., (1997).* Abbreviations: EFC = Eastern Finland Complex, IC = Iisalmi Complex, KaB = Kainuu Belt, PC = Pudasjärvi Complex, SA = Saimaa Area, PiB = Pirkkala Belt, TaB = Tampere Belt, CFGC = Central Finland Granitoid Complex, PoB = Pohjanmaa Belt, OA = Outokumpu Area, SaB = Savo Belt, SaR = Satakunta Rapakivi area, UuB = Uusimaa Belt, HaB = Hämeenlinna Belt, KuB = Kuhmo Belt.

TE and TM modes simultaneously leads to a fairly poor fit with the RMS error above 10.0%. The best fitting model with the RMS error below 3.0% is obtained when the determinant average impedance data are used. Determinant average of the magnetotelluric impedance tensor is defined as $Z_{det} = \sqrt{Z_{xx}Z_{yy} - Z_{xy}Z_{yx}}$, where Z_{xx}, Z_{yy}, Z_{xy} and Z_{yx} are the elements of the impedance tensor. Determinant is rotationally invariant and therefore the use of it does not require any a priori knowledge on geoelectric strike directions. It also allows us to invert data simultaneously from adjacent geoelectric units having different geoelectric strike. The disadvantage of using determinant is the loss of directionality information (e.g. information on geoelectric strike direction and anisotropy).

Inversion was performed using the REBOCC code (*Siripunvaraporn and Egbert, 2000*). The inversion algorithm is based on an efficient variant on the OCCAM algorithm of *deGroot-Hedlin and Constable (1990)* for generating smooth conductivity models. The minimum RMS of the final model (2.6 %) was reached by rejecting poor data (5%) from the profile. In order to study the stability of the final inversion model, various starting models were used. The main features of the final model are not dependent on the starting model. Furthermore, several tests were made in order to study the resolution of the inversion. Depth extent of the model was studied by fixing the resistivity of the starting model at various depths and observing the resulting total RMS after several iterations. The depth extent of the model, i.e. the depth at which the resolving power of the data vanishes, without long period data is c. 60-90 km with acceptable low RMS of 3%. The use of the long period data increases the depth extent of the model to 400 km.

3. Resistivity models and discussion

The main features of the conductivity model along the profile 1 (Fig. 2) include upper and mid-crustal conductors (ρ = 5 - 100 Ωm) at both sides of the Pirkkala and Tampere Belts i.e a northward dipping conductor in the northern side beneath the Central Finland Granitoid Complex and a southward dipping/sub-horizontal conductor in the southern side of the PiB-belt. These conductors possibly indicate two subduction zones in the target area.

Highly resistive (ρ = 1000 - 30000 Ωm) upper crust at the depths of 0-30 km and a conductive (ρ = 10-100 Ωm) lower crust at the depths of 30-60 km is seen in the CFGC in all models. These are compatible with the previous results obtained through forward modelling of magnetotelluric data (*Korja and Koivukoski, 1994*).

The inversion models along profiles 1 and 2 are quite similar, in general, but some distinct differences are observed. The conductance of the crust (0-60 km) beneath PiB and TaB is higher beneath the profile 2 than the profile 1 being 10000 S and 3000 S in the uppermost 60 km, respectively. Furthermore, the northward dipping conductor beneath the CFGC along the profile 1 is in conflict with nearly vertical conductor beneath the profile 2 (e.g. *Korja and Koivukoski, 1994*) possibly reflecting a change from a more steep geometry in east (in Tampere-Hämeenlinna region) to a more gentle geometry towards west.

A southwestward dipping conductor beneath the Archaean Iisalmi Complex and the Kainuu Belt is revealed from the inversion. These conductive structures suggest the presence of Palaeoproterozoic sedimentary rocks beneath the Archaean rocks of the Iisalmi complex. Interestingly, only minor conductors are seen at the lithological boundary between the Karelian Province and the Svecofennian Domain (i.e. Ladoga-Bothnian Bay Zone) having a dip towards southwest while the strongest conductivity boundary is detected beneath the Kainuu Belt. The conductive lower crust is absent northeastwards of the Kainuu Belt while it is present southwestwards from the Kainuu Belt. Thus, the final inversion model shows that a conductivity boundary of the resistive Archaean and more

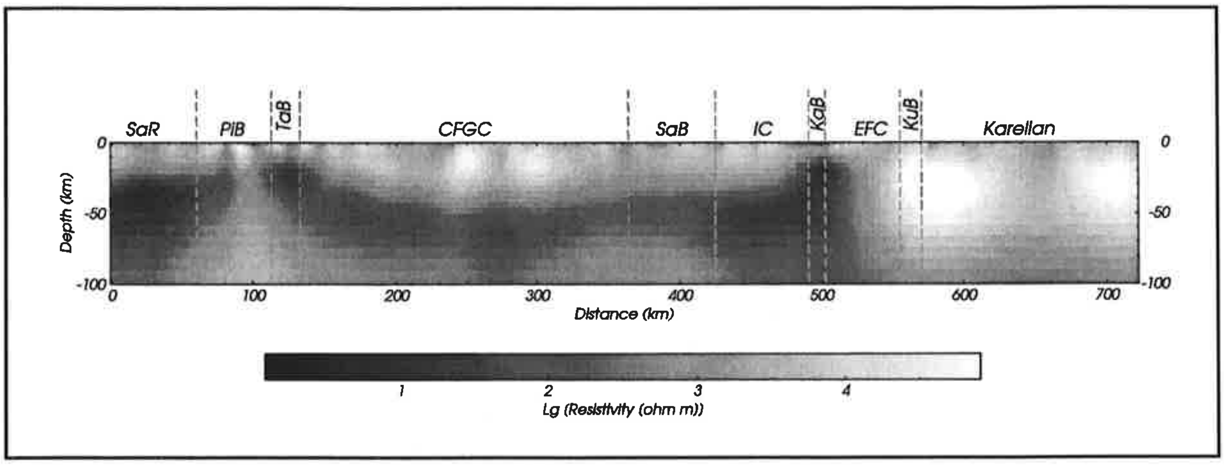


Figure 2. Smooth two-dimensional inversion model for the crust and uppermost mantle lithosphere along the profile 1 (RMS = 2.6%). Abbreviations: EFC = Eastern Finland Complex, IC = Iisalmi Complex, KaB = Kainuu Belt, PiB = Pirkkala Belt, TaB = Tampere Belt, CFGC = Central Finland Granitoid Complex, SaB = Savo Belt, SaR = Satakunta Rapakivi area, KuB = Kuhmo Belt.

conducting Palaeoproterozoic lithosphere is located beneath the Kainuu Belt rather than beneath Ladoga-Bothnian Bay Zone. One explanation for the conductivity difference could be 'pervasive' saturation of Palaeoproterozoic crust with carbon (graphite).

The use of long period data allowed us to study conductivity of the mantle lithosphere. The long period inversion model along the profile 1 shows increased conductivity below the depths of 230-280 km along the entire profile. This may indicate the presence of an electrical asthenosphere beneath the craton.

References

deGroot-Hedlin, C., Constable, S., 1990. Occam's inversion to generate smooth, two-dimensional models from magnetotelluric data. *Geophysics* **55**, 1613-1624.

Korja, T., Koivukoski, K., 1994. Crustal conductors along the SVEKA Profile in the Fennoscandian (Baltic) Shield, Finland. *Geophys. J. Int.* **116**, 173-197.

Korsman, K., Korja, T., Pajunen, M., Virransalo, P. and GGT/SVEKA working group, 1999 The GGT/SVEKA transect: Structure and evolution of the continental crust in the Palaeoproterozoic Svecofennian Orogen in Finland. *International Geology Review* **41**, 287-333.

Korsman, K., Koistinen, T., Korhonen, J., Wennerström, M., Ekdal, E., Honkamo, M., Idman, H., Pekkala, Y. (eds.), 1997. Suomen kallioperäkartta 1:1000 000, Geological Survey of Finland, Espoo, Finland.

Siripunvaraporn, W., Egbert, G., 2000. An Efficient data-subspace inversion method for 2-D magnetotelluric data. *Geophysics* **65**, 791-803.

Geochemistry and Metamorphic Petrology of the Archaean Pudasjärvi Granulite Area, Finland

Katja Lalli[1], Heikki Juopperi[2] and Pekka Tuisku[1]

[1]Department of Geosciences, P.O. Box 3000, FIN-90401 University of Oulu, Finland
[2]Geological Survey of Finland, P.O. Box 77, FIN-96101 Rovaniemi, Finland
E-mail: pekka.tuisku@oulu.fi

During the fieldwork at summer 2000 in Pudasjärvi, seventy kilometres northeast from Oulu, it came apparent that there is a folded, magnetic granulite belt in the western part of the Pudasjärvi County. The subject of this Master Thesis is these Archean granulites in Pudasjärvi Complex.

The Pudasjärvi Complex is poorly known and consists mainly of Archean gneisses and granitoids and abundant amphibolites. Proterozoic granites and diabase dikes have intruded the Archean gneisses. The Pudasjärvi Granulite Belt is over fifty kilometres long and is scattered by several fault and shear zones into smaller pieces. We will concentrate on the biggest slab, which is best exposed in Isokumpu area.

The granulites in Isokumpu are mainly felsic enderbites and charnockites, which are highly metamorphosed, orthopyroxene-bearing equivalents to tonalites and granites. In addition to orthopyroxene, these rocks contain quartz, plagioclase, garnet and alkali feldspar (in granites over 10 %).

Intercalated with the felsic granulites there are mafic two pyroxene granulites, in which two geochemically distinctive series can be separated. The first one consists mainly of rocks containing brownish hornblende, and abundant plagioclase, orthopyroxene and clinopyroxene. The REE-pattern to these rocks is very flat indicating to mantle origin. The composition of these rocks corresponds to Mg-rich tholeites. The second series consist of rocks, which in addition to pyroxenes contains only plagioclase and minor or not at all hornblende. The REE-patterns of these rocks are more differentiated and indicate to influence of crustal material. These rocks correspond to Fe-rich tholeites by their chemical composition. There are also third kind of rocks consisting of plagioclase, quartz, orthopyroxene and clinopyroxene. These rocks are intermediate and considering their geochemistry could represent a more differentiated part of the second series.

The granulites in Isokumpu area represent fairly low-grade granulite facies metamorphism. In felsic granulites metamorphic pressures vary from 2,5 to 5,5 kbars and temperatures 500-650 °C. In mafic granulites metamorphic temperature is 700-800 °C and the pressure is at the most 6,5-7 kbars.

Considering the Nb and Ti minimum in the mantle normalized multi-element diagrams of the granulites it is possible that they have formed in a volcanic arc environment.

2.44 Ga Bimodal Magmatism in Koillismaa and Adjacent Russia: the Nd Isotopic Story

Laura S. Lauri[1], O. Tapani Rämö[1], Tuomo Karinen[2] and Hannu Huhma[3]

[1]Department of Geology, P.O. Box 64, FIN-00014 University of Helsinki, Finland
[2]Department of Geology, FIN-90014 University of Oulu, Finland
[3]Geological Survey of Finland, P.O. Box 96, FIN-02151 Espoo, Finland
E-mail: laura.lauri@helsinki.fi

Nd isotopic characteristics were determined for the ~2.44 Ga Koillismaa layered igneous complex, the Kynsijärvi quartz alkali feldspar syenite, volcanic rocks in the Koillismaa area and the Nuorunen granite in adjacent Russia. Samples from the surrounding gneiss complex were also analyzed to determine the isotopic composition of the surrounding Archean crust. The Nd isotope characteristics of the mafic layered intrusion may reflect the isotopic composition of the subcontinental mantle at ~2.44 Ga.

Keywords: Koillismaa, layered intrusion, Nd isotope, Karelian Province, Belomorian Belt, Paleoproterozoic, continental rifting

1. Introduction

The Archean Domain of the Fennoscandian Shield consists of at least two major units, the Karelian Province in the central part of the shield and the Belomorian Belt on its northeastern side. The Karelian Province is a typical granite-greenstone terrain the age of which is 3.1–2.7 Ga whereas the 2.8–2.5 Ga Belomorian Belt is a high-grade gneiss terrain (*Gaál and Gorbatschev, 1987; Gorbatschev and Bogdanova, 1993*). Although it is not clear when the Archean blocks amalgamated, it is obvious they were part of the same continent by the end of the Archean.

Early Paleoproterozoic rifting of the Archean craton at 2.50–2.44 Ga led to the emplacement of several mafic layered intrusions and a few silicic plutons. The Koillismaa layered igneous complex (*Alapieti, 1982*) and the Kynsijärvi quartz alkali feldspar syenite (*Lauri and Mänttäri, 2002*) were emplaced in Koillismaa area and the Oulanka layered complex (*Lavrov, 1979*) and the Nuorunen granite (*Buyko et al., 1995*) in adjacent Russia, near the Karelian–Belomorian suture (Fig. 1). Volcanic activity also occurred in both areas at this time (*e.g., Lauri et al., in prep.*).

2. Nd isotopic data

Samples for Nd isotopic studies were collected from all the ~2.44 Ga rock types in the Koillismaa area and from the Archean granite gneiss complex surrounding the layered intrusions. Gabbroids are from the layered series of the Porttivaara block, Koillismaa complex, granophyres from Murtolampi, Kuusijärvi, Porttivaara and Pirinvaara areas, and mafic and felsic volcanic rocks from outcrops close to the granophyre in Porttivaara and Kuusijärvi. Samples from the Kynsijärvi quartz syenite were taken from drill cores as the pluton is poorly exposed, and three samples were analyzed from the Nuorunen granite (Fig. 1). The Sm-Nd determinations were made at the isotope laboratory of the Geological Survey of Finland.

The results of the Nd isotope studies are presented in Table 1 and Fig. 2. The analyzed Archean samples show ε_{Nd} (at 2440 Ma) values mostly between –8.5 and –5 and T_{DM} model ages (*after DePaolo, 1981*) of 2900–3100 Ma. This is, in general, consistent with the data acquired from other Archean rocks in Finland (*Huhma, 1986; O'Brien et al., 1993; Hölttä et al., 2000; Hanski et al., 2001*). However, one Archean sample is more radiogenic with ε_{Nd} (at 2440 Ma) value of –2.1 and T_{DM} model age of 2800 Ma.

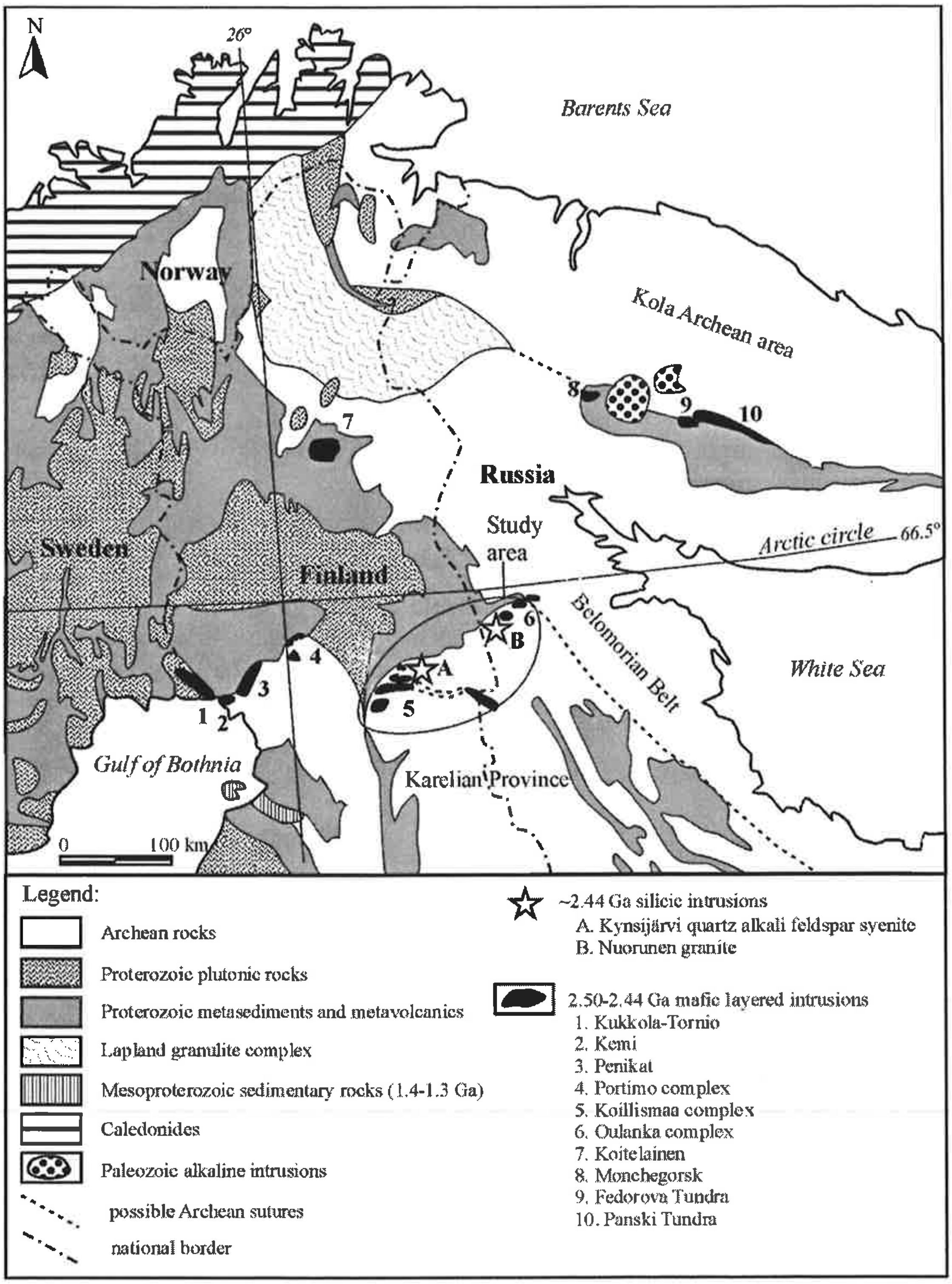


Figure 1. General bedrock map of the northern Fennoscandian Shield and locations of the ~2.44 Ga mafic and silicic intrusions (modified from Lauri and Mänttäri, 2002).

Table 1. Nd isotopic compositions from Koillismaa and Nuorunen

Sample	Rock type	ε_{Nd}(2440 Ma)
Koillismaa silicic and volcanic rocks		
LSL-98-20	granophyre	-1.8
LSL-98-92	granophyre	-2.1
LSL-98-47	mafic volcanite	-2.4
LSL-00-14	felsic volcanite	-1.9
LSL-00-8	granophyre	-1.9
LSL-00-10	granophyre	-2.3
LSL-00-65	granophyre	-5.3
LSL-00-83	granophyre	-3.3
LSL-00-97	granophyre	-2.1
LSL-99-262	granophyre	-1.1
LSL-99-263	granophyre	-2.3
A1585	q-afsp-syenite	-4.4
Kynsi	q-afsp-syenite	-7.5
R351	q-afsp-syenite	-4.9
R356	q-afsp-syenite	-5.2
368-TTK-00	felsic volcanite	-4.6
Nuorunen granite		
Nuo-1	granite	-1.9
Nuo-2	granite	-2.1
Nuo-3	granite	-1.8
Archean rocks		
A415	granite	-5.0
A1656	granite	-6.7
A1657	granite	-8.5
A1642	granite	-6.1
A1644	gneiss	-2.1
A1661	px-tonalite	-7.6
A1662	tonalite	-5.3
Koillismaa layered igneous complex		
TTK-00-335	UZb mtgb	-0.8
TTK-00-320	MZ gb	-2.0
TTK-00-309	MZ gbnor	-1.3
TTK-01-14	LZc	-1.7
TTK-00-300	LZb gbnor	-2.0
TTK-00-277	LZb gbnor	-2.1
TTK-00-265	LZa olgbnor	-1.5

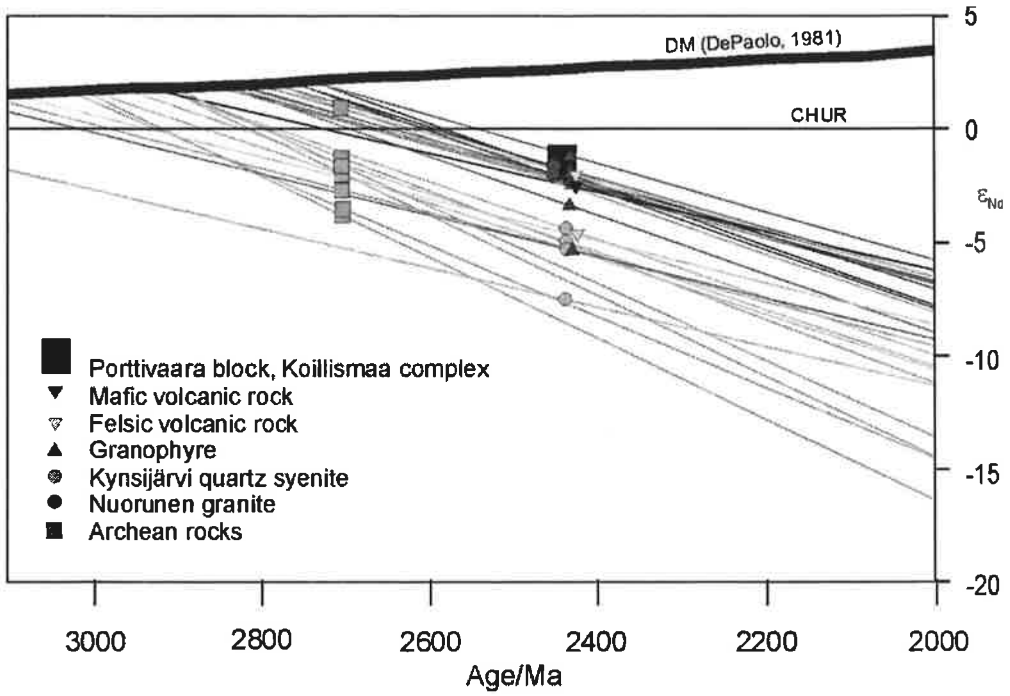


Figure 2. ε_{Nd} vs. age –diagram on the Koillismaa Archean and Paleoproterozoic rocks. Depleted mantle (DM) evolution curve after *DePaolo (1981)*.

The Kynsijärvi quartz syenite falls within the evolution path of the Archean crust with ε_{Nd} (at 2440 Ma) values between –4.4 and –5.2 and T_{DM} model ages of 2900–3000 Ma. One sample with a more negative ε_{Nd} value (–7.5) and model age in excess of 3400 Ma is probably disturbed, as the Sm/Nd ratio and elemental concentrations are anomalous. The Nuorunen granite shows notably less negative values with ε_{Nd} (at 2440 Ma) between –1.8 and –2.1 and T_{DM} model ages of around 2750 Ma.

The Koillismaa complex layered sequence shows quite homogenous ε_{Nd} (at 2440 Ma) values between –0.8 to –2.1 (mean value –1.6 ± 0.5 [1 s.d.], n = 7) and T_{DM} model ages of 2800–3000 Ma. The values in the granophyre and the volcanic rocks are much more scattered with ε_{Nd} (at 2440 Ma) between –1.1 and –5.3 and T_{DM} model ages of 2800–3200 Ma.

3. Discussion

The Archean gneiss complex in the Koillismaa area is poorly dated. The only age so far published is by *Lauerma (1982)*, who describes a 2.7 Ga granite body, giving a minimum age for the gneiss the granite intrudes. The Nd isotopes define a "typical" Archean trend with ε_{Nd} (at 2440 Ma) between –5 and –8.5 and model ages of 2900–3100 Ma. One sample shows a more radiogenic Nd isotope composition (–2.1) and slightly younger T_{DM} model age of 2800 Ma and may reflect age differences of the Archean basement in the Koillismaa area.

The Nd isotope characteristics of the Koillismaa silicic rocks seem to follow the heterogeneity of the Archean basement in the area. The intrusive Kynsijärvi quartz syenite, shows relatively little scatter but the volcanic rocks show a range of values possibly reflecting the characteristics of their source, as most of the samples do not seem to show disturbedisotope systems. The Nuorunen granite, that has intruded close to the suture between the Karelian Province and the Belomorian Belt (Fig. 1), shows a more radiogenic Nd isotopic composition and younger T_{DM} model ages.

The initial Nd isotopic composition of Porttivaara layered series is astonishingly uniform. The seven whole-rock samples analyzed so far have ε_{Nd} values that barely exceed the experimental error (Table 1, Fig. 2) and define an initial ε_{Nd} (at 2440 Ma) of ~−1.5. Whether this represents the composition of the concurrent subcontinental mantle or reflects crustal contamination of the mafic magma is yet to be determined. In view of the large volume of the mafic magma (Alapieti 1982) and homogeneous initial composition, the former hypothesis seems more likely.

4. Conclusions

The Kynsijärvi quartz syenite and the Nuorunen granite show different initial Nd compositions that may reflect the age differences in the surrounding Archean. It may be possible to define the boundary between the Karelian Province and the Belomorian Belt by Nd isotopic studies. The initial composition of the Koillismaa layered igneous complex may represent that of the subcontinental mantle at 2.44 Ga.

References

Alapieti, T. 1982. The Koillismaa layered igneous complex, Finland – its structure, mineralogy and geochemistry, with emphasis on the distribtion of chromium. *Geol. Surv. Finland, Bull.319,* 116 p.

Buyko, A. K., Levchenkov, O. A., Turchenko, S. I. and Drubetskoi, Je. R.,1995. Geologija I izotopnoje datirovanije ranneproterozoiskogo sumiisko-sarioliiskogo kompleksa Severnoi Karelii (Paanajarvi-Tsipringskaja struktura). *Stratigrafija Geologicheskaja korreljatsija, T.3, No 4,* 16–30 (*in Russian*).

DePaolo, D. J. 1981. Neodymium isotopes in the Colorado Front Range and crust–mantle evolution in the Proterozoic. *Nature 291,* 684–687.

Gaál, G. and Gorbatschev, R. 1987. An outline of the evolution of the Baltic Shield. *Precambrian Res. 35,* 15–52.

Gorbatschev, R. and Bogdanova, S., 1993. Frontiers in the Baltic Shield. *Precambrian Res. 64,* 3–21.

Hanski, E., Walker, R. J., Huhma, H. and Suominen, I. 2001. The Os and Nd isotopic systematics of the c. 2.44 Ga Akanvaara and Koitelainen mafic layered intrusions in northern Finland. *Precambrian Res. 109,* 73–102.

Hölttä, P., Huhma, H., Mänttäri, I. and Paavola, J. 2000. P-T-t development of Archaean granulites in Varpaisjärvi, central Finland. II. Dating of high-grade metamorphism with the U–Pb and Sm–Nd methods. *Lithos 50(1–3),* 121–136.

Huhma, H. 1986. Sm–Nd, U–Pb and Pb–Pb isotopic evidence for the origin of the Early Proterozoic Svecokarelian crust in Finland. *Geol. Surv. Finland, Bull. 337,* 48 p.

Lauerma, R. 1982. On the ages of some granitoid and schist complexes in northern Finland. *Bull. Geol. Soc. Finland 54, 1-2,* 85–100.

Lauri, L. S. and Mänttäri, I. 2002. The Kynsijärvi quartz alkali feldspar syenite, Koillismaa, eastern Finland – silicic magmatism associated with 2.44 Ga continental rifting. *Precambrian Res.* (*in press*).

Lauri, L. S., Karinen, T. and Räsänen, J. The Koillismaa granophyre – observations on its characteristics and relations to adjacent rocks. *Bull. Geol. Soc. Finland* (*submitted*).

Lavrov, M. M., 1979. Giberbasity I rasslojennie peridotit-gabbro-noritovie intrusii dokembria Severnoi Karelii. *NAUKA, Leningrad, Soviet Union* (*in Russian*).

O'Brien, H., Huhma, H. and Sorjonen–Ward, P. 1993. Petrogenesis of the late Archean Hattu schist belt, Ilomantsi, eastern Finland: geochemistry and Sr, Nd isotopic composition. In: Nurmi, P. A. and Sorjonen–Ward, P. (eds.) Geological development, gold mineralization and exploration methods in the late Archean Hattu schist belt, Ilomantsi, eastern Finland. *Geol. Surv. Finland, Spec. Paper 17,* 147–184.

Postglacial Rebound and Crustal Deformation in Finland: Recent Geodetic Results

Jaakko Mäkinen[1], Joel Ahola[1], Jorma Jokela[1], Hannu Koivula[1], Martti Ollikainen[1], Markku Poutanen[1], Veikko Saaranen[1], Mikko Takalo[1] and Martin Vermeer[2]

[1]Finnish Geodetic Institute, Geodeetinrinne 2, FIN-02430 Masala, Finland
[2]Department of Surveying, P.O. Box 1200,
FIN-02015 Helsinki University of Technology, Finland
E-mail: Jaakko.Makinen@fgi.fi

We summarize recent geodetic results obtained in the study of postglacial rebound and intra-plate tectonics in Finland.

Keywords: Postglacial rebound, glacial isostatic adjustment, intra-plate tectonics, GPS, precise levelling

1. Postglacial rebound—vertical

The classical methods to determine vertical crustal motion are repeated precise levellings and tide gauge records. In Finland we have three country-wide precision levellings: The First Levelling (1892–1910) which only extended to the line Kemi–Oulu–Kajaani, the Second Levelling (1935–1975) which covered all of Finland, and the Third Levelling which was started in 1978 and will be completed in 2003. Figure 1 shows the vertical velocities determined from all three levellings.

For geodynamical studies, continuous GPS on permanent stations has quickly become the method of choice. From five years of data (1996–2001) in the permanent Finnish GPS network FinnRef®, uplift rates can be determined with the same precision as from the levellings spanning one hundred years. Figure 2 shows a comparison between levelled rates, tide gauge rates and GPS rates (*Mäkinen et al., 2002*). From the comparison, all three methods are good to at least 0.4 mm/yr (one-sigma).

Strictly speaking, the methods have different reference surfaces. Repeated precise levelling gives velocity differences relative to the geoid ("levelled rates"). Tide gauges give velocities relative to mean sea level ("apparent rates"). GPS rates formally refer to the Earth's center of mass ("absolute rates").

The difference between "levelled" rates and apparent rates is the eustatic rise in mean sea level, 1…2 mm/yr for 1890–1990, plus possible change in mean sea surface topography. The difference between "absolute" rates and "levelled" rates is the uplift rate of the geoid. For the glacial isostatic adjustment process, it is less than 10% of the absolute rates.

"Absolute" for GPS rates is misleading; in practice GPS results are always obtained relative to something. For instance, relative to the frame defined by the coordinates and velocities of an ensemble of reference stations, or relative to a fixed point in the network.

2. Postglacial rebound—horizontal

GPS gives 3-D velocities. The horizontal motion in the glacial isostatic adjustment (GIA) is nearly one order of magnitude smaller than the vertical, but on the other hand the accuracy of the GPS in the horizontal is about three times better. Note that due to the curvature of the Earth, even purely radial motion (i.e., no tangential components) will result in the change in the lengths of the vectors between the stations. Elementary geometrical considerations show that for elevation changes Δh_1 and Δh_2 at two points, the relative change in baseline length s

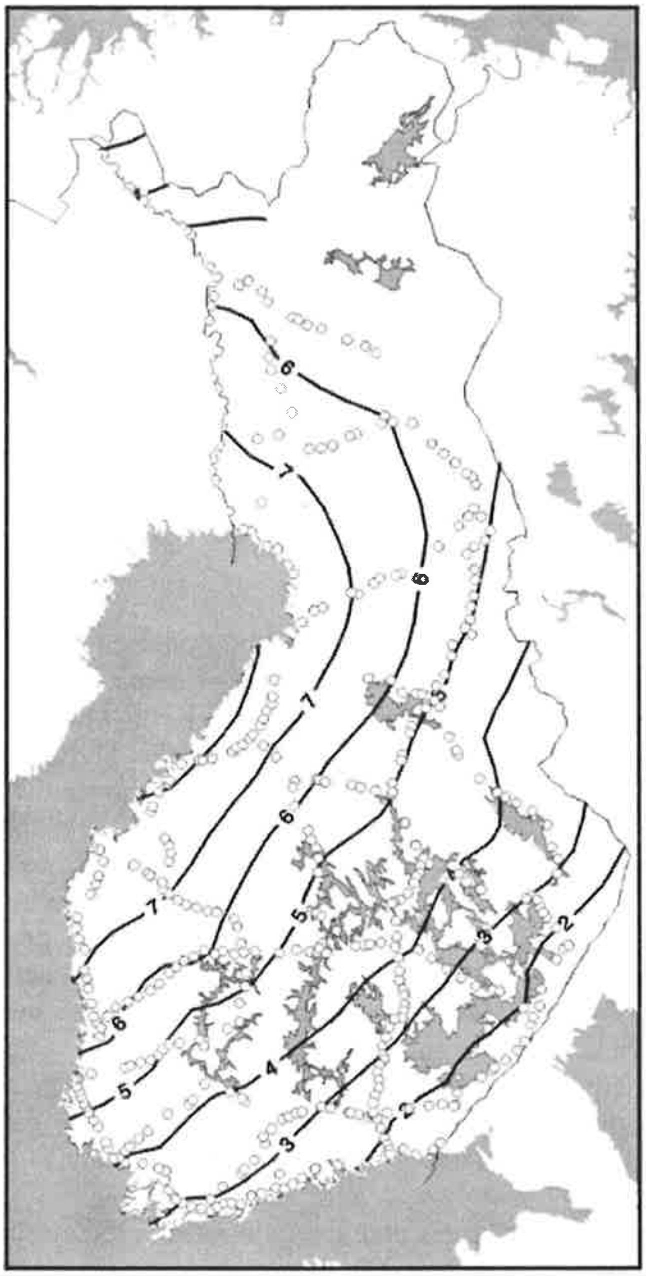


Figure 1. Postglacial rebound rates relative to mean sea level from three precise levellings. Hanko tide gauge (2.73 mm/yr from *Vermeer et al., 1988*) provides the zero.

between them is $\Delta s / s = (\Delta h_1 + \Delta h_2)/2R$ where R is the radius of the Earth. This is valid as soon as $|(\Delta h_1 - \Delta h_2)|/s << s/R$, i.e. already at very short distances. Thus if two stations have uplift 6 mm/year, the strain rate between them is 1×10^{-9}/yr.

Figure 3 shows the horizontal motion from 5 years of FinnRef® (1996–2001). Metsähovi was kept fixed, and the time series of the 3-D vectors relative to it were transformed to horizontal and vertical motion at the end points. This method of plotting confounds radial and tangential motion for the reasons just explained, but it gives the horizontal strain rates relative to the hub point.

Figure 4 shows an example of FinnRef® time series.

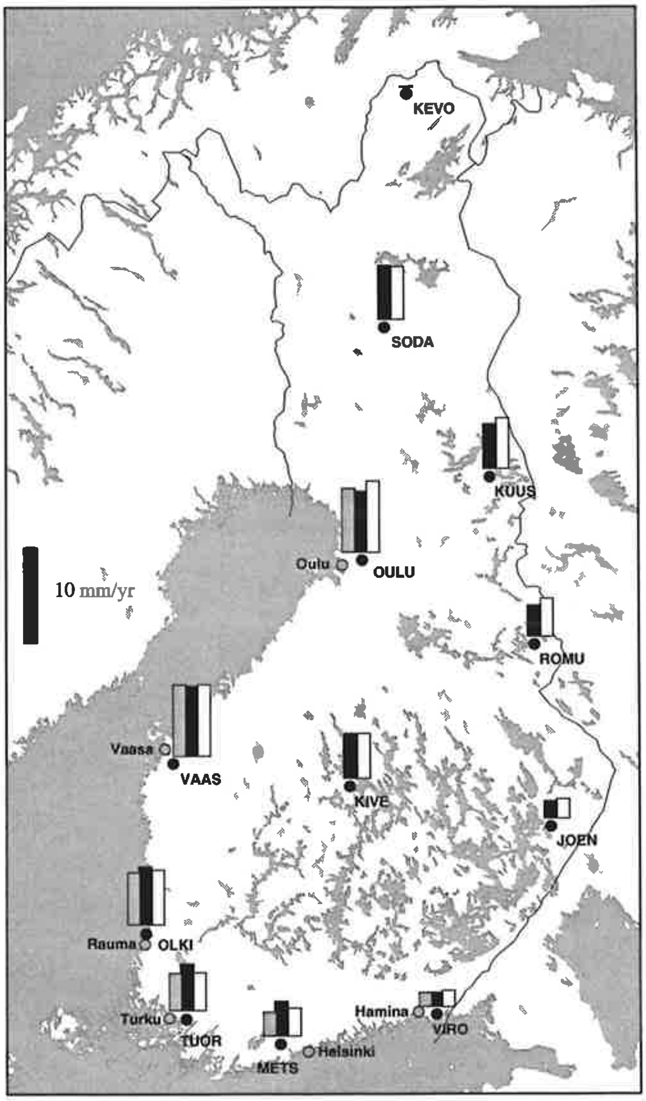


Figure 2. Postglacial rebound rates at FinnRef® stations from three precise levellings (white columns), FinnRef® (black columns), and nearest tide gauge (grey columns). FinnRef® rates are shifted to have the same mean as levelling rates. Tide gauge rates of Ekman (1996) are transferred to the GPS station using precise levelling.

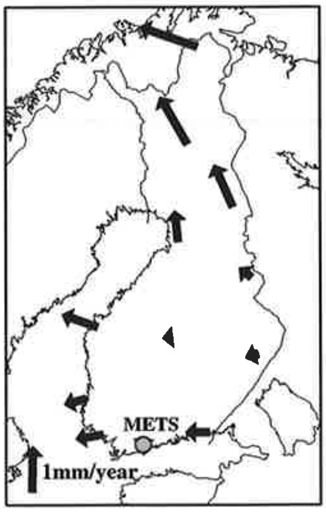


Figure 3. Horizontal velocities from FinnRef® relative to Metsähovi. The rotation of the Eurasian plate according to NNR-NUVEL-1A has been subtracted. The rates are larger than purely vertical motion can explain (see text). Referring them to Vaasa would produce an approximately concentric pattern away from the uplift center.

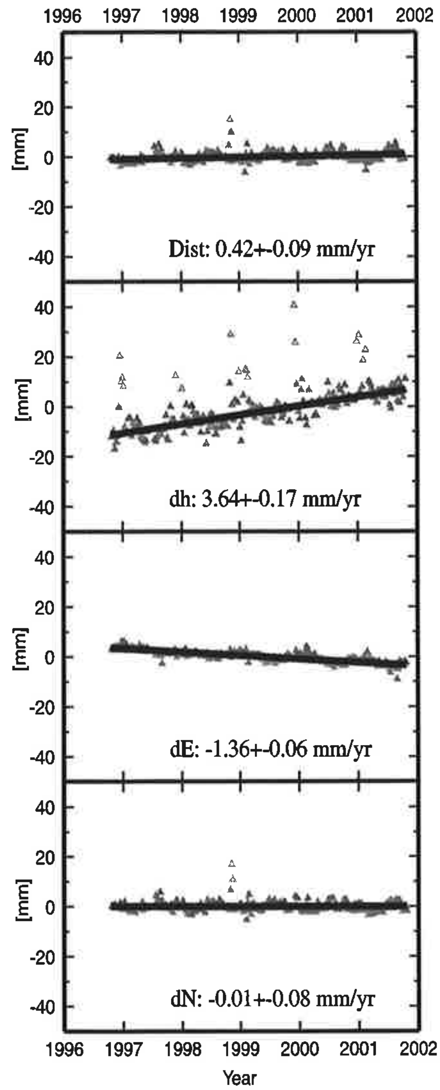


Figure 4. Finnref® time series at Vaasa relative to Metsähovi 1995–2001, weekly coordinate series. From top to bottom: Vector length, vertical, east, north. Zeroes are arbitrary. Outliers in the vertical series are due to snow on the antenna in winter. The annual period (Poutanen et al., 2002) is probably due to loading by atmosphere, oceans, and continental water storage.

3. Geophysical models of the glacial isostatic adjustment

The purpose of geophysical models of the glacial isostatic adjustment (GIA) is usually to draw inferences on the rheological parameters of the Earth and/or on ice history (which then opens the road to paleoclimatology, among other things; see *Peltier, 1998*). Typically, the parameters are lower and upper mantle viscosity (sometimes with finer stratification) and the (effective elastic) thickness of the lithosphere. Other relevant Earth properties are fixed at their values from seismology. Linear Maxwell rheology of the mantle is ubiquitous. Even today, lateral variation in the unknown parameters (viscosity, thickness) is hard to handle. The main data have been relative sea level (RSL) histories since deglaciation. Either a Bayesian inversion is done, or/and a larger number of forward calculations, searching the parameter space for the minimum of the error function.

Only recently contemporary motion results have been used as input data. *Lambeck et al. (1998)* fine-tune their model (primarily derived from RSL histories) on tide gauge and lake tilt results in Fennoscandia. In the work by *Milne et al. (2001)*, the 3-D velocities, obtained in the project BIFROST from FinnRef® and the Swedish permanent GPS network SWEPOS are the only observational data but the ice model is fixed at that of *Lambeck et al. (1998)*.

Within their uncertainties, the solutions using contemporary motion are consistent with those obtained using RSL histories: *Milne et al. (2001)* find the effective lithospheric thickness 120 km and the upper mantle viscosity 8×10^{20} Pa s . The uncertainties are still large, reflecting in part the shortness of the GPS record. However, with increased time span and perhaps improved spatial coverage it is obvious that the GPS data have great potential. Horizontal motion provides new information as it "sees" upper mantle viscosity and ice history differently from vertical motion. Combining RSL records and contemporary motion data one can reduce the trade-off between ice history and rheological parameters in the solution. Comparisons of RSL and contemporary solutions might make it possible to further test whether mantle viscosity is the same at different timescales.

For references to work on gravity change and geophysical models, see *Mäkinen (2000).*

4. Tectonic motion

Finland, as a part of the Fennoscandian shield participates in the motion of the Eurasian plate. Here we are only interested in intra-plate tectonics. A hotly debated subject for some time has been whether there actually is conclusive geodetic evidence for such present day motion in the Fennoscandian shield (*Kakkuri, 1997*).

Comparing classical triangulation and trilateration, *Chen (1991)* found evidence for horizontal strain of up to 5×10^{-7}/yr (at the scale of 25-30 km and upwards). *Veriö et al. (1993)* performed re-levellings over tens of fracture lines and discovered irregular height changes of up to 3 cm. *Lehmuskoski (1996)* screened the network of the three precision levellings and found 34 sites where the change in uplift rate exceeds three times its a-priori standard error. *Mäkinen and Saaranen (1998)* pointed out that nevertheless the population of all changes is compatible with the normal distribution where a-priori errors are scaled by 1.4.

So far, the GPS time series in the FinnRef® network do not show any irregular motion, changes in trend etc. (Figure 4)

The Finnish Geodetic Institute is currently pursuing a number of local deformation studies which are shortly reported here.

5. Nuottavaara fractures

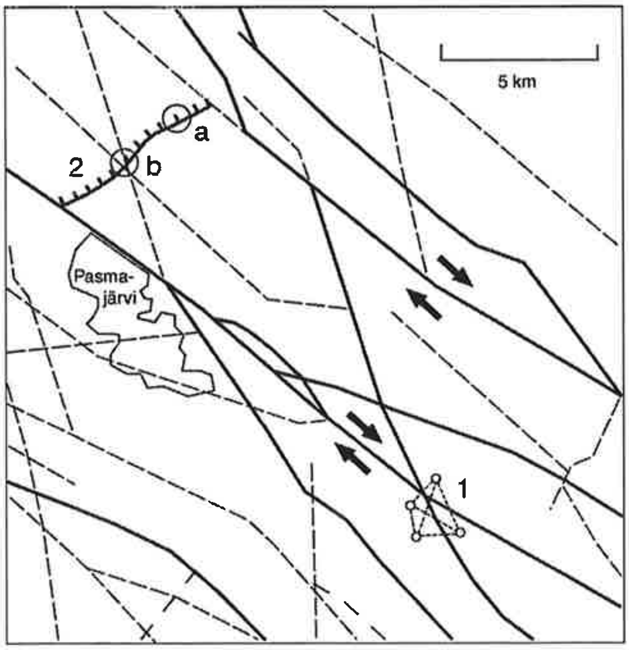
An old fracture zone complex is located in the North Finland about 50 km from Kolari in the Nuottavaara area (Figure 5). The movement of the fracture zone is suspected to have caused the birth of the nearby Pasmajärvi postglacial fault (*Kuivamäki et al. 1998*), see next section. A four-point network was established in Nuottavaara (Figure 5) in 1991 in cooperation with the Geological Survey of Finland. The distances between the markers in bedrock are 1.5 to 3 km.

The network was measured with GPS in 1991, 1992, 1995 and 2000 (*Poutanen and Ollikainen 1995, Ahola 2001)*. No detectable movements have taken place. Possible motions must be smaller than 1 mm/yr.

6. Pasmajärvi postglacial fault

This well-known fault is 4 km long and the height of the fault scarp is 5 to 12 m (see *Kuivamäki et al., 1998*, and the references therein). In 1987–1990 two geodetic networks were established across the fault (Figure 5). The NE network has benchmarks in bedrock. In the SW network the benchmarks are on pillars of concrete poured on the bedrock, which is at 2…5 m depth in till.

High precision angle and length measurements were performed in 1988, 1990, 1992, 1994 and 2002. Precise levelling was done in 1988, 1990, 1994, 2000, and 2001. Trigonometric



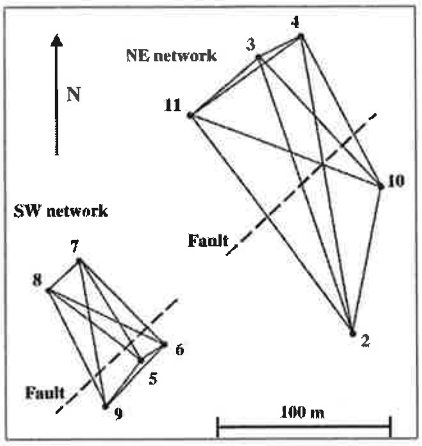


Figure 5. Left: (1) Nuottavaara fractures with network, at about 67°04'N, 24°36'E. (2) Pasmajärvi fault with (a) NE network and (b) SW network. The picture is due to *Kuivamäki et al. (1998).* Right: The Pasmajärvi networks.

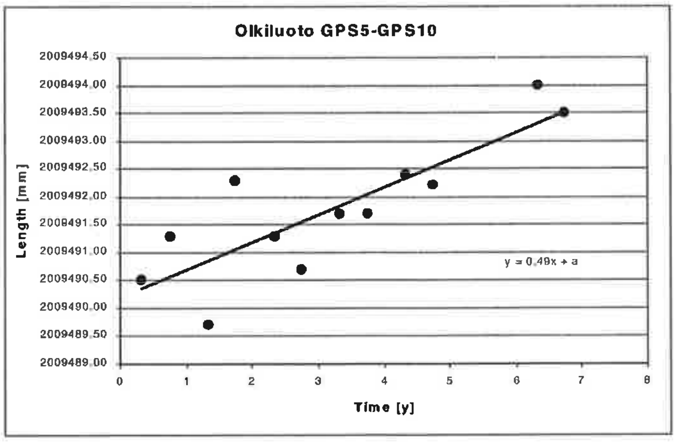


Figure 6. The change rate between the pillars GPS5 and GPS10 at Olkiluoto is statistically significant (0.49 mm/y ± 0.10 mm/y). In the later campaigns the scatter is much smaller than in the beginning. This is due to 24-hour GPS sessions instead of 8 hours.

levelling was performed in 1988, 1990, and 1992. In addition there is a separate two-point gravity line measured in 1987, 1989, 1991, 1993, 1995 and 2000. The 2001–2002 round of measurements was in part motivated by a nearby earthquake on May 2, 2001 (magnitude 2.9, epicentre at 67°11 N', 24°40' E, depth 7 km).

First results were reported by *Kiviniemi et al. (1992).* Horizontal movements in the SW network between 1988 and 1990 were initially found by *Konttinen (op.cit.)*, but could have been due to settling down of the recently poured pillars. Preliminary comparisons between the angle and length measurements of 1990, 1992, 1994 and 2002 show 1...2 mm differences in adjusted coordinates in both networks.

Precise levelling constrains possible vertical motion to be a fraction of millimetre, with the exception of 1 mm motion of one point in the SW network. This is preliminarily interpreted as pillar movement. A report on all geodetic observations at Pasmajärvi and Nuottavaara will be published shortly (*Takalo et al., 2002*).

7. Posiva micronetworks

In cooperation with Posiva Ltd., high precision GPS networks were established at Olkiluoto (61°14'N, 21°28'E), Kivetty (62°49'N, 25°42'E) and Romuvaara (64°13'N, 29°55'E) in 1994–1995. The purpose is to study local crustal deformation at places selected as candidates for disposal of spent nuclear fuel. The studies are now concentrated at Olkiluoto.

The GPS networks include ten reinforced concrete pillars with antenna mounts in Olkiluoto and seven at Kivetty and at Romuvaara. The stations are on solid bedrock and they are located on different geological blocks. The distances between the pillars are 0.5 to 3.5 km. One station at each investigation area belongs to FinnRef®. Every station has two reserve markers and pillar stability is controlled regularly.

The GPS networks have been observed twice a year since 1995. No measurements were made in 2000 because ionospheric activity was close to the maximum, a major disturbance for GPS. The most recent annual report is by *Ollikainen et al. (2002)*. Up till now 13 campaigns have been performed at Olkiluoto, and 10 at Kivetty and Romuvaara. The rms accuracy of the GPS results from a single campaign is about 1 mm. At a few baselines change rates are around $0.5 \pm 0.1 \ldots 0.2$ mm/yr (Figure 6), but generally the networks are quite stable. We will continue GPS observations at Olkiluoto twice a year and at Kivetty and Romuvaara annually.

References

Ahola, J., 2001. Nuottavaaran siirroksen deformaatiotutkimus GPS-mittausten avulla. Deformation studies of the Nuottavaara fault with GPS measurements. In Finnish with an English abstract. *M.Sc. Thesis at the Helsinki University of Technology*. Espoo.

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

Ekman, M., 1996. A consistent map of the postglacial uplift of Fennoscandia. *Terra Nova* 8, 158–165.

Kakkuri, J., 1997. Postglacial deformation of the Fennoscandian crust. *Geophysica* 33(1), 99–109.

Kiviniemi, A., Konttinen, R., Kuivamäki, A., Takalo, M. and Vuorela, P., 1992. Geodetic observations on the Pasmajärvi postglacial fault. *Rep. Finn. Geod. Inst.* 92:2. Helsinki.

Kuivamäki, A., Vuorela, P. and Paananen, M., 1998. Indications of postglacial and recent bedrock movements in Finland and Russian Karelia. *Geological Survey of Finland, Nuclear Waste Disposal Research, Report YST-99*. Espoo.

Lambeck, K., Smither, C., Ekman, M., 1998. Tests of glacial rebound models for Fennoscandia based on instrumented sea- and lake-level records. *Geophys. J. Int.* 135, 375–387.

Lehmuskoski P., 1996. Active fault line search in Southern and Central Finland with precise levellings. *Rep. Finn. Geod. Inst.* 96:5. Kirkkonummi.

Mäkinen, J., 2000. Geodynamical studies using gravimetry and levelling. Academic Dissertation. Yliopistopaino, Helsinki, 82 p. Summary *http://ethesis.helsinki.fi*, 18 p.

Mäkinen J. and Saaranen V., 1998. Determination of postglacial land uplift from the three precise levellings in Finland. *J. Geodesy* 72, 516–529.

Mäkinen, J., Koivula, H., Poutanen, M., Saaranen, V., 2002. Vertical velocities in Finland from permanent GPS networks and from repeated precise levelling. *J. Geodynamics*, accepted for publication.

Milne, G.A., Davis, J.L., Mitrovica, J.X., Scherneck, H.-G., Johansson, J.M., Vermeer, M., Koivula, H., 2001. Space-Geodetic Constraints on Glacial Isostatic Adjustment in Fennoscandia. *Science* 291, 2381–2385.

Ollikainen, M., Ahola, J. and Koivula, H., 2002. GPS operations at Olkiluoto, Kivetty ja Romuvaara for 2001. *Working Report 2002-16. Posiva Oy*. Helsinki.

Peltier, W., 1998. Postglacial variations in the level of the sea: Implications for climate dynamics and solid-earth geophysics. *Rev. Geophysics* 36(4), 603–689.

Poutanen, M. and Ollikainen, M., 1995. GPS-measurements at the Nuottavaara postglacial fault. *Rep. Finn. Geod. Inst.*, 95:6. Kirkkonummi.

Poutanen, M., Koivula, M., Ollikainen, M., 2002. On the periodicity of GPS time series. *Proceedings of IAG Scientific Assembly, Budapest, Hungary 2–7.9.2001*. In print.

Takalo, M., Ahola, J., Jokela, J., Rouhiainen, P. and Ruotsalainen, H., 2002. Geodetic observations at the Pasmajärvi and Nuottavaara faults. *Rep. Finn. Geod. Inst.*, in preparation.

Veriö, A., Kuivamäki, A., Vuorela, P., 1993. Kallioperän murroslinjojen nykyliikunnoista. Maanmittauslaitoksen murroslinjavaaitukset 1974–1992. Recent displacements within reactivated fracture systems in Finland; Results and analysis of levelling measurements carried out by the National Land Survey of Finland in 1974–1992. In Finnish with an English summary. *Geological Survey of Finland, Nuclear Waste Disposal Research, Report YST-84*. Espoo.

Vermeer, M., Kakkuri, J., Mälkki, P., Boman, H., Kahma, K.K, Leppäranta, M., 1988. Land uplift and sea level variability spectrum using fully measured monthly means of tide gauge readings. *Finnish Marine Research* 256, 3–75.

Paleoproterozoic Tectonic Evolution of the Fennoscandian Shield – Comparison to Modern Analogues

Mikko Nironen[1], Raimo Lahtinen[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: mikko.nironen@gsf.fi

A plate tectonic model for the evolution of the Fennoscandian Shield, involving >2.0 Ga microcontinents, pre-1.92 Ga island arcs, and 1.90-1.83 Ga arcs that accreted against an assembly of three Archean cratonic units is presented. The complex evolution is divided into five orogenies and involved a microcontinent accretion stage, a continental extension stage, a continental collision stage, and an orogenic collapse and stabilization stage. Each stage is compared with one or several modern analogues.

Introduction

The Precambrian bedrock of Finland forms the core of the Fennoscandian Shield, which is part of the East European Craton. Based on an integrated study of lithological, petrological, geochronological, potential field, deep seismic reflection and refraction, and geoelectric data we propose a tectonic model for the Paleoproterozoic evolution of the Fennoscandian Shield (Lahtinen et al., 2003). We reject the concept of a semi-continuous Svecokarelian/ Svecofennian orogeny and define five orogenies for the time period 1.92-1.79 Ga instead. The orogenies that overlap partly in time and space have different structural grains. The orogenic evolution is divided into 1) a microcontinent accretion stage (1.92-1.88 Ga), 2) a continental extension stage (1.88-1.85 Ga), 3) a continent-continent collision stage (1.85-1.79 Ga), and 4) an orogenic collapse and stabilization stage (1.79-1.77 Ga). We present selected examples from the tectonic model and compare them with modern analogues. The focus is in the crustal-scale compressional and extensional periods that occurred recurrently and partly simultaneously.

Geochemical and isotopic evidence indicate that microcontinents, now seen as crustal domains, started to form already 2.1-2.0 Ga ago. The abundant 1.95-2.1 Ga ages in detrital zircon data also suggest that pre-1.92 Ga crust-forming processes were important. Based on lithological, geochemical, isotopic, and geophysical data we suggest the following pre-1.92 Ga components (Fig. 1): Karelian, Kola and Norrbotten Archean cratons; Keitele, Bergslagen and Bothnia >2.0 Ga microcontinents; Kittilä ~2.0 Ga island arc; and Savo, Knaften, Inari and Tersk ~1.95 Ga island arcs. Some of these units, e.g. Keitele and Bothnia, are hidden and have no identified surface expressions.

The microcontinent accretion stage is suggested to explain the complex assembly history between 1.92-1.87 Ga. This situation is similar to that noticed in the Banda Sea (Lee and Lawver, 1995). The Banda Sea area is one example of the occurrence of microcontinents and many coeval subduction zones. The subduction directions may be at high angle and even with opposite directions to each other as in the Mollucca Sea between Sulawesi and Halmahera in Indonesia (Hall et al., 1995). Also the current tectonic setting of the southernmost Andes, Scotia plate and northern part of Antarctic Peninsula is very complex (Diraison et al., 2000) and it has similarities to some stages in our model. The shoehorn-shaped curve of the Banda Arc shows an extremely complicated subduction system that is caused by either two separate and distinct lithospheric slabs or a rapidly eastward retreating lithospheric slab (see discussion in Milsom, 2001).

At the end of the microcontinent accretion stage a large continental plate (Fennoscandia) had developed. An attempt of orogenic collapse at the Archean-Proterozoic boundary occurred at 1.88-1.87 Ga. Large-scale extension at 1.86-1.85 Ga took place in the central part

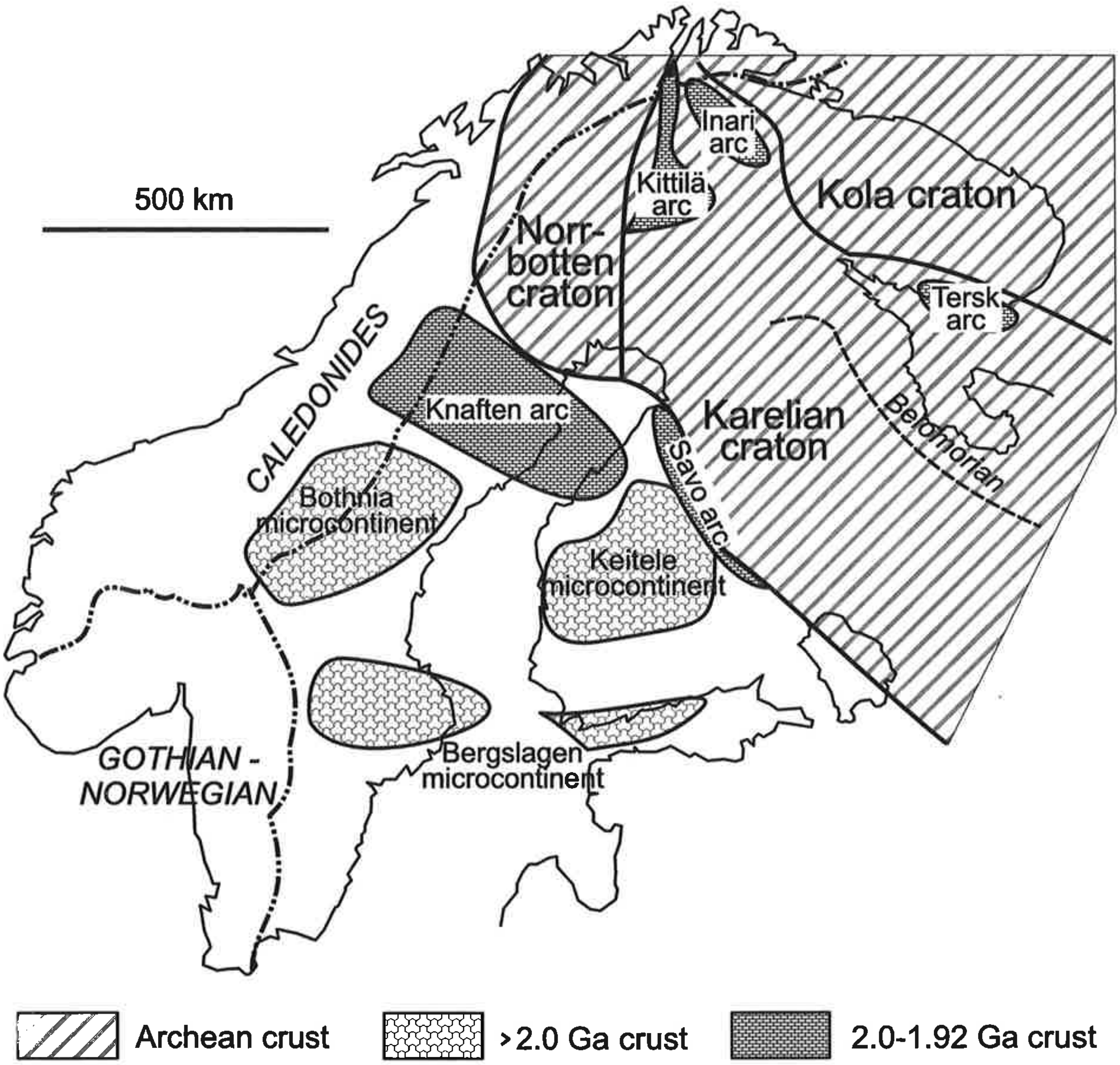


Figure 1. The present position of the pre-1.92 Ga crustal components of the Fennoscandian Shield (Lahtinen et al., 2003).

of the continent, simultaneously with subduction at the western and southern margins of the continent. The extension, resulting in the development of retro-arc or 'back-arc' basins, is compared with the Basin and Range province.

Two subduction zones, in the south and in the west, were active between 1.86 and 1.81 Ga. These intervening subduction zones at high angles are the key to the evolution of the Fennoscandian Shield; they account for the observed magmatic and structural grains. The Ryukyu and Manala Trenches around Taiwan (Chemeda et al., 2001) as well as the trenches in the Sulawesi – Halmahera area in Indonesia (Lee and Lawver, 1995) are examples of active subduction zones at high angles. The large-scale crustal extension was followed by oblique collision of Fennoscandia with Sarmatia at 1.85-1.80 Ga. A partly coeval collision of Amazonia with Fennoscandia affected the central and northern parts of the western edge of Fennoscandia at 1.82-1.80 Ga. These collisions are compared with the Himalayan and Alpine evolution.

Large-scale orogenic collapse and the stabilization of Fennoscandia occurred between 1.79 and 1.77 Ga. A modern analogue to the orogenic collapse is the presently extending Tibetan plateau (England and Houseman, 1989). The development of the stable Fennoscandian Shield is based on the thermal resetting and late tectono-magmatic episodes at 1.80-1.77 Ga on the flanks of the Fennoscandian crustal segment.

References

Chemenda, A.I., Yang, R.-K., Stephan, J.-F., Konstantinovskaya, E.A., Ivanov, G.M., 2001. New results from physical modeling of arc-continent collision in Taiwan: evolutionary model. Tectonophysics 333, 159-178.

Diraison, M., Cobbold, P.R., Gapais, D., Rossello, E.A., Le Corre, C., 2000. Cenozoic crustal thickening, wrenching and rifting in the foothills of the southernmost Andes. Tectonophysics 316, 91-119.

England, P., Houseman, G., 1989. Extension during continental convergence, with application to the Tibetan Plateau. J. Geophys. Res. 94, B12, 17561-17579.

Hall, R., Ali, J.R., Anderson, C.D., Baker, S.J., 1995. Origin and motion history of the Philippine Sea Plate. Tectonophysics 251, 229-250.

Lahtinen, R., Korja, A. and Nironen, M., 2003. Paleoproterozoic tectonic evolution of the Fennoscandian Shield – a plate tectonic model. In: Lehtinen, M., Nurmi, P. and Rämö, T. (Eds.) The Precambrian Bedrock of Finland - Key to the Evolution of the Fennoscandian Shield. Elsevier Science, B.V. Precambrian Series (in print).

Lee, T.-Y., Lawver, L.A., 1995. Cenozoic plate reconstruction of Southeast Asia. Tectonophysics 251, 85-138.

Milsom, J., 2001. Subduction in eastern Indonesia: how many slabs? Tectonophysics 338, 167-178.

Evolution of the Lithospheric Mantle in Eastern Finland at 3.0–1.95 Ga Based on NORDSIM Dating of the Jormua Ophiolite Complex

Petri Peltonen[1], Irmeli Mänttäri[1] and Asko Kontinen[2]

[1]Geological Survey of Finland, P.O. Box 96, FIN-02151 Espoo, Finland
[2]Geological Survey of Finland, P.O. Box 1237, FIN-70211 Kuopio, Finland
E-mail: petri.peltonen@gsf.fi

Continental rifting at c. 1950 Ma has provided us samples from the subcontinental lithospheric mantle which originally was underlying the Karelian craton. The mantle rocks exposed within the Jormua Ophiolite Complex represents the uppermost (<50 km) sub-crustal lithosphere of the craton. Intriquingly, mantle xenoliths representing this same continental mantle but of significantly greater depths have been recovered from 600 Ma kimberlite pipes that intrude the craton margin 100 km SSE of Jormua. Combined, an astonishing window to understand the evolution of the Karelian mantle for over ~3.5 Ga period and over its entire vertical depth seems to be at hand.

Keywords: lithosphere, ophiolite, mantle, Archean, zircon, Finland

1. Introduction

The Jormua Ophiolite Complex is an allochtonous mafic-ultramafic rock complex, thrusted onto the Karelian Craton margin. For some time the Jormua Ophiolite was considered as a slow-spreading –ridge ophiolite, i.e. a mafic-ultramafic complex consisting of oceanic crustal unit and remnants of oceanic mantle diapir that was formed in embryonic oceanic basin. More recentlly, the ideas of origin of the Jormua have undergone significant revision (*Peltonen et al., 1998; Tsuru et al., 2000; Peltonen & Kontinen, 2002*). Now we believe that the Jormua Ophiolite Complex represents a practically complete sample of seafloor from an ancient ocean–continent transition (OCT) zone – strikingly similar to that reported from e.g. the Cretaceous West Iberia non-volcanic continental margin (e.g. *Whitmarsh et al., 2001*) – and mainly consists of Archean upper continental lithospheric mantle rocks (Fig. 1).

Samples from the crustal unit of the Jormua Ophiolite, one high-level gabbro and one plagiogranite, yielded U-Pb zircon ages of 1960 ± 12 Ma and 1954 ± 11 Ma, respectively *(Kontinen, 1987)*. To define the age relationship between the crustal unit of the Jormua Ophiolite and gabbroic dykes that intrude the mantle tectonites deeper in the ophiolite stratigraphy, *Peltonen et al. (1998)* dated pegmatitic variety of gabbroic feeder dyke which yielded an high-precision age of 1953 ± 2 Ma. This age is consided to most reliably date the formation of oceanic crust in Jormua.

2. Dating the mantle rocks

The well-exposed mantle section makes the Jormua unique among ancient ophiolites. It permits the direct study of the processes that took place in the upper mantle during early Proterozoic continental break-up and formation of a new oceanic basin. Mantle lithologies in fact cover approximately 70% of the total exposure of the Jormua Ophiolite (i.e., > 30 km^2). The mantle section consists of mantle peridotites and various types of intrusive rocks types. In addition to sheeted dykes and gabbro pods the residual peridotites are intruded by chromitite pods and abundant clinopyroxenitic and hornblenditic-carbonatitic dykes and veins that do not have extrusive counterparts in the crustal unit. To constrain their emplacement age and to provide minimum age for the separation of the Jormua mantle rocks from the convective mantle we have dated zircons from 4 such dykes using the NORDSIM facility at Stockholm. These results are summarised in Table 1.

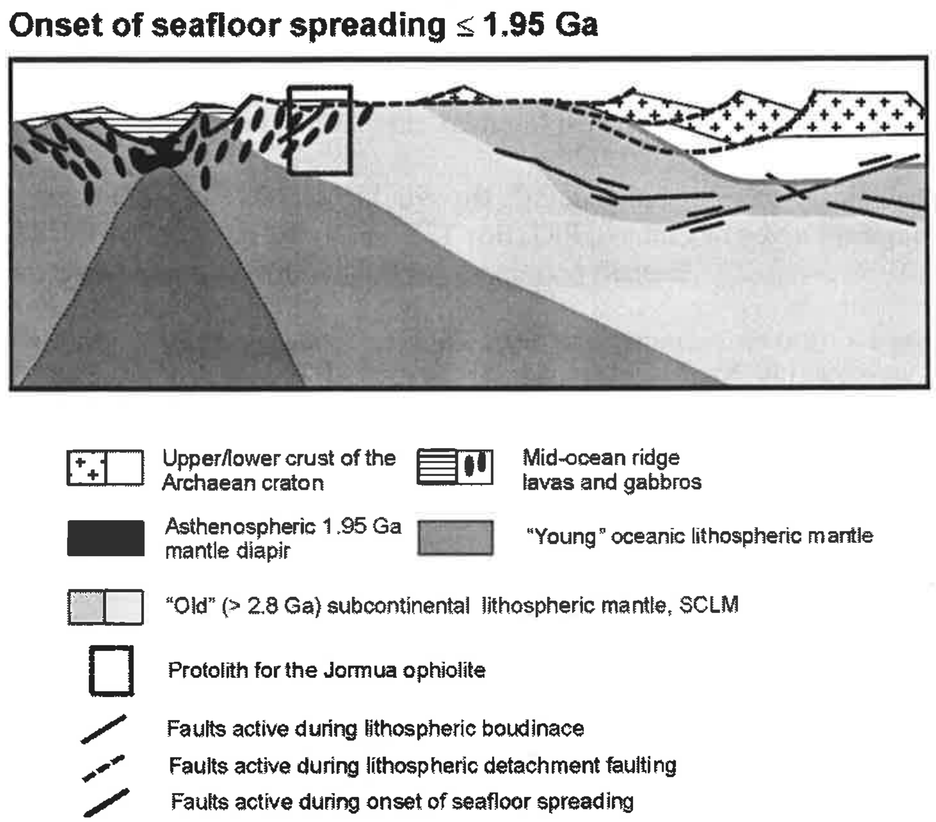


Figure 1. Proposed tectonic setting for the protolith of the Jormua Ophiolite Complex within an ocean–continent transition zone. Modified after *Whitmarsh et al. (2001)* emphasizing the situation at the present West Iberia passive margin.

Table 1. Summary of the U-Pb zircon ages of the Jormua mafic-ultramafic complex

Lithology	Sample	mineral/method	Age	Ref.
Jormua				
Gabbro	A729	zr/conventional	1960 ± 12 Ma	1
Plagiogranite	A196	zr/conventional	1954 ± 11 Ma	1
Gabbroic feeder dike	A1402	zr/microcapsule	1953 ± 2 Ma	2
Hornblenditic mantle dike	A1403	zr/microcapsule	> 1.94 Ga	2
Clinopyroxenitic mantle dike	24-ATK	magmatic zr/SIMS	2747 ± 8 Ma	3
		recrystallized zr/SIMS	~2040–1960 Ma	3
Clinopyroxenitic mantle dike	JCX-23B	magmatic zr/SIMS	~2800 Ma (3100 Ma?)	3
OIB mantle dike	106-JMD	recrystallized zr/SIMS	~2020–1960	3
Carbonatitic mantle vein	60-L	purple, high-U zr/SIMS	~2.1 Ga	3
		bright, low-U zr/sims	1948 ± 30 Ma	3

References: 1) *Kontinen, 1987*; 2) *Peltonen et al., 1998*; 3) *Peltonen and Mänttäri, unpubl.*

Two distinct types of zircon grains were recovered from clinopyroxene-dominated cumulate dyke 24-ATK (Table 2). First subgroup consists of euhedral, brownish grains with clear magmatic growth zonation. These zircons yielded Archean ages with a concordia age of 2747 ± 8 Ma (n = 7). One grain recorded even an older $^{207}Pb/^{206}Pb$ age of 2858 ± 14 Ma. The second subgroup of zircons consists of colorless rounded grains which lack magmatic growth textures. They recorded concordant ages between 2020 and 1940 Ma. An obvious interpretation of such correlation between zircon morphology and ages is that these dykes are indeed Archean in age, but went through pervasive recrystallization at ~1950 Ma due to emplacement of hot asthenospheric diapir and related magmatism. Archean age for clinopyroxenite dykes was confirmed by another clinopyroxenite dyke JCX-23 which yielded $^{207}Pb/^{206}Pb$ ages between 2.9 and 2.8 Ga. Again, one older grain (c. 3100 Ma) suggests that the true protolith age may be significantly older.

One OIB-type basaltic dyke and zircon-rich carbonatitic segregation also yielded ages in excess of that of the gabbros. These dykes (106-JMD & 60-L) contained only non-zoned rounded zircons and yielded Proterozoic ages with an spread towards higher ages. For example, purple, uranium-rich grains from the carbonatitic vein yielded ages close to 2100 Ma while bright uranium-poor grains gave significantly younger ages around 1950 Ma. The zircon age data suggests that the host peridotites for these dykes represent Archean subcontinental lithospheric mantle. The scatter in the ages of Proterozoic zircons reflects incomplete recrystallization of the grains older than 1950 Ma. Altogether, the zircon chronology is indicative that the Jormua Ophiolite includes both Archean and Proterozoic units.

Furthermore, Archean origin for the Jormua peridotites is supported by the Re-Os study of *Tsuru et al (2000)*. They demonstrate that isotopic composition of chromite separated both from the serpentinites (i.e. altered residual peridotites) and local chromitite boulders are consistent with closed-system behaviour. Chromite from serpentinites yielded very depleted present-day $^{187}Os/^{188}Os$ with an average calculated $\gamma_{Os\ (1950\ Ma)}$ of -5.1 ± 0.8. Such a negative value requires that the peridotites were strongly depleted already at least 1 billion years before the time of the formation of the gabbros and lavas at c. 1.95 Ga. This suggests that true oceanic mantle (1.95 Ga asthenospheric diapir) is probably not exposed at Jormua, but that all peridotites represent streched slivers of the (Archean) subcontinental lithospheric mantle. This does not contradict with the presence of Proterozoic E-MORB basaltic rocks (dykes, lavas, gabbro pods) within the peridotites. In such a scenario listric faulting has exposed SCLM at the incipient oceanic basin which subsequently became intruded by basalts fed from the underlying asthenospheric diapir.

3. The model

The mantle peridotites had yielded melt already before they were intruded by the oldest suite of dykes at ~2800 Ma. These old clinopyroxene cumulate dykes are products of some Archean magmatic episode. Later – during the initial stages of continental break-up at ~2080 Ma – this same piece of mantle became extensively invaded by hydrous alkaline magmas that resulted in formation of high-pressure hornblendite-carbonatite dykes deep in the ophiolite stratigraphy and OIB-type dykes at more shallow levels. Alkaline magmatism was soon followed by lithospheric detachment faulting that exposed the subcrustal peridotites at the seafloor, where they became intruded by gabbros and sheeted dykes and covered by tholeiitic (EMORB) pillow and massive lavas.

This model implies that the Jormua Ophiolite represents the transition zone where continental lithosphere grades into oceanic regime. Elsewhere, Phanerozoic analogues include e.g. the Northern Apennines ophiolites, which were formed in conjunction with the opening of the Jurassic Tethys Ocean *(Rampone & Piccardo, 2000)*. In modern terrains such a lithological succession is seldom exposed for direct study, but is present e.g., in the Zabargad Island of the Red Sea where continental mantle became exhumed due to extreme crustal thinning and detachment faulting during the final stages of continental break-up (e.g., *Bonatti et al., 1981*). Also passive margins of modern major oceans expose strikingly similar lithological sequences as the Jormua ophiolite. One such location is the West Iberia Margin where the ocean – continent transition zone between the rifted and thinned continental crust and true oceanic crust has been studied in detail. There, seismic studies have identified a ~100 km wide zone of partially serpentinized peridotites exposed at the seafloor. Sampling of this zone has indicated that scarce basaltic rocks, locally pillowed, were deposited – as in Jormua – on a substratum of continental lithospheric peridotites which enclose strongly sheared and metamorphosed gabbro intrusions and alkaline pyroxenites. Not all these intrusions are believed to be comagmatic with the volcanic rocks but are interpreted to represent magmas that underplated the continental crust already before the final rifting (*Cornen et al., 1999*).

References

Bonatti, E., Hamlyn, P.R., Ottonello, G., 1981. The upper mantle beneath a young oceanic rift: peridotites from the island of Zabargad (Red Sea). Geology 9, 474–491.

Cornen, G., Girardeau, J., Monnier, C., 1999. Basalts, underplated gabbros and pyroxenites record the rifting process of the West Iberian margin. Mineral. Petrol. 67, 111–142.

Kontinen, A., 1987. An early Proterozoic ophiolite – the Jormua mafic-ultramafic complex, northern Finland. Prec. Res. 35, 313–341.

Peltonen, P., Kontinen, A., Huhma, H., 1998. Petrogenesis of the mantle sequence of the Jormua ophiolite (Finland): Melt migration in the upper mantle during Palaeoproterozoic continental break-up. J. Petrol. 39, 297–329.

Peltonen, P. & Kontinen, A. 2002. The Jormua Ophiolite: a mafic-ultramafic complex from an ancient ocean – continent transition zone. In: Kusky, T. (ed.), Precambrian Ophiolites and Related Rocks, Developments in Precambrian Geology, Elsevier, Amsterdam. In press.

Tsuru, A., Walker, R.J., Kontinen, A., Peltonen, P., Hanski, E., 2000. Re-Os isotopic systematics of the 1.95 Ga Jormua ophiolite complex, northeastern Finland. Chem. Geol. 164, 123–141.

Whitmarsh, R.B., Manatschai, G., Minshull, T.A., 2001. Evolution of magma-poor continental margins from rifting to seafloor spreading. Nature 413, 150–154.

Palaeomagnetic Configuration of Continents During the Proterozoic

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

[1]Department of Geophysics, P.O. Box 64, FIN-00014 University of Helsinki, Finland
[2]Geological Survey of Finland, P.O. Box 96, FIN-02151 Espoo, Finland
E-mail: lauri.pesonen@helsinki.fi

Keywords: Lithosphere, Baltica, Fennoscandia, palaeomagnetism, supercontinents, Hudsonia, Rodinia, Columbia

1. Introduction

Supercontinents are important in our understanding of the Earth. They play a vital role in concepts like mantle plumes, low latitude glaciations, carbon isotope excursions, fragmentations of dyke swarms, true polar wander and peaked age distributions of geological events and orogenic belts. Three supercontinent assemblies (Pangaea, Gondwana and Rodinia) have existed since the Proterozoic. Palaeomagnetism, coupled with isotope ages, is the only method, which provides direct knowledge of the positions of the continents. In spite of this, the role of palaeomagnetism in many previous reconstructions has been limited, because of poor age control of the poles, their large scatter, or because the poles from different continents are not coeval. In this paper we use high quality updated palaeomagnetic data, combined with geological information, to define the positions of the continents during the Proterozoic. For a detailed account of the data, results and references, see Pesonen et al. (2002).

2. Data and techniques

Palaeomagnetic data were compiled from fourteen cratons covering 2.45-1.0 Ga. Applying reliability grades as a filter, we selected the most reliable poles for analysis. The ages of individual poles are generally < 30 My from the chosen binning age (e.g., 2.45, 2.0 Ga etc) and the departure from this is an additional source of error in the reconstructions and arises from the continuous drift of the continents. The other errors arise from the uncertainties in the pole positions (Pesonen et al. 2002). The continents were plotted in their ancient latitudes and orientations as defined by the poles using the GMAP program. The palaeolongitude is undetermined due to axial symmetry of the dipole field. Geological features, such as orogenic belts, mafic dyke swarms, large igneous provinces and major lineaments were plotted on the continents to see if meaningful geological matching at each time bin can be seen. Occasionally, in seeking better matching, the continent is plotted in its antipodal hemisphere (with inverted orientation). This is allowed since we do not know the true polarity of the Precambrian poles. This technique led to thirteen configurations of continents during the Proterozoic. Four examples (2.45, 1.83, 1.25 and 1.05 Ga) are shown in Figures 1-3. Figure 4 shows the drift histories of four continents during 2.45-1.0 Ga. The rest of the configurations are seen in Pesonen et al. (2001; 2002).

3. Reconstruction at 2.45 Ga

Figure 1a shows the positions of Laurentia, Baltica, Kalahari and Australia at 2.45 Ga. Laurentia and Baltica are united as based on poles of the 2.45 Ga old dyke swarms. In this configuration, the Karelian and the Matachewan/Hearst dyke swarms become parallel supporting the existence of a presumably global mantle plume. The igneous events at 2.45 Ga

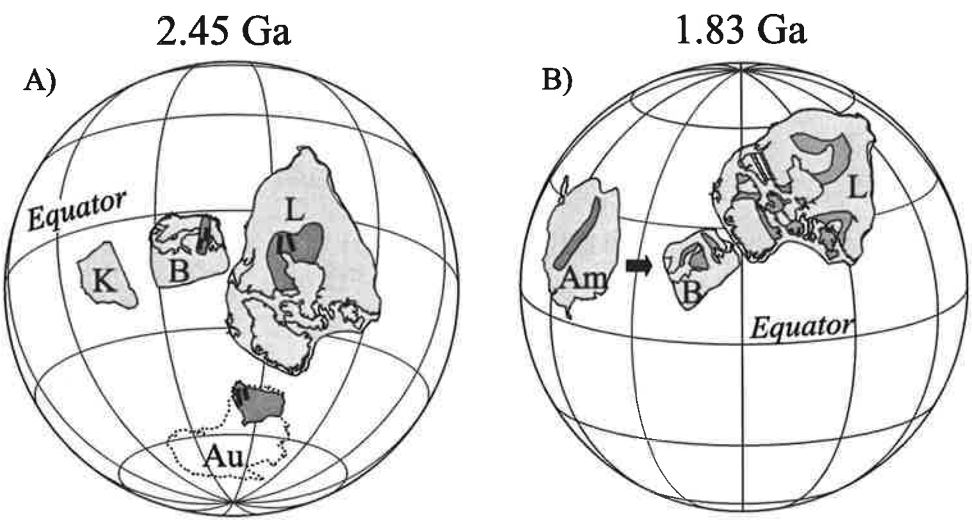


Figure 1. a) The reconstruction of continents at 2.45 Ga with data from Laurentia, Australia, Baltica and Kalahari. Note that Greenland has been rotated to North America using the Roest and Srivastava (1989) fit at 92 Ma with Euler pole at 66.6^{O}N, 240.5^{O}E and rotation angle of -12.2^{O}. The bold sticks denote the trends of the 2.45 Ga old dyke swarms. b) The reconstruction of continents at 1.83 Ga. Data available from Laurentia, Baltica and Amazonia in the supercontinent Hudsonia.

mark the onset of rifting of a large Archaean supercontinent called Kenorland, of which the Superior, Wyoming and Karelia cratons were members. The Widgiemooltha dyke swarm of Yilgarn has a similar trend as the Matachewan-Karelia dyke swarms (Fig. 1a). Its distance, however, to the Superior and Karelia cratons is more than 30 degrees and its relation to these remains to be studied. We note that Kalahari may have been attached with Baltica but since there are no geological evidence to support this, other positions are possible.

During 2.4-2.3 Ga the cratons experienced glaciations (Marmo and Ojakangas, 1984, Bekker et al., 2001). Present palaeomagnetic data support the growing evidence of early Palaeoproterozoic glaciations at nearly equatorial latitudes. Various models have been presented to explain the low latitude glaciations, e.g. the "snowball Earth", the Earth's obliquity, the non-dipole field, or the remagnetisation explanation (see Pesonen et al., 2002). The low latitude positions of these cratons is supported by sedimentological studies which show, in addition to the glaciogenic deposits, zones of palaeoweatherings indicating tropical climate conditions during 2.4-2.3 Ga. Probably very rapid climatic changes took place during a relatively short time interval from 2.4-to 2.3 Ga (Bekker et al., 2001).

The Baltica-Laurentia landmass broke up in rifting events at ca. 2.2-2.1 Ga as evidenced by the mafic dyke swarms with passive margin sediments in Baltica and in Laurentia. This fragmentation (including probably Australia) is probably linked to the global peak in the carbon-13 curve at 2.15 Ga ago (Bekker et al., 2001).

4. Reconstructions at 2.0-1.27 Ga

Reconstruction at 2.0 Ga ago reveal that continents occupy moderate to shallow latitudes, but probably as independent blocks because there are no compelling geologic reasons for putting them together (Pesonen et al., 2002). From 2.45 Ga to 2.00 Ga Laurentia drifted from southern to northern latitudes and rotated ~ 100 degrees anticlockwise. Unfortunately, there are no reliable palaeomagnetic data from Baltica between 2.45-2.00 Ga to seek if it has a similar drift pattern. However, the position of Baltica at 1.93 Ga is different to that at 2.45 Ga ago.

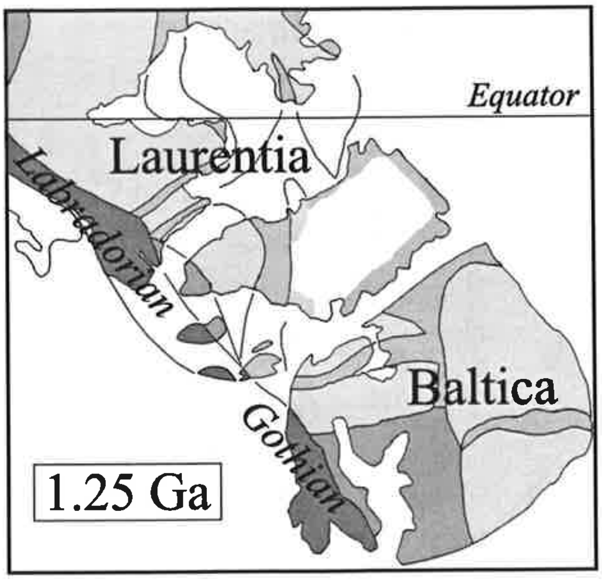


Figure 2. The reconstruction of Baltica-Laurentia continents at 1.25 Ga showing the continuation of the Gothian-Labradorian belts via Scottish isles from Baltica to Laurentia.

It is suggested that the amalgamation of Baltica with Laurentia began at ca. 1.93 Ga ago when Laurentia collided with Baltica in the present north resulting to the Nagssugtoqidian/Törngat orogens in Laurentia and the Kola/Lapland orogen in Baltica. The exact plate tectonic processes causing these belts are not known but the complexity of the continent-continent collisions is manifested by the network of the 1.93-1.88 Ga old orogenic belts within the Archeaen terranes in both Laurentia and Baltica.

Palaeomagnetic data for the 1.83 Ga reconstruction come from Laurentia, Baltica and Amazonia and is shown in Figure 1b. From 1.88 Ga to 1.83 Ga Laurentia has remained nearly in the same latitude but rotated some 80 degrees clockwise, while Baltica has drifted only slightly northwards without significant rotation before docking with Laurentia. Palaeomagnetic data thus indicate that relative rotations and latitudinal shifts will take place during the docking process.

The geotectonic scenario for the evolution of the middle Proterozoic orogenic belts in Laurentia, Baltica and Amazonia might be as follows. At the same time when the Nagssugtoqidian/Törngat orogens in Laurentia and the Kola/Lapland orogen in Baltica were generated in the north, the juxtaposed Baltica-Laurentia unity is affected by a third continent, Amazonia in the south (e.g., Karlström et al. 1999). Palaeomagnetic data show that Amazonia was close to Baltica at 1.83 Ga and remains in close connection with it at least up to 1.6 Ga. We propose that not only the coeval Svecofennian orogeny of Baltica and the Ventuari-Tapajos orogeny in Amazonia but also the successively younger marginal orogenies in Baltica (the Transcandinavian Igneus Belt and the Gothian belt) and in Amazonia (the Rio-Negro Juruena belt) are caused by long lasting accretional processes in the margins of Baltica and Amazonia (see Tassinari et al., 2000; Pesonen et al., 2002).

5. Supercontinent Hudsonia

The paleomagnetic data showed that Laurentia, Baltica and Amazonia formed a huge landmass at ca. 1.83 Ga. The geological data from these continents and from other continents (e.g., India, Australia, Congo/São Francisco, West Africa; see Rogers and Santosh, 2002) support the idea that this is a time of major crustal formation during the Proterozoic history of the Earth, when large amounts of new juvenile crust is added to continents at their margins. Palaeomagnetic data are too sparse to show that all continents were involved in a

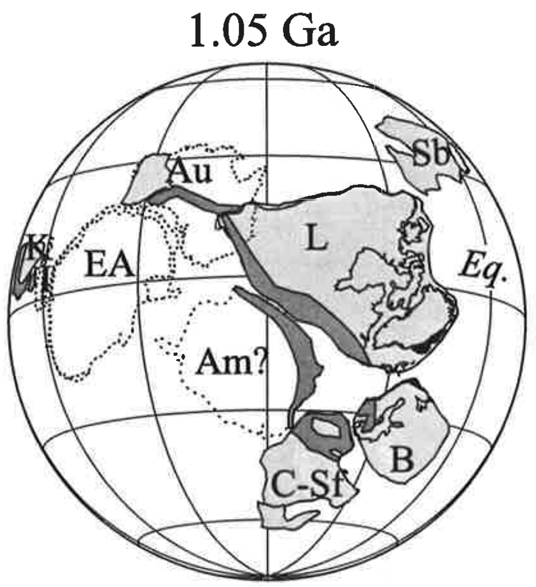


Figure 3. A global reconstruction of continents at 1.05 Ga ago. Data available from Laurentia, Baltica, Congo/SãoFrancisco, Kalahari and Siberia. The dark belts indicate Genvillian collisional events.

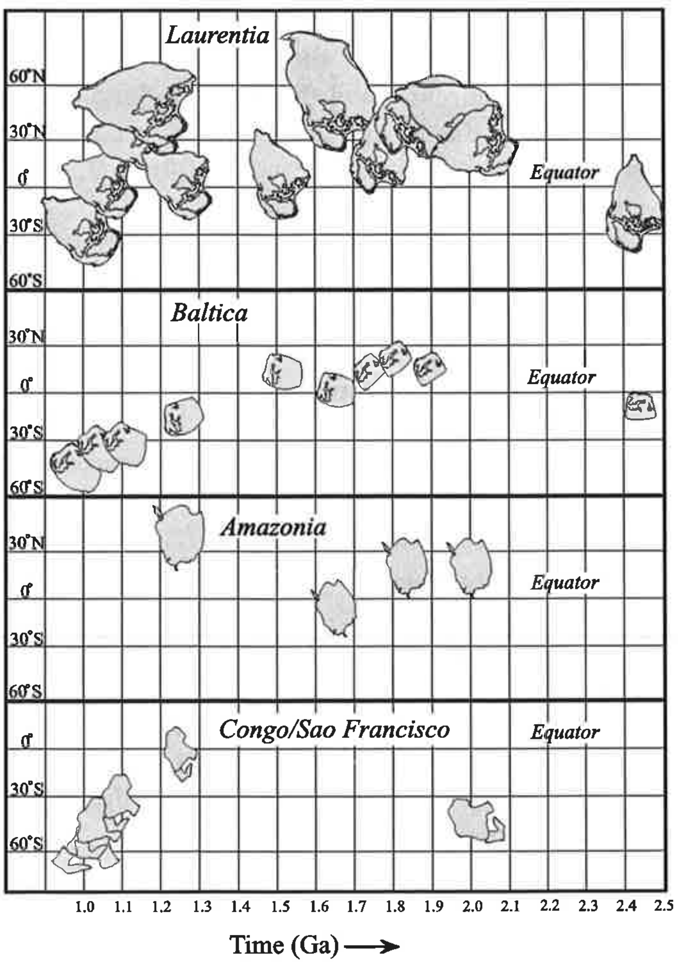


Figure 4. An animation of the drifts of Laurentia, Baltica, Amazonia and Congo/São Francisco during 2.45-1.0 Ga. Horizontal axis is age (Ga). Vertical axis is paleolatitude. The continents have been moved artificially sideways to show up.

supercontinent assembly but at least Laurentia, Amazonia and Baltica were members. We name this assembly as Hudsonia (or "Hudsonland", see Pesonen et al., 2002). It is also known as Columbia (Rogers and Santosh, 2002). The existence of a supercontinent is manifested by coeval global orogenic events, associated with juvenile island-arc magmatism. The global tectonomagmatic events taking place in or around this supercontinent are seen for example in peaks of the ages (ca. 1.88-1.80 Ga) of the juvenile continental crust, in ages of the banded iron formations and black shales, and in many other geological features (see Bekker et al. 2001).

The 1.9-1.8 Ga Hudsonian (Penokean) belt, and the younger Labradorian belt in Laurentia are more difficult to explain. The oblique network of the Hudsonian belts in Laurentia (e.g., Trans-Hudsonian Orogen, Ketilidian and Makkovik belts) and the coeval belt in Baltica (Svecofennian) suggest a close connection of Baltica and Laurentia during the formation of these belts. However, it is also possible that at the same time as Amazonia is colliding with Baltica, a sufficiently large unknown continent is colliding with Laurentia. We do not have 1.9-1.8 Ga palaeomagnetic data from other continents than Amazonia and Baltica to test this scenario.

6. Reconstruction from 1.6 to 1. 27 Ga

It is proposed that at about 1.65 Ga the Baltica and Amazonia collided resulting in the Gothian orogenic belt in SW Baltica and the Rio Negro-Juruena belt in Amazonia. It is also possible that the Labradorian orogenic belt was caused by continent-continent collision but we do not have reliable palaeomagnetic data from any of the large continents to collide with Laurentia, despite that Amazonia may have collided with Laurentia more or less simultaneously with Baltica.

The close juxtaposition of Amazonia with Baltica, and perhaps also with south Australia is supported by the coeval 1.57-1.54 Ga magmatism of rapakivi-anorthosite intrusions and mafic dyke swarms in these continents. This rapakivi-anorthosite magmatic activity is consistent with a mantle plume underlying the Baltica-Amazonia-Australia (?) landmass. The long-lasting juxtaposition of Baltica and Amazonia ends in a rifting episode between Baltica and Amazonia which will probably take place at sometimes during 1.57-1.4 Ga when Amazonia separates from Baltica. This coincides with the youngest pulse of rapakivi-anorthosite magmatism in Amazonia and Baltica which may be genetically linked to the rifting (and death) of Hudsonia.

Figure 2 shows the assembly of Laurentia and Baltica at 1.25 Ga when the continents are located at shallow to intermediate latitudes. This configuration of Laurentia-Baltica is somewhat similar to previous models (e.g., Karlström et al., 1999) at ca. 1.3 Ga. The relative position of Baltica-Laurentia at 1.25 Ga is roughly the same as at 1.83 Ga ago although this large landmass rotated as a unity by 80 degrees anticlockwise and drifted southwards since 1.83 Ga. This Baltica-Laurentia configuration is independent of the polarity ambiguity since the data of 1.25 Ga old mafic dykes from both continents are of single polarity. If this unity of Baltica-Laurentia appears to be valid it persisted some 600 million years. Reliable palaeomagnetic poles of 1.8-1.3 Ga are needed from both continents to prove this (Buchan et al. 2000).

The 1.25 Ga assembly of Baltica-Laurentia is supported by numerous geological observations. Buchan et al. (2000) have shown that the 1.65-1.4 Ga old Gothian-Labradorian belts from southern Scandinavia to Labrador will be aligned in this configuration (Fig. 2). The rift-related post-Jotnian dykes and sills in central Baltica are probably genetically associated with the huge Mackenzie rifting and dyking event in Canada. The fan-shaped Mackenzie dyke swarm could be a signature of the mantle plume occurring at 1.25 Ga ago in NW Canada. Elming and Mattson (2001) have shown that the Central Scandinavian dolerite activity is more widespread in Baltica than previously thought and is now extending further north in

Sweden, towards Greenland. The dyke trends in Baltica (the Vaasa and Satakunta dolerites) and Laurentia (the Mackenzie swarm) become contiguous in this reconstruction, although the former is distinctly smaller swarm than the huge Mackenzie swarm. The 1.25 Ga dyke activity is a global one and is well documented in several continents (see Rogers and Santosh, 2002).

7. The birth of Rodinia

Figure 3 shows the configuration of the continents at 1.05 Ga. This time marks the assembly of Rodinia with possible minor adjustments still taking place during 1.05-1.0 Ga. Data are available from Laurentia, Baltica, Kalahari, Congo/São Francisco, Australia and Siberia. The so-called Congo Sea (see Mertanen and Pesonen, 2000, Pesonen et al., 2002) between the landmasses of Laurentia/Amazonia and Congo/São Francisco-Baltica began to close sometimes after 1.10 Ga. The two huge landmasses aggregated together at ca. 1.05 Ga producing the "late " Grenvillan collisions in these continents. The scenario predicts that late Grenvillian events might have taken place in NW Baltica, in Svalbard and in E Greenland due to the collision of Laurentia with Baltica. Recent U-Pb age data from E. Geenland and Svalbard indeed show "late" Grenvillian ages consistent with the model (Henriksson, 2002 and references therein). The known Grenville collisional belt in SW Laurentia may continue from Laurentia across central Australia (Musgrave and Albany-Fraser belts) and then around East Gondwana to Kalahari (Namaqua-Natal belt) and to India, where Grenville age belts are also present.

8. Summary

As a summary, figure 4 shows the continental movement of four landmasses, Laurentia, Baltica, Amazonia and Congo/São Francisco at 2.45-1.0 Ga. Similarities in plate movement between Laurentia, Baltica and Amazonia especially at 1.9-1.6 Ga envisage the close connection of these continents during this time interval. Likewise, a similar drift pattern of Laurentia, Baltica and Congo/São Francisco after ca. 1.2 Ga suggest a common movement of these cratons before their amalgamation to supercontinent Rodinia.

Acknowledgements

We want to thank the participants of the Precambrian Group of the Fourth Nordic Palaeomagnetic Workshop; Sten-Åke Elming, Sergei Pisarevski, Manoel D'Agrella-Filho, Phil Schmidt, Niels Abrahamsen, Göran Bylund and Håkan Mattsson for their contribution in compiling the reconstructions. Fabio Donadini is acknowledged for assisting in the final stages of the compilation.

References

Bekker, A., Kaufman, A.J., Karhu, J.A., Beukes, N.J., Swart, Q.D., Coetzee, L.L., Eriksson, K.A., 2001. Chemostratigraphy of the Paleoproterozoic Duitschland Formation, South Africa: Implications for coupled climate change and carbon cycling. Am. J. Sci. 301, 261-285.

Buchan, K., Mertanen, S., Elming, S.-Å ., Park, R.G., Pesonen, L.J., Bylund, G., and Abrahamsen, N., 2000. The drift of Laurentia and Baltica in the Proterozoic: a comparison based on key paleomagnetic poles. Tectonophysics 319, 167-198

Elming, S.A. & Mattson, H., 2001. Post-Jotnian basic intrusions in the Fennoscandian shield, and the break up of Baltica and from Laurentia: a paleomagnetic and AMS study, Precam. Res., 108, 215- 236.

Karlström, K.E., Åhäll,K-I., Harlan, S.S., Williams, M.L., McLelland, J., Geissman, J.W., 2001. Long-lived (1.8-1-0 Ga) convergent orogen in southern Laurentia, its extension to Australia and Baltica, and implications for refining Rodinia. Prec. Res., 111, 5-30.

Marmo, J. & Ojakangas, R., 1984. Lower Proterozoic glaciogenic deposits, eastern Finland. Geol. Soc. of America Bull., v. 95, 1055-1062.

Mertanen, S. and Pesonen, L.J., 2000. Fennoscandia as a part of a Proterozoic supercontinent. In: Pesonen, L.J., Korja A. and Hjelt, S.E., 2000 (editors). Litosphere 2000 - A symposium on the Structure, Composition and Evolution of the Lithosphere in Finland. Program and Extended Abstracts, Espoo, Finland, October 4-5, 2000. Institute of Seismology, University of Helsinki, Report S-41, 137-141.

Pesonen, L.J., Mertanen, S., Elming, S.-Å., 2001. Reconstructions of Continents During the Proterozoic With a

Special Focus on Fennoscandia. *In*: M. Poutanen (editor), Abstracts of the SRCM´ 01, an IAG International Symposium on Recent Crustal Movements, Helsinki, Finland, 79-81.

Pesonen, L.J., Elming, S.-Å., Pisarevsky, S., Mertanen, S., D´Agrella-Filho, M., Meert, J., Schmidt, P., Bylund, G., 2002. A pre-Rodinia Supercontinent - A paleomagnetic survey. In: C. Powell Memorial Volume, Austr. Geol. Soc. (submitted).

Rogers, J.J.W. and Santosh, M., 2002. Configuration of Columbia, a Mesoproterozoic supercontinent, Gondwana Res., 5, 5-22.

Tassinari, C.C.G., Bettencourt, J.S., Geraldes, M.C., Macambira, M.J.B. & Lafon, J.M., 2000. The Amazonian craton. In: Cordiani, U.G.,Milani, E.J., Thomaz Filho, A. & Campos, D.A.(eds), Tectonic evolution of South America, Rio de Janeiro, 41-95.

Magnetic Overprints in the Silurian Dolomites, Central Estonia

Jüri Plado[1], Lauri J. Pesonen[2], Kalle Kirsimäe[1], Väino Puura[1], Tõnu Pani[1], and Tiiu Elbra[1]

[1]Institute of Geology, University of Tartu, Vanemuise 46, 51014 Tartu, Estonia.
[2]Division of Geophysics, University of Helsinki, P.O. Box 64, FIN-00014 Helsinki, Finland.
E-mail: jplado@ut.ee, lauri.pesonen@helsinki.fi

Mineralogical, petrophysical, and palaeomagnetic studies of Silurian dolomites from the Rõstla quarry, central Estonia were carried out. Hematite and goethite were found to be the main magnetic carriers. The paragenetic relationships suggest that hematite crystals have formed during dolomitization of the primary carbonate sediments whereas goethite is the youngest phase, which precipitated in the walls of microscopic fractures. Two magnetic components (A: D = 54.5°, I = 25.6°, N = 6, α_{95} = 15.4, and B: D = 25.0°, I = 62.3°, N = 5, α_{95} = 11.0) were identified. When compared with the Apparent Polar Wander Paths of Baltica and Europe, these pinpoint to Late Devonian to Carboniferous (component A) to Cretaceous to present ages (component B), respectively.

Keywords: dolomites, Estonia, Fennoscandia, goethite, hematite, overprints, palaeomagnetism, Silurian

1. Introduction

Recent palaeomagnetic data show that the supposedly stable Fennoscandian Shield has suffered from multiple, Late Precambrian to recent, overprintings. This emphasizes need for the reanalysis of previous paleomagnetic data with sophisticated multicomponent analysing techniques to verify the existence and the extent of these remagnetization events in order to find an explanation for them (e.g., fluid motions, tectonothermal events, etc.). The Estonian sedimentary sequence, ranging in age from Vendian to Devonian is in unique position since it lies close to "exposed" Fennoscandian shield and may have recorded regional overprinting events, which have taken place in Fennoscandia or nearby. Moreover, the strata are undeformed and unaltered so that both primary (sedimentary) and secondary events can be accurately identified, measured, and dated. Here we describe new palaeomagnetic results of Estonian Silurian data, which show evidences of post-Silurian overprinting. The new results have several implications to the drift history of Baltica and can be used to test the various supercontinent models during the Phanerozoic.

2. The Estonian dolomites and samplings

Estonia is located on the southern slope of the Fennoscandian Shield. Here, the sedimentary rocks of Vendian to Devonian cover the Svecofennian Crustal Province. The sedimentary sequence is sporadically dolomitized (see, e.g., *Raukas and Teedumäe, 1997*). Several models, (i) syngenetic (*Kaljo, 1970; Kaljo and Orviku, 1960*), (ii) early diagenetic mixing zone (*Kiipli, 1983; Teedumäe et al., 2001*), and (iii) hydrothermal (*Vaher et al., 1962*), of origin of dolomites are proposed, as based in their different structure, texture, mineralogical, and chemical composition. However, the age and reasons of dolomitization are still purely studied. For example, *Vingisaar and Taalmann* (*1974*) proposed late diagenetic age of Late Silurian - Early Devonian for the whole variety of Estonian dolomites, whereas *Jürgenson* (*1970*) suggested that dolomitization lasted for a long time, but that it mainly took place after the Devonian.

We used rock magnetic and mineralogical methods to specify the dolomitization age of carbonate rocks from Rõstla quarry, central Estonia (Fig. 1). The dolomites of the Mõhküla formation, which belongs to the youngest dolomite beds of the Raikküla Stage, and dating to

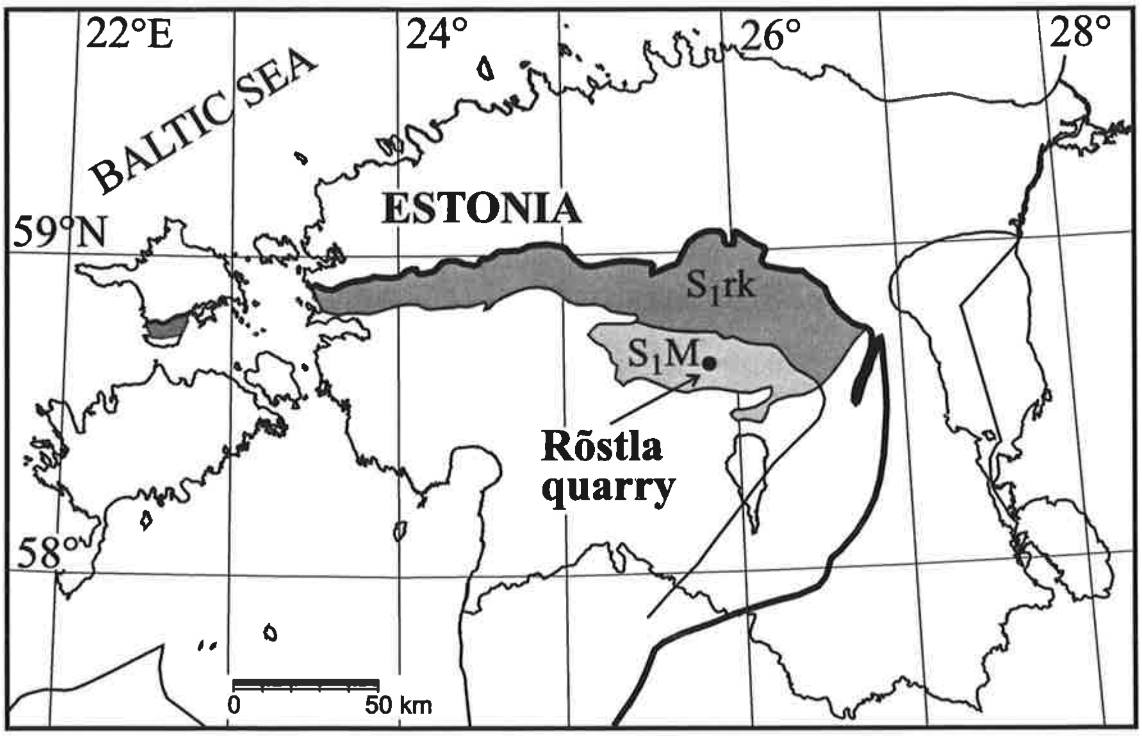


Figure 1. Outcrop area of the Raikküla Stage (S_1rk, dark grey) and Mõhküla Beds (S_1M, light grey) in central and western Estonia. Thick black line contours distribution of the carbonates of the Raikküla Stage, thin line the same of the Mõhküla Beds. The map is redrawn after *Nestor (1997)*. Location of the Rõstla quarry is shown.

middle Llandovery (443-428 Ma), early Silurian were investigated. A description of the Rõstla quarry section is given by *Teedumäe et al. (2001)*.

Nine oriented hand-samples, which represent three varieties of dolomites, were collected from the Rõstla quarry. *The first variety* represents fine-crystalline, nodular-like light-yellow dolomite with reddish tint. The reddish nodules-like masses of the samples are coarser but more porous, and have sharp violet-figured rim at the contact with the more fine-crystalline part. *The second variety* is micro- to fine crystalline, mainly light-beige, wavy- or nodular-bedded argillaceous dolomite. In places, this variety is ferruginous owing black, red or violet colour mostly within the lenses of sceletal interbeds of silicified grainstone, or, in nodules. A vertical massive dolomite body that permeates the southern wall of the quarry and shows no sedimentary features represents *the third variety*. It consists of medium to coarse-grained, (locally greenish)-grey dolomites that contain partly of relatively voluminous (up to several hundreds cm^3) caverns. The walls of caverns show large (up to few mm) rhombs of dolomite crystals with rare dark or rusty impregnation of rare iron minerals.

3. Results

Based on XRD data the samples are dolomites to argillaceous dolomites. The dolomite content in *varieties I and III* is from 92 to 95%, whereas in the *variety II* the dolomite content is only 71-76%, the rest being clay minerals, quartz and K-feldspar. Only magnetic mineral recognized in whole rock samples in a detectable amount (*i.e.* > 0.5 %) is hematite, which compromises in the argillaceous dolomites 1-2 % of the rock composition. In all other samples, the content of hematite, as well as other magnetic mineral phases, is under the limit of detection of XRD. However, the mineral fraction studies reveal apart from abundant hematite, the presence of goethite. Binocular microscopic observations show that goethite

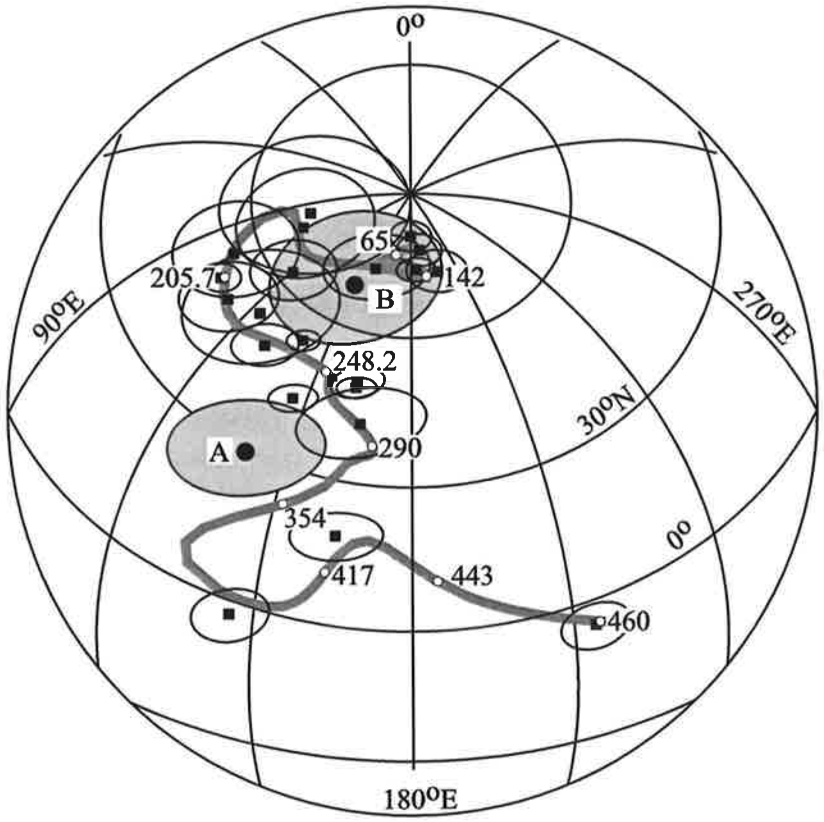


Figure 2. Combination of (i) selected (*Van der Voo (1990)* Q-factor ≥ 6) individual Ordovician to Paleocene poles (black squares) with α_{95} confidence circles (white ovals) from Europe and/or Baltica, (ii) APW path (a dark grey line with white dots of ages) constructed with the spherical spline method (smoothing parameter = 200; poles with 5 ≤ Q ≤ 7 only) weighted according to their performance in the Q-factor) by the GMAP programme of Torsvik and Smethurst (http://www.geophysics.ngu.no), and (iii) two poles ('A' and 'B') of the present study (black dots) with α_{95} confidence circles (grey ovals).

occurs as secondary reddish to dark-brown spherolitic aggregates at the hematite impregnated fracture surfaces.

Two distinct remanent magnetization components (hereafter called A and B) were identified by directional groupings and alternating field (AF) coercivity spectra. Both components are present in *varieties I and II* with NRM intensities generally > 0.2 mA m^{-1}. The *variety III* has extremely weak NRM and, therefore, no detectable remanent magnetization components were identified. We did not observe any significant differences in palaeomagnetic behaviour between *varieties I and II* being from different stratigraphic levels. The remanence components A and B differ in their stability against the AF demagnetization. The component A is soft and removed at fields ≤80 mT (as maximum). Component B is more resistant to AF, and, in most of specimens it has not yet been totally removed at 160 mT, the maximum field applied. By the knowledge of magnetic mineralogy, and coercivity spectra of hematite and goethite (see, *e.g.*, *Dekkers and Rochette, 1992*; *McElhinny and McFadden, 2000*), we propose that component A is carried by hematite and component B by goethite.

Both polarities are present in B with antipodal symmetry. The combined site-mean direction of B (D = 25.0°; I = 62.3°; N = 5; α_{95} = 11.0) is only slightly shifted from the present Earth' magnetic field (PEF) at the Rõstla site (D ≈ 6°; I ≈ 72°). We suggest B represents a magnetic overprint that is, most likely, of Cenozoic origin. It is carried by goethite that is generally thought (*Dekkers and Rochette, 1992*) to have a chemical origin due to weathering processes. The low-coercivity component A, of single polarity, has a mean direction (D = 54.5°; I = 25.6°; N = 6; α_{95} = 15.4) that is clearly different from PEF and B.

Component A has lower precision than B, caused mainly by considerable variation of declinations, whereas the inclination being nearly constant.

To find the possible ages for A and B, we plotted the palaeomagnetic poles of A and B on the Apparent Polar Wander Path (APWP) for Baltica (Fig. 2). This APWP includes poles from stable Europe of late- and post-Carboniferous times (≤ 300 Ma; *Torsvik et al., 2001*). The APWP was constructed with the spherical spline method with a smoothing parameter of 200 weighted according to the reliability criteria (Q-factor; *Van der Voo, 1990*) of the available poles. We used the GMAP programme of Torsvik and Smethurst (*http://www.geophysics.ngu.no*). The poles with Q-factor of 5, 6, and 7 selected from the database in *Torsvik et al.* (*1996, 2001*) were used.

Comparison of poles A and B of this study with the APWP suggests Late-Palaeozoic age for A, and Cretaceous to present age for B (Fig. 2). However, both poles, especially A, plot slightly aside of the Baltic-European APWP. There exist several possible reasons for this. *First*, the number of samples is too small, therefore, the pole could be offset due to unaveraged secular variation. *Second*, because the demagnetization spectra of two components slightly overlap, the two components may have been separated imperfectly. Therefore, the components, especially the low-coercivity component A, may be contaminated by another component. The individual sample-mean A poles align subparallel to the Late-Devonian-Carboniferous section of the APWP-curve suggesting that the A poles may reflect considerable time of acquisition. Neither A nor B poles pinpoint towards the Silurian, therefore, we deduce that a primary sedimentary magnetic component of Silurian age is lost and overprinted by younger events. Both A and B are most likely of chemical origin.

These data are consistent with the geological data on the post-Devonian age of the dolomitization and sulphide mineralization in the Võhma-Navesti area. In a larger regional context, the results support an idea of the Carboniferous-Permian tectono-thermal event (*Puura et al. 1996*).

4. Conclusions

Two magnetic minerals, hematite and goethite, dominate in the dolomites of the Mõhküla Beds in the Rõstla quarry. Hematite was formed simultaneously with the dolomitization event at the end of the Paleozoic, presumably at Late Devonian to Carboniferous times. Goethite is more recent, most likely Cenozoic. Hematite and goethite do not carry the primary Silurian or any early diagenetic directions. However, these results do not exclude any possible syn-and/or diagenetic dolomitizations, but suggest that the latest event took place from Late Devonian to Carboniferous overprinting the earlier events.

Acknowledgements

We are thankful to Dr. Satu Mertanen and Matti Leino for their permission to use the equipments at the Laboratory for Palaeomagnetism, Geological Survey of Finland. The Estonian Science Foundation (grant # 4417) supported the study.

References

Dekkers, M.J. and Rochette, P., 1992. Magnetic properties of chemical remanent magnetization in synthetic and natural goethite: Prospects for a natural remanent magnetization / thermoremanent magnetization ratio paleomagnetic stability test? *Journal of Geophysical Research*, **97**, 17,291-17,307.

Jürgenson, E., 1970. Secondary changes of the Silurian sequence (in Russian). In: Kaljo, D. (ed.) *Silurian of Estonia*. Valgus, Tallinn, pp 96-101.

Kaljo, D., 1970. Adavere Stage (in Russian). In: Kaljo, D. (ed) *Silurian of Estonia*. Valgus, Tallinn, pp. 232-243.

Kaljo, D.L. and Orviku K.K., 1960. History of the geological evolution and palaeogeography of the Estonian SSR (in Russian). In: Orviku K.K. (ed.) *Geology of the SSSR, vol. XXVIII, Estonian SSR*. Gosgeoltehizdat, Moscow, pp. 231-250.

Kiipli, T., 1983. On the genesis of Ordovician and Silurian dolomites at the contact with Devonian deposits (in Russian). *Proceedings of the Estonian Academy of Sciences*, **32**, 110-117.

McElhinny, M.W. and McFadden P.L., 2000. Paleomagnetism: Continents and Oceans. In: *International Geophysics Series*, vol. 73. Academic Press, San Diego, 386 pp.

Mertanen, S. and Pesonen, L.J., 1994. Preliminary results of palaeomagnetic and rock magnetic study of the Proterozoic Tsuomasvarri intrusions, northern Fennoscandia. *Precambrian Research*, **69**, 25-50.

Nestor, H., 1997. Silurian. In: Raukas, A. and Teedumäe, A. (eds.) *Geology and Mineral Resources of Estonia.* Estonian Academy Publishers, Tallinn, pp. 89-106.

Puura, V., Amantov, A., Tikhomirov, S. and Laitakari, I., 1996. Latest events affecting the Precambrian basement, Gulf of Finland and surrounding areas. In: Koistinen, T. (ed) Explanation to the map of Precambrian basement of the Gulf of Finland and surrounding area 1 : 1 mill. *Geological Survey of Finland, Special Paper,* **21**, 115-125.

Raukas, A. and Teedumäe, A. (eds.), 1997. *Geology abd Mineral Resources of Estonia.* Estonian Academy Publishers, Tallinn, 436 pp.

Teedumäe, A., Kallaste, T. and Kiipli, T., 2001. Aspects of the dolomitization of the Mõhküla beds (Silurian, Estonia). *Proceedings of the Estonian Academy of Sciences* **50**, 190-205.

Torsvik, T.H., Smethurst, M.A., Meert, J.G., Van der Voo, R., McKerrow, W.S., Brasier, M.D., Sturt, B.A. and Walderhaug, H.J., 1996. Continental break-up and collision in the Neoproterozoic and Palaeozoic - A tale of Baltica and Laurentia. *Earth-Science Reviews*, 40, 229-258.

Torsvik, T.H., Van der Voo, R., Meert, J.G., Mosar, J. and Walderhaug, H.J., 2001. Reconstructions of the continents around the North Atlantic at about the 60th parallel. *Earth and Planetary Science Letters*, **187**, 55-69.

Vaher, R.M., Puura, V.A. and Erisalu, E.K., 1962. Tectonism of NE Estonia (in Russian). In: *Geology of Paleozoic.* Valgus, Tallinn, 351 pp.

Van der Voo, R., 1990. The reliability of paleomagnetic data. *Tectonophysics* **184**, 1-9.

Vingisaar P.A. and Taalmann V.A., 1974. An overview on the dolomitization of lower-Paleozoic carbonate rocks in Estonia (in Russian). *Proceedings of the Academy of Sciences of theEstonian SSR. Ghemistry. Geology* **3**, 237-243.

Petrophysical Characteristics of Igneous Rocks in Finland Compared to Igneous Rocks in Subareas of the Seismic Reflection Profile FIRE-1

Tapio Ruotoistenmäki

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

The poster considers the susceptibility – density relations of the igneous rocks of Finland. The total number of samples is ca 52000. The data are classified using peak values in scatter diagrams resulting in characteristic susceptibility-density pairs of the samples. The peak values can further be used in gravimetric and magnetic modeling of the study area. The advantage of the method is to tie together the corresponding susceptibility-density pairs, which is not achieved if the distributions are studied separately. The susceptibility-density diagram of Finnish igneous rocks is given in Fig. 1. In the figure is also shown 13 peaks representing values around which are the main data clusters. In Table 1 are shown petrophysical characteristics of the main rock types connected with the peaks.

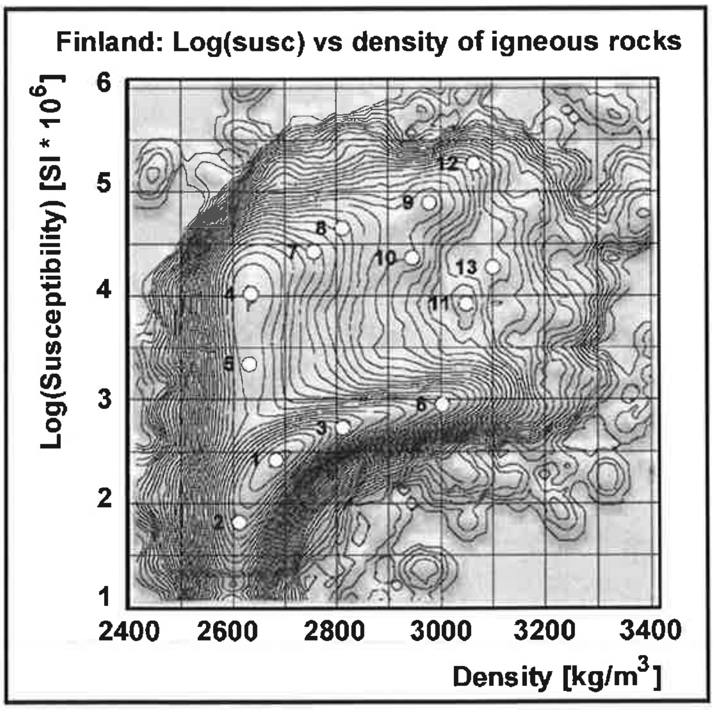


Figure 1. Susceptibility-density diagram of igneous rocks of Finland.

From figure 1 and table 1 it can be seen that some rocks are grouped in several separate susceptibility-density classes (peaks). This refers that rock types can only poorly be classified according to their petrophysical properties (even density).

When making regional scale potential field modelling it is possible to use the susceptibility - density pairs of the peak values as model parameters. After obtaining a satisfactory model combination the probable rock types connected with the parameters can be evaluated from the main rock types connected with the peaks.

For comparison, the distributions of data of the igneous rocks in the area covered by the deep seismic reflection profile FIRE-1 are studied. The area is divided into six sub-areas: 1. Archaean Eastern Finland Complex, 2. Early Proterozoic Kainuu Belt, 3. Archaean Iisalmi Complex, 4. NW part of the Proterozoic Ladoga-Bothnian Bay Zone, 5. SW part of the Ladoga-Bothnian Bay Zone, 6. Proterozoic Central Finland Granitoid Complex. The Ladoga-Bothnian bay zone is thus divided in two separate parts based mainly on their different geophysical characteristics.

The distributions of the petrophysical parameters of the sub-areas show following characteristics and variations:

- The density distributions of the NW and SW parts of the Ladoga-Bothnian Bay Zones are roughly identical (both are relatively dense compared to Finnish averages). However, the susceptibilities of the NW part are higher.
- The densities and susceptibilities of rocks in Eastern Finland Archaean complex are relatively low.
- The densities of the igneous rocks within Kainuu Belt are also low, but they are relatively magnetic.
- The density and susceptibility distributions of the igneous rocks in the Central Finland Granitoid Complex are close to the average of all Finnish igneous rocks, felsic rocks being slightly denser than Finnish average.
- The rocks in Iisalmi complex are denser and more magnetic than the Finnish average.

Table 1. Characteristics of the main rock types connected with peaks in Fig. 1. The iron content of paramagnetic rocks and magnetite content of ferromagnetic rocks has been calculated using formulas given by Puranen (1989).

Class (peak)	Peak density [kg/m**3]	Peak susceptibility [SI*10**6]	Iron content [%]	Magnetite content [%]	Samples in class (total)	Main rock types	Samples in class	prc of all samples in class
1	2683	264	3.44		15698	GRANODIORITE GRANITE PORPHYRITIC GRANODIORITE	2878 1729 1700	18.33 11.01 10.83
2	2612	66	0.88		12024	GRANITE MICROCLINE GRANITE	3224 2772	26.81 23.05
3	2811	536	6.67		5972	GRANODIORITE GABBROID QUARTZ DIORITE	1103 765 742	18.47 12.81 12.42
4	2635	10514		0.60	5087	GRANITE MICROCLINE GRANITE OLIGOCLASE GRANITE GRANODIORITE	1368 756 633 572	26.89 14.86 12.44 11.24
5	2632	2211		0.13	5378	GRANITE MICROCLINE GRANITE GRANODIORITE	1131 830 599	21.03 15.43 11.14
6	3003	901	10.50		2050	GABBROID	942	45.95
7	2757	26626		1.45	1868	GRANODIORITE QUARTZ DIORITE	282 208	15.1 11.13
8	2811	44773		2.39	512	GRANODIORITE GABBROID DIORITE	73 65 61	14.26 12.7 11.91
9	2978	78929		3.98	332	GABBROID	167	50.3
10	2946	23193		1.18	513	GABBROID	229	44.64
11	3049	8434		0.41	249	GABBROID PERIDOTITE	97 26	38.96 10.44
12	3064	187678		9.19	105	GABBROID	61	58.1
13	3101	18830		0.91	76	GABBROID PERIDOTITE HORNBLENDITE	23 12 11	30.26 15.79 14.47

References

FIRE Working Group, 2002. Reflection seismic data on transect FIRE-1 in eastern and central Finland.

Puranen, Risto 1989. Susceptibilities, iron and magnetite content of Precambrian rocks in Finland. Geological Survey of Finland. Report of Investigation 90. 45 p. + 6 app. pages

Earthquakes and Seismotectonics in Finland

Jouni Saari[1] and Marja Uski[2]

[1]Fortum Engineering Ltd, P.O. Box 10, FIN-00048 Fortum, Finland
[2]Institute of Seismology, P.O. Box 26, FIN-00014 University of Helsinki, Finland
E-mail: jouni.saari@fortum.com, marja.uski@seismo.helsinki.fi

The dominating stress field in Finland and in Fennoscandia as a whole follows the pattern of NW-SE oriented compression attributed to sea-floor spreading along the mid-Atlantic Ridge. Local variations observed in the stress pattern in Finland are most probably caused by local geology and postglacial rebound, which is delineated by the area of the present day lithospheric uplift.

Keywords: Earthquakes, seismotectonics, stress field, Finland

1. Introduction

Earth's surface and crust are subject to deformation caused by external forces such as relative plate movements and glaciation/deglaciation cycles. The nearest plate boundary, the North Atlantic Ridge, is located north-west from the territory of Finland. Sea-floor spreading outwards from the ridge causes a uniform horizontal compression in the Fennoscandian crust. The slowly accumulated strain energy in the crust may be released in the form of a sudden slip, earthquake, or as a continuous displacement within time, aseismic creep. The movements generally occur along pre-existing zones of crustal weakness.

Seismotectonics defines the knowledge of the causative processes behind the current regional deformations. Earthquake data (location, depth, size, focal mechanism), geodetic observations, and in situ stress measurements are the main sources of information in the seismotectonic context. Earthquakes sample deeper parts of the crust while the other observations provide information on stress states at the upper few kilometers of the crust. An earthquake focal mechanism gives estimates of the faulting plane, the direction of slip within that plane, and the causative stress directions. Seismotectonics of Finland has been studied for many years, but the most recent digitally recorded earthquakes have enabled a detailed seismotectonic interpretation (*Saari, 1998; Uski et al., 2003*).

2. Earthquake data

The earliest indications of earthquake activity in Fennoscandia are related to the paleoseismic period after the end of the last glaciation, about 10000 years ago. The land uplift immediately after the retreat of ice was much more pronounced than today. Earthquakes assumed to be responsible for the large thrust faults found in northern Finland would have magnitudes varying between 5.5 and 7.5 (e.g., *Kuivamäki et al., 1998*).

Observations of present-day seismicity in Finland cover the years from 1610 to the present. The older data are based on macroseismic observations, but since mid 1960's the earthquake source parameters have been determined mainly from instrumental recordings. Due to the different degree of completeness and accuracy of the source parameters, the earthquake data have been divided into historical and instrumental catalogs (*Ahjos and Uski, 1992*). A statistical analysis of the present-day seismicity suggests that the maximum magnitude of earthquakes in Finland is M_L=5 (*Ahjos et al., 1984*).

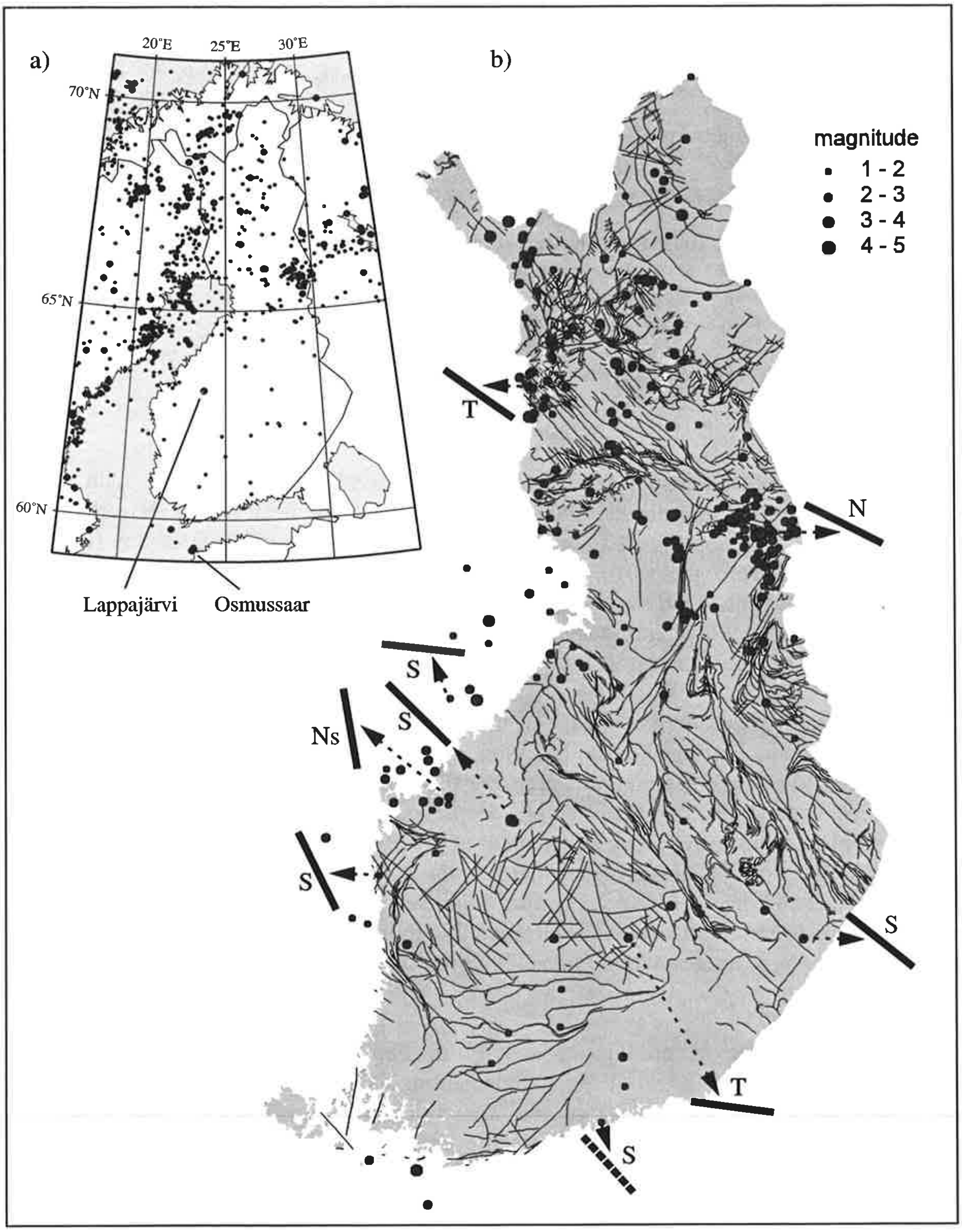


Figure 1. Distribution of earthquake epicenters (filled circles) for the period 1965-2001.
a) Regional seismicity map. b) Earthquakes in Finland and orientation of the largest horizontal compression as derived from the available fault plane solutions: Solid line=individual focal mechanism solution; dotted line=composite solution for 15 microearhquakes around the Loviisa nuclear power plant. Type of faulting: N=normal; S=strike-slip; T=thrust or reverse; Ns=predominately normal with a component of strike-slip. Structural features are modified from *Korsman and others* (1997).

The distribution of earthquake epicenters for the instrumental period (1965-2001) is shown in Fig.1. The figure also displays the orientation of the largest horizontal compression as derived from available fault plane solutions (*Slunga and Ahjos, 1986; Saari 1998; Uski et al., 2003*). Local concentrations of epicenters in the Bothnian Bay area, Kuusamo district, and western Finnish Lapland are clearly seen in Fig. 1. The seismic activity in the Bothnian Bay area correlates well with the major faults and shear zones, e.g. the NW-SE-striking Quark Shear zone extending approximately from Umeå to Vaasa. The strongest instrumentally recorded earthquake in Finland, the 17 February 1979 Lappajärvi earthquake (M_L= 3.8) occurred inland from this seismic zone. Bothnian Bay area is the center of postglacial land uplift in Fennoscandia. Postglacial rebound with a maximum uplift rate of about 10 mm/year is regarded as a significant local stress generator in the region.

Kuusamo is the seismically most active area in Finland. The earthquakes there are confined to the topographically elevated Kuusamo block (*Airo, 2000*). The block is transected by several NE-oriented shear zones starting from central Finland and ending to the White sea. Earthquake activity seems to follow the trend of those fractures and to cluster around the intersections with NW-SE-striking fault-lines.

In nortwestern Finland earthquake epicenters appear to concentrate in a broad N-S-directed zone running from the Bothnian Bay to the Atlantic Ocean. The zone intersects prominent NW- SE-striking Precambrian shear zones traversing the Central Lapland granite area and the known postglacial faults in the Kolari district. Part of the current seismic activity in Kolari as well as in northern Norway and Sweden is associated with the postglacial faults in the region (*Arvidsson, 1996; Bungum and Lindholm, 1996; Uski et al., 2003*).

The earthquakes in southern and central Finland have been weak and randomly scattered. We however point out that the 25 October 1976 event (M_L=4.9) in Osmussaar, northwestern coast of Estonia (Fig. 1), may be associated with a fracture zone running from the Ålands archipelago through Estonia.

3. Current stress field

Figure 1 shows that a NW-SE-oriented horizontal compression is dominating the stress release for the earthquakes in Finland. The normal faulting mechanism of the Kuusamo event deviates from the general pattern. However, with only one focal mechanism solution at hand, it is too early to draw conclusions on the local stress regime in Kuusamo.

The World Stress Map (available at: www.worl-stress-map.org) confirms that the dominating horizontal compressional stress in Fennoscandia is oriented NW-SE, but the local variation is significant. The impression is that the ridge-push compression alone may not be sufficient to cause earthquakes. Instead, a superposition of plate tectonic forces and local/regional stresses, interacting with favourably oriented zones of crustal weakness, is needed to explain the uneven distribution of seismic activity and the spatial inhomogeneity of the stress field.

The relative importance of postglacial rebound in triggering current seismicity is yet uncertain.The rebound model by *Wu and others* (1999) predicted that although tectonic stress is dominating the current stress orientation, the rebound stress available today is sufficient to to reactivate optimally oriented pre-existing faults. However, the predicted dominance of thrust faulting and concentration of major earthquakes near the uplift center are not consistent with observations.

4. Conclusions

Earthquakes in Finland occur in old Precambrian faults and shear zones, which have been reactivated. The present dominating stress field in in Finland is most likely of plate tectonic origin. The local stress field related to a single earthquake is a combination of the ridge push compression, postglacial rebound and local geology. Obviously, the relative significance of the three factors vary from region to region. Further regional investigations on earthquake mechanisms and other stress determinations are therefore needed. In a low seismicity area, microearthquake analysis can provide reliable data on local stress field and slip pattern as well as on active faults and their geometry (*Saari, 1998*). It is thus possible, in a relatively short term, to get regionally and statistically representative information on the present-day stress states and earthquake triggering -mechanisms.

5. Acknowledgements

Research assistant Kati Karkkulainen from the Institute of Seismology helped producing the structural geological map.

References

Ahjos, T., Saari, J., Penttilä, E. and Korhonen, H., 1984. Earthquakes and seismic hazard in Finland. Engineering geology, 20:1-12.

Ahjos, T. and Uski, M., 1992. Earthquakes in northern Europe in 1375-1989. Tectonophysics, 207, 1-23.

Airo, M-L., 1999. Aeromagnetic and petrophysical investigations applied to tectonic analysis in the northern Fennoscandian shield. Geological Survey of Finland, Report of Investigation 145, 51 p.

Arvidsson, R. and Kulhanek, O., 1994. Seismodynamics of Sweden deduced from earthquake focal mechanisms. Geoph. J. Int., 116, 377-392.

Bungum, H. and Lindholm, C., 1996. Seismo- and neotectonics in Finnmark, Kola and the southern Barents sea, 2: Seismological analysis and seismotectonics. Tectonophysics, 270, 15- 28.

Korsman, K., Koistinen, T., Kohonen, J., Wennervirta, M., Ekdahl, E., Honkamo, M., Idman, H. and Pekkala, Y. (Editors), 1997. Bedrock map of Finland 1:1,000,000. Geological Survey of Finland, Espoo, Finland.

Kuivamäki, A., Vuorela, P. and Paananen, M., 1998. Indications of postglacial and recent bedrock movements in Finland and Russian Karelia. Geological Survey of Finland, Nuclear waste disposal research, Report YST-99, 97 p.

Saari, J., 1998. Regional and local seismotectonic characteristics of the area surrounding Loviisa nuclear power plant in SE Finland. Institute of Seismology, Univ. Helsinki. Report 1998.

Slunga, R. and Ahjos, T., 1986. Fault mechanism of Finnish earthquakes, crustal stresses and faults. Geophysica, 22 (1-2), 1-13.

Uski, M., Hyvönen, T., Korja, A.-K. and Airo, M.-L., 2003. Focal mechanisms of threee earthquakes in Finland and their relation to surface faults. Accepted to Tectonophysics.

Wu, P., Johnston, P. and Lambeck, K., 1999. Postglacial rebound and fault instability in Fennoscandia. Geophys. J. Int., 139,657-670.

Earth Dynamics and Environmental Change are the Headlines in the IODP, the Successor to the Successful ODP

Kari Strand[1], Carl Ehlers[2] and Karla Tiensuu[3]

[1]Thule Institute, P.O. Box 7300, FIN-90014 University of Oulu, Finland
[2]Department of Geology and Mineralogy, Åbo Akademi University, FIN-20500 Turku, Finland
[3]Department of Geophysics, Institute of Geosciences, P.O. Box 3000, FIN-90014 University of Oulu, Finland
E-mail: kari.strand@oulu.fi

The Ocean Drilling Program (ODP), which comes to an end after 2003, has been a highly successful international cooperation in all of the geosciences. Finland has a membership in this program through the European Science Foundation Consortium for Ocean Drilling (ECOD). Since 1985 cruise and drilling operations have been made by a modernized drill-ship, *JOIDES Resolution.* The successor of the ODP, the Integrated Ocean Drilling Program (IODP) which will use multiple drilling platforms and new drilling technology is one of the main tasks of the international geoscientific community in year 2003.

Keywords: Earth's dynamics, climatic change, resources and deep biosphere

1. ODP 1985-2003

The Ocean Drilling program (ODP) is a large international geoscientific program with up to 22 member countries. This long-running partnership of research institutions and geoscientists is presently planning the schedule for its final year of operations. Finland has a membership in this program through the European Science Foundation Consortium for Ocean Drilling (ECOD) since 1986 (http://www.geo.vu.nl/~esco). Cruise and drilling operations are made by a modernized 144 metres long drill-ship, *JOIDES Resolution.* The cruise participants are offered a unique possibility to participate in international scientific work, and to explore and create new knowledge on climatic history and dynamics of the Earth's interior. Further information can be obtained at the ODP website: http://www.oceandrilling.org and a special issue of the JOIDES Journal vol. 28 (2002).

2. Finnish participation to ODP

The ODP related projects in Finland (Table 1) have concentrated to high-latitude climate history studies in North Pacific (Leg 145), North Atlantic (Leg 151) and Antarctica (Leg 188). The Earth's tectonic and interior processes have been studied in Chile Triple Junction (Leg 141) and Côte D'Ivoire-Ghana Transform Margin (Leg 159). The ODP is now going to execute its final year operations in 2003 (Becker, 2002).

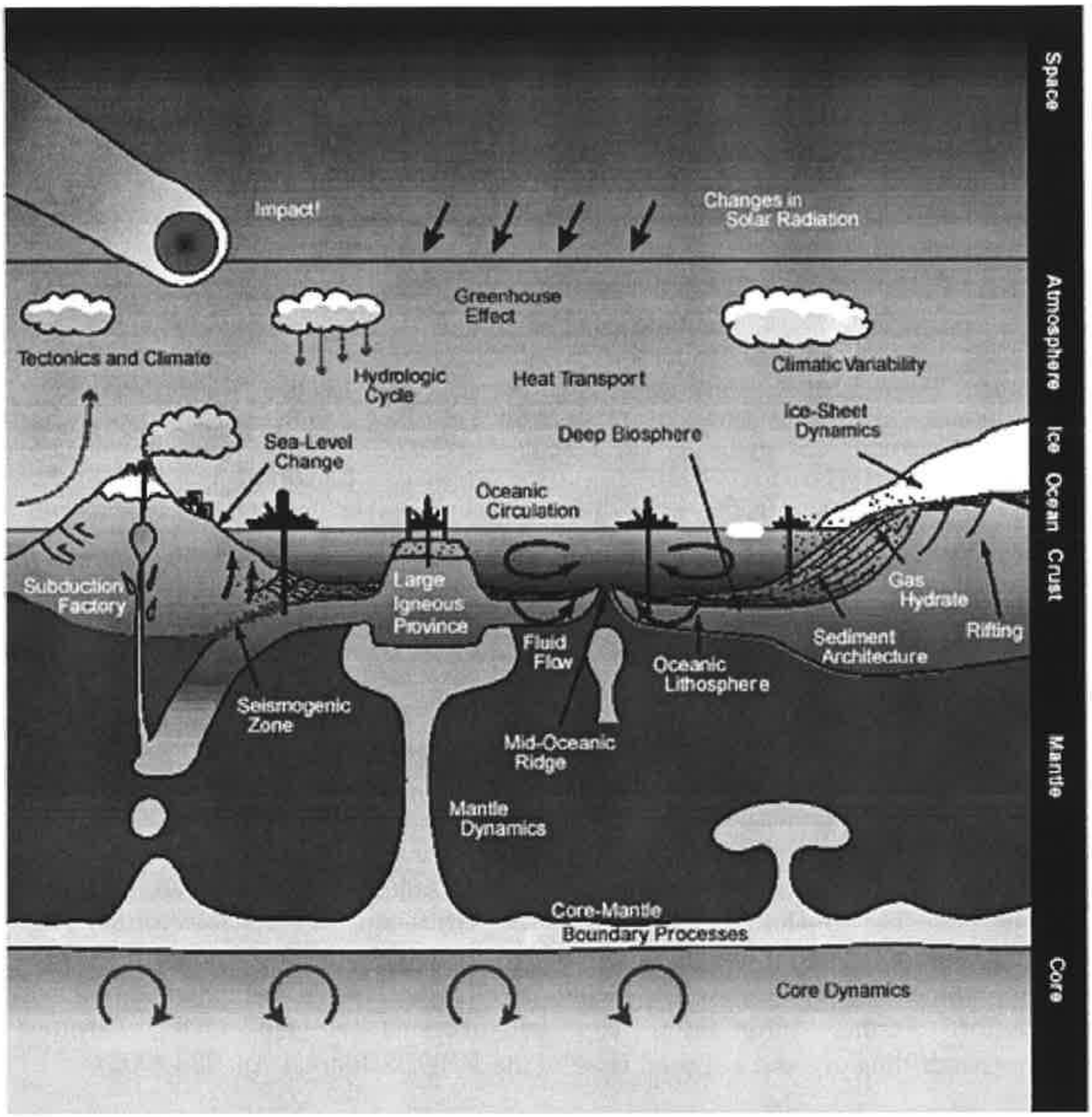


Figure 1. Earth system components, processes, and phenomena are targets for multiplatform ocean research drilling (figure from IODP Initial Science Plan, 2001).

Table 1. Finnish participation to the ODP.

Cruise	Leg description	Researchers involved	Publications
Leg 141; Chile Triple Junction	Represents the only presently active ridge-trench collision where the overriding plate is composed of continental lithosphere.	Kari Strand, Marko Matinlassi	1 MSc Thesis, 9 scientific publications, 8 conference abstracts
Leg 145; North Pacific Transect	The objectives were to collect high-resolution records of Miocene to Quaternary changes in ocean circulation, biological activity, and global climate.	Aarno Kotilainen, Mia Kotilainen, Arja Hokkanen	1 PhD Thesis, 6 scientific publications, 2 conference abstracts
Leg 151; North Atlantic-Arctic Gateways I	During the Arctic summer of 1993, *JOIDES Resolution*, accompanied by the Finnish icebreaker MSV Fennica, recovered first scientific drill cores from the eastern Arctic Ocean, including material which records the earliest history of the onset of glacial climate in the Arctic.	Mattiina Ruikka, Kari Strand	1 MSc Thesis, 1 scientific publication, 2 conference abstracts
Leg 159; Cöte D'Ivoire-Ghana Transform Margin Eastern Equatorial Atlantic	Drilled within continental crust adjacent to the continent-ocean transition along the transform passive margin. Represents the first deep-sea drilling to the tectonics of transform-margin development.	Kari Strand, Titta-Mia Kaivola	1 MSc Thesis, 7 scientific publications, 6 conference abstracts
Leg 188; Prydz Bay-Cooperation Sea, Antarctica: Glacial History and Paleoceanog raphy	Drilled to decipher Cenozoic glacial history and paleoenvironments of Antarctica. The drilled sites provide (1) records of the transition from East Antarctic preglacial to glacial conditions, (2) the variability of onshore erosion and glaciomarine depositional settings during latest Neogene, and (3) the transition from temperate to cold-climate glaciation since early Miocene time.	Kari Strand, Mattiina Ruikka, Jari Näsi, Jussi Peuraniemi, Katri Vaittinen, Hanna Silvennoinen, Karla Tiensuu, Juho Junttila	Under work: 1 PhD Thesis, 2 MSc Thesis, so far 2 scientific publications, 7 conference abstracts
Leg 208; Walvis Ridge	The proposed drill sites in the southern Atlantic Ocean will be used to reconstruct in detail the paleoceanographic variations associated with several prominent episodes of early Cenozoic extreme climate change.	Henry Wallius (official ESCO nominee to this Leg)	To be drilled 8 March to 9 May 2003

3. IODP will replace ODP

The future of the ODP towards the Integrated Ocean Drilling Program (IODP) is presently one of the main subjects in present international geoscientific planning. The JOIDES Resolution has not the capabilities for the drilling projects of the future, and must be succeeded by new (multiple) platforms with new technology. Scientific opportunities for the new drilling platforms under the IODP are listed as follows (http://www.iodp.org, IODP Initial Science Plan, 2001):

THE DEEP BIOSPHERE & THE SUBSEAFLOOR OCEAN

- The subseafloor ocean in various geological settings
- The deep biosphere
- Gas hydrates

ENVIRONMENTAL CHANGE, PROCESSES AND EFFECTS

- Internal forcing of environmental change
- External forcing of environmental change
- Environmental change induced by internal and external processes

SOLID EARTH CYCLES & GEODYNAMICS

- Formation of rifted continental margins, oceanic large igneous provinces and oceanic lithosphere
- Recycling of oceanic lithosphere into the deeper mantle and formation of continental crust

Eight initiatives are identified that are ready to be addressed within the next decade of drilling. These initiatives represent areas of study that have been identified as high priority by either workshop reports or strong proposal pressure within the ODP.

- Deep Biosphere
- Gas Hydrate
- Extreme Climates
- Rapid Climate Change
- Continental Breakup & Sedimentary Basin Formation
- Large Igneous Provinces
- 21st Century Mohole
- Seismogenic Zone

4. European participation to IODP

Joint European Ocean Drilling Initiative (JEODI) has now established the basis of a management and operational structure for Europe as part of the international IODP which will commence in October 2003 The project is a "Thematic Network" of the 15 European member countries of ODP funded by the European Commission with the objective of providing the basis for European participation in IODP, as a single European member, and for operating "Mission Specific Platforms" (MSPs) as part of IODP. JEODI is also defining a science plan for oceanic drilling through to 2010 (Brussels workshop report, 2001). This plan will draw on the success (and failures) of ODP and will include all potential platforms for future scientific drilling and will underline key scientific objectives from an European perspective, such as the development of a scientific rational for drilling and definition of the technological requirements for scientific drilling in the Arctic (http://www.jeodi.org).

The JEODI management team comprises 9 coordinated work-package groups focusing on:

- Implementation plan of MSP projects as early as October 2003,
- Implementation of a network of logging facilities in Europe including physical properties of core-samples,
- Undertaking an inventory of shore- and ship-based facilities for drilling and core handling and storage in Europe,
- Management plan for Europe in IODP,
- Outreach and publicity programs.

The USA is in the process of reviewing the requirements for a new drilling vessel, similar to the *JOIDES Resolution*, but with enhanced station keeping and coring capabilities. The funding for this vessel will probably be approved in 2003 and the vessel will be available for drilling by December 2005.The Japanese JAMSTEC are in the process of constructing a drilling vessel with riser capabilities. This vessel is specifically designed for scientific drilling in over-pressured and other technically challenging rock formations. The vessel will be equipped with drilling capabilities and laboratories between 2003 and 2005. It will be available for the scientific community by 2007. The two platforms identified above will be unable to drill effectively in shallow water, in sedimentary environments dominated by sand and silt lithologies, and in ice-covered regions. The European operational component will be to provide the technological know-how to undertake drilling in these environments by using a suite of platforms (MSPs) to undertake scientific drilling in these environments.

European national partners in JEODI created, in January 2002, the European Consortium for Ocean Research Drilling (ECORD), which will serve as the official European structure for a the new era of scientific ocean drilling. The broad aims of the project is to identify, promote and endorse areas of novel and exciting science that are of particular interest to Europe and require global ocean drilling and to explore new ways of linking European technological capabilities with these scientific requirements. Also to work towards implementation and funding of "European Infrastructure" for scientific ocean drilling, which would be part of IODP.

References

Achievements and opportunities of Scientific Ocean Drilling – the Legacy of the Ocean Drilling Program, 2002. A special Issue of the JOIDES Journal vol. 28, 1., 96 p.
Becker, K., 2002. Ocean drilling program plans final year operations. EOS, Trans. AGU, 83, 157, 159.
European Consortium for Ocean Drilling (ECOD) website: http://www.geo.vu.nl/~esco (cited 2002-09-15).
Integrated Ocean Drilling Program (IODP) initiative website: http://www.iodp.org (cited 2002-09-15).
Earth, Oceans and Life., 2001. Scientific Investigation of the Earth System using multible Drilling platforms and new technologies- Integrated Ocean Drilling Program - Initial Science Plan, 2003-2013, 110 p.
Joint European Ocean Drilling Initiative (JEODI) website: http://www.jeodi.org (cited 2002-09-15)
Ocean Drilling Program (ODP) website: http://www.oceandrilling.org (cited 2002-09-15).
Workshop on Alternate Drilling Platform: Europe as the third leg of IODP, 2001. European Science Foundation Reports and working papers – Workshop – Brussels, 8-9 January 2001, 126 p.

Radiogenic Leads from Fissures and Sandstones Within the Fennoscandian Shield: Indications of Paleozoic Weathering-Related Enrichment of Base Metals

Krister Sundblad[1], Matti Vaasjoki[2] and Elisabeth Alm[3]

[1]Department of Geology, FIN-20014 University of Turku, Finland
[2]Geological Survey of Finland, P.O. Box 96, FIN-02151 Espoo, Finland
[3]University of Stockholm, Sweden
E-mail: krister.sundblad@utu.fi

Close to 40 occurrences of grossly radiogenic lead are known from various parts of the Fennoscandian Shield (Fig. 1). The types of occurrences include both 1) fluorite-calcite-galena (±sphalerite and/or pyrite) filled fissures in both Svecofennian rocks and rapakivi granites and 2) sandstone-hosted occurrences ranging from major ore deposits (Laisvall) to boulders encountered at the shoreline of the Bothnian Bay (Raahe).

Both analyses on new samples as well published results indicate that the data form a sublinear trend on the uranium-related $^{207}Pb/^{204}Pb$ vs. $^{206}Pb/^{204}Pb$ diagram, whereas there is definite scatter on the $^{208}Pb/^{204}Pb$ vs. $^{206}Pb/^{204}Pb$ diagram, where the effects of changes in the Th/U ratio become apparent Fig. 2). As no uranium minerals have been detected in either the fissure or the sandstone deposits, it is obvious that the radiogenic part of the lead cannot have been generated in situ, and hence the observed lead isotopic compositions must reflect the properties of their source areas. The sublinear trend observed for the uranogenic leads is consistent with a Paleozoic remobilization of lead generated in Svecofennian host rocks during a time period of well over 1 billion years. For the sandstone-hosted Laisvall type occurrences (Fig. 3) a fine structure, related on basement geology and local nappe structures is evident.

We conclude that the available results can be interpreted as indicating that:

1) The erosion of the Fennoscandian Shield had reached practically its present level already at the onset of the Phanerozoic eon.

2) Deposition of sedimentary piles thick enough to create diagenetic conditions at their base occurred in Paleozoic times.

3) Supergene fluids leached radiogenic lead (as well as other elements) from the basement and deposited them under favorable local conditions either in the cover rocks or in fissures within the basement.

4) Variations in the data show that the lead in the calcite-fluorite-galenas fissures is of rather local derivation, where as the sandstone hosted leads show a lesser range of variation, suggesting of fluid circulation over a larger area.

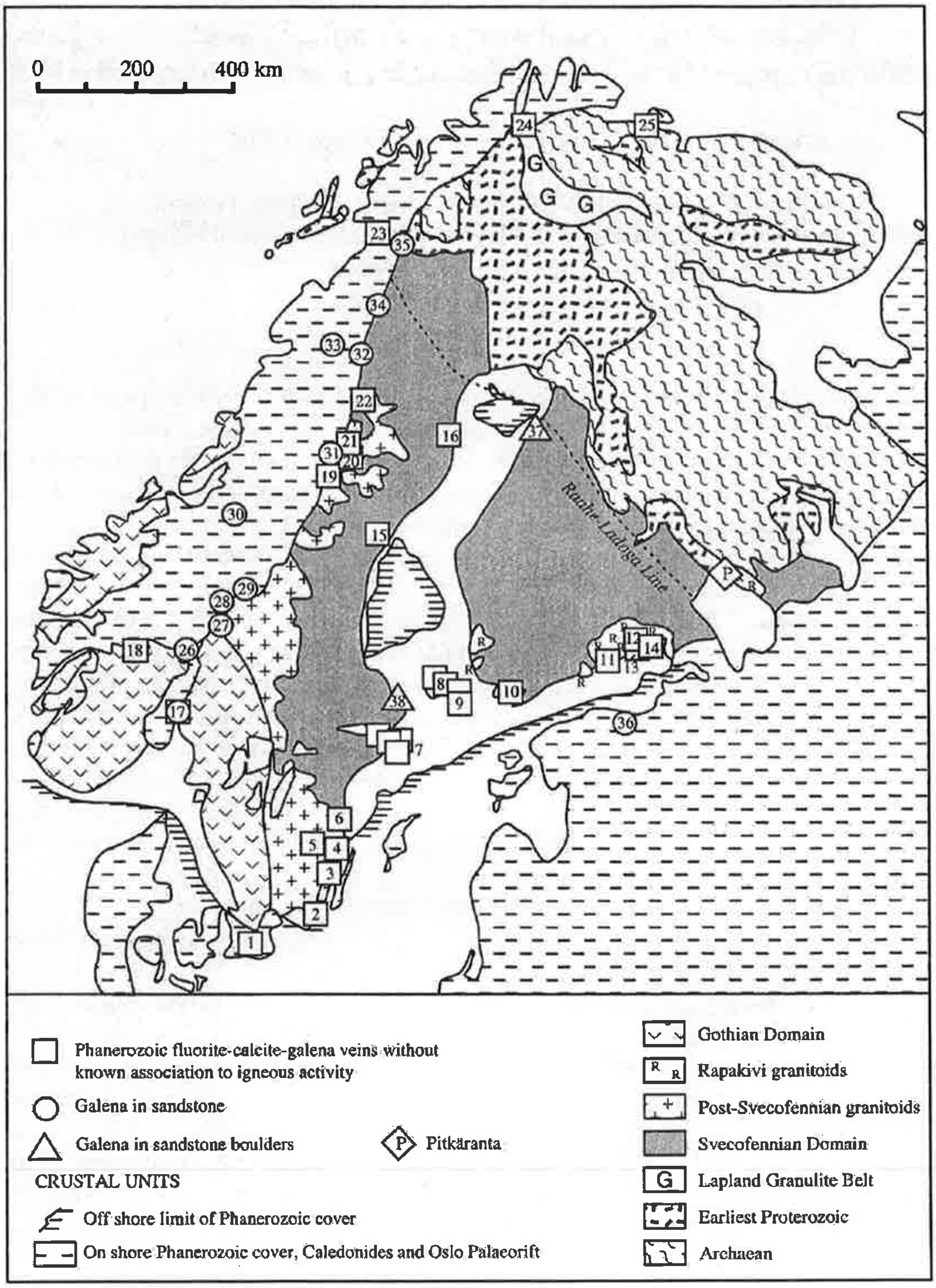


Figure 1. Locations of radiogenic galena occurrences in sandstones and calcite-fluorite-galena fissures.

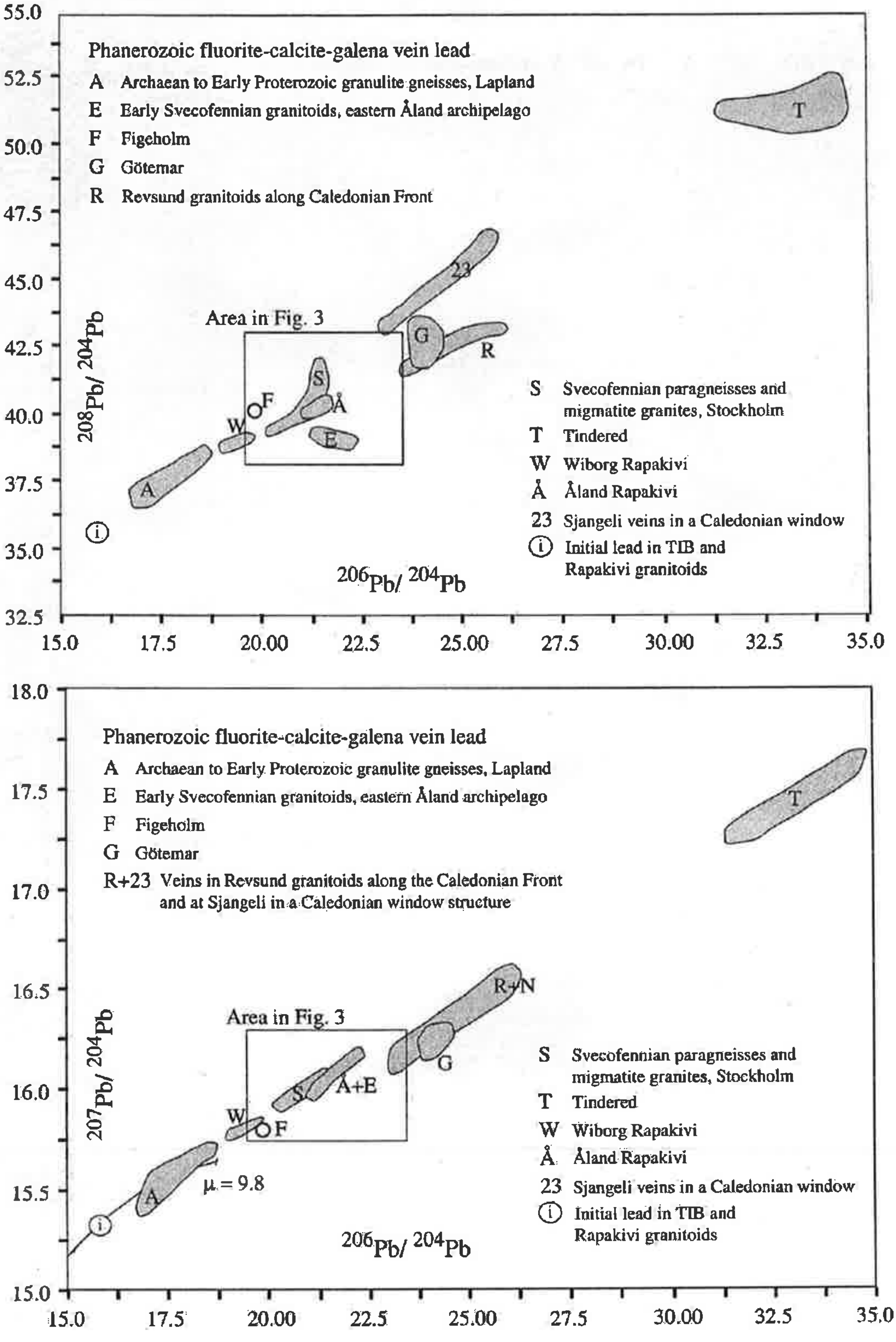


Figure 2. Summary diagrams for lead isotopic analyses from fluorite-calcite-galena filled fissures occurring both in Svecofennian rocks and rapakivi granites.

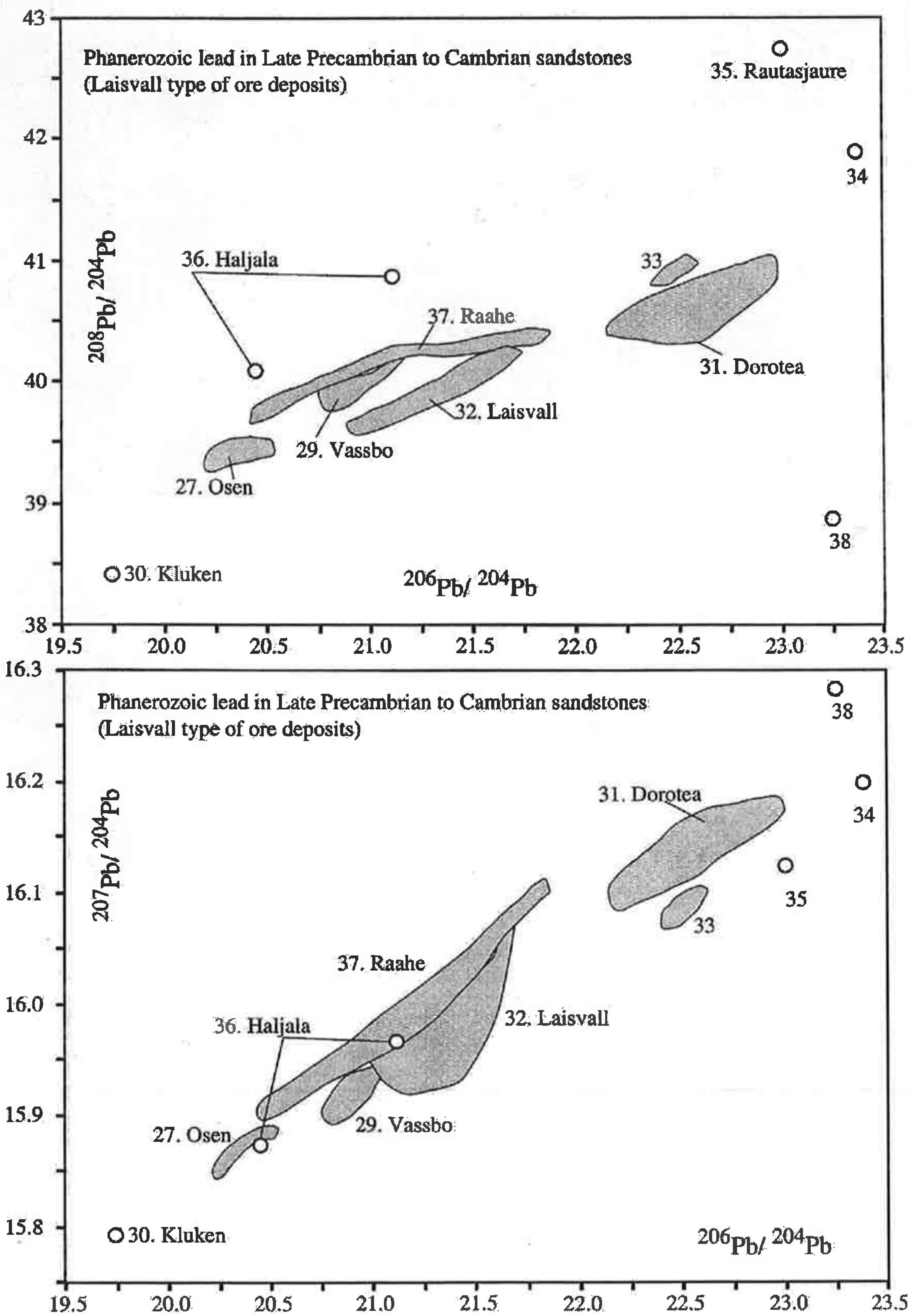


Figure 3. Summary diagram for lead isotopic data from sandstone-hosted base metal (Laisvall type) deposits and sulfide-cemented sandstone boulders from the Raahe area.

The Mineralogy of Pegmatitic Cavities in a Fractionated Rapakivi Granite – the Influence of Fluids in Elemental Mobility in a Crystallizing Front, Vehmaa Rapakivi Batholith, SW-Finland

Vesa Toropainen and Olav Eklund

Department of Geology, FIN-20014 University of Turku, Finland
E-mail: vesa.toropainen@utu.fi

Pegmatitic pockets, about 1 m in size, were studied to determine the mobility of elements in a crystallizing front, to examine in what phases the elements will participate and to investigate if certain elements are mobile with specific fluids in the crystallizing front. The result with this study is that certain elements concentrate in specific zones of the pegmatite in cases correlated with specific fluids. A particular result is that there is a higher concentration of HREE than LREE in the central part of the pegmatitic pockets.

1. Introduction

The Vehmaa batholith occupies an area of approx. 700 km^2 in SW Finland and appears to be built up concentrically *(Lindberg and Bergman 1993).* The major part of the batholith consists of a coarse K-feldspar porphyritic (up to 4 cm across) granite variety. Towards the central parts of the batholith, later pulses of a more even-grained variety appear. The even-grained porphyritic rapakivi is strongly differentiated compared to the major phase of the Vehmaa batholith (enriched in Rb, Y, Nb, Th and Li). Pegmatitic cavities, about 1 m in size occurs in the even-grained porphyritic granite. The pegmatitic cavity represents the final stage of magmatic activity in the batholith. Cavities principally represent a closed system into which late fluids transported incompatible elements. This took place after approximately 60 – 70 % crystallization of the magma when volatiles were separated into an independent phase in the magma at 2 kbar and approx. 700°C. The major and trace mineral association (beryl, micas, fluorite, sulphides, fluorocarbonates, phosphates) indicates that fluids (at least H_2O, F, S, CO_2) have had major influence in the late stage crystallization, and the fluid – rock interaction. The goal of this study is to determine the mineralogy of the different concentric zones of the cavities and examine the correlation between distinct fluids, mineralogy and geochemistry in the crystallization front.

2. Crystallizing conditions

We have been investigating the formation of new accessory phases as a result of the partitioning of elements between minerals-melts and minerals-fluids in the crystallizing front from one to two meter sized pegmatitic cavities in a fractionated granite variety. This even-grained variety is geochemically more fractionated than other varieties in the batholith (enriched in Rb, Y, Nb, Th and Li). The investigated pegmatitic cavities are found in this fractionated variety. The emplacement level of the batholith is considered to approx 2 kbar. At such pressures volatiles are released from the magma with 70% crystals as a separate phase at approx. 700°C. The solidus temperature of rapakivi magmas at 2 kbar has been determined to approx. 620°C *(Eklund and Shebanov 1998).* Measured temperatures from fluid inclusions in rapakivi cavities have been determined to approx. 420- 600°C (for references, see *Eklund and Shebanov 1998*).

The investigated cavities are zoned in a concentric pattern with zones either depleted or enriched in Al-rich phases. The principal Al-depleted zone is the quartz rich outermost part of the cavity in contact with the host granite. The zone inside the quartz-rich zone consists mainly of beryl, fluorite, K-feldspar and micas.

3. Sample preparation

To identify the major and accessory mineral phases and to trace the chemical variation in the pegmatite, the following samples were made. Polished thin sections were prepared from all zones of the cavity, all the way from the host granite to the central parts of the cavities (7 thin sections). The back-pieces of the thin sections were analyzed by ICP-MS (major elements, research quality of trace elements and H_2O+ and H_2O-, S, Cl, F, CO_2 and B). From the same zones, mineral separations were done to extract small amounts of minor heavy and light phases (unidentifiable in thin sections). These minor phases were analyzed with SEM-EDS for preliminary results and with a microprobe for quantitative results (EU geochemical facilities, Bristol). Also major mineral phases were analyzed with a microprobe.

4. Results, mineralogy

With distance from the host rock, the following zones appear

1) quartz enriched granite, zone 1
2) beryl - quartz – fluorite, zone 3
2) beryl - fluorite rich rock, zone 4
3) beryl - K-feldspar rock, zone 6

Minor, and accessory phases that were found in the zones: Biotite, muscovite, plagioclase (albite), apatite, zircon, titanite, pyrite, galena, chalcopyrite, magnetite, calcite, Ca-Mn-carbonate, secondary Fe-oxide, REE-xenotime, REE-monazite and LREE-fluorocarbonates (bastnäsite and synchysite; table 1).

Table 1. Compositions of REE-minerals in pegmatite pockets in the Vehmaa rapakivi batholith.

	Xenotime	Monazite	Bastnäsite	Synchysite	Fe-Y-rich Synchysite
Y_2O_3 %	42,47	0,21	0,46	1,55	2,17
Ce_2O_3 %	n/a	33,07	36,39	25,14	21,11
La_2O_3 %	n/a	16,02	17,06	11,06	9,46
Pr_2O_3 %	n/a	3,59	3,9	2,77	2,42
Nd_2O_3 %	n/a	12,27	13,72	9,67	8,44
Sm_2O_3 %	n/a	1,09	1,42	1,45	1,22
Gd_2O_3 %	1,36	n/a	n/a	n/a	n/a
Dy_2O_3 %	5,09	n/a	n/a	n/a	n/a
Ho_2O_3 %	1,28	n/a	n/a	n/a	n/a
Er_2O_3 %	4,79	n/a	n/a	n/a	n/a
Tm_2O_3 %	0,5	n/a	n/a	n/a	n/a
Yb_2O_3 %	6,32	n/a	n/a	n/a	n/a
Tb_2O_3 %	0,26	n/a	n/a	n/a	n/a
Lu_2O_3 %	0,98	n/a	n/a	n/a	n/a
Nb_2O_5 %	0,01	0,04	0,01	0,02	0,02
SrO %	0	0	0,03	0,03	0,06
FeO %	0,37	0,07	0,12	0,16	3,69
CaO %	0,04	0,05	0,66	17,21	16,84
MgO %	n/a	0	0	0,01	0,03
Na_2O %	n/a	0	0	0	0,01
UO_2 %	0,04	0,11	0,02	0,04	0,05
ThO_2 %	0,88	1,41	0,2	0,09	0,89
P_2O_5 %	35,56	30,68	0,02	0,02	0,04
F %	0,35	0,6	6,91	4,36	5,25
Oxide Totals	100,29	99,21	80,91	73,72	71,69

Note: The oxide totals of REE-fluorocarbonates lack the proportion of CO_3 and possibly some F. Synchysite (Ce,La,Nd)Ca$(CO_3)_2$F, Bastnäsite (Ce,La,Nd)CO_3F.

5. Geochemistry

With the distance from the host rock towards the beryl dominated rock:

- there is a strong enrichment of Be, Sc, Cs, HREE, Y, F and a slight enrichment of Al_2O_3, LREE, Sb, B, Cl (In, As)
- there is a depletion of SiO_2, MgO, FeO, P_2O_5, Ba, Rb, Ta, Nb, Tl, W, CO_2
- almost constant concentrations was found for Cu, U, Th, Zr, Sn
- enrichment with subsequent depletion MnO, Ce, Pb, Ni, Cr, Co, S
- in the sample most remote from the host rock, there is abrupt decrease in otherwise enriching trends of HREE, CaO, Sb, Y, F, S
- in the sample most remote from the host rock there is an abrupt increase in otherwise depleting trends of K_2O, Na_2O, Rb, Ba, Tl

6. Discussion

The enrichment of particular elements in certain zones in the pegmatitic cavity is clearly seen in mineralogy and geochemistry of different zones.

Transition elements (Mn, Co, W, Mo, Ni, Cr and Pb) are enriched in the quartz-dominant zone (1). Also the highest sulphur concentration is found in the zone 1, indicating those elements tendency to form sulphide phases.

Fluid compatible RE-elements, as well as Ca, have enriched towards the middle zones (3 and 4). The LREEs reach their maximum earlier (in zone 3) with the highest F content, compared to the heavy RE-elements (in zone 4). The LREEs and Ca are situated in fluorite, fluocerite, ksenotimes and fluorocarbonates in opposition to the HREEs, which are concentrated in monazites. The LREE/HREE partitioning may also have been affected by the changing concentrations of Cl and CO_2 during crystallization, as the LREE have slightly bigger D values in low temperature and pressure *(Wendtland and Harrison, 1979),* and the HREE form stronger complexes with chlorine ions *(Henderson, ed., 1984).*

Some trace elements (As, Ga, Sc and In) are seen to follow increasing boron concentration towards the innermost zone (6). Although the mineral phase for boron is unknown, as for example tourmaline was not found in the rock.

Large ion lithophile elements (K, Rb and Ba) and Na concentrated in K-feldspar and albite, have high values in the background granite (zone 1) and the innermost zone (6), and show low concentration in the beryl and fluorite dominated zones 3 and 4.

The geochemistry was compared to the Väkkärä granite in the Eurajoki rapakivi granite stock *(Haapala 1977).* Both of the granites represent the latest phases of rapakivi magmatism, and are the most differentiated parts of the batholiths they belong to. The major element geochemistry of the background Helsinki-granite is similar compared to the even grained Väkkärä granite. In the Väkkärä granite, the mineralogy of the greisen includes the same mineral phases as in the Be-rich cavity studied here: beryl, fluorite, carbonate, chalcopyrite, secondary iron oxide, pyrite, magnetite, galena, monazite, xenotime and bastnaesite.

Table 2. The trace element concentrations in Helsinki and Väkkärä granite *(Haapala 1977)* varieties.

	Helsinki cavity Even grained	Väkkärä even grained	Väkkärä porphyritic	Väkkärä coarse gr.	Be- Greisen in Väkkärä	Be- in Helsinki
Be	6	3,4	<3 - 42	<3 -5	1400	32640
Sn	5	18	84	75	77	4
Ga	24	29	67	63	90	31
Rb	327	440	885	900	1830	120
Sr	86	37	<40	<40	50	22
Ba	445	150	<100	<100	<100	111
Zr	346	220	68	83	74	322
Nb	51	< 40	54	63	50	20

The trace element concentrations (Table 2.) show that the Helsinki granite is highly differentiated, but not comparable to the most differentiated porphyritic and coarse-grained parts of the Väkkärä granite. *Haapala (1977)* observed, that Be and Sn do not show good correlation in their enrichment, and that that they may enrich in different greisenization processes. This may be the case also within the Helsinki granite, as also Sn enriched greisens have been found in the Helsinki granite.

References

Eklund, O. and Shebanov, A.D. (1998) The origin of rapakivi texture by sub-isothermal decompression. *Precambrian Research 95, 129-146.*

Haapala, I. (1977) Petrography and geochemistry of the Eurajoki stock, a rapakivi-granite complex with greisen-type mineralization in southwestern Finland. *Geological Survey of Finland, Bulletin* ***286.***

Henderson P. ed. (1984) Rare earth geochemistry. *Developments in Geochemistry;* **2.** Elsevier Science Publishers B.V., The Netherlands.

Lindberg, B. and Bergman L. (1993) Pre-quarternary rocks of the Vehmaa map-sheet area. *Geological maps of Finland 1:100 000.* Explanation to the maps of pre-Quarternary rocks, sheet 1042. *Geological Survey of Finland, Espoo, Finland.*

Wentland, R.F. and Harrison, W.J., 1979. Rare earth partitioning between immiscible carbonete and silicate liquids and CO_2 vapor: results and implications for the formation of light rare earth enriched rocks. *Contrib. Mineral. Petrol.,* ***70*** *: 341 - 356*

The South Finland Shear Zone - Ductile Shearing of Paleoproterozoic Crust in SW Finland

Taija Torvela and Carl Ehlers

Åbo Akademi University, Department of Geology and Mineralogy,
Domkyrkotorget 1, FIN-20500 Åbo, Finland
E-mail: ttorvela@abo.fi, cehlers@abo.fi

The ca 1.9 Ga old Paleoproterozoic crust in southern Finland was inhomogeneously deformed along several E-W / NW-SE trending shear zones. The South Finland shear zone (SFSZ) shows evidence of a prolonged activity that started with a wide zone of ductile shearing and ended with strain partitioning along distinct mylonitic and pseudotachylitic bands. The field data indicate a dextral shearing along the zone as well as an uplift of the southern part relative to the northern side of the zone (preliminary results).

Keywords: Svecofennian, Paleoproterozoic, ductile, shear zone, Finland, mylonite

The ca 1.9 Ga old Paleoproterozoic crust in southern Finland was inhomogeneously deformed along several E-W / NW-SE trending shear zones. A major ductile shear zone, "the South Finland shear zone" (SFSZ) can be traced for almost 200 km in the archipelago off the southern and southwestern coast of Finland. Our study area is situated in the archipelago of SW Finland (fig. 1). The more than a kilometre wide zone shows traces of a long history of activity starting with a wide zone of ductile dextral shearing followed by more concentrated mylonitic and pseudotachylitic episodes. The latest phases of activity took place in a late-orogenic stage, contemporaneously with the emplacement of the ca 1.8 Ga old Mosshaga granite and related (shear controlled?) 1.79 Ga old microgranitic intrusions (*Ehlers & Skiöld 2001*). The earliest possible age (ca 1.65 Ga) for the SFSZ is determined by amphibolitic metadiabase dykes that cut through the bedrock south of the shear zone, while the maximum age is app. 1.86 Ga determined by early Svecofennian granitoid gneisses (*Ehlers & Skiöld 2001*).

The shearing along the SFSZ resulted in different lithologies on both sides of the NW-SE trending zone. The northern side of the shear zone consists of gently dipping (ca 1.83 –1.84 Ga old) migmatites with abundant traces of metamorphosed supracrustal and volcanic rocks while to the south the bedrock is dominantly (ca 50 Ma older) granodioritic and amphibolitic gneiss metamorphosed at up to upper amphibolite facies. The contemporary app. 1.79 Ga granites and later anorogenic rapakivi granites cut the shear zone in many places (fig. 1).

The field data implicate a dextral shearing along the zone. The field data further preliminary suggest that there was an uplift of the southern part of the bedrock relative to the northern side of the shear zone. The most important evidence for this is the regularly, gently to steeply eastward plunging stretching lineations observed in the rocks throughout the study area. Lineations are, however, poorly preserved in the app. 1-2 km wide zone that shows the most intensive shearing.

The shear zone itself is mainly mylonitic, metamorphosed at up to upper amphibole facies with abundant secondary quartz, feldspar and later epidotic veins. There are sporadic, thin sections that show granulite facies metamorphism but the main metamorphism did not seem to induce extensive melting of the bedrock along the shear zone. There is also evidence of prolonged activity along the zone: later movements within the early wide ductile zone show strain partitioning along distinct mylonitic bands followed by later pseudotachylytes. This could indicate continued rapid movements along the zone during the crustal uplift after the main shearing event.

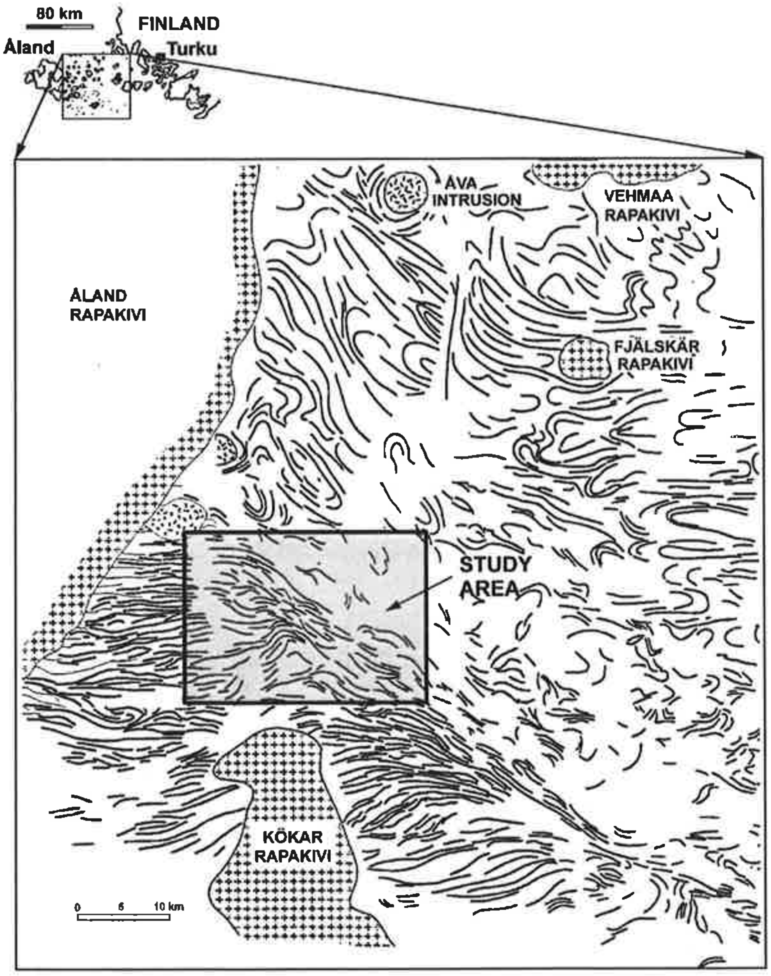


Figure 1. Study area and a structural map of Åland archipelaco (a compilation of observed foliations). A dextral shear zone cuts through the archipelago: the same zone can be followed for another 150 km along the southern coastline of Finland (SE of the map). The largest partly contemporary granitic intrusions as well as later rapakivi intrusions are shown. Modified from Ehlers & Skiöld 2001.

The foliation at the margins of the zone is folded along E-W trending, fold axes plunging gently to steeply eastwards. The zone with the most intensive shearing shows a regular NW-SE foliation and is about 1-2 km wide.

According to the field observations both pure shear and simple shear components have been involved. In the area of the most intensive shear a stretching component is clearly visible as abundant amphibolitic boudines. Further research is, however, required before a more detailed strain analyses can be constructed.

The SFSZ continues E-SE of the study area all the way to the area off the coast of West Uusimaa. The zone may be related to similar ca 1,8 Ga ductile shear zones in Central Sweden (*Högdahl 2000*). There are also several other NW-SE trending shear zones that cut the Svecofennian domain in Western and Central Finland that very likely have a connection to the shearing in Southern Finland. Also the abundant, mainly sinistral N-S trending shear zones in Southern Finland, Åland archipelago and Central Sweden are possibly a part of the one major shearing event that has affected the whole Svecofennian domain.

References:

Ehlers, C. & Skiöld, T. (2001): SIMS-dating of ductile shearing in SW Finland. EUG XI, Journal of Conference Abstracts, Vol. 6, nr.1, Abstract Volume, p 624.

Högdahl, K. (2000): Late-orogenic, ductile shear zones and protolith ages in the Svecofennian domain, central Sweden. Ph.D. thesis, Stockholm University, Department of geology and Geochemistry.

Paleozoic Remobilisation and Enrichment of Proterozoic Uranium Mineralisation in the East-Uusimaa Area, Finland

Matti Vaasjoki, Hannu Appelqvist and Kari A. Kinnunen

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

Mineralogical and fluid inclusion studies suggest that secondary remobilisation of uranium at low temperatures and pressures has enriched sheared U-mineralisation in southern Finland. U-Pb dating of various minerals indicates that the emplacement of original low-grade U-mineralization occurred during the late orogenic Svecofennian migmatizing event c. 1.83 Ga ago and during the intrusion of the anorogenic Wiborg rapakivi batholith c. 1.64 Ga ago. U-Pb analyses of uraninites and allanites from the sheared deposits indicate that the remobilisation and enrichment of uranium occurred in Paleozoic times, most likely during the Ordovician c. 0.45 Ga ago.

It is suggested that this was caused by supergene oxidizing solutions in fracture zones which extended across the Precambrian-Paleozoic unconformity from the crystalline basement to its now eroded and removed sedimentary cover. Thus the high-grade uranium occurrences found in shear zones in the East-Uusimaa area may be classified as unconformity-related uranium deposits.

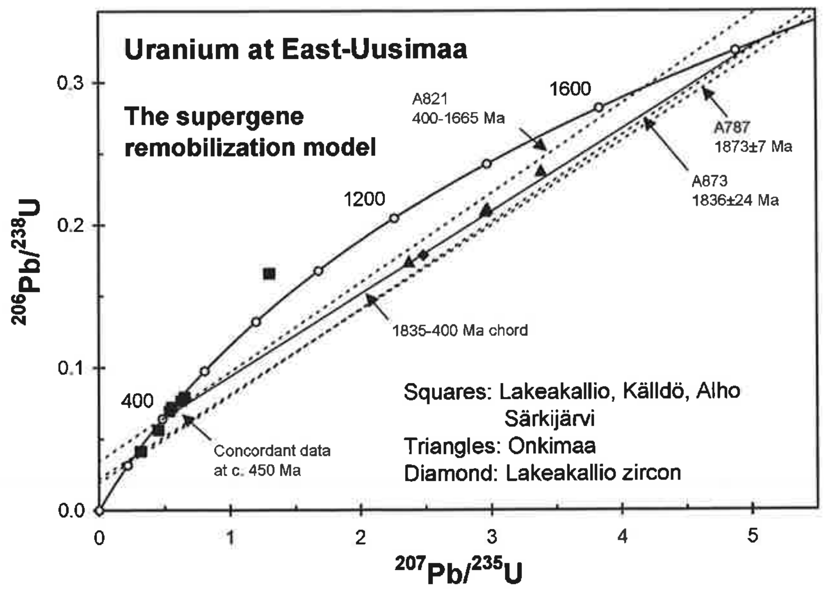


Figure 1. U-Pb analyses of uraninites from the East-Uusimaa area on a concordia diagram. The discordia chords represent possible lead loss scenarios.

Ore Lead Isotope Systematics During the Svecofennian Orogeny

Matti Vaasjoki[1] and Krister Sundblad[2]

[1]Geological Survey of Finland, P.O. Box 96, FIN-02151 Espoo, Finland
[2]Department of Geology, FIN-20014 University of Turku, Finland
E-mail: matti.vaasjoki@gsf.fi

The isotopic composition of ore lead is primarily a function of two parameters; 1) time; the age of the metal sources as well as the time of the ore forming process and 2) geochemical character of the source rocks. For volcanogenic ore deposits, the ore forming process is closely related in time to major igneous activity, which governs hydrothermal leaching of trace amounts of metals from large volumes of rocks. Subsequently, the metals are redeposited in sulfide ores and the lead isotopic compositions of the metal sources are preserved in minerals where lead is a principal cation. Thus if the age of the host rocks (i.e., the ore forming event) is independently known, the observed variations in ore lead isotopic composition in volcanogenic ore deposits reflect the geochemical character and crustal maturity of the metal sources of the ores.

Both published and new lead isotopic data (Fig. 1) from more than 250 samples representing 97 base metal occurrences within the Svecofennian domain of the Fennoscandian shield are considered in their regional geological context. Based on geological, geophysical and geochemical criteria, the Svecofennian domain can be divided into eight different terranes (Fig. 2), which are:

1) the Arjeplog-Vihanti-Joroinen belt, 2) the Skellefte basin, 3) the Central Finland granitoid complex, 4) the Bothnian Basin, 5) the Tampere schist belt, 6) the Rockliden-Häme belt, 7) the Bergslagen-Orijärvi-Savo block and 8) the Fröderyd inlier.

Most of these terranes contain ores with rather typical lead isotopic characteristics, although local deviations may arise from the immediate geological situation. Thus, e.g., within the Skellefte district, the stratigraphically higher deposits in metasediments dominated lithologies exhibit more radiogenic leads than deposits in the lower, volcanic dominated formations.

A time-dependent three-way mixing model with Archean U-poor (lower) crust, Archean U-rich (upper) crust and Paleoproterozoic mantle as end members is presented and it is apparent that the model could account for all variation found in the Svecofennian ore leads (Fig. 3). Examination of the geological implications of the model shows that this interpretation of lead isotopic data from the Fennoscandian Shield is consistent with recent plate tectonic scenarios (e.g., Lahtinen 1994, Nironen 1997, Korsman et al. 1999) suggesting that several new terranes were accreted to the active continental margin from a southern/southwesterly direction.

The rocks of the Arjeplog-Vihanti-Joroinen belt, representing a primitive arc system, were first welded to the Archean craton, and by about the same time the next arc formation in the area encompassing the Skellefte district and the central Finnish granitoid complex was reaching maturity. Magmatic activity started with the extrusion of rhyolitic and dacitic rocks, which were deformed and partly consumed by syntectonic granitoids intruded from 1895 to 1885 Ma within the back-arc basin. This led to the generation of a massive continental segment in central Finland and north-central Sweden, resulting in a continental margin setting in the Bergslagen-Orijärvi-Rantasalmi Belt.

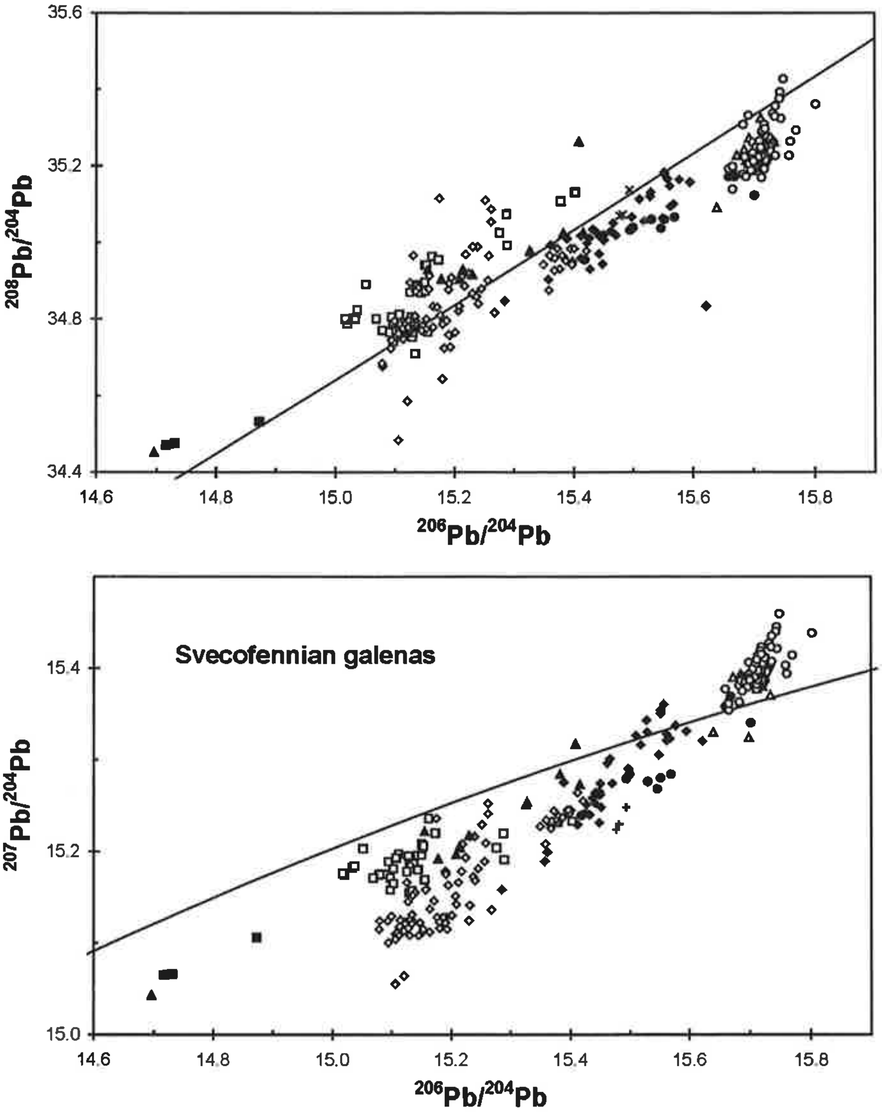


Figure 1. Lead isotopic compositions of Svecofennian galenas from Finland and Sweden. Closed squares: Outokumpu. Open squares: Arjeplog-Vihanti-Joroinen. Closed diamonds: central Finland granitoid complex. Open diamonds: Skellefte district. Closed circles: Rockliden-Häme belt. Open circles: Bergslagen-Orijärvi-Savo block. Crosses: Fröderyd inlier. Lead evolution curves according to Stacey and Kramers (1975).

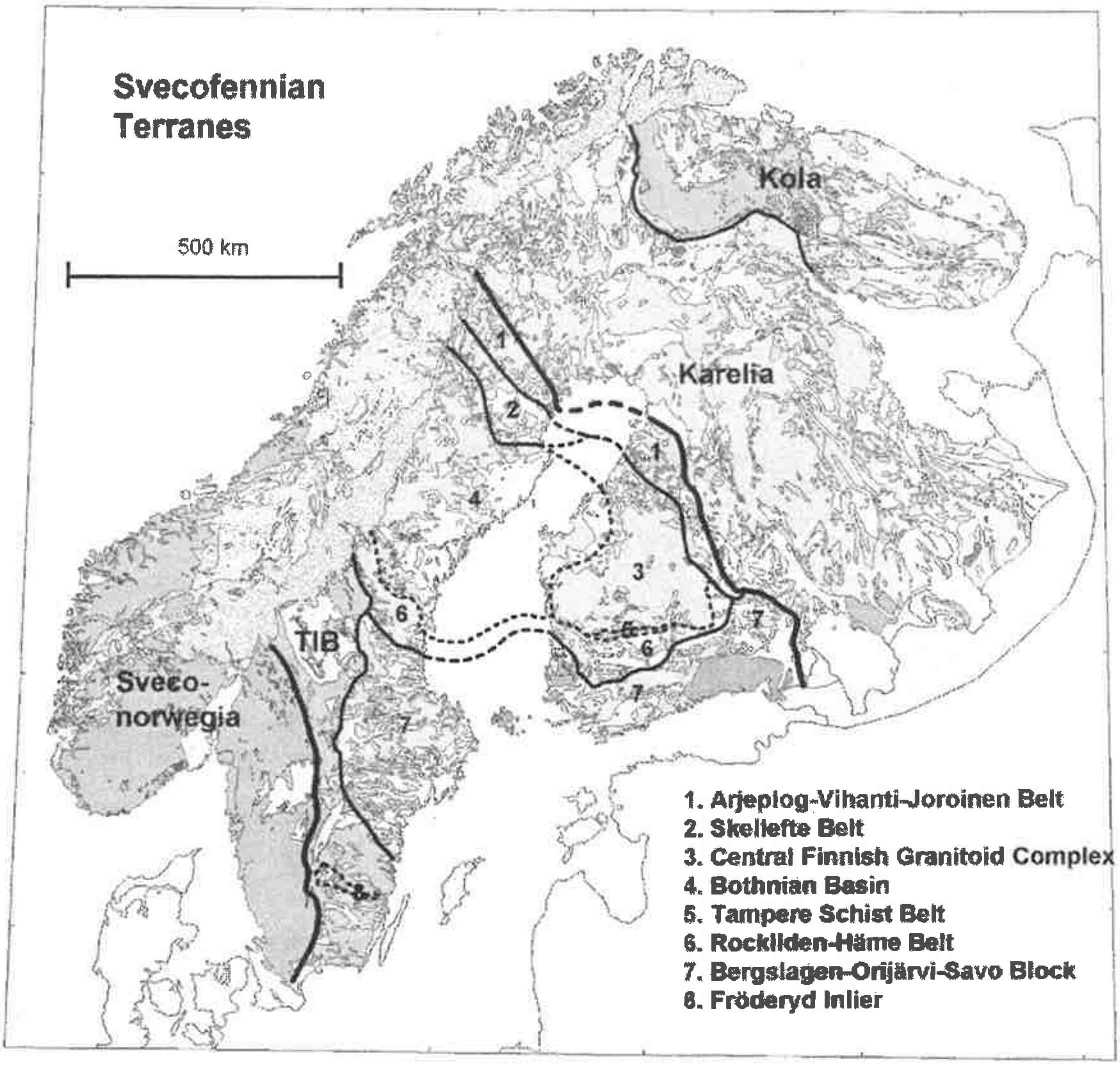


Figure 2. A terrane division of the Svecofennian based on geological, geochemical and lead isotopic criteria. Base map simplified from Gorbunov and Papunen (1985).

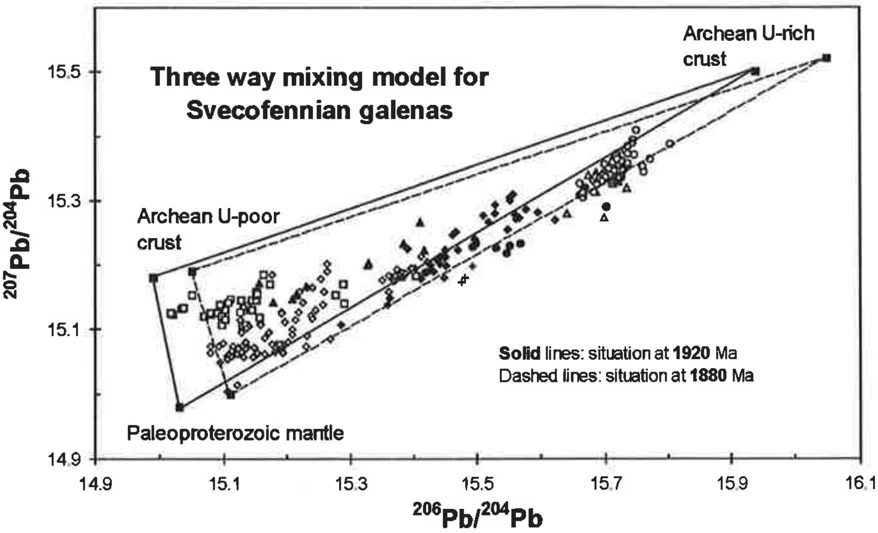


Figure 3. Relationship of the analytical data (Outokumpu and Haveri omitted) in relationship to the proposed three way mixing model.

References

Gorbunov, G. & Papunen, H. (eds.) 1985. General geological map of the Baltic Shield 1:2 500 000. Geological Survey of Finland, Bulletin 333.

Korsman, K., Korja, T., Pajunen, M. & Virransalo, P. 1999. The GGT/SVEKA transect: structure and evolution of the continental crust in the Paleoproterozoic Svecofennian orogen in Finland. International Geology Review 41, 287-333.

Lahtinen, R. 1994. Crustal evolution of the Svecofennian and Karelian domains during 2.1-1.79 Ga, with special emphasis on the geochemistry and origin of 1.93-1.91 Ga gneissic tonalites and associated supracrustal rocks in the Rautalampi area, central Finland. Geological Survey of Finland, Bulletin 378, 128 p.

Nironen, M. 1997. The Svecofennian Orogen: a tectonic model. Precambrian Research 86, 21-44.

Stacey, J. & Kramers, J. 1975. Approximation of terrestrial lead isotope evolution by a two-stage model. Earth and Planetary Science Letters 26, 207-221.

Comparison of Crustal Models of the Sarmatia and Fennoscandian Shields

Jukka Yliniemi[1], Timo Tiira[2] and Kari Komminaho[3]

[1]Sodankylä Geophysical Observatory, P.O. Box 3000, FIN-90014 University of Oulu, 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: jukka.yliniemi@oulu.fi

During the years 1979-2002 Finnish seismologists have participated in many international Deep Seismic Soundings (DSS) projects in Finland and abroad. In Finland the international collaboration studies started with FINLAP profile in northern Finland in 1979. The second profile was the well-known SVEKA'81. It was followed by BALTIC (1982), POLAR (1985), SVEKA'91 and FENNIA (1994). In 1989, a reflection experiment BABEL was carried out on the Baltic Sea. A simultaneous refraction study was carried out with continuously recording temporary land stations along the coasts. The latest international co-operation was SVEKALAPKO (1998-1999) Seismic Tomographic Experiment in the southern and middle Finland.

The first European DSS-project where the Finnish seismologist participated was the Seismic Refraction Sounding (LT-7) in NW-Poland in 1987. The cooperation with the Polish seismologist had already started from SVEKA'81 project and the intensive joint work has continued more than 20 years not only in Poland and Finland but also in other countries. Examples of the joint studies are TTZ (1993, Poland), EUROBRIDGE'95 (1995, Lithuania), EUROBRIDGE'96 (1996, Belarus), POLONAISE (1997, Poland), EUROBRIDGE'97 (1997, Belarus), CELEBRATION 2000 (2000, Poland) and ALP2002 (2002, Austria). The locations of the profiles have been presented in figure 1.

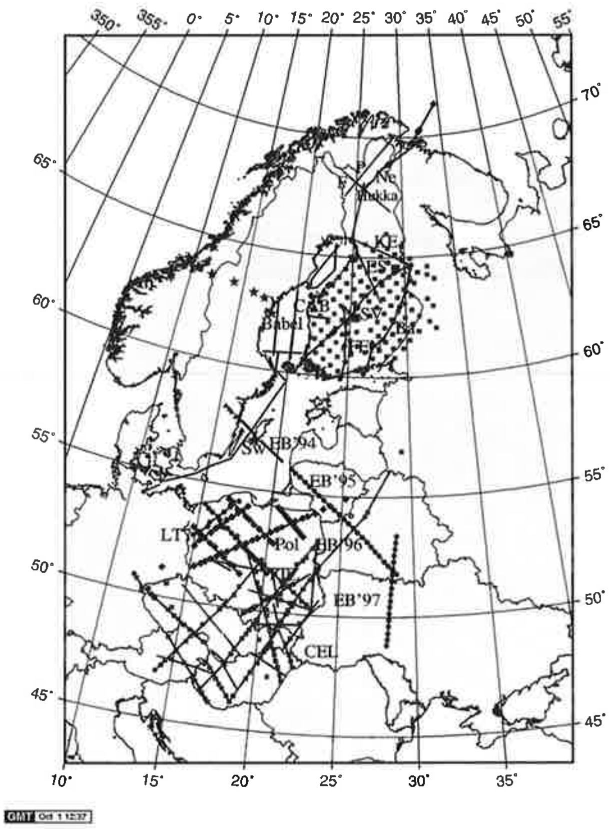


Figure 1. Location map of DSS profiles, to which Finnish seismologists has participated during 1979-2001.

Selected references

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

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

EUROBRIDGE Seismic Working Group, 2001. EUROBRIDGE `95: deep seismic profiling within the East European Craton. Tectonophysics 339,153-175.

EUROBRIDGE Seismic Working Group, 1999. Seismic velocity structure across the Fennoscandia-Sarmatia suture beneath the EUROBRIDGE profile through Lithuania and Belarus. Tectonophysics, 314, 193-217.

Fennia Working Group, 1998. P- and S-velocity structure of the Baltic Shield beneath the FENNIA profile in southern Finland. Inst. Seimology, Univ. Helsinki, Rep. S-38, 14 p.

Grad, M. and Luosto, U., 1987. Seismic models of the crust of the Baltic shield along the SVEKA profile in Finland. Ann. Geophysicae, 5B, 639-650.

Grad, M., Janik, T., Yliniemi, J., Guterch, A., Luosto, U., Tiira, T., Komminaho, K., Sroda, P., Höing, K., Makris, J., and Lund, C.-E., 1999. Crustal structure of the Mid-Polish Trough beneath the TTZ seismic profile. Tectonophysics 314,145-160.

Guterch, A., M. Grad and G.R. Keller, CELEBRATION 2000 Organizing Committee - the authors mentioned above, and K. Posgay, J. Vozar, A. Spicak, E. Brueckl, Z. Hajnal, H. Thybo, O. Selvi and CELEBRATION 2000 Experiment Team - T. Janik, P. Sroda, W. Czuba, E. Gaczynski, M. WidePiórko, K.C. Miller, S. Harder, E. Hegedus, P. Hrubcova, K. Aric, F. Kohlbeck, I. Aseduh, R. Clows, P. Joergensen, S.L. Kostiuchenko, G. Jentzsch, D. Krake, T. Tiira, J. Yliniemi, A. Belinsky. Seismologists Celebrate the New Millenium with an Experiment in Central Europe. EOS, Trans., American Geophysical Union, Vol. 82, No. 45, pp. 529-534-535, 2001.

Guterch, A., M. Grad, H. Thybo, R. Keller and the POLONAISE Working Group, 1999. POLONAISE `97 - International seismic experiment between Precambrian and Variscan Europe in Poland. Tectonophysics 314,101-121.

Luosto, U., Flueh, E.R., Lund, C.-E., 1989. The crustal structure along the POLAR Profile from seismic refraction investigations. Tectonophysics 162, 51-85.

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

Luosto, U., Heikkinen, P., Komminaho, K.,Yliniemi, J., 1994. Crustal structure along the SVEKA'91 profile in Finland. 24th General Assembly of the European Seismological Commission, 1994 September 19-24, Athens, Greece: Proceedings and activity report 1992-1994. - Athens: University of Athens, 1994. 2, 974-983.

Luosto, U., Heikkinen, P., 2001. Ylävaipan seismisestä rakenteesta Etelä-Suomessa. In: Meri-Liisa Airo and Satu Mertanen (Eds.). Geofysiikan päivät : Helsingissä 15.-16.5.2001. Geofysiikan seura, Helsinki, p. 117-122.

Sroda, P. and POLONAISE Working Group (including T. Tiira), 1999. P- and S-wave velocity model of the southwestern margin of the Precambrian East European Craton; POLONAISE'97, profile P3. Tectonophysics 314,175-192.

ISBN 951-45-7992-5
ISSN 0357-3060

Helsinki 2002
Helsinki University Press