33 IGC excursion No 15, August 15th – 21st 2008

33 IGC, The Nordic Countries

Metallogeny and tectonic evolution of the Northern Fennoscandian Shield

Guidebook edited by V. Juhani Ojala and Markku Iljina

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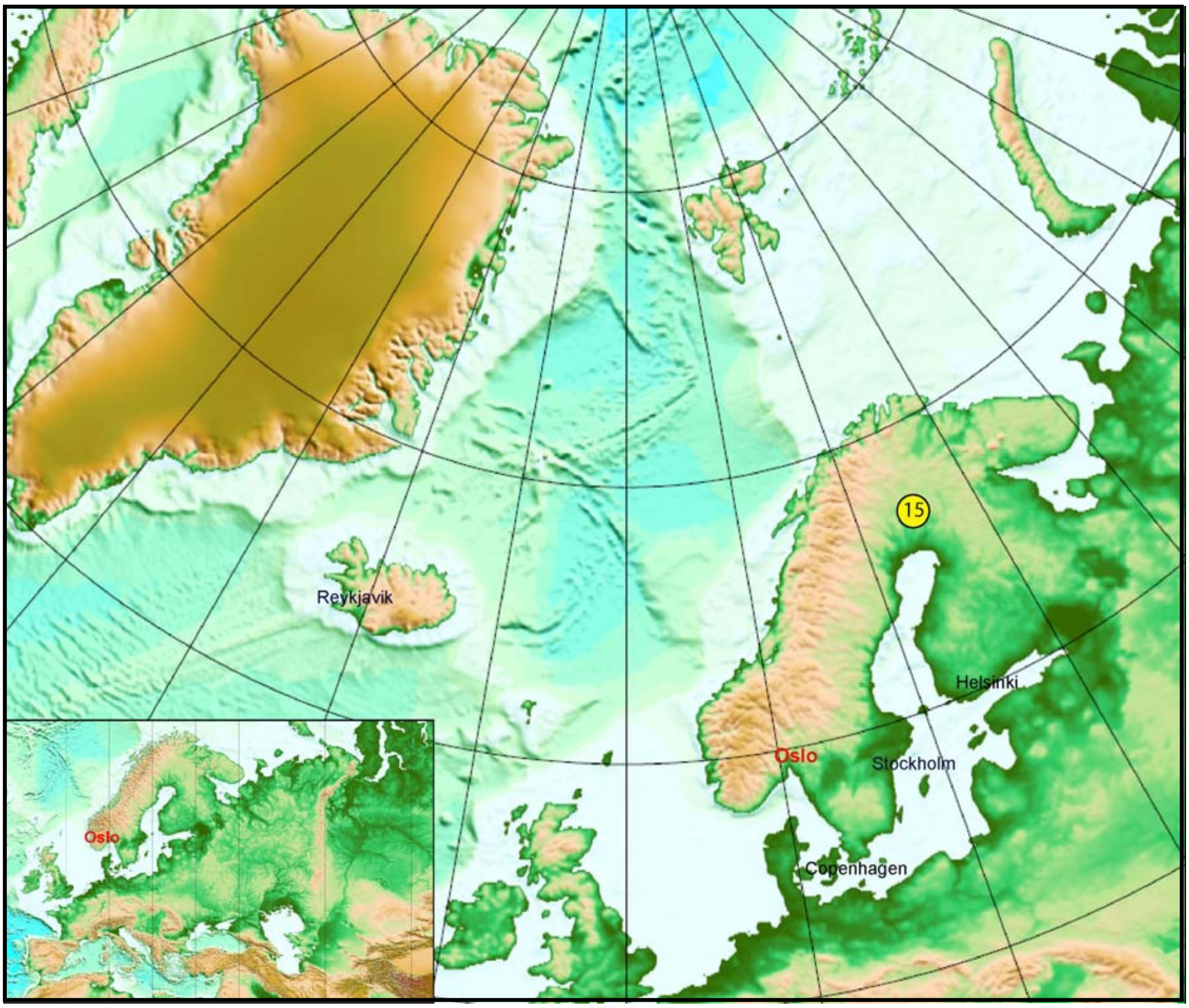

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Abstract

The Fennoscandian Shield is one of the most important mining areas in Europe. Mineral deposit types include VMS, Kiruna-type apatite-iron, orogenic Au, epigenetic Cu-Au ore, mafic and ultramafic-hosted Cr, Ni-(Cu), PGE and BIF. Palaeoproterozoic parts of the shield are better mineralized than the Archaean areas.

The Portimo Complex is exceptional in hosting a variety of styles of PGE mineralization. Economically most potential styles are the contact type and reef-type PGE deposits, and offset base-metal and PGE deposits in the footwall rocks. Other PGE enrichment include in the Portimo Dykes below the Konttijärvi and Ahmavaara marginal series, PGE concentrations near the roof of the Suhanko Intrusion, a Pt-anomalous pyroxenitic pegmatite pipe, and chromite and silicate-associated PGE enrichments in the lower parts of the Narkaus Intrusion and MCU II.

Pahtavaara is an active gold mine, with a total in situ size estimate of 15 t gold. It is sited in an altered komatiitic sequence at the eastern part of the Central Lapland greenstone belt and has many of the characteristics orogenic gold deposits, but has an anomalous barite-gold association and a very high fineness (>99.5 % Au) of gold.

The Kevitsa Ni-PGE deposit is a large, low-grade disseminated sulphide deposit located in the upper part of the ultramafic zone, in the NE part of the Kevitsa intrusion (2.057±5 Ga). Distribution of Cu, Ni, PGE+Au, and S within the deposit is complex and variable. The deposit has been divided into two bodies, the main ore body (or Main Ore) and the overlying Upper Ore. Four main ore types have been defined, based on the metal and sulphur contents: Regular ore, false ore, Ni-PGE ore, and transitional ore. As distribution of Cu, Ni, PGE+Au, and S within the deposit is complex and variable, the different ore types tend to grade into another.

The Suurikuusikko gold deposit is the largest known gold resource in northern Europe. Current resource estimate is about 80 t gold (16 million tonnes at 5.1 g/t). Host rocks are dominantly mafic volcanic rocks within over a 25-kilometre long strike-slip shear zone. Gold is refractory, occurring within arsenopyrite and pyrite.

Iron ores in the Kolari area contain significant amounts of copper and gold. The ores are hosted by diopside skarn and quartz-albite rocks. Hannukainen deposit produced 1.96 Mt iron, 40,000 t copper and 4300 kg gold in 1978-1992. The present in situ resource estimate is 16 t Au, 125,000 t Cu and 26 Mt Fe. Typical ore mineral association is magnetite-chalcopyrite-pyrite±pyrrhotite.

The Sahavaara iron ore comprises three lenses of skarn-rich iron formation. Resources at Stora Sahavaara are 145 Mt with 43.1 % Fe and 0.076 % Cu. The ore zone consists of serpentine-rich magnetite ore including lenses and layers of serpentine-diopside-tremolite skarn. Pyrrhotite and pyrite occur disseminated in the ore together with minor chalcopyrite. The Kiruna apatite-magnetite-hematite deposit comprises about 2000 Mt of ore. The present production is over 20 Mt per year with 46.2 % Fe. The ore body is 5 km long, up to 100 m thick, and it extends at least 1500 m below the surface. It follows the contact between a thick pile of trachyandesitic lava and overlying pyroclastic rhyodacite. Granophyric dikes cut the ore and give the minimum age for the ore (U-Pb zircon age of 1880±3 Ma).

The Gruvberget Cu-mines in Norrbotten produced about 1000 ton Cu during 1657–1684. The nearby Gruvberget apatite iron ore is estimated to contain 64.1 Mt with 56.9 % Fe and 0.87 % P to the depth of 300 m. The host rocks are strongly scapolite- and K feldspar-altered intermediate to mafic volcanic rocks. The apatite iron ore consists of magnetite in the northern part and hematite in the middle and southern part of the deposit.

Aitik is Sweden's largest sulphide mine with an annual production of 18 Mt of ore with 0.38 % Cu and 0.22 ppm Au. Reserves are at 244 Mt, and there is an additional mineral resource of 970 Mt. Chalcopyrite and pyrite are the main ore minerals with minor magnetite, pyrrhotite, bornite, and molybdenite. The host rock is garnet-bearing biotite schist and gneiss in the footwall, and quartz-muscovite schist in the hanging wall. Intermediate footwall subvolcanic c. 1.873±24 Ga intrusion is weakly mineralised.

The Kemi Chrome mine is hosted by a 2.4 Ga mafic-ultramafic layered intrusion. The mine's current proven ore reserves are 40 Mt plus 85 Mt in resources. The average chromium oxide content of the ore is about 26 % and its average chrome-iron ratio is 1.6. The chromitite layer, which parallels the basal contact zone of the Kemi Intrusion, is known over the whole length of the complex. In the central part of the intrusion, the basal chromitite layer widens into a thick (up to 160 m) chromitite accumulation.

Logistics

Dates and location

***Timing*:**	**Friday 15th (evening) – Thursday 21st (16:00 hours) August 2008**
***Start location*:**	**Rovaniemi, Finland**
***End location*:**	**Rovaniemi, Finland**

Travel arrangements

Participants should organise flights to arrive Rovaniemi on Friday August 15th, preferentially on the flight AY429 departing at 16:20 hours. The last flight to Rovaniemi is AY 355 from Helsinki via Oulu to Rovaniemi will depart at 20.10 hours, it will to arrive in Rovaniemi at 22:10 hours. The airport taxis (about 6 Euros) are the easiest way to get to the hotel. The excursion will end at Rovaniemi Thursday 21st about 16:00 hours and, if necessary, participants can be dropped at the airport for the flight AY430 which leave 18:05 hours to Helsinki.

Accommodation

We are staying in good standard hotels in towns and off season ski resorts. These have all normal hotel facilities with towels, linen etc. provided. The normal price of the excursion is based on shared accommodation; please indicate if you wish to have a single room at the time of registration. If you are staying in Rovaniemi after the excursion, you should make you own booking e.g. before we leave on Saturday 16th morning.

Safety Rules - IMPORTANT

NOTE: The instructions of your guides MUST be followed at all times. When we are visting mines pay special attention to the movement of the very large machinery. If you are taking samples, make sure that the location(s) are safe. In the mine sites hard hats and safety goggles must be worn at all times.

Field logistics

The first day:
We'll start with an introductory lecture at 08:30 hours at the conference room of the hotel (the name of the hotel will be announced later).

Workshops:
We will book conference rooms at the hotels for casual wrap ups of each day and intros for the next day before dinner (except kemi, as we will arrive late). The length depends on how keen we are, lectures may take anything between 15 mins and 2 hours.

Food:
Lunches are included, we will have packed (in some hotels we will pack ourselves) field lunches from the hotels in the morning. In the Suurikuusikko, Aitik and Kemi mines lunches are sponsored by the mines.

Dinners are not included, but we will book tables for the group in the restaurants of the accommodating hotels. The hotels have buffet dinners for about 20 Euros. We are staying in ski resorts and as it is out of the season, so a la carte menus might be limited. Typical prices: starters 10 Euros, main course 15-30 Euros, desserts 5 Euros, beer 5 Euros, bottle of wine 20-50 euros.

Weather:
Typical day temperatures between 10-15°C are expected in the mid August, mornings may be chilly 2-5°C. Long term statistics indicate that temperatures can vary between -5°C and +25°C. It can be sunny all the time, but there can be rain or even snow if we are lucky. Take clothes accordingly. Good field boots are necessary but there are no difficult terrains, steep slopes or long walks during the excursion. Even it is the end of the summer; there can be mosquitoes, midges or horse flies: repellent may be needed.

Transport during the excursion:
There is a large 50 person bus all the way.

General Introduction to Geology and Metallogeny of Fennoscandian Shield

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The Fennoscandian Shield forms the north-westernmost part of the East European craton and constitutes large parts of Finland, NW Russia, Norway, and Sweden (Fig. 1). The oldest rocks yet found in the shield have been dated at 3.5 Ga (Huhma et al. 2004) and major orogenies took place in the Archaean and Palaeoproterozoic. Younger Meso- and Neoproterozoic crustal growth took place mainly in the western part, but apart from the anorthositic Ti-deposits in SW Norway, no major ore deposits are related to rocks of this age. The western part of the shield was reworked during the Caledonian Orogeny.

Economic mineral deposits are largely restricted to the Palaeoproterozoic parts of the shield. Although Ni–PGE, Mo, BIF, and orogenic gold deposits, and some very minor VMS deposits occur in the Archaean, most economic examples of these deposit types are related to Palaeoproterozoic magmatism, deformation and fluid flow. Besides these major deposit types, the Palaeoproterozoic part of the shield is also known for its Fe-oxide deposits, including the famous Kiruna-type Fe-apatite deposits. Large-tonnage low-grade Cu–Au deposits (e.g., Aitik), are associated with intrusive rocks in the northern part of the Fennoscandian Shield. These deposits have been described as porphyry style deposits or as hybrid deposits with features that also warrant classification as iron oxide–copper–gold (IOCG) deposits (Weihed 2001, Wanhainen et al. 2005).

A generalised geological map of northern Fennoscandia is provided in Appendix 1, major deposits are indicated on this map. During this field trip to northern Fennoscandia (Appendix 2), we will emphasize deposit characteristics, their diversity, and speculate on temporal and spatial relationship between different deposits. The deposits are discussed in terms of their tectonic setting and relationship to the overall geodynamic evolution of the shield. Also considered are deposit-scale structural features and their relevance for the understanding of the ore genesis.

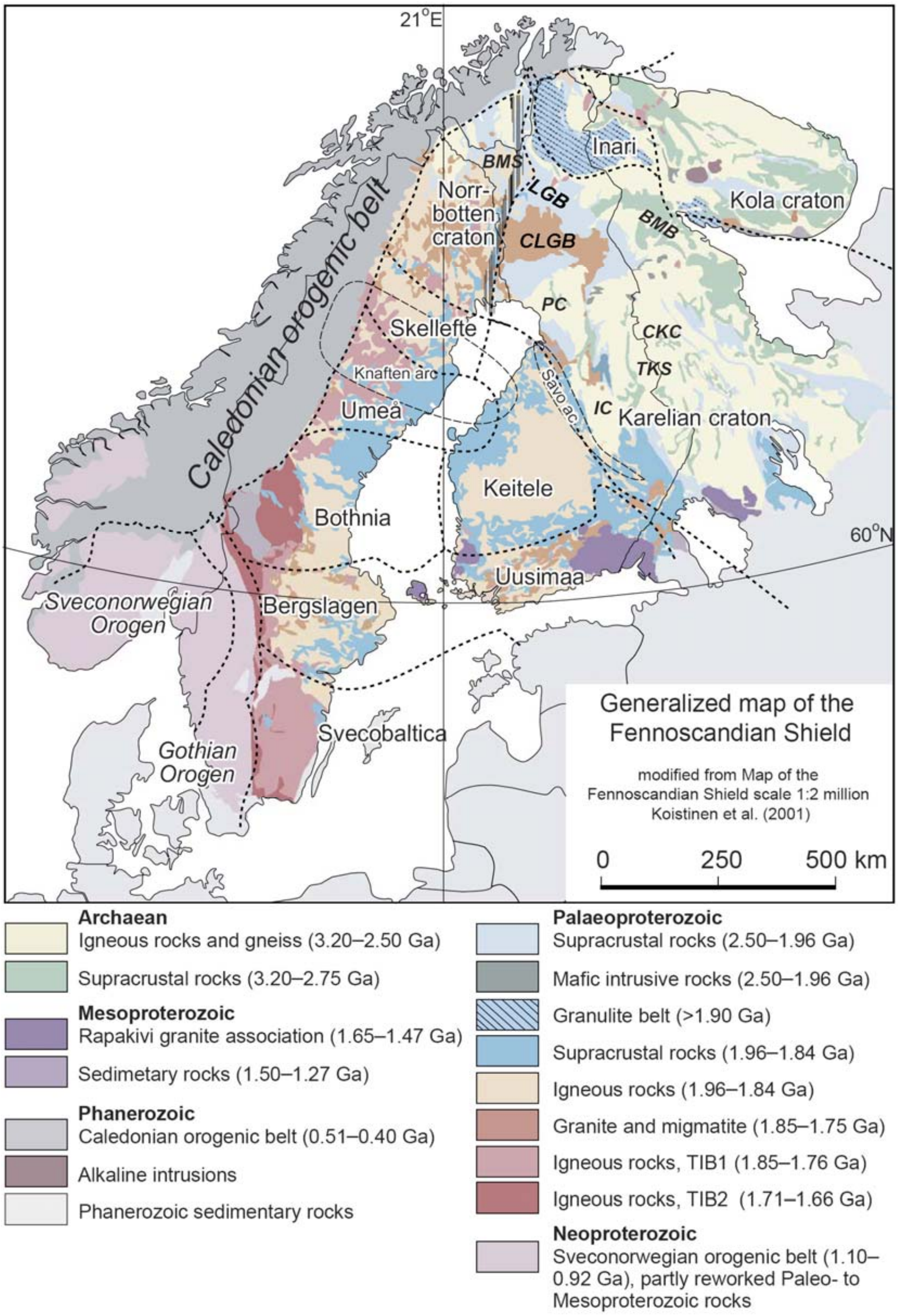

Figure 1. Simplified geological map of the Fennoscandian Shield with major tectono-stratigraphic units discussed in text. Map adapted from Koistinen et al. (2001), tectonic interpretation after Lahtinen et al. (2005). LGB = Lapland Greenstone Belt, CLGC = Central Lapland Granitoid Complex, BMB = Belomorian Mobile Belt, CKC = Central Karelian Complex, IC = Iisalmi Complex, PC = Pudasjärvi Complex, TKS = Tipasjärvi–Kuhmo–Suomussalmi greenstone complex. Shaded area, BMS = Bothnian Megashear.

Regional Geology

Introduction

The Fennoscandian Shield is one of the most important mining areas in Europe, and the northern part, including Sweden and Finland, (Fig. 1, Appendix 1) is intensely mineralised. Mineral deposit types include VMS, Kiruna-type apatite-iron ores, mesothermal (orogenic) Au ore, epigenetic Cu-Au ore, mafic and ultramafic-hosted Cr, Ni-(Cu), PGE and BIF. Unlike most other shield areas, the Fennoscandian Shield is more mineralised in its Palaeoproterozoic than the Archaean areas.

The oldest preserved continental crust in the Fennoscandian Shield was generated during the Saamian Orogeny at 3.1–2.9 Ga (Fig. 1) and is dominated by gneissic tonalite, trondhjemite and granodiorite. Rift- and volcanic arc-related greenstones, subduction-generated calc-alkaline volcanic rocks and tonalitic-trondhjemitic igneous rocks were formed during the Lopian Orogeny at 2.9–2.6 Ga. Only a few Archaean economic to subeconomic mineral deposits have been found in the shield, including orogenic gold, BIF and Mo occurrences, and ultramafic- to mafic-hosted Ni-Cu (Frietsch et al. 1979, Gaál 1990, Weihed et al. 2005).

During the Palaeoproterozoic, Sumi-Sariolian (2.5–2.3 Ga) clastic sediments, intercalated with volcanic rocks varying in composition from komatiitic and tholeiitic to calc-alkaline and intermediate to felsic, were deposited on the deformed and metamorphosed Archaean basement during extensional events. Layered intrusions, most of them with Cr, Ni, Ti, V and/or PGE occurrences, represent a major magmatic input at 2.45–2.39 Ga (Amelin et al. 1995, Mutanen 1997, Alapieti & Lahtinen 2002). Periods of arenitic sedimentation preceded and followed extensive komatiitic and basaltic volcanic stages at c. 2.2, 2.13, 2.05 and 2.0 Ga in the northeastern part of the Fennoscandian Shield during extensional events (Mutanen 1997, Lehtonen et al. 1998, Rastas et al. 2001). Associated with the subaquatic extrusive and volcaniclastic units, there are carbonate rocks, graphite schist, iron formation and stratiform sulphide occurrences across the region.

Svecofennian subduction-generated calc-alkaline andesites and related volcaniclastic sedimentary units were deposited around 1.9 Ga in the northern Fennoscandia in a subaerial to shallow-water environment. In the Kiruna area, the 1.89 Ga Kiirunavaara Group rocks (formerly Kiruna Porphyries) are chemically different from the andesites and are geographically restricted to this area. The Svecofennian porphyries form host to apatite-iron ores and various styles of epigenetic Cu-Au occurrences including porphyry Cu-style deposits (Weihed et al. 2005).

The up to 10 km thick pile of Palaeoproterozoic volcanic and sedimentary rocks was multiply deformed and metamorphosed contemporaneously with the intrusion of the 1.89–1.87 Ga granitoids. Anatectic granites were formed during 1.82–1.79 Ga, during another major stage of deformation and metamorphism. Large-scale migration of fluids of variable salinity during the many stages of igneous activity, metamorphism and deformation is expressed by regional scapolitisation, albitisation and albite-carbonate alteration in the region. For example, scapolitisation is suggested to be related to felsic intrusions (Ödman 1957), or to be an expression of mobilised evaporates from the supracrustal successions during metamorphism (Tuisku 1985, Frietsch et al. 1997, Vanhanen 2001).

Since Hietanen (1975) proposed a subduction zone dipping north beneath the Skellefte district, many similar models have been proposed for the main period of the formation of the crust during the Svecokarelian (or Svecofennian) orogeny roughly between 1.95 and 1.77 Ga (e.g. Rickard & Zweifel 1975, Lundberg 1980, Pharaoh & Pearce 1984, Berthelsen & Marker 1986, Gaál 1986, Weihed et al. 1992). This orogeny involved both strong reworking of older crust within the Karelian craton and, importantly, subduction towards NE, below the Archaean, and the accretion of several volcanic arc complexes from the SW towards NE. Recently, substantially more complex models for crustal growth at this stage of the evolution of the Fennoscandian Shield have been proposed (e.g. Nironen 1997, Lahtinen et al. 2003 2005). The most recent model for the Palaeoproterozoic tectonic evolution of the Fennoscandian Shield involving five partly overlapping orogenies was presented by Lahtinen et al. (2005). This model builds on the amalgamation of several microcontinents and island arcs with the Archaean Karelian, Kola and Norrbotten cratons and other pre-1.92 Ga components. The Karelian craton experienced a long period of rifting (2.5–2.1 Ga) that finally led to continental break-up (c. 2.06 Ga). The microcontinent accretion stage (1.92–1.87 Ga) includes the Lapland-Kola and Lapland-Savo orogenies (both with peak at 1.91 Ga) when the Karelian craton collided with Kola and the Norrbotten cratons, respectively. It also includes the Fennian orogeny (peak at c. 1.88 Ga) caused by the accretion of the Bergslagen microcontinent in the south. The following continental extension stage (1.86–1.84 Ga) was caused by extension of hot crust in the hinterlands of subduction zones located to the south and west. Oblique collision with Sarmatia occurred during the Svecobaltic orogeny (1.84–1.80 Ga). After collision with Amazonia, in the west, during the Nordic orogeny (1.82–1.80 Ga), orogenic collapse and stabilization of the Fennoscandian Shield took place at 1.79–1.77 Ga. The Gothian orogeny (1.73–1.55 Ga) at the southwestern margin of the shield ended the Palaeoproterozoic orogenic development.

Despite these new, refined models of the Palaeoproterozoic evolution between 1.95 and 1.77 Ga, the tectonic evolution of the northern part of the Karelian craton, i.e. the part north of the Archaean-Proterozoic palaeoboundary, is still rather poorly understood in detail.

Palaeoproterozoic 2.45–1.97 Ga greenstone belts

The Palaeoproterozoic Lapland greenstone belt, which overlies much of the northern part of the Archaean craton, is the largest coherent greenstone terrain exposed in the Fennoscandian Shield (Fig 1). It extends for over 500 km from the Norwegian northwest coast through the Swedish and Finnish Lapland into the adjacent Russian Karelia in the southeast. Due to large lithostratigraphic similarities in different greenstone areas from this region and the mainly tholeiitic character of the volcanic rocks, Pharaoh (1985) suggested them to be coeval and representing a major tholeiitic province. Based on petrological and chemical studies of the mafic volcanic rocks and associated sediments, an originally continental rift setting is favoured for these greenstones (e.g. Lehtonen et al. 1985, Pharaoh et al. 1987, Huhma et al. 1990, Olesen & Sandstad 1993, Martinsson 1997). It includes the Central Lapland greenstone belt in Finland and the Kiruna and Masugnsbyn areas in Sweden, all of which are visited during this field trip. The lithostratigraphy of the Finnish part of the Lapland greenstone belt, the Central Lapland greenstone belt, is presented in Figure 2.

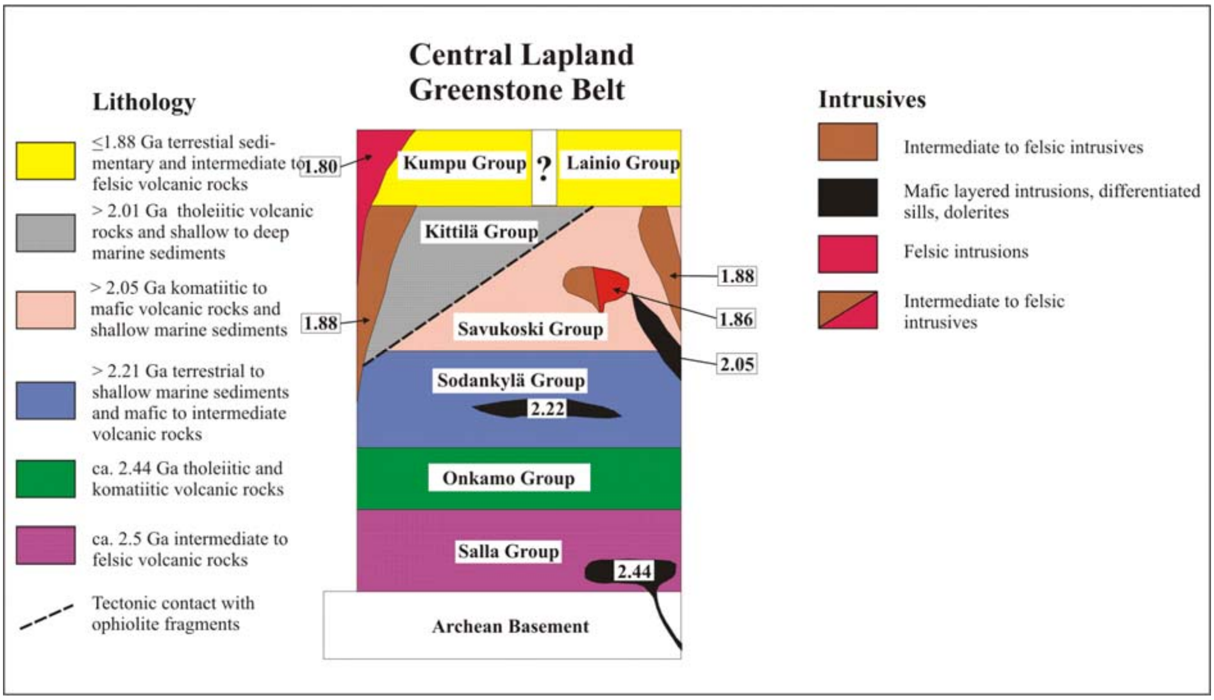

Figure 2. Stratigraphy of the Central Lapland greenstone belt. Ages given as Ga. Compiled by Tero Niiranen, after Lehtonen et al. (1998) and Hanski et al. (2001).

In northern Sweden, a Palaeoproterozoic succession of greenstones, porphyries and clastic sediments rests unconformably on deformed, 2.7–2.8 Ga, Archaean basement. Stratigraphically lowest is the Kovo Group. It includes a basal conglomerate, tholeiitic lava, calc-alkaline basic to intermediate volcanic rocks and volcaniclastic sediments. Sedimentary rocks were deposited along a coastline of a marine rift basin, and material input was provided through a number of alluvial fans (Kumpulainen 2000). The Kovo Group is overlain by the Kiruna Greenstone Group which is dominated by mafic to ultramafic volcanic rocks. An albite diabase (albitised dolerite), intruding the lower part of the Kovo Group, has been dated at 2.18 Ga (Skiöld 1986), and gives a minimum depositional age for this unit. The Kovo Group is suggested to be c. 2.5–2.3 Ga in age (Sumi-Sariolan) whereas the Kiruna Greenstone Group is suggested to be 2.2–2.0 Ga in age (Jatulian and Ludikowian). The upper contacts of the Kovo Group and the Kiruna Greenstone Group are characterised by minor unconformities and clasts from these units are found in basal conglomerates in overlying units.

In Finland, the lowermost units of the greenstones also lie unconformably on the Archaean, and are represented by the Salla Group rocks in the Central Lapland greenstone belt (CLGB; Fig. 2), a polymictic conglomerate in the Kuusamo schist belt and the Sompujärvi Formation of the Peräpohja schist belt. This is followed by sedimentary units which precede the c. 2.2 Ga igneous event and comprise the Onkamo and Sodankylä Group rocks in the CLGB. The latter lithostratigraphic group also hosts most of the known Palaeoproterozoic syngenetic sulphide occurrences in the CLGB.

The Savukoski Group mafic to ultramafic volcanic and shallow-marine sedimentary units were deposited between 2.2 and 2.01 Ga in the CLGB, and similar units were also formed in the Kuusamo and Peräpohja belts (Lehtonen et al. 1998, Rastas et al. 2001). Age determinations of the Palaeoproterozoic greenstones exist mainly from Finland (e.g. Perttunen & Vaasjoki 2001, Rastas et al. 2001, Väänänen & Lehtonen 2001) and suggests a major magmatic and rifting event at c. 2.1 Ga with the final break up taking place at c. 2.06 Ga.

Extensive occurrence of 2.13 and 2.05 Ga dolerites also support these dates. Thick piles of mantle-derived volcanic rocks including komatiitic and picritic high-temperature melts are restricted to the Kittilä-Karasjokk-Kautokeino-Kiruna area and are suggested to represent plume-generated volcanism (Martinsson 1997). The rifting of the Archaean craton, along a line in a NW-direction from Ladoga to Lofoten, was accompanied by NW-SE and NE-SW directed rift basins (Saverikko 1990) and injection of 2.1 Ga trending dyke swarms parallel to these (Vuollo 1994). Eruption of N-MORB pillow lava occurred along the rift margins as exemplified by occurrences at Tohmajärvi, Kuopio, Ostrobothnia and Piteå (Åhman 1957, Kähkönen et al. 1986, Lukkarinen 1990, Pekkarinen & Lukkarinen 1991). The Kiruna greenstones and dyke swarms north of Kiruna outline a NNE-trending magmatic belt extending to Alta and Repparfjord in the northernmost Norway. This belt is almost perpendicular to the major rift, and may represent a failed rift arm related to a triple junction south of Kiruna (Martinsson 1997). The rapid basin subsidence, accompanied by eruption of a 500–2000 m thick unit of MORB-type pillow lava is suggested to be an expression of the development of this rift arm.

Rifting culminated in extensive mafic and ultramafic volcanism and the formation of oceanic crust at c. 1.97 Ga. This is indicated by the extensive komatiitic and basaltic lavas of the Kittilä Group of the CLGB in the central parts of the Finnish Lapland (Fig. 2). The 1.97 Ga stage also included deposition of shallow- to deep-marine sediments, the latter indicating the most extensive rifting in the region. Fragments of oceanic crust were subsequently emplaced back onto the Karelian craton in Finland, as indicated by the Nuttio ophiolites in central Finnish Lapland and the Jormua and Outokumpu ophiolites further south (Kontinen 1987, Gaál 1990, Sorjonen-Ward et al. 1997, Lehtonen et al. 1998).

Svecofennian complexes

The Palaeoproterozoic greenstones are overlain by volcanic and sedimentary rocks comprising several different but stratigraphically related units. Regionally, they exhibit considerable variation in lithological composition due to partly rapid changes from volcanic- to sedimentary-dominated facies. Stratigraphically lowest in the Kiruna area are rocks of the Porphyrite Group and the Kurravaara Conglomerate. The former represents a volcanic-dominated unit and the latter is a mainly epiclastic unit (Offerberg 1967) deposited as one or two fan deltas (Kumpulainen 2000). The Sammakkovaara Group in northeastern Norrbotten comprises a mixed volcanic-epiclastic sequence that is interpreted to be stratigraphically equivalent to the Porphyrite Group and the Kurravaara Conglomerate, and the Pahakurkio Group, south of Masugnsbyn. The Muorjevaara Group in the Gällivare area is also considered to be equivalent to the Sammakkovaara Group in the Pajala area and is dominated by intermediate volcaniclastic rocks and epiclastic sediments. In the Kiruna area, these volcanic and sedimentary units are overlain by the Kiirunavaara Group that is followed by the Hauki and Maattavaara quartzites constituting the uppermost Svecofennian units in the area.

In northern Finland, pelitic rocks in the Lapland Granulite Belt were deposited after 1.94 Ga (Tuisku & Huhma 2006). Svecofennian units are mainly represented by the Lainio and Kumpu Groups in the CLGB (Lehtonen et al. 1998) and by the Paakkola Group in the Peräpohja area (Perttunen & Vaasjoki 2001). The molasse-like conglomerates and quartzites comprising the Kumpu Group were deposited in deltaic and fluvial fan environments after 1913 Ma and before c. 1800 Ma (Rastas et al. 2001). The Kumpu rocks apparently are equivalent to the Hauki and Maattavaara quartzites, whereas the sedimentary and volcanic units of the Lainio Group could be related to the Porphyrite Group rocks and the Kurravaara Conglomerate of the Kiruna area.

With the present knowledge of ages and petrochemistry of the Porphyrite, Lainio and Kumpu Groups, it is possible to attribute these rocks partially (Kumpu) to completely (Porphyrite and Lainio) to the same event of collisional tectonics and juvenile convergent margin magmatism. This period of convergence was manifested by the numerous intrusions of Jörn- (south of the craton margin) and Haparanda- (within the craton) type calc-alkaline intrusions, as described by Mellqvist et al. (2003). Within a few million years, this period of convergent margin magmatism was followed by a rapid uplift recorded in extensive conglomeratic units, more alkaline and terrestrial volcanism (Vargfors-Arvidsjaur Groups south of the craton margin and the Kiirunavaara Group within the craton) and plutonism (Gallejaur-Arvidsjaur type south of the craton margin, Perthite Monzonite Suite within the craton). This took place between 1.88 and 1.86 Ga and the main volcanic episode probably lasted less than 10 million years.

The evolution after c. 1.86 is mainly recorded by an extensive S-type magmatism (c. 1.85 Ga Jyryjoki, and 1.81–1.78 Ga Lina-type and the Central Lapland Granitoid Complex) derived from anatectic melts in the middle crust. In the western part of the shield, extensive I- to A-type magmatism (Revsund-Sorsele type) formed roughly N-S trending batholiths (the Transcandinavian Igneous Belt) coeval with the S-type magmatism. Scattered intrusions of this type and age also occur further east (e.g. Edefors in Sweden, Nattanen in Finland). The period from c. 1.87 to 1.80 Ga possibly also involved a shift in orogenic vergence from NE-SW to E-W in the northern part of the Shield as suggested by Weihed et al. (2002).

Palaeoproterozoic magmatism

Early rifting and emplacement of layered igneous complexes

The beginning of the rifting period between 2.51 and 2.43 Ga is indicated by intrusion of numerous layered mafic igneous complexes (Alapieti et al. 1990, Weihed et al. 2005). Most of the intrusions are located along the margin of the Archaean granitoid area, either at the boundary against the Proterozoic supracrustal sequence, totally enclosed by Archaean granitoid, or enclosed by a Proterozoic supracrustal sequence. Most of the intrusions are found in west - east trending Tornio-Näränkävaara belt of layered intrusions (Iljina & Hanski 2005). Rest of the intrusions are found in NW Russia, central Finnish Lapland and NW Finland. These Palaeoproterozoic layered intrusions are characteristic to northern Finland as only one of them, the Tornio intrusion, being partly on the Swedish side of the border. Alapieti and Lahtinen (2002) divided the intrusions into three types, (1) ultramafic–mafic, (2) mafic and (3) intermediate megacyclic. They also interpret the ultramafic–mafic and the lowermost part of the megacyclic type to have crystallised from a similar, quite primitive magma type, which is characterised by slightly negative initial ε_{Nd} values and relatively high MgO and Cr, intermediate SiO_2, and low TiO_2 concentrations, resembling the boninitic magma type. The upper parts of megacyclic type intrusions and most mafic intrusions crystallised from an evolved Ti-poor, Al-rich basaltic magma.

Amelin et al. (1995) suggested two slightly different age groups of the intrusions for Fennoscandian Shield, the first with U–Pb ages between 2.505 and 2.501 Ga, and the second of a slightly younger period, 2.449 to 2.430 Ga. All Finnish layered intrusions belong to the younger age group. The intrusions were later deformed and metamorphosed during the Svecokarelian Orogeny.

Mafic dykes

Mafic dykes are locally abundant and show a variable strike, degree of alteration and metamorphic recrystallisation which, with age dating, indicate multiple igneous episodes. Albite diabase (a term commonly used in Finland and Sweden for any albitised dolerite) is a characteristic type of intrusions that form up to 200 m thick sills. They have a coarse-grained central part dominated by albitic plagioclase and constitute laterally extensive, highly magnetic units north of Kiruna. Similar to the greenstone-related albite diabases also occur in eastern Finland (Vuollo 1994, Lehtonen et al. 1998), and they have an age of c. 2.2 Ga (Skiöld 1986, Vuollo 1994).

Extensive dyke swarms occur in the Archaean domain north of Kiruna; the swarms are dominated by 1–100 m wide dykes with a metamorphic mineral assemblage but with a more or less preserved igneous texture (Ödman 1957, Martinsson 1999a,b). The NNE-trending dykes that are suggested to represent feeders to the Kiruna Greenstone Group (Martinsson 1997, 1999a,b). Scapolite-biotite alteration is common in the dykes within Svecofennian rocks (Offerberg 1967) and also in feeder dykes within the lower part of the Kiruna Greenstone Group (Martinsson 1997).

In northern Finland, albite diabases, both sills and dykes, form age groups of 2.2, 2.13, 2.05 and 2.0 Ga (Vuollo 1994, Lehtonen et al. 1998, Perttunen & Vaasjoki 2001, Rastas et al. 2001). These dates also reflect extrusive magmatism in the region. The dykes vary in size from <1 m to one km wide, nearly all show internal differentiation and igneous textures but metamorphic and altered mineral assemblages (carbonate, sericite, epidote, biotite or scapolite), and in areas with greenschist-facies regional metamorphism are commonly surrounded by albitised and carbonated country rocks (Eilu 1994).

Granitoids

A major part of the bedrock in the northernmost Sweden and Finland is composed of various types of granitoids. The major suites are: 1) Haparanda Suite, calc-alkaline, 1.90–1.86 Ga, granite-granodiorite-tonalite-diorite-gabbro, 2) Perthite Monzonite Suite, 1.88–1.86 Ga, granite-monzonite-diorite-gabbro-peridotite, 3) Lina Suite, S-type, minimum melt, anatectic, migmatites-associated, 1.82–1.78 Ga, granite-pegmatite, and 4) A-, I-type intrusions, 1.80–1.77 Ga, granite-monzonite-granodiorite-diorite-gabbro. In the Lapland Granulite Belt arc magmatism with norite-enderbite series rocks intruded the supracrustal sequence at 1920–1905 Ma (Tuisku & Huhma 2006).

Haparanda Suite

The name Haparanda Suite was originally assigned to intrusions in southeastern Norrbotten (Ödman et al. 1949). Later, it was extended to comprise petrographically similar rocks in northern Norrbotten and Finland (Ödman 1957, Hiltunen 1982). These intrusions are medium- to coarse-grained, even grained, moderately to intensely deformed, grey tonalites and granodiorites, which are associated with gabbros, diorites and rare true granites (Ödman 1957).

The geochemical signature of the Haparanda suite is typical for ”volcanic arc granitoids” with low Rb, Y, and Nb (Mellqvist et al. 2003). They define a calc-alkaline trend and are metaluminous to slightly peraluminous. Reported age determinations for Haparanda type intrusions from the southeastern Norrbotten show a range of 1.89–1.87 Ga (Öhlander et al. 1987a, Wikström et al. 1996, Witschard 1996, Persson & Lundqvist 1997, Wikström & Persson 1997a, Mellqvist et al. 2003). An age range of 1.90–1.86 Ga has been defined for the suite in the western parts of northern Finland (Huhma 1986, Perttunen & Vaasjoki 2001, Rastas et al. 2001, Väänänen & Lehtonen 2001).

The Haparanda Suite intrusions are regarded as comagmatic with extrusive phases of the early Svecofennian arc magmatism. This is supported by their calc-alkaline character (Bergman et al. 2001) and, also, by the contemporaneous timing with subduction modelled for the shield (Lahtinen et al. 2003, 2005).

Perthite Monzonite Suite

Perthite Monzonite Suite intrusions occur as large plutons in the northwestern part of Norrbotten in Sweden, but are rarer in eastern Norrbotten (Geijer 1931, Witschard 1984, Bergman et al. 2001). The plutons are generally undeformed, although magmatic foliation may occur at the contacts. Three major clusters are outlined by their silica content, at 38–52, 57–66 and 70–76 %. The Perthite Monzonite Suite can be classified as a quartz monzonite–adamellite–granite suite, which is peraluminous to metaluminous with alkaline trends (Ahl et al. 2001). Zoned composite intrusions, typically with a mafic to intermediate outer part and a felsic centre, are common (Kathol & Martinsson 1999). The main magmatic event can probably be set at 1.87–1.88 Ga with the emplacement of the composite monzonitic-syenitic-granitic intrusions (Skiöld & Öhlander 1989, Martinsson et al. 1999), whereas some granites formed as late as at c. 1.86 Ga (Skiöld 1981, Skiöld & Öhlander 1989).

Intrusions of the Perthite Monzonite Suite are suggested to be comagmatic with the Kiirunavaara Group volcanic rocks. Both display a compositional variation from mafic to felsic combined with a relatively high content of alkali and HFS-elements. The intra-plate setting suggested for the Kiirunavaara Group is indicated by the chemical characteristics of the Perthite Monzonite Suite intrusions. Mantle plume origin is supported by the abundant occurrence of mafic-ultramafic complexes northwest of Kiruna, which possibly define the plume centre.

Lina Suite

Intrusions of the Lina Suite are extensive in northern Norrbotten where they typically occur as granite, pegmatite and aplite of mainly minimum melt composition generated by crustal melting. In Finland, they appear to form most of the volume of the Central Lapland Granitoid Complex (Fig. 1), and are also present as smaller intrusions in many areas across northern Finland (Lehtonen et al. 1998). However, the seismic appearance of the Central Lapland Granitoid Complex is inconsistent with this area as an intrusion-rich belt, and it may have a composition comparable with the supracrustal belts to the north and south (Patison et al. 2006).

The Lina Suite is composed of monzo-, syeno-granites, and adamellite, and is characterised by its restricted $SiO2$ range between 72 and 76 wt. %. It is peraluminous and a high content of Rb and depletion of Eu are characteristic.

The heat source generating the magmas might be the continent-continent collision events to the south and west (Öhlander et al. 1987b, Öhlander & Skiöld 1994, Lahtinen et al. 2003 2005) or the contemporaneous TIB 1 magmatism (Åhäll & Larsson 2000). Age determinations indicate a relatively large span in the emplacement age from 1.81 to 1.78 Ga for the Lina Suite (Huhma 1986, Skiöld et al. 1988, Wikström and Persson 1997b, Perttunen & Vaasjoki 2001, Rastas et al. 2001, Väänänen & Lehtonen 2001, Bergman et al. 2002).

A- and I-type intrusions

This is the youngest of the described intrusive suites and, in the west; it forms part of the Transcandinavian Igneous Belt (TIB). Two generations (c. 1.8 and 1.7 Ga) of intrusions belonging to the TIB exist in northern Sweden and adjacent areas of Norway. They commonly show quartz-poor monzonitic trends, and gabbroic-dioritic-granitic components are relatively common. (Gavelin 1955, Romer et al. 1992, 1994, Öhlander & Skiöld 1994)

Across northern Finland, the suite is represented by the Nattanen-type granitic intrusions dated at 1.80–1.77 Ga (Huhma 1986, Rastas et al. 2001). They form undeformed and unmetamorphosed, multiphase, peraluminous, F-rich plutons which sharply cut across their country rocks. Their Nd and Hf isotopic ratios indicate a substantial Archaean component in their source.

In northern Norrbotten, monzonitic to syenitic rocks give ages between 1.80 and 1.79 (Romer et al 1994, Bergman et al. 2001), whereas granites range from 1.78–1.77 and 1.72–1.70 Ga (Romer et al. 1992). Further south, the age of the granitic Ale massif in the Luleå area is 1802±3 Ma and 1796±2 Ma for the core and the rim of the massif, respectively (Öhlander & Schöberg 1991). This is similar to the 1.80 Ga age of Edefors type monzonitic to granitic rocks (Öhlander & Skiöld 1994).

This suite can be classified as a quartz monzodiorite–quartz monzonite–adamellite–granite suite and shows a metaluminous to peraluminous trend with alkaline affinity (Ahl et al. 2001). Lithophile elements are enriched in this suite, e.g. Zr is strongly enriched in the Edefors granitoids (Öhlander & Skiöld 1994).

Characteristic for the 1.8 Ga monzonitic to syenitic rocks is the occurrence of augite and locally also orthopyroxene and olivine demonstrating an origin from dry magmas (Ödman 1957, Öhlander & Skiöld 1994, Bergman et al. 2001). The Transscandinavian Igneous Belt (TIB) has been suggested to have formed in response to eastward subduction (Wilson 1980, Nyström 1982, Andersson 1991, Romer et al. 1992, Weihed et al. 2002), possibly during a period of extensional conditions (Wilson et al. 1986, Åhäll & Larsson 2000). The Edefors granitoids are interpreted as products of plate convergence and a mantle source is suggested for these rocks based on Sm-Nd isotopic characteristics. Mafic magmas may have formed by mantle melting in an extensional setting caused by a 1.8 Ga collisional event following northward subduction. These magmas were subsequently contaminated with continental crust and crystallised as monzonitic to granitic rocks (Öhlander & Skiöld 1994).

The related plate-tectonic setting may also be that of the final orogenic collapse, decompression and/or thermal resetting in the terminal stages of the orogenic development, following the continent-continent collisional stage (Lahtinen et al. 2003, 2005).

Deformation and metamorphism

The Palaeoproterozoic rocks in the northern part of the Fennoscandian Shield have undergone several phases of deformation and metamorphism. Metamorphic grades vary from greenschist to granulite facies. A sequence of ductile deformation events in central Finnish Lapland is reported in Hölttä et al. (2007) and Patison (2007) and references therein. The earliest foliation (S1) is bedding-parallel and can be seen in F2 fold hinges and as inclusion trails in andalusite, garnet and staurolite porphyroblasts. The main regional foliation S2 is axial planar to tight or isoclinal folds. It is mostly gently dipping to flat-lying, and suggested to have been caused by horizontal movements related to thrust tectonics, e.g. along the Sirkka Shear Zone. The elongation lineation generally trends NNE-SSW, and the movement direction was from SSW to NNE. The S-dipping Sirkka Shear Zone is composed of several sub-parallel thrusts and fold structures at the southern margin of the Central Lapland Greenstone Belt. This NNE-directed thrusting occurred during D1-D2, with a maximum age of c. 1.89 Ga (Lehtonen et al. 1998), and was contemporaneous with S- to SW-directed thrusting of the Lapland Granulite Belt in the north. This thrusting geometry is consistent with data from recent seismic reflection studies (Patison et al. 2006). The D2 and earlier structures are overprinted by sets of late folds, collectively called F3-folds, with a variety of orientations. It is possible that some earlier-formed structures were reactivated during D3. A minimum age for the D3 deformation is given by post-collisional 1.77 Ga Nattanen-type granites. This age is also the maximium age for D4, which is characterised by discontinuous brittle shear zones.

Ductile deformation in Sweden includes at least three phases of folding and also involves the formation of major crustal-scale shear zones. The intensity of deformation varies from a strong penetrative foliation to texturally and structurally well preserved rocks both regionally and on a local scale. Axial surface trace of the folds mainly trends in a SE or a SSW direction (Bergman et al. 2001). Locally, they interfere in a dome and basin pattern but more commonly either trend is dominant. The difference in the intensity of deformation shown by intrusions of the Haparanda Suite and the Perthite Monzonite Suite suggests an event of regional metamorphism and deformation at c. 1.88 Ga in northern Norrbotten (Bergman et al. 2001), corresponding to D1–D2 in Finland. Evidence for an episode of magmatism, ductile deformation and metamorphism at c. 1.86–1.85 Ga from the Pajala area in the northeastern part of Norrbotten has been presented by Bergman et al. (2006). A third metamorphic event at 1.82–1.78 Ga is recorded by chronological data from zircon and monazite in the same area. Movement along the Pajala-Kolari Shear Zone occurred during this event.

Major ductile shear zones in Sweden are represented by the NNE-trending Karesuando-Arjeplog deformation zone, the N to NNE-directed Pajala-Kolari Shear Zone and the NNW-directed Nautanen deformation zone (Appendix 1). The Pajala-Kolari Shear Zone has been given a major significance as representing the boundary between the Karelian and Norrbotten Cratons (Lahtinen et al. 2005). These major shear zones show evidences to have been active at c. 1.8 Ga. In general the shear zones in the western part show a western-side-up movement whereas the shear zones in the eastern northern Norrbotten are characterised by an eastern-side-up movement (Bergman et al. 2001).

One striking feature is that several of the crustal-scale shear zones are associated with abrupt changes in metamorphic grade, indicating that these zones have been active after the peak of regional metamorphism. Moreover, many of the epigenetic Au and Cu-Au deposits also show a strong spatial relationship with these major shear zones, although their local control are the second- to fourth-order faults and shear zones. Geochronology and structural evidence indicate late- to post-peak metamorphic conditions for many of the epigenetic Cu-Au occurrences in Sweden, whereas close to syn-peak metamorphic timing has been suggested for most of the occurrences in Finland (Mänttäri 1995, Eilu et al. 2003), although very few age dates exist for mineralisation in Finland

The metamorphic grade mainly is of low- to intermediate-pressure type, in Sweden generally varying from upper-greenschist to upper-amphibolite and in Finland from lower-greenschist to upper-amphibolite facies . Granulite facies rocks are only of minor importance, except for the northern Finnish Lapland and Kola Peninsula with the arcuate Lapland Granulite Belt (Fig. 1).

Regional metamorphic assemblages in metaargillites and mafic metavolcanic rocks, interpreted to be of Svecofennian age and generally indicate that the metamorphism is of low to medium pressure type, 2–4 and 6–7.5 kbar, under temperatures of 510–570°C and 615–805°C, respectively. High T – low P regional metamorphism characterise large areas of Norrbotten, but as pointed out by Bergman et al. (2001), the measured pressures and temperatures are not constrained in time and could be related to different metamorphic events. Still the geochronology of the metamorphic history in northern Sweden is rather sparse and the distribution in time and space is not well-known.

Bergman et al. (2001) divided the pre-1.88 Ga rocks in northernmost Sweden into low-, medium- and high-grade areas following the definitions of Winkler (1979). It is interesting to note that most of the low-grade areas there (i.e. Kiruna, Rensjön and Stora Sjöfallet) are located in the westernmost part of Norrbotten whereas the majority of medium to high grade metamorphic rocks are located in the central to eastern part where also the vast majority of the Lina type granites (c. 1.81 to 1.78 Ga) are situated. The strong spatial relationship between the higher-grade metamorphic rocks and the S-type granites is either a result of deeper erosional level of the crust in these areas or reflects areas affected by higher heat flow at c. 1.8 Ga.

In central Finnish Lapland the following metamorphic zones have been mapped (Hölttä et al. 2007): I) granulite facies migmatitic amphibolites south of the Lapland Granulite Belt, II) high pressure mid-amphibolite facies rocks south of the zone I, characterised by garnet-kyanite-biotite-muscovite assemblages with local migmatisation in metapelites, and garnet-hornblende-plagioclase assemblages in mafic rocks, III) low-pressure mid-amphibolite facies rocks south of the zone II, with garnet-andalusite-staurolite-chlorite-muscovite assemblages with retrograde chloritoid and kyanite in metapelites, and hornblende-plagioclase-quartz±garnet in metabasites, IV) greenschist facies rocks of the Central Lapland Greenstone Belt, with fine-grained white mica-chlorite-biotite-albite-quartz in metapelites, and actinolite-albite-chlorite-epidote-carbonate in metabasites, V) prograde metamorphism south of the zone IV from lower-amphibolite (andalusite-kyanite-staurolite-muscovite-chlorite±chloritoid schists), to mid-amphibolite facies (kyanite-andalusite-staurolite-biotite-muscovite gneisses, and upper amphibolite facies garnet-sillimanite-biotite gneisses, VI) amphibolite facies pluton-derived metamorphism related with heat flow from central and western Lapland granitoids.

The present structural geometry shows an inverted gradient where pressure and temperature increase upwards in the present tectonostratigraphy from greenschist facies in the zone IV through garnet-andalusite-staurolite grade in the zone III and garnet-kyanite grade amphibolite facies in the zone II to granulite facies in the zone I. The inverted gradient could be explained by crustal thickening caused by overthrusting of the hot granulite complex onto the lower grade rocks. Metamorphism in the Lapland Granulite Belt occurred at 1.91–1.88 Ga

(Tuisku & Huhma 2006), but the present metamorphic structure in central Finnish Lapland may record later, postmetamorphic thrusting and folding events (Hölttä et al. 2007).

Ore deposits

Introduction

The northernmost Finland, Norway and Sweden are characterised by mafic to ultramafic intrusion-hosted Cr, Fe-V-Ti and Ni±Cu±PG ores, VMS Cu-Zn, epigenetic Cu±Au and Au, and Fe oxide ± apatite ores (Table 1; Weihed et al. 2005). Based on the style of mineralisation, alteration and structural control, the region has been regarded as a typical Fe oxide Cu-Au (IOCG) ore province (e.g. Martinsson 2001, Williams et al. 2003). Similarly, especially the southern part of the region in Finland can be seen as a globally major mafic intrusion-hosted magmatic ore province (e.g. Peltonen 1995, Lamberg 2005).

Table 1. Examples of major metal deposits in northern Finland and Sweden. Tonnage gives the pre-mining resource.

Ore type	Occurrence	Tonnage (Mt)	Grade	Ore minerals	Age of ore (Ga)
Mafic-ultramafic igneous rocks hosted	Kemi[1]	158	19.7 % Cr	Cr	2.44
	Siika-Kämä[2]	43.1	3.51 ppm PGE+Au	Cp, Ml, Bo, Pn	2.44
	Ahmavaara[3]	60.0	0.27 % Cu, 1.86 ppm PGE+Au	Po, Cp, Pn, Pg	2.44
	Konttijärvi[4]	38.8	0.17 % Cu, 2.32 ppm PGE+Au		
	Mustavaara[5]	43.4	21.5 % Fe, 5 % Ti, 0.2 % V	Mt, Il	2.44
	Kevitsa[6]	66.8	0.30% Ni, 0.43 % Cu, 0.64 ppm PGE+Au	Po, Cp, Pn, Pg	2.057
	Koitelainen[7]	130	15 % Cr, 1.1–1.6 ppm PGE	Cr, Il, Pg	2.44
	Ruossakero[8]	4.2	0.52% Ni	Mi	2.7?
	Liakka[9]	0.25	0.37 % Ni, 0.78 % Cu	Cp, Po, Pn	
	Lappvattnet[10]	1.1	1.0 % Ni, 0.2 % Cu		1.88
	Karenhaugen[11]	1.0	0.6 Cu, 1.2 ppm PGE	Bn, Cc, Cp, Cr	c. 2.0
VMS	Viscaria[12]	12.54	2.29 % Cu, 0.50 % Zn	Cp, Mt, Po, Sp	2.05-2.20
	Huornaisen-vuoma[13]	0.56	4.8 % Zn, 1.7 % Pb, 0.2 % Cu, 12 ppm Ag	Sp, Gn, Mt, Cp	2.05-2.20
	Pahtavuoma[14]	21.4	0.3 % Cu, 0.67 % Zn, 10 ppm Ag	Po, Sp, Cp, Ap	2.05-2.20
Sediment hosted Cu	Repparfjord[15]	3.1	0.66 % Cu	Cp, Bo, Cc	2.05-2.20?
BIF	Bjørnevatn[15]	140	31 % Fe	Mt	>2.5
Fe oxide-apatite±REE	Kiirunavaara [12]	2180	47.7 % Fe	Mt, Ht	1.88
	Malmberget[13]	838	44.9 % Fe	Mt, Ht	1.88?
Fe oxide-Fe-sulphide-Cu	Stora Sahavaara[16]	145	43 % Fe, 0.08 % Cu	Mt, Po, Py, Cp	1.80
	Hannukainen[17,22]	66	42.5 % Fe, 0.25 % Cu, 0.3 ppm Au	Mt, Py, Po, Cp	1.80
	Rautuvaara Cu[17,22]	2.8	21.8 % Fe, 0.48 % Cu, 0.3 ppm Au	Mt, Py, Po, Cp	

Fe oxide-apatite-Cu±Au	Tjårrojåkka[18] Nautanen[18]	52.6 0.12	51.5 % Fe 1.87 % Cu, 1.1 ppm Au, 9 ppm Ag	Mt, Cp, Py, Bo Mt, Cp, Py, Bo	1.78 1.78?
Porphyry(?)	Aitik[12]	1616	0.38 % Cu, 0.2 ppm Au, 3.5 ppm Ag	Cp, Py, Po, Bo	1.89
Orogenic gold, normal	Suurikuusikko[19] Pahtavaara[20,22]	24.3 3.5	4.75 ppm Au 3.38 ppm Au	Ap, Py, Au Py, Au	1.89-1.80 1.89-1.80
Orogenic gold, atypical metal association	Pahtohavare[12] Saattopora[14,22] Bidjovagge[21]	1.72 2.163 2.0	0.9 ppm Au, 1.89 % Cu 2.9 ppm Au, 0.25 % Cu 3.6 ppm Au, 1.2 % Cu	Py?, Po, Cp, Au Po, Cp, Au Py?, Po?, Cp, Au	1.89-1.80
Atypical orogenic or syngenetic gold	Kuusamo deposits[22]	<01–0.8	1–14 ppm Au, 0.05–0.3 % Co, <0.05–0.3 % Cu, 0.003–0.1 % U	Po, Py, Au, Cp, Pn, Co, Un	1.89-1.80

Abbreviations: Ap = arsenopyrite, Au = native gold, Bo = bornite, Cc = chalcocite, Co = cobaltite, Cp = chalcopyrite, Cr = chromite, Ht = hematite, Il = ilmenite, Ml = millerite, Mt = magnetite, Pg = PG minerals, Pn = pentlandite, Po = pyrrhotite, Py = pyrite, Sp = sphalerite, Un = uraninite.

References: 1) Outokumpu Oyj (2005), Saltikoff et al. 2006, 2) Gold Fields Ltd press release, July 2003, 3-4) Gold Fields Ltd, July 2004, 5) Puustinen 2003, Adriana Resources (2006), 6) Mineral reserve (Scandinavian Minerals 2007), 7) Mutanen 1997, 8-9) GTK nickel database, 10) SGU Exploration Newsletter, Nov. 2004, 11) NGU web site, deposit id 2020.004, 12) Weihed et al. (2005), 13) SGU deposit database 2007, 14) Korvuo (1997), 15) NGU deposit database 2007, 16) Northland Resources (www.northlandresourcesinc.com/s/Stora.asp), 17) Puustinen (2003), 18) Edfelt et al. 2006, 19) Riddarhyttan Resources, press release 19 July 2005, 20) ScanMining, press release 15 Oct 2003, 21) Lindblom et al. (1996), 22) FINGOLD (2007).

Economically the most important for the region have been the apatite-iron ores with an annual production of about 31 Mt of ore from the Kiirunavaara and Malmberget mines and a total production of about 1600 Mt from 10 mines during the last 100 years. Equally important is the Kemi Cr mine which has produced about 8 Mt of chromium since the start of mining in 1966, and is the main cause for the presence of the Tornio stainless steel plant.

Iron ore has also been produced in smaller scale from the Rautuvaara and Misi areas in northern Finland and Sydvaranger in northeastern Norway. Copper has been produced intermittently during the 17th and 18th centuries in northern Sweden and Norway, but during the last 40 years copper and gold has been mined in larger scale in Sweden (Aitik, Viscaria, Pahtohavare), Finland (Saattopora, Pahtavaara) and Norway (Repparfjord, Bidjovagge). All the sulphide deposits are hosted by Palaeoproterozoic greenstones and are small to medium sized except for Aitik which is in Svecofennian volcaniclastic rocks and is a world class deposit with the current annual production of 18 Mt of ore and with a total tonnage >1000 Mt; the production is planned to double in year 2010.

Magmatic deposits of economic size have, so far, been defined only from the 2.44 Ga intrusions, in northern Finland. In addition to the Kemi Cr deposit, the Mustavaara Fe-Ti-V deposit has been exploited. In total, 13.6 Mt @ 0.2 % V was mined from Mustavaara, and the remaining reserves are at least 30 Mt at a similar grade (Adriana Resources 2006). In addition, large PGE and Cu-Ni occurrences have been found in several of the 2.44 Ga layered intrusions (Huhtelin 1991, Halkoaho 1993, Iljina 1994, Alapieti & Lahtinen 2002, Iljina & Hanski 2005) and in the 2.06 Ga Kevitsa ultramafic intrusion (Mutanen 1997). So far, none of the latter has been taken into production.

Mafic and ultramafic igneous rocks hosted deposits

As typical for shield areas, nickel, copper, platinum-group elements (PGE), chromium, vanadium, and titanium occurrences are hosted by mafic and ultramafic rocks in various geological settings in the Fennoscandian Shield. These metals occur as basal accumulations or stratiform horizons of chromite, PGE, and Fe-Ti-V oxides in layered intrusions and as disseminations of Fe-Ni-Cu-PGE sulphides in volcanic rocks. Four principal groups can be identified in the northern Fennoscandian Shield: (1) Archaean komatiite hosted deposits and showings in NW Finland, (2) the Tornio-Näränkävaara Belt of layered intrusions (Fig. 3) and related intrusions (c. 2.44 Ga) in the Central Lapland and NW Finland, (3) younger intrusions and volcanic rocks hosted deposits in the Central Lapland and (4) Haparanda Suite intrusion (1.88 Ga) hosted deposits. The Archaean komatiite hosted nickel deposit type is represented by the Ruossakero deposit in the northwesternmost Finland (Table 1).

Several styles of mineralisation have been discovered in the Tornio-Näränkävaara Belt (c. 2.45–2.42 Ga), and two economic oxide deposits have so far been mined: the Cr deposit in the Kemi intrusion and the Mustavaara Ti-V deposit in the Porttivaara block of the Western intrusion of the Koillismaa Complex (Alapieti & Lahtinen 2002, Iljina & Hanski 2005). There are PGE and chalcophile element occurrences in the Penikat and the Portimo and Koillismaa Complexes. These can be classified into three main types: (1) disseminated and massive PGE-enriched Cu-Ni sulphides of the marginal series, (2) reef-type PGE deposits, and (3) offset base-metal and PGE deposits in the footwall rocks. The first type is almost exclusively confined to well-developed, thicker marginal series in the Suhanko and Konttijärvi intrusions of the Portimo Complex (Figs 10-11 in Section 'Day 1') and in the Western intrusion of the Koillismaa Complex. In the former, both disseminated and massive concentrations of sulphides have been detected, whereas only disseminated sulphides have been discovered from the latter. The reef-type PGE enrichments in the Penikat intrusion and Portimo Complex can be divided into the major or principal enrichments which are laterally continuous and have PGE concentrations at least several ppm, and PGE showings which are less continuous and rarely grade above one ppm. The principal PGE reefs include the Sompujärvi (SJ), Ala-Penikka (AP) and Paasivaara (PV) reefs in the Penikat intrusion, and Siika-Kämä (SK) and Rytikangas (RK) reefs in the Portimo Complex. The SJ, PV, and SK are considered highly viable for economic exploitation. These and other reef-type PGE deposits have low, barely visible, concentrations of Cu, Ni and Fe sulphides; in places, the sulphides are virtually absent and, instead, chromite is present, as in the Sompujärvi and Siika-Kämä reefs.

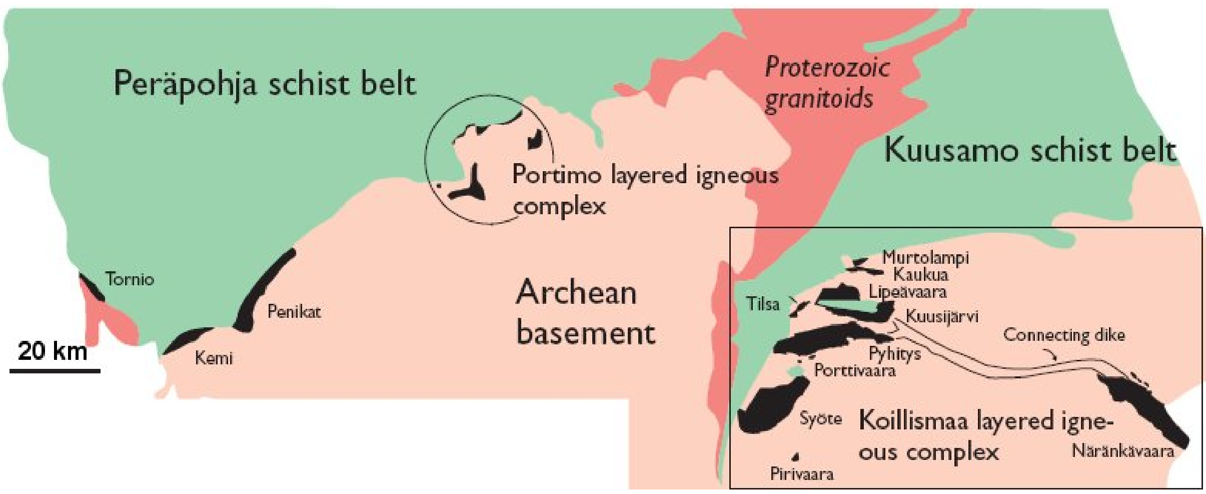

Figure 3. Mafic to ultramafic layered intrusions (black) forming the Tornio-Näränkävaara Belt. Modified from Iljina & Hanski (2005).

Tornio-Näränkävaara type of intrusions of similar age also occur in Central Lapland and NW Finland. The Central Lapland intrusions are represented by the Koitelainen and Akanvaara intrusions, which both have a very similar igneous stratigraphy as composing of an ultramafic lower part followed by gabbroic cumulates. The mineral showings within these two intrusions include the lower and upper chromitite layers and a magnetite gabbro layer, the latter resembling the one located in the Koillismaa Complex. In addition, also the Kaamajoki intrusion, close to the Ruossakero deposit in the northwesternmost Finland, contains copper-palladium showings (Heikura et al. 2004).

The Kevitsa Ni-Cu-PGE deposit represents a major mineral resource hosted by the younger mafic-ultramafic intrusions of 2.06 Ga in age. The reported resources are 66.8 Mt @ 0.30 wt.% Ni, 0.43 Cu and 0.64 ppm 2PGE+Au (Scandinavian Minerals 2007). Other reported mineral resources of about the same age are the Porsvann and Karenhaugen, Karasjok Belt, Norway (Karenhaugen: 1 Mt @ 0.6 wt.% Cu and 1.2 ppm Pt+Pd, NGU 2004), and the Lomalampi platinum-dominated PGE-Ni showing in komatiite, Central Lapland (Räsänen 2004).

The Haparanda Suite (1.90–1.86 Ga) intrusions range in modal composition from peridotite through gabbro, diorite and granodiorite to tonalite and are accompanied also by true granites; ultramafic cumulates are rare. The Haparanda Suite forms a calc-alkaline series instead of the older tholeiitic-komatiitic ones described above. Small and low grade, subeconomic, Cu-Ni-PGE mineralisation has taken place in many intrusions: Liakka (Finland) and Notträsk (Sweden) being located in the northern Fennoscandian Shield close the towns of Tornio and Luleå, respectively. At Liakka, the reported Ni-Cu deposit is hosted by the peridotitic cumulates at the bottom of the intrusion and the reserves are 0.25 Mt @ 0.37 wt.% Ni and 0.78 Cu (Inkinen 1990). The Lappvattnet intrusion (Table 1) in Sweden represents a different type of 1.88 Ga intrusions as resembling more the Kotalahti type (zone of intrusions with numerous nickel deposits in central Finland, Makkonen 1996) than the Haparanda Suite (Weihed et al. 2005).

Stratiform-stratabound sulphide deposits

Stratiform deposits of base metals are restricted to the Palaeoproterozoic greenstone successions where they occur in volcaniclastic units. The sulphide occurrences are tabular to blanket-shaped and consist of varying proportions of chalcopyrite, pyrrhotite, pyrite, sphalerite, galena and magnetite. The ore minerals occur disseminated, in breccia veins, and as massive intercalations in tuffite, black schist and carbonate rocks. Chert may occur as extensive beds up to 20 m thick.

The only stratiform sulphide deposit so far having had an economic importance is Viscaria at Kiruna with a production of 12.54 Mt @ 2.29 % Cu mined during 1982–1997 (Martinsson et al. 1997a). Other significant occurrences include the test-mined Pahtavuoma Cu-Zn deposit at Kittilä and the unmined Huornaisen¬vuoma Zn-Pb-Ag deposit in northeastern Norrbotten (Table 1). All these are suggested to be syngenetic in origin and formed by exhalative activity (Inkinen 1979, Martinsson et al. 1997a, Bergman et al. 2001). Genetically significant features at Viscaria include the blanket shaped and partly laminated style of mineralisation, the pronounced zonation defined by Cu and Zn, and the extensive footwall alteration zone. These characteristics are suggested to reflect deposition from a brine pool at the sea floor, in a situation similar to the Atlantis II Deep in the Red Sea (Martinsson et al. 1997a). In contrast to most of the epigenetic Cu deposits in the region, Au is almost absent and Zn is significantly enriched at Viscaria.

At Huornaisenvuoma, the ore is at the top of a thick dolomite unit in the uppermost part of the greenstone succession. Calc-silicates are developed as metamorphic minerals in the mineralised zone and in a restricted area in the footwall below the central part of the deposit. The ore minerals mainly occur in thin and stratiform massive layers but also disseminated in the mineralised zone. (Bergman et al. 2001)

The Pahtavuoma deposit is dominantly stratabound, with several ore bodies within a sequence of greywacke, phyllite, black schist, mafic tuffite and lava, and chert. The Cu-rich ore bodies contain 6.5 Mt @ 0.84 % Cu and 21 ppm Ag, 17 Mt @ 0.81 % Zn and the separate uranium ore bodies 0.14 Mt @ 0.39 % U, 24 ppm Ag, 0.24 % Cu (Inkinen 1979, Korkalo 2006). The ore minerals occur as dissemination and breccia veins. Copper is enriched in the central and stratigraphically lower part of the deposit, whereas Zn is enriched in the hanging wall, in the lateral extension from the copper ores and as separate ore bodies. The vein-type U mineralisation is mostly closely related to the Cu ore bodies which also are weakly enriched in Co, As, Ag and Mo (Inkinen 1979, Korkalo 2006).

Skarn-like iron deposits

Lens- and irregular-shaped iron occurrences consisting of magnetite, and Mg and Ca-Mg silicates are common within the greenstones in Sweden and the westernmost northern Finland (Table 1). Some of the deposits are appears as being spatially associated to oxide- and silicate-facies BIF. These skarn or skarn-like iron deposits occur in association to tuffite, black schist and dolomitic marble and in Sweden are mainly located to the upper parts of the greenstone piles. They have a size of up to 145 Mt and an iron content of 35–50 % (eg. Stora Sahavaara, Table 1). Disseminated pyrite, pyrrhotite and chalcopyrite are commonly present, and the sulphur content is in the range of 1–5 %. The concentration of Cu typically is less than 0.1 %. Phosphorous varies from 0.02 to 0.1 % with a few more P-rich exceptions (Grip and Frietsch 1973, Hiltunen 1982, Niiranen 2005). Some of the deposits appear to grade into BIF towards the hanging wall and/or along strike. The occurrences have been suggested to be metamorphic expressions of originally syngenetic exhalative deposits (Grip and Frietsch 1973, Bergman et al. 2001) or intrusion-related skarn deposits (Hiltunen 1982). The latest work for the Kolari deposits strongly suggests that they, and similar deposits at Pajala in Sweden, are epigenetic, and would best fit into the IOCG category (Niiranen et al. 2007).

Two skarn-like iron deposits have been mined in the Kolari area in northwestern Finland, where the original reserves were reported to have been about 85 Mt of ore, and in the Misi region in southern Finnish Lapland (Nuutilainen 1968, Hiltunen 1982, Niiranen et al. 2005, Niiranen et al. 2007). Significant amounts of Cu and Au have been recovered from the magnetite rock of the Laurinoja ore body of the Hannukainen mine (Hiltunen 1982). Ore-grade Au and Cu is also reported from the magnetite-rock hosted Rautuvaara and Kuervitikko deposits in Kolari (Niiranen et al. 2007). The deposits in the Kolari area are discussed in more detail in the excursion locality descriptions of this guide book.

Apatite-iron deposits

Kiruna is the type area for apatite-iron ores (iron ores of Kiruna type, defined by Geijer 1931) with the Kiirunavaara deposit as the largest and most well known example (Table 1). In total, about 40 apatite-iron deposits are known from northern Norrbotten and about 1600 Mt of ore have been mined from 10 deposits during the last 100 years. This type of deposits is mainly spatially restricted to areas occupied by the Kiirunavaara Group rocks, and very few occurrences exist outside the Kiruna-Gällivare area (Fig. 4). Individual deposits have an average content of Fe and P varying between 30–70 % and 0.05–5 %, respectively. Besides magnetite and hematite, most deposits contain significant amounts of apatite and they are generally strongly enriched in LREE (Frietsch & Perdahl 1995).

The apatite-iron ores exhibit a considerable variation in host rock composition, host rock relationship, alteration, P content, and associated minor components. It is possible to distinguish two rather distinct groups of deposits: breccia type and stratiform-stratabound type. A third and less distinct group has features similar to both the other two groups, and is in many respects intermediate in character between them (Bergman et al. 2001).
Breccia-type apatite-iron ores mainly are hosted by intermediate to mafic volcanic rocks, in a stratigraphically low position of the Kiirunavaara Group or within the underlying Porphyrite Group. Amphibole is always present as a minor component and accessory amounts of pyrite, chalcopyrite and titanite may be encountered. Host rock alteration is not reported to be a prominent feature, but albite and scapolite alteration seem to be rather common, whereas sericite, epidote and tourmaline are less important alteration minerals. Characteristic for the breccia-type deposits is their low P content (typically in the range of 0.05–0.3%) and an average iron content of only about 30%. However, the central parts of larger deposits can be higher-grade (e.g. Mertainen). With a few exceptions, magnetite is the only iron oxide present. (Martinsson 2003)

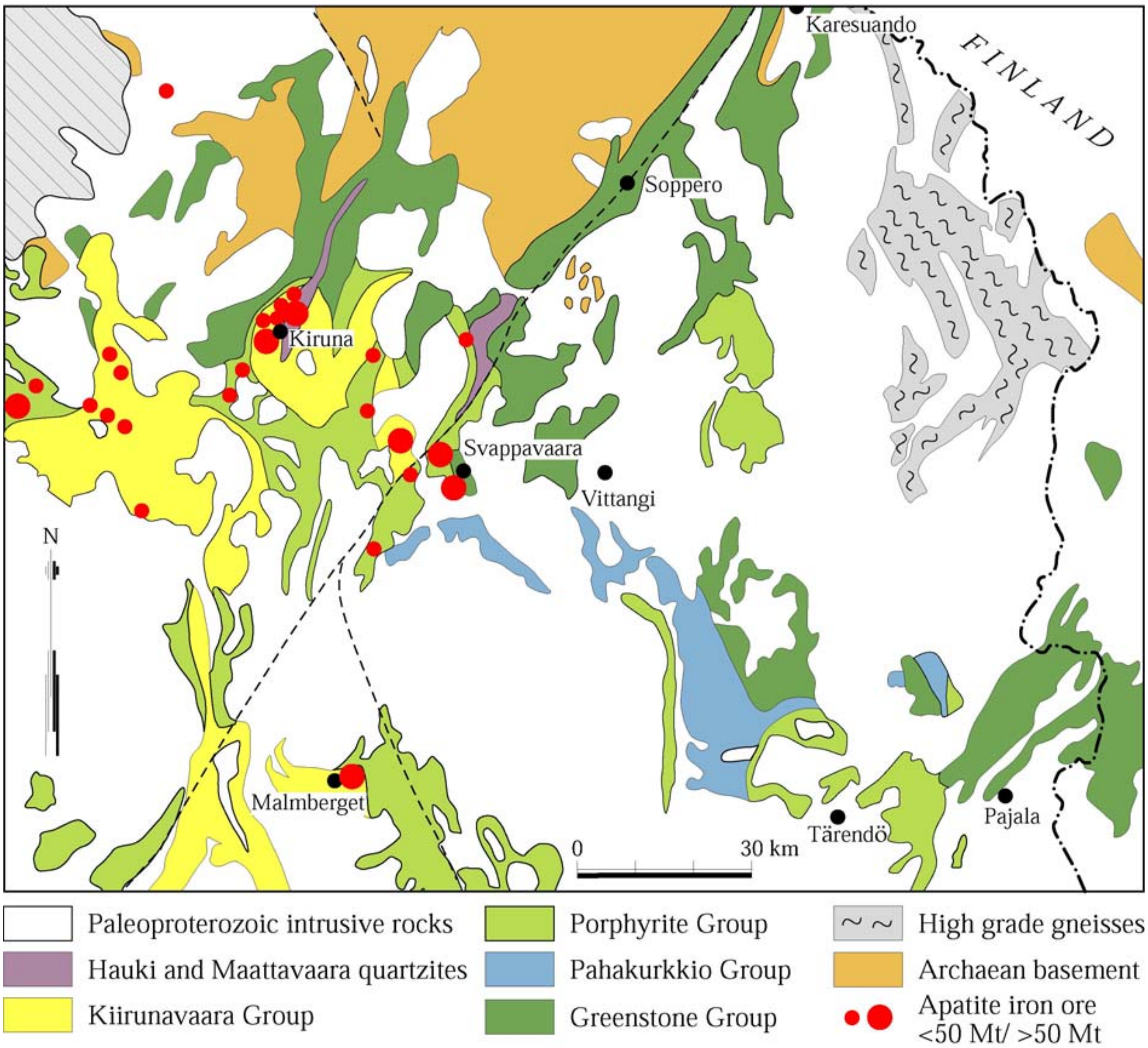

Figure 4. Generalised geology and apatite-iron deposits in northern Norrbotten, Sweden.

The stratiform-stratabound type comprises lenses at stratigraphically high positions within the Kiirunavaara Group. They have hematite as the major iron oxide together with varying amounts of magnetite. These deposits also have a high P content, at 1–4.5%. Amphibole is absent, and the main gangue minerals are apatite, quartz and carbonate. Host rock alteration is common with sericite, biotite, tourmaline and carbonate as typical products. The hanging wall rocks may also be strongly silicified. Sulphides are rare and mainly found in small amounts in the altered footwall or as crosscutting late veinlets within the ores. Ore breccia is absent or restricted to the footwall. Included into this group are the Per Geijer ores (i.e. Nukutus, Henry, Rektorn, Haukivaara, and Lappmalmen in the central Kiruna area) and Ekströmsberg. (Martinsson 2003)

The intermediate types of apatite-iron ore are dominantly stratabound in character but also have ore breccia developed along the wallrock contacts (e.g. Kiirunavaara and Tjårrojåkka). Magnetite is the dominant, or the sole iron oxide. Amphibole is a characteristic minor component and titanite may be present in accessory amounts. The Fe content is high (55–67 %), and the average P-content generally low, although it may vary considerably (from 0.01 to

> 5 %) within individual deposits. Alteration is not well defined, but includes the formation of albite, amphibole, biotite, sericite and, locally, scapolite or tourmaline (Bergman et al. 2001). Ages for Kiruna-type magnetite-apatite ores are only published from the Kiirunavaara area. In the footwall to the Luossavaara deposit, titanite exists together with magnetite in the form of coarse-grained magnetite dykes within biotite-chlorite altered trachyandesite. Large platy crystals of titanite from these veins have given an U-Pb age of 1888±6 Ma (Romer et al. 1994), whereas granophyric to granitic dykes crosscutting the Kiirunavaara ore have an U-Pb zircon age of 1880±3 Ma (Cliff et al. 1990). This suggests that the Kiruna-type magnetite-apatite ores were formed between 1.89 and 1.88 Ga.

Apatite-iron ores have been suggested to represent an iron-dominated and sulphide-poor end member of the Fe oxide Cu-Au class of metallic ore deposits (Hitzman et al. 1992) which makes them important not only as sources of iron, but also for the metallogenetic understanding of the Northern Norrbotten Fe-Cu-Au province. The genesis of the apatite-iron ores has been discussed for more than 100 years and is still controversial. Suggested models include sedimentary, hydrothermal or magmatic processes. Most features of the ores are compatible with either a magmatic intrusive origin or a hydrothermal origin. Probably both magmatic and hydrothermal processes have been active explaining the large variation in mineralisation style recognised within and between individual deposits. Most of the massive deposits are suggested to have a mainly magmatic origin with a minor overprinting hydrothermal phases altering the wall rocks and forming veins. Some deposits may represent transitional forms between a magmatic and hydrothermal origin similar to that at Lightning Creek in the Cloncurry area in Queensland, Australia (Perring et al. 2000).

Sulphides are mostly rare constituents in the apatite-iron ores and occur disseminated or in veinlets. Significant Cu mineralisation is found spatially associated with apatite ores in only a few places (e.g. Tjårrojåkka and Gruvberget). A genetic relationship between Cu and iron oxide mineralisation is suggested at Tjårrojåkka (Edfelt 2007). At Gruvberget, the relationship might be more of a coincidence with the Cu occurrence representing a separate and later event with the iron ore only acting as a chemical-structural trap (Lindskog 2001). The U-Pb titanite ages indicate that Cu mineralisation at Tjårrojåkka and Gruvberget is of ca. 1.8 Ga in age (Billström & Martinsson 2000, Edfelt & Martinsson 2005) which is significantly younger than the suggested 1.9 Ga emplacement age for apatite-iron ores in the central Kiruna area.

Epigenetic Au and Cu-Au deposits

Epigenetic sulphide deposits in the northern part of the Fennoscandian Shield form a heterogeneous group with extensive variation in the style of mineralisation, metal association, host rock, and in the variation in possible genetic types. Most deposits are hosted by tuffitic units of the Palaeoproterozoic greenstones (mostly in Finland, e.g., in CLGB and Kuusamo) and mafic to intermediate volcanic rocks within the Svecofennian porphyries (mostly in Sweden, e.g., in Porphyrite and Kiirunavaara Groups). For the latter, a number of the latter display a close genetic and/or spatial relation to orogenic, 1.9–1.8 Ga, felsic to intermediate intrusive rocks, and magnetite is a common minor component in some occurrences and locally they occur adjacent to, or are hosted by, major magnetite deposits. On the other hand, for the greenstone-hosted deposits, there appears no connection to intrusion and most of the deposits show total destruction of iron oxides during mineralisation. A close spatial relationship with regional shear zones is common with second- to fourth-order structures typically controlling the location of an occurrence. Besides structural traps, also chemical traps may be important with redox reactions involving an originally high graphite or iron content of the host rock to trigger sulphide precipitation. Many are gold-only occurrences, but almost equally as many contain significant Cu in addition to gold. Also, some occurrences contain Co in economic to subeconomic amounts. Other elements that are significantly enriched in a few cases include Fe, LREE, Ba, U and Mo. The latter elements typically are enriched in the Au-Co±Cu occurrences, but very rarely in cases where gold is the sole major commodity. There also are deposits, in the Alta area in northern Norway, where Cu is the sole important commodity. (Ettner et al. 1994, Eilu et al. 2003, Sundblad 2003, Eilu & Weihed 2005, Weihed et al. 2005, FINGOLD 2007)

The importance of saline hydrothermal fluids to explain the origin of regional albite-scapolite alteration and the nature of the Au deposits also significantly enriched in other metals has been emphasized by Frietsch et al. (1997) and Vanhanen (2001). Highly saline fluid inclusions with 30–45 eq. wt % NaCl and depositional temperatures of 300–500°C are recorded for the Cu-Au deposits in this region (Ettner et al. 1993, 1994, Lindblom et al. 1996, Broman and Martinsson 2000, Wanhainen et al. 2003a, Edfelt et al. 2005, Niiranen et al. 2007). However, a prominent E-W trend in Au/Cu ratio exhibited by the epigenetic Cu-Au occurrences in a profile from Kiruna to Kuusamo may reflect some regional differences in fluid composition (Fig. 5). The more abundant occurrence of scapolite as an closely ore-related alteration mineral in Sweden and the more frequent carbonate alteration in Finland may also suggests fundamental differences in fluid characteristics going from west to east, but could also be due to general change in regional metamorphic grade between the mineralised belts.

Age data from Cu-Au deposits and related hydrothermal alteration from northern Norrbotten demonstrates two major events of ore formation at c. 1.87 Ga and 1.77 Ga, respectively (Billström & Martinsson 2000). Similar results are obtained from deposits in the northern parts of Norway and Finland with a third probable stage of mineralisation at ca. 1.84–1.80 Ga (Bjørlykke et al. 1990, Mänttäri 1995, Niiranen et al. 2007). These events are temporally related to magmatic, deformation and metamorphic episodes of regional importance, that is, directly related to major orogenic stages of the evolution of the Fennoscandian Shield (Lahtinen et al. 2005, Patison et al. 2006).

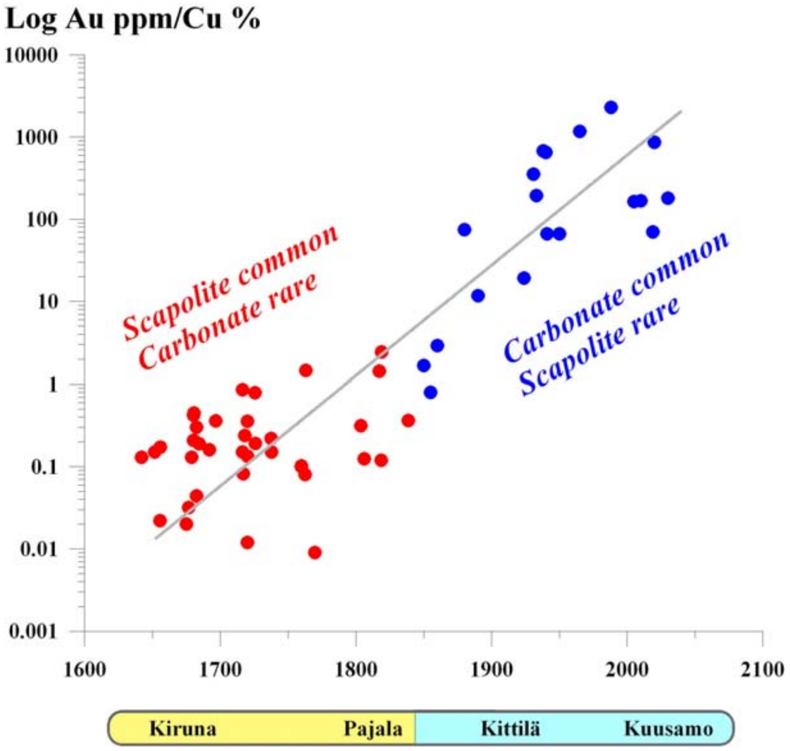

Figure 5. Au-Cu ratio in epigenetic Cu-Au occurrences in an E-W transect from Kiruna to Kuusamo. Note, however, that in greenstone or schist-belt scale there is abundant scapolite also across northern Finland, also in the Kuusamo schist belt. Data from Nurmi et al. (1991), Eilu (1999), and unpublished data. Horizontal scale in kilometres.

Greenstone-hosted deposits

A number of Palaeoproterozoic greenstone-hosted Cu±Au deposits have been mined since the 17th century in the northern Sweden and Norway. Most of them are very small but during the last 20 years three deposits have been mined in a larger scale producing both Cu and Au (Bidjovagge Au-Cu in Norway, Pahtohavare Cu-Au in Sweden and Saattopora Au-Cu in Finland). These three deposits and several other subeconomic occurrences are characterised by the metals Cu-Au±Co±U and the lithological association of mafic to intermediate tuffite, black schist, carbonate rocks, chert and dolerite.

The Pahtohavare deposit (Martinsson et al. 1997b) is in may ways similar to Bidjovagge (Ettner et al. 1993, 1994) and Saattopora (Korvuo 1997). Typical is the strong premineralisation albitisation of the host rocks which include graphitic schist, mafic to intermediate volcaniclastic rocks and mafic sills. Biotite-scapolite alteration typically envelopes the albite-rich zone that contains chalcopyrite and pyrite as dissemination or breccia filling and veins together with carbonate, quartz, albite and, locally, scapolite.

These occurrences share many features with both IOCG and orogenic gold styles of mineralisation, such as structural and lithological control. Features similar to IOCG deposits (as defined by Hitzman 2000), but different to typical orogenic gold deposits, include metal and sulphide association, saline fluids, and multistage alteration. However, the deposits share more features with the orogenic deposits (sensu Groves et al. 1998): there is no direct timing or genetic relationship to intrusion, style of alteration directly related to the mineralisation stage, style of structural control, destruction of all Fe oxides during mineralisation, most of the mass transfer (gains and losses of metals, semimetals and volatiles), and many occurrences, including the by far largest one (Suurikuusikko) are gold-only deposits (Ettner et al. 1993 and 1994, Lindblom et al. 1996, Eilu et al. 2003, Weihed and Eilu 2005, Eilu et al. 2007, Patison et al. 2007). For details on the Pahtavaara and Suurikuusikko deposits, see the deposit description below in this guide book.

Carbonate-quartz vein-type deposits containing chalcopyrite or locally chalcocite in a gangue of quartz, ferro-dolomite or calcite are common within the greenstones. They typically are hosted by dolerite, mafic to ultramafic volcanic rocks and intermediate sedimentary rocks. Ore-related alteration is characterised by carbonatisation, sericitisation and biotitisation. The vein-type deposits lack economic importance in Sweden although they attracted prospectors in the 17th and 18th centuries due to the locally high Cu-grade within these deposits. In northern Norway, there are a few more important vein-type deposits (Porsa and Kåfjord) with mining during the years 1825–1931 (Bugge 1978). In Finland, the small Kivimaa Cu-Au deposit in the Peräpohja schist belt was mined in 1969 (Rouhunkoski & Isokangas 1974).

Albitisation and carbonatisation are common and extensive in greenstones in the CLGB (Eilu 1994, Eilu et al. 2007) and Kuusamo (Vanhanen 2001). Tens of Au-only, orogenic, shear-zone controlled occurrences have been discovered in these areas (FINGOLD 2007). Several of them show similarities to the Bidjovagge-Pahtohavare type as they occur in albite- and carbonate-altered komatiites and basaltic rocks. Gold occurs together with pyrite, arsenopyrite, pyrrhotite and chalcopyrite in quartz veins typically hosted by altered komatiites, basalts, phyllites and tuffites, but also disseminated in the host rocks. Albite-carbonate altered felsic porphyry dikes may also be host rocks to these Au deposits (Härkönen & Keinänen 1989). In Finland, the most significant of the shear zones related to epigenetic mineralization is the W- to NW-trending, Sirkka Shear Zone traversing across the central Finnish Lapland and the CLGB (Lehtonen et al 1998, Eilu et al. 2003). More than 25 Au and Au-Cu deposits and occurrences have so far been discovered within the Sirkka Shear Zone and lower-degree faults branching from this crustal-scale, >100 km long, structural break.

Genetic considerations on greenstone-hosted deposits

The obvious problem in fitting base metal-enriched occurrences, multistage regional alteration and saline fluids of the northern Fennoscandian gold deposits into the orogenic gold category could, perhaps, be best explained by applying the 'atypical metal association' concept defined by Goldfarb et al. (2001). In that scenario, deformation of older, intracratonic basins are included into the processes of orogenic mineralisation, the ore fluids may become anomalously saline, and produce base metal-rich orogenic gold deposits. Goldfarb et al. (2001) suggest this as an explanation for the formation of base metal-enriched gold deposits in the Sabie–Pilgrim's Rest in South Africa, and in Tennant Creek, Pine Creek and Telfer in Australia. Also the northern part of the Fennoscandian Shield is characterised by Palaeoproterozoic rifted basins of intracratonic setting, extensive supracrustal sequence and probable evaporates. These sequences accumulated, compacted, were intruded by magmas, and were subject to several major stages of alteration predating gold mineralisation. Hence, there was an exceptionally wide range of rock types, structures, and fluid and metal sources to be subjected to the orogenic processes. When an orogenic fluid met such an sequence, during 1.92–1.88 and/or 1.85–1.79 Ga (Lahtinen et al. 2005), it became more saline when meeting the possible evaporates and connate brines, became able to leach and transport both gold and base metals. Eventually, the metals were precipitated when the brines met structural and chemical traps in the greenstone belts. On the other hand, where the fluids did not become saline, gold-only occurrences were formed.

Most difficult is to put the epigenetic deposits of the Kuusamo schist belt (KSB) into the orogenic category. In the KSB, there are several Au-Co-Cu±U occurrences which by their metal association and alteration are clearly different to the Bidjovagge-Pahtohavare-Saattopora type. Orogenic gold with atypical metal association, iron oxide-copper-gold, and syngenetic style have been suggested for the gold-cobalt-copper ± uranium occurrences at Kuusamo. Structural control and timing seem to fit with the orogenic hypothesis, alteration, metal association, necessary mineralising fluid(s) and structural control with both the IOCG and orogenic gold with atypical metal association hypothesis, whereas mineralising fluid(s) and the rift to self and host rock settings with the syngenetic (metamorphosed) hypothesis. Gold fineness may fit with any of the genetic styles proposed. (Pankka and Vanhanen 1992, Eilu 1999, Vanhanen 2001, David Groves pers. comm. May 2006)

Cu-Au deposits in Svecofennian rocks

Number of minor prospects for Cu, Mo, and Au in northern Sweden are hosted by felsic to intermediate intrusive rocks (Walser and Einarsson 1982). At this time, it was realised that the formation of porphyry Cu-style mineralisation was not exclusive to Phanerozoic terranes, that under favourable conditions they could be preserved in Precambrian orogenic belts, too (Weihed 1992, Sikka and Nehru 1997 and references therein). In the Fennoscandian Shield, there are now several examples of low-grade, intermediate-tonnage, occurrences of Cu±Mo±Au that have been described as porphyry Cu-style deposits (e.g., Gaal and Isohanni 1979, Weihed 1992 and references therein, Weihed 2001, Wanhainen et al. 2003a). Some of these have been attributed to alternative genetic models and some have been suggested to be related to the IOCG family of deposits. The Aitik Cu-Au-Ag-Mo deposit is the only major sulphide deposit hosted by Svecofennian rocks in northern Sweden and Finland. It has recently been interpreted as a metamorphosed Porphyry Cu deposit overprinted by later IOCG-style of mineralization (Wanhainen et al. 2005).

Deposits vary in character from the large disseminated ore body at Aitik to small high-grade vein occurrences such as Lieteksavo. In Sweden, they are in the Porphyrite Group and the Kiirunavaara Group rocks and are characterised by alteration producing K feldspar, scapolite, biotite, minor tourmaline and, in places, sericite. Extensive albitisation is only locally developed and then mainly in association with intermediate to felsic intrusions. The main ore minerals are pyrite and chalcopyrite with magnetite as a minor to major constituent in most occurrences. Several deposits also contain bornite and minor amounts of molybdenite. Pyrite with a high Co content occurs in a few deposits and hematite may be present as a minor component. The ore minerals occur as disseminated, in quartz-tourmaline veins, veinlets and breccias. Generally, occurrences are within areas dominated by K feldspar alteration, whereas scapolite-biotite alteration may be more important outside the mineralised area. The paragenetic sequence from oldest to youngest is mostly: scapolite + biotite → K feldspar → sericite → tourmaline. Stilbite and chabazite may occur as the latest phases in druses and veins together with calcite. Ore minerals mainly are associated with the intermediate or late stages of alteration. Bornite and chalcocite are commonly paragenetically late and related to tourmaline and zeolites. A low sulphur content, with bornite, chalcocite and magnetite as important ore minerals, characterises some of the occurrences. (Frietsch 1966, Bergman et al. 2001, Wanhainen et al. 2003a, 2005)

Excursion Route and Road Log

The excursion route starts (Fig. 1.) from the Rovaniemi town which is located at the joining point of two large rivers: Kemijoki and Ounasjoki. First we drive south over the Peräpohja schist belt to the Suhanko area just on the Archaen Pudasjärvi block. Then we drive north back to Rovaniemi and continue to Luosto ski resort over the Central Lapland granitoid complex. Most of the fells, like the Luosto fell, and higher hills are quartzites. Next day we drive further north through the Sodankylä town to the Pahtavaara Au mine which in the foot of a carbonatised komatite hill. Then we drive to the Kevitsa PGE project. From Kevitsa we head first back to Sodankylä and then turn west to Kittilä. We stay overnight in the Levi ski resort at Sirkka next to the E-W trending Levi quartzite fell chain. Following day we go to the Suurikuusikko Au deposit (Kittilä mine), which is middle of the Kittilä group of the Central Lapland greenstone belt. From Suurikuusikko we head to SW to Kolari and go over the N-S trending Ylläs fells which mark the boundary between Karelian and Norbotten cratons. After overnight in the Ylläs ski resort we cross the border to Pajala in Sweden. After stops in an old iron works and Fe mine in Masungsbyn and stops at the Pahakurkkio rapids in the Kalix river, we continue to the higher ground to Kiruna mining town. Next day stop at the historic Gruvberget Cu mine on the way to the Aitik Cu mine near Gällivare. After Aitik we have the longest drive of the excursion to Kemi. On the last day we visit the Kemi Crome mine and head to Rovaniemi in early afternoon.

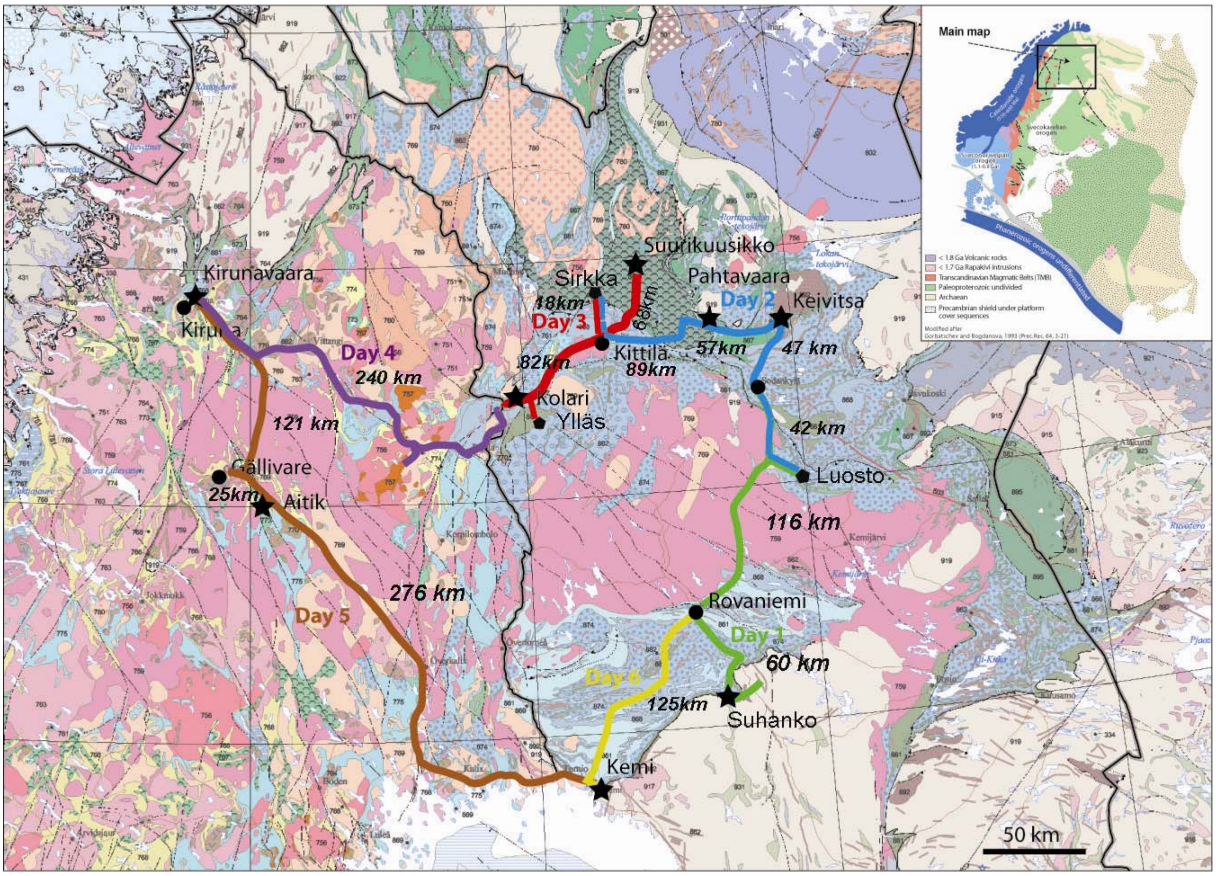

Figure. 1 Excursion route on the geological map of the northern Fennoscandia.

Excursion Stops

Day 1: The Suhanko-PGE prospect and the Portimo layered intrusion

Portimo Layered Igneous Complex

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Structural units of the Portimo Complex

The Portimo Layered Igneous Complex (Portimo Complex, Fig. 1) belongs to the Tornio-Näränkävaara Belt of c. 2.44 Ga layered intrusions and is composed of four principal structural units (Alapieti et al. 1989a, Iljina 1994, Iljina & Hanski 2005):

- Narkaus Intrusion
- Suhanko Intrusion
- Konttijärvi Intrusion
- Portimo Dykes

The Konttijärvi Intrusion is separated from the western end of the Suhanko Intrusion (Ahmavaara Block) by 3.5 km of Archaean rocks. The Konttijärvi mafic body is regarded as a faulted offset and subsequent erosion of the Suhanko mafic body, because the stratigraphic succession and style of mineralisation is similar to that in the Ahmavaara Block which is the westernmost tip of the Suhanko body. Each intrusion contains a marginal series and an overlaying layered series (Fig. 2). The marginal series of the Suhanko and Konttijärvi Intrusions differ from that at Narkaus in thickness and the prevailing rock types. The Narkaus marginal series generally varies from 10 to 20 m in thickness, whereas the Suhanko and Konttijärvi marginal series may reach several tens of metres. The Narkaus marginal series is mainly composed of pyroxenite, with some plagioclase-bearing rocks in its lower parts, whereas olivine cumulates commonly constitute the upper half of the Suhanko and Konttijärvi marginal series.

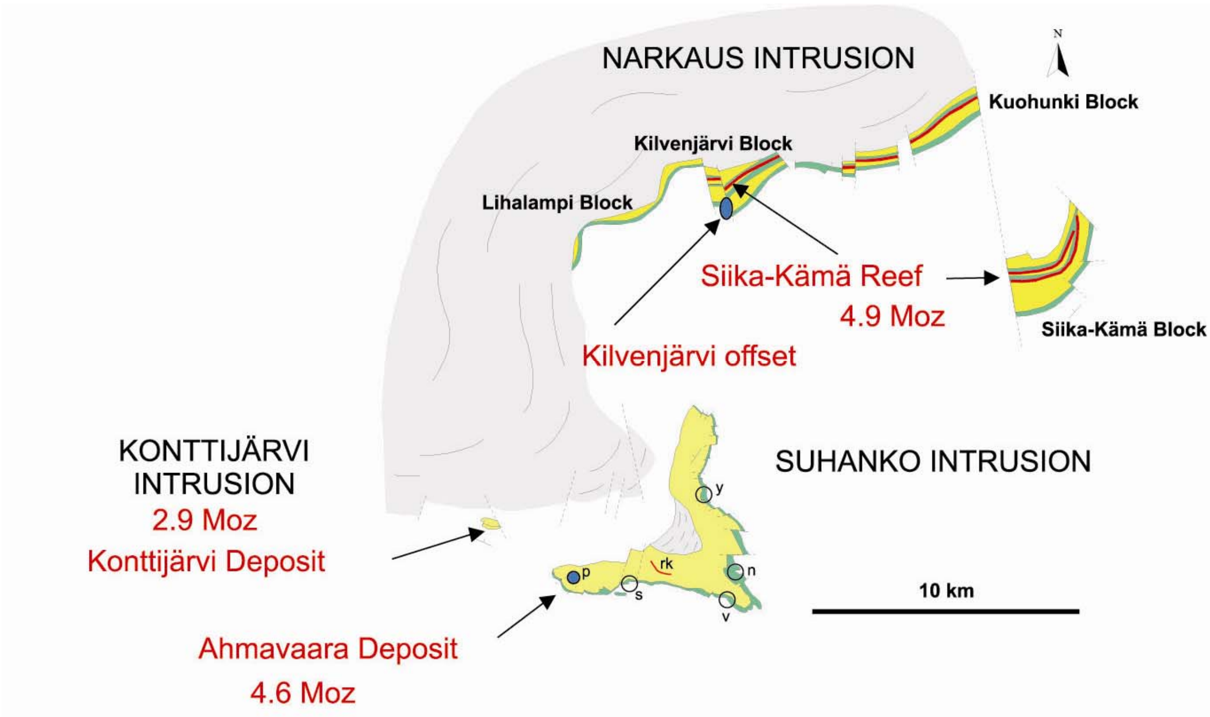

Figure 1. General geological map of the Portimo layered Igneous Complex (Iljina 1994) and published Pt-Pd-Au resources (Gold Fields press releases in July 2003). rk, Rytikangas PGE Reef. The massive sulphide deposits in the Suhanko Intrusion are also shown: s, Suhanko proper; v, Vaaralampi; n, Niittylampi and y, Yli-Portimojärvi.

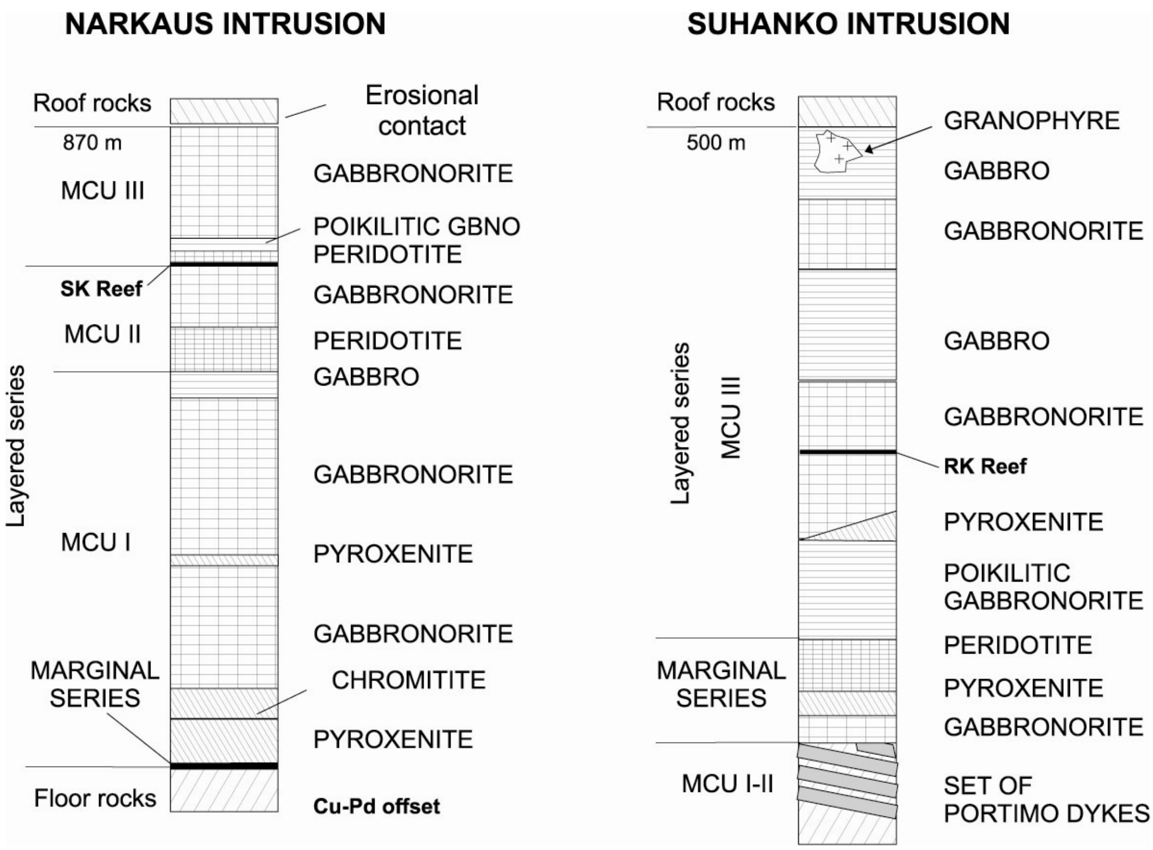

Figure 2. Cumulus stratigraphies of the Narkaus and Suhanko Intrusions and the locations of the principal PGE occurrences. MCU, megacyclic unit. Modified after Iljina (1994).

A striking difference between the layered series of the intrusions is the presence of marked reversals in the Narkaus Intrusion, as shown by the thick ultramafic olivine-rich cumulate layers, whereas in the Suhanko and Konttijärvi Intrusions crystallisation continued without any notable reversals. The layered series of the Suhanko Intrusion (except in the Ahmavaara Block, see below) commences with plagioclase-bronzite orthocumulates (with poikilitic augite) that also contain some bronzite cumulate interlayers. This poikilitic rock is separated from the overlying, rather monotonous plagioclase-bronzite-augite adcumulates by pyroxenite that is a few metres thick. About midway in the stratigraphy, bronzite disappears as a cumulus mineral, but returns higher up in the Suhanko sequence. Four poikilitic anorthosite layers also occur in the upper Suhanko layered series. Granophyric material is limited to discontinuous patches and cross-cutting dykes in both the upper Suhanko and upper Konttijärvi layered series.

The major reversals in the Narkaus layered series resemble those in the Penikat Intrusion and enable its layered series to be divided into three megacyclic units (Fig. 2). The lowermost (MCU I) commences with a thick (~80 m) bronzite cumulate layer with a massive chromitite layer close to the top. The rest of MCU I, and the gabbroic parts of MCU II and MCU III, comprises mainly plagioclase-bronzite-augite adcumulates, with the exception of a poikilitic plagioclase cumulate layer above the ultramafic basal part of MCU III. MCU unit II, however, is found only in the Kilvenjärvi Block and fades away eastwards.

Mafic and ultramafic dykes, known as the Portimo Dykes, occur in the basement below the Konttijärvi Intrusion and in the Ahmavaara area of the Suhanko Intrusion. They have also been detected as fragments in the Konttijärvi marginal series (Fig. 3A). The dykes have not been dated and their association to the main intrusions is based on geochemical observations, as discussed below. The dykes are subparallel to the basal contact of the intrusion and locally merge with it.

Fine-grained, non-cumulate-textured gabbroic bodies up to a few tens of metres thick and several hundred metres long, and fragments from a centimetre to a metre in size, occur in many places in the Suhanko marginal series (Figs. 4A and 6). The chemical composition of these fragments seems to vary along the strike, as the Ahmavaara bodies turned out to have a distinctly higher Cr and slightly higher MgO content than bodies in the SE tip of the Suhanko Intrusion, Niittylampi area (Table 3). The chemical composition of the Niittylampi fragments is similar to the mean composition of the Suhanko Intrusion. The chemical features and the mode of occurrence of these plagioclase–two pyroxene rocks have led to the interpretation that the bodies are autoliths and representatives of chilled margin rocks that were disrupted and entrained by subsequent magma pulses (Iljina 1994).

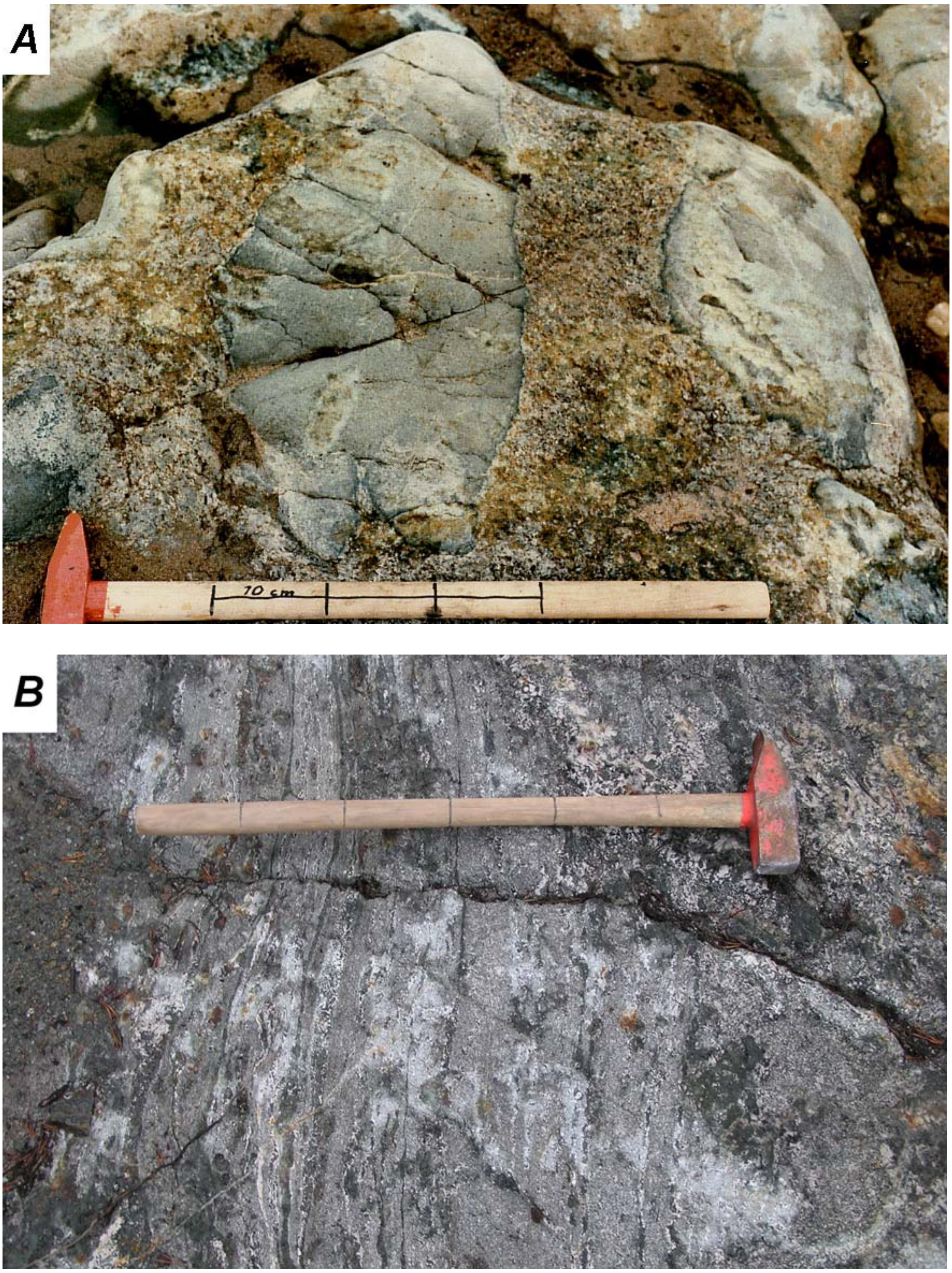

Figure 3. A, large blocks of the Portimo Dykes in the olivine cumulate (darker) of the Konttijärvi marginal series. B, banded gabbro, varitextured zone, Konttijärvi. Photo M. Iljina.

Special stratigraphic features at Konttijärvi and Ahmavaara

The cumulus sequences in the layered series of the small Konttijärvi Intrusion and the western end of the Suhanko Intrusion, the Ahmavaara Block, resemble each other. Pyroxenite, which separates the lowermost poikilitic orthocumulate from the overlying gabbroic adcumulate, attains a thickness of tens of metres in both sections. This pyroxenite is a tenth of the thickness elsewhere in the Suhanko Intrusion. A laterally persistent olivine cumulate layer about 10 m thick is found in the lower part of the Ahmavaara Block (Fig. 6). This peridotite layer is separated by a roughly 20 m thick poikilitic layered series gabbronorite from the underlying peridotite of the marginal series. The layered series peridotite is not interpreted as referring to a new megacyclic unit but merely as reflecting smaller-scale cyclicity in the lower Ahmavaara stratigraphy.

The gabbroic rocks in the Konttijärvi marginal series, indicated in Figure 5, are partly pyroxene cumulates with variable portions of felsic material introduced by floor rock contamination. When cumulus terminology is used, this and the thick layered series pyroxenite make the present-day Konttijärvi stratigraphy rather ultramafic. The lower contact of the Konttijärvi Intrusion is also rather unique. Below the lowermost more homogenous cumulate, there is a thick mixing zone made of varitextured gabbro, which in some places is up to 150 m wide (Fig. 10). The combined thickness of the homogenous Konttijärvi marginal and layered series is only slightly greater (about 160 m) than the varitextured gabbro zone.

The varitextured gabbro zone (also called Transition or Mixing Zone) comprises a rock type termed 'hybrid gabbro' and of banded gabbro (Fig. 10). The 'hybrid gabbro' is characterised by grain-size variations from fine to medium and also contains an almost assimilated felsic contaminant. Further away from the intrusion, the hybrid gabbro turns into banded gabbro (Fig. 3B), which in its outcrop appearance looks like recrystallised banded Archaean quartz dioritic gneiss which still has a primary folded texture but gabbroic mineralogy. Some of the banding is due to the turbulent flow of an unhomogenised mixture of the mafic and felsic (melted basement) melts. Contacts between homogenous gabbro, hybrid gabbro and banded gabbro are arbitrary, but the hybrid and banded gabbros are mapped to form a domain of their own. This division is due to a pattern in which many drill holes contain several sections of hybrid and banded gabbros (1–20 m in length) right next to each other, so that the two gabbro types together form a distinctive, mappable unit. This unit also contains basement gneiss blocks up to several tens of metres in size. The varitextured gabbro zone appears to be a result of the mechanical and metasomatic mixing of melted Archaean gneiss and mafic magma, indicating dynamic intrusion of the mafic magma.

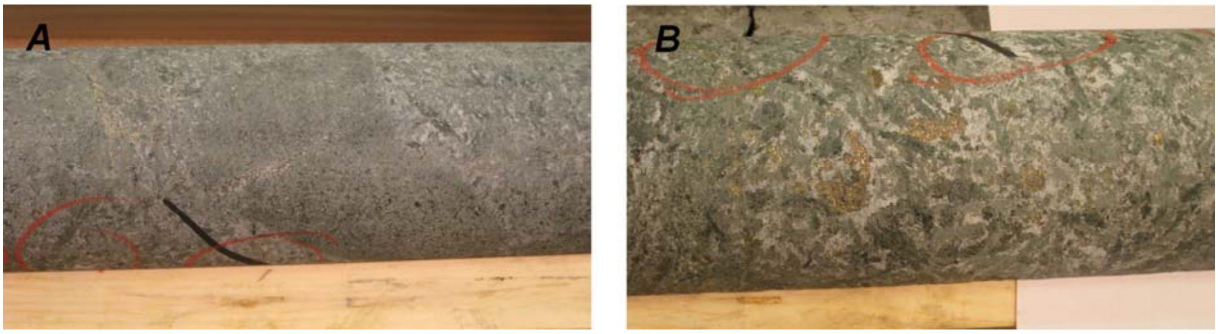

Figure 4. A, Fine-grained and sharp-edged chill margin fragments in the Ahmavaara drill core. For structural position see Figure 6. B, cresscumulate with sulphides in the lower part of the Ahmavaara marginal series. Photo M. Iljina.

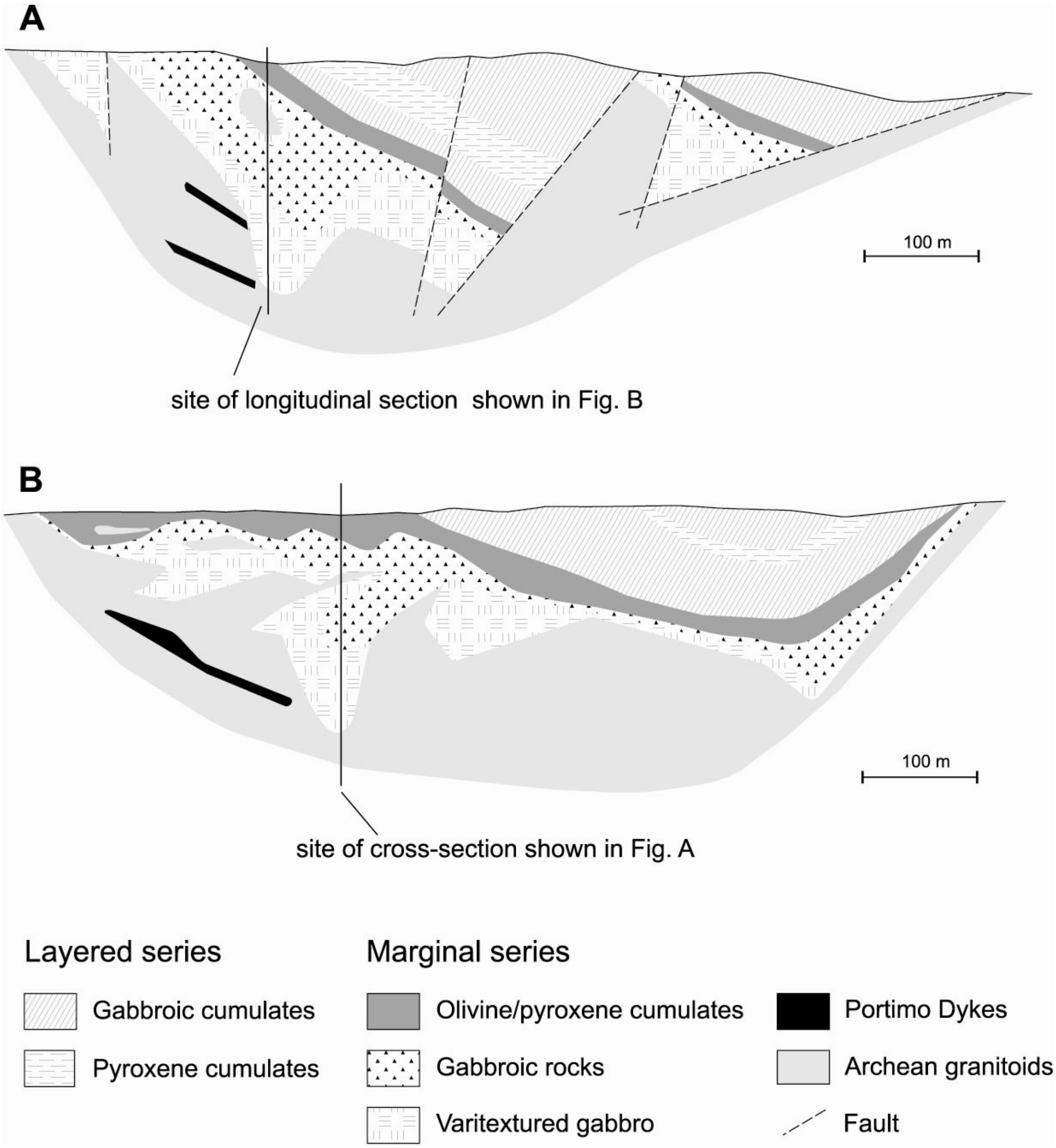

Figure 5. Cross-section (A) and longitudinal section (B) of the Konttijärvi Intrusion. Modified after Iljina and Hanski (2005).

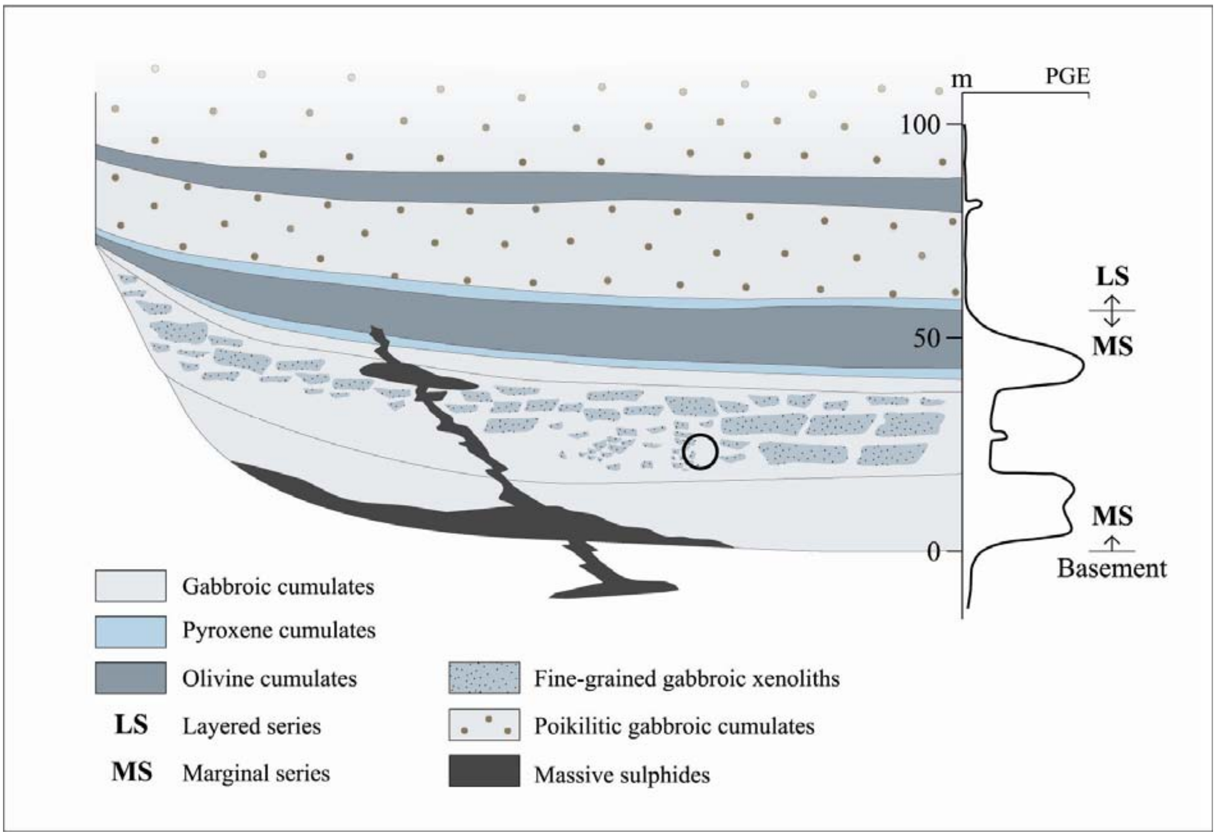

Figure 6. Schematic cross-section of the Ahmavaara marginal series and lowermost layered series. The circle shows the location of the photograph in Figure 4A.

Three-dimensional structure of the Portimo Complex

Various types of geophysical measurements carried out on the Suhanko Intrusion reveal its three-dimensional structure fairly well. The Suhanko Intrusion has an estimated present-day volume of about 10 km^3. Figures 7 and 8 present horizontal sections through the Suhanko Intrusion at three depths and three vertical cross sections. The significant feature is that the central and northern parts of the Suhanko Intrusion plunge under the roof rocks at a low angle, reaching a depth of about one kilometre relative to the present erosional surface. The lower contact of the intrusion is 'transgressive' with the base of the southeastern tip, which is much higher than that of the western and northern limbs if we rotate the intrusion into its original position (Fig. 9). The Ahmavaara block is relatively shallow and can be divided into southern and northern embayments, separated from each other by an east-west-trending 'anticline' (Fig. 8A). The southeastern tip of the Suhanko Intrusion is even shallower, and only marginal series cumulates are preserved, the overlying cumulates having been eroded away.

In view of their cumulus stratigraphy and chemistry, the Ahmavaara section is interpreted as representing the deepest part of the original Suhanko Intrusion.

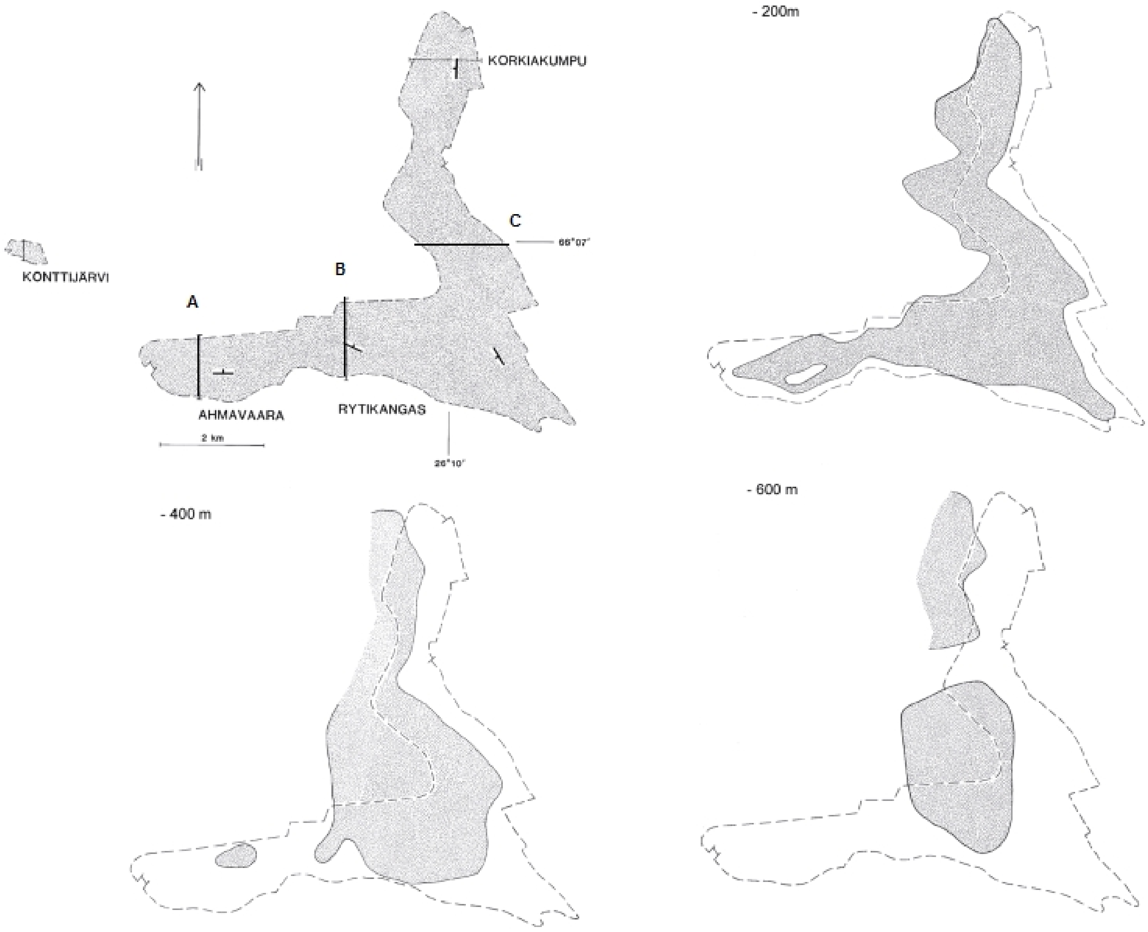

Figure 7. Suhanko Intrusion: plan view and horizontal cross-sections at depths of 200, 400 and 600 metres. Sites of the vertical cross-sections A-C depicted in Fig. 8 are also shown. Modified after Pernu et al., 1986.

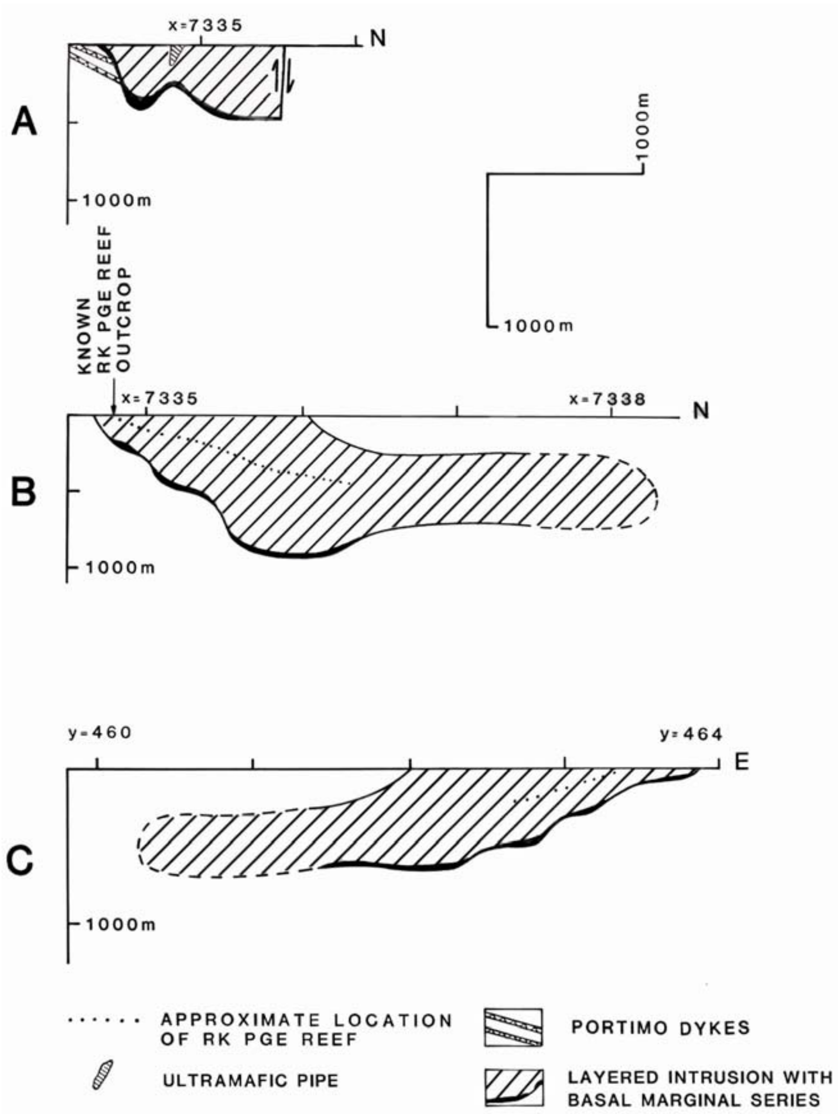

Figure 8. Three vertical cross-sections through the Suhanko Intrusion. Sites marked on Fig. 7 Modified after Pernu et al., 1986.

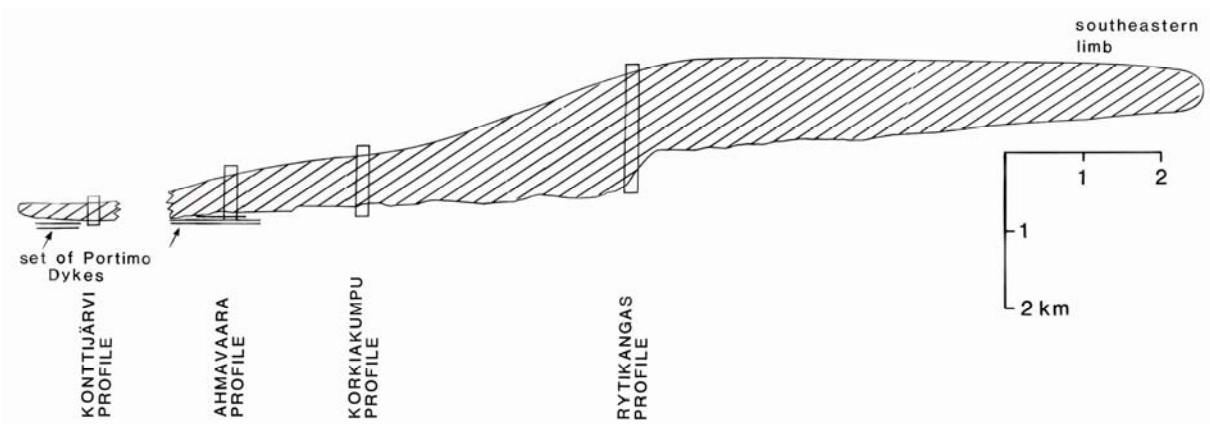

Figure 9. Schematic cross-section of the initial Suhanko Intrusion. Modified after Iljina (1994).

Cu-Ni-PGE mineralised zones in the Portimo Complex

Among the layered intrusions, the Portimo Complex is exceptional in hosting a variety of styles of PGE mineralisation (Figs. 10–15). The principal mineralisation types are (Iljina 1994):

- PGE-bearing Cu-Ni-Fe sulphide dissemination in the marginal series of the Suhanko and Konttijärvi Intrusions
- Predominantly massive pyrrhotite deposits located close to the basal contact of the Suhanko Intrusion
- Rytikangas PGE Reef in the layered series of the Suhanko Intrusion
- Siika-Kämä PGE Reef in the Narkaus layered series
- Offset Cu-PGE mineralisation below the Narkaus Intrusion

The first two styles represent a mineralisation type defined recently as a 'contact type' mineralisation. Five other PGE enrichment types are depicted in Figure 15. These are 1) the PGE enrichment in the Portimo Dykes below the Konttijärvi and Ahmavaara marginal series, 2) the PGE concentrations near the roof of the Suhanko Intrusion, mostly associated with pegmatites, 3) a Pt-anomalous pyroxenitic pegmatite pipe in the western limb of the Suhanko Intrusion, and 4) chromite and silicate-associated PGE enrichments in the lower parts of the Narkaus Intrusion and MCU II.

Figure 15 shows the structural model for the Portimo Complex and the positions of the mineralisations described above, as interpreted by Iljina (1994). Taking the boundary of the two parental magmas as a reference level, it can be seen that the Siika-Kämä Reef and the highly mineralised Ahmavaara and Konttijärvi marginal series are located in the same positions in terms of magmatic stratigraphy.

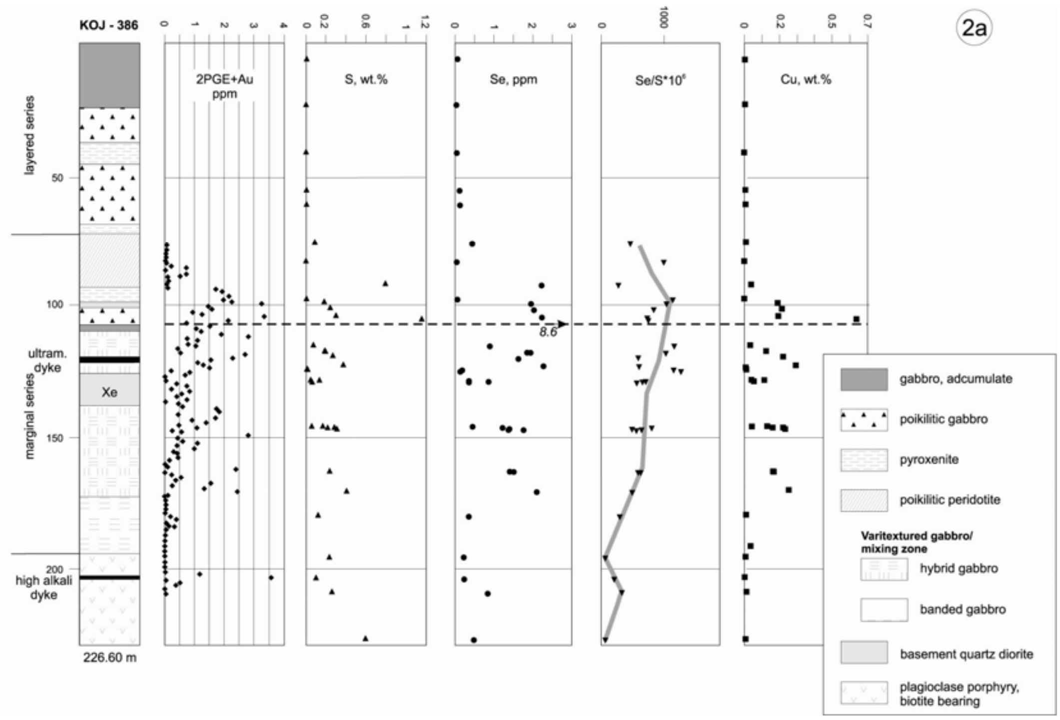

Figure 10. Stratigraphic sequence of the Konttijärvi marginal series showing variations in bulk Pt+Pd+Au, S, Se, Se/S and Cu. For structural position see 2a in Figure 15. Xe, basement xenolith. Modified after Iljina (2005).

Disseminated PGE-bearing base-metal sulphide mineralised zones

Disseminated PGE-bearing base-metal sulphide mineralised zones, normally 10–30 m thick, occur throughout the marginal series of the Suhanko and Konttijärvi Intrusions (Figs. 10 and 15). Their distribution is erratic and they generally extend from the lower peridotitic layer downwards for some 30 m into the basement. The PGE contents vary from only weakly anomalous values to 2 ppm in most places in the marginal series of the Suhanko Intrusion but rise to >10 ppm in several samples in the Konttijärvi and Ahmavaara.

Figure 10 depicts the variation in the copper, precious metals and Se/S ratio in one drill hole across the Konttijärvi marginal series. Whole-rock PGE seems to have a good correlation with copper, Se and Se/S. Also, as a whole, half of the metal content of the entire Konttijärvi deposit is hosted lithologies below (varitextured gabbro, basement) the marginal series proper. Moreover, especially in the Suhanko Intrusion the lowermost marginal series has cresscumulate texture (Fig. 4B).

Massive sulphide mineralisation

Massive sulphide mineralisation is characteristic of the marginal series of the Suhanko Intrusion. These zones have the form of dykes and obviously also plate-like bodies conformable to layering, and generally vary in thickness from 20 cm to 20 m. The mineralised zones also vary in location from 30 m below the basal contact of the intrusion to a position 20 m above it (Fig. 6), and range in size from less than 1 million tonnes to more than 10 million tonnes. The sulphide paragenesis is composed almost exclusively of pyrrhotite, except in the Ahmavaara deposit which also contains chalcopyrite and pentlandite. The massive pyrrhotite deposits show relatively low PGE values with the maximum Pt + Pd normally reaching to a few ppm (exemplified by circle 3, Figs. 11 and 15). However, similarly to the marginal series disseminated sulphide mineralisation of the same intrusion (see above), the PGE concentrations are generally much higher in the Ahmavaara deposit, attaining a level of 20 ppm (exemplified by circle 2b, Figs. 11 and 15). Drilling also shows that the low PGE grade Suhanko (proper) massive sulphide deposit locates physically above the Rytikangas Reef due to the 'transgressive' nature of the Suhanko marginal series (Iljina et al. 1992).

Rytikangas PGE Reef

The Rytikangas PGE Reef represents the main PGE occurrence in the layered series of the Suhanko Intrusion (Figs. 1, 12 and 15) where it is located in the middle of the western limb, about 170 m above the base of the intrusion. Its position is known over a distance of 1.5 km. The Rytikangas Reef is hosted by poikilitic plagioclase, plagioclase-bronzite, and plagioclase-bronzite-augite orthocumulates, all containing augite oikocrysts. This cumulate series overlies a 70 m sequence of monotonous plagioclase-bronzite-augite adcumulates and underlies 10 m of homogeneous plagioclase-bronzite mesocumulates with nonpoikilitic intercumulus augite. The orthocumulate layer varies in thickness from 30 cm to 10 m. The thickness of the Reef itself is 30–50 cm and it typically occurs on top of the poikilitic orthocumulate layer. The cumulus stratigraphy and drop in the whole-rock Cr content across the Rytikangas Reef are practically identical to those of the Ala-Penikka Reef in the Penikat Intrusion (Fig. 12).

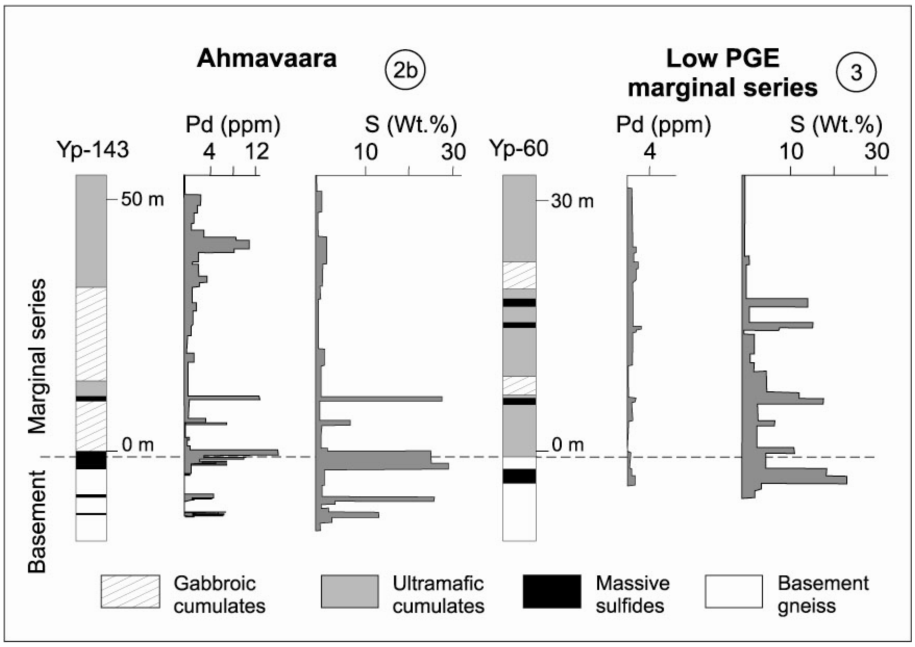

Figure 11. Comparison of the Ahmavaara deposit and one low grade, disseminated and massive sulphide deposit of the Suhanko Intrusion (Suhanko proper, Fig. 1). For structural position see circles 2b and 3 in Figure 15. Modified after Iljina et al. (1992).

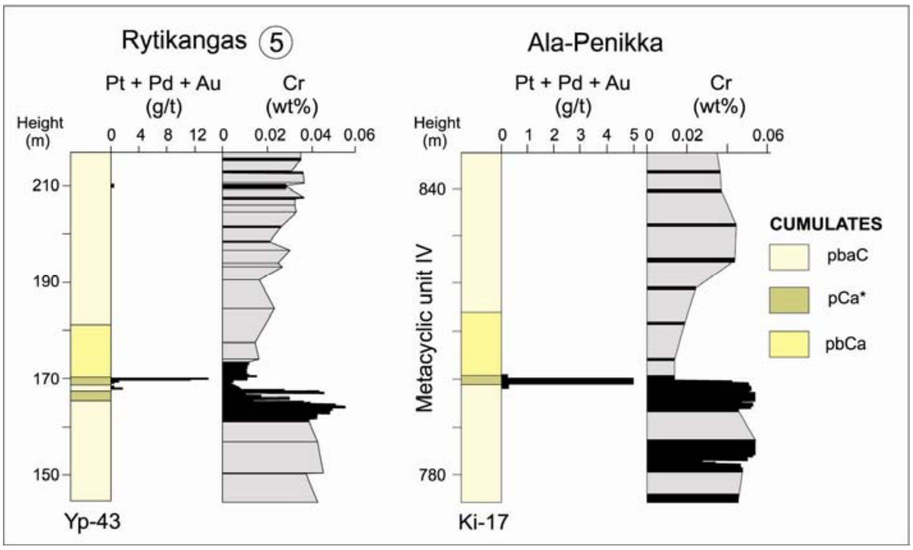

Figure 12. Comparison of the Rytikangas (RK) and Ala-Penikka PGE Reefs from the Suhanko and Penikat Complexes, respectively. For the structural position of the RK Reef see circle 5 in Figure 15. Modified after Iljina et al. (1992).

Siika-Kämä PGE Reef

The Siika-Kämä PGE Reef of the Narkaus Intrusion is most commonly located at the base of MCU III (Figs. 13 and 15), but it may lie somewhat below that or in the middle of the olivine cumulate layer of MCU III. Chlorite-amphibole schist similar to that in the Sompujärvi PGE Reef in the Penikat Intrusion commonly hosts the Siika-Kämä Reef. In some parts of the reef, the PGE mineralisation is accompanied by chromite seams or chromite dissemination. The thickness of the reef varies from less than one metre to several metres, and many drill holes penetrate a number of mineralised layers separated by PGE-poor layers which can be several metres thick. The PGE concentration varies from several hundred ppb to tens of ppm. Some gabbroic pegmatites, abundant in the uppermost gabbroic adcumulates tens of metres below MCU III, are also mineralised and can contain several ppm of Pd and Pt. The Siika-Kämä mineralisation is one of the most sulphide-deficient PGE mineralisations in the Portimo Complex, in some places containing no visible sulphides and rarely exceeding a whole-rock sulphur content of 1 wt.%.

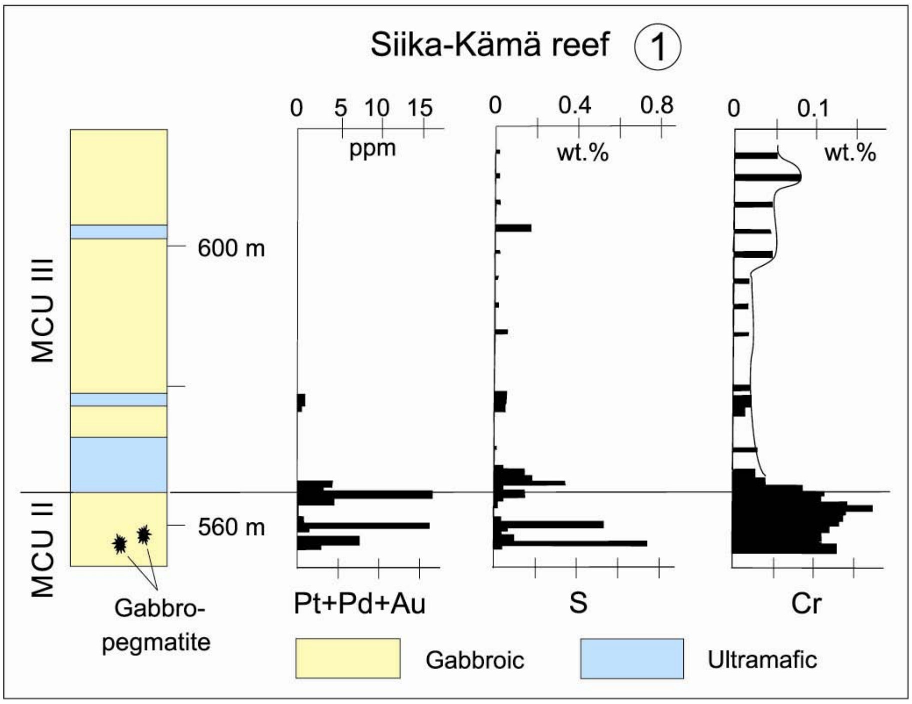

Figure 13. Cumulus stratigraphy and variation of the precious metals and S and Cr, Siika-Kämä PGE Reef, Narkaus Intrusion. For the structural position see circle 1 in Figure 15. Metal data from Huhtelin et al. (1989a).

Offset Cu-Pd mineralisation

The offset mineralisation is sporadically distributed in the basement gneisses and granites below the Narkaus Intrusion. The largest deposit, and also the best-known, is situated below the Kilvenjärvi Block (Figs. 1, 14 and 15). This deposit is composed of a cluster of closely grouped ore bodies and is located in and near a N-trending major fault zone some tens of metres wide, against which the Kilvenjärvi Block terminates. The offset mineralisation represents the richest PGE deposit type within the Portimo area, with Pt+Pd contents reaching up to 100 ppm. The offset mineralisation is predominantly a Pd deposit, as it has a much higher Pd/Pt ratio than the other Portimo deposits (or any other Tornio-Näränkävaara Belt PGE deposit) and is extremely low in Os, Ir, Ru, and Rh (Table 2). Furthermore, it is extremely irregular in form, containing disseminated sulphide-PGM 'clouds', massive sulphide veins or bodies, and breccias in which sulphide veins brecciate granitoids. The proportions of base metal sulphides and PGM are highly variable, but the massive sulphide bodies are always rich in PGE, whereas some samples containing almost no visible sulphides can carry several tens of ppm of Pd. In general terms, the more sulphide-rich occurrences are situated closer to the basal contact of the intrusion and those poorer in sulphides are encountered in a wider zone below the intrusion (Fig. 14).

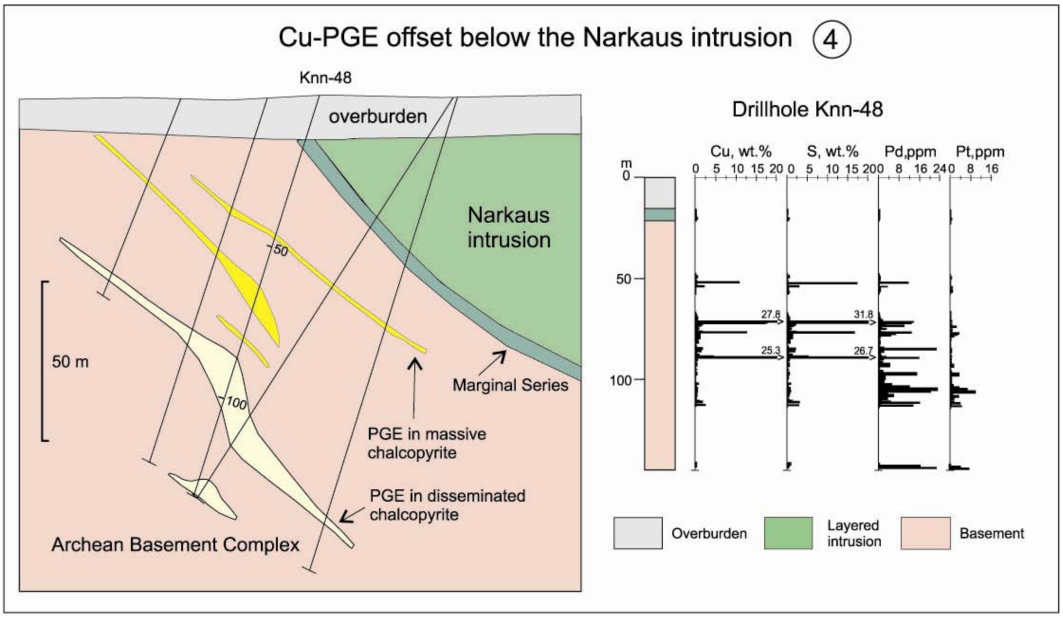

Figure 14. Off-set Cu-Pd deposit at Kilvenjärvi below the Narkaus Intrusion. For the structural position see circle 4 in Figure 15. Metal data from Huhtelin et al. (1989a).

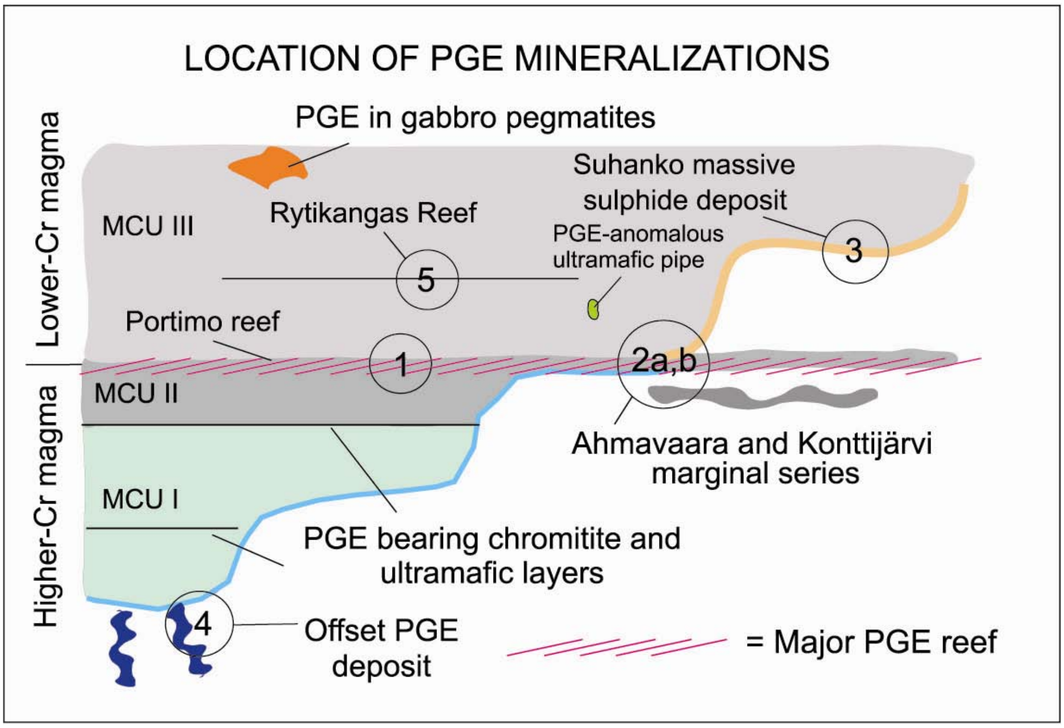

Figure 15. Schematic presentation of the locations of the various PGE enrichments encountered in the Portimo Layered Igneous Complex. The circled numbers refer to Figs. 10-14 in this paper. Modified after Iljina (1994).

Composition of the sulphide fraction, PGE ratios and chondrite-normalised distribution patterns

The concentrations of sulphur and base and noble metals in the type samples (A) and their values when recalculated to 100% sulphides (B), are presented in Table 1. The element concentrations, also recalculated to 100 % sulphides, marked by C in Table 1, represent the means of a large number of samples.

The massive sulphide deposits at the base of the Suhanko Intrusion, except for Ahmavaara, proved to be poor in nickel and copper, as their concentrations in the sulphide fraction ranged from 0.48 to 2.2 wt.% Ni and from 0.37 to 2.4 wt.% Cu. In places, the olivine cumulates above have an even higher nickel content than the massive sulphides below. Conversely, the Ahmavaara deposit has a markedly higher nickel content than the others, 2.7 wt.% in sulphide fraction.

Table 1. A: Average whole-rock Ni, Cu, S, PGE and Au concentrations for selected type samples, with standard deviation in parentheses; B: concentrations in the type samples (n, number of samples), recalculated to 100% sulphide; C: metal concentrations in a large number of samples, recalculated to 100% sulphide. See text for further information. Data from Iljina (1994) and from references therein.

		Ni(wt.%)	Cu	S	Os(ppb)	Ir	Ru	Rh	Pt	Pd	Au
Portimo Dykes and sulphide disseminated marginal series											
Portimo Dykes,	A	0.025	0.093	0.183	3.0	0.5	-	2.0	510	2200	47
n=1	B	5.0	18.5	36.5	598	99.7		399	101 700	438 700	9370
Konttijärvi	A	0.056	0.239	0.323	8.6	19.0	14.2	92	1 300	4 070	240
marginal series,		(0.040)	(0.170)	(0.200)	(8.3)	(12.9)	(10.0)	(72)	(710)	(3 260)	(110)
n=5	B	6.1	25.9	35.0	930	2 060	1 540	9 970	140 800	440 900	26 000
	C'	5.4	14.4	36.7							
Massive sulphide deposits of marginal series											
Ahmavaara,	A	2.00	0.719	25.8	20	50	44	357	1 510	11 030	104
n=3		(0.46)	(0.400)	(4.5)	(11)	(18)	(23)	(81)	(710)	(2 870)	(90)
	B	3.0	1.1	39	30	76	67	540	2 280	16 700	160
	C	2.7	2.4	37					2 120	15 200	
Suhanko,	A	0.919	0.807	20.7	30	64	64	222	207	1 230	7.3
n=3		(0.140)	(0.240)	(3.8)	(20)	(27)	(41)	(60)	(76)	(184)	(2.5)
	B	1.7	1.5	39.1	57	120	120	420	390	2 320	14
	C	1.5	2.4	37							
Vaaralampi[1]	A	0.284	0.143	23.3	25	8.5	89	24		485	5.0
n=2		(0.010)	(0.060)	(7.2)	(2.8)	(3.5)	(22)	(21)	-	(21)	(2.8)
	B	0.48	0.24	39.5	42	14	151	41		820	8.5
	C	0.94	0.63	37							
Niittylampi,	A	1.67	0.305	32.7	23	79	36	550	136	835	19
n=2		(0.06)	(0.030)	(0.7)	(5.7)	(9.9)	(11)	(14)	(83)	(7.1)	(11)
	B	2.0	0.37	39.2	28	95	43	660	160	1 000	23
	C	2.2	1.6	37					870	2 190	
Yli-Portimojärvi,	A	0.456	0.575	24.9		111		199	171	930	7.0
n=2		(0.160)	(0.540)	(12.1)	-	(154)	-	(271)	(127)	(636)	(5.7)
	B	0.72	0.91	39.4		180		310	270	1 470	11
Layered series											
Siika-Kämä Reef[2]	A	0.080	0.360	0.454	32	65	47	330	2 850	9 980	337
n=4		(0.074)	(0.191)	(0.229)	(14)	(35)	(36)	(151)	(1 125)	(6 120)	(324)
	B	6.2	27.7	34.9	2 460	5 000	3 610	25 370	219 000	766 900	25 910
Rytikangas Reef,	A	0.063	0.249	0.291	28	32	24	171	1 630	7 240	207
n=5		(0.080)	(0.190)	(0.330)	(27)	(24)	(18)	(123)	(427)	(2 460)	(145)
	B	7.4	29.3	34.2	3 290	3 760	2 820	20 100	191 600	850 900	24 300
	C"	6.4	38.7	33.0							
Mineralized upper	A	0.035	0.231	0.492	23	27	76	53	605	1270	40
Suhanko layered		(0.020)	(0.090)	(0.110)	(3.5)	(5.0)	(14)	(7.0)	(105)	(290)	(0)
series, n=2	B	2.7	17.9	38.2	1 790	2 100	5 900	4 120	47 000	98 600	3 100

The Pd/Pt, Pd/Ir and (Pt+Pd)/(Os+Ir+Ru) ratios are presented in Table 2. All the mineralisations are characterised by a predominance of palladium over platinum, only the Siika-Kämä Reef shows platinum domination in some places, as well as in the chromite and silicate-associated PGE enrichments of the MCU II. The Konttijärvi high-PGE grade marginal series, the Ahmavaara massive sulphide deposit and the Rytikangas and Siika-Kämä PGE Reefs have similar Pd/Ir and (Pt+Pd)/(Os+Ir+Ru) ratios. The above metal ratios were higher in the Portimo Dykes, where they were 3840 and >267, respectively, but distinctly lower in the low PGE-grade massive sulphide ores.

Table 2. Pd/Pt, Pd/Ir and (Pt+Pd)/(Os+Ir+Ru) ratios in the mineralised zones. The limits of variation are in parenthesis. N, number of samples. Data from Iljina (1994) and references therein.

	N	Pd/Pt	Pd/Ir	(Pt+Pd)/(Os+Ir+Ru)
Offset, Portimo Dykes and Konttijärvi marginal series				
Offset	1	6.7	>12 900	>687
Portimo Dykes[3]	2	3.4 (2.6-4.3)	3 840 (3 270-4 400)	>267
Konttijärvi marginal series	5	3.0 (1.4-4.0)	212 (138-429)	141 (67-300)
Massive sulphide deposits of marginal series				
Ahmavaara	3	8.0 (6.2-10.3)	243 (114-335)	116 (64-175)
Suhanko	3	6.6 (4.5-9.3)	20.6 (15.2-24.3)	10.5 (6.6-14.2)
Vaaralampi	1	3.6	42.7	4.2
Niittylampi	2	7.6 (4.3-10.9)	10.7 (9.7-11.7)	7.1 (6.8-7.3)
Yli-Portimojärvi	1	5.3	6.3	2.3
Layered series				
Rytikangas Reef	5	4.4 (3.7-5.1)	270 (147-382)	132 (77-184)
Siika-Kämä Reef	8	1.7 (0.8-3.0)	83.6 (32.1-132)	62.2 (41.7-93.3)
Mineralized upper Suhanko layered series	2	2.1 (2.0-2.2)	46.6 (44.5-48.6)	14.9 (14.4-15.4)
Chromitite layer, MCU I	1	3.8	17.1	5.6
Chromite and, silicate-associated PGE, MCU II	3	0.9	49.8	43.1

The chondrite-normalised distribution patterns are presented in Figure 16. A comparison of the PGE-mineralised zones on or slightly above the transition zone between the Cr-MgO richer and Cr-MgO poorer parental magmas indicates that the Portimo Dykes, Konttijärvi and Ahmavaara disseminated high PGE-grade marginal series, the Ahmavaara massive sulphides, and the Rytikangas and Siika-Kämä Reefs have very similar patterns, all possessing quite a steep positive slope and a depletion in Ru content, thus differing greatly from the low-PGE grade pyrrhotite deposits which do not have such a steep positive slope. The latter seem to have a negative Pt anomaly instead of a negative Ru anomaly (Fig. 16).

Parental magma composition

In order to evaluate the composition of 'Portimo' parental magmas, Table 3 gives some analyses of the Portimo Dykes and the fine-grained marginal rocks, Konttijärvi varitextured gabbros and weighted averages of the Suhanko Intrusion and MCU I of the Narkaus Intrusion. For comparison, the same table shows the composition of some sills in the Bushveld and Stillwater Complexes.

Two groups can be distinguished on the basis of whole-rock Cr and MgO content: an earlier magma type richer in Cr and MgO and a subsequently intruded magma type poorer in Cr and MgO. The whole-rock main and trace elements (except REE) of the Portimo Dykes and Ahmavaara autoliths resemble each other and the MCU I of the Narkaus Intrusion. Conversely, the composition of the Niittylampi autoliths is similar to the weighted average of the Suhanko Intrusion (Cr-MgO poorer type).

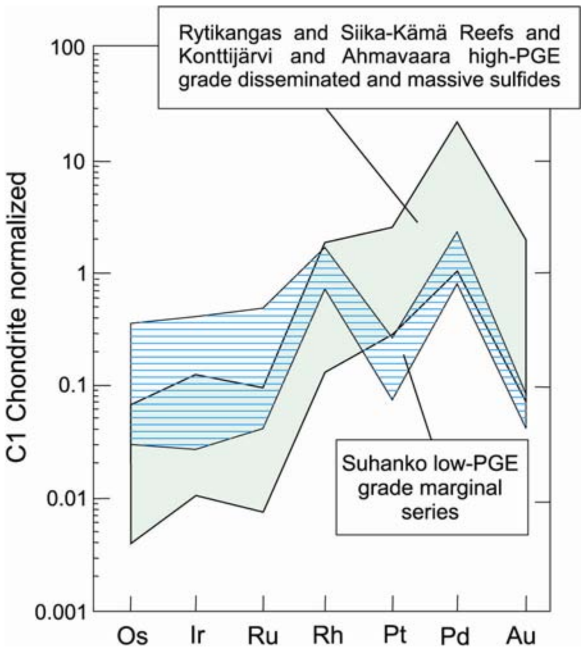

Figure 16. Chondrite-normalised PGE and Au patterns of the Siika-Kämä and Rytikangas PGE reefs, Konttijärvi and Ahmavaara high-grade PGE disseminated and massive sulphide deposits and lower-grade PGE massive sulphide deposits of the Suhanko marginal series. Data from Iljina (1994).

Table 3. Chemical compositions for the evaluation of the parental magmas. Data from Iljina 2005 unless otherwise indicated.

wt.%	**1**	**2**	**3**	**4**	**5**	**6**	**7**	**8**	**9**	**10**	**11**	**12**	**13**	**14**	**15**	**16**	**17**	**18**	**19**	**20**
SiO_2	52.2	51.3	51.7	52.7	53.5	53.4	53.4	54.0	51.4	56.1	50.7	49.9	50.4	52.6	45.1	53.5	54.5	55.7	57.0	57.4
TiO_2	0.15	0.12	0.19	0.33	0.21	0.24	0.30	0.16	0.40	0.34	0.22	0.53	0.23	0.19	0.25	0.44	0.44	0.63	0.29	0.47
Al_2O_3	16.1	17.1	16.8	16.9	13.4	12.2	9.8	3.2	8.0	11.5	14.2	13.6	16.9	11.5	20.1	15.4	13.0	15.6	6.0	19.4
FeO_{tot}	7.4	5.9	7.1	7.3	9.7	9.9	10.5	7.9	12.0	9.5	7.1	11.6	6.1	8.7	9.1	8.0	8.5	7.8	10.0	6.3
MnO	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.3	0.2	0.1	0.2	0.1	0.2	0.3	0.1	0.2	0.1	0.2	0.1
MgO	9.8	10.6	7.9	9.0	12.6	13.7	15.1	23.3	13.8	13.0	14.7	11.3	11.7	18.5	13.9	8.4	9.7	8.2	17.3	6.2
CaO	10.9	12.2	11.8	10.7	8.0	8.0	9.2	9.9	11.0	6.7	11.6	10.1	13.0	7.1	9.5	10.6	10.4	5.4	8.3	2.1
Na_2O	2.84	2.26	2.31	2.38	2.14	2.05	1.29	0.11	0.62	1.68	1.22	1.34	1.26	0.71	0.48	2.89	2.42	4.41	0.07	6.50
K_2O	0.42	0.18	0.24	0.57	0.10	0.19	0.11	0.01	0.45	0.80	0.22	0.27	0.18	0.34	1.23	0.55	0.64	1.92	0.65	1.36
P_2O_5	0.01	0.01	0.02	0.04	0.01	0.01	0.02	0.03	0.07	0.07	0.02	0.15	0.02	0.02	0.02	0.04	0.13	0.14	0.02	0.24
ppm																				
V	150	150	130		127	126	157	50	230	167			108	98	96	185	171	147	102	121
Cr	290	380	550	280	1137	951	1000	1800	2300	1240	3330	640	583	754	355	264	477	354	1030	81
Ni	160	190	130	280	375	440	318	610	250	295	270		255	399	1038	162	456	703	893	80
Zn	60	58	56		83	81	79	85	160	78		86	48	63	113	75	106	101	177	86
Sc	33	35	31		28	29	30	23	37			32	26	27	25	32	39	31	37	12
Sr	340	500	330	210	382	308	157	10	20	160	150		249	71	319	907	681	316	10	263
Y	2.5	1.0	11		4.1	4.9	6.2	4.8	7.0	13			4.7	3.6	4.6	6.3	11	13	10	10
Zr	30	1.0	26	47	10	10	15	20	40	77	25		13	10	10	10	10	81	27	76
Nb	2.7	1.9			<0.2	0.34	0.22	1.9	2.0	3.6			<10	<10	<10	<10	<10	<10	<10	<10
Ba	110	52		180	68	108	63	53	47		75	60	54	82	70	160	190	491	146	224
La	2.50	1.10	3.80		1.83	3.81	2.51		10.9	14.8		3.28	3.04	2.06	3.78	3.79	7.43	10.9	5.58	25.1
Ce	7.00	2.00	8.30		3.37	7.44	5.83		25.0	29.5		8.00	6.11	4.12	6.56	8.04	17.3	21.6	12.6	45.2
Nd	4.00	1.00			1.76	3.95	4.04		12.0	12.1		8.90	3.03	2.24	3.51	5.85	12.1	12.3	7.09	17.2
Sm	0.75	0.37	0.86		0.26	0.53	0.77		2.50	2.70		1.29	0.70	0.41	0.72	1.30	2.64	2.83	1.73	2.56
Eu	0.32	0.26	0.52		0.37	0.52	0.38		0.61	0.72		0.58	0.30	0.22		0.69	0.88	0.98	0.36	1.20
Tb	0.20	<0.10			<0.1	0.13	0.19		0.30	0.30		0.29	0.14	0.10	0.12	0.20	0.36	0.40	0.30	0.34
Yb	0.59	0.30	0.64		0.58	0.55	0.48		1.08	1.16		1.29	0.48	0.34	0.38	0.58	1.06	1.20	0.91	0.83
Lu	0.09	0.04	0.07		0.10	0.1	<0.10		0.16	0.15		0.20	<0.10	<0.10	<0.10	0.10	0.16	0.17	0.14	0.12
U	<0.10	<0.10			<0.2	<0.2	<0.2	0.80	1.20				0.20	0.20	0.20	0.20	0.20	0.83	0.59	0.66

1. Niittylampi autolith. Iljina (1994).
2. Niittylampi autolith. Iljina (1994).
3. Fine-grained marginal rock adjacent to the Main Zone/ Bushveld Complex. Hatton and Sharpe, 1989.
4. Weighted average of the Suhanko Intrusion (Alapieti et al. 1990).
5. Ahmavaara autolith. yp-517/109.9m.
6. Ahmavaara autolith. yp-517/115.5m.
7. Ahmavaara autolith. yp-517/126.4m.
8. Westerite Portimo Dyke. Konttijärvi. Iljina (1994).
9. Westerite Portimo Dyke. Ahmavaara. Iljina (1994).
10. Bushveld B1 Sill (Sharpe & Hulbert 1985).
11. Weighted average of Narkaus MCU I (Alapieti et al. 1990).
12. High-Mg gabbronorite sill. Stillwater Complex. (Heltz 1985).
13. Gabbronorite. Konttijärvi layered series. koj-386/6.1 m.
14. Gabbronorite. Konttijärvi layered series. koj-386/23.3 m.
15. Homogenous gabbronorite. Konttijärvi marginal series. koj-386/98.9 m.
16. Hybrid gabbro. Konttijärvi koj-386/145.9 m.
17. Hybrid gabbro. Konttijärvi. koj-386/147.0 m.
18. Hybrid gabbro. Konttijärvi. koj-386/170.6 m.
19. Ultramafic dyke. koj-386/119.5 m.
20. High alkali dyke. koj-386/203.7 m

The normalised REE pattern in the Portimo Dykes almost equals that of the Bushveld B1 sill and the patterns of the Niittylampi autoliths resemble Bushveld sills adjacent to the Main Zone (~B3 magma). The Ahmavaara autoliths differ slightly from the Niittylampi ones in that they have slightly higher LREE contents. For further discussion of the REE patterns, see Iljina (1994, 2005).

Concluding remarks on the marginal series-hosted mineralisation

The following west-to-east trends can be observed in the Konttijärvi and Suhanko Intrusion areas (Fig. 9):

- The spatial frequency of massive sulphide accumulations increases from west to east.
- The lower contact of the Suhanko Intrusion is 'transgressive' with the base of the southeastern tip which is much higher than that of the western limb of the Suhanko Intrusion. Drilling shows that the Rytikangas Reef is located physically below the Suhanko proper massive sulphide deposit.
- The most extensive interaction (varitextured gabbro) between mafic magma and footwall rocks is in the west (Konttijärvi).
- The Cu, Ni and precious metal concentrations in the whole-rock and the sulphide fraction are highest in the west (Konttijärvi and Ahmavaara areas).
- The relative proportion of ultramafic cumulates in the entire stratigraphic section through the marginal series and overlying layered series is highest in the west (Konttijärvi and Ahmavaara sections).
- Observations of the Portimo Dykes, suggested as representing the earlier more Cr-MgO enriched parental magma, are restricted to the Konttijärvi and Ahmavaara areas.
- Chill margin rocks (autoliths) become less Cr- and MgO-rich to the east. The 'west to east' change in the whole-rock main and trace element concentrations of the autoliths (including REE) resembles the change from B1 to ~B3 type parental magma in the Bushveld Complex.
- Iljina and Lee (2005) demonstrated that the whole-rock Se/S ratio drops from west to east. This is in line with the observation of Hattori et al. (2002), who concluded that high Se/S ratios are characteristic features of boninitic high-MgO second stage melts.

The last three of the above features support the conclusion that the westernmost part of the Suhanko-Konttijärvi area had an influx of the earlier Cr-MgO richer magma type ('more boninitic') and that the metal enrichments are related to the change-over stage of the two magma types.

Grades and tonnages of the Konttijärvi and Ahmavaara deposits

[As provided by Gold Fields Arctic Platinum Oy]

The Konttijärvi deposit has a strike length of 1,000 metres. The thickness of open pittable mineralisation varies from 30 to 100 metres. The orebody dips at 30°–40° to the N in the central and western parts, and 10°–20° to the N in the eastern part. The down dip continuity of the main ore body (Main Block) is terminated by a normal fault which displaces the mineralised zone back to the surface north of the fault.

The Ahmavaara deposit has a total strike length of 2,700 metres, the eastern 1,000 metres of which is called Ahmavaara East. The deposit comprises two slab-shaped mineralised units with a combined total thickness varying from 20 to 60 metres. These well-mineralised zones are mostly separated from each other by a poorly mineralised gabbroic unit some 5–30 metres in thickness. The western and southern limits of the deposit have partly primary, partly fault-disrupted lithological contacts with Archaean basement rocks. The deepest known marginal series mineralisation at Ahmavaara is approximately 500 m below the surface.

Mineral resource estimates based on July 2004 resource models for Konttijärvi and Ahmavaara are given in Table 4. The total resources above a 1 g/t 2PGE+Au cut-off total 119 Mt @ 1.97 g/t 2PGE+Au for 7.54 million contained 2PGE+Au Oz. In addition, the deposits contain significant concentrations of copper and nickel.

Table 4. Metal resources at Portimo.

Deposit	Mt	2PGE +Au(g/t)	Metal (000's oz)	Pd (g/t)	Pt (g/t)	Au (g/t)	Cu (%)	Ni (%)
Konttijärvi[1]	38.8	2.32	2,903	1.72	0.48	0.12	0.17	0.07
Ahmavaara[1]	60.0	1.86	3,592	1.40	0.30	0.16	0.27	0.10
Ahmavaara East[1]	20.1	1.61	1,043	1.20	0.28	0.14	0.23	0.08
Total Konttijärvi and Suhanko[1]	118.9	1.97	7,538	1.47	0.35	0.15	0.23	0.09
Siika-Kämä Reef[2]	43.1	3.51	4.9	2.70	0.72	0.08	0.11	0.08

1 Classified resources reported at 1.0g/t 2PGE+Au cut-off grade, 2004.
2 Resources at July 2003, Gold Fields press release, cut-off 1.0 g/t to depth of 100 m and 2.0 g/t for greater depth.

Acknowledgement
The authors are grateful to Gold Fields Arctic Platinum Oy for permission to publish this paper. We also thank the whole personnel of the company for their help with various phases of the work.

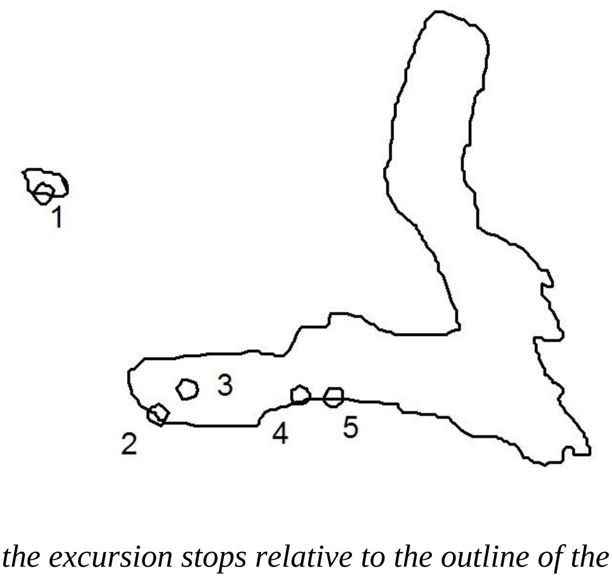

Figure 17. Location of the excursion stops relative to the outline of the Suhanko Intrusion, day 1.

Stop 1 Konttijärvi marginal series and related sulphide mineralisation

-footwall contact
-footwall gneiss, varitextured gabbro, gabbro, pyroxenite and peridotite
-Portimo Dyke xenoliths
-Cu-Ni-PGE mineralisation

Stop 2 Ahmavaara marginal series and related sulphide mineralisation.

-footwall contact
-footwall gneiss, chill margin xenolith, gabbro, pyroxenite and peridotite
-disseminated sulphides and massive sulphide veins
-drill core presentation

Optional stops

Stop 3. Ultramafic pipe

A Pt-anomalous pyroxenitic pegmatite pipe, c. 200 m in diameter, is found in the western limb of the Suhanko Intrusion. Center of the pipe (Ø c. 10 m) is iron enriched containing magnetite up to 10 vol.%. The outcropping outer margin is coarse grained augite-dominated augite-bronzite pegmatite. The only observations of fresh igneous augites (mg 70) are from this pipe.

Stop 4. Structural relationships.

A bus stop presentation demonstrating the structural position of one massive Fe-sulphide deposit (circle 3, Fig. 15) and the Rytikangas PGE Reef (circle 5, Fig. 15).

Stop 5. Vertical marginal series and angular discordance in layering.

A road side stop demonstrating thin entirely pyroxenitic marginal series and an angular discordance in the overlying layered series.

Day 2: The Pahtavaara gold mine and Kevitsa Ni-PGE deposits

Pahtavaara Au deposit

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Introduction

Pahtavaara is an active gold mine (in production 1996-2000, 2003-2007, in care and maintenance in early 2008) with a total in situ size estimate of 15 t gold (production + resource, February 2006). It is an underground mine but it started as an open cut operation. It is sited in an altered komatiitic sequence at the eastern part of the Central Lapland greenstone belt (see maps in Figs. 1 and 2 in the section Day 3, the Suurikuusikko gold deposit). It comprises of a swarm of subparallel lodes; nearly all gold is free native. It has many of the alteration characteristics of amphibolite-facies orogenic gold deposits and an obvious structural control, but has an anomalous barite-gold association and a very high fineness (>99.5 % Au) of gold. The geometry of high-grade quartz-barite lenses and amphibole rock bodies relative to biotite-rich alteration zones is also anomalous, as is the $\delta^{13}C$ of alteration carbonate minerals. Pahtavaara is best interpreted as a metamorphosed seafloor alteration system with ore lenses as either carbonate- and barite-bearing cherts or quartz-carbonate-barite veins. The gold may have been introduced later, but its grain size, textural position (nearly all is free, native, and occurs with silicates, not sulphides) and high fineness point to a pre-peak metamorphic timing which is highly anomalous for orogenic gold.

Geology and hydrothermal alteration

Following description is mainly extracted from Korkiakoski (1992). Pahtavaara gold deposit is hosted by the predominantly pyroclastic Sattasvaara komatiite complex. The present mineral composition of the least altered komatiites consists of an amphibole-chlorite assemblage resulting from regional greenschist facies metamorphism. The intensively altered rocks form a subvertically dipping alteration domain about 100 m x 500 m in extent (Figs. 1 - 3), comprised by two heterogeneous and intercalated lithological types: (1) biotite schists with talc-carbonate ± pyrite ± magnetite veins, and (2) coarse-grained and non-schistose amphibole rocks with associated quartz±barite veins and pods.

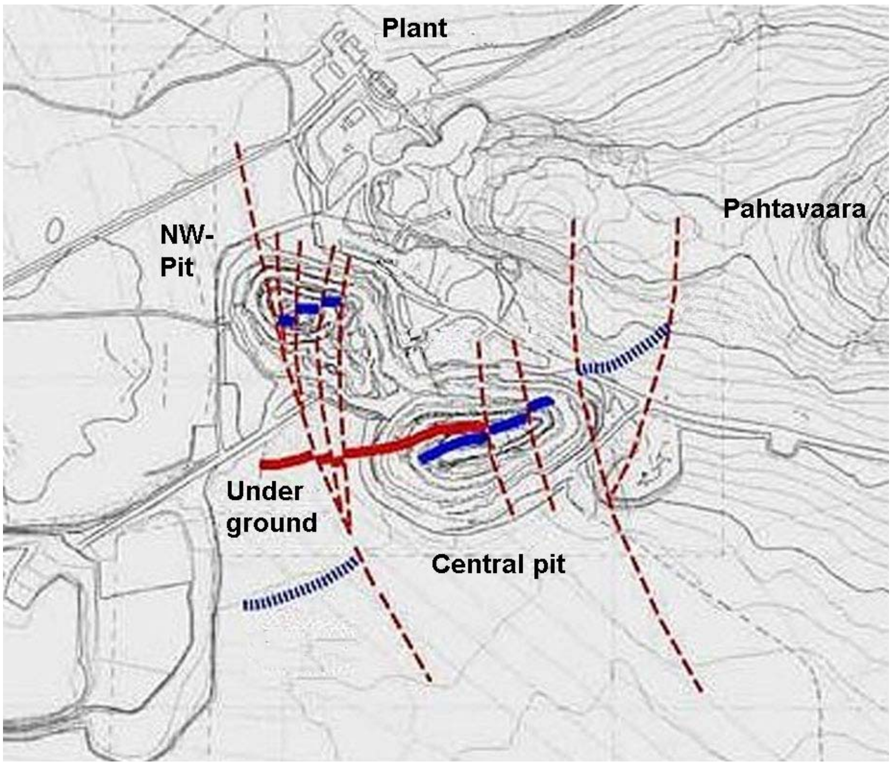
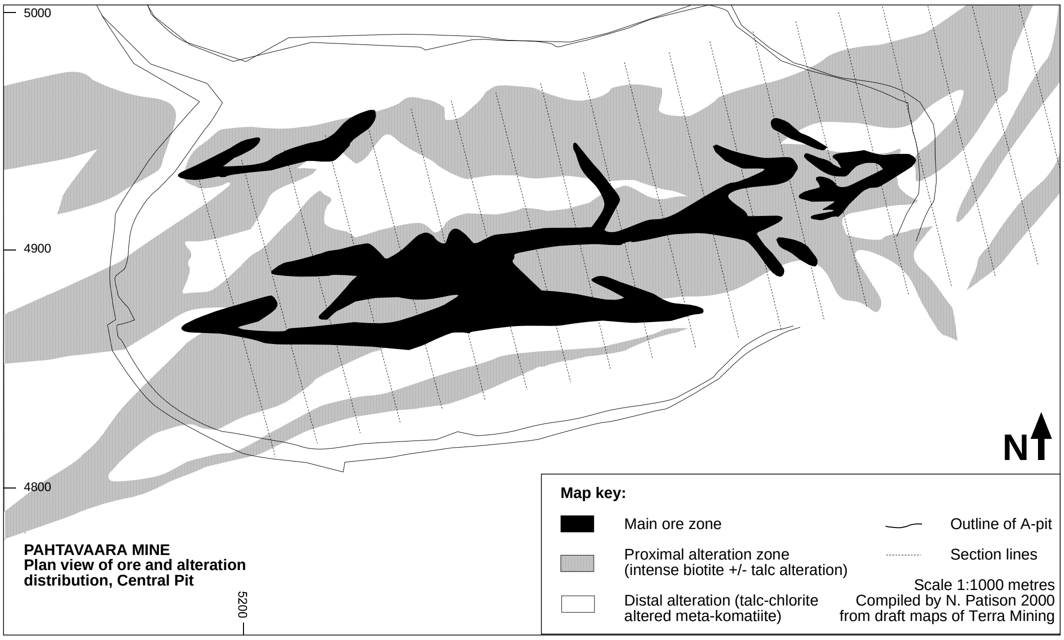


Figure 1. Map showing the Pahtavaara mine infrastructure and open cuts. Pahtavaara is a prominent hill and the landmark of the area. The rocks in the map area are Sattasvaara komatitic rocks. Blue lines show ore mined in the open pits. The thick solid red line shows the underground resource and dashed lines cross cutting faults (Source www.scanmining.se press release 28/04/2003).

The least altered amphibole-chlorite schists correspond compositionally to Geluk-type basaltic komatiites. The original komatiitic nature of the altered rock types is indicated by (1) the similarity in homogeneous immobile element ratios (Al/Ti) compared to those of less altered type, (2) mineralogical and geochemical gradations between rock types, and (3) similar REE patterns to those of the Sattasvaara komatiites.

Mass balance calculations have shown that biotite schists have been enriched in CO_2, K, Fe, Au, Ba, S, W, Te, Sr, and Mn, and depleted in Mg, Ca, Co, Si, and Zn, accompanied by a 10–30 % decrease in net volume. Amphibole rocks record a marked increase in volume, with gains in Ca, Si, Au, Na, Ba, Te, S, W, Sr and P, and losses in CO_2, Co, Mg, Fe and Zn.



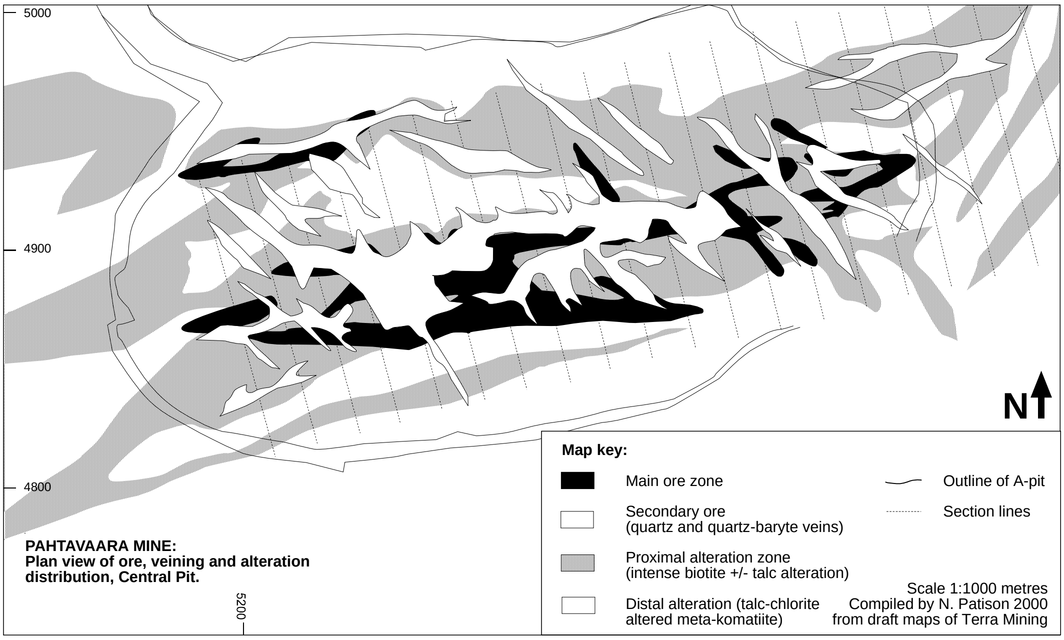


Figure 2. Geological maps of the Pahtavaara Mine open pit (compiled by N. Patison). North up.

The two major altered rock types reflect two stages of hydrothermal alteration (Fig. 1), which, on the basis of textural and geochemical evidence, include: (1) earlier biotitisation (K alteration), and (2) later amphibole overgrowth (Ca-Si alteration). The former has been interpreted to have taken place during or immediately after the peak of regional metamorphism, and during ductile deformation. Its distribution was controlled by a combination of high permeability in the originally pyroclastic komatiites, and NE-SW trending deformation zones. Later amphibole growth was related to the NNE-trending shearing resulting in the formation of zones of dilation into which hydrothermal fluids were focused under conditions straddling the brittle-ductile transition.

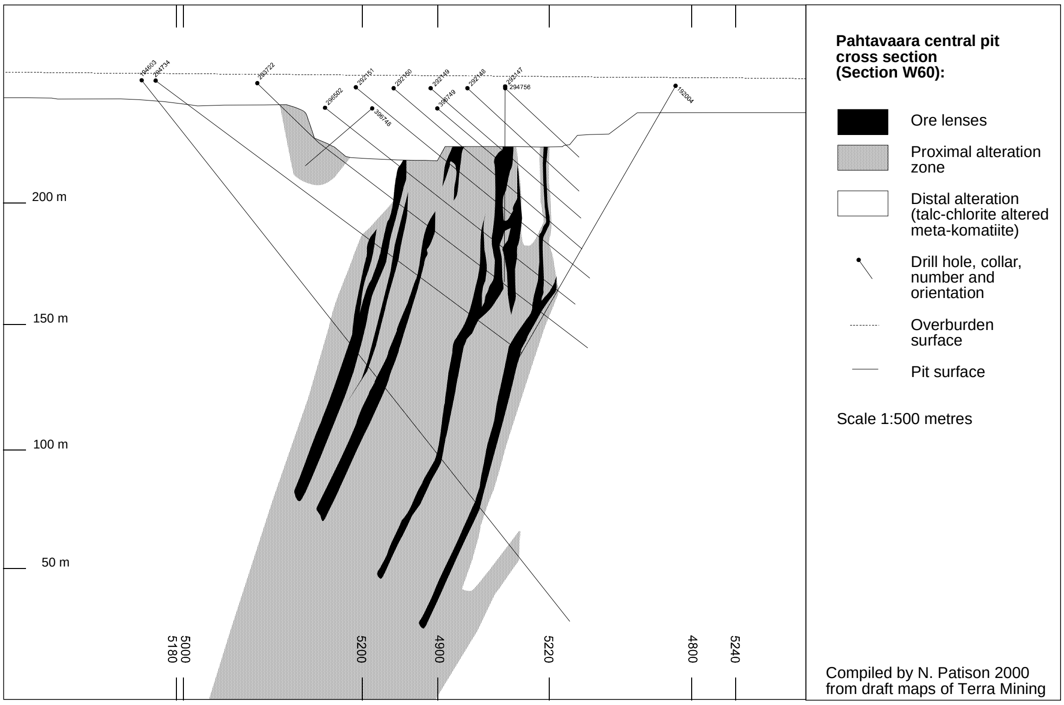


Figure 3. Cross section through the central pit of the Pahtavaara gold mine.

Mining

The gold ore at Pahtavaara forms narrow lodes generally 5–10 m wide, trending almost E-W and dipping northwards at about 70–80° (Fig. 2). For mining, the ore has been divided into the A+, A- and E-zones. A- zone ores are characterised by biotite-talc breccias that are typically surrounded by a more massive tremolitic amphibole rock characterised by irregular dilatational arrays of barite-carbonate-quartz veins. The A+ zone contains abundant barite and the A- zone veins also typically contain barite, in addition to quartz and carbonate. The E-zone comprises smaller lodes associated with quartz-carbonate-barite veins trending predominantly E-W and NNW-SSE. The only economically recoverable metal is gold, sulphides being relatively rare, with pyrite being the most abundant, comprising about 1% of the ore. Magnetite can constitute up to 5–10% of ore grade material, particularly in the biotite schists. Gold is free milling, has a very high fineness (>99.5 % Au), and it occurs as discrete grains, highly variable in size, between silicate grains and along fracture surfaces; some 50–60 % of gold grains are less than 50 μm in diameter.

The ore is concentrated at the Pahtavaara plant using gravimetric and flotation processes. Autogenic grinding, where amphibole-rich gangue rock is used as the grinding agent, precedes hydrocyclonic separation, followed by a gravimetric circuit including Reichert cones and spirals, with two shaking tables, and followed by flotation to enhance recovery at the final stages.

Stop 1 Pahtavaara mine

Ore and alteration zone in handspeciamens, boulders and drill core. The exact open cut and/or underground localities to be visited in the open cut depend on the accessibility to different parts of the mine which changes rapidly as a consequence of the mining activities.

The Kevitsa intrusion and associated Ni-Cu-PGE deposit

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Preface

The following description is mostly based on the works of Mutanen (1997, 2005) unless otherwise indicated. For a more detailed description see those extensive reports.

Location and exploration history

The Kevitsa mafic layered intrusion is located in Central Lapland, some 35 km north of Sodankylä, 0.8 km from the southern margin of the 2.44 Ga Koitelainen intrusion. It forms the western part of the larger Kevitsa-Satovaara Complex (Fig. 1).

The Kevitsa intrusion was explored by the Geological Survey of Finland between the years 1984 and 1995, in three phases. The presence of magmatic sulphides within the intrusion was indicated by early drilling in 1984 which intersected several meters of pyrrhotite-rich sulphides with low base and precious metal values, near the basal contact. The discovery hole (R326) was drilled in 1987 to the western part of the currently known mineralised domain, where it intersected about 30 meters of disseminated sulphides. Subsequent drilling programs delineated a large, low-grade Ni-Cu-PGE-Au occurrence, the Kevitsa deposit (formerly known by the names Keivitsa and Keivitsansarvi). The deposit was subsequently held and evaluated by the Outokumpu Company (1995–1998). In 2000, Scandinavian Gold Prospecting AB (subsidiary of Scandinavian Minerals Ltd) acquired the deposit.

General geology

The rocks in the vicinity of the Kevitsa intrusion belong to the 2.0–2.2 Ga old Savukoski Group (SKG) of Central Lapland. The Savukoski Group is subdivided into four formations; Matarakoski (MkF), Linkupalo (LpF), Sotkaselkä (SoF), and Sattasvaara (SaF) Formation, where MkF is the lowermost and SaF the uppermost unit (Fig. 2 in the Introduction; Lehtonen et. al 1998). The Kevitsa intrusion is surrounded by mica schists with graphite- and sulphide bearing interlayers, felsic volcanic rocks, magnesian metapelites and calcareous metasediments of the Matarakoski Formation (Fig. 1). These are overlain by komatiitic volcanic rocks intercalated with sulphide-rich Mg-pelites. Differentiated komatiitic sills occur close to the base of the intrusion and on the northern side of the intrusion. Pelitic rocks near the intrusion contacts were altered to hornfels due to thermal metamorphism caused by the intrusion. Regional metamorphism reached amphibolite facies grade and affected especially the country rocks and the upper parts of the intrusion the central part of the intrusion being less metamorphosed.

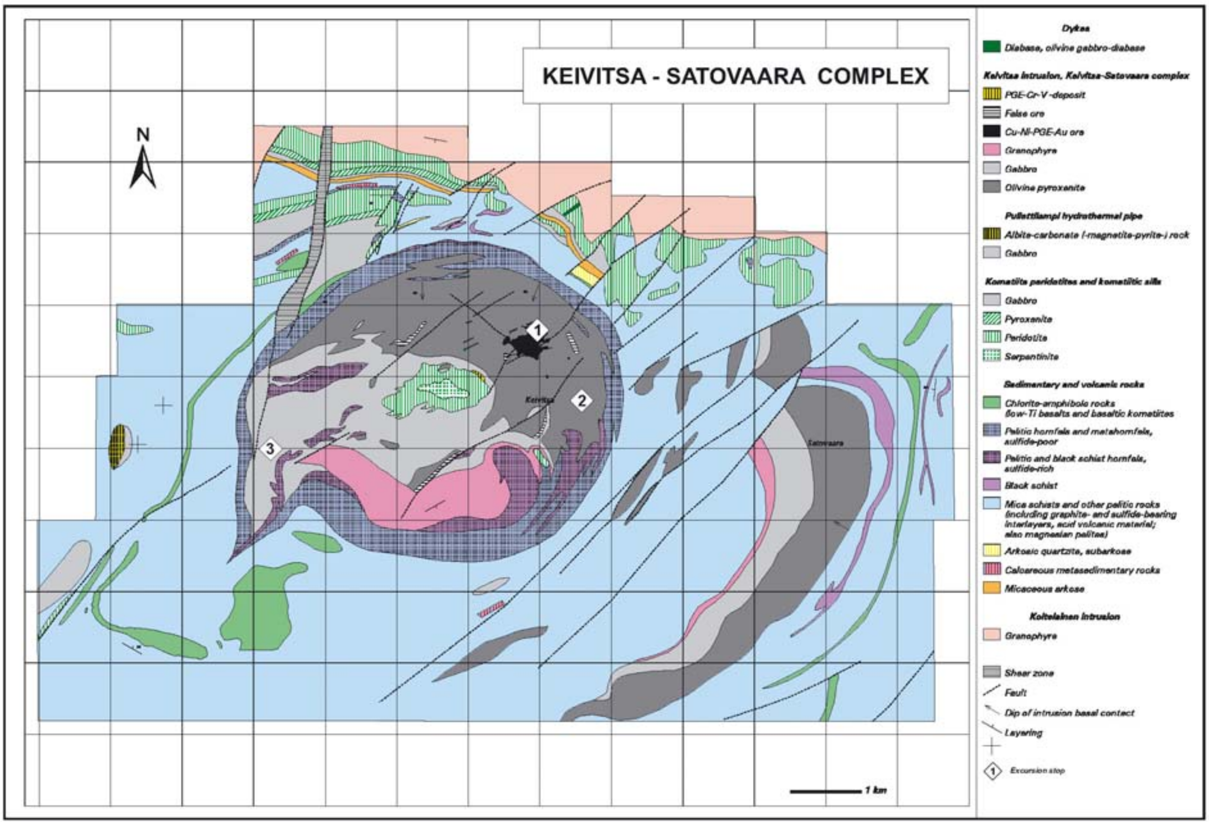

Figure 1. Geological map of the Kevitsa area, with excursion stops indicated by numbers. From Mutanen (2005).

Kevitsa intrusion, age and structure

The zircon U-Pb age of the Kevitsa intrusion is 2.057±5 Ga (Huhma et al. 1996, Mutanen & Huhma 2001). This correlates well with the Sm-Nd age of 2.05 Ga determined from primary igneous minerals (Huhma et al. 1996).

The 4x5 km^2 sized Kevitsa intrusion is funnel-shaped and dips to S-SW. The contacts cut across the surrounding metasedimentary strata, with basal contact dipping 45–50° to the S. Contacts are commonly interfingered with the country rocks. Igneous layering is parallel to the basal contact in the lower parts of the intrusion, 20–30° in the upper part of the ultramafic zone, and almost horizontal in the gabbro and granophyre zones. The intrusion has been divided into four zones (from base upwards): 1) marginal zone, 2) ultramafic zone, 3) gabbro zone, and 4) granophyre zone (Fig. 1).

The marginal (chill) zone is 0–8 m thick and consists of microgabbro, contaminated quartz gabbros and quartz-rich pyroxenites which grade rapidly to olivine pyroxenites of the ultramafic zone.

The ultramafic zone is most prominent in the NE part of the intrusion. The thickness of the zone is not known but is at least 1000 meters, possibly 2000 meters or more. The rocks are mostly olivine-augite mesocumulates (wehrlites and olivine websterites, here generally named as olivine pyroxenites), locally with plagioclase and/or orthopyroxene as cumulus or intercumulus phases with minor hornblende and phlogopite. Altered counterparts of olivine pyroxenites are named as metaperidotites which are composed of amphibole, serpentine, chlorite and talc. Within the ultramafic zone, there are discontinuous layers of pyroxenites and gabbros. Various types of komatiitic xenoliths are common in the ultramafic zone, especially within the mineralised part, whereas pelitic xenoliths are common closer to the contacts of the intrusion.

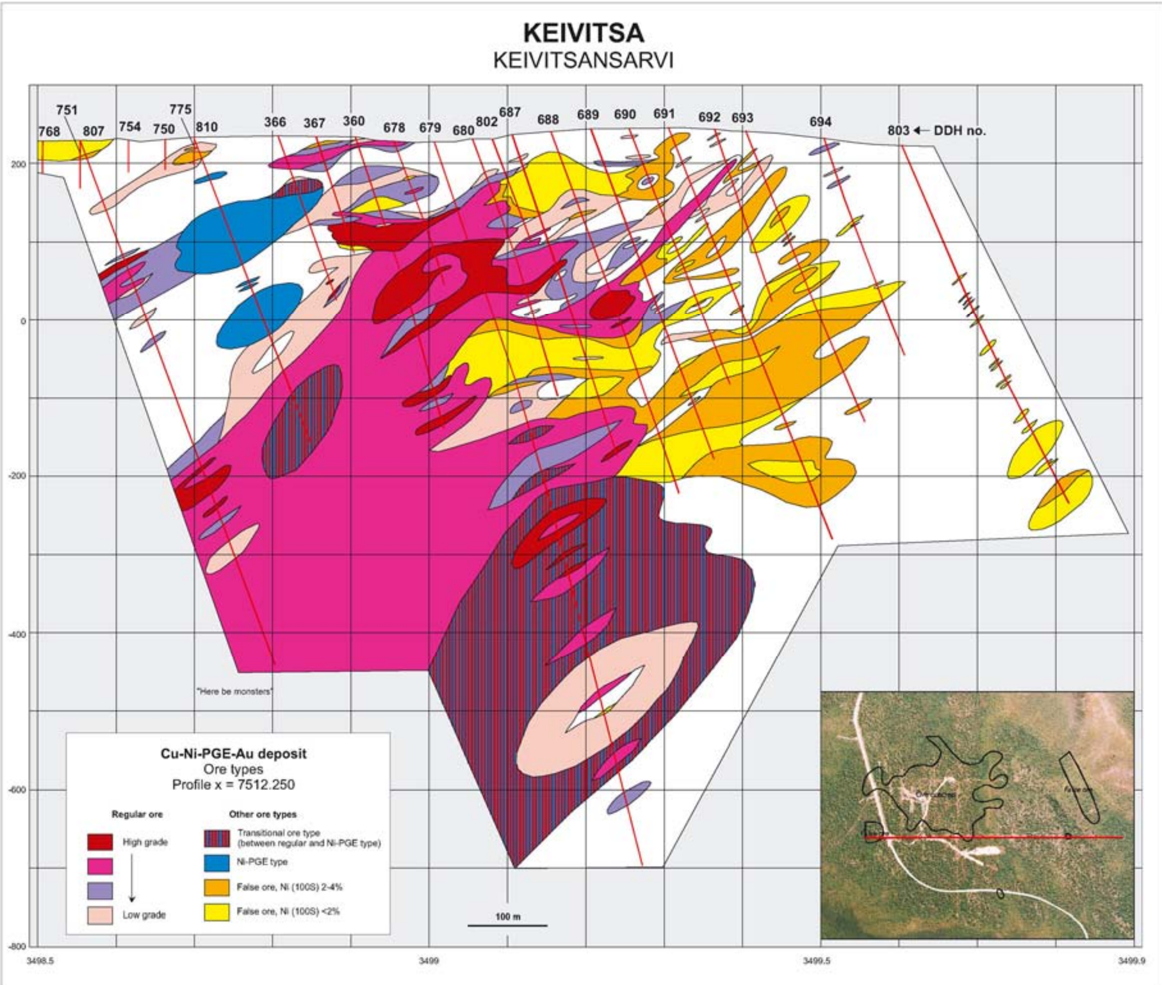

Figure 2. Cross section of the Kevitsa deposit, with different ore types indicated. DDH profile x=7512.250. From Mutanen (2005).

Rocks belonging to the gabbro zone are most prominent in the SW-part of the intrusion. They consist mainly of pyroxene gabbro, ferrogabbro (with pigeonite and fayalite), magnetite gabbro (with V-rich magnetite), and their metamorphic (hydrated) equivalents. Discontinuous units of Fe-rich, Mg-, Ni-, Cr-poor olivine pyroxenites occur in the upper parts of the gabbro zone. The gabbro zone contains large pelitic and minor komatiitic xenoliths. The thickness of the gabbro zone is not known, but drilling indicates that it is at least 500 m thick.

The magnetite gabbro of the upper part of the gabbro zone rapidly grades into the granophyre which represents the uppermost magmatic unit of the Kevitsa intrusion. The granophyre is mainly composed of sodic plagioclase, quartz, and secondary amphibole, with abundant magnetite, ilmenite, fluorapatite and sulphides. The granophyre contains small pelitic xenoliths.

As can be seen from the above, various types of xenoliths are common and they are encountered in various parts of the intrusion. The most common types are komatiitic and pelitic xenoliths. Komatiitic xenoliths occur as massive, banded, or layered rocks that have been mechanically disintegrated to a variable degree. They are composed of variable amounts of olivine, clinopyroxene, orthopyroxene and chromite. Komatiitic xenoliths are especially common in the ore zone and there is a 4–10 m thick xenolith-rich layer in the upper part of the ore that has been traced for 300 m from north to south. It is interesting to note that komatiitic xenoliths within the ore zone contain fine-grained disseminated sulphides, while those from the barren parts of the intrusion do not contain sulphides. Pelitic rocks, now pyroxene-plagioclase hornfels, occur as large xenoliths which are often partially digested ("rotten" xenoliths). Small (5–10 cm across) graphitic xenoliths indicate assimilation of graphite-rich black schist material. Graphite-rich pelitic hornfels xenoliths are also associated with pyrrhotite-rich sulphides 2 km west of the Kevitsa deposit.

Various types of dyke rocks cut the Kevitsa intrusion. They can be broadly classified into three categories: gabbro, diorite-felsite, and diabase. Porphyric gabbroic veins with sharp contacts represent the earliest phase. They have been interpreted as local evolved intercumulus liquids, based on chemical and mineralogical composition. The diorite-felsite veins show a paragenetic and compositional continuum and, indeed, form also composite veins with felsite occurring in the middle of diorite veins. These rocks are made of variable amounts of plagioclase, hornblende, and quartz. U-Pb zircon gives a comagmatic age of 2.054±5 Ga (Mutanen & Huhma 2001). Diabase and related olivine gabbro-diabase dykes are younger than the intrusion with a Sm-Nd mineral age of 1.916 Ga (Mutanen 2005). The ENE-striking dykes have fine-grained chilled contacts with the intrusion rocks. A typical feature of the olivine gabbro-diabase dykes is the presence of coarse-grained (up to 2 cm) olivine crystals in the mid-parts of the dykes.

The Kevitsa Cu-Ni-PGE deposit

The Kevitsa deposit is a large, low-grade disseminated sulphide deposit located in the upper part of the ultramafic zone, in the NE part of the intrusion (Fig. 1). The surface cross-section of the ore body is about 13.4 hectares and it extends to the depth of >400 meters. The host rocks are olivine pyroxenites and their metamorphic equivalents (metaperidotites). The deposit has been divided into two bodies, the main ore body (or Main Ore) and the overlying Upper Ore. There are four main ore types, based on the metal and sulphur contents: 1) regular ore, 2) false ore, 3) Ni-PGE ore, and 4) transitional ore. As distribution of Cu, Ni, PGE+Au, and S within the deposit is complex and variable, the different ore types tend to grade into another. Figure 2 shows a section trough the Kevitsa deposit and gives some indication on the distribution of different ore types: Regular ore makes up most of the main ore, whereas the false ore mainly occurs in the eastern part of the deposit. The Ni-PGE type mostly occurs in the upper parts of the ore, forming N-S trending pipe-like bodies, 30–50 m long, 10–30 m wide and extending to a depth of up to 400 m (Gervilla et al. 2003, Kojonen et al. 2004). The Ni-PGE ore is shown in blue in the centre of the open pit model in Fig. 3 (by J. Parkkinen, courtesy of Scandinavian Minerals Ltd).

The regular ore typically contains 0.4–0.6 % Cu, 0.2–0.4 % Ni, 0.015 % Co, 0.5–3.0 % S, and about 0.5–1.0 ppm of combined Pt+Pd+Au, giving an average Ni/Cu ratio of 0.6–0.8 and Ni/Co ratio of 15–25. The Ni content of the sulphide fraction typically is 4–7 %. Precious metals show fairly good positive correlation with the Cu+Ni values. Compared to the regular ore, the *false ore* typically is much more sulphur-rich (>5 % S) and grades locally into sulphide vein network. However, the metal contents are much lower, for example Ni is generally less than 0.1 % (< 4 % Ni in sulphide fraction). In the leanest false ore (0.3–0.5 % Ni in sulphide fraction), the Ni/Co ration is 1–2 which is similar to sulphides in the metasedimentary rocks surrounding the intrusion. The *Ni-PGE type* has a variable but generally fairly low sulphur content of about 0.5–1 % S, high nickel (>0.5 % Ni, 40–60 % Ni in sulphide fraction), high PGE (>1 ppm, up to 26.75 ppm (Gervilla et al. 2003)), and low copper (<0.1 % Cu) and gold contents (max. 0.13 ppm Au). The high Ni and low Cu contents give rise to high Ni/Cu ratio (generally >5, up to 50–90) and Ni/Co ratio (25–80). The *transitional ore* represents an intermediate ore type which is gradational in metal content to both the regular ore and the Ni-PGE ore. Its Ni/Cu ratio normally is 1.5–2.5 or higher and sulphide fraction nickel content 15–23 %. The changes in precious metal and sulphur contents between the different ore types is depicted in the sulphur vs. PGE+Au diagram, along with other data in Fig. 4.

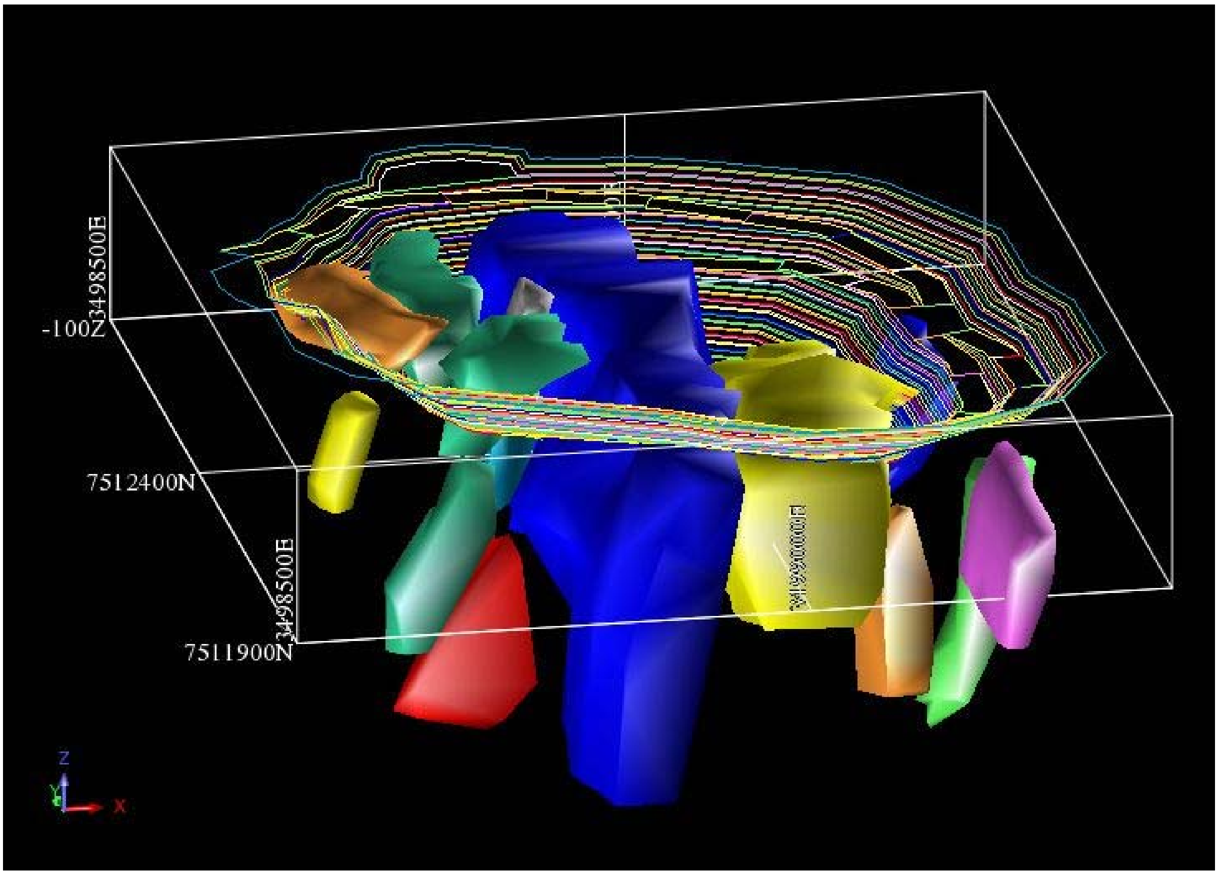

Figure 3. 3D block model of the Kevitsa deposit with Ni-PGE ore depicted in blue. Image by J. Parkkinen, Kevitsa Mining Oy.

Ore mineralogy

The main sulphide minerals at Kevitsa are troilite, hexagonal pyrrhotite, pentlandite, and chalcopyrite, with subordinate amounts of cubanite, talnakhite and magnetite, and a number of minor to trace mineral phases (Table 10 in Mutanen 1997). The Ni-PGE type of ore has a somewhat different paragenesis with pentlandite, pyrite, and chalcopyrite as the main phases, with subordinate, but locally abundant pyrrhotite, millerite, heazlewoodite, nickeline, maucherite and gersdorffite. The Ni-PGE type also contains graphite, whereas magnetite is rarer.

Altogether, about 40 platinum group minerals (PGM) have been identified from the deposit. The PGE mineralogy is dominated by various Pd-Pt-Te-Bi phases and speryllite, whereas PGE sulphides such as cooperite and braggite are more rare. However, the distribution of the PGMs is quite heterogeneous, as is evident from the study of Gervilla et al. (2003). They studied the PGE mineralogy of Ni-PGE ore from four different drill cores which intersected the ore at different depths. In their study, braggite was the most abundant PGE mineral and also geversite ($Pt(Sb,Bi)_2$) was locally abundant, with highly variable distribution of PGMs form hole to hole. About 55 % of the PGE minerals occur as inclusions in silicates (amphibole, serpentine, chlorite, pyroxene), 8–13 % as inclusions in sulphides, and 32–39 % are at silicate-sulphide grain boundaries.

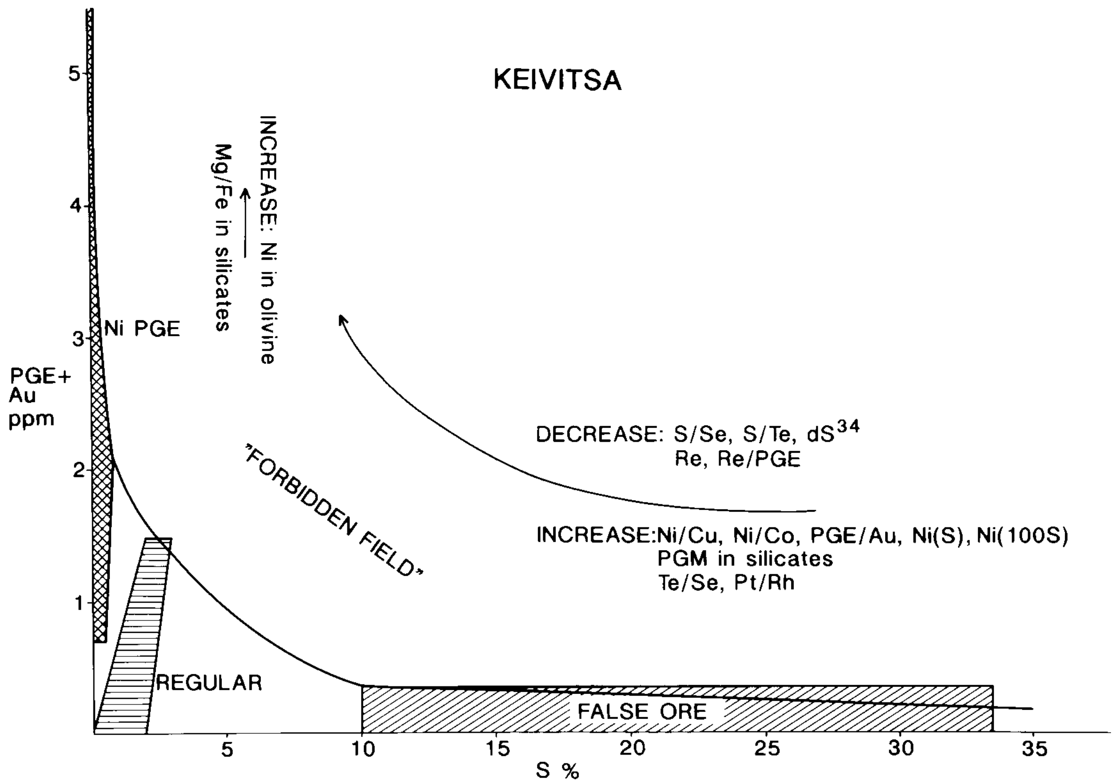

Figure 4. Precious metals vs. sulphur diagram showing geochemical trends between different ore types. From Mutanen (2005).

Contamination and ore genesis

The abundance of various supracrustal xenoliths attest to strong contamination by country rocks during the emplacement of the intrusion. Contamination is reflected in the isotope composition of the magma and different ore types. The initial epsilon Nd value of -3.5 and gamma Os value of +19.1 indicate substantial crustal contamination (Huhma et al. 1995). The average $\delta^{34}S$ value for regular ore is +4.0 ‰, for false ore it is +8.9 ‰, and for Ni-PGE ore it is +6.0 ‰ (Hanski et al. 1996). One analysis from a sulphur-rich sample in the marginal zone has $\delta^{34}S$ at +6 ‰. Gabbroic rocks have highly variable $\delta^{34}S$ values ranging from +5 ‰ in lower gabbros to a high of +24.4 ‰ in overlying graphite-bearing gabbros and ferrogabbros. Dunite inclusions have $\delta^{34}S$ between +5 to +9 ‰, whereas various metasediments have values between +1 to +24.4 ‰, with most values clustering between +17 to +20 ‰. All the intrusion rocks have high positive $\delta^{34}S$ values that are outside the range of values for typical mantle-derived sulphur, indicating variable contamination by heavy crustal sulphur. Of the different ore types, the false ore has the highest positive $\delta^{34}S$ values indicating the most substantial contamination by sedimentary country rocks (Fig. 4). Contamination is also reflected in the high Cl content of all of the ore types as well as barren ultramafic rocks and the presence of primary Cl apatite and Cl amphibole (dashkesanite).

Two models have been proposed for the ore genesis. Mutanen (1997) attest the formation of the regular and false ore to contamination by variable amounts of komatiitic material and S- and C-rich metasediments, wherein the regular ore received some additional sulphur from the metasediments and additional nickel from the komatiitic material, whereas the false ore was more heavily contaminated by S-rich metasediments, which led to dilution of the ore (Fig. 4). The Ni-PGE ore type has many peculiar features, such as a high REE content (Fig. 5), high Ni content both in sulphide fraction and in olivine (about 1.5 % NiO in olivine, Fig. 6), low S, and a very low Cu content, which make the origin of this ore type more enigmatic. Furthermore, the Ni-PGE ore type formed in a highly reducing, S-poor environment caused by assimilation graphite-rich metasediments, with also some S coming from metasedimentary material (reflected in the S isotopes), and residual olivine from disintegrated komatiites contributing most of the Ni in the ore. The high olivine Ni content is explained by olivine equilibrium with metallic Ni in a highly reducing environment. A different kind of genetic model was proposed by Gervilla et al. (2003), whereby the Ni-PGE ore type is the product of leaching of S and Cu and/or remobilisation of PGE and Ni by metamorphic Cl-rich fluids resulting in the deposition of Ni-rich sulphides and, for instance, unusually Ni-rich braggite.

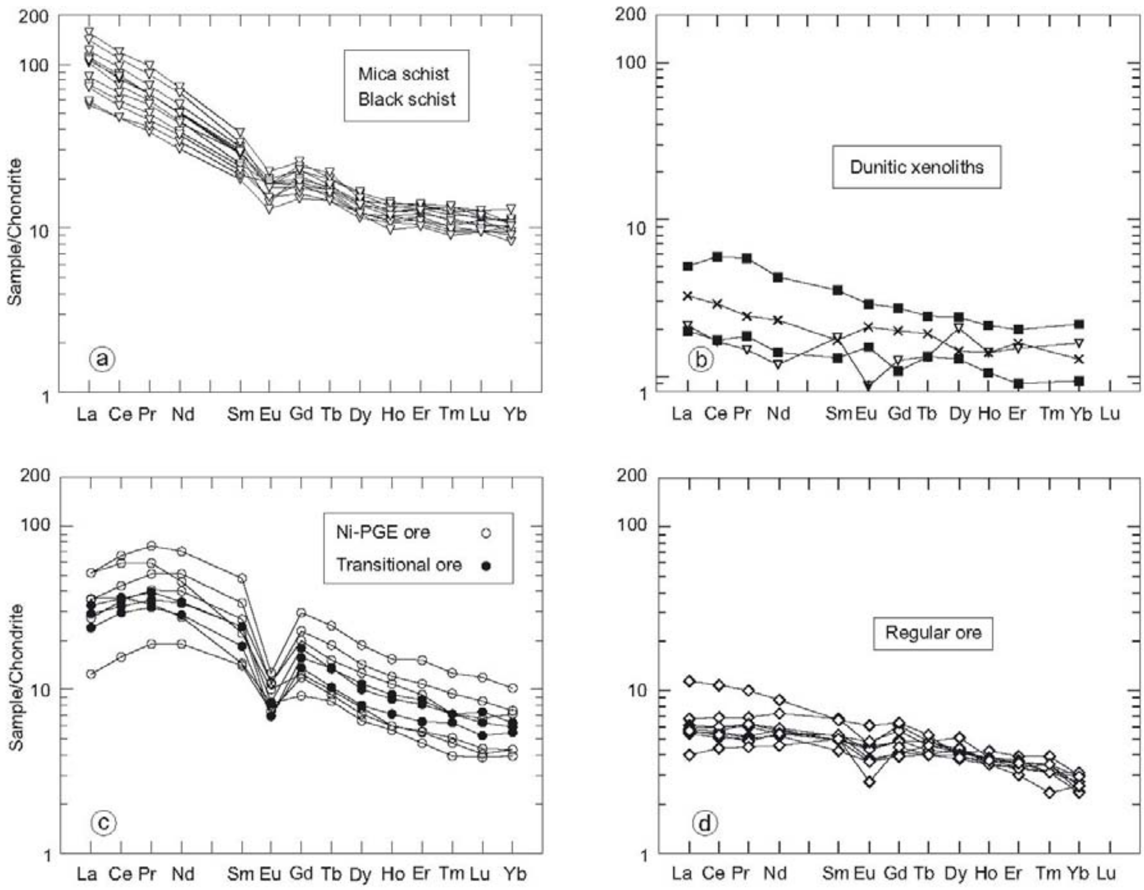

Figure 5. Chondrite-normalised REE diagram for various rock types of the Kevitsa complex. From Mutanen (2005).

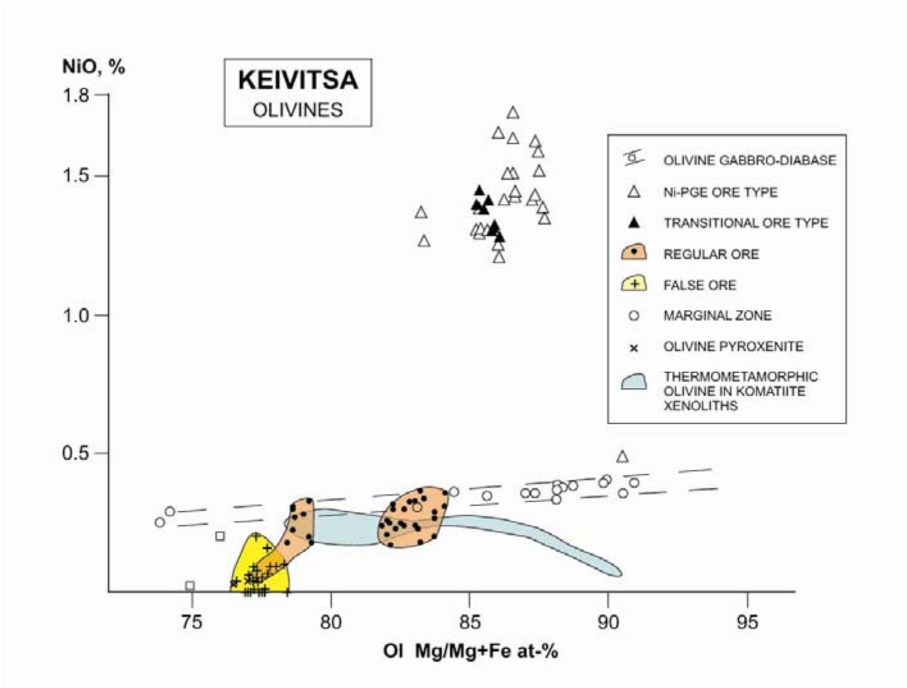

Figure 6. Kevitsa intrusion olivine NiO (%) content vs. olivine Mg/Mg+Fe (at-%) ratio. From Mutanen (2005).

Mineral resources of the Kevitsa intrusion

Table 1 shows the current mineral resource of the Kevitsa deposit, to a depth of 400 meters, based on the pre-feasibility study of 2006. A full bankable feasibility study is currently underway.

Table 1. Mineral reserve to 400 meters depth at 0.18 % Ni cut-off, July 2006 (Scandinavian Minerals 2007).

	Tonnage	Ni %	Cu %	Co %	Au ppm	Pd ppm	Pt ppm
Proven	56.2 Mt	0.295	0.415	0.014	0.141	0.201	0.310
Probable	10.6 Mt	0.295	0.492	0.015	0.142	0.171	0.267
Total	66.8 Mt	0.295	0.427	0.014	0.141	0.196	0.303

Acknowledgements

The paper was reviewed and improved by Dr. Tapani Mutanen who is thanked for his help. The authors also thank Scandinavian Minerals Ltd and Kevitsa Mining Oy for providing material for this report.

Excursion stops

Stop 1. Kevitsa Cu-Ni-PGE deposit

Variably mineralised olivine pyroxenite and metaperidotite are visible in the recent test pit and rock piles. We will also see some drill core of the various rock and ore types.

Stop 2 Pitted peridotite

Pitted, heterogeneous feldspathic olivine websterite. Orthopyroxene, biotite and sulphide containing parts are eroded into pits, whereas cpx-rich parts are preserved as knobs.

Optional stop

Stop 3. Hanhilehto Hill. Gabbro and roof hornfels

Strongly altered, medium-grained rocks of the gabbro zone. Also pelitic and black schist metahornfels of the roof rocks of the intrusion.

Day 3: The Suurikuusikko gold deposit (Kittilä Mine) and the Kolari IOCG deposit

Suurikuusikko gold deposit

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Introduction

The orogenic Suurikuusikko gold deposit occurs within the Palaeoproterozoic Central Lapland Greenstone Belt, approximately 50 km northeast of the town of Kittilä in Finnish Lapland (Figs. 1 and 2). The host rocks, timing of ore formation relative to regional deformation, metamorphic grade, alteration assemblages present, and structurally controlled nature of the deposit make it analogous to better known deposits in greenstone belts throughout the world (e.g., Yilgarn of Australia, Superior Province of Canada). Gold is refractory, occurring within arsenopyrite (>70%) and arsenian-pyrite as lattice-bound gold or sub-microscopic inclusions. A 13-year mining operation is planned to start in 2008 targeting a gold resource of 16 million tonnes (2.6 million ounces) averaging 5.1 grams per tonne gold (Agnico-Eagle Ltd. 21.2.2007).

Exploration History

Visible gold was discovered SSW of Suurikuusikko by the Geological Survey of Finland (GTK) in 1986 (Härkönen & Keinänen 1989). Subsequent ground geophysical surveys and geochemical sampling lead to the identification of the Kiistala Shear Zone (KiSZ), the deposit's host structure. Suurikuusikko was discovered in 1986 during diamond drilling by GTK. A total of 77 diamond drill holes (9,320 metres) were completed by GTK, outlining a resource of 1.5 million tonnes with an average grade of 5.9 grams per tonne (285,000 ounces of gold) by 1997. In April 1998, the deposit was acquired by Riddarhyttan Resources AB and the company's exploration activities increased the resource size to over 2 million ounces of gold (Bartlett 2002). Ore-grade mineralisation was found over a five-kilometre strike length of the KiSZ in similar structural and stratigraphic positions. Mine feasibility studies on Suurikuusikko began in winter 2000. In 2004, Agnico-Eagle Mines Limited acquired a 14 % ownership interest in Riddarhyttan, and in 2005 acquired the remaining Riddarhyttan shares. In June 2006 a decision was made to begin mine development.

Resource

The resource estimate for the deposit as at 21.2.2007 was 2.6 Moz gold (16 Mt @ 5.1 g/t gold). Production is estimated to commence in the second half of 2008 generating 150,000 ounces of gold each year for 13 years (Agnico-Eagle 2007). Ore intersections have very even grade distribution due to the 'disseminated sulphide-like' nature of the ore (i.e. negligible nugget effect). Table 1 shows examples of typical ore intercepts in drill core.

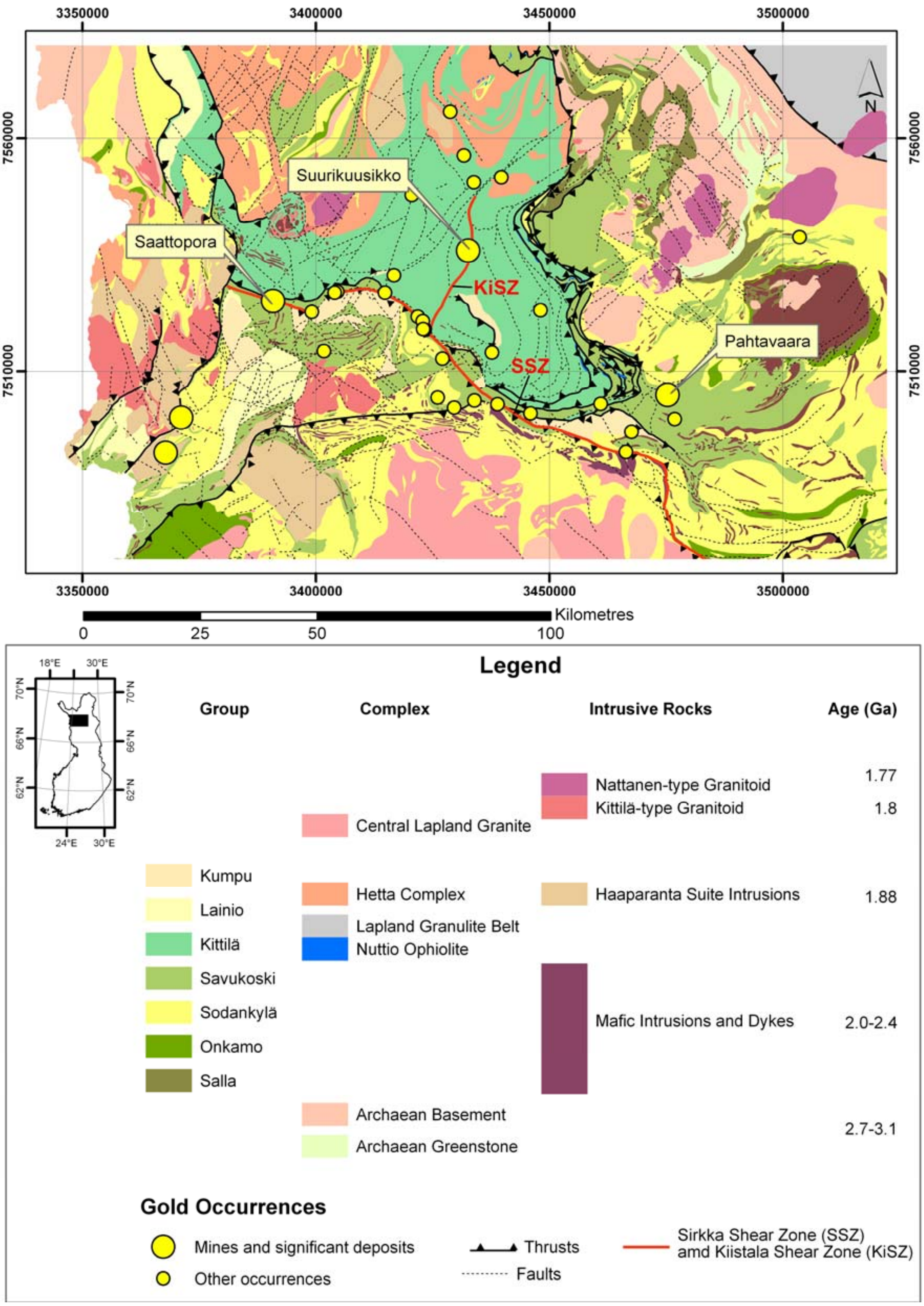


Figure 1. Formation map of the Central Lapland Greenstone Belt (after Lehtonen et al. 1998) showing the location of the gold deposits and occurrences in thea area, with the three largest known deposits named.

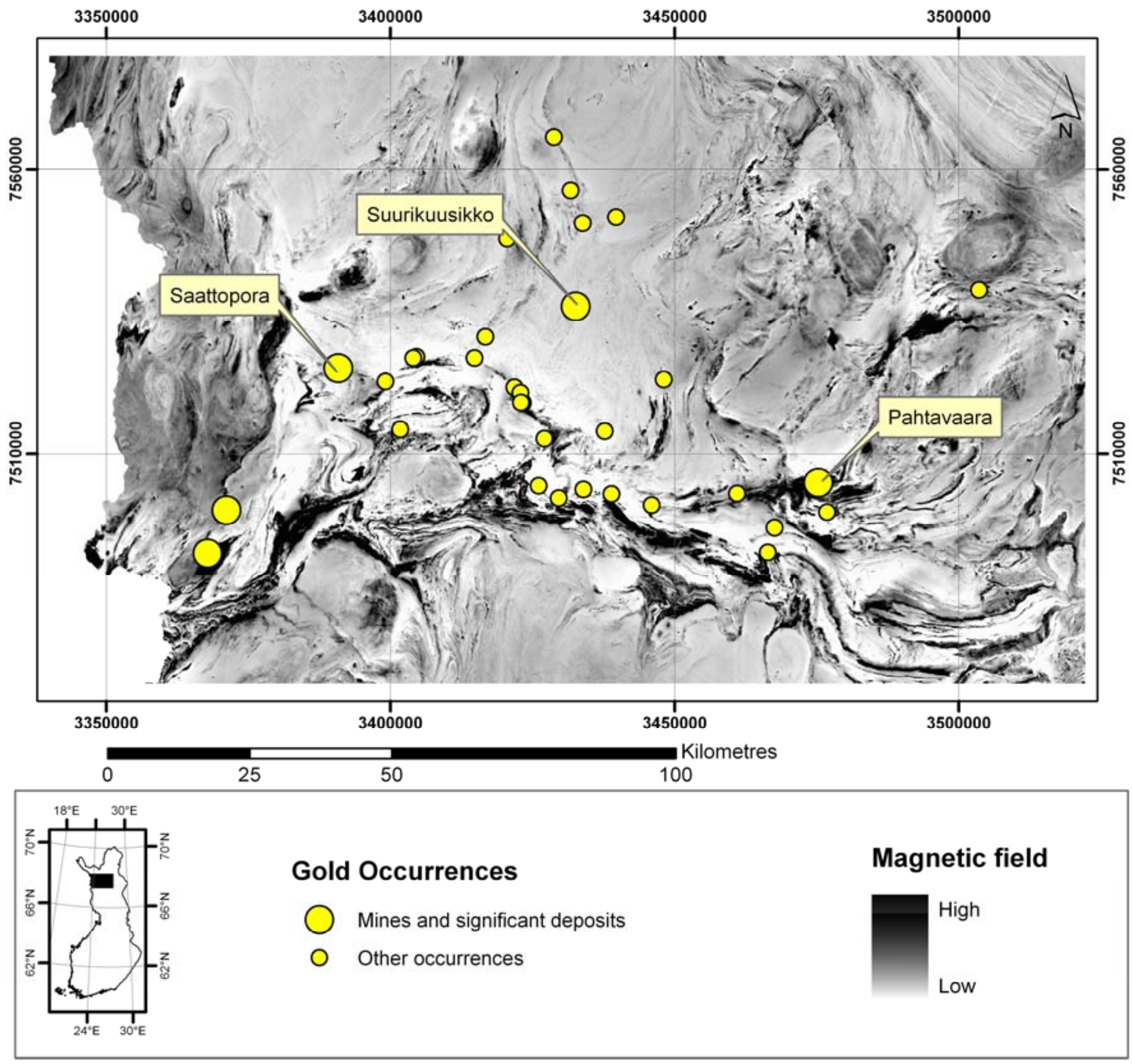


Figure 2. Greyscale aeromagnatic map of the Central Lapland Greenstone Belt (area as in Figure 1). Data from Geological Survey of Finland.

Table 1. Suurikuusikko. Examples of gold intercepts from drill core.

Zone	Drill hole number	Mineralised section length (m)	Averaged grade of section (g/t Au)
Ketola	02114	6.40	4.20
Ketola	02107	7.00	11.10
Ketola	02107	3.20	7.10
Ketola	02104	10.70	4.00
Etelä	R407	7.00	7.50
Etelä	01802	5.60	8.60
Etelä	02039	8.10	9.50
Main	R473	14.00	10.40
Main	R504	10.80	9.10
Main	00717	14.30	10.60
Main	R478	18.20	5.10
Main	99002	18.20	16.50
Main	R479	26.80	17.30
Main	00730	18.90	9.10
Main	98004	29.60	11.90
Main	00903	46.20	8.90

Geology

Suurikuusikko occurs within c. 2.0 Ga (Lehtonen et al. 1998) greenschist-facies metavolcanic rocks of the Kittilä Group (Fig. 1). Geochemical heterogeneity among Kittilä Group rocks has been interpreted to indicate that the Group is a composite of arc terranes and oceanic plateaux amalgamated during oceanic convergence (Hanski & Huhma 2005). Significant variations in metamorphic grade within the Group also suggest that a number of distinct elements could be present within the area currently mapped as Kittilä Group, and seismic surveys across central Lapland indicate a number of distinct crustal blocks (Patison et al. 2006a). The maximum current thickness of the Kittilä Group is between six and seven kilometres (Luosto et al. 1989) in the Suurikuusikko area.

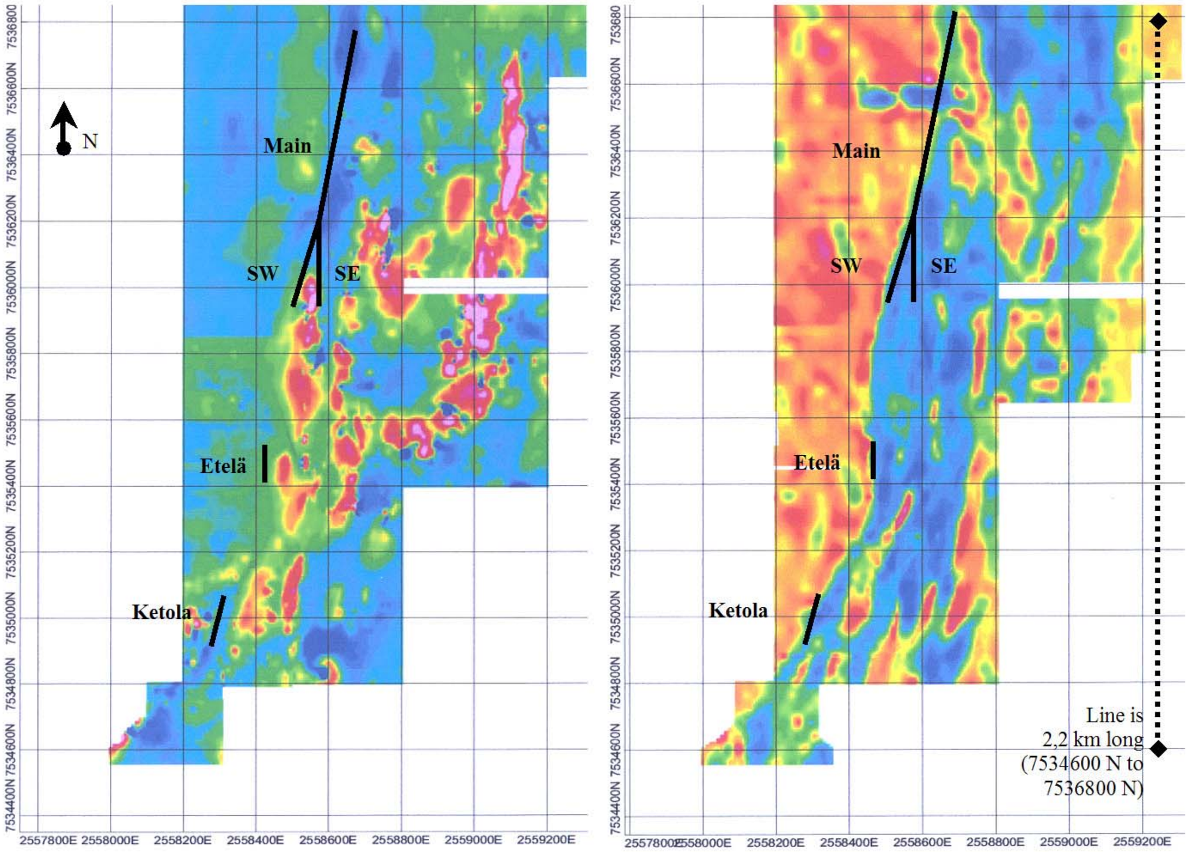


Figure 3. Total magnetic field (3a) and electromagnetic (slingram out-of-phase, 3b) images for the southern part of the Suurikuusikko area. The blue colour represents magnetic lows and conductivity highs respectively in Figures 3a and 3b. Names refer to individual ore zones

The mineralisation typically occurs in a transitional horizon between two thick (several 100 metres) mafic lava sequences (Figs. 3 and 4). The N- to NNE-trending host structure for the deposit (KiSZ) coincides with this contact between western and eastern lava packages. In the area of the 'Main' ore zone, host rocks change from mafic pillow and massive lavas west of the mineralised zones to mafic transitional to intermediate lavas (andesite flows of Powell 2001) and minor pyroclastic material within mineralised zones. Graphitic sediment intercalations containing chert, argillitic material and BIF occur within mafic volcanics at the eastern margin of mineralised zones, followed further east by mafic lava packages and ultramafic volcanic rocks. The extent of intermediate and felsic rock compositions present at this deposit is not studied. The variation in appearance (and hence the logging and mapping terminology for rock compositions used here) may also alternatively result from progressive alteration of mafic rocks. Most ore is hosted by mafic rocks and those mapped as intermediate or felsic. Metasedimentary units including BIF typically have low to no gold grade, and ultramafic rocks are unmineralised.

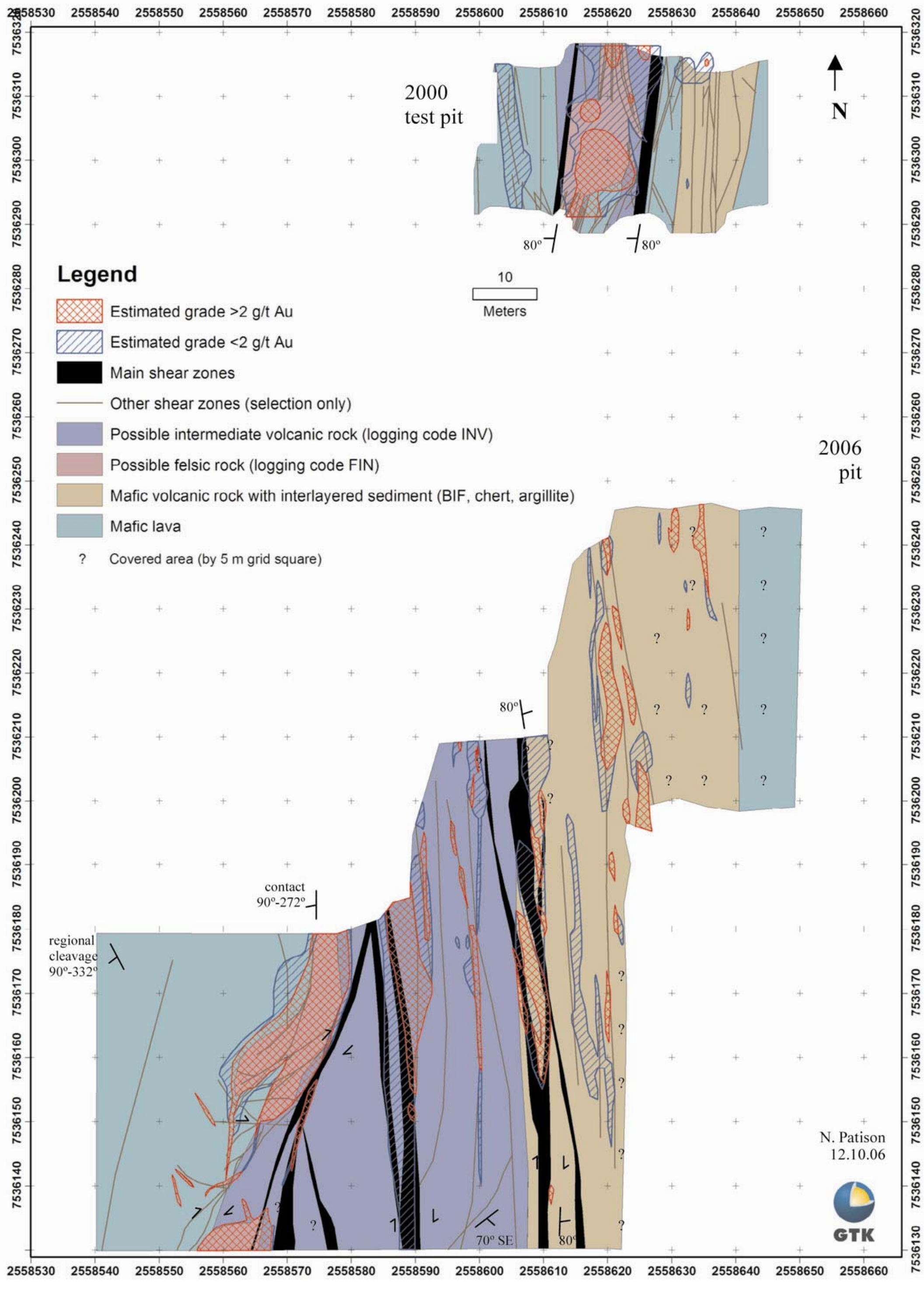


Figure 4. Pit map showing the main rock types and structures (after Patison et al. 2006b). The grade estimates shown are visual estimates based on arsenopyrite abundance. Exposure of the deposit prior to 2007 was limited to the two pits shown in this Figure.

Orogenic events relating to CLGB development generated several phases of deformation. The earliest deformation phases preserved (D_1, D_2) involved roughly synchronous N- to NNE- and S- to SW-directed thrusting at the southern and northeastern margins of the CLGB (Ward et al. 1989). Northwest-, N-, and NE-trending D_3 strike-slip shear zones, including the Suurikuusikko Shear Zone hosting the Suurikuusikko deposit, cut early folding and thrusting, but may also reflect reactivation of older structures. Post-D_3 events are limited to brittle, low-displacement faults.

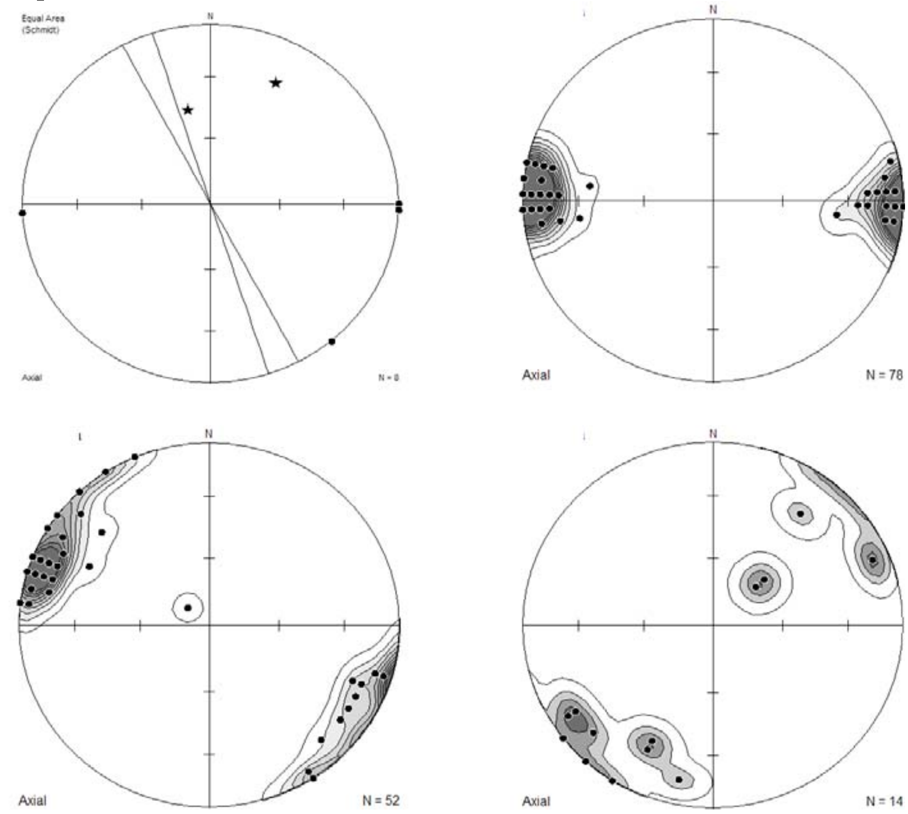
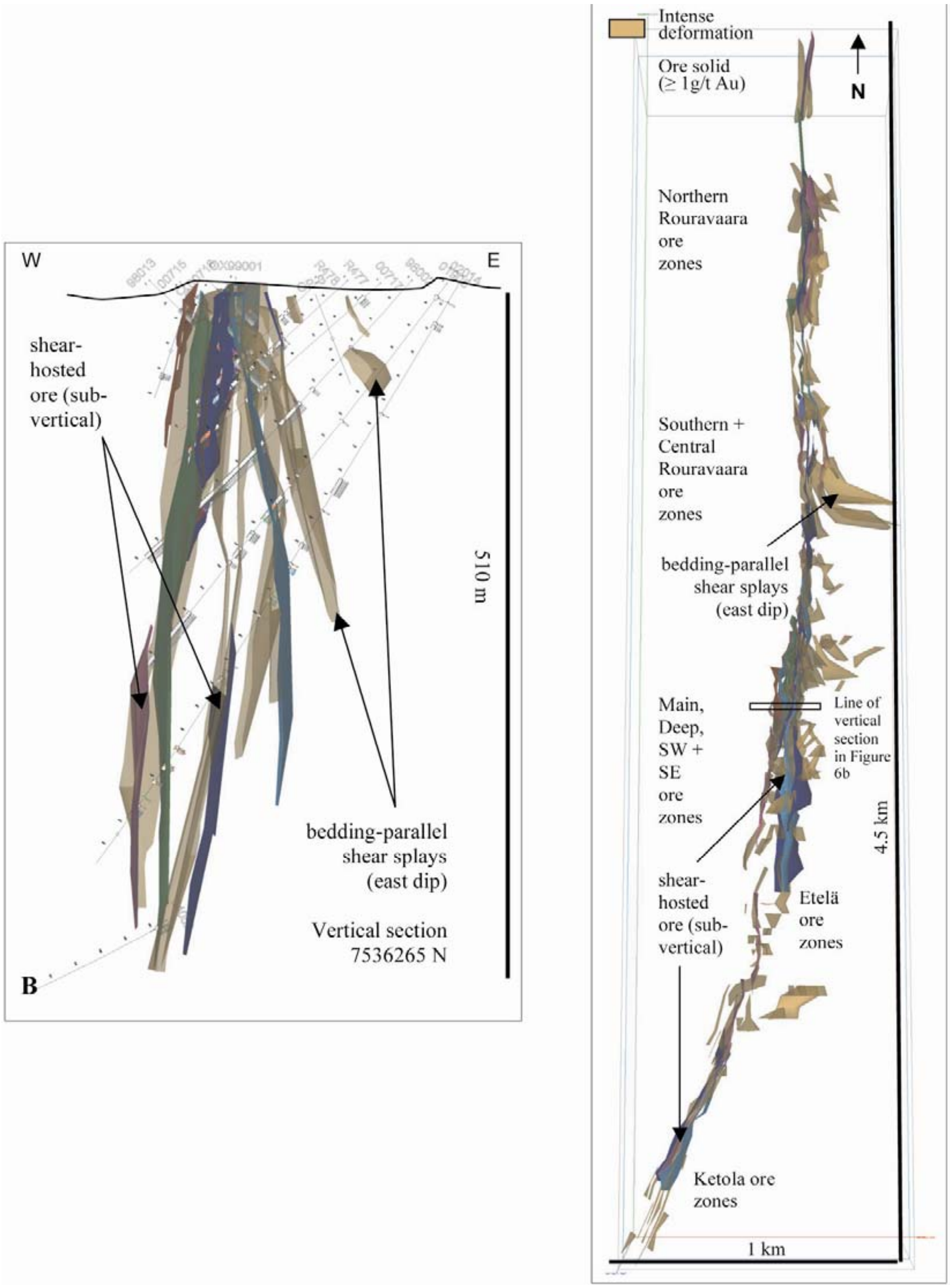


Figure 5. These stereoplots show the orientations of deformation features observed for Suurikuusikko (ordered from oldest to youngest). Figure 5a shows bedding (dots), the trend of the typical regional foliation (lines) formed prior to movements of the KiSZ related to mineralisation, and fold axes measured in the deposit area (stars). Figures 5b and 5c show the orientation of the 'graphitic' shear zones (e.g., Figure 4) associated with the KFZ and ore zones. Figure 5d shows the common orientation of post-mineralisation faults, although NE- (e.g., Figure 6a) and E-striking faults and veins are also seen. Plots are lower hemisphere projections on equal area nets; point symbols are poles to planes with frequency contours, stars in Figure 5a are plunging lines; lines are planes). Plots after Patison et al. 2006b and Patison 2001

Representative structural data for the deposit are shown in Figures 5a to 5d. The Kiistala Shear Zone has a strike length of at least 25 kilometres (Figs. 1 and 2). The dip of this shear zone in the Suurikuusikko area is steeply west to sub-vertical (Figs. 5b and 5c). Known mineralisation occurs within N-trending and less frequently NE-trending (e.g., Ketola ore bodies, Fig. 3) shear zone segments. The KiSZ is a complex structure, recording several phases of movement. A minor degree of west-up movement has occurred, but most deformation has occurred by flattening accompanied by some sinistral and dextral strike-slip movements. Aeromagnetic images of the KiSZ indicate early sinistral strike-slip movement along the zone. Immediately above the widest mineralised zones, late dextral strike-slip movements are recorded on shear planes bounding mineralised zones. It is not yet clear if the timing of mineralisation coincides with a combination of early and late shearing or only to the later dextral shearing event which now delineates the limits of gold mineralisation in most ore zones. An apparent correlation exists between points of more intense shearing within the KiSZ and the amount of gold present in host rocks (Figs. 6a and 6b).



Figures 6a and 6b. Suurikuusikko 3D solid geology models (Patison 2006b). Coloured solids are assay-based ore solids for gold grade (≥ 1 g/t). The brown solid is the host shear zone constructed to show zones were deformation intensity is highest. Figure 6a (near plan view with slight N plunge) and Figure 6b (vertical section) show the shear-bound nature of the ore zones. Sheared bedding contacts which are also mineralised (unmineable grades at the time *of model creation) are illustrated by the moderately east-dipping solids. Truncation by cross faults is also evident in Figure 6a*

The envelopes of ore bodies strike N and have a moderate N plunge. The control on the northerly plunge is not completely resolved: factors to be explored include the role of intersections between multiple shear planes, and of the intersections of depositional surfaces and shear planes. The orientation of regional fold axes (similar to axes in Fig. 5a) may also have a role in determining favourable sites for mineralisation during shearing. Sulphides and host rocks show some evidence for deformation relating to post-mineralisation movements on host shear planes. Post-mineralisation brittle faults crosscut mineralised zones but are not known to cause significant displacement of ore lenses.

Alteration in and around the deposit has not been studied in great detail, but appears typical for deposits of this type. Visually, intense carbonate and albite alteration are associated with gold-bearing arsenopyrite and pyrite. Albite occurs as a matrix overprint that typically extends less than two metres into barren rock, and as brecciating micro veinlets. Barren carbonate alteration includes distal calcite veins, and dolomite/ankerite veins and infilling tectonic and/or hydrothermal breccia proximal and within ore zones, respectively. Table 2 presents a summary of progressive alteration of mafic pillow lavas.

Table 2. Table shows an example of the minerals present in progressively altered mafic pillow lava. The data used are modal weight percentages of mineral phases calculated using Mineral Liberation Analysis data collected at GTK. The thickness of line is proportional to the relative volume of each mineral present in the sample. A mafic pillow lava sequence was used for this example to ensure a constant rock type, although pillow lavas do not host significant volumes of ore. The 'felsic' mineralised sample is included for comparison and may, in fact, be the most altered end-member of a mafic rock alteration sequence.

Alteration Zone	Distal	Intermediate	Proximal / Ore	Ore	Ore
Rock type	Mafic Pillow lava	Mafic Pillow lava	Mafic Pillow lava	Mafic Pillow lava	'Felsic'
Sample	F5-001	F5-007	00404 189.90	F5-003	F5-002
SILICATES					
Actinolite		▬▬▬			
Epidote		▬▬▬▬▬			
Titanite	------------------	---------------		-------------	
Chlorite	▬▬	▬▬▬	------------------	▬▬▬	
Muscovite			------------------	-------------	------------
Albite		▬▬▬	▬▬	▬▬	▬
Microcline		---------------	------------------		------------
Plagioclase	▬▬▬				
Clinopyroxene (matrix)		=========			
Quartz	------------------	—————	------------------	-------------	------------
CARBONATES					
Calcite	========	—————		▬▬▬▬	
Dolomite			▬▬▬	- - - - - - - -	▬▬▬▬
PHOSPHATES					
Apatite	- - - - - - - - - -	- - - - - - - - -			
OXIDES					
Rutile		---------------	------------------	-------------	------------
SULPHIDES					
Arsenopyrite			==========	========	=======
Pyrite			——————	—————	=======
Pyrrhotite			------------------	-------------	------------
GOLD GRADE (g/t)	0	0	5.16	3.3	8.71

Absent formTable 2 is amorphous carbon. The abundance of this ‘graphitic’ carbon correlates with the intense shearing that bounds most mineralised zones. The presence of such carbon suggests extremely reducing fluid conditions during shearing and possibly mineralisation. Gold-bearing sulphides commonly nucleated on shear planes, stylolitic cleavage, and fractures bearing amorphous carbon (Figs. 7a and 7b). Carbon isotope data indicates that this material is sourced from carbon-rich sediments within the host sequence (Patison, unpublished data). Argillite-rich horizons intercalated with volcaniclastic material have high primary carbon contents, and may have been chemically important for localising gold-rich phases given the association between amorphous carbon alteration and mineralisation. Other alteration and ore mineral phases include rutile and less abundant sericite, tetrahedrite, chalcopyrite, gersdorffite, chalcocite, sphalerite, bornite, chromite, galena, talnakhite, and Fe-hydroxides (the latter produced by weathering) in varying abundances (Chernet et al. 2000).

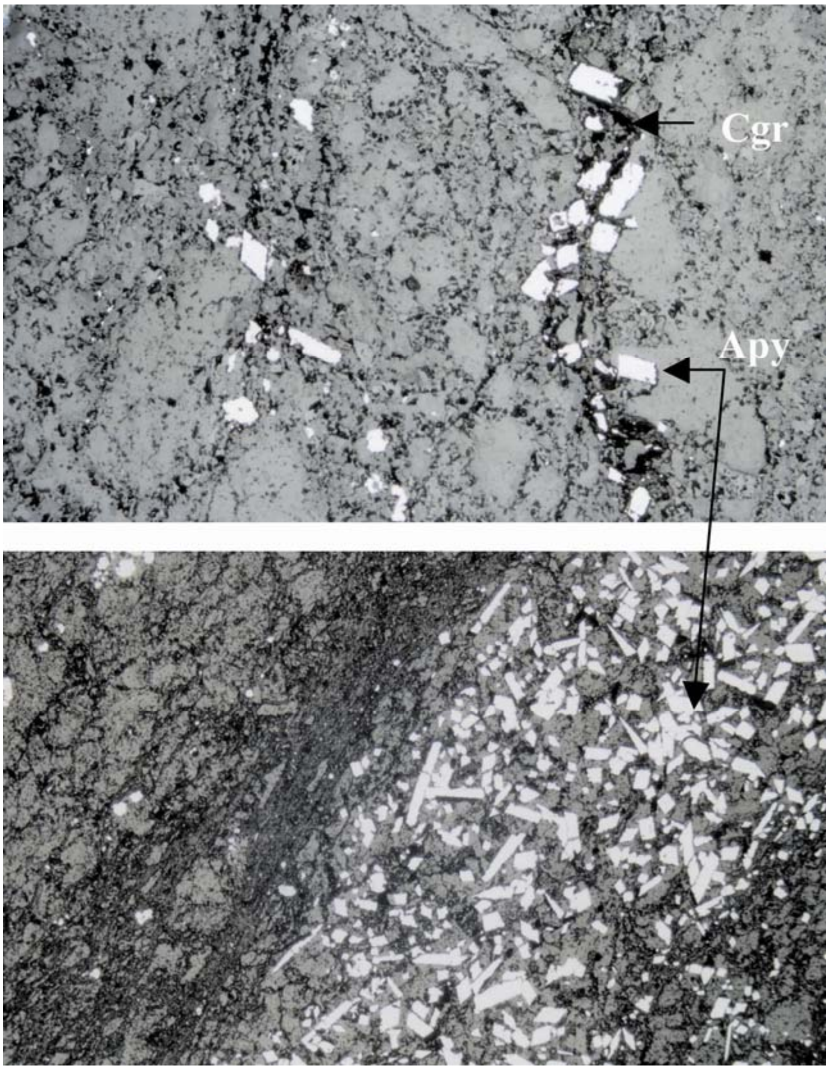


Figure 7a and 7b. Back-scattered electron microprobe images of ore samples in mafic host rocks. Figure 7a shows nucleation (or recrystallisation) of arsenopyrite (Apy) in graphitic (black phase, Cgr) milled zones relating to shearing. Figure 7b shows a fractured competent rock fragment with arsenopyrite associated with fracture infill. A penetrative shear boundary is seen at the edge of this mineralised fragment (center of photograph). Magnification for both images is times 2.5

The gold-rich sulphides appear to have a late timing within the paragenetic sequence at Suurikuusikko. The majority (71 %) of gold occurs within arsenopyrite, and visible arsenopyrite is a reliable indication of the presence of gold within samples. Remaining gold occurs in arsenian-pyrite (22 % of gold), and infrequently as free gold (Kojonen & Johanson 1999). Sub-microscopic gold is found as inclusions or solid-solution lattice substitutions within arsenopyrite and pyrite (Chernet et al. 2000). Gold as inclusions is common in pyrite but rare in arsenopyrite (typical grain size from <1 to 100 mm; Kojonen & Johanson 1999). The composition of gold inclusions includes various alloys with silver and mercury (Chernet et al. 2000). Rare stibnite veins and amorphous grains contain extremely high gold grades and overprint the main ore-bearing sulphides.

Acknowledgements

Agnico Eagle is gratefully acknowledged for the permission to publish the data. The information in this summary reflects the opinions of the author only unless otherwise referenced. The majority of information is based on data collected prior to 2005 (during the exploration phase of the deposit preceding mine development).

Stop 1 Suurikuusikko gold deposit

Ore and alteration zone in handspeciamens, boulders and drill core. Open cut and/or underground visits depend on the access.

Hannukainen, Kolari

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Introduction

The Kolari region is in the western part of the Central Lapland greenstone belt (CLGB) which was formed during prolonged stages of rifting, sedimentation and magmatism in cratonic margin and intracratonic rift settings between 2.5 and 1.9 Ga (Hanski et al. 1997, Sorjonen-Ward et al. 2003). The general geology of the region is shown in Figure 8.

The bedrock of the Kolari region consists of Onkamo, Savukoski, Sodankylä and Lainio group supracrustal rocks, and ca. 1.86 Ga monzonite and diorite intrusives (Hiltunen 1982, Niiranen et al. 2007), as well as of a smaller amount of ca. 1.80 Ga granitoids. A few ca. 2.2–2.0 Ga dolerites crosscut the Sodankylä and uppermost Lainio Group supracrustal rocks in the area.

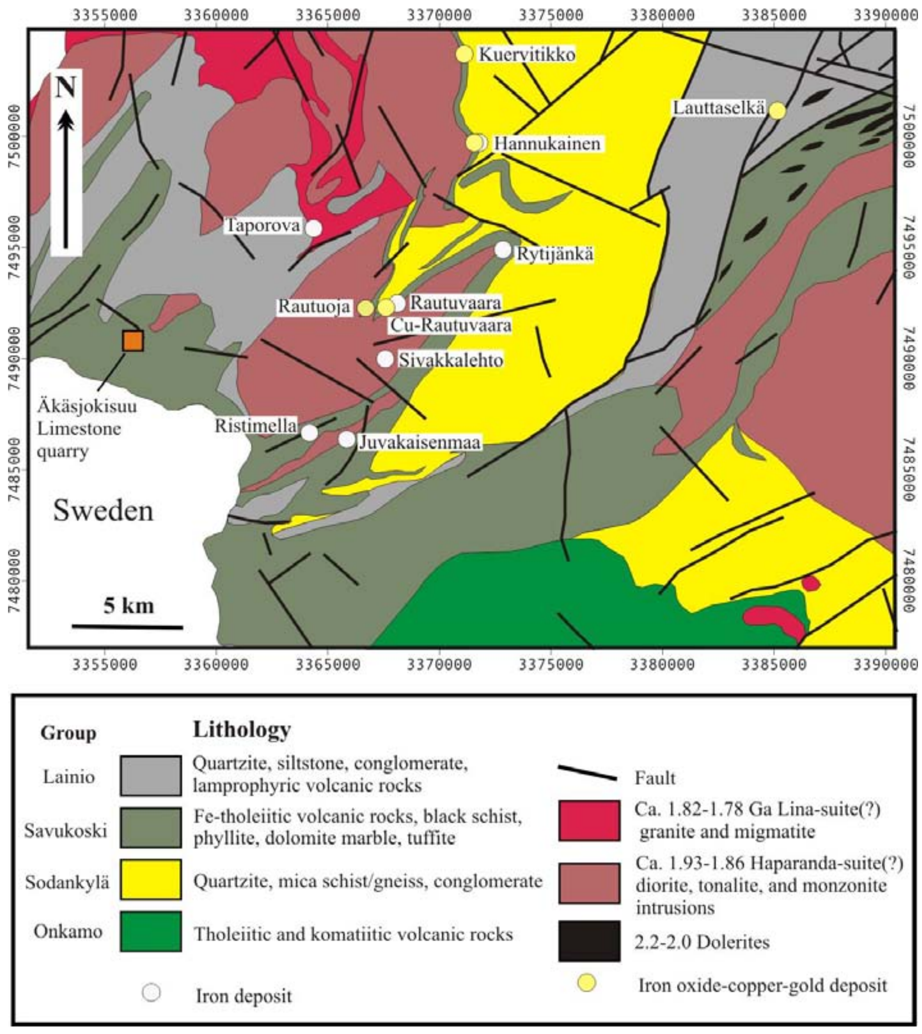


Figure 8. General geology of the Kolari region and location of Fe and Fe-Cu-Au deposits. Modified after Korsman et al. (1997). Lithostratigraphic groups as in Figure 9 in the main section Introduction.

Several iron oxide, and iron oxide-copper-gold deposits occur in the Kolari region (Fig. 8). Most of these deposits are near the contact between syn-orogenic monzonite or diorite intrusions and the supracrustal, 2.20–2.05 Ga, Matarakoski Formation rocks of the Savukoski Group. All of the iron oxide-copper-gold deposits are located near to fault or shear zones that crosscut or lie along the contact between the intrusions and supracrustal rocks.

Iron oxide-copper-gold deposits

Five of the known deposits and prospects in Kolari contain significant amounts of copper and/or gold: Cu-Rautuvaara, Hannukainen, Kuervitikko, Rautuoja, and Lauttaselkä (Figs. 8 and 9). The geological setting in all of these is similar, except at Lauttaselkä, which seems to completely be in Lainio Group rocks. The general characteristics, grades, and tonnages for the first four are given in Table 3.

The ores are hosted by diopside skarn and in quartz-albite rocks. Variably altered diorite intrusion forms the hanging wall, and mafic metalava, quartz feldspar schist, mica gneiss, and quartzite comprise the footwall rocks. Other rock types in the ore zones are granitic pegmatite and albitite. Figures 10 and 11 show typical wall and host rocks and ore samples from the known Kolari IOCG deposits.

Table 3. Grades tonnages and general characteristics of the Laurinoja, Cu-Rautuvaara, and Kuervitikko IOCG occurrences.

Deposit/Prospect	Hannukainen	Rautuvaara	Kuervitikko	Rautuoja
Size & Grade	Total ca. 66 Mt @ 0.2–4 g/t Au, 0.1–2 % Cu, 40–55 % Fe. Laurinoja ore body, Au-Cu production: 4.56 Mt @ 1 g/t Au, 0.88% Cu	Cu-Rautuvaara: 2.8 Mt @ 0.48 % Cu; NE-Rautuvaara: 13.3 Mt @ 0.20 % Cu, 46.8 % Fe; SW-Rautuvaara: 4.5 Mt @ 42.7 % Fe, 0.81 % Mn	1.2 Mt @ 1 ppm Au, 0.3 % Cu	1.9 Mt @ 0.34 ppm Au, 0.19 % Cu, 36.7 % Fe
Enriched elements in the ore	Fe, Cu, Au, S ± Ag, Bi, Ba, Co, Mo, Sb, Te, LREE, Zn	Fe, Cu, S ± Au, Ba, Bi, Mo, Se, Th, Te, U, LREE, Zn	Fe, Cu, Au, S ± Ba, Bi, Mo, Se, Te, Zn	Fe, Cu, Au, S
Ore minerals	Mgt, Py, Po, Cpy ± Moly, Au, Tell	Mgt, Po, Py, Cpy ± Ura	Mgt, Py, Cpy, Po	Mgt, Po, Py, Cpy
Gangue	Di, Bt, Ab, Hbl, ± Sca, Oli, Cc, Ap, Ep	Di, Ab, Atf, ± Hbl, Bt, Chl, Ap	Di, Ab, Hbl, ± Bt, Chl, Cc, Ep, Ap	Not reported
Host rock(s)	Diopside skarn	Diopside skarn, Ab-Atf rock	Diopside skarn	Diopside skarn
Major wall rocks	Diorite, Mafic meta-lava, Qtz-Ab rock	Diorite, Mafic meta-lava, Ab-Atf rock	Diorite, Mafic meta-lava, Qtz-Ab rock	Diorite

Data from Hiltunen (1982), Keinänen (1995), Puustinen (2003), Korkalo (2006), Niiranen (2005), and Niiranen et al. (2007). Mineral abbreviations: Ab = albite, Ap = apatite, Atf = anthophyllite, Au = gold, Bt = biotite, Cc = calcite, Chl = chlorite, Cpy = chalcopyrite, Di = diopside, Ep = epidote, Hbl = hornblende, Oli = olivine, Mgt = magnetite, Moly = molybdenite, Po = Pyrrhotite, Py = Pyrite, Tell = tellurides, Sca = scapolite, Ura = uraninite.

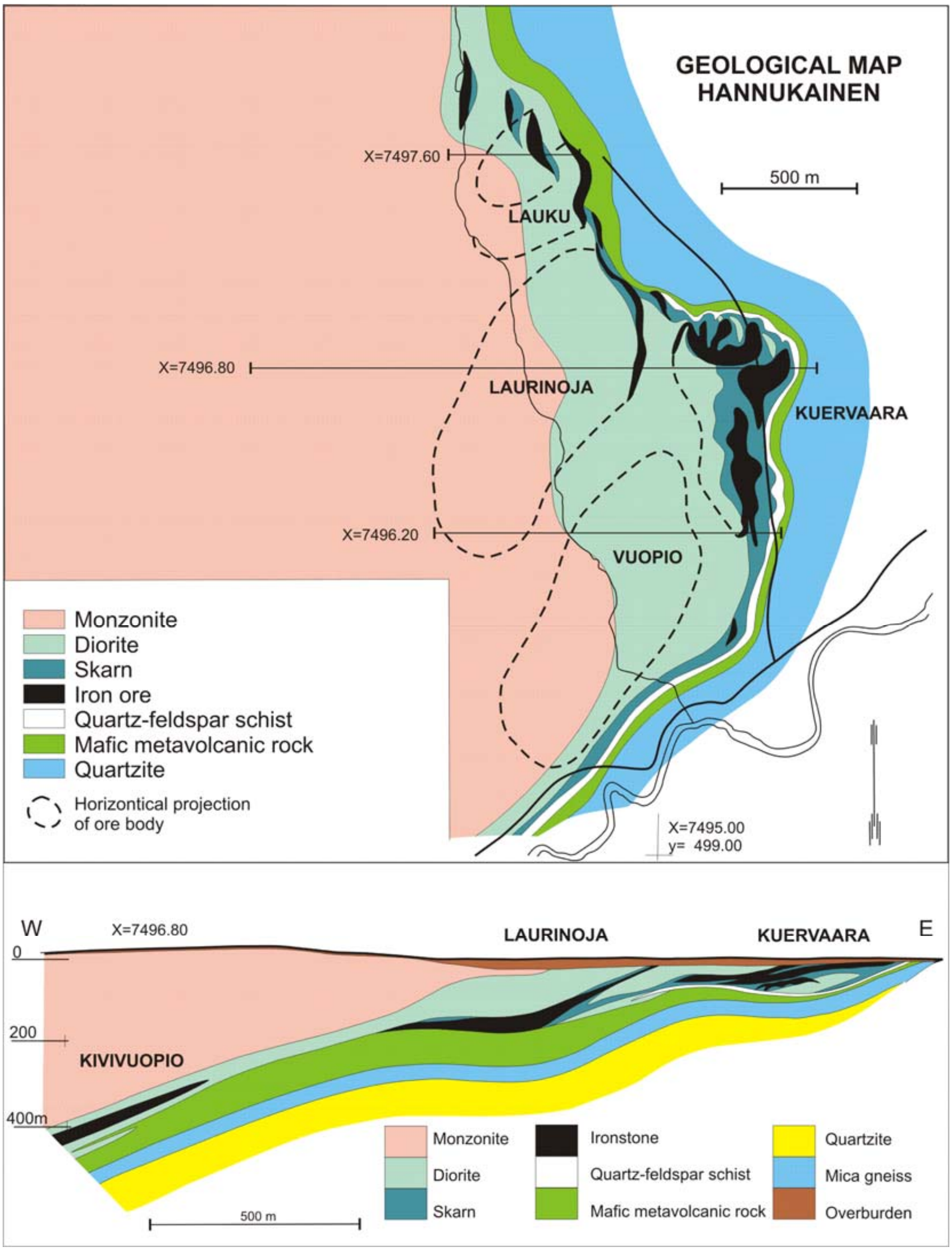
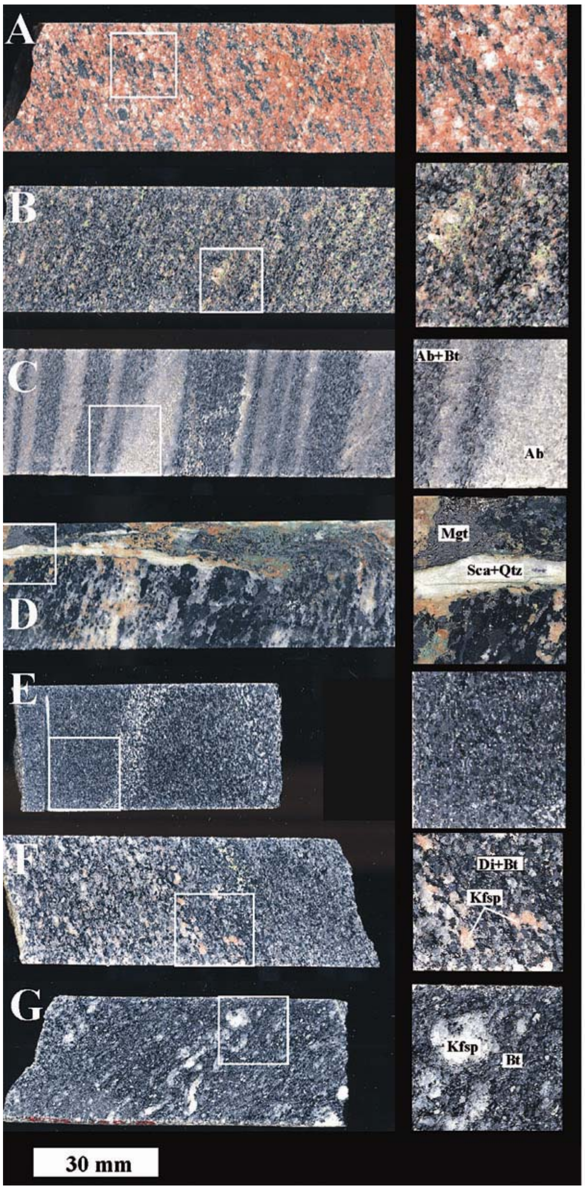


Figure 9. Surface geology and a cross section of the Hannukainen ore field. After Hiltunen (1982), redrawn by P. Kurki.

Hannukainen

Hannukainen deposit mined (open pit) in 1978-1992 when 1.96 Mt iron, 40,000 t copper and 4300 kg gold was produced. Total measured plus indicated resources are 84.6 million tonnes with an average grade of 34.6% Fe, 0.20% Cu, and 0.093 g/t Au using a cut-off of 15% Fe. An additional 81.6 million tonnes @ 35.7 % Fe + 0.13% Cu + 0.036 g/t Au have been defined in the inferred category (Northland Resources press release August 23, 2007). Five ore bodies comprise the Hannukainen deposit: Lauku, Laurinoja, Vuopio, Kuervaara, and Kivivuopio (Fig. 9) (Hiltunen 1982). Of these ore bodies, only Laurinoja and Kuervaara have been partially mined, during 1978–1990. Typical ore mineral association at Laurinoja is magnetite-chalcopyrite-pyrite±pyrrhotite. In places, the amount of pyrrhotite exceeds that of pyrite. Alteration styles noted at Hannukainen are sodic to calcic in the hanging wall diorite, Ca-Fe in the ore-skarn zone, and potassic to calcic in the footwall. A lithological and geochemical profile across the Laurinoja ore body is shown in Figure 12. The Laurinoja open pit is now flooded and we see ore boulders and host rocks next to the open pit and in waste rock piles.



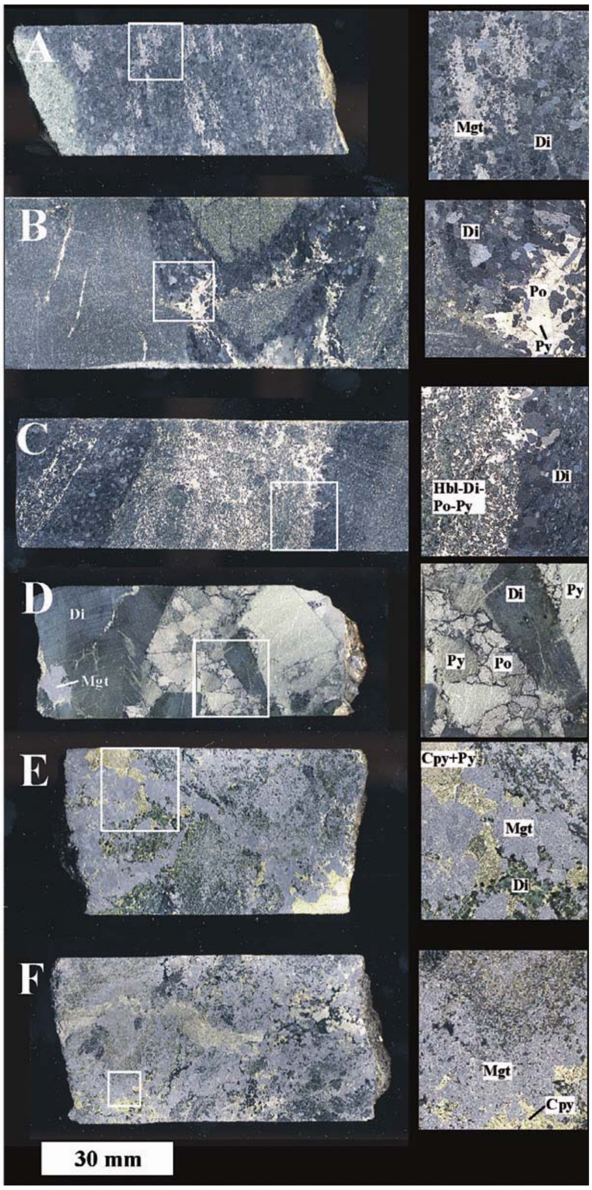


Figure 10. (left) Typical wall rocks of the Kolari ores. All samples from Laurinoja at Hannukainen, except the C from Kuervitikko. White rectangle indicates location of the blowup images on the right. A. Typical monzonite, B. Slightly Na-K altered diorite, C. Albitised diorite, D. Albite-diopside skarn with a late mylonite seam. Altered mafic meta-lava. E. Weakly K-altered mafic metalava. F. K-Ca altered mafic metalava. G. Intensely K-altered mafic metalava.

Figure 11. (right) Typical skarns and ores of the Kolari region. B and C from Kuervitikko, all others from Laurinoja at Hannukainen. Location of the blowup images on the right marked with the white rectangle. A. Diopside-magnetite skarn. B. Diopside skarn band with sulphides in metalava. C. Diopside skarn with pyrite-pyrrhotite dissemination. D. Coarse-grained

diopside-pyrite-pyrrhotite skarn. Locally these rocks contain a significant amount of chalcopyrite. E & F. Laurinoja Fe-Cu-Au ore. Note the difference in grain size for magnetite between non-sulphidic parts and in association with sulphides.

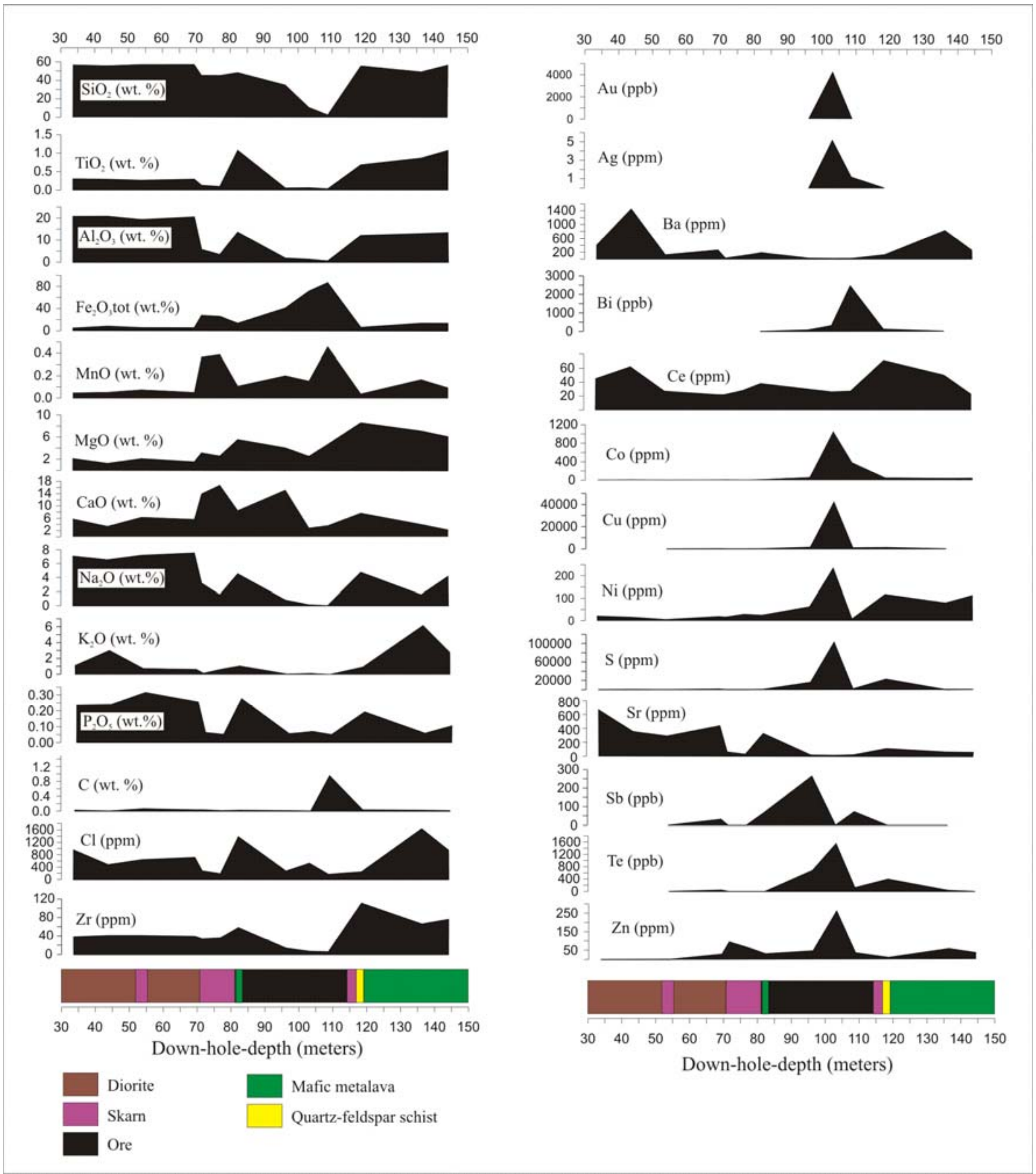


Figure 12. Lithology and geochemistry of the drill hole R75 from Laurinoja. After Niiranen et al. (2007).

Fluids and O-, and C-isotope data

The fluid inclusion data from Laurinoja suggest for a two-staged system (Niiranen et al. 2007). First stage fluids were moderately saline (12–22 wt.% NaCl eq.), Na-Ca ± K, Fe - bearing H_2O fluids. The second stage fluids consisted of highly saline (32–56 wt.% NaCl eq.), Na-Ca ± K -bearing H_2O-CO_2 fluids. The temperature was 450–550°C and 290–370°C for the first and second stage, respectively. The pressure during both stages was 1.5–3.5 kbar. (Niiranen et al. 2007)

Based on the oxygen isotope data on oxides, silicates, and carbonates from Laurinoja, the δ18Ofluid at 500°C was +7.7 – +12.7 ‰ VSMOW. The δ13C of the carbonates ranges from -3.4 – -6.9 ‰ VPDB. (Niiranen 2005, Niiranen et al. 2007)

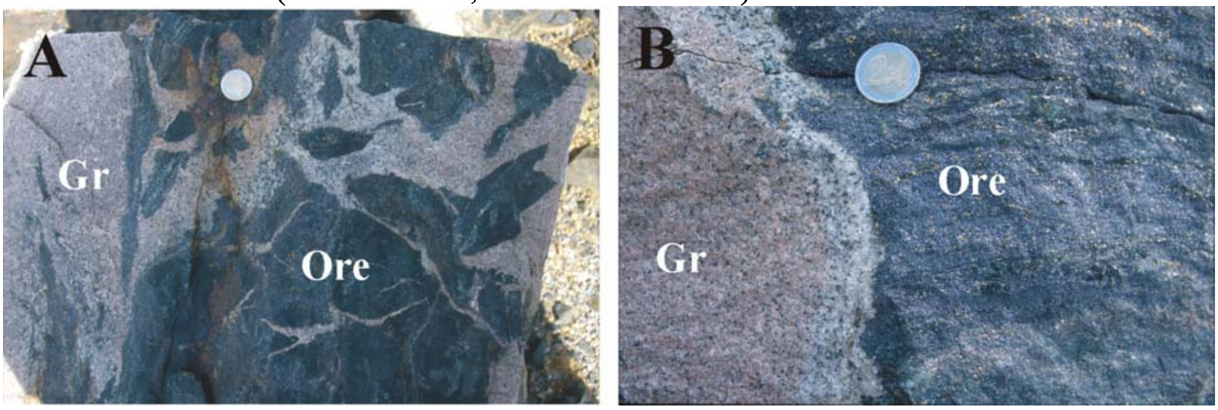


Figure 13. Granite brecciates(dated at 1766 Ma) (A) and crosscuts (B) the ore at Laurinoja. Note the chilled margin in granite (B) and how the granite appears to have no effect on the sulphides in the ore.

Timing constraints

Magmatic zircons from the hanging wall monzonite gave an U-Pb age of 1862±3 Ma, and from diorite 1864±6 Ma (Hiltunen 1982, Niiranen et al. 2007). Since ore-hosting skarns overprint partly the hanging wall diorite at Laurinoja, the latter age can be considered to be the upper age limit for the mineralisation. The magmatic age for the granite brecciating and crosscutting the ore (Figs. 13 and 14) at Hannukainen, 1766±7 Ma (U-Pb zircon) gives the minimum age for mineralisation. The 1797±5 Ma U-Pb age of zircon in skarn, and 1810–1780 Ma ages of metamorphic titanite in altered wallrocks and skarn, suggest that the mineralisation took place around 1800 Ma (Hiltunen 1982, Niiranen 2005, Niiranen et al. 2007). This age and the fact that dominantly only the D4 brittle deformation has affected the ore, indicate that the mineralisation is a late- to post-D3 event within the Pajala-Kolari Shear Zone, unrelated to Haparanda suite intrusions.

Stop 1 Hannukainen iron oxide copper gold deposit

Ore and alteration zone in handspeciamens, boulders and drill core. Open cut visit depend on the access.

Stop 2 Limestone quarry, Äkäsjokisuu

Shallowly dipping, folded and sheared calcitic marble. This stop is to demonstrate the regional and structural geology in the area where natural outcrops are rare.

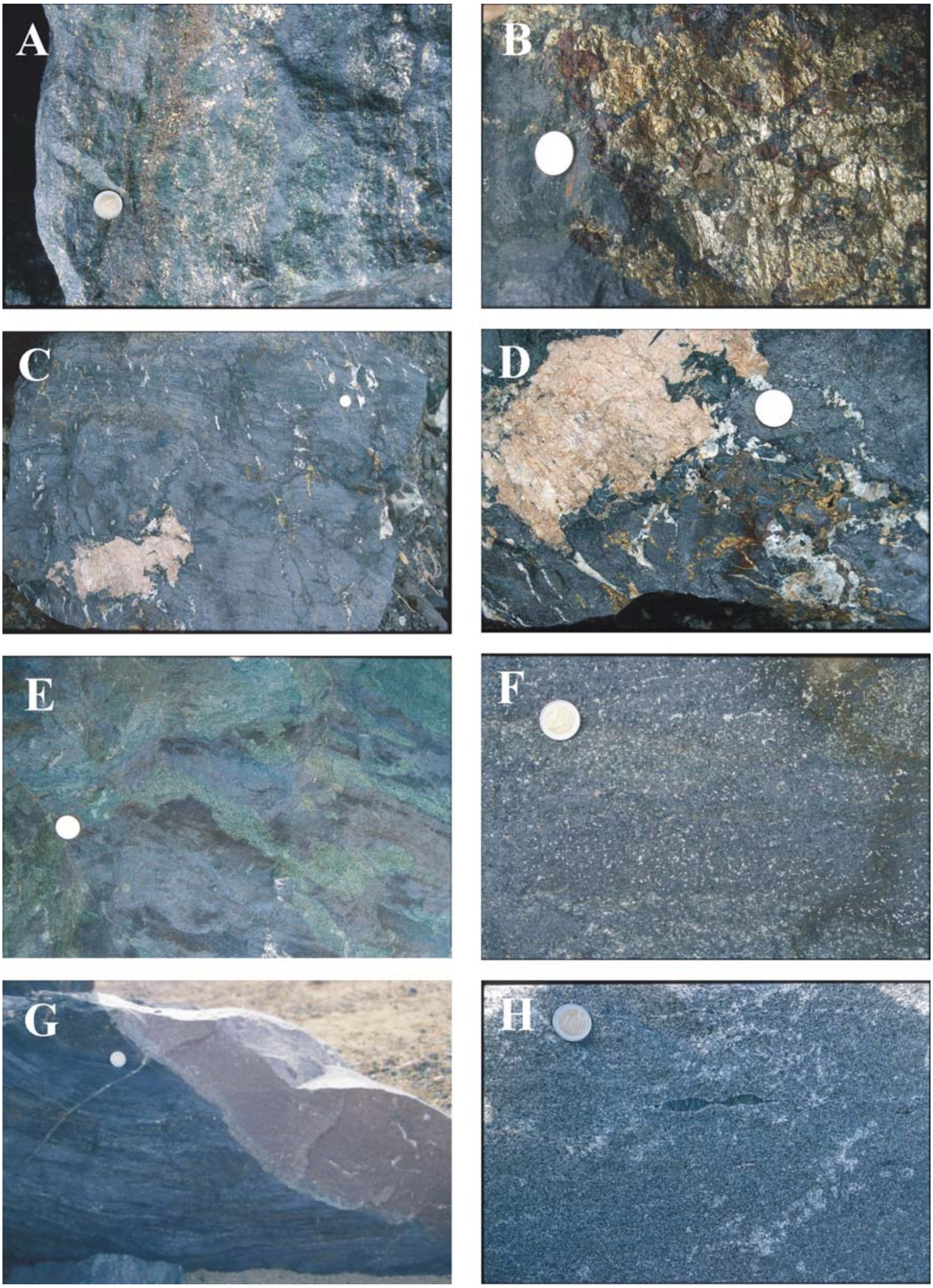


Figure 14. Selected boulders expected to be seen during the excursion at Hannukainen. A. Fe-Cu-Au ore. B. Coarse-grained pyrite-pyrrhotite-chalcopyrite in magnetite ore. Compare to Figure 4. C-D. Brittle fractures in ironstone, chalcopyrite-pyrite-scapolite-quartz infill. Relationship with felsic pegmatite? Second generation of sulphides? E. Sulphide-free iron ore with diopside gangue. F. Coarse-grained ore with diopside-albite-scapolite gangue. G. 1766 Ma granite crosscutting the Fe-Cu-Au ore. This is the very boulder from which the granite was dated! F. Hanging wall diorite. Note the dark mafic enclave(?).

Day 4: Regional geology of Norrbotten, Sweden, skarn iron ores and the Kiirunavaara apatite Fe-deposit

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Regional Geology

The bedrock of the Pajala-Masugnsbyn area shows examples of two main supracrustal units and several suites of intrusive rocks (Fig. 1). The Veikkavaara greenstones (Ludikovian, 2.06-1.96 Ga) are dominated by tholeiitic tuffites intruded by mafic sills. Graphitic schist and silicate-oxide facies BIF are common at high stratigraphic levels, and the uppermost unit is a dolomite (Stop 3c: see description for excursion stops below). The greenstones contain skarn iron ore deposits that occur as stratigraphic equivalents to BIF (Stop 3a, Magnetgruvan) or the top dolomite (Stop 1, Stora Sahavaara).

The Veikkavaara greenstones are overlain by andalusite-bearing mica schist and quartzite of the Svecofennian Pahakurkio group (Stop 2). These rocks were deposited in a shallow marine environment and record two consecutive cycles of deepening and shallowing (Kumpulainen 2000). The Pahakurkio group is associated with intermediate metavolcanic rocks of the Porphyrite group (c. 1.88 Ga, e.g. Edfelt et al. 2006). Towards the south the metamorphic grade increases and the aluminium silicate is sillimanite. Metamorphic monazite from the mica schist has a U-Pb age of 1.86-1.85 Ga (Bergman et al. 2006), which overlaps with the intrusive age of the Masugnsbyn granite (Stop 3b). This granite has been assigned to the Perthite Monzonite Suite, which has its main outcrop area in western Norrbotten.

In the Kiruna area sediments (Kurravaara Conglomerate), equivalent to the Pahakurkio Group are overlain by the Kiirunavaara Group which is the host rock to iron ores of the Kiruna Type. These apatite-magnetite-hematite ores includes the world class Kiirunavaara iron deposit.

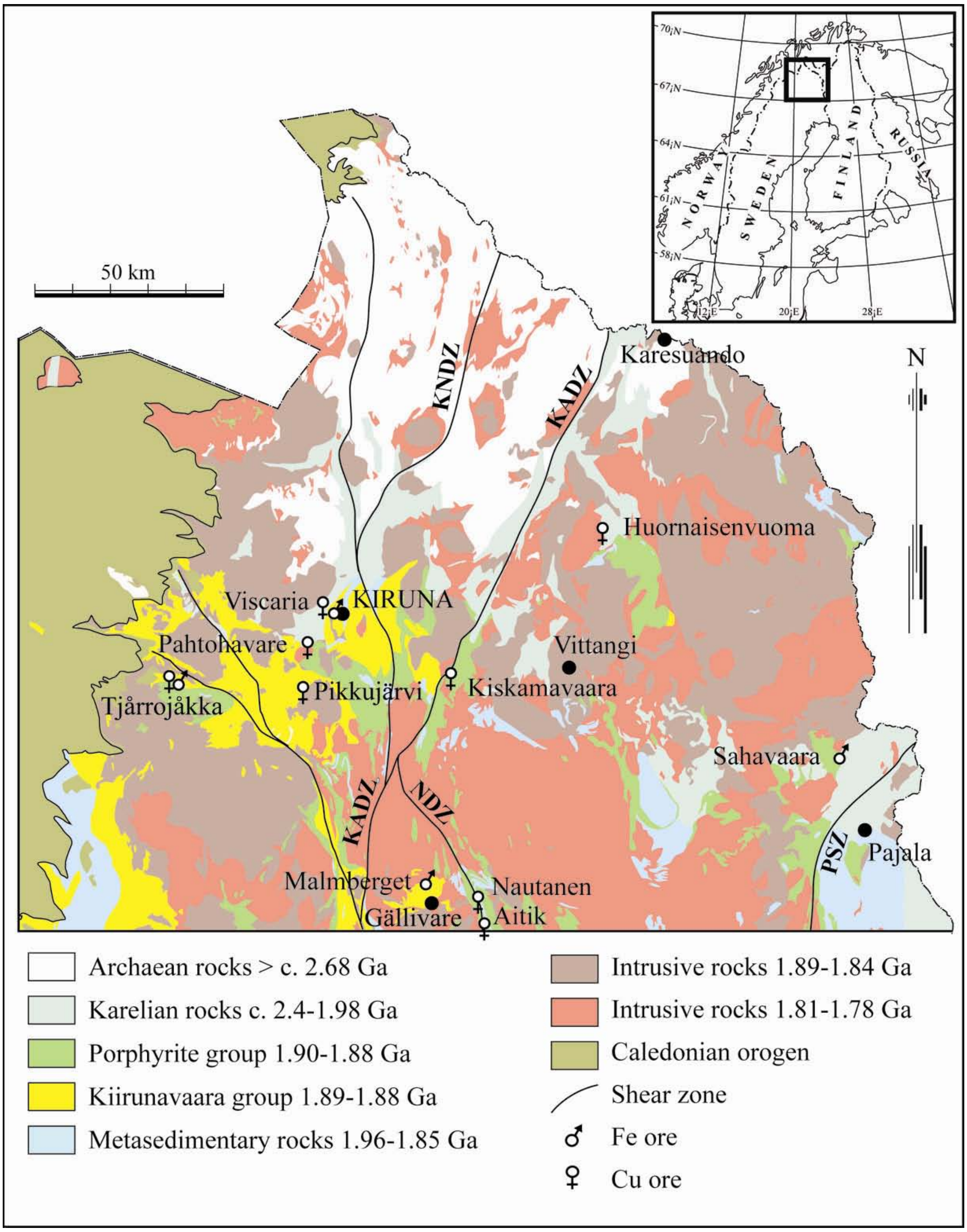

Figure 1. Geology of Northern Norrbotten with selected mineral deposits; modified from Bergman et al. (2001).

Stop 1 Stora Sahavaara

Several iron occurrences have been detected northeast of Pajala (Fig. 2). The largest one is the Sahavaara deposit, which comprises three lenses of skarn-rich iron formation. They were investigated by the Geological Survey of Sweden in the 1960's by geophysical measurements and drilling. The largest one, Stora Sahavaara, was then estimated to contain 82 Mt with 41 % Fe, 2.5 % S, 0.07 % P and 0.08 % Cu. Södra Sahavaara and Östra Sahavaara, situated in a stratigraphically slightly lower position, comprise 19.6 and 2 Mt of ore, respectively (Grip & Frietsch 1973). Recent investigations, with drilling, by Northland Resources have increased the resources (measured, indicated and inferred) at Stora Sahavaara to 145 Mt with 43.1 % Fe and 0.076 % Cu.

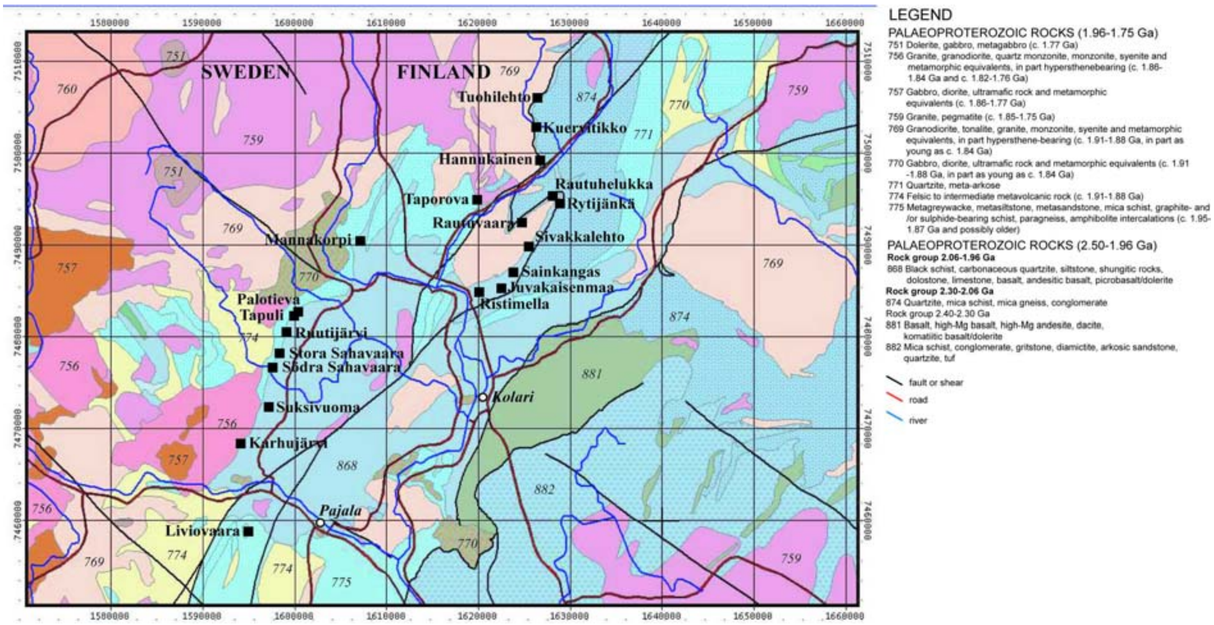

Figure 2. IOCG-style deposits in the Pajala-Kolari area at Sweden-Finland border on a geological map (Excerpt from the geological map of the Fennoscandian Shield, Koistinen et al. 2002)

The Sahavaara deposit is in the upper part of the greenstone pile at the contact between volcaniclastic greenstones in the footwall and clastic sediments in the hanging wall (Figs. 2 and 3). The footwall tuffites, which are deposited upon lapilli tuff of picritic to high-Mg basalt composition. However, close to the ore, the tuffites becomes more felsic and they are generally rich in graphite and scapolite. The hanging wall is dominated by impure quartzite with some intercalations of andesitic volcaniclastic rocks (Martinsson 1995).

The ore zone at Stora Sahavaara is up to 80 m thick and it consists of serpentine-rich magnetite ore including lenses and layers of serpentine-diopside-tremolite skarn (Lundberg 1967). Pyrrhotite and pyrite occur disseminated in the ore together with minor chalcopyrite. A zoned skarn unit usually caps the ore. Close to the ore it is dominated by serpentine, but it gradually changes to diopside-amphibole and at the top of the ore zone it ends up in a calc-silicate bearing chert. The highest contents of Fe and S are in the central part of the deposit, and this area is also slightly enriched in Cu and Co.

An alteration zone extends to about 200 m below Stora Sahavaara and it also encloses the smaller Östra Sahavaara horizon. It is variably rich in biotite, resulting in enrichment in Mg, K, Ba and Rb, whereas Na, Ca and Sr are slightly depleted. However, locally Na is strongly enriched in felsic albite rocks. The occurrence of scapolite in the footwall of the Stora Sahavaara deposit may be a feature unrelated to the ore as similar scapolite-rich rocks are present at the same stratigraphic position elsewhere in the region (Martinsson 1995).

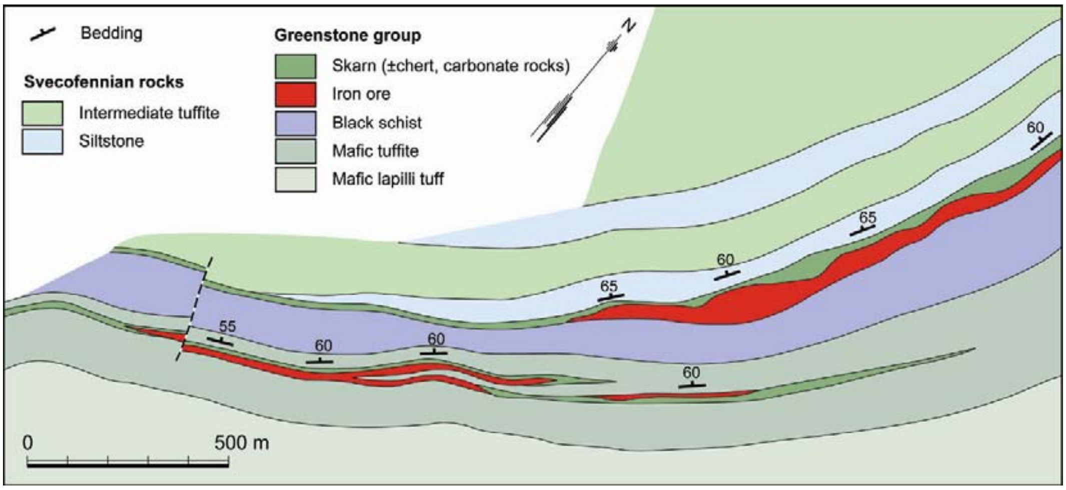

Figure 3. Geology of the Sahavaara deposit (modified from Lundberg 1967).

Stop 2 Pahakurkio

Outcrops of Svecofennian metasedimentary rocks of the Pahakurkio group along River Kalixälven:
a. Gently dipping, andalusite-porphyroblastic mica schist. Minor folds are consistently north vergent. (RT90 7484800/1770220)
b. Quartzite with layers rich in heavy minerals, cross bedding and ripple marks. (RT90 7484800/1769100).

Stop 3 Masugnsbyn

Masugnsbyn skarn iron ore with granite intruding ore. Skarn-rich iron formation and dolomite in the upper part of the Veikkavaara greenstones.
a. Skarn iron ore at Magnetgruvan, discovered in 1642, mine production until the early 19th century. Later investigations have shown a total tonnage of 28 Mt ore containing 32 % Fe. Magnetite is affiliated with amphibole-pyroxene-serpentine skarn and some Fe-sulphide. Locally relatively large amounts of chondrodite. Uranium-mineralised fractures have been discovered with an Pb-Pb age of 1845 Ma for uraninite (Welin & Blomqvist 1966). (RT90 7497480/1767130).

b. Red, fine to medium grained granite belonging to the Perthite Monzonite Suite. The U-Pb zircon age is 1858±9 Ma (Skiöld & Öhlander 1989), and $\varepsilon_{Nd}(1.87) = -2.5$. The Rb-Sr age is 1535±30 Ma, MSWD = 5.9, I.R. = 0.713±0.004 (Gulson 1972). (RT90 7497450/1768400)
c. Dolomite quarry run by Norrbottens Järnverk AB from 1952 to 1972, now run by LKAB. Total production is about 3 Mt. The dolomite is used as additive in pellets. The SiO_2 content is as low as 1.5 %, which is essential for industrial purposes. Olivine, amphibole, chlorite, pyrite and calcite exist in low amounts. The normally 100–200 m thick dolomite is the uppermost unit in the Veikkavaara greenstones, and occurs between the greenstones and the Svecofennian supracrustal rocks. At the quarry, the dolomite is thickened, and the thickness exceeds 300 metres. (RT90 7497400/1767235)

Kiirunavaara

Kiirunavaara is the largest of the apatite iron ores in Sweden, comprising about 2000 Mt of iron ore with 60 to 68 % Fe. It was found in outcrop in 1696, but regular mining started not until 1900 when a railway was built from the coast to Kiruna. Open pit mining ceased in 1962, with a total production of 209 Mt. Underground work started in a small scale during the 1950's and the ore is now mined by large-scale sublevel stoping. The present main haulage level is at 1045 m and the production in 2005 was 23.4 Mt with 46.2 % Fe. Combined reserves and resources were 1242 Mt at the end of 2005 (LKAB 2006).
The tabular ore body is roughly 5 km long, up to 100 m thick, and it extends at least 1500 m below the surface (Fig. 4). It follows the contact between a thick pile of trachyandesitic lava (traditionally named syenite porphyry) and overlying pyroclastic rhyodacite (traditionally named quartz-bearing porphyry). Towards north, the much smaller Luossavaara ore is situated in a similar stratigraphic position.

The trachyandesite lava occurs as numerous lava flows which are strongly albite-altered and rich in amygdales close to the flow tops. An U-Pb age of 1876±9 is given for titanite occurring together with actinolite and magnetite in amygdales (Romer et al. 1994). A potassic granite is present at deeper levels in the mine on the footwall side of the ore and several dikes of granophyric to granitic character cut the ore. Some of these dikes are composite in character also including diabase. An U-Pb zircon age of 1880±3 Ma (Cliff et al., 1990) has been obtained for the granophyric dikes and gives the minimum age of the ore. This is consistent with an U-Pb titanite age of 1888±6 Ma for magnetite-titanite veins in the footwall to the Luossavaara deposit (Romer et al. 1994).

The phosphorus content of the ore exhibits a pronounced bimodal distribution with either less than 0.05 % P (B ore) or more than 1.0 % P (D ore). The B-ore may contains up to 15 % disseminated actinolite in a 5 to 20 m wide zone along its borders where it is in contact with the wall rocks. The magnetite is mostly very fine-grained (<0.3 mm), but in the central part of larger B ore lenses a zone of coarser magnetite (up to 2 mm) exists together with some calcite and small amounts of pyrite. The D ore locally has a banded structure and the proportions of apatite and magnetite is widely varied. The age relation between B and D ores is ambiguous, and both ore types can be seen cutting each other. Columnar and dendritic magnetite are locally developed in the ore suggesting a rapid crystallisation in a supercooled magma (Geijer 1910, Nyström 1985, Nyström & Henriquez 1989, 1994). Veins of anhydrite, anhydrite-pyrite-magnetite and coarse-grained pyrite are occur in the ore and its wall rocks.

Magnetite-actinolite veining (ore breccia) is developed both in the footwall and hanging wall along the contacts of the Kiirunavaara ore body. Close to the hanging wall contact, the ore typically is rich in angular to subrounded clasts of rhyodacitic tuff. Actinolite is a common alteration mineral both at the footwall and the hanging wall contacts and it may form massive skarn bordering the ore. Actinolite also replaces, partly or completely, clasts of wallrocks in the ore and in the ore breccia. Besides actinolite and magnetite veining close to the ore, the hanging wall is in some areas affected by biotite-chlorite alteration, which commonly is accompanied by disseminated pyrite and a weak enrichment of Cu, Co and Mo.

Stop 1 Kiirunavaara Fe deposit

Under the underground visit in the northern par of the ore (Sjömalmen), we'll see the ore and its host rocks. The exact localities to be visited depend on the accessibility to different parts of the mine which changes rapidly as a consequence of the mining activities.

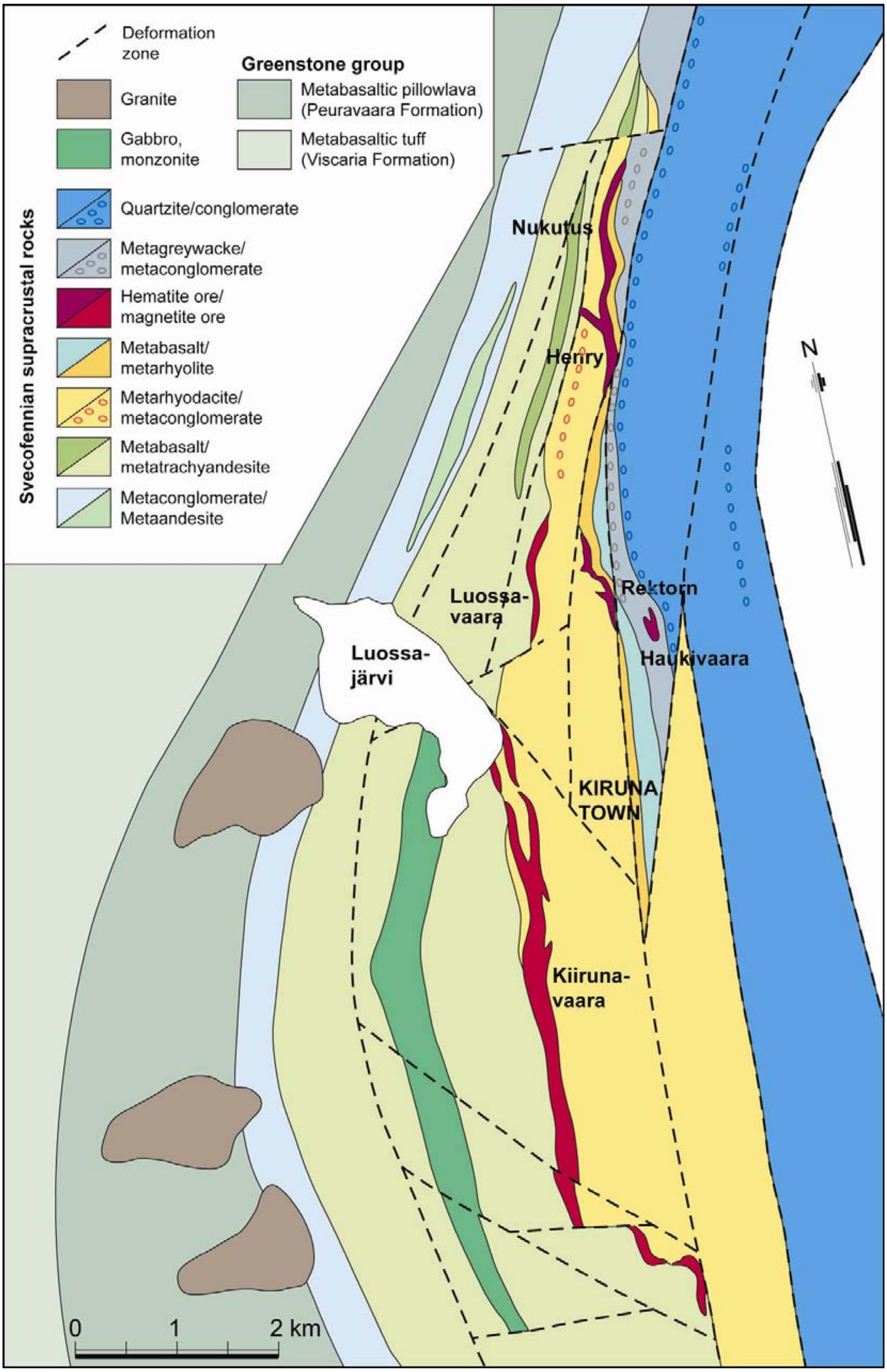

Figure 4. Geology of the Kiruna area with location of the Kiirunavaara deposit (modified from Bergman et al. 2001).

Day 5: Gruvberget and Aitik deposits.

Apatite iron ore and old Cu-mines at Gruvberget

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Introduction

Gruvberget, located close to Svappavaara (Fig. 4 in Introduction), is the largest of the old Cu-mines in Norrbotten. It was found 1654 and during the period 1657–1684 about 1000 ton Cu was produced (Tegengren 1924). The Cu mines occur close to the Gruvberget apatite iron ore, which is 1300 m long and up to 65 m thick. It is calculated to contain 64.1 Mt with 56.9 % Fe and 0.87 % P to a depth of about 300 m where the ore still has the approximately same thickness as at the surface. The bedrock consists of intensely scapolite- and K feldspar-altered intermediate to mafic volcanic rocks. Several dikes of metadiabase with a NE direction cut the ore and its wall rocks (Frietsch 1966).

The apatite iron ore is mostly massive, consisting of magnetite in the northern part and hematite in the middle and southern part of the deposit. Apatite, calcite, actinolite and garnet are gangue minerals occurring in small amounts. In the northern part, the ore is bordered by a narrow zone of garnet, amphibole and epidote towards the hanging wall. Veins and schlieren of magnetite, hematite, apatite and amphibole form an extensive ore breccia in the footwall at the middle part of the deposit. The richer part of the breccia is calculated to contain 9.7 Mt with 40.9 % Fe and 0.88 % P.

Copper sulphides are scattered through the Gruvberget area, with zones of richer mineralisation mainly developed in the footwall to the iron ore. Chalcopyrite and, in lesser volumes, bornite are the main ore minerals, occurring disseminated together with magnetite in altered rocks, or as rich ore shoots at the contact to the iron ore. Molybdenite occurs locally in small amounts. Druses with epidote, magnetite, pyrite, Cu sulphides and desmine are common within the sulphide mineralisations. Several of the mines are close to metadiabases, and the Cu mineralisation seem to be controlled by the same structures as the dikes. The more competent iron ore probably has acted both as a structural and chemical trap. Cu is the only metal reaching economic grades and the Au content is generally very low. The Cu mineralisation is suggested to be genetically unrelated to the iron ore and of younger age (Lindskog 2001).

The host rocks to the Gruvberget deposit commonly are strongly scapolite altered. K-feldspar alteration is extensively developed east of the iron ore, resulting in a high K_2O content (up to 9.8 %). This area is also affected by sericite alteration in narrow schistose zones. However, in association with bornite mineralisation, intense K-feldspar alteration is locally developed west of the iron ore replacing the earlier scapolite. An U-Pb titanite age of c. 1.8 Ga is given for the alteration and Cu mineralisation (Billström & Martinsson 2000).

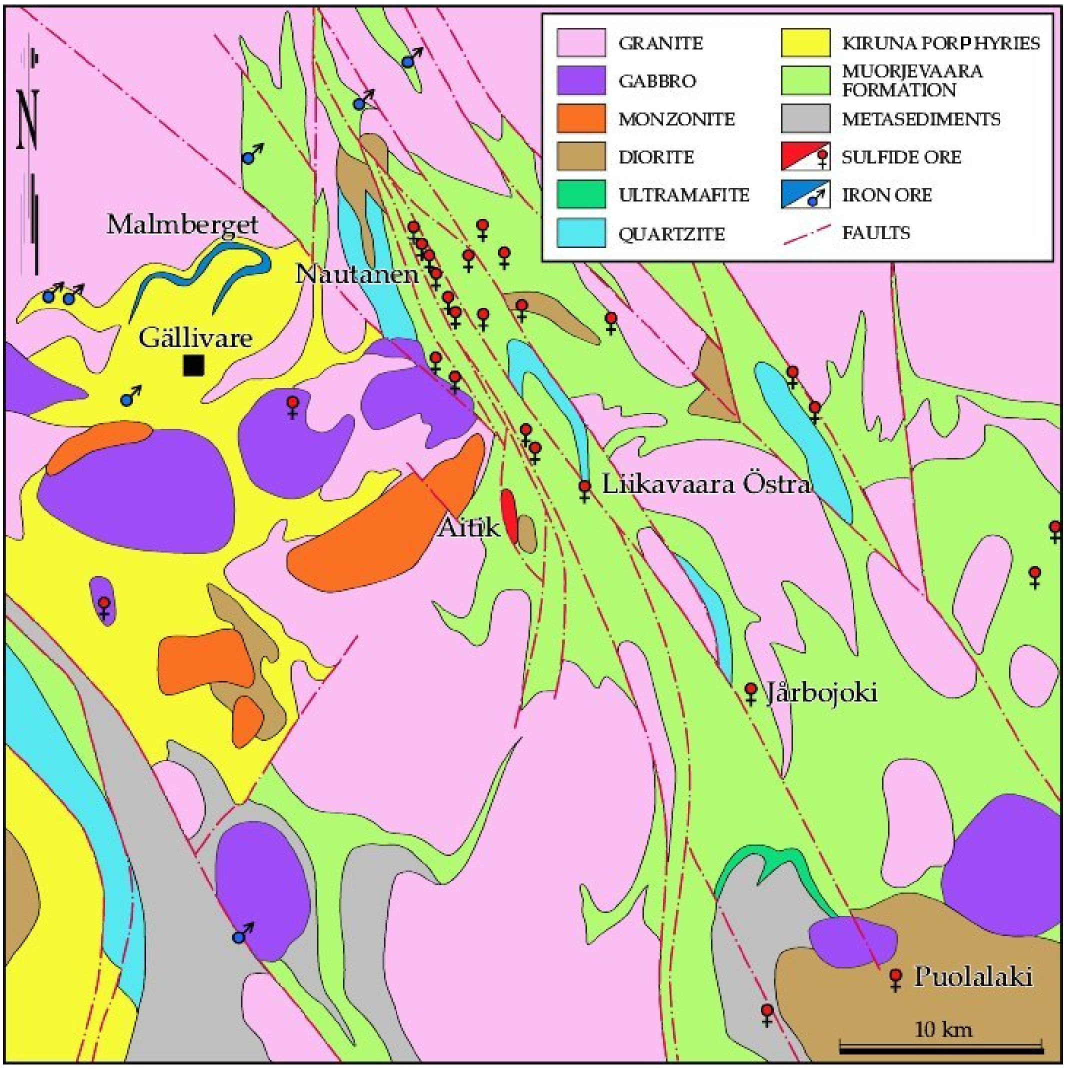

Figure 1. Geology of the Gällivare-Aitik area; from Martinsson and Wanhainen (2000).

Stop 1. Historic Gruvberget mines

The excavated northern part of the iron ore with old Cu mines from the 17th century is visited. When walking towards west across the iron ore, a strongly altered metadiabase is seen to cut the ore in a northeastern direction. The fine-grained mafic rock has sharp contacts to the ore and consists mainly of biotite and scapolite. At the western contact of the iron ore, two types of Cu mineralisation is seen. A small pod of chalcopyrite occupies the contact and a few meters towards south K-feldspar alteration with minor bornite mineralisation is overprinting the earlier scapolite alteration in the footwall to the iron ore. Small amounts of molybdenite, stilbite and chabazite accompany the mineralisation. A 100 meters further southwards, a strongly scapolite altered inclusion with pyroxene veins can be seen in the magnetite ore. Walking in a SSW direction, the iron ore becomes dominated by hematite and the open pit of Storgruvan (means “the big mine”) is seen at the contact of the ore and its footwall. It was mined between 1655 and 1673 and has a depth of 91 m.

Aitik Cu-Au-Ag mine

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Introduction

The Aitik Cu-Au-Ag mine is situated in Norrbotten County, northern Sweden, some 100 km north of the Arctic Circle and 17 km east of Gällivare town (Fig. 1). The mine started operating in 1968 at a capacity of 2 Mt of ore annually. Subsequent expansions to 5 Mt (1970–72), 11 Mt (1979–81), have brought the capacity up to 16 Mt (1989–91). The next expansion will be operational in 2010–2011 and will bring the capacity up to 33 Mt of ore in 2010, which will be ramped up to 36 Mt annually.

Mining

The Aitik mine (Figs. 2 and 3) is a conventional large open-pit operation with an in-pit crusher (18.4 Mt of ore mined 2006). The Cu-Au-Ag ore is moved by trucks carrying 240 tonnes of ore to the crushers. The ore is crushed, milled and processed in the flotation plant yielding a chalcopyrite concentrate. The economic product is a Cu-(Au-Ag) concentrate with an average grade of 27–29 % Cu, 8 ppm Au and 250 ppm Ag. The concentrate is transported by truck to Gällivare and then railed 400 km to the Rönnskär Cu smelter east of Skellefteå, where LME (London Metal Exchange) grade Cu cathodes are produced. By-product gold and silver are also extracted at Rönnskär to produce metallic Au and Ag. Sulphur is captured by the smelter and converted into sulphuric acid. In 2006, Aitik produced about 29 % of the required feed of the Rönnskär smelter, or 240,000 tonnes of Cu concentrate. An average year at Aitik would yield some 60,000 tonnes of Cu-in-concentrate, 1.5–2 tonnes of Au, and some 40–50 tonnes of Ag, from 17–18 Mt of ore.

Since the start of mining at Aitik in 1968, approximately 450 Mt of ore have been mined from a 3 km long, 1 km wide and 390 m deep open pit. In addition, some 400 Mt of waste rocks have been removed to expose the ore body. Proven and probable ore reserves at the start of 2007 were 625 Mt with 0.28 % Cu, 0.2 ppm Au and 2 ppm Ag. Additional measured and indicated mineral resources were 858 Mt with 0.24 % Cu, 0.2 ppm Au and 2 ppm Ag, with an additional 66 Mt of inferred resources grading 0.25 % Cu, 0.2 ppm Au and 2 ppm Ag (Boliden AB 2006). This makes Aitik the largest Cu deposit in the Fennoscandian Shield and one of the largest Au-rich porphyry copper deposits in the world. The current mine life, including the expansion to up to 36 Mt/a, will allow the mine to continue to operate until 2026. The final dimensions of the open pit in 2026 will be 5000 m long by 1400 m wide and 600 m deep. Exploration in the area is ongoing.

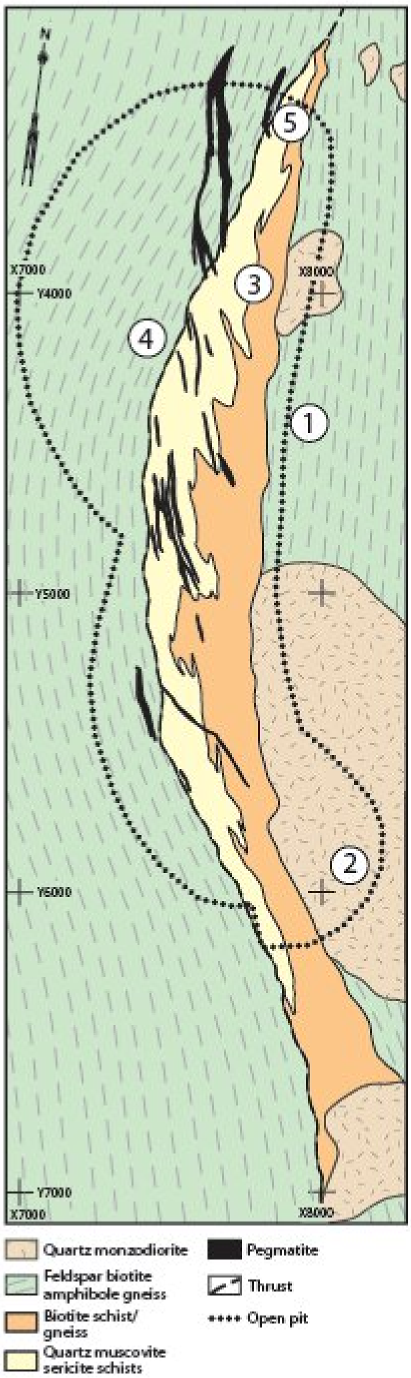

Figure 2. Local geology and excursion stops at the Aitik mine. Geology from Wanhainen and Martinsson (1999).

Mine geology

The local mine geology at Aitik (Figs. 2 and 4) is divided into 3 main parts, i.e. the hanging wall, main ore zone and the footwall complex. The hanging wall is basically one unit of strongly banded hornblende gneisses. The main ore zone consists of three main units, a muscovite schist, biotite schist and biotite gneisses. These rocks are strongly deformed and altered which obscure their primary character. However, their chemical character suggests a magmatic precursor of intermediate composition and, based on the knowledge from areas outside the mine, a volcaniclastic origin (Wanhainen & Martinsson 1999). The most important footwall unit is the quartz monzodioritic intrusive. Other intrusives of interest are the pegmatite dykes which cross cut the hanging wall, main ore zone and the footwall complex.

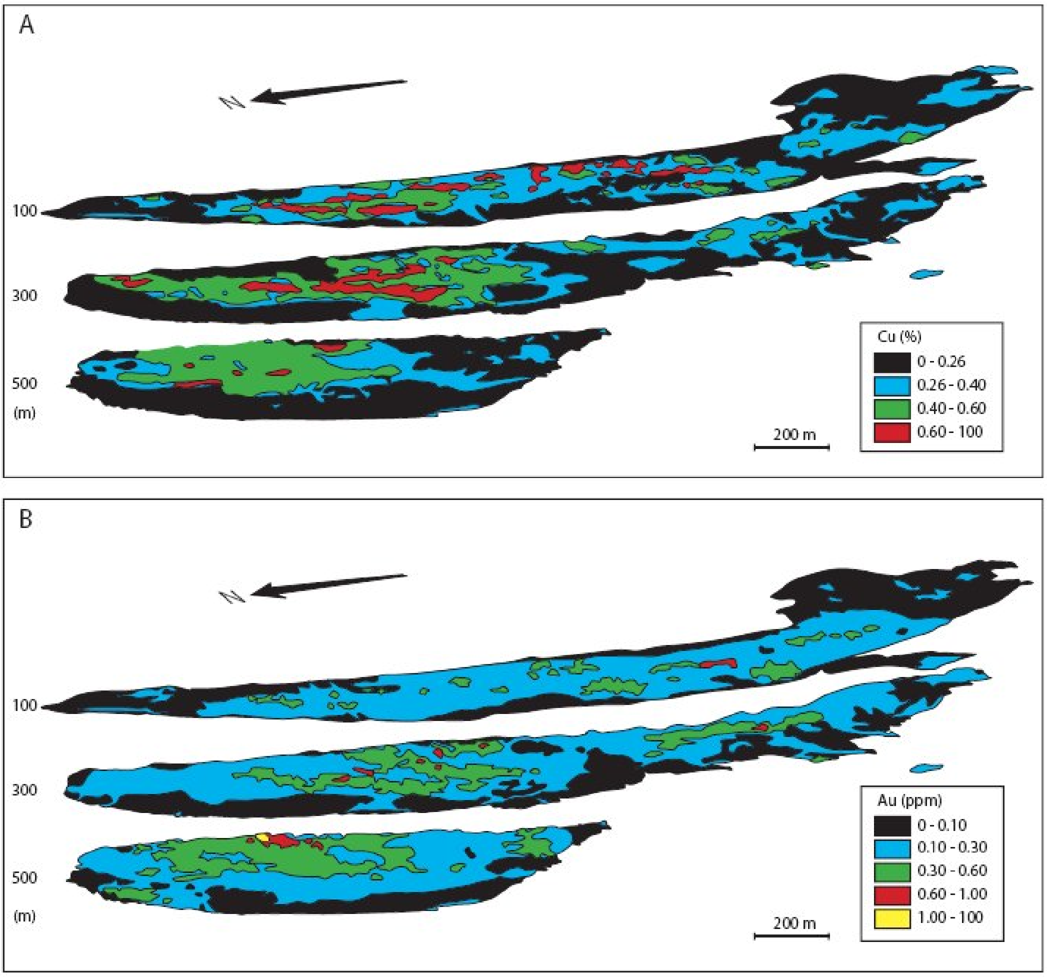

Figure 3. Metal distribution at Aitik for copper (A) and gold (B) for the 100, 300 and 500 m horizontal levels. Class limits are chosen after the classification of mineable to waste rock and low- to high-grade ore used by Boliden AB. From Wanhainen et al. (2003b).

The main ore zone dips roughly 45° to the west (Figs. 3 and 4), and the lower ore contact consists of a gradational weakening of the copper grade at roughly 50° to the west. The lower contact is approximately where biotite gneisses change into regional biotite-amphibole gneiss. Sporadic Cu mineralisation of no economic interest exists in these footwall gneisses. The footwall quartz monzodiorite in the southern part of the mine is part of newly started series of push backs. Below follows a detailed description of the Aitik mine rock units (see also the rock types depicted in Fig. 5):

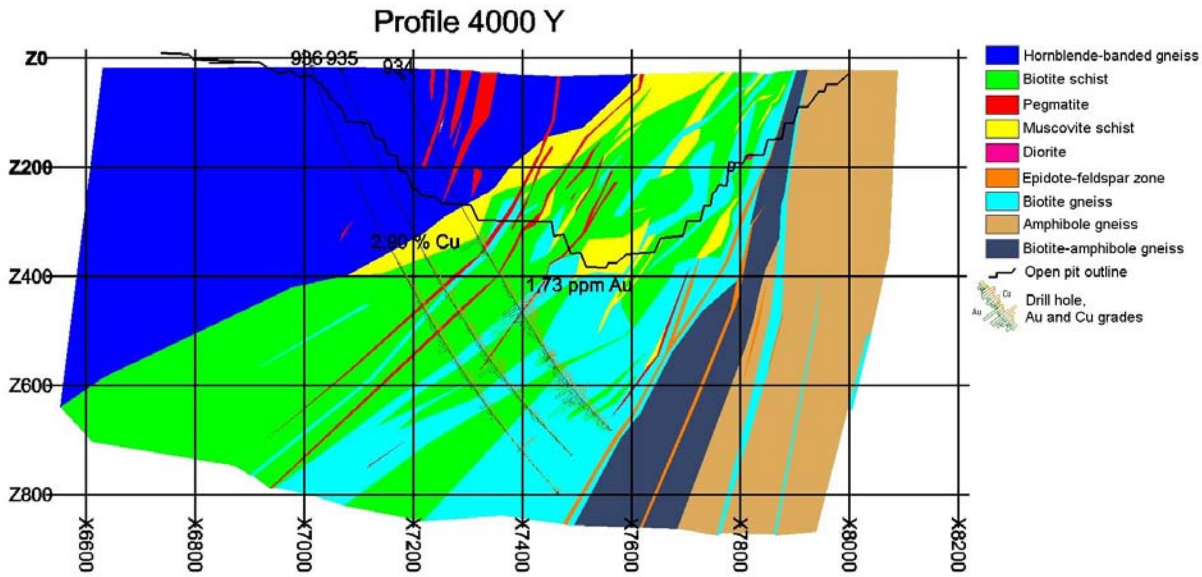

Figure 4. Section across the Aitik deposit, view to the north, 200 m grid.

Hornblende-banded gneiss is a finely banded unit (mm to cm wide bands) with alternating dark olive green and light grey layers. This unit is more than 250 m thick, and overlies the main ore zone. It is devoid of sulphides. Mineralogically it is dominated by hornblende, with biotite, quartz and minor plagioclase. The light grey bands have weak to moderate sericitic and chloritic alteration. The unit also shows a red-green microcline-epidote-alteration. Scapolite porhyroblasts of 1–5 mm in diameter occur throughout the unit. Other accessory minerals are magnetite and tourmaline. The fine-grained unit likely represents original compositional variations, even though it is strongly metamorphosed. Based on field evidence, it is suggested that the fining upwards of the layering shows that way-up is towards the west. The unit appears to have been tectonically emplaced over the main ore zone. The fault at the contact appears to be a thrust. The boundary between the main ore zone and the hornblende-banded gneiss is in places highly fractured, causing problems for drilling. The border zone between hornblende banded gneisses and the main ore zone is also intruded by several pegmatite dykes up to 40 m wide.

Quartz-muscovite (sericite) schist constitutes the upper part of the main ore zone. The unit is roughly 200 m thick, and consist of a strongly foliated muscovite-rich matrix with quartz, biotite, microcline and plagioclase. It is a light-buff coloured unit showing a sharp contact with the overlying hornblende-banded gneiss and a gradational lower contact grading into biotite schist. Accessory minerals in this unit are epidote, tourmaline, magnetite and garnet. Magnetite occurs as occasional mm-scale porphyroblasts, and also as fine dissemination (1–3 % magnetite). The sulphide minerals are dominated by pyrite and chalcopyrite (py > cpy > po). Total sulphur content can reach 5–7 %, corresponding to 15–20 vol-% of sulphides. The muscovite schist has a Cpy:Py ratio ranging from 1:2 to 1:7. The upper contact of the muscovite schist contains a sulphide rich zone, 5–40 m wide with up to 20–25 % sulphides. Gold and copper zonation is shown in Figure 3. Gold and copper grades increase at depth in the northern part of the pit. Pyrrhotite and molybdenite occur as less common sulphides. Pyrite typically occurs as large blebs, or along foliation planes, and as small veinlets. The Ba content of the unit is quite high, in the order of 1,000 – several 1,000s of ppm.

Biotite schist constitutes the middle part of the main ore zone. It is gradational into the biotite gneisses below as well as to the muscovite schists above. The thickness is on average 150 m.

This unit is strongly foliated and sheared in a roughly north-south direction. It contains pyrite and chalcopyrite dissemination and veinlets, and chalcopyrite clots, with pyrite and chalcopyrite as equal volumes. Magnetite occurs as a fine dissemination with grains commonly enclosed within amphibole and/or garnet porphyroblasts. Molybdenite is present in the northern part of the mineralisation. Biotite dominates over muscovite, and defines a strong foliation. Thin veinlets of quartz, commonly deformed, occur in this unit. Undeformed veinlets with late zeolites and epidote occasionally occur within the unit.

Biotite gneisses constitute the lowermost part of the main ore zone, although the rock type is not always present. They commonly display zones of red garnet (spessartine-almandine) and more gneissic, coarser-grained character than the strongly foliated biotite schist. Mineralisation is of the same style as in the biotite schist.

Quartz monzodiorite is the dominant footwall unit, being up to 600 m thick. It shows medium-grained equigranular, 2–5 mm phases as well as strongly porphyritic phases. Transition between these phases (= subphases of the quartz monzodiorite) is almost always gradational. The quartz monzodiorite contains plagioclase phenocrysts being up to 7–9 mm in size. The plagioclase show compositional zoning. The matrix of the quartz monzodiorite consists of a fine-grained mixture of plagioclase, quartz, biotite and minor sericite. Alteration is commonly present as weak silicification and pinkish potassic alteration. Mineralisation is dominated by fracture-controlled py-cpy±MoS_2, but finely disseminated sulphides are also present. A minor accessory mineral is epidote, which can contain fine grained cpy. Hornblende and quartz-tourmaline veinlets occur throughout this unit. Veining of quartz, quartz-tourmaline, gypsum, gypsum-fluorite and zeolites occur as mm–cm wide veinlets. The zeolites present are stilbite, chabazite and desmine, and calcite and baryte have also been observed in this association. These stockwork veins cut each other at high angles, but zones of deformation are also present. The quartz monzodiorite has a zircon U-Pb age of ca. 1.89 Ga (Wanhainen et al. 2006), which fits well with reported ages for regional Haparanda suite granitoids (Bergman et al. 2001).

Feldspar-porphyritic andesitic intrusives occur as large dykes and occasionally show chilled margins. These types of intrusives occur throughout the entire stratigraphic column, but are more common in the footwall area. These dykes are strongly porphyritic in character, with large feldspar phenocryst laths, up to 25 mm long and 4–5 mm wide. They are set in a dark olive green matrix of hornblende, biotite, chlorite, and occasionally actinolite or tremolite. The fine-grained, equigranular variety of this rock is termed amphibolite in the mine. Sulphides, when present, are typically pyrite-chalcopyrite at a 1:1 ratio, and they appear to be both remobilised from the adjacent rocks and to be present within the feldspar porphyritic unit.

Amphibole and amphibole-biotite gneisses constitute a major part of the footwall unit. These rocks typically exhibit an anastomosing weak network of 5–30 mm wide hornblende veinlets or schlieren with a light-coloured feldspar (albite) rim. Biotite defines a weak foliation, and porphyroblastic garnet is commonly present forming 1–5 vol-% of the rock. Sporadic scapolite is present as small grains and as zones of intense scapolitisation. Magnetite is a common accessory (1–3 %), and occurs as small porphyroblasts and as veinlets.

Thin **pegmatite dykes** are common; they may reach a maximum width of 40 m. Their distribution is varied within the mine area with the largest frequency of the dykes in and around the hanging wall contact, where they are unmineralised. At the hanging wall contact, they are oriented roughly N-S and dip about 60º to the west. In the main ore zone, the pegmatite dykes occur less frequently, and one series of the dykes show a NNW orientation and a steep dip. The pegmatites commonly have been intruded forcefully since they can contain large fragments of the adjacent country rock. When they intrude mineralised host rock they can also exhibit py-cpy mineralisation. Mineralogically they are dominated by very-coarse grained microcline, quartz and typically long prismatic black tourmaline. Greenish muscovite flakes also are common. Accessory minerals within the pegmatites are molybdenite and fluorite.

Genetic model

The Aitik host rocks belong to the regionally widespread Haparanda suite of intrusions and Porphyrite group of comagmatic volcanic rocks (Wanhainen & Martinsson 1999, Wanhainen et al. 2006) which were generated during subduction of oceanic crust beneath the Archaean craton around 1.9 Ga, during the Svecokarelian orogeny (Weihed 2003). High-salinity fluids (30–38 eq. wt. % $NaCl+CaCl_2$) responsible for chalcopyrite-pyrite mineralisation in Aitik were released contemporaneously with quartz monzodiorite emplacement and quartz stockwork formation at ca. 1.89 Ga and caused potassic alteration of the intrusive and surrounding volcaniclastic rocks. The mineralised quartz monzodiorite in the footwall is suggested to represent an apophyse from a larger intrusion at depth consistent with the porphyry copper model presented by Lowell and Guilbert (1970). Furthermore, zonation and alteration patterns, although disturbed, fit quite well with this model (Yngström et al. 1986, Monro 1988, Wanhainen 2005).

However, all features of the main ore zone are not typical for a porphyry system, and Aitik is suggested to be hybrid in character with an affinity to both IOCG and porphyry-copper mineralisation based on the character of the high salinity ore fluids, the alteration and mineralisation styles, and on the 160 Ma (Re-Os molybdenite and U-P titanite and zircon dating) evolution of the deposit (Wanhainen et al. 2003a, Wanhainen et al. 2005, Wanhainen et al. 2006), including a regional mineralising event of IOCG-type at ca. 1.8 Ga.

Stop 1 Aitik open cut

The exact localities to be visited in the open cut depend on the accessibility to different parts of the mine which changes rapidly as a consequence of the mining activities.

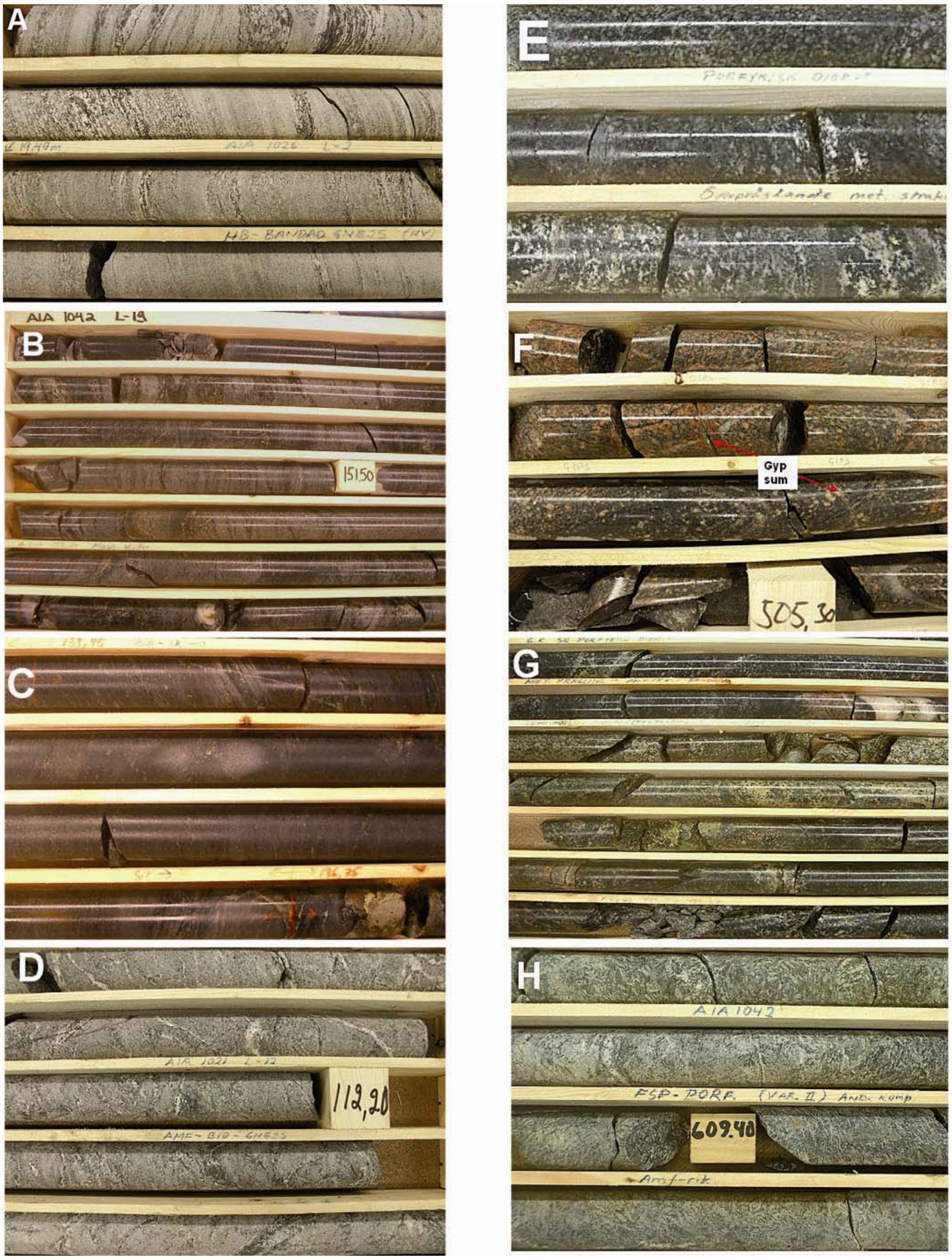

Figure 5. Rock types at Aitik. A. Hornblende-banded gneiss – AIA1026 (HWC – at 19.40 m). Finely banded unit with minor coarser bands. B. Muscovite schist – AIA1042 (MOZ) at 151.50 m. Displaying mixed nature with alternating muscovite and biotite-bands. C. Biotite schist – AIA1042 (MOZ – 133.95 m) Dark grey rock with biotite bands, displaying some muscovite. Garnet-porphyroblasts and dissemination of chalcopyrite, pyrite and pyrrhotite. D. Amphibole-biotite-gneiss – AIA1021 (MOZ – at 112.20 m). Metamorphic hornblende patches and schlieren. E. Diorite – AIA1042 (MOZ/FWC). Coarse-grained and porphyritic. Overprinting metamorphic amphibole alteration and silicification causes diffuse textures. F. Diorite – porphyritic with altered plagioclase phenocrysts (potassic alteration) – AIA1042 (at 505.30 m) Dioritic matrix. Potassic alteration and late gypsum veinlets. G. Mineralised diorite. H. Feldspar porphyritic gabbro – AIA1042 (stratigraphic footwall at 609.40 m). Andesitic matrix. Plagioclase laths (5-15 mm).

Day 6: The Kemi Layered Intrusion and the Kemi chrome mine

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Kemi Layered Intrusion

Introduction

The Kemi Layered Intrusion can be considered the most significant of the Fennoscandian layered intrusions, since it contains the sole mine presently active within these intrusions (Alapieti et al. 1989b). The chromitite deposit is hosted by a layered intrusion extending some 15 km northeast of Kemi, a town on the coast of the Gulf of Bothnia (Figs. 1 and 2). U-Pb zircon data yield an age of 2.44 Ga for the Kemi Intrusion (Patchett et al. 1981), and the whole rock Pb-Pb data define an age of 2.44±0.16 Ga (Manhes et al. 1980).

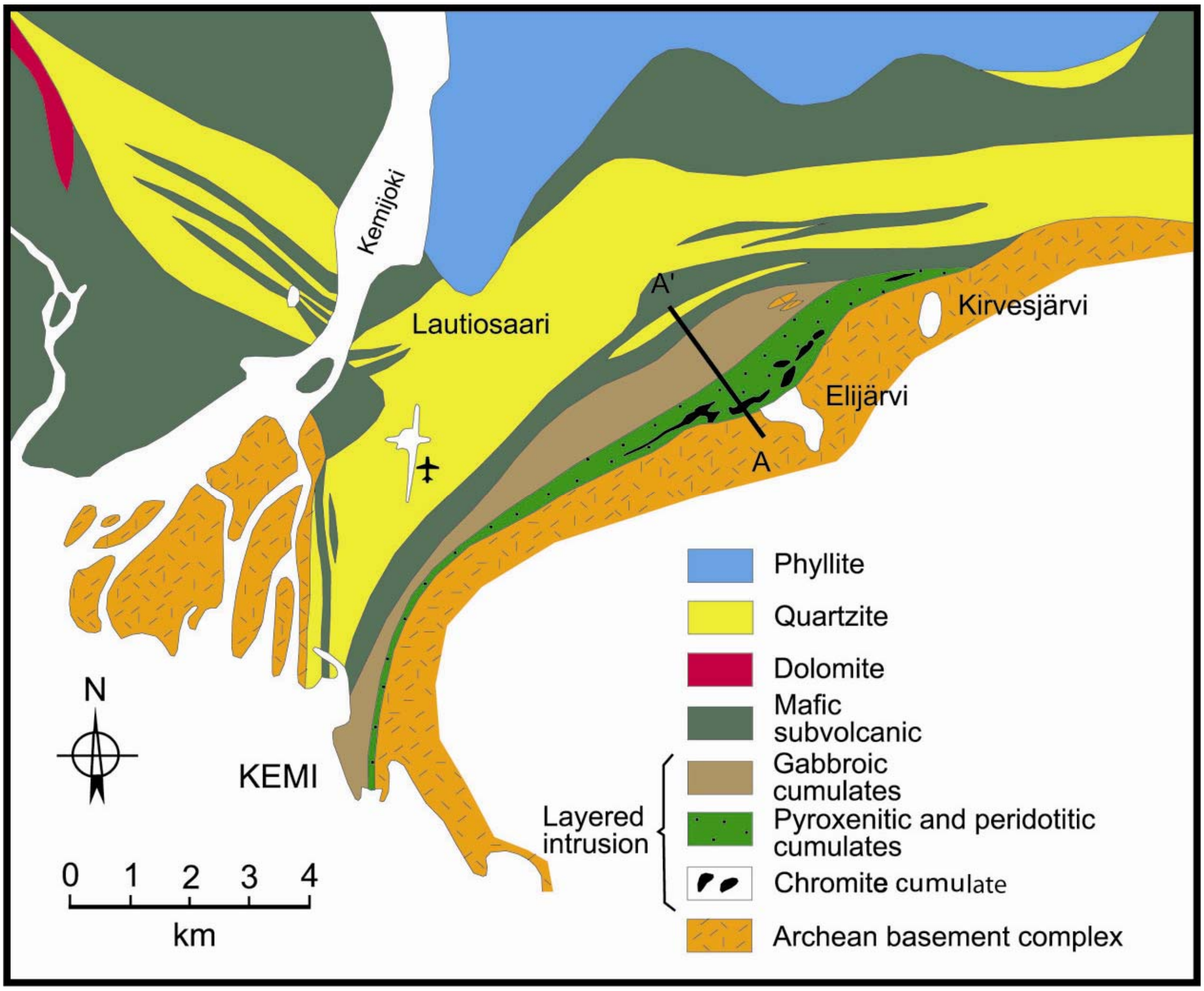

Figure 1. Generalised geological map of the Kemi region. The section depicted in Figure 2 is indicated by the line A–A', after Alapieti et al. (1989b).

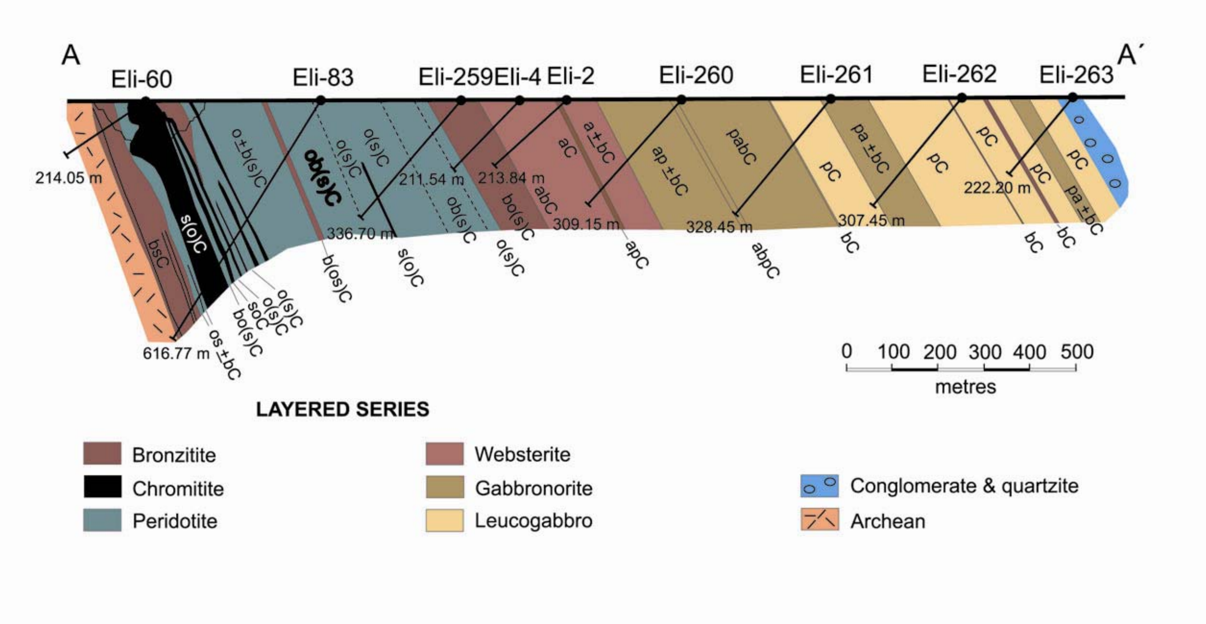

Figure 2. Cross section of the Kemi Intrusion based on the drilled profile A–A' indicated in Figure 1 (Alapieti & Huhtelin 2005).

Layering of the Kemi Intrusion

The present surface section of the Kemi Intrusion is lenticular in shape, being about 15 km long and 0.2–2 km wide (Figs. 1 and 2). It represents a cross-section of an originally funnel-shaped intrusion which was tilted by tectonic movements during the Svecokarelidic orogeny to form a body dipping about 70^o to the NW and, according to geophysical survey, extending down for at least 2 km (Alapieti & Huhtelin 2005). The intrusion comprises an ultramafic lower part and a gabbroic upper part. The individual cumulate layers are thickest in the middle part of the intrusion and become thinner toward the ends. This feature is well established from the variation in thickness of the ultramafic cumulates (Fig. 1). Exploration form underground inclined tunnels has recently proved that the magmatic conduit which fed the magma chamber was located just below the thickening, as already suggested by Alapieti et al. (1989b). This feeder dike comprises fine-grained, uralitised rock types close to the contacts which grade into more coarse grained ones inward, and the middle part of the dyke is composed of a few meter thick chromitite (Alapieti & Huhtelin 2005).

The footwall of the intrusion consists of Neoarchaean granitoids, and the hanging-wall rocks are either younger mafic volcanics or subvolcanic sills 2,150 Ma in age (Sakko 1971), or a polymictic conglomerate of unknown age but younger than the intrusion. This indicates that the present upper contact is erosional, implying that the original roof rocks and the uppermost cumulates of the layered series, together with the possible granophyre layer, have been obliterated by erosion. The feeder dikes of the subvolcanic sills, referred to as albite diabase, intersect the Kemi Intrusion (Alapieti et al. 1989b).

The area of the Kemi Intrusion underwent lower-amphibolite facies metamorphism during the Svecokarelidic orogeny, probably during 1.9–1.8 Ga. The original magmatic silicates have been completely altered to secondary minerals in the lower and upper parts of the intrusion, whereas those in the middle have been preserved and are fresh in appearance. Many chromites have nevertheless preserved their original composition in their cores, even though the silicates of the same rock have been completely altered. The altered rocks have preserved their cumulate textures fairly well and despite alteration, many of the primary minerals can still be recognised by means of pseudomorphs, thus enabling the cumulate sequences to be determined (Alapieti et al. 1989b).

The basal contact series in the Kemi Intrusion is highly altered. It is at present mainly represented by a mylonitic talc-chlorite-carbonate schist, and in places by a talc-carbonate rock, in contact with the basement granitoid. Recently, in the underground mine from level –400 downwards, an approximately 10 m thick metagabbro has also been encountered in some of the new tunnels intersecting the basal contact. This metagabbro does not form a continuous layer, however, its contacts commonly being quite irregular. It can mainly be encountered within the mylonite and the talc-carbonate rock. In a tunnel at level –450, where the basal contact series is quite well preserved, a metapyroxenite is also encountered, overlaying the metagabbro. It seems possible, in accordance with Alapieti et al. (1989b), that the basal contact series may have been partially obliterated by erosion during the magmatic stage, as suggested by the sporadic appearance of the metagabbro.

This sequence is overlain by a markedly altered sequence 50 to 100 m thick, the lower part of which is composed of a bronzite-chromite cumulate and the upper part of an olivine-chromite ± bronzite cumulate, with chromitite interlayers from 0.5 to 1.5 m thick. The bronzitic cumulates typically are characterised by gneissic xenoliths from the underlying basement complex (Alapieti et al. 1989b). In addition to chromitite interlayers, varying amounts of chromite dissemination and roundish lumps of chromitite, whose diameter varies from a few centimetres up to several meters, can be encountered in this sequence.
The sequence described above is followed by the main chromitite unit which in many intersections is composed of two parts with a more silicate-rich rock between them. The total average thickness of the main chromitite unit is 40 m. Its cumulus minerals are chromite and olivine, and the intercumulus minerals comprise poikilitic bronzite and to a lesser extent augite. The abundance of cumulus olivine in relation to chromite is relatively low in the upper part. Bronzite locally occurs as the cumulus phase in the more silicate-rich interlayer which is typified by annular textures constituting accumulations of small chromite grains around the larger cumulus silicate minerals (Alapieti et al. 1989b).

The main chromitite unit is overlain by about 550 m of peridotitic cumulates (Fig. 2) with olivine, chromite and occasional bronzite as the cumulus minerals. This thick cumulate sequence contains about 15 chromite-rich interlayers varying in thickness from 5 cm to 2.5 m, the uppermost being about 370 m above the main chromitite layer (Alapieti et al. 1989b). Three, approximately 10–30 m thick pyroxenitic interlayers occur in the lower part of the peridotitic sequence. The uppermost pyroxenite is situated between sequences of well-developed rhythmic layering patterns approximately 30–50 m thick. The rhythmic pattern is composed of alternating layers of olivine-(chromite) cumulates and olivine-bronzite-(chromite) cumulates, augite being the main intercumulus mineral in both rock types. A similar type of rhythmic sequence occurs in the third megacyclic unit of the Penikat Layered Intrusion, situated some 10 km northeast of the Kemi Intrusion (Huhtelin et al. 1989b). Bronzite becomes the dominant cumulus mineral about 700 m above the basal contact of the intrusion, with olivine and chromite as the other cumulus phases. Then, about 100 m higher up, augite becomes the dominant cumulus mineral alongside bronzite, but olivine and chromite disappear. Even bronzite is so low in abundance in places that the rock could be referred to as a diallagite (Alapieti et al. 1989b).

At about 1,000 m above the basement, plagioclase becomes the cumulus phase alongside augite and bronzite. These plagioclase cumulates continue for about 800 m upward to the contact with the hanging wall. In the upper part of the sequence, augite occurs as the intercumulus phase, and there is little or no Ca-poor pyroxene. In conventional terms, these rocks are therefore leucogabbros or anorthosites (Alapieti et al. 1989b).

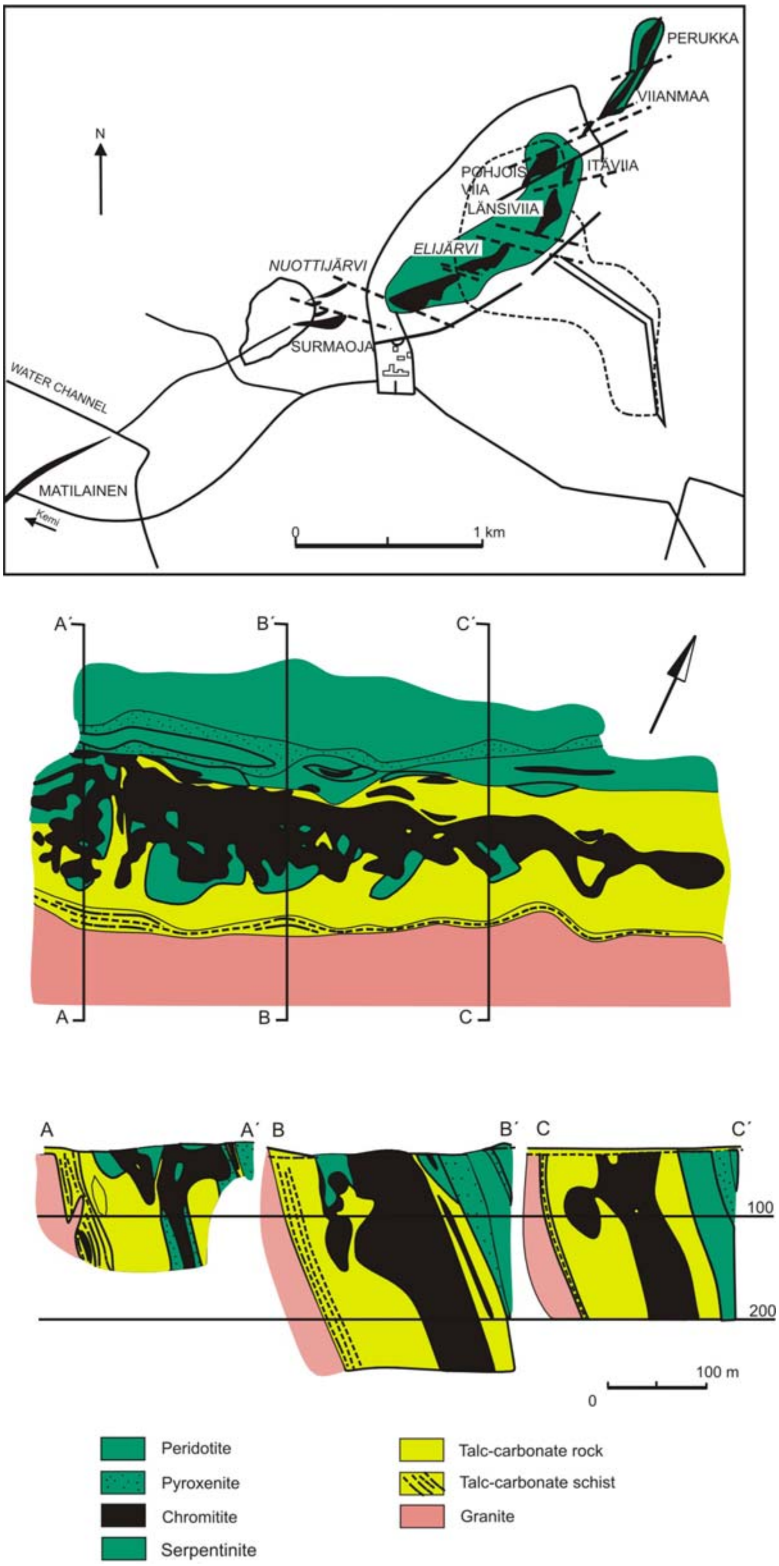

Figure 3. Surface plan and sections across the Elijärvi ore body, after Alapieti et al. (1989b).

The chromite ores

The chromitite layer which parallels the basal contact zone of the Kemi Intrusion is known over the whole length of the complex. In the central part of the intrusion, the basal chromitite layer widens into a thick chromitite accumulation (Fig. 3). The chromite-rich unit has an average dip of 70° to the NW (Alapieti et al. 1989b).

The thickness of the main chromitite unit averages 40 meters, but it varies in thickness from a few meters to over 160 meters. The upper contact of the chromitite unit lies stratigraphically 100 to 150 m above the basal contact of the complex, but its position has been changed by several strike-slip faults (Alapieti & Huhtelin 2005). The top of the main chromitite unit is layered in structure, the hanging wall contact of the ore being sharp, whereas the lower part is non-layered and brecciated, characterised by gradually diminishing chromite dissemination towards the bottom of the intrusion accompanied by irregular ore lumps. The ore boundary at the footwall is, therefore, quite complex. The chromitite unit contains abundant barren ultramafic inclusions, especially in its lower part. Gneissic xenoliths are also locally encountered within the chrome ore.

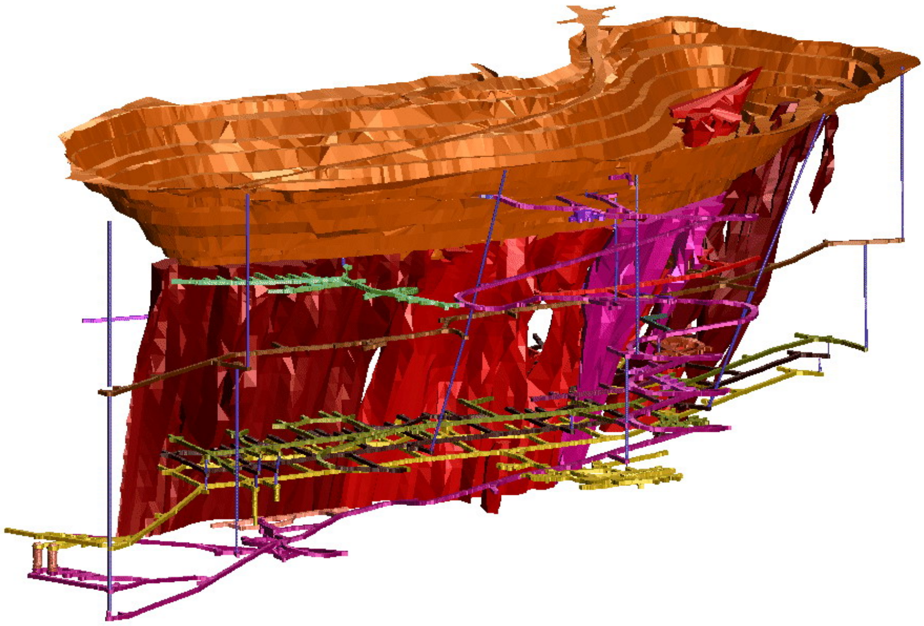

Figure 4. A 3D model of the Kemi mine (Alapieti & Huhtelin 2005).

Kemi Chrome Mine

Mine history and resource

The chromite mineralisation was discovered in 1959, when a fresh-water channel was being excavated in the area. Mr. Martti Matilainen, a local diver who was interested in ore prospecting, discovered the first chromite-bearing blocks. Geological Survey of Finland (GTK) immediately began exploration which led to the discovery of a chromitite-rich layer. From 1960 onwards, the exploration was conducted by Outokumpu Oy under contract from the government of Finland. The decision to exploit the deposit was made in 1964, by which time 30 million metric tons of chromite ore had been located in the basal part of the intrusion (Alapieti et al. 1989b).

The mine's current proven ore reserves total approximately 40 Mt. In addition, it is estimated that there are some 85 Mt of mineral resources (Outokumpu Oyj 2005). The average chromium oxide content of the ore is about 26 percent and its average chrome-iron ratio is 1.6.

Mining and processing

Full-scale open pit mining commenced in 1968, and was carried out until December 2005, when the main open pit had reached a depth of approximately 200 meters. Construction of the underground mine began in 1999, and the underground mine was officially opened in September 2003 (Figs. 3 and 4). Annual production during recent years has amounted to around 1.2 Mt of ore.

Economically, the most important portion of the chromitite layer extends from the Elijärvi ore body in the west to the Pohjois-Viia ore body in the east (Fig. 3) situated below the main open pit. The length of the chromitite unit under mining is about 1.5 km. The chromitite layer is cut into several ore bodies by numerous faults, and these bodies are treated, to a certain degree, as separate units for the purposes of mining, beneficiation, and metallurgy. The reason for this is that the separate ore bodies, due to variation in the degree of metamorphism, host different secondary gangue minerals in the interstices of chromite grains, which has an effect on hardness and specific gravity of the ore. Also, the natural size of chromite grains, which varies greatly, is characterised by extensive microcracking and brokenness in certain ore bodies (Leinonen 1998), thus reducing the size of purely-ground chromite in concentration.

The ore feed is concentrated into upgraded lumpy ore and fine concentrate. In the first stage of the process, the ore is crushed and screened to a diameter of 12–100 millimetres. After crushing, ore lumps are processed by heavy medium separation. In this process, upgraded lumpy ore is separated from the ore. Further processing takes place in the concentrating plant where the ore is first ground in rod and ball mills. The fine concentrate is produced by gravity separation using spirals and Reichert cone separators. In addition, high-gradient magnetic separation is used for fine material unsuitable for gravity concentration. Annual production of lumpy ore and fine concentrate is about 210,000 tonnes and 390,000 tonnes, respectively.

The associated industrial facilities were built close to the harbour of Tornio, a town situated some 25 km from the Kemi mine. Production of ferrochrome commenced in 1968. The ferrochrome works was expanded with the second smelting furnace in 1985. The construction process of the stainless steel works at Tornio was completed in 1976. The unique production chain from chrome ore to stainless steel coils and plates was completed in 1988, when the hot rolling mill was inaugurated. Further investments in the 1990's and during recent years have increased the annual production capacity of Tornio Works to 1.7 Mt of rolled products.

The Kemi chrome mine is a good example of the exploitation of a low-grade ore, distinctly lower in grade than in the stratiform deposits in southern Africa. The success of the operation is due to the convenient location of the deposit relative to existing infrastructure, combined with advanced mineral processing and ferrochrome production technology.

Stop 1. Kemi chrome mine

The exact localities to be visited in the open cut and undergroung depend on the accessibility to different parts of the mine which changes rapidly as a consequence of the mining activities.

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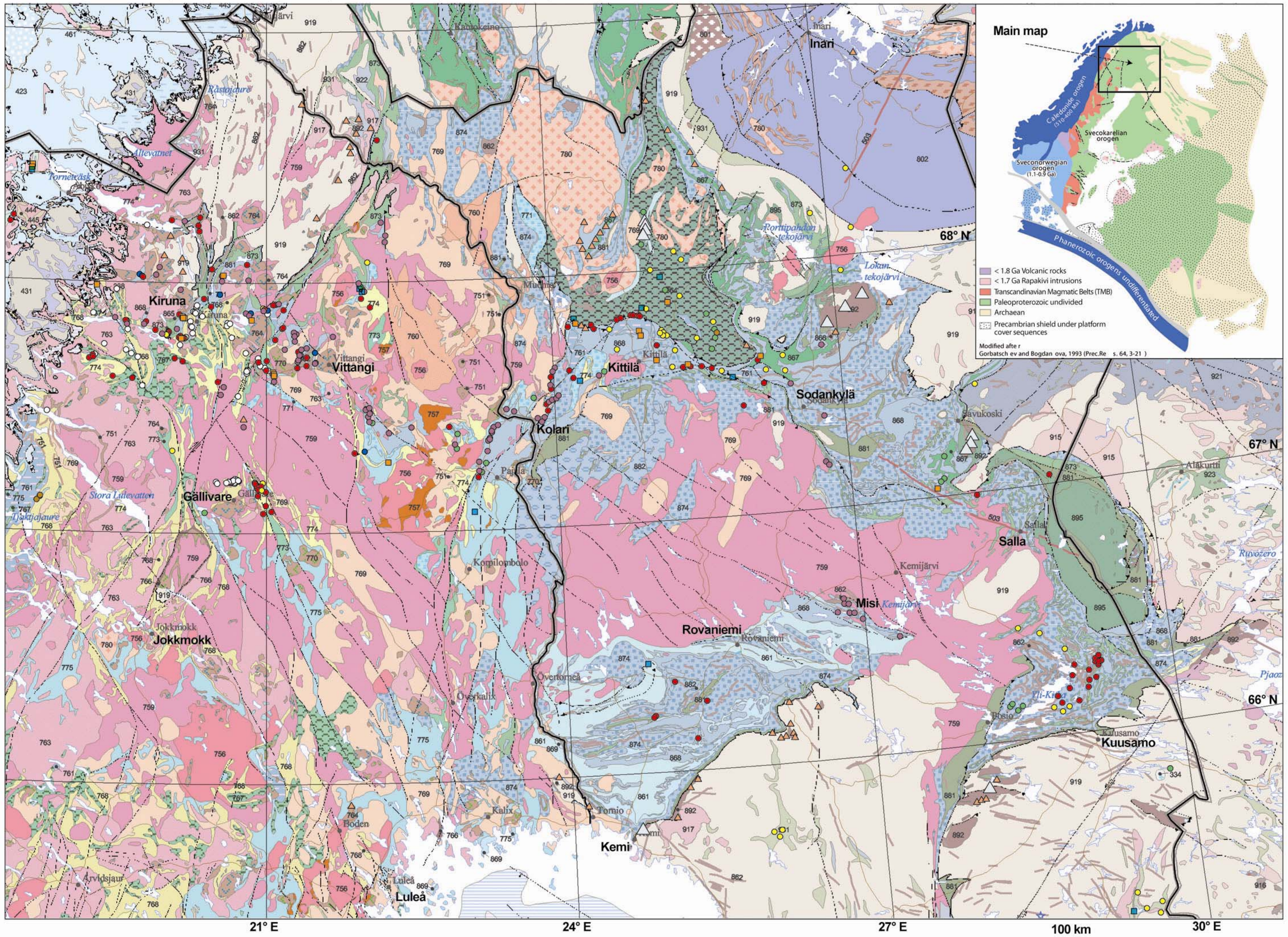
Appendix 1

Excerpt from **GEOLOGICAL MAP OF THE FENNOSCANDIAN SHIELD**, scale 1:1 million



○ Apatite iron ore

Epigenetic deposits
- Copper-gold deposit
- Zinc-lead-silver deposit
- Gold deposit

Stratiform-stratabound base metal and iron deposits
- Skarn-rich iron formation
- Quartz-banded iron formation
- Manganese-iron deposit
- Stratiform copper deposit
- Stratiform zinc-lead-silver deposit

Deposits in mafic to ultramafic rocks
- Ni ± Cu ± PGE deposits
- Cr or Fe - Ti -V deposit

Legend

NEOPROTEROZOIC (AND POSSIBLY MESOPROTEROZOIC) AND PHANEROZOIC ROCKS OUTSIDE THE CALEDONIAN OROGENIC BELT

334 Nepheline syenite, foidolite, carbonatite, melilitolite, pyroxenite, peridotite (Devonian)

CALEDONIAN OROGENIC BELT

Exotic and outboard terranes in Upper and Uppermost Allochthons

423 Metagreywacke, phyllite, conglomerate, quartzite, limestone, felsic and mafic metavolcanic rocks in Upper Allochthon (Lower Palaeozoic)

Continent-ocean transition zone in Upper Allochthon

Neoproterozoic sedimentary cover and igneous rocks

431 Metadolerite including sheeted dyke complex, amphibolite, gabbro, eclogite, ultramafic rock

Continental margin in Middle Allochthon

Neoproterozoic sedimentary cover and Neoproterozoic to Cambrian intrusive rocks

443 Feldspathic metasandstone, meta-arkose, quartzite, metagreywacke, marble, tillite (Neoproterozoic)

NEOPROTEROZOIC (TO MESOPROTEROZOIC) ROCKS

503 Dolerite, metadolerite (c. 1.18-0.93 Ga)

MESOPROTEROZOIC (TO PALAEOPROTEROZOIC) ROCKS

604 Lamprophyre, lamproite (c. 1.23-1.15 Ga)

PALAEOPROTEROZOIC ROCKS (1.96-1.75 Ga)

751 Dolerite, gabbro, metagabbro (c. 1.77 Ga
756 Granite, granodiorite, quartz monzonite, monzonite, syenite and metamorphic equivalents, in part hypersthene-bearing (c. 1.86-1.84 Ga and c. 1.82-1.76 Ga)
757 Gabbro, diorite, ultramafic rock and metamorphic equivalents (c. 1.86-1.77 Ga)
759 Granite, pegmatite (c. 1.85-1.75 Ga)
760 Granodiorite, granite, pegmatite, subordinate gabbro (c. 1.86-1.85 Ga)
761 Red sandstone and mudstone, conglomerate, metasandstone, quartzite, phyllite, volcanic and metavolcanic rocks
763 Granite, monzonite, syenite, in part pyroxene-bearing (c. 1.88-1.87 Ga)
764 Gabbro, diorite, monzodiorite, ultramafic rock (c. 1.88-1.87 Ga)
766 Dolerite dyke complex (c. 1.88 Ga)
767 Mafic to intermediate volcanic and metavolcanic rocks (c. 1.88-1.86 Ga, in part possibly younger)
768 Felsic to intermediate volcanic and metavolcanic rocks (c. 1.88-1.86 Ga, in part possibly younger)
769 Granodiorite, tonalite, granite, monzonite, syenite and metamorphic equivalents, in part hypersthene-bearing (c. 1.91-1.88 Ga, in part as young as c. 1.84 Ga)
770 Gabbro, diorite, ultramafic rock and metamorphic equivalents (c. 1.91-1.88 Ga, in part as young as c. 1.84 Ga)
771 Quartzite, meta-arkose
773 Mafic metavolcanic rock (c. 1.91-1.88 Ga)
774 Felsic to intermediate metavolcanic rock (c. 1.91-1.88 Ga)
775 Metagreywacke, metasiltstone, metasandstone, mica schist, graphite- and/or sulphide-bearing schist, paragneiss, amphibolite intercalations (c. 1.95-1.87 Ga and possibly older)
780 Granodiorite, tonalite, granite, gabbro and metamorphic equivalents; alkaline gneiss (c. 1.96-1.91 Ga)

LAPLAND-WHITE SEA GRANULITE BELT (rocks of uncertain age, in time range 2.30-1.90 Ga)

801 Anorthosite
802 Felsic to intermediate granulitic rock
803 Mafic to intermediate granulitic rock
804 Mafic granulitic rock, amphibolite

PALAEOPROTEROZOIC ROCKS (2.50-1.96 Ga)

Rock group 2.06-1.96 Ga
861 Mica schist, metagreywacke, black schist, conglomerate
862 Gabbro and dolerite, of variable age
864 Tholeiitic basalt, rhyolite, chert, jasper, banded iron formation
865 Tholeiitic basalt, ferropicrite, picrite, peridotite, pyroxenite, gabbro, wehrlite/dolerite
866 Gabbro, peridotite
867 Komatiite, picrite, tholeiitic basalt
868 Black schist, carbonaceous quartzite, siltstone, shungitic rocks, dolostone, limestone, basalt, andesitic basalt, picrobasalt/dolerite
869 Tholeiitic basalt, arkosic sandstone, quartzite, greywacke, dolostone, black schist
Rock group 2.30-2.06 Ga
873 Tholeiitic basalt, subordinate quartzite and conglomerate
874 Quartzite, mica schist, mica gneiss, conglomerate
Rock group 2.40-2.30 Ga
881 Basalt, high-Mg basalt, high-Mg andesite, dacite, komatiitic basalt/dolerite
882 Mica schist, conglomerate, gritstone, diamictite, arkosic sandstone, quartzite, tuffite
Rock group 2.50-2.40 Ga
892 Layered intrusion: gabbro, gabbro-norite, anorthosite, dunite, peridotite, pyroxenite
895 Tholeiitic, komatiitic and andesitic basalt, andesite, dacite, peridotite, gabbro, siltstone, quartzite, arkosic sandstone

ARCHAEAN ROCKS

Plutonic rocks and undifferentiated gneiss and migmatitic rock complexes
915 Granite, pegmatite (c. 2.70-2.65 Ga)
916 Gabbro, monzodiorite, syenite, granodiorite (c. 2.74-2.65 Ga)
917 Granite, granodiorite, diorite, quartz diorite, porphyritic granite (c. 2.75-2.65 Ga)
919 Tonalite-trondhjemite-granodiorite gneiss, quartzo-feldspathic gneiss, enderbite, migmatitic gneiss, with mafic and felsic enclaves (c. 3.20-2.65 Ga and possibly older)
Amphibolite – schist – gneiss belts.
921 High-Al mica schist, mica gneiss, hornblende gneiss with amphibolite enclaves
922 Mica schist and mica gneiss, migmatitic gneiss, amphibolite, banded iron formation
923 Amphibolite, amphibole gneiss
Volcanic-dominated greenstone belts (c. 3.20-2.75 Ga and possibly older)
931 Komatiite, basalt, andesite, dacite, rhyolite
933 Tholeiitic, komatiitic and Fe-rich tholeiitic basalt, peridotite, gabbro, dacite, rhyolite, conglomerate