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LITHOSPHERE 2010

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

PROGRAMME AND EXTENDED ABSTRACTS

edited by

Pekka Heikkinen, Katriina Arhe, Toivo Korja , Raimo Lahtinen, Lauri J. Pesonen and Tapani Rämö

University of Helsinki
Kumpula Campus
Helsinki, October 27-28, 2010

Helsinki 2010

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Keywords (GeoRef Thesaurus, AGI): lithosphere, crust, upper mantle, Fennoscandia, Finland, Precambrian, Baltic Shield, symposia

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PREFACE

The Finnish National committee of the International Lithosphere Programme (ILP) organises every second year the LITHOSPHERE symposium, which provides a forum for lithosphere researchers to present results and reviews as well as to inspire interdisciplinary discussions. The sixth symposium - LITHOSPHERE 2010 - will have 41 presentations. Extended abstracts (this volume) will give a good overview on current research on structure and processes of solid Earth. This symposium has two special themes; evolution of the Archaean lithosphere and international lithosphere research programs. Within the general theme of the meeting – structure, composition and evolution of the lithosphere – contributions on four topics were especially encouraged: major features of the lithosphere, crustal deformations, supercontinents & mantle plumes and natural resources. The symposium will focus at these issues in the following sessions:

Session 1: Lithosphere: major features
Session 2: Crustal deformations
Session 3: Natural resources
Session 4: Other lithosphere studies
Session 5: Posters
Session 6: Supercontinents and mantle plumes
Session 7: Evolution of the Archaean lithosphere
Session 8: International lithosphere research programs

The two-day symposium is hosted by the University of Helsinki and it will take place at the Kumpula campus in October 27-28, 2010. The participants from the Universities of Helsinki, Turku and Oulu, Åbo Akademi, the Geological Survey of Finland, the Finnish Geodetic Institute, Rice University and Universities of Aberdeen and Tartu will present their results in oral and poster sessions. Posters prepared by graduate or postgraduate students will be evaluated and the best one will be awarded. The traditional invited talk is given by Prof. Alan Jones from the Dublin Institute of Advanced Sciences.

This special volume ***"LITHOSPHERE 2010"*** contains the programme and extended abstracts of the symposium in alphabetical order.

Helsinki, October 14, 2010

Pekka Heikkinen, Toivo Korja, Raimo Lahtinen,
Lauri Pesonen and Tapani Rämö

Lithosphere 2010 Organizing Committee

LITHOSPHERE 2010 Symposium Programme

Wednesday, October 27

09:00 - 10:00	Registration at the University of Helsinki, auditorium D101, Kumpula campus area, Physicum
10:00 - 10:05	Opening of the symposium
10:05- 11:50	**Session 1: Lithosphere: major features** Chair: Pekka Heikkinen
10:05- 10:50	**A.G. Jones [Invited]** Lithospheric structures and geometries revealed by deep-probing magnetotellurics
10:50- 11:10	**T. Tiira, T. Janik, E. Kozlovskaya, M. Grad, K. Komminaho, E. Hegedűs, C. A. Kovács, E. Brückl, A. Korja and H. Silvennoinen (given by E. Kozlovskaya)** Preliminary P- and S-wave velocity model of HUKKA 2007 wide-angle reflection and refraction profile: an evidence for an unknown terrain boundary?
11:10 - 11:30	**T. Korja** Electrical conductivity of mantle lithosphere in Fennoscandia
11:30 - 11:50	**H.E. O'Brien** Craton mantle roots: Do we really know how they form and how old they are?
11:50 - 13:00	**Lunch**
13:00- 15:30	**Session 2: Crustal deformations** Chair: Raimo Lahtinen
13:00- 13:20	**T. Hyvönen, T. Tiira, A. Korja and K. Komminaho** Velocity blocks and anisotropy in the crust of the central Fennoscandian Shield
13:20- 13:40	**A. Kärki, C. Ehlers, S. Paulamäki, T. Torvela and P. Tuisku** Internal structures of ductile shear zones
13:40- 14:00	**T. M. Lammi and T. Kilpeläinen** Lineation map of Finland – a project
14:00- 14:20	**A. Korja, T. Kilpeläinen, T. Lammi, T. Hyvönen, K. Nikkilä and T. Torvela** Consequences of orogenic spreading – examples from the Svecofennian orogen
14:20 - 14:50	**Coffee/Tea**
14:50- 15:10	**J. Mattila** Paleostress and 3D slip and dilation tendency analysis of fault zones – a case study from Olkiluoto, SW Finland.

15:10- 15:30 **S. Nyberg, M. Poutanen and U. Kallio**
Monitoring crustal deformations in Satakunta using GPS measurements

15:30- 17:10 Session 3: Natural resources
Chair: Lauri Pesonen

15:30- 15:50 **W.D. Maier, P. Peltonen, A. Kontinen, I. McDonald and S.-J. Barnes**
Comparison between PGE contents in Kaapvaal and Karelian sub-continental lithospheric mantle, based on kimberlite xenoliths and ophiolite samples

15:50 - 16:10 **S. Heinonen, M. Imaña Osorio, I. Kukkonen and P. Heikkinen**
Seismic imaging of deep massive sulphides in Pyhäsalmi, Finland

16:10- 16:30 Break

16:30 - 16:50 **I.T. Kukkonen, P. Heikkinen, S. Heinonen, J. Laitinen, P. Peltonen and HIRE Working Group**
HIRE Seismic Reflection Survey in the Hannukainen-Rautuvaara Fe-Cu-Au exploration area, Northern Finland

16:50 - 17:10 **E. Laine, K. Saalmann and E. Koistinen**
3D modeling of polydeformed and metamorphosed rocks - an old Outokumpu Cu-Co-Zn mining area as an example

17:10- 17:30 Session 4: Other lithosphere studies
Chair: Markku Poutanen

17:10 - 17:30 **M. Väisänen, O. Eklund, Y. Lahaye, H. O'Brien, S. Fröjdö and M. Lammi**
Intra-orogenic Svecofennian magmatism in SW Finland: evidence from LA-ICP-MS zircon dating and geochemistry

17:30- Session 5: Posters
Chair: Toivo Korja

17:30 -18:00 Poster introductions (one min talks with one slide in lecture room)

P01 **S. Basan, S. Sjöblom and O. Eklund**
Intraorogenic mafic dykes in the accretionary arc complex of southern Finland

P02 **M. Grad, T. Tiira and ESC Working Group**
The Moho depth map of the European Plate

P03 **R. Klein, L.J. Pesonen, S. Mertanen and H. Kujala.**
Paleomagnetic study on Satakunta sandstone, Finland

P04 **T. Korja, M. Smirnov, E. Sokolova, I. Varentsov, N. Palshin, K. Vaittinen, and I. Lahti**
Archaean-Proterozoic boundary in the East European Craton: Crustal conductivity in the Fennoscandian and Ukrainian Shields

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Depth distribution of earthquakes under the Finnish part of Archean Karelian craton

P16 **K. Vaittinen, T. Korja, P. Kaikkonen and I. Lahti**
Magnetotelluric studies of the collisional and extensional processes in the central Fennoscandian Shield

P17 **T. Veikkolainen, K. Korhonen and L.J. Pesonen**
Testing the GAD model of the geomagnetic field by using igneous rock data

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- 21:00 **Networking including refreshments**

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Chair: I. Kukkonen

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Late Palaeozoic remagnetization in the Baltic Plate

09:40- 10:00 **F. Karell**
Magnetic studies as indicators for ascent and emplacement of rapakivi granites

10:00- 10:20 **S. Mertanen and F. Karell**
Magnetic data constraining the 1.63-1.45 Ga rifting episodes in the southern Fennoscandian shield

10:20- 10:40 **E. Koivisto and R.G. Gordon**
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11:10- 12:30 Session 7: Development of the Archaean lithosphere
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11:10- 11:30 **E. Heilimo, J. Halla, P.S. Hölttä, H. Huhma, T. Andersen and P. Mikkola**
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13:30- 13:50 **E. Kozlovskaya, O. Usoltseva and POLENET/LAPNET Working Group**
POLENET/LAPNET: project status and first results

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14:10 - 14:30 **I.T. Kukkonen, M.V.S. Ask, O.Olesen and PFDP Working Group**
Postglacial Fault Drilling Project: A new initiative for an ICDP project

14:30- 15:30 Short communications, discussions and poster award
Chair**:** Pekka Heikkinen

EXTENDED ABSTRACTS

Intraorogenic mafic dykes in the accretionary arc complex of southern Finland

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Keywords: Mafic dykes, intraorogenic magmatism, Svecofennian

General

Väisänen et al. (2010, this issue) present intraorogenic mafic – felsic magmatism in SW Finland. They conclude that in the time frame 1865 – 1850 Ma the Svecofennian crust was injected by shoshonitic magmas and subsequently the crust was partially melted resulting in granitic magmas. This took place before the regional metamorphism at approx. 1830.

Our research is to study amphibolite dykes crosscutting the Svecofennian pre- to synorogenic formations. Our hypothesis is that the dykes were puncturing the Svecofennian crust before the regional metamorphism, since they are in the amphibolite facies. The dykes may be associated with the pre- to synorogenic magmatism, but in places, the dykes crosscut the regional foliation why we consider them younger.

Age determination of the dykes will put them on the evolutionary scheme of the Svecofennian in southern Finland and the geochemistry will hopefully give some information about the tectonic setting for the dykes.

12 samples of amphibolite dykes, varying in width between 10 cm to 2 m, were collected in the Åland archipelago in an area covering 60 km from Eckerö in west towards Föglö in east. The samples were prepared for petrographic classification, geochemistry and age determination. However, by the time for the abstract deadline, no results were received. The results will be presented by a poster at LITO-2010.

This is a study within the consortium Timing and P-T-conditions for the orogenic extension in the Fennoscandian shield

Reference:

Väisänen, M., Eklund, O., Lahaye, Y., O'Brien, H., Fröjdö, S., Lammi, M. 2010. Intra-orogenic Svecofennian magmatism in SW Finland: evidence from LA-ICP-MS zircon dating and geochemistry. LITHOSPHERE 2010, this issue.

The Moho depth map of the European Plate

M. Grad[1], T. Tiira[2] and ESC Working Group

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The European Plate has a 4.5 Gy long and complex tectonic history. This is reflected in the present-day large-scale crustal structures. A new digital Moho depth map is compiled from more than 250 data sets of individual seismic profiles, 3-D models obtained by body and surface waves, receiver function results and maps of seismic and/or gravity data compilations. We have compiled the first digital, high-resolution map of the Moho depth for the whole European Plate, extending from the mid-Atlantic ridge in the west to the Ural Mountains in the east, and from the Mediterranean Sea in the south to the Barents Sea and Spitsbergen in the Arctic in the north. In general, three large domains within the European Plate crust are visible. The oldest Archean and Proterozoic crust has a thickness of 40–60 km, the continental Variscan and Alpine crust has a thickness of 20–40 km, and the youngest oceanic Atlantic crust has a thickness of 10–20 km.

Keywords: Crust, Moho, European Plate

1. Introduction

In 1910, the Croatian seismologist Andrija Mohorovičić (1857–1936) published his important paper ‘Potres of 8.X.1909’ (Earthquake of 8 October 1909). In this paper, he studied seismograms of an earthquake in the Kupa Valley (Croatia), together with other events from this region and he discriminated two distinct pairs of compressional (*P*) and shear (*S*) waves. He writes in his paper: ‘When I was sure, based on data, that two kinds of first preliminary waves exist, both kinds reaching all locations from 300 to 700 km distance, and that from the epicentre to approximately 300 km distance only the first kind arrives, whereas from 700 km distance onward only the second kind arrives, I tried to explain this until now unknown fact. In today’s nomenclature, the first kind of the arrivals correspond to crystalline basement *Pg* and *Sg* phases and overcritical crustal phases *P*crustal and *S*crustal, while the second kind of arrivals correspond to mantle *Pn* (*Sn*) and *P* (*S*) phases. The interpretation of the two sets of arrivals led Andrija Mohorovičić to discover the existence of the velocity discontinuity in the uppermost Earth. He evaluated the depth to be at 50 km, with *P*-wave velocities 5.60 km s^{-1} above and 7.747 km s^{-1} below (respectively, 3.27 and 4.182 km s^{-1} for *S* waves). Below the boundary surface, the velocity ratio was Vp/Vs=1.852, which was significantly larger than in the upper layer were it was 1.710. Studies during the next 100 yr showed that the sharp seismic discontinuity discovered by Mohorovičić was found worldwide, and that it separates crust from underlying upper mantle. It was named the Mohorovičić discontinuity or Moho in abbreviated form, or even M-discontinuity.

2. Data and the new Moho depth map

The oldest data come from early the 1970s and the 1980s, and most of them were compiled in regional maps that were published in the last 20 yr. A huge amount of new data were obtained in last decade, for example, in Central Europe, particularly within the refraction and wide-angle reflection projects POLONAISE'97, CELEBRATION 2000, ALP 2002 and SUDETES 2003. The results of different projects were evaluated, and they were given relative weighs in our compilation. The highest weigh was given to data from modern refraction and wide-angle reflection profiles with a dense system of observations and good reciprocal coverage. Moho depths were directly extracted from 2-D numerical models and transferred to geographical coordinates: latitude ϕ, longitude λ, and Moho depth h below the sea level. Older profiles were digitized by hand from published papers. Only those parts of the models that were sufficiently sampled by rays were included in the database. In case of significant disagreement of depth at crossing points of two profiles, either the results of the profile with better quality were used, or both profiles were removed from the data set. For some areas, we used regional Moho depth maps, compiled using deep seismic data, usually both refracted and reflected. Some of them we got in digital form, for example, for Germany and for the Barents

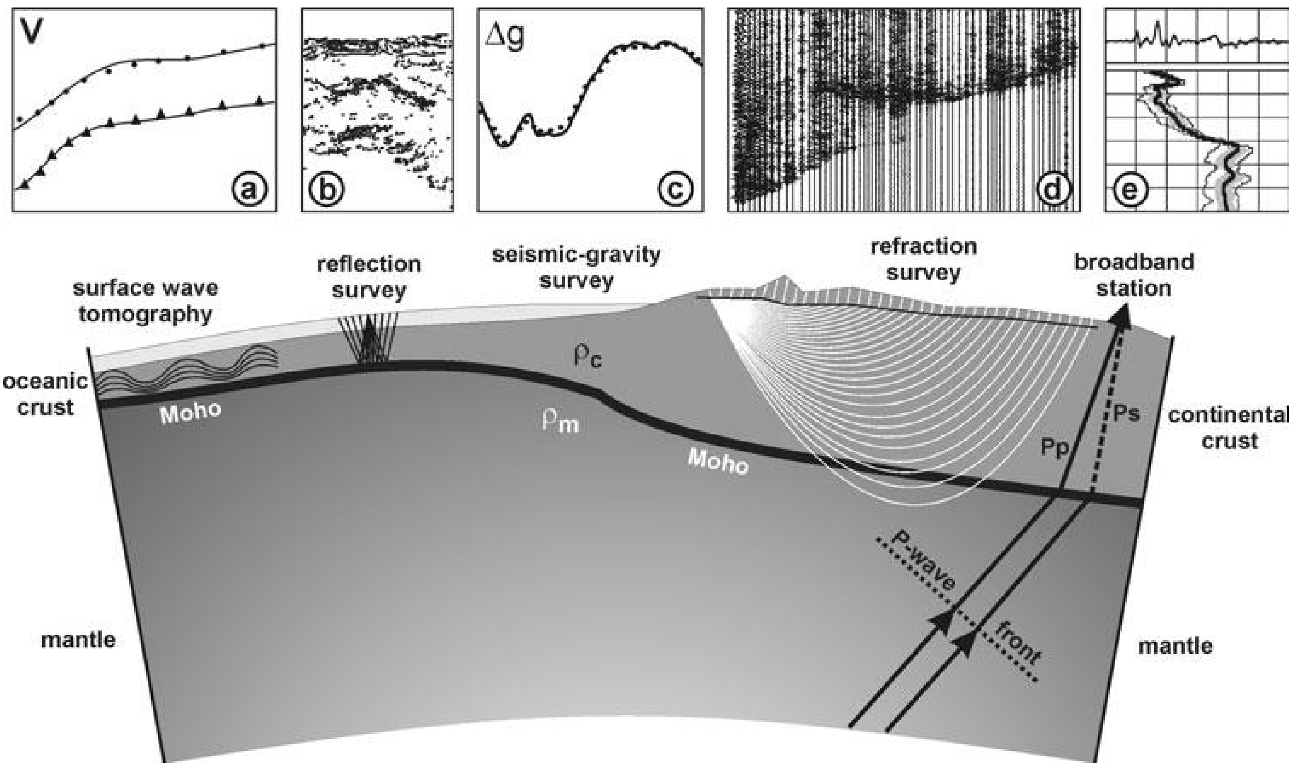

Figure 1. An illustration of the methods used to study crustal structure and a sketch of the waves penetrating into crust/mantle model: (a) dispersion curves of surface waves in surface wave tomography; (b) a seismic section in a near-vertical reflection profiling and reflected rays from the Moho; (c) a gravity model obtained using the contrast of densities of the crust (ρ c) and mantle (ρ m); (d) a controlled source seismic refraction profiling record section and rays of refracted waves in the crust and uppermost mantle; (e) a plot illustrating the RF technique with rays converted at the Moho—*Pp* (solid) and *Ps* (dashed)—the front of the teleseismic *P* wave is marked by dotted line.

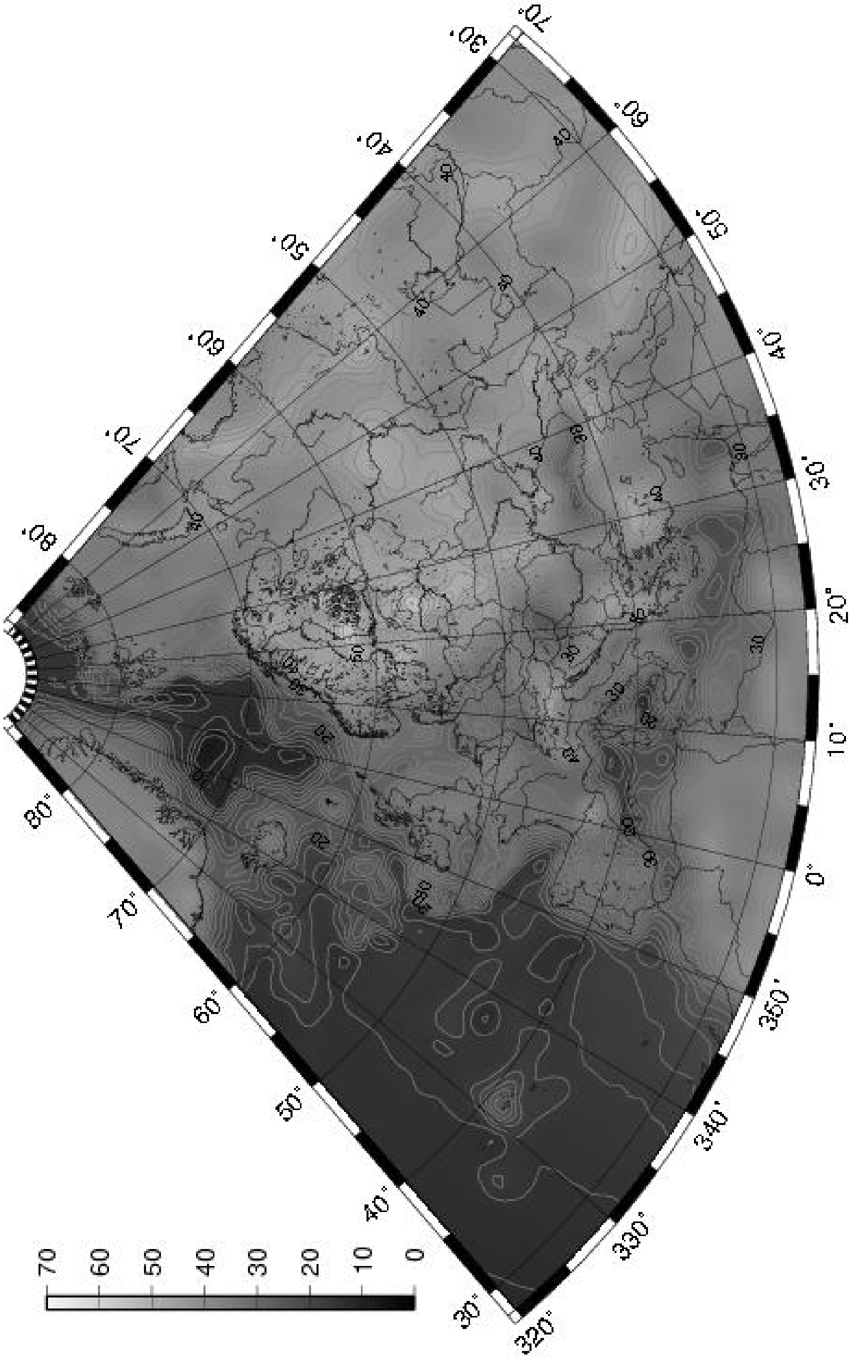

Figure 2. The Moho depth map of the European Plate. The database for this compilation comprises more than 250 data sets from individual seismic profiles, 3-D models obtained by body and surface waves, RF and maps of seismic and/or gravity data compilations.

Sea. For the territory of Russia a few tens of deep seismic sounding profiles, surveyed mainly by GEON, Moscow, were compiled into Moho depth map. We used also Moho depths in digital form from 3-D models, derived from local seismic tomography, surface waves and seismic/gravity modeling. Another class of seismic data were RF Moho depth estimations beneath permanent or temporary seismic stations. Areas without regional seismic or gravity data (usually at European Plate surroundings) were filled using more general, lower-resolution global models. Altogether, the Moho map database comprises more than 250 data sets.

3. Moho map for the European Plate

The new European Plate Moho depth map is a compilation of data, published before September 2007. The complex tectonic history of Europe reflects the breakup of a Neoproterozoic supercontinents Rodinia/Pannotia to form the fragment of Baltica and the subsequent growth of continental Europe, beginning with the Caledonian orogeny. Caledonian and younger Variscan orogenesis involved accretion of Laurentian and Gondwanan terranes to the rifted margin of Baltica during the Palaeozoic. The suite of sutures and terranes that formed, the so-called Trans-European suture zone (TESZ) adjacent to the rifted margin of Baltica, extends from the British Isles to the Black Sea region. The TESZ is far more complex than a single suture, but in a broad sense, it is the boundary between the accreted Phanerozoic terranes and Proterozoic Baltica. Understanding its structure and evolution is one of the key tectonic challenges in Europe and is certainly of global importance to studies in terrane tectonics and continental evolution. The younger Alps, Carpathian Mountains arc and Pannonian backarc basin in the south form interrelated components of the Mediterranean arc–basin complex and are the result of intricate Mesozoic/Cenozoic plate interactions in the Mediterranean region as the Tethys Ocean, closed during the convergence of Europe and Afro–Arabia. All tectonic processes and geological structures mentioned above have their images in the Moho depth map. Regional and local properties can be studied using existing models and data for relatively limited area. Here, we would like to concentrate on the continental scale of visible structures. In Moho map the thick crust of the East European platform (with the Baltic shield) is separated in the west from the Atlantic by the Caledonides. In the south, between TESZ and the Mediterranean Sea, accreted terranes form a much shallower crust (ATA, Iberia, Pannonian Basin), with the exception of the thick collisional Alpine crust. In general, three large domains within European Plate crust are visible. The oldest Archean and Proterozoic crust of thickness 40–60 km, continental Variscan and Alpine crust of thickness 20–40 km, and the youngest oceanic crust of Atlantic of thickness 10–20 km.

Moho map is available at webpages of the University of Helsinki and the University of Warsaw as a graphic (tiff, pdf, eps and jpg formats), as well as in data files (ASCII text and GMT format), with latitude, longitude, Moho depth and Moho depth uncertainty. We hope that new data and contributions could give an opportunity to update our map after 5–7 yr. It will be particularly useful to get new data sets from the mid-Atlantic ridge and Europe–Africa–Arabia transition, where the resolution of present map is the lowest. We also look forward to your contributions to our future map. The map can be found at:

http://www.igf.fuw.edu.pl/mohomap2007/ and
http://www.seismo.helsinki.fi/mohomap/

Grad, M., Tiira, T., and ESC Working Group, 2009. The Moho depth map of the European Plate, Geophysical Journal International, **176,** 279–292 doi: 10.1111/j.1365-246X.2008.03919.x

Neoarchean sanukitoid series magmatism in the Karelian Province, eastern Finland

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Sanukitoid series are minor and divergent group of Neoarchean igneous rocks. Sanukitoid series rocks have been described from Archean cratons worldwide. A group of Neoarchean intrusions in eastern Finland are distinguished as a part of sanukitoid series based on their whole-rock composition. U–Pb age determinations of this group gives an age range from ~ 2715 to 2745 Ma for the sanukitoid magmatism in Finland. We describe also whole-rock Sm-Nd isotopes, and *in situ* Hf-O isotopes from zircons of the sanukitoid series.

Keywords: sanukitoid, magmatism, Neoarchean, Karelia, Finland

1. Introduction

The term sanukitoid series encompasses granitoids, which are enriched in both compatible elements (Mg, Ni and Cr) and incompatible elements [LILE (K, Ba, Sr) and LREE] and have high Mg# at a given SiO_2 content. Neoarchean late to post-tectonic sanukitoid intrusions are found in the Finnish and Russian part of the Karelian Province and are often related to major shear zones (Lobach-Zhuchenko et al., 2005; Heilimo et al., 2010; Figure 1). Variable sized, even-grained to K-feldspar porphyritic intrusions form a series of diorites, tonalites, and granodiorites, that are calc-alkaline to alkali-calcic, magnesian, and mostly metaluminous. The major and trace element geochemistry of the intrusions show typical sanukitoid affinities that distinguish them from the TTGs: a mantle signature (high content of Mg, Ni, Cr and high Mg#) and enrichment in LILE (especially K, Ba, and Sr).

2. Geochronology

A Single-grain zircon U–Pb secondary ion mass spectrometer (SIMS) study shows temporally limited emplacement for the sanukitoid intrusions in eastern Finland as the ages vary between ~2715 and 2745 Ma (Heilimo et al., in revision). When these results are compiled with single-grain age data of the Russian (Bibikova et al., 2005) parts of the Karelian Province they confirm the occurrence of two temporally separate sanukitoid zones throughout the Province (Figure 1). The western zone (~2718 Ma) is on average ~20 Ma younger than the eastern zone (~2740 Ma).

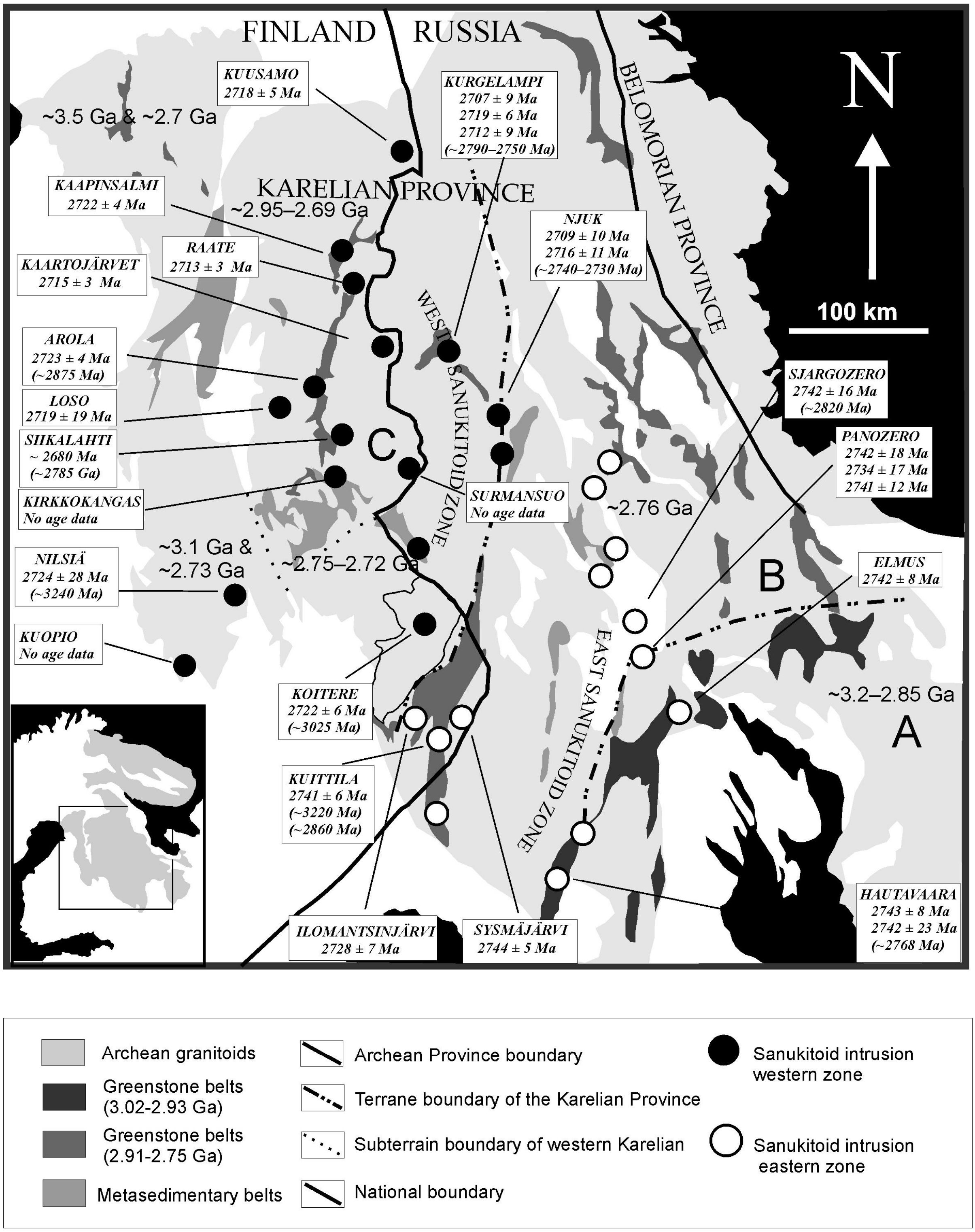

Figure 1. Compiled map of the sanukitoid intrusions in the the Karelian Province (Lobach-Zhuchenko et al., 2005; Heilimo et al., 2010; Mikkola unpublished data), and U–Pb ages and inherited cores. Ages in the Russian side after Bibikova et al., (2005), Finnish side after Heilimo et al., (in revision) except Loso (Kontinen et al., 2007), Siikalahti (Käpyaho et al., 2006), Raate (Mikkola et al., in revision). Symbols for the Karelian terranes: A — the Vodlozero terrane, B — the Central Karelian terrane, and C — the Western Karelian terrane.

3. Isotope geology

The available whole-rock Sm-Nd data from sanukitoids measured at the Laboratory for Isotope Geology at GTK with thermal ionization multiple collector mass spectrometer (TIMS), show generally positive initial εNd values between +0.1 – +2.0. The only exceptions from positive initial εNd values is from the Kaapinsalmi intrusion that show negative initial εNd values between -2.0 and -1.4. T_{DM}s for samples are between 2.75 and 2.89 Ga, excluding Kaapinsalmi (T_{DM}=2.97 – 3.08 Ga). (O'Brien et al., 1993; Halla, 2005; Käpyaho et al., 2006; Heilimo et al., unpublished data)

In situ Lu-Hf data from zircon analysed with laser ablation multicollector inductively coupled mass plasma spectrometry (LAM-ICPMS) at the University of Oslo show relatively homogenous initial $^{176}Hf/^{177}Hf$ ratios (0.28095 – 0.28110; $\varepsilon_{Hf} \approx$ -4 – +2.5). Zircon grains from the Kaapinsalmi intrusion have marginally less radiogenic composition in initial $^{176}Hf/^{177}Hf$ ratios (0.28089 – 0.28101; $\varepsilon_{Hf} \approx$ -5.8 – -0.9).

Surprisingly the SIMS *in situ* oxygen isotope analyses of zircons from sanukitoids are scattered and yield values from 5.02 ± 0.26 ‰ to 8.58 ± 0.29 ‰ (Heilimo et al, unpublished data; Mikkola et al., in revision). These results are not in line with the interpretation that the sanukitoids always show higher $\delta^{18}O$ values compared to TTG (King et al., 1998). The results might be interpreted to as a result of complex petrogenetic processes that formed the sanukitoids in Karelia.

4. Petrogenetic remarks

The petrogenesis of sanukitoids is usually explained by a two-stage process. Firstly, fluids and/or melts from a subducting oceanic slab and possibly sediments enriched the mantle wedge peridotite. Secondly, the enriched (metasomatized) mantle wedge was partially melted, melting was triggered by a slab break off at the end of subduction. The composition and sudden appearance of sanukitoids near the Archean-Proterozoic boundary indicate a distinct plate dynamic process. Our Sm-Nd and Lu-Hf data favour well mixed homogenized source, while compositional and O-isotope data indicate more complex co-genetic type process where contribution of different components (e.g. slab melts and sediments) varies from intrusion to intrusion.

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Seismic imaging of deep massive sulphides in Pyhäsalmi, Finland

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Reflection seismic data is used in help of the deep exploration of the massive sulphides in Pyhäsalmi area. Network of six seismic profiles was acquired in 2007 as part of the HIRE-project of Geological survey of Finland. The seismic data reveals a complicated reflectivity of subsurface caused by the complex geology of the area. The known Pyhäsalmi ore deposit is not directly imaged in the reflection seismic data but the strongly reflective structure hosting it can be followed throughout the region. The acoustic logging shows, that reflections are caused in contacts between mafic volcanites and other common rocks of the area. The average acoustic properties of the ore differ dramatically of those of the hosting rocks. The interpretation of the seismic data will be used for 4D-modelling of the Pyhäsalmi in ProMine-project by European Union.

Keywords: Pyhäsalmi, seismic exploration, massive sulphides

1. Introduction

The giant (>70Mt) Pyhäsalmi Zn-Cu ore deposit in Central Finland is a Volcanogenic massive sulfide (VHMS) that has been actively mined since 1962, and expected mine life extends to 2018. The architecture and structure of the deposit indicate a substantial ore potential beyond known reserves which motivates active exploration using a combination of deep penetrating geophysical techniques and drilling. The Pyhäsalmi deposit is enclosed within the rocks of contrasting acoustic properties, such as mica-cordierite schists and felsic and mafic volcanics. Contacts between these rock units are likely detected with reflection seismic method.

HIRE (High Resoution Reflection Seismics for ore Exploration 2007-2010) is a project led by the Geological Survey of Finland and participated by 11 industrial partners. Pyhäsalmi is one of the 16 targets of the reflection survey. Pyhäsalmi mine is also part of the ProMine project of the European Union with aims to build a 4D model of the mineralized region. The HIRE seismic data will also be incorporated in this model.

2. Acoustic logging

Astrock conducted drill hole logging measurements in Pyhäsalmi mine area during the spring 2010. PYS133 is 1,3 km long hole penetrating mainly felsic and mafic volcanites. About 500 m long R2229 is drilled from mine level 1.2 km. From short hole R2285 the acoustic properties of the Pyhäsalmi massive suphide itself were derived. Sonic and density logging results show that, in general, the contacts between mafic volcanites and any other rocktypes cause detectable reflections as difference in acoustic impedance is more than 2.5 x 10^5 kg/m^2s (e.g. Salisbury et al., 2003). The density and seismic P-wave velocity values measured from ore differ dramatically from any other measured values and thus the massive sulphides of Pyhäsalmi should be seen as bright reflectors or diffractors at reflection seismic data.

3. Seismic data

The reflection seismic survey in Pyhäsalmi contained a network of 2D seismic reflection lines with the total length of almost 30 km. The shot and receiver group intervals were 50 m and 12.5 m, respectively. The aims of the Pyhäsalmi survey are to gain understanding about geological structures at depth, to study the seismic signal produced by the known ore deposit, to better understand the structural control of the ore deposit and to provide new drilling targets for exploration.

The commercially processed seismic data of the HIRE-survey revealed a complicated reflectivity patterns caused by the complex geological structure of the area. The main feature of the area is a strongly reflective 1-2 km thick structure which dips towards E-NE in the eastern part of the study area, but which is subhorizontal in the western part. The deep ore body lies at an interface between mica schist and unaltered rocks on top of this large-scale structure but the ore body is not clearly imaged in current seismic profiles. Due to strong deformation, lithological contacts close to surface are mostly subvertical causing difficulties to reflection seismic technique better suited for imaging the horizontal and moderately dipping structures. However, recently the deep drill holes (>1000 m) have penetrated also flat laying structures. Adjustment and validation of the seismic interpretation will be aided by new drill hole data, geological modelling, and the results of the full waveform sonic and gamma-gamma density logging obtained from exploration drillholes.

Although the commercially processed data are of good quality, reprocessing using unconventional techniques and careful parameter selection tailored for Pyhäsalmi case may improve the resolution and imaging power of the seismic data. Reprocessing of the seismic data aims at (i) gaining more information about structures and lithological contacts from shot gathers, (ii) enhancing the stack quality especially in the shallow part of the section, (iii) imaging better steeply dipping structures and (iiii) making an improved velocity model that could be utilized in the interpretation of the data

4. Conclusions

The reflection seismic method is a powerful tool for regional scale mapping of subsurface structures. Even though the direct detection of the ore bodies is in limits of the resolution of the method, the information about the subsurface is valuable for both exploration and understanding of the geological structures and evolution of the area. In Pyhäsalmi, the seismic data together with other geophysical data is used to obtain a geologically and geophysically robust 3D model of the area with indications of promising exploration targets.

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Age and Sm-Nd isotopes on the Archean greenstone belts in Finland

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Reliable concordant U-Pb data obtained recently for volcanic and sedimentary rocks in the greenstone belts in Finland indicate distinct age groups for each belt: Suomussalmi 2.82, 2.87, 2.94 Ga; Kuhmo 2.75, 2.80, 2.82, 2.84 Ga, Tipasjärvi 2.75, 2.80, 2.83 Ga, Ilomantsi-Kovero 2.75, 2.88 Ga, Oijärvi 2.80, 2.82 Ga. Sm-Nd data show that rocks in the Kuhmo and Tipasjärvi belts represents largely newly mantle-derived material whereas in Suomussalmi belt the crustal residence ages are significantly older (>3 Ga). Altogether the isotope results suggest that the greenstone belts store a long-lived (>190 Ma) fragmentary record of geological evolution.

Keywords: greenstone belts, Archean, Finland, age, U-Pb, Sm-Nd isotopes

1. Introduction

The age and other characteristics of the greenstone belts, especially their relationship to the surrounding granitoids, are essential in modelling the evolution of the Archean lithosphere. From the Finnish Archean multi-grain zircon U-Pb TIMS analyses have been available for decades (Wetherill et al 1962, Hyppönen 1983). Although often imprecise these data have provided useful constraints for the geological evolution. Precise multigrain dating has often been difficult because of inherited zircon and/or effects of metamorphism. A frequent problem is also that suitable material for dating is scarce, and also identification of the protolith can be difficult due to thorough reworking. The Sm-Nd method has turned out to be even more problematic in the dating of the Archean greenstone belts. Primary igneous minerals are not available, and the very strict assumptions required for whole rock dating are usually difficult to verify.

Recently, a number of new data have been obtained using SIMS, LA-MCICPMS and TIMS. The results quoted here give a new, much improved basis for constraining the Archean evolution in Finland.

2. Geochronology

Volcanic rocks providing U-Pb ages in two groups at 2.94 Ga (samples A1467, A1593) and 2.82 Ga (A1428) occur in the main *Suomussalmi Greenstone belt*, but the relative proportion of these units remains unclear. The same age groups are also found for granitoid rocks west from the greenstone belt (Mikkola et al, submitted). In the eastern Tormua branch of the Suomussalmi belt, mafic rocks were formed at 2.87 Ga ago (A1821) and magmatic activity close to 2.82 Ga (A1429) is also evident.

From the *Kuhmo greenstone belt* new, precise U-Pb data are available on 16 samples. Igneous rocks from the Kellojärvi area yield ages close to 2800 Ma (A1346, A1377, A1560, A1771), and similar ages were also obtained from felsic volcanic rocks at Katerma (Ontojärvi, A511) and Polvilampi (A788), ca. ten km south and north from Kellojärvi, respectively. When the sample is characterised by a single age population consistent results have been obtained by different methods. For example, for the Lampela andesite (A1346, Figure 1) the U-Pb zircon ages are 2797±2 Ma (concordant CA-TIMS analyses), 2798±4 Ma (SIMS) and 2794±8 Ma (LA-MCICPMS).

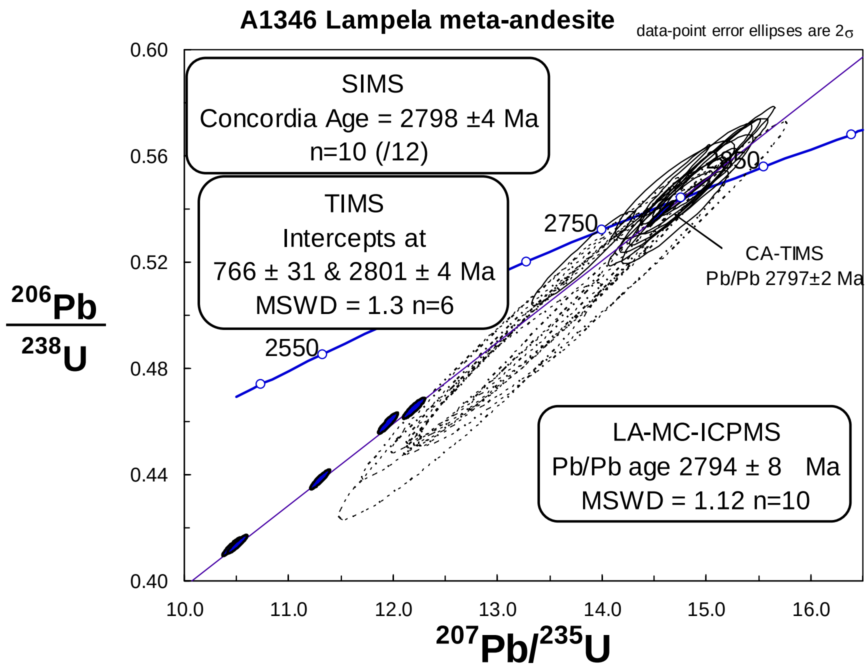

Figure 1. Concordia diagram of zircon analyses from the Lampela andesite (A1346): Relatively small error ellipse at concordia refer to SIMS data, small filled ellipse TIMS data, large error ellipse using dotted line ICPMS data.

The U-Pb zircon age for the gabbroic rock within the Kuhmo belt at Moisiovaara (A976) is close to 2.82 Ga, which is also the age for the felsic volcanic rocks at Ruokojärvi (A1000, A120) slightly east from the northern main belt. In addition to the 2.82 Ga zircon grains, sample A120 contains abundant xenocrystic zircon up to 3.13 Ga. The ages obtained for felsic volcanic rocks at Hetteilä (A1773) and Pitkäperä (A1213, A1254) at the eastern margin of the main Kuhmo belt are close to 2.84 Ga.

Four samples of felsic volcanic rocks (A1174, A1886, A1921, A1922) dated from the *Tipasjärvi greenstone belt* give ages from 2.80 to 2.83 Ga. Both Kuhmo and Tipasjärvi greenstone belts contain large areas of sedimentary rocks (Ronkaperä Fm and Kokkoniemi Fm). The main detrital zircon population in these rocks is ca. 2.75 Ga old, constraining the age of deposition significantly younger than the underlying volcanic rocks. Most granitoids surrounding the Kuhmo belt also postdate the volcanic rocks (e.g., Käpyaho et al 2006).

The U-Pb zircon age of 2754±6 Ma from the Poikapää meta-andesite (A1038) has been considered as the best estimate for the volcanism in the *Ilomantsi greenstone belt* (Vaasjoki et al 1993). The ages of the intruding granitoid plutons and dykes are also close to 2.75 Ga. Abundant inherited zircon grains up to 3.2 Ga have been found in few samples (Silvevaara granodiorite A284, Sorjonen-Ward and Claoué-Long, 1993; Vehkavaara porphyries A282, A301).

In *the Kovero greenstone belt*, SW from Ilomantsi, ages of 2.75 Ga are obtained from dykes (A1155, A1520, A1625) and gabbros (A1626), whereas zircon from two felsic volcanic rock samples (A1624, A1627) give significantly older ages at 2.88 Ga. The data (from A1627) by SIMS reveal that the zircon population is homogeneous and the old ages can thus be considered defining the age of the felsic volcanism. The relative abundance of rocks within these age groups remains unclear.

Two samples (A1782, A1783) from the *Oijärvi greenstone belt* in the Pudasjärvi complex provide U-Pb zircon ages of 2.80-2.82 Ga.

3. Sm-Nd isotopes

Sm-Nd isotopic data from ca. 400 samples provide a general overview of the formation of the Archean crust in Finland. Despite of problems related to secondary REE mobility, the Sm-Nd results have been considered to indicate that a mantle with time-integrated depletion in LREE was an important source reservoir during Neoarchean time. The data show that most Archean felsic rocks in Finland have depleted mantle model ages of ca. 2.8-3.0 Ga, suggesting together with the U-Pb zircon ages that much of the Archean crust consists of relatively juvenile Neoarchean material. Model ages in excess of 3.3 Ga are few. The rocks which have yielded the oldest reliable Sm-Nd model ages up to ca. 3.7 Ga are the 3.5 Ga Siurua gneisses in Pudasjärvi, which also are the oldest rocks in the Fennoscandian Shield.

Concerning the greenstone belts, many felsic rocks in Kuhmo, Tipasjärvi and Ilomantsi represent juvenile additions to the crust, whereas in Suomussalmi involvement of material from old LREE enriched lithosphere (crust) is significant (Figure 2).

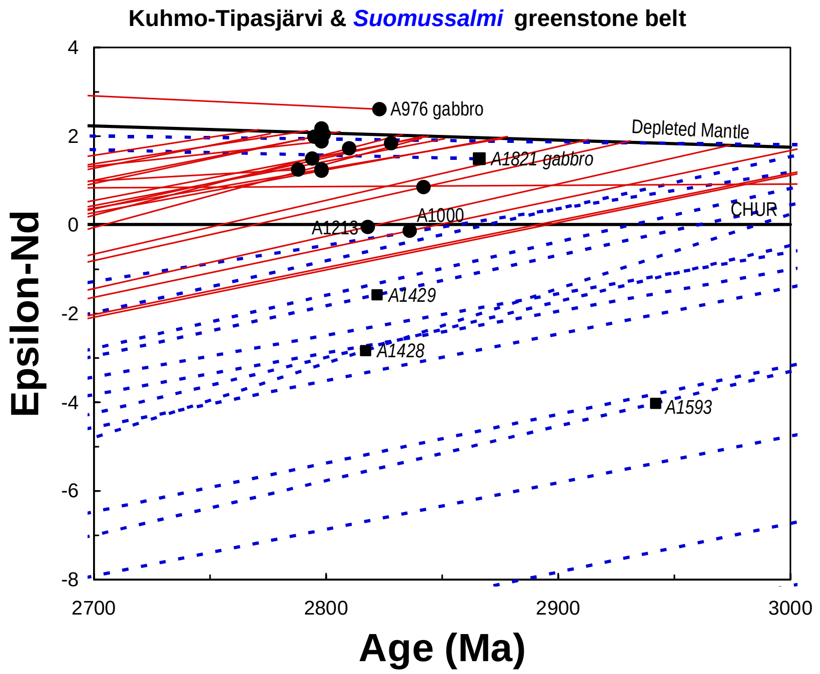

Figure 2. Epsilon-Nd vs. age diagram showing evolution lines and initial epsilon values for rocks from the Kuhmo-Tipasjärvi (circle, solid line) and Suomussalmi (square, dotted line) greenstone belts. Initial epsilon values are shown for 18 samples which have dated using U-Pb. Felsic rocks from the main Kuhmo-Tipasjärvi belt represent largely juvenile new crust, whereas significant contribution from older crustal material is evident in the Suomussalmi felsic rocks.

4. Conclusions

Precise and accurate U-Pb ages obtained for volcanic and sedimentary rocks from the main Archean greenstone belts in Finland show age grouping distinct for each belt: Suomussalmi 2.82, 2.87, 2.94 Ga; Kuhmo 2.75, 2.80, 2.82, 2.84 Ga, Tipasjärvi 2.75, 2.80, 2.83 Ga, Ilomantsi-Kovero 2.75, 2.88 Ga, Oijärvi 2.80, 2.82 Ga. These results call for revision of some traditional views of these belts (e.g. Papunen et al. 2009). In the case of the Kuhmo belt, the results show that the age of igneous rocks in the central part of the belt (Kellojärvi area) is close to 2800 Ma, which is the minimum age also for the local mafic-ultramafic magmatism, including komatiites. Mafic rocks in the Kuhmo belt represented by the Moisiovaara gabbro are ca. 2.82 Ga old, which is also the age of the Ruokojärvi felsic volcanic rocks and a significant age group also in the Suomussalmi belt. Plentiful Sm-Nd data show, that much of the crust in the Kuhmo belt represents newly mantle-derived material whereas crustal residence ages in Suomussalmi belt are significantly older, mostly exceeding 3 Ga. Altogether the isotope results suggest the greenstone belts record long-lived and spatially over short distances varied record of geological evolution. Interpretation and modelling of the tectonic framework of this evolution remains a major challenge.

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Velocity blocks and anisotropy in the crust of the central Fennoscandian Shield

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The central Fennoscandian Shield has a complex structure. After several tectonic events, forming up island arcs, progressive thrusting and extensional phases, the crust is composed of varying size blocks and slanting belts. These blocks and belts are visualized by seismic tomography method. The tomography data comprises of P-wave and S-wave arrival data from controlled source and seismic tomography experiments, and from chemical explosions. The inverted seismic velocities vary locally in the whole crust expressing several distinct anomalies. The boundary zone between the Archean and the Proterozoic terranes of the Shield appears as an upper crustal low velocity zone. Lower velocities (Vp/Vs<1.70) are associated with schist belts, and higher velocities (Vp/Vs >1.76) with rapakivi batholiths and with the Central Finland Granitoid Complex suggesting hidden mafic blocks in the lower crust.

In the isotropic tomographic velocity models of the crust, anisotropic component is embedded in the residual component. Horizontal azimuthal dependency of residual component images the horizontal component of tectonic transport. An azimuthal anisotropy in tomographic model is verified when fast P-wave velocity direction is matched with slow one in the orthogonal direction. The resulting anisotropy directions coincide with transport directions drawn from structural observations.

Keywords: lithosphere, crust, velocity tomography, anisotropy, Fennoscandia

1. Velocity blocks

The tectonic history of the central Fennoscandian Shield contains various recurrent events, i.e. thrusting, extension and building up island arcs. Consequently, the crust has complex structure composed of varying size blocks and slanting belts. These blocks and belts are visualized by seismic tomography models.

The tomography inversion uses P-wave and S-wave arrival data from controlled source refraction and reflection experiments, from a passive seismic tomography experiment, and from chemical explosions. Addition of the data from the BABEL sea reflection lines and land stations enabled to include the Gulf of Bothnia to the study area. The number of receivers increased to 2895 and seismic sources to 565 giving 19180 first P-wave and 15146 S-wave crustal travel times. The resulting rays covered larger area and revealed smaller-scale structural blocks than the previous tomography model by Hyvönen et al. (2007; Korja et al. 2009).

The distribution of the P- and S-wave velocities, and the velocity ratio, Vp/Vs, varies locally in the whole crust to the depth of 40 km. The anomalous velocity behavior expresses several distinct blocks, belts, and zones which can be associated with the main geological units. The border zone between the Archean and the Proterozoic terranes of the Shield is expressed as a chain of low velocity segments in the upper crust. The schist belts are associated with velocity minima, Vp/Vs <1.70, whereas higher velocities, Vp/Vs >1.76, characterize granitoid formations such as Vyborg and Laitila rapakivi areas and the Central Finland Granitoid Complex suggesting hidden mafic blocks in the lower crust.

2. Anisotropy component

In the isotropic tomographic velocity models of the crust anisotropic component is embedded in the residual component. Horizontal azimuthal dependency of residual component images the horizontal component of tectonic transport (Figure 1). The plate tectonic movement makes the crustal blocks move more laterally than vertically. Thus, in many geological problems the direction of the horizontal component is a satisfying approximation. This assumption is even more valid with growing temperatures at deeper levels (>15 km). An azimuthal anisotropy in the tomographic model by Tiira and others (2010a, 2010b) is verified when fast P-wave velocity direction is matched with slow one in the orthogonal direction. The resulting anisotropy directions coincide with transport directions drawn from structural observations.

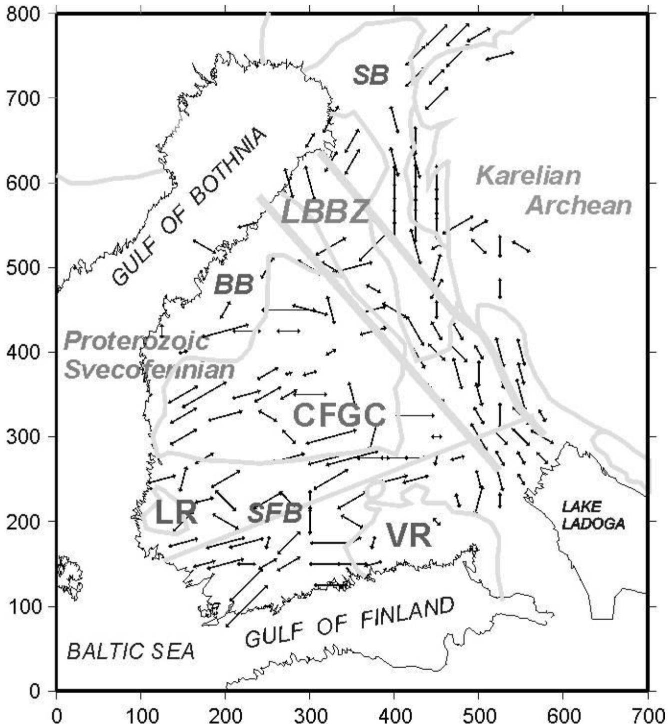

Figure 1. The fast P-wave azimuth directions for the middle crust (at depths of 15-30 km) in south-central Finland are plotted with the main geological units [Vyborg rapakivi (VR), Laitila rapakivi (LR), Central Finland Granitoid Complex (CFGC), Ladoga-Bothian Bay Zone (LBBZ), southern Finland belt (SFB), Bothnian belt (BB), and Savo belt (SB)].

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Neoarchaean evolution of the Western Karelian Province

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The Archaean of Fennoscandia is traditionally divided into Murmansk, Kola, Belomorian and Karelian Provinces (Fig. 1, Slabunov et al., 2006; Hölttä et al., 2008). Large parts of these provinces were reworked in the Paleoproterozoic Svecofennian and Lapland–Kola collisional orogens (Kontinen et al., 1992; Daly et al., 2006; Bibikova et al. 2001). Lobach-Zhuchenko et al. (2005) and Slabunov et al. (2006) divided the Karelian Province into three terranes that have many lithological, structural and age differences. These are the Vodlozero terrane, the Central Karelian terrane and the Western Karelian terrane. Mesoarchaean 2.8-3.2 Ga rocks are common in the Vodlozero and Western Karelian terranes, whereas on the basis of existing age determinations both granitoids and greenstones in the Central Karelian terrane are Neoarchaean, c. 2.76-2.72 Ga, although some of them contain reworked Mesoarchaean crustal material. The Archaean of the Fennoscandian Shield mainly consists of rocks that belong to the TTG association (covering about 80% of the area), with greenstone belts, paragneiss areas and migmatitic amphibolites.

The Western Karelian terrane comprises much of eastern Finland and the westernmost part of Russian Karelia. Migmatitic TTG orthogneisses and amphibolites are the most common rock types in the area, with small medium to low pressure granulite areas being present in the western part of the terrane. Crustal architecture inferred from seismic data, as well as bedrock structural data, suggest thrust stacking and transpression during late orogenic deformation around 2.70 Ga. Tectonic transport was towards NE and SE, with evidence that the Western Karelian terrane was emplaced eastwards over the Central Karelian terrane (Kontinen & Paavola, 2006).

In the Western Karelian terrane rocks whose zircon ages are > 3.0 Ga are rare, existing only from the Iisalmi (3.2-3.1 Ga) and Pudasjärvi (3.5 Ga) terranes (Mänttäri & Hölttä, 2002; Mutanen & Huhma, 2003). The oldest volcanic rocks, 2.96-2.94 Ga in age, are found in the northern part of the Suomussalmi greenstone belt (Luukkonen et al., 2002), and some nearby TTGs and migmatites are of the same age (Käpyaho et al., 2007; Mikkola et al., in review). Existing data indicate that melting of basaltic protoliths at various depths, either in thickened lower crust or subduction setting, followed by Neoarchean collision and remelting of the crust explains most of the observed features of the Archean bedrock.

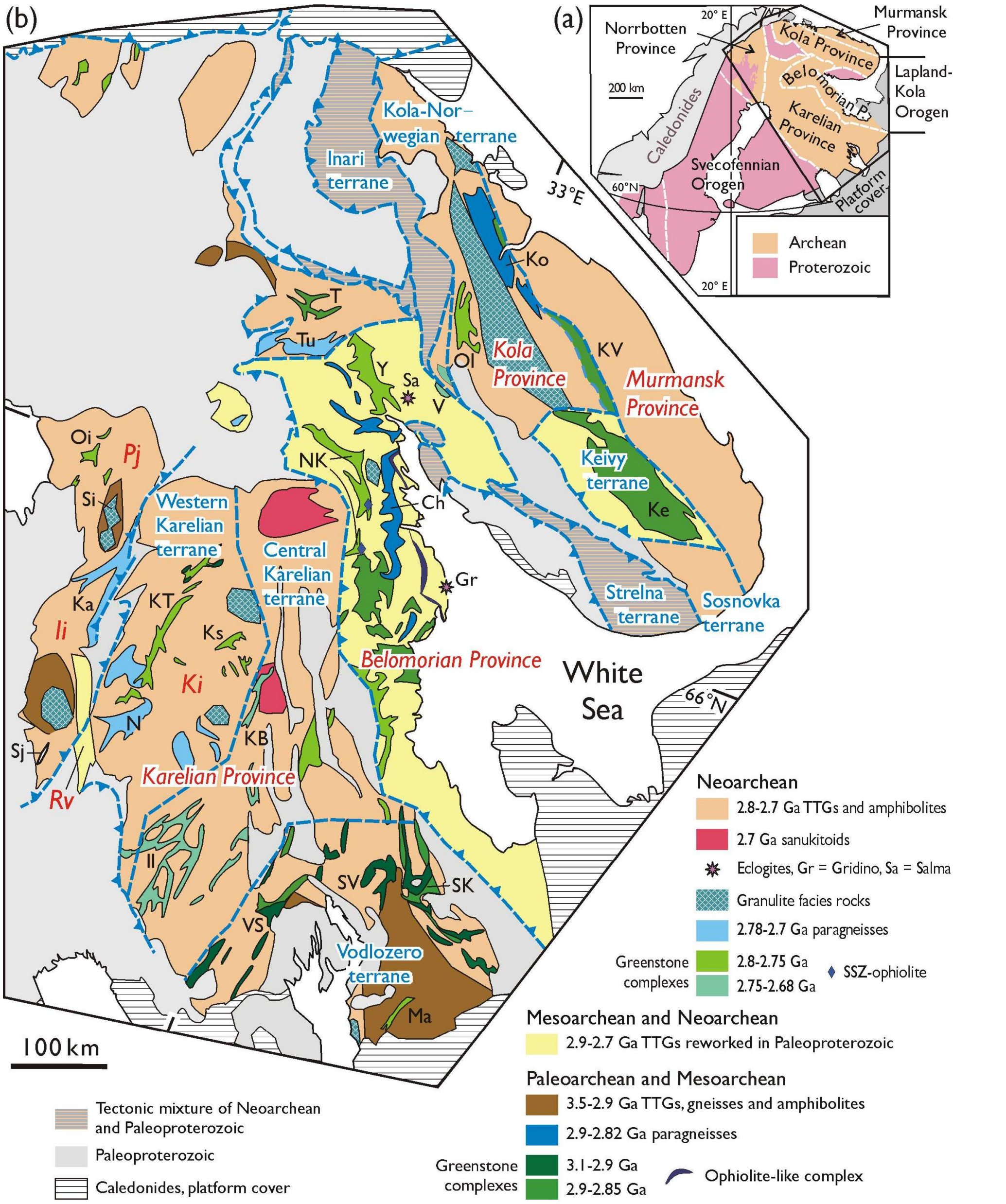

Figure 1. Generalised geological map of the Archaean of Fennoscandia (Hölttä et al., 2008).

However, most of the dated volcanic rocks in the Tipasjärvi, Kuhmo, Suomussalmi and Oijärvi greenstone belts yield ages of 2.84-2.75 Ga. Mesosomes of migmatites and TTG orthogneisses outside the greenstone belts give similar ages (Hyppönen, 1983; Luukkonen 1985; Vaasjoki et al., 1999; Käpyaho et al., 2006, 2007; Lauri et al., 2006; Lauri et al., 2010; Mikkola et al., in review), but display a distinct pause between 2.79 and 2.76 Ga. Komatiites of all greenstone belts are Al-undepleted or "Munro-type", and geochemistry of tholeiitic basalts indicate rather plume-type than non-plume-type sources. However, in the Ilomantsi greenstone belt there are high abundances of greywackes and 2.74-2.72 Ga island arc-type basalt-andesite-dacite-rhyolite series volcanic rocks that may indicate some kind of plume-arc interaction (Sorjonen-Ward, 1993; Vaasjoki et al., 1993; Puchtel et al., 1998).

The ion probe and TIMS U-Pb age determinations on zircon grains from mesosomes of paragneisses from various parts of the Western Karelian terrane constrain the deposition of the protolith wackes to c. 2.71 - 2.69 Ga. Trace element and U–Pb data from the Nurmes paragneisses suggest that the source terrains comprised mainly 2.75–2.70 Ga TTG and/or sanukitoid-type plutonic and mafic volcanic rocks. The presence of MORB-type volcanic intercalations in Nurmes wackes suggests they were deposited in a back- or intra-arc setting (Kontinen et al., 2007).

Archean granitoids form four major groups that are TTG (tonalite-trondhjemite-granodiorite), sanukitoids, QQ (quartz diorite-quartz monzonite) and GGM (granodiorite-granite-monzonite) groups. Furthermore the TTG group can be subdivided into four groups (low HREE TTG, low HREE TTG with positive Eu/Eu* and high HREE TTG) based on the REE patterns slope (Halla et al., 2009). The low HREE group has high SiO_2, low Mg, low HREE and high Sr/Y and La_N/Yb_N ratios that suggest garnet-bearing basaltic source and melting in high pressure. Rocks belonging to the second group have positive Eu anomalies, strongly fractionated REE patterns with HREE mostly below detection limits, high Sr/Y, La_N/Yb_N and Zr/Sm ratios and low abundances of compatible elements and Fe and Mg. These geochemical features probably indicate both deep melting and plagioclase fractionation. The high HREE group has relatively low SiO_2, elevated Mg, low Sr/Y and La_N/Yb_N ratios and high HREE indicating garnet-free source. Fourth TTG group is the transitional TTGs, which is present at least in northern part of the Kianta complex. This group has K_2O/Na_2O>0.5 and higher LILE concentrations than other TTG groups and can not therefore regarded as TTGs sensu stricto. The compositional features suggest at least partial derivation from pre-existing felsic crust (Mikkola et al., in review). Geochemical changes between these three groups are gradual and the groups cannot be distinguished on the basis of either Sm-Nd model ages or Ub-Pb ages on zircon. It seems that apart from the melting pressure the REE fractionation is controlled by the chemical composition of the basaltic protolith that has a strong influence on the abundance of garnet in the restite.

The second granitoid group consists of high-Ba-Sr sanukitoids that are generally younger (2.74.- 2.72 Ga) than other TTGs. Their low SiO_2 and high K_2O and MgO contents suggest an origin by melting in an enriched mantle source (Bibikova et al., 2005; Halla, 2005; Lobach-Zhuchenko et al., 2005; Käpyaho et al., 2006; Heilimo et al., 2010).

The QQ group is 2.71-2.70 Ga in age. Rocks of this group have lower SiO_2 content than TTGs, HREE is relatively high and consequently Sr/Y and La_N/Yb_N ratios lower than in most TTG groups. The Ranua diorite, Iisalmi enderbites and some smaller intrusions in the western part of the Karelian province belong to this group. Unlike sanukitoids QQ rocks are not enriched in LIL elements, Ni and Cr. They probably represent garnet free melting in relatively low pressure in lower crust or upper mantle.

Granite-granodiorite-monzonite (GGM) series rocks that are dated at c. 2.71-2.66 Ga are the youngest Neoarchean rocks occurring in large volumes. Granites are mostly magnesian, calc-alkalic and alkali-calcic. Most granites have highly fractionated REE patterns with negative or no Eu anomaly but many of them are less fractionated with high HREE. Some have low REE abundances with positive Eu anomaly, resembling Eu positive TTGs. Granites could be the melting products of both sedimentary gneisses and sodic TTGs. The ε_{Nd} (2700 Ma) of granites is generally c. -4. 0 c. -1.5, but values as high as +1.0 exist, indicating that some of the granites might represent melting of juvenile 2.72-2.78 TTGs whereas others could originate from sediments that contain older detrital material.

Migmatitic amphibolites occur as interlayers in the TTG complexes, their regional abundance and degree of migmatisation being very variable. They fall into two major geochemical groups (Hölttä, 1997; Nehring et al., 2009). Basaltic rocks belonging to the first group have flat or HREE depleted trace element patterns, resembling those of mid-ocean ridge basalts. In the second group basaltic rocks are enriched in LREE and LIL elements. They have a negative trough in Nb, like island arc basalts. Compatible elements, especially Ni but also Cr are low in LREE enriched group. Andesitic amphibolites have similar trace element contents with LREE enriched basaltic amphibolites, indicating that they belong to the same magma system, which may represent island arc magmatism or dykes that intruded into TTGs and were contaminated with the crustal material. On the basis of field relations some rocks classified as TTGs are highly melted diatexitic amphibolites where the percentage of neosome may be up to 90 %, representing therefore crustal melting.

The Western Karelian terrane was metamorphosed in upper amphibolite facies and granulite facies conditions. Exceptions are the inner parts of greenstone belts which still often have mid or lower amphibolite facies mineral assemblages, well preserved primary structures and only a little or no migmatisation. Medium pressure granulites, metamorphosed at c. 9-11 kbars and 800-900^{o}C are found only in the Iisalmi block (Hölttä & Paavola, 2000). In the Siurua area there are mafic granulites whose metamorphic pressures and temperatures were c. 5-6 kbars and 700-800^{o}C (Lalli, 2002). On the basis of U-Pb ages on titanites, monazites and zircons from granulites and migmatite leucosomes this took place at 2.71-2.62 Ga, coevally with the emplacement of the GGM granitoids. Also the QQ rocks underwent granulite facies metamorphism, at least in the Iisalmi area (Hölttä et al., 2000; Mänttäri & Hölttä, 2002; Mutanen & Huhma, 2003; Käpyaho et al., 2006; Lauri et al., 2006; Käpyaho et al., 2007; Lauri et al., 2010). In the Belomorian Province there are 2.72 Ga eclogites that were metamorphosed at 14-17 kilobars occur (Volodichev et al. 2004). They have not been found in the Karelian Province, but near the eastern border of the Kuhmo greenstone belt TTGs have interlayers of garnet-clinopyroxene amphibolites that lack plagioclase. This assemblage indicates high pressures; garnet in these rocks have rare albite inclusions that, when used in barometry together with pyroxene and amphibole, give pressures of 13-14 kbars. These pressures indicate thickening of the crust or they may even suggest subduction at around 2.7 Ga.

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Lithospheric structures and geometries revealed by deep-probing magnetotellurics

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This paper describes the state of knowledge of the continental upper mantle obtained primarily from the natural source magnetotelluric technique, and outlines how hypotheses and models regarding the development of cratonic lithosphere can be tested using deep-probing electromagnetic surveying. Examples of data and interpretations from various regions around the globe are discussed to demonstrate the correlation of electromagnetic and seismic observations through petro-physical arguments and of the two main boundaries – the Moho and the Lithosphere-Asthenosphere boundary. Particular attention is paid to electrical anisotropy and the information that those observations yields.

Keywords: lithosphere, crust, upper mantle

1. Introduction

The plate tectonic paradigm is a remarkably successful model in describing the Earth's dominant tectonic process - the movement of Earth's tectonic plates around the globe - in a kinematic sense. These plates comprise the crustal and mantle lithosphere, and are riding on top of the weaker and more "fluid–like" (on geological time scales) asthenosphere. What is lacking though in the model is complete knowledge and understanding of dynamic processes occurring within the Earth's mantle, particularly its upper mantle, which account for how the plates move and particularly how they interact with each other. On a fundamental level, we do not know how lithosphere is formed, deformed, or consumed. We do not know the nature of the coupling between the lithosphere and the underlying asthenosphere at their boundary – the Lithosphere-Asthenosphere Boundary (LAB). We do not know how the topology of the LAB shapes and focuses stresses on the base of the lithosphere.

To understand how these processes operate, we must determine and understand the nature of the lithosphere and asthenosphere, and particularly the transition between the two – the LAB, beneath ancient cores of cratons, their bounding belts, and currently-active regions at plate boundaries. The nature of the mantle – the spatial variation in chemistry, petrology and temperature; the geometry of boundaries and internal structures – provides a direct record of the dynamic processes taking place.

This paper describes the state of knowledge of the continental upper mantle obtained primarily from the natural source magnetotelluric technique, and outlines how hypotheses and models regarding the development of cratonic lithosphere can be tested using deep-probing electromagnetic surveying. The resolution properties of the method show the difficulties that can be encountered if there is conducting material in the crust. Examples of data and interpretations from various regions around the globe are discussed to demonstrate the correlation of electromagnetic and seismic observations through petro-physical arguments and of the two main boundaries – the Moho and the Lithosphere-Asthenosphere boundary. Particular attention is paid to electrical anisotropy and the information that those observations yields.

2. Imaging the electrical Lithosphere-Asthenosphere Boundary - eLAB

The nature of the Lithosphere-Asthenosphere Boundary (LAB) is poorly understood, notwithstanding the fact that it is the most pervasive plate tectonic boundary on the planet underlying both oceanic and continental regions. Although the LAB is an active plate boundary, beneath the continents it is relatively cryptic compared to other first-order structural subdivisions of Earth. Specimens of the LAB and environs, in the form of xenoliths and xenocrysts brought to the surface in alkaline magmas such as kimberlite, are sparsely and unevenly distributed and, as argued below, virtually nonexistent beneath cratons.

Furthermore, since the differential motion between lithosphere and asthenosphere is generally accommodated aseismically, there is no seismological technique to map the LAB directly. In addition, global gravity observations show that the lithosphere is approximately in large-scale isostatic equilibrium (Shapiro et al., 1999), rendering long-wavelength gravity inversion ineffective as a mapping tool. These limitations have spawned many indirect proxies for the LAB (Eaton et al., 2009) that depend, in a practical sense, on the type of measurement made. These proxies are derived from disparate types of observations, including petrologic, geochemical, thermal, seismic-velocity, seismic-anisotropy and electrical-conductivity characteristics of the LAB and regions directly above and below it.

The easiest lithospheric target for MT studies is the depth to the base of resistive lithosphere at the transition to conductive asthenospheric-mantle across the LAB, where at least an order of magnitude change in conductivity occurs (Jones, 1999). Although the long-held interpretation of partial melt as the cause of low resistivities in the asthenosphere beneath the continents is being challenged by the possible role of dissolved hydrogen (water) (Hirth et al., 2000), both of these mechanisms dramatically enhance bulk electrical conductivity in the asthenosphere with respect to the overlying lithosphere. Laboratory studies demonstrate that at the onset of partial melting, even at very low-order melt fractions (sub 1% fractions), melts are well interconnected, resulting in conductivity enhancements of around two orders of magnitude (see Jones, 1999). Both the LAB and regions of melt generation and accumulation are resolvable targets in good quality MT data.

In addition, work pioneered on the Southern African Magnetotelluric Experiment (SAMTEX) has shown that electrical anisotropy can also be used as a distinctive marker of the transition into the asthenosphere (Hamilton, 2008; Hamilton et al., 2006), in a similar manner to the method of Plomerova and Babuska using teleseismic arrivals (Plomerova and Babuska, 2010; Plomerova et al., 2002). Electrical anisotropy through lattice preferred orientation of olivine in the mantle may potentially provide evidence of strain localization in the early stages of rifting. Recent comparisons of seismically-determined and electrically-determined depths to the LAB for Europe (Jones et al., 2010) demonstrate that in places they are in agreement, but in other places they are dissimilar, and the differences provide significant insight into the nature of the lowermost lithosphere.

3. Imaging the electrical crust-mantle boundary – eMoho

The transition from the crust to the mantle is characterized petrologically in terms of mafic to ultra-mafic assemblages, and seismologically by a velocity increase from around 6.8-7.2 km/s to 7.8-8.2 km/s. Laboratory studies on mafic and ultramafic rocks suggest that there should be an <u>increase</u> in electrical resistivity at the petrological crust-mantle transition (Haak, 1982). Controlled-source ocean studies demonstrate that oceanic uppermost mantle is as highly resistive as laboratory studies suggest (Constable and Cox, 1996).

However, the almost ubiquitous conducting lower crust in the continents (Jones, 1992; Jones, 1998) makes resolution of the resistivity of the uppermost mantle impossible (Jones, 1999). Only in places where the lower crust is resistive do we have a window on the uppermost mantle resistivity. The first definitive resolution of uppermost mantle resistivity was by Jones and Ferguson (2001) for a region of the Slave Archean craton in northwestern Canada. The region displays highly resistive lower crust, of some 40,000 Ω.m, beneath which is a less resistive uppermost mantle, of some 4,000 Ω.m. Thus we see perplexingly a resistivity decrease at the electrical Moho, or eMoho.

Resistivities of a few thousand ohm.m are difficult to explain – they are too high to use arguments about interconnected graphite but some conducting phase is required otherwise the resistivity should be some tens to a hundred thousand ohm.m.

Other observations of an eMoho have been inferred in various regions of the globe, including in the Rae craton of Canada (Jones et al., 2002) where the uppermost Rae mantle is highly resistive (>65,000 Ω.m), and beneath the Eastern Indian Craton (Bhattacharya and Shalivahan, 2002) where again there is a decrease from some 30,000 Ω.m in the lower crust to 8,500 Ω.m in the uppermost mantle. Most recently, Moorkamp and Jones (Moorkamp et al., 2007; Moorkamp et al., 2010) have used stochastic modelling (genetic algorithm) to search for models that were self-consistent between MT and seismic (Receiver Functions and Surface Waves) data, and were able to find models for the Slave and Kaapvaal cratons that showed eMohos.

4. Structures within the lithosphere

A number of MT studies have imaging structures within the lithospheric mantle that have geometries indicative of tectonic processes. One of the most comprehensive used the conductivity structure imaged in the centre of the Slave craton, northern Canada (Jones et al., 2001; Jones et al., 2003) to describe a tectonic history and scenario for the craton that incorporated MT, seismic, petrological, geochemical and geological data (Davis et al., 2003).

5. Lithospheric anisotropy

Rock fabrics induced by tectonic processes, particularly, but not exclusively, by lateral plate tectonic translations, provide important clues about petrogenesis and the deformation history of the region. Unfortunately, as discussed above our direct knowledge of subcontinental lithospheric fabrics is severely limited by the scarcity and bias of mantle samples. This leaves a significant observational gap in our understanding of dynamics of tectonic processes – in particular, how continents formed and interacted with underlying mantle regions in the past, and how they do so today. This gap can be filled by appropriate geophysical observations of lithospheric anisotropy.

Continental seismic anisotropy has been studied for two decades, with shear wave splitting analyses of core-traversing teleseismic waves (SKS arrivals) providing the most comprehensive observational dataset (Park and Levin, 2002; Savage, 1999; Silver, 1996). More recently, seismic anisotropy deduced from surface-wave studies and receiver functions are also providing important insights (e.g., Gung et al., 2003). Seismic anisotropy in the lithospheric and sublithospheric mantle is readily explained in terms of aligned olivine crystals. However, given our understanding of the likely processes of lithospheric mantle formation, not only today but also in the past (particularly the Archean), that interpretation does test the bounds of credulity, and alternative explanations are urgently needed.

Electromagnetic observations of long-period magnetotelluric (MT) signals provide the best available method to measure electrical anisotropy of the mantle. However, in contrast to seismic anisotropy, electrical anisotropy interpreted from MT observations, from Mareschal et al.’s (1995) key contribution to more recent studies (Gatzemeier and Moorkamp, 2005; Simpson, 2002) is often significantly higher than expected from intrinsic crystal anisotropy.

Without a ready explanation the origin of electrical anisotropy remains controversial; processes that have been proposed to explain it include melt lenses, conductive films along grain boundaries, and anisotropic diffusion of hydrogen, but none are without considerable objection.

Both SKS and MT methods suffer from significant inherent weaknesses. SKS has no intrinsic depth resolution; the results are generally interpreted in terms of either fossil structures in the lithosphere representative of past formation and deformation processes (the "Silver" school of thought, championed by Paul Silver; Silver and Chan, 1988), or present-day structures in the lowermost lithosphere and underlying asthenosphere resulting from mantle flow (the "Vinnik" school of thought, of Lev Vinnik; Vinnik et al., 1992). A raging debate is occurring between these two schools, often without recognition that both mechanisms are likely at play but to differing degrees at differing depths, and progress on understanding the observations is hindered by this lack of depth resolution.

Equally, MT interpretations are beset by intrinsic limitations. Jones (2006) draws attention to the consequences of the vast range of electrical conductivity in rocks, and that attenuation in anisotropic crustal conductors may severely limit mantle penetration. Crustal attenuation and anisotropy effects in Fennoscandia (Lahti et al., 2005) call into question the mantle paleoflow interpretation of Bahr and Simpson (2002) based on their MT observations.

6. Petro-physical considerations

When the value of electrical resistivity of a rock is driven by interconnectivity of a minor phase, one should not necessarily expect there to be a relationship between seismic and electrical parameters. However, in certain circumstances where the resistivity is from the bulk properties, then one might expect there to be a definable relationship. Jones et al. (2009) review the literature and define petro-physical relationships for the moduli and electrical conductivity from an assemblage of four minerals, namely olivine, ortho- and clino-pyroxene and garnet.

A comparison between seismic velocities and log(resistivity) at 100 km beneath southern Africa demonstrates that a quadratic relationship can be defined between the two that is self-consistent for 70% to 95% of the region. The major implication for this is that seismic velocity is driven almost entirely by temperature variation and not by compositional variation.

7. Conclusions

Deep-probing magnetotellurics can add significantly to our understanding of lithospheric formation and deformation processes by defining geometries and structures within it.

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Magnetic studies as indicators for ascent and emplacement of rapakivi granites

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The present study deals with the ascent and emplacement of ~1.6-1.5 Ga rapakivi granite batholiths in Finland in context of the amalgamation of the Columbia supercontinent. The spatial and temporal variety within rapakivi granite batholiths in Fennoscandia can be interpreted by examining fault and shear zones shown on aeromagnetic and gravity maps together with petrophysical classification and investigation of the magnetic fabric of intrusions within rapakivi granite batholiths.

Keywords: rapakivi granite, magnetic susceptibility, aeromagnetics, gravity, Fennoscandian Shield, supercontinents

Globally, rapakivi granites are found in most old cratons of the world. They occur most frequently in the East European craton and North and South American platforms but also in platforms in Siberia, North China, Africa, Australia and Antarctica. Their temporal distribution set these rocks in three major time periods (Larin, 2009). The minority of rapakivi granites were formed during the Archaean, at 2.8-2.6 Ga ago but the most abundant volumes of rapakivi granites are early to middle Proterozoic (1.8 to 1.0 Ga). The youngest rapakivi granites are late Proterozoic (0.6-0.5) or even Phanerozoic 0.4-0.05 (Larin, 2009; Rämö and Haapala, 1995).

The genesis of rapakivi granites have been recently linked to the assemblage of the Columbia supercontinent (Vigneresse, 2005; Kukkonen and Lauri, 2008). Åhäll et al. (2000) link the rapakivi granites in the Svecofennian domain directly to subduction related magmatism within the Gothian orogenesis with three westward younging stages (Wiborg suite, Åland & Riga Suite and Salmi & Ragunda suite). Kukkonen and Lauri (2008) relate the rapakivi magmatism to the Svecofennian orogeny as post-orogenic due to collisional thickening and long-term warming of the crust without inputs from mantle plumes or magmatic underplatings. Vigneresse (2005) suggested that Columbia was formed by successive aggregation of smaller continents above/from a descending mantle convective cell and the heat problem is explained with the antiplume model where the mantle material moves downward but the heat flow still goes upward through the lithosphere. The rapakivi granites ascended at shear zones between juvenile crust and surrounding older cratons within the supercontinent and which then partly split through strike-slip deformation and rifting. New palaeomagnetic and isotopic data suggests that after Sarmatia collided with Fennoscandia at 1.8 Ga the Sarmatian plate still rotated ca 43º until 1.7 Ga (Elming et al., 2010). These studies indicate that the “anorogenic” rapakivi granites generated slowly and ascended in the crust as dykes and finally spread horizontally to form large and thin tabular batholiths.

Aeromagnetic and gravity data reveal structures relevant to the ascent and emplacement of rapakivi granites in Finland. These large batholiths have earlier been linked mainly by their temporal varieties. However, a new approach to distinguish and classify these rapakivi granites can be considered by their magnetic properties. Previously Dall`Agnol and Oliveira (2007) classified the classical rapakivi granites of Fennoscandia as ilmenite-bearing reduced

A-type granites in comparison with the magnetite-bearing oxidized A-type granites of Laurentia. This classification however must be revised since the magnetic susceptibility of Finnish rapakivi granites indicate that at least the majority of Åland rapakivi granites belong to magnetite series with a mean magnetic susceptibility of 10 000 µSI, while the rapakivi granites of the Wiborg batholith show lower magnetic susceptibilities (mean 1 600 µSI). Vehmaa rapakivi granite batholith on the other hand has a clear separation of both paramagnetic and ferromagnetic granites (mean 4 200 µSI).

Aeromagnetic and gravity maps reveal structures that control the ascent and internal structures relevant to the emplacement of these large rapakivi batholiths in Finland. On the magnetic map of the Fennoscandian Shield both Wiborg and Salmi rapakivi granite batholiths show low magnetic anomalies whereas Åland and Riga have higher magnetic anomalies. Anisotropy of magnetic susceptibility (AMS) studies shows the internal structures of selected intrusions. These studies indicate that the batholiths were emplaced as succession of gently dipping sheets above collapsing cauldron shaped structures (Karell et al., 2009). Together, data from AMS measurements as well as local anomalies seen on aeromagnetic and gravity maps give us new understanding to the emplacement of Finnish rapakivi granites as part of the Columbia supercontinent.

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Paleomagnetic study on Satakunta sandstone, Finland

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The objectives of the present study are to obtain more accurate paleomagnetic data from Satakunta sandstone and subsequently determine a more accurate magnetization age. Paleomagnetic measurements have been carried out on sandstone samples and a bedding tilt correction has been applied for each site. Preliminary results show that the sandstone carries three remanence components. One is clearly a viscous component due to PEF. Another component has a steep upward direction and is most likely caused by diabase baking. The third component has a lower inclination and NE declination, and occurs mainly in minerals with high coercivities. The results from the tilt-correction test are negative for both steep and shallow components, indicating that these components were acquired post tilting. This is an ongoing study and we aim to present additional results at the Lithosphere symposium 2010.

Keywords: Satakunta sandstone, paleomagnetism, Fennoscandia, tilt-correction.

1. Geological setting

The Proterozoic Satakunta sandstone is situated in the Satakunta region in SW Finland and stretches from Yläne in the SE to Reposaari in the NW and extents into the Bothian Sea. It formed as an alluvial basin 1.6 - 1.3 Ga ago in a graben (Kohonen *et al.* 1993). The exact age of formation is uncertain. The poorly exposed sandstone is cut by diabase sheets (*ca.* 1260 Ma) which baked the sandstone at most outcrop sites.

2. Previous research

A previous paleomagnetic study on Satakunta sandstone by Neuvonen (1973) yielded a mean direction of D=24; I=-41 which is close to the "post-Jotnian" diabase direction of D=35.5; I=-34 (Neuvonen 1974). Neuvonen (1973) applied single step demagnetization of 400°C which is lower than the unblocking temperatures of the magnetite and hematite carrying the magnetization of the sandstone. This led us to believe that the sandstone direction obtained by Neuvonen is in fact a secondary component resulting from baking by adjacent diabase intrusions.

3. Preliminary results

Paleomagnetic measurements were carried out on sandstone samples collected in 1980, 1996 and 1998. Step wise alternating field (AF) and thermal demagnetization was applied. Mean remanence directions of components were calculated using Fisher statistics and a bedding-tilt test was applied.

Preliminary results show that the sandstone carries three remanence components. One is a viscous component due to PEF. Another component (steep component) has a steep upward direction and is most likely caused by diabase baking. This component corresponds with the one obtained by Neuvonen (1973). The third component (shallow component) has a lower inclination and NE declination, and occurs mainly in minerals with high coercivities.

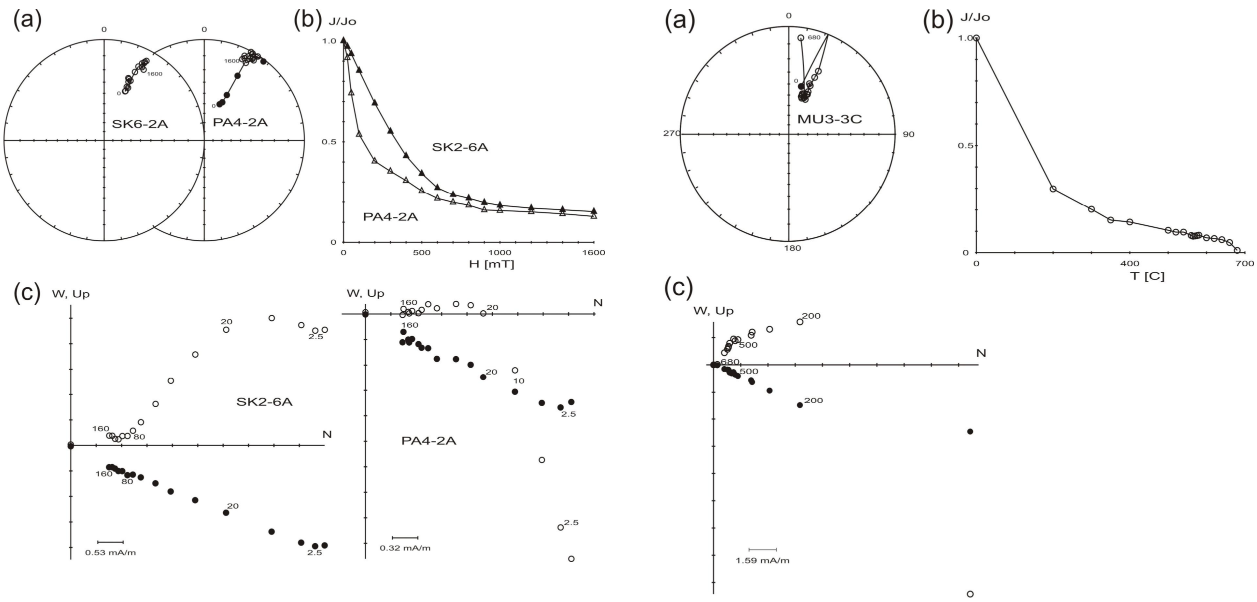

Figure 1. Examples of AF (left) and thermal (right) demagnetization behaviour of sandstone specimens**. (a)**Stereographic projections of directional data upon demagnetization, **(b)** relative intensity decay, **(c)** orthogonal (Zijderveld) demagnetization diagrams. Open and closed symbols respectively denote vertical and horizontal planes.

The statistical parameters for site means before and after tilt-correction show that the data derived from both steep and shallow components are slightly more scattered after tilt-correction, which suggests post-tilting aquisition (Figure 2). This supports our notion that the steep component is secondary, obtained from "post-Jotnian" diabase baking.

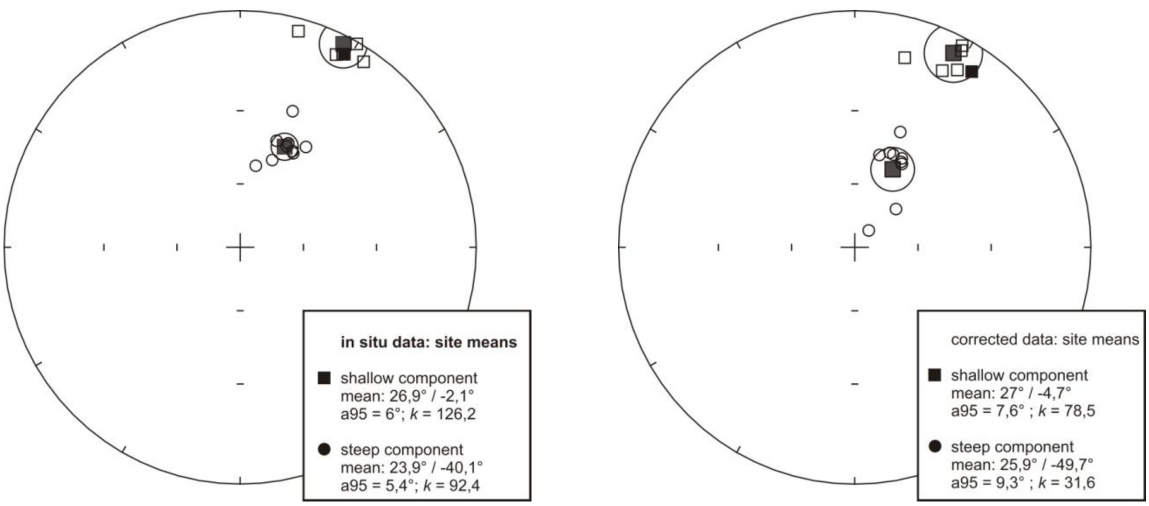

Figure 2. Mean directions of site-mean magnetic components and statistical parameters before (in-situ) and after tilt-corrections

4. Conclusions

Based on the preliminary results and the proximity of sandstone outcrops to diabase intrusions, the deep component is interpreted as a secondary component obtained through baking by diabase. The shallow component is interpreted as either the primary sandstone component or a secondary component resulting from partial baking.

At the time of abstract submission we were carrying out detailed thermal demagnetization treatment to better isolate the high T_b-component, and we intend to present these new findings in addition to the above results at the Lithospere 2010 Symposium. Current measurements on samples from Harjavalta show promising results so far, with evidence of reversed directions.

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Tests of inter-hotspot motion and of hotspot motion relative to the spin axis

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Herein, we present updated reconstructions of the Pacific plate relative to the hotspots for the past 68 million years, with the uncertainties in the reconstructions. Plate-circuit reconstructions are used to predict the tracks of some major Indo-Atlantic hotspots (Tristan da Cunha, Réunion and Iceland) from the Pacific-hotspot plate motion, and the rates of relative motion between the Pacific and Indo-Atlantic hotspots are estimated. Within the uncertainties, motion between the hotspots is found insignificant for the past 48 million years. For earlier times, a systematic error in the plate circuit used to make the predictions is inferred and which may be due to unmodeled motion between East and West Antarctica. If the observed discrepancy can be shown to correspond to an error in the plate circuit, the southward motion of the Hawaiian hotspot of 13° since ≈72 Ma, derived from an updated Pacific paleomagnetic skewness pole for chron 32 (72 Ma), can likely be attributed to true polar wander. Finally, we present a globally self-consistent model of plate motions relative to the global hotspots for the past 48 million years is. This globally self-consistent set of reconstructions can be used as a fixed frame of reference for absolute plate motions, and true polar wander, for the past 48 million years.

Keywords: mantle plumes, hotspots, plate-circuit reconstructions, paleomagnetism, Pacific plate

1. Introduction

Hotspots are volcanic anomalies, either in an intraplate setting or in the form of excessive volcanism along the plate boundaries, not explained by classic plate tectonics. In the early 70's, along with a deep mantle origin, hotspots were proposed to move so slowly relative to one another such that they could be used as a reference frame fixed in the deep mantle for describing plate motions in an "absolute" sense (e.g. Morgan, 1972). In this scenario, the stationary plumes leave trails of age-progressive volcanism on the plates as the plates move over them. These trails can then be used to track the history of absolute global plate motions. Ever since the idea was first introduced, however, the rates of relative hotspot motion, and thus the limits of the hotspot frame of reference, have remained a source of heated debate with a range of suggestions from apparent fixity (e.g. Morgan, 1971a; 1971b; 1972; Duncan, 1981; Müller et al., 1993; and also our position) to rapid motion between the hotspots (e.g. Molnar and Stock (1987) and up to 80 mm a^{-1} by Raymond et al. (2000)).

The question of inter-hotspot motion is closely related to the estimation of true polar wander—rotation of the whole solid earth relative to the spin axis. A fundamental problem of global tectonics and paleomagnetism is determining which part of apparent polar wander—the apparent movement of age-progressive paleomagnetic poles relative to the continent in question—is due to plate motion, and which part is due to true polar wander. One approach for separating these is available if the hotspots are indeed tracking the motion of the mantle beneath the asthenosphere and are moving slowly relative to one another. In this case, a model of plate motion relative to the hotspots can be used to predict the positions of past paleomagnetic poles relative to the spin axis and thus estimate the amount of true polar wander.

The Pacific plate provides an excellent opportunity to study these questions because it hosts two of the most prominent and best sampled hotspot tracks, the Hawaiian-Emperor and

Louisville chains, and paleomagnetic poles can be estimated from skewness of the marine magnetic anomalies. Thus, to make progress on answering these questions, high-quality paleomagnetic poles for the Pacific plate are needed, as well as estimates of the Pacific plate motion relative to the hotspots, along with the uncertainties in such motion. The work presented herein addresses all these questions.

2. Main results

We believe that many of the discrepancies in the suggested rates of relative hotspot motion can be attributed to shortcomings in the methods used in the previous studies, in particular to shortcomings in quantifying the inherent uncertainties. Additionally, recent improvements in the age progression along the hotspot tracks (e.g. Sharp and Clague, 2006; Koppers et al., 2004) and geomagnetic reversal time scale, lead to significant changes in results. Herein, we build on a new method for objectively estimating plate-hotspot rotations and their uncertainties (Andrews et al., 2006), and present updated reconstructions of the Pacific plate relative to the hotspots for the past 68 million years with the uncertainties in the reconstructions. To investigate the question of relative hotspot motion, plate-circuit reconstructions are used to predict the tracks of some major Indo-Atlantic hotspots (Tristan da Cunha, Réunion and Iceland) from our Pacific-hotspot rotations and estimate the rates of relative motion between the Pacific and Indo-Atlantic hotspots. Besides the uncertainties in plate-hotspot rotations, uncertainties in relative plate motions are accumulated through the plate circuit to obtain the final uncertainty (in the form of two-dimensional 95 per cent confidence regions) in the predicted positions.

It is found that the predicted and observed tracks agree much better than found in many of the previous studies, for the past 48 million years, and that within the uncertainties, motion between the hotspots is insignificant. For the discrepancy observed at earlier times, a systematic error in the plate circuit used to make the predictions is suggested, most likely unmodeled motion between East and West Antarctica. If the observed discrepancy can be shown to correspond to an error in the plate circuit, the southward motion of the Hawaiian hotspot of 13° since ≈72 Ma, as derived from an updated high-quality Pacific paleomagnetic pole for chron 32 (≈72 Ma), can likely be attributed to true polar wander.

In line with these results, we finally present a globally self-consistent model of plate motions relative to the hotspots for the past 48 million years. In this study, we have used the most up-to-date reconstructions together with radiometric dates along the major hotspot tracks to derive a plate motion model relative to the major hotspots in the Pacific, Atlantic and Indian Oceans. The new set of rotations presented provide a firm basis for estimating absolute plate motions for the past 48 million years and, in particular, can be used to separate paleomagnetically determined apparent polar wander into the component due to plate motion and the component due to true polar wander.

3. Conclusions

We show that the Indo-Atlantic hotspots have had no significant motion relative to Pacific hotspots since 48 Ma. Prior to 48 Ma, however, the apparent rates of inter-hotspot motion increase significantly. A possible cause for the pre-48 Ma apparent motion is a systematic error in the global plate circuits used to make the predictions.

It is further shown that it is possible to find a set of rotations of all plates relative to the hotspots for the past 48 million years that is both constrained to and consistent with known relative plate motions and is further consistent with fixed hotspots within uncertainties. The new set of plate reconstructions presented here provides a firm basis for estimating absolute plate motions for the past 48 million years. They can be used to separate paleomagnetically determined apparent polar wander into the part due to plate motion and the part due to true polar wander.

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Consequences of orogenic spreading – examples from the Svecofennian orogen

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The evolution of the accretionary Paleoproterozoic Svecofennian orogen encompasses three evolutionary phases: accretion, plateau and collapse. During the plateau phase the upper and middle crust have been laterally spreading towards thinner areas. Several kinematic indications suggest that the collisional crustal architecture was rearranged and a new super-infrastructure was developed during lateral spreading and flow of the crust. We will show examples of post-collisional spreading with the help of seismic reflection data, seismic anisotropy maps, lineation maps, analogue models and structural observations.

Keywords: middle crust, flow, orogenic spreading, Svecofennian

1. Introduction

In collision zones terranes with variable density, velocity and thickness are merged together resulting in an orogen with thickened crust and lithospheric mantle.The orogenic evolution has three distinct phases, (1) construction of a crustal accretionary wedge, (2) development of a continental plateau, and (3) gravitational collapse - each of which will leave different tectonic and magmatic markers (Beaumont et al., 2004; Vanderhaeghe and Teyssier, 2001a,b). The wedge to plateau transition is caused by the thermal maturation of the thickened crust and the associated genesis of a partially molten, magma bearing, low-viscosity layer within the thickened crust (Vanderhaeghe et al., 2003). The development of a continental plateau is associated with lateral gravity-driven flow of the thermally matured low-viscosity mid- to lower crust composed of migmatites and granulites. Escaping magmas are emplaced as “new granite” and volcanic suites in the plateau. The crustal flow may take place at one or more layers simultaneously; separated by crustal scale detachment zones (Culshaw et al.2006; Rey et al. 2001). The crust begins to thin only after a plate tectonic setting changes from convergent to extension – driving the thickened orogenic system in to the collapse phase. A stable crust is formed when something in the post-collisional process prohibits further extension and invokes freezing of the flow structure.

The crustal structures and the rocks record the kinematic evolution of the Earth’s crust. Theoretically, midcrustal flow (Fig. 1) should result in a crustal anisotropy (Tatham et al., 2008; Yang et al., 2008). The kinematic indicators should subsequently “freeze” and be detectable also in ancient orogens (stretching lineations, shear zones, azimuthal anisotropy of seismic velocity, and magnetic susceptibility). Midcrustal flow should also result in distinct anisotropic seismic velocities in the different crustal layers.

2. Geological background

The evolution of the accretionary Paleoproterozoic Svecofennian orogen (1.89-1.80 Ga) (Korja et al., 2006) encompasses all three evolutionary phases: accretion, plateau and collapse. At present, the upper parts of the orogen are exhumed and deeper parts (10 -15 km in depth) are exposed. The exposed crust that host representatives from palaeo lower upper crust and upper-middle crust is characterized by granitoid intrusions and highly deformed supracrustal units metamorphosed under upper amphibolite or lower granulite facies.

Our study area is central and southern Finland, where two terranes (Keitele, and Bergslagen microcontinents) and intervening basins and arcs collided with a continental plate (Karelia) around 1900 billion years ago, forming the core of the Svecofennian orogen (Korja A. et al., 2006, 2009; Fig. 1). Lateral spreading of the orogen was possible towards the west and north, where the crust was thinner. The nature of the southern boundary is enigmatic. Crustal structures suggest crustal flow towards the south (Korja A. et al., 2009) indicating another free boundary (Torvela & Ehlers, 2010).

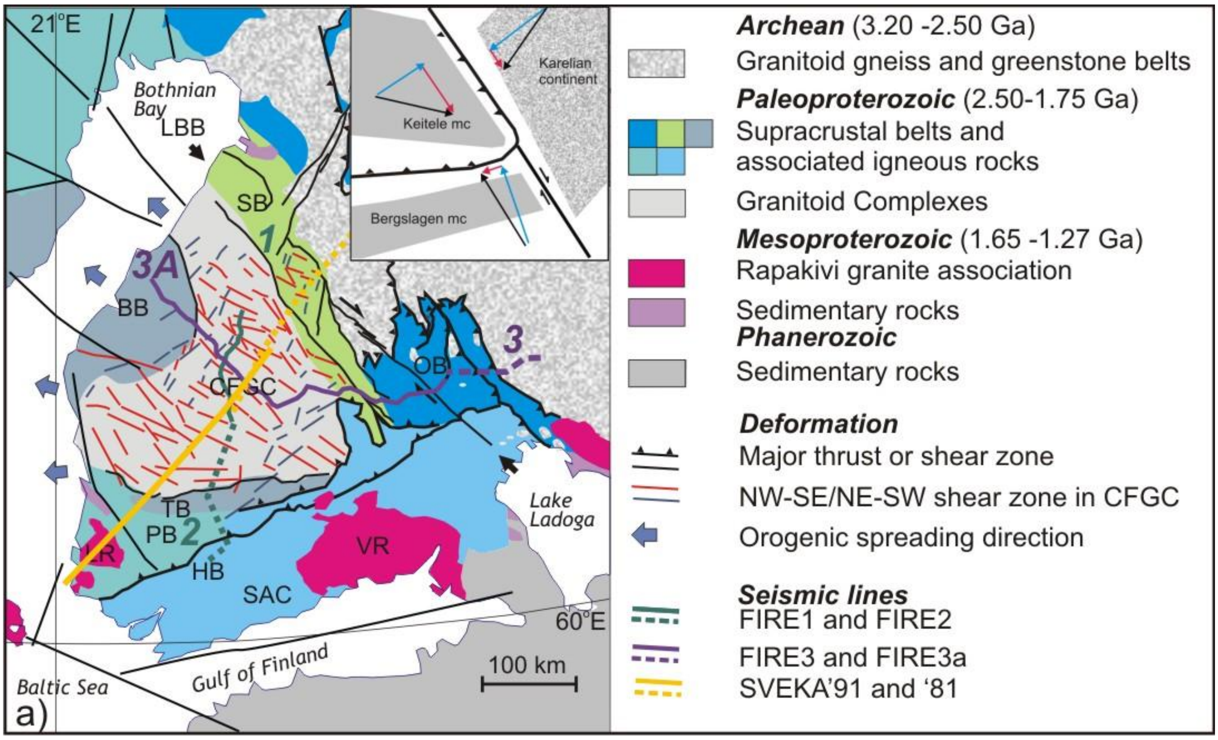

Figure 1. Seismic lines on a map of tectonic setting of the Svecofennian orogen modified after Korja et al. (2009). **a)** A map of major terranes and deformation patterns. Abbreviations: BB – Bothnian belt; CFGC – Central Finland granitoid complex; HB - Häme belt KA – Archean Karelian domain; OB – Outokumpu belt; PB -Pirkanmaa belt; SAC – Southern Finland arc complex; SB – Savo belt and TB – Tampere belt. LBB – Ladoga-Bothnian Bay wrench fault zone is indicated with arrows. **b)** Precollisional plate tectonic setting at 2.0 Ma.

3. Indications of detachments and flow

The deep seismic reflection profiles (FIRE) crossing the study area in two perpendicular directions suggest a three-fold layering of the thick crust (>55 km) that was developed after accretion of the orogeny (Fig. 2). The crustal layers are separated by décollement zones on to which crustal reflection sole out and across which velocity steps occur.

The upper-middle crust décollement zone is exposed at the NE edge of the Central Finland granitoid complex. The detachment is identified from seismic sections as a 2-3 km thick highly reflective zone, separating the upper and middle crust. It is exposed at Elämäjärvi, where low-angle shear zones exhume it. The Elämäjarvi deformation zone is a 10 km wide and 70 km long macro-structure, where highly strained rocks are found in low-angle orientations (Kilpeläinen et al., 2008). The lithologies range from sheared coarse-porphyritic granite to augengneiss, mylonite and to ultramylonites cutting mylonites suggesting that the shear zone was active for a long time and at various depths.

FIRE profiles show complementary crustal scale structures of compression and extension in orthogonal directions (Korja, A. and Heikkinen 2008; Korja, A. et al., 2009). The geometric relationships of the crustal layers and deformation patterns suggest compression in WNW – ESE direction and extension in N-S direction. The upper crust spread in a brittle to ductile regime along listric, low angle and transfer shear zones, and the middle crust thinned via ductile flow and extrusion. The middle crust displays typical large scale extensional flow structures: herringbone and anticlinal ramps (Culshaw et al., 2006). Both are rooted to large scale listric surfaces.

Nikkilä et al. (2009) were able to produce the main crustal structures described from the FIRE profiles in an analogue modeling experiment testing convergence coupled with perpendicular extension. The results of the high-T experiment support layered crustal model consisting of brittle upper crust, plastic middle crust and rigid lower crust. An additional heating event further enhanced the midcrustal extensional structures. The models also suggest that an additional heating event was required to produce the intensive and well-preserved extensional structures of the Svecofennian middle crust.

The three dimensional tomography model of southern and central Finland (Hyvönen et al., 2007) suggests that the Svecofennian crust is composed of large scale blocks. The residual component of the tomographic model hosts azimuthal velocity anisotropy that shows consistent patterns over large areas (Hyvönen et al., 2008; 2010(this volume)). Because the anisotropy varies with depth it is suggested that different layers of the crust have moved independently after collision indicating lateral spreading during plateau phase.

Another example of post-collisional movement of the bedrock blocks is the formation of persistent low angle lineations in most of the rock types (Kilpeläinen et al., 2008). The new lineation map of Finland (Lammi and Kilpeläinen, 2010, this volume) that contains 78 000 lineation measurements across the country displays consistent regional movement directions. The patterns suggest large-scale block-movements and deformation in the upper and middle crustal after collision. It is suggested that lineations are often related to lateral spreading processes of the Svecofennian orogeny.

Granulites and migmatites, which are partially molten rocks of the middle crust, are well exposed in West Uusimaa Complex (WUC), where they can be studied for midcrustal flow mechanisms and directions. Our preliminary observations (Torvela and Korja, 2010, this volume) from the area; E-dipping low to moderately dipping schistosity, E-dipping low angle lineations and a possible low-angle detachment shear zone coupled with the E-W- directed steeply dipping shear zones are compatible with a flow structure where the movement is in east-west direction. Especially the large-scale, steeply dipping shear zones play a central part in the mid-crustal flow architecture as already pointed out by Cagnard et al. (2006).

4. Conclusions

Several kinematic indications suggest that the collisional crustal architecture was rearranged during plateau stage and a new super-infrastructure was developed during lateral spreading and flow of the crust. The current surface exposure is interplay between upper crustal superstructure and middle crustal infrastructure.

Acknowledgements:

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Electrical conductivity of mantle lithosphere in Fennoscandia

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Keywords: electrical conductivity, magnetotellurics, continental lithosphere, Fennoscandia

The first long period electromagnetic soundings in Fennoscandia were completed in early 1980's (e.g. *Jones, 1980, 1983;* Adam et al., 1982; *Jones et al., 1983; Krasnobaeva et al.,1981; Kaikkonen et al., 1983*). Since then several additional deep probing studies were carried out (e.g. *Kovtun et al., 1988, 1994; Pajunpää, 1988; Rasmusen, 1988; Viljakainen 1996*) before the end of the last century. Recent improvements of magnetotelluric instrumentation have made it possible to obtain reliable and good quality data from magnetotelluric profile and array measurements. This is important, in particular, for upper mantle studies because long recording times are needed for deep probing soundings to obtain long period data and to correct for source field effects. As an example, in the recent EMMA work (*Smirnov et al., 2006*), simultaneous nine months long recordings at 12 sites were carried out from Aug 2005 to Jun 2006. These recordings provide information to the depths of several hundreds of kilometres.

In 1998 a large MT array was employed in Fennoscandia as a part of the BEAR research (*Korja et al., 2002*), which, for the first time, provided data over the entire shield (e.g. *Lahti et al., 2005*). Since then several extensive data sets have been collected both in the Fennoscandian Shield (Jämtland - *Korja et al., 2008*; EMMA – *Smirnov et al., 2006*; MT-FIRE – *Vaittinen et al., 2006*) as well as on its margins (TOR – *Smirnov and Pedersen, 2006*; EMTESZ-Pomerania – *Brasse et al., 2006*). A list of references to original work used in the current work are given in the reference list.

Results show that in Fennoscandia (and in the East European Craton) electrical asthenosphere is either very deep or is absent (or cannot be detected by magnetotellurics) whereas in Central and Southern Europe electrical asthenosphere is much shallower. Rapid transition from thick East European Craton to thinner Phanerozoic Europe coincides with the Trans European Suture Zone.

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Archaean-Proterozoic boundary in the East European Craton: Crustal conductivity in the Fennoscandian and Ukrainian Shields

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Crustal conductivity structure is examined at the Archaean-Proterozoic boundary in the Fennoscandian and Ukrainian Shields. Earlier large scale magnetometer array studies are used to delineate the conductors while recent magnetotelluric studies are used to examine the detailed structure of the conductors. Results from both shields yield similar conductivity structures. The structures may not coincide spatially rather than suggest similar genetics for the conductors associated with Ar-Pt boundary.

Keywords: Crustal conductivity anomalies, East European Craton, Fennoscandian and Ukrainian Shields, Archaean-Proterozoic boundary, collisional tectonics

Ancient cratons, such as the East European Craton (EEC), have been assembled by the amalgamation of continental lithospheric plates resulting in the relatively homogeneous interiors of the plates and complex boundary zones between the plates. One of the major boundaries is the Archaean-Proterozoic (Ar-Pt) boundary, which was formed when, in the Palaeoproterozoic time, young Palaeoproterozoic plates or microcontinents collided with the Archaean cratons in modern type subduction/collision processes. We will investigate the nature and structure of the Ar-Pt boundary zone using electrical conductivity models across the boundary zone in the EEC. The EEC is characterized by a set of elongated conductivity anomalies that separate large resistive regions. We have collected electrical conductivity models representing the Ar-Pt boundary zone from several regions in the EEC (Fig, 1). Two of these areas are associated with the ca. 1000 km long Ladoga - Bothnian Bay zone (LBBZ) in the Fennoscandian Shield. The LBBZ has been studied by magnetometer arrays and magnetotellurics since 1980 in several profiles crossing the zone (Figs. 1 and 2.). We will use models from the older GGTSVEKA transect and the recent FIRE-MT profiles (Fig. 3) in Finland as well as the Suoyarvy-Vyborg profile in Russia provided by the groups from the universities of Oulu and St. Petersburg, respectively. The third area is associated with the Kirovograd anomaly (KvA) area in the Ukrainian Shield (Fig. 4). The anomaly was originally discovered by the Ukrainian geophysicists. Its northern continuation under the sedimentary cover of the EEC is currently the target of a collaborative magnetotelluric study by researchers from the GEMRC, the Moscow State University and the Geophysical Institute of the Ukrainian National Academy of Sciences. We will compare the conductivity structures across the Ar-Pt boundary zone and illustrate their similarities and differences together with the geological properties of the target areas. Supplementary geological/geophysical information has helped us to discern between the competing hypothesis of the origin of the enhanced conductivity and conditions of its formation. We suggest that the hypothesis attributing the high conductivity to the Palaeoproterozoic graphite/sulphide/iron-oxide -

bearing metasedimentary rocks and reworked volcanic rocks in the sutures between the colliding plates looks the most plausible. We will also investigate whether these two conductive belts form a single, extremely long conductor or represent separate conductors formed in similar processes at the same time (Palaeoproterozoic) but separated in space. This is the first detailed comparison between the two distant outcrops of the Precambrian basement in the EEC and will provide a basis for the further deep investigations to check the existing hypotheses and to contribute to the understanding about the ancient tectonic processes in the making of the East European Craton.

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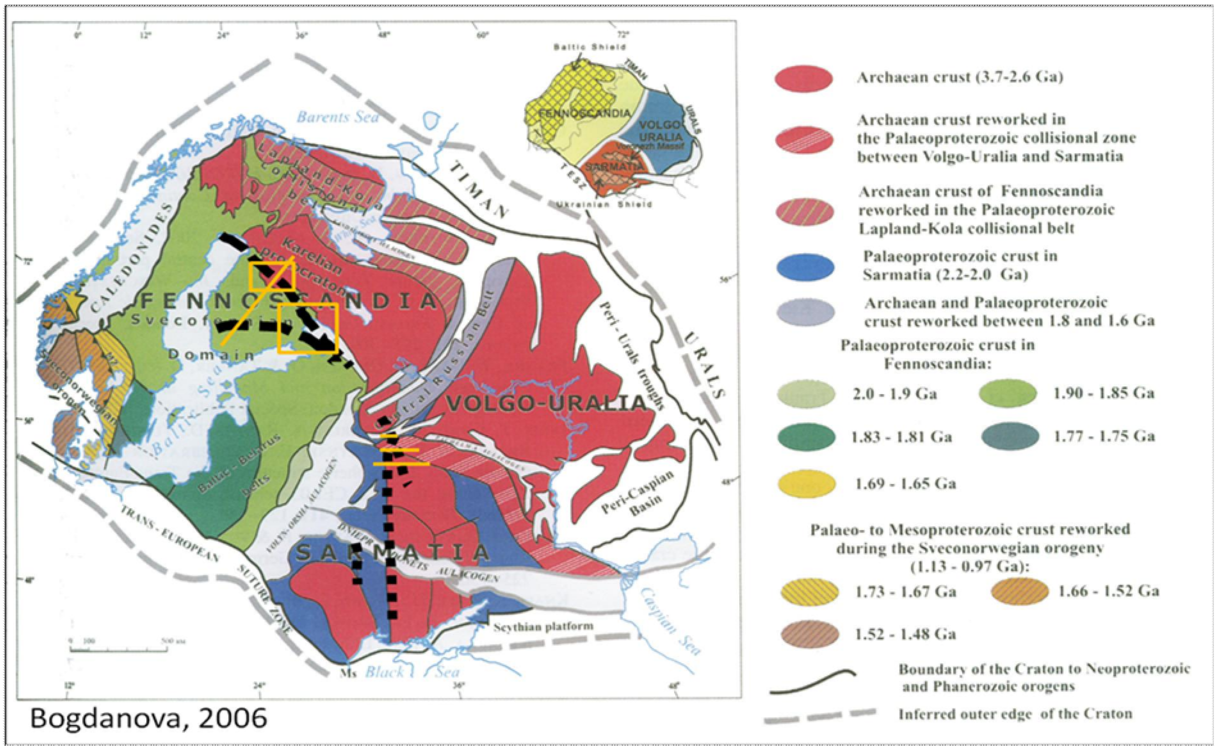

Figure 1. Approximate location of elongated crustal conductors (black dashed lines) at the Archaean-Proterozoic boundary in the Fennoscandian and Ukrainian Shield. Geological map from Bogdanova et al., 2005. Yellow lines and rectangular show the location of detailed crustal conductivity models.

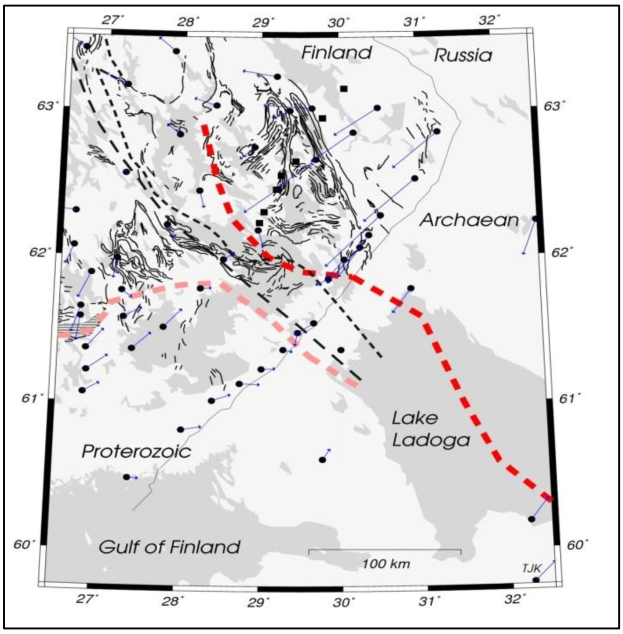

Figure 2. Location of deep (upper to middle crustal) conductors in the Lake Ladoga area (red dashed lines). Background map (black line drawings) shows the location of near-surface conductors from airborne electromagnetic mapping (Airo 2005). Induction arrows are from Pajunpää (1987). Figure from Korja et al., 1999.

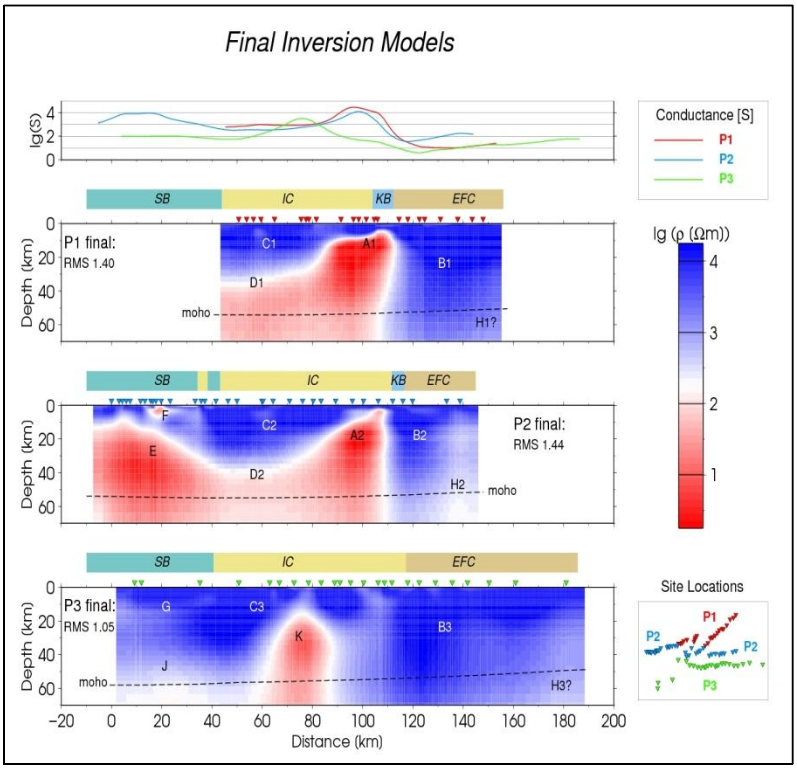

Figure 3. Final 2D inversion models from the east-west directed MT-FIRE profiles across the Archaean-Proterozoic border in central Fennoscandian Shield (Vaittinen et al., 2010).

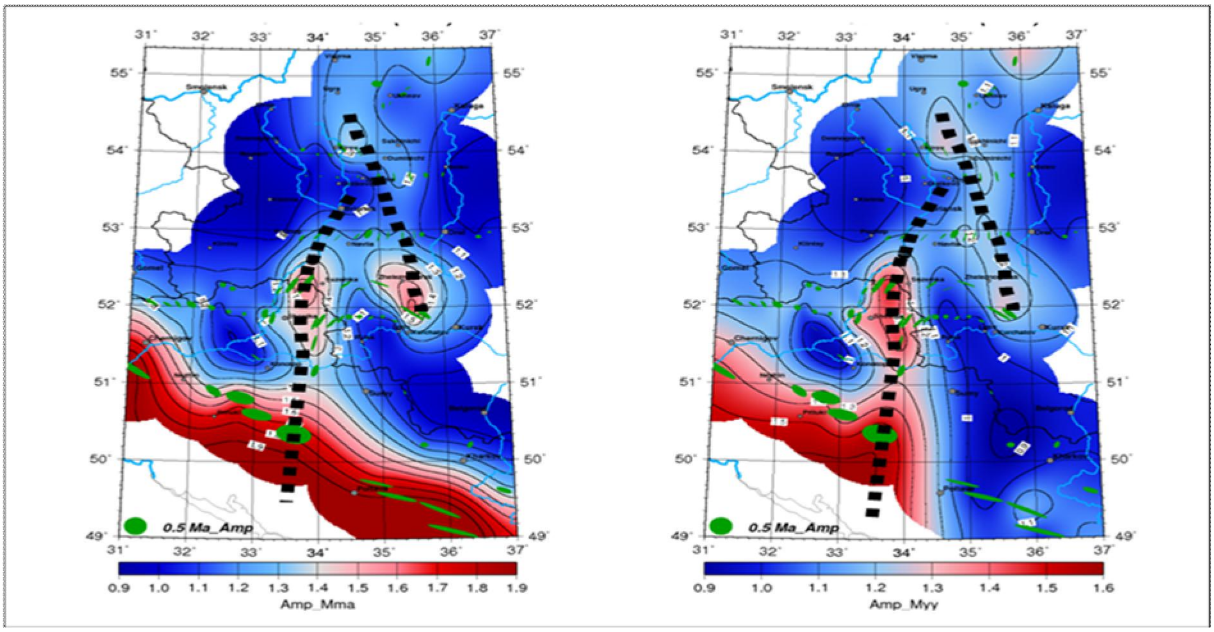

Figure 4. Horizontal magnetic responses locate the northward extension of the Kirovograd anomaly (Ukrainian Shield) under the sedimentary cover of the East European Craton.

Ultrasonic measurements of P- and S-wave velocities in lower crustal rocks under uniaxial compression

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Longitudinal (V_p) and shear (V_s) wave velocities of unheated lower crustal rock samples were measured using an ultrasonic method. The measurements were carried out on cylindrical samples under uniaxial compression corresponding to hydrostatic pressures up to 300 MPa. Preliminary velocity estimates, as well as calculated reflection coefficients between different rock types are presented.

Keywords: V_p, V_s, ultrasonic, pressure, lower crust, reflection

1. Introduction

Detailed seismic reflection and refraction studies require knowledge of reflection coefficients between various crustal and upper mantle layers. Here we present new ultrasonic velocity (Vp, Vs) and reflection coefficient data for various lower crustal rocks from Central Finland and the Belomorian Province, Russia.

2. Samples

For this study we chose 8 cylinders drilled from unoriented rock samples. Samples 02D (ultramafic rock), 03A (enderbite) and 04C (amphibolite) were provided by Jorma Paavola, and were collected by hand from Varpaisjärvi and Sonkajärvi (Fig. 1). Eclogite sample B2-A was acquired from the Belomorian Province, near the town of Gridino (Volodichev et al., 2004) and it was provided by Pentti Hölttä.

The cylinders were (22.0-22.3) mm in height and (25.0±1) mm in diameter. To see the effect of anisotropy the samples were drilled in several directions. This was however difficult due to the small size of the samples, which is why only sample 02D was drilled in all orthogonal directions with z being the vertical axis. The dry bulk density of the samples was also measured and is presented in Table 1.

3. Ultrasonic method

The device used for the measurements (Fig. 2) has been developed in the Electronics Research Laboratory of the University of Helsinki. It consists of two custom-built transducers, a hydraulic cylinder and a load cell residing inside a metal cradle. The piezoelectric crystals for both longitudinal and shear waves (1,0 MHz and 1,1 MHz respectively) are excited simultaneously with a pulser (5058PR, Panametrics). Seismic velocities are then calculated from the time of flight through the sample. The method has been described in detail by Karlqvist (2009) in his MSc thesis and Lassila et al. (2010). The correspondance of uniaxial load measurements to triaxial stress measurements has been showed (Lassila et al., 2010).

The samples were measured in 25 points under loads of 250-15000 kg. This corresponds to hydrostatic pressures of 5-300 MPa and crustal depths of < 10 km.

4. Results

A summary of the results has been given in Table 1. To illustrate one whole measurement, the velocity of sample B2-A has been plotted as a function of pressure. The reliability of the measurements increases with added compression, because depressions caused by the rock samples on the transducer surfaces may impede on the wave transfer under lower pressures. This will be taken into account for further development of the method.

Reflection coefficients between the samples have been calculated from acoustic impedance. Some values high enough (> 0.1) to cause strong reflections can be seen (Warner, 1990).

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Warner, M., 1990. Absolute reflection coefficients from deep seismic reflections. Tectonophysics 173, 15-23.

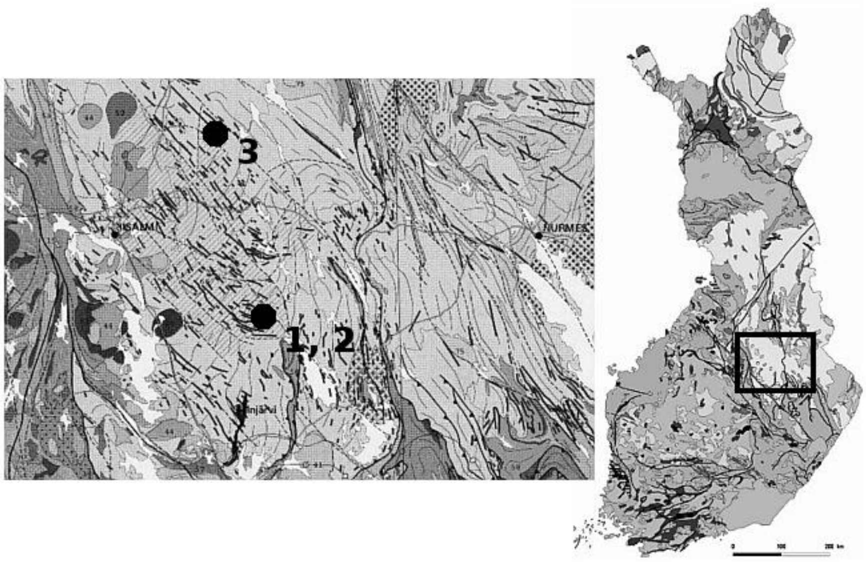

Figure 1. Samples 1) 02D and 2) 03A were collected from Jonsa, Varpaisjärvi, and sample 3) 04C from Kulvemäki, Sonkajärvi.

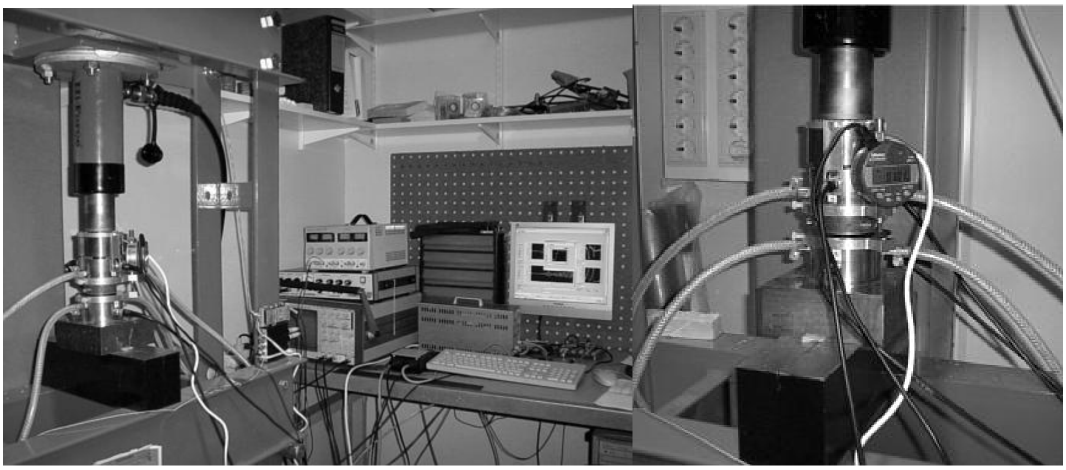

Figure 2. Measurement device.

Table 1. Summary of the results. Rock type and dry bulk density of the samples, as well as seismic velocities measured under the lowest and highest compression. V_p/V_s -ratio for all the measurement points.

Sample	Rock type	Density (kg/m3)	V_p (m/s) P= 5MPa	V_p (m/s) P=300MPa	V_s (m/s) P= 5MPa	V_s (m/s) P=300MPa	V_p/V_s
04C-x	amphibolite	3167	5922	7145	3259	3881	1.53-1.84
03A-x	enderbite	2824	5803	6154	2722	3447	1.68-1.79
02D-x	ultramafic rock	3065	6432	6927	3648	4032	1.57-1.69
02D-y	(ortoamphibole-	3085	6474	7705	3628	4450	1.45-1.73
02D-z	kornerupine)	3084	5566	7344	3032	3716	1.48-1.95
B2-Ay	eclogite	3180	4406	6594	2307	3468	1.27-1.90
B2-Aza	eclogite	3160	5110	6741	2180	3411	1.50-1.98
B2-Azb	eclogite	3167	4059	6649	2209	3339	1.22-1.99

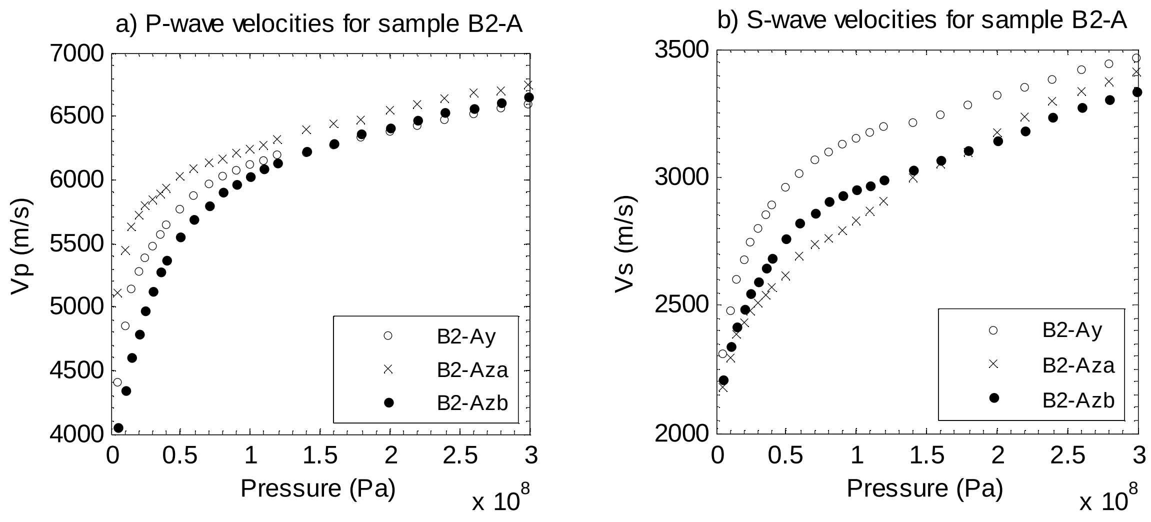

Figure 3. Longitudinal (a) and shear (b) wave velocities for eclogite sample B2-A.

Table 2. Reflection coefficients between different rock samples calculated from acoustic impedance. Higher values are bolded.

	Reflection coefficients (P=300 MPa)							
Sample	04C-x	03A-x	02D-x	02D-y	02D-z	B2-Ay	B2-Aza	B2-Azb
04C-x	0.00	**0.13**	-0.03	-0.02	-0.00	0.04	0.03	0.04
03A-x	**-0.13**	0.00	**-0.10**	**-0.16**	**-0.13**	**-0.09**	**-0.10**	**-0.10**
02D-x	0.03	**0.10**	0.00	-0.06	-0.03	0.01	-0.00	0.00
02D-y	0.02	**0.16**	0.06	0.00	0.01	**0.06**	0.05	**0.06**
02D-z	0.00	**0.13**	0.03	-0.01	0.00	0.04	0.03	0.04
B2-Ay	-0.04	**0.09**	-0.01	**-0.06**	-0.04	0.00	-0.01	-0.00
B2-Aza	-0.03	**0.10**	0.00	-0.05	-0.03	0.01	0.00	0.01
B2-Azb	-0.04	**0.10**	-0.00	**-0.06**	-0.04	0.00	-0.01	0.00

POLENET/LAPNET: project status and first results

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We present the first results of the POLENET/LAPNET seismic array research in northern Fennoscandia (northern parts of Finland, Sweden, Norway and Russian Karelia) during the International Polar Year 2007-2009.

Keywords: IPY, POLENET, glacial earthquakes, local seismic events, Fennoscandia, Greenland

1. Introduction

POLENET/LAPNET is a sub-project of the IPY 2007-2009 POLENET consortium related to seismic studies in the Arctic. The main target of the project is to study dynamics of the lithosphere in northern Fennoscandia based on data of an temporary broadband seismic array research. The POLENET/LAPNET array, with the average spacing between stations of 70 km, was designed to solve specific tasks of polar seismology. The collected POLENET/LAPNET dataset includes high-frequency continuous data (sampling rate from 50 to 100 sps) of 37 temporary stations, which were in operation during the time frame from 01.05.2008 to 31.09.2009, and of 21 stations of selected permanent networks in Fennoscandia. Most of the stations of the array were equipped with broadband sensors. The data of broadband stations, pre-processed into the standard seismological miniSeed format, is now deposited into the database of FOSFORE Data Centre at the University of Grenoble (France) and backup copy of all continuous data is stored at the Sodankylä Geophysical Observatory (Finland). In addition, the data of several short period stations is archived at the Sodankylä Geophysical Observatory and at the Geophysical Centre RAS, Institute of Physics of the Earth RAS, Russia.

After the necessary data validation period (until June, 2012), the POLENET/LAPNET data will be published and made freely available via international ORFEUS and IRIS data centres. The data validation will include not only formal control on data formats and data continuity, but also detailed work with the waveform data of seismic events.

The POLENET/LAPNET scientific program includes several major topics:
1) studying the deep structure of the Earth (in particular, the Earth’s core);
2) studying the lithosphere structure and evolution in the northern part of the Fennoscandian shield and estimation of realistic distribution of mantle viscosity in the areas of glacial isostatic adjustement;
3) studying the cryosphere-lithosphere interaction and glacial seismic events.

Selected results of the POLENET/LAPNET project are presented below.

2. Studying local seismic events in northern Fennoscandia

The POLENET/LAPNET array recorded a number of local seismic events (both earthquakes and explosions). Particularly interesting was a group of earthquakes originating from the area around the Tornio river. The area is cut by numerous faults, stretching both from NNE to SSE and from N to S. The events were relocated using first arrivals of P-and S-waves picked from the POLENET/LAPNET data at stations with distance less than 200 km from the epicentre.

Different location techniques were applied and master events (blasts from the Hukkavaara hill, for which the coordinates are known with high precision) were used to test precision of location procedures. The relocated events can be separated into two groups according to the hypocentre depth: one group includes events with the hypocentre depths of 0-15 km, which is a typical depth for earthquakes in intraplate Fennoscandia, and the second one includes events with the hypocentre depths of 15-30 km, which is not typical for this region. Epicentres of relocated events show good coincidence with the known faults in the area.

4. Monitoring of glacial earthquakes from Greenland

Monitoring of glacial earthquakes from Greenland was one of the major targets of the POLENET/LAPNET experiment. The first results of the data analysis have shown that the POLENET/LAPNET array, located at regional distances from Greenland, recorded more such events than it has been recorded by the Global Seismographic Network during the same observation period. The waveforms of events recorded by the array differ from both long-period waveforms of glacial earthquakes recorded at teleseismic distances and from short-period glacial rumblings recorded at local distances. Recordings of glacial earthquakes obtained by the POLENET/LAPNET array contain the long-period energy only. In many cases the events were recorded in groups within the time interval of up to 1 hour. Generally, the waveforms of events within the same group are different and the events not always originate from the same location. For some of the events it was possible to recognize not only dispersed long-period surface waves, but also the first arrival of a very long-period P-wave. This observation suggests that source duration of these events was long. Diversity of the waveforms of glacial events recorded by the array can be considered as evidence for diversity of source mechanisms of these events.

5. Acknowledgements

POLENET/LAPNET Working Group: Elena Kozlovskaya, Helle Pedersen, Jaroslava Plomerova, Ulrich Achauer, Eduard H Kissling, Irina Sanina, Teppo Jämsen, Hanna Silvennoinen, Catherine Pequegnat, Riitta Hurskainen, Helmut Hausmann, Petr Jedlicka, Igor Aleshine, Ekaterina Bourova , Reynir Bodvarsson, Ewald P Brueckl, Tuna Eken, Pekka J Heikkinen, Gregory A Houseman, Helge Johnsen, Kari Komminaho, Helena Munzarova, Roland Roberts, Bohuslav Ruzek, Zaher Hosein Shomali1, Johannes Schweitzer, Artem Shaumyan, Ludek Vecsey, Sergei Volosov

Institutions participating in the POLENET/LAPNET experiment:

Sodankylä Geophysival Observatory of the University of Oulu(FINLAND)
Institute of Seismology of the University of Helsinki (FINLAND)
University of Grenoble (FRANCE)
University of Strasbourg (FRANCE)
Institute of Geodesy and Geophysics, Vienna University of Technology (AUSTRIA)
Geophysical Institute of the Czech Academy of Sciences, Prague (CZECH REPUBLIC)
Institute of Geophysics ETH Zürich (SWITZERLAND)
Institute of Geospheres Dynamics of the Russian Academy of Sciences, Moscow (RUSSIA)
Geophysical Centre of the Russian Academy of Sciences, Schmidt Institute of Physics of the Earth of the Russian Academy of Sciences (RUSSIA)
(11) Swedish National Seismological Network, University of Uppsala (SWEDEN)
(12) Institute of Solid Earth Physics, University of Bergen (NORWAY)
(13) NORSAR (NORWAY)
(14) University of Leeds (UK)

Postglacial Fault Drilling Project: A new initiative for an ICDP project

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The Postglacial Fault Drilling Project (PFDP) is a drilling project in development of the International Continental Scientific Drilling Program (ICDP). Finland, Sweden and Norway are all member countries of ICDP. The suggested PFDP project aims to drill deep research boreholes into postglacial faults in northern Fennoscandia which were generated in the last stages of the Weichselian glaciation. The research is anticipated to advance science in neotectonics, hydrogeology and deep biosphere studies, and provide important information for nuclear waste disposal, seismic risks of hydropower dams and mines, petroleum exploration on the Norwegian continental shelf and studies of mineral resources in PG fault areas. In October 2010, an international ICDP supported workshop was arranged in Skokloster, Sweden. 40 participants from 8 countries exchanged ideas and planned the further development of the project.

Keywords: Postglacial faults, Fennoscandian Shield, drilling, glaciation, tectonics, intraplate seismicity, earthquakes, rock stress, groundwater flow, groundwater chemistry, deep biosphere

1. ICDP and Nordic countries

The International Continental Scientific Drilling Program (ICDP) is an international scientific organization under the umbrella of International Union of Geodesy and Geophysics (IUGG), International Union of Geological Sciences (IUGS) and International Lithosphere Program (ILP). ICDP promotes scientific drilling studies to provide exact, fundamental and globally significant knowledge of the composition, structure and processes of the Earth's crust. ICDP has 18 member countries including Finland, Norway and Sweden. ICDP projects with Nordic co-ordinators and significant participation include the Outokumpu Deep Drilling Project (Kukkonen, 2009) and the FAR-DEEP project (see http://far-deep.icdp-online.org/). Furthermore, the Swedish Deep Drilling Program (SDDP) (Lorenz, 2010) is planning to acquire a special drilling rig for 2-3 km deep holes and two deep holes have been suggested to be drilled in the COSC project in the western Sweden (Gee et al., 2010) as a part of the SDDP.

2. Postglacial Fault Drilling Project

During the last stages of the Weichselian glaciation (ca. 9,000 - 15,000 years B.P.), reduced ice load and glacially affected stress field resulted in active faulting in Fennoscandia with fault scarps up to160 km long and up to 30 m high. These postglacial (PG) faults are usually SE dipping, SW-NE oriented thrusts, and represent reactivated, pre-existing crustal discontinuities (Fig.1). The faults were probably formed in single large-magnitude earthquakes, and current seismic activity is commonly observed in these structures. Postglacial faulting indicates that the glacio-isostatic compensation is not only a gradual viscoelastic phenomenon, but includes also unexpected violent earthquakes, suggestively larger than other known earthquakes in stable continental regions. The PFDP project aims to investigate, via scientific drilling, the characteristics of postglacial faults in northern Fennoscandia, including their structure and rock properties, present and past seismic activity and state of stress, as well as hydrogeology and associated deep biosphere (Kukkonen et al., 2010). The presentation introduces the PFDP project and the present status of the project development.

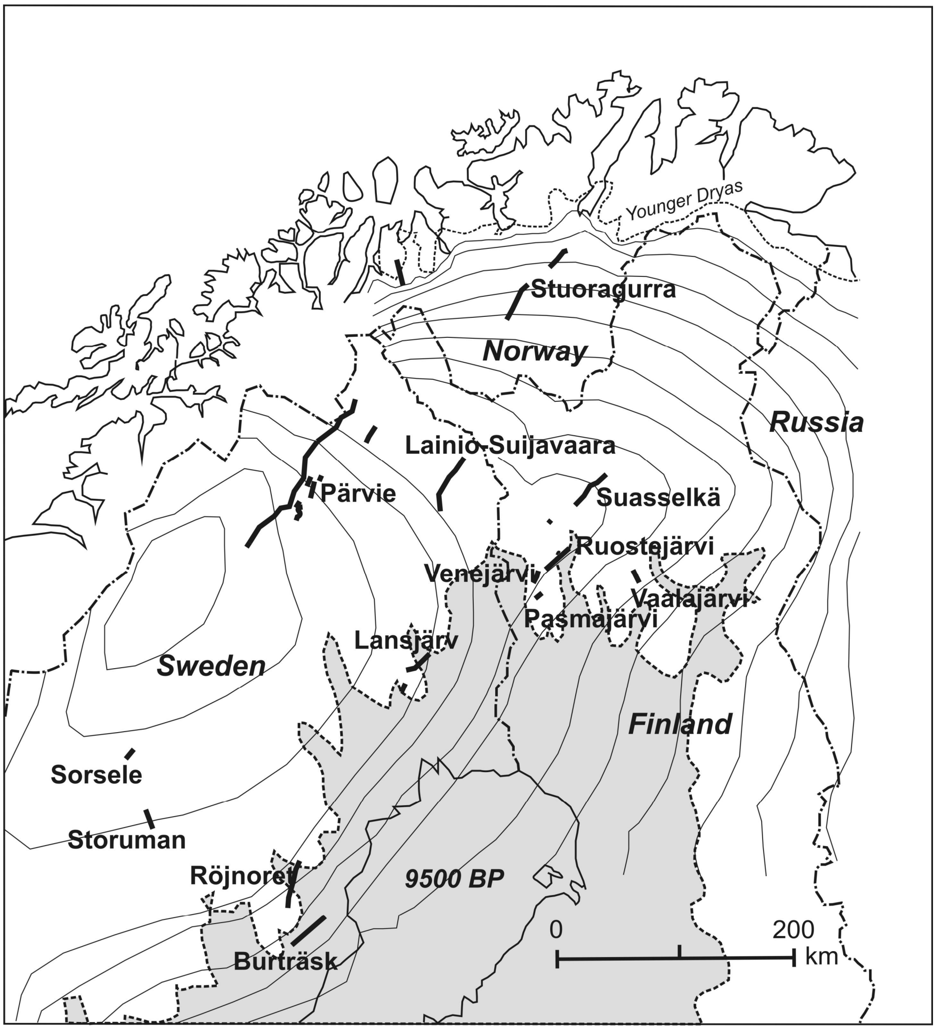

Figure 1. Location of postglacial faults in northern Fennoscandia (thick lines), and successive ice-marginal lines between ca. 10,000 and 9,000 years B.P. (thin lines). The grey area shows the highest shoreline of the Baltic. Figure adapted from Kukkonen et al. (2010)

References:

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HIRE Seismic Reflection Survey in the Hannukainen-Rautuvaara Fe-Cu-Au exploration area, Northern Finland

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A seismic reflection survey comprising six Vibroseis lines (total length of 71.7 km) and two explosion seismic lines (total length 8.7 km) was carried out in the Hannukainen-Rautuvaara Fe-Cu-Au exploration area, northern Finland, in March-April 2008. The survey is a part of the project HIRE (High Resolution Reflection Seismics for Ore Exploration 2007-2010) of the Geological Survey of Finland (GTK). The Hannukainen-Rautuvaara survey was done in co-operation with the Northland Exploration Finland Oy.

Keywords: Seismic reflection surveys, Fe-Cu-Au deposits, Hannukainen, Rautuvaara, Fennoscandian Shield

1. HIRE project

The goals of the HIRE (High Resolution Reflection Seismics for Ore Exploration 2007-2010) project are to (1) extend reflection surveys to exploration of the Precambrian crystalline bedrock of Finland, (2) apply 3D visualization and modelling techniques in data interpretation, and (3) improve the structural database on the most important mineral resource provinces in Finland.

The seismic contractor responsible for the field acquisition and basic processing of the data was Vniigeofizika (together with Machinoexport), Moscow, Russia. The Institute of Seismology, University of Helsinki, is GTK's research partner and subcontractor in the project and responsible for the more detailed post-stack processing of the results. As many of the targets are located in areas of active exploration and/or mining, 12 industrial partners have also participated in the project. The HIRE surveys altogether comprise 700 line-km of 2D reflection surveys in 16 target areas. The targets consist of exploration and mining camps in a very diverse selection of geological environments with Cu, Ni, Cr, PGE, Zn, and Au deposits, most of them economic, as well as the Finnish site for nuclear waste disposal.

HIRE fieldwork was carried out in the 16 target areas in 2007- 2008. Typically, a target was covered with a network of connected survey lines, with a total length of lines varying from about 10 km to 90 km per target.

The present paper presents results from the HIRE survey in the Hannukainen-Rautuvaara Fe-Cu-Au exploration area in northern Finland.

2. The HIRE survey results in the Hannukainen-Rautuvaara area

The Hannukainen-Rautuvaara area in northern Finland hosts several Fe-oxide-Cu-Au deposits, mostly in clinopyroxene-dominated skarn or albitite. The study area is characterized by strong, high-amplitude reflectivity in the uppermost 5 km. The most prominent regional structure is an extensive system of three major reflective layers which range in thickness from less than 200 m to about 1 km. The layers can be correlated laterally over distances of 10-20 km. These reflectors form an open asymmetric antiform which has one limb dipping about 20º

SW under the monzonite intrusions in the west, and the other limb in a subhorizontal position under quartzite and other metasediments in the eastern part of the survey area. The large-scale structure can be interpreted as thrusting from SW to NE (the Äkäsjoki Thrust Plane) and be related to a major bedrock structure in western Lapland, the Kolari Shear Zone. Thrusting has taken place in the same direction as the well-developed lineation of the area. The SW-NE Äkäsjoki shear zone is interpreted as a subvertical strike-slip fault developed in the direction of thrusting.

Reflectors in the Hannukainen area correlate with the known Fe-Cu-Au deposits, and the results suggest that the strong reflectivity is due to iron stone, sulphide, skarn and amphibolite within monzonite, diorite and metasediments (Fig. 1). Reflectors also correlate with magnetic and electrically conductive layers. There are plenty of reflectors beneath the known deposits in the Hannukainen and Kuervitikko areas. Several potentially interesting targets for further exploration can be identified in the already known deposit areas, but also in areas covered by quartzite and other metasediments. In Rautuvaara too, reflectors correlate with amphibolite and skarn layers, but the structures are steeper than in Hannukainen, which complicates the modelling of seismic sections. The seismic data suggest that the structures in Rautuvaara become much less steep at the depth of about 1 km.

The HIRE results suggest that the ore potential of the Hannukainen-Rautuvaara area may be higher than earlier anticipated.

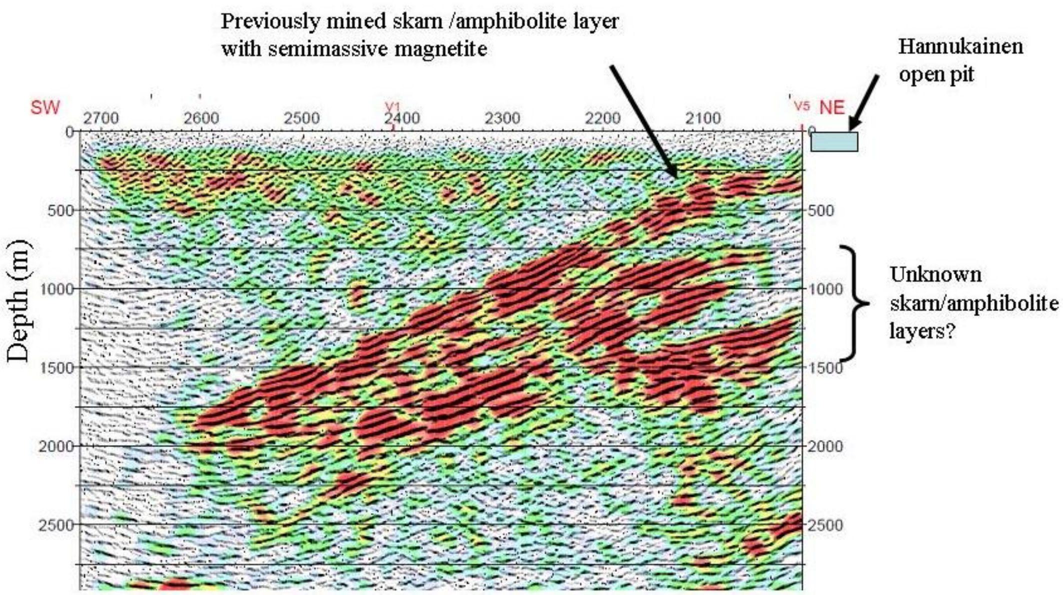

Figure 1. A migrated seismic section in the Hannukainen area. The Fe-oxide-Cu-Au deposits mined in the Hannukainen open pit mine are associated with the uppermost strongly reflective layer. Beneath this layer, similar bright reflector packages exist.

Reference:

Kukkonen, I., Heikkinen, P., Heinonen, S., Laitinen, J. & HIRE Working Group of the Geological Survey of Finland, 2009. HIRE Seismic Reflection Survey in the Hannukainen-Rautuvaara Fe-Cu-Au exploration area, northern Finland. Geological Survey of Finland, Report Q23/2009/50, 55 p.

Internal Structures of Ductile Shear Zones

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Finnish Archean and Paleoproterozoic domains are intersected by numerous shear zones which are, in detail, complicated systems created by many shear deformation events. Indication of multistage evolution is visible in the varying types of fault rocks as well as in the variation of shear sense. This presentation gives three examples of shear zone systems developed at different tectonic settings and composed of elements created at different crustal depths, in diverse metamorphic environments. Thus, a record of impacts and environments prevailed during individual stages of polyphase tectonic evolution can be seen in these shear structures.

Keywords: shear zone, internal structure, fault rock, cataclastite, mylonite, blastomylonite, shear migmatite, annealed fault rock.

1. Introduction

The bedrock of Finland includes numerous ductile shear zones that have been formed as a consequence of different stages of tectonic events. This presentation gives examples of internal structures of three Finnish shear zones developed in different tectonic settings. One of those defines the border zone of Archean and Paleoproterozoic domain, one is located in the Finnish Karelian domain and one in the Svecofennian domain. Evolution of each zone comprises multiple stages of shear deformation, and the sections exposed at the present surface contain products created at various crustal levels in different metamorphic environments.

2. Shear Zone Examples

Hirvaskoski Shear Zone (Kärki et al. 1993, 1995) defines the border between the Archean Pudasjärvi complex and Kainuu belt (Vaasjoki et al. 1998) in the middle part of Finland (Fig. 1). The zone strikes in the north – south direction and individual shear structures are often steeply westward dipping. As a whole, the deformation zone is ca. 15 km wide and both the Archean migmatites and the Paleoproterozoic supracrustal units were deformed by diverse mechanisms during Paleoproterozoic and later tectonic events. Intrusive-like pegmatoids are important constituents of this zone and they build elongated bodies of different sizes following the strike of the zone.

Most of the pegmatoids within the HSZ seem to represent migrated leucosomes generated simultaneously with certain stages of Paleoproterozoic deformation, in relatively high grade metamorphic environment, at least in upper amphibolite facies (Tuisku et al. 1990). Shear structures with totally brittle, incohesive character are omitted from this review but, as expected, earlier shear structures are overprinted by brittle structures in many places. More ductile shear bands can be tens of metres wide and internally quite homogeneous but also narrow, mm-scale shear bands are significant products of certain deformation phases. Fault rocks vary from semi-brittle elements to ductile, partially molten types. Kinematics indicators are similarly variable and relative horizontal displacement in different shear bands can be either sinistral or dextral.

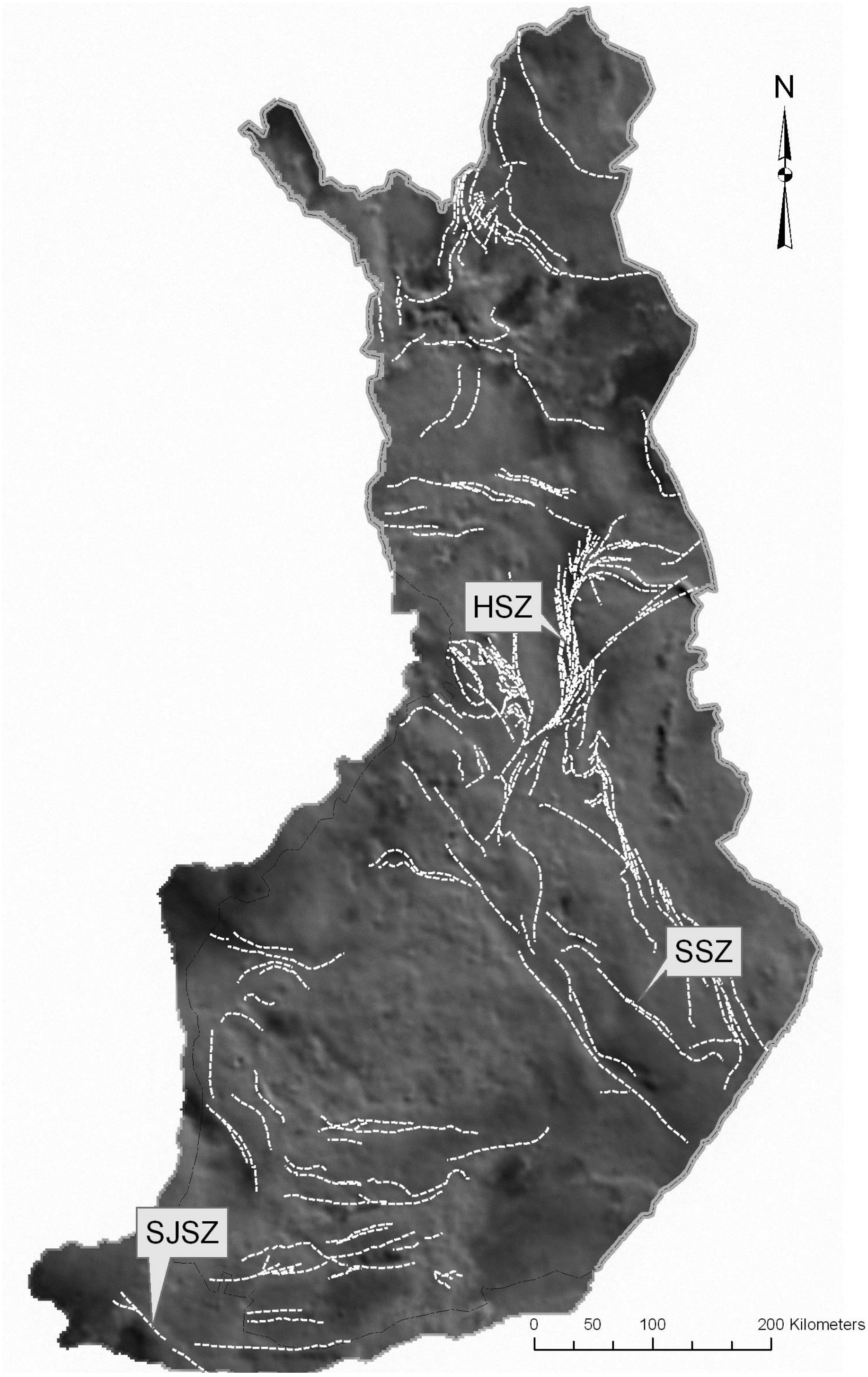

Figure 1. Ductile shear zones in Finland and location of Hirvaskoski Shear Zone (HSZ), Sottunga - Jurmo Shear Zone and Savonranta Shear Zone (SSZ). Base map: Gravimetric map of Finland, Geological Survey of Finland.

Thus, the final structure of the HSZ seems to be a complicated mixture of elements created in different crustal depths as a consequence of multiple stages of shear deformation. In addition, the shear structures created by earlier stages may be metamorphosed and overprinted by later deformations, which increases the complexity of their identification.

Savonranta shear zone (SSZ) splits the Karelian Outokumpu area (Vaasjoki et al. 1998) in the NW – SE direction (Fig. 1). The zone is strongly affected by late-stage brittle deformation thus being detectable as a geographic lineament (Talvitie 1979). However, the more remarkable elements within this deformation zone are the diverse ductile elements, mylonites and blastomylonites with migmatitic higher grade fault rocks (Halden 1982, 1983). Wide variation in the fault rock type is an obvious evidence of polyphase evolution at different crustal depths.

The Sottunga-Jurmo shear zone (SJSZ) in SW Finland is a part of a large-scale deformation zone that transects the Palaeoproterozoic Fennoscandian shield in a NW-SE direction. The SJSZ shows polyphase deformation as it was active and reactivated several times during the prolonged orogenic history of Fennoscandia. Deformation occurred in several separate phases between 1.85-1.79 Ga within the ductile regime, and at least once after 1.79 Ga within the brittle-ductile transition zone (Torvela et al. 2008).

The peak metamorphic P-T conditions within the SJSZ are c. 650°C and 7 kbar (Torvela & Annersten 2005) differing from the general P-T trends in southern Finland. The present erosion surface represents mid- to lower mid-crustal levels of the Palaeoproterozoic Svecofennian crust. The shear zone represents a discontinuity between the amphibolite-to-granulite facies, dome-and-basin style crustal block to the north (LSGM, Ehlers et al 1993) and the amphibolite facies rocks with dominantly steeply dipping structures to the south (Torvela & Ehlers 2010a, b.). The different structural styles, thermal histories and lithologies on both sides of the SJSZ thus imply that the SZ outlines a significant crustal discontinuity, and it possibly also represents a terrane boundary.

3. Discussion

Final structures of the above mentioned crustal-scale shear zones are complicated mixtures of products of numerous tectonic events dominated by shear related deformation. Once developed large-scale shear zones seem to have been reactivated over and over again and the products of those deformations offer a useable record for evaluation of dynamics and regional impacts of those tectonic events. Careful examination of kinematic evolution together with determination of prevailed metamorphic conditions and assessment of absolute age relations appear to be worthwhile procedures in this evaluation.

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3D modeling of polydeformed and metamorphosed rocks – an old Outokumpu Cu-Co-Zn mining area as an example

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The present paper summarizes the results of the project “3D/4D modeling – Outokumpu area as a case study” in years 2007-2009 at GTK. The aim of this project was to apply and evaluate 3D/4D modeling software and formulate principles for visualizing and analyzing complicated geological structures typical for the Finnish bedrock. The project aim was to make recommendations for the use of 3D-modelling tools and to recommend processes for 3D / 4D modeling at GTK. The tools and methods were tested by building geological 3D models of targets within Outokumpu area by combining geological and geophysical data, application of geomathematical methods (e.g. statistics) and by using of different geological software. The Outokumpu area was chosen because it includes several old ore deposits which have been studied and mined since the beginning of 1910. An important data source was the GEOMEX project done between 1998 and 2003 which studied the geological character, evolution and metallogenic potential of the Outokumpu region in Northern Karelia.

Keywords: 3D modelling, Outokumpu, Fennoscandia

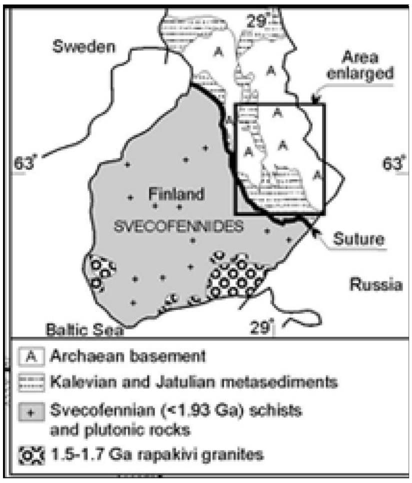
1. Data and tools

The Outokumpu area (Figure 1.) was chosen as a case study for the 3D/4D project (at GTK 2007-2009) because of a long history of research, and hence, availability of data from several sources. The mining history of this area extends from 1913 to 1988 involving the exploitation of three major deposits Outokumpu, Vuonos and Luikonlahti. Geological and geophysical data the were used in 3D modeling of the Outokumpu assemblage, the Keretti and Vuonos ore bodies and rock walls in an open pit at Horsmanaho. The aim was to present 3D models using different approaches.

The applied software packages have different means for interpolation and inference of geological structures such as lithological interfaces and faults. Interpolation methods included simple triangulation in connecting features from geological sections into 3D objects as well as geostatistical methods applied to potential fields in order to model subsurface geological structures in 3D.

It was concluded that different approaches were needed in order to model complex contact relationships within Precambrian rocks. It is not always possible to find suitable sections for realistic 3D modeling of complicated fold patterns. On the other hand, tectonic data along drill holes are usually missing in the descriptions. The possibility of importing different data sets and models into the same platform easies the 3D interpretation of structures. On the other hand, the use of different approaches and data from several sources prerequisites a detailed planning and documentation of the 3D modeling process.

A)



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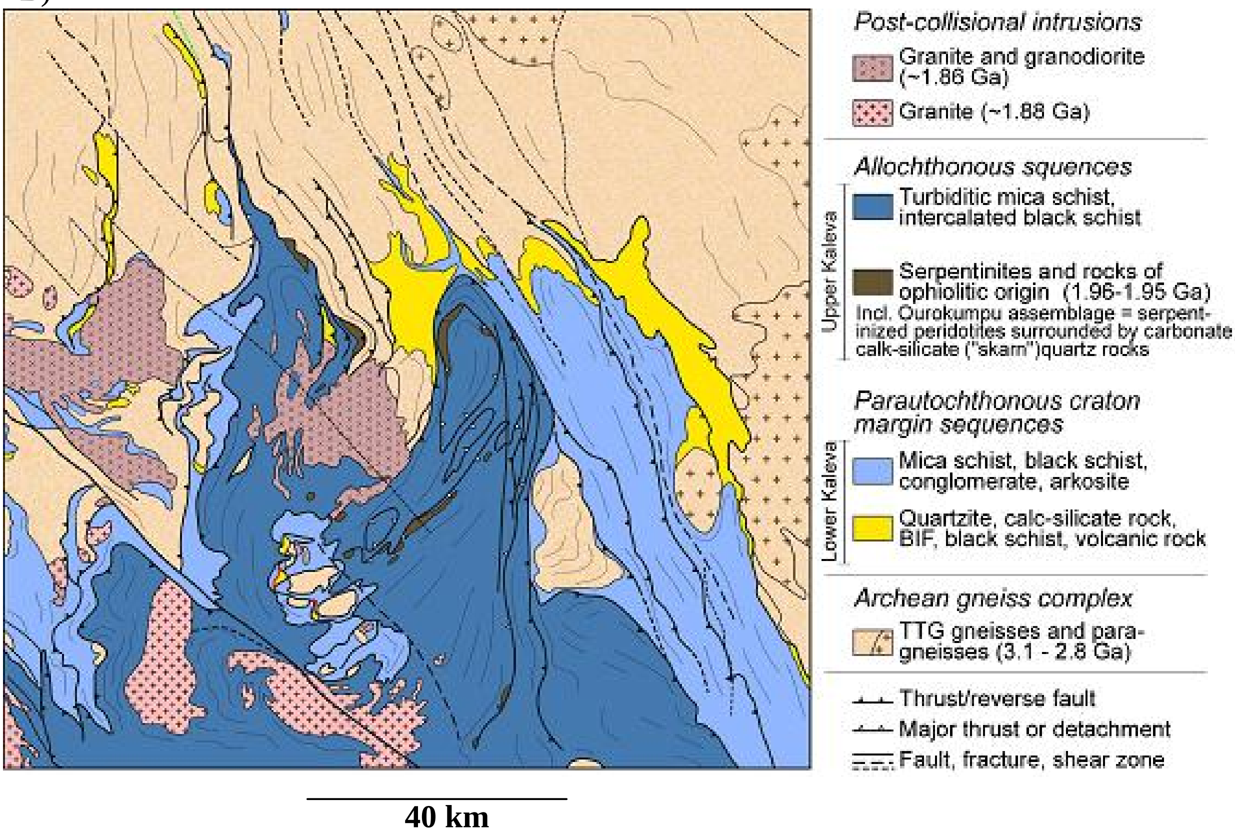

Figure 2. A) Location of the Outokumpu area; B) Regional geology after Korsman et al. (1997).

2. 3D geological modelling process

The starting point of a 3D modeling process is most often an interpretation or knowledge of the subsurface geological structures. Before any further steps the major discontinuity structures, faults and shear zones, should be known and their relative ages defined. These structures are the key elements because they divide the bedrock into separate units.

Collection and georeferencing geological data is perhaps the most time consuming part of the 3D modeling. Unifying and correlating the geological rock units from maps with those appearing in drill holes may be a difficult task. The geological map of the Outokumpu area (Gáal and Koistinen, 1973), for example, contains six rock types, whereas about 55 different rock names were found for drill hole lithologies. In our study we used already unified rock types from the GEOMEX project (Peltonen et al. 2008). In order to make a 3D model the geology had to be simplified to some extent, this included the selection of the main structures such as major faults and folds, leaving out minor features.

The Outokumpu area has been affected by a multiple-phase tectono-magmatic history with polyphase folding and shearing of different orientations causing complex interference structures followed by polyphase fault tectonics. The Outokumpu association was modelled by using the geological model of Koistinen in Gáal and Koistinen (1973) as a base. The modeling would have been very difficult by using solely drill hole data. The knowledge of the tectonic orientation and the geological map helped to connect the serpentinite in the sections.

Areas build up of strongly metamorphosed Precambrian rocks often lack an established stratigraphy. Still, there can be found rules how the rock types are distributed relative to each other. Geometrical difficulties are caused by overturned tight folds and lense shaped bodies. Nevertheless, relatively regular metamorphic layered structures are found in Finnish bedrock.

Geological 3D models seldom are fully realistic and this is not the aim either. This is because of lack of data. The right idea of the form of geological bodies helps in order to build a reliable model. The way the software connects the points into a model also affects the result. Polygonal shapes are caused by triangulation and on the other hand smooth interpolated surfaces may leave important details out. It should be kept in mind that all model parts outside the hard data points are a result of interpretation.

In the case of Finnish bedrock and available data the following work flow could be recommended:

1. Compilation of data and georeferencing.
2. Modeling surfaces of the youngest discontinuities that cut the geological formation under study
3. Digitizing of sections in case of plenty of data and fairly linear bodies. In the case of complicated bodies the more sophisticated use of tectonic orientation and contact points should be used. The latter approach is also recommended if there is very little knowledge of the subsurface and the model is based solely on the geological intuition or theory.
4. Use of geophysical data through forward modeling in checking whether or not the obtained model reproduces the measured potential fields or signals. This is especially important in the case of seismic surveys.
5. Improvement of the geological model using geophysical inversion, i.e. optimization of the model.

6. Saving the model together with hard data and full step-by-step documentation of the modeling process. Sometimes the 3D model result is treated almost as hard data. As far we are continuing a model outside the actual data points the model is uncertain. It is a result of interpolation, extrapolation and interpretation and also reflects the ideas and theories of the time it is done. Therefore, it is more important to save the associated hard data and the document the 3D modeling process together with underlying theories than saving the geometric 3D objects alone.

In 3D modeling it is important that geologists, 3D modelers, geophysicists are closely working together. The organization of the team work is not just to collect a group of people. This demands the learning of the new way of team work, communication, and dealing with sometimes contradictory ideas.

3D modeling demands a lot of time. Fast computers may create an illusion of easy and fast building of geological 3D models. However, in order to make a 3D model as reliable as possible a lot of data collection, georeferencing and digitizing has to be done.

3. Conclusions

The Outokumpu area has been studied since 1910 and, therefore, the data sources consists of old maps and section drawings. Because the mining areas are covered by thick layers of Quaternary deposits, there are only very few outcrops available. Hence, section drawings and drill hole data formed the main geological data source. The Keretti and Vuonos mines are closed and unaccessible.

The Outokumpu area has a complex geology. The structural architecture, the tectonic evolution, and the ore development are far from resolved. A purely technical visualization using data sets without involving geological ideas and interpretations and considering different view points is impossible. It is also evident that the Outokumpu 3D model strongly depends on the interpretation of the regional deformation and alteration history so that a different interpretation of the geological history could lead to a quite different 3D model.

However, the interpretation of structural features is easier by using modern 3D modeling software. The data and models from different sources may be studied on the same platform. Tectonic orientations may be used directly in 3D modeling. Still the data from different sources in various scales and resolutions do not make the modeling process easier but instead demands a close co-operation of individual specialists of different expertise in geology in a research team. The understanding and documentation of the each step of the modeling process is essential.

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Lineation map of Finland – a project

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The purpose of the lineation map project was to produce a lineation map of Finland. Almost 74 000 lineation measurements were used and the maps cover the whole of Finland with some minor gaps mostly due to insufficient or incoherent data. The final interpretations of the map are yet to be made.

Keywords: lineation, map, Finland

1. Introduction

The lineation map project is a part of a large-scale 3D deformation project, which studies the crustal effects caused by the collapse of the Svecofennian orogeny. The original project area covered Finnish territory south of Tornio, but soon expanded to cover Finland completely.

It is suggested that the stretching lineations in Finland are often related to extensional processes during the Svecofennian orogeny (Korja et al., 2010). Therefore the lineation map can help determining the overall kinematics of the extensional processes and structures in Finland.

2. Data processing

The project was started by drawing 90x90 km, 30x30 km and 10x10 km square grids over the map area with AutoCAD Map 3D. Each grid square was labelled with a letter-number combination (Fig. 1). The lineation data was divided into squares according to their coordinates. All the coordinates were converted into Uniform Coordinate System, Zone 3 for easier data processing.

AB-7 (1-9)	AB-8 (1-9)	AB-9 (1-9)
AB-4 (1-9)	AB-5 (1-9)	AB-6 (1-9)
AB-1 (1-9)	AB-2 (1-9)	AB-3 (1-9)

Figure 1. Square AB (90x90 km) divided into 9 smaller squares (30x30 km). Each of these is divided into 9 smaller squares (10x10 km).

The data handling started with the KALPEA-database lineation measurements received from the Geological Survey of Finland (approx. 31 000 measurements). The paper archives at the Geological Survey produced ~12 000 additional measurements. The OKU-database contained approx. 27 000 lineation measurements. Almost 2000 measurements were collected from old 1:400 000 map sheets and ~2600 from scanned old Rautaruukki field observation forms. Additional observations in smaller amounts were received from people connected to the project. Finally, the Geological Survey's web-published reports were used for finding lineation references. The final map contains nearly 74 000 lineations from all over Finland.

The digitized measurements were plotted as stereonet diagrams with the GEOrient program. These diagrams were made for the 30x30 km and 10x10 km squares (Fig. 2). The mean azimuths and plunges from the diagrams were drawn as arrows into the AutoCAD Map 3D with the corrected magnetic declinations (received from the Finnish Meteorological Institute for June 2009). The square was left blank if the mean azimuth and plunge in the diagram was incoherent or contained insufficient observations.

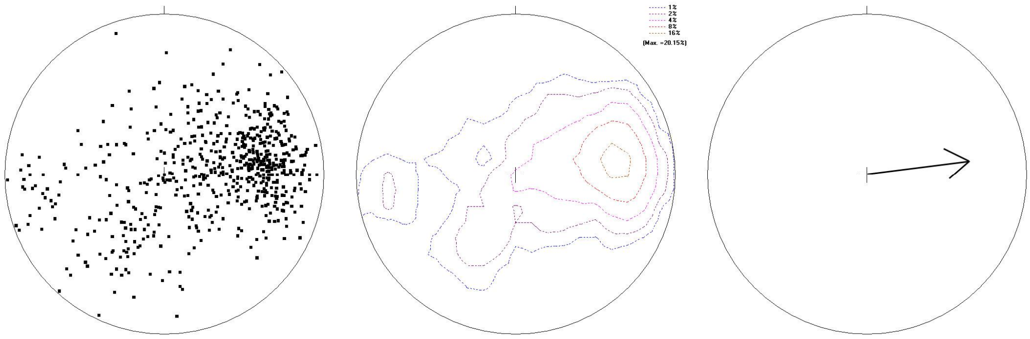

Figure 2. An example from square AB-9 with 650 observations. Left: azimuths and plunges, middle: contour lines, right: mean azimuth and plunge.

3. Results

The final maps contain around 74 000 measurements and cover most of Finland. The biggest gaps in the 30x30 km and 10x10 km maps are in northern Ostrobothnia (Fig. 3). The Wiborg batholith and areas in northern Lapland also contained only very few observations. In western Finland and around the eastern border the gaps are fewer. The final interpretations of the maps are yet to be made.

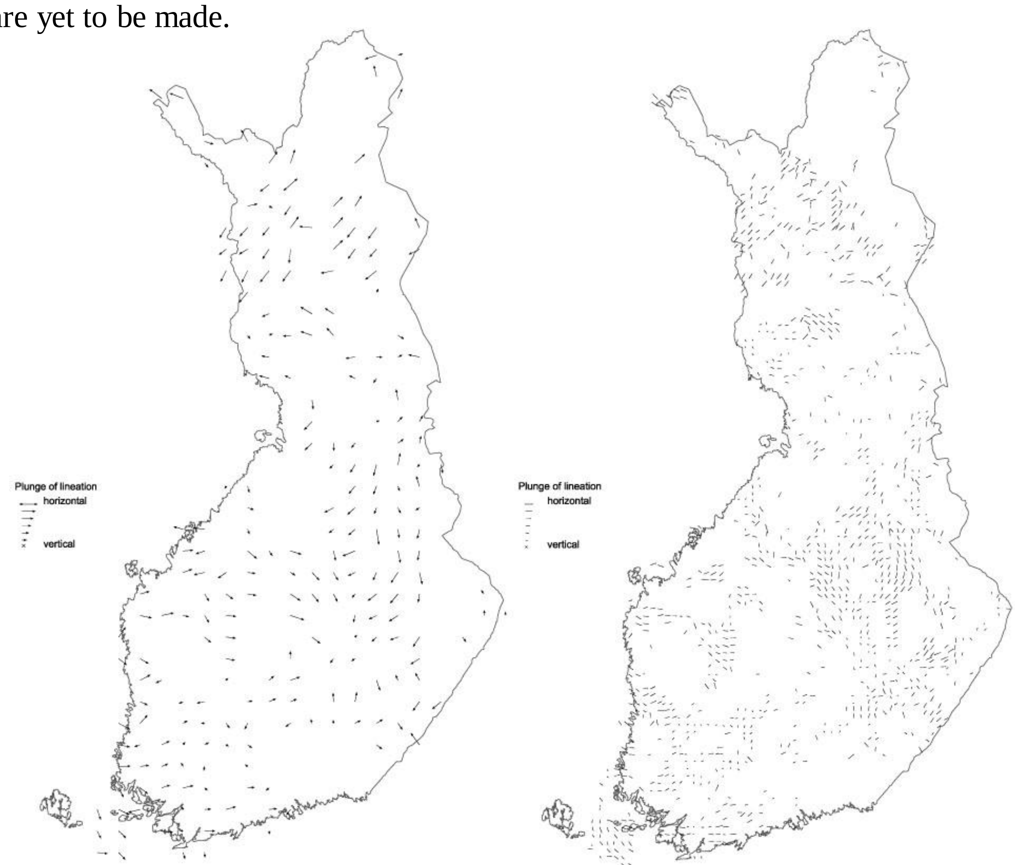

Figure 3. Lineation maps of Finland. Left: 30x30 km grid, right: 10x10 km grid. Arrow direction indicates azimuth and arrow length indicates plunge (the longer the arrow, the shallower the plunge).

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Geophysical Studies of the El'gygytgyn Impact Structure

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Impact cratering is a very important process in planetary evolution. The exact mechanisms by which impact craters form are yet to be fully understood. The study of impact structures is done by observations of such structures formed on terrestrial bodies in the solar system. There are about 180 known impact structures on Earth. From these less than five have formed on volcanic targets. The 3.6 Myr, 15 km diameter El'gygytgyn impact structure in northeast Russia, is the only known impact structure which has been formed on a siliceous volcanic target on Earth. In 2008/2009 the International Continental Scientific Drilling Program (ICDP) drilled the El'gygytgyn crater. The University of Helsinki group, which has also been involved in previous ICDP projects, is currently undertaking the measurements of the physical properties of samples obtained from the ICDP drill core from El'gygytgyn.

Keywords: Impact cratering, volcanic target, El'gygytgyn, ICDP

1. Impact Structures on Earth

Hypervelocity interplanetary collisions have played a fundamental role in the formation of the solar system. Evidence of these collisions can be seen on the surface of planets and moons which have been scared by these impacts. The Canadian Earth Impact Database (http://www.unb.ca/passc/ImpactDatabase/) gives a list of proven impact sites on Earth. The close to 180 impact structures on the Earth, (Figure 1), is much less than the number found on the Moon, Mars or Mercury. This is a result of the active geological and tectonic processes which have reshaped the Earth's surface and erased large parts of the cratering record. The craters on Earth range from less than 10 km up to 300 km in diameter. The oldest known impact structure on Earth, the Suavjärvi crater, is over 2 billion years old, but most of the known structures are less than 100 million years old, as can be seen from database given above.

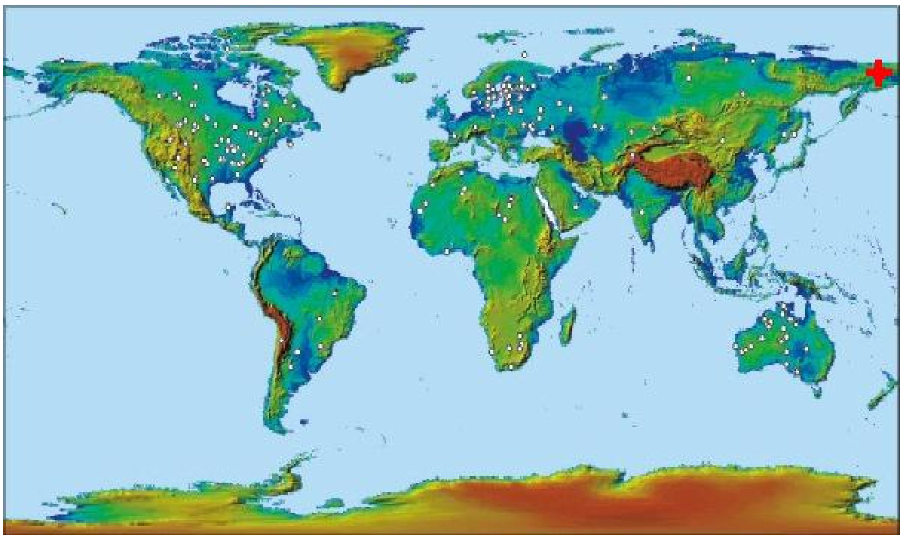

Figure 1. Location of impact structures on Earth (modified from the Canadian Earth Impact Database). The cross indicates the location of El'gygytgyn.

There are two basic types of impact craters, simple and complex. Simple craters are bowl shaped and have diameters ranging from 2 to 4 km (Koeberl and Milkereit 2007). Both types of craters have an outer rim and are filled by fallback ejecta and slumped materials from the walls and crater rim during the early phases of crater formation. This infill usually includes brecciated and fractured rocks as well as impact melts. Complex craters are characterised by a central uplift, formed from the rebound of material which may have been initially buried at significant depth (Koeberl and Milkereit 2007). This central uplift usually consists of dense basement rocks and severely shocked rocks. As it is usually more resistant to erosion, the central uplift structure is sometimes the only remnant structure of a crater which can still be identified (Koeberl and Milkereit 2007). Larger complex craters may have multiple rings.

2. ICDP Impact Drilling

About one third of the impact structures on Earth are not exposed on the surface (Koeberl and Milkereit 2007) and can only be studied by drilling or geophysical methods. The confirmation of the impact origin of a structure can only be done by petrographic and geochemical studies. Hence it is necessary to obtain samples of the subsurface. Even in the case of structures which may be exposed at the surface, subsurface samples are required to investigate the deeper structure and the shock penetration. Therefore the drilling of impact structures is necessary to obtain important information about the structures.

The International Continental Scientific Drilling Program (ICDP) is a scientific organisation, which partly aims to assist geological and geophysical research. It undertakes both shallow and deep drillings to address specific research topics. To date the ICDP has drilled 4 impact structures; (i) the K-T boundary aged Chicxulub structure in Mexico in 2001/2002, (ii) the Bosumtwi structure in Ghana in 2004, (iii) the Chesapeake Bay structure in USA in 2005 and (iv) the El'gygytgyn structure in northeast Russia in 2008/2009.

The Solid Earth Geophysics group at the University of Helsinki has been involved in all four projects. Elbra et al. (2007) present the findings from the Bosumtwi structure and Elbra et al. (2009) present those from the Chesapeake Bay structure.

3. The El'gygytgyn Impact Structure

The 3.6 Myr, El'gygytgyn impact structure is located in the northeast of Siberia, Russia (Figure 1). It has a rim to rim diameter of 15 km. In the basin of the crater there is the Lake El'gygytgyn. This impact structure is one of only three confirmed impact structures formed on volcanic targets. The other two structures are the Lonar structure, India, and the Logancha structure, Russia. Currently investigations regarding a fourth possible structure in Japan are ongoing (D. Badykov, personal communication). El'gygytgyn is unique in that the target sequence was made of siliceous, volcanic rocks (Gurov et al. 2006).

Broadly speaking, impact craters are formed due to the effect of shock mechanisms generated by the hypervelocity collisions. In order to test and refine present theories about shock propagation, studies on different target sequences are required and so the uniqueness of the El'gygytgyn target sequence is highly valuable.

Prior to drilling, many studies on the El'gygytgyn structure have been made. Gurov et al. (2006) present the known characteristics of the structure. Figure 2, after Gebhardt et al. (2006), shows a model for the formation and structure of the El'gygytgyn crater. The ICDP drilling occurred at the central uplift. The aim was to investigate in situ impactites and melt bearing breccias. Since the target sequence is made from volcanic melt, one of the main challenges of El'gygytgyn is to distinguish between impact and volcanic melt. Geophysical and petrophysical data can be useful for this.

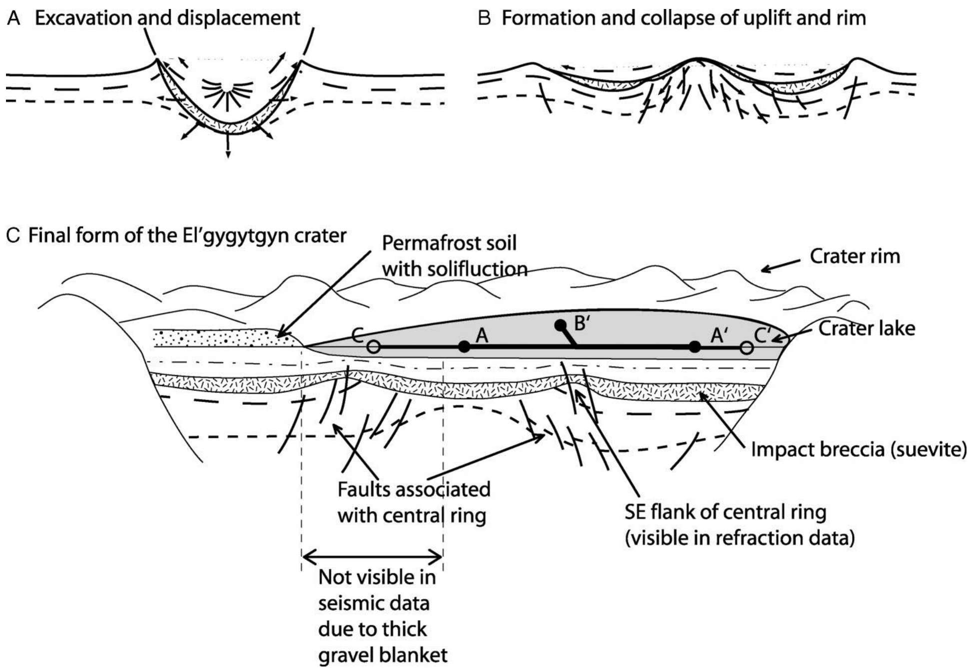

Figure 2. Model for the formation and structure of the El'gygytgyn crater. (After Gebhardt et al. 2006)

4. Petrophysical Measurements

In March, 2010 the ICDP accepted the University of Helsinki Solid Earth Geophysics group's research proposal. The following physical properties are currently being measured by the group:

- dry and wet bulk densities,
- porosities,
- seismic velocities and impedances and their attenuation,
- magnetic susceptibility, its anisotropy and its dependence and temperature,
- natural remanent magnetization (NRM), its nature, origin and demagnetization, and
- electrical conductivity.

The knowledge of the physical properties through the drill core is crucial, since models strongly depend on the contrasts in these properties, both between the impact structure and its unshocked surroundings and within the structure itself. These properties are extremely relevant in the study of the shock penetration of the impact. In the case of El'gygytgyn they can also help to distinguish between impact and volcanic melt.

The results of these studies will be presented in the first common ICDP meeting on the El'gygytgyn structure.

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Comparison between PGE contents in Kaapvaal and Karelian sub-continental lithospheric mantle, based on kimberlite xenoliths and ophiolite samples

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Keywords: platinum-group elements, Kaapvaal craton, Karelian craton, Jormua, sub-continental lithospheric mantle

1. Introduction and analytical procedure

The Kaapvaal craton represents the best studied cratonic nucleus on Earth, owing to an abundance of mantle xenoliths in kimberlites. Study of mantle xenoliths from the Karelian craton has begun during the last decade, after the discovery of several diamondiferous kimberlites (Peltonen et al., 1999). Additional mantle samples are available from the Jormua and Outokumpu ophiolites. Many mantle samples derived from the Kaapvaal and Karelian sub-continental lithospheric mantle (SCLM) are depleted in basaltic components (e.g. Al_2O_3, CaO, Sc). Based in part on Re/Os studies, this has been modeled by predominantly Archean extraction of komatiitic magma (Boyd, 1989; Griffin et al. 2009; Walker et al., 1989; Peltonen et al., 2003). Subsequently, the SCLM of both cratons has seen multiple episodes of refertilisation with LREE-LILE enriched fluids and/or alkaline basaltic and kimberlitic melts of asthenospheric and lithospheric derivation (Erlank et al., 1987; Peltonen et al., 1999).

One of the problems in constraining the origin of the SCLM is that our knowledge of its composition is almost entirely based on xenoliths brought to surface by kimberlite, and the assumption that such xenoliths are representative of the craton as a whole. Moreover, relatively few (<60) Kaapvaal xenoliths have been analysed for the complete PGE spectrum, mostly from Premier (Maier et al., 2005) and Lesotho (Pearson et al., 2004). In the present study, we aim to establish whether it is possible to produce a representative average PGE composition of the Archean SCLM of the Kaapvaal and Karelian cratons. This could potentially allow us to constrain any secular variation in the composition of the mantle (c.f. Maier et al., 2009). We have thus determined whole rock PGE contents of more than 103 xenoliths from 15 kimberlite pipes on the Kaapvaal craton. In order to evaluate inter-cratonic PGE variability we have also determined PGE in 20 xenoliths from 5 pipes in the Karelian craton, and ca 15 samples from the Jormua ophiolite. The PGE in most Kaapvaal samples were determined by ICP-MS after nickel sulfide fire assay and tellurium co-precipitation at Cardiff University. The PGE in the samples from Gibeon, Lesotho, and Premier were analysed by the same method, but at the Geological Survey of Finland.

2. Results

The samples have highly variable concentrations of Pt and Pd, but average Pt and Pd contents of the peridotites are similar in both cratons examined and they also show considerable similarity with SCLM peridotites from the Slave craton (Irvine et al., 2003)(Table 1). In most Kaapvaal and Karelian peridotites Pt and, in particular, Pd are strongly depleted relative to the IPGE. Average Pt/Pd ratios in the samples are twice those of PUM (Pt/Pd = 2.2 - 2.3, vs 1.07

in PUM), but in our Lesotho xenoliths Pt/Pd is much higher. Rhodium is less depleted relative to PUM than Pt and Pd. All PPGE are depleted relative to shallow mantle rocks (basalt-borne xenoliths) and modern orogenic peridotites. The concentrations of the IPGE in our xenoliths are equally highly variable (Fig. 1), but on average these elements occur in similar concentrations in the Kaapvaal and Karelian (and Slave) cratons, and modern primitive mantle. Amongst the few Kaapvaal and Karelia samples with higher Ir and Ru contents than PUM, most are from Jagersfontein (Fig. 1). Several samples have negative Ru anomalies, and a small number are highly depleted in all PGE (< 0.1 x PUM). The correlation between PGE and indices of melt depletion (i.e. Al_2O_3) is poor. This may be due to contamination with host kimberlite or mantle metasomatism by silicate and/or sulfide melt (Pearson et al., 2004; Maier et al., 2005). We found no systematic variation in PGE contents and distribution patterns between xenoliths from Group 1 and Group 2 kimberlite pipes, or between on-craton and off-craton (Gibeon, Markt, Melon Wold) localities.

3. Review of previous work

Past PGE studies on mantle xenoliths from the Kaapvaal and other cratons showed a pattern of distinct depletion in Pt and Pd, and to a lesser degree Rh, relative to the IPGE (Os-Ir-Ru) and primitive mantle values (e.g. Pearson et al., 2004; Maier et al., 2005). The data were interpreted to reflect incompatible behaviour of Pt, Pd and, somewhat less so, Rh, during mantle melting, and compatible behaviour of the IPGE. The IPGE and Rh are largely hosted by monosulfide solid solution (mss) included within olivine, whereas Pt and Pd tend to be mainly hosted by interstitial Cu-rich sulfides (Alard et al. 2001). The mss may represent residual sulfide shielded from dissolution during mantle melting by the olivine, whereas the Cu sulfides crystallized from metasomatic fluids/melts.

Modelling of melting of sulfide-bearing mantle (e.g. Mungall 2007) suggests that much of the variance in PPGE abundance in the mantle restite occurs within a critical melting range, corresponding to the final disappearance of sulfide. Hence, there should be a broadly bimodal “all or nothing” distribution of PPGE contents of mantle samples, ranging from those with close to their original endowment, representing samples containing residual sulfide but largely independent of how much, to those highly depleted in Pt and Pd representing complete extraction of the sulfide component. IPGE variation in mantle restites should be more complex, and more sensitively dependent on degree of mantle melting, depending on the degree to which the IPGE-rich MSS is shielded by unmelted olivine. This combination of factors leads to the observed patterns of wide continuous variation in IPGE, coupled with discontinuous wholesale depletion of PPGE.

It remains puzzling why there are no terrestrial magmas with IPGE contents higher than PUM. Such magmas should have formed in response to high degree melting and complete dissolution of mantle sulfides. In general, it is our impression that quantitative modeling of PGE behaviour during melting and refertilisation of the SCLM has so far been moderately successful, partly because the nature of the metasomatic agents (particularly their PGE contents) and the high pressure D values of the PGE (Righter et al., 2008) remain incompletely understood, because of retention or entrainment of refractory IPGE-rich phases, and because multiple events of melting and refertilisation that occured over a span of 3 billion years are difficult to unravel. An alternative approach to improve the understanding of the petrogenesis of the mantle consists of compiling the bulk composition of the mantle (e.g. Becker et al. 2006). Their estimate differs from previous estimates mainly in that it has almost twice the amount of Pd (~7 as opposed to ~4). It has recently met with some sceptisism (e.g. Lorand et al., 2008) on the grounds that mantle samples may have had their PGE (particularly PPGE) contents modified by processes of mantle metasomatism.

4. PGE mass balance of the SCLM

We want to take a step back and examine whether a purely mass balance approach can provide some constraints on the PGE budget of the Archean (i.e. rather than the modern) mantle. Maier et al. (2009) suggested that the early Archean mantle had lower PGE contents than the modern mantle. If this were true, then PUM estimates would be of limited use in constraining core-mantle differentiation. Knowledge of the composition of the Archean mantle comes mainly from SCLM derived xenoliths, and interpreting these samples is more complicated than modern samples of oceanic mantle because the former underwent much longer and more complex melting and refertilisation histories that may have affected their PGE contents. In order to gain as representative a composition as possible the sample population has to be large, and we also have to consider the composition of Archean mantle derived magmas (Fig. 2).

The two most important results of our study may be summarised as following: Firstly, The modern mantle (as represented by the PUM estimate) and most primitive mantle-derived magmas (komatiites) have Pt/Pd approximately at unity (Maier and Barnes, 2004; Becker et al., 2006; Barnes and Fiorentini 2008), yet most of our SCLM xenoliths and ophiolite samples have Pt/Pd significantly above unity (average Pt/Pd = 2.2). The mass balance can be satisfied in the following way: Komatiites represent ca 50% partial melting and have ca 10 ppb Pt and Pd each. Assuming 7 ppb Pd in the mantle, this indicates that Pd is not entirely incompatible, as otherwise the magmas should have ca 14 ppb Pd, which is never observed. Retention of some Pd in the mantle during melting is also consistent with the lack of any terrestrial magmas that have more than ca 20 ppb Pd; At 20% partial melting and complete incompatibility of Pd, one would expect to find basalts with up to 35 ppb Pd. Thus, ca 2 ppb Pd remain behind, consistent with observations, and indicating little reintroduction of Pd. Platinum is apparently equally not entirely incompatible during mantle melting: Assuming 7.5 ppb Pt in PUM, extraction of komatiite (50% partial melting) with 10 ppb Pt would result in 5 ppb Pt to remain in the mantle residue, consistent with observations. The somewhat more compatible behaviour of Pt relative to Pd could be explained by the presence of insoluble, refractory Pt-Fe alloys (Alard et al., 2001). It is apparent that the model mantle residues (2ppb Pd and 5ppb Pt) have broadly similar Pt and Pd contents as the observed SCLM samples, suggesting that the SCLM has not been significantly refertilised with Pt.

Secondly, the average IPGE contents of the Kaapvaal and Karelian (and Slave) SCLM are broadly similar to PUM (Fig. 1). This result is particularly remarkable because basaltic and komatiitic magmas contain IPGE contents systematically and significantly below PUM (Fig. 1b), due to the compatible behaviour of the IPGE during mantle melting. As a result, IPGE contents in residual mantle samples should be higher than in PUM, no matter how complex the melting and refertilisation history of the mantle was. Several possible explanations may be considered to explain this mismatch. (i) Most of our samples represent mantle that has been refertilised by alkaline or basaltic melts. If these magmas were IPGE poor, as expected for small degree melts, this could have led to dilution of IPGE contents. However, significant resetting would require introduction of tens of percent of metasomatic material, inconsistent with the relatively high MgO contents of the samples (Fig. 1). (ii) The missing IPGE of the SCLM could be hosted in chromite pods that were disaggregated during kimberlite ascent. However, chromite control of the IPGE is inconsistent with the lack of correlation between Cr and IPGE in our samples. (iii) Some of the more IPGE depleted samples could represent cumulates from magmas ascending through the mantle (Niu et al., 1997), or from the oceanic lithosphere subsequently accreted to the craton roots (Peltonen et al., 1999; Canil and Lee, 2009). (iv) The SCLM could represent buoyant residues of originally more fertile mantle residues from which Fe-rich components had segregated gravitationally (Arndt et al., 2009). However, it is not apparent why the IPGE would be preferentially concentrated in the more Fe rich component. (v) The Archean mantle could have been depleted in IPGE relative to modern mantle (Maier et al., 2009). (vi) The present PUM estimate could be too high. This is suggested by the fact that the IPGE contents of modern sub-oceanic mantle, as sampled by orogenic peridotite massifs, are also lower than expected from melt residues; Many oceanic peridotites are residues of ca 8-18% melting of MORB basalt (Lorand et al. 2008) and thus, if the IPGE behaved in a compatible manner, the peridotites should have ca 10-20% higher IPGE contents than PUM. The observed levels are slightly below PUM (Fig. 1) suggesting that current PUM estimates of the IPGE are ca. 10-20% too high. The mass balance mismatch is less significant than in the case of the Archean mantle, but larger than analytical error. (vii) The mantle could contain a restititc eclogite component (Bedard, 2005).

Griffin et al. (2009) recently argued that kimberlites preferentially sample refertilised and thus non-representative sections of the sub-continental lithospheric mantle. If this were true then our estimate of the composition of the SCLM would be erronous. The model of Griffin et al. (2009) can be tested by a detailed examination of PGE contents in the Jormua ophiolite of Finland, believed to represent a contiguous sub-continental mantle section (Peltonen et al. 2003). This work is in progress, with preliminary data from more than 20 samples showing broadly similar PGE concentration patterns as the Karelian and Kaapvaal xenoliths.

5. Summary

PGE data from more than 120 samples of kimberlite-hosted mantle peridotite xenoliths and samples from the Jormua SCLM indicate that the SCLM contains significantly lower Pt and Pd contents than the presently accepted PUM estimate, and broadly similar IPGE contents. The Pt and Pd depletion can largely be explained by the incompatible behavior of these elements during mantle melting with complete consumption of sulfides. In contrast, experimental results and data from terrestrial mantle-derived magmas indicate that the IPGE behave in a compatible manner during mantle melting. Therefore, these elements should be significantly enriched in mantle residues. The mismatch could suggest that many SCLM samples represent cumulates. Alternatively, the current PUM estimate for the IPGE could be erronous.

Table 1: Average PGE contents of Kaapvaal and Karelia xenoliths, and the Jormua SCLM section

	Os	Ir	Ru	Rh	Pt	Pd	Au	n
Kaapvaal								82
avg	3.12	3.67	6.62	0.95	4.34	1.87	1.14	
stdev	1.63	1.62	3.76	0.46	3.01	2.10	1.21	
Karelia								14
avg	4.06	3.68	6.44	0.91	4.11	1.69	1.09	
stdev	3.21	2.28	2.20	0.39	3.24	1.85	1.02	
Jormua								19
avg	3.82	3.40	7.03	0.90	3.13	1.47	1.02	
stdev	1.21	0.95	1.61	0.36	2.06	1.26	0.52	
Somerset Island (Irvine et al., 2003)								
avg	3.56	3.52	6.69		3.40	1.91		
stdev	1.30	1.24	2.29		2.02	1.53		

(1) Several samples contained Pd below the dl (0.5 ppb). Ignoring these samples would result in a misleadingly high average Pd content. We therefore assumed a value of 0.5 ppb for these samples.

(2) Kaapvaal average excludes MARID samples

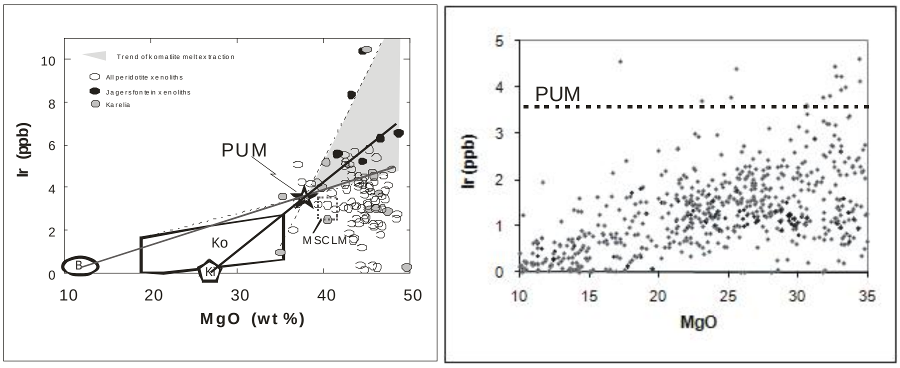

Figure 1a: Iridium and Ru contents of Kaapvaal and Karelian mantle xenoliths, plotted vs MgO. Extraction of basaltic (B) or komatiitic (Ko) magma results in Ir and Ru contents higher than PUM. Refertilisation with basalt or kimberlite (K) can lower Ir and Ru contents, but results in lower than observed MgO contents. MSCLM=modern sub-continental mantle. **1b:** PGE contents in terrestrial basalts and komatiites. See text for explanation. Data from Barnes and Fiorentini (2008) and Maier et al. (2009).

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Paleostress and 3D slip and dilation tendency analysis of fault zones – a case study from Olkiluoto, SW Finland

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Paleostress analysis of fault zone BFZ099, located at Olkiluoto Island, SW Finland, reveals four distinct slip events and corresponding stress states. These events correspond to NNW-SSE-compression in a strike-slip regime, NE-SW-compressions in a strike-slip regime, radial extension, with the main extension direction approximately NE-SW and finally, to WNW-ESE-compression in a thrust regime. The slip tendency analyses indicate that the period of NE-SW extension may correspond to the primary faulting event as the shear traction values were generally high for the whole fault geometry at the time of the slip. The lower slip tendency values for the other three slip events suggest that these are later reactivations and slips may have occurred under conditions of reduced fault strength. The use of 3D slip and dilation tendency maps show the locations of high dilation potential on the investigated faults and is may therefore be a useful tool in mineral prospecting and for the estimation of hydraulic behaviour of the faults under present day stress conditions.

Keywords: Paleostress, fault zone, slip tendency, dilation tendency, Olkiluoto, Finland

1. Introduction

When attempting to predict the properties and behaviour of fault zones it is essential to understand the structural history and evolution of the zones, but in complex geological settings, such as Precambrian shields, poor exposure commonly hinders acquisition of sufficient and relevant structural data. However, at Olkiluoto Island in SW Finland, which is the site planned as a deep repository of spent nuclear fuel, major low-angle fault zones have been in intersected by numerous drill holes and an extensive fault slip data set has been collected from a volume corresponding to the damage zone of the faults.

In this study, these fault slip data were further used in assessing the reactivation history of fault zone BFZ099 at Olkiluoto (Mattila et al. 2008) and in the computation of paleostresses related to these reactivations. Furthermore, based on the computed stresses, slip and dilation tendencies were calculated for the whole fault zone geometry and the results were used to build 3D shear trajectory and slip and dilation tendency maps for the fault.

2. Methodology

Paleostress analyses aim to define the state of stress during the time of faulting from fault slip data and are based on the simple assumption that slip on a plane occurs in the direction of resolved shear stress (Wallace 1951, Bott 1959). However, under natural conditions the slip directions on fault planes are typically dispersed and therefore algorithms developed for paleostress analysis try to find the best-fit stress tensor by minimising the angular misfits between the observed and theoretical slip directions. This is accomplished by iteratively searching for stress directions (σ_1, σ_2, σ_3) and stress ratios ($R=(\sigma_2-\sigma_3)/(\sigma_1-\sigma_3)$) that fill the minimising criteria. It is noteworthy, however, that the analyses result in reduced stress tensors, i.e. the actual magnitudes of the principal stresses are unknown and other means are needed to constrain them.

According to Amontov's law, a cohesionless fault will slip if the resolved shear traction on the fault equals or exceeds friction. The slip tendency of a fault is a derivative of this law and is defined as the ratio of resolved shear traction to normal traction on that plane (Morris et al. 1996):

$$T_s=\tau/\sigma_n$$

If this ratio equals or exceeds the coefficient of static friction, there is a high probability that the fault will slip; the value of slip tendency is only dependant on the geometry of a fault and the given stress state and can also be calculated from a reduced stress tensor acquired through paleostress analysis (Lisle and Srivastava 2004). Dilation tendency depicts the potential of a fault or fracture to dilate in a given stress state and is defined as the ratio of normal traction to maximum differential stress:

$$T_d=(\sigma_1-\sigma_n)/(\sigma_1-\sigma_3)$$

The distributions of both the slip tendency and dilation tendency values can be calculated for the whole 3D geometry of the investigated fault and the results given as 3D distribution maps which can be further utilised in the estimating which parts of the fault are prone to slip or dilate in the applied stress state.

3. Results of the analyses

Fault zone BFZ099 is a major low-angle fault located at the Olkiluoto Island and its geometry has been modelled in great detail by the use of several drillhole intersections, 3D seismics and mise-à-la-masse measurements (Mattila et al. 2008). The fault slip data from the zone cannot be explained by a single slip increment or stress state and the computed paleostresses reveal that the zone has gone through at least four distinct slip events, as shown in Figure 1. Stress state A corresponds to NNW-SSE-compression in a strike-slip regime, stress state B to NE-SW-compressions, similarly in a strike-slip regime, C to radial extension, with the main extension direction approximately in NE-SW-direction and D to WNW-ESE-compression in a thrust regime.

The 3D slip tendency maps for the BFZ099 reveal that the highest shear traction values for the zone are given by the stress state C (Figure 2), suggesting that this phase may represent the primary slip increment on the fault. The other three stress states induce in general relatively low slip tendency values, which indicates that these are later reactivations and slip may have occurred under the condition of reduced fault strength. Dilation tendencies are correspondingly higher for the stress states A, B and D than for C. An example of a dilation tendency map is given in Figure 3 for the stress state B, together with the resolved shear directions. The map indicates that the dilation tendency in the given stress state was generally high and the lowest values were located in the NE-portion of the fault.

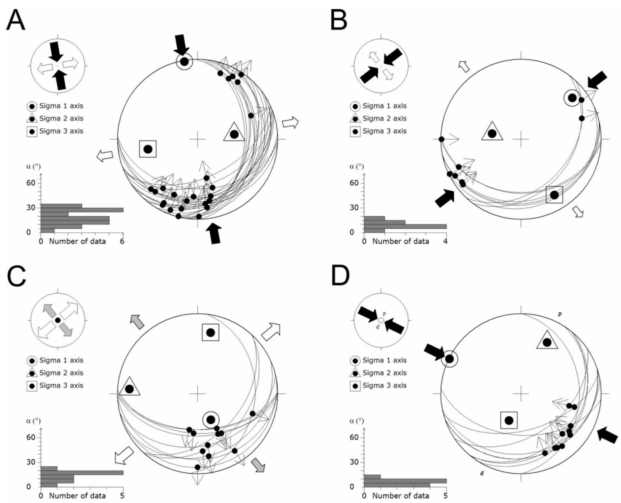

Figure 1. Paleostresses computed for the fault zone BFZ099

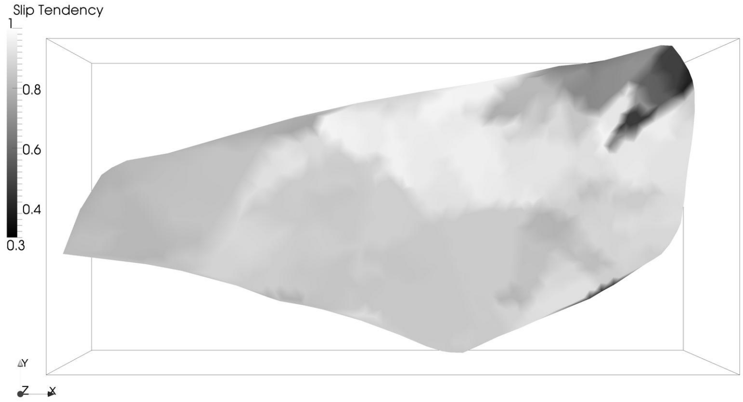

Figure 2. Slip tendency values on the fault zone BFZ099 in 3D, corresponding to the stress state C in Figure 1. View from above.

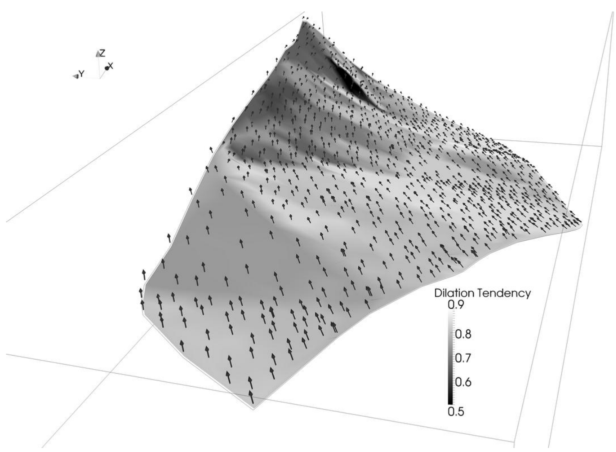

Figure 3. Shear directions and dilation tendency values on the fault zone BFZ099 in 3D, corresponding to the stress state B in Figure 1. View towards NE.

4. Application of the methodology

Slip tendency analysis have been typically applied in the analysis of slip potentials of faults in earthquake prone areas. However, by the combination of paleostress analysis and both the slip and dilation tendency analyses, it is also possible to estimate which parts of a fault were likely to dilate and therefore more prone to be locations for fluid flow and mineralisations. The methodology presented here can therefore be considers as potential tool in mineral prospecting, and if present day stress states are used, for the estimation of hydraulic behaviour of the faults, which is an important aspect of safety evaluation of planned radioactive waste disposal sites.

5. Summary

Paleostress analysis of fault zone BFZ099, located at Olkiluoto Island, SW Finland, reveals four distinct slip events and corresponding stress states. These events correspond to NNW-SSE-compression in a strike-slip regime, NE-SW-compression, also in a strike-slip regime, radial extension, with the main extension direction approximately NE-SW and lastly, WNW-ESE-compression in a thrust regime. The slip tendency analyses indicate that the period of NE-SW extension may correspond to the primary faulting event as the shear traction values were generally high for the whole fault geometry at the time of the slip. The lower slip tendency values for the other three slip events suggest that these are later reactivations and slips may have occurred under the condition of reduced fault strength.

The presented methodology is also considered as a potential tool in mineral prospecting, and in the estimation of hydraulic behaviour of the faults under present day stress conditions.

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Magnetic data constraining the 1.63-1.45 Ga rifting episodes in the southern Fennoscandian shield

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In order to investigate if the Kopparnäs diabase dyke swarm in southern Finland forms part of the 1.45 Ga rifting event seen in Lake Ladoga area in NW Russia and in Dala area in Sweden, we have compared the magnetic properties (magnetic susceptibility, remanent magnetization, anisotropy of magnetic susceptibility (AMS) and magnetic mineralogy) and paleomagnetic poles with these from the 1.45 Ga diabase dykes at Lake Ladoga, with the 1.46 Ga Tuna diabase dykes in Sweden and with the 1.63 Ga Sipoo diabase dykes in Finland. Rock magnetic tests and SEM studies show that the Fe-oxide minerals are different in the Kopparnäs and Lake Ladoga dykes. Also the magnetic susceptibilities and remanence intensities of the Kopparnäs dykes deviate from those of the Lake Ladoga dykes but show similarities with the Sipoo dykes. The paleomagnetic pole of the Kopparnäs dykes is in closest agreement with the pole from the Sipoo dykes, but deviates from the poles of Lake Ladoga formations and from the poles of the Tuna dykes. The results thus imply that despite to geochemical similarities of the Kopparnäs dykes with the Lake Ladoga basalts, the Kopparnäs dykes are most probably related to the much older rapakivi magmatism.

Keywords: diabase, paleomagnetism, AMS, Mesoproterozoic, Fennoscandian shield

1. Introduction

Based on geochemical data from the Kopparnäs diabase dyke swarm in the southern coast of Finland, it has been suggested that the dykes belong to a regional scale 1.45 Ga dyking event that extends about 1 000 km from Lake Ladoga area in NW Russia through Finland to the Tuna dykes in Sweden (Luttinen and Kosunen, 2006) (Fig. 1). The Lake Ladoga dykes are dated at 1452±12 Ma (Lubnina et al., 2010) and the Tuna dykes at 1.46 Ga (Söderlund et al., 2006), but the Kopparnäs dykes are not dated. Attempts of isotopic dating were carried out in the dating laboratory of the Lund University, Sweden, but due to fine grain size and lack of baddeleyite, the datings were not successfull. Therefore, one of the main aims of our studies has been to get a paleomagnetic dating for the dykes. The dating is done by comparing the paleomagnetic poles with the Lake Ladoga (Lubnina et al., 2010) and Tuna poles, (Bylund, 1985, Piper, 1992) and with the poles from the Sipoo dykes that are related to the 1.63 Ga Onas rapakivi granite (Mertanen and Pesonen, 1995). As the Kopparnäs dykes are located close to the Obbnäs rapakivi granite intrusion, rapakivi age magmatism may form one source also to the Kopparnäs dykes. Magnetic property measurements and investigations on anisotropy of magnetic susceptibility (AMS) were carried out to get new information on the internal structures and magma flow direction of the dykes. The rock magnetic data from the Kopparnäs dykes are compared with those from the Lake Ladoga dykes (Lubnina et al., 2010, Scherbakova et al., 2008).

2. Geology, sampling and laboratory measurements

The E-W trending Kopparnäs dolerite dyke swarm outcrops in a very limited area extending about two kilometers in length and about half a kilometer in width, but according to GTK's aeromagnetic data, similar E-W trending weak anomalies can be seen on two locations under the water. Width of the dykes is about 2 m at maximum, and some dykes show en echelon structures. The dykes are undeformed, vertical or subvertical and cut sharply the Svecofennian 1.90-1.82 Ga gabbros and granodiorite-tonalites. They are typically fine

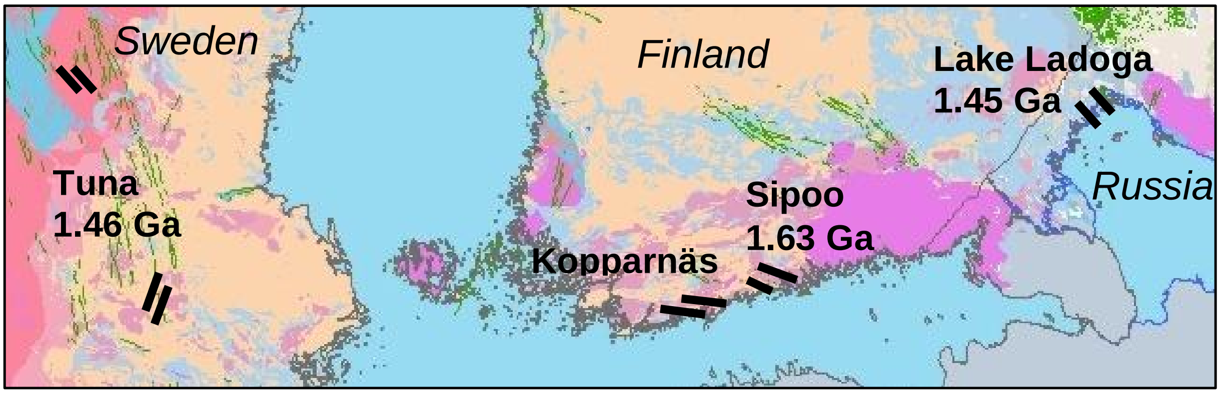

Figure. 1. Locations of the Kopparnäs, Sipoo, Lake Ladoga and Tuna dykes, relevant for this study. Geological map modified after Koistinen et al. (2001).

but in more wider dykes the central parts are coarse grained and show plagioclase laths and ophitic textures. The dykes show low-grade metamorphism which is believed to be related to the late stages of dyke emplacement (Luttinen and Kosunen, 2006).

From the dykes and their baked and unbaked host rocks we took altogether 225 samples from 22 sites by using a mini-drill. The samples were oriented with sun and magnetic compasses. In the laboratory, the samples were cut into two to three specimens. Density and magnetic susceptibility was measured from each specimen before remanence measurements, AMS determinations and measurements of three axis isothermal remanent magnetization (IRM). Polished thin section studies and SEM observations were performed for selected specimens.

3. Results and discussion

3.1. Magnetic mineralogy

According to SEM studies the main Ti-Fe-oxide mineral in the Kopparnäs dykes is ilmenite, which occurs as large separate laths. Only very little amounts of Ti-magnetite was observed. Pyrite is the other common opaque mineral and pyrrhotite is found sporadically. The magnetic mineralogy of the Kopparnäs dykes differs clearly from the Lake Ladoga dykes which contain exsolution lamellae of ilmenite in primary titanomagnetite (Scherbakova et al., 2008).

Rock magnetic tests (Lowrie, 1990) and thermal demagnetizations of the Kopparnäs dykes indicate that the carrier of magnetic susceptibility and remanent magnetization is titanomagnetite. According to these tests the magnetic grain sizes of different Kopparnäs dykes are highly variable while in the Lake Ladoga dykes they are very constant and show the dominance of high coercivity grains that are able to carry stable remanence.

3.2. Magnetic properties and AMS

Magnetic susceptibilities of the Kopparnäs samples vary from weakly magnetic (1 200) to strongly magnetic (94 000 µSI). There is a positive correlation between the thickness of the dykes and magnetic susceptibility, as the thicker dykes (100 cm or more) tend to have higher susceptibilities (> 20 000 µSI) in the centre of the dyke but lower near the contact. Narrow dykes (less than 40 cm) normally do not have high susceptibilities (> 20 000 µSI) or great variation. In the Lake Ladoga samples the average susceptibilities range from 20 000 to 40000 µSI being thus more compact compared to the Kopparnäs dykes. Likewise, the remanence intensities and Koenisgberger ratios (Q-value) of the Kopparnäs dykes show much greater variation than the Lake Ladoga dykes.

The anisotropy degree of the Kopparnäs dykes range between 1.2 and 10% (average 5.4%). Also the magnetic anisotropy shows a variation in correlation to the thickness of the dyke. Profiles across the dykes show that the anisotropy degree decreases towards the centre of the dyke where the susceptibility is higher. Also the AMS shape varies according to the location in the dyke. Foliation dominates in the centre of the dyke whereas lineation dominates in the dyke margins. The studied samples from Lake Ladoga have similar anisotropy degrees as the Kopparnäs samples, but the shapes are predominantly controlled by foliation.

Directional AMS data reveal that in the Kopparnäs dykes the magma flow is in general in the E-W direction, parallel to the direction of the dykes. The magnetic lineation, as referred to the direction of the magma flow, differs from site to site depending on the degree and shape of anisotropy. The foliation plane varies from horizontal to vertical along the direction of the dykes. It is also strongly dependent on the thickness of the dykes and whether the samples are taken from near the contact or from the centre of the dyke. The complexity of the Kopparnäs dykes may indicate that the intruding magma was cooled rather fast and emplaced locally, possibly in association with the rapakivi granite. The directional AMS data from Lake Ladoga dykes on the other hand reveal that the magnetic foliation strikes vertically in the direction of the dykes and magnetic lineation is scattered along the dyke.

3.3. Paleomagnetism

In the Kopparnäs dykes we isolated a normal polarity NE pointing low inclination component in 14 sites (85 samples) and an opposite polarity SW pointing component in three (13 samples) sites. In many cases both polarities occur in a single specimen, shown as a great circle curve from normal to reversed polarity. This behaviour resembles the one that was previously observed in the Sipoo dykes (Mertanen and Pesonen, 1995) where the great circle trajectories go from the Present Earth's Field direction (PEF) to the reversed direction. In the Lake Ladoga dykes the remanence of the dykes is very stable and shows typically only a single component, either of normal or reversed polarity.

The paleomagnetic pole of the Kopparnäs dykes (Fig. 2) is closest to the pole from the 1630 Ma Sipoo diabase dykes (Mertanen and Pesonen, 1995). It deviates clearly from the Lake Ladoga poles obtained from the Valaam sill (Lubnina et al, 2010, Salminen et al., 2007), Salmi basalt and dolerite dykes (Lubnina et al., 2010). The poles from the Tuna dykes (Bylund, 1985, Piper, 1992) in Sweden are highly scattered so that no firm conclusions can be made based on that data.

4. Conclusions

Differences in the paleomagnetic poles and rock magnetic properties of the Kopparnäs dykes with those from the 1.45 Ga Lake Ladoga diabase dykes suggest that the two dyke swarms were not formed in the same rifting event, despite to geochemical similarities. On the other hand, resemblance of the magnetic properties and paleomagnetic poles of the Kopparnäs dykes with those of the Sipoo diabase dykes suggests that the dykes are related to the 1.63 Ga rapakivi magmatism. The present results thus imply that in the Lake Ladoga area in Russia and in the Dala area in Sweden, rifting was a more local event that didn't affect the bedrock of southern Finland.

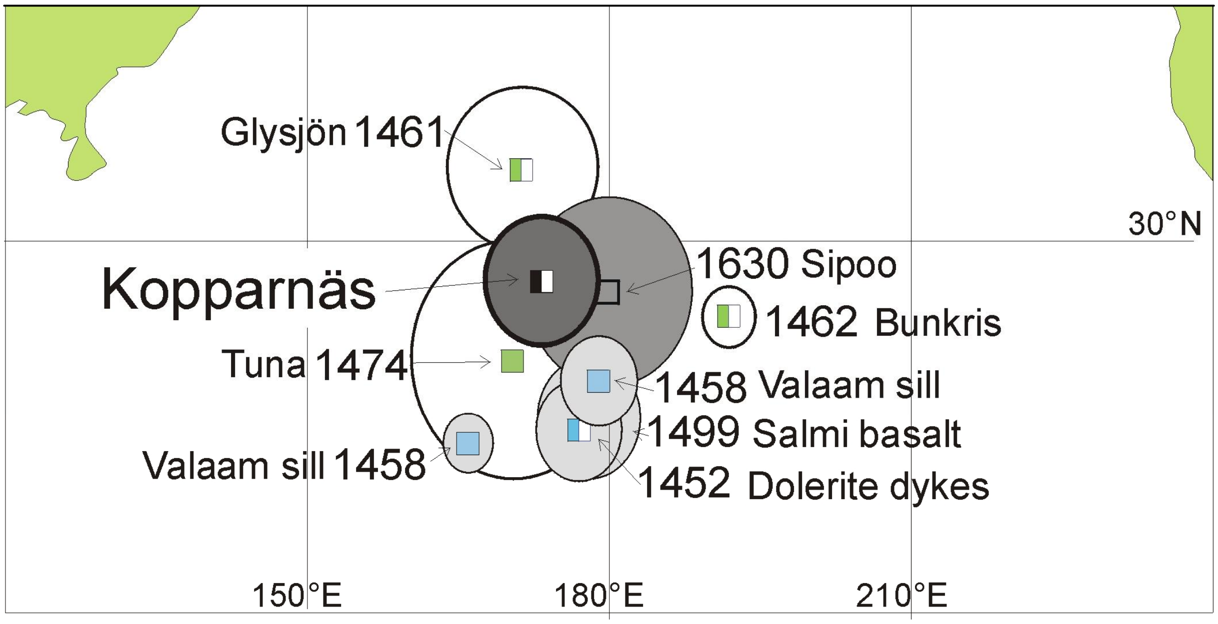

Figure 2. Paleomagnetic poles of the Kopparnäs, Sipoo and Tuna (Glysjön, Bunkris, Tuna) dykes and Lake Ladoga formations (Valaam sill, Salmi basalt, Dolerite dykes). The circles show the A95 confidence limits of the poles. See text for references. Ages are in million years.

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Monitoring crustal deformations in Satakunta using GPS measurements

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The GeoSatakunta GPS network was established in 2003 to obtain information on crustal deformations in the Satakunta region. The network consists of 13 concrete pillars for episodic GPS campaigns and the Olkiluoto permanent GPS station. We have processed the GPS data of three annual campaigns from years 2003-2008 using Bernese 5.0 GPS software. The results were analysed based on the time series of baseline lengths and their temporal variation. On the average the baseline length varied less than 2.0 mm without any evident trend. The estimated velocities were (max 0.3 mm/a) mostly statistically insignificant because of the relatively short time series. For illustration of the results the station velocities were interpolated in a grid and overlaid on the geological data. The possible change seems to take place in the vicinity of geologically interesting features, like the major shear zones and the Jotnian sandstone. More measurement campaigns are anticipated in the future to better constrain the limits for the deformation in the area.

Keywords: GeoSatakunta, GPS time series, crustal deformation

1. Introduction

The Finnish Geodetic Institute, the Geological Survey of Finland, Posiva Oy and municipalities in the district of Satakunta launched the GeoSatakunta research program in 2002 to carry out interdisciplinary studies on the regional bedrock stress field and to apply the results, e.g., in land use planning in the Satakunta area.

The bedrock in the Satakunta area offers exceptional possibilities for the stress field study. The structure and geological evolution of the area is well-known (Paulamäki et al., 2002). It is characterised by younger bedrock units like Rapakivi bodies, Jotnian sandstone and olivine diabases, which cause marked changes in the stress field through variations in densities and structural properties (Huhta and Korsman, 2005). On the other hand, the bedrock experiences a postglacial uplift, about 6-7 mm/a (Mäkinen et al., 2003). The uplift rate changes about 1 mm/a from NW to SE, which should cause some effect over time on the stress field of the crust. In addition, there are other processes listed in (Lambeck and Purcell, 2003), which can contribute to the regional stress field, like spreading of the Atlantic seafloor and its push against the Fennoscandian shield.

As a part of a larger co-operation project we have established a regional GPS network in 2002 to obtain information about contemporary crustal deformation. Geodetic GPS measurements are the most accurate method to determine horizontal movements over a large area. Following the Olkiluoto deformation studies we can achieve a detection threshold of 0.2 mm/a for single-site motion from ten year time series (Kallio et al., 2009).

This paper describes the measurements in the GeoSatakunta GPS network, presents the results of the time series analysis and illustrates them in connection with the geological data.

2. GPS measurements and data processing

GeoSatakunta network was established in 2002, and it was later (2005-2006) expanded southwards (Figure 1). We have carried out three annual campaigns in 2003-2008. The network consists of 13 pillars and of the Olkiluoto permanent GPS station. The site selection was based on geological considerations by the Geological Survey of Finland.

The GPS pillars are made of reinforced concrete on-site. All the pillars are attached to the solid bedrock with iron bars. Ashtech Z12 and μZ dual frequency receivers and Ashtech Dorne Margolin Choke Ring antenna were used.

The computation was carried out using Bernese 5.0 software (Dach et al., 2007). The data processing strategy was based on the double-difference approach. We first computed entire set of baselines in each session. After that a network adjustment was applied, where the Olkiluoto permanent station was kept fixed. As a result we obtained coordinates of the pillars for each campaign.

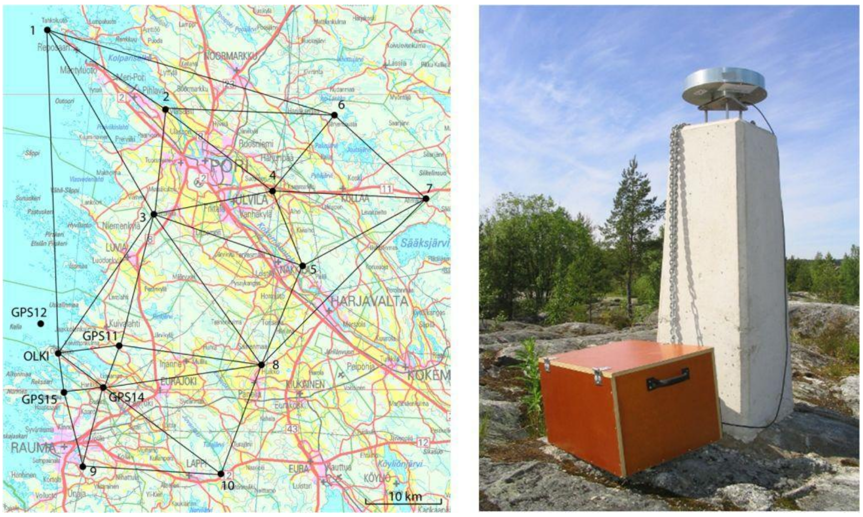

Figure 1. Measurements at the pillar GPS 15 of the GeoSatakunta network.
Base map © National Land Survey, license number 51/MML/09.

3. Baseline length time series

We computed baseline lengths from the three-dimensional coordinates obtained from the GPS processing in order to get an insight into the temporal variation of the solutions (Figure 2). On the average the baseline lengths varied less than 2.0 mm without any evident trend. In a few cases the scatter was as low as 1.0 mm.

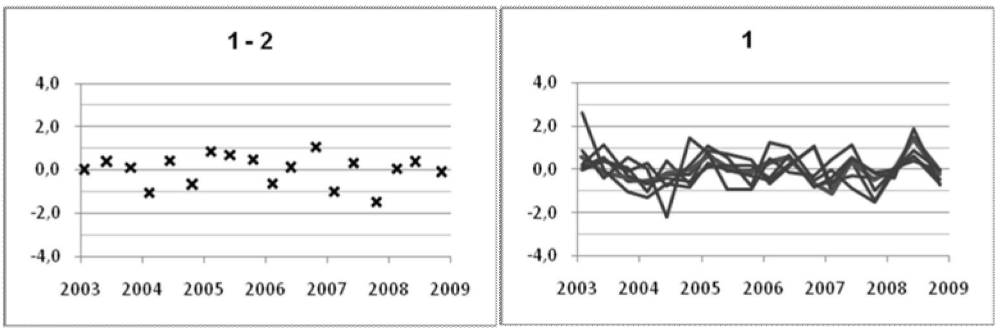

Figure 2. Example of baseline length time series (deviation from the mean in mm). A single baseline on the left, all baselines from a single pillar on the right.

4. Horizontal velocities

The deformation analysis was based on the coordinate differences between adjacent pillars, which were determined automatically using Delaunay triangulation. A least squares free network adjustment was applied with station specific velocity parameters as additional parameters (Kallio et al., 2009). The 3σ criterion was used to test the statistical significance of the estimated velocities, which roughly corresponds the 99% confidence level.

In the first approach the deformation analysis was performed for the entire network. The estimated velocities were (max 0.3 mm/a) mostly statistically insignificant, but of the same order of magnitude as detected in Olkiluoto deformation studies (e.g., Kallio et al., 2009). The velocities for the southern part of the network were more uncertain because of shorter time series.

In the second approach the deformation analysis was performed using only the northern pillars (1–7) and the Olkiluoto station, which have longer and therefore more reliable time series. The estimated velocities and the standard deviations were slightly smaller than in the first analysis for the entire dataset. For the illustration of the results the horizontal velocities were interpolated in a grid and overlaid on the geological data. The possible changes seem to take place in the vicinity of geologically interesting features, like the major shear zones and the Jotnian sandstone. The results do not allow us to make any further interdisciplinary interpretations.

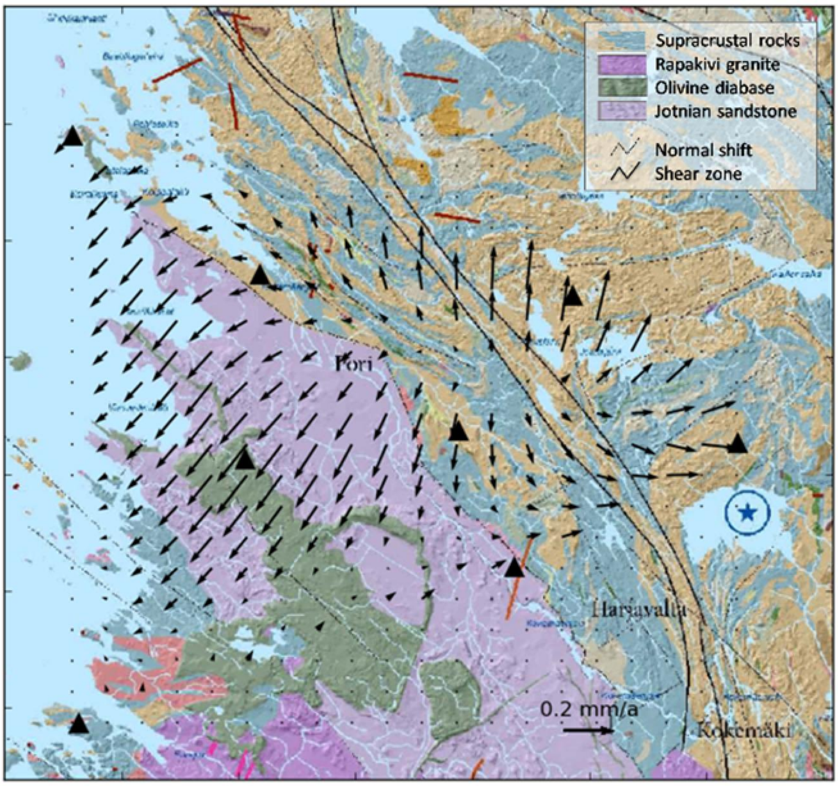

Figure 3. The horizontal velocities interpolated in a grid and overlaid on the geological data. The uncertainties of the velocities are on the average 0.07 and 0.05 mm/a in the North and East components, respectively. Basemap © Geological Survey of Finland.

5. Conclusions

On the basis of six years of measurements we have achieved a detection threshold of 0.3 mm/a for single-site motion. Using the subnet with the longest history the threshold diminishes to about 0.2 and 0.1 mm/a in the N and E components, respectively. A more sophisticated statistical analysis is needed to draw any definite conclusions about motions, as well as longer time series.

GPS observing campaigns in the network, combined with the activity at Olkiluoto, will be the basis of the upcoming research. We also plan to use existing gravity data and GPS-levelling results to compute a local geoid model and to fit it better in the sharp geoid features at the River Kokemäki. The Satakunta network is the first high-precision network in Finland of this size, and therefore, a valuable test field for geodetic techniques to detect minor deformations and network stability.

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Craton mantle roots: Do we really know how they form and how old they are?

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Formation of the mantle that underlies cratons is an active area of research today. This ancient mantle has several unique characteristics, including low density relative to average peridotite, concomitant infertile compositions, highly magnesian olivine and consequently high Vs and Vp values. The buoyancy and rigidity of this type of mantle has allowed Archean crust to remain stable and "float" in a way that keeps it (mostly) from being reworked into the convecting mantle. All models so far proposed for the generation of this type of mantle are based on partial melting, and that this partial melting reflects the age of formation. However, the ever growing body of evidence for extensive refertilization of the mantle effectively removes this limitation, and leaves open the possibility that craton mantle roots are much older than the present 3.5 Ga upper limit.

Keywords: Hadean, Archean, lithospheric mantle, kimberlites, craton

1. Introduction

Cratons are pieces of the earth's lithosphere that are over 2.5 Ga in age. Most of what we know about cratons comes from the study of the crustal rocks to which we have relatively easy access. The mantle portion of this lithosphere is less readily visited or sampled. However, it because of the great value of diamond, and the fact that there has been an enormous amount of money spent finding diamond deposits, that we know as much as we do about the cratonic mantle lithosphere. The link here is that diamond deposits are almost exclusively found in Archean terrains in the remnants of small volcanoes of a rock type called kimberlite (and a few cases of lamproite and UML). Along with the odd diamond, these volatile-rich, explosively emplaced magmas have carried with them an abundance of mantle material, typically several weight percent (more if the estimates of macrocrysts of olivine are correctly identified as mantle-derived). It is this material from which we extract the bulk of our information about the mineralogical and chemical stratigraphy of the mantle lithosphere.

From such studies of continental lithospheric mantle rock samples (xenoliths) and disaggregated rock samples (xenocrysts), and from exposed orogenic massifs around the world, three main models have been proposed to explain the formation of the thick, chemically depleted mantle roots under Archean crust and all three are based on large degrees of partial melting to produce extremely depleted mantle material. Published models have recently been reviewed by Arndt et al. (2009) and essentially vary in terms of tectonic setting: 1. Subduction and stacking of ocean lithosphere (e.g., Helmstaedt and Gurney, 1995); 2. partial melting due to heating by an impinging plume leaving a highly depleted residue (e.g., Griffin et al., 2003); 3. Processes in a subduction zone (e.g., Jordan 1988, Lee, 2006) involving wet melting of arc wedge peridotites due to water derived from the subducted slab.

2. Pros and Cons of the current craton mantle root formation models

Subduction and stacking – Based mostly on geochemical arguments, a number of researchers (e.g. Canil, 2004; Parman et al., 2004; Lee, 2006; Pearson and Wittig, 2008) are adamant that the lithospheric mantle, all of it, including continental and oceanic varieties, must have formed at relatively shallow depths of melting, < 120 km. However in a comparison of xenoliths from Archean mantle from all major cratons of the world, Wittig et al. (2008) conclude that there is such a large error in the estimation of primitive mantle composition (PRIMA) that major element melting models are in fact ambiguous with regards to differentiating deep vs. shallow melting. Nevertheless they maintain that uniformly low HREE concentrations in these peridotites require little to no garnet in melt residues, and this is strong evidence that much of the melting happened in the spinel field, above the pressure stability field of garnet. A strong corroborating piece of evidence for the model comes from the large number of eclogite xenoliths carried by kimberlites, and the fact that these eclogites fit nicely as high pressure versions of subducted oceanic crust. Diamonds found in some of these xenoliths with strongly negative $\delta^{13}C$ (as low as -30, thought to fingerprint biogenic C) and inclusions of coesite with ($\delta^{18}O_{VSMOW}$ ‰ up to +17; Schulze et al., 2003) are most likely reflecting surface processes.

However, all of these lines of evidence are potentially refutable and/or ambiguous. For example, Griffin et al. (2009) point out that high Cr-numbers in pyrope and chromite diamond inclusions and as free grains in some diamond-bearing xenoliths are not unambiguous signatures of low pressure melting (as has often been quoted) when metasomatic effects are factored in (see later mantle refertilization discussion). Lee (2006) points out that the wide variety of rock types that make up the oceanic lithosphere, and the relative volume ratios of these rock types, simply cannot add up to the depleted mantle roots petrologically or chemically.

Mantle plumes - Rather than subduct ocean crust to generate the Archean mantle root, Griffin and coworkers (Griffin et al., 2003, 2009) rather point to the Norwegian dunites/harzburgites rocks of the Western gneiss region (Beyer et al., 2006) that formed at high pressure (5-6 Gpa, op. cit.) as very close analogues of the type of peridotites formed at the base of the lithosphere by plume activity (so-called subcretion). As a corroborating piece of evidence for the plume model, rare diamonds are found in kimberlites that contain inclusions of phases that are only stable in the lower mantle (e.g., periclase, CaSi-perovskite). Griffin et al. (1999) suggest that diamonds of this type in Lac de Gras area kimberlites of Canada were first carried from the lower mantle and underplated to the SCLM by ascending plumes. The kimberlites later picked up these diamonds and transported them to the surface during emplacement.

Subduction zone processes – The main idea here is that wet melting in the mantle wedge overlying the subduction zone can produce very refractory harzburgites and dunites. The main problem with this model are that the resulting residues most probably should have arc-type trace element signatures indicative worldwide of arc-related magmatism (LILE-enrichment and HFSE-depletions), and this is generally not the case (Lee, 2006). Moreover, it would seem this model would produce more depleted residues below less depleted, forming an unstable buoyancy problem and the need to invert this column to match the stratigraphy observed in most craton mantle roots (Arndt et al., 2009).

3. Cratonic mantle older than 3.5 Ga?

Evidence for remnant Hadean mantle reservoirs - At least five pieces of evidence suggest a very old component in the mantle once existed or still exists, suggesting that craton mantle roots may be older than current direct age dating suggests.

1. Jack Hills zircons (up to 4.37 Ga in age) have inclusions of apatite, quartz, xenotime, monazite, rutile, biotite, amphibole, K-feldspar and plagioclase; rather similar to what is expected from a granite host rock (Wilde et al., 2001). Diamond inclusions are probably related to impact processes on the earth’s surface, possibly at some earlier time

(Nemchin et al., 2008). The petrogenesis of the inclusion suite is interpreted to require significant crust-mantle separation, at least locally already at this time (Hopkins et al., 2008).

2. Hf analyses from these same zircons require that some crust separated from the mantle 4.4-4.5 Ga based on Hf isotopes (Harrison et al., 2005).

3. Nuvvuagittuq greenstone belt mafic samples, Hudson, bay show evidence of ^{142}Nd anomalies (from decay of the short-lived ^{146}Sm-^{142}Nd isotopic system with a 103 million year half life) that suggest fractionation from the mantle at about 4.2 Ga based on a ^{142}Nd /^{144}Nd – Sm/Nd whole rock isochron (O'Neil et al., 2008).

4. Similar ^{142}Nd anomalies in the Khariar alkaline complex in eastern India require that these magmas sampled a mantle that separated from the bulk earth prior to 4.1 Ga and suggest that this reservoir is the continental lithospheric mantle (Upadhyay et al., 2009).

5. Cenozoic Baffin Island – W. Greenland lavas reported by Jackson et al. (2010) with a clear 4.5 Ga mantle signature as shown by combined He-Pb-Nd isotopic data represent one of the first magma suites that clearly requires a Hadean-aged mantle reservoir. Although the authors do not suggest it, one likely candidate for such a reservoir is continental lithospheric mantle.

Evidence against remnant Hadean mantle reservoirs – The main line of evidence against a ca. 4.4 Ga age for any of the craton mantle roots is the age dating of mantle xenoliths using the Re-Os system. Other isotopic systems, including Nd and Sr model ages give younger or similar ages in some cases (e.g., Richardson et al., 1984, 1993, 2001, 2004). In terms of Re-Os T_{RD} or T_{MA} ages, however, almost on every continent the oldest ages obtained on mantle samples, either whole rock or single grain sulfides is ca. 3.5 Ga (Pearson et al., 1998). This is a large body of data to refute in the claim for extant Hadean mantle, but to do so first requires discussion of one other important aspect of the craton mantle roots – mantle refertilization.

4. Mantle refertilization, diamond-sulfide formation and Re-Os T_{RD}

There is considerable evidence that craton mantle roots have been extensively refertilized by later percolating melts and/or fluids. This concept has been well known for some time (e.g., Dawson (1984) wherein cryptic and modal metasomatism were described). A particularly good example of the type of metasomatic transformation relevant to this discussion is shown by the dunite and garnet lherzolite bodies from Western Norway (Beyer et al., 2006). In this example, depleted dunites form the host rock for meter-scale thick lenses of garnet lherzolites, where the latter are thought to represent melt channelways along which the original dunites were infiltrated and metasomatised. A similar melt infiltration process has been ascribed to the harzburgite to lherzolite metasomatic transformation suspected for the lherz massif (Le Roux et al., 2007). A corollary of this process is that potentially all garnet, clinopyroxene and possibly orthopyroxene present in continental mantle formed through the process of fertilization of dunite, i.e., they are all secondary minerals.

Evidence is building that metasomatic processes control the formation of peridotitic diamonds in the mantle, in this case by the oxidation of asthenosphere-derived CH_4 ± H_2O fluids as they invade and react with subcontinental mantle peridotites (Stachel et al., 2004, Malkovets et al., 2007). At the same time this reaction produces metasomatic horizons in the mantle with subcalcic harzburgitic pyropes and explains the fact that harzburgitic garnets are the dominant silicate inclusion in diamond. This also implies that even rocks that have been thought to be quite refractory (harzburgite) can, in fact, be products of metasomatism. It appears that only lithospheric mantle roots contain sufficiently refractory dunites for this diamond-forming reaction to proceed. This provides a link between the ancient age of diamonds, and their restriction to cratonic areas. Implicitly of course it implies that any age dating of the diamond is the date of a metasomatic event, and requires that the host rock be somewhat, if not significantly, older.

By far the most common inclusion in diamond is sulfide, and it has been speculated that diamond precipitation is facilitated by the presence of sulfide (e.g, Palyanov et al., 2007). The suggestion is that as for diamonds, sulfides are precipitated as fluids migrate into the mantle and the process of fertilization proceeds. Gurney et al. (2010) recently reviewed diamond forming events through history, and showed that in the Kaapvaal, Siberian and Slave cratons, the earliest peridotite-type diamond forming event occurred between 3.5 to 3.2 Ga. Most of the age dates for these diamonds come from Re-Os measurements on single sulfide inclusions encapsulated within the diamonds.

Bringing all these loose threads together, the T_{RD} (and T_{MA}) ages collected thus far from mantle samples (mostly from sulfide grains, and many as inclusions in diamond) are concrete evidence of an important event that has occurred in the mantle, seemingly a massive influx of (plume-derived?) melts and/or fluids at 3.5 Ga that formed diamonds and sulfides in the depleted dunite of the craton mantle roots all over the world, among other effects. Only after these highly refractory mantle roots were fertilized, were they able to produce melts. The Re-Os model ages therefore either record the timing of the sulfide/diamond formation or subsequent melting events, but probably not the true age of the host rocks. The comment of Rudnick and Walker (2009, p.1084) regarding the interpretation of Re-Os model ages is very relevant here, “If one assumes that continental mantle lithosphere formation occurs at the time of melt depletion, then Re-Os has the potential to date lithosphere formation.” One could add the corollary “But if the continental mantle lithosphere did not form simply by partial melting, then Re-Os dates put no limits on the age of its formation.”

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Karelian Craton mantle root formation

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1. Introduction

The Karelian craton covers an area of ~ 400 000 km^2 in eastern and northern Finland, and NW Russia. It represents the largest portion of the Archean Karelian-Kola-Kuloi megacraton (>2.3 million km^2), comprised of the Fennoscandian shield and parts of the adjacent Russian platform. Within the area of the Karelian craton, explosively emplaced alkalic magma types kimberlite and olivine lamproite have sampled mantle material, in the form of xenoliths and xenocrysts, from all levels of the lithospheric mantle underlying the craton and carried this material to the earth's surface. The suite of magmas not only have a wide areal distribution but also show a wide range of ages, with clusters of 1800 Ma (Kemozero), 1200 Ma (Lentiira-Kuhmo-Kostamuksha), 760 Ma (Kuusamo), and 600 Ma (Kaavi-Kuopio). Based on xenoliths and xenocrysts derived from these rocks, a cross section of the Karelian craton mantle root has been produced. We report here a review and update of the most recent work on these Karelian craton mantle materials and implications for the assembly of the craton mantle root.

2. The mantle section

Previous studies (Peltonen et al., 1999; Peltonen et al., 2003; O´Brien et al., 2003; Lehtonen et al., 2004; Lehtonen, 2005; Peltonen & Brügmann, 2006; Lehtonen & O´Brien, 2009) of mantle-derived xenolith and xenocryst studies indicate that the mantle root of the Karelian craton varies considerably from margin to core (summarized here and shown in figure 1). At the margin, in the **Kuopio and Kaavi area**, the mantle is stratified into at least three distinct layers: (A) A shallow, 60-110 km, garnet-spinel peridotite layer, (B) A variably depleted peridotitic horizon from 110 to 180 km containing diamond-indicative subcalcic harzburgitic pyropes, (C) A deep layer, >180 km, composed largely of fertile peridotites. Shallow *layer A* contains peridotites that have "ultradepleted" arc mantle-type compositions, and have been metasomatised by radiogenic $^{187}Os/^{188}Os$, possibly from slab-derived fluids. Xenoliths derived from the middle *layer B* (at ~110–180 km depth), which is also the main source of harzburgitic garnets (G10) xenocrysts in Kaavi-Kuopio kimberlites, are characterised by an unradiogenic Os isotopic composition. In these xenoliths $^{187}Os/^{188}Os$ shows a good correlation with indices of partial melting implying an age of ~3.3 Ga for melt extraction (Peltonen & Brügmann, 2006). This age is slightly younger than the oldest formation ages of the overlying lower crust, as judged by zircons of up to 3.5 Ga in mafic granulite xenoliths (Peltonen et al., 2006) from the Lahtojoki kimberlite in the Kaavi area. The underlying G10-pyrope-free *layer C* (at 180–250 km depth) is the main source of Ti-rich pyropes of megacrystic composition, which however, lacks G10 pyropes. The osmium isotopic composition of the layer C xenoliths is more radiogenic compared to layer B, yielding only Proterozoic T_{RD} ages. Layer C is interpreted as a melt metasomatised equivalent of layer B. This metasomatism most likely occurred at c. 2.0 Ga when the present craton margin formed following break-up of the proto-craton.

The mantle stratigraphy of the craton core, in the **Kuhmo, Lentiira and Kostomuksha** areas, shows less variation. There is no evidence of Layer A. Layer B begins with the lowest temperature pyropes at an inferred depth of 70 km and continues to a depth of about 250 km showing a relatively homogenous distribution of harzburgite and lherzolite pyropes throughout. Differences in craton core layer B compared to craton edge layer B include: 1. Wehrlite garnet is very rare, as is chrome diopside. 2. The G10 to G9 ratio in both exploration samples and hardrock sources is considerably higher implying core layer B is relatively harzburgite-rich. However, Ti-rich megacryst composition pyropes are very common, so better evidence for a deep 250-300 km Layer C may become available with further sampling. 3. The overall Mg# of the mantle lithosphere in this area is extremely high relative to the worldwide average and to the mantle at Kaavi-Kuopio. Coupled with the rarity of mantle-derived chrome diopside, the implication is that this portion of the mantle underwent unusually high levels of partial melting to produce such a refractory residua or was assembled from originally extremely melt-depleted source rocks. 4. The amount of eclogite pyrope garnet is extremely low in the craton core exploration and hard rock samples samples. 5. At least within the limitations of the present database there is only a weak indication of a G10 pyrope-free, metasomatically enriched Layer C.

The position of the **Kuusamo** area well within the Karelian craton would imply it represents craton core and the distribution and composition of mantle xenocrysts derived from this area confirms the existence of depleted, and therefore ancient mantle underlying this area. Xenocryst pyrope chemistry demonstrates that lherzolite and harzburgite are roughly equally distributed down to depths of roughly 180 km. Although the data are relatively sparse, we interpret the existence of harzburgite throughout the mantle section, without any obvious layering, to represent lithospheric mantle similar to that in the Kuhmo region, albeit slightly thinner. Bolstering this interpretation is the fact that the amount of eclogite material in exploration and hard rock samples is similarly low quite low. The existence of abundant harzburgitic rocks, some with ultra depleted pyrope compositions and extremely high *mg* numbers, implies that the age of this mantle is Archean.

3. Craton mantle chemical evolution -- Pyrope major and trace element signatures

Mg and Ti contents – Variations in pyrope Ti and Mg# can be used to determine the fertility of the mantle peridotites from which they have been derived. The following pyrope compositional relationships have been observed (Lehtonen and O'Brien, 2009): 1. the Mg# of the overall pyrope population is significantly lower at Kaavi-Kuopio relative to Kuhmo-Lentiira and Kuusamo, 2. Ti-variation is wide throughout the Kaavi-Kuopio section (TiO_2 0.01-0.90 wt%) whereas similar Ti-variation only occurs in the deeper section (>180 km) at Kuhmo-Lentiira and Kuusamo, 3. Mg# shows no correlation with depth at Kaavi-Kuopio (except being uniformly lower in upper layer A) whereas there is a uniform increase in Mg# with depth at Lentiira-Kuhmo (i.e., a function predominantly of temperature). The latter implies a closed, well-equilibrated, and long-lived system.

Trace elements – The elements Y, Zr and the rare earth elements provide considerable information on chemical influences on individual pyrope grains by metasomatic agents that are not buffered by the bulk rock peridotite. For the same pyrope dataset discussed above (minus Kuusamo) the following observations have been made (Lehtonen, 2005; Lehtonen and O'Brien, 2009): 1. the highest temperature at which a low Zr or Y pyrope exists (representing the base of layer B mantle) is around 1150 °C at Kaavi-Kuopio and 1400 °C at Lentiira-Kuhmo, 2. high Zr without concomitant increase in Ti or Y is limited to pyropes from

Lentiira-Kuhmo, 3. the type of LREE-enrichment at Kaavi-Kuopio is different than Lentiira-Kuhmo, with higher $(Nd/Ce)_N$ (i.e., less steep negative LREE slopes) across all pyrope varieties.

All of these lines of detailed geochemical evidence point toward a stronger melt metasomatic signature at Kaavi-Kuopio and a stronger fluid metasomatic influence at Kuhmo-Lentiira (also with partial evidence at Kuusamo). Nevertheless some examples of fluid metasomatized pyropes exist at Kaavi-Kuopio, suggesting that early fluid metasomatism was overprinted by later melt metasomatism. Therefore the Kuhmo-Lentiira mantle is thought to represent the mantle section least affected by metasomatism subsequent to formation.

4. Time for a caveat - Diamond formation and mantle refertilization

Evidence is building that metasomatic processes control the formation of peridotitic diamonds in the mantle, in this case by the oxidation of asthenosphere-derived $CH_4 \pm H_2O$ fluids as they invade and react with subcontinental mantle peridotites (Stachel et al., 2004, Malkovets et al., 2007). At the same time this reaction produces (subvertical?) metasomatic horizons (veins) in the mantle with subcalcic harzburgitic pyropes as one byproduct and explains the fact that harzburgitic garnets are the dominant silicate inclusion in diamond. It is also likely that melts, including kimberlites and lamproites discussed here, are preferentially channelled into pre-existing fractures within the mantle. Given the reactivity of fluids derived from volatile charged kimberlitic magmas, it follows that the conduits through which precursor melts have ascended are metasomatized. This leads to the inevitable conclusion that kimberlitic magmas probably are **not** carrying a representative sample of the bulk mantle of an area, but rather the average of a certain metasomatized domain. How biased kimberlite sampling of the mantle might be, is hard to assess, but with some certainty we can say the bulk of the mantle lithosphere is more refractory than the samples we have to study.

5. Implications for the Karelian craton assembly

Three main models have been proposed to explain the formation of the thick, chemically depleted mantle roots under Archean crust and all three are based on large degrees of partial melting to produce extremely depleted mantle material. These models have recently been reviewed by Arndt et al (2009) and essentially vary in terms of tectonic setting: 1. Subduction and stacking of ocean 2. partial melting due to heating by an impinging plume leaving a highly depleted residue 3. Processes in a subduction zone involving wet melting of arc wedge peridotites due to water derived from the subducted slab. Our present interpretation of the Karelian craton mantle cross section is that the original craton (closely resembling the section at Kuhmo-Lentiira-Kuusamo) may have formed by plume activity, high degrees of partial melting and komatiite extraction. The time of this stabilization must be as least as old as the oldest crustal rocks, and we take this to be 3.5 Ga based on garnet granulite xenoliths from Lahtojoki (Peltonen et al., 2006). However, at the craton edge at Kaavi-Kuopio, the uppermost mantle layer A, representing Proterozoic arc-wedge mantle produced by subduction processes (although where and with what polarity this subduction occurred is unknown) was subsequently accreted at approximately 1.9 Ga to the existing craton core during collision between the Svecofennian arc complex and Karelian craton. The relative abundance of eclogite component at Kaavi-Kuopio essentially absent in the craton core suggests eclogitic material was also accreted during this collisional event.

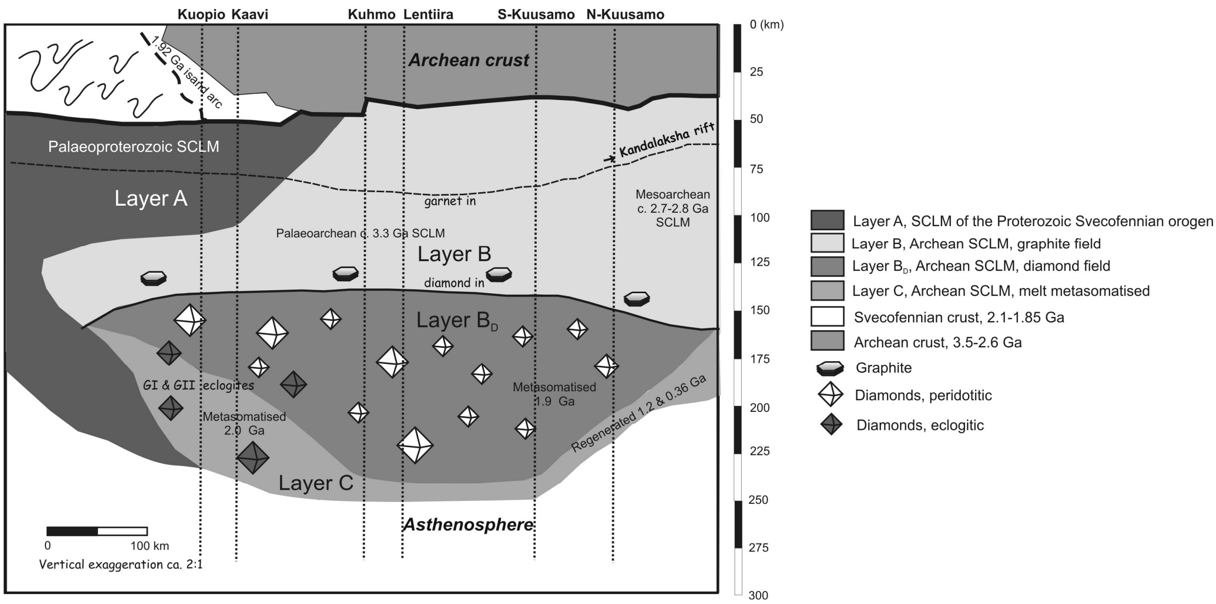

Figure 1. Simplified geological cross-section of the Karelian craton crust and mantle across as 600 km transect. After Lehtonen and O'Brien, 2009.

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An integration of the paleomagnetism and geochemistry of Proterozoic dykes, Dharwar Craton, Southern India

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The integration of paleomagnetic data with geochronological and geochemical analyses of mafic dyke swarms is a powerful tool for correlating the dyke swarms on regional and global scales. The multidisciplinary approach may also provide important insights into various problems of Earth evolution such as paleoreconstructions, the morphology of the geomagnetic field and mantle partial melting properties. We present the results of our ongoing paleomagnetic and geochemical investigations of several Proterozoic dykes from the Dharwar craton (Southern India). When combined with the existing geochronological data, our data support the presence of at least two different E-W trending dyke swarms (~2370 and ~1890 Ma) in the Dharwar craton. Our results are also consistent with the possibility that the Bastar and Dharwar cratons were amalgamated already before ~2370 Ma.

Keywords: paleomagnetism, geochemistry, Proterozoic, Dharwar Craton

1. Introduction

The integration of paleomagnetic and geochemical data with precise U-Pb geochronology has proven to be a powerful tool to study large igneous provinces (LIPs), continental reconstructions, the morphology of the geomagnetic field and the source mantle properties (e.g. Halls 2008). The Proterozoic mafic dyke swarms exposed in the Dharwar craton (Southern India) are of special interest because the craton has been a principal constituent of ancient supercontinents (e.g. Rogers 1996). We report new results of our paleomagnetic and geochemical analyses of several Proterozoic Dharwar dykes and discuss their geodynamical implications.

2. Geology and sampling

Multiple cross-cutting mafic dyke swarms transect the Dharwar craton, especially in its eastern part, trending in three dominant directions: NW-SE, E-W and NE-SW (e.g. Murthy 1995). Their composition ranges from tholeiites or alkali-olivine basalts to metadolerites/metanorites (e.g. Rao and Pupper 1996). Radiometric age data and observed cross-cutting relationships indicate that most of the Dharwar dyke swarms were emplaced during the Proterozoic with the peak of emplacement activity in the Paleoproterozoic:

1) ~2366 Ma E-W trending swarm i.e. the Bangalore LIP (e.g. Halls et al. 2007). The NE-SW trending Karimnagar swarm has been suggested to be a part of this LIP (Kumar et al. 2010).
2) ~2214 Ma N-S trending Kandlamadugu swarm and ~2209 Ma NW-SE trending Somala swarm (French 2007).
3) ~2180 Ma NW-SE to E-W dykes with a ~60° fanning i.e. the Mahhubnagar LIP (French et al. 2004). We speculate that based on the fanning geometry and nearly

coeval emplacement ages, the Somala and Kandlamadugu dykes could be a part of the Mahhubnagar LIP event.

4) ~1890 Ma E-W, ENE-WSW trending dykes in the Dharwar craton being part of the Southern Bastar–Cuddapah LIP (Ernst and Srivastava 2008).
5) ~1027 Ma age was reported from an E-W trending great dyke of Penukonda, southwest of the Cuddapah basin (Pradhan et al. 2009). However, a much older age of ~2365.9±1.5 Ma was determined from supposedly the same dyke by French (2007).

We collected 230 hand samples from 30 dykes (12 E-W, 5 NW-SE, 6 NE-SW and 5 N-S trending) and from their baked and unbaked host rocks when possible. The dykes vary in width from 0.5 to 100 m with the mode of 20-30 m. The samples were taken from quarry and roadside outcrops, river cuts, and small field exposures. The paleomagnetic samples were oriented using both sun and magnetic compass.

3. Results and discussion

Rock magnetic analyses suggest pseudo-single domain (PSD) magnetite (~585°C Curie temperature and PSD region in Day plot after Day et al. 1977) as the principal magnetic carrier in studied dykes. We excluded thirteen dykes from further consideration because of either their very low susceptibility values (<1000 µSI) indicating strong alteration or very large Q values (> 60) indicating lightning strikes. Paleomagnetic directions were measured by detailed thermal and/or alternating field (AF) demagnetization using a 2G SQUID magnetometer. The remanence directions and site mean directions were calculated using the principle component analysis (Kirschvink 1980) and the Fisher statistics (Fisher, 1953), respectively. We were able to obtain well-defined paleomagnetic directions from 17 dykes. The characteristic remanent magnetization (ChRM) was usually isolated within a narrow (10-25°C) unblocking temperature range above 550°C or within the high-coercivity part (>80 mT) of AF demagnetization spectra.

The geochemical samples were analyzed by Activation Labs Canada using lithium metaborate/tetraborate fusion followed by ICP for whole rock analyses and ICP/MS for trace elements. Loss on Ignition (LOI) is low for all samples consistent with a lack of alteration as suggested by review of thin sections. Samples in this study are sub-alkaline and their extrusive equivalents vary from basalt to andesite. When plotted in an AFM diagram (FeO^t – Na_2O+K_2O – MgO; after Irvine and Baragar 1971) they follow a pattern of iron enrichment indicative of tholeiitic magmas. Srivastava (2006) has demonstrated trace element differences between NW-SE trending dykes of different ages in Bastar craton. Our NW-SE trending dykes from Dharwar craton show the highest enrichment of trace elements similar to that found in the ~1890 Ma dykes in Bastar craton. The rest of the samples in this study have a trace element geochemical signature that falls intermediate between the two older age groups from the Bastar craton. We note also that the younger ~1894 Ma dyke of Halls et al. (2007) closely follows the REE pattern of ~1890 Ma dykes from the Bastar craton (Srivastava 2006).

Most of the magnetization components that we determined from the E-W trending dykes are consistent with the directions reported in prior works (e.g. Halls et al. 2007 and references therein). The steep up direction was identified as the component A from the ~2366 Ma E-W dykes and the steep down direction as the component Ar (Halls et al. 2007). The geochemical characteristics of the ~2366 Ma E-W dykes with steep up directions (this work, French 2007, Halls et al. 2007) do not significantly differ from one another. The observed minor variation may reflect the differences in analytical techniques and/or temporal evolution of magma source over a protracted period of time. We note that the geochemistry of the older NW-SE dyke swarms from the Bastar craton (Srivastava 2006) is similar to that of the ~2366 Ma E-W dykes. Although no paleomagnetic data are available for the older Bastar craton dykes, this observation invites speculation about a similar/shared magma source for the oldest dyke swarms in the Dharwar and Bastar cratons. This in turn could indicate that these cratons were geographically contiguous already at ~2366 Ma.

The steep down direction that we observed from the N-S to NW-SE dykes is close to that observed from two of the E-W dykes. One possibility is that the latter dykes belong to the younger ~2180 Ma Mahhubnagar LIP. However, in our opinion this possibility is not consistent with the observed dyke swarm geometries. Based on their trend and location, these dykes are more likely to belong to the older Bangalore LIP in which case their paleomagnetic directions may represent the reversed A-component as discussed by Halls et al (2007). If true, this would indicate a considerably slow plate movement between ~2370 Ma to ~2200 Ma. We note however that the steep down-component was also observed in one N-trending ~1192 Ma alkaline dyke (e.g. Pradhan et al. 2008), so the possibility of an overprint cannot be ruled out.

The shallow NE component observed from the NE-SW dykes is similar to the directions reported from the same area (Kumar and Bhalla 1983; Pradhan et al. 2009) and from the Karimnagar dykes (Subba Rao and Radhakrishna Murthy 1985). We note that the zigzag-shaped Zijderveld plots indicate the presence of a chemical remanent magnetization (CRM). The origin of this component remains unknown. The geochemical characteristics of NE-SW trending Karimnagar dyke swarm (Rao et al. 1990) show a largely bimodal distribution overlapping with both the ~2366 Ma (Halls et al. 2007; French 2007) and ~1890 Ma dyke swarms (Srivastava 2006; Halls et al. 2007) (Fig. 3). The Karimnagar dykes also yield two different paleomagnetic directions: a NE shallow up and steep up (e.g. Subba Rao and Radhakrishna Murthy 1985). These observations indicate that the Karimnagar swarm may actually represent two generations of dyke emplacement. Precise age determinations are crucial to test this hypothesis (Kumar et al. 2010).

We feel that the apparent contradiction may be explained with the co-existence of three different nearly parallel E-W trending dyke swarms of different age (~2370Ma, ~1890 Ma and ~1030 Ma). Such a possibility can be supported by visual inspection of satellite images and earlier studies indicating episodic magmatic intrusive activity along preferred directions (Chatterjee and Bhattacharji 2001).

We note that the geochemical fingerprints of the ~1890 Ma dykes and the NE-SW dykes studied here are similar to those of the ~1890 Ma NW-SE trending dyke swarm from the Bastar craton (Srivastava, 2006; French et al. 2008). Therefore these dykes and the ~1890 Ma E-W to NE-SW trending Dharwar dykes with shallow NE directions can be related to the formation of Cuddapah basin. A detailed paleomagnetic study of the Bastar craton dykes is needed to clarify this problem.

4. Conclusions

The results of our paleomagnetic and geochemical investigations are consistent with the results reported for the Dharwar mafic dykes by other authors. Our study hints that the Bastar and Dharwar cratons could have been amalgamated before ~2370 Ma. In addition, an integrated analysis of the paleomagnetic and geochemical data performed here supports the existence of at least two different ~E-W-trending dyke swarms (~2370 and ~1890 Ma) in the Dharwar craton, both possibly reaching the Bastar craton. While geochemistry seems to differentiate these two dyke swarms fairly well, the primary origin of their magnetization is still debatable. Our results, currently based only on limited sampling, should therefore be considered preliminary. Additional paleomagnetic and geochemical analyses as well as precise age determinations are needed to decipher the complex geological history of the Dharwar craton. Our study, nevertheless, demonstrates the potential of combining paleomagnetism with geochemistry and accurate age data.

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ILP Regional Coordination Committee DynaQlim: Upper Mantle Dynamics and Quaternary Climate in Cratonic Areas

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The ILP Regional Coordination Committee DynaQlim (Upper Mantle Dynamics and Quaternary Climate in Cratonic Areas) was established in 2007 for studying the relationship between Glacial Isostatic Adjustment (GIA), upper mantle structure, dynamics and Quaternary climate. It integrates existing data and models from a variety of disciplines that consider processes over a range of spatial and temporal scales relating to the Quaternary evolution of cratonic regions. A key aim will be to facilitate the development of various models in order to generate more accurate predictions of Earth and ice sheet evolution during the Quaternary, and thus to understand the past and contemporaneous evolution of topography in previously glaciated terrains.

Keywords: lithosphere, upper mantle, GIA, Quaternary climate

1. Introduction

The Glacial Isostatic Adjustment (GIA) allows retrieving information about structure, dynamics and rheology of the lithosphere and upper mantle. GIA related phenomena affect also on Earth rotation, polar motion and crustal deformation. In geodetic point of view, it affects on global, regional and local reference frames, heights and gravity. Glacial ice sheet dynamics is one of the key elements constrained by coupled process of the deformation of viscoelastic Earth, and ocean and climate variability. The main focus of DynaQlim (Upper Mantle Dynamics and Quaternary Climate in Cratonic Areas), a regional coordination committee of the International Lithosphere Program since 2007, is to study these phenomena (Poutanen et al., 2009).

2. Observational basis

Postglacial uplift observations over last 100 years are based on geodetic data of precise levelling, tide gauges, gravity observations and postglacial fault activity monitoring. Horizontal motions could not be observed accurately over large areas without modern GNSS (Global Geodetic Observing Systems, including GPS) techniques. As an example, the project BIFROST (Baseline Inferences for Fennoscandian Rebound Observations, Sea Level, and Tectonics), initiated in 1993, is using tens of permanent GPS stations in Finland and Sweden to produce maps of the 3-D crustal velocity field of improved spatial fidelity and accuracy (Milne et al., 2001, Johansson et al., 2002, Lidberg et al., 2010)

The gravitational uplift signal can be detected by absolute and relative gravimetry (e.g., Mäkinen et al., 2005) or by the gravity satellite mission like GRACE (e.g. Tamisiea et al., 2007). GRACE data show temporal gravity variations both in Fennoscandia and North America (Tamisiea et al., 2007, Steffen et al. 2008).

The Fennoscandian uplift history is visible in ancient shorelines (e.g. Lambeck et al. 1998, Tikkanen and Oksanen, 2002). Over the last decades attempts were made to reconstruct palaeo-shorelines worldwide. In the periglacial regions the palaeo-sea level is dominated by regional lithospheric flexure due to glacial loadings. Therefore, in such coastal regions there is a close link between GIA and reconstructions of palaeoclimate (Poutanen and Ivins, 2010).

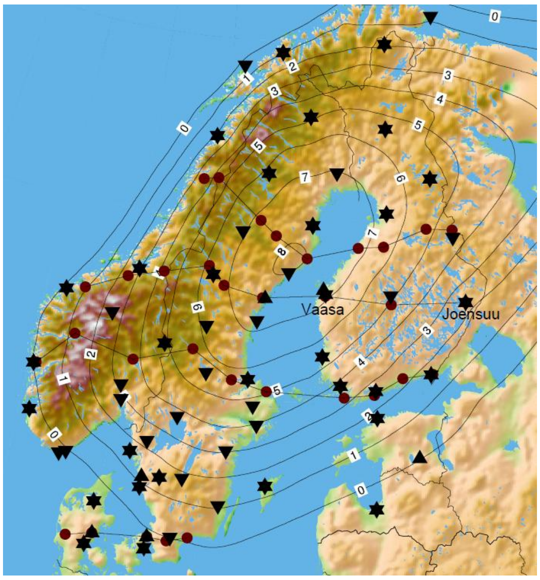

Figure 1. GIA in Fennoscandia. The upside-down triangles on the map are permanent GNSS stations, triangles stations where absolute gravity is regularly measured, and dots with joining lines are the land uplift gravity lines, measured since the mid-1960's. Contour lines show the apparent land uplift relative to the Baltic mean sea level 1892-1991, based on Nordic uplift model NKG2005LU (Vestøl, 2006; Ågren and Svensson, 2007)

The viscoelastic response of the solid Earth and palaeotopography are influenced by behaviour of the lithosphere and its lateral variations. Our knowledge of the rheology and structure of the lithosphere is based on rock deformation experiments, petrophysical inference from seismology and heat flow. The surface heat flow impacts the flow dynamics of the ice sheet (Näslund et al., 2005) and thus the deglaciation history.

Continuous GNSS observations of plate-wide strain, accompanied by seismological investigations, and followed by continuum mechanical modelling of GIA, studies of seismic source and wave propagation, and studies of the postglacial faults offer new insights into properties of the lithosphere. Observations and models of glacial and postglacial faulting can help to better understand crustal stress fields and crustal rheology.

Rebound processes may have its effect in seismicity. The intraplate seismicity in Fennoscandia is quite low and the epicentres are concentrated along ancient tectonic deformation zones in areas which are favourably orientated with respect to the regional stress field, and are thus capable of being reactivated (Uski et al., 2003, Gregersen and Voss, 2010).

Advances in studies of the glacial history of northern Europe and Eurasia have significantly improved our understanding of the ice history during the Late Weichselian (~ 100 – 21 kyr BP), glacial transition times and during the Holocene (11 kyr to present-day) (Svendsen et al. 2004). New models of the evolution of ice thickness would improve also our understanding of climate prior to and during these times.

3. GIA and GIA models

There are also several ongoing multi-disciplinary projects utilizing data collected over decades and which will benefit on the studies of the mechanisms of GIA. These projects include BIFROST, DynaQlim, COST Action ES0701 "Improved Constraints on Models of Glacial Isostatic Adjustment" (COST, 2010), and SPP1257 "Mass transport and mass distribution in the system Earth" (Ilk et al, 2005).

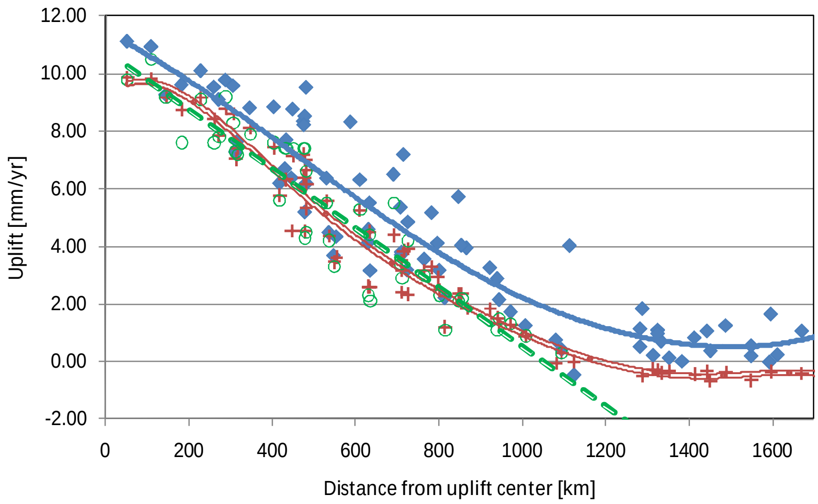

Figure 2. Vertical rates of Fennoscandian rebound as a function of distance to the uplift maximum. GPS data (diamonds) are based on latest BIFROST results of Lidberg et al. (2010), tide gauge data (open circles) are from Ekman and Mäkinen (1996), and GIA model (crosses) is based on revised model constructed by Glenn Milne (Lidberg et al., 2010). (Courtesy Poutanen and Ivins, 2010)

GIA offers information both on the Earth's upper mantle and crust, and the mass variation of glaciers. Due to the slow processes, we see not only contemporary effects but we can also retrieve information about the climate change through the Holocene back to the late Pleistocene. The GIA signal, however, is contaminated by signals of several other spatially and temporally varying mass changes and crustal deformation. These include seismic deformation, mantle convection and plate tectonics (e.g. van Dam et al., 2008). Separating GIA-induced contributions from the other sources is not straightforward. As an example, the deep geoid depression in Northern Canada is partly due to the lithosphere and mantle heterogeneities and is only partly related to GIA (Tamisiea et al., 2007).

Connection of the glacial-to-transition and Holocene climates however is complex, and only a long-term, sustained, inter-disciplinary science effort will allow GIA to be connected appropriately to global ocean-atmosphere reorganizations. Understanding the connection is essential because without these climate changes the GIA phenomenon would simply not exist.

4. Conclusion

The DynaQlim initiative aims to integrate existing data and models of GIA processes, including geological and geodetic observations combined with climatology and other relevant geo-data, such as seismological or sea-level data. The two formerly glaciated regions, Fennoscandia and Laurentia, are the most well known areas for our model development. Combining historical and modern terrestrial and space-borne geodetic observations with seismological investigations, studies of the postglacial faults and continuum mechanical modelling of GIA, the research goal of DynaQlim is to offer new insights into properties of the lithosphere and upper mantle.

The joint inversion of different types of observational data is an important step toward providing a better understanding of GIA on all levels of Earth sciences. One of the objectives of the DynaQlim is to produce better models both of the glacial history and of the Earth response in order to enhance our understanding of the phenomenon of glacially induced faulting and to better predict where and when it occurs.

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Late Palaeozoic remagnetization in the Baltic Plate

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Palaeomagnetic studies performed in different localities in Estonia and Finland have revealed a common secondary magnetization of Permian age. However, the reason for this remagnetization is still unknown. We discuss the possible causes of the remagnetization that link it with local and/or regional geological events.

Keywords: palaeomagnetism, Permian remagnetization, Baltic plate, Pangaea

1. Introduction

During the rock and sediment formation the ferromagnetic constituents can record the direction of the Earth's magnetic field. However, this primary magnetization may be partly or completely destroyed by post-sedimentational geological processes. The secondary magnetic overprint, acquired at any geological time after the formation of a rock, often replaces the primary magnetization. Because remagnetizations are common in all different rock types, the palaeomagnetic method has proven to be useful in the search for evidences of younger alterations.

Recent palaeomagnetic studies performed in Estonian Palaeozoic carbonates and Finnish Precambrian crystalline shear and fault zone rocks have revealed a common secondary magnetization of Permian age (Preeden et al., 2008, 2009; Plado et al., 2010). Such a late Palaeozoic remagnetization is also reported by earlier studies from Fennoscandia and NE Russia (Elming and Bylund, 1991; Perroud et al., 1992; Torsvik et al., 1992; Smethurst et al., 1998; Andersen et al., 1999; Torsvik and Rehnström, 2003; Lubnina 2004; Gurevitch et al., 2005; Mertanen et al., 2008; Khramov and Iosifidi, 2009; Lubnina et al. 2010).

2. Possible reasons for late Palaeozoic remagnetizations

Different authors have attributed the late Palaeozoic component to different geological processes. In the inside and outside areas of the Siljan impact structure it was interpreted to be caused by oxidation due to circulating meteoric water (Elming and Bylund, 1991). Relation of Permian remagnetization to the surface and/or subsurface processes of weathering in Lower Ordovician and Cambrian sedimentary rocks was proposed by Khramov and Iosifidi (2009). Based on studies of Ordovician limestones of St.Petersburg area, Lubnina (2004) interpreted the Permian overprint as being related to tectonic events in the Urals and Western Europe. Torsvik and Rehnström (2003) tied the secondary magnetization in Komstad limestones of Bornholm with cooling/uplift after sedimentary burial. In western and southern Norway, a significant stage in the formation and rejuvenation of faults occurred also during Permian (Torsvik et al., 1992; Andersen et al., 1999). This tectonic activity was probably related to the major Permo-Triassic rifting which resulted to the formation of Oslo rift.

In the area around the Finnish Gulf, a significant thermal influence on the magnetization is not likely as there are no geological evidences that the rocks were heated to considerably high temperatures in Phanerozoic. In Estonia, the maximum diagenetic temperatures suggested are 50 to 80˚C (Männik and Viira, 1990). Thermoviscous or partial thermoremanent magnetization origin for the secondary late Palaeozoic remagnetization is therefore not likely and the observed remagnetizations are interpreted to be of chemical (CRM) origin (Preeden, 2009).

Fluid interaction is a common mechanism invoked to explain many chemical remanent magnetizations (McCabe and Elmore, 1989). Migration of orogenic fluids during late Palaeozoic can be one mechanism that was triggered by post-folding processes either due to the Hercynian orogeny at the southern margin of Baltica or due to the effect of the collisional Uralian orogeny. Also, the magmatic Oslo rifting event in southern Norway can be one source for fluid migration.

However, even if the fluid movement mechanism due to orogenic events is a compelling explanation, environmental changes in the late Palaeozoic can be one possible explanation for the Permian remagnetization as well. Weathering can affect the original ferromagnetic minerals and result in the formation of new ferromagnetic minerals with attendant CRM components. In the early Palaeozoic the Fennoscandian Shield was buried under marine sediments and uncovered subsequently, giving rise to near-surface meteoric fluid circulation (Puura et al., 1999). Widespread Permian regression of the seas from old continental areas (see Zharkov and Chumakov, 2001 and references therein) coupled with the presence of hematite as the main carrier of the Permian overprint in the studied samples indicate that the base level of erosion lowered and the oxidizing fluids could have reached the older rocks.

As the secondary late Palaeozoic magnetization has been observed also world-wide (Zwing, 2003), we suggest that the formation of supercontinent Pangaea is the overall cause for all the processes described above (Preeden, 2009).

3. Conclusions

The area around the Finnish Gulf lacks the late and post-Palaeozoic deposits making the direct sedimentary studies for the last ~350 Ma unfeasible. Thus, palaeomagnetic studies can give new knowledge and largely contribute to the detection of younger events that have affected the region. In the area under discussion, Carboniferous to Neogene has been considered as a relatively stable continental period. However, several palaeomagnetic studies have revealed indications of younger remagnetizations. The occurrence of a late Palaeozoic overprint in different rock types (with different ages) on the Baltic Plate seems to be common. The migration of orogenic and/or meteoric fluids can be the reason for this remagnetization. Determination of the exact mechanism of remagnetization should be a focus of future geochemical studies.

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Breccia found from Vilppula drill core: Connection to the Keurusselkä impact structure?

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Vilppula drill core no. 2 is located within the Keurusselkä impact structure, central Finland. The drill core lithology consists of mica schist, mica gneiss and hornblende mica gneiss. At the depth of 100-110 meters we have identified a thin breccia vein. Here we describe the petrophysical and rock magnetic properties of these breccias. In addition, optical analysis of thin-sections was performed, in order to identify impact shock features in breccia.

Keywords: Keurusselkä, impact structure, breccia, Vilppula drill core

1. Introduction

Generally, during the crater formation, the subcrater rocks e.g. lithic breccias are formed within the lower displaced zone of the transient crater (French, 1998). Shock pressures in this zone are relatively low and shock features are limited to fracturing, shatter coning and brecciation, although deformation features in minerals can be produced in small volume by higher pressures.

The Keurusselkä impact structure, center 62°08’N, 24°37’E, is situated within Central Finland Granitoid Complex (Figure 1), which formed during Svecofennian orogeny (1890-1860 Ma) (Lehtinen et al., 2005). Keurusselkä structure is among the recently discovered impact craters in Fennoscandia (Dypvik et al., 2008). Keurusselkä impact structure is deeply eroded complex crater (D ~30 km, age ~1140 Ma) with impact features, such as, shatter cones with planar deformation features (Ferrière et al., 2010) and pseudotachyltic breccias (Schmieder at al. 2009).

2. Vilppula drill core and sampling

Vilppula drill cores are located in the edge of the central uplift of Keurusselkä impact structure (Figure 1). The drill core no. 2 consists of mica schist, mica gneiss, hornblende mica gneiss with occasional FeS_2 rich veins. The cores were drilled in 1968 by Suomen Malmi Oy in order to find ore deposits, as the area had noticeable magnetic anomalies. Unfortunately, the drill cores did not contain sufficient amounts of ore minerals and the cores were storaged and forgotten.

The interest towards Vilppula drill cores rise again with the discovery of Keurusselkä structure in 2004 (Hietala and Moilanen, 2004). In 2009 our group collected 44 samples from drill core no. 2 (Loppi, Geological Survey of Finland) for further analysis. Drill core rocks were generally fractured and gneisses had porfyroblasts. Breccias were found within the gneisses in depth of 100-110 metres indicating a brecciated target rock vein cutting the core (Figure 2).

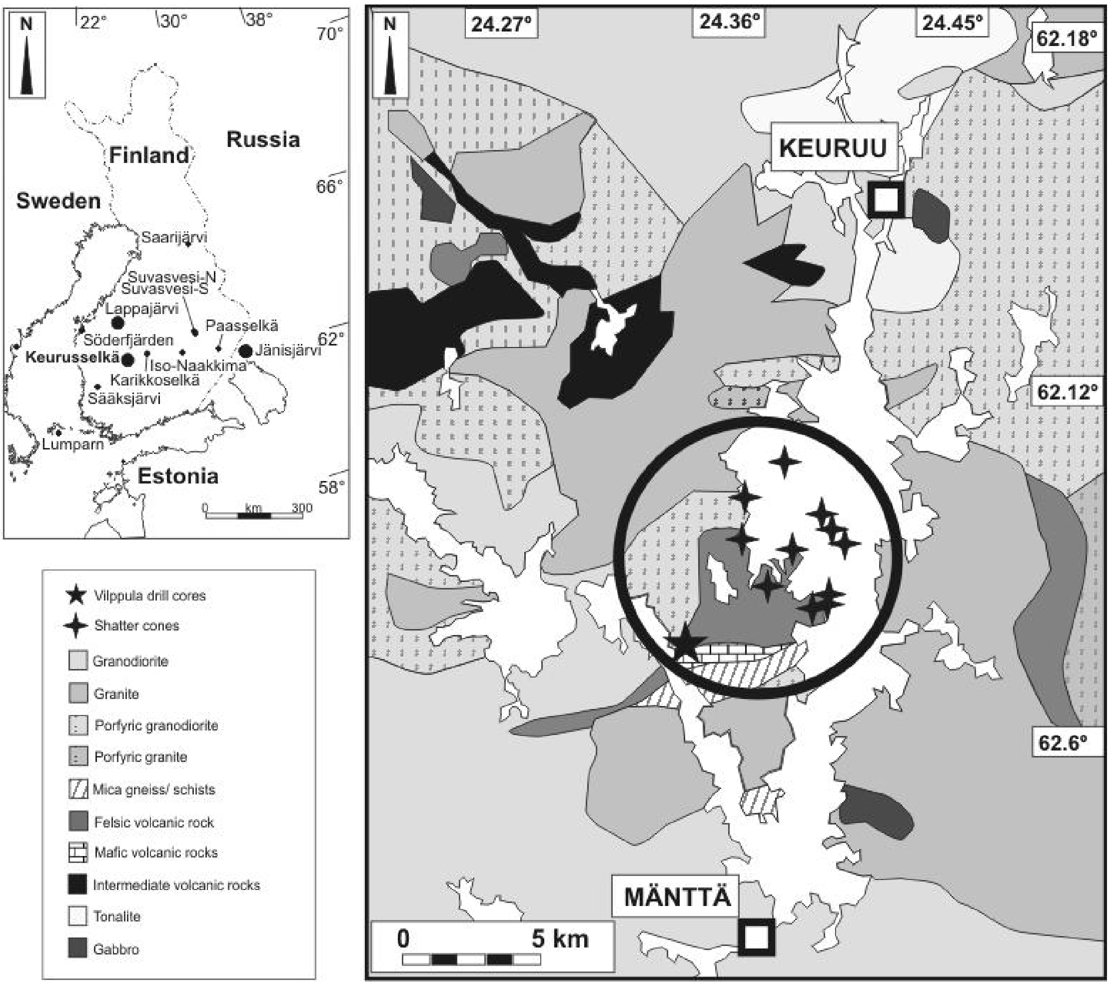

Figure 1. Geological map of the Keurusselkä area with location of Vilppula drill core. Circle represents the area of proposed central uplift with reported impact features (map with the locations of Finnish impact craters is modified after Dypvik et al., 2008; geological map modified after Nironen, 2003).

3. Methods

All petrophysical measurements were done in Solid Earth Geophysics laboratory at the University of Helsinki. Density, magnetic susceptibility and NRM were conducted using the Risto-5 Kappabridge for all the samples. Rock magnetic measurements for 13 selected bulk specimens (powder) included Curie temperatures and hysteresis. Curie temperatures were measured with AGICO's KLY-3S Kappabridge by heating the specimens in argon gas environment up to 700°C and back to room temperature. The magnetic hysteresis was measured using Princeton Measurement Corporation MicroMag™3900 Vibrating Sample Magnetometer (VSM).

4. Petrophysical and rock magnetic properties of drill core rocks and breccia

Target rock (mica schist, mica gneiss and hornblende mica gneiss) in Vilppula drill core is fractured influencing the densities of the samples. In general, schists and gneisses are also magnetically weak depending on the distribution of the magnetic fraction (Figure 2). Breccia, found between the depths of 100 to 110 meters, showed further decrease in magnetic properties. The Curie temperatures of 330°C indicated that the main magnetic mineral in the samples is pyrrhotite. The abundance of pyrrhotite, however, is low as the magnetic hysteresis, in addition to susceptibility, showed mainly paramagnetic behaviour. Interestingly, surface samples of mica gneiss showed also magnetite (Raiskila et al., 2010). Similar tendency has been reported for Bosumtwi impact structure, where drill core samples showed the presence of pyrrhotite while magnetite was observed in surface rocks (Elbra et al., 2007), which was explained by near-surface alteration of pyrrhotite into pyrite and magnetite.

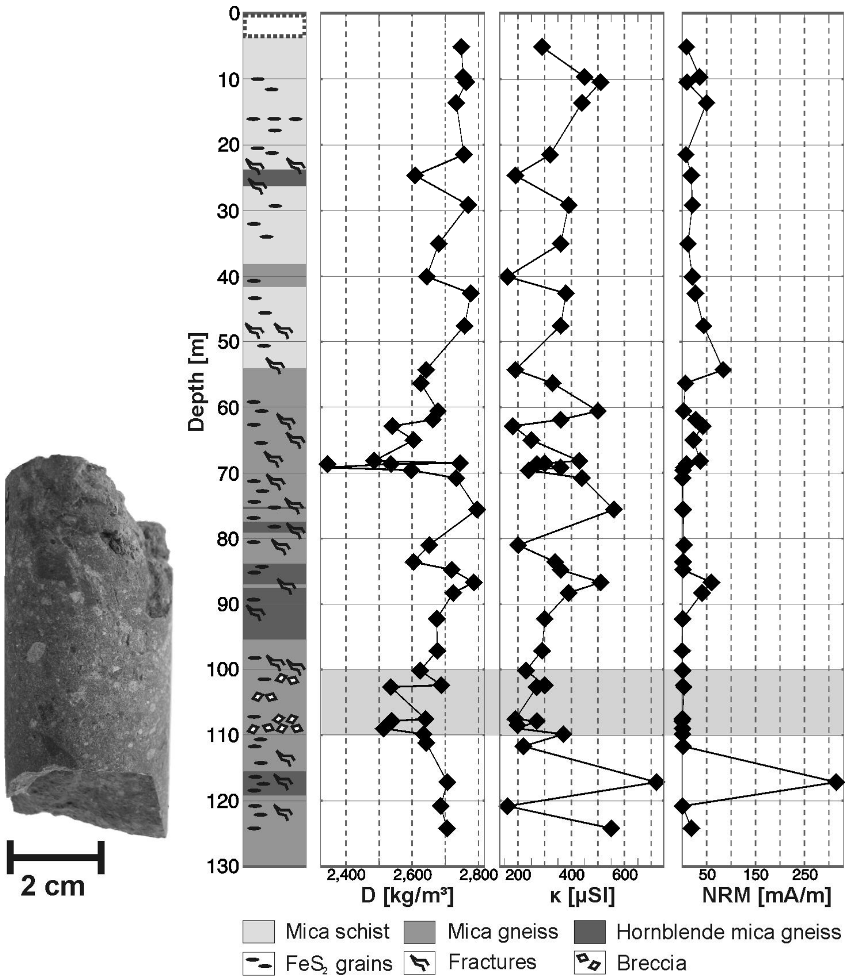

Figure 2. (Left) Brecciated mica gneiss from Vilppula drill core. (Right) Petrophysical properties of drill core samples with different lithologies.

5. Optical and SEM analysis of impact breccia

Studied breccias indicated that the fragments of primary target mica gneiss are surrounded by a fine-grained matrix of pulverized grains from the same rocks. Both lithic and mineral clasts with different size and shapes can be found.

Thin section petrography of four core samples of mica gneiss breccia samples show the dominance of the assemblage of plagioclase, K-feldspars, quartz, dark minerals like biotite and hornblende. Scanning electron microscopy (SEM) studies showed that biotite and hornblende are mostly heavily chloritized, which indicates a wide hydrothermal alteration. Chloritization in such extent has not been observed in non-brecciated target rocks (Poikolainen, 2008). Furthermore, the presence of monazite–(Ce) that usually forms lamellae and irregular aggregates within chlorite grains or fills some voids is clearly indicating secondary hydrothermal/metamorphic origin. In addition, different sulphides like chalcopyrite, pyrite and pyrrhotite were detected, with alteration of sulphides, seen as oxidation of the edges in some grains.

Planar deformation features in quartz, indicating impact shock pressures, were not found. However, studied rocks are heavily brecciated and altered by secondary processes possibly related to the impact.

6. Conclusions

So far, the reported impact features of Keurusselkä structure are shatter cones with planar deformation features, pseudotachylitic breccia and geophysical anomalies. The analysed samples of breccias from Vilppula drill core are similar to impact related lithic parautochthonous impact breccias in general. However, the Vilppula breccia zone is very thin representing only a vein cutting the drill core.

Impact-produced breccias may form beneath the crater floor, but these breccias are produced in relatively low pressures and late in the impact process (French and Koeberl, 2010). Such breccias generally lack distinctive impact features (planar deformation features, impact melt) and are difficult to distinguish from endogenic breccias. Nevertheless, Vilppula breccia is associated to the impact crater center and is within the area of previously determined clear impact features, such as planar deformation features in granitic rocks with peak pressures up to 20 GPa (Ferrière et al., 2010), and could, thus, be impact generated.

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Paleomagnetic and Rock Magnetic Studies on the 2.45-2.1 Ga Diabase Dykes of Karelia, East Finland - Key for Testing the Proposed Superia Supercraton

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We present here paleomagnetic data from Paleoproterozoic diabase dykes in the Taivalkoski area in the eastern Fennoscandian Shield. The majority of dykes show overprinted remanent magnetization directions caused by the Svecofennian orogeny. Three other magnetic components which may represent primary magnetization were separated. Components **D’** (D: 125°, I: -3°, α_{95}: 5°), **D** (D: 116°, I: 50°, α_{95}: 27°) and **E** (D: 295°, I: 44°, α95: 11.2°) are interpreted to correspond to magnetization ages ≥ 2.45 Ga; ≈2.45 Ga; and 2.06 Ga, respectively. Paleomagnetic data from Karelia compared to similar aged paleomagnetic data from Superior negates the tight Superia fit of Karelia and Superior cartons between 2.5-2.1 Ga.

Keywords: supercontinents, paleomagnetism, Paleoproterozoic, Superia, Karelia, mafic dykes

1. Introduction

Mafic dyke swarms provide good material for paleomagnetic and supercontinent studies due to their extensive spatial and temporal coverage. Recently, Bleeker and Ernst (2006) have presented a model of a long-lived (2.7-2.0 Ga) supercraton Superia where the Superior, Hearne, Karelia and Kola cratons are joined based on matching of two or more coeval 2.5–2.1 Ga dyke swarms on each craton. Due to the complex paleomagnetic data from the ca. 2.45 Ga units of the Karelian craton Bleeker and Ernst (2006) did not include the paleomagnetic information from the Karelian craton in their model. So far most of the paleomagnetic data (e.g. Mertanen et al., 1999, 2006) obtained for Karelia negate the Superia fit of Karelia and Superior. In order to get more evidence on the paleoposition of the Karelian craton, and to test the proposed model (Bleeker and Ernst, 2006) paleomagnetic and rock magnetic studies on several Paleoproterozoic mafic dyke swarms, especially in the Taivalkoski area in northern Finnish Karelia, have been carried out.

2. Sampling and geology

Altogether 44 Paleoproterozoic diabase dykes from the Taivalkoski region in the Karelian Province have been sampled. The sampled mafic dykes show two major trends, either E-W or NW-SE. These dykes extend from Finland to Russia. U-Pb zircon results indicate several dyke emplacement events in Karelia between 2.5 and 1.98 Ga (Vuollo and Huhma, 2005). Most of the dykes are well-preserved and undeformed on outcrop scale, but aeromagnetic data shows that the Taivalkoski area has suffered some deformation. Host rocks of the studied dykes are mainly Archean TTG gneisses. These were sampled in several sites for a baked contact test (Everitt and Clegg, 1962).

3. Results and discussion

Paleomagnetic results show that the dykes carry several remanence components. The majority of the dykes show an intermediate downward northwesterly component **A**. This is interpreted to be caused by the Svecofennian orogeny at ca. 1.88-1.84 Ga and it is typical for Karelian Paleoproterozoic formations (e.g. Mertanen et. al., 1999). Numerous dykes also show intermediate northeasterly component **B**. This also is typical for other formations in Karelia and it is interpreted to be related to the last stages of the Svecofennian orogeny, or even to younger events (Preeden et al., 2009). Here we discuss only those dykes that gave a result where we are most probably seeing trough the pervasive Svecofennian metamorphic event.

3.1. Component E

Three dykes show an intermediate westerly component **E** (D: 295°, I: 44°, α_{95}: 11.2°). This component has a high coercivity and high unblocking temperature. One baked sample shows similar results. Unblocking temperatures and Curie point measurements of dykes indicate magnetite to be the carrier of the remanence. Temperature vs. susceptibility curves and hysteresis property measurements favour single domain (SD) - pseudo single domain (PSD) magnetite, both domain sizes being able to carry stable paleomagnetic directions. These dykes have not been dated, but the virtual geomagnetic pole **E** plotted on the apparent polar wander path (APWP) of Baltica indicates an age of 2.1-2.06 Ga. Previously, a primary remanence direction corresponding to component **E** has been obtained from 2058 ± 6 Ma Kuetsyarvi formation in Karelia (Torsvik and Meert, 1995). Moreover, similar component was also seen as an overprint in the 2.44 Ga Koillismaa layered intrusions (Mertanen et al., 1989).

3.2. Component D

One dated dyke (2407 ± 35 Ma, Sm-Nd –age; (Vuollo and Huhma, 2005) and its baked host rock show a high coercivity and high unblocking temperature easterly-south-easterly component **D**. Both the dyke and the baked host rock show SD-PSD magnetite grains. The remanent magnetizations of the unbaked host rocks are weak and unstable, but clearly different from the baked host rocks. We interpret that the remanence in the baked host rock (D: 116°, I: 50°, α_{95}: 27°) may represent the primary magnetization. It is in close agreement with the direction obtained from from 2.45 Ga gabbronorite dykes in Russian Karelia (Mertanen et al., 1999, 2006). The remanence direction gives an intermediate latitudinal position for the Karelian craton at the time of remanence acquistions.

3.3. Component D’

A shallow (up, down) SE directed component **D’** (D: 125°, I: -3°, α_{95}: 5°) was obtained from the tonalitic baked host rock of one undated dyke. The host rock beyond baking zone gives a clearly different direction. **D’** component gives an equatorial latitude for the Karelian craton at the time of remanence acquisition. A corresponding low inclination component was previously obtained in two gabbronorite/ Fe-tholeiitic dykes at Lake Pääjärvi area (Mertanen et al., 1999) and it was suggested that it can be one candidate for the primary 2.45 Ga remanence. In the 2510 Ma (Bleeker, 2008) Shalskiy gabbronorite dyke in the Burakovka area, a low inclination component with higher declination was interpreted to represent the primary 2.5 Ga remanence (Mertanen et al., 2006).

3.4. Proximity of Karelia and Superior between 2.5 and 2.1 Ga

In order to study the suggested continuous proximity of the Karelian and Superior cratons during 2.5-2.1 Ga, we have compared similar aged paleomagnetic poles of 2.5 Ga, 2.45 Ga and 2.1 Ga from both cratons. As there are two possibilities for the primary 2.45 Ga remanence (component D or D'), we have tested reconstructions by using both alternatives (Fig. 1a and Fig. 1b, respectively) as a pair for the 2.45 Ga data of Superior. The reconstructions are compared with the Superia model of Bleeker and Ernst (2006) (Fig. 1c).

The reconstruction with D pole (Fig. 1a) separates the Karelian and Superior cratons about 45° at maximum. The 2.5 Ga and 2.1 Ga pole pairs support this reconstruction, implying that this separation lasted for the whole time period from 2.5 Ga to 2.1 Ga.

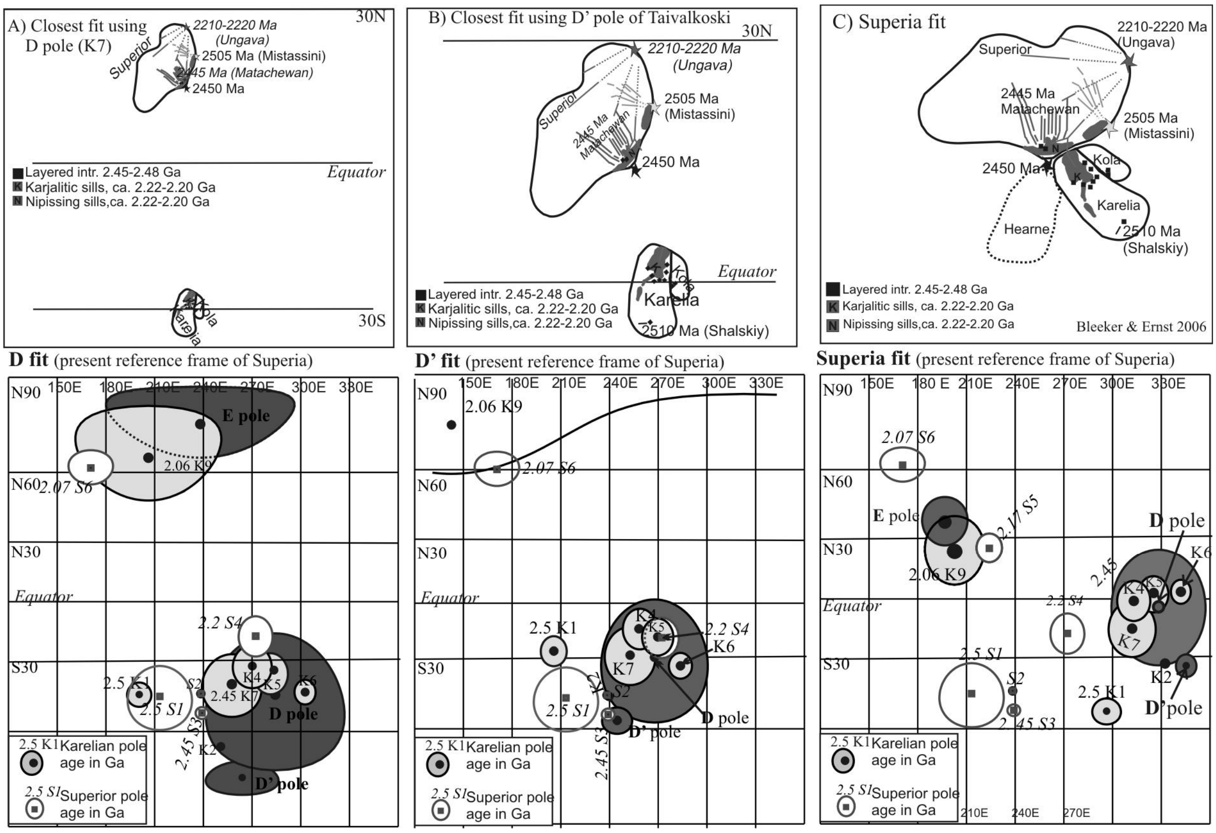

Figure 1. a) Closest fit of Karelia and Superior using D pole (K7) from Karelia (Mertanen et al., 2006) and Matachewan R pole (S2) from Superior (2446 Ma, Evans & Halls 2010). Karelian poles were rotated to present Superior reference frame according to Euler Pole (E.P.) 58°, 286°, -160°, **b)** Closest fit using D' pole of Taivalkoski and Matachewan N pole (S3). Karelian poles were rotated by using E.P. 63.5°, 299.5°, -173.6°, **c)** Superia fit (modified from Bleeker and Ernst, 2006). Karelian poles were rotated by using E.P. 72°, 251°, -114°. Poles: S1 = Ptarmigan-Mistassini 2505 Ma, S2, S3 = Matachewan R, N 2446 Ma, S4 = Nipissing N1 2217 Ma, S5 = Biscotasing 2170 Ma, S6 = Lac Esprit 2069 Ma. (S1-S6: see Evans and Halls, 2010), K1 = Shalskiy 2510 Ma, K2 = Pääjärvi D' ≥ 2450 Ma; K4,K5= Burakovka dykes and intrusion 2449 Ma; K6 Pääjärvi D; K7 Burakovka gabbro-norites ≈ 2.45 Ga; K9 = Kuetsyarvi 2058 Ma. (K1-K7: see Mertanen et al., 2006 and K9: Torsvik and Meert 1995).

However, most of the Karelian D-type poles are between the 2.45 Ga and 2.2 Ga poles of Superior suggesting that the magnetization age of the Karelian 2.45 Ga formations is between these ages. The reconstruction by using D'- pole (Fig. 1b) leaves a minimum of 10° (1111 km) distance between the cratons. According to this configuration, the 2.5 Ga poles do not overlap and thus do not support the presented model. The 2.06 Ga poles overlap to some extent. Interestingly, the 2.2 Ga Nipissing pole (S4) coincides with the D-type Karelian poles (K4-K7) suggesting the magnetization age of 2.2 Ga for the 2.45 Ga formations in Karelia. This would mean that the 2.45 Ga formations were remagnetized at 2.2 Ga in a geological event that affected the whole Karelian Archean craton. As there are no such age evidences, this configuration seems to be questionable. In the third case, by rotating the Karelian poles according to the Superia model (Fig. 1c) the Superior and Karelian poles do not overlap in any pole pair cases. The E pole of this study is close and the 2.06 Ga pole (K9) overlaps with the 2.17 Ga Biscotasing pole of Superior which is not reasonable. Therefore, based on these pole positions, the model (Fig. 1c) of Bleeker and Ernst (2006) is not supported by paleomagnetic data. From the two other configurations, the one with D pole (Fig. 1a) is supported by paleomagnetic data, but leaves the question of the reason for separation of the Superior and Karelia cratons at 2.5-2.1 Ga, as their unity is anyhow supported by geological similarities. The possibility that the D pole is a younger remagnetization formed in a widespread geological event between 2.4-2.2 Ga still has to be studied more thoroughly. The same applies if reconstruction with pole D' is valid (Fig. 1b).

4. Conclusions

Five remanence components have been isolated from the dykes and their baked host rocks in the Taivalkoski area. Component **A** and **B** represent deep penetrating Svecofennian remagnetization at ca. 1.88 Ga and possibly 1.75 Ga, respectively, as previously seen in many Archean rocks. Component **E** may represent the primary Paleoproterozoic ca. 2.06 Ga magnetization since a similar pole is seen also in the 2.06 Ga Kuetsyarvi formation (Torsvik and Meert, 1995). Component **D** is still interpreted to represent the primary magnetization of the age of ca. 2.45 Ga based on well-defined data from formations of that age. However, other possibilities have to be considered. The origin of component **D'** is open as we do not have yet any solid explanation for it, though some evidences point to its 2.45 Ga age. Paleomagnetic data negates the Superia fit of Karelia and Superior cartons between 2.5-2.1 Ga.

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The compilation of initial 3-D crustal model of POLENET/LAPNET research area, northern Fennoscandian shield

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The aim of this work was to build an initial 3-D crustal model for the use on upper mantle studies with POLENET/LAPNET data. The data used were models along previous 2-D controlled source seismic profiles and receiver function analysis at SVEKALAPKO stations.

Keywords: Moho, Fennoscandia, crust, 3-D modelling

1.POLENET/LAPNET project

POLENET/LAPNET project is a passive seismic array experiment in northern Finland with some stations also in northern Sweden, Norway and Russia. The experiment was a part of International Polar Year (IPY) 2007-2009. The data acquisition period of POLENET/LAPNET experiment was May, 2007 – September, 2009. One of the main targets of the experiment is to obtain a seismic model of the upper mantle using tomographic inversion of teleseismic travel times. In order to correct teleseismic travel times for crustal effect a seismic crustal 3-D model of the POLENET/LAPNET research area is needed.

The aim of this work was to compile a 3-D crustal model of northern Fennoscandian shield centred in northern Finland and extending to surrounding areas in Sweden, Norway, and Russia. The modelled area is located between 64˚ – 70˚ N and 18˚ – 34˚ E.

2. Available data

The new crustal model may be regarded as a northward extension of the 3D crustal model by Sandoval et al. (2003). It is based mainly on published models of previous 2-D controlled source seismic experiments. There are four main seismic profiles in our research area: wide-angle reflection and refraction profile FENNOLORA in Sweden (Guggisberg, 1991), wide-angle reflection and refraction profile POLAR and near-vertical reflection profile FIRE4 in Finland (Janik et al., 2009), and wide-angle reflection and refraction profile Kostomuksha-Pechenga in Russia (Azbel et al, 1989). All of these profiles are approximately north-south directed and POLAR and FIRE4 profiles are almost co-located. In addition to main previous profiles there are some earlier one shot-point profiles, but there are also quite large areas with no previous information at all.

In addition to previous 2-D models also receiver function information on Moho depth is available at the southern part of our research area, where receiver functions have been calculated for SVEKALAPKO station (Kozlovskaya, 2004).

3. The Method of compiling 3-D crustal model with CRUST3D program

CRUST3D program that is a new version of MakeUp3D by Waldhauser et al. (1998), is used for compiling a 3-D crustal model. By CRUST3D we may establish a 3-D crustal model based on data from controlled source seismic experiments, local source seismic tomography, and information from receiver function studies. The program is designed to take advantage of the different methodological strengths and to compile a 3-D crustal model that fits all available data within it's appropriate individual and methodological uncertainty limits.

The a priori information on Moho depth and crustal velocities is based mostly on existing 2-D near-vertical and wide-angle reflection and refraction models. The original data leading to the published models were carefully analysed to ascertain only information from locations where Moho reflectors/refractors were actually observed is used. In these locations Moho depth as well as average crustal velocities and, if possible, upper mantle velocities were derived from 2-D controlled source seismic models and quality estimates were obtained from the data.

During the compilation of 3-D crustal model, Moho reflector/refractor elements are off-line migrated in space based on a preliminary interpolation and mainly relying on information on nearby profiles. After 3-D migration of all derived Moho elements, final Moho surface is derived by a second interpolation round. With the Moho interface defined, all available CSS velocity information can be locally appointed to the corresponding grid.

The Moho interface is obtained by application of the principle of simplicity: the aim is to find a smoothest Moho interface that satisfies all reflector data within their a priori estimated error bars. It is important to note that we do not suggest our such derived 3D crustal model to be "correct" or the even geologically most likely model. Rather, our a priori model denotes the seismic model of least complexity fitting all presently available seismic information as this is the best initial reference model for further updates with new and additional data.

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Western margin of Fennoscandia: Electrical conductivity of the lithosphere

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Western margin of Fennoscandia is investigated by long period magnetotelluric study along a 350 km long profile across the Scandes mountain range in Jämtland-Tröndelag region, Sweden and Norway. Results show that lithosphere is thinning from 250-300 km in east in the Precambrian basement to 150-200 in the west at the coast of the Norwegian Sea.

Keywords: magnetotelluric, continental lithosphere, Caledonides, Fennoscandian Shield

We have carried out 90 broad-band magnetotelluric soundings along a 350 km long profile in Jämtland, Sweden and Tröndelag, Norway across the Central Scandinavian Caledonides (Fig. 1). We aim to determine the electrical conductivity of the (near-surface) accretionary wedge of the Caledonian orogen and the underlying autochthonous/parautochthonous carbonaceous alum shales as well as the electrical conductivity of the Precambrian crust beneath the Caledonides and the deep western margin of the continental Fennoscandian lithosphere towards the oceanic lithosphere beneath the Atlantic Ocean. Dimensionality analysis and regional strike estimates of the magnetotelluric data indicate that the conductivity structure can be approximated by a 2D model having a N40°E strike direction consistent with the dominant geological strike. The determinant averages of the magnetotelluric impedance tensor data together with the tipper transfer functions from the best 59 sites were inverted. The main features of the resulting conductivity model are (Fig. 2): (1) An electrically highly conducting layer beneath the Caledonides images alum shales, the autochthonous Cambrian carbon-bearing black shales on top of the Precambrian basement. Based on the comparison of electrical conductivity and seismic reflectivity models, we suggest that the Caledonian accretionary wedge above the conductor thickens in a step-wise manner from c. 1 km to 15 km towards the west, where, as suggested by the conductivity model, the wedge contains a considerable amount of resistive allochthonous basement rocks. (2) The upper crust of the autochthonous Precambrian basement below the Caledonides is homogeneous and resistive from the surface down to the depth of 15 km and can be associated with the TIB-granites. (3) The lower crust and the uppermost mantle in the easternmost part of the profile are very resistive whereas in the west they are two to three orders of magnitude more conductive. The increase in average crustal conductivity is related to the Caledonian processes or later opening of the Atlantic Ocean that have affected also the lower crust. (4) The depth to the lithosphere-asthenosphere border (LAB) changes from 250 – 300 km in the east to 150-200 km in the west. This transition maps the western limit of the lithosphere of the cratonic Fennoscandian Shield and indicates a rather steep thinning of the cratonic continental lithosphere at its margins. (5) A c. 100 km wide region of enhanced conductivity is detected under the Caledonides at the depth of c. 100-150 km. The nature of this feature, required by data, is unclear since other geophysical methods are not imaging anomalous mantle structures under the Caledonides. New studies, however, suggest that thermal processes may account for the recent uplift of the Caledonides (Pascal and Olesen, 2009) and consequently also for the electrical anomaly.

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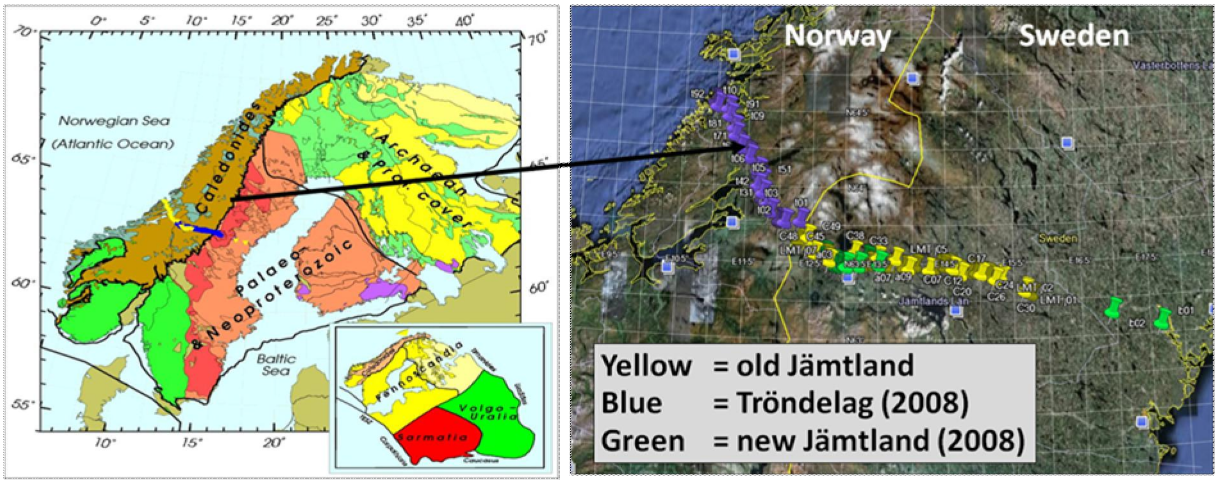

Figure 1. Location of the Jämtland-Tröndelag profile and the distribution of sites.

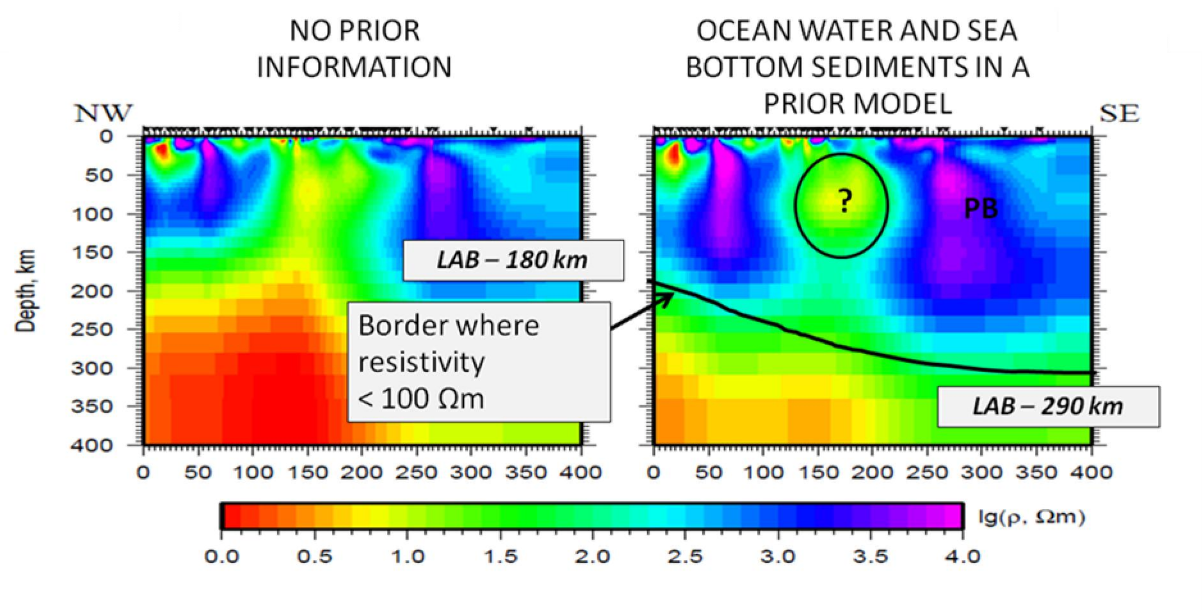

Figure 2. 2D inversion models. Left: unconstrained inversion. Right: A priori model used in the inversion. The prior model has a true ocean model (ocean water and ocean bottom sediments from Korja et al., 2002) to the left of the study area (not shown here).

Preliminary P- and S-wave velocity model of HUKKA 2007 wide-angle reflection and refraction profile: an evidence for an unknown terrain boundary?

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We present the preliminary 2-D seismic velocity model (Vp and Vp/Vs ratio in the crust, depth to the Moho and depth to the intracrustal reflectors) along a new HUKKA 2007 wide-angle reflection and refraction profile in northern Finland. The profile was measured using different commercial and military chemical explosions. The new velocity model suggests existing of previously unknown terrain boundary in northern Finland.

Keywords: Fennoscandia, lithosphere, crust, upper mantle, P- and S- wave 2-D raytracing modelling, Vp/Vs ratio

1. Introduction

Following Vaughan et al. (2005) the geological terrain can be defined as a fault-bounded package of rocks of regional extent characterized by a geologic history that differs from that of neighbouring terranes. Recently it was demonstrated that such features as the depth of the seismic Moho, the abruptness of the change in seismic velocity across the Moho, the velocity profile through the crust and the seismic velocity of the upper mantle may be used to infer the differences between geological terrains and trace the faults separating them (Bogdanova et al., 2006).

In our study we use the same approach in order to investigate crustal terrains and their boundaries in the northern part of the Fennoscandian Shield. For this we utilize results of a new HUKKA 2007 wide-angle reflection and refraction experiment that was carried out in autumn, 2007. We developed seismic P-and S-wave velocity models and estimated distribution of the Vp/Vs ratio along the HUKKA 2007 profile using forward raytracing modelling of refracted and reflected phases of both P- and S-waves.

2. Data acquisition

The HUKKA 2007 is a deep seismic sounding (DSS) profile which uses different commercial and military chemical explosions as sources of the seismic energy. The location of the 455 km long line (Fig. 1) of the profile running in NNW-SSE direction was defined by two large main sources (SP H and SP K) of seismic energy. The SP H (named after the Hukkakero hill) is the site where Finnish Army destroys old explosive materials. The SP K is located at the Kostamush open iron mine in the territory of Russia, where explosions of many tons of explosive materials are made few times per week. Few additional sources, open mines or quarries, located up to 14 km off the line of the profile, were also utilized.

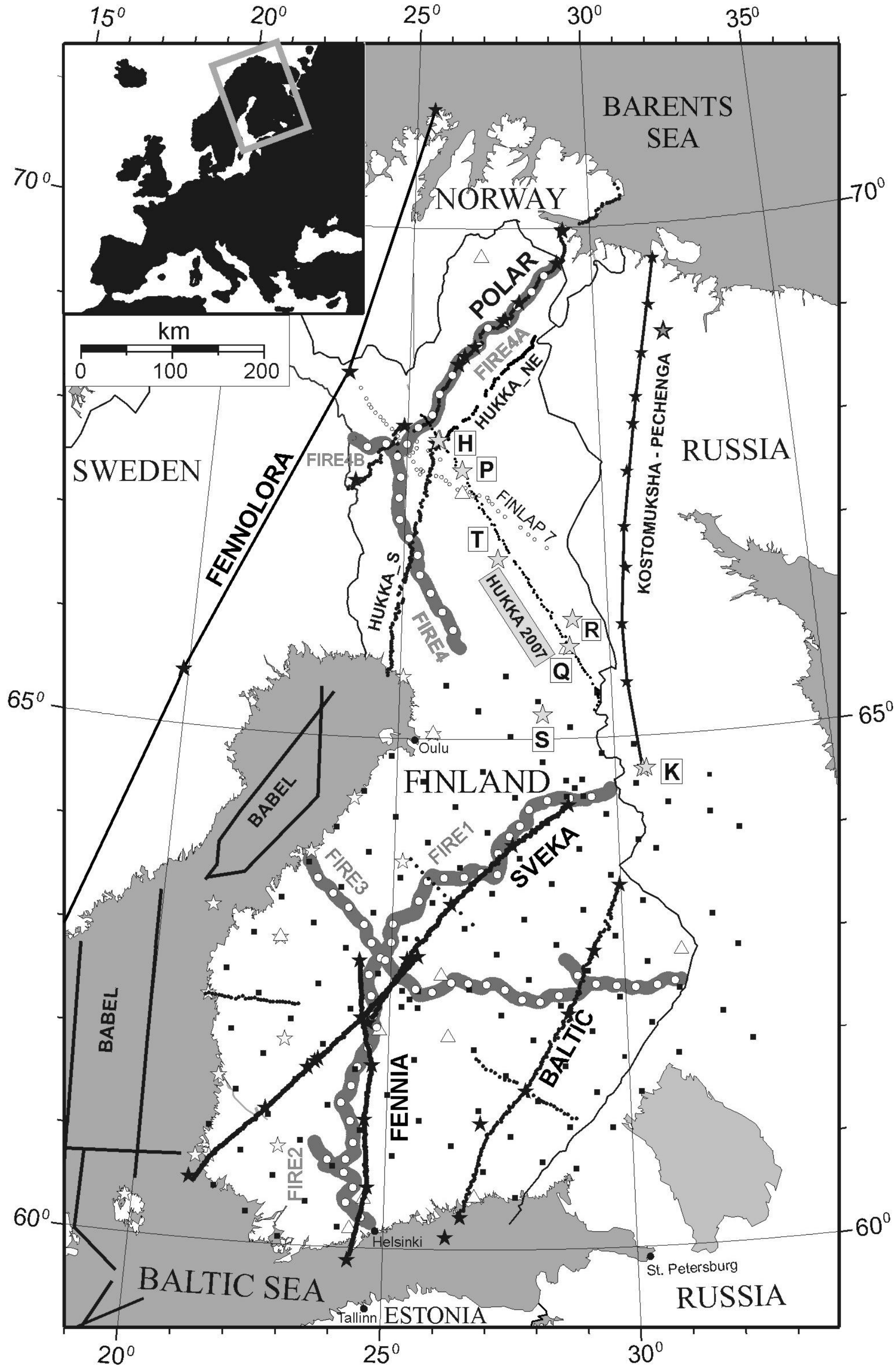

Figure 1. Location of main onshore and offshore seismic experiments in the eastern part of the Fennoscandia. The shots of the HUKKA 2007 profile are denoted with large stars and shot names. Explanations: stars represent the shot points of the profiles, H – Hukkavara, P – Pahtavaara; T – Tohmo, R – Ruka, Q – Kuusamo, S – Suomussalmi, K – Kostamush; black dots represent recording stations of DSS experiments; grey dots (seen as lines) represent of receiver stations and white dots represent selected CMP points of the deep seismic reflection experiment FIRE. Positions of record stations of the SVEKALAPKO temporary array are denoted by small black rectangles (temporary broadband and short period stations) and by open triangles (permanent stations). Inset map shows the location target area in Europe.

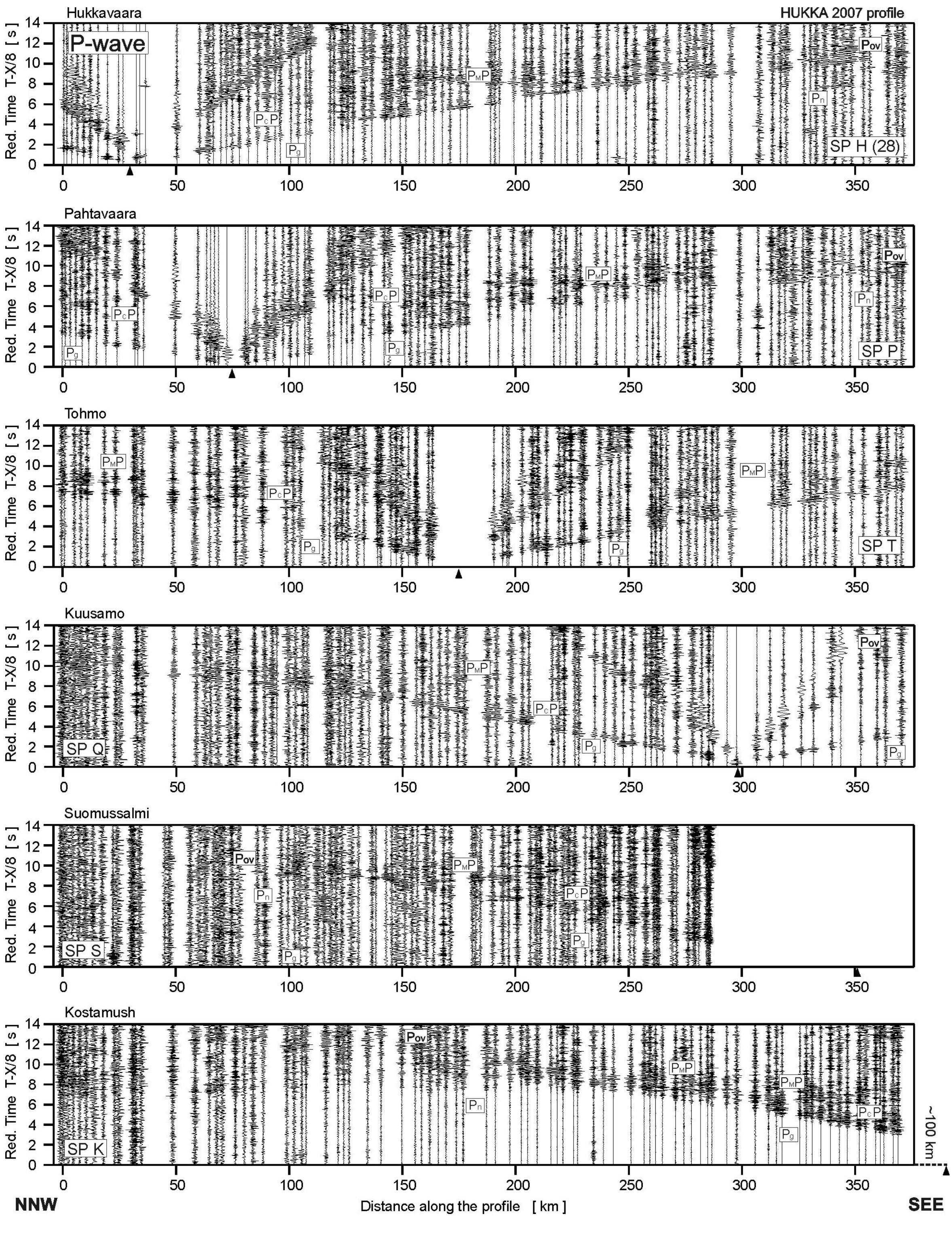

Figure 2. Examples of trace-normalized, vertical-component seismic record sections for P-waves along the HUKKA 2007 profile (SP H, SP P, SP T, SP Q, SP S, SP K) filtered by a band-pass filters (5-20 Hz). Abbreviations: Pg - seismic refractions from the upper and middle crystalline crust; Pov – overcritical crustal reflections; PcP – reflections from the middle crustal discontinuities, PmP - reflected waves from the Moho boundary; Pn - refractions from the sub-Moho upper mantle. The reduction velocities are 8.0 km/s.

Exception was the SP S located ca. 65 km off the line of the profile. The profile was extended to NNW, up to crossing with the POLAR profile. Measurements along the profile were carried out in August of 2007 by the Institute of Seismology, University of Helsinki using equipment and resources provided by several institutions from Finland, Poland, Hungary and Austria. The shots were recorded by 115 recording stations deployed along the profile with an average station spacing of 3.45 km. Recordings were made on the territory of Finland only; that is why the first 75 km from SP K were missed. The example of seismic records for all shots with exception SP R, are shown in Fig. 2.

3. Modelling results

The 2-D velocity model of the HUKKA 2007 profile was developed by SEIS83 forward raytracing package using arrivals of major refracted and reflected P- and S-wave phases. The velocity of the basement is 6.2–6.35 km/s. An interesting feature is a low velocity layer with Vp about 6.15 km/s at a depth of 5 to 20 km that can be followed in northern 350 km along the profile. In addition, the velocity model along the HUKKA 2007 profile shows significant difference in crustal velocity structure between the northern (up to 120 km) and southern (from 200 up to 400 km) parts of the profile. The difference in P-wave velocities and Vp/Vs ratio can be followed through the whole crust down to the Moho boundary. This suggests that the HUKKA 2007 profile crosses different crustal terrains. The central part of the profile (between 120 and 200 km) has a complicated structure in the middle and lower crust and has a zone of low (lower than 6 km/s) P-wave velocity in the uppermost crust. This zone is about 4 km thick and 100-150 km wide. It may correspond to a fault zone separating these two terrains.

4. Preliminary conclusion

The preliminary velocity model along the HUKKA 2007 DSS profile suggests existing of a previously unknown terrain boundary in the northern part of the Fennoscandian Shield. However, position of this new boundary with respect to major crustal units is controversial. It can be the boundary that that separates the non-reworked part of the Archean Karelian craton from its part reworked in Proterozoic. Alternatively, it can be the boundary that separates the Karelian craton from the Belomorian mobile belt.

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Mid-crustal flow in southern Finland

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The partial melting within the mid-crust leads, theoretically, to the rheological weakening and plastic flow of the middle crust. Models for mid-crustal flow are yet to be tested in the exposed orogenic mid-crust of the Fennoscandian shield. This presentation displays preliminary results and hypotheses of a pilot study in southern Finland. The results show a presence of many features compatible with the models for mid-crustal flow, which has implications to the temporal and spatial evolution of the Svecofennian orogeny. The results highlight the role of large-scale shear zones in the spatial distribution of the flow. Discussion about competing melting/recrystallisation processes during shearing in upper mid-crust is also invoked.

Keywords: core complex, middle crust, orogenesis, detachment, flow, shear zone

The solidus of rocks is approached and melts start to form at the upper boundary of the continental, thickened orogenic middle crust, leading to considerable rheological weakening of the mid-crust. Especially the so-called metamorphic core complexes (MCC) have proved to be crucial for the understanding of orogenic processes and the behaviour of the continental lithosphere during and after orogenesis. The term MCC is used for a variety of structures with different relationships and scales; we consider MCC at a lithospheric scale and in orogenic settings. In these settings, MCC usually refers to a situation where the middle-crustal high-grade metamorphic rocks are exposed, due to steep normal faults and/or low-angle detachments forming close to the base of the ambient, lower-grade crust (e.g. Lister & Davis 1989; Rey et al. 2009). The process leading to detachment-related MCC development in orogenic settings is often considered to involve the mobilisation (plastic flow) of the mid-crust, weakened due to partial melting, during a horizontal extension of a thermally mature orogen (e.g. Rey et al. 2009). The mid-crustal flow and/or extension is thought to be closely related to the collapse and/or (lateral) spreading of orogens, although the causal relationship remains unclear.

The knowledge of the nature and the behaviour of the middle crust during and after the Svecofennian orogeny is essential to the understanding of the processes that are responsible for the present crustal geometry of central and southern Finland. The present exposure level of the Svecofennian orogenic crust in southern Finland is very close to the upper-middle crust transition and, therefore, offers an excellent opportunity to study the behaviour of middle crust during and after orogenesis, and the formation and subsequent exhumation of core complexes. Furthermore, the deep seismic reflection profile 2A of the Finnish Reflection Experiment (FIRE) gives additional valuable information about the structure and properties of the Svecofennian crust.

The aim of the project is to test different models for mid-crustal flow against the Svecofennian rocks if southern and central Finland. In this poster, we present preliminary results and hypotheses based on field and petrographic studies, combined with interpretation of deep seismic reflection data from the West Uusimaa granulite complex (WUC). In the WUC, the dome-like structures have often been allocated to interference folding resulting in dome-and-basin style structures (e.g. Pajunen et al, 2008), and/or to magma diapirism (e.g. Scheurs & Westra, 1986). The aspects of possible flow processes in the Svecofennian mid-crust have not yet been widely considered.

The results show that the WUC displays many features compatible with the models for mid-crustal flow, which has implications to the temporal and spatial evolution of the Svecofennian orogeny. In addition, the structural grain of the mid-crust appears to be established relatively early during the orogenesis, controlling the spatial distribution of the flow. Especially the large-scale, steeply dipping shear zones play a central flow in the mid-crustal architecture. A possible low-angle detachment shear zone is also identified, containing evidence of competing melting and (dynamic) recrystallisation processes. In addition to the geological models of the Fennoscandian shield, the project has important implications to orogenic studies worldwide: analogues to e.g. the Himalayan middle crust from exposed, ancient, deep orogenic roots have been relatively scarce so far, although such analogues enhance the understanding the development and behaviour of modern orogenies.

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Thermodynamic modelling of pelite melting in Olkiluoto

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The bedrock of Olkiluoto consists mostly of pelitic migmatites. The typical biotite-sillimanite-cordierite assemblage of pelitic migmatites in Olkiluoto has a relatively large stability field from 560 ^{o}C and 2 kbar to about 850 ^{o}C and at least 8 kbar. The assemblages of other rock types as tonalitic gneisses in the region have even larges stability limits and the conditions of metamorphism are thus difficult to predict from assemblages alone. Further constraints may be set by phase equilibria of two simple systems, which may applied to metapelites, meta-arenites or silicic igneous rocks of Olkiluoto. Low- and medium-grade metapelites commonly contain muscovite, oligoclase (plagioclase), biotite and quartz and may thereby suffer migmatization due to dehydration melting of muscovite or biotite. On the other hand crystallization of water-bearing granitic magma often produce muscovite-, oligoclase-, quartz- and K-feldspar-bearing granite or pegmatite, by the reverse reaction of muscovite dehydration. Another simplified model system, which represents especially the pelitic migmatites of Olkiluoto, is biotite – sillimanite – plagioclase (An_{20}) – quartz – H_2O.There melting reactions in these two, relatively simple system may thus be used to predict 1) conditions of first melting, 2) conditions of peak temperature and 3) the amount of melt produced (rock/melt ratio). Modern thermodynamic data sets allow also modelling of natural systems with larger number of components. In this work we have modelled the melting reactions by the two simple systems as well as variable natural rock compositions in Olkiluoto. The results are compared to calculation made by standard mineralogical thermo-barometers and may be further applied in various field oriented studies of migmatites as observing melt/rock ratios (i.e. grade of migmatization), separation of melts, interaction between melting and rock structure generation etc.

Keywords: metapelite, migmatite, melting, reaction, dehydration, Olkiluoto, Finland

1. Introduction

The bedrock of Olkiluoto consists of pelitic migmatites, granite-tonalite gneisses, pegmatite granites and subordinate quartzites, biotite-hornblende-clinopyroxene gneisses and amphibolites (Fig 1.). The supracrustal rocks are deformed in three deformation phases (D_1 –

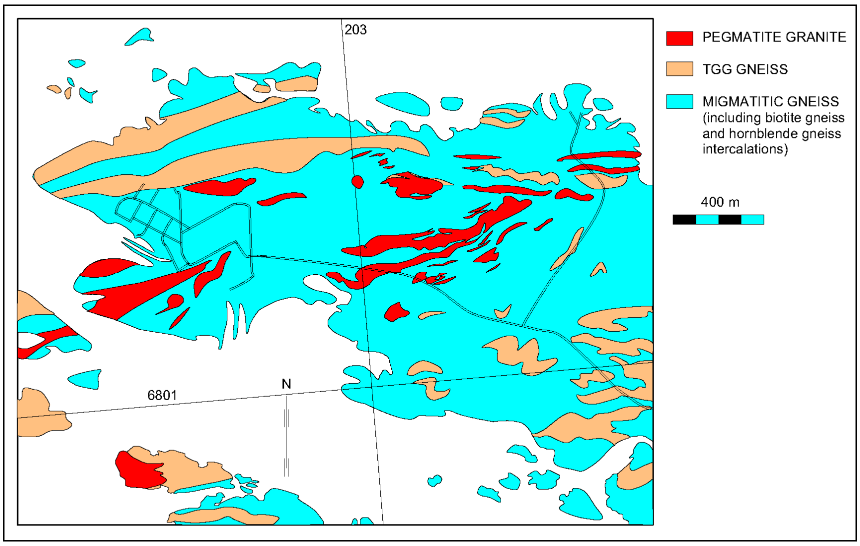


Figure 1. Geological map of Olkiluoto. Modified from Kärki and Paulamäki (2006).

D_3) during collisional and post-collisional stages of Svecofennian orogeny. Tonalites and granodiorites were intruded before or during the phase D_2 and potassium granites during the phase D_3 (Kärki and Paulamäki, 2006). Mänttäri et al. (2006) dated zircons from a tonalite. The zoned prismatic zircons gave an crystallization age of 1863 ± 6 Ma which is in accordance with the TIMS age from the southern part of the Rauma map sheet (Suominen et al., 1997) and sets the maximum age for the deformation phase D_2 (Kärki and Paulamäki, 2006). Mänttäri et al. (2006) obtained an 1822 ± 13 Ma age for a homogeneous zircon grain which is in agreement with the monazite age of Suominen et al. (1997) indicating post-intrusive thermal diffusion. The mineral assemblages produced during the peak of regional metamorphism (i.e. cordierite) overprint the S_2 foliation (Fig. 2). The peak of regional metamorphism accordingly took place between 1860 and 1820 Ma, the latter being the age of granite deformed by D_3 deformation (Tuisku and Kärki, 2010).

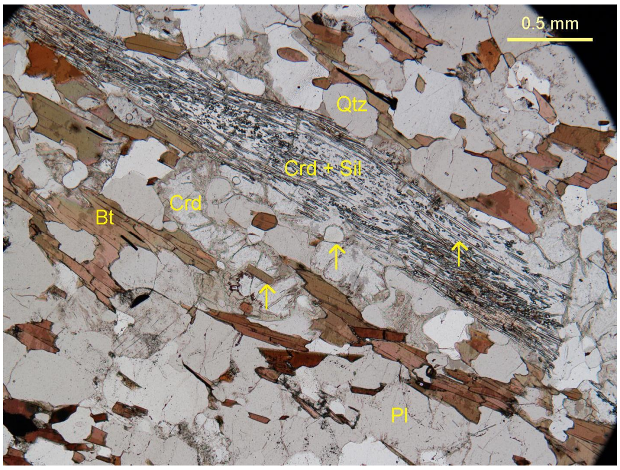


Figure 2. Micrograph of the mesosome in migmatitic gneiss Arrows from left to right point to biotite relic, quartz inclusion and sillimanite relic in cordierite porphyroblast. All inclusions are oriented parallel to S_2-foliation, while cordierite is undeformed (from Tuisku and Kärki, 2010).

2. Melting of model systems

Melting was modelled in systems A) 1 muscovite – 1 plagioclase (An_{20}) – 4 quartz – (0-10) H_2O and B) biotite – 2 sillimanite – plagioclase (An_{20}) – 4 quartz – (0-10) H_2O. The former illustrates muscovite dehydration melting in pelitic rocks as well as crystallization of leucogranites and pegmatites and the latter biotite dehydration melting in typical Olkiluoto metapelite. Results of the calculation, using modified Powell and Holland (2001, THERMOCALC 2003 version) with melt data of White et al. (2007) are shown in Figures 3 and 4, respectively. Melting of muscovite in system A begins at 652 °C at 4 kbar and all muscovite is consumed at 658 °C (Fig. 3). Excess water will retard formation of K-feldspar and in the higher temperatures assist dissolution of quartz from palaeosomes. Melting of biotite in system B begins at 666.5 °C at 4 kbar and all biotite is consumed at 825 °C, depending slightly on bulk water content (Fig. 4). Similarly to previous system, excess water will retard formation of K-feldspar and in the higher temperatures assist dissolution of quartz from paleosomes. Both systems clearly demonstrate that petrographic observation of migmatite mineralogy is fundamental for understanding such processes as fluid inflow, structural channelling etc. during migmatization processes and should be in key role in any study of migmatite terrains.

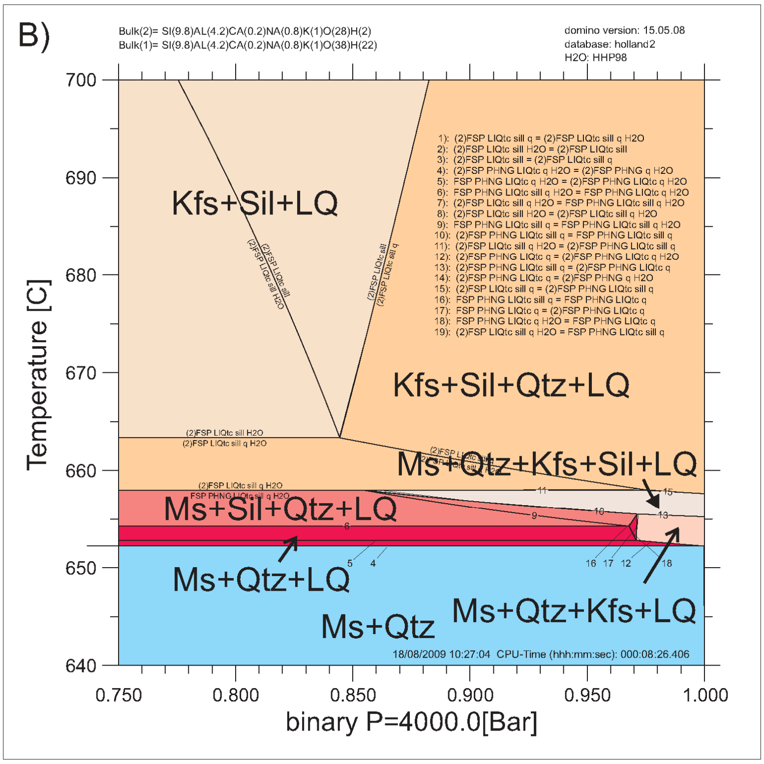


Figure 3. Dehydration melting in a model systems muscovite – 1 plagioclase (An_{20}) – 4 quartz – (0-2.5) H_2O.

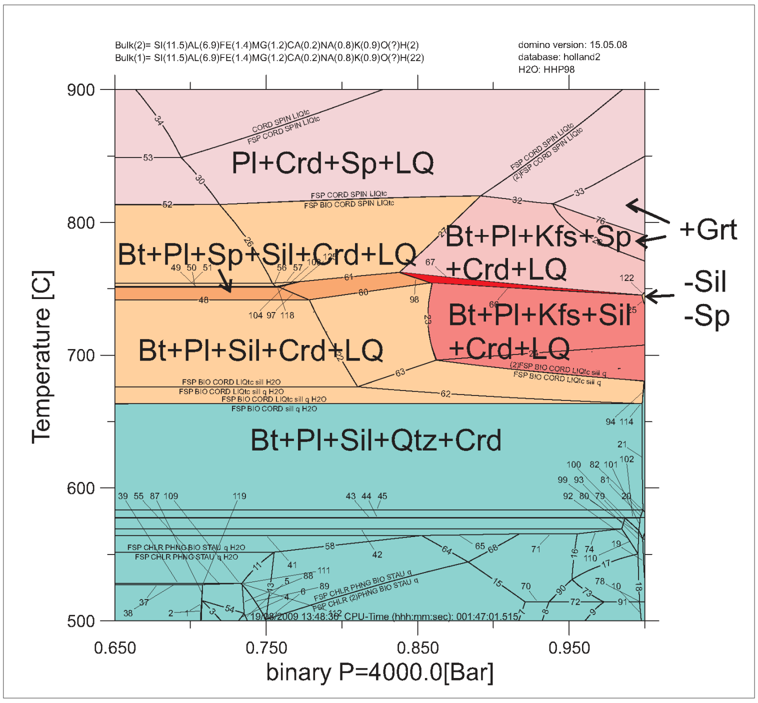
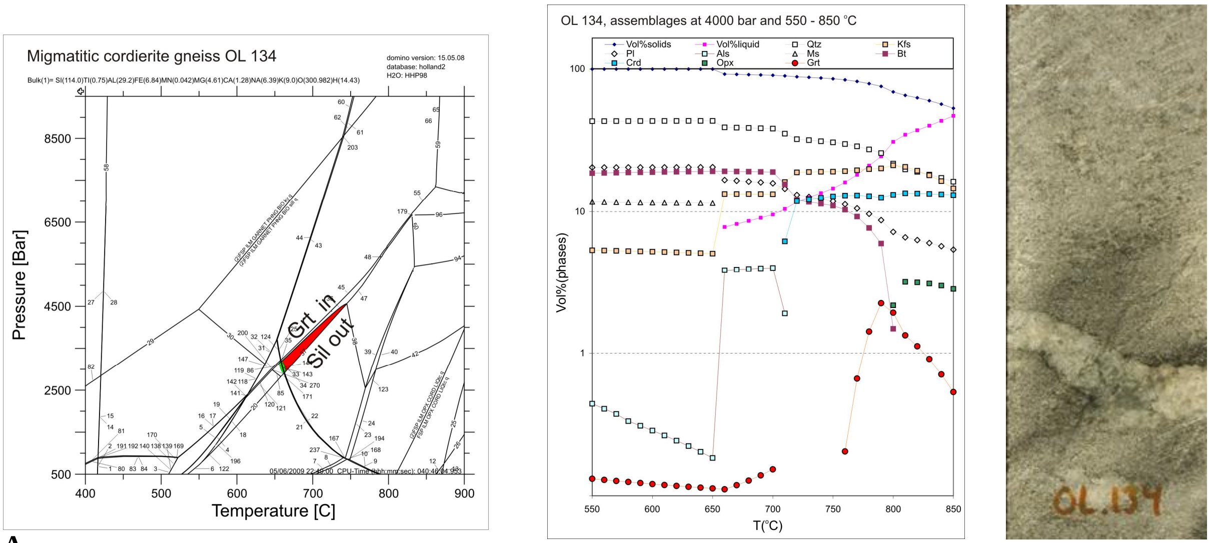


Figure 4. Dehydration melting in a model systems biotite – 2 sillimanite – plagioclase(An_{20}) – 4 quartz – (0-3.5) H_2O.

3. Melting of natural rocks

We calculated phase equilibria in several compositions of natural rocks from Olkiluoto migmatite complex. Figure 5 gives an example of such a calculation (sample OL134). Fig. 5A is calculated phase diagram of the sample. The dark red area is the stability field of the natural assemblage. 5B gives equilibrium volume ratios of phases during heating at 4000 bar from 550 to 850 °C. 5C is the drill core sample of the rock. In this sample, there is almost perfect fit between theoretical and observed volume ratios of all solid phases and as well as melt. The PT range indicated by natural assemblage versus phase equilibria calculation also quite well coincides with the results of standard thermobarometric calculations from several samples of this particular drill core.



A B C

Figure 5. Example of dehydration melting in a natural system from in Olkiluoto. The mode 14% leucosome, 35 % Qtz, 10 % Pl, 10 % Kfs, 18% Bt, 13 % Crd 1% Sil corresponds 700 – 715 °C at 4 kbar just above Grt out and below Sil out equilibria.

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Depth distribution of earthquakes under the Finnish part of Archean Karelian craton

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The thickness of seismogenic layer under the Finnish part of Archean Karelian craton (~64°-68°N, 24°-30°E) is outlined with the help of new velocity models, temporary seismic network and synthetic waveform modeling. The two-dimensional seismic velocity models are derived from passive seismic observations. In addition, an updated Moho depth map is constructed for the area by integrating the results of this study with previous observations from surrounding regions.
Lack of seismic activity within the high-velocity body in the uppermost 7 km of crust and occurrence of mid-crustal (20-30 km deep) earthquakes are characteristic features of the Kuusamo seismicity zone. In other parts of the study area, the uppermost crust is also seismically active. The focal depth distribution supports the "jelly-sandwich" rheological model of eastern Finland, which predicts that the middle crust is close to the transition from brittle to (semi-)ductile behaviour at about 30 km depth.

Keywords: seismogenic layer, crustal structure, Moho map, Archean Karelian craton

1. Introduction

Both the historical and the instrumental seismicity of Finland are concentrated to two seismically active zones in the Archean part of the bedrock. One is the Kuusamo district in north-eastern Finland and the other a broad zone running form the Bothnian Bay along the Tornionjoki river valley towards northern Norway. Earthquakes in these zones seem to occupy unusually wide depth range: from few km down to ~30 km. Some events on the Swedish side of the zone have even been located close to the Moho boundary (Arvidsson and Kulhanek, 1994; Arvidsson, 1996).

In this study we outline the depth distribution of earthquakes in the Finnish part of Archean Karelian craton (FKC). We will show that the divergent focal depth pattern observed in this area reflects real characteristics of local seismicity, not merely deficiencies in model parameters and station geometry.

2. Seismic data acquisition

To better constrain the structural data, we have set up a passive refraction seismic experiment and constructed two-dimensional (2-D) crustal velocity models for FKC. The layout of the experiment is presented in Figure 1a. Seismic sources for lines 1, 2, 4 and 5 are local earthquakes and explosions recorded by both the Kuusamo seismic network and the nearest permanent stations in 2003-2007. Line 3 is based on local events recorded by a subnet of broad-band stations during the SVEKALAPKO passive seismic experiment in 1998-1999 (e.g., Hjelt et al., 1996; Hyvönen et al., 2007). The survey has been complemented with five passive refraction profiles (lines 6-10 in Fig. 1a) acquired in 1980's and early 1990's (Yliniemi, 1991). The new P wave velocity models together with available observations from the surrounding regions have been used to compile an updated Moho depth map of the area (Fig. 1a; Grad et al., 2009).

2. Results and discussion

2.1 Structural models

The P wave models reveal a three-layer structure typical of the Archean segment of the Fennoscandian Shield. The upper crust is 12-20 km thick and associated with P wave velocities of 6.1-6.4 km/s. The high velocities are explained by mafic layered intrusions and volcanic rocks. The middle crust is a fairly homogeneous layer associated with velocities ranging from 6.5 to 6.8 km/s. The middle crust – lower crust boundary is generally located at depths between 30 and 35 km. The thickness of lower crust increases from 5-15 km in the Archean part to 15-22 km in the Archean-Proterozoic transition zone. The layer is associated with P wave velocities of 6.9-7.3 km/s. None of the 2-D models encompasses a distinct high-velocity layer (Vp >7.3 km/s) at the bottom of the lower crust. The average Vp/Vs ratio increases from 1.71 in the upper crust to 1.76 in the lower crust.

2.2 Moho depth and tectonic implications

The crustal thickness varies between 40 km and 52 km (Fig. 1a). The crust is thinnest (40-42 km) beneath the Pudasjärvi block bordering the Bothnian Bay. This Archean block has experienced rift-related extension and crustal thinning in the early Paleoproterozoic, but it is not significantly affected by the later Paleoproterozoic orogenic events (Korja and Heikkinen, 2008).

Three Moho depressions are located in the area: a crustal depression bulging out from the Archean-Proterozoic transition zone towards the Kuusamo region, and locally thickened crust under the Salla and Kittilä greenstone belts. The depressions are tentatively associated with collision zones that have been initiated or reactivated during the Paleoproterozoic orogenic events (e.g., Kontinen and Paavola, 2006; Patison et al., 2006; Korja and Heikkinen, 2008).

2.3 Depth extent of seismogenic layer

Figure 1b shows hypocentres of 27 earthquakes from the Kuusamo-White Sea seismicity zone. The data set has been relocated by employing the new structural models and the focal depths are either verified or fixed with the help of synthetic waveform modeling (Figs 1c-d).

The cross-section reveals a south-westward deepening of the seismogenic layer: earthquakes are clustering at about 10 km and 20 km depth in the Russian and Finnish part of the zone, respectively. Within the Kuusamo network, no hypocentres shallower than 7 km are found and the seismogenic layer extends down to a depth of about 25-30 km, i.e., to the lower part of the middle crust. Lack of seismicity in the uppermost crust seems to be a characteristic feature of the Kuusamo area, because near-surface earthquakes are common in other parts of the Archean Karelian craton. The seismically quiet upper crustal segment roughly coincides with the mafic high-velocity body detected beneath the local Kuusamo network. The deepest earthquakes appear to cluster in the area where the local Moho depression is found.

The depth extend of the seismogenic zone is in line with the rheological profile of lithosphere derived by Kukkonen and Peltonen (1999) from xenolith-controlled geotherm in eastern Finland (at the SE end of line 10; Fig. 1a). The model predicts that the middle crust is close to the transition from brittle to (semi-)ductile behaviour at about 30 km depth. The model further suggests that the upper part of the lower crust may be brittle, whereas the lower part behaves in a ductile way. Furthermore, the FIRE 4 sections reveal a vertical change in crustal reflectivity properties at the same depth range beneath the neighbouring Pudasjärvi block (Kukkonen et al., 2006; Patison et al., 2006).

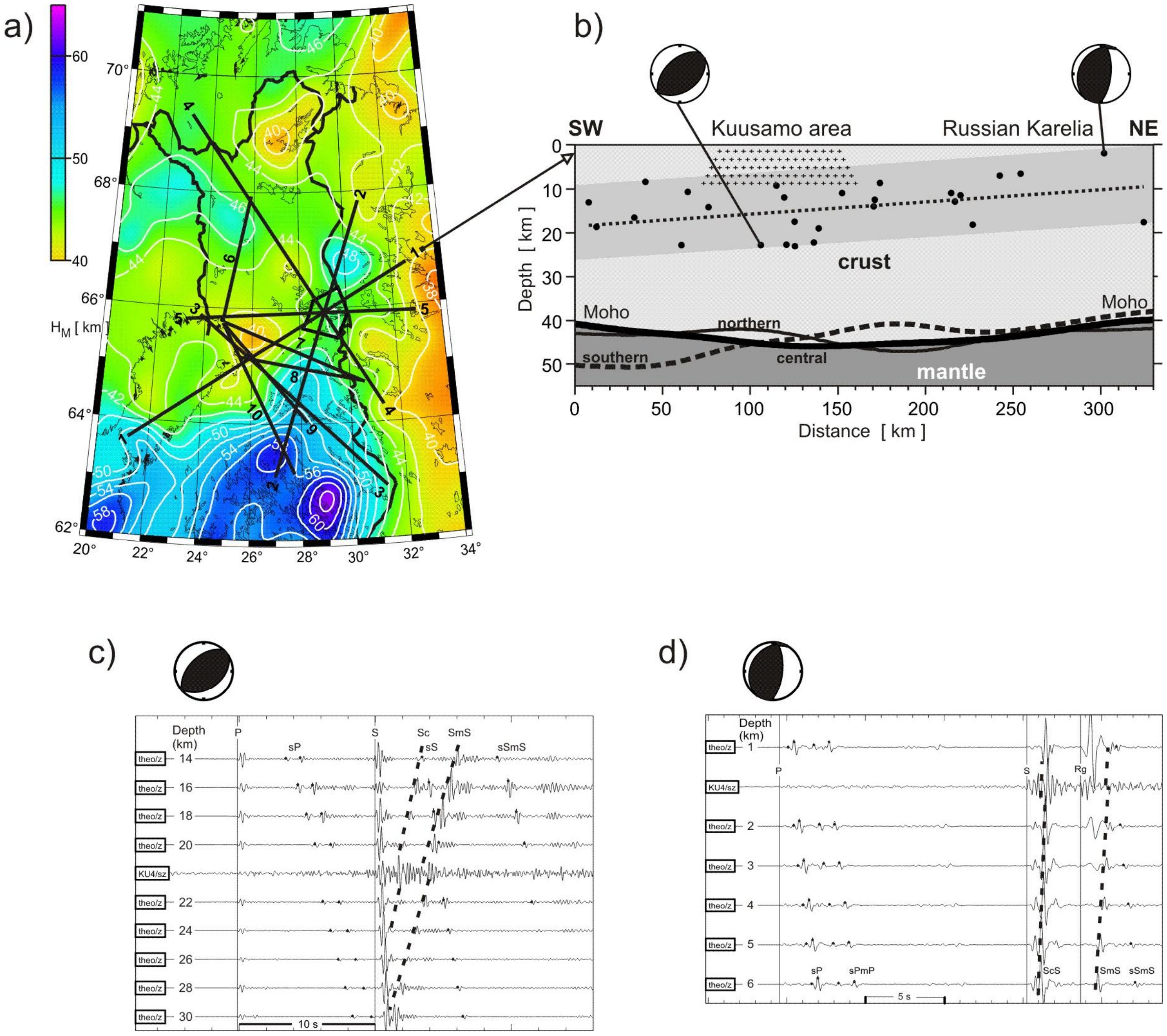

Figure 1. Updated Moho depth map (a) and depth distribution of earthquakes (b-d) for FKC**.**

a) Isolines of Moho depth (white lines) are drawn at every 2 km. Black lines (1-10) display the profile layout and black polygon outlines the limits of the local Kuusamo network.

b) A SW-NE profile crossing the area (NE end of line 1; Fig. 1a). The 27 earthquakes (black dots) located less than 50 km away from the profile are projected onto it. Moho depth curves along the main profile and profiles 100 km to the SE and NW are superimposed. A high velocity body beneath the Kuusamo area is marked with crosses.

c-d) Examples showing vertical-component recordings of station KU4 for an earthquake at 89 km distance and 22 km depth (c) and 138 km distance and 2 km depth (d). The theoretical seismograms are computed with the reflectivity method (Kind, 1978, 1979) using the new structural models and preliminary focal mechanisms (top). The numbers superimposed on synthetic waveforms denote focal depth. Vertical lines show the onset of first P, S and Rg on the original records, whereas dashed curves and black dots mark intracrustal (Sc, SmS) and surface (sP, sS, sPmP, sSmS) reflections traced on the theoretical seismograms. All traces are aligned with the first P, scaled uniformly and filtered with a narrow band-pass filter centered at 3 Hz.

3. Conclusions

In the Kuusamo seismicity zone, the seismogenic layer extends from about 7 km below the surface down to a depth of 25-30 km. In other parts of the study area, the uppermost crust is also seismically active.

The depth distribution is in agreement with the "jelly-sandwich" rheological model of eastern Finland. The model suggests that a (first) change in rheological properties of the crust occurs at this depth range.

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Magnetotelluric studies of the collisional and extensional processes in the central Fennoscandian Shield

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We study the architecture of the Archaean-Proterozoic collisional zone in the Fennoscandian Shield. We also and investigate the difference between the reworked Archaean crust close to the border zone and the intact Archaean crust further to the east. The third target area of this study is the western margin of the Central Finland Granitoid Complex (Bothnia), where we can observe crustal structures caused by the collision of the Bothnian mc with the Keitele mc and extensional collapse of the Svecofennian orogen. We use broadband magnetotelluric data and perform two-dimensional inversions along several profiles. The main features from the Kainuu area models include the graphite and/or sulphide-bearing metasedimentary rocks formed on surface and transported to deeper crustal levels in the collision of Karelian craton and Keitele mc (Savo orogeny). They are imaged within the Kainuu and Savo Belts as highly conductive structures flanking the resistive Archaean Iisalmi Complex. The main features in the models of the Bothnia region include a set of SE-dipping conductors in the upper crust representing the collisional structures of the Bothnia Belt. Conductors coincide very well with listric reflectors seen in seismic results. These conductors are abruptly cut by a large resistive region under the Central Finland Granitoid Complex. This may indicate that originally conductive graphite- and sulphide-bearing upper crustal metasedimentary rocks and conductive lower crustal rocks have been deformed by later processes. It is suggested that these processes destroyed the connectivity of a conductive phase and made the rocks resistive for magnetotellurics.

Keywords: magnetotellurics, crust, collision, extension, Svecofennides, Fennoscandia

According to Korja, A. et al. (2006) the Svecofennian orogen was formed in four stages which partly overlapped in time and space: 1) microcontinent accretion (1.92-1.88 Ga), 2) the large-scale extension of the accreted crust (1.87-1.84 Ga), 3) continent-continent collision (1.87-1.79 Ga) and 4) gravitational collapse (1.79 and 1.77 Ga). During the first phase, the island arcs, which were formed after the Archaean rifting, started to assembly and make microcontinents (mc) such as Keitele, Bothnia and Bergslagen. These microcontinents collided against the Archaean Karelian nucleus from several directions, e.g. at the western margin as the Lapland-Savo orogeny (Bothnia mc and Keitele mc). The latter peaked at the collision of the Keitele microcontinent and the Karelian craton. The crustal growth continued southwestwards in the Fennian orogeny as a collision between the Keitele and Bergslagen microcontinents. When the active compression within the newly formed terrain ceased, the gravitationally and thermally unstable interior of the continent was subject to extensional processes. This phase produced large scale extensional basins. The third phase began with the collision of Fennoscandia and the Kola craton in the northeast and was followed by the docking of Sarmatia in the southeast. The latter started the Svecobaltic orogeny, which evolved with another yet unknown continent docking in the southern margin. The Nordic orogeny at the western edge marked the collision of the Fennoscandia and Amazonia continents, and left Fennoscandia finally positioned in the centre of a supercontinent. The last phase for the Fennoscandian evolution was a stabilization period with gravitational collapse, thermal resetting and late tectono-magmatic episodes.

We use the broadband magnetotelluric data and two-dimensional determinant inversions of several profiles (104 sites) to study the architecture of the crust at the Karelian-

Svecofennian collisional zone (Kainuu area) and in the Karelian craton far from the collisional zone (Kuhmo area). The third target area of this study is the western margin of the Central Finland Granitoid Complex (Bothnia), where we can observe crustal structures caused by the collision of the Bothnian mc with the Keitele mc and extensional collapse of the Svecofennian orogen. Both of these areas are also imaged with reflection seismic studies (Kukkonen and Lahtinen, 2005). Our primary aim is to study the tectonic evolution of the central part of the Svecofennides by combining electromagnetic data with seismic data.

We have used three approaches for dimensionality and strike analysis, viz. Swift (1967), Bahr (1988) and Q-function (Zhang et al., 1987; Smirnov and Pedersen, 2009) methods. Q-function analysis suggests that data are 2D at shorter periods but at periods above 100 s do not fit with the underlying assumption of the 2D regional structure. We have therefore restricted further analysis to periods below 100 s, i.e. for crustal structures. Various tests with the Q-function analysis suggest that the best strike for 2D inversion is N-14ºW, which has been used for 2D determinant inversions. Determinant inversion was selected (1) to increase the amount of data suitable for inversion since in nearly half of the sites one polarization is out-of-quadrant whereas determinant responses remain in the first quadrant, (2) to ease with the strike determination since the determinant is rotationally invariant and the strike direction is needed only to project profiles in a proper direction for inversion and (3) to suppress 3D effects on data since, according to Pedersen and Engels (2005), determinant response is the smoothest and robuster against 3D distortions. In inversion, we have used the REBOCC code (Siripunvaraporn and Egbert, 2000) modified for determinant inversion by Pedersen & Engels (2005). The stability and quality of the final inversion model has been assessed with various inversion tests to, for example, investigate the effect of data errors (or error floors) or to investigate the stability of certain elements in the model. Finally, we have tested to what extent it is possible to replace smooth models with geologically more appropriate rough models.

The main features from the Kainuu area models include the graphite and/or sulphide-bearing metasedimentary rocks formed on surface and transported to deeper crustal levels in the collision of Karelian craton and Keitele mc (Savo orogeny). They are imaged within the Kainuu and Savo Belts as highly conductive structures flanking the resistive Archaean Iisalmi Complex. The main features in the models of the Bothnia region include a set of SE-dipping conductors in the upper crust representing the collisional structures of the Bothnia Belt. Conductors coincide very well with listric reflectors seen in seismic results. These conductors are abruptly cut by a large resistive region under the Central Finland Granitoid Complex. This may indicate that originally conductive graphite- and sulphide-bearing upper crustal metasedimentary rocks and conductive lower crustal rocks (Lahti et al., 2005) have been deformed by later processes (a gravitational collapse of the over-thickened Svecofennian crust and intrusion of granites). It is suggested that these processes destroyed the connectivity of a conductive phase and made the rocks resistive for magnetotellurics.

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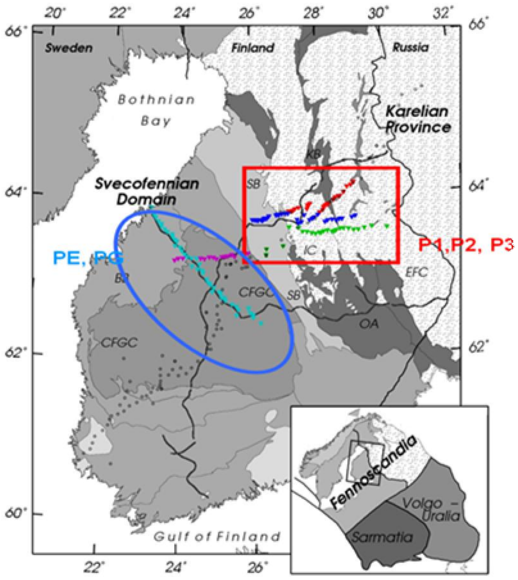

Figure 1. Location of the research areas. Red rectangular: MT-FIRE profiles across the Archaean-Proterozoic boundary. Different colours (red, blue green) the three east-west directed 2D inversion profiles. Blue ellipse: Region to study the Bothnian Belt and Central Finland Granitoid Complex.

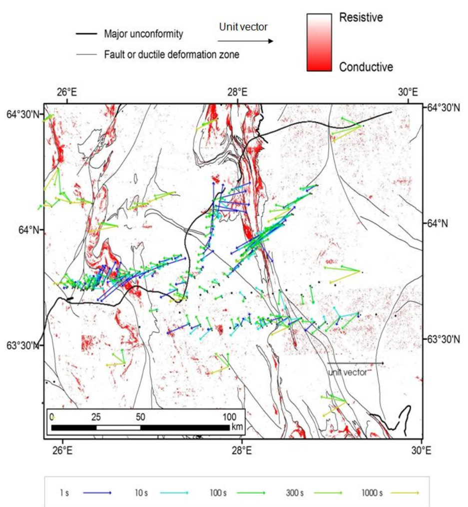

Figure 2. Near-surface conductors (red) from airborne electromagnetic mapping (Ario, 2005) and induction arrows from this study. Arrows point towards conductors.

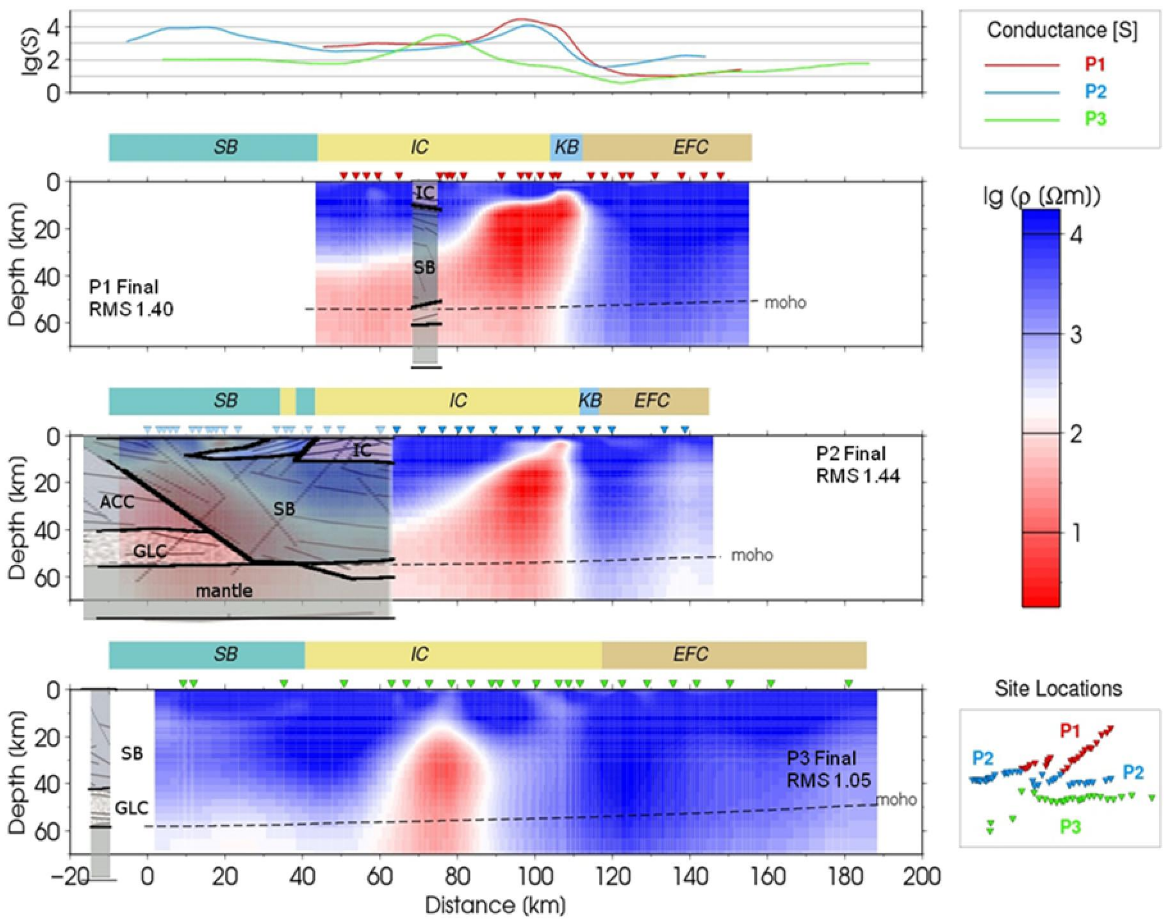

Figure 3. Final 2D inversion models from the Archaean-Proterozoic boundary zone (Proterozoic Pyhäsalmi Belt, Archaean Iisalmi Complex, Proeterozoic Kainuu Belt and Archaean Eastern Finland Complex).

Testing the GAD model of the geomagnetic field by using igneous rock data

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According to the fundamental assumption used in paleomagnetic studies, the time-averaged magnetic field of the Earth is described by a geocentric axial dipole model (GAD). However, high-quality data from well-mapped cratons shows that the field actually deviates from that hypothesis in long timescales. Out of various testing methods of the GAD, this study used inclination data from Precambric rocks. It was concluded that an octupolar component with the strength of 10 % of GAD must be added to fit a reasonable model to observations. A field like this is much closer to GAD than fields obtained from previous studies.

Keywords: GAD, paleomagnetic, model, inclination, distribution, craton.

1. Introduction

The magnetic field of the Earth is described by a vectorial sum of dipole and non-dipolar components, such as quadrupole, octupole etc. Currently, the dipole forms only ca. 75 % of the total field. When the field is averaged over a timescale long enough, it is assumed that the non-dipolar fields vanish according to the geocentric axial dipole hypothesis (Merrill et al. 1995). There are several ways to test this theory, e.g. paleoclimate, pelagic sediments and reversals of the field. This study was based on global precambric data from 23 different geologic units (British Isles, Greenland, Scotland, Svalbard, Ukraine, Ural, Baltica, other Europe, Antarctica, Grunehogna, Arabia, Australia, India, North China, South China, Laurentia, South America, Africa, Madagascar, Seychelles, Siberia, Tarim and Taimyr). Particular focus was on high-quality data from igneous rocks of five best-mapped cratons with large bare shield areas. Some other cratons, prominently Eurasian ones, contain less useful sedimentary data only.

2. Modelling

The strength of the global geomagnetic field undulates with respect to location, so it is reasonable to describe it with spherical harmonics. Models, such as the IGRF (International Geomagnetic Reference Field) use this theory. When the degrees of the field are concerned, those smaller than 14 (2^{14}=16384 poles) are generated from the main field and higher ones from crustal and external fields. In most tests of the dipole model, including this study, only fields up to 2^3=8 poles have been considered, as smaller ones cannot be properly modelled due to errors in the data.

Each combination of dipole (G1), quadrupole (G2) and octupole (G3) field follows a certain inclination (I) distribution. Here the latitudinal binning has been made with the interval of 10 degrees. Actually, I spans from -90° to +90°, but in studies like this, the sign doesn't have any influence. Distributions of G2 and G3 are flatter than that of GAD, and generally, G3 prefers low inclinations and has a sign opposite to the sign of GAD. For distributions of G1, G2 and G3, see Table 1. In GAD, the median of |I| should be at 49°, but in datasets from many continents, it is closer to 40° or even 30°. For the purposes of modelling, inclination is better than declination, because the |D| vector is poorly determined in the proximity of geomagnetic poles.

3. Data and analysis

The paleomagnetic database collected by University of Helsinki contains 2821 observations. All of them are from rocks older than 542 million years. When the data is sorted by rocktype, igneous rocks are the dominant ones, followed by first sediments and then metamorphics. Every observation has an AV number that tells its quality. In the scale, 6 is best and 0 is worst. In this case, the data was filtered with the criteria AV≥3. As the geologic units wander randomly through the surface of the globe, all cratons with a sufficient coverage of data are applicable. For further analysis, Australia (AU), Baltica (BA), Greenland (GR), Laurentia (LA) and Scotland (SC) were selected, with the total number of 623 observations. Rocks other than igneous ones were neglected. The reason for that was the shallowing of inclinations of sedimentary rocks, as seen in Figure 1. In addition to that, sedimentary data is too scarce in oldest temporal intervals (rocks older than 2 billion years). Igneous observations were sorted by using 10-degree intervals of |I| and the percentage of observations in every interval was calculated. A model with certain values of G1, G2 and G3 was fit into the data and results were compared with GAD. Calculations and sketching of distributions was done by using a Python script.

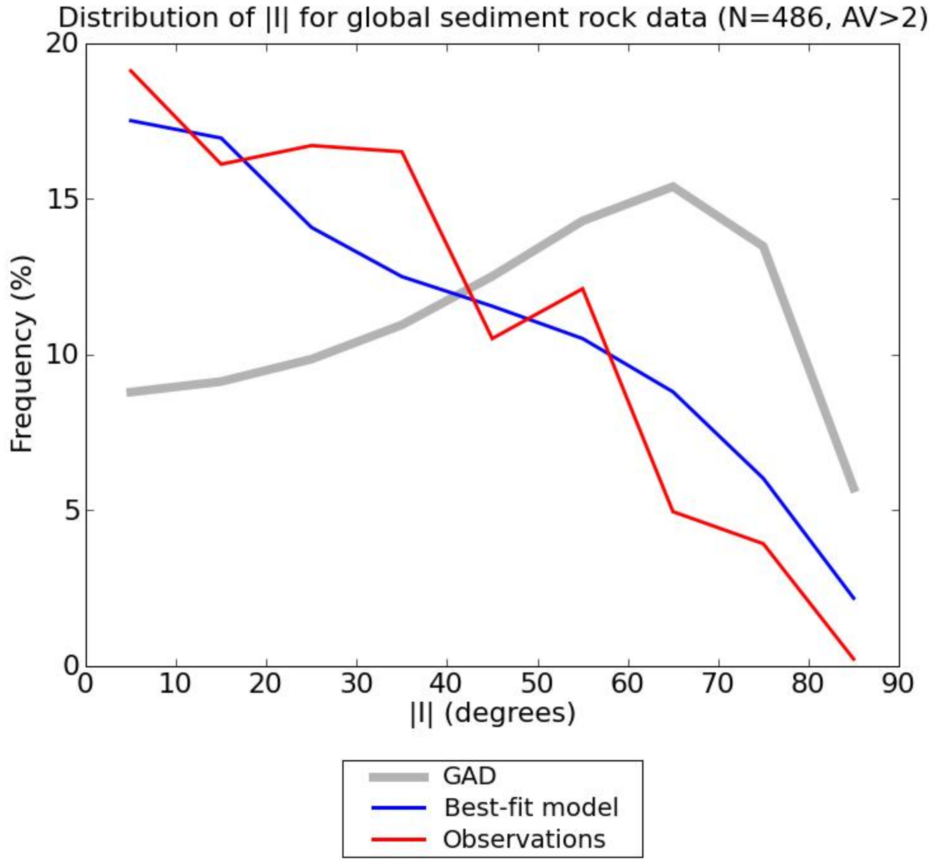

Figure 1. Rocktype does matter. The density function of |I| shows that the global sedimentary data is strongly biased towards low inclinations, with the parameters G2 = 5 % and G3 = 25 % in the best fit. However, this is in accordance to results from previous studies.

4. Results and conclusions

The modelling shows that the long-term geomagnetic field is best described by the GAD combined with a small octupole (10 % of GAD) and no quadrupole at all. Kent and Smethurst (1998), who used very similar methods, had a quadrupole 10 % of GAD and octupole 25 % of GAD. A newer study by Tauxe and Kodama (2009) found respective values: G2 = 19 % of GAD and G3 = 15 % of GAD.

The results presented here are closer to the GAD than previous ones, and this may be due to the strict filtering procedure. In general, igneous rocks observed during over half a century in Europe, North America and Australia have proved to be a crucial source of information of the behavior of the geomagnetic field, but still further data is needed to prove the validity of the model shown. Models are not unique, and actually, a model with G2 = 8 % of GAD is indistinguishable from the GAD. Also a new random-walk test must be done to ensure that all continents have properly sampled the whole of the globe.

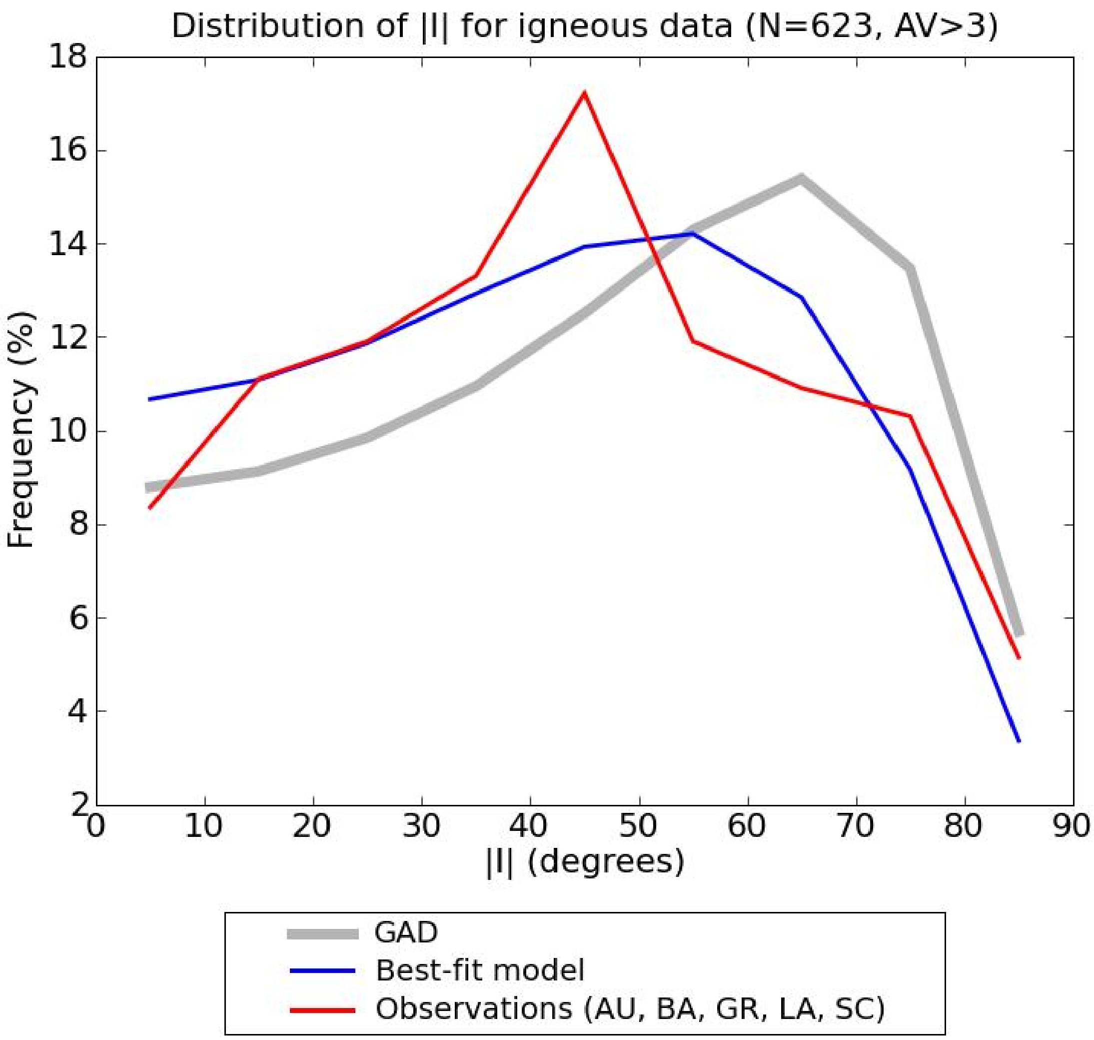

Figure 2. Distribution of the absolute value of inclination of filtered paleomagnetic data. The best-fit model has a zero quadrupole and an octupole of 10 % of GAD. The peak visible on mid-latitudes cannot be explained by any model. It is probably due to sampling mistakes in the original dataset.

interval	G1 (dipole)	G2 (quadrupole)	G3 (octupole)
0° – 10°	8.78	9.54	9.92
10° – 20°	9.12	9.78	10.08
20° – 30°	9.84	10.26	10.46
30° – 40°	10.94	10.96	10.94
40° – 50°	12.50	11.76	11.50
50° – 60°	14.28	12.54	11.98
60° – 70°	15.38	12.88	12.18
70° – 80°	13.46	12.16	11.90
80° – 90°	5.70	10.12	11.04

Table 1. Distributions of |I| for pure dipolar, quadrupolar and octupolar fields. These values are independent of the strength of the field.

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Intra-orogenic Svecofennian magmatism in SW Finland: evidence from LA-ICP-MS zircon dating and geochemistry

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We have studied plutonic rocks that apparently can be categorised as "intra-orogenic". The diorite from Rauma yielded an age of 1865 ± 9 Ma and the diorite from Korpo an age of 1851 ± 5 Ma. The adjacent garnet-bearing Korpo granite was 1852 ± 10 Ma in age. The mafic rocks are high-K and shoshonitic mantle-derived intrusions whereas the granites show typical features of crustal melting. Both the mantle-derived and crust-derived intra-orogenic magmatism are considered to have had a causal effect on the late orogenic thermal evolution in southern Finland.

Keywords: Svecofennian, intra-orogenic magmatism, mantle-crust interaction, crustal melting

1. Introduction

A new tectonic model for the formation the Fennoscandian Shield was provided by Lahtinen et al. (2005). In the model they divided the Svecofennian evolution into the accretionary Fennian orogen at 1.92-1.87 Ga, followed by an extensional period prior to the subsequent Svecobaltic continent-continent collision at 1.84-1.79 Ga. The proposed intervening extensional period is an interesting concept, but challenging to study in the Southern Svecofennian Arc Complex because of the extensive tectono-metamorphic events that subsequently took place during the Svecobaltic stage: i.e. major crustal shortening with upright to overturned folding and granulite facies metamorphism with major crustal melting producing anatectic granites and migmatites (e.g. Ehlers et al., 1993; Korsman et al., 1999). In fact, there are different opinions of the existence, magnitude and time constraints of the proposed extension (c.f. Lahtinen et al., 2005; Cagnard et al., 2007; Nironen and Kurhila, 2008; Skyttä and Mänttäri, 2008; Torvela et al., 2008; Kukkonen and Lauri, 2009).

Suominen (1991) dated garnet- and pyroxene-bearing plutonic rocks and obtained ages between c. 1.87 and 1.84 Ga and considered them to be intraorogenic magmatism. Van Duin (1992) dated charnockites in the Turku granulite area and interpreted them to be mantle-derived and c. 1.85-1.82 Ga in age, thus adding more age data to this age group. Later, when SIMS dating at Nordsim became available, at least the charnockite ages from the Turku area turned out to be the result of mixed zircon populations and they were reinterpreted to be synorogenic 1.87 Ga rocks metamorphosed during the late orogenic high heat flow event at c. 1.82 Ga (Väisänen et al. 2002). At the time, this severely questioned the existence of mantle derived intra-orogenic magmatism. Lately, however, an increasing number of single zircon age data have revealed that magmatism with ages within the intra-orogenic (or extensional) interval is quite common (Ehlers et al., 2004; Kurhila et al., 2005, 2010; Mänttäri et al., 2006, 2007; Pajunen et al., 2007; Nironen and Kurhila, 2008; Skyttä and Mänttäri, 2008). In addition, detrital single zircon data suggest that sedimentary basins opened at that time, indicating an extensional period (Bergman et al., 2008).

In this contribution we present laser-ablation single zircon age data and whole-rock geochemistry from the plutonic mafic-intermediate and felsic rocks in the Korpo and Rauma areas. We suggest that these examples provide evidence for both mantle and crustal magmatism during the intra-orogenic period.

2. Study targets

Our samples come from two study ares in SW Finland: the Rauma area and the Korpo area (Figure 1).

Olkiluoto is the nuclear power plant site, where the bedrock has been mapped and studied in considerable detail, including age data. We collected samples from the tonalitic gneisses that occur as narrow dyke- or sill-like intrusions, tens to hundreds of metres wide. Their ion microprobe (SIMS) concordia ages are 1863 ± 6 Ma, 1856 ± 5 Ma and 1851 ± 5 Ma (Mänttäri et al., 2006, 2007). The *Rauma* samples come from the E-W trending approximately 2-3 x 20 km intrusion consisting of granodiorites and quartz diorites (Suominen and Torssonen, 1993). The samples were collected from the city area where the rocks apparently were more mafic than shown on the geological map. The rock is deformed and displays a steep E-W trending tectonic foliation.

The *Korpo* diorites occur in the northern part of Korpo island in an area of ~ 5 x 10 km. It comprises several E-W trending intrusions which show distinct aeromagnetic anomalies. The diorites are surrounded by pink garnet-bearing porphyric granites which are the main rock types of the area (Suominen, 1987). The diorites are deformed and display a steep E-W trending tectonic foliation. The diorites and granites also form hybrids. Granites dominate at *Nagu*, E of Korpo, but many small mafic “inclusions” are shown on the geological map (Edelman, 1973). One of these shows quite a strong aeromagnetic anomaly, much larger than the exposed mafic body, indicating a larger volume of mafic rocks at depth below the granite.

Figure 1. General geological map southern Finland (modified after Korsman et al., 1997) with study targets indicated. CSAC=Central Svecofennian Arc Complex, K=Korpo, N=Nagu, O=Olkiluoto, PG=Perniö Granite, R=Rauma, SSAC=Southern Svecofennian Arc Complex.

3. Analytical methods

Zircons were separated using standard methods, mounted on an epoxy disc and imaged with BSE at Topanalytica, Turku, Finland. The zircons were analysed using a 193 nm solid state laser ablation system (New Wave) attached to a multi-collector ICP mass-spectrometer (Nu Instruments) at the Finland Isotope Geosciences Laboratory, GTK, Espoo.

All the geochemical analyses were purchased from AcmeLabs Ltd, Canada. The major elements and Cr were analysed by ICP-AES, whereas trace elements were analysed by ICP-MS and F by assay method.

4. Results

The zircons from the Rauma sample were homogeneous without any obvious inner cores or outer rim domains. The measurements yielded a concordia age of 1865 ± 9 Ma. The zircons from the Korpo diorite sample (from Galtby) were clearly of two different morphologies: rounded and prismatic ones. Within errors, however, no age difference was distinguished and all zircons combined from this sample yielded a concordia age of 1851 ± 5 Ma. The Korpo granite sample (from Dimanskär) had a variety of zircon morphologies and many zircons were metamict and altered. Analyses revealed two older, probably inherited, ages between 1.91-1.87 Ga and three analyses were discordant. The remaining concordant analyses yielded a concordia age of 1852 ± 10 Ma. Field relationships that indicated contemporaneous magmatism by mingling and mixing between the mafic and felsic magmas are therefore also confirmed by these age data.

The mafic-intermediate rocks from Korpo, Olkiluoto and Rauma all show similar geochemical characteristics. The SiO_2 contents range between 47 and 60 wt %. Most of the samples show elevated Fe, P, Ti and F contents and plot in the high-K and shoshonitic fields in the K_2O *vs* SiO_2 diagram. LREEs are enriched and trace element diagrams also suggest a shoshonitic affinity of these rocks. The Korpo granite is in many aspects similar to those often described as “late orogenic” granites of southern Finland, i.e. products of crustal melting.

5. Discussion and conclusions

The new single zircon age data in this and in some previous studies indicate that there are no long-lasting gaps in mantle-derived magmatism in the Svecofennian orogen in southern Finland. The synorogenic c. 1.87 Ga stage is followed by c. 1.86−1.85 Ga mafic-intermediate magmatism (Mänttäri et. al., 2006, 2007; this study). Pajunen et al. (2007) presented a SIMS age of 1838 ± 4 Ma for the Jyskelä gabbro. To date, that is the youngest age from mafic magmatism before the so called post-collisional magmatism that started at 1815 ± 2 Ma (Väisänen et al., 2000). Geochemically, the intra-orogenic and post-collisional mafic magmatism are quite similar, apart from the latter showing even more pronounced LILE and LREE enrichment (Rutanen et al., 2010 and references therein).

The 1.85 Ga Korpo diorites are associated with large volume crustal-derived granites. Within the same zone to the east similar granites occur as the Perniö granite. The known single zircon ages of the Perniö granite are 1835 ± 12 Ma (Kurhila et al, 2005) and 1853 ± 18 Ma (Kurhila et al., 2010) which, within errors, overlap with the Korpo granite age reported in this study. This result supports the idea that the Korpo granite is a western extension of the Perniö granite.

Our hypothesis is that mantle-derived magmatism is the main driver of the high grade “late orogenic” 1.85-1.81 Ga metamorphism in southern Finland, whereas crustal stacking, radioactive decay and conductive heating, as modelled by Kukkonen and Lauri (2009), may have set the stage for further wide-spread crustal melting. Timing of the intra-orogenic magmatism coincides with the extensional period in the model of Lahtinen et al. (2005).

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