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Ore Geology Reviews xx (2005) xxx–xxx

ORE GEOLOGY REVIEWS

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Precambrian geodynamics and ore formation: The Fennoscandian Shield

Pär Weihed [a,*], Nicholas Arndt [b], Kjell Billström [c], Jean-Clair Duchesne [d], Pasi Eilu [e], Olof Martinsson [a], Heikki Papunen [f], Raimo Lahtinen [e]

[a] *Division of Ore Geology and Applied Geophysics, Luleå University of Technology, SE-971 87 Luleå, Sweden*
[b] *LGCA, Université Joseph Fourier, Grenoble, Maison des Géosciences 1381, rue de la Piscine 38400 Saint Martin d'Heres, France*
[c] *Swedish Museum of Natural History, Stockholm, Box 50007, SE-10405, Sweden*
[d] *University of Liège Bat. B20, B-4000 Sart Tilman, Belgium*
[e] *Geological Survey of Finland, PO Box 96, FIN-02151 Espoo, Finland*
[f] *Department of Geology, University of Turku FIN-20014 Turku, Finland*

Received 2 July 2004; accepted 6 July 2005

Abstract

Compared with present-day global plate tectonics, Archaean and Palaeoproterozoic plate tectonics may have involved faster moving, hotter plates that accumulated less sediment and contained a thinner section of lithospheric mantle. This scenario also fits with the complex geodynamic evolution of the Fennoscandian Shield from 2.06 to 1.78 Ga when rapid accretion of island arcs and several microcontinent–continent collisions in a complex array of orogens was manifested in short-lived but intense orogenies involving voluminous magmatism. With a few exceptions, all major ore deposits formed in specific tectonic settings between 2.06 and 1.78 Ga and thus a strong geodynamic control on ore deposit formation is suggested.

All orogenic gold deposits formed syn- to post-peak metamorphism and their timing reflects the orogenic younging of the shield towards the SW and west. Most orogenic gold deposits formed during periods of crustal shortening with peaks at 2.72 to 2.67, 1.90 to 1.86 and 1.85 to 1.79 Ga.

The ca. 2.5 to 2.4 Ga Ni–Cu $\pm$ PGE deposits formed both as part of layered igneous complexes and associated with mafic volcanism, in basins formed during rifting of the Archaean craton at ca. 2.5 to 2.4 Ga. Svecokarelian ca. 1.89 to 1.88 Ga Ni–Cu deposits are related to mafic–ultramafic rocks intruded along linear belts at the accretionary margins of microcratons.

All major VMS deposits in the Fennoscandian Shield formed between 1.97 and 1.88 Ga, in extensional settings, prior to basin inversion and accretion. The oldest "Cyprus-type" deposits were obducted onto the Archaean continent during the onset of convergence. The Pyhäsalmi VMS deposits formed at 1.93 to 1.91 Ga in primitive, bimodal arc complexes during extension of the arc. In contrast, the Skellefte VMS deposits are 20 to 30 million years younger and formed in a strongly extensional intra-arc region that developed on continental or mature arc crust. Deposits in the Bergslagen–Uusimaa belt are similar in age to the Skellefte deposits and formed in a microcraton that collided with the Karelian craton at ca. 1.88 to 1.87 Ga. The Bergslagen–

* Corresponding author.
E-mail address: par.weihed@ltu.se (P. Weihed).

doi:10.1016/j.oregeorev.2005.07.008

Uusimaa belt is interpreted as an intra-continental, or continental margin back-arc, extensional region developed on older continental crust.

Iron oxide–copper–gold (IOCG) deposits are diverse in style. At least the oldest mineralizing stages, at ca. 1.88 Ga, are coeval with calc-alkaline to monzonitic magmatism and coeval and possibly cogenetic subaerial volcanism more akin to continental arcs or to magmatic arcs inboard of the active subduction zone. Younger mineralization of similar style took place when S-type magmatism occurred at ca. 1.80 to 1.77 Ga during cratonization distal to the active N–S-trending subduction zone in the west. Possibly, interaction of magmatic fluids with evaporitic sequences in older rift sequences was important for ore formation.

Finally, the large volumes of anorthositic magmas that characterize the Sveconorwegian Orogeny formed a major concentration of Ti in the SW part of the Sveconorwegian orogenic belt under granulite facies conditions, about 40 million years after the last regional deformation of the Sveconorwegian Orogeny, between ca. 930 and 920 Ma.

Keywords: Palaeoproterozoic; Fennoscandian Shield; VMS; Orogenic gold; Iron oxide–copper–gold (IOCG); Anorthosite; Titanium deposit

1. Introduction

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 described in this paper, 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 deposits, orogenic gold deposits, and some very minor VMS deposits occur in the Archaean, virtually all 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. The genesis of these deposits is still much debated and both magmatic segregation (e.g., Nyström and Henriquez, 1994) and hydrothermal (e.g., Hitzman et al., 1992) origin have been advocated. The Kiruna-type Fe-apatite deposits have recently also been suggested to form an end member of the disparate family of iron oxide–copper–gold (IOCG) deposits (cf. Hitzman et al., 1992; Hitzman, 2000; Porter, 2000; Williams et al., 2003). Cu–Au deposits, with a large tonnage and low grade (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., 2003).

In this paper, we discuss the deposit types mentioned above, including the important Mesoproterozoic anorthosite-hosted Ti deposits. Although still much debated both the Kiruna-type Fe-apatite deposits and the intrusion related Cu–Au deposits are here discussed as IOCG deposits. The emphasis is on the geodynamic aspects of ore formation. 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.

2. The geodynamics of ore formation in the Precambrian

Precambrian ore deposits have much in common with more recent deposits. There are some notable differences, such as the virtual restriction of komatiite-hosted deposits to the Archaean and Early Proterozoic, and the absence of Pb-rich VMS deposits and the scarcity of economic porphyry copper deposits in Archaean and Palaeoproterozoic granite–greenstone

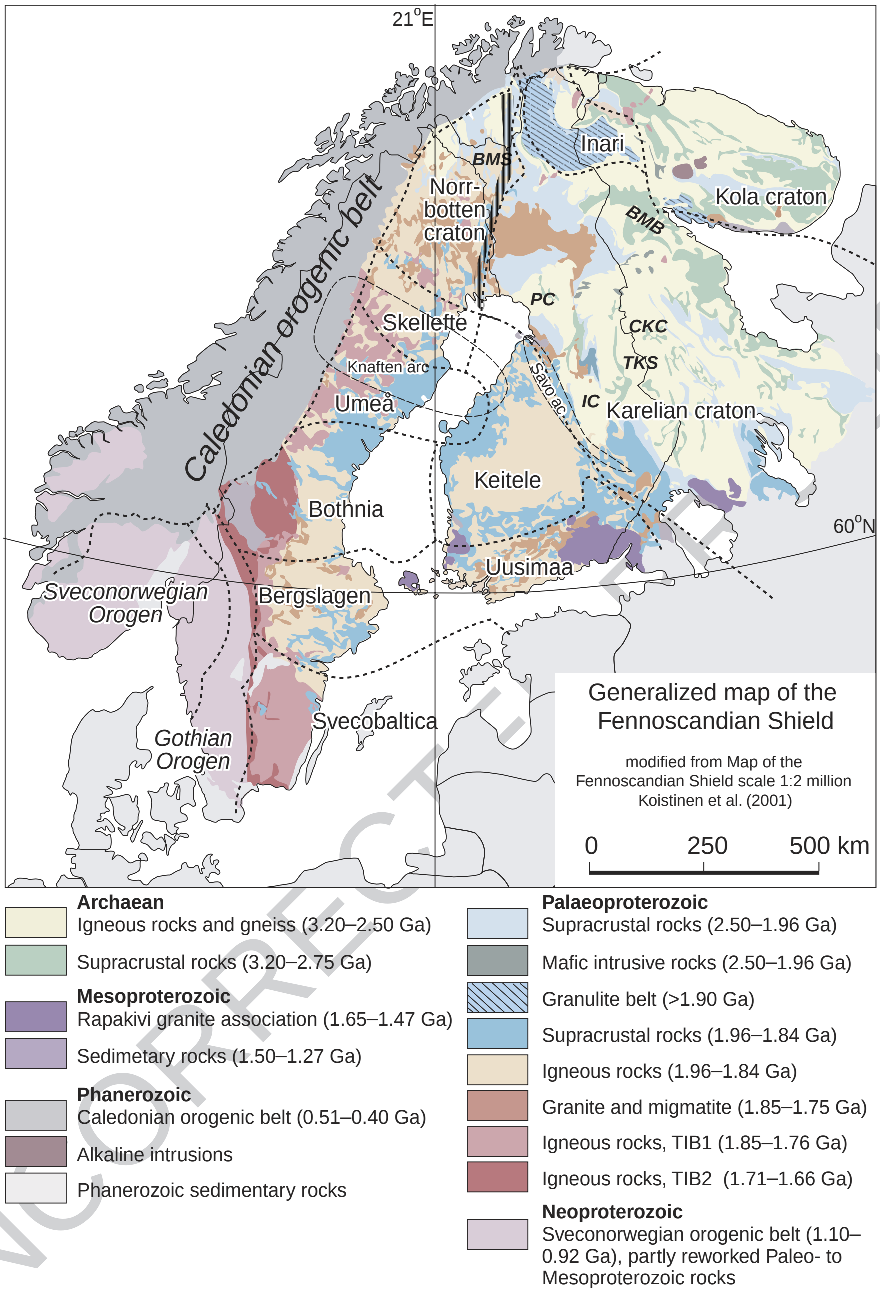


Fig. 1. Simplified geological map of the Fennoscandian Shield with major tectono-stratigraphic units discussed in text (after Lahtinen et al., 2005). Map adapted from Koistinen et al. (2001). BMB = Belomorian Mobile Belt, CKC = Central Karelian Complex, IC = Iisalmi Complex, PC = Pudasjärvi Complex,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

terranes. However, in terms of their overall characteristics, there is little to distinguish an Archaean or Palaeoproterozoic VMS and an Archaean or Palaeoproterozoic orogenic gold deposit from their more modern counterparts. At first sight, this is surprising because physical and chemical conditions in the

Archaean (and Palaeoproterozoic?) mantle and crust should have been very different from those of today. Although we can expect the geodynamic processes to have remained constant through time, the conditions in which they have operated have changed. The mantle was hotter and this would have led to more active and voluminous magmatism (Arndt et al., 1997). At spreading centres (Sleep and Windley, 1982), in mantle plumes (Herzberg, 1992; Nisbet et al., 1993) and above subduction zones (Parman et al., 2001; Arndt, 2003), the degree of mantle melting would have been greater and the magmas produced would have had more magnesian compositions. At spreading centres, the higher degree of mantle melting produced thicker oceanic crust. According to the estimates of Sleep and Windley (1982) and Vlaar (1986), Archaean oceanic crust was 20 to 30 km thick, compared with 6 to 8 km today. We are not sure how this crust subducted, but subduct it probably did, because rocks similar to those found in modern island arcs and convergent margins are recognized in most Archaean greenstone belts. Plate tectonics, though evolving under different conditions, appear to have been operating in the Archaean.

Thus, it is important when interpreting the formation of Archaean and Palaeoproterozoic ore deposits that the likely conditions in the crust and upper mantle at the time, and the uncertainties about them, are taken into account. Inevitably, the evidence is speculative, but the following paragraphs offer a reasonable scenario.

At an Archaean oceanic ridge or back-arc basin spreading centre, with no significant sub-crustal lithosphere, the geothermal gradient across a 30 km-thick Archaean oceanic crust, from <5 °C to at most 50 °C on the ocean floor, to approximately 1400 °C at the base of the crust, would have been 45 to 46 °C/km, far less than the ca. 200 °C/km gradient, from <5 to 1200 °C, across the 6 km thickness of modern oceanic crust. In a modern mid-ocean ridge or back-arc setting, magma chambers located at shallow levels at the spreading centre drive the fluid circulation. Although the geometry of the magmatic plumbing system in the thicker oceanic crust of an Archaean spreading centre is unknown, the thicker crust, lower geothermal gradient and faster spreading rate would probably have combined to distribute magmatic heat sources both deeper and more dispersed. Only the uppermost few km of the oceanic crust interact with seawater and become hydrated (de Wit et al., 1987). This interval represents a significant fraction of 6 km-thick modern oceanic crust but only a small fraction of thicker Archaean oceanic crust. Thus the temperature of fluid circulating in the crust at Archaean spreading centres would probably have been lower than nowadays.

Just before subduction, the modern oceanic plate consists of a thin layer of sediments, thin, largely hydrated oceanic crust and a thick section of mantle lithosphere. Faster moving, hotter Archaean plates accumulated less sediment and contained a thinner section of mantle lithosphere (Sleep and Windley, 1982). To initiate subduction, the basaltic section of the crust must convert, at least in part, to dense eclogite. It is not clear how this happens, even in the modern situation. If the lower parts of 30 km-thick Archaean crust were transformed to eclogite, this large volume of relatively cold, dense rock would have acted as a sinker, dragging down the rest of the slab, despite the thinner mantle portion of the lithosphere. The dips of Archaean subduction zones may have been high (Russell and Arndt, 2005), not low, as is commonly proposed (e.g., Karsten et al., 1996; Foley et al., 2003). Hinge retreat or slab roll-back might have been common. In a subduction zone, the relatively small proportion of hydrated crust would have become sandwiched between hot overlying mantle and the cold, thick lower part of the subducted crust. Perhaps Archaean oceanic crust was not always totally subducted. The partially hydrated, basalt-rich and less dense upper parts may have been obducted onto growing arcs or continents while the dense eclogitized lower parts plunged back into the mantle or, alternatively, the lower ultramafic cumulate part of a differentiated crust may have delaminated and the upper section, composed of hydrated basaltic rocks, may have accreted to the growing island arc (Foley et al., 2003).

Fluids released by dehydration of subducting crust would have passed into a hot, overlying Archaean mantle wedge. These fluids contained little of the sedimentary component that dominates modern fluids and they interacted with peridotite in the mantle wedge that was highly depleted in the basaltic component because of the generally high degree of partial melting. This resulted in water-fluxed melting of a refractory mantle source. The magmas produced may have been boninitic in character, rich in Mg, and poor in Fe and incompatible trace elements. These magmas would have ascended until they reached the crust of the underthrust plate. There they pooled, fractionated, degassed and perhaps provoked partial melting of the crustal rocks.

Some aspects of Archaean island arc successions are consistent with this scenario. Sequences in the Abitibi Belt of Canada (see Thurston and Ayres, 2004, for a recent review) contain both tholeiitic and calc-alkaline members, but the calc-alkaline rocks are poorer in alkalis than their modern counterparts. More significantly, Archaean arcs contain a bimodal volcanic suite comprising basalts and dacite–rhyolites. True calc-alkaline andesites, and especially rocks containing abundant, complexly zoned or eroded plagioclase phenocrysts like modern calc-alkaline andesites, are rare to absent. Boninites, though reported, also seem to be rare, perhaps because the crust formed an effective filter that prevented the passage to the surface of dense, water-rich magmas.

The characteristics of hydrothermal systems on Archaean and Proterozoic island arcs and back-arc basins are a matter of speculation. Fluids may have been hotter or cooler than in modern subduction-related settings, depending on the geometry of the crust and the distribution of heat sources. Fluid fluxes may have been high, due to rapid dehydration of the upper part of the magnesian, serpentine-rich crust as it was dragged rapidly into hot Archaean or Proterozoic mantle. A clear difference would be a scarcity of sediment-hosted deposits like those in Japan and New Brunswick (Allen et al., 2002).

Continental crust formed episodically during the Archaean and Proterozoic. Compilations of reliable ages obtained on crustal rocks, and from detrital zircons in large rivers, produce age spectra characterized by the presence of large peaks at 2.7, 2.5, 2.1 and about 1.9 to 1.8 Ga (Goldstein et al., 1997; Condie, 2004; Nelson, 2004). Although some authors interpret these peaks as the times of supercontinent formation, arguments can be advanced that suggest that they represent periods of accelerated mantle convection and enhanced crustal growth, as could have been the case in the Fennoscandian Shield. In many regions, the peak of granitoid emplacement, the event that is recorded by U–Pb zircon ages, is preceded by about 30 million years by the eruption of voluminous mafic–ultramafic volcanic rocks (Nelson, 2004). In the southern Superior Province of Canada, in the Yilgarn of Australia, in Brazil, Zimbabwe and in parts of the Aldan Shield in Siberia, komatiites and basalts are dated from 2.73 to 2.70 Ga and the peak of granite emplacement at around 2.70 Ga. In the Birimian of West Africa, mafic volcanism is dated at 2.13 Ga and the granites intruded mainly around 2.07 Ga. Abouchami et al. (1990) equate the Birimian mafic volcanism with the enormous ocean plateaux emplaced in the Pacific in the Cretaceous and Boher et al. (1991) interpret the granites as accelerated subduction-related magmatism triggered by the period of enhanced plume activity. This idea has been taken further by Stein and Hofmann (1994) and Condie (2004) who have developed a model in which periods of enhanced plume-dominated mantle convection alternate with periods of plate tectonics.

The physical conditions in the crust and upper mantle when the Archaean and Palaeoproterozoic greenstone belts formed probably correlate with the types of ore deposits they contain. For example, the Abitibi belt in the Canadian Superior Province, which is believed to have formed through the accretion of oceanic crust, oceanic plateaux and island arcs (Card, 1990; Kimura et al., 1993), is the host of numerous large VMS and gold deposits but few Ni–PGE sulphide deposits. In contrast, the greenstone belts of the Yilgarn craton in Western Australia apparently formed through flood volcanism on an older continental basement. This region contains few VMS deposits, abundant gold deposits and hosts, at Kambalda, the type examples of Ni–PGE sulphide deposits in komatiite lava flows (Lesher, 1989). The link between geodynamic context and Ni–PGE mineralization lies in the role that assimilation of continental crust played in the formation of the Kambalda deposits.

The formation of major Neoarchaean VMS, orogenic gold, and Ni–PGE mineralization seems to be restricted to the period ca. 2.74 to 2.69 Ga and appears to correspond to a period of intense intrabasinal mantle plumes and a subsequent global plume-breakout event (Barley et al., 1998). Apart from the low degree of exploration, especially in the eastern part, the obvious lack of major Neoarchaean mineralization in the Fennoscandian Shield could possibly be explained by the age of Neoarchaean greenstone belts. Most of the Fennoscandian greenstone belts seem to be slightly older than the global Neoarchaean peak in mineralization and mantle plume activity (e.g., Huhma et al., 1999) and, hence, magmatism and hydrothermal activity might have been less intense and unable to form major metal deposits. This could also explain the lack of major lode gold deposits related to subsequent accretion during peak orogeny. It is also possible that the Fennoscandian Shield exhibits a smaller window into Archaean greenstone terrains, simply due to the local peculiarities of late orogenic evolution, when the thermal regime influences the extent of crustal remelting and degree of exhumation (Peter Sorjonen-Ward, pers. comm. 2005). Hence, although there is evidence for Neoarchaean greenstone belts formed through the accretion of oceanic crust, oceanic plateaux and island arcs and those formed through flood volcanism on an older continental basement (see below), they apparently do not contain major ore deposits in the Fennoscandian Shield. This is in contrast to the Palaeoproterozoic period of ca. 1.95 to 1.80 Ga in the Fennoscandian Shield, which is intensely mineralized and constitutes one of the major Palaeoproterozoic metallogenetic provinces on Earth. Most of the Fennoscandian ore deposits (i.e., VMS, IOCG, and orogenic gold) in the Palaeoproterozoic are related to the formation of island arcs, continental margin arcs, or subsequent accretion to the Archaean (Karelian) craton. This is discussed further below.

3. Tectonic evolution of the Fennoscandian Shield

3.1. Archaean geology

The Archaean bedrock in the Fennoscandian Shield includes two cratonic nuclei, Karelia and Kola, that were fragmented and further reassembled during the Palaeoproterozoic (Fig. 1). The Karelian craton can be divided into the Belomorian mobile belt and three complexes: Central Karelian, Iisalmi, and Pudasjärvi (Fig. 1). The boundary zone between the Central Karelian Complex and the Belomorian mobile belt was formed by accretion of island arc and continental fragments to the Karelian core in the Neoarchaean (Mints et al., 2001). This boundary zone and the whole Belomorian mobile belt were strongly reactivated in the Palaeoproterozoic (Gaál and Gorbatschev, 1987; Bibikova et al., 2001).

Extensive areas of the granitoid and gneiss terrain of eastern Karelia are dated at 2.85 Ga, whereas rocks in the central part of the Karelian Province yield ages of ca. 3.05 Ga (Sm–Nd method, Slabunov and Bibikova, 2001). The Vodlozero terrane east of the Lake Onega and the central parts of the Pudasjärvi Complex, with ages up to 3.50 Ga, are the oldest parts of the Karelian Craton (Lobach-Zhuchenko et al., 1993; Huhma et al., 2004).

Geochronological studies in NW Russia indicate four generations of greenstones, with age groups of >3.20 to 3.10, 3.10 to 2.90, 2.90 to 2.80 and 2.80 to 2.75 Ga (Slabunov and Bibikova, 2001). Based mainly on trace element and isotope geochemistry of the rocks, Svetov (2001) and Puchtel et al. (1999) inferred that the greenstones formed at a convergent ocean–continent boundary by tectonic accretion, and the final collision stages resulted in asymmetric structures of the greenstone belts.

Detailed stratigraphic studies have been conducted in the Archaean Tipasjärvi–Kuhmo–Suomussalmi (TKS) greenstone complex (Fig. 1) in Finland, which therefore serves as a good example of the Archaean evolution (Piirainen, 1988; Papunen et al., 1998; Halkoaho et al., 2000). A bimodal volcanic sequence at the eastern margin of the Suomussalmi greenstone belt evidently belongs to an older supracrustal formation. For example, Vaasjoki et al. (1999) reported a U–Pb age of 2966 ± 9 Ma for the zircons of felsic volcanic rocks, and consider that there is a time gap of ca. 150 million years between the bimodal volcanic sequence and the lowermost felsic volcanism of the TKS greenstone belt. This time gap is also characterized by deformation, metamorphism, and resetting of U–Pb ages of the tonalite–trondhjemite–granitoid (TTG) complex at 2.83 Ga (Luukkonen, 1992).

The lowermost stratigraphic unit of the TKS proper consists of 2.81 to 2.79 Ga, felsic to intermediate, subaerial to shallow water-deposited, calc-alkaline volcanic rocks. They occur at the margins of the belt and are geochemically similar to TTG rocks. The overlying volcanic formation is composed of pillowed and massive tholeiitic basalts with interlayers of oxidic and sulphidic Algoma-type banded iron formations. A mafic sill, considered to be comagmatic with the tholeiitic basalts, has been dated at 2790 ± 18 Ma and a gabbroic variety of a komatiitic cumulate yielded a zircon U–Pb age of 2757 ± 20 Ma that is so far the only direct age obtained from the ultramafic sequence (Luukkonen, 1992; H. Huhma pers. comm., 1998). A sequence of Al-depleted komatiites and komatiitic basalts overlies the tholeiitic basalts. The komatiite formation includes a thick ultramafic olivine cumulate body and several ultramafic lenses interpreted as channel facies olivine ad- and meso-cumulates of komatiitic volcanic flows. The cumulate body displays pyroxenitic marginal series against the TTG wallrock and also contains enclaves of tholeiitic basalt and felsic TTG gneisses. The komatiites are overlain by basaltic volcanic rocks and finally by detrital and lahar-like felsic to intermediate volcaniclastic deposits, indicating rapid erosion and infilling of basins. The volcanic sequence is linked with a mantle hotspot activity in an intracratonic basin. The continental crust partially melted in the incipient phase to form the felsic volcanic rock and subsequently contaminated the voluminous eruptions of komatiitic lavas.

The TKS greenstone belt was intruded by granodiorites and tonalites with zircon U–Pb ages ranging from 2.75 to 2.69 Ga (Luukkonen, 1992) and was isoclinally folded to form a wide synclinorium. Ductile shear zones deformed the belt and acted as conduits for ascending metamorphic fluids at ca. 2.75 to 2.67 Ga.

3.2. Palaeoproterozoic tectonic evolution from 2.5 to 1.9 Ga

Several periods of sedimentation and magmatism characterize the Palaeoproterozoic evolution of the shield before the Svecokarelian Orogeny. The Archaean craton of Fennoscandia consolidated after the last major phase of granitoid intrusions at 2.69 Ga. During the period 2.5 to 1.9 Ga, it underwent several episodes of continental rifting and related, dominantly mafic, magmatism, denudation and sedimentation. These resulted in the formation of volcano-sedimentary sequences, which were deformed during the Svecokarelian Orogeny between 1.9 and 1.8 Ga. The lowermost stratigraphic units consist of clastic sedimentary rocks and bimodal volcanism. An unconformity characterized by polymict conglomerates separates these groups from the overlying epicontinental sediments and basalts and denotes a period of weathering and quiescence in the tectonic history.

3.2.1. Incipient rifting, volcanism, and emplacement of layered igneous complexes and mafic dykes

The beginning of the rifting period between 2.51 and 2.43 Ga is indicated by intrusion of numerous layered mafic igneous complexes (Alapieti and Lahtinen, 2002). 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.

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 crystallized from a similar, quite primitive magma type, which is characterized by slightly negative initial $\varepsilon_{Nd}^{(T)}$ 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 crystallized from an evolved Ti-poor, Al-rich basaltic magma.

Amelin et al. (1995) emphasize the existence of two slightly different age groups of the intrusions, 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. The first includes intrusions along the Polmak–Pechenga–Imandra–Varzuga greenstone belt, such as Mt. Generalskaya in Pechenga and Imandra, Monchegorsk, Pana and Feodor Tundras in central Kola Peninsula and the second the intrusions in Karelia and the Kola–Finnish Lapland areas (Fig. 2). However, the timing of these intrusions is disputed, as contrasting U–Pb ages of 2496 ± 10 and 2447 ± 10 Ma were recently determined from two different stratigraphic levels of the Mt. Generalskaya intrusion (Bayanova et al., 1999) and similar results have been obtained from the Panski Tundra intrusions (Mitrofanov and Bayanova, 1999).

Amelin et al. (1995) bracketed the U–Pb age of the oldest basaltic volcanism between 2.443 and 2.440 Ga, as the volcanic rocks are intruded by a 2441.3 ± 1.2 Ma pluton but overlie a 2442.1 ± 1.4 Ma pluton. Other age determinations of volcanism in eastern Karelia indicate ages between 2.45 and 2.43 Ga. Manninen

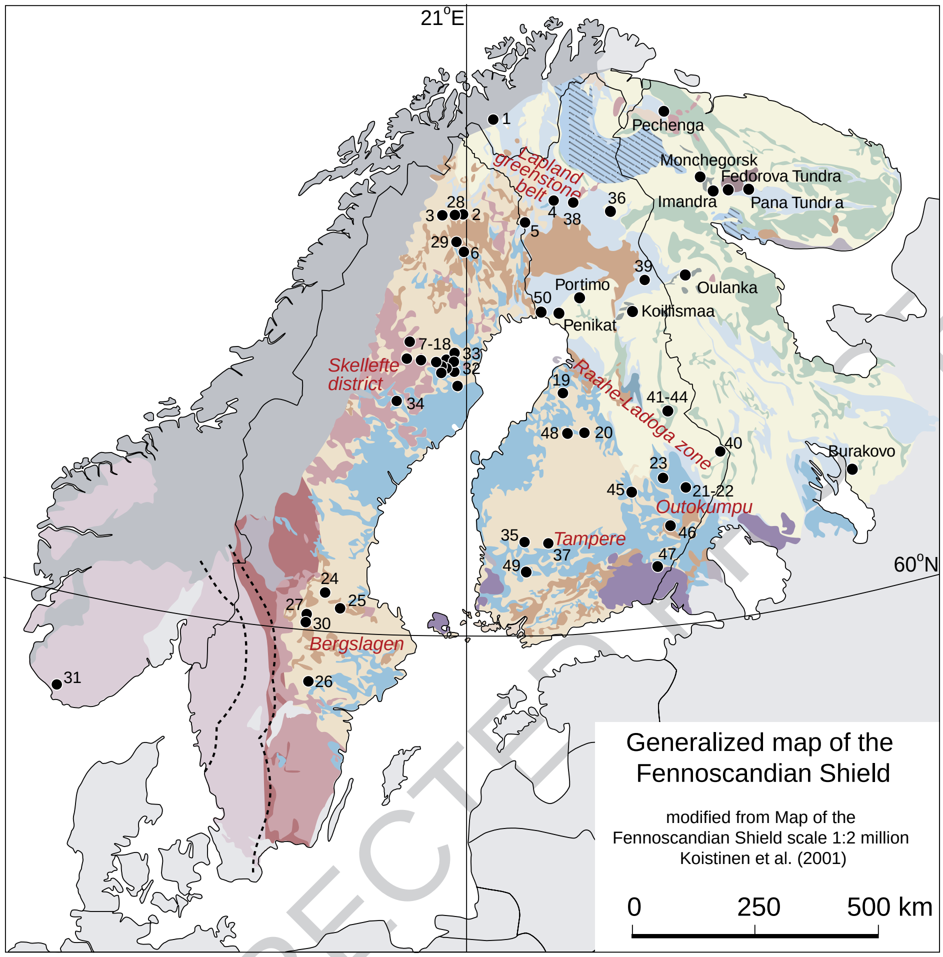


Fig. 2. Simplified geological map of the Fennoscandian Shield with major ore deposits discussed in text indicated. Map adapted from Koistinen et al. (2001). Numbers refer to deposits listed in Table 1. Deposits without numbers are discussed in the text but are not included in Table 1.

and Huhma (2001) reported similar ages from the Central Lapland greenstone belt. Geochronological data from Finland and from the Russian part of the shield thus show that large volumes of continental volcanic rocks were erupted contemporaneously with the 2.44 to 2.45 Ga layered intrusions. A number of coeval mafic dykes in the Karelian Craton have ages around 2.44 Ga (Vuollo, 1994). Sm–Nd isotope studies indicate that boninitic dykes have negative $\varepsilon_{Nd}^{(T)}$ values from -1.2 to -2.5, consistent with the values obtained from the layered intrusions (Huhma et al., 1990; Alapieti and Lahtinen, 2002).

Layered intrusions, mafic dyke swarms, and coeval continental volcanism together suggest mantle plume activity in an extensional setting (Amelin et al., 1995). This model implies extension and crustal uplift before volcanism and subsequent subsidence and formation of graben structures during and after volcanism. Poorly sorted sediments characteristic of graben infill are recorded at this stratigraphic level (Gorbunov et al., 1985). The intense faulting and displacement are related to the late stages of rifting, probably to the cooling and subsidence of the brittle layer of the lithosphere after the buoyant rise of the base of the lithosphere during the active rifting phase (see Alapieti and Lahtinen, 2002). The intrusions were later deformed and metamorphosed during the Svecokarelian Orogeny.

3.2.2. Evolution between 2.33 and 1.87 Ga

The Polmak–Pechenga greenstone belt extends for about 200 km from northern Finland to NW Russia. The area has been intensely studied since the discovery of the Pechenga Ni–Cu deposits in the 1920s, and the review here is mainly based on the work of Hanski (1992), Melezhik (1996), Green and Melezhik (1999) and Barnes et al. (2001). The 2.3 Ga, subaerially deposited, andesitic basalts of the lowermost volcanic unit of the Pechenga sequence represent the next episode of magmatic activity in the Fennoscandian Shield (Melezhik, 1996). Conglomerate and a regolith on the Archaean crust underlie these basalts. The conglomerate contains clasts derived from the Mt. Generalskaya layered complex (2505 ± 1.6 Ma U–Pb age of baddeleyite, Amelin et al., 1995), indicating that the crust had been deeply eroded to expose the intrusion before the onset of volcanism. The Pechenga sequence developed with cyclic sedimentation and volcanism in an intracratonic rift zone, but the volcanism at ca. 2.1 to 2.0 Ga represents a transition from an intracratonic to an intercontinental oceanic rift environment that developed between 2.00 and 1.97 Ga into an oceanic rift with voluminous sedimentation and volcanism. The Pechenga sequence was subducted between 1.97 and 1.87 Ga and a continent–continent collision followed from 1.87 to 1.80 Ga in the region (Melezhik, 1996). Ultramafic, ferropicritic volcanic rocks (Hanski and Smolkin, 1989; Hanski, 1992), dated at 1,977 ± 52 Ma (Sm–Nd, Pb–Pb, Hanski, 1992) host the Pechenga Ni–Cu deposits. The ferropicrites are enriched in LILE, LREE and HFSE and display similarities with the picritic volcanic rocks of the Central Lapland greenstone belt discussed below. Although the idea of a core signature has been suggested for the Fe-rich Archaean to Proterozoic ultramafic volcanic rocks (Puchtel et al., 1999), it has also been argued that the Archaean/Proterozoic picrites were 30% richer in Fe compared with modern OIB since the mantle was more Fe-rich at that time (Francis et al., 1999; Barnes et al., 2001).

The Central Lapland greenstone belt (CLGB) is the most extensive belt of mafic volcanic rocks and related sedimentary units in Fennoscandia (Fig. 2). The CLGB is divided into seven lithostratigraphic groups (Lehtonen et al., 1998; Hanski et al., 2001a; Vaasjoki, 2001), of which the oldest belong to the intracratonic hotspot-related rift evolution of 2.45 to 2.43 Ga discussed above. After this phase, epiclastic sedimentary rocks, quartzites, mica schists and minor carbonate rocks were deposited. Räsänen and Huhma (2001) reported ages between 2.45 and 2.22 Ga for mafic magmatism interlayered in the sedimentary sequence, but the extensive 2.2 Ga-old mafic sills represent the main phase of igneous activity, indicating renewed rifting of the underlying craton. The depositional basin became deeper and fine-grained sediments, phyllites and carbonaceous sulphide-bearing black schists were deposited between 2.2 and 2.06 Ga. These are characterized by isotopically heavy carbon with $\delta^{13}C$ values up to +18‰, which correlate with the global heavy C-isotopes of carbonates of this time span (Karhu, 1993; Melezhik, 1996). The 2058 ± 4 Ma Kevitsa mafic–ultramafic intrusion intersects the sequence and provides a minimum age for the sedimentary sequence. The sediments in the basin are overlain by a voluminous, 2.05 Ga, ultramafic komatiitic and picritic volcanic belt extending for 350 km from central Lapland to northern Norway. The komatiites are exceptional in their chemical composition with high Ti concentrations and flat LREE-depleted patterns (Barnes and Often, 1990), whereas the picrites are LREE and HFSE enriched, which indicate mantle hotspot activity (Hanski et al., 2001b).

The overlying mafic submarine volcanic sequences of the CLGB display tectonic contacts against the 2.2 to 2.05 Ga unit, with the contact zone being characterized by serpentinites interpreted as dismembered and overthrust pieces of ocean-floor ophiolites (Hanski, 1997). The submarine volcanic sequence is composed of tholeiitic mafic volcanic rocks and mafic to intermediate volcaniclastic rocks interlayered with fine-grained detrital and chemical sediments. This indicates the existence of oceanic crust, which was obducted to its present position before the intrusion of 1.91 Ga granites. A U–Pb age of ca. 2.015 Ga has been obtained for the upper part of this unit (Rastas et al., 2001).

3.2.3. Summary of evolution between 2.5 and 1.87 Ga

Intracratonic basin evolution with intermittent volcanism lasted for up to 500 million years with no indications of accretionary phases or formation of major new felsic crust. Recurrent mantle hotspot activity characterizes the period 2.5 to 2.05 Ga with numerous layered intrusions, komatiite and picrite eruptions in Finnish Lapland at 2.05 Ga, and finally the extrusion/intrusion of the Pechenga ferropicrites at 1.97 Ga. Oceanic crust only formed during the last phase of the extension and was finally obducted during accretion and continent–continent collision at ca. 1.9 to 1.8 Ga.

Such a long period for the evolution of intracratonic, transient and oceanic basins deviates from the normal time span of Phanerozoic plate tectonic processes where a major intracratonic rifting is followed by opening of an ocean in a relatively short period of time. Also the time span of about 30 million years that commonly separates the major mafic volcanic episodes from the intrusion of granitoids, as noted above, did not hold in the geotectonic evolution of the Fennoscandian Shield between 2.5 and 2.0 Ga, as several phases of volcanism and mafic intrusions formed without related felsic magmatism. This must have been the result of a profound change in mantle convection and interaction between asthenosphere and lithosphere at this time, but discussion of these processes is beyond the scope of this paper.

3.3. The 1.9 to 1.8 Ga Svecokarelian Orogeny

The most intense crustal growth in the Palaeoproterozoic took place during the Svecokarelian/Svecofennian Orogeny (both names occur in the literature) at ca. 1.9 to 1.8 Ga. Hietanen (1975) presented the first plate tectonic interpretation of the Svecofennian Orogeny based on the comparison between western North American Cordilleran and the Svecofennian. Gaál (1982) presented a plate tectonic model with a subduction towards ENE and collision at ca. 1.9 Ga. Gaál and Gorbatschev (1987) later extended this model. Ward (1987) argued that the scarcity of subduction-related magmatism in the Archaean craton margin and the easterly-directed tectonic transport implied westerly-directed subduction before collision. Gaál (1990) adopted this idea and included a subduction reversal in his model to account for the voluminous magmatism in central Finland.

The ca. 1.95 Ga rocks in the Knaften area (Wasström, 1993), south of the Skellefte District in Sweden, and the 1.92 Ga primitive island arc rocks in the Savo Belt (Korsman et al., 1997), adjacent to the Archaean craton in Finland (Fig. 1), are the oldest documented Svecofennian units in the shield, but older protoliths (~2.1 to 2.0 Ga) are inferred from Nd isotope geochemistry and detrital zircon studies (Lahtinen and Huhma, 1997). Island arc-type volcanic rocks and coeval calc-alkaline granitoids aged 1.90 to 1.87 Ga dominate in the central Fennoscandian Shield. Plutonic rocks in Sweden, aged 1.80 to 1.78 Ga with mixed I- to A-type characteristics, represent the youngest major Palaeoproterozoic magmatism in the central shield. Migmatites with tonalitic leucosome in Finland formed from immature psammites at 1.89 to 1.88 Ga, whereas younger migmatites with granite leucosome, and associated S-type granites, formed at 1.86 to 1.82 Ga in the Bothnian Basin and in the southern part of the Svecofennian domain (Lundqvist et al., 1998; Korsman et al., 1999; Rutland et al., 2001; Weihed et al., 2002). The southern Svecofennian domain includes the 1.90 to 1.89 Ga Bergslagen–Uusimaa belt (Fig. 1), which formed, in part, in an intra-arc basin of a mature continental arc (e.g., Kähkönen et al., 1994; Allen et al., 1996a). Metapelite-dominated sedimentary rocks, quartzites and carbonate rocks characterize the southern part of the Svecofennian domain. Plutonism in that area occurred between 1.89 and 1.85, 1.84 and 1.82, and 1.81 and 1.79 Ga. S-type granites and migmatites aged 1.84 to 1.82 Ga form a belt that extends from SE Finland to central Sweden (e.g., Korsman et al., 1999).

Reflection seismic studies in the 1980s indicated possible fossil subduction zones and remnant slabs from subduction immediately south of the Skellefte District in Sweden (BABEL Working Group, 1990). Korja et al. (1993) proposed a mantle underplating model to account for the thick crust in central Finland, whereas Lahtinen (1994) presented a model for the Svecofennian of Finland involving several accretionary units and three collisional stages at 1.91 to 1.90, 1.89 to 1.88, and 1.86 to 1.84 Ga. Korja (1995) introduced the concept of orogenic collapse to account for the variation in crustal thickness in southern Finland, and Nironen (1997) presented a kinematic plate tectonic model for the Svecofennian Orogen starting with the opening of an ocean at 1.95 Ga, followed by progressive accretion of two arc complexes on to the Archaean craton between 1.91 and 1.87 Ga. The accretionary orogens are progressively younger towards the west, with the subsequent Gothian Orogeny between 1.75 and 1.55 Ga. Later reworking of the crust occurred during the Sveconorwegian/Grenvillian Orogeny at ca. 1.15 to 0.9 Ga (e.g., Gorbatschev and Bogdanova, 1993; Åhäll and Larson, 2000).

According to Lahtinen et al. (2003, 2004, 2005) the ca. 2.00 to 1.92 Ga evolution of the shield involved the amalgamation of several microcontinents and island arcs. This included several pre-1.92 Ga cratons, >2.0 Ga microcontinents, and ca. 2.0 to 1.95 Ga island arcs. Previous models have proposed a semi-continuous Svecokarelian/Svecofennian Orogeny whereas Lahtinen et al. (2004, 2005) define five orogenies for the time period 1.92 to 1.79 Ga and divide this period into a microcontinent accretion stage (1.92 to 1.88 Ga), and a continent–continent collision stage (1.87 to 1.79 Ga).

The Palaeoproterozoic tectonic evolution of the Karelian craton, the Archaean nucleus of the shield, involved a long period of intracontinental extension between 2.5 and 2.1 Ga, finally leading to a continental break-up at 2.06 Ga as described above. Subsequent convergence and microcontinent accretion (1.92 to 1.88 Ga) resulted in the collision of the Kola and Karelian cratons (cf. Fig. 1) that, according to Lahtinen et al. (2004, 2005), led to the Lapland–Kola Orogeny. The collision of the Karelian craton with the Norrbotten craton (the Archaean rocks west of the Bothnian megashear as defined by Berthelsen and Marker, 1986) and the Keitele microcontinent, and the docking of the Bothnia microcontinent (Fig. 1), led to the Lapland–Savo Orogeny. The collision of the Bergslagen microcontinent with the newly formed Archaean–Palaeoproterozoic complex led to the Fennian Orogeny.

During subsequent continent–continent collision, two subduction zones, in the south and in the west, were active between 1.86 and 1.81 Ga. According to this model (Lahtinen et al., 2004, 2005), subduction was followed by oblique collision of Fennoscandia with Sarmatia between 1.84 and 1.80 Ga, defining the Svecobaltic Orogeny. A crustal-scale shear zone divided the Svecobaltic Orogen into two distinct compressional regimes; (1) a retreating subduction zone at an Andean-type margin in the SW, and (2) a transpressional regime in the SE. A collision between Amazonia and Fennoscandia affected the central and northern parts of the western edge of the Fennoscandian Shield at 1.82 to 1.80 Ga and is defined as the Nordic Orogeny. Orogenic collapse and the stabilization of the Fennoscandian Shield essentially occurred between 1.79 and 1.77 Ga. This was followed by younger orogenies in the SW and a westward growth of the shield.

4. Ore-forming processes—relationship between ore deposits and geodynamic setting

In the Fennoscandian Shield, the major deposit types are excellent guides to the geodynamic processes that operated in the Archaean and Proterozoic. Here we discuss five important deposit types and emphasize their relationship with the evolution of the shield. All deposit types contain ores that are or have been economic and that today are actively explored for in the shield. The five deposit types are (1) Ni–Cu–PGE deposits, (2) VMS (Zn–Cu–Pb ± Au ± Ag) deposits, (3) orogenic gold deposits, (4) iron oxide–copper–gold deposits (IOCG), including Kiruna-type Fe deposits and (5) Fe–Ti oxides in anorthosites. Major deposits in the Fennoscandian Shield are listed in Table 1 and their distribution is shown in Fig. 2.

4.1. Ni–Cu–PGE deposits

Ni–Cu ± PGE deposits occur in several different settings within the shield. Mining of Ni as the main commodity has mainly occurred in NW Russia, (e.g., Pechenga), in Finland (e.g., Kotalahti, Hitura and Vammala) and only to a lesser extent in Sweden (e.g., Lainejaur). It is possible to subdivide these deposits on the basis of their geodynamic setting into the following types: (1) deposits in Archaean greenstone belts (2.74 Ga), (2) deposits in mafic layered intrusions (2.49 to 2.45 Ga), (3) deposits in Palaeoproterozoic greenstone belts (2.2 to 2.05 Ga), (4) deposits in Palaeoproterozoic ophiolite complexes (1.97 Ga), (5) deposits associated with rift-related ultramafic volcanism (1.97 Ga), (6) deposits in Svecofennian orogenic mafic–ultramafic intrusions (1.88 Ga) and (7) deposits in post-orogenic diabase dykes. Of these, type four is discussed in the VMS section and type seven is minor and will not be discussed further.

Table 1
Grade and pre-mining tonnage of selected[a] ore deposits in the Fennoscandian Shield

	Type[b]	Ton. (Mt)	Cu %	Zn %	Pb %	Co %	Au ppm	Ag ppm	Fe %	Ni %	Cr %	TiO_2 %	Status	Reference[c]
1) Bidjovagge	Cu–Au	2	1.2	–	–	–	3.6	–	–	–	–	–	Closed mine	Ettner et al. (1994)
2) Viscaria	Strat. Cu	12.54	2.29	–	–	–	–	–	–	–	–	–	Closed mine	Martinsson et al. (1997)
3) Pahtohavare	Cu–Au	1.7	1.9	–	–	–	0.9	–	–	–	–	–	Closed mine	Lindblom et al. (1996)
4) Saattopora	Cu–Au	2.2	0.28	–	–	–	3.29	–	–	–	–	–	Closed mine	Grönholm (1999)
5) Laurinoja	Cu–Au	4.6	0.88	–	–	–	0.95	–	–	–	–	–	Closed mine	Hiltunen (1982)
6) Aitik	Porph.	1600	0.4	–	–	–	0.2	4	–	–	–	–	Active mine	Wanhainen et al. (2003)
7) Boliden	VMS	8.3	1.4	0.9	0.3	–	15.5	50	–	–	–	–	Closed mine	Bergman Weihed et al. (1996)
8) Långdal	VMS	4.4	0.1	5.8	1.7	–	1.9	149	–	–	–	–	Closed mine	Allen et al. (1996a)
9) Långsele	VMS	12.0	0.6	3.9	0.3	–	0.9	25	–	–	–	–	Closed mine	Allen et al. (1996a)
10) Renström	VMS	> 9	0.8	6.5	1.5	–	2.8	155	–	–	–	–	Active mine	Allen et al. (1996a)
11) Petiknäs S	VMS	6.5	1.1	4.8	0.9	–	2.3	108	–	–	–	–	Active mine	Allen et al. (1996a)
12) Udden	VMS	6.7	0.4	4.3	0.3	–	0.7	36	–	–	–	–	Closed mine	Allen et al. (1996a)
13) Maurliden W.	VMS	6.9	0.2	3.4	0.4	–	0.9	49	–	–	–	–	Active mine	Allen et al. (1996a)
14) Näsliden	VMS	4.6	1.1	3.0	0.3	–	1.3	35	–	–	–	–	Closed mine	Allen et al. (1996a)
15) Rakkejaur	VMS	> 20	0.3	2.4	–	1.0	50	–	–	–	–	–	Closed mine	Allen et al. (1996a)
16) Kristineberg	VMS	> 22	1.0	3.2	0.4	–	1.0	32	–	–	–	–	Active mine	Allen et al. (1996a)
17) Rävlidmyran	VMS	7.5	1.0	3.9	0.6	–	0.8	51	–	–	–	–	Closed mine	Allen et al. (1996a)
18) Rudtjebäcken	VMS	4.7	0.9	2.9	0.1	–	0.3	10	–	–	–	–	Closed mine	Allen et al. (1996a)
19) Vihanti	VMS	28.1	0.48	5.12	0.36	–	0.49	25	–	–	–	–	Closed mine	Helovuori (1979)
20) Pyhäsalmi	VMS	71	0.79	2.47	–	–	0.4	15	–	–	–	–	Active mine	Weihed (2001)
21) Keretti	VMS	28.5	3.8	1.07	–	0.24	0.8	8.9	–	0.17	–	–	Closed mine	Gaál (1985)
22) Vuonos	VMS	5.9	2.45	1.6	–	0.15	0.1	11	–	0.17	–	–	Closed mine	Gaál (1985)
23) Luikonlahti	VMS	7.5	0.99	0.5	–	0.11	–	–	–	0.09	–	–	Closed mine	Gaál (1985)
24) Falun	VMS	28.1	~3	4	1.5	–	~3[d]	~20	–	–	–	–	Closed mine	Allen et al. (1996b)
25) Garpenberg	VMS	21.5	0.3	5.3	3.3	–	0.65	98	–	–	–	–	Active mine	Allen et al. (1996b)
26) Zinkgruvan	VMS	60?	–	10	2	–	–	50–100	–	–	–	–	Active mine	Allen et al. (1996b)
27) Saxberget	VMS	6.8	0.9	7.1	2.2	–	0.4	42	–	–	–	–	Closed mine	Allen et al. (1996b)
28) Kiirunavaara	FeOx	> 2000	–	–	–	–	–	–	> 60		–	–	Active mine	Martinsson (1997)

29) Malmberget	FeOx	> 660	–	–	–	–	–	51–61	–	–	–	–	Active mine	Martinsson (1997)
30) Grängesberg	FeOx	> 198	–	–	–	–	–	–	58–64		–	–	Closed mine	Allen et al. (1996b)
31) Tellnes	Ti	> 300	–	–	–	–	–	–	–	–	–	18	Active mine	Charlier (this volume)
32) Björkdal	Active mine	> 20	–	–	–	–	2.6	–	–	–	–	–	Active mine	Weihed et al. (2003)
33) Åkerberg	Active mine	1.1	–	–	–	–	3	–	–	–	–	–	Closed mine	Sundblad (2003)
34) Svartliden	Active mine	2.5	–	–	–	–	5.4	–	–	–	–	–	Active mine	Sundblad (2003)
35) Haveri	VMS	1.5	0.37	–	–	–	2.8	–	–	–	–	–	Closed mine	Eilu et al. (2003)
36) Pahtavaara	Active mine	> 3	–	–	–	–	3	–	–	–	–	–	Active mine	Eilu et al. (2003)
37) Kutemajärvi	Epith. Au	2.0	–	–	–	–	9	–	–	–	–	–	Closed mine	Poutiainen and Grönholm (1996)
38) Suurikuusikko	Active mine	17	–	–	–	–	5.2	–	–	–	–	–	Prospect	Eilu et al. (2003)
39) Juomasuo	Active mine	1.8	–	–	–	0.2	3	–	–	–	–	–	Closed mine	Eilu et al. (2003)
40) Pampalo	Active mine	1.2	–	–	–	–	8	–	–	–	–	–	Closed mine	Eilu et al. (2003)
41) Peura-aho	Ni–Cu	0.26	0.24	–	–	–	–	–	–	–	0.58	–	Prospect	Kurki and Papunen (1985)
42) Hietaharju	Ni–Cu	0.24	0.43	–	–	–	–	–	–	0.86	–	–	Prospect	Kurki and Papunen (1985)
43) Arola	Ni–Cu	1.52	–	–	–	–	–	–	–	0.56	–	–	Prospect	Kurki and Papunen (1985)
44) Sika-aho	Ni–Cu	0.18	–	–	–	–	–	–	–	0.66	–	–	Prospect	Kurki and Papunen (1985)
45) Kotalahti	Ni–Cu	13	0.27	–	–	–	–	–	–	0.72	–	–	Closed mine	Papunen and Gorbunov (1985)
46) Laukunkangas	Ni–Cu	6.7	0.22	–	–	–	–	–	–	0.76	–	–	Closed mine	Papunen and Gorbunov (1985)
47) Telkkälä	Ni–Cu	0.6	0.35	–	–	–	–	–	–	1.41	–	–	Closed mine	Papunen and Vorma (1985)
48) Hitura	Ni–Cu	13	0.2	–	–	–	–	–	–	0.59	–	–	Closed mine	Papunen and Gorbunov (1985)
49) Vammala	Ni–Cu	7.4	0.4	–	–	–	–	–	–	0.69	–	–	Closed mine	Papunen and Gorbunov (1985)
50) Kemi	Cr	> 236	–	–	–	–	–	–	–	–	26	–	Active mine	Alapieti et al. (1989)

The genesis of some deposits is still debated, this is discussed further in the text.

[a] This table lists all major deposits, or deposits discussed in the text, of each type in the Fennoscandian Shield where deposit data are available. Data for deposits 4, 5, 19, 20, 21, 22, 23, 35–50 from the Geological Survey of Finland deposit database, data for deposit 1, 2, 7–18 from Weihed (2001), data for deposit 6 from Boliden Mineral AB, data for deposits 24–27 from Allen et al. (1996b), data for deposits 28–30 from Geological Survey of Sweden mineral deposit database and data for deposit 31 from Charlier (2005).

[b] Abbreviations: Strat. Cu = Stratiform Cu deposits; Porph. = Porphyry type deposit; VMS = Volcanogenic massive sulphide deposit; FeOx = Fe-oxide deposit; Orog. Au = Orogenic gold deposit; Epith. Au = Epithermal Au deposit.

[c] There is only one reference for each deposit listed in the table. Where possible this is a recent reference containing more references to the deposit concerned.

[d] Grade for epigenetic quartz vein hosted gold part of deposit.

4.1.1. Archaean greenstone belts

Unlike many other Archaean areas, the greenstone belts of the Fennoscandian Shield include only minor occurrences of Ni–Cu sulphides of which the disseminated Ni sulphide mineral deposits of Vaara and Kauniinlampi (Halkoaho and Pietikäinen, 1999) in a komatiitic cumulate of the Suomussalmi greenstone belt are the most notable. The Ni/S ratios are high and Cu tenors very low, similar to the Mt. Keith type of komatiitic sulphides (Naldrett, 1989). The prospects of Peura-aho and Hietaharju (Kurki and Papunen, 1985; Table 1) are composed of massive and disseminated Ni–Cu sulphides at the contact zone between a basal cumulate serpentinite lens of a komatiitic basalt flow and underlying sulphide-bearing felsic volcanic rock. The Arola and Sika-aho prospects (Table 1) are tectonically remobilized Ni-sulphides in shear zones, whereas the Tainiovaara deposit is located in a small, intensely metamorphosed and deformed ultramafic lens totally surrounded by Archaean granitoids.

The geological environment and stratigraphic sequences of the TKS greenstone complex (see above) are quite similar to those of the Norseman–Wiluna belt, Western Australia (Hill, 2001), although the dimensions of the TKS belt are much smaller. Thermal erosion of sulphidic substrates and channelized ultramafic volcanic flows are considered prerequisites for the contamination and accumulation of Ni-bearing sulphides in komatiites (Huppert et al., 1984; Lesher et al., 1984; Hill, 2001; Naldrett, 2001). In the Kuhmo belt, sulphide-bearing chert layers that locally underlie the komatiitic flows and cumulates in the lowermost komatiitic flow are depleted in Ni, indicating sulphide contamination and segregation somewhere in the passage of the flow (Papunen et al., 1998).

4.1.2. Mafic layered intrusions

The 2.5 to 2.4 Ga period of igneous activity that resulted in the emplacement of numerous layered mafic–ultramafic intrusive complexes was important in terms of major chromitite and Ni–Cu–PGE deposits. According to Alapieti and Lahtinen (2002), approximately two dozen layered mafic intrusions and intrusion fragments are scattered within the Fennoscandian Shield. One belt extends along the Archaean–Proterozoic boundary and includes the Tornio–Kukkola intrusion at the Finnish–Swedish border, the Kemi and Penikat intrusions and scattered remnants of the wide Portio and Koillismaa complexes. Another belt trends in a SE direction through Finnish Lapland into Russia, and includes the Kaamajoki–Tsohkkoaivi, Koitelainen and Akanvaara intrusions in Finland and the Oulanka complex in Russia. The largest layered intrusion in the shield, the Burakovo intrusion in Russia, may be regarded as a continuation of this belt (Alapieti and Lahtinen, 2002). Several layered mafic intrusions also follow the margins of the Polmak–Pechenga–Imandra–Vazuga–Ust'Ponoy belt.

A number of the layered igneous complexes in Finland host Ni–Cu and PGE occurrences. In NW Russia the intrusions, such as Mt. Generalskaya, Monchegorsk, Imandra, Feodor Tundra and Pana Tundra, are extensive and display high potential for Ni–Cu and PGE deposits. Alapieti and Lahtinen (2002) classified the PGE occurrences into six categories: (1) disseminated base-metal sulphide–PGE deposits, (2) PGE-bearing offset deposits, (3) base-metal sulphide-bearing PGE reefs, (4) sulphide-poor PGE reefs, (5) disseminated base-metal sulphide–PGE deposits associated with microgabbronorites and (6) PGE enrichments associated with "upper chromitites".

Reef type PGE deposits characterize, for example, the Penikat complex, where the megacyclic units were interpreted to be the result of replenishment of the magma in the crystallizing magma chamber, and the lower contacts of the units correlate with both sulphide-bearing and sulphide-poor PGE reefs. The disseminated base-metal sulphide–PGE deposits occupy the marginal part of the intrusion where the crystallization sequence is inverted and the rock is heterogeneous with abundant wall-rock fragments and breccia structures. The Kemi intrusion hosts a world-class chromite deposit (see Table 1). The deposit consists of stratiform, massive or semi-massive chromitite in the ultramafic basal cumulate of a layered igneous complex. The ore deposit is anomalously thick, up to 90 m, and it extends subvertically to at least a depth of 500 m. According to Alapieti et al. (1989), contamination of parental magma caused the crystallizing evolved melt to move to the primary liquidus field of chromite, and the dynamic conditions and tectonically-shaped form of the magma chamber accumulated chromite into a thick pile around the magma vent.

4.1.3. Ni–Cu deposits in Palaeoproterozoic greenstone belts associated with rift-related ultramafic volcanism

Rift-related subaerial to submarine volcanism, ranging in composition from ultramafic to mafic and intermediate, characterizes the Palaeoproterozoic greenstone belts within the Karelian Craton. The majority of the associated mineral deposits are iron formations, but there is also a notable low-grade Ni–Cu and PGE deposit related to the intrusion of the 2.06 Ga Kevitsa layered igneous complex (cf. Fig. 2) in Central Lapland (Mutanen, 1997). The best sections of the disseminated mineralization are related to tectonic remobilization, which upgraded the primary igneous sulphides.

The Pechenga Ni–Cu deposits (Fig. 2) in NW Russia are hosted by both ultramafic, weakly differentiated ferropicritic flows and differentiated gabbro–wehrlite intrusions within the 600 to 1000 m thickness of the Pilgujärvi volcano-sedimentary formation (Hanski, 1992; Melezhik, 1996). The mineralized ultramafic bodies are structurally controlled by the West Rift Graben and related palaeotectonic setting. Two eruptive centres, Kaula and Kierdzhipori, have been identified in the Pilgujärvi formation on the western and eastern sides of the graben, respectively (Melezhik et al., 1994). The ultramafic bodies are further divided into the western, located higher up in the stratigraphy, and eastern group. Green and Melezhik (1999) suggested that the western group is composed of several flows, or portions of one flow, interlayered with sediments, ferropicritic flows, and tuffs, whereas the eastern group is composed of fractionated gabbro–wehrlite sills emplaced close to the base of the sedimentary sequence. According to Green and Melezhik (1999), 226 differentiated ultramafic–mafic bodies can be distinguished: 25 contain Ni–Cu deposits of economic interest, 68 are classified as “Ni–Cu-bearing”, and the remaining 113 are described as “barren”. There are four Ni–Cu sulphide ore types: massive ultramafic-hosted, brecciated, disseminated, and “black shale”-hosted, and predominantly Cu-rich stringer ores (Gorbunov et al., 1985). The ores, which extend up to 400 m away from the gabbro–wehrlite intrusions, are Cu-rich, containing 2% Ni and up to 10% Cu. The ultramafic bodies of Souker, Raisoaivi, Mirona, Kierddzhipori, Pilgujärvi and Onki in the eastern group host disseminated and massive Ni–Cu deposits in differentiated gabbro–wehrlite intrusions, and the Pilgujärvi intrusion is particularly voluminous (500 m thick, 6 km strike length) and well differentiated. The ultramafic mineralized ferropicritic flows of Semiletka, Kammikivi, Kotselvaara and Kaula of the western ore group are thin (<100 m) and contain all ore types (Green and Melezhik, 1999). Hanski (1992) and Melezhik (1996) consider that the host rocks of all the Ni–Cu sulphide mineral deposits originated from the same ferropicritic parental magma (Fig. 3). This magma was derived from the stem of a mantle plume and erupted along a graben structure in the western ultramafic group as a flow on top of the sediments, whereas in the eastern group the magma intruded within a sedimentary pile where the magma cooled slowly and fractionated. Based on the sulphide textures and geochemistry, sulphur isotope studies (Hanski, 1992; Melezhik et al., 1994, 1998; Abzalov and Both, 1997), Re–Os isotope data (Walker et al., 1997) and PGE data (Abzalov and Both, 1997), Barnes et al. (2001) modelled the formation of ores and concluded that the ferropicritic magma reached sulphide saturation prior to ore formation. Sulphur was derived from the unconsolidated sediments and reacted with the ultramafic flow, collected metals and accumulated sulphide melt in structural traps where sulphides were fractionated and separated Mss-rich crystals from Cu-rich residual melt. Finally the breccia sulphides formed during deformation.

4.1.4. Ni–Cu deposits in Svecofennian orogenic mafic–ultramafic intrusions

A number of mafic–ultramafic intrusions were emplaced during the Svecofennian Orogeny at 1.89 to 1.87 Ga. Peltonen (2005) divides them into three Groups (I, II and III), of which the Group I intrusions, derived from hydrous arc-type tholeiitic basalts, were emplaced close to the peak of the Svecofennian Orogeny (at ~1.89 Ga). Group II intrusions are large synvolcanic layered gabbro complexes in the southern Finland arc complex and represent low-pressure crystallization products of relatively juvenile subalkalic tholeiitic basalts, within an oceanic arc. These intrusions have low potential for Ni deposits. Group III intrusions include Ti–Fe–P-rich anorogenic gabbros within the central Finland granitoid region, and host a few Ti–P deposits.

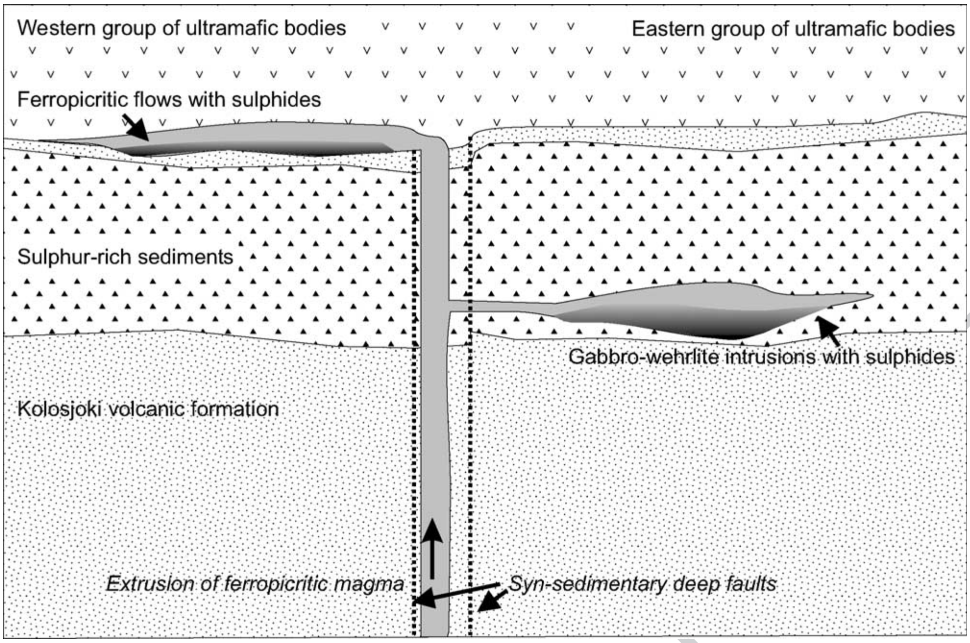


Fig. 3. Depositional model of the formation of the Pechenga Ni–Cu deposits after Melezhik (1996) and Barnes et al. (2001).

The Group I intrusions host several magmatic Ni–Cu sulphide occurrences, and nine of them have been mined in central Finland since the 1960s, producing a total of 0.28 Mt of contained nickel metal and 0.1 Mt copper (Papunen, 1989, 2003; Puustinen et al., 1995). This group is described below in more detail.

There are three main nickel ore belts in the Svecofennian area (Papunen and Gorbunov, 1985), which can be further divided into several subzones (Puustinen et al., 1995). Two of the belts, the Kotalahti belt and the Vammala belt, are located in central Finland, whereas the Lappvattnet belt is located in northern Sweden. The Kotalahti belt roughly parallels the Archaean–Proterozoic boundary and is hosted by sedimentary–volcanic formations between the Keitele microcontinent and the Archaean craton, whereas the Vammala belt parallels the Tampere schist belt along the southern margin of the Keitele microcontinent against the Bergslagen microcontinent (see Section 5 below). The Lappvattnet belt is situated along the southern margin of the Knaften area against the Bothnia microcontinent in Sweden (Fig. 1). Intensely deformed metasediments with black schist interlayers characterize the environment of the intrusions and the wall-rocks are commonly migmatized to neosome-rich schollen migmatites. The neosome intersects the intrusions as random granitic vein networks. The felsic veins metasomatized the ultramafic olivine-bearing parts of the intrusions, forming zoned margins composed of talc, tremolite, and chlorite against the phlogopite-bearing vein fill (Papunen, 1971; Marshall et al., 1995; see also Menard et al., 1999). The zoned margins were not developed around the veins in pyroxenitic and gabbroic parts of the intrusions. Structural analysis implies that the intrusion of mafic magma took place before or at the peak of D_2 deformation and the intrusions were deformed and brecciated during late F_2 folding (Kilpeläinen, 1998). The contact zones against felsic migmatites experienced metasomatic alteration and peridotites were altered to serpentinites and pyroxenites to amphibole–chlorite rocks. In the Vammala belt the metamorphic conditions reached upper amphibolite to lower granulite facies (i.e., 600 to 700 °C and 5 to 6 kbar; Peltonen, 1990) and the cooling from peak conditions was slow, as evident from the subsolidus re-equilibration of olivine and chromite spinel and redistribution of Ca between pyroxenes (Peltonen, 1995b).

Two main types of Group I mafic–ultramafic intrusions host Ni–Cu sulphides: a) differentiated peridotite–gabbro ± diorite bodies (e.g., Kotalahti, Laukunkangas and Telkkälä) and b) weakly differentiated ultramafic olivine-dominated cumulate bodies (e.g., Vammala; Fig. 4, Kylmäkoski, and Hitura; Mäkinen, 1987). The Group Ia intrusions are situated at the craton margin in the eastern part of the Kotalahti belt. Ni–Cu deposits hosted by differentiated intrusions have been mined in Kotalahti, Laukunkangas, Hälvälä, and Tekkälä in SE Finland (Fig. 5; Papunen and Vorma, 1985), of which Kotalahti is the largest deposit that has been mined so far (see Table 1). The host intrusions vary in shape, size and composition, and the rock suite ranges from peridotites to diorites (Makkonen, 1996; Papunen, 2003). In Kotalahti (Fig. 5) and Telkkälä the most ultramafic members are located in the central parts of the intrusive bodies, but in Laukunkangas the ultramafic rock is located at the base of the predominantly mafic body. Locally, gabbroic and dioritic members of the differentiation series display anomalously low Ni content of mafic silicates.

Olivine is the earliest cumulus mineral in all mineralized Group Ia intrusions, followed by orthopyroxene and plagioclase. Chromite is rare or totally absent, but primary magmatic amphiboles are common intercumulus minerals. In gabbroic intrusions orthopyroxene is the dominant cumulus phase, followed by plagioclase. Sulphides are of disseminated and breccia type, partly outside the intrusion as offset orebodies. The parental magmas of the Ia intrusions were tholeiitic basalts, with MgO contents ranging from 8% to 11% and an Al_2O_3/TiO_2 ratio of around 10 (Makkonen, 1996). Their concentrations of compatible elements, notably Ni, are relatively high, but the PGE contents are low (Papunen, 1989). Evidence for contamination includes elevated LREE and Zr and low average $\varepsilon_{Nd}^{(1.9\ Ga)}$ value, +0.7 (Makkonen, 1996).

The Group Ib intrusions and related Ni–Cu deposits are situated in the Vammala belt, although the Group Ia intrusions also occur in the western extension of the belt (e.g., Hyvelä). Group Ib intrusions also characterize the Lappvattnet belt and the Hitura area. The rock types range from dunites to wehrlites and early-crystallized chromian spinel is a common accessory mineral. Early crystallization of clinopyroxene and lack of cumulus plagioclase in Group Ib are the main distinctive features. The sulphides are mainly of disseminated type and accumulated at the basal contact zones of primary olivine ± clinopyroxene cumulates (Fig. 4). The average $\varepsilon_{Nd}^{(1.9\ Ga)}$ value of the type Ib intrusions in the Vammala belt is +1.7, which is lower than the corresponding value of +2.7 obtained from the mafic intrusions and extrusions of the Group II intrusions south of the Vammala belt (Peltonen, 2005).

The geodynamic setting of orogenic Ni–Cu deposits at convergent boundaries of microplates is greatly obscured by severe tectonic deformation, metamorphism and metasomatic overprints. Gaál (1972, 1985) and Puustinen et al. (1995) inferred that the Group Ia intrusions were emplaced into a subvertical D_3 wrench lineament, which is clearly visible along the Kotalahti belt in tectonic and geophysical maps. However, in detail most of the intrusions occur outside the shear zone and a genetic relationship is ambiguous. According to Peltonen (2005), the plutonism occurred over a wide zone due to westward subduction during the final stages of the closure of the basin between the primitive arc complex and the Archaean craton. Synchronous transtensional shear systems, developed at the continental margin, facilitated the ascent of melts locally along subvertical shear zones. During D_3 the zones were reactivated and the intrusions were deformed and brecciated. Early assimilation of felsic sedimentary material and related increase of silica in the ascending magma resulted in the crystallization sequence olivine–orthopyroxene–plagioclase typical of type Ia intrusions (Haughton et al., 1974). Accordingly, low $\varepsilon_{Nd}^{(T)}$ values and elevated LREE and Zr abundances indicate contamination (Makkonen, 1996). The primary parental magma was sulphide unsaturated and the concentrations of chalcophile elements, notably Ni, were high. Contamination with sulphide-bearing sediments turned the magma to sulphide saturation and accumulation of sulphides took place. Mutually intrusive breccias between the members of the differentiation series, lack of compositional layering, breccia type sulphides also as offset orebodies outside the intrusion proper, and complicated shapes of the intrusive complexes are all evidence of polyphase intrusion, where the melt fractionated intermittently in a magma chamber before final emplacement (Papunen, 2003). Syn-magmatic deformation squeezed the melt and early-crystallized silicates from high stress to low stress areas. At this stage the evolved, depleted and barren melt from the

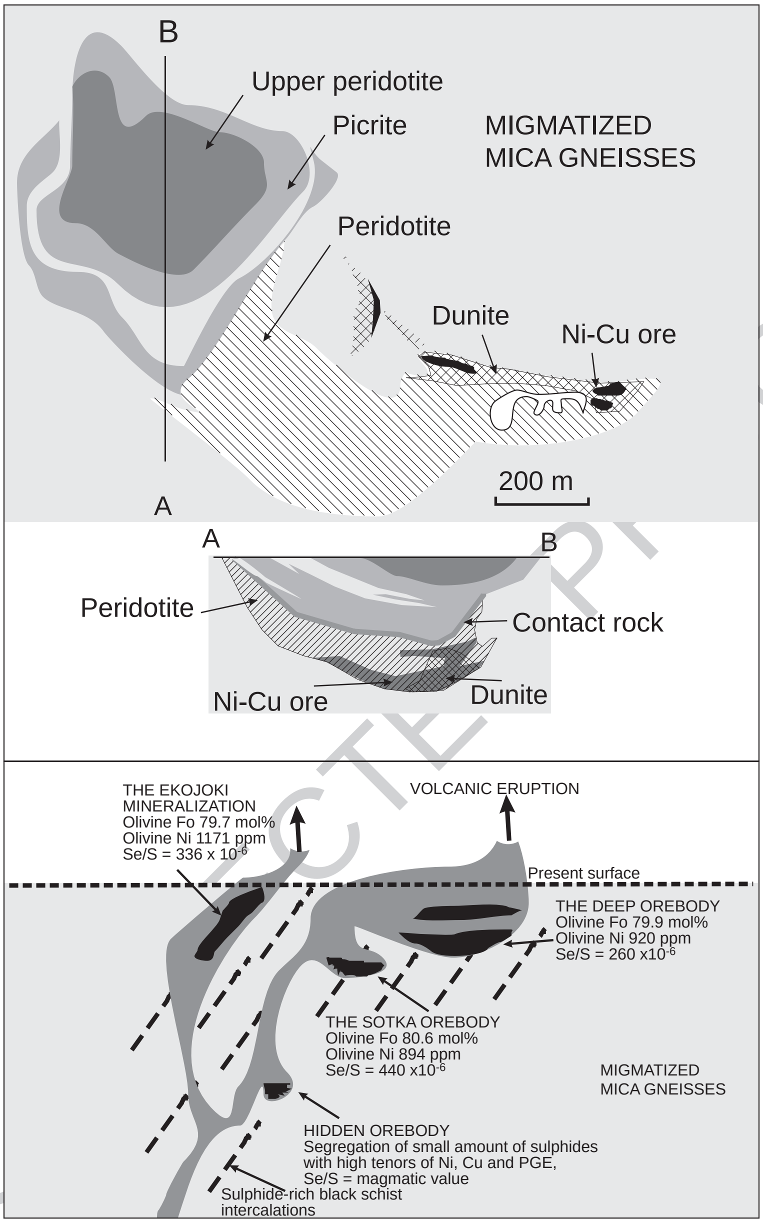


Fig. 4. The Vammala Ni–Cu deposit: the host rock of the Vammala Ni–Cu deposit is the Stormi ultramafic intrusion, which has the shape of a shallow bowl with mineralized dunitic portions at the base. Upper part of the intrusion consists of picrite and "upper peridotite", which are barren and geochemically different from the mineralized lower dunite–peridotite intrusion. The picritic portion represents an ultramafic volcanic formation, which belongs to the supracrustal sequence and pre-dates the intrusion of the fertile dunite–peridotite body. The wall-rocks are migmatized mica gneisses, which also exist as inclusions and tongues inside the picritic layer and locally also between the lower peridotite and picrite. The whole ultramafic body is metamorphosed and intersected by felsic pegmatite dykes. The deposit was mined between 1973 and 1995 and produced 7.4 Mt ore at 0.69% Ni and 0.4% Cu. The Vammala genetic model: according to Peltonen (1995a) the Vammala-type intrusions are feeder channels of basaltic lava flows. The magma became contaminated by wall-rock sulphides and accumulated disseminated and semi-massive sulphides together with early crystallized olivine, spinel and pyroxenes in suitable parts of the intrusion channel. Composition of olivine can be used to follow the evolution of magma in the feeder channel.

upper part of the magma chamber separated and formed Ni-depleted regions of the intrusive complex. In the final phase, the sulphide-bearing ultramafic cumulates and magma from the lower part of chamber intruded to form mineralized ultramafic bodies. Finally the accumulated massive sulphides were squeezed out from the deeper parts of the magma chamber and deposited breccia offset orebodies in a zone of tectonic weakness.

Peltonen (2005) considers that the Group Ib ultramafic cumulate bodies represent former feeder channels for mafic shallow intrusions, sills or volcanic eruptions (Fig. 4). Calculations based on the composition of the most magnesian olivine in the intrusions and Fe/Mg distribution between olivine and melt indicate that the MgO contents of the parental magmas ranged from 8% up to 12% MgO. Common magmatic intercumulus amphibole suggests a hydrous parental magma indicative of arc-type basalt. Accordingly, the mass calculations based on the high Mg content of cumulates compared to the composition of parental magma suggest that the ultramafic cumulates visible at the present erosion level represent only a minor portion of the total igneous complex from which the upper part was eroded away (Peltonen, 1995a). Trace element composition, low Se/S in sulphides, lower than mantle $\varepsilon_{\mathrm{Nd}}^{(1.9\ \mathrm{Ga})}$ values, and common graphite in ultramafic cumulates show that the trace element composition of the parental magma for the Group Ib intrusions was strongly modified during emplacement through the crust. The mantle-derived magma was sulphide unsaturated, but became saturated at the level of crust due to interaction with sulphur derived from the black schists (Peltonen, 1995a). The sulphides accumulated at suitable traps in feeder channels together with early crystallizing olivine and spinel.

The differences between Group Ia and Ib intrusions are due to the more profound contamination of Group Ia intrusions, their fractionation in an intermittent magma chamber, and intrusion as fractionated batches to their present positions in the crust. The Group Ib intrusions represent feeder channels of a more open intrusive–volcanic system, and probably also a more voluminous intrusion of magma into an environment where the crustal contaminant was slightly different and the tectonic evolution more tranquil than for Group Ia intrusions.

4.2. VMS deposits

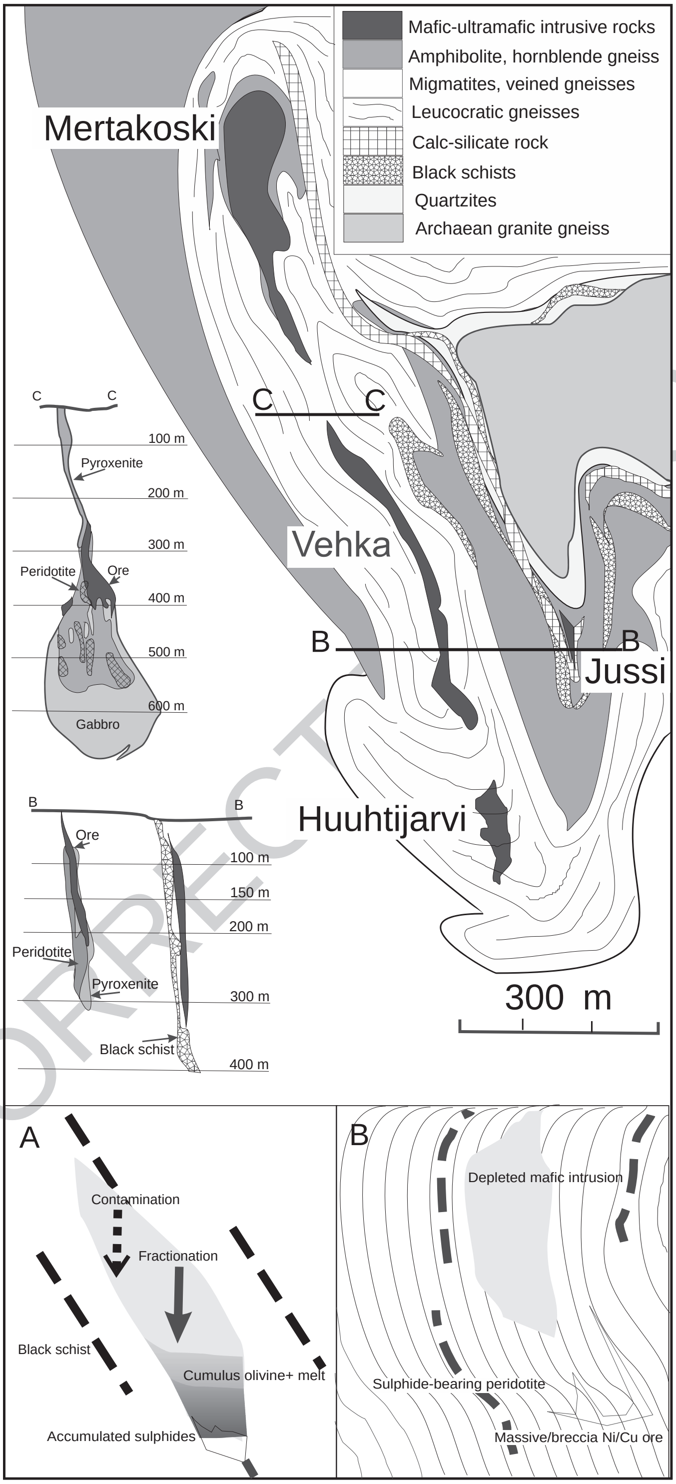
Volcanogenic massive sulphide (VMS) deposits are the ore type that is currently the most exploited in the Fennoscandian Shield. Five deposits are currently mined in the Skellefte district in northern Sweden, one deposit in the Pyhäsalmi area in central Finland and two deposits in the Bergslagen region of south-central Sweden. However, it is unclear whether or not some of the major deposits in the Bergslagen region should be classified as VMS deposits (e.g., Garpenberg and Zinkgruvan), see below. Also an open question is whether or not the Outokumpu deposits (Kontinen, 1998; Sorjonen-Ward et al., 2004), although discussed in this section, really are VMS deposits sensu stricto.

4.2.1. Geodynamic setting

Significant VMS deposits (Table 1) are associated exclusively with Palaeoproterozoic volcanic arc terranes in the Fennoscandian Shield (Box 8-1, Weihed and Eilu, 2005, this volume). With recent improvements in radiogenic dating techniques it has been shown that the host volcanic arcs have different ages and were accreted to the old Karelian craton at different stages during the evolution of the Svecokarelian orogen, between ca. 1.95 and 1.85 Ga. The earliest expressions of Svecokarelian accretionary processes are the Jormua, Outokumpu, and possibly Nuttio ophiolitic sequences, which formed at ca. 1.97 Ga and were emplaced onto the Karelian Craton between 1.94 and 1.89 Ga.

The 1.93 to 1.92 Ga Pyhäsalmi arc was the next to be accreted and contains Kuroko-style VMS deposits, as does the Skellefte volcanic arc that formed 20 to 30 million years later than the Pyhäsalmi arc. The Bergslagen–Uusimaa belt in south-central Sweden and southern Finland contains VMS deposits of a more continental arc affinity that formed roughly at the same time as the Skellefte deposits.

The Palaeoproterozoic arc assemblages also seem in detail to represent slightly different tectonic settings, apart from their age differences (Fig. 6). The Pyhäsalmi bimodal volcanic rocks formed during rifting of a Palaeoproterozoic oceanic island arc, whereas the younger Skellefte volcanic rocks were formed during extension of an immature continental volcanic arc. The Bergslagen province formed by volcanism in


Mafic-ultramafic intrusive rocks
Amphibolite, hornblende gneiss
Migmatites, veined gneisses
Leucocratic gneisses
Calc-silicate rock
Black schists
Quartzites
Archaean granite gneiss
Mertakoski
Vehka
Jussi
Huuhtijarvi
C C
B B
300 m
C C
100 m
Pyroxenite
200 m
300 m
Peridotite
Ore
400 m
500 m
600 m
Gabbro
B B
Ore
100 m
150 m
200 m
Peridotite
Pyroxenite
300 m
Black schist
400 m
A
Contamination
Fractionation
Black schist
Cumulus olivine+ melt
Accumulated sulphides
B
Depleted mafic intrusion
Sulphide-bearing peridotite
Massive/breccia Ni/Cu ore

a continental margin setting and is fundamentally different in the composition of its host rocks and associated metallogeny compared to the Pyhäsalmi and Skellefte provinces. The Vihanti–Pyhäsalmi and Skellefte deposits show many similarities with Kuroko-type deposits, whereas the Outokumpu deposits have been described as "Cyprus-type" deposits (Helovuori, 1979; Ekdahl, 1993; Allen et al., 1996a). Recently, however, the idea of exhalative ore formation for the Outokumpu deposits has been questioned, as the chemical composition of the host rocks indicates that they are subcontinental lithospheric mantle rocks (Kontinen, 1998; Sorjonen-Ward et al., 2004). Sorjonen-Ward et al. (2004) propose that mineralization instead may represent deep levels of subseafloor hydrothermal convection.

The apparent restriction of Palaeoproterozoic VMS mineralization to the Fennoscandian Shield is somewhat enigmatic. The deposits of the Fennoscandian Shield have ages that are similar, within error, to those of the Trans-Hudson and Penokean orogenies in North America (e.g., in the Flin-Flon and Snow Lake areas) and share many other features with these deposits (see Syme et al., 1982). The scarcity of known VMS mineralization in the Archaean of the Fennoscandian Shield can be attributed to two factors: (1) the lack of suitable host sequences of the appropriate age, as explained above, and (2) the unexplored nature of the Archaean areas, especially in Russia.

4.2.2. Timing and relation to regional tectonic evolution of the shield

The *Pyhäsalmi area* is characterized by bimodal volcanic sequences. The mafic volcanic rocks are low-K, island-arc tholeiite metabasalts and basaltic meta-andesites, whereas the felsic volcanic rocks include low-K, transitional to calc-alkaline rhyodacites, rhyolites, and high-silica rhyolites (Rasilainen, 1991; Kousa et al., 1994; Lahtinen, 1994). The local basement to the bimodal volcanic sequence is suggested to comprise a collage of ca. 2.0 to 1.94 Ga volcanic rocks (Lahtinen, 1994). Within the bimodal volcanic sequence, massive sulphide deposits occur either at the transition from mafic to felsic volcanic rocks, or are entirely hosted by felsic pyroclastic rocks (Huhtala, 1979; Mäki, 1986; Rasilainen, 1991). Deformed 1.93 to 1.91 Ga tonalites and trondhjemites are spatially associated with ore-associated bimodal volcanic sequences (Lahtinen, 1994). Isotope and trace element data from these rocks show that they were produced by partial melting of ca. 2.0 Ga primitive island-arc tholeiite basalts (Lahtinen, 1994; Lahtinen and Huhma, 1997). The intrusive rocks are geochemically similar to and of the same age as the ore-associated metarhyolites, and are therefore considered to represent subvolcanic equivalents of the latter (Kousa et al., 1994; Lahtinen, 1994).

In places, the bimodal volcanic sequence is overlain by migmatitic metasedimentary rocks with intercalated calc-silicate and graphite-bearing interlayers. In turn, this sequence is succeeded by a younger (~1.88 Ga) calc-alkaline volcanic suite formed in a mature arc setting (Kousa et al., 1994).

According to Lahtinen (1994), the 1.93 to 1.91 Ga volcanic rocks that host Zn–Cu deposits in the Vihanti–Pyhäsalmi district formed within a rifted primitive island arc. Intra-arc rifting in a juvenile setting is implied by the occurrence of bimodal volcanism, low-K tholeiitic basalts, and minimal hornblende fractionation in the generation of rhyolites. The felsic subvolcanic intrusions have juvenile ε_{Nd} values, indicating an origin by partial melting of a primitive low-K island-arc basalt source (Kousa et al., 1994; Lahtinen, 1994).

Fig. 5. Kotalahti: this is an example of fractionated intrusions, which host Ni–Cu deposits in the Svecofennian area of Finland. The main rock types are peridotite and pyroxenite in the upper part of the subvertical, plate-shaped intrusion and in the southern Huuhtijärvi vertical pipe-shaped body, but gabbros abound in the deep part of the Vehka body. Contacts between different rock types are commonly sharp or gradual over a short distance. The disseminated and breccia ores occur mainly in peridotites and pyroxenites, and gabbros in the deeper part of the intrusion, in particular, are barren and even depleted in nickel. The intrusion is metamorphosed and intersected by felsic dykes. The massive Jussi orebody with high-grade Ni–Cu sulphides exists about 150 m east of the intrusion as a vertical slab in black schist and calc-silicate rock environment. The mine produced 13 Mt of ore at 0.7% Ni and 0.27% Cu. The Kotalahti model: the model of the Kotalahti-type deposits is based on early contamination of mantle-derived melt by sedimentary sulphides and felsic country rocks, which caused sulphide immiscibility and fractionation in a magma chamber where the upper part became depleted in chalcophile elements and sulphides and mafic crystals accumulated at the lower part of the chamber (A). In subsequent orogenic deformation the different parts of the fractionated chamber remobilized and intruded separately to form depleted mafic and intermediate intrusions and sulphide-bearing peridotite and pyroxenite bodies. Sulphides remobilized with the ultramafic portion and could also be intruded as offset ores outside the intrusion proper (B).

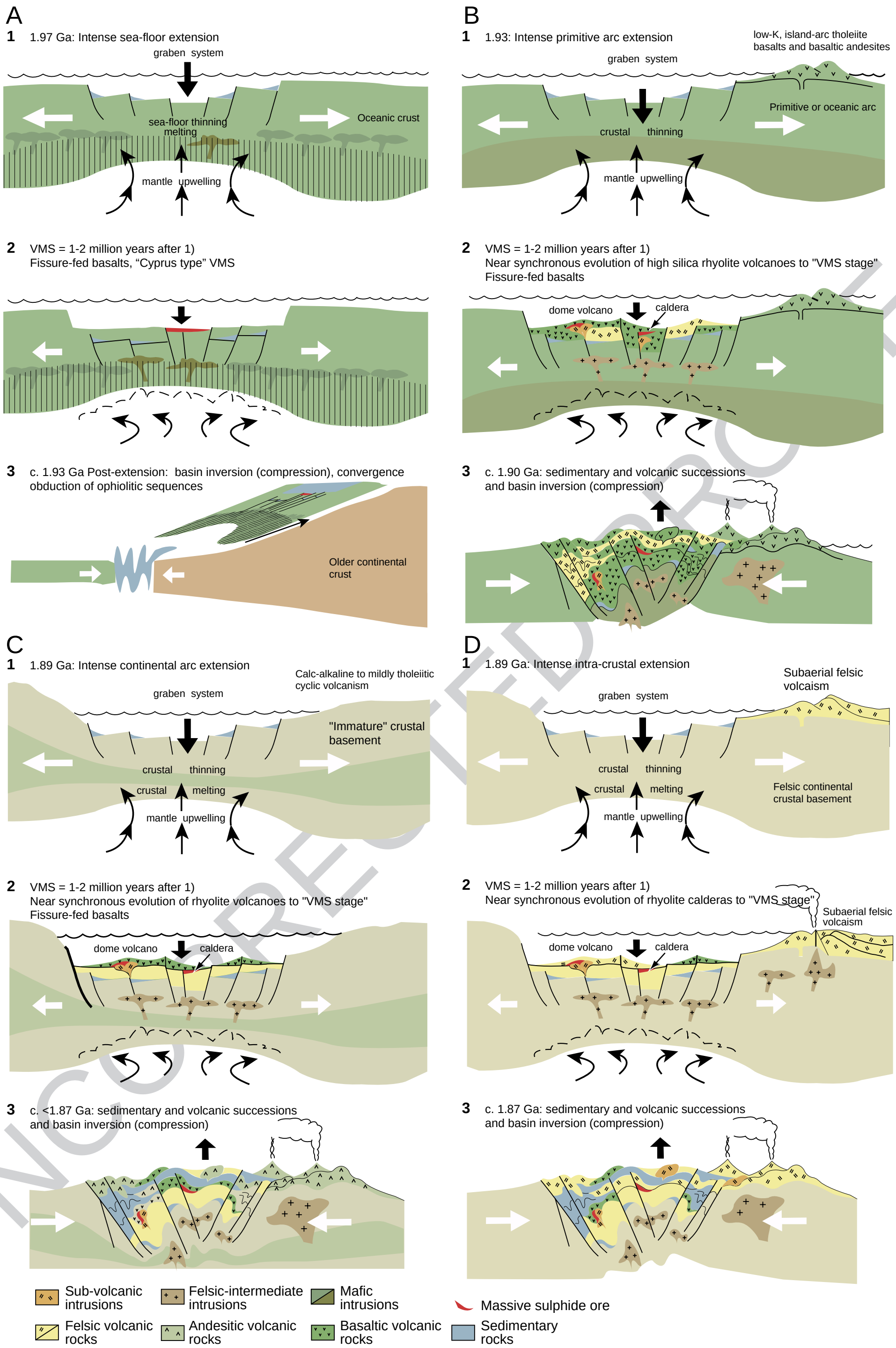


Fig. 6. Key geodynamic features of VMS deposits. Model adapted from Allen et al. (2002). Tectonic setting during formation of (A) ophiolitic VMS deposits (Outokumpu type), (B) primitive arc VMS deposits (Vihanti–Pyhäsalmi type), (C) mature arc or continental margin VMS deposits (Skellefte type) and (D) intracontinental VMS deposits (Bergslagen type).

Roberts (2002) suggested that the basalts and rhyolites in the immediate host sequence to these deposits did not form in a proto-arc setting. Instead, Roberts (2002) interprets the district as a rifted primitive volcanic arc. Roberts (2002) also suggests that an association with tholeiitic basalts and transitional to calc-alkalic rhyolites, including high-silica varieties, broadly fits within the "bimodal mafic type" classification of Barrie and Hannington (1999) and argues that the association of Zn–Cu mineralization within a relatively mature arc assemblage is analogous to the observations of Bailes and Galley (1999) in the Flin-Flon Belt, Canada, where Zn-dominant deposits occur within mature arc assemblages, and most Cu-dominant deposits occur within primitive arc assemblages.

The *Skellefte district* is a 120 × 30 km Palaeoproterozoic volcanic-dominated belt that contains over 80 massive sulphide deposits (Rickard, 1986; Weihed et al., 1992; Allen et al., 1996b). The volcanic stratigraphy is composed of calc-alkaline basalt–andesite–dacite–rhyolite, tholeiitic basalt–andesite–dacite, high-Mg (komatiitic) basalt and subordinate sedimentary rocks, and is intruded by syn- and post-volcanic granitoids (Vivallo and Claesson, 1987; Allen et al., 1996b, 2002). About half the volcanic rocks are rhyolites. Primitive isotopic signatures suggest that magmas were mainly mantle-derived (Billström and Weihed, 1996). The stratigraphy is very complex, laterally variable, and diachronous, and marker horizons are rare. On a regional scale, however, a lower >3 km thick, ore bearing, marine volcanic complex (~1.90 to 1.88 Ga) is overlain by a >4 km thick, mixed sedimentary and volcanic sequence. According to Allen et al. (2002), the dominance of marine depositional environments throughout the lower volcanic complex indicates strong extension and subsidence. The overlying mixed sedimentary and volcanic sequence records rapid uplift, erosion and renewed rifting, and includes medial–distal facies of voluminous, ca. 1.88 to 1.87 Ga, subaerial felsic magmatism. Allen et al. (2002) point out that the stratigraphic architecture, range of volcanic compositions and abundance of rhyolites indicate that the district is a remnant of a strongly extensional intra-arc region that developed on continental or mature arc crust in contrast to the primitive volcanic arc setting for the Vihanti–Pyhäsalmi belt. Most VMS ores occur in near-vent facies associations at the top of local volcanic cycles, especially rhyolitic dome–tuff cone volcanoes (Allen et al., 1996b). Regionally, these VMS ores occur on at least two stratigraphic levels, most commonly near the upper contact of the lower volcanic complex.

The *Bergslagen district*, located in south-central Sweden, and the extension into SW Finland, the Uusimaa belt, are 1.90 to 1.87 Ga in age and characterized as a felsic magmatic region. The volcanic succession is 1.5 km thick and overlies turbiditic metasediments in the east, and is over 7 km thick with no exposed base in the west (Lundström, 1987; Allen et al., 1996a). The basement is interpreted to be older, unexposed, continental crust.

In contrast to the Vihanti–Pyhäsalmi and Skellefte districts, Bergslagen contains a diverse range of ore deposits, including banded iron formations, magnetite skarns, manganiferous skarns and marble-hosted iron ores, apatite–magnetite iron ores, stratiform and stratabound Zn–Pb–Ag–(Cu–Au) sulphide ores, and W skarns (Hedström et al., 1989; Sundblad, 1994; Allen et al., 1996b).

Although traditionally described as massive sulphide ores divided into the Åmmeberg and Falun types (Sundblad, 1994), and historically among the first ore deposits in Sweden described as exhalative (Koark, 1962), their classification as typical VMS deposits can be questioned. The ore deposits conform, with some exceptions, to a regional ore stratigraphy. Based on physical characteristics, the main base-metal deposits span a range between two end-member types (Allen et al., 1996b). The first type comprises sheet-like, bedded, stratiform Zn–Pb–Ag–rich, Fe and Cu sulphide-poor deposits such as Zinkgruvan. These deposits are hosted by rhyolitic ash–siltstones with marble, skarn, and siliceous chemical sediment beds, and have intense footwall potassic alteration, silicification, and subordinate Mg-rich alteration (Hedström et al., 1989). Allen et al. (1996b) named these deposit "stratiform ash siltstone"-hosted Zn–Pb–Ag sulphide deposits (SAS type) and included some deposits of the Åmmeberg type (see Sundblad, 1994 and references therein). The second end-member type according to Allen et al. (1996b) includes irregular multi-lens and pod-like, strata-bound, massive and disseminated Zn–Pb–Ag–Cu mineralization such as Garpenberg, and more massive pyritic Cu–Zn–Pb–Ag–Au mineralization such as Falun. These deposits straddle marble beds within felsic metavolcanic rocks, are closely associated with Mg-rich tremolite–diopside skarn and dolomite zones within the marbles, and have intense footwall Mg-rich alteration (phlogopite–biotite–talc–almandine–cordierite–amphibole–quartz schists). These deposits are designated "strata-bound", volcanic-associated, limestone–skarn Zn–Pb–Ag–Cu–Au sulphide deposits, SVALS type, by Allen et al. (1996b) and correspond approximately to the Falun type (see Sundblad, 1994 and references therein).

Allen et al. (1996b) suggested that most of the sulphide ores at Garpenberg and Stollberg formed as synvolcanic stratabound subsea-floor replacements and that the thin sheet-like superficially stratiform massive sulphide layers are tectonically mechanically remobilized ore. They also suggested that much of the ore deposition occurred by reaction of the ascending hydrothermal solutions with limestones below the sea floor. This distinguishes them from VMS deposits sensu stricto. Also the Zinkgruvan type stratiform Zn–Pb–Ag sulphide deposits are difficult to classify as VMS deposits sensu stricto (Allen et al., 1996b). They have some similarities to VMS deposits and to some stratiform sediment-hosted Pb–Zn deposits. However, they appear most similar to the Broken Hill-type deposits; sheetlike ore lenses, Zn–Pb–Ag-rich and Cu-poor, mainly pyrite-poor composition, stratigraphic succession grading from coarse-grained metavolcanic and metasedimentary rocks below the ores to pelites above, proximity to banded iron-formations, association with limestone, calc-silicate rocks, mafic sills (amphibolites) and unusual chemical sedimentary rocks such as garnet quartzite.

The volcanic succession in this district is dominated by calc-alkaline rhyolites with minor calc-alkaline dacite and andesite, and chemically unrelated, probably tholeiitic basalts (Allen et al., 1996a, 2002). The stratigraphy is composed of: (1) a lower, 1 to 5 km thick, poorly stratified felsic complex dominated by a proximal–medial facies of interfingering and overlapping large caldera volcanoes, and minor interbedded marble, (2) a middle, 0.5 to 2.5 km thick, well-stratified interval dominated by medial–distal juvenile volcaniclastic facies and marble sheets, and (3) an upper, >3 km thick, post-volcanic meta-argillite–turbidite sequence (Allen et al., 1996a). Depositional environments fluctuated mainly between shallow and moderately deep subaqueous throughout the lower and middle stratigraphic intervals, and became consistently deep subaqueous in the upper interval (Allen et al., 1996a).

The supracrustal succession has been intruded by syn- and post-volcanic granitoids. The stratigraphy reflects an evolution from intense magmatism, thermal doming, and crustal extension, followed by waning extension, waning volcanism, and thermal subsidence, then reversal from extension to compressional deformation and metamorphism (Allen et al., 2002). The region is interpreted as an extensional intra-continental or continental margin back-arc, region (Allen et al., 1996a).

4.3. Orogenic gold

Orogenic gold deposits, following the terminology of Groves et al. (1998), are present in both Archaean and Proterozoic units of the Fennoscandian Shield (Fig. 2). Some gold-only deposits have been described with alternative genetic models; for example, the Enåsen (Hallberg, 1994) and Kutemajärvi (Poutiainen and Grönholm, 1996) have been interpreted as metamorphosed epithermal deposits. The largest gold deposit in the Fennoscandian Shield, the Boliden deposit, has also been described as a hybrid epithermal VMS deposit (Bergman Weihed et al., 1996). In the northernmost part of the Fennoscandian Shield Cu–Au deposits such as Bidjovagge (Ettner et al., 1994), Pahtohavare (Lindblom et al., 1996) and Saattopora (Grönholm, 1999) have been described as orogenic gold deposits, but the high content of Cu makes these deposit more akin to IOCG deposits (see Weihed, 2001).

Although the currently economic deposits are strongly concentrated in the Palaeoproterozoic domains (cf. Sundblad, 2003), tens of occurrences have also been identified in Archaean areas that have been explored for gold by modern methods (Eilu et al., 2003). The apparent scarcity of Archaean economic deposits is probably due to the fact that little exploration for gold has been performed in the Russian part of the shield and that exploration in the Finnish part is relatively recent.

Age data on orogenic gold mineralizing events are scarce, but it is possible to constrain three major periods of mineralization; 2.72 to 2.67, 1.90 to 1.86 and 1.85 to 1.79 Ga (Box 8-2, Eilu and Weihed, 2005, this volume), excluding a few minor younger events (Luukkonen, 1992; Sorjonen-Ward, 1993; Mänttäri, 1995; Bark and Weihed, 2003; Eilu et al., 2003; Weihed et al., 2003). The age data appear to define a rough zonation from NE to SW, which seems to be related to the south-westward growth of the Fennoscandian Shield with time.

4.3.1. Geodynamic setting

The periods of orogenic mineralization in the Fennoscandian Shield fit into two of the main global periods of orogenic gold mineralization during the Precambrian, at ca. 2.7 to 2.6 and ca. 1.9 to 1.8 Ga, which correlate with the major episodes of juvenile continental growth discussed above (Stein and Hofmann, 1994; Goldfarb et al., 2001). More specifically, the periods of orogenic gold mineralization can be correlated with the main compressional to transpressional events, with peak regional metamorphism and the latest main stage of deformation during major orogenies in the shield (Luukkonen, 1992; Sorjonen-Ward, 1993; Lahtinen et al., 2003; Eilu et al., 2003; Sundblad, 2003; Weihed et al., 2003):

(i) The 2.72 to 2.67 Ga stage coincides with the global period of accelerated crustal growth near the end of the Archaean.

(ii) The 1.90 to 1.86 Ga stage relates to microcontinent accretion (see Section 5) that formed part of a second peak of crustal growth. For the Archaean units, this includes the collision of the Karelian craton with the Kola craton in the north and Norrbotten craton in the west, as discussed below. Simultaneously, to the SW of the Archaean units, the Palaeoproterozoic Keitele microcontinent collided with the Karelian craton (Lapland–Savo Orogeny), eastward subduction under the Norrbotten craton started at 1.89 to 1.88 Ga, the Bergslagen microcontinent started to accrete (Fennian Orogeny), and finally, in the south, the Bothnia and Bergslagen microcontinents amalgamated at 1.87 to 1.86 Ga. The stage resulted in the formation of a large continental plate, the Fennoscandia Plate.

(iii) The 1.85 to 1.79 Ga stage involved collision of the Fennoscandia continental plate with Sarmatia in the SE and Amazonia in the west (Svecobaltic and Nordic orogenies, respectively).

4.3.2. Timing and relation to regional tectonic evolution of the shield

Most of the greenstone belts in the Archaean Karelian craton formed after the earliest well-recorded magmatic and metamorphic event at 2.84 Ga and were deformed and intruded by tonalitic to granitic magmas between 2.76 and 2.70 Ga (O'Brien et al., 1993; Vaasjoki et al., 1993, 1999). In the Ilomantsi greenstone belt in the easternmost part of Finland, there is textural and structural evidence that gold mineralization slightly preceded the peak of the regional metamorphism (Sorjonen-Ward, 1993; Sorjonen-Ward and Luukkonen, 2005). Large areas of the Archaean domain were reheated by burial beneath a sequence of nappes in the foreland of the Palaeoproterozoic orogenies around 1.9 Ga. Despite also indications of Proterozoic fluid activity in the region (e.g., Poutiainen and Partamies, 2003), there is no evidence for a distinct Proterozoic gold mineralization event in the area (Kontinen et al., 1992; O'Brien et al., 1993). The lack of Proterozoic gold in the Archaean areas could simply be due to the style of the Proterozoic tectonic processes in the area: a system of subhorizontal nappes typically does not contain deep, transcrustal structures necessary to tap the fluid and metal sources and focus enough mineralizing fluids for orogenic gold deposits to form (Goldfarb et al., 2001).

For the Palaeoproterozoic greenstone belts in northern Finland, most of the radiometric age dating and structural evidence suggest mineralization between ca. 1.85 and 1.79 Ga (Sorjonen-Ward et al., 1992; Mänttäri, 1995). On the other hand, structural evidence and a few radiometric ages of the wall-rocks from some of the gold occurrences point towards mineralization between ca. 1.90 and 1.88 Ga (Mänttäri, 1995). In any case, most of the available evidence supports the occurrence of orogenic mineralization in northern Finland during the peak deformation stage of the collisional epoch at 1.85 to 1.79 Ga. Two different styles of epigenetic gold occurrence in the region are displayed in Figs. 5 and 6, respectively. The Suurikuusikko deposit (Fig. 7) exhibits features that are typical of shear zone-hosted orogenic gold deposits, whereas the Saattopora Cu–Au deposit (Fig. 8) exhibits many features that are typical of orogenic gold lode deposits, but also includes some features characteristic of IOCG deposits (see Box 8-2, Eilu and Weihed, 2005, this volume).

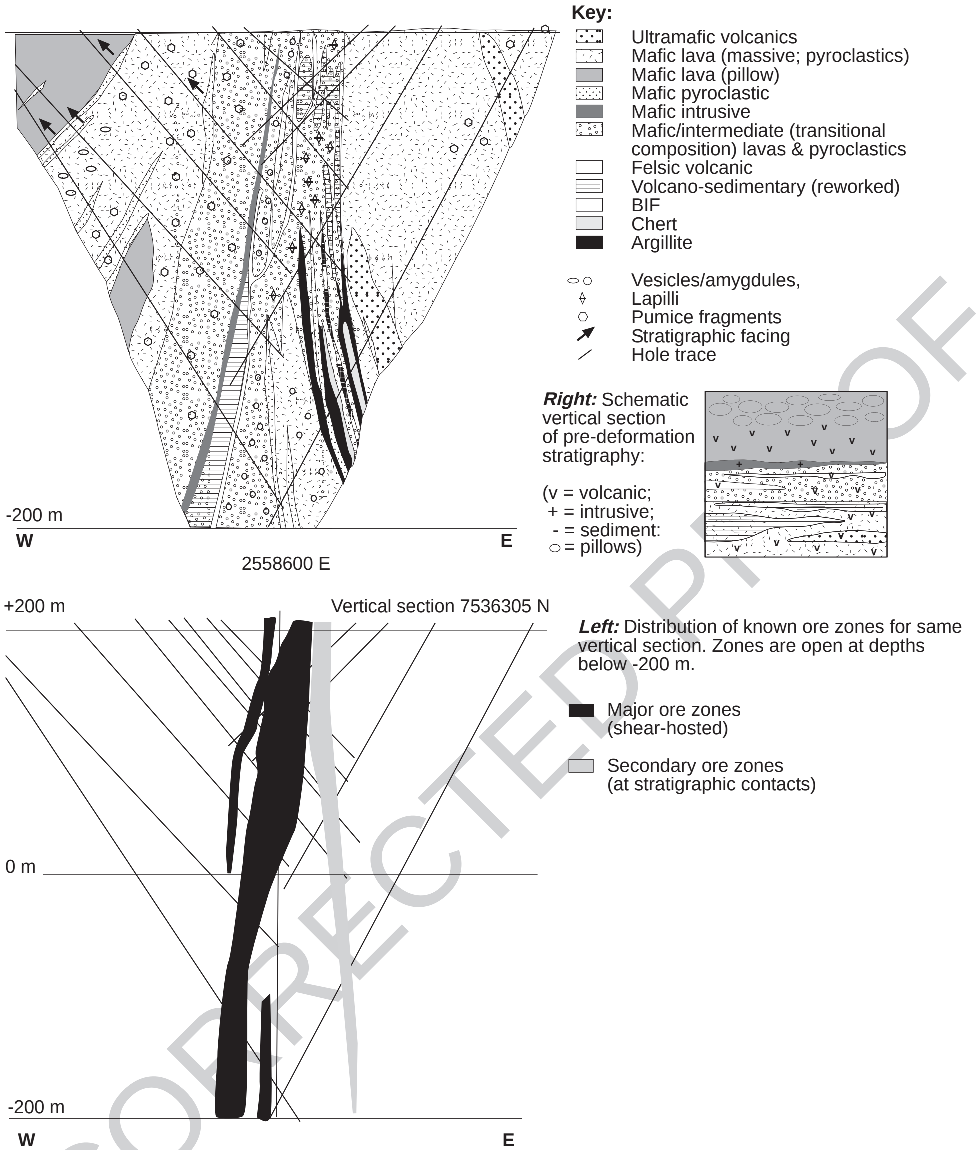


Fig. 7. Simplified vertical cross-section showing the host rocks of the Suurikuusikko gold deposit. Major ore zones typically occur within the central part of the host stratigraphy (Mafic/intermediate (transitional composition) lavas and pyroclastics). These volcanic to volcanogenic sedimentary rocks have a geochemical composition transitional between mafic and intermediate. Primary volcanoclastic textures and thin felsic volcanic layers are also more abundant in this part of the local stratigraphy. Secondary ore zones occur on stratigraphic contacts throughout the entire host rock package. All parts of the local stratigraphy appear to have a geochemical affinity with the 2.012 Ga Vesmajärvi Formation (as defined in Lehtonen et al., 1998). In the structure hosting Suurikuusikko (Kiistala Shear Zone, not shown), the most intense shearing associated (spatially and temporally) with mineralization was focused within the central units of the local stratigraphy.

In the Svecofennian domain of Finland, orogenic mineralization post-dates the earliest deformation. However, some of the occurrences were recrystallized and deformed to varying degrees after mineralization between 1.84 and 1.80 Ga (e.g., Kontoniemi, 1998). This indicates that the timing of mineralization was either ca. 1.90 to 1.86 or 1.84 to 1.80 Ga.

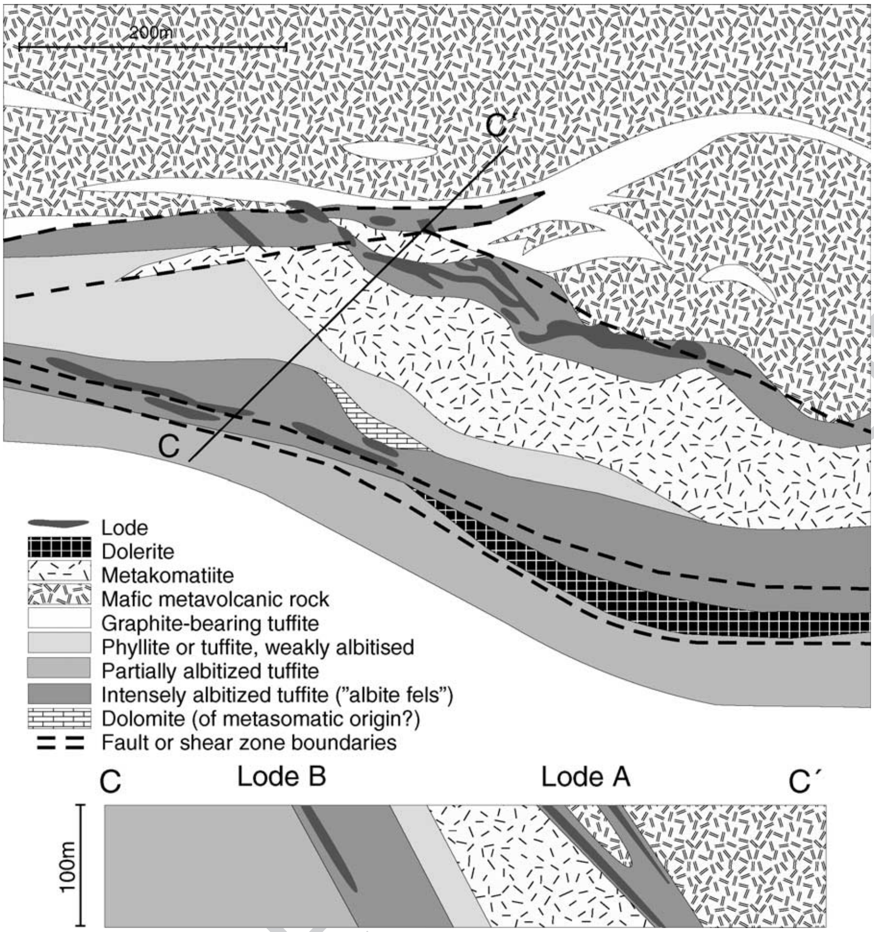


Fig. 8. Plan view and cross-section of the Saattopora deposit. Geology is based on Korvuo (1997); stratigraphy on Lehtonen et al. (1998).

Recent age data on orogenic gold deposits in the western part of the shield indicate that many of the deposits, at least in part, formed slightly after peak metamorphism, as late as ca. 1.79 to 1.77 Ga (Bark and Weihed, 2003; Weihed et al., 2003). Later remobilization may also have occurred and there is abundant evidence for young, post-1.78 Ga, hydrothermal activity in northernmost Sweden (see Section 4.4).

4.3.3. Controls on mineralization

Orogenic gold deposits in the Fennoscandian Shield are structurally controlled. All occurrences are in second- to lower-order shear or fault zones, at their intersections, or at the intersections between antiforms and crosscutting fault and shear zones, all indicative of a compressional to transpressional regime at the time of mineralization, as exemplified in the Palaeoproterozoic Central Lapland and Kuusamo greenstone belts. In Central Lapland, the W- to NW-trending Sirkka Shear Zone, a major crustal-scale structure, is located close to most of the gold occurrences, which in many cases are in lower-order shear or fault zones branching from the main structure and at localities where younger faults cut across the latter (Eilu et al., 2003). In the Kuusamo greenstone belt, nearly all occurrences are located within two NE- to NW-trending antiforms located in the central part of the greenstone belt, at or near intersections between antiforms and crosscutting faults (Pankka, 1992; Vanhanen, 2001). Similarly, deposits in central and southern Finland display a close association with major faults, but are hosted by second- and third-order structures branching from the large faults. In the Björkdal deposit in the Skellefte district in Sweden, the gold is related to third-order structures. There, the gold precipitated in or adjacent to conjugate quartz veins in the footwall of a major duplex structure that formed during roughly E–W crustal shortening at ca. 1.80 Ga (Weihed et al., 2003).

On a local scale, favoured sites for gold mineralization are (1) pre-gold albitized units, (2) competent units enveloped by softer rocks, and (3) contact zones between chemically-reactive rocks with a significant competency difference. There are examples of each of these in Lapland: in the Saattopora mine (Fig. 8), the pre-gold albitization significantly increased the competency of the host tuffites and phyllites, providing pathways for the mineralizing fluids where these units were brecciated (Grönholm, 1999). Grönholm (1999) further suggests that precipitation of gold at Saattopora was induced by reduction–oxidation reactions between the mineralizing fluid and graphite in the albitized wallrock. Increased competency due to pre-gold albitization has also been suggested for most of the Kuusamo deposits (Pankka, 1992; Vanhanen, 2001). A large number of occurrences in Central Lapland are located in contact zones between chemically reactive rocks with a significant competency difference, for example between intensely carbonated and more competent metakomatiites and more plastic graphitic tuffite or phyllite (Eilu et al., 2003). At Björkdal, the gold mineralization appears to be related to the competency contrast between the host intermediate intrusion and the surrounding supracrustal rocks, including a mylonitized marble unit (Weihed et al., 2003). In the Fäbodliden deposit, south of the Skellefte district in Sweden, the gold is spatially related to graphite-bearing pelitic metasedimentary rocks, possibly implying a strong redox control on gold precipitation.

4.4. IOCG deposits

The northern region of the Fennoscandian Shield, including parts of Finland, Norway and Sweden, is an economically important metallogenic province dominated by Fe-oxide and Cu ± Au ores. Based on the style of Fe and Au–Cu mineralization and the extensive albite and scapolite alteration, the region has been regarded as a typical IOCG province (e.g., Martinsson, 2001; Williams et al., 2003). These deposits are quite variable in character and include four major types: stratiform Cu ± Zn deposits, skarn-rich iron deposits, Kiruna type Fe-oxide deposits (apatite iron ores), and epigenetic and porphyry style Cu ± Au and Au deposits. In strictly genetic terms only some of these deposits may be classified as typical IOCG deposits whereas others only share a few characteristic features with this rather loosely defined ore type (see Hitzman et al., 1992; Hitzman, 2000). In addition, orogenic gold deposits and subeconomic banded iron formations (BIF) also occur in the region. Some examples of potential IOCG occurrences and their characteristic features are listed in Table 2.

Economically, the most important deposit type for the region has been the apatite–iron ores, presently

Table 2
Examples of deposits potentially belonging to the IOCG category in northern Finland and Sweden

Ore type	Occurrence	Character	Main ore minerals	Alteration	Hosting sequence	Approx dep. age (Ga)
Fe-oxide–Fe-sulphide–Cu	Laurinoja, Kuervitikko	Massive lenses	Mt, Py, Po, Cp	Di, Bi, Ab, Am	SG	1.86 to 1.76
Fe-oxide ± Co–Cu–Au	Vähäjoki	Breccia	Mt, Py, Cp, Co	Am, Bi	TF	1.9 to 1.8?
Fe-oxide	Mertainen	Breccia	Mt, (Ht)	Ab, Am, Sc	KiG	1.88
Fe-oxide–apatite ± REE	Kiirunavaara, Rektorn	Massive lenses	Mt, Ht	Am, Ab, Bi, Kf	KiG KiG	1.88
Fe-oxide–apatite–Cu ± Au	Tjårrojåkka Nautanen	Disseminated, veins	Mt, Cp, Py, Bo	Kf, Sc, Bi, To	PoG PoG	1.77?
Fe-oxide–Cu ± Co ± Au	Kiskamavaara	Disseminated, breccia	Mt, (Ht), Py, Cp	Kf, Bi, Sc	PoG PoG	1.86?
Cu ± Au ± Fe-oxide	Aitik, Pikkujärvi Pahtohavare	Disseminated, veins	Cp, Py, Po, Bo, Cc, Mt	Ab, Sc, Bi, Kf, To	KiG KiG KGG	1.89 1.88 1.88 to 1.86?
Cu–Au	Lieteksavo Ferrum	Vein, disseminated	Bo, Cp	Sc, To, Bi	KiG	1.76

Mineral abbreviations: Ab = albite, Am = amphibole, Bi = biotite, Bo = bornite, Cc = chalcocite, Co = cobaltite, Cp = chalcopyrite, Di = diopside, Ht = hematite, Kf = K feldspar, Mt = magnetite, Po = pyrrhotite, Py = pyrite, Sc = scapolite, SG = Savukoski Group, To = tourmaline.
Host sequence abbreviations: KiG = Kiirunavaara Group, KGG = Kiruna Greenstone Group, PoG = Porphyrite Group, TF = Tikanmaa Formation.

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. Copper and gold have been mined on a large scale in Sweden (Aitik, Viscaria, Pahtohavare), Finland (Saattopora, Pahtavaara) and Norway (Bidjovagge). All of the sulphide deposits are hosted by Palaeoproterozoic greenstones and are small to medium sized except for Aitik, which occurs in Svecofennian volcaniclastic rocks and is a world-class deposit with a total tonnage >1000 Mt and an annual production of 18 Mt.

4.4.1. Geodynamic setting

Fe-oxide and Cu ± Au ores in Norrbotten in northernmost Sweden formed during the evolution of a Palaeoproterozoic continental margin arc from ca. 1.89 to 1.75 Ga (e.g., Juhlin et al., 2002). The deposits occur where the margin developed above Archaean continental crust and are hosted both by juvenile rocks and by older Palaeoproterozoic volcano-sedimentary sequences formed during rifting of the Archaean craton (Martinsson and Weihed, 1999). The rift-related, 2.2 to 2.0 Ga Karelian greenstones comprise mafic and ultramafic volcanic rocks, graphitic schists and sedimentary carbonate rocks. The lower part of the sequence contains clastic sedimentary and inferred, evaporitic units (Martinsson, 1997; Vanhanen, 2001). At ca. 1.9 Ga, subduction of oceanic crust at the SW margin of the Karelian craton involved both strong reworking of older crust and the formation of juvenile crust by accretion of several volcanic arc complexes with the cratonic nucleus (see Section 4.2). This ca. 1.90 to 1.88 Ga magmatism is represented by the calc-alkaline and andesite-dominated volcanic successions and the co-magmatic, intrusive Haparanda Suite within the NW part of the craton. In the Kiruna area, these rocks are succeeded by the bimodal Kiirunavaara Group volcanic rocks and the coeval and chemically similar Perthite Monzonite Suite (Bergman et al., 2001). This 1.88 to 1.86 Ga magmatic activity has a more alkaline character that suggests an extensional intraplate setting.

The magmatism related to the Svecokarelian orogen after ca. 1.86 Ga is mainly of S-type (1.81 to 1.78 Ga Lina suite intrusions) derived from anatectic melts in the middle crust. In the western part of the shield, extensive I- to A-type magmatism formed a roughly N–S-trending belt of batholiths (the Trans-scandinavian Igneous Belt) coeval with the S-type magmatism, possibly as a result of eastward subduction (Weihed et al., 2002), followed by collision between Fennoscandia and Amazonia (see Section 5). Although the metamorphic history of the northern part of the shield is not well-constrained, in time or in space, it appears that metamorphic peaks more or less overlap with the main magmatic events at about 1.88 and 1.80 Ga. A model for the tectonic setting of the IOCG deposits is presented in Fig. 9.

4.4.2. General features and controls on mineralization

Skarn-rich and lens- to irregularly-shaped Fe-oxide occurrences consisting of magnetite, and Mg and Ca–Mg silicates are common within the greenstones of northern Sweden and the westernmost part of northern Finland. Some of the deposits are spatially associated with oxide- and silicate-facies BIF. These skarn or skarn-like iron deposits occur in association with tuffite, graphitic schist, and dolomitic to calcitic marble, and are located mainly in the upper parts of the greenstone sequences. These occurrences have previously been suggested to be metamorphic expressions of originally syngenetic exhalative deposits (Frietsch, 1977; Bergman et al., 2001) or intrusion-related skarn deposits (Hiltunen, 1982). Several skarn-rich iron deposits have been mined in the Kolari area in NW Finland and in the Misi region in southern Finnish Lapland (Nuutilainen, 1968; Hiltunen, 1982; Niiranen et al., 2003). In addition, significant amounts of Cu and Au have been recovered from the Laurinoja orebody in the Kolari area (Hiltunen, 1982; Geological Survey of Finland, 2004).

Kiruna is the type area for apatite–iron ores (Kiruna type Fe-oxide ores) with the Kiirunavaara deposit as the largest and best known example (Table 1). In all, about 40 apatite–iron deposits are known from the northern Norrbotten ore province in northernmost Sweden. This type of deposits is mainly spatially restricted to areas occupied by the Kiirunavaara Group and very few occurrences exist outside the Kiruna–Gällivare area. The apatite–iron ores exhibit a considerable variation in host rock composition and relationship, alteration, P content, and associated minor components. It is possible to distinguish two subgroups: breccia deposits and stratiform–stratabound deposits. A third and less distinct group has

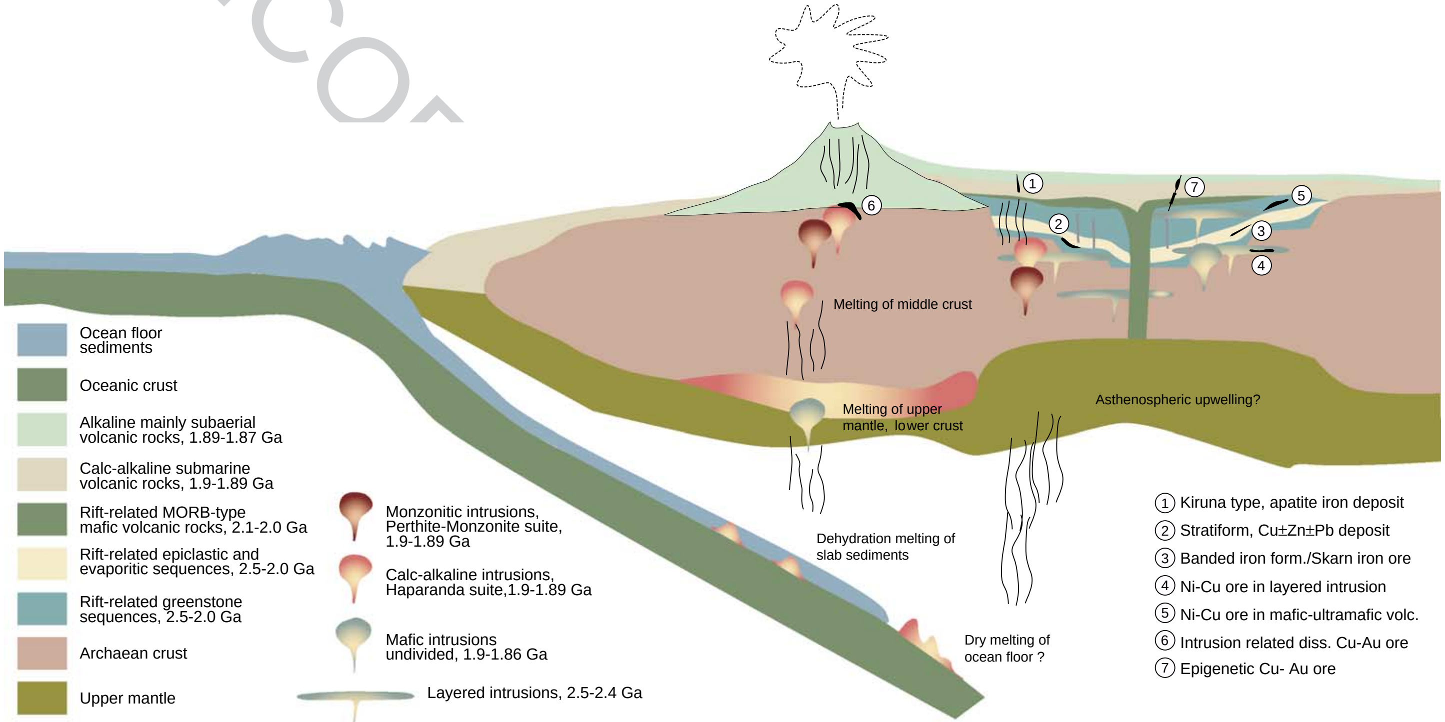


Fig. 9. Key geodynamic features of IOCG deposits. The generalized tectonic section illustrates the relationship between host rocks, tectonic setting and IOCG deposits and other greenstone-related ore types. The section is not meant to illustrate specific temporal relationships, but is a generalization over 500 million years of basically intracontinental to continental margin evolution of the rifted Karelian craton between 2.45 and 1.85 Ga.

features similar to both of the other two groups (Bergman et al., 2001). Breccia-type apatite–iron ores (e.g., Mertainen) are mainly associated with intermediate to mafic volcanic rocks, in a stratigraphically low position of the Svecofennian Kiirunavaara Group or within the underlying Porphyrite Group. The stratiform–stratabound type (e.g., Nukutus, Henry, Rektorn, Lappmalmen and Ekströmsberg) comprises hematite-dominated lenses at stratigraphically high positions within the Kiirunavaara Group. The intermediate types of apatite–iron ore (e.g., Kiirunavaara and Tjårrojåkka) are dominantly stratabound in character, but have breccia ores developed along the wall-rock contacts (Box 8-3, Edfelt and Martinsson, 2005, this volume). Magnetite is the dominant, or the sole Fe-oxide.

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. Suggested genetic models include sedimentary, hydrothermal or magmatic processes (Parák, 1975; Hitzman et al., 1992; Nyström and Henriquez, 1994). Most features of the ores are compatible with either a magmatic intrusive origin or a hydrothermal origin. Probably both magmatic and hydrothermal processes were involved, explaining the large variation in mineralization style recognized within and between individual deposits. Most of the massive deposits are suggested to have a mainly magmatic origin with minor overprinting hydrothermal phases altering the wall-rocks and forming the veins. Some deposits (e.g., Tjårrojåkka) may represent transitional forms between a magmatic and hydrothermal origin similar to that at Lightning Creek in the Cloncurry area, Queensland, Australia (Perring et al., 2000).

Sulphides are mostly rare constituents in the apatite–iron ores and occur disseminated or in veinlets. Significant Cu mineralization spatially associated with apatite ores occurs only in a few places (e.g., Tjårrojåkka and Gruvberget). A genetic relationship between Cu and Fe-oxide mineralization has not been proved, but is probable at Tjårrojåkka (Edfelt and Martinsson, 2004). At Gruvberget, the relationship might be more of a coincidence with the Cu occurrence representing a later separate event and the iron ore only acting as a chemical–structural trap (Lindskog, 2001). U–Pb titanite ages indicate that Cu mineralization at Tjårrojåkka and Gruvberget is ca. 1.8 Ga in age (Billström and Martinsson, 2000), which is significantly younger than the suggested ca. 1.9 Ga emplacement age for apatite–iron ores in the Kiruna area.

Epigenetic Cu ± Au deposits form a heterogeneous group with extensive variation in the style of mineralization, metal association and host rock. Most deposits are hosted by tuffitic units of the Karelian greenstones (e.g., in Central Lapland and Kuusamo in Finland) and mafic to intermediate volcanic rocks within the Svecofennian porphyries (e.g., in Porphyrite and Kiirunavaara Groups in Sweden). Some of them display close genetic links (e.g., Aitik), supported by stable isotope and fluid inclusion evidence (Yngström et al., 1986; Wanhainen et al., 2003), and/or spatial relation to intrusive rocks of the Haparanda and Perthite Monzonite suites, varying in composition from monzodiorite to granite. The deposits with a close genetic link to intrusive rocks are regarded by many authors as porphyry style deposits (see Weihed, 2001; Wanhainen et al., 2003). Magnetite is a common minor component in many occurrences and locally they occur adjacent to major magnetite deposits. A close spatial relationship with regional shear zones is common, with second- to fourth-order structures controlling the location of an occurrence (Eilu et al., in press). In addition to structural traps, chemical traps may also be important with redox reactions involving an originally high graphite or magnetite content of the host rock to trigger sulphide precipitation. In addition to Cu, several occurrences also contain Co and/or Au in economic to subeconomic amounts. Other elements that may be significantly enriched include LREE, Ba, U, and Mo (e.g., Martinsson, 2001; Vanhanen, 2001).

Highly saline fluid inclusions with 30 to 45 eq. wt.% NaCl and depositional temperatures of 500 to 300 °C are recorded for the epigenetic Cu ± Au deposits in the region (Ettner et al., 1993; Lindblom et al., 1996; Broman and Martinsson, 2000; Wanhainen et al., 2003; Williams et al., 2003; Edfelt et al., 2004; Niiranen pers. comm. 2004).

Extensive hydrothermal alteration systems producing, for example, scapolite, albite, K-feldspar, biotite, amphibole and tourmaline characterize the northern part of the Fennoscandian Shield, indicating extensive interaction between hydrothermal fluids and the crust. These systems occur both on a regional scale and directly coupled with different kinds of mineralization (Frietsch et al., 1997; Bergman et al., 2001). The importance of evaporites in the genesis of Fe-oxide–Cu–Au deposits has been argued by Barton and Johnson (1996) and evaporitic units are suggested to have been present in the Karelian greenstones based on the abundance of albite and scapolite in certain stratigraphic units (Tuisku, 1985; Martinsson, 1997; Vanhanen, 2001; Eilu et al., in press) and the Br/Cl ratio of the ore bearing fluids (Wanhainen et al., 2003; Williams et al., 2003). On a broad scale, most of the Kiruna-type Fe-oxide and epigenetic Cu ± Au deposits are located in areas where both strong Na–Cl alteration and calc-alkaline to alkali-calcic, monzonitic magmatism occur (Fig. 9). In areas where either regional scapolitization or monzonitic magmatism is lacking, only minor deposits are found. Thus, it seems that continental margin magmatism interacting with evaporitic sequences is a metallogenetically important feature in the Fennoscandian Shield (cf. Fig. 9).

A fairly comprehensive U–Pb age data set exists for titanite from Kiruna-type Fe-oxide and epigenetic Cu ± Au occurrences, but it should be noted that in more complex systems with overprinting alteration assemblages, different titanite age determinations may occasionally yield disparate ages. Age data displayed in Fig. 10 from epigenetic Cu ± Au deposits and related hydrothermal alteration systems in the northern Norrbotten ore province in Sweden indicate two major events of ore formation at ca. 1.81 to 1.76 and 1.88 to 1.85 Ga (Billström and Martinsson, 2000). Deposits in the northern parts of Norway and Finland exhibit the same two events plus a third probable stage of mineralization at ca. 1.84 to 1.81 Ga (Bjørlykke et al., 1990; Mänttäri, 1995). Ages for Kiruna-type Fe-oxide ores are only published from the Kiruna area and suggest that these ores were formed between 1.89 and 1.88 Ga (Romer et al., 1994; Cliff et al., 1990). Thus, there is both a strong spatial and temporal correlation with the Kiirunavaara Group and a temporal correlation with the older intrusions of the Perthite–Monzonite Suite, co-magmatic with the Kiirunavaara Group rocks.

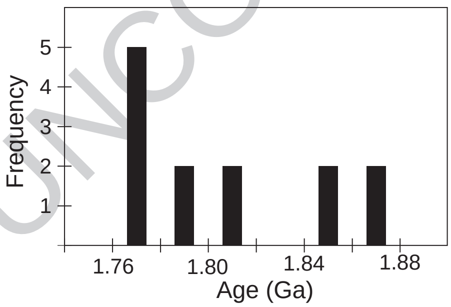


Fig. 10. Histogram showing U–Pb titanite ages for epigenetic Fe–Cu–Au occurrences from northern Sweden. Data from Billström and Martinsson (2000) and Billström (unpublished).

Fe-oxide–Cu–Au style mineralization in the Fennoscandian Shield seems to be a product of multistage magmatic, metamorphic and tectonic processes, invariably involving saline hydrothermal fluids. Several different age groups of mineralization, 1.89 to 1.88 (Kiruna-type Fe-oxide), 1.89 to 1.85, 1.84 to 1.82, and 1.81 to 1.76 Ga (epigenetic Cu ± Au), imply that mineralization took place in different tectonic settings. The older Kiruna type Fe-oxide occurrences formed in a continental margin arc, possibly from Fe-oxide magmas genetically related to monzonitic intrusions. High-salinity fluids formed as a result of magma–crust interaction and were responsible for epigenetic Cu ± Au mineralization. The younger deposits formed largely during or slightly after the peak of the orogeny, during convergence, and at least partly by remobilization of older mineralization. The hydrothermal fluids responsible for the younger deposits were focused along crustal-scale shear zones with mineralization in second to fourth order structures.

4.5. Fe–Ti oxides in anorthosites

Magmatic ilmenite deposits are typically hosted by massif-type anorthosites. The largest deposit in the world is the Lac Tio ilmenite body in the Havre St Pierre anorthosite complex in Quebec. The second largest, the Tellnes ilmenite norite deposit, sits in the Åna-Sira anorthosite (Fig. 11) in Rogaland, southern Norway. Understanding of the formation of ilmenite deposits thus relies on the nature of the anorthosites, the nature of their parent magma, and the tectonic setting in which they were emplaced.

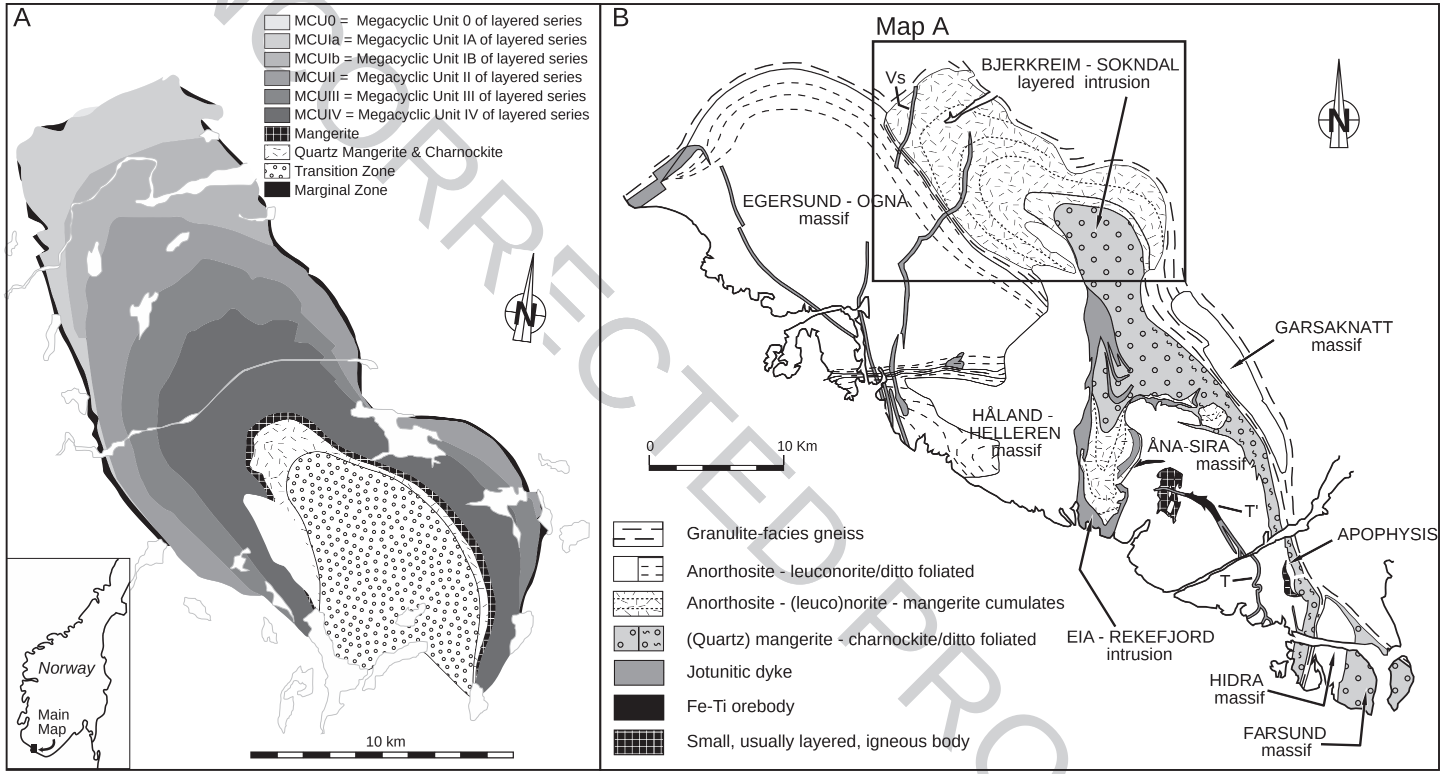


Fig. 11. The geology of the Bjerkreim–Sokndal intrusion, SW Norway, after Robins and Wilson (2001), illustration A; and Duchesne and Bingen (2001), illustration B. T = Tellnes "main dyke"; T' = Tellnes ilmenite deposit.

The consensus model for anorthosite petrogenesis, summarized by Ashwal (1993), involves a deep-seated magma chamber in which a mantle-derived mafic magma ponds at the crust–mantle interface. Differentiation produces a lower density plagioclase cumulate at the roof of the chamber, from which, due to gravity instabilities, blobs of plagioclase mush detach and rise diapirically through the lower crust to be emplaced at mid-crustal levels, where they coalesce to form anorthosite plutons.

In the last decade, this model has been improved and partly modified to account for new field, experimental and geochemical constraints. The typical high-alumina orthopyroxene megacrysts (with plagioclase exsolutions) that form subophitic aggregates with plagioclase megacrysts is stable at 10 to 13 kbar in contrast to the <5 kbar pressure of final emplacement (Fram and Longhi, 1992; Longhi et al., 1993), confirming the polybaric character of the crystallization. The diapirism mechanism was verified by finite element modelling, taking into account the thermo-mechanical properties of anorthosite and mid- and lower crustal rocks (Barnichon et al., 1999).

Experimental data for dry basaltic systems (Longhi et al., 1999) show that the parent magma compositions of anorthosite massifs (high-alumina basalt at Harp Lake, Nain; hypersthene monzodiorite or jotunite in Rogaland) occur on a thermal maximum of the plagioclase + orthopyroxene + clinopyroxene cotectic, in the pressure range (10 to 13 kbar) typical of the high-alumina orthopyroxene megacrysts. Consequently, the parent magmas cannot be generated by fractionation of a mantle-derived basaltic magma, but result from the melting of a mafic rock (containing plagioclase and two pyroxenes) between 40 and 50 km depth. Re–Os isotope data strongly support a mafic lower crust origin (Stein et al., 1998; Wiszniewska et al., 2002), particularly in Rogaland where high Os isotopic ratios can only be accounted for by a mafic source, because there is no significantly older crust in SW Scandinavia (Schiellerup et al., 2000).

Detailed field studies in several anorthosite complexes have shown that anorthosites are frequently associated with zones of weakness in the crust that may have favoured their emplacement at mid-crust levels (Emslie et al., 1994; Scoates and Chamberlain, 1997). In SW Scandinavia, terrane boundaries have been traced in deep seismic profiles to Moho offsets or to tongues of lower crustal material (the so-called Telemark Craton Tongue) underthrust to depths greater than 40 to 50 km (Andersson et al., 1996). In Poland the Suwalki anorthosite was emplaced in the Svecofennian platform along the E–W-trending Mazury lineament (Wiszniewska et al., 2002). In Ukraine, the Korosten pluton occurs at the intersection of two lithospheric-scale faults, imaged by geophysical methods (Bogdanova et al., 2004). It has been suggested that such tongues of underthrust, mafic lower crust can reach sufficiently high temperatures to melt and produce the parent magma of massive anorthosites, if accompanied by delamination along the weakness zone and asthenospheric uprise (Duchesne et al., 1999). This hypothesis seems superior to the classical hot-spot hypothesis, which is unable to explain how several generations of anorthosites, separated by hundreds of millions of years, can occur in the same province (Scoates and Chamberlain, 1997; Hamilton et al., 1998). An alternative crustal tongue melting model (Duchesne et al., 1999) links anorthosite production to a local geochemical property of the lower crust, which permits melting each time heat is added.

In the light of these new developments, the concept of an intraplate setting, typical of the consensus model of anorthosite formation, has also been reconsidered. Emplacement along weakness zones can be favoured by relative movements between terranes either in a post-collisional setting or, if a local source of extra heat is provided, during tectonic rejuvenation (Duchesne et al., 1999), possibly due to distant collisions at the margins of the craton.

Ilmenite deposits are hosted in andesine anorthosites (Anderson and Morin, 1969; Ashwal, 1993) and their genesis has been interpreted to reflect complex diapiric evolution of the host anorthosite and fractional crystallization of the parent magma (Duchesne, 1999; Duchesne and Schiellerup, 2001). The diapiric mechanism has been verified by finite element modelling, taking into account the thermo-mechanical properties of anorthosite and mid- and lower-crustal rocks (Barnichon et al., 1999). There are multiple lines of evidence to indicate that the parental magmas of andesine anorthosites are enriched in Fe and Ti, whichever model is invoked for their genesis. In the underplating model, crystallization of olivine and high alumina orthopyroxene megacrysts in the deep-seated magma chamber enriches the residual melts in Fe, Ti and P to produce ferrodiorite (Ashwal, 1993). In the crustal tongue melting model, a Fe–Ti–P rich melt, hypersthene monzodiorite (jotunite), is directly produced by dry melting of gabbronorite in the lower crust (Longhi et al., 1999). In the deep-seated magma chamber, this melt can generate andesine anorthosite mushes liable to rise diapirically. The melt can also rise through dykes to shallower magma chambers where it can differentiate at lower pressure conditions. The Bjerkreim–Sokndal layered intrusion in Rogaland (Duchesne and Hertogen, 1988; Vander Auwera and Longhi, 1994; Wilson et al., 1996) is an example of such a hypersthene monzodiorite (jotunite) magma chamber (Fig. 11).

These Ti-rich magmas are, however, unable to produce ilmenite concentrations of economic value under normal conditions. Norite layers rarely contain more than 8% TiO_2 (Charlier and Duchesne, 2003). Ilmenite must therefore be concentrated to generate an orebody. Ilmenite–silicate liquid immiscibility has long been invoked (Philpotts, 1967) because it seems to account for field observations of ilmenite dykes with sharp contacts, suggesting a liquid behaviour of pure ilmenite (Duchesne, 1996). Although this mechanism is appealing it has not been confirmed experimentally (Lindsley, 2003), because of the high temperature (1400 °C) required to produce pure ilmenite melt. The apparent liquid behaviour of ilmenite therefore probably results from its ability to creep in the solid state in response to stress (Paludan et al., 1994; Duchesne, 1996). Fractional crystallization of hypersthene monzodiorite (jotunite), illustrated by the succession of cumulates in the Bjerkreim–Sokndal intrusion (Wilson et al., 1996) and well constrained experimentally (Vander Auwera and Longhi, 1994), shows that ilmenite is the second mineral to appear on the liquidus, after plagioclase but before orthopyroxene. Provided that an adequate mechanism (crystal sorting, delayed nucleation, oscillation of the cotectic with pressure, or some other more enigmatic mechanism) was involved, ilmenite might be separated from plagioclase to form monomineralic layers. In this respect, the debate is similar to that on the origin of the chromitite layers of the Bushveld Complex (Wager and Brown, 1968; Cawthorn, 1996). Cotectic crystallization of ilmenite and plagioclase, however, raises another problem. In layered bodies such as the Bjerkreim–Sokndal intrusion, the amount of plagioclase–ilmenite cumulates is small relative to plagioclase–ilmenite–orthopyroxene cumulates and plagioclase–ilmenite–magnetite–orthopyroxene–clinopyroxene–apatite cumulates. In a large deposit such as Lac Tio, the inverse situation is observed; the mass of ilmenite is much greater than the mass of norite and gabbronorites. Selective erosion of the ilmenite-poor rocks is possible, but not completely convincing. Flushing of a magma through an unconsolidated crystal mush in a conduit can account for the large quantity of primitive cumulates, but also begs the question of the norite and gabbronorite counterparts. Finally, solid-state creep can also be invoked to explain the ilmenite enrichment by separating it from the more competent plagioclase (Duchesne, 1999). The diapiric environment of anorthosite in which crystallization and deformation occur at high temperature is suitable for such a mechanism in massif anorthosite-hosted deposits (Duchesne, 1999).

The trace element contents of ilmenite and accompanying magnetite are an important issue because they drastically constrain the economic value of a deposit; Cr and Mg are deleterious in ilmenite and V is beneficial in magnetite. Cr and V occur in the magma with different valence states, essentially 3+, 4+, and 5+. Only the trivalent species can substitute in ilmenite and magnetite. It can thus be anticipated that the oxygen fugacity will be a controlling factor of the trace element contents (Toplis and Corgne, 2002). The other factors controlling the partitioning of Cr and V into oxides are the contents in the parental magma, the degree of fractional crystallization, and the intensity of subsolidus readjustment between oxide minerals (Duchesne, 1999). The mineral/melt partition coefficient for Mg in ilmenite does not vary with pressure (Vander Auwera et al., 2003) and thus the most critical factor remains the composition of the magma. The content of Cr and V in magnetite drastically depends on the degree of evolution of the magma when magnetite appears at the liquidus. Toplis and Corgne (2002) have calculated that in ilmenite-free systems a decrease of oxygen fugacity by two orders of magnitude (e.g., from NNO+1 to NNO-1) will increase the V content of magnetite by a factor of 3 (e.g., from 1% V_2O_5 to 3% V_2O_5). Clearly, in ilmenite-bearing systems, the ilmenite content will also be indirectly influenced by such behaviour.

5. Discussion

The formation of Precambrian ore deposits has been studied extensively in most shields around the world. Genetic models of these deposits have developed largely in parallel with the evolving concepts of Precambrian plate tectonics. Today there is little argument about the existence of subduction processes in the Proterozoic and the late Archaean, but there is a debate about the rate of subduction, how efficient the subduction processes were, and the effects of subduction on magmatism. As discussed above, geodynamic processes must have operated very differently from those of the modern Earth since the mantle was hotter, leading to more active and voluminous magmatism, and the degree of mantle melting was probably greater, leading to more Mg-rich magmas. Faster moving, hotter Archaean (and Palaeoproterozoic) plates also probably accumulated less sediment and contained a thinner section of lithospheric mantle (Sleep and Windley, 1982).

Precambrian continental crust formed episodically around ca. 2.7, 2.5, 2.1 and 1.9 to 1.8 Ga (Goldstein et al., 1997; Condie, 1999). As noted above, these peaks are interpreted by some authors as the times of supercontinent formation: for Fennoscandia, the data simply suggest that they represent periods of enhanced crustal growth and possibly accelerated mantle convection. However, economic mineral deposits in the Fennoscandian Shield are mainly hosted in Palaeoproterozoic rocks. Only a few significant examples of Archaean orogenic gold and Ni–Cu deposits are known. In Ilomantsi, the gold mineralization slightly preceded or was synchronous with the peak of the regional metamorphism (Sorjonen-Ward, 1993). There is no really good understanding as to why the Archaean seems to be less prospective in the Fennoscandian Shield compared with most other shield areas. One possible reason is that large parts of the Archaean greenstone belts are located in Russia and are under-explored by modern standards. Also parts of the Archaean in Finland, especially under thick overburden, are under-explored. Large areas of the Archaean have also experienced a substantial Palaeoproterozoic thickening, which might have produced an erosional level devoid of greenstone belts. The period from ca. 2.74 to 2.69 Ga corresponds to a period of intense intrabasinal mantle plumes and a subsequent global plume-breakout event (Barley et al., 1998). The greenstone belts in the Fennoscandian Shield seem to be slightly older than the global Neoarchaean peak in mineralization and mantle plume activity (e.g., Huhma et al., 1999) and, hence, magmatism and hydrothermal activity might have been less intense and unable to form major Ni–PGE and VMS deposits. This could also explain the lack of large orogenic gold deposits related to subsequent accretion during peak orogeny.

In contrast to the Archaean, the Palaeoproterozoic sequences are intensely mineralized and contain a wealth of economic mineral deposits. It is therefore suitable here to discuss the Palaeoproterozoic geodynamic evolution and metallogeny in more detail, especially for the time period from ca. 2.1 to 1.8 Ga when the vast majority of known economic deposits formed. We will utilize a slightly modified version of the recent geodynamic model for this period by Lahtinen et al. (2005), which is displayed in Fig. 12 as twelve generalized cartoons (a–l), showing the plate tectonic geometry at different stages between 2.06 and 1.78 Ga. The ore deposit types discussed in this paper are also indicated for those stages during which they are suggested to have formed. The cartoons are accompanied by schematic cross-sections (Fig. 13, adapted from Lahtinen et al., 2005), displaying the tectonic environments of ore formation. The geodynamic settings of ore deposits are discussed below within the framework of (1) Palaeoproterozoic rifting of the Archaean continent at 2.5 to 2.06 Ga, (2) microcontinent accretion at 1.96 to 1.88 Ga, (3) continent–continent collision at 1.87 to 1.79 Ga and (4) oblique continent–continent collision at 1.84 to 1.78 Ga.

5.1. Palaeoproterozoic rifting of the Archaean continent at 2.5 to 2.06 Ga

The Archaean cratonic nucleus of the Fennoscandian Shield rifted in several stages during the Early Proterozoic. The hot spot-related rifting event at 2.4 Ga is manifested in layered igneous complexes, mafic dyke swarms, intracontinental and continental margin volcanism, and epicontinental sedimentation. Erosion of the Archaean craton is indicated by numerous coarse clastic units deposited within rift basins during the time interval between 2.5 and 2.06 Ga.

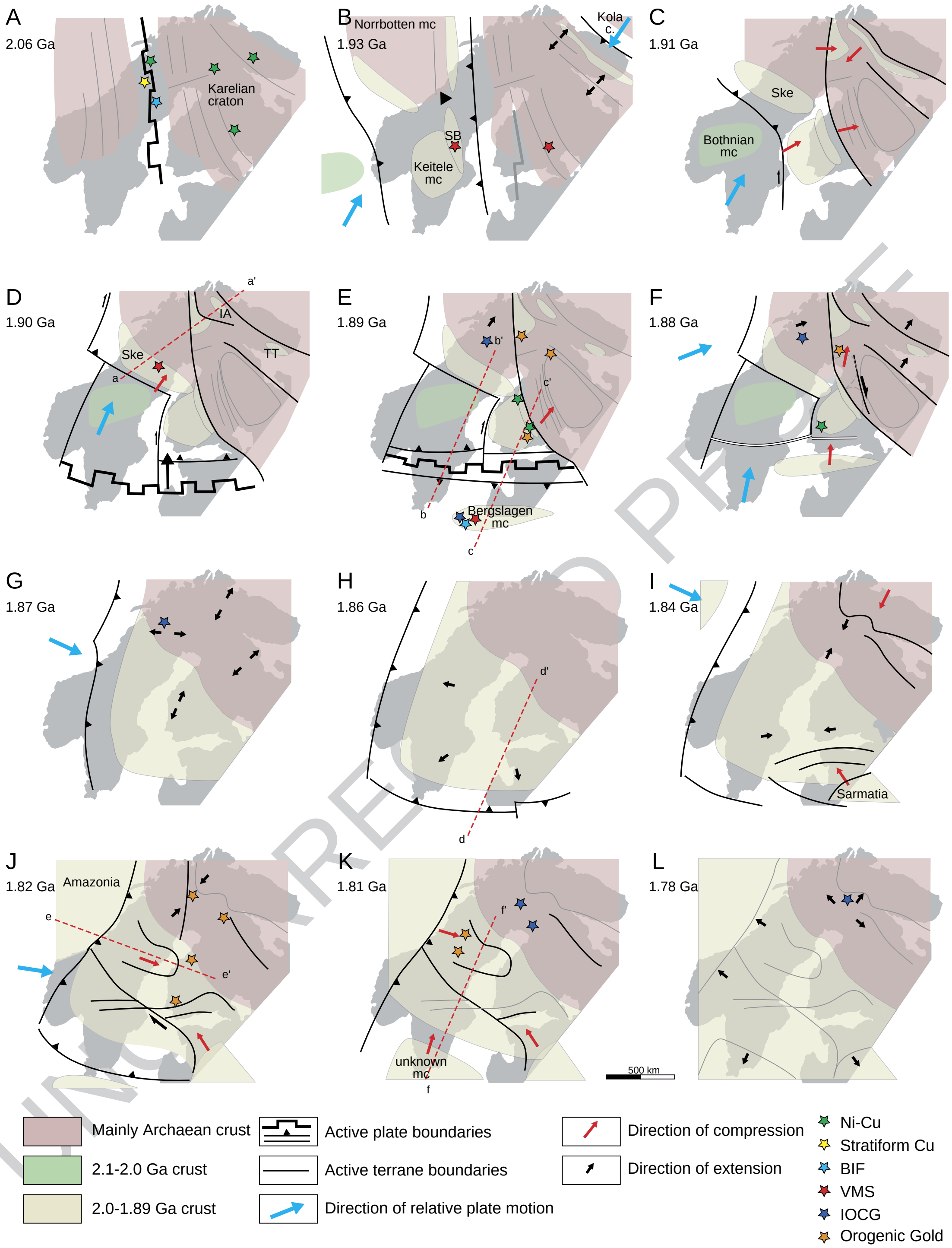


Fig. 12. The geodynamic evolution of the Fennoscandian Shield between 2.06 and 1.78 Ga after Lahtinen et al. (2005). The temporal and spatial relationship between tectonic setting and ore types is indicated and discussed in the text. Generalized cross-sections a–a' to f–f' are shown in Fig. 13.

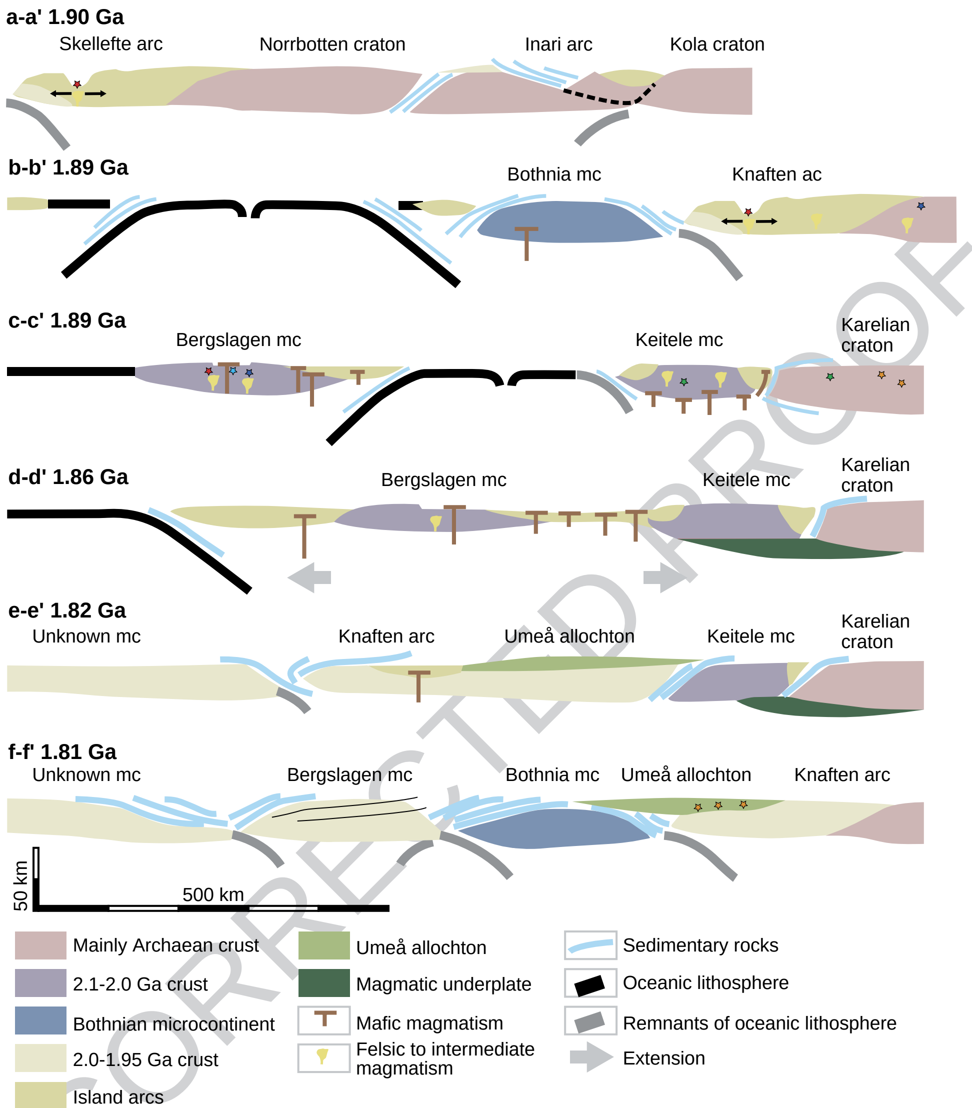


Fig. 13. Generalized cross-sections indicated in Fig. 12 after Lahtinen et al. (2005). Stars denote mineralization styles as in Fig. 12. mc = microcontinent. According to Lahtinen et al. (2005) lithologic, geochemical, isotopic and geophysical data suggest the following pre-1.92 Ga components in the Fennoscandian Shield, illustrated in the cross-sections:

- The Karelian, Kola and Norrbotten Archaean cratons
- The Keitele, Bergslagen and Bothnia > 2.0 Ga age microcontinents
- The Kittilä ~2.0 Ga island arc and oceanic crust
- The Savo, Knaften, Inari and Tersk ~1.95 Ga island arcs

The Karelian and Kola cratons are well exposed, whereas the Norrbotten craton is not exposed. The Paleoproterozoic microcontinents Keitele, Bergslagen and Bothnia have no identifed surface expressions. The Kittilä and Savo arcs are partly exposed and only small slivers of the Knaften arc are found at surface. The relationship of the Inari and Tersk arcs with the Archaean crust is not well known. The Umeå allochthon consists of a pre 1.9 Ga rock sequence, which is overlain by younger rocks in the Bothnian basin.

In this environment, the mafic layered intrusions that were emplaced during the early phases of the rifting of the craton contain major chromitite and Ni–Cu sulphide–PGE deposits. In the rift basins, continental to submarine volcanism, ranging in composition from ultramafic to mafic and intermediate, is associated with iron formations, but also with 2.2 to 2.05 Ga, low-grade, Ni–Cu and PGE deposits within layered igneous complexes (e.g., Keivitsa) and high-grade Ni–Cu deposits in ultramafic volcanic rocks and fractionated mafic–ultramafic intrusions (e.g., Pechenga). In some of the rift basins, stratiform Cu ± Fe-oxide deposits formed within volcaniclastic units of the greenstone belts (Inkinen, 1979; Martinsson, 1997). The largest and sole economic deposit of this category is the Viscaria Cu deposit in Kiruna. The tectonic situation at the end of extension and the onset of convergence, leading to basin inversion, and the Svecokarelian orogen (below subdivided into several discrete tectonic events) are illustrated in Fig. 12.

5.2. Microcontinent accretion at 1.96 to 1.88 Ga

The first prominent evidence of convergence is the obduction of ca. 1.97 to 1.96 Ga ophiolitic sequences at Jormua and Outokumpu in Finland. These ophiolitic slices contain VMS-like deposits. However, as they have recently been described as hosted by mantle rocks (Kontinen, 1998; Sorjonen-Ward et al., 2004), an exhalative or shallow subseafloor replacement origin is excluded and they cannot be classified into the VMS category sensu stricto. The tectonic situation after the obduction of these ophiolites is illustrated in Fig. 13 where subduction and back-arc rifting in the Lapland–Kola area, westward subduction under the Keitele microcontinent (Savo Belt) and Norrbotten microcontinent (Kittilä), and NE subduction under the Norrbotten microcontinent are the main tectonic features at ca. 1.93 Ga. A rifted primitive island arc complex (the Savo belt), formed by north-eastward subduction west of the craton, contains the VMS deposits formed in the Pyhäsalmi area in Finland (Lahtinen, 1994).

The Savo belt was accreted to the craton at ca. 1.91 to 1.90 Ga during the peak of the Lapland–Kola and Lapland–Savo orogenies (Fig. 12C, D). The initial stage of collision of the Bothnian microcontinent with the Norrbotten and Keitele microcontinents also occurred at this stage. Large areas of the Archaean domain were reheated by burial beneath a sequence of nappes in the foreland of the Palaeoproterozoic orogens around 1.9 Ga, but there is no evidence for a distinct Proterozoic gold mineralization event in the area at this stage. The Skellefte arc formed during the collision of the Bothnian microcontinent with the Norrbotten and Keitele microcontinents (Fig. 12D), either as a continental margin arc or possibly as an accreted island arc. There is evidence that this extensively mineralized arc was under extension during the formation of the VMS ores (Allen et al., 2002). Extension in the Skellefte district was followed by basin inversion and rapid uplift and erosion of the arc. Porphyry copper deposits were formed at this stage, and also on the continent side of the arc (Weihed et al., 1992). Docking of the Bothnian microcontinent with the Norrbotten and Keitele microcontinents and differences in relative plate motions resulted in a transform fault between the Keitele and Bothnian microcontinents (Fig. 12D). Polarity reversal of subduction, and the onset of subduction towards the north under the Keitele microcontinent, was also initiated at ca. 1.90 Ga (Fig. 12D).

During subduction switchover and the onset of subduction towards the north under the Bothnian microcontinent, magmatism was intense along the craton margin. During crustal shortening, orogenic gold mineralization occurred in the thickened crust within accreted terranes at ca. 1.89 to 1.88 Ga, during which period metamorphism peaked within the craton. Subduction under the Keitele microcontinent locked up and the ocean basin was consumed by subduction towards the south under the combined Uusimaa island arc and the Bergslagen microcontinent (Fig. 12E, F). Magmatism along the craton margin produced mafic intrusion-hosted Ni–Cu deposits. Magmatic hydrothermal fluids and fluids derived from heating of evaporitic sequences in the rift-related greenstone sequences within the craton, possibly related to extension of the continental crust, formed iron-oxide and IOCG deposits in areas underlain by older Palaeoproterozoic cratonized crust. The Bergslagen microcontinent started to accrete from the south due to consumption of the ocean by subduction at ca. 1.88 Ga. In extensional settings within continental margin arcs within the Bergslagen continent, VMS, iron-oxide and banded iron formations formed, closely related in space and time.

The peak of the Svecokarelian Orogeny at ca. 1.88 to 1.87 Ga in the eastern part of the shield involved a strong compressional stage. The Keitele–Bergslagen collision resulted in substantial shortening within the collision zone, overthrusting at the western margin of the Karelian craton, basin inversion in Lapland, and reactivation of the Lapland–Savo suture zone. Subduction beneath the Bothnian microcontinent was still active. Subduction towards the east under the Norrbotten microcontinent commenced and local extensional domains in the Kola and Belomorian areas were initiated (Fig. 12F, G). The continued crustal shortening within the craton resulted in orogenic gold mineralization, especially in the eastern part of the shield.

5.3. Continent–continent collision between 1.87 and 1.79 Ga, and possible orogenic collapse

The period between 1.87 and 1.83 Ga is still rather poorly understood in terms of tectonic evolution. There is evidence of scattered magmatism within the craton as well as deformation and metamorphism that may have peaked later in the western part of the shield. Lahtinen et al. (2005) propose an attempted orogenic collapse of the Svecokarelian orogen at ca. 1.87 to 1.85 Ga, whilst at the western margin the subduction zone migrated southwards (Fig. 12G, H). There is, however, little hard evidence for orogenic collapse and, in fact, many orogenic gold occurrences, notably in Finland, possibly formed within this timeframe (Eilu et al., 2003). At the southern margin of the craton, subduction towards the SE and the NE was initiated (c.f., sections d–d' and f–f' in Fig. 13) and, according to Lahtinen et al. (2005), large-scale extension took place in the hinterland. After ca. 1.87 Ga there is little evidence of any other major ore formation until crustal shortening related to E–W subduction and subsequent continent–continent collision at ca. 1.84 to 1.78 Ga.

5.4. Oblique continent–continent collision at 1.84 to 1.78 Ga

The Sarmatian crustal segment collided with the SE margin of Fennoscandia at ca. 1.84 Ga (Fig. 12I). This initiated what Lahtinen et al. (2005) refer to as the Svecobaltic Orogeny, expressed as basin inversion and thrusting. Subduction in the W and SW was still active and docking of Laurentia to Fennoscandia in the NE led to the final emplacement of the Lapland Granulite Belt and reactivation of the Belomorian mobile zone. The peak of the Svecobaltic Orogeny and the onset of the Nordic Orogeny occurred at ca. 1.82 Ga (Lahtinen et al., 2005). The oblique collision of Fennoscandia with Sarmatia resulted in the migration of a transform fault within the continent. This crustal-scale shear zone divided the Svecobaltic orogen into two different compressional regimes, a retreating subduction zone active in the SW, whilst a transpressional regime prevailed in the SE. The Nordic Orogeny started with collision of Amazonia and Fennoscandia in the NW with crustal-scale thrusting occurring in the hinterland (Fig. 12J). This collision caused crustal shortening in the hinterland and fluids circulating in the middle to upper crust formed orogenic gold deposits in, at least, the western part of the shield. Far-field effects of this collision may also have caused reactivation of shear zones further east and orogenic gold mineralization may also have occurred in the eastern part of the craton in Finland. The crustal shortening caused extensive deformation and metamorphism as well as partial melting of the middle continental crust. These processes also led to late-stage remobilization and formation of Cu–Au occurrences within the NW part of the shield in the hinterland to the active collision. Subsequent amalgamation of Laurentia, Fennoscandia, Amazonia, Sarmatia and an unknown continent in the SW came to an end at ca. 1.81 to 1.79 Ga, forming the Palaeoproterozoic supercontinent (Fig. 12J, K). The westward growth of the Fennoscandian Shield was initiated and the Fennoscandian Shield was stabilized between 1.79 and 1.77 Ga (Fig. 12L).

The Anorthosite-hosted Ti-deposits in the SW part of the shield are related to continued westward growth and orogenic stacking in the craton. The Tellnes Ti-deposit occurs in part of the 930 to 920 Ma (Schärer et al., 1996) Rogaland Anorthosite Province, (Charlier, 2005, Box 8-4). This magmatic province was formed in the SW part of the Sveconorwegian orogen during granulite facies metamorphism, post-dating the last regional deformation by ca. 40 million years (Box 8-4, Charlier, 2005, this volume).

6. Conclusions

The Palaeoproterozoic, not the Archaean, hosts most of the economic mineral deposits in the Fennoscandian Shield. Although the reason for this is not fully understood, the fact that a major part of the Archaean located in Russia is under-explored by modern standards may be one reason why not many major deposits have been found in the Archaean units. Another reason might be that most of the Archaean greenstone belts seem to be slightly older than the global Neoarchaean peak in mineralization and mantle plume activity (e.g., Huhma et al., 1999) and, hence, magmatism and hydrothermal activity might have been less intense and unable to form major deposits.

In contrast, the Palaeoproterozoic is well endowed with metallic resources. It appears that Precambrian geodynamics involved faster moving, hotter plates that accumulated less sediment and contained a thinner section of lithosphere mantle than the present plate-tectonic processes. This scenario also fits with the complex evolution of the Fennoscandian Shield between 2.06 and 1.78 Ga, when rapid accretion of island arcs and several microcontinent–continent collisions in a complex array of orogens was manifested in short-lived but intense orogenies. Most of the major ore deposits in the area also formed during this evolutionary stage and thus a strong geodynamic control on ore deposit formation is suggested.

Throughout the evolution of the Fennoscandian Shield the orogenic gold deposits, where direct or indirect constraints on age are available, also reflect the orogenic younging of the shield towards the SW and west. Most orogenic gold deposits formed during periods of crustal shortening and mineralization processes peaked at 2.72 to 2.67, 1.90 to 1.86, and 1.85 to 1.79 Ga.

At ca. 2.5 to 2.4 Ga the Archaean craton rifted for the first time, facilitating the emplacement of extensive layered intrusions and mafic dyke swarms. At his stage and, to some extent, also during the later rifting stages at 2.2 to 2.05 Ga, Ni–Cu ± PGE deposits formed both as part of layered igneous complexes and associated with mafic volcanism in the rift basins. Synorogenic mafic–ultramafic intrusions formed during the peak of the Svecokarelian orogen at ca. 1.89 to 1.88 Ga. These host numerous Ni–Cu deposits and are confined to linear belts that slightly post-date arc volcanism.

Nearly all VMS-style deposits in the Fennoscandian Shield formed during the time span between 1.97 and 1.88 Ga in extensional settings during basin inversion and accretion. The oldest, the Outokumpu-type ophiolitic Cu–Co–Au deposits were apparently formed at ca. 1.97 Ga in mantle rocks and were obducted on to the Archaean continent during onset of convergence. The next, more typical VMS deposits formed at 1.93 to 1.91 Ga in an accreted, primitive, bimodal arc setting formed during extension of only slightly older volcanic crust in the Pyhäsalmi area in central Finland. Their host rocks are tholeiitic basalts and transitional to calc-alkaline rhyolites, including high-silica varieties, and the deposits broadly fit within the "bimodal mafic type" classification of Barrie and Hannington (1999). The Skellefte VMS deposits are 20 to 30 million years younger and Allen et al. (2002) suggest that, in contrast to the VMS deposits of the Pyhäsalmi area, the district is a remnant of a strongly extensional intra-arc region that developed on continental or mature arc crust where the basement was only slightly older. The Bergslagen–Uusimaa belt, with a much more diverse metallogeny compared to the Skellefte and Pyhäsalmi areas, is coeval with the Skellefte area, but was formed within or at the margin of a microcontinent that collided with Fennoscandia at ca. 1.88 to 1.87 Ga. The Bergslagen region is interpreted as an intra-continental extensional or continental margin back-arc region developed on older continental crust.

IOCG occurrences in the Fennoscandian Shield are diverse in style. At least the oldest mineralizing stages at ca. 1.88 Ga are coeval with magmatism having a monzonitic fractionation trend and calc-alkaline to alkaline subaerial volcanism more akin to continental arc or magmatism inboard of an active arc. There is also evidence for multiple metal introduction or remobilization between ca. 1.80 and 1.77 Ga related to late- to post-orogenic magmatism distal to the active N–S subduction zone in the west. Models have also been suggested where the interaction of magmas with evaporitic sequences in the older Palaeoproterozoic rift sequences is important for forming fluids that have the right composition to carry large amounts of Fe, Cu and Au.

Large volumes of anorthositic magmas characterize the Sveconorwegian Orogeny, in the SW part of the

Fennoscandian Shield. The best example of a major concentration of Ti associated with these anorthosites is the Tellnes deposit. The Tellnes ilmenite deposit belongs to the Mesoproterozoic (930 to 920 Ma) Rogaland Anorthosite Province in SW Norway. The rocks of this province were emplaced in the SW part of the Sveconorwegian orogenic belt under granulite facies conditions, ca. 40 million years after the last regional deformation of the Sveconorwegian Orogeny.

This paper has demonstrated the intimate interplay between Precambrian geodynamics and metal concentrations. All ore types discussed in this paper ultimately have their genesis determined by their tectonic setting and therefore the understanding of geodynamic processes in the Precambrian will be a critical part in future sustainable exploration and exploitation of metal resources in shield areas around the world.

Acknowledgements

We would like to thank Nicole Patison of Riddarhyttan Resources AB for providing the cross-section of the Suurikuusikko deposit. Discussions with Tero Niiranen and Juhani Ojala were valuable during the process of producing this paper. We would also like to express our sincere gratitude to Derek Blundell for continuous encouragement while writing this manuscript and also for his skills and enthusiasm that made the GEODE project a success. The reviewers Mark Barley and Michael Lesher are warmly thanked for thoughtful comments and suggestions that improved the manuscript substantially.

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