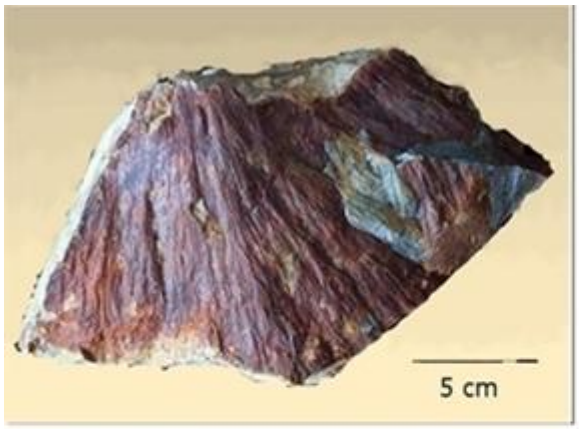

Supercontinent Symposium 2012

Pre-Symposium Excursion Guidebook

Keuruu – Jormua – Taivalkoski – Siurua – Tervola

Sept. 21-25, 2012

Edited by

Lauri J. Pesonen, Satu Mertanen, Pathamawan Sangchan and Ella Koljonen

GTK

UNIVERSITY OF HELSINKI

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Cover images:
Up left: Shatter cone of Keurusselkä (Stop 1) by L. J. Pesonen;
Below left: mafic dyke of Taivalkoski (Stop 8) by M. Melamies;
Below right: Rantamaa stromatolites (Stop 12) by V. Perttunen

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Geological Survey of Finland
Espoo 2012

Preface

This excursion guidebook provides a background for the seven geological landmark sites in Finland which will be visited during the pre-symposium (Supercontinents 2012) excursion. During the four excursion days (Sept. 22-25, 2012) twelve stops constitute the excursion programme. Descriptions of the stops are preceded by a general introduction to the topics followed by geology and geophysics of the stops in the framework of Precambrian supercontinents - the topic of the symposium. The guidebook is organized in day-by-day principle. In addition to geology, short descriptions of Finland and in particular the city of Helsinki and the other cities, towns and villages we are passing through, is provided. For those who are not familiar with the geology of the Baltic or Fennoscandian Shield a short description is added to provide a framework for the tour.

This guidebook is edited by Lauri J. Pesonen, Satu Mertanen, Pathamawan Sangchan and Ella Koljonen. Each chapter has their own editors including, in addition to the afore mentioned, Robert Klein and Selen Raiskila (Keurusselkä), Asko Kontinen (Jormua), Jouni Vuollo and Johanna Salminen (Taivalkoski), Hannu Huhma and Katja Lalli (Siurua) and Juha Karhu, Nina Hendriksson and Vesa Perttunen (Rantamaa).

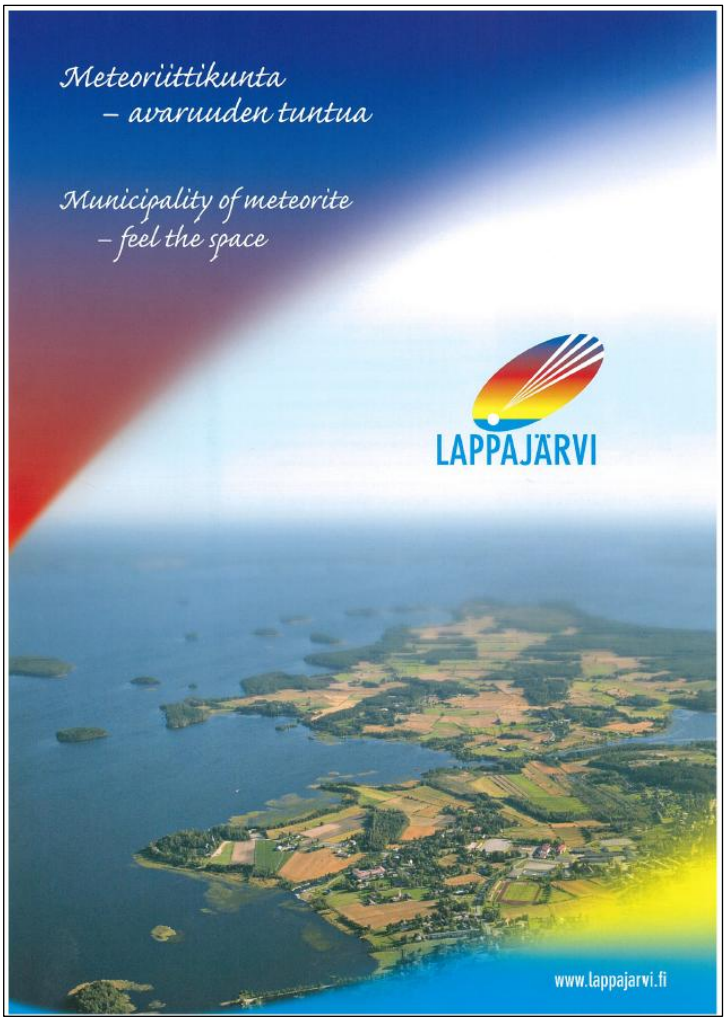
Several cities, towns, and communes have supported us, notably, Helsinki, Keuruu, Lappajärvi, Rovaniemi and Tervola. We thank all of them. We also thank Kari Jääskeläinen, Tuire Laine, Toni Veikkolainen, Harri Matikainen, Markku Montonen and Tommi Vuorinen for their help in preparing this guidebook and the excursion. The publishing of the guidebook was made possible with the help of the Geological Survey of Finland.

Welcome to the Excursion of the Supercontinent Symposium 2012!

Kumpula, Sept. 19, 2012

Lauri J. Pesonen
Solid Earth Geophysics Laboratory
University of Helsinki

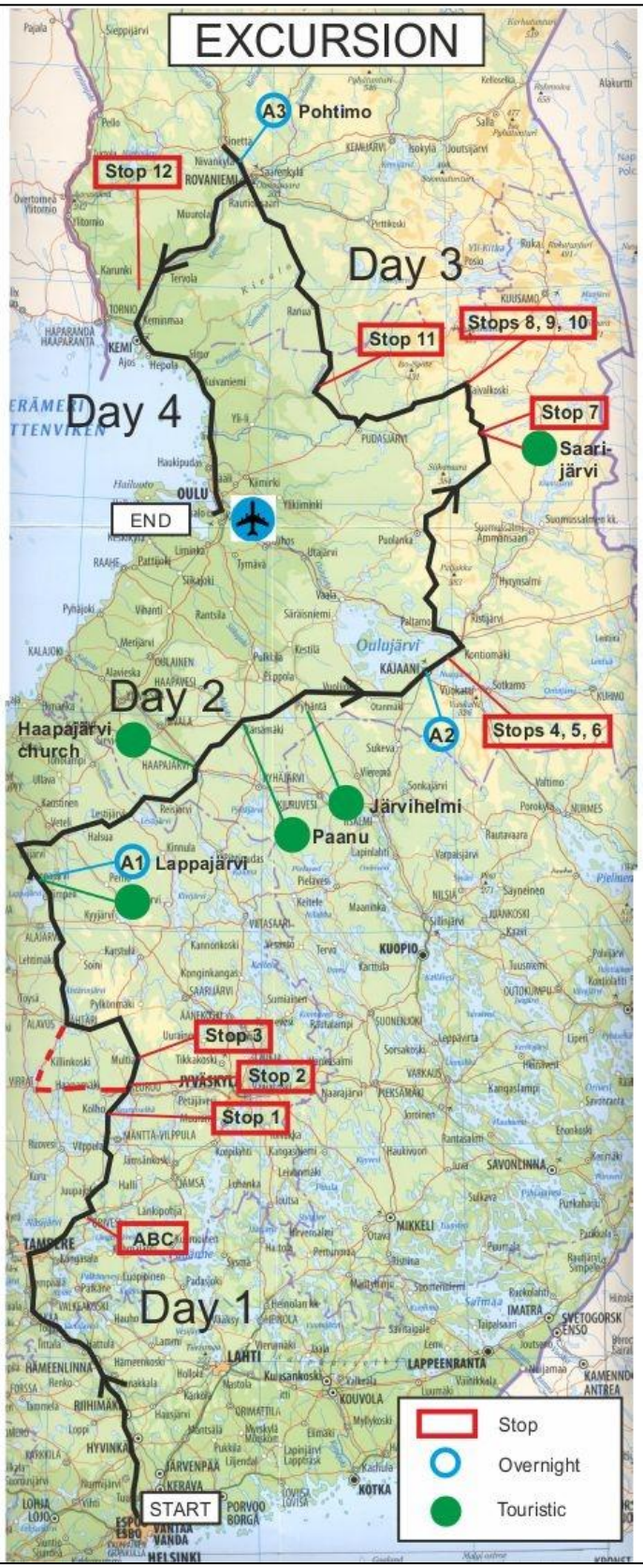

Keuruu
- the town by the lake!
The town centre of Keuruu has been built by the beautiful lake Keurusselkä.
Enjoying life on the lakes, railroad culture, the five churches and the historically valuable area of old Keuruu bring the atmosphere of past times to present time.
Have a wonderful time as our guest!
KEURUU.fi


Meteoriittikunta
– avaruuden tuntua
Municipality of meteorite
– feel the space
LAPPAJÄRVI
www.lappajarvi.fi

Pre-Symposium Excursion Guidebook

Contents

Excursion route



Finland in focus

Finland in focus

The Republic of Finland (i.e. “Suomi”) is an independent (since 1917) country in northern Europe at the shores of the Baltic Sea. Finland is part of Nordic Countries and European Union. The country is surrounded by neighbors such as Russia in the east, Norway in the north, Sweden in the west and across the Gulf of Finland in the south is Estonia. Ahvenanmaa, an archipelago in the Baltic Sea, is part of Finland but has autonomy and is demilitarized.

Figure 1. Geographical location of Finland within the European Union. Finland is outlined by black contour.

Climatically, Finland is situated in the northern part of the *temperate* zone of the world. It is also part of the Boreal forest zone or Taiga.

Finland has a relatively low population density (18 inhabitants per km^2) and most of them live in the southern and mid parts of the country. There are two official languages, Finnish and Swedish. Some 5.4% of the 5.4 million people living in Finland speak Swedish as their mother tongue and generally they live in the south or southwestern areas of Finland.

Finland in figures

Population

- 5.4 million with annual growth rate 0.5%.
- Life expectancy: men 76.5, women 83.2 yrs.
- Average household size: 2.1 persons. 55% of the households live in single-family houses. 44% in apartment blocks. 84.4% are urban-dwellers, with 1 million in the Helsinki area, which includes Espoo and Vantaa.
- Other major cities in Finland include Tampere, Turku, Oulu and Jyväskylä.
- Religion: 78% are Lutheran; 1% Orthodox.
- Education: 78% of the population aged 25 to 64 have completed upper secondary or tertiary education and 35% (the highest percentage in the EU countries) have university or other tertiary qualifications.
- Wired: 82% of Finnish households own a personal computer and 70% broadband; 92% own a digital television and 98% of households have cell phones.

Government

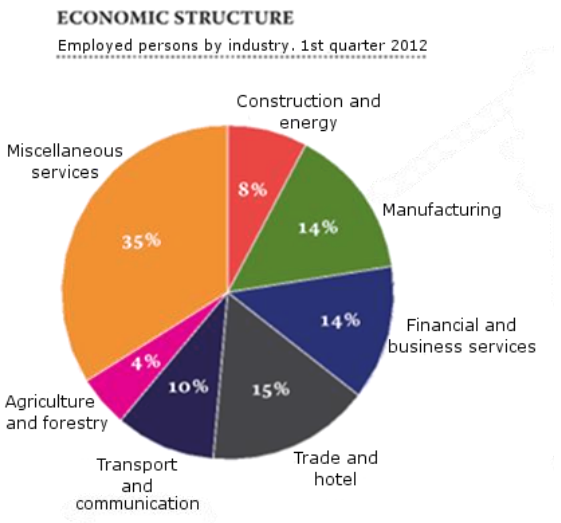
- Sovereign parliamentary republic since 1917. From 1809-1917, autonomous Grand Duchy within the Russian Empire; before that part of the Kingdom of Sweden for centuries.
- The president is elected every six years. The new president of Finland Sauli Niinistö took office in March 2012.
- The 200 members of Parliament are elected for four-year terms. The most recent general election produced the following result: National Coalition Party 44; Social Democratic Party 42; The Finns 39; Finnish Centre 35 seats; Left Alliance 14; Greens 10; Swedish People's Party 6; Åland 1.
- Finland has been a member of the European Union since January 1995.

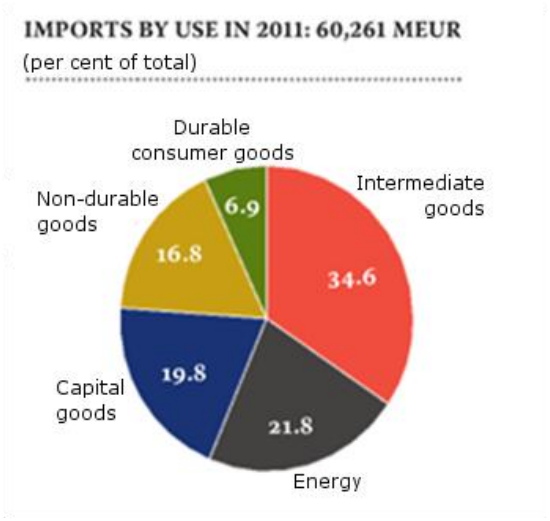
Working life and Economy

- 85% of women aged 25-54 are employed outside the home. Average monthly earnings (2011): men 3487 €, women 2841 €. Current unemployment rate 8.4%.
- GDP 2011; 192 billion euros.
- Annual inflation rate (2012): 3.1%.
- Currency: Euro €.

Area

- 390 920 km^2 of which 9% is fresh water; land area is 303 909 km^2. There are 188 000 lakes, 6% of the land is under cultivation with barley and oats the main crops. Forests (mainly pine and spruce) cover 68% of the country.





I Short overview of the geology of the Fennoscandian Shield

Introduction

The Fennoscandian (or Baltic) Shield represents the largest outcropping domain of Precambrian bedrock in Europe, covering more than a million km^2 throughout Norway, Sweden, Finland, Bornholm island of Denmark and northwestern Russia (Fig.2a).

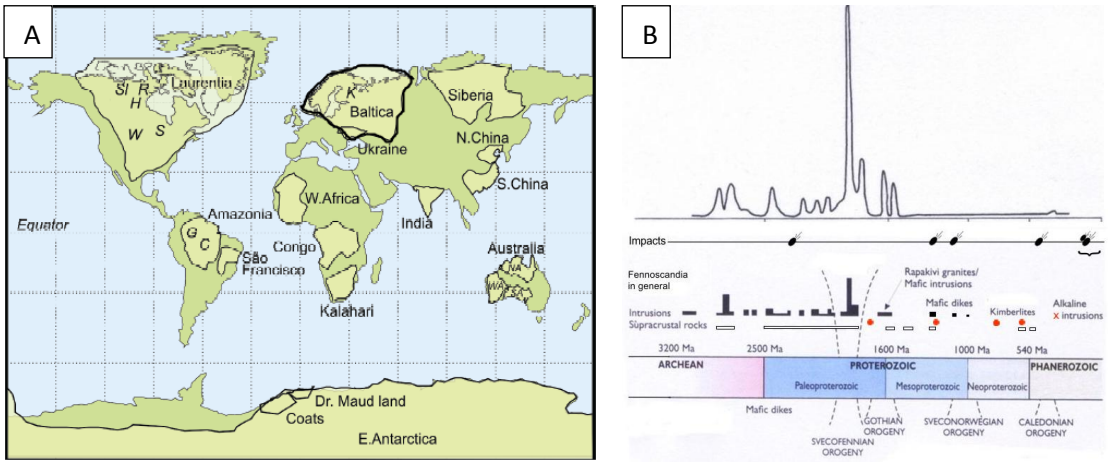


Figure 2. (A) Precambrian continental cratons of the world are outlined in yellow shading (Mertanen and Pesonen, 2012). Fennoscandia or Baltica is shown with a black outline. (B) Major geological events in the Fennoscandian Shield including magmatic dyking events, ages of kimberlite clan rocks, meteorite impacts and deposition of sedimentary basins. Modified after Kohonen and Rämö (2005).

The bedrock of Finland belongs to the East European Craton (EEC) of northern and eastern Europe and northwestern Russia. Precambrian crystalline rocks crop out only in the northern and southwestern parts of the craton, in the Fennoscandian and Ukrainian shields, respectively; elsewhere they are covered by platform sediments. In Sweden and Norway, the Fennoscandian Shield is delimited by the Caledonides. In Estonia in the south and Russia in the south-east, the Precambrian bedrock plunges at a shallow angle under Phanerozoic sedimentary rocks.

The most important events during the evolution of the Finnish bedrock occurred at 2800-2700 Ma and 1900-1800 Ma (Fig. 2b). In those times, continental crust was segregated from the Earth's mantle in two major (probably multiphase) orogenies. The resultant Archean and Paleoproterozoic crust of Finland is divided into 25 areas with characteristic lithologic units (Fig. 3). This Chapter gives an overview of Finland's bedrock and its evolution from the Mesoarchean to the present time.

Finland forms about one third of the *Fennoscandian Shield* which crops out among younger sedimentary rocks and the Caledonian mountain belt. By age, it can be divided into four main areas: the Archean, the Svecofennian and the Sveconorwegian domains, and the Transscandinavian igneous belt lying between the latter two (Fig. 3). The northern and eastern parts of Finland belong to the >2.5 Ga *Archean* block, divided usually into the Kola

and Karelia cratons, while the central and southern parts comprise the Paleoproterozoic *Svecofennian* rocks, ca. 1.93-1.80 Ga in age. Only a small part of the Finnish bedrock is younger than 1.8 Ga: the most significant of the younger formations are the 1.65-1.54 Ga *rapakivi granites* and associate gabbros, anorthosites and dykes. After the intrusion of the rapakivi batholiths no major magmatism, except a few dyking events (1.26 Ga, 1.12 Ga and 1.05 Ga), has occurred in Finland. Considerable graben formation took place during the Mesoproterozoic and at least southern Finland was covered by Paleoproterozoic-Mesozoic sediments.

Figure 3 gives an outline of the various “block” names used for the EEC and Fennoscandian Shield. The dotted line shows the excursion route, which starts from Paleoproterozoic rocks (Days 1-2), traverses shortly to the Archean (Karelian craton) rocks in North Finland (Day 3) and ends again in Paleoproterozoic area (Day 4).

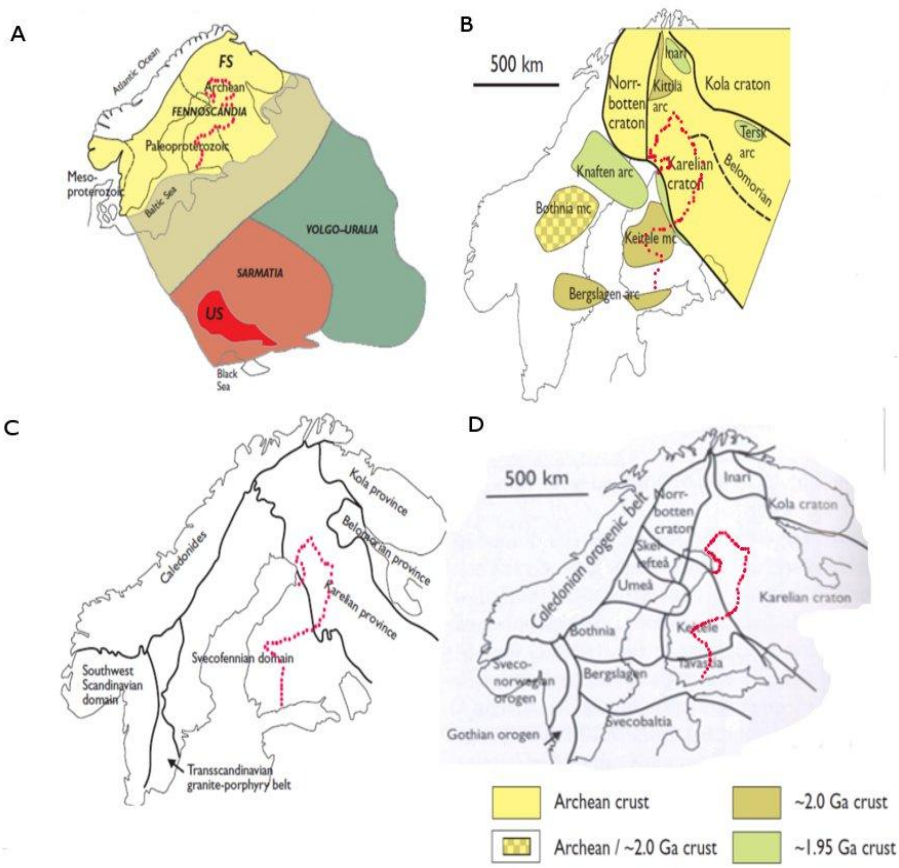


Figure 3. A) Crustal segments of the East European Craton. FS – Fennoscandian Shield; US – Ukrainian Shield. B) Pre-1.92 Ga crustal components and major terranes/units in the Fennoscandian Shield. Exposed and hidden pre-1.92 Ga components are outlined with a broken line; juvenile crust, crust without known mantle separation age or crust dominated by accreted sediments is shown in white. C) Major tectonic units of the Fennoscandian Shield. D) Major geological units with Paleo-proterozoic and younger boundaries. The dotted line denoted the excursion route. Modified after Gorbatschev and Bogdanova (1993) and Lahtinen et al. (2005).

Geophysics along the excursion route

The excursion traverses several remarkable geophysical anomalies in Finland. Figure 4 shows the (a) aeromagnetic, (b) Bouguer gravity, (c) earthquake activity with land-uplift and (d) Moho-depth maps of Finland.

Diabase dyke swarms (Fig. 5), kimberlites and alkaline dykes provide useful information concerning crustal tectonics and supercontinent reconstructions, since dykes of different ages occur within sedimentary basins, greenstone belts, and gneissose to granitic shield areas.

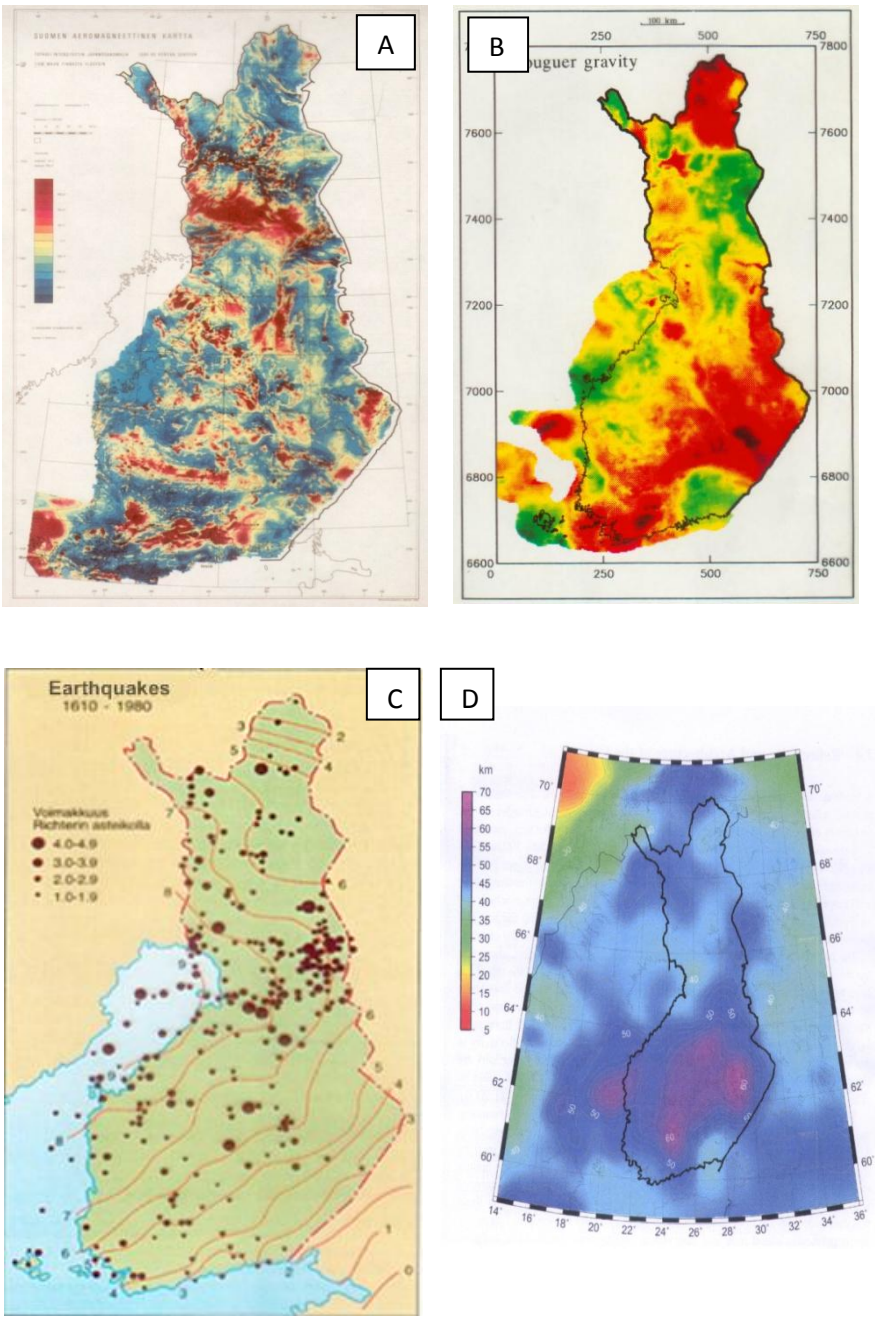


Figure 4. Major Geophysical anomalies in Finland along the excursion route. A) Aeromagnetic, courtesy by J. Korhonen, GTK, B) Bouguer gravity, courtesy by S. Elo, GTK, C) Earthquake activity with land uplift contours and D) Moho depth in Finland (Heikkinen, 2012).

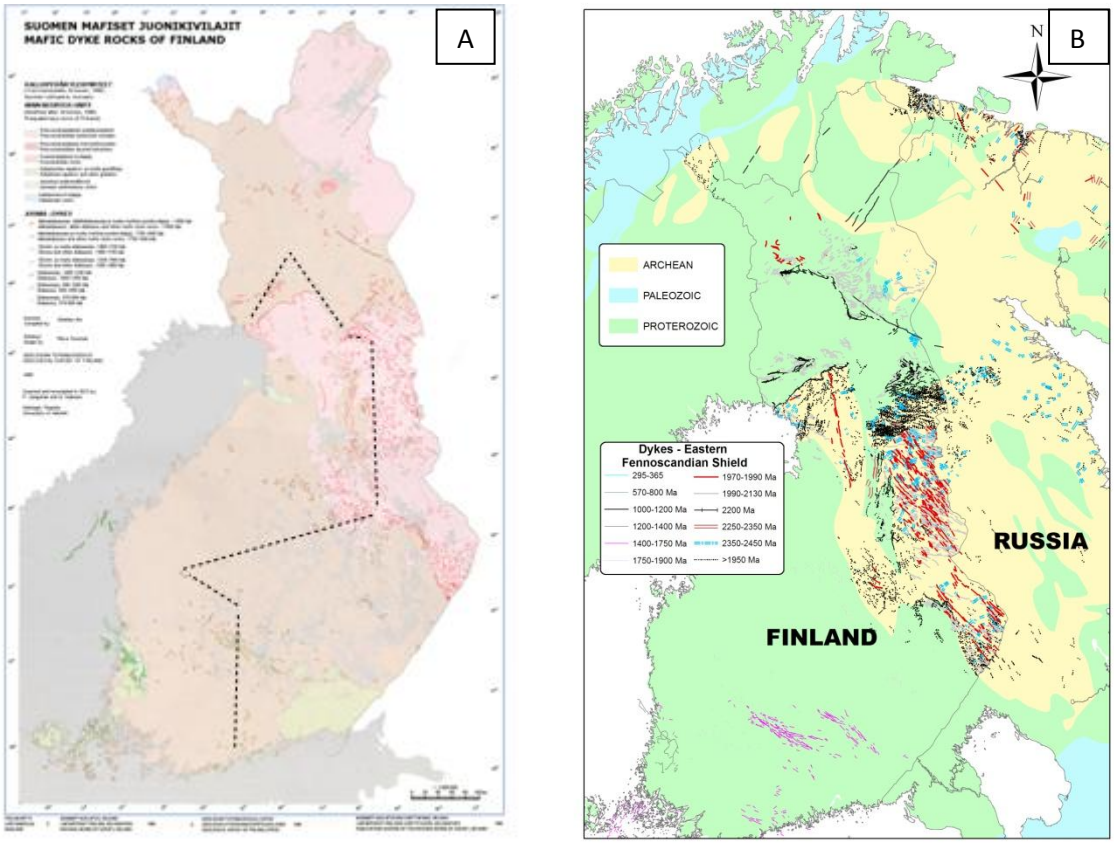


Figure 5. A) Simplified map of mafic dyke swarms of Finland after Aro and Laitakari (1987). The excursion route has been schematically drawn. B) Updated Fennoscandian dyke map by J. Vuollo (GTK).

II Impact cratering: a short overview with Finland as a focus

Introduction

Impact cratering is one of the most important surface-modifying processes on the Earth. Especially large meteorite impact events are the cause of major crustal deformation with important economic mineral and hydrocarbon deposits. Rapidly radiating shock pressures in target rocks may produce e.g. rock melting (≥ 60 GPa) with post-shock temperatures up to 2000°C, diaplectic glass phases (30-45 GPa), high-pressure minerals (12-30 GPa) and planar deformation features (PDFs) in quartz (10-25 GPa) (*French and Koeberl, 2010*). Shatter cones, which are striated conical structures in rock formations, occur usually at lower pressures (1-5 GPa; up to 45 GPa) and are coupled with extensive fracturing of the target rock notably in the central uplift area. The impact processes also form breccias, suevites, pseudotachylite veins and ejecta deposits, which are very abundant in larger impact craters. Most of the subcrater breccias are formed late in the cratering stage at relatively low shock pressures (< 1 GPa) lacking distinctive features of shock metamorphism.

The Earth has been covered by a comparable number of impact scars, but due to active geological processes, such as plate tectonics, weathering and erosion etc, the number of preserved and recognized impact craters on the Earth is limited. Nevertheless, the study of the impact structures on the Earth provides a method to understand the evolution of our solar system planetary geology!

Approximately 180 impact structures have been presently confirmed on the Earth (*Earth Impact Database 2012*). Fennoscandia is statistically one of the most densely crater-populated parts of the Earth (*Dypvik et al. 2008*) as most of the craters are exposed. Altogether, 31 impact structures, with ages from ca. 2400 Ma to approximately 0.004 Ma, have been discovered in Fennoscandia. The high density of identified craters is due to the considerable research activity, coupled with a deterministic view of what we look for. In spite of these results, many Nordic structures are poorly understood due to the lack of 3D-geophysical interpretations, isotope or other datings and better knowledge of the amount of erosion and subsequent tectonic modifications (Dypvik *et al*. 2008). Studies of paleomagnetic properties of several Fennoscandian impact structures have been performed successfully, e.g., Lappajärvi, Karikkoselkä, Iso-Naakkima, Suvasvesi doublet and Keurusselkä (Raiskila et al. 2012) in Finland, Kärdla in Estonia (*Plado et al. 1996*), Jänisjärvi in Russian Karelia (*Salminen et al. 2006*), and Siljan (*Elming and Bylund. 1991*) in Sweden.

Eleven proven impact structures have so far been found in Finland (Fig 6). These are (with discovery year in parenthesis): Lappajärvi (1967), Sääksjärvi (1969), Söderfjärden (1978), Iso-Naakkima (1993), Lumparn (1992), Suvasvesi North (1993), Karikkoselkä (1996),

Saarijärvi (1997), Paasselkä (1999), Suvasvesi South (2001) and Keurusselkä (2003). The ages vary from ca. 1200 Ma (Iso-Naakkima) to about 73 Ma (Lappajärvi), but the majority is poorly dated. The present diameters vary from ~23 km in Lappajärvi to 1.4 km in Karikkoselkä.

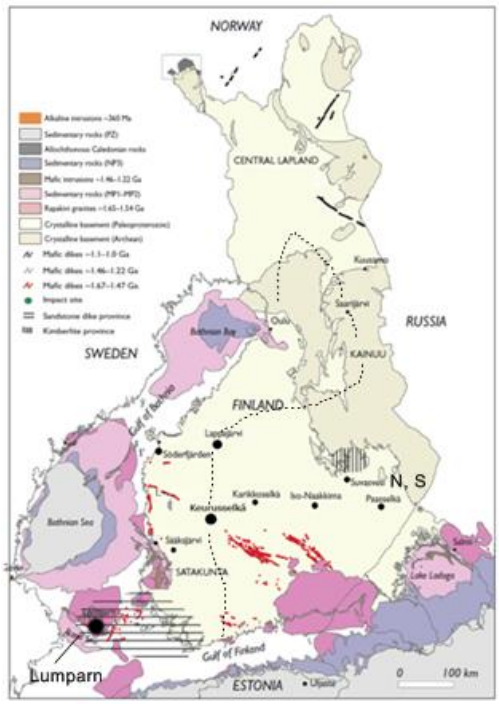


Figure 6. Simplified geological map of Finland (Koistinen *et al.*, 2001). The eleven proven meteorite impacts are shown as closed spheres the size of which are relative to the diameter of the structure. This excursion will visit Keurusselkä, Lappajärvi and Saarijärvi structures.

These are minimum estimates since the structures are moderately to strongly eroded. In this excursion guidebook we discuss shortly of the three impact structures relevant to our excursion: Keurusselkä, Lappajärvi and Saarijärvi.

The *Keurusselkä* structure (<1800 Ma) is located 120 km southeast from Lappajärvi (Fig. 6). It is most likely a deeply eroded complex or multiring impact structure and the preliminary estimate of the original diameter, based on shatter cone findings, points to >30 km. Keurusselkä will be discussed in more details in next chapter since it is our first excursion stop (Stop 1).

Lappajärvi is the largest impact structure (D 23 km) in Finland and covers the present lake Lappajärvi. Although the lake itself is elliptical in shape, the subsurface structure is nearly circular as seen in aeromagnetic (Fig. 7a) and gravity maps (Ables et al. 2002). The Ar-Ar and U-Pb dating resulted in ages of 77.3±4 Ma to 71 Ma (Dypvik et al., 2008). Dark and dense impact melt, so-called kärnäite, occurs in Lappajärvi's central island and nearby islands. The kärnäite layer is about 145 m thick and displays enrichments of Ni and Ir, most likely of meteoritic origin. Below the kärnäite, the few meters thick layer of suevite rests on a

clastic impact breccia. PDFs in quartz are common. Lappajärvi hosts a “meteorite museum” inside the Kivitippu (“Rock fall”) Spa Hotel (our first overnight place).

The *Saarijärvi* impact structure (age < 500 Ma, D~1.5 km) is a drop-shaped lake located in Archean basement in Kainuu, north-central Finland. The drill-core samples revealed PDFs in quartz from the impact breccias. Saarijärvi will be shortly visited with an “en-route” sightseeing stop (Stop 7) on Monday, Sept. 23.

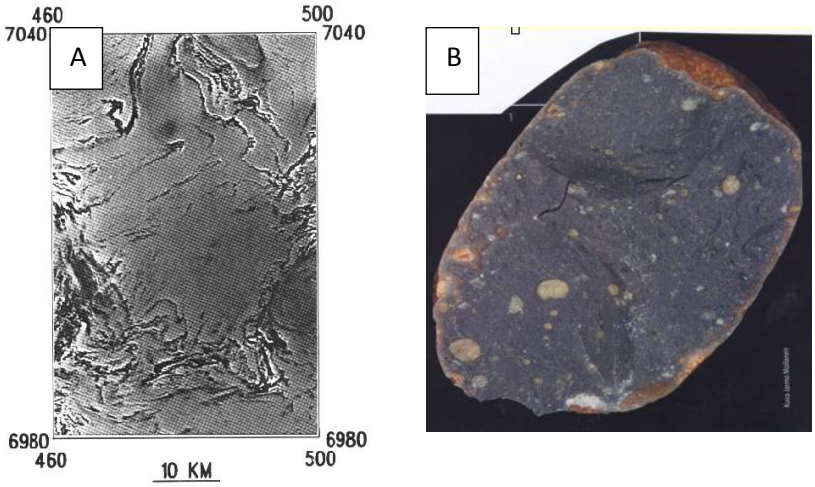


Fig. 7. A) Aeromagnetic map of Lappajärvi impact structure (GTK). B) Melt boulder from Lappajärvi, central-west Finland. (Photo: Jarmo Moilanen)

Impact structures form an important factor in the research of shield areas, since these rocks have recorded “events” for billions of years. It is noteworthy that they “constrain” also supercontinent reconstruction studies since, unlike “episodic” igneous activity, the impacts can occur at any time (Fig. 1b). The global paleomagnetic database (Pesonen et al. 2012, Abstract Volume of Symposium) contains several reliable (dated) poles of impact rocks (see e.g., the case of Keurusselkä), which have been used in making supercontinent reconstruction using paleomagnetic techniques (see Part IV).

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III Day “0”, Friday, Sept. 21, 2012

City of Helsinki

University of Helsinki

The excursion starts from Helsinki and ends there.

Helsinki: the “Daughter of the Baltic Sea”

Helsinki has been the capital of Finland for the past 200 years. The new capital was appointed in 1812 and now it is celebrated with many types of events; street art, tours into the history of the city, lectures etc., not forgetting the Supercontinent Symposium 2012! This year Helsinki (together with cities of Espoo, Kauniainen, and Lahti) is also the World Design Capital.

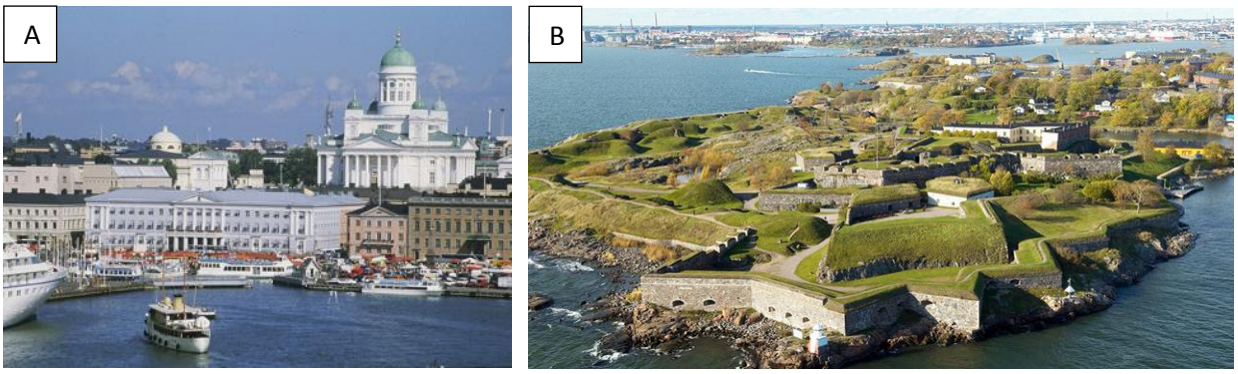


Figure 8. A) Helsinki from the sea with Tuomiokirkko (Lutheran Cathedral). B) Suomenlinna Fortress (1748-1973)

Attractions

There are numerous touristic attractions that are worth of visiting, such as the Tuomiokirkko (Fig. 8a) and the Suomenlinna Sea Fortress, a World Heritage Site located off the coast of Helsinki (Fig. 8b). It is also a garrison town and a base for archipelago navy.

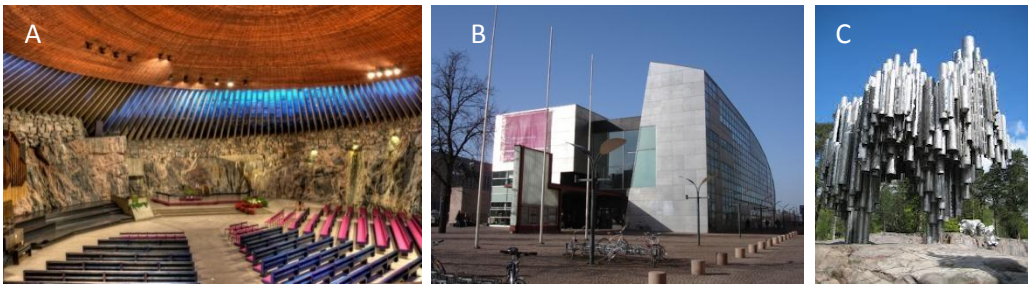


Figure 9. A) Church in a “Svecofennian” rock – Temppeliaukion kirkko. B) Kiasma contemporary art museum. C) Sibelus monument.

The other attractions include the Temppelinaukio church within Sverofennian rocks (Fig. 9A), the Orthodox Uspenski Cathedral, the Kiasma contemporary art museum (Fig. 9B), Sibelius (famous Finnish composer Jean Sibelius 1865-1957) momument (Fig. 9C) and Ateneum art museum. If you like there are also special types of museums such as the open air museum of Seurasaari where one can see beautiful nature and get familiar with the history of Helsinki.

Several walking tours and city bus tours are also provided in Helsinki and neighborhoods: with the “hop-on&hop-off”- tourist bus you can see all the main attractions Helsinki has to offer. However, if you are on a budget and would like to see as much as possible you can, take the tram 3B or 3T that circles the central Helsinki, and find your own way around. Moreover, on Thursday evening, Sept. 27, you have unigue chance to see some Helsinki touristic attraction from the “Streecar named Larambia, “travelling from Kumpula to University of Helsinki Main Building.

University of Helsinki

Figure 10. (A) Main building of the University of Helsinki in Downtown Campus. Here, in the Press Hall we will have the Rector’s buffet dinner on Thursday, Sept. 27. (B) Physicum building in Kumpula Campus of UH hosting the Symposium in (C) lecture hall D101.

The University of Helsinki (since 1829) was founded in the city of Turku in 1640 as The Royal Academy of Turku when Finland was part of the Swedish Empire. It is the oldest and largest university in Finland with the widest range of disciplines available. Around 35,000 students are currently enrolled in various degree programs, spread across several faculties and 11 research institutes.

As of August 1, 2005, the University complies with the standards of the Europe-wide Bologna Process and offers Bachelor, Master, Licenciate, and Doctoral degrees. Admission to degree programmes is usually determined by entrance examinations, in the case of bachelor degrees, and by prior degree results, in the case of master and postgraduate degrees. The university is bilingual, with teaching provided both in Finnish and Swedish. Teaching in English is extensive throughout the university, making it a *de facto* third language of instruction.

Remaining true to its traditionally strong Humboldtian ethos, the University of Helsinki places heavy emphasis on high-quality teaching and research of a top international standard. It is a member of various prominent international university networks, such as Europaeum, UNICA, the Utrecht Network, and is a founding member of the League of European Research Universities.

Our Symposium will take place in high-standard lecture hall D101 in the entrance level of Physicum building (Fig. 10B, C). Although not part of the Excursion programme we suggest you to visit three highlights of this building as outlined in Fig 11.

Figure 11. Highlights of Physicum building. (A) Rock momument in front of the building, (B) rotating (granite) sphere and (C) geophysical data and co-ordinates in a sign plate on the wall.

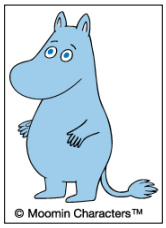
Geology of Helsinki

Helsinki area belongs to the Paleoproterozoic Svecofennian Uusimaa schist belt where schists and gneisses are accompanied by younger granites. The schists and gneisses are partly sedimentary and partly volcanic in origin. The rocks are dated at ~1.9 - 1.8 Ga. The main igneous rocks are granites: mafic intrusives such as diorites or gabbros, are scarce. Several fracture zones cut through the rocks of Helsinki and the most significant ones are aligned to north-northeast such as the famous Kluuvi-fracture in central Helsinki.

Famous people of Helsinki

Artturi Ilmari (A. I.) Virtanen, a chemistry nobelist (1945), lived and worked 1895-1973 in Helsinki.

Jörn Donner, a writer and movie director, born in 1933.



Tove Jansson (1914-2001), a writer and artist who created the Moomin characters, lived in Helsinki.

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City of Helsinki: www.visithelsinki.fi/juuri-nyt/tapahtumia/helsinki-200-vuotta-paakaupunkina-2012

SUPERCONTINENT EXCURSION "DAY BY DAY"

short outline of the Excursion

Excursion route

Day "0" (Sept. 21, 2012)

Short outline of the Excursion will be given in the evening at Symposium Hotel (Park Hotel Käpylä).

Day 1 (Sept. 22, 2012): Keuruu and Keurusselkä

PDF's

Extraterrestrial exhibition at Spa Hotel Kivitippu

The first day consists of three stops at Keuruu to see:
1) The Keurusselkä impact structure, discovered in 2004 by Hietala and Moilanen as based on remarkable *in situ* shatter cone occurences with striking PDF's. The crater is deeply eroded and paleomagnetic and Ar-Ar datings suggest an age of ~1.1 Ga.
2) Keuruu diabase dyke swarm. There are some 20 dykes with a roughly NW-SW strike and vertical dips. U-Pb (zircon) dating yields an age of ~1.8 Ga confirmed by paleomagnetic data.
The night is spent in Lappajärvi, which is a 73 Ma old impact crater itself. All impact lithologies such as kärnaites (impact melt), suevites, breccias and target rocks can be seen in the "meteorite exhibition" at the Kivitippu Spa Hotel ("Rock Fall") where we overnight.

Day 2 (Sept. 23, 2012): Jormua

Pillow lavas of JOC

Sheeted dykes of JOC

In Jormua we will see a Paleoproterozoic ophiolite complex (JOC) on the southern edge of Lake Oulujärvi. It is the most complete ophiolite outcrop in Finland displaying all the elements of an ophiolite: mantle tectonites, gabbros, sheeted dikes, pillow lavas and hyaloclastites. The night is spent at the famous Hotel Kajanus in the historical "castle" city of Kajaani.

Day 3 (Sept. 24, 2012): Saarijärvi, Taivalkoski and Siurua

Day 3 will be very busy as three sites are visited:
1) The Saarijärvi impact structure will be the first stop en route. The structure has a diameter of 4 km, and an age of ~ <500 Ma. The impact origin is verified by PDF's in breccias of a drillcore as well as macroscopic shatter cone occurences and circular geophysical anomalies.
2) Taivalkoski, displaying a Paleoproterozoic (~2.4 Ga) mafic dyke swarm of eastern Finland. The Taivalkoski dykes are the target of an ongoing paleomagnetic study bearing importance in testing the so called Superia-Supercraton model. One of the Taivalkoski sites forms a cornerstone of the house of famous Finnish writer Kalle Päätalo.
3) Siurua, showing a trondhjemite gneiss exposure, is one of the oldest rocks in Europe (3.5 Ga). The chemical composition is close to the average Archean TTG gneisses. The REE pattern of Siurua and felsic granulite of Pudasjärvi Granulite Belt are similar with negative Eu anomaly; however, the REE concentrations of the Siurua trondhjemite are about an order of magnitude higher.

The overnight takes place in the Bear's Lodge, a Lapland style wilderness resort.

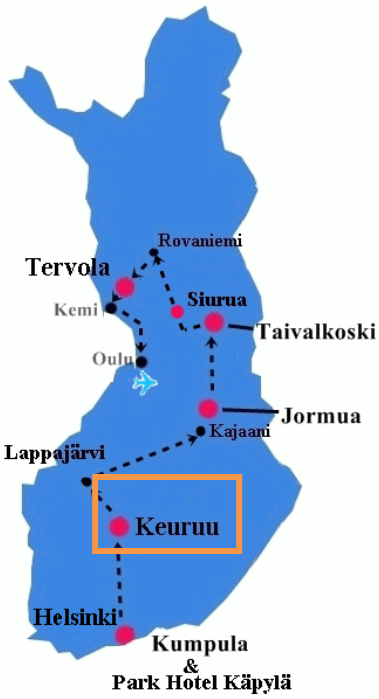
Day 4 (Sept. 25, 2012, morning): Tervola

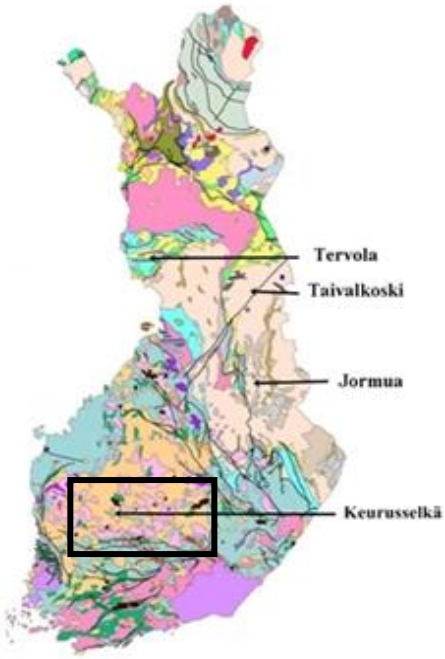
The Paleoproterozoic (~2.2-2.16 Ga) Rantamaa quarry in Tervola, part of the Peräpohja schist area, provides an excellent site for stromatolites. The Paleoproterozoic carbonates are strongly enriched by the ^{13}C isotope and represent a positive anomaly that is global in carbonates of this era. The sedimentary structures of these carbonates are very well preserved due to low metamorphism. The Peräpohja area has abundant reserves of stromatolite structures that have been formed in shallow seas by microbial agents.

Day 4 (Sept. 25, 2012, afternoon): Flying back from Oulu to Helsinki

... Just in time to participate in the ice-breaker of the Supercontinent Symposium 2012 taking place at the Park Hotel Käpylä at 19.30!

This poster is available at http://supercontinent2012.helsinki.fi/





IV Day 1 (Sept. 22, 2012)

IV.1. Keurusselkä impact structure

Lauri J. Pesonen, Selen Raiskila,
Satu Mertanen and Robert Klein

IV.1. Keurusselkä impact structure

Lauri J. Pesonen, Selen Raiskila, Satu Mertanen and Robert Klein

After driving from Helsinki, with a short coffee break in Orijärvi, we will arrive in Keurusselkä. The bus will be parked at the Valkeeniemi farmhouse yard (Fig. 12) from where we have about 25 minutes walk to Jylhänniemi, SW Lake Keururusselkä. At the shore of the lake we'll see well preserved shatter cones at the outcrops (Stop 1, Fig. 14, 16) and on boulders. After walking back, we'll have lunch, kindly served by City of Keuruu and, the host and hostess (the Valkeeniemi family) of the house. Then we'll continue our trip to the Keuruu dykes (IV.2).

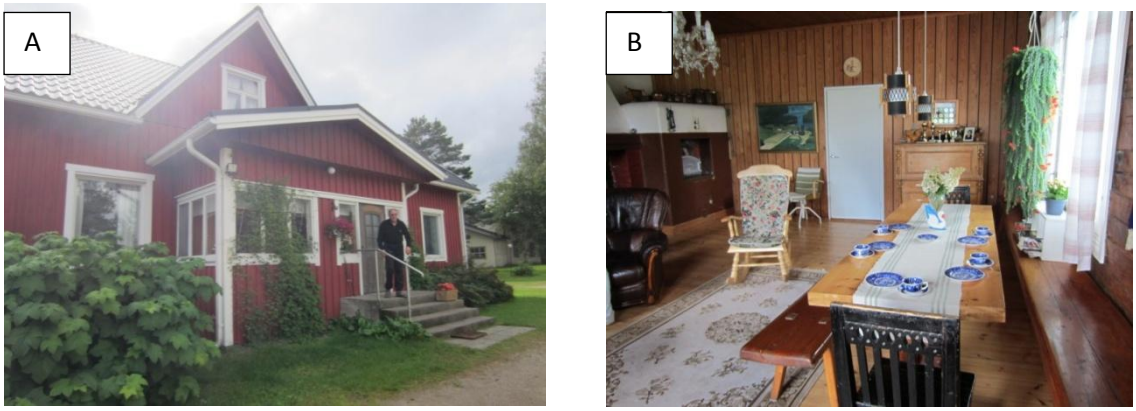


Figure 12. (A) The Valkeeniemi farmhouse at lake shore of Keurusselkä. (B) Inside of the farmhouse. Photos: Satu Mertanen, 2012.

The Keurusselkä impact structure

The Keurusselkä impact structure, found in 2004 (*Hietala and Moilanen, 2004*), is among the latest (11th) in Fennoscandia (Fig. 13, *Dypvik et al., 2008*). The impact origin for Keurusselkä was confirmed with a petrographic investigation of shatter cones (*Ferrière et al., 2010*). The shock metamorphic planar deformation features (PDFs) found in quartz indicated shock pressures from 2 GPa to 20 GPa (Fig. 26). The structure is located about 30 km west from the small Karikkoselkä impact structure (~230 Ma) and 120 km SE from the ~73 Ma Lappajärvi impact structure (see p.80) within the Central Finland Granitoid Complex (CFGC). The CFGC was formed 1890–1860 Ma ago during the peak phase of the orogeny, where the growing Svecofennian island arc system accreted against the Archaean continent (*Nironen, 2005*). The granitoids are predominant rocks in the CFGC together with schists and gneisses which are mainly exposed close to the shoreline of Lake Keurusselkä.

According to Kähkönen (2005), the Svecofennian granitic bedrock resulted from remelting of 2100 Ma crust of several microcontinents. Based on ε_{Nd} of the Tampere schist belt, southern parts of CFGC, includes an Archaean component as well as abundant approximately 2.0 Ga detritus from the suggested microcontinents.

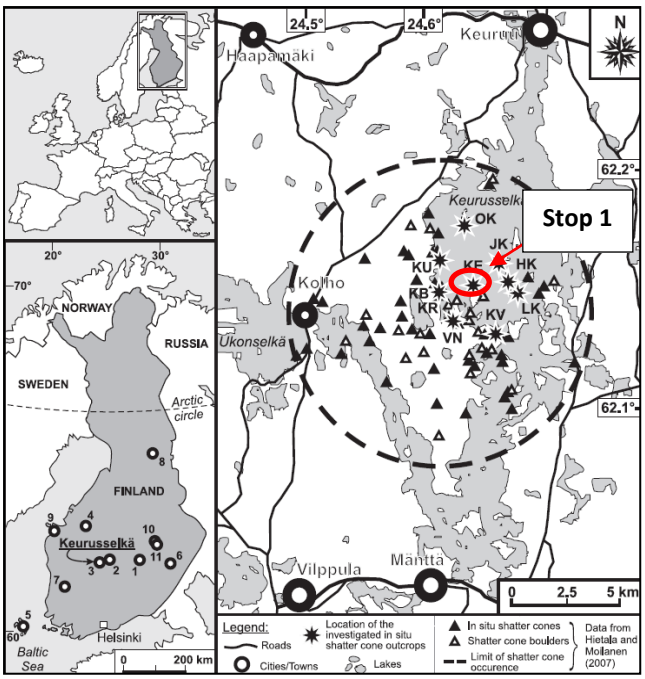


Figure 13. Left: General map of Finland (inset shows its position in Europe) with location of the Keurusselkä impact structure, as well as the position of the 10 other confirmed impact structures from Finland. Right: Detailed geographical map of the Keurusselkä structure, including roads, nearby cities and towns, lakes, sample locations of the investigated *in situ* shatter samples (marked with stars), and the occurrence of shatter cone boulders, as reported in Hietala and Moilanen (2004). The 11 confirmed impact structures from Finland are listed in p.80. S. Stop 1 (site KE) consists of in-situ shatter cones.

Supracrustal rocks of Svecofennian domain are, in many cases, separated by faults, fractures and intrusions dividing them into minor belts. The subseguent indicators, in additions to shatter cones and their PDF's, that support the impact origin was a discovery of a thin pseudotachylitic breccia vein showing planar formations (PFs) in quartz. (Schmieder et al., 2009)

Ages

According to database of Geological Survey of Finland (GSF), U-Pb (zircon) ages of the granitoids in the region show crystallization age of 1883 Ma (Keuruu granodiorite), 1882 Ma (Karikkoselkä granite), 1880 Ma (Multia granite; northern part of the Keurusselkä) and 1876±12 Ma (Keuruu dyke). Volcanic rocks in CFGC are somewhat older, approximately 1900 Ma. The Keurusselkä target rocks show also significant alteration of minerals due to possibly impact related or post impact processes. Three major shear zones cross the area, which are probably originally related to late Svecofennian events (*Nironen 2003*), but, at least, the NW-SE fracture appears to be cut by the impact structure. This fracture is clearly recognized as a weakly magnetic lineament in the aeromagnetic map (Fig.19).

Indications of impact: shatter cones

In general, the allochthonous breccia lens with fractured and shocked rocks is the main macroscopic feature present in impact craters (*Melosh 1989*). In the Keurusselkä area, the bedrock is strongly fractured, including *radial fracturing* in outcrops which can be distinguished from more linear, old fracture directions (Nironen, 2003) and from the common NW-SE pointing linear Holocene glacial striations detected across Finland (Fig. 14A).

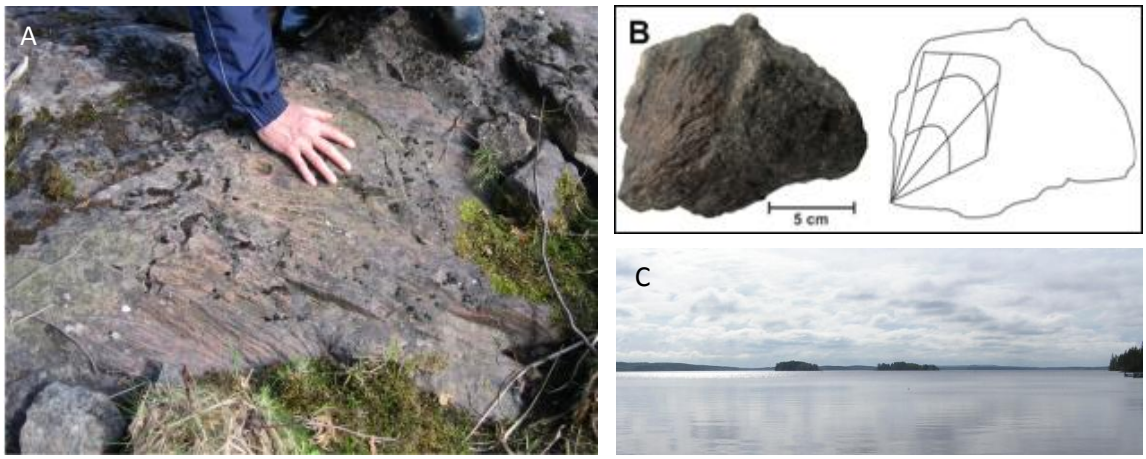


Figure 14. (A) Glacial striations and their direction in bedrock shown by Martti Lehtinen, (B) shatter cone boulder and a line-drawing of downward pointing apex. (C) Photographic image across the lake Keurusselkä.

The locations of clearly recognizable *in situ* shatter cones show in Fig. 16. The cones are abundant and well developed, principally in the northeastern part of the impact structure. Shatter cones are best exposed as clusters in dense and fine-grained metavolcanic rocks, but they are also observed in granites and in mica gneisses (Figs. 14, 15). At the Jylhänniemi peninsula (site KE; Fig. 13, 14, Stop 1), the target metavolcanites are highly fractured and shatter cones occur as large swarms with several different orientations. Shatter cones are generally poorly developed in coarse and weathered granite; in some cases, it is difficult to recognize them as they do not show clear cone apices and, thus, could be misidentified as slickensides.

Fig. 16 shows the sites of Ferrière *et al.* (2010) with clear in situ shatter cones. Boulders of shatter cones were found in several locations near the in situ ones like at site KE (Stop 1, Fig. 15). In situ shatter cones are mainly 15 cm long and 5–10 cm wide displaying pointing apices (Fig. 15). The cones are well developed, although they occasionally appear only as conical, striated surfaces.

The shatter cones are exposed in an area of diameter of 5 km and considered to represent the central uplift (CU) of the structure as they appear in a circular area close to the presumed center point of the impact (*Ferrière et al. 2010*). Figure 16b demonstrates that a majority of shatter cones are pointing upward, and have a tendency to point away from the central uplift area. The steep upward direction and the lack of well-defined intersection point of the apices may suggest that the shatter cones are peripherally located, around the central uplift. The planar deformation features in quartz, found from shatter cones in the central uplift, are decorated with fluid inclusions indicating that alteration by post-impact processes was present (Fig. 18).

Figure 15. (A) Damaged metavolcanic rocks with shatter cones on the shoreline of the Keurusselkä structure (site KE, Jylhänniemi, Stop 1; coordinates N62°9'9.3", E24°39'4"). (B) Large well developed almost full cone in metavolcanic rock from site KE. (C and D) Lauri Pesonen, Robert Klein and Harri Matikainen on the shore of Lake Keurusselkä, with block samples showing shatter cones.

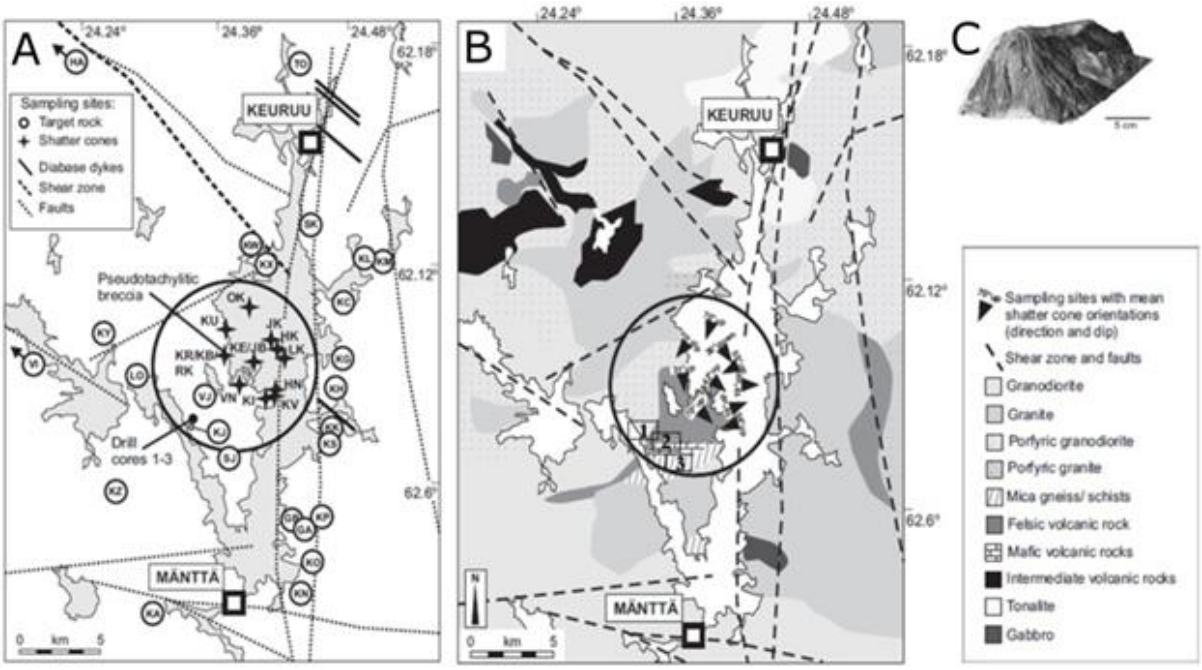


Figure 16. A) The Keurusselkä sampling sites. Solid ring marks the area of proposed central uplift. Sites with an arrow are sampled outside the map region along a shear zones/faults. B) Geological map showing shatter cone sites and the orientations of their apices and drilling locations (1, 2, 3). Geological map modified after Nironen 2003). C) Shatter cone boulder of metavolcanic rock.

Large impact breccia deposits or impact melt, however, were not found, which further indicates that Keurusselkä is deeply eroded, and possibly much older compared with the other Finnish impact structures (*Dypvik et al. 2008*).

Shallow drillings in the area of Keurusselkä impact structure: core V-002

Shallow cores (V-001, V-002, V-003) were drilled in the vicinity of the central uplift near Vilppula (Fig. 16b). The core lithologies consist of schists (metagraywackes), metavolcanic rocks, gneisses and breccias. Amphiboles and micas in the breccias are strongly altered and replaced by secondary chlorite. Chloritization may indicate widespread impact-induced hydrothermal alteration of the target rocks or it may be related to regional tectonic shearing.

The most interesting shallow drill core is the V-002 (Fig. 17), since it reveals distinct brecciation.

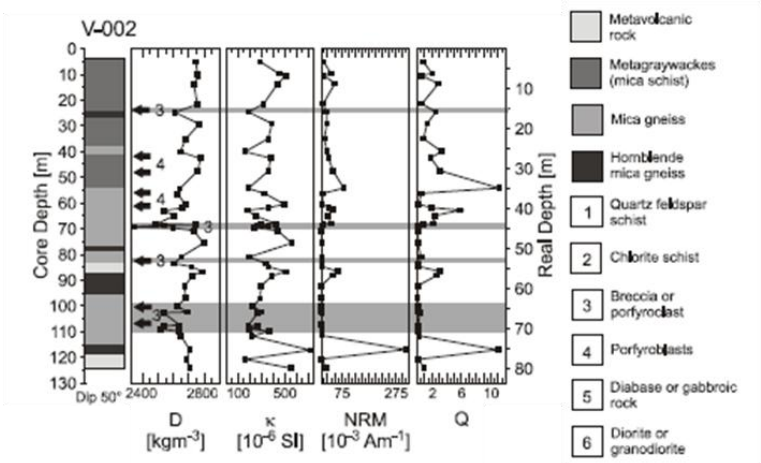


Figure 17. Petrophysical properties (density *D*, susceptibility *K*, *NRM* and *Q*) of Vilppula drill core V-002 with core lithologies. Gray sections within the data indicate the breccia or porfyroblast layers in the analysed cores.

Fig. 17 depicts narrow monomictic breccia veins in V-002 The density (average 2662 kg/m^3) has been observed to vary both due to fracturing and differences in lithology but the breccia vein cutting the parautochthonous subcrater floor has a very low density (~2576 kgm^{-3}), typical to suevite-breccia found in many impact structures (e.g. Lappajärvi, Karikkoselkä).

Based on topographic and satellite images, the estimate for the original size of the structure is 30 km (Raiskila et al. 2012), which is consistent with a 5 km wide central uplift. Therefore, the Vilppula cores are situated inside the impact region. Despite the location of the cores the samples did not reveal apparent impact features except perhaps the breccia veins (Fig. 17) surrounding target.

Post-impact hydrothermal alteration of crater rocks is a common impact-related phenomenon and this kind of activity is known from over 60 impact structures on Earth (*Naumov, 2005*). Occurrence of secondary mineral assemblages with ore-forming minerals and chemical alteration combined with fluid inclusions in the planar deformation features in

quartz (*Ferrière et al., 2010*) are typical evidences of hydrothermal alteration on Keurusselkä impact structure. The secondary minerals in Vilppula drill cores are mainly chlorite and pyrite. The first assumption for the origin of the studied Vilppula monomictic breccia veins would be endogenic. However, impact origin is likely due to the fact that these breccias are situated approximately 3 km from the assumed crater center showing impact pressures of 20 GPa. We note here in passing that petrophysical, rock magnetic and petrographic data of Vilppula drill core sample show different behavior of the breccia samples compared to background bedrock samples (see Raiskila et al. 2012). They suggest (but do not prove) a shock influence.

However, pseudotachylitic breccia veins are quite common in geological environments, and they are usually also associated with tectonic events. Yet, the discovery of weak shock features (PFs) in quartz (*Schmieder et al. 2009*) could point to the impact origin for this pseudotachylitic breccia and suggest a Mesoproterozoic age for the impact event. We return to this point in chapter, “Paleomagnetism”.

Shatter cones and planar deformation features

The shock produced planar deformation features (PDFs) in quartz and feldspar are clear evidence of a hypervelocity impact, as they are distinct from non-impact deformation features (*French and Koeberl 2010*). PDFs are also critical in recognizing deeply eroded impact structures, such as the Keurusselkä structure lacking other impact lithologies.

Ferrière et al. (2010) have reported results from selected samples (e.g. shatter cone of site KE (our Stop 1) using universal-stage measurement method. These samples revealed clear impact-related shock features, such as PDFs and PF’s in quartz and also feldspar and other mineralogical changes due to the shock. Examination of these thin sections showed that quartz grains with PDFs are mainly granulated. Overall, PDFs were found in a variety of minerals including quartz, plagioclase, and also apatite; specimens from site KE contain weak PDFs in plagioclase with set of PFs and kink bands in biotite grains. Thin-section sample (site KE) showed similar features to those of thin-sections of the Lappajärvi impact rocks (*Lehtinen 1976*) and the Karikkoselkä borehole samples (*Pesonen et al. 1999*).

The shatter cone samples from Jylhänniemi site (KE, Stop 1) consist of fine-grained granodiorite, showing slight variations in mineral composition and in grain size, reflecting the inhomogeneous nature of the basement rocks of the Keurusselkä structure, even at a local scale.

PDFs occur in quartz grains as narrow, nondecorated or slightly decorated, planes, but more commonly as bands of aligned, tiny fluid inclusions (decorated PDFs), spaced 2–10 µm apart. The PDFs commonly occur as multiple sets per grain, generally two sets (Fig. 18) under the flat-stage optical microscope, and up to three to five sets when seen under the U-stage microscope. The PDFs are generally decorated with tiny fluid inclusions, usually less than 1–2 µm in diameter (in some cases up to 3–4 µm). The decoration of the PDFs in Keurusselkä quartz grains indicates that they were subjected to postshock alteration (such as due to a saturated target or fluid circulation). Detailed description is given in Ferriere et al. (2008).

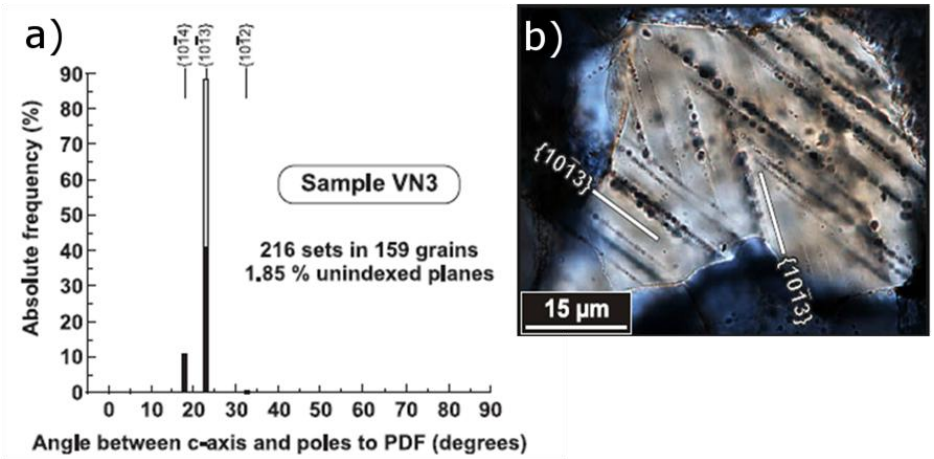


Figure 18. Histogram of the absolute frequency percent of indexed PDFs in quartz grains from shatter cone samples from the Keurusselkä impact structure, as determined using the new stereographic projection template (NSPT; *Ferrière et al. 2009a*) for the indexing. Note that PDF planes that fall into the overlap zone between {10$\bar{1}$4} and {10$\bar{1}$3} crystallographic orientations are considered as {10$\bar{1}$3} orientations in this figure, as suggested by Ferrière *et al.* (2009), but are reported in gray on top of the uniquely indexed {10$\bar{1}$3} orientations. a) Plot of combined PDF data from thin sections VN3A and VN3B. b) sample VN3B

Shatter cones and crater size

Shatter cones occur usually in the central part of complex impact structures and in a few rare cases, as isolated clasts in impact breccias (e.g., at the Haughton impact structure, Canada). The distribution of shatter cones at a minimum impact site has been used as a parameter for estimating the original size of a structure, particularly for old and eroded impact sites such as at Keurusselkä. It has been previously proposed that when restored to their original position prior the impact, shatter cones apices indicate the point of impact (e.g., *Manton 1965*). However, shatter cones at the Keurusselkä structure are generally found as clusters (rarely as single specimens) of partial or rarely complete cones, with frequently opposite orientations at the decimeter scale. Thus, the use of shatter cone apex orientation to determine the center of the Keurusselkä structure is likely to yield incorrect results. This is consistent with observations of shatter cone orientations from Vredefort Dome (see Raiskila et al. 2012 and references therein).

A minimum crater size can, however, be estimated based on the distribution of in situ shatter cones. These calculations yield to a minimum estimate of the original diameter ≥ 30km. Keeping in mind that we do not know about the amount of erosion in Keurusselkä case.

Geophysics

Aeromagnetic data from the Keurusselkä impact area shows prominent high amplitude magnetic anomalies (Fig. 19B). High-amplitude (up to 500 nT) short wavelength anomalies occur at the ~6 km wide area, which coincides with locations of the in-situ shatter cones and is distinguished from the overall regional field. This circular positive magnetic anomaly coincides with the negative Bouguer anomaly, but does not extend as far to the East. Circular anomalies further away around the central magnetic high-amplitude anomaly are likely of regional origin.

In the Keurusselkä structure, the shatter cones within crystalline rocks are exposed in an area of 19 km^2, and show distinctively higher magnetization features on aeromagnetic map (Fig. 19b). This anomaly is interpreted to form the circular central uplift of the structure. On the eastern side of the central anomaly is a circular magnetic low for which we do not yet have a definite explanation. The magnetic pattern in the Keurusselkä area shows also linear and folded schist belts and roundish granitoid intrusions, which are seen as magnetic lows. Raiskila et al. (2011) proposed that results from the drill core samples and the Keurusselkä shatter cones imply magnetic enhancement of the schist belt at the central uplift due to the impact.

The gravity data of the Keurusselkä region show a circular gravity minimum of ~9 m Gal around the central uplift (Fig. 19B). Chain of positive gravity anomalies with a diameter of 25 km is recognized around the central uplift as possible traces of the rim. Gravity influence of the Keurusselkä lake water was reduced from all gravity observations in the area by calculating vertical gravitation of lake mass with 6.4 meter mean depth of Keuruselkä lake area and with standard water density of 1000 kgm^{-3}. Other smaller circular negative Bouguer anomalies in the area were also found.

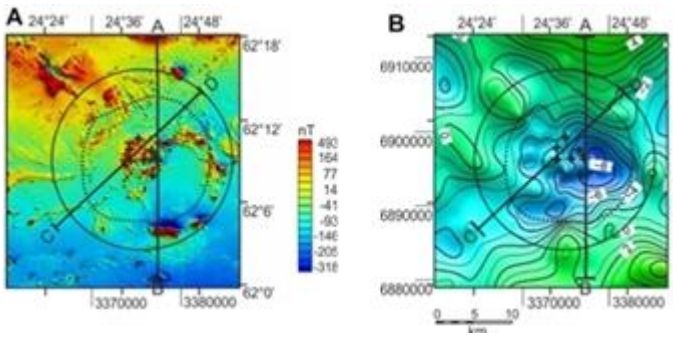


Figure 19 (A) Aeromagnetic map (courtesy of GTK). Shatter cones are marked as crosses. (B) Local Bouguer gravity map (courtesy of FGI).

These circular shaped anomalies are, however, mainly caused by younger granitic intrusions (~1860 Ma), which are recognized also in the geological map by Nironen (2003) and in aeromagnetic map as low amplitude regions (Fig. 19A).

The FIRE2 seismics profile (Kukkonen and Laitinen 2006) crosses eastern margin of the Keurusselkä structure from NS direction (Fig. 20). Recently, a tomographic analysis of the seismic data was done to study the possible impact features (Institute of Seismology, University of Helsinki, unpublished data). The tomographysic velocity model of FIRE 2 did not show any clear boundaries that could be tied with the Keurusselkä structure, however, within the structure area somewhat slower seismic velocities were recognized from the depth range of 570 meters.

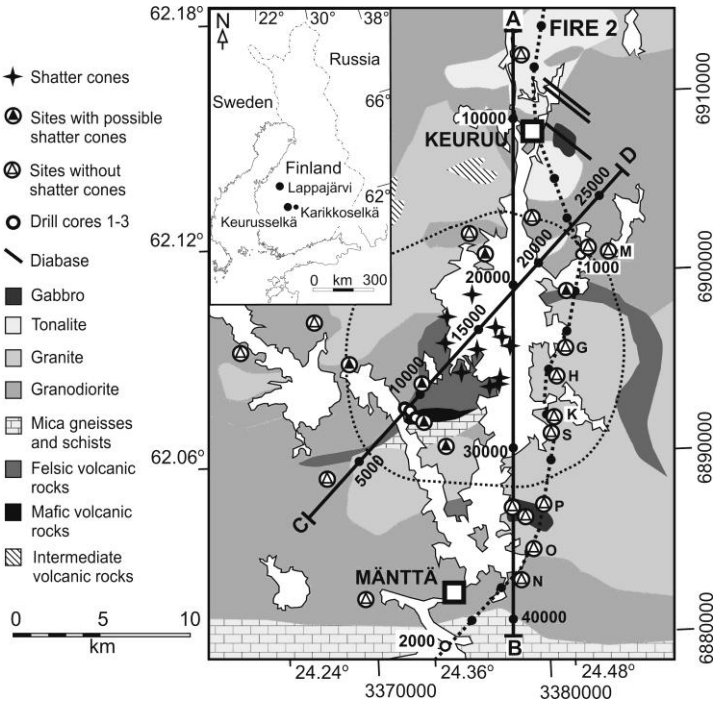


Figure 20. Dotted line shows the FIRE 2 seismic profile. The geological setting is by Nironen (2003) and the two marked modeling profiles (A-B and C-D) are shown as solid lines.

Crater dimensions from geophysics

To estimate the dimensions of the crater structure, scaling laws and dimensions outlined in Fig. 21 were applied. A medium size impact crater (with rim-to-rim diameter D ≈ 4–50 km) is of complex type with a central uplift (CU) (Fig. 21A). Assuming the shatter cones are located within the eroded CU of Keurusselkä structure, and based on the approximate diameter of their coverage area (D_{CU} ≈ 6 km), original diameter would be between 16-23 km. The structural uplift would be from 1.5 to 2.2 km and the diameter of transient cavity D_{TC} from 11 to 21 km (see Raiskila et al. 2012 and references therein). To adapt these theoretical parameters to the Keurusselkä structure and its central uplift, would results with a minimum crater diameter of 16 km, although the diameter of central uplift increases with the erosion.

Since the central peak collapse may start when the crater diameter exceeds 24 km (Abels 2003), it would suggest that the central uplift of Keurusselkä is not collapsed.

A digital elevation model (DEM) over Keurusselkä region (Fig. 21B) shows elevation points, which define a ring of hills, with anomalous heights over regional average, which could point the crater rim. The possible rim is partly missing as gaps along the ring, especially in southwest.

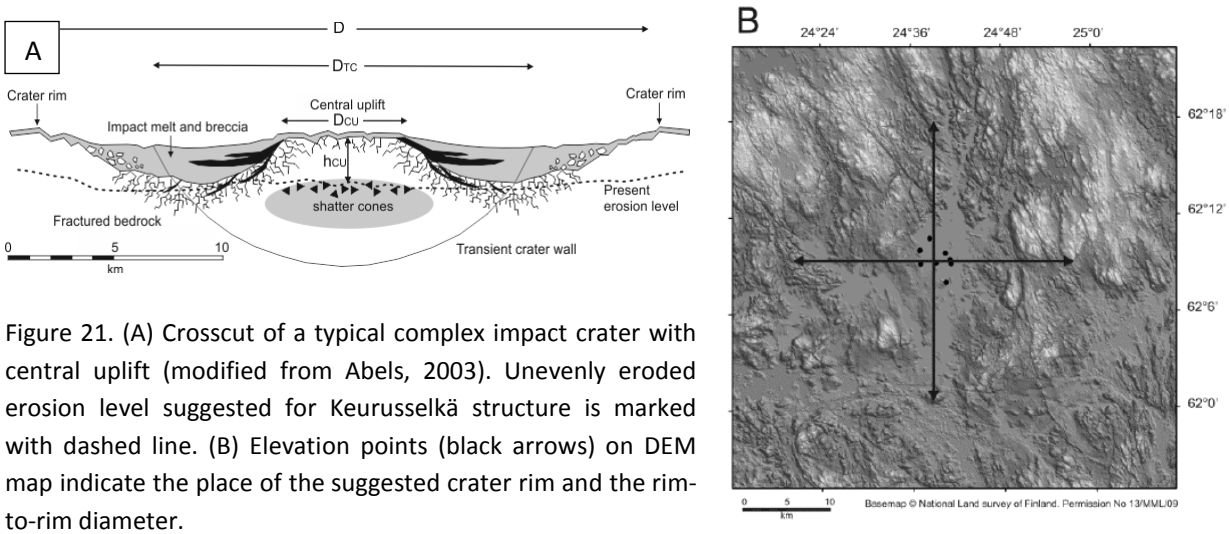


Figure 21. (A) Crosscut of a typical complex impact crater with central uplift (modified from Abels, 2003). Unevenly eroded erosion level suggested for Keurusselkä structure is marked with dashed line. (B) Elevation points (black arrows) on DEM map indicate the place of the suggested crater rim and the rim-to-rim diameter.

Modelling methods

A straightforward two-dimensional magnetic and gravity model is introduced to interpret the observed geophysical anomalies. Based on the models an estimate of the dimensions (diameter, depth) of Keurusselkä impact structure can be made.

The joint (gravity, magnetics) modeling was carried out using a 2.5-dimensional cross-sections with modelled source bodies of target rocks and impact generated anomalous sources. Details are presented in Raiskila et al. (2012) including petrophysical contrasts (model vs. background). The 2.5-dimensional models describing of a 5-km-deep section of the upper crust, are shown in Fig. 22. All the geological bodies in the model are related to the known geological features. The striped area represents a bowl shaped depression of gravity minimum with density of 2500 kgm^{-3} as partly based on Vippula drill core studies. This area shows the crushed and damaged crater basement, which exceeds to the depth of ~2 km.

They suggested that the strain model, where stresses, strains, and strain rates are all highest near the impact site and decrease with radial distance, should correlate with observed variations in bulk density (see also Pesonen, 2012). However, potential field

anomalies strongly depend on the erosion level and the deformation degree of the impact structures (Plado *et al*., 1999).

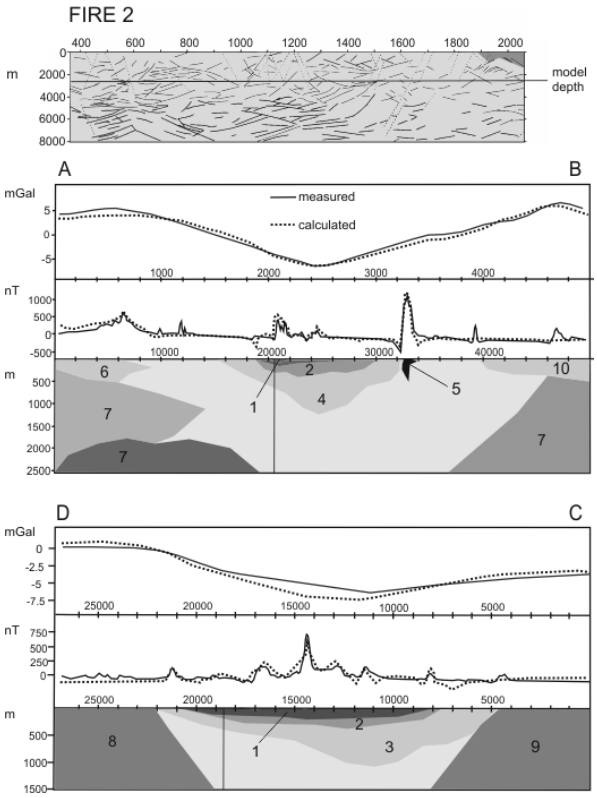


Figure 22. The line drawing interpretation of the FIRE2 seismic profile Nironen et al. (2006) Darker grey area represents the gneissose schists (see Fig. 24) and combined gravity and aeromagnetic models of a SW-NE (A to B) and NS (C to D) sections over Keurusselkä impact structure. Vertical dashed line in both models marks the cutting point of the profiles.

The NS profile (A-B) along the FIRE 2 seismic profile highlights the eastern margin of the circular central magnetic anomaly. The numbered source bodies from 1 to 5 have petrophysical values of (1= tonalitic) k=2900×10^{-6} SI, D=2720 kgm^{-3}, (2=granitic) k=1000×10^{-6} SI, D=2700 kgm^{-3}, (3=metavolcanic rock) k=3000×10^{-6} SI, D=2650 kgm^{-3}, (4=sill/granitic) k=1000×10^{-6} SI, D=2760 kgm^{-3} and (5=mica gneiss/schist) k=2000×10^{-6} SI, D=2900 kgm^{-3}. The darker gray objects (positions 11000 and 32000) are gabbro intrusions and the black rod feature (position 10000) indicates possible the continuation of the Keuruu diabase dykes or a gabbro intrusion.

Keurusselkä is deeply eroded: the rim and the impact rock units have been removed almost completely and only the upper part of the fractured basement is exposed. Yet, weak morphological features can be seen around the central uplift, which are partly asymmetric (Fig. 22). Reasons for uneven erosion could be an oblique impact, geological anisotropy, or both. The consistent bowl shaped region, down to the depth of 3 km, in Keurusselkä area can be identified from the seismic profile data (Nironen *et al*., 2006; Fig. 22.). Our model (NS-profile) highlights this area of less dense and fractured bedrock. This area has a diameter of 30 km and it explains the observed circular 6 mGal negative local Bouguer anomaly (Fig. 22).

Contrast to the Lappajärvi impact structure (diameter ~23 km, age 73 Ma), which has -11 mGal gravity anomaly, Keurusselkä could represent a ~0.5–1 km deeper section of the bedrock than Lappajärvi, based on the erosion estimation introduced by Plado *et al.* (1999). Heterogeneity of the target could also be the reason for the reduced anomalies. Also subsequent geologic processes are able to weaken the negative gravity anomaly, such as postimpact thermal alteration.

The measured magnetic anomalies of the Keurusselkä impact structure are complex. Impact-induced changes in the aeromagnetic anomaly pattern have typical impact features

of a strong circular anomaly in the center of the structure and an annular magnetic “halo” around the center (Pesonen et al., 2002), which is irregular due to geological heterogeneities and partly missing in the west side of the Keurusselkä structure. The positive central magnetic anomalies are clearly related to the area with in-situ shatter cones, where the rocks differ from un-shocked target rocks having increased magnetizations. The high-amplitude magnetic anomalies in the aeromagnetic data are associated with the outer margin of the central uplift (Fig. 22; A-B profile). Few gabbro intrusions create also sharp anomalies, while negative magnetic anomalies are linked to the granitic intrusions. The SW-NE trending profile highlights a polygonal shape body of a strongly magnetized area, revealing a circular central uplift with a peak anomaly in its center (Fig. 23).

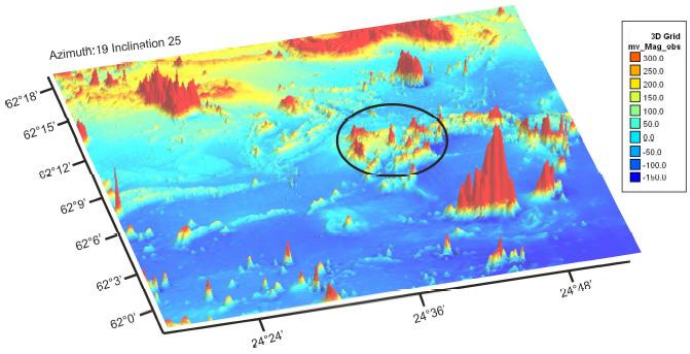


Figure 23. A 3D-image of aeromagnetic map. Ring area marks the magnetic anomaly associated to the central uplift. (Raiskila et al. 2012)

Paleomagnetic results

Paleomagnetic techniques and results are summarized in Raiskila et al. (2012) as based on alternating field (AF) and thermal treatments. Due to weak intensities and often also unstable demagnetization behavior, only less than half of the selected and demagnetized specimens showed stable remanence directions.

As the studied samples present the deeply eroded impact crater basement, we were not able to use paleomagnetic field tests for verifying the primary nature of isolated remanent magnetization directions. We attempted to apply the impact test of Pesonen (2001) in seeking evidence of decay of shock as a function of distance but the results were inconclusive. Four new remanece components (A, B, C and E) were isolated in Keurusselkä rocks. The directions are summarized in Fig. 24.

Component A. This component is best preserved on sites KL and KM ~10-20 km away from the impact center. It is especially observed during the thermal demagnetization but is also isolated during the AF treatments. This “A” direction is also clearly shown on sites, which are close to the faults and fractures.

The mean A direction is a typical Svecofennian direction (D~335º, I~46°) (Fig. 24, 25), observed in numerous locations within the Svecofennian terrain in Finland and Sweden (*Pesonen et al.1991*). Moreover, it is strikingly similar to direction isolated from unshocked granites near the nearby Karikkoselkä impact structure (*Pesonen et al. 1999*).

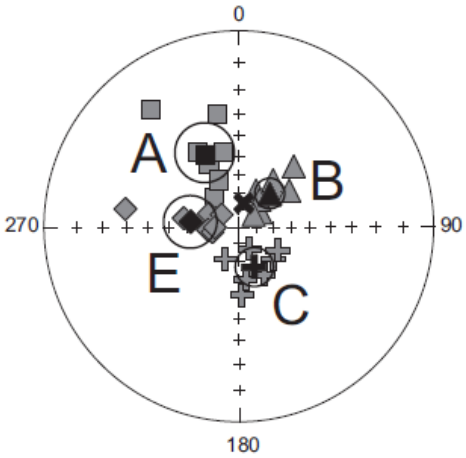


Figure 24. Site mean paleomagnetic directions for target rock and shatter cones. Component A refers to characteristic Svecofennian direction. Components B, C, and E denote the secondary directions. Black symbols display site means with α_{95} confidences about the means and gray symbols overall mean directions. Present Earth Field direction is marked as black X.

Component B. The NW upward intermediate inclination component B is isolated for nine sites occasionally occurring in the same samples with component A. It is best preserved in the sites with shatter cones showing mostly stable magnetizations and low-to-intermediate coercivities during the AF treatment. Blocking temperature ranges between 150 °C and 580 °C. B is in most cases isolated in the shatter cones the central uplift area with impact pressures around 20 GPa.

The mean direction of component B (D~42°, I~64°) is distinct from PEF (D~7º, I~73º) and yields a pole for impact-affected basement samples and shatter cones (Fig. 25).

The other components occasionally observed (C, E) are discussed in Raiskila et al. (2012).

Discussion

Petrophysical properties

Petrophysical properties of the Keurusselkä impact structure are generally typical for the granitic bedrock (CFGC). However, shocked crystalline rocks with shatter cones within the central parts of the structure show enhanced magnetic susceptibilities, which are interpreted to be associated with the central uplift. Increased susceptibilities in Keurusselkä structure could be a result of modification of magnetic carriers due to high P-T-conditions during the impact and later in hydrothermal processes. Hydrothermal activity has been significant after the impact event and it has caused fluid inclusion trails decorating the originally amorphous planar deformation features (PDFs) in quartz (*Ferrière et al. 2010*). Linear and folded schist

belts dominate the overall magnetic pattern over Keurusselkä. The magnetic signal of pyrrhotite-bearing schist belts is possibly also enhanced due to the impact (*Raiskila et al. 2011*).

Target rocks in impact craters are subjected to extreme pressure (up to 100 GPa) and temperature (up to 1000 °C) during the impact event. Typically, the porosity and fracture density decay gradually from the center of the crater to outer regions of the impact structure (*Cockell et al. 2005*). In the Keurusselkä area, the average porosities of shatter cones are low because of the weakly shocked crater basement and lack of highly porous impact rocks, such as suevites and impact melt breccias. The degree of fracturing is associated distinctively with the appearance of shatter cones at the central uplift area, and it is in resemblance to fracturing of the Karikkoselkä impact structure formed on the same granitic target rock (*Pesonen et al. 1999*).

Paleomagnetic Ploes and APWP

The characteristic component A obtained for different lithologies of unshocked target rocks and occasionally from shatter cones shows the typical Svecofennian remanent magnetization direction. This component is interpreted to represent the primary magnetization acquired during the formation and slow cooling of the CFGC at 1880–1860 Ma (Fig. 25A). Field tests for the CFGC area, to confirm the primary nature of magnetization, are not possible except in the case of nearby Keuruu dykes (~1876 Ma), with positive contact test. The magnetization direction “A” has been widely observed throughout the Fennoscandian Shield including the unshocked granitic target rock (1880 Ma) from Karikkoselkä impact structure.

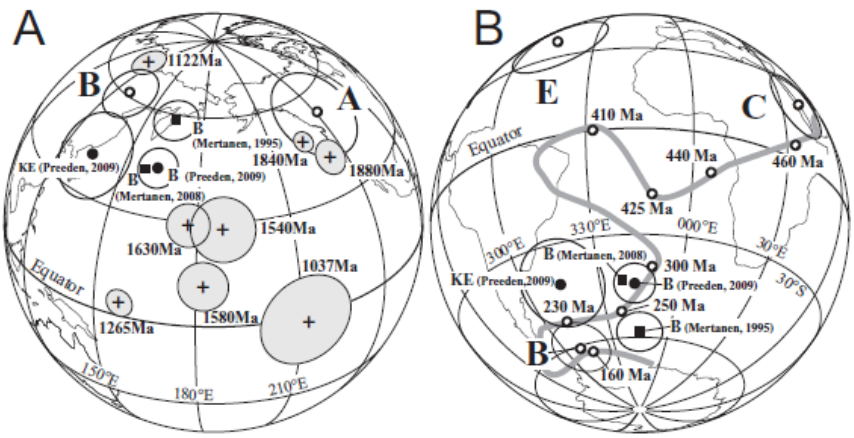


Figure 25. A) Mean paleomagnetic poles for characteristic Svecofennian component A and secondary component B calculated from shatter cone sites and unshocked target rock sites. Baltica “key poles” and chosen well-dated poles are marked as crosses with A_{95} circles (*Buchan et al. 2000; Salminen et al. 2009*). B) Fennoscandian Phanerozoic APW path (*Torsvik et al. 1996; Smethurst et al. 1998; Torsvik and Rehnström 2001, 2003*) and mean paleomagnetic poles for the south pole of component B and secondary components C and E.

Pole B is obtained mainly from shatter cones. This pole does not agree with the Precambrian key poles of Baltica (*Elming et al. 1993*). However, it is close to the pole of a well-dated (1122 ± 7 Ma) Salla diabase dyke (*Salminen et al. 2009*), indicating late Mesoproterozoic magnetization age. This agrees with the U-Pb age of 1140 ± 6 Ma of the pseudotachylitic breccia vein (*Schmieder et al. 2009*), supporting the idea that component B represents the magnetization due to the impact event. Component B has also low coercivity, which could be an indication of shock remanent magnetization (*Halls 1979)*. Keurusselkä and Salla poles are far from the Precambrian APWP of Baltica, thus creating a large clockwise loop from 1265 Ma to 1036 Ma, Salla pole (1122 Ma) being the apex of the loop. Salminen et al. (2009) discussed the similarities between the pre-Sveconorwegian (approximately 1.3–1.0 Ga) APWP loops of Baltica, Laurentia (including Logan Loop), and Kalahari-Grunehogna. These cratons seem to reach their apexes at slightly different times. Salminen *et al.* (2009) proposed that Baltica and Laurentia drifted together until 1.12 Ga and later separated at 1.04 Ga. If indeed all continents will reveal the 1.15-1.05 Ga "loop" a possibility of Mesoproterozoic true polar wander event (Evans 2003) will arise. However, better dated paleomagnetic poles for this interval are required to test this idea. Nevertheless, the ~1.12 Ga poles Finland provide hints that the Mesoproterozoic Baltica–Laurentia unity in the Columbia supercontinent may have lasted until 1.12 Ga.

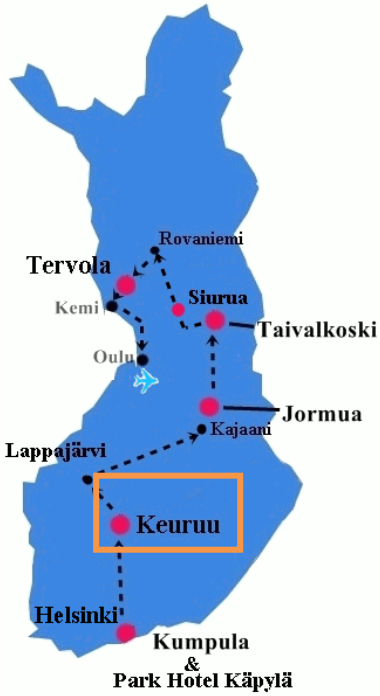
The opposite polarity of Keurusselkä pole B fits also to the Phanerozoic APW path (230 Ma). Mertanen and Pesonen (1995), Mertanen *et al.* (2008), and Preeden *et al*. (2009) have presented similar paleopoles (265 Ma) obtained from South Finland shear zones. They favored Phanerozoic age for their "B"-paleopole and interpreted them as remagnetizations caused by oxidizing fluids.

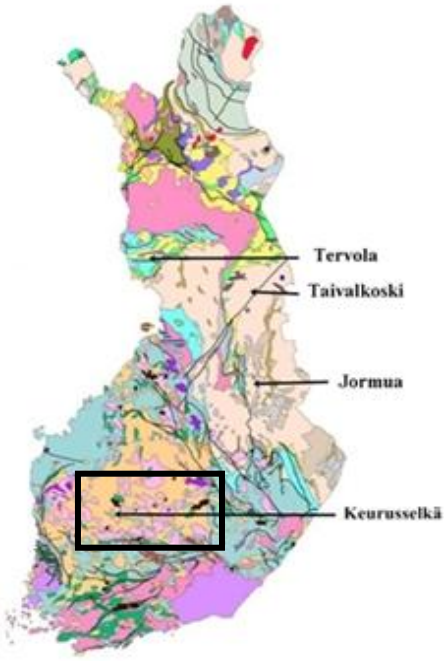
Conclusions

The Keurusselkä structure is a deeply eroded complex crater lacking the impact melt and rock lithologies, such as suevite and breccia. Impact features include (i) shatter cones in crystalline basement rocks with distinct planar deformation features (PDFs) in quartz, heavily fractured basement, (iii) discoveries of breccia veins in drillcores and a pseudotachylite with "young" age of ~1.14 Ga.

Petrophysical and rock magnetic properties imply that increased magnetization within the area, where shatter cones were observed, represent the central uplift with diameter of 5–10 km. Alteration of minerals in hydrothermal processes in the crater basement could be the reason for increased magnetic properties.

Four paleomagnetic components for the Keurusselkä rocks are interpreted to record the regional Mesoproterozoic and possible Paleozoic geological events during the evolution of the Fennoscandian Shield. Notably, component B is isolated in crater basement and shatter cones and is interpreted to represent magnetization related to the impact event at ~1.12 Ga supported by the $^{40}Ar/^{39}Ar$ age of the pseudotachylite vein (1140 ± 6 Ma). This new pole from the Keurusselkä impact structure gives further support for the proposed large APWP loop of Baltica during 1.26-1.12 Ga.





IV Day 1 (Sept. 22, 2012)

IV.Ib. Paleomagnetism and petrophysics of the Keuruu dykes

Lauri J. Pesonen, Robert Klein,Selen Raiskila,
Satu Mertanen and Pathamawan Sangchan

IV.Ib. Paleomagnetism and petrophysics of the Keuruu dykes

Lauri J. Pesonen, Robert Klein, Selen Raiskila, Satu Mertanen and Pathamawan Sangchan

The Keuruu diabase dyke swarm in Central Finland forms the second target of the excursion and comprises Stops 2 and 3. The dykes are of Paleoproterozoic age (~1876 Ma) and cut the Svecofennian gneisses, granodiorites and gabbros. The nearly vertical dykes strike NNW-SSE, have maximum length of a few kilometers and widths of ~ 0.3-2 m. More than one hundred samples were collected from 17 dykes in order to constrain the Paleo- to Mesoproterozoic APW-path of Fennoscandia, to estimate the ancient latitude of Fennoscandia at the time of Svecofennian orogeny and to study the magnetic fabric of the dykes and its relation to the Natural Remanent Magnetization (NRM). Chemical composition and petrophysical properties of the dykes are also presented in addition to a few paleointensity determinations.

Multicomponent analysis of NRM's, reveal three components: (i) a primary magnetization of dual polarity (N and R), (ii) a secondary component (B) and (iii) a viscous PEF-overprint. Positive baked contact test strongly suggest that the acquisition time of primary magnetization is close to the time of dyke emplacements during or slightly after the peak of the Svecofennian orogeny. The origin and nature of the secondary component is unclear but its connection to the Keurusselkä impact event at ~1120 Ma is not ruled out.

The schematic view of the Paleoproterozoic Keuruu dyke swarm is outlined in Fig. 26. where the NNW-SSE trending Keuruu dykes appear as an "isolated" swarm within Central Finland Granitoid Complex (CFGC). It is possible that Keuruu dykes show fanning to SSE. However, they are not an offshoot of the variable trending undated Kuru-Orivesi dykes, nor of the WNW-ESE trending Häme or Suomenniemi dyke swarms of late Paleoproterozoic age (~1.66 Ga). The latter ones show distinct fanning to ESE and seem to merge with the Wiborg or Suomenniemi rapakivi massifs.

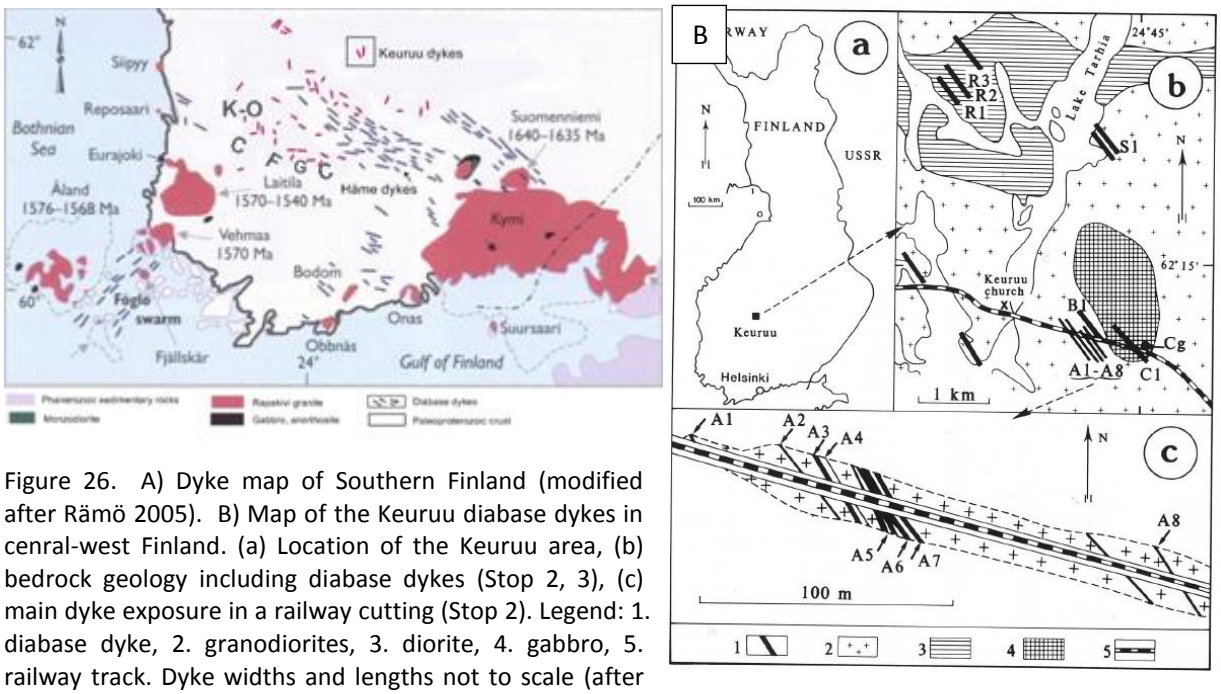


Figure 26. A) Dyke map of Southern Finland (modified after Rämö 2005). B) Map of the Keuruu diabase dykes in cenral-west Finland. (a) Location of the Keuruu area, (b) bedrock geology including diabase dykes (Stop 2, 3), (c) main dyke exposure in a railway cutting (Stop 2). Legend: 1. diabase dyke, 2. granodiorites, 3. diorite, 4. gabbro, 5. railway track. Dyke widths and lengths not to scale (after Puranen et al. 1992).

Fig. 26b outlines the Keuruu dyke swarm in more detail. Three distinct clusters are visible: (i) Keuruu railroad cluster (Stop 2), (ii) Pirttivuori cluster (dated dyke) and Multia cluster (Stop 3).

Age of Keuruu dykes

O. Kouvo (GTK) made several attemps in trying to date the Keuruu diabases using U-Pb techniques applied to zircon and other minerals. In spite of these attempts (only few zircon grains isloted with sufficient amount of U) no reliable age estimate was found. The ^{207}Pb-^{206}Pb age of slightly discordant U-Pb analysis on zircon of the Pirttijärvi dyke (N polarity) yield an age of 1876±9 Ma, which can be regarded as the best estimate for the age of Keuruu N-polarity dykes. This age is consistent with field observations showing that the dykes cut both the Keuruu gabbro and granodiorite (1883±14 Ma) (Fig.39b). No age data are available for the R-polarity dykes.

Petrophysics

Figure 27 summarizes the petrophysical data of Keuruu diabase dykes in two plots: density vs. susceptibility (A) and NRM vs. susceptibility (B). Petrophysical data are shown according to their paleomagnetic polarity (N, R or M (=mixed)) since most likely the polarities represent two separate pulses of dyking (see also geochemistry and paleomagnetism).

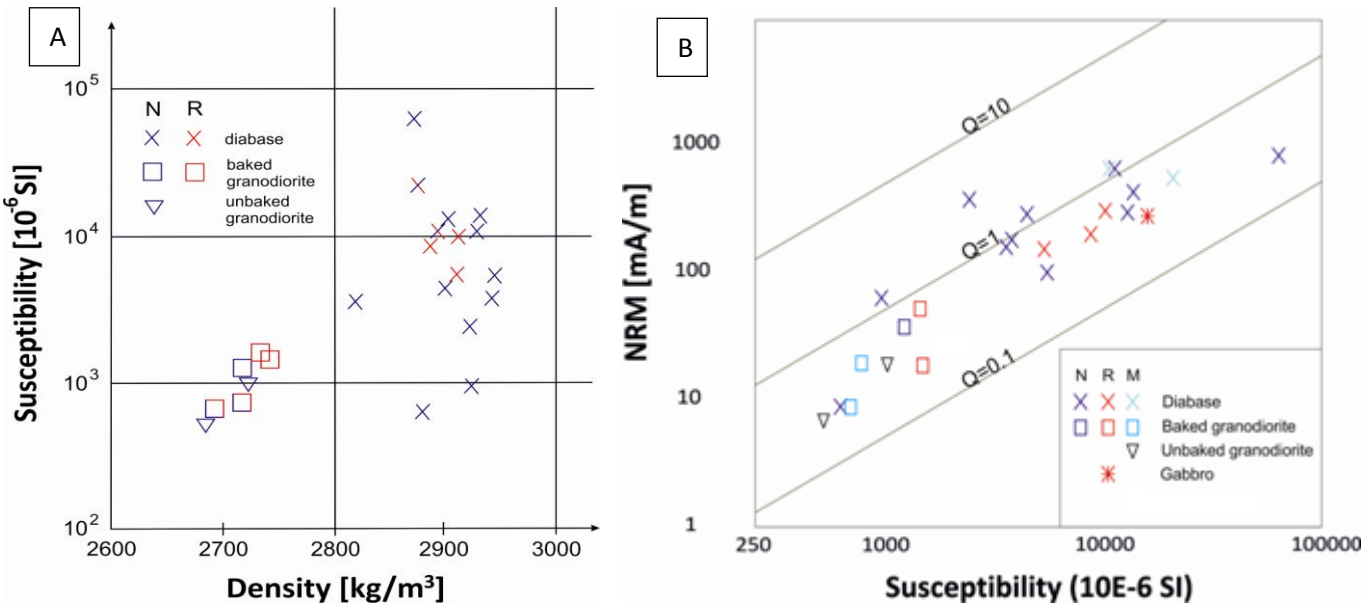


Fig 27. Summary of petrophysical properties of selected specimens from Keuruu diabases, baked and unbaked granodiorites and gabbro. N(R) denotes normal (reversed) magnetic polarity and M cases where they are superimposed. (A) susceptibility vs. density, (B) NRM vs susceptibility.

From Figure 27 we observe that:

1. Densities of diabases vary from ~2870 to ~2950 kgm^{-3} being typical or slightly lower than in mafic dykes in Finland as a whole (GTK database).

2. There is a tendency that the density of N-polarity dykes is slightly higher than that in R dykes. There are also slight differences in chemical and mineralogical compositions between the two polarity groups.

3. Susceptibility and its anisotropy do not show any trend with respect to polarity.

4. Unbaked granodiorites, as expected, have much lower densities (more Si) than mafic dykes. The baked granodiorites have “enhanced” densities, susceptibilities, NRM’s and Q-values presumably due to baking. These samples were selected to paleointensity determinations.

5. There is no trend in Q-values with respect to polarity: Q varies from 0.6 to ~4.5, typical for mafic dykes.

Hysteresis results are summarized in the Day-plot (Fig. 28) and suggest that the remanence carriers are in the borderline of the PSD-MD grain size consistent with AF-demagnetization behavior (Figs. 34-36).

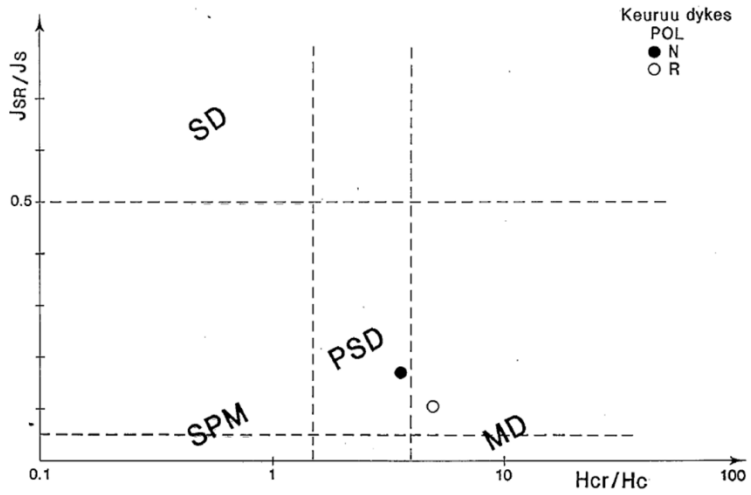


Figure 28. Hysteresis data of Keuruu dykes plotted in the standard Day-plot. (Day et al., 1997)

Magnetic fabric

A detailed review of the magnetic fabric of Keuruu dykes was carried out by Puranen et al. (1992) using the method by Lister and Kerr (1991). The dykes, as based on magnetic fabric can be devided into three groups: (i) group h with horizontal, (ii) group v with vertical, and (iii) group m with intermediate magma flow fabric. The anisotropy data of Keuruu dykes are plotted in Fig. 29 as L vs. F-plots.

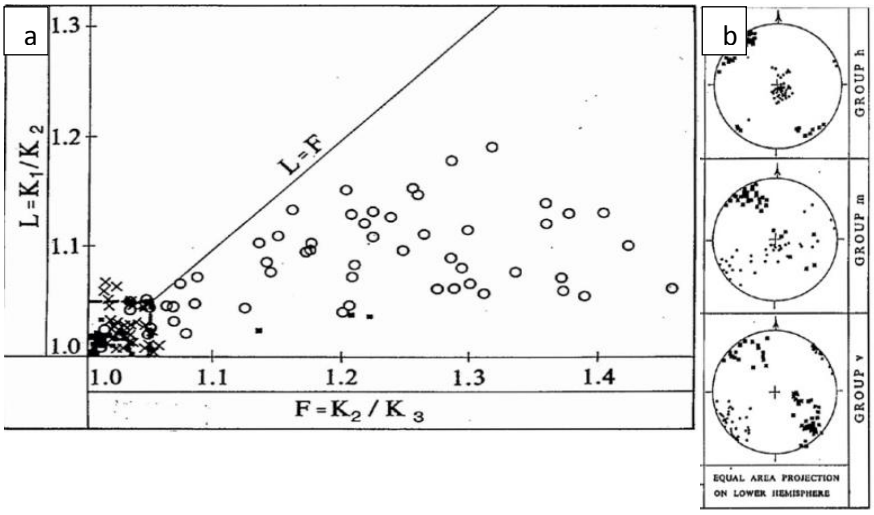


Figure 29. (a) Depencence between magnetic lineation L and factor F for dyke groups. The area defined by dashes in the lower left corner characterizes the typical field of magnetic flow fabrics. (b) Anisotropy directions which was the basis for dividing the dykes into h, v, and m groups. See Puranen et al. (1992)

Magnetic mineralogy, the Curie-point determinations (Fiof Keuruu diabase samples with h, v or m fabric types reveal the following characteristics:

1. All fabric types show thermal hysteresis: the heating and cooling curves differ indicating chemical alterations during heating.

2. Although several phases are seen in the heating curves: e.g., one at ~320-360°C (pyrrhotite?), all of them show Tc's near 570°C indicating Ti-poor magnetite as the main carrier. This is supported by distinct Hopkinson-effects at ~530-560°C in all fabric types.

3. Weak evidences for hematite is present consistent with microscope observations.Figure XX. Three examples of thermomagnetic curves (susceptibility vs. temperature) of Keuruu diabases. Left: fabric type h; Middle: type m and Right: type v.

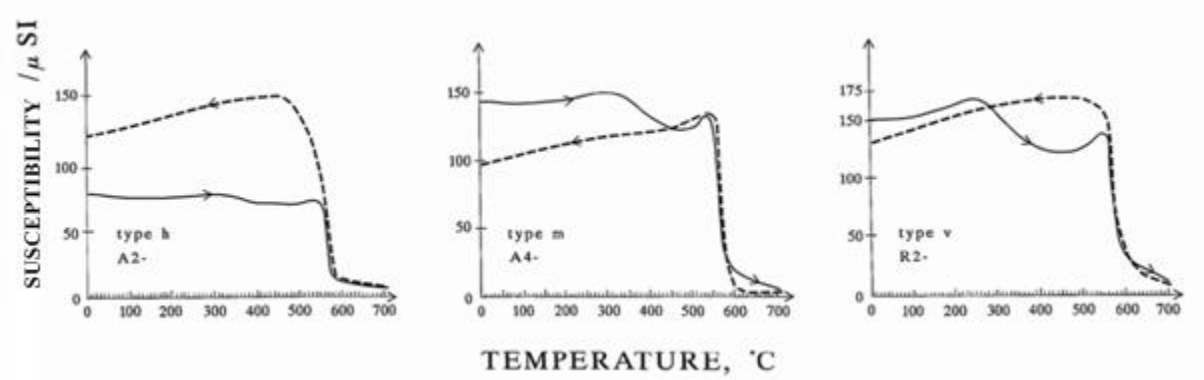


Figure 30. Three examples of thermomagnetic curves of Keuruu dyke samples with h, m and v fabric types.

Petrography and geochemistry

Microscopic studies of the Keuruu diabases reveal that the predominant ferromagnetic mineral is titanomagnetite, which is rather evenly distributed within the matrix as separate grains with sizes typically in the PSD-MD range 20-200 μm (see Puranen et al. 1992 for details).

Stop 2. This "railway exposure (Fig. 31) stop comprises several distrinct diabase dykes cutting the Keuruu granodiorite (Fig. 31) as well as the Keuruu gabbro. In the former case we can observe several cutting dykes (width from 30 cm to 1 m), sharp contacts with chilled margins and also dyke offshoots or apophyses (Fig. 31b).

The lack of secondary mineral inclusions in the titanomagnetite grains and their euhedral shape indicate that they are of primary origin, although some of the larger magnetic grains are partially martitized. Secondary magnetite is also commonly present as chains of platy grains that fill microcracks, and as dustlike concentrations of almost submicroscopic particles produced during the alteration of mafic minerals. The secondary alteration may be deuteric (late crystallization) but it can also be post emplacement alteration. Candidates for the latter one are late Svecofennian tectonic events (1.78-1.7 Ga) or, as pointed out in the previous

Chapter, the Mesoproterozoic impact induced hydrothermal event at ca. ~1.12 Ga. The paleomagnetic data give further support for this latter possibility.

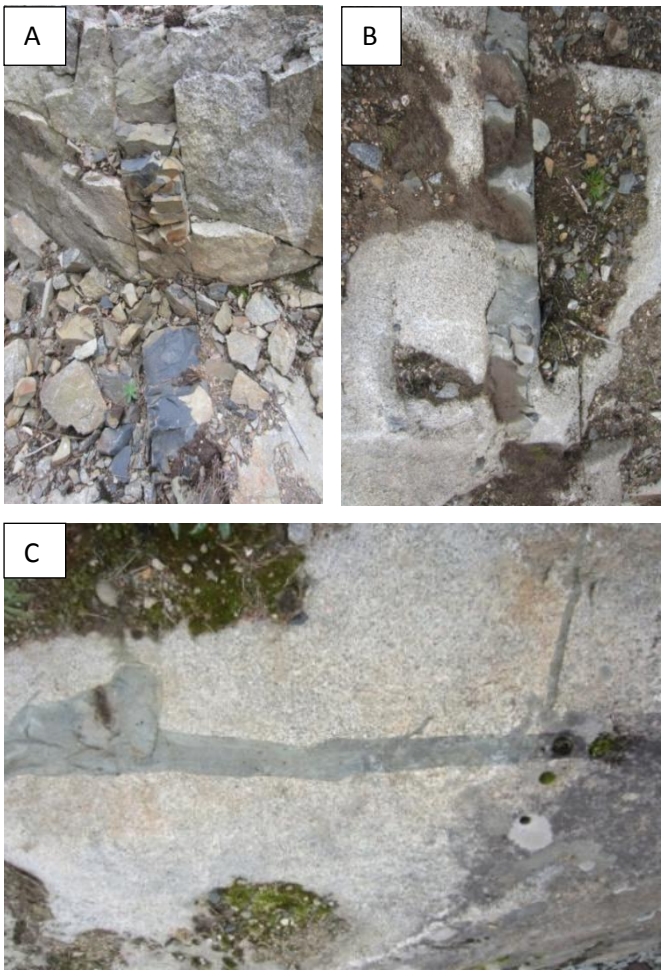


Figure 31. (A) and (B) Vertical dykes (widths 80 cm and 30 cm) with distinct contacts and chilled margins cutting Svecofennian grano-diorite; Railway exposure (Stop 2). (C) Off shooting dykelet (width 10 cm); Railway exposure on the side of the railway where the bus stops. Photos by Satu Mertanen.

The coarser central parts of the wider Keuruu dykes are ophitic and finer marginal parts are often porphyritic in texture. The ophitic parts are mainly composed of weakly aligned plagioclase laths (An_{50-55}) and hornblende grains. The porphyritic parts contain sporadic phenocrysts of plagioclase and pyroxenes (hypersthene and augite). In addition to these minerals, the matrix commonly includes small amounts of biotite, chlorite, apatite, epidote, titanite and pyrite (Marmo and Mikkola, 1963). The dated Pirttivuori dyke (1876±9 Ma) belongs to this type. The overall mineral composition and the degree of alteration do not change systematically between the different dyke groups and dyke widths.

The chemical composition of the Keuruu diabase is summarized in four discrimination diagrams (see data in Puranen et al. 1992). The results indicate that the basaltic magma was not intruded directly from the mantle but has undergone extensive near surface (low pressure) fractionation.

The diabase dykes are tholeiitic in character and were intruded in a continental setting. The dyke groups with different magnetic fabrics (h, v, m) are chemically quite similar, their degree of alteration is low, and they bear a close chemical resemblance to the coarse-grained gabbro of Keuruu. On the other hand, the overall mineral composition and the chemical composition of these dyke groups are not significantly different, suggesting a common or similar magma source. It is noteworthy that when viewed with respect to polarity (N, R) one can see slight differences in chemistry. The N polarity dykes appear to have higher Al_2O_3, CaO and MgO-contents but lower Na_2O and P_2O_5 contents than in R-polarity dykes. Similarly the N-polarity dykes have higher Cr, Cu and Ni contents than their R-polarity counterparts. These slight differences, as also observed in petrophysical data, may reflect age or source difference, keeping in mind that only two R-polarity samples were analysed.

Figure 32. Dyke R2 in Multia, on a roadcut exposure (width 30 cm) with somewhat fractured contacts. Photo by Satu Mertanen.

Baked contact test

We carried out baked contact test in four localites including both polarity cases. The tests were applied as profile samplings from the dyke onto the baked contact zone (generally within 15 cm from dyke wall) up to unbaked Svecofennian granodiorite.

Fig. 33 summarizes the results. Because the dykes (~1876 Ma) are nearly coeval with the age of the granodiorite (~1883 Ma) we expect no significant difference in paleomagnetic directions between N-polarity data of dykes and unbaked granodiorite. This appears to be the case (Fig. 33A): the directions of dykes, baked and unbaked granodiorites are quite similar showing a typical Svecofennian A-type direction. It is possible that the mean dyke direction is slightly shallower and northerly from the typical Svecofennian A-direction as seen in the unbaked granodiorite. We note here in passing that although paleomagneticlly the baked granodiorites yield similar directions as the unbaked granodiorites, they reveal higher

NRM's, magnetic hardnesses and Q-values, consistent with the idea that they have been baked by the dykes (see also thermal demagnetization data).

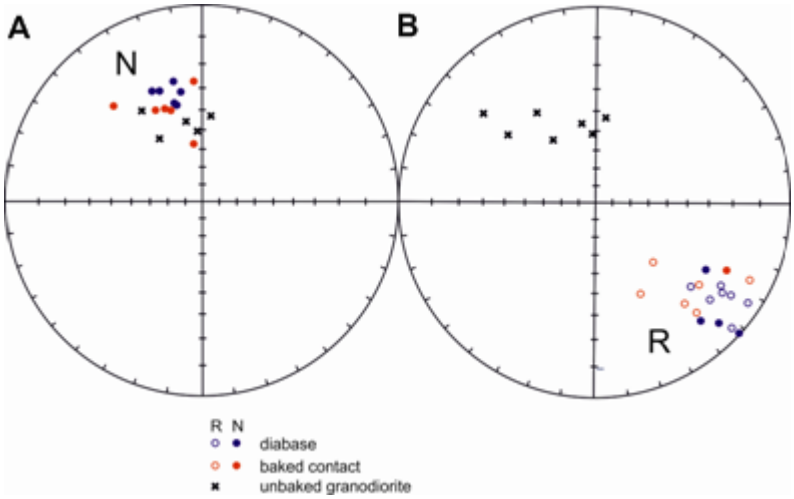
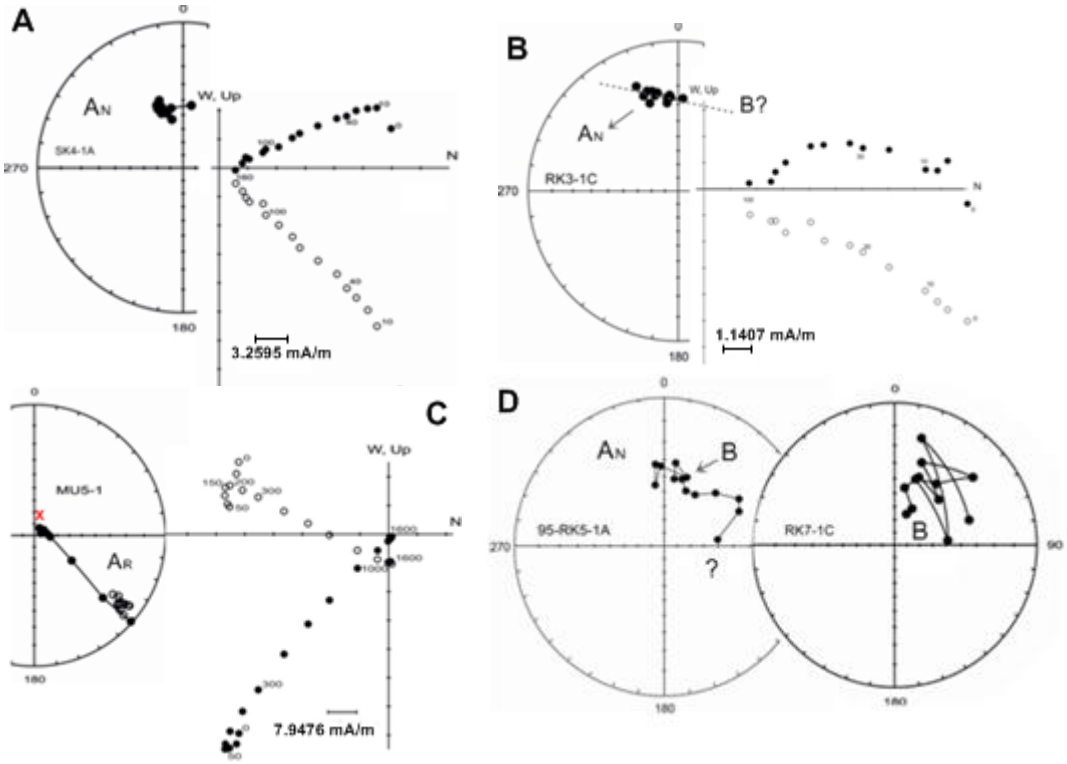


Figure 33. Baked contact test results in stereoplots for (A) N- and (B) R-polarity dykes.

The reversed polarity case (Fig. 33B) reveals clearly a positive baked contact test: the ChRM directions of baked granodiorites yield consistent directions with the ChRM's of reversed (R) V dykes whereas the unbaked granodiorites near R-dykes reveal N-polarity directions, as was the case with N-dykes. We thus have a fully positive baked contact test.

AF demagnetizations

Figure 34 shows representative examples of demagnetization behaviour of Keuruu dykes where components can be isolated using Zijderveld orthogonal plots.



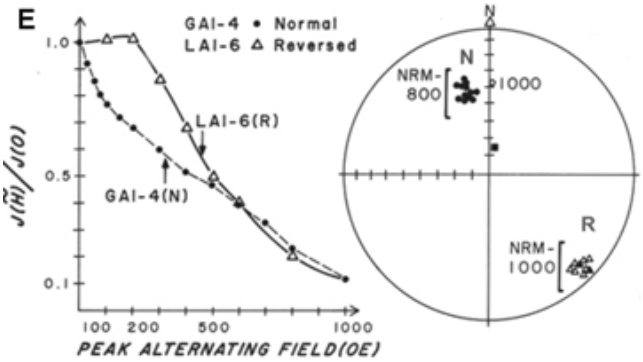


Figure 34. examples of AF-demagnetization of Keuruu dyke specimens. In A-C, Left: steroplot and Right: Zijderveld diagrammes. In D, two specimens with steroplots only, closed (open) symbol N-S (up down-E-W) projection.

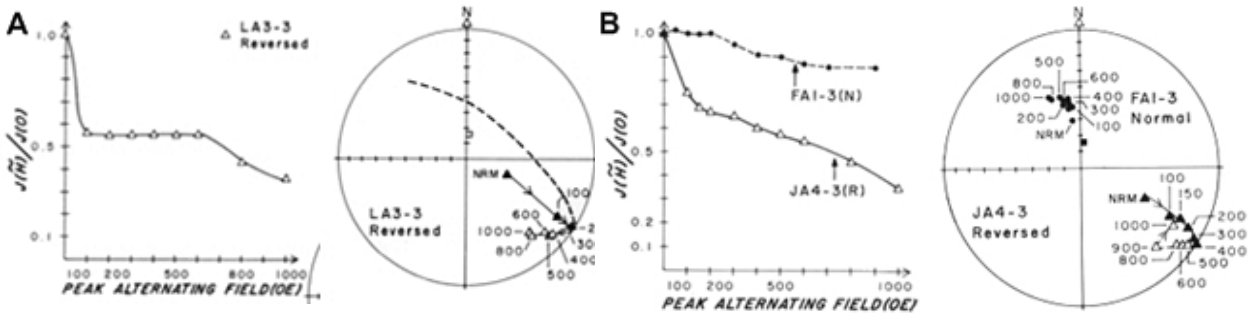


Figure 35. Two example AF-demagnetization behavior where end points serve as estimates of ChRM.

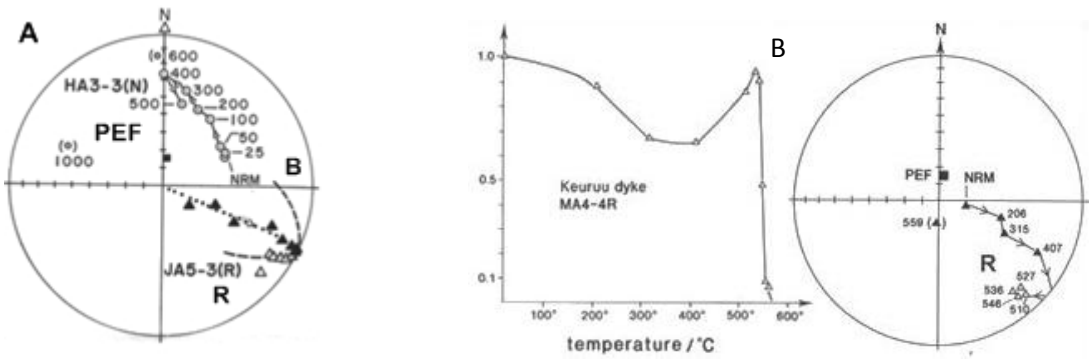


Figure 36. (A) Shows two examples (stereoplot) where two great-circles can be isolated in the same specimen. Sample JA5-3, where great-circle "B" join R and B components, and great-circle "PEF" joins PEF and R directions. Sample HA3-3(N) has A_N as the end-point and "B" or A_R as a low coercivity contaminant. (B) One example of thermal demagnetization revealing component A_R as endpoint.

Remagnetization circle analysis

The AF-and thermal demagnetization treatments of Keuruu rocks reveal great circle-type trajectories indicating superimpositions of two (or more) remanence components. In our analysis we used the great-circle tracing technique developed by Matti Leino (1994). Neglecting here those trajectories which involve PEF as a trajectory begin-point, the remaining trajectories revealed two intersecting great-circle families: (i) one moving from B to component A_N (Fig. 34D) and one moving from B to A_R Fig.(35A). Generally the movement

starts somewhere near B and moves towards either A_N or A_R, but opposite trends were also observed. Thus, in general the B is softer both in terms of coercivity or blocking temperature spectra. B component is also more scattered than A leading to fanning towards A_N or A_R. The intersections of the great circles provided us another way (in addition to tube find analysis of Leino (1994), to define the A_N and A_R. Results are shown in Table 1.

The great circle families ("A_N" and "A_R"), when superimposed into same stereoplot, intersect each other (Fig. 37c). This enables us to get another estimate for the secondary component B which yield to D≈47º, I≈64º (α95º=11, n=16 specimens). This estimate was obtained by pairwise counting the great circle intersections. This direction is close to the value (D≈34º, I≈45º, α95º=4.8, n=68) obtained by Zijderveld-analysis of individual specimens using orthogonal diagrams, vector subtraction or end-point methods (see Figs 34-36).

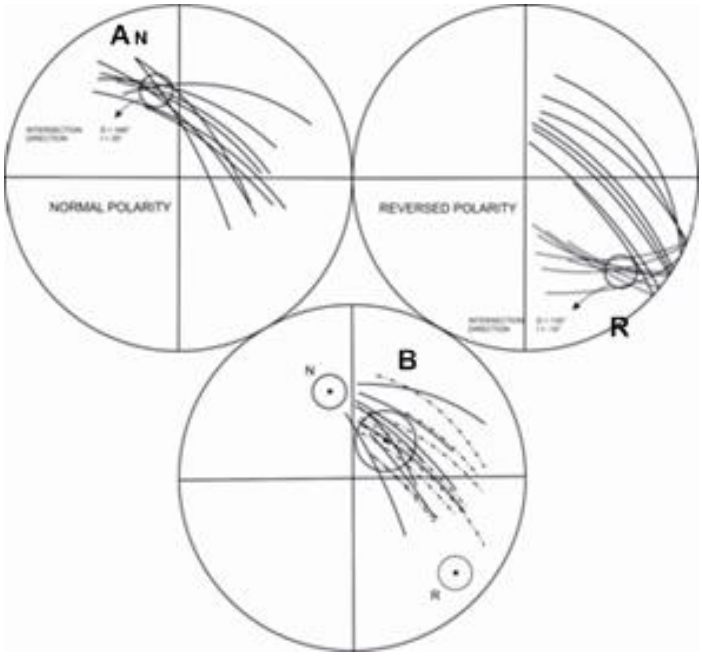


Figure 37. Remagnetization circles delineating (A) A_N, (B) A_R and (C) B components

Mean directions

The Keuruu paleomagnetic data are summarized in Table 1 and plotted in Figs. 38 and 39.

(i) Component A of dual polarity (A_N, A_R) in majority of samples
(ii) Component B (of N-polarity) in several samples as well as in using great-circle analysis
(iii) PEF as low coercivity component

Component A_N is typical Svecofennian "A-type" direction (D≈341º, I≈31º). Component B is more difficult to isolate. Since the dykes are younger than 2.1 Ga it cannot be the "same" B

as observed in several 2.2-2.1 Ga dykes (Varpaisjärvi dykes, layered intrusions, unless it is a late Svecofennian overprint of ca. 1.7 Ga (see discussion in Pesonen et al. 1989, 1990).

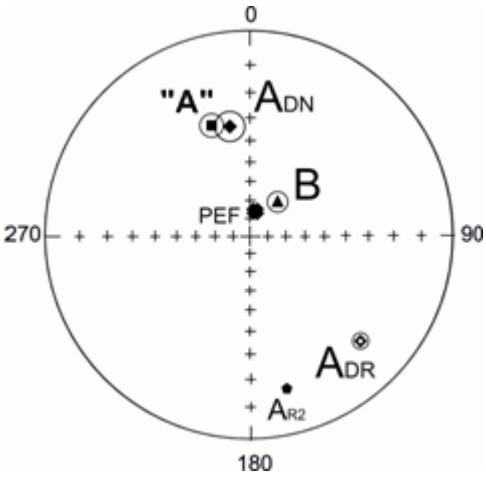


Fig.38 Mean paleomagnetic directions of this work.

Table 1. Paleomagnetic results of Keuruu dykes (Lat 62.3°N, Long 24.7°E)

Unit	P	*B*/n*	*D*(°)	*I*(°)	*k*	*α95*	Plat (N°)	Plon (E°)	A_{95} (°)
1.Keuruu N-dykes (z)	N	*11/34	349.9	33.4	79	6.5	45.4	218.4	2.17
2.Keuruu R-dykes (Z)	R	*3/12	133.6	-15.9	1694	2.9	26.3	257.8	1.52
3.Keuruu R2-dyke (z)	R	1/*2	166.2	-14.3	-	(35.3)	34.0	221.3	25.9
4a.Keuruu B (g.c)	N	/*40	41.0	196	18.1	5.5	63.5	127.3	5.29
4b.Keuruu B (z)	N	6/*68	34.6	55.0	13.9	4.8	55.7	149.7	4.83
5.Mean Svecof."A"	N	*4/	341.0	31.0	196	5.0	41.0	233.0	5.0

B/N Number of dykes/specimens, D declination, I inclination, k Fisher (1953) precision parameter, α95 95% confidence circle of mean direction, Plat, Plon Latitude, Longitude of the paleomagnetic pole, A95 95% confidence circle of the pole

APWP

The poles are plotted on the Fennoscandian APWP in Fig. 39.

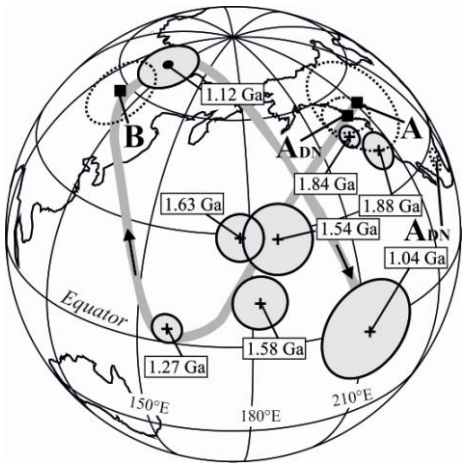


Figure 39. Paleomagnetic poles of this study plotted on the APWP of Baltica during 1.9-1.1 Ga. B denoted the secondary "B" component (see text).

The APWP-interpretation shows that N polarity poles lie in the segment of the Svecofennian APWP (1.9-1.8 Ga) consistent with the age (1876 Ma). In contrast, most poles of R polarity

lie significantly off from the APWP. This asymmetry could be caused by data bias, magnetic anisotropy, crustal tilting, secondary overprint, age difference or anomalies in the geomagnetic field. A fabric anisotropy study of magnetic *susceptibility* (Puranen *et al.* 1992) reveals that it is not the cause of asymmetry (see next paragraph).

Since a similar offset in N and R poles is observed in some other Svecofennian rocks from Sweden and Finland (Pesonen et al. 1989, 1990), the APW (i.e. the continental drift during the polarity crossing) or the asymmetric Earth's magnetic field interpretations are more likely candidates for the pattern of poles.

However, one has to remember that fabric anisotropy is determined using a weak field (45 μT) kappabridge, the results of which do not nessary tell whether the remanent magnetization will be deflected from the applied field direction by anisotropy.

To study the effect of anisotropy to remanence acquisition we performed a detailed IRM study of three specimens with low (l: $P \leq 0.03$), intermediate (i: $0.03 \leq P \leq 0.10$) and high (h: $P \geq 0.10$) fabric anisotropy. The results show that the anisotropy can only slightly (less than 10º) deflect the NRM but is unlikely to cause the polarity asymmetry seen in Fig. 38, which is ≥ 15º (Table 1).

On the APWP of Fennoscandia, the pole of B may suggest a Precambrian age between 1.7-1.0 Ga or a Phanerozoic (Triassic to Jurassic?) age. Given the lack of any Phanerozoic widespread hydrothermal event in the area, we propose here that the B-component is caused by the Keurusselkä impact event at ~1.12 Ga as discussed in the previous chapter. We speculate that the impact has triggered fracturing (radial or peripherial) enhancing widespread hydrothermal fluids to penetrate into the dykes and target rocks causing the partial overprinting B. If this turns out to be real, it is equally likely that the microscopically observed evidences of secondary alterations link to this same impact generated alteration process, as discussed in previous Chapter (see Raiskila et al., 2012a, b).

Paleointensity

Using the Thellier-technique on Keuruu rocks we attemped to estimate the intensity of the Earth's magnetic field at 1.88 Ga. Only three baked granodiorite samples yielded valuable data. The results suggest a weak paleointensity between 10.0-12.5 μT yielding a mean VDM of $2.5 \pm 0.1 \cdot 10^{22}$ Am^2 (Fig. 40).

The result supports the idea that the field during Paleoproterozoic was very weak as shown in Fig. 40B.

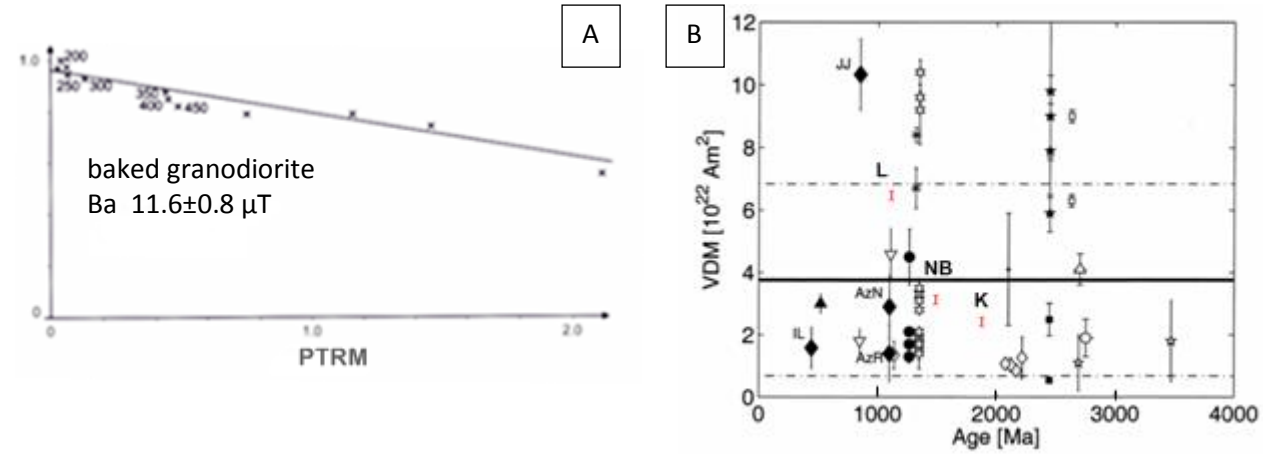


Figure 40. (A) One example of successful Thellier-paleointensity determination of a baked granodiorote (KR 11). (B) Mean VDM of Keuruu dykes (=K) plotted on global Precambrian VDM-plot of Donadini (2007). Also included are previously unpublished data of Åland Subjotnian (Norra Betesön dyke = NB) dyke and of Laanila (1.05 Ga) dyke (=L). JJ is VDM of Jänisjärvi impact structure (Salminen et al. 2007), IL is Ilyinets impact structure (Pesonen et al. 2003) and AZ, AN are data of Arizona dykes (~1.1 Ga) (Donadini et al. 2011).

Position of Baltica at 1.88 Ga: joined with Laurentia

The mean Keuruu paleomagnetic pole (only N-polarity accepted here) is in good agreement with the mean Svecofennian pole for which an age 1.88 Ga has been assigned. This pole was used in several Baltica-Laurentia reconstructions (Pesonen et al. 2003; Pesonen et al. 2012; Mertanen and Pesonen, 2012). The reconstruction of Baltica-Laurentia is reproduced here including reliable poles of the age of about 1.88 Ga available from three Nuna cratons (Baltica, Laurentia and Siberia), from two Ur cratons (Australia and Kalahari), and from one Atlantica cratons (Amazonia). The reconstruction is shown in Fig.41. All continents have moderate to low latitudinal positions with the exception of Kalahari which seems not to belong to this “Early Nuna” landmass. The so called Ur continent is thus not supported due to significant separation between Australia and Kalahari.

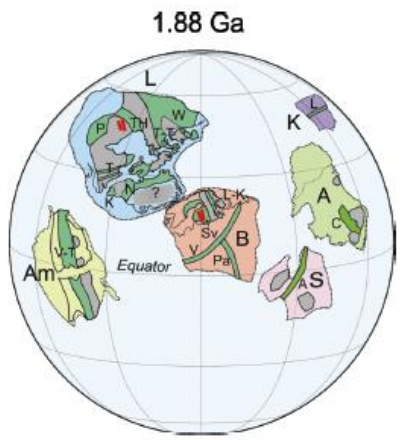


Figure 41. Reconstruction of cratons and orogenic belts (green) at 1.88 Ga. The Archean cratons are shown in gray. Data available from Laurentia (L), Baltica (B), Amazonia (Am), Siberia (S), Australia (A) and Kalahari (K). The ca. 1.90 - 1.80 Ga orogenic belts are shown in dark green and they are: Laurentia: Nagssugtoqidian (N), Ketilidian (K), Torngat (T), Trans-Hudson (TH), Penokean (P), Woopmay (W), Taltson-Thelon (T-T), Baltica: Lapland-Kola (L-K), Svecofennian (Sv), Amazonia: Ventuari-Tapajos (VT), Siberia: Akitkan (A), Australia: Capricorn (C), Kalahari: Limpopo (L). For explanation.

The assembly of Laurentia and Baltica at 1.88 Ga, together with Australia and Siberia marks the onset of the supercontinent Columbia although the final amalgamation occured as late as ~1.53 Ga (Pesonen et al. 2012). The position of Baltica against Laurentia is rather well established as paleomagnetic data are available from several Svecofennian 1.88 - 1.87 Ga gabbros and Keuruu dykes. However, the age of the mean pole is somewhat uncertain as discussed in Mertanen and Pesonen (2012). The uncertainly of Laurentia's position is due to complexity related to the "B" pole of the Molson dykes (Halls and Heaman, 2000). We note that the Laurentia-Baltica unity (Fig. 41) departs from the 2.45 Ga configuration, consistent with separation of Superior from Karelia between 2.45 Ga and 2.05 Ga. The data further suggest that a considerable latitudinal drift from 2. 45 Ga to 1.88 Ga took place for Laurentia but much less for Baltica. Although the data from Amazonia are not of the best quality (Q ≈ 2-3), Amazonia was probably not yet a part of the 1.88 Ga Laurentia-Baltica assembly (Fig. 41).

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City of Keuruu

Keuruu became an independent parish in 1628, but only gained a town status in 1986. Today there are approximately 20 000 inhabitants of which 7000 live in the central Keuruu.

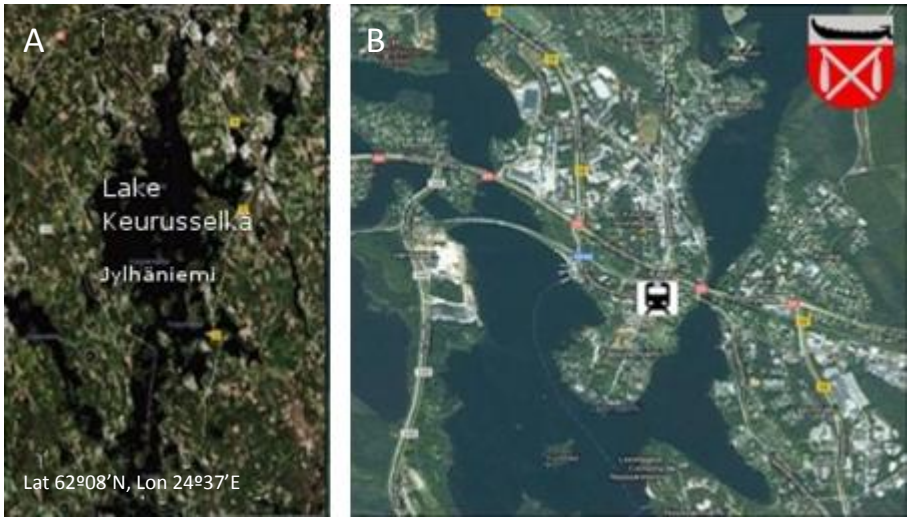

Fig 42. Keuruu is located in the middle of western Finland, and most of the center of Keuruu is built on four islands by the Lake Keurusselkä.

The building base at Old Keuruu (Vanha Keuruu) area is nationally recognized. The first church at Keuruu was built and designed by Antti Hakola between 1756-1759. The church was inaugurated in 1758 and was decorated by Johan Tilén.

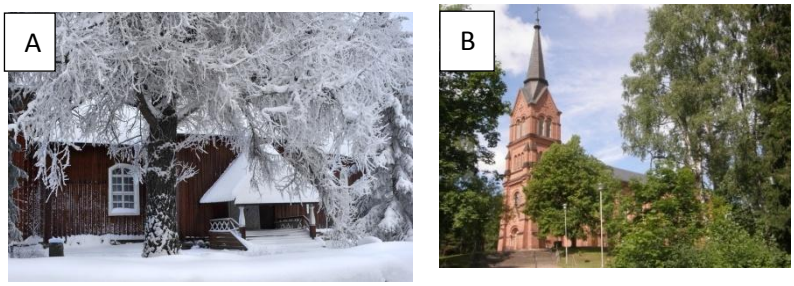

Fig. 43. A) The old wooden church of Keuruu from outside. B) Keuruu new church (anno 1892)

The old church was replaced by the new church of new-gothic style, designed by Theodor Granstedt (1842-1927). The altar painting is made by Eero Järnefelt (1863-1937), a famous Finnish painter. The inside of the new church is airy and cathedral like, and the foundations of the brick church are made of granite.

The annual Keuruu market takes place in mid-July and is the biggest event during mid-summer in western central Finland. There are over 200 sellers and stands and the event draws about 20000 visitors each year to buy unique objects and to enjoy beautiful Keuruu.

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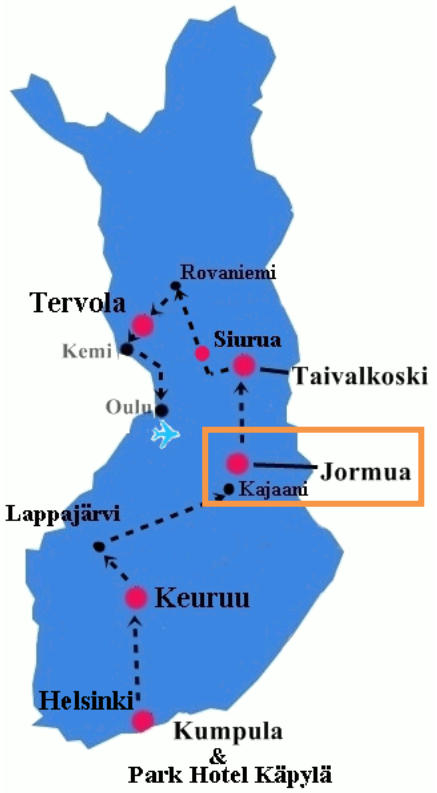
Lappajärvi town

Figure 44. (A) Lake Lappajärvi, which gives the name to the municipality, is a meteor crater, one of the few meteor crater lakes found in Finland. (B) Meteorite exhibition at Kivitippu Spa Hotel (Our 1st overnight place).

Lappajärvi is best known for its connection to a meteorite impact and is rightly called the ”municipality of the meteorite”. It is a blue jewel in the SW Finnish countryside and also the biggest lake in Ostrobothnian Finland. The meteorite has been estimated to be nearly one kilometer long and it hit the place roughly 73 Ma ago. The deepest point of the lake is at 38 m depth, however the average depth is between 5-20 metres. The impact gave rise to kärnäiitti impact melt which is a typical rock in this area (also the Ostrobothnian provincial rock!). The kärnäiitti-rocks can be seen on the lakeshores and islands.

At the downstairs of Kivitippu Spa there is Meteorite Exhibition. Here you can explore meteorites and effects of meteorites (impacts) and how the lake Lappajärvi was born.

City of Lappajärvi– *Meteorite municipality – feel the space*: http://www.lappajarvi.fi



Day 2 (Sept. 23, 2012)

V Jormua Ophiolite

Asko Kontinen, Lauri J. Pesonen, Satu Mertanen,

Robert Klein, and Ella Koljonen

During the morning of Sept. 23 we drive from Lappajärvi through Ostrobothnia all the way to Kajaani and Jormua.

We will pass several historical cities and villages full of history from Swedish Kingdom time (1397-1809) followed by Russian domination (1809-1917), not forgetting the events occurring during our civil war (1917-1918), during the winter war (1939-1940) and Continuation War (1941-1944).

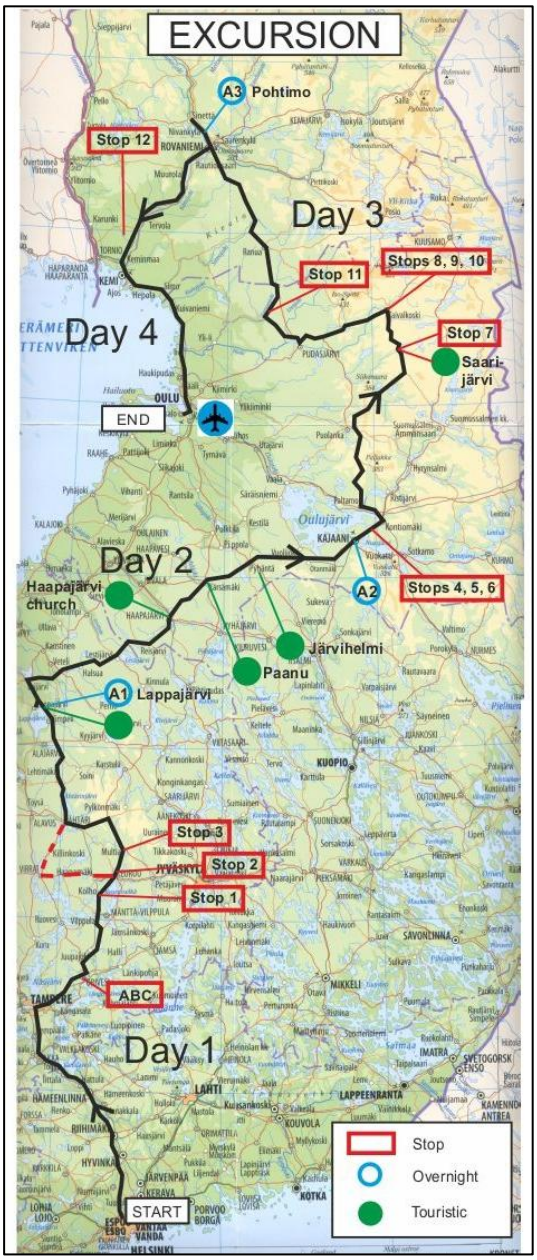


Figure 45. (a) Haapajärvi wooden church, (B) Ring bell of Kärsämäki shingle church "Paanukirkko" (here we have 30 min stop), (C) Kärsämäki shingle church (Paanu), (D) Restaurant Järvihelmi, our lunch place in Pyhäntä on Sept. 23.

Figure 46. Bus route from Lappajärvi (A1) -Haapajärvi-Kärsämäki-Pyhäntä to Jormua.

Fig. 46 shows our scientific excursion stops, overnight places and some tourism attractions along the route.

The Jormua Ophiolite Complex

Location, main geological features and tectonig setting of origin

Three Precambrian (2.0-1.95 Ga) mafic–ultramafic rock complexes are presently known from Finland: Jormua, Outokumpu, and Nuttio – that reasonably can be interpreted as ophiolites and fragments of ancient oceanic crust. From these the Jormua and Outokumpu occur in eastern Finland along the Karelian craton margin, enclosed in deep sea turbidites of the allochthonous Kaleva (Fig. 47). In terms of rock assemblage the Jormua Ophiolite Complex (JOC) is the most complete one containing all the salient lithologic units of an ophiolite: mantle tectonites, gabbros and plagiogranites, mafic sheeted dikes and pillow lavas and hyaloclastites. Although tectonically disrupted the JOC also has a reconstructable stratigraphic structure similar to ophiolites and oceanic crust (Fig. 48). The sheeted dykes and extrusive basalts have a distinct major and trace element affinity with E-MORB (Kontinen 1987, Peltonen et al. 1996). Also the gabbros and plagiogranites are of clear oceanic E-character being characterised by relatively high HFSE but very low Rb and Th contents.

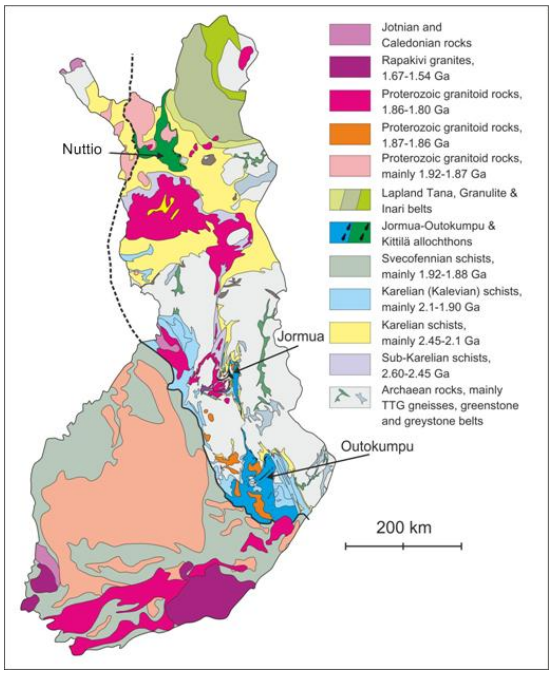


Figure 47. Loaction of the Jormua, Outokumpu and Nuttio ophiolites and host allochthons on a generalised geological map of Finland. The abrupt boundary or "suture" between the Svecofennian arc domain in the west and Karelian cratonic domain in the east is shown by thick black line. The dashed black line indicates possible continuation of the suture in northern Sweden and Finland. The Jormua–Outokumpu allochton consists >90 % of deep sea 1.92–1.90 Ga turbidite wackes-shales (Lahtinen et al. 2009), whereas the Kittilä allochton consists >80% of c. 2.0 Ga MOR and IAT type basaltic rocks (Hanski & Huhma, 2005).

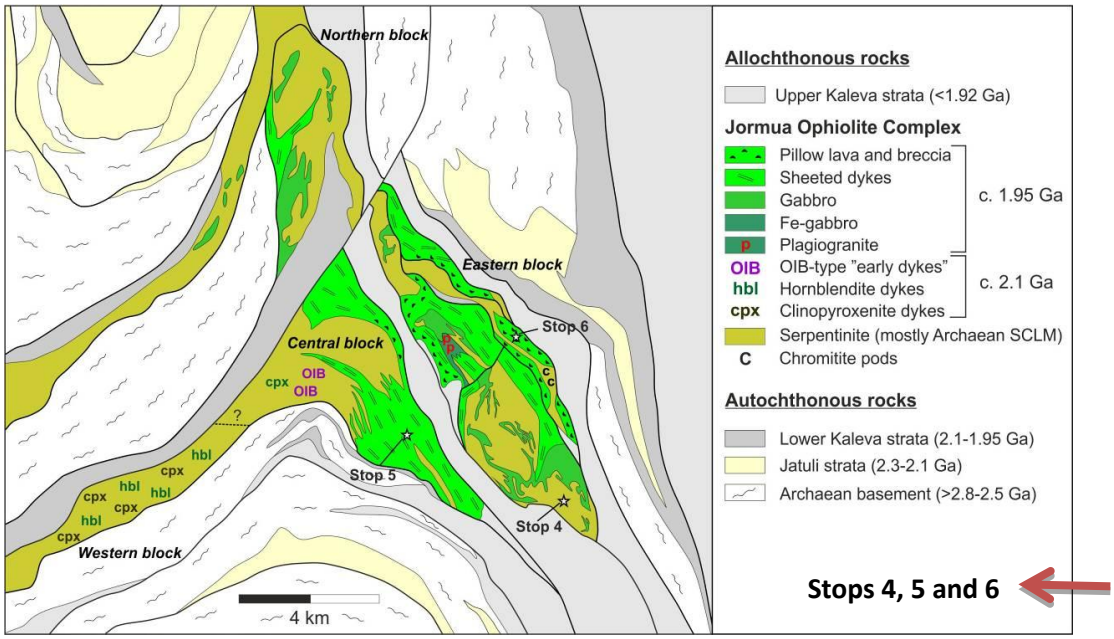


Figure 48. Generalised geological map of the Jormua Ophiolite Complex showing locations of the Jormua stops 4, 5 and 6 of the Supercontinent Symposium 2012 excursion. The Upper Kaleva comprises dominantly deep water metaturbiditic (amphibolite grade) greywacke–shale–black shale locally enclosing ophiolite fragments. The Lower Kaleva is characterised by riftogenous marine sediments such as conglomerates, quartz rich wackes, phyllites, iron formations and black shales. The black shales are locally metal rich; at Talvivaara a c. 1.6 Gt deposit @ 0.22 % Ni, 0.49 % Zn, 0.13 % Cu and 0.02 % Co is currently utilized by bioheapleaching for c. 30 Mt of ore per annum (Kontinen, 2012). The Jatuli strata contains mainly cratonic–epicratonic mature quartz arenites with some dolomities and mafic volcanic rocks in the top part of the sequence. The Archaean basement is dominated by 2.85-2.70 Ga TTG gneisses enclosing locally minor inclusions of komatiite, basalt, wacke and iron formation.

However, the mantle unit of the JOC is somewhat atypical for ophiolites as it seems to consist mainly of subcontinental lithospheric mantle peridotites (Peltonen et al. 1998). In support for a SCLM origin the Jormua peridotites are compositionally closer to ancient Archaean than young abyssal peridotites (Fig. 51) and consistently carry Re-Os evidence of melting from as far back as 3.0 Ga ago (Tsuru et al. 2000). It is to be noted also that pyroxenitic dykes (Fig. 49A), which are common along with hornblendite dykes in the mantle peridotites of the western block of the JOC, and OIB-type (pervasively altered ultramafic lamprophyre) dykes in the central block (Fig. 49B), both seem to have crystallised possibly already at 2.1 Ga and contain inherited in-mantle-generated Archaean zircon crystals (Peltonen et al. 2003), while the gabbros and plagiogranites crystallised at c. 1.95 Ga (Fig. 52), mostly preceding the broadly coeval sheeted dyke and extrusive units. These features together with the OIB to E-MORB character of the mafic units suggest that the JOC probably formed within an ocean–continent transition zone, similar to that found in modern magma-poor passive margins such as the present West Iberian margin, where old subcontinental

lithospheric mantle becomes unroofed along low-angle detachment faults and intruded by basaltic melts from an asthenospheric source (Cornen et al. 1999, Withmarsh et al. 2001).

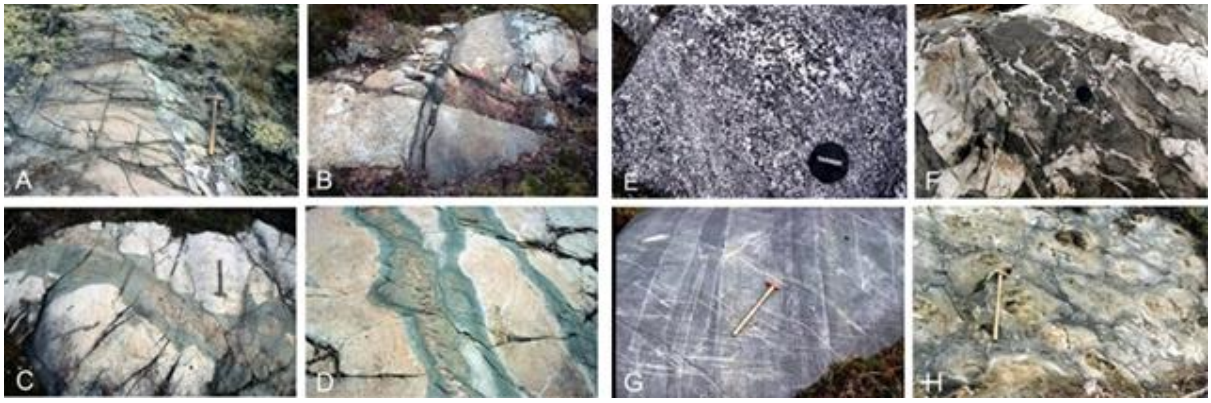


Figure 49. Photographs of main rock types in the Jormua Ophiolite Complex. (A) Clinpyroxenite dyke in metaserpentinitised mantle tectonite of the western (Hannusranta) block of the JOC, (B) OIB-type early dyke in metaserpentinised chromite knobby mantle tectonite of the central (Lehmivaara) block, (C) Gabbro–gabbro pegmatite dyke in metaserpentinised mantle tectonite of the eastern (Antinmäki) block, (D) Deep basaltic dykes in metaserpentinised mantle tectonite of the central block, (E) Varied-textured gabbro of the Sarvikangas gabbro–Fe-gabbro–plagiogranite stock in the eastern block, (F) Plagiogranite veins in microdiorite of the Sarvikangas gabbro–Fe-gabbro–plagiogranite stock in the eastern block, (G) Sheeted dykes in the Sammakolampi sheeted dyke unit of the central block, (H) Pillow basalt of the Kylmä tectonic slice of the eastern block.

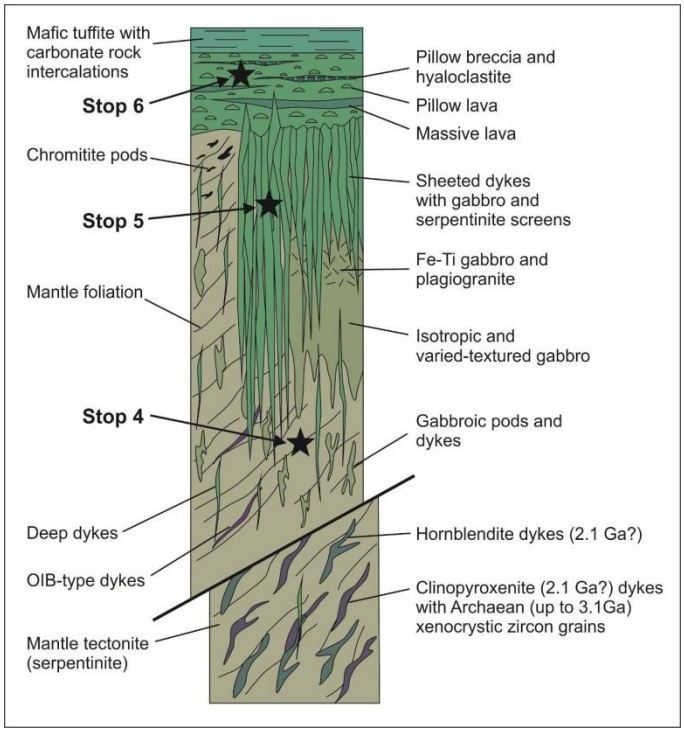


Figure 50. Reconstructed original pseudostratigraphy of the now dismembered Jormua Ophiolite Complex. Approximate stratigraphic positions of the stops 4, 5 and 6 of the present excursion are indicated in the sketch.

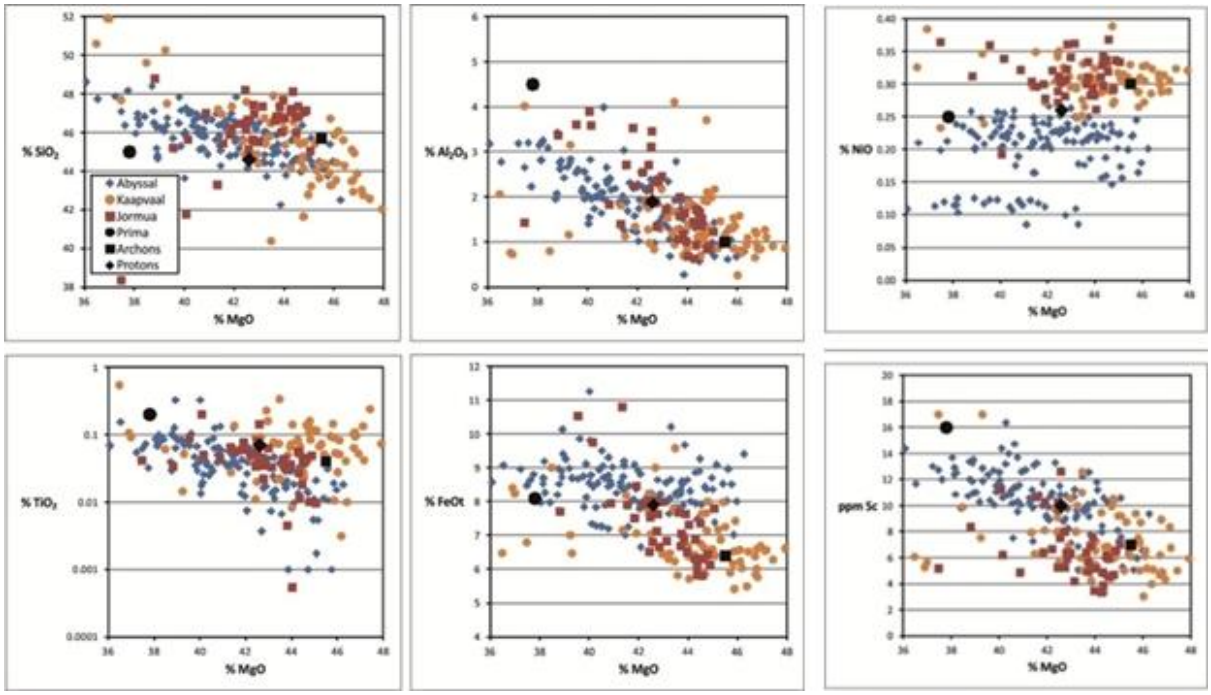


Figure 51. Variation of selected major and trace elements vs. MgO in Jormua serpentinites, which seem to have closer compositional matching with Kapvaal Archon SCLM xenoliths (data provided by Wolfgang Maier, pers. comm. 2012) than abyssal peridotites (data from Niu 2004). Compositions of primitive mantle, average Archon and Proton SCLM are from McDonough and Sun (1995) and Griffin et al. (2003), respectively. All data plotted similarly normalised to 100% on an anhydrous basis.

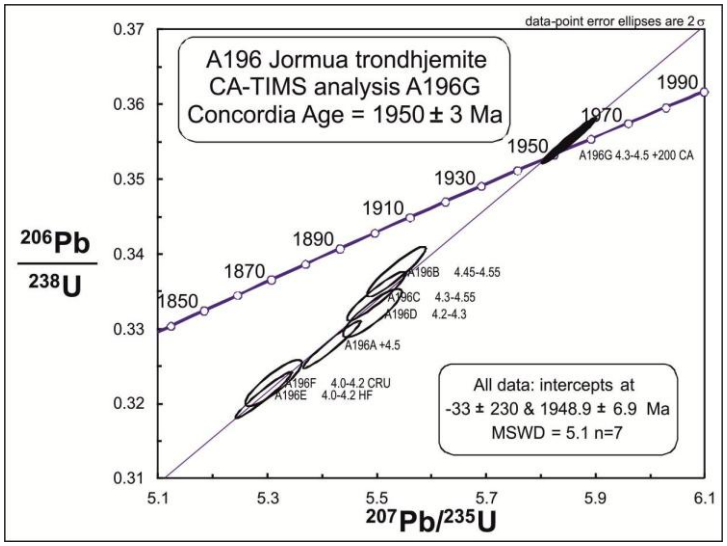


Figure 52. U-Pb concordia diagram of isotopic analyses for zircon grains from the Jormua plagiogranite sample A196 (data provided by Hannu Huhma). The concordant CA-TIMS (chemical abrasion TIMS) result at 1950±3 Ma may be considered a good estimate of the crystallisation age of this zircon rich (700 ppm Zr) leucotonalitic rock. Near identical results obtained for gabbro samples, including 1952±2 Ma for a coarse gabbro (A729) and 1953±3 Ma pegmatoid gabbro (A1402), suggest the whole mafic lid of the JOC was near coeval and aged c. 1950 Ma.

The tectonostratigraphically correlative but frequently highly peridotite-dominated Outokumpu ophiolitic massifs south of Jormua have been interpreted to represent mantle from the same passive margin environment as the JOC but from a more ocean-ward setting (Peltonen 2005, Peltonen et al. 2008). The more residual character of the Outokumpu peridotites, high-Mg and low-Ti nature of the rare mafic lavas and the presence of Cu-rich sulphide mineralisation all suggest a higher heat flow environment (at least locally) for

Outokumpu than Jormua (Peltonen et al. 2008). The Outokumpu massifs may represent oceanic mantle and gabbros from ridge-axis discontinuities or leaky transforms, which are particularly favorable sites for both peridotite exposure and enhanced hydrothermal activity.

The initially covering, now enclosing host “Upper Kaleva” c. 1.95–1.92 Ga turbidites of the Jormua and Outokumpu massifs comprise extremely well mixed material from Archaean and Proterozoic felsic plutonic dominated sources (Kontinen and Ward 1991, Lahtinen et al. 2009). The extremely thorough provenance mixing and total absence of igneous intercalations in the host turbidites suggest they deposited in a magma-tectonically passive setting (Kontinen and Ward 1991), which is reinforcing the idea of origin of the Jormua and Outokumpu massifs as fragments of incipient oceanic crust formed in a Red Sea type setting and their nature as a “passive margin ophiolites”. Soon after its c. 1.95-1.92 Ga burial below the Kalevian turbidites the Jormua-Outokumpu passive margin seafloor became obducted onto the Karelian craton margin, probably as a corollary to the presumably c. 1.90 Ga collision of the Svecofennian terrain to the craton margin.

Excursion stops at Jormua

Our visit to the Jormua Ophiolite Complex comprises three stops (Stops 4, 5 and 6) representing the most diagnostic rock units of the Jormua complex in terms of the ophiolite interpretation. Brief descriptions of the stops are given below. All the three stops are roadside targets and no special clothing such as rubber boots is needed. Walking required per stop is at all stops less than 300 m.

Stop 4. Serpentinised harzburgitic mantle peridotite (Fig. 53), Jormua, Shell Kontiomäki (N7135914, E3550586).

Lherzolitic–harzburgitic serpentinite is the dominant rock type within the Jormua mantle blocks (Peltonen et al. 1998). Stop 4 shows a variant of these rocks particularly well-preserved for its pseudomorphic peridotite structure. Pale, slightly elevated knobs characterising the weathered outcrop surface are pseudomorphs (bastite largely recrystallised to antigorite) after orthopyroxene, whereas the intervening “groundmass” of somewhat darker serpentine plus magnetite dust is after olivine (Fig. 53). The coarser oxide grains (black), which comprise predominantly ferrichromite and Cr-magnetite replacing chromite, which is preserved in rare relic cores, define a weak foliation, which is of probable mantle origin. The serpentinite is cut by very coarse-grained, rodingitized metagabbro dyke in which clinopyroxene is partly preserved but plagioclase thoroughly replaced mainly by epidote-group minerals and grossular garnet.

Serpentine both in the orthopyroxene pseudomorphs and after the intervening olivine is mainly lath to felty-textured prograde antigorite. Sporadic patches of serpentine with magnetite dust mesh typical for low-T serpentinisation (lizardite-chrysotile) of olivine, but now extensively replaced by the prograde antigorite, are observed under the microscope. This suggests, together with the metarodingitic nature of the cross-cutting gabbro dyke, that the prograde metamorphism to the present antigorite-dominated rock was preceded by an extensive low-T serpentinisation event, which possibly occurred already in the oceanic stage of the JOC or at the latest during its obduction.

Figure 53. Pseudomorphic structure typical of the serpentinitised harzburgite–dunite at Stop 4, Jormua, Shell Kontiomäki. The pale grey "knobs" are after orthopyroxene and intervening brownish material and black grains after olivine and chromite, respectively.

The Al_2O_3 content of the Stop 4 serpentinite is 1.15 wt.% for an average sample. The rare chromites preserved in cores of the coarser oxide grains have Cr and Mg numbers around 75 and 50, respectively. These values are consistent with the texture-based interpretation that the rock had a harzburgite origin. Despite the apparent immobility of trivalent cations in the relict chromite cores, high ZnO (up to 3.6 wt.%) and MnO (up to 0.9 wt.%) contents of the core chromites, greatly exceeding the range of the pristine mantle chromites, reveal that the chromite cores are altered for their divalent cations. High ZnO (typically 0.50-2.50 wt.%), and MnO (typically 0.50-1.0 wt.%) in chromite (relic cores) is an ubiquitous feature for the Jormua and other ophiolitic serpentinites in eastern Finland, reflecting high affinity of Zn and Mn to chromite in the conditions of metaserpentinisation in the prograde lower amphibolite facies.

Stop 5. Sheeted dykes (Fig. 54), Jormua, Särkijärvi (N7137750, E3546280), c. 5 km NW from Stop 4.

Sheeted dykes are one of the more diagnostic units of ophiolite complexes, attesting to an extensional formative setting similar as in oceanic spreading ridges. Sheeted dykes are well represented in Jormua in several localities in numerous outcrops. Stop 5 shows a typical

example of well-preserved, only little strained Jormua sheeted dykes. The outcrop locates in the centre of the up to 2 km wide and 4 km long Lehmilampi dyke complex, which consists nearly 100 % of subparallel metadolerite and metabasalt dykes with only occasional narrow serpentinite and gabbro screens (septa). Several dyke generations can be observed; the younger, mostly 20-80 cm wide whole dikes intrude and split the older dykes, which have been transformed to interdyke screens of “half” (“one-way chilled”) and “marginless” dykes. The dyke contacts are generally very sharp and often with distinct chilled margins. Branching of dykes and apophyses along dyke margins are common. Most of the dykes appear subparallel but there are slight strike–attitude variations seen in that the older half and marginless dykes that tend to pinch/wedge out at low angles. Some of the dykes contain abundant plagioclase phenocrysts/xenocrysts, in some cases concentrated by magmatic flow in the dyke interiors. The occasional presence of widely separated serpentinite and gabbro screens/septa (not seen at Stop 5) in the Lehmilampi dyke complex attest in a striking way to the great magnitude of extension (from 100% peridotite to >95% sheeted dykes) during the formation of this part of the JOC and the JOC in general.

Figure 54. Sheeted dykes at Stop 5, Jormua, Särkilampi. The outcrop consists 100% of <2 m wide dyke-in-dyke lithology. Some of the relatively older dykes are marginless screens (septa), some are “half-dykes”, the youngest are whole dykes with both (chill) margins present.

Stop 6. Pillow lava (Fig. 55), Jormua, Kylmä, c. 5 km NE from Stop 5.

The Kylmä stop locates within the largest occurrence of metavolcanic rocks in the JOC consisting of a fault-bound slice with a maximum thickness of c. 400 m and length of c. 2.5 km. The c. 45-65 degrees southwest-dipping lava slice has its foot-wall against a thin slice of serpentinites and talc–carbonate rocks in the west and hanging-wallagainst Upper Kaleva black schists and metagraywackes in the east. The included basaltic sequence comprises mainly pillow lavas with minor intercalations of pillow breccias and hyaloclastites, but also massive flows or flow parts, basaltic–doleritic dykes and pyroxenite–gabbro intrusions do occur. Intercalations of terrigeneous sedimentary materials have not been observed.

The Stop 6 outcrop shows typical of many Kylmä slice outcrops tightly packed pillows with only relatively little interpillow hyaloclastite material, and in this case also a cross-cutting narrow basaltic–andesitic dyke.

Figure 55. Pillow lava and a cross-cutting c. 30–50 cm wide basaltic–andesitic dyke at Stop 6, Jormua, Kylmä. The brown colour of the left part of the outcrop reflects its more recent clearance from moss.

The Kylmä basalts have 5.3–6.9 wt% MgO, 1.02–1.48 wt% TiO_2, 430–680 ppm Cr and 90–230 ppm Ni and a clear E-MORB affinity with flat to slightly LREE chondrite-normalised REE patterns and 6.5–15 ppm Nb but only 0.5–0.9 ppm Th. Post-extrusive chemical alteration has affected the basalts to varying degrees, causing e.g. LREE loss and thereby resetting of the Sm–Nd isotope systematics to yield an “isochron” at 1.72±0.12 Ga (Peltonen et al. 1996). Most of the alteration was probably related to infiltration of the same CO_2-rich metamorphic fluids that caused extensive silica–carbonate and talc–carbonate alteration in the nearby mantle peridotites, which locally host economically feasible talc reserves.

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City of Kajaani

Fig. 56. The Kajaani Castle ruins. The castle was built between 1604-1619 by the order of the King Kaarle II Adolf. In 1660 the castle was completely renovated, however the castle was destroyed in 1716 during the Great Northern War (Iso Pohjan sota) by the Russian army. Today the excavated ruins are open to visitors and the Kajaani river valley area is appreciated as nationally important cultural site.

Kajaani was announced as a city as early as 1651. The world's only working tar channel is also in Kajaani. This was built in the 1840's and was taken off use in 1915 and opened again in 1984. At summer time there are tar-rowing demonstrations in the channel. Kajaani also arranges different cultural events including dance, plays are, poetry events.

Geology of Kajaani

The rocks of Kajaani have mostly formed of Archean basement gneisses (3-2.65 Ga) and granites of an age 1.8 Ga. In places uranium mineralizations are found in the 1.8 Ga granites which may cause increased radon content in wells and houses. The eastern and northern parts of Kajaani belong to the early Proterozoic Kainuu schist belt that is characterized by 2.5-1.90 Ga quartzites, phyllites, mica schists and gneisses and serpentinites. The quartzite areas have better survived the erosion and thus are usually located topographically higher. At Lehtovaara area there are dolomites as layers within the quartzite and mica schists which is shown as florally particularly rich vegetation.

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Visit Kajaani. *Sights and places to visit*: http://www.visitkajaani.fi

Day 3 (Sept. 24, 2012)

VI: Saarijärvi impact structure, Taivalkoski dykes and Siurua gneiss

Lauri J. Pesonen, Satu Mertanen, Jouni Vuollo,

Johanna Salminen and Ella Koljonen

VI.1. Saarijärivi impact structure

VI.1. Saarijärvi impact structure (“en route” Stop 7).

The Lake Saarijärvi structure (centered 65°17.4’N 28°23.3’E) in Taivalkoski, northern Finland is a small and highly eroded impact structure situated in the Archean basement. Petrographic analysis of drill core samples, supported by geophysical and geological data, confirm that Lake Saarijärvi, is a remnant of a deeply eroded impact structure. It is the eighth impact structure found in Finland.

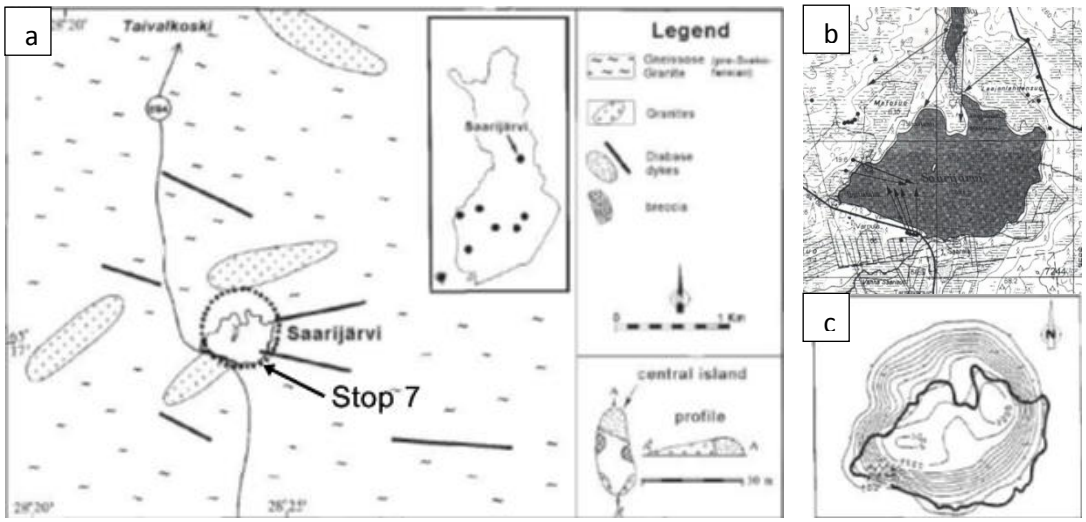

Figure 57. (a) Inset: The location and general geology of the Lake Saarijärvi impact structure in northern Finland (also shown are the ten other known impact structures of Finland). For rock types, see legend. The “en-route“ Stop 7 is shown as an arrow. Note that there is a central island in the lake where two breccia outcrops have been discovered. The cross denotes the drill site. (b) Topographic map of area. Note the huge fracture “Julmaöykky”, N of the lake. (c) Ground electromagnetic (Slingram, in-phase component) anomalies associated with the Saarijärvi structure (Pesonen et al. 1997).

The Saarijärvi impact site is visible in Landsat satellite images as a roundish structure with a diameter of ~1.5 km. The structure occupies the present lake Saarijärvi extending slightly to the north into a topographically low area. Geologically the structure lies in the Archaean basement of the Suomussalmi block. The main target rock type is tonalitic, gneissose granite rich in oligoclase. The bedrock is cut by diabase dykes of different ages as seen by high-resolution aeromagnetic maps and in outcrops including the central island (Fig. 57a). Within the structure proper, on top of the basement, there exists an up to 156 m thick sedimentary unit with varieties of sandstones and claystones. The central drilling penetrated the sedimentary pile up into fractured basement gneiss. A few centimeter thick breccia layer was found at the top of the fractured basement just below the sediment. Moreover, two anomalous mafic veins cut the fractured basement, which is below them less fractured. Two breccia outcrops were also discovered on the central island.

Geophysical observations. Strikingly circular aeroelectromagnetic and ground electromagnetic anomalies are associated with the Lake Saarijärvi structure (Fig. 57b). They

are partly related to the lake water and bottom mud of Quaternary age, but partly reflect deeper located conductors, presumably breccia layers or fractured bedrock. The aeromagnetic data show a generally weak signature over the lake, and some of the linear aeromagnetic anomalies due to mafic dykes appear to be truncated by the structure. The gravity data reveal a negative Bouguer anomaly of about -1.5 mGal over the structure (Fig.57). These geophysical data are consistent with an impact origin of the structure.

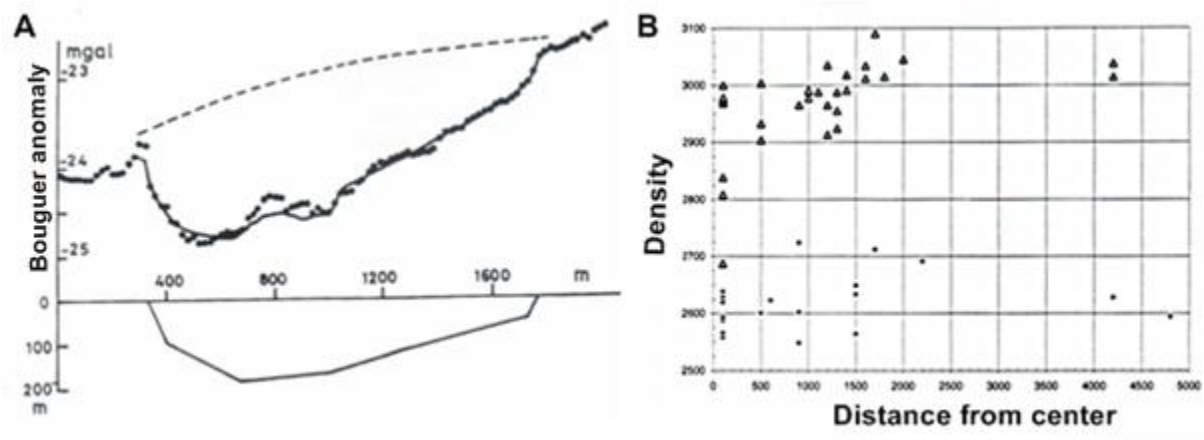

Figure 58. (A) Gravity low as measured across the Saarijärvi structure: Profile N-S. (B) Density vs. distance across the Saarijärvi structure. Note the possible radial increase of density (Öhman, 2002).

The Saarijärvi structure has been a target of investigation for more than 60 years. Three explanations for its origin have been offered: (i) a tectonically downfaulted sedimentary unit, (ii) a kimberlite pipe, or (iii) a meteorite impact. The deep drillcore showed (in 1997) that the target rock mainly consists of gneissose granitoids. Cross cutting Paleoproterozoic (most probably 2440 - 2100 Ma) diabase dykes are also typical for the area. The drilling resulted in a discovery of fragile granitic breccias with multiple sets of planar deformation features (PDF's) from the depth of 156.38 m, thus providing the impact origin for the structure. Support for the impact interpretation comes from the work by Öhman (2002) with discoveries of shatter cones (boulders but probably also in-situ), breccia boulders and sediments with kinkbands and PF's (Fig. 60). The impact interpretation is also supported by gravity, airborne and ground magnetic and electromagnetic data and by seismic reflection surveys. The age of the impact is not known but is thought to be between 2450 Ma to 600 Ma based on age of the diabase dykes and Vendian microfossils discovered from drill-core samples of sand- and claystones preserved in the crater (Öhman, 2002).

Stop 7. The bus stops shortly (15 min) at the southern shoreline of the lake Saarijärvi (Fig. 59) but due to time limitations no outcrops will be visited.

Figure 59. View over Saarijärvi impact lake. Photo: Saarijärven kaupunki

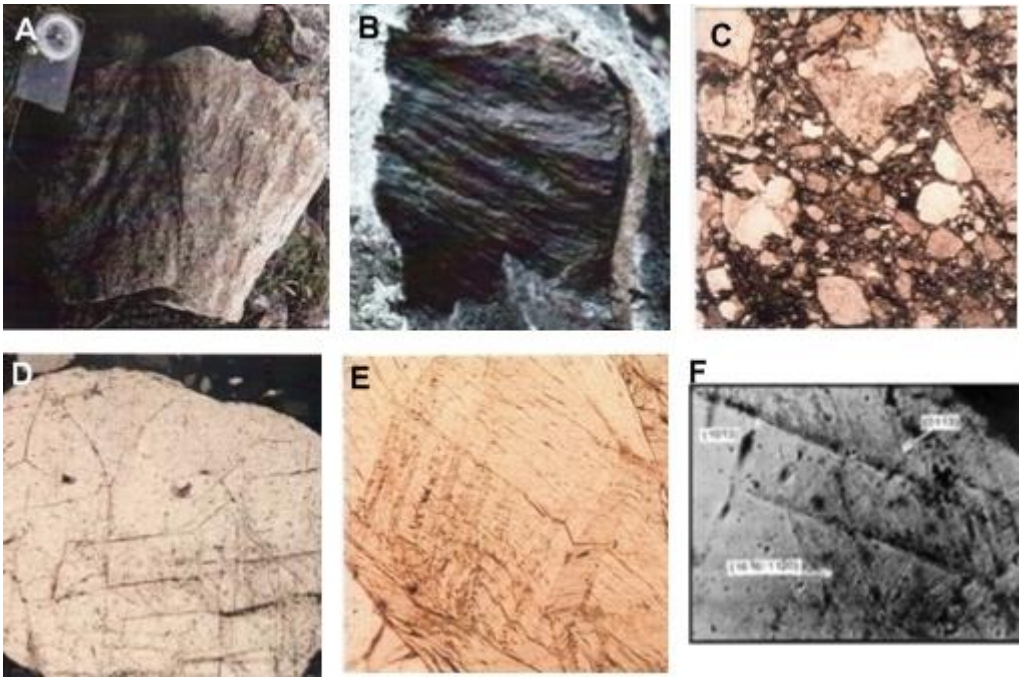

Figure 60. Shock features in rocks of the Saarijärvi impact structure, Taivalkoski. (A) Typical granitic shatter cone (boulder). The size of compass is ca. 12 cm. (B) Small in-situ shatter cone in a metadiabase. The size is same as in A., (C) Breccia structures asseen in thin section. The size of the image ~4.1mm. (D) Planar deformations in quartz clast (three to four orientation). Size of grain ca 1 mm. (E) Kinkbanding in a breccia boulder. The size of image is ~0.7 mm. (F) Multiple sets of PDFs in the breccia layer of the drill core at 156.38 m depth. See text and Öhman (2002).

Geological mapping of the area in 1999 revealed weaker developed shatter cones with the apices of the cones pointing roughly to the center of the structure. The most distant shatter cones were approximately 1 km north from the center. Therefore it could be possible that the original diameter of the crater was at least 2 km. Results from density measurements of diabases are also consistent with the hypothesis of a larger crater. New microfossil studies were also carried out, and according to them also Cambrian sediments (543-500 Ma) are present. Petrographic, electron probe microanalyzer and SEM-EDS research of brecciated granitoids from the central island showed narrow, partly glassy veins (Öhman, 2002). These

together with amphibolitic veins discovered in diabases gives the rocks of the central island resemblance to pseudotachylites found in impact structures. However, even the proposed 2 km diameter is too small for any significant melting, or to the explanation of the island as a central uplift. A tectonic explanation is more likely. Paleomagnetic results of samples from the central island (diabases, granite breccias and granites) and rocks in the NE part of the crater yielded three remanence components (Pesonen, L.J., unpublished data). The first most likely indicates the time (~2440 Ma) of intrusion of the diabase dykes. The second, ~1870 Ma, is a clear overprint of the Svecofennian orogeny. The third component is anomalous, and can be caused by the impact. This component plots slightly off of the apparent polar wander path of Fennoscandia but could have been acquired either during ~2100 Ma (reversed polarity option) or ~1200 Ma (normal polarity option), or is a late Cambrian age. All of these ages are consistent with the very deeply eroded appearance of the present Lake Saarijärvi impact structure. If the shock features in the Saarijärvi sediment (PF's and Kinkbands) are truly of shock origin a late Cambrian age (~530-500 Ma) is likely.

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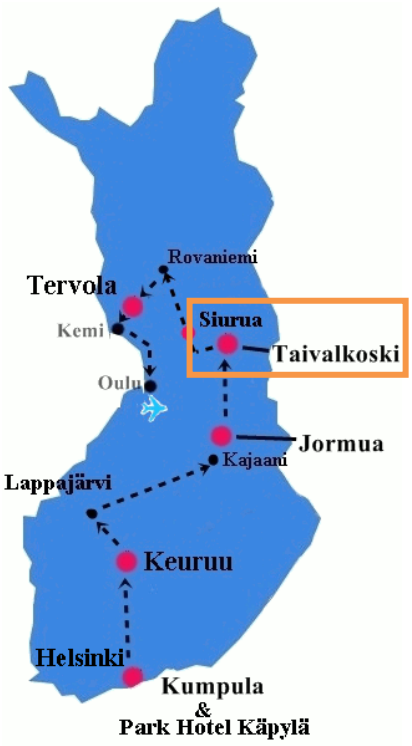


Day 3 (Sept. 24, 2012)

VI.2: Taivalkoski mafic dykes

Jouni Vuollo, Johanna Salminen, Satu Mertanen, Lauri J. Pesonen,

Pathamawan Sangchan and Robert Klein

VI.2: Taivalkoski mafic dykes

From Saarijärvi impact structure we'll drive to Taivalkoski area where at least three outcrops (Stops 8, 9 and 10) of Paleoproterozoic Taivalkoski dykes are seen. One of the outcrops is located at the Päätalo yard and we'll have coffee and tea while looking at the dyke. The other outcrops are at road cuts. Lunch will be served in the long-established 'Jalavan kauppa' (Jalava Shop) in the center of Taivalkoski.

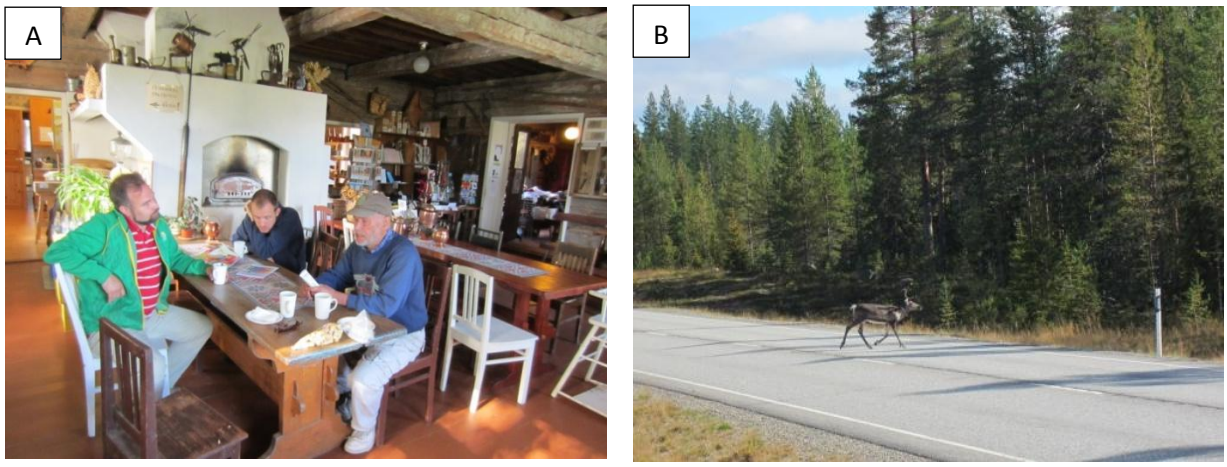


Figure 61. (A) Lauri Pesonen, Robert Klein and Harri Matikainen having coffee at Jalava Shop where we'll stop for lunch. (B) Reindeer is a local animal and can often be seen to cross the roads. Photos: S. Mertanen

Geological background

The Karelian Supergroup, which rests unconformably on the Archean basement, is divided into three groups: Sumi-Sariola, Jatuli and Kaleva. This first Paleoproterozoic continental rifting phase of the cratonized Archean basement involved emplacement of layered intrusions and dykes of ages ~2.5 Ga (Kola Province) and ~2.45 Ga (Kola and Karelian Provinces: Boninite-noritic-gabbronoritic-tholeiitic to Fe-tholeiitic dykes found in eastern and northern Finland (Fig. 2). The next rifting phase occurred at ~2.32 Ga and was represented by small intrusions and dykes (2.32 Ga). The sedimentation of the Jatuli formations was accompanied by three basic igneous events. The low-Al tholeiitic or karjalitic magmas intruded through the basement at 2.2 Ga. Most of the karjalites (gabbro-wehrlite association) occur as layered sills. The extensive set of dyke swarms cutting through the entire Archean crust and the Jatulian Group is Fe-tholeiitic. It has been dated at ~2.1 Ga and trending E–W in the Kuhmo block and pointing to pronounced extensional development. The Keivitsa intrusion (Mutanen and Huhma, 2001) in central Lapland and Otanmäki layered intrusion (Nykänen et al. 2005) in Kainuu represent the next extensional phase and at the same time there are signs of dyke swarms of age ~2.05 Ga.

At the top of the Karelian sequence are turbiditic rocks of the Kaleva Group, which are separated from the Jatulian Group by an unconformity. The 1.95 Ga slices of ophiolites at Outokumpu and Jormua (Chapter V, this volume) are encountered in the upper Kaleva. A significant sign of the break-up event, which predates the ophiolites, is the existence of the 1.98 Ga tholeiitic and Fe-tholeiitic dyke swarm intersecting the Jatulian Group supracrustal rocks (North-Karelia and Central Lapland) and the Archean craton (NW-trend; Kuhmo block), whereas there are no indications of this magmatic activity in the Kaleva turbidites.

All the Paleoproterozoic units were deformed and metamorphosed in several stages during the Svecofennian orogeny at ~ 1.8–1.9 Ga. The Archean basement was involved in these and earlier processes, as shown by the numerous basic dyke swarms cutting across it.

Paleoproterozoic dyke swarms of the Fennoscandian Shield

Dyke formation has taken place episodically over the last 3000 million years in all continents, and a considerable proportion of the continental mafic dyke swarms are either Proterozoic or Late Phanerozoic in age. Several mafic dyke swarms are also found in the eastern and northern parts of the Fennoscandian Shield (Gorbatschev *et al*. 1987; Vuollo *et al.* 1995b, 2001; Vuollo and Huhma, 2005 and references therein). Thus Finland, Russian Karelia, and the Kola Peninsula offer a healthy ground for the study of Paleoproterozoic dykes. The entire Archean craton and Karelian Supergroup up to the Kalevian Group are intersected by voluminous NW-, E-, and NE-trending dyke swarms (1-3 dykes/km^2). The first general picture of dyke swarms was provided by the map of Aro and Laitakari (1987). The new version of the Fennoscandian mafic dyke swarm map, now based on GIS-databases (Vuollo and Salmirinne, 2005) is shown in Fig. 5.

Since 1993, joint studies of Paleoproterozoic dyke swarms have been carried out at the University of Oulu, University of Toronto, Royal Ontario Museum, Canada, Geological Survey of Finland and Russian Academy of Science, Apatity and Petrozavodsk, utilizing geochronology, geochemistry and paleomagnetism (Mertanen and Pesonen, 2005; Salminen *et al*. 2011). The aim has been to identify various dyking events in the eastern part of the Fennoscandian Shield and their relationship to economically important layered intrusions and ophiolites, and at establishing the earliest part of the Proterozoic apparent polar wander path for Fennoscandia (Mertanen and Pesonen, 2005). These studies have provided valuable information on the Paleoproterozoic geological evolution of the Fennoscandian Shield and have enabled continental reconstructions to be made, e.g. in the North Atlantic area (Pesonen *et al*. 2003, Salminen *et al.* 2012). Dyke swarms studies have been done in regions of the Karelian province notably in Taivalkoski and Pudasjärvi blocks.

A map of the areal distribution of mafic dyke swarms in the eastern Fennoscandian Shield is shown in Fig. 5B. Together with previous zircon ages (summarized by Vuollo, 1994), the results indicate that there were several dyke emplacement events between 2.5 and 1.98 Ga. It is suggested that these dyke swarms and intrusions/sills can be divided into at least six main groups based on their geochemical composition, age, and mode of occurrence. These groups have approximate ages of 2.5, 2.45, 2.32, 2.2, 2.1, and 1.98 Ga (Table 2). The 2.5 Ga age-group occurs only in the Kola Peninsula, but the others are found throughout the province of Karelia. These six magmatic events are assigned here to age-groups in order to facilitate handling of the data, although each age-group may contain events of slightly different ages and involve rocks of different geochemical compositions. The names chosen for the main dyke swarms indicate their geochemical characteristics. The U-Pb and Sm-Nd ages for mafic dykes/intrusions in northern and eastern Finland gathered from the literature and from Vuollo and Huhma (2005) are summarized in Fig. 63.

The 2.45 Ga dyke swarms can be divided into five subgroups based on their field relationships and geochemical and isotopic characteristics: 1) NE-trending boninite-noritic dykes, 2) NW-trending gabbronorite dykes, 3) NW-trending tholeiitic dykes, 4) NW- and E-W–trending Fe-tholeiitic dykes, and 5) E-W-trending orthopyroxene-plagioclase phyric dykes. The younger dyke swarms (2.32 – 1.98 Ga) form a homogenous group in terms of their geochemical composition, resembling that of continental tholeiitic basalts. This means that subclassification according to their chemistry is virtually impossible.

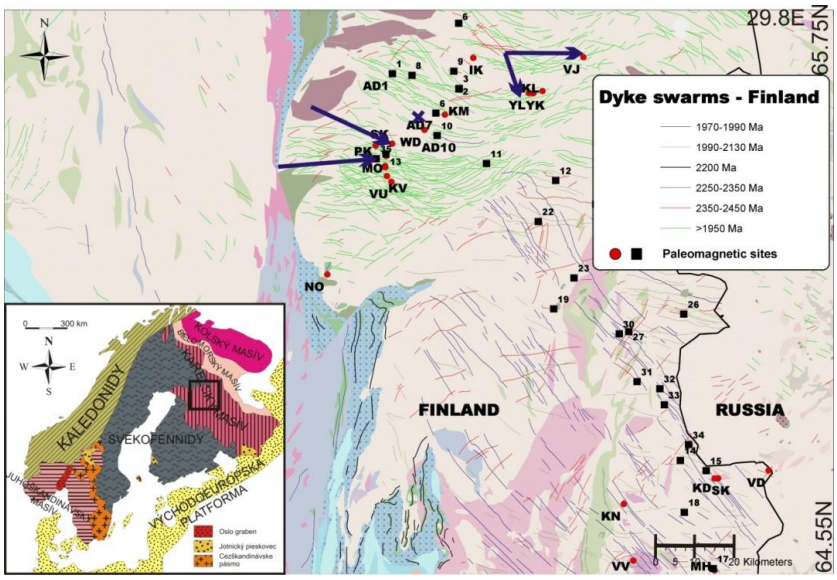


Figure 62. A map of the areal distribution of mafic dyke swarms in the Taivalkoski area. Age results of diabase dykes indicate that there were several dyke emplacement events between 2.45 and 1.98 Ga (Vuollo and Huhma, 2005). All numbered sites (black square) are part of the AD series that was originally collected by Jouni Vuollo and Henry Halls in 1993 for paleomagnetic study; remaining sites (red circles), where few duplicate to AD series are collected by Lauri Pesonen, Satu Mertanen and Jouni Vuollo. Arrows point to dykes giving other than overprinted paleomagnetic results. Taivalkoski village is marked with cross.

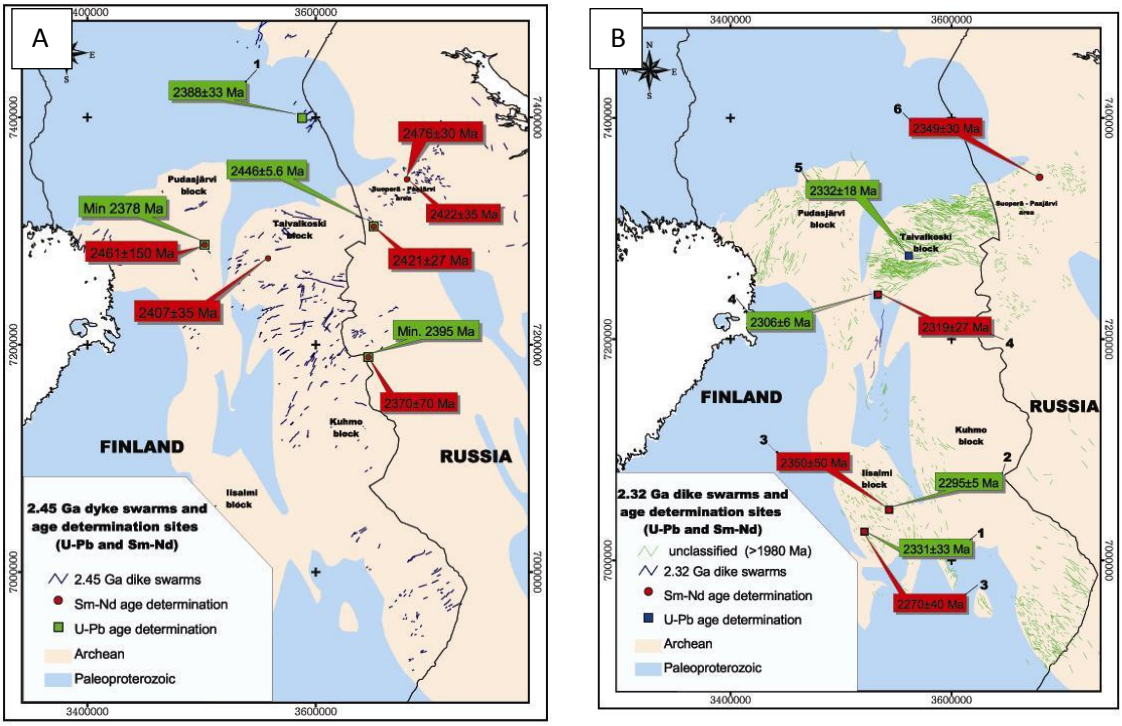


Figure 63. (A) Areal distribution and dating locations (green rectangle = U-Pb zircon or baddeleyite age, red rectangle = Sm-Nd age) of the ~2.45 Ga dyke swarms in the eastern and northern Fennoscandian Shield, Finland and Russian Karelia. Age data: Vuollo and Huhma (2005 and reference therein). (B) Areal distribution and dating locations of the ~2.32 Ga dyke swarm in the eastern and northern parts of the Fennoscandian Shield, Finland and Russian Karelia. Age data: Vuollo and Huhma (2005).

Taivalkoski mafic dykes

On the third day we will examine three outcrops of Taivalkoski mafic dykes of various ages. All age determinations published by Vuollo *et al*. (2000) are from the Archean basement area (Kuhmo, Taivalkoski and Pudasjärvi blocks and Russian Karelia). One U-Pb baddeleyite result with four mineral fractions gives an age of 2446 ± 5.6 Ma (gabbronorite – NW trend), while other U-Pb (baddeleyite) ages provide only estimates: a minimum age of 2395 Ma for a boninite-norite dyke (NE trend) and 2378 Ma for a tholeiitic dyke (NW trend). New U-Pb (baddeleyite) age datings have been carried out on some of the dykes, and the results will be published later.

Paleomagnetic study has been carried out for 50 different Paleoproterozoic dykes (Fig. 62) from Taivalkoski area. In the following, paleomagnetic results combine two studies: the first one was started by Jouni Vuollo and Henry Halls and was then followed by a Finnish team (Lauri Pesonen, Satu Mertanen, Jouni Vuollo and Johanna Salminen). Henry Halls carried out paleomagnetic studies for 36 dykes (AD series) and the Finnish team for 20 dykes, where 6 are duplicates of the AD series of Halls and Vuollo. Sampling sites were chosen according to K-Ar data that indicated minimal Svecofennian overprinting of the otherwise

pervasive 1.8-1.9 Ga Svecofennian orogeny (Kontinen et al. 1992). Despite of this selection rationale, the majority of samples show remagnetization of Svecofennian age. However, part of the area is better preserved and primary Paleoproterozoic paleomagnetic data can be obtained.

STOP 8: 2.4 Ga Fe-tholeiitic dyke (WD)

So far, only one ca. 2.4 Ga Fe-tholeiitic dyke has been dated (Sm-Nd, model age of 2407 ± 35 Ma and ε_{Nd} +1.6, Vuollo el al. 2005 and Table 2). Nevertheless, we propose that this is a new subgroup among the 2.45 Ga dyke swarms. The age determination site is situated just near the village of Taivalkoski at a road cut and will be our excursion stop. This fresh dyke is ~10 m wide and both contacts with glassy chilled margins are exposed. The modal composition of this dyke is the same as in the younger dykes; plagioclase (60%), clinopyroxene (30%), and Fe-Ti-V oxides (5–10%) with minor amounts of quartz and biotite.
Paleomagnetic data of this dyke reveal three components: ESE pointing intermediate remanence component (so called component “D”) and either a Svecofennian overprint (so called component “A”) or a late Svecofennian type overprint “B”, which can also be of Phanerozoic age (Salminen *et al*. 2012). Baked basement rock samples (migmatites) for dyke WD also show component D. The remanent magnetization of the unbaked basement rocks is weak and unstable, but clearly different from the baked basement rocks. For the dyke WD component D is separated in narrower temperature range of 520-560°C, but as wide coercivity range (30-100/160 mT) as components A and B. Because migmatitic baked basement rocks show clearer distinction between coercivity spectras and unblocking temperatures of component D and overprinted components, we interpret that baked migmatites reveal the original direction for component D (D = 116°, I = 50°, α_{95} = 3°) with a virtual geomagnetic pole (VGP) Plat = -19°N, Plon = 266°, A_{95} =4°). This differs from the remanence of the Archean basement rocks. A similar magnetization has been earlier interpreted to represent the primary 2.45 Ga old magnetization in the mafic layered intrusions and associated dykes in the Karelia Province (e.g. Mertanen et al., 1999, 2006).

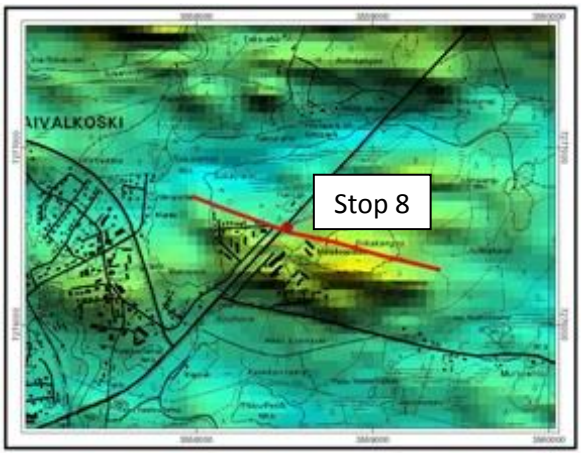


Figure 64. A. Stop 8 – Fe-tholeiitic dyke – Sm-Nd age dating site where we can see a sharp contact between 2.4 Ga diabase and Archean country rock (photo M. Melamies). B. Aeromagnetic map of Stop 8.

2.3 Ga Fe-tholeiitic dykes

Until recently, there have been only a few indications of the ~2.3 Ga magmatic events in this part of the Fennoscandian Shield. New age determinations (e.g. 2349 ± 30 Ma and 2332 ± 18 Ma, Vuollo *et al*., 2000 and Vuollo and Huhma 2005; 2306 ± 6 Ma and 2319 ± 27 Ma, Nykänen *et al*., pers. comm., 2003; 2295 ± 5 Ma Paavola *et al*., pers. comm., 2003) together with the previous results clearly indicate that the ~2.32 Ga magmatism represents a significant magmatic event in the eastern part of the Shield. Figure 63 shows the age determination sites reached so far. Four Sm-Nd results are mainly (2349–2270 Ma) just over 2.3 Ga within errors and U-Pb ages are approximately 2300 Ma (intrusions) and 2320 Ma (dykes).

The areal distribution of the 2.32 Ga dykes and intrusions is currently difficult to estimate, because we have only a few age determinations and geochemically these igneous rocks are inconspicuous (see geochemistry pictures). The new age detereminations will hopefully shed new light on these questions, too. The dykes consist of plagioclase (50%), clinopyroxene (30–40%), quartz (5%), Fe-Ti-V oxides (5%), and minor amounts of olivine (< 2%), biotite, apatite, and epidote. Aeromagnetic maps and field observations show that the trend of the dyke swarm is roughly E (95°) and that the dykes are broken and can be traced only for a few kilometers.

STOP 9: 2.32 Ga Fe-tholeiitic dyke (Museum of Kalle Päätalo)

At Kallioniemi Museum (Our Stop 9) we will see a nice outcrop of well-preserved diabase with plagioclase and clinopyroxene (Fig. 65). Preliminary Sm-Nd analyses (2352±62 Ma) from the Kallioniemi site is available and the results show that these dykes are nearly of the same age as obtained from U-Pb baddeleyite analyses from another dyke, AD10 (Fig. 66).

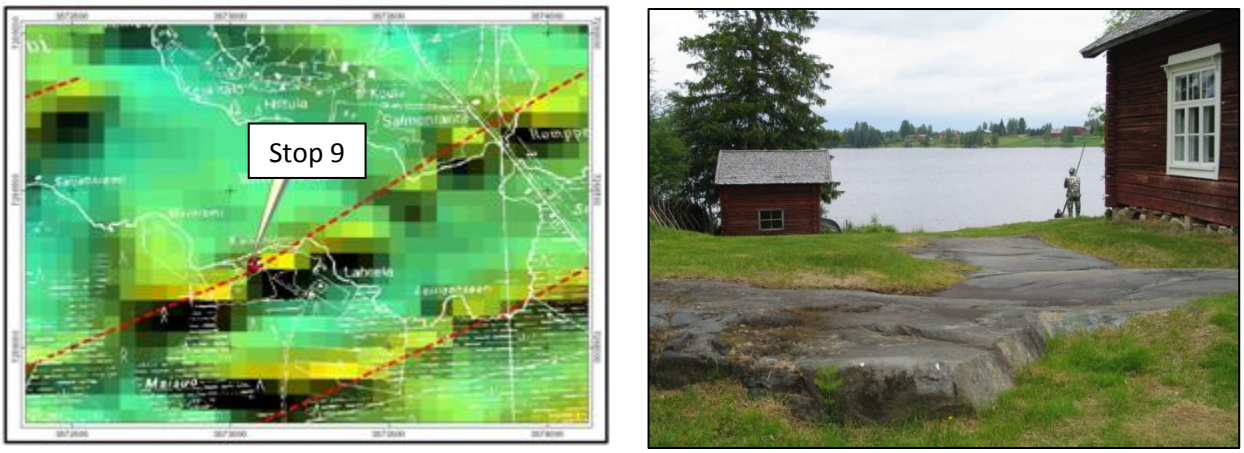
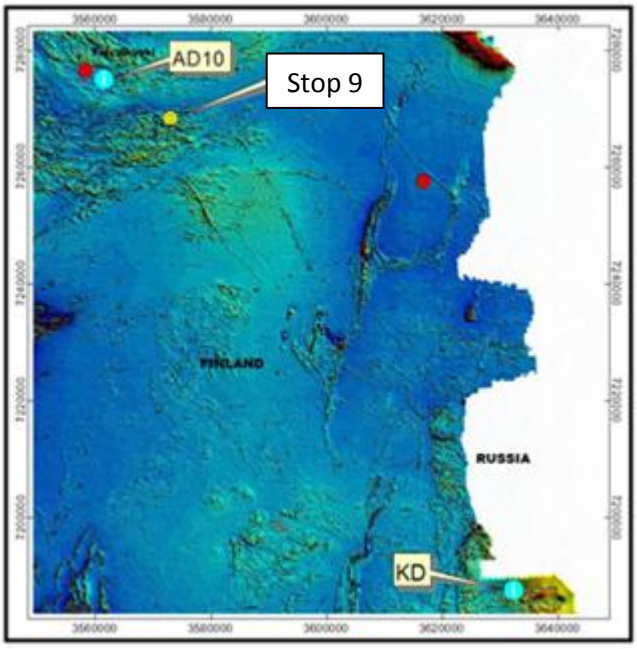


Figure 65. a) Kalle Päätalo's house and diabase dyke. B):The Fe-tholeiitic dyke (red line, Stop 9) in aeromagnetic map. Base map (c) National land survey of Finland, licence No 13/MYY/05.

Geochemically this dyke is very similar to all Fe-tholeiitic dyke swarms. Near Lake Jokijärvi, at least five c. 1 – 2 km long dykes can be seen and their trend is c. 35°. So far, contacts with the country rocks have not been observed. Paleomagnetic analyses of this dyke show only component A (Svecofennian overprint).

The Kallioniemi (promontory) is writer Professor Kalle Päätalo's childhood home just near the coastline of Lake Jokijärvi. Kallioniemi is one of the most important tourist attractions at Taivalkoski area and more than 10'000 visitors go there yearly. We will visit the Museum, and at the farmyard where there is this nice diabase outcrop (Fig. 65). **No sampling!**



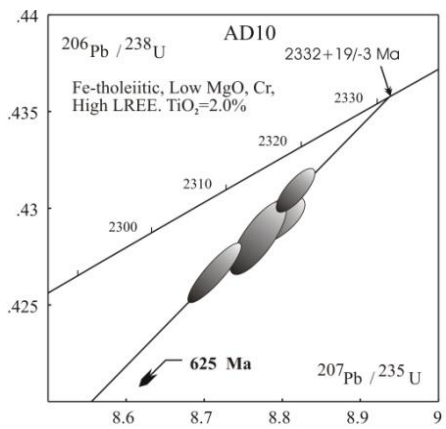


Figure 66. A) Aeromagnetic map showing location of Stop 9 and AD10. B) U-Pb dating of dyke AD10. Base map © National land survey Finland, licence No 13/MYY/05.

STOP 10: Fe-tholeiitic diabase (Majavanoja, AD13)

Stop 10 is a road cut where a Fe-tholeiitic diabase (width of 38 m) with trend of 275º is clearly observable. Primary mineralogy is well preserved – plagioclase and clinopyroxene. The aeromagnetic anomaly is weaker than at the previous stop and is seen just 1-2 km away (Fig. 67). A new U-Pb baddeleyite age of the dyke will be published later.

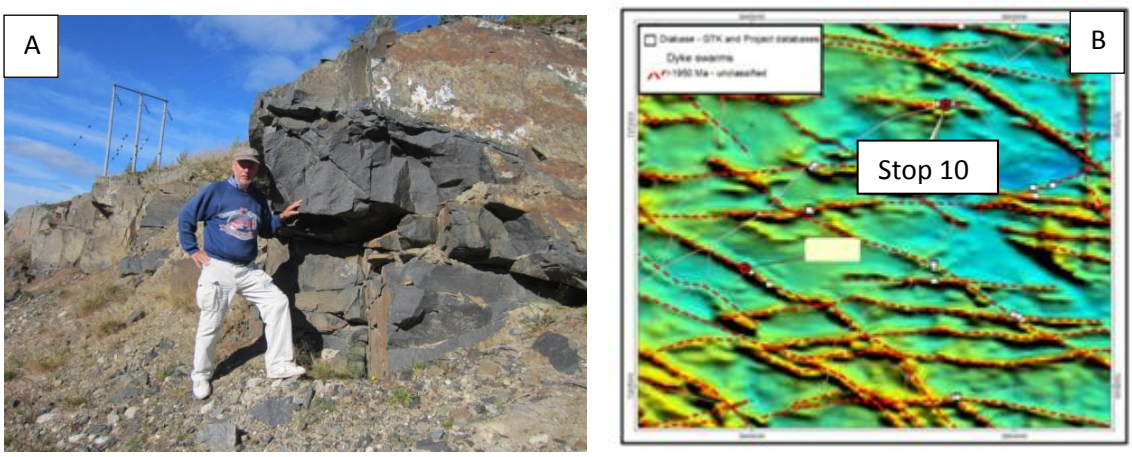


Figure 67.A) Lauri Pesonen at dyke Majavanoja, Stop 10. B) Aeromagnetic map with different trends of dyke swarms.

Paleomagnetic poles obtained from the Taivalkoski area are shown in Figure 68. From the baked host of the dyke we obtained a shallow SE pointing component called component "D'" (D=125°, I=-3°, α_{95}=5°, number of samples 5) while the dyke itself carries secondary components A and B. D´ component is similar to a component separated in ca. 2300 Ma gabbro in the Karelian Province. Previously it has been observed also at the Pääjärvi and Vodlozero areas (Mertanen et al., 1999; 2006). This component yields the VGP of Plat = 15°, and Plon = 266° with A_{95} = 5° and its relative location compared to other Precambrian poles of Baltica suggests a magnetization age younger than 2.45 Ga but older than 2.12 Ga.

According to present and previous paleomagnetic interpretations, Karelia occupied equatorial latitudes at 2.5 Ga, moved on to the intermediate latitudes at 2.45 Ga, and back to the equator when it acquired the magnetization component D' (Fig. 69). At 2.3-2.1 Ga Karelia was located on latitudes of 20-25°. The paleomagnetic data from Karelia compared to similar-aged paleomagnetic data from the Superior Province (Fig. 69) yields a looser fit of the Karelian and Superior cratons compared to a recently proposed configuration, based upon dyke swarm trajectories (Bleeker and Ernst, 2006; Bleeker, 2006). However, as discussed in Mertanen et al. (2006) and Salminen et al., (2010b), the 2.45 Ga data for

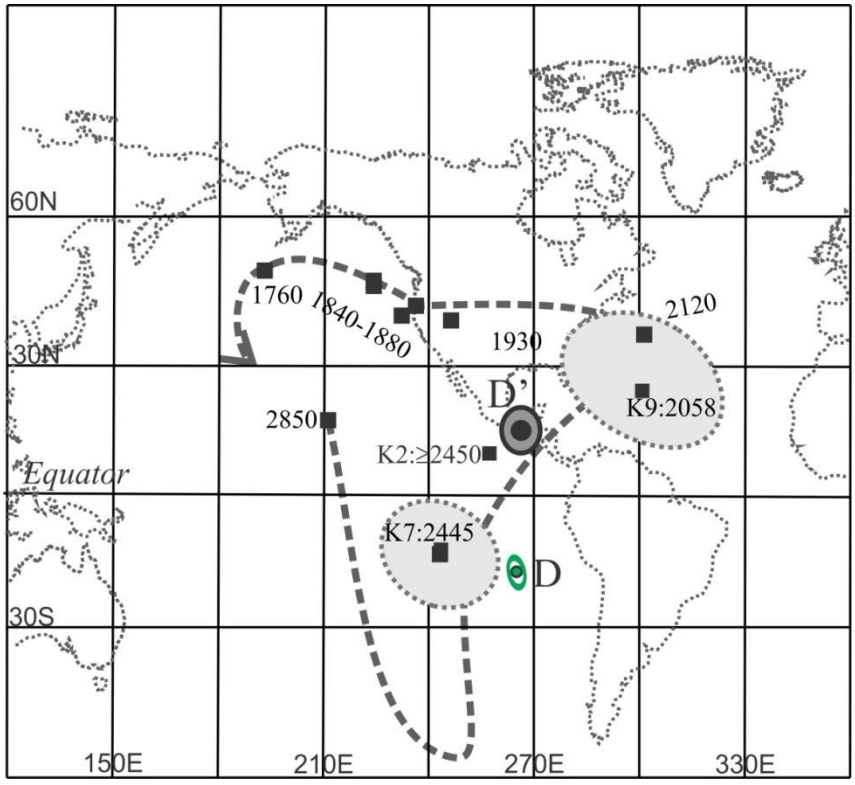


Figure 68. Poles D and D' from Taivalkoski dykes and their 95% confidence circles plotted on Precambrian apparent polar wander path (apwp) of Baltica based on keypoles (Buchan et al. 2000).

Karelia is problematic because of paleomagnetic overprints mostly during Svecofennian orogeny at ca. 1.9-1.8 Ga.

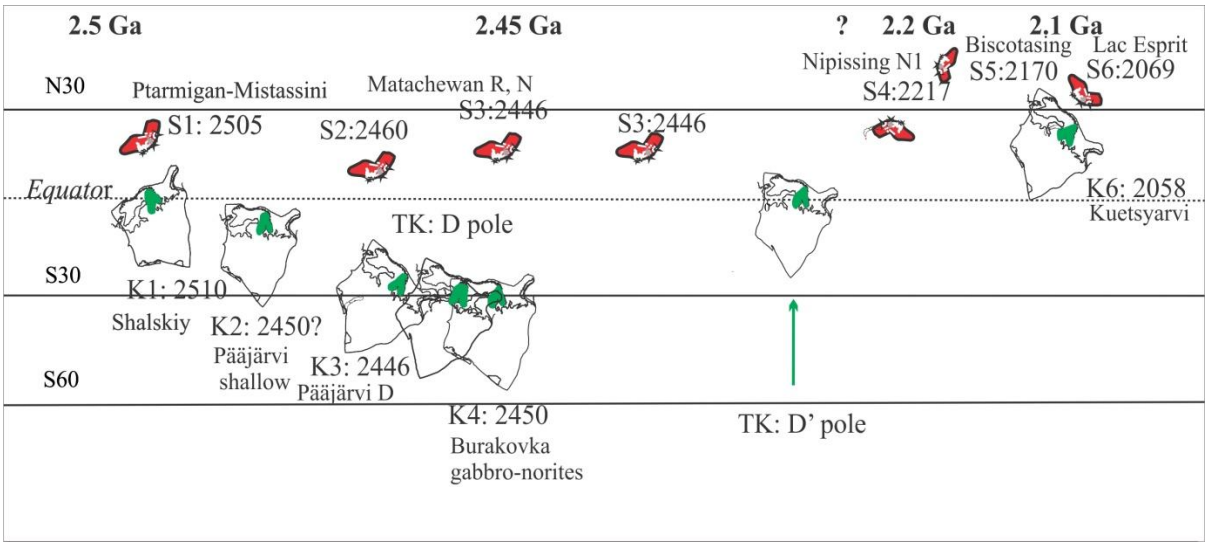


Figure 69. Cartoon showing latitudinal drifts of Superior (red) and Kola and Karelia (green) within Baltica from 2.5 to 2.1 Ga. Numbers denote ages in Ma. K1-K5: see Mertanen et al. (2006). K6: Torsvik and Meert (1995). K7: Pesonen et al. 2003. Age for Shalskiy dyke (K1) is from Bleeker, (2008). S1-S7: see Evans and Halls (2010). Note: Time scale shown across top is not in scale. Superior is drawn twice with pole S3 in order to compare with various 2.45 Ga data from Karelia.

These three stops demonstrate the different types, trends and ages of Fe-tholeiitic diabases. Some U-Pb and Sm-Nd age data are now available from most sites (Table 1).

Table 1. U-Pb and Sm-Nd age determinations and geographical orientations (Trd) of mafic dyke swarms in Taivalkoski and Pudasjärvi blocks.

	U-Pb				Sm-Nd			
Block	Taivalkoski	Trd	Pudasjärvi	Trd	Taivalkoski	εNd	Trd	
1.1. Bon-norite	-	60°	-	280°	-		60°	
1.2. Gabbronorite	-	-	-	-	-		-	
1.3. Tholeiite	-	-	2378	330°	-		-	
1.4. Fe-tholeiite			**-**	**-**	**2407 ± 35**	**+1.6**	**285°**	**Stop 8**
2. Fe-tholeiite	2332 ± 18	278°	-	-	**2352 ± 62**		**35°**	**Stop 9**
	2306 ± 6	340°			2319 ± 27	+1.8	340°	
3. Fe-tholeiite	-	-	2118 ± 14	300°	-		-	
4. Fe-tholeiite	-	-		350°?	-		-	

Other dykes

Vuollo et al. (2005 and references therein) give a comprehensive description of other dykes (e.g.gabbro noritic, boninite and 2.1 Ga Fe-tholeitic dykes) which are not visited during this excursion.

Geochemical and isotopic characteristics of Paleoproterozoic dyke swarms are also summarized in numerous geochemical discrimination plots and will not be repealed here with the exception of the major and some trace element analyses of different dyke swarms (from Kuhmo and Taivalkoski blocks and Russian Karelia) which are presented in Figs 70-71. One distinctive feature of the 2.45-Ga dyke swarm compared with younger dyke swarms, is a wider range of geochemical compositions.

Geochemical and isotopic characteristics

A Ti/V diagram clearly shows that the 2.45 Ga dykes (both BN and TH) are distinct from the Fe-tholeiites. They show an island arc (IAT) signature, whereas the Fe-rich tholeiites fall in the within-plate basalt field (WPB). All Fe-tholeiitic dykes fall into the continental flood basalt field in the Al_2O_3/TiO_2 vs. Ti/Zr diagram. The BN dykes form a more diverse group plotting on both sides of the boundary between the boninite and volcanic arc basalt field, whereas the TH dykes fall in the mid-ocean ridge basalt (MORB) field.

Figures 70-71 show selected analyses from the dykes of Kuhmo and Taivalkoski blocks and Russian Karelia. It is evident that the eastern Fennoscandian Paleoproterozoic dykes are

compositionally similar to the dyke swarms in other shields (Condie, 1997). The BN dykes are mainly in the noritic field and tholeiitic and Fe-tholeiitic dykes (2.45–1.98 Ga) plot in the tholeiitic field regardless of their age (2.45–1.98 Ga).

Chondrite-normalized REE patterns for various dyke swarms are shown in Figure 71. The REE patterns for Fe-tholeiitic dyke swarms (2.45–1.98 Ga) are homogenous showing moderate LREE enrichment [(La/Yb)N ~2–3)]. The BN dykes usually have relatively high LREE/HREE ratios, with (La/Yb)N of 3–9 for boninite–norites, 4–5 for gabbro norite dykes, and 7–11 for orthopyroxene-plagioclase-phyric dykes. The TH dykes are usually slightly LREE-enriched but may also show flat or LREE-depleted patterns (Fig. 71).

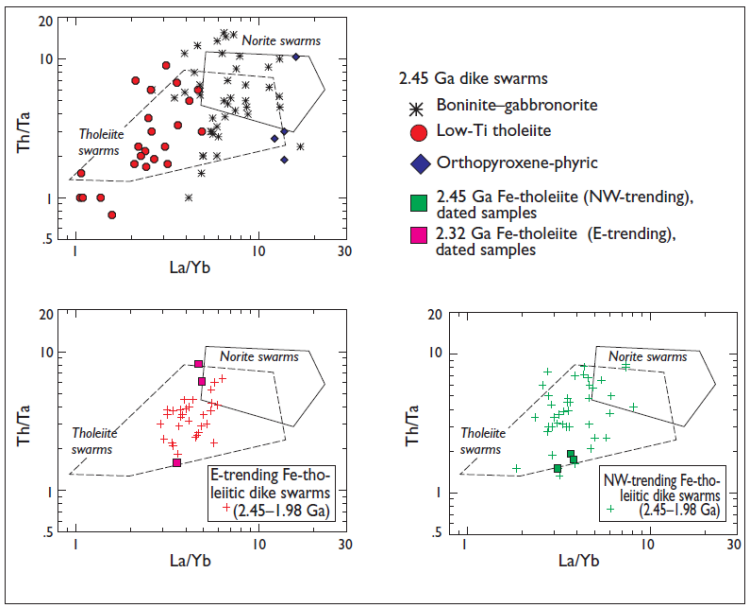


Figure 70. Selected chemical analyses (158 samples) of the Paleoproterozoic dyke swarms (2.45 Ga, 2.32 Ga, 2.1 Ga, and 1.98 Ga) from the Kuhmo block and Russian Karelia plotted on the La/Yb vs. Th/Ta diagram (Condie 1997).

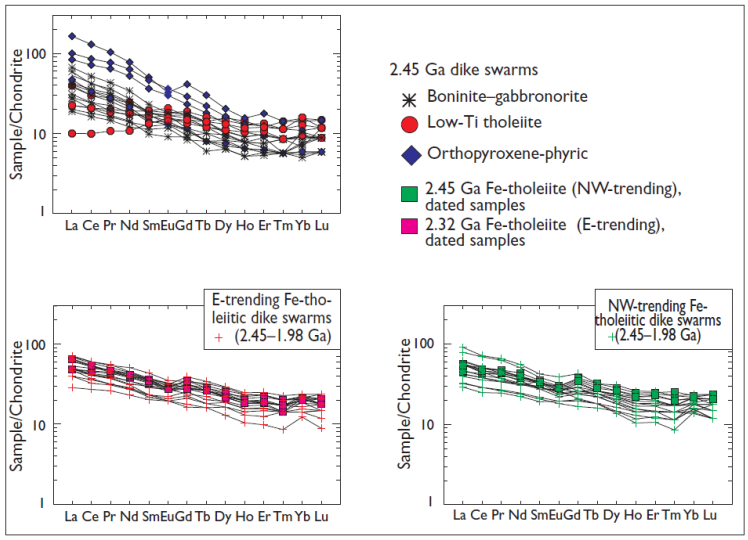


Figure 71. Examples of chondrite-normalized REE patterns for 2.45 Ga, 2.32 Ga, 2.1 Ga, and 1.98 Ga dykes from the Kuhmo block and Russian Karelia.

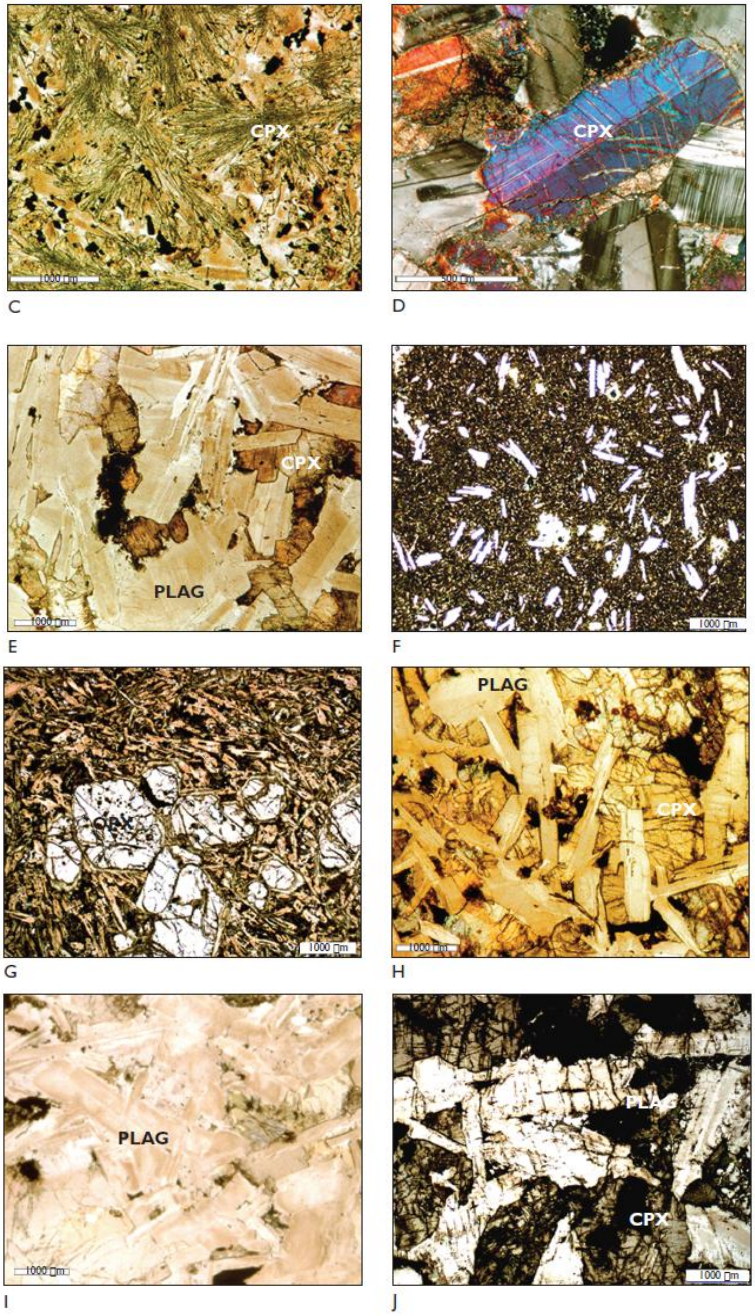


Figure 72. Photomicrographs of rock types from Paleoproterozoic dyke swarms. 2.45 Ga: (C) Gabbronorite dyke, sample near the contact, quench-texture (sample XD20); (D) Gabbronorite dike, clinopyroxene grain with pigeonite exsolution lamellae, fish-bone texture (sample XD6), crossed nicols; (E) Tholeiitic dyke with a cumulus texture (sample 1-UD-93); (F) Fine-grained Fe-tholeiitic dyke (sample WD9); (G) Orthopyroxene-phyric dyke (sample 42-VEN-94). 2.32 Ga: (H) Fe-tholeiitic dyke with cloudy feldspar (sample XD17). ~2.1 Ga: (I) Fe-tholeiitic dyke with faintly cloudy feldspar (sample VEPO28–12.55). 1.98 Ga: (J) Fe-tholeiitic dyke with faintly cloudy feldspar (sample KD12).

Abbreviations:PLAG–plagioclase, OLIV–olivine,CPX–clinopyroxene, OPX–orthopyroxene. If not mentioned, the scale bar is 500 µm in length. Photos: Jouni Vuollo.

Geochemically, the Fe-tholeiitic dykes form a relatively homogenous group from North Karelia through Kainuu to Lapland (Vuollo *et al*. 2005). They are quartz-normative, sub-alkaline tholeiitic basalts and form a set of continental dyke swarms of the type frequently found in shield area.

Dyke database work

The first general account map of mafic dyke swarms was provided by Aro and Laitakari (1987). A new version of the Fennoscandian dyke swarm map now builds on GIS databases (Vuollo *et al*. 2001). Figure 5B shows the newest version of the Eastern Fennoscandian Shiel mafic dyke swarms map compiled from current GIS database.

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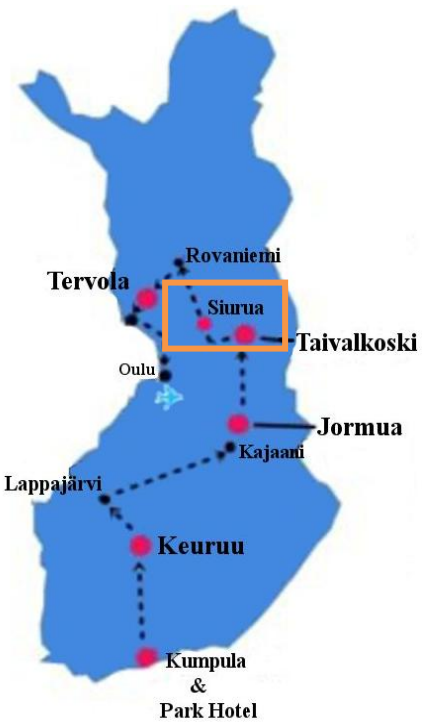

Day 3 (Sept. 24, 2012)

VI.3. Siurua gneiss - One of the oldest rocks of Europe

Hannu Huhma, Katja Lalli, Lauri J. Pesonen,

Pathamawan Sangchan, Satu Mertanen and Ella Koljonen

Part VI: Siurua trondhjemite gneiss

Hannu Huhma, Katja Lalli, Lauri J. Pesonen, Pathamawan Sangchan, Satu Mertanen and Ella Koljonen

In the afternoon of the third excursion day we will shortly visit the Siurua trondhjemite gneiss. The isotopic data available reveal that the gneisses in Siurua are among the oldest rocks in Europe. The U-Pb ages on zircon obtained by SIMS and LA-MCICPMS are mostly 3.4-3.5 Ga, suggesting an igneous age of ca. 3.5 Ga. Some zircon cores register ages up to 3.73 Ga. Involvement of such old crustal material is supported by Sm-Nd analyses on whole rocks and Lu-Hf on zircon.

Most of the dated Archean rocks in Finland have ages of 2.68-2.76 Ga (57%) or 2.79-2.84 Ga (25%), and only few samples are older with some clustering seen in ages at ca. 2.86 Ga, 2.95 Ga and 3.1-3.2 Ga (Huhma et al. 2011). Tonalitic gneisses with the bulk U-Pb ages exceeding 3.1 Ga have been found in four localities in the Fennoscandian Shield: 1) Tojottamanselkä, Central Lapland (Kröner et al. 1981, Kröner & Compston 1990), 2) Lapinlahti, Eastern Finland (Paavola 1986, Hölttä et al. 2000, Lauri et al. 2011), 3) Vodlozero, Russian Karelia (Lobach-Zhuchenko et al. 1993) and 4) Siurua, Pudasjärvi, northern Finland (Mutanen & Huhma 2003, Lauri et al. 2011).

Stop 11

The Siurua tonalite belongs to the Pudasjärvi Complex (Fig. 73) consisting mainly of Archean gneisses and granitoids and abundant amphibolites. Proterozoic diabase dykes have also intruded into the Archean gneisses.

The north-south trending Pudasjärvi Granulite Belt is located in the middle of the gneiss complex. The Granulite Belt is approximately sixty kilometres long and is scattered by several fault and shear zones into smaller pieces. The metamorphic grade varies from amphibolite to granulite facies between blocks.

The Siurua tonalite gneiss (Stop 11) is the dominant rock type in a large outcrop area (Fig. 73). The rock is granoblastic, medium-grained and consists of plagioclase, quartz, potassium feldspar, dark biotite and accessory fluorapatite, zircon, monazite, ilmenite, magnetite and metamict allanite. The chemical composition is close to the average Archean TTG gneisses. The major differences are the depletion of phosphorus and strong enrichment of Th and LREE. The REE pattern of Siurua and felsic granulite of Pudasjärvi Granulite Belt is similar with negative Eu anomaly; however, the REE concentrations of the Siurua trondhjemite are about an order of magnitude higher. Gneissic banding is seen on an outcrop but penetrative deformation is not noticeable under a microscope. Gneissosity-parallel veins of granite

pegmatite and roundish boudins of mafic granulite appear on the outcrop, although the gneiss is not in its present state on granulite facies.

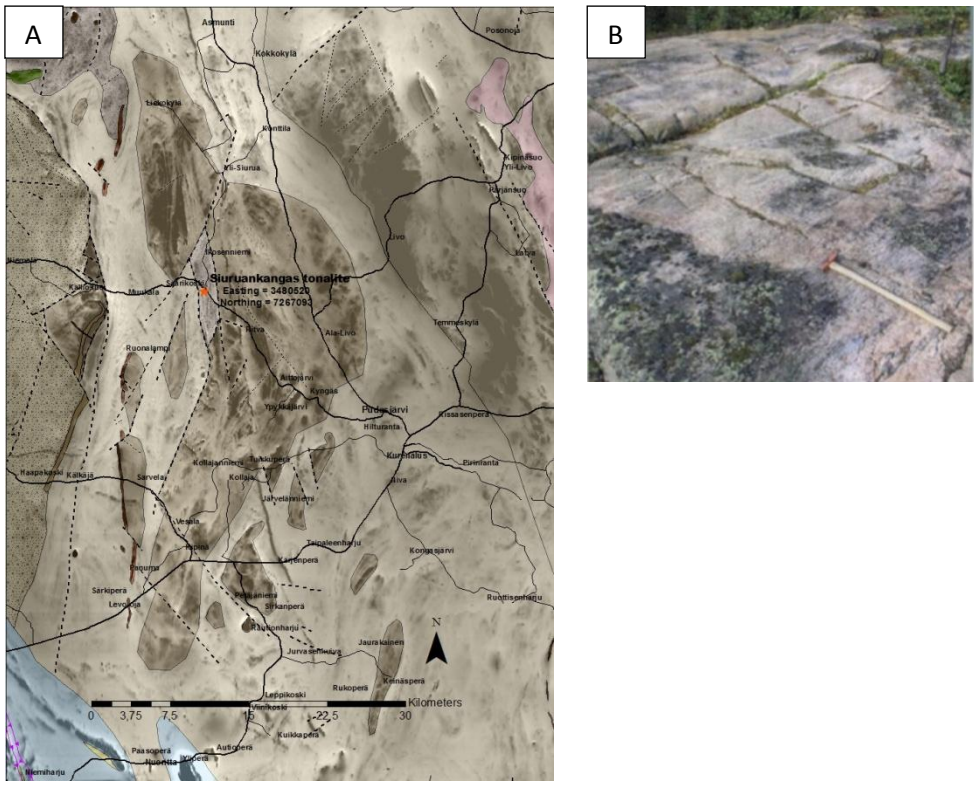

Figure 73. (A) The Pudasjärvi Granulite Belt is about 60 kilometers long and is scattered by several fault and shear zones into tectonic blocks. These blocks consist mainly of felsic enderbites (darkish brown) and are surrounded by Archean tonalities (beige). The fold structures visible in the aeromagnetic data in large scale can be seen in outcrops in smaller scale. Extract from DigiKP200 ©GTK in scale 1:250 000 (30.8.2012) with GTK aeromagnetic data. (B) Photograph of the Siurua gneiss exposure (Stop 11) from which the age dating sample was collected. Photo © Helsingin sanomat, Oct. 19, 2006.

The first U-Pb TIMS analyses on zircon showed that the Siurua gneiss is significantly older than any other rocks so far identified in the Fennoscandian Shield. This was confirmed by ion microprobe (NORDSIM) U-Pb analyses on zoned magmatic zircons, which suggested an age of ca. 3.5 Ga (Mutanen & Huhma 2003). The U-Pb SIMS data are, however, heterogeneous and suggest several stages of zircon growth, mostly at 3.5-3.4 Ga. An inherited core in one crystal provided an age of 3.73 Ga, whereas the youngest analyses yield ages of 3.1 and 3.3 Ga (Fig.74). The U-Pb results obtained by LA-MCICPMS on the same and another sample are very similar (Lauri et al. 2011).

Complex crustal evolution is further registered by metamorphic monazite, which formed in the Siurua gneiss ca. 2.66 Ga ago, roughly contemporaneously with the high-grade metamorphism recorded by zircon in a mafic granulite in the belt (Mutanen & Huhma 2003).

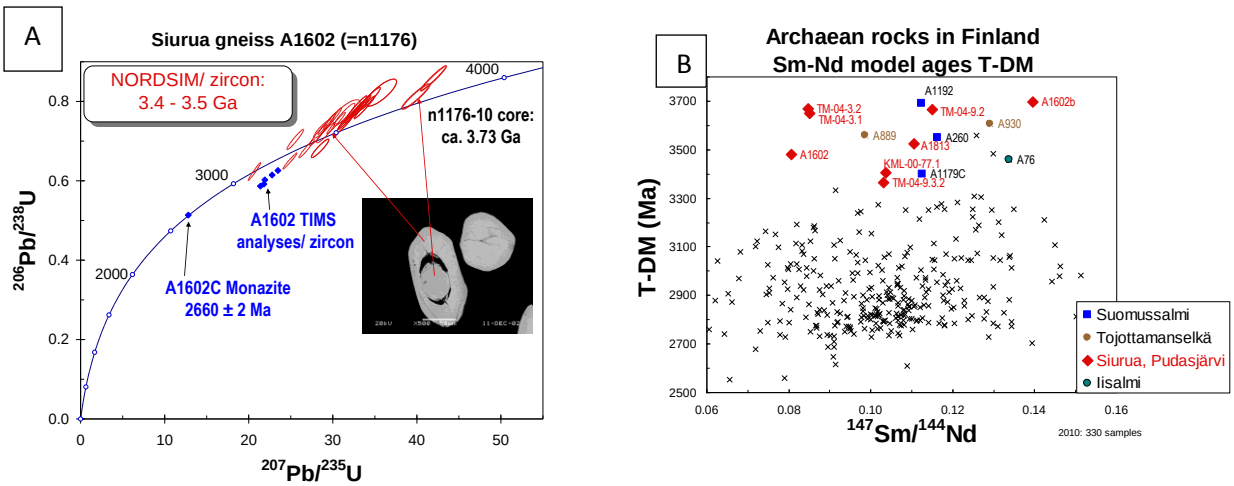

Figure 74. (A) U-Pb. U-Pb concordia plot of zircon data from the Siurua gneiss A1602 (Mutanen & Huhma 2003). (B) Nd. Sm-Nd model ages T_{DM} for Archaean whole rock samples from Finland (Huhma et al, in press).

The old age is supported by the Sm-Nd analyses on whole rocks, which give depleted mantle model ages up to 3.7 Ga (Fig. 74). Initial Hf isotope compositions on zircon also strongly support major involvement of such old crustal material in the genesis of Siurua gneisses, and unradiogenic Hf measured in few spots have been interpreted as evidence of crustal history even back to ca. 4 Ga (Lauri et al. 2011).

The Sm-Nd data on 50 whole rock samples from the Pudasjärvi Complex reveal that in addition to Siurua outcrops, Sm-Nd model ages exceeding 3.3 Ga were obtained also from migmatitic gneisses in Kolkkoaho, located 20 km north of the Siurua locality. Samples from other parts of the Pudasjärvi Complex, in contrast, yield model ages mostly between 2.7 and 3.1 Ga. In terms of Sm-Nd model ages, the Siurua gneisses are also distinct when comparing all data on Archaean areas in Finland (Fig. 74, Huhma et al., in press).

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Taivalkoski town

Taivalkoski is part of the boreal zone and the landscape is dominated by alternating forests and swamps. There is a multitude of ridges and water ways that have been shaped by the last Ice Age to align from east to west. Most beautiful examples of these ridges are Kylmäluomanharju and Loukusanharju. The biggest lakes are Tyräjärvi and Kostonjärvi, while Iijoki and Kostonjoki are the most well known rivers in the area. The fauna is composed of typical species for the northern coniferous zone; moose, reindeer, rabbits, and also bear are present. Newer species that have arrived into the area include mink, raccoon and hedgehog. In the older parts of the forests there are many species that can only live in these types of forests such as Siberian jay, three-toed woodpecker and capercaillie to name a few. Also the river Korvuanjoki that runs through the protected area has rare fresh water pearl mussels and their revival project is part of EU's Life-Nature project. The Saarijärvi impact lake described previously is also located in Taivalkoski area.


Figure 75. Kalliomäki, Päätalo's childhood home.

One of the most famous cultural sites at Taivalkoski is the Päätalo house located directly on top of an ~2.4-2.3 Ga diabase (Stop 10). Kalle Päätalo (1919-2000) was a famous writer who spent his childhood at Taivalkoski and wrote the so-called "Taivalkoski epos". The house is called Kalliomäki (rock hill) and has been renovated as closely as possible to replicate times gone by. The area is a nationally important cultural site. Every year Taivalkoski arranges a Päätalo week, with a theme dedicated to log floating and logging culture. The town of Taivalkoski was built around a saw mill at the Iijoki River at Taivalkoski rapid and it became the centre of the area at the end on 1800's. People lived mostly from agriculture and forestry, but hunting and fishing were important as well. The area grew rapidly in 1970's when Mustavaara mine was opened and housing was needed for workers.

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Day 4 (Sept. 25, 2012)

VII Paleoproterozoic Rantamaa formation with abundant stromatolites, Tervola, Northwest Finland

Juha A. Karhu, Vesa Perttunen, Nina Hendriksson, Lauri J. Pesonen,

Satu Mertanen and Robert Klein

Rantamaa formation

The target of the last excursion day, Sept. 25, 2012 is a dolomite quarry situated at Rantamaa, between Tornio and Tervola, Northwest Finland (Fig. 76). The quarry is owned by SMA Minerals Oy. The exploitation at Rantamaa started in the early 1970´s. The quarry produces crushed marble products with the trade mark Lappia Clear. Attempts to achieve dimensional stone have been unsuccessful. The company has another, large dolomite quarry, Kalkkimaa, some 10 km to the south from Rantamaa.

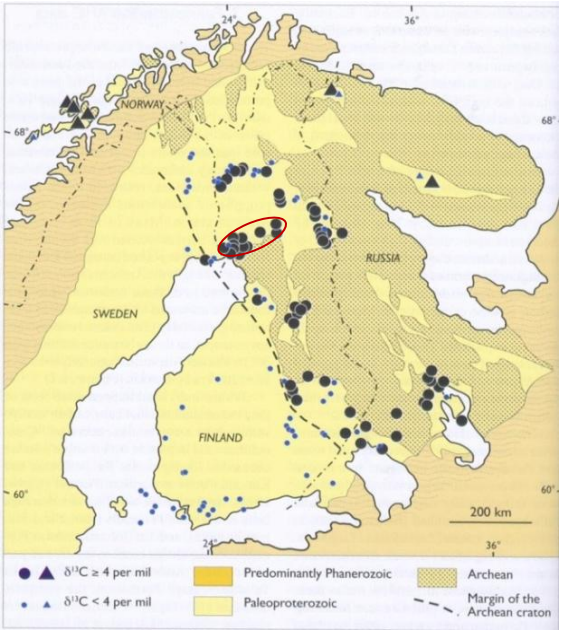

Figure 76. Areal distribution of ^{13}C-enriched, Paleoproterozoic sedimentary carbonate rocks on the Fennoscandian Shield (big circles and triangles, Karhu 2005). The location of the Peräpohja Belt is circled.

Peräpohja Belt

Geologically, the Peräpohja area (Fig. 77) can be divided into three distinct geological units: the Pudasjärvi Granite Gneiss Complex, the Peräpohja Belt, and early Proterozoic intrusions cutting the supracrustal units. The Rantamaa dolomite quarry is situated in the Rantamaa Dolomite Formation within the Paleoproterozoic Peräpohja Belt. The supracrustal rocks of the belt rest unconformably on the Archean Pudasjärvi Granite Gneiss Complex. The basement rocks include granitoids and rare remnants of greenstone belts as well as early Paleoproterozoic layered intrusions. To the north the belt is bounded by younger Paleoproterozoic plutonic rocks of the Central Lapland Granitoid Complex and in the west by the plutonic rocks of the Haaparanta Suite.

The supracrustal rocks at of the Peräpohja Belt have been metamorphosed at a low grade greenschist facies. The belt is separated by a weathering horizon from the Archean Granite Gneiss Complex. U-Pb age data for the rocks of the Peräpohja Belt have been published by Perttunen and Vaasjoki (2001).

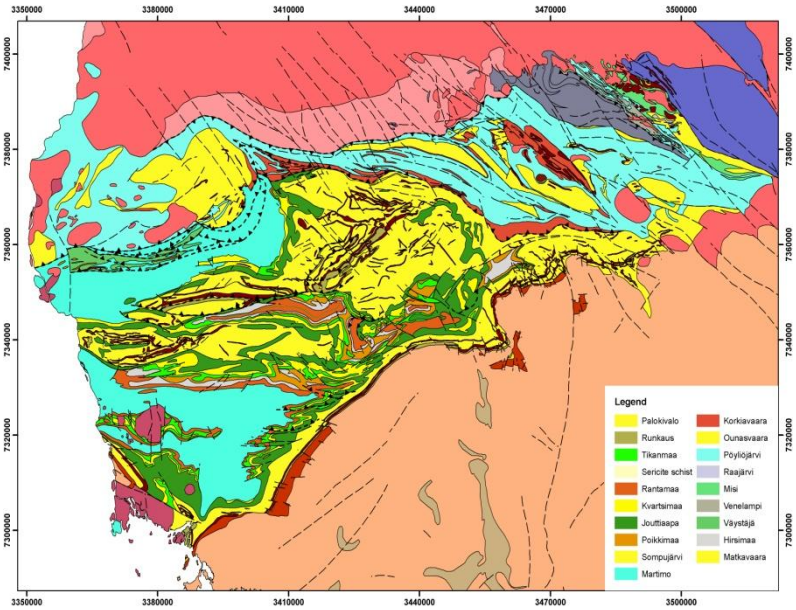

Figure 77. Geological map of the Peräpohja area, Northwestern Finland. Geological Survey of Finland.

Lithostratigraphy

The supracrustal rocks of the Peräpohja Belt are divided into the Kivalo and Paakkola Groups (Perttunen and Hanski 2003). The sedimentary rocks of the Kivalo Group form a dolomite-orthoquartzite association, intercalated by volcanic units at several stratigraphic levels. The Runkaus and the Jouttiaapa Formations are the lowermost volcanic units (Figure 77). The volcanic rocks consist mainly of subaerial amygdaloidal tholeiitic basalt flows, but the two formations have strikingly different trace element contents. The former displays the effects of crustal contamination, whereas the latter is uncontaminated and has an ultradepleted chemical and isotopic signature. Between them is a thick, wide-spread quartzitic Palokivalo Formation. The Jouttiaapa Formation is overlain by a thin Kvartsimaa Orthoquartzite Formation, with some distinct dolomitic interlayers. Most of the dolomitic rocks of the Peräpohja area belong to the Rantamaa Formation on top of the Tikanmaa Tuffitic Formation.

The Tikanmaa, Hirsimaa, and Lamulehto Formations, separated from each other by dolomitic sedimentary formations, are mafic tuffitic rocks with distinct geochemical characteristics varying from depleted tholeiites to OIB-like trachyandesites.

The rocks of the overlying Paakkola Group are divided into the sedimentary Martimo Formation and the predominantly volcanic Väystäjä Formation. The Martimo Formation consists of turbiditic mica schists and phyllites with black schist intercalations. The volcanic rocks of the Väystäjä Formation are pillowed tholeiitic basalts accompanied with minor tuffites and acid volcanic rocks.

Stromatolite structures

Stromatolites are organo-sedimentary structures formed in shallow water by micro-organisms, especially cyanobacteria (blue-green algae). Blue-green algae are primitive organisms that live in colonies, and they bind matter easily. As a result, layered structures are formed that are rounded when cut in horizontal sections and domed or columnar and possibly forked in vertical sections. Stromatolites are commonly met in Precambrian sedimentary carbonates but are more rare in more recent sediments.

The dolomite successions at Peräpohja are characterized by abundant, well preserved stromatolite structures (Perttunen, 1970). Most of the occurrences belong to the Rantamaa Formation, but distinct stromatolitic layers have also been met in the Kvartsimaa Formation (Figs. 78 and 79).

Figure 78. Stromatolites in the Kvartsimaa Formation. Kvartsimaa, Tornio. Photo V. Perttunen

Figure 79. Stromatolites in the Rantamaa Formation, Tervola. Photo V.Perttunen

Dolomite at the Rantamaa quarry represents sedimentation in an intertidal to supratidal sedimentary environment (Kortelainen, 1998). As a result of drying, teepee and rosette structures are locally present. Laminar dolomite, stromatolite domes and columns, quartzite interlayers with herring bone cross bedding and ripple marks give indications of variations in water depth and different energy environments on the carbonate platform.

Carbon isotope signature of dolomites

Paleoproterozoic sedimentary carbonates are characterized by a distinct positive carbon isotope excursion. Between 2.2 and 2.1 Ga, the $\delta^{13}C$ of carbonate sediments reached high values of 10 ‰ and even higher (Fig. 80). The excursion is global in character (Karhu and Holland, 1996).

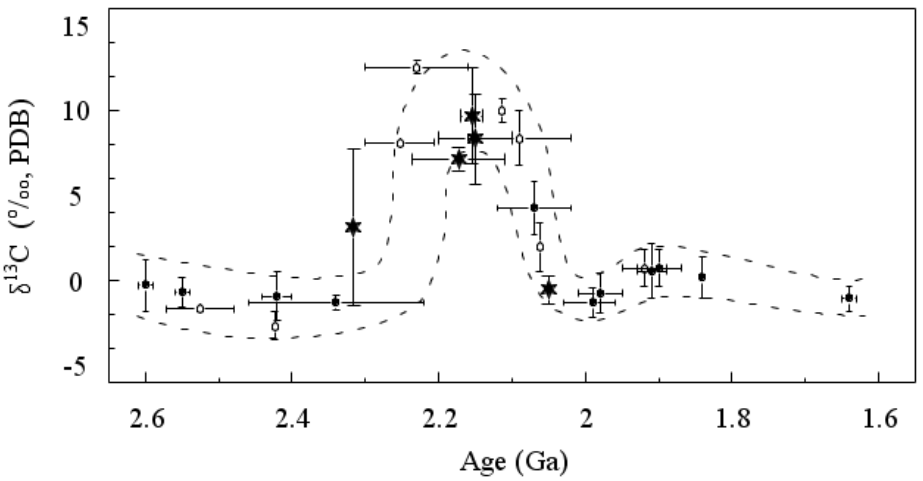

Figure 80. Isotopic composition of carbon in Paleoproterozoic sedimentary carbonates. Data points representing the Fennoscandian Shield are shown by open circles and stars (Karhu 2005).

The Rantamaa Formation is one of the best stratigraphic sections covering the latter part of the excursion at ca. 2.1 Ga. Dolomite samples collected from the lowermost dolomites at the contact of the Tikanmaa Tuffite Formation have yielded positive $\delta^{13}C$ values from 9.6 to 11.4 ‰ (Karhu, 1993). A stratigraphic section of dolomite samples from the Rantamaa quarry are characterized by an oscillating trend of $\delta^{13}C$ values, first showing a upward decrease from 7 to 3 ‰ in the white to light gray dolomites and continuing with an increase back to 5 ‰ in the uppermost laminated dark grey dolomites (Fig. 81). A zircon sample from the pyroclastic Hirsimaa Formation within the dolomite units has yielded a ion microprobe U-Pb age of 2106±8 Ma, constraining the time of deposition of the Rantamaa Dolomite Formation.

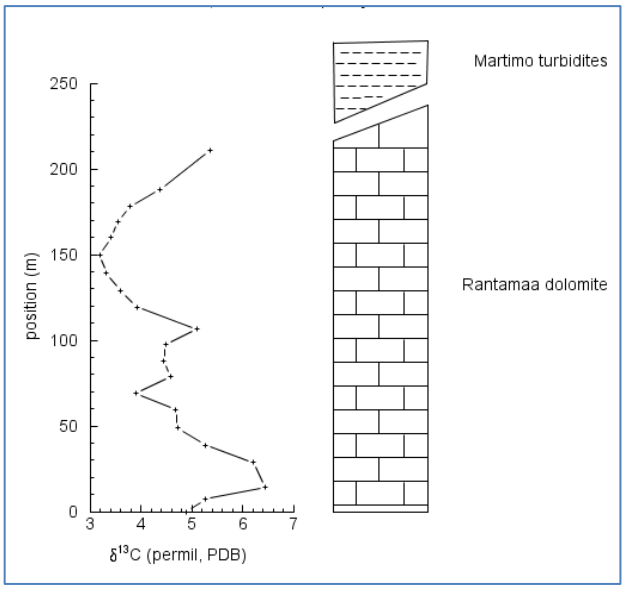

Figure 81. Isotopic composition of carbon in the Rantamaa quarry section (unpubl., Karhu)

The occurrence of mass independent fractionation (MIF) of sulfur isotopes in Archean sedimentary rocks has provided the strongest evidence for the lack of oxygen in the Archean

atmosphere. Papineau et al. (2005) investigated the sulfur isotope composition of pyrite and other sulfide minerals occurring as accessory phases in samples collected from the dark grey Dolomite beds of the Rantamaa quarry. No MIF was observed, indicative of the oxidizing nature of the atmosphere at the time of the deposition of the Rantamaa Formation.

Conclusions

The following conclusions can be made on the basis of field observations and new isotopic data from the Peräpohja belt.

1. The dolomitic carbonate rocks of the Rantamaa Formation at Peräpohja have sedimentary structures providing evidence for deposition in a supratidal to intertidal marine environment.
2. The Rantamaa Formation represents an extensive stratigraphic section covering the latest part of the global Paleoproterozoic carbon isotope excursion in sedimentary carbonates.
3. An ion-microprobe U-Pb date of 2106±8 Ma provides a new constraint for the termination of the Paleoproterozoic carbon isotope excursion at a time, when the marine $\delta^{13}C$ values were at 4 – 6 ‰.

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Rovaniemi and Pohtimolampi

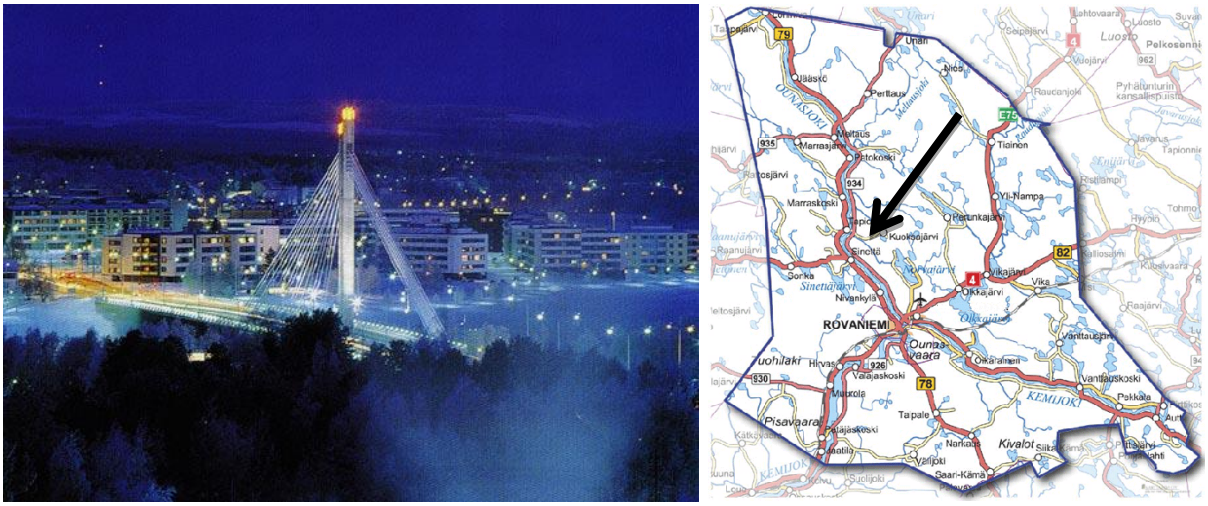

Figure 82. (A) Jätkänkynttilä bridge with its eternal flame over the Kemijoki river. It is a landmark of Rovaniemi. (B) Map of Rovaniemi area. The star indicates the location of Bear's Lodge at Pohtimolampi.

In the lap of two big rivers, Ounasjoki and Kemijoki, has developed an international city of commerce, administration and education. Rovaniemi has always been a gateway into Lapland and it became the admistrative centre of Lapland province in 1938. In 1960 Rovaniemi got a town status and in 2006 the town and the county surrounding it became a city. Due to the location of the city it has become the capital of Lapland that has its own rhythm – the pounding heart of Lapland. The city is an excellent place to enjoy all four seasons in astounding surroundings.

The city centre is surrounded by lively countryside with many villages. This type of landscape with much country around is typical to Rovaniemi; forest, bogs, fields, rivers and lakes create a unique atmosphere. The most well-known person from Rovaniemi is Santa Claus, or at least they like to think so, who you can meet at Santa Claus Office in Santa's land a few km from Rovaniemi. The excursion overnight will take place at Pohtimolampi Bear's Lodge wildness hotel (Fig. 83) north of Rovaniemi (see arrow in map of Fig. 82).

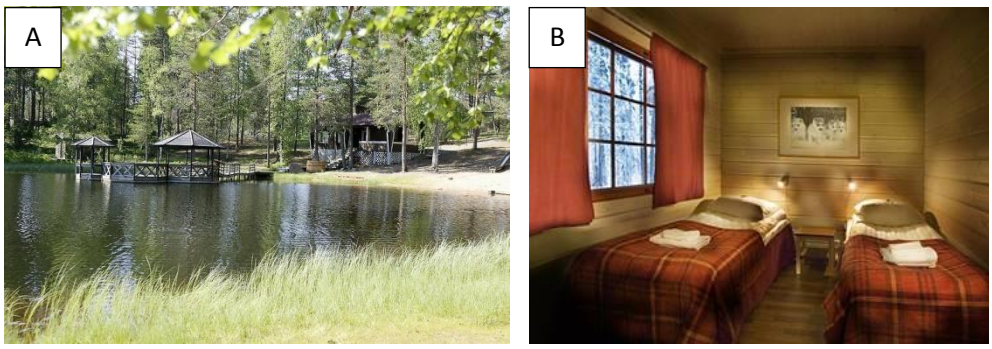

Figure 83. (A) Pohtimolampi (pond) and (B) Bear's Lodge from inside

Rovaniemi: http://www.visitrovaniemi.fi/

City of Tervola

Figure 84. (A) View over Tervola, wildness. (B) The wooden Tervola church built in 1687 on the banks of Kemijoki river at idyllic surroundings.

Tervola is from a short drive from Rovaniemi, Kemi and Tornio. The municipality is largely rural and forestry and agriculture are prominent. Tervola is famous for wood-, rock- and metal industry. The landscape is dominated by Kemijoki river and several brooks and hills.

Outdoors play a vital role with the tourism of the Tervola area; there are activities such as fishing, hiking, boating and canoeing for which the Kemijoki and its rapid tributaries offer many opportunities.

Tervola has been important already in the prehistorical times. Over 800 residential wells have been found which are among the top such sites in Europe. There is also an exhibition park where several plant species, some rare ones too, can be seen.

Tervola: http://www.tervola.fi/matkailu.htm

From Rantamaa (Tervola) we will drive to Oulu Airport and take late afternoon flight to Helsinki. The excursion thus gets to its end at Helsinki-Vantaa Airport.

I hope you enjoyed this excursion visiting several geological landmark sites of Finland.

Kumpula, Sept. 19, 2012

Lauri