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Metallogeny of the Northern Norrbotten Ore Province, northern Fennoscandian Shield with emphasis on IOCG and apatite-iron ore deposits

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Abstract The Northern Norrbotten Ore Province in northernmost Sweden includes the type localities for Kiruna-type apatite iron deposits and has been the focus for intense exploration and research related to Fe oxide-Cu-Au mineralisation during the last decades. Several different types of Fe-oxide and Cu-Au±Fe oxide mineralisation occur in the region and include: stratiform Cu±Zn±Pb±Fe oxide type, iron formations (including BIF's), Kiruna-type apatite iron ore, and epigenetic Cu±Au±Fe oxide type which may be further subdivided into different styles of mineralisation, some of them with typical IOCG (Iron Oxide-Copper-Gold) characteristics. Generally, the formation of Fe oxide±Cu±Au mineralisation is directly or indirectly dated between ~2.1 and 1.75 Ga, thus spanning about 350 m.y. of geological evolution.

The current paper will present in more detail the characteristics of certain key deposits, and aims to put the global concepts of Fe-oxide Cu-Au mineralisations into a regional context. The focus will be on iron deposits and various types of deposits containing Fe-oxides and Cu-sulphides in different proportions which generally have some characteristics in common with the IOCG style. In particular, ore fluid characteristics (magmatic versus non-magmatic) and new geochronological data are used to link the ore-forming processes with the overall crustal evolution to generate a metallogenetic model.

Rift bounded shallow marine basins developed at ~2.1-2.0 Ga following a long period of extensional tectonics within the Greenstone-dominated, 2.5-2.0 Ga Karelian craton. The ~1.9-1.8 Ga Svecofennian Orogen is characterised by subduction and accretion from the southwest. An initial emplacement of calc-alkaline magmas into ~1.9 Ga continental arcs led to the formation of the Haparanda Suite and the Porphyrite Group volcanic rocks. Following this early stage of magmatic activity, and separated from it by the earliest deformation and metamorphism, more alkali-rich magmas of the Perthite Monzonite Suite and the Kiirunavaara Group volcanic rocks were formed at ~1.88 Ga. Subsequently, partial melting of the middle crust produced large volumes of ~1.85 and 1.8 Ga S-type granites in conjunction with subduction related A-/I-type magmatism and associated deformation and metamorphism.

In our metallogenic model the ore formation is considered to relate to the geological evolution as follows. Iron formations and a few stratiform sulphide deposits were deposited in relation to exhalative processes in rift bounded marine basins. The iron formations may be sub-divided into BIF- (banded iron formations) and Mg-rich types, and at several locations these types grade into each other. There is no direct age evidence to constrain the deposition of iron formations, but stable isotope data and stratigraphic correlations suggest a formation within the 2.1-2.0 Ga age range. The major Kiruna-type ores formed from an iron-rich magma (generally with a hydrothermal over-print) and are restricted to areas occupied by volcanic rocks of the Kiirunavaara Group. It is suggested here that 1.89-1.88 Ga tholeiitic magmas underwent magma liquid immiscibility reactions during fractionation and interaction with crustal rocks, including metaevaporites, generating more felsic magmatic rocks and Kiruna-type iron deposits. A second generation of this ore type, with a minor economic importance, appears to have been formed about 100 Ma later. The epigenetic Cu-Au±Fe oxide mineralisation formed during two stages of the Svecofennian evolution in association with magmatic and metamorphic events and crustal-scale shear zones. During the first stage of mineralisation, from 1.89-1.88 Ga, intrusion-related (porphyry-style) mineralisation and Cu-Au deposits of IOCG affinity formed from magmatic-hydrothermal systems, whereas vein-style and shear zone deposits largely formed at c. 1.78 Ga.

The large range of different Fe oxide and Cu-Au±Fe oxide deposits in Northern Norrbotten is associated with various alteration systems, involving e.g. scapolite, albite, K feldspar, biotite, carbonates, tourmaline and sericite. However, among the apatite iron ores and the epigenetic Cu-Au±Fe oxide deposits the character of mineralization, type of ore- and alteration minerals and metal associations are partly controlled by stratigraphic position (i.e. depth of emplacement). Highly saline, NaCl+CaCl₂ dominated fluids, commonly also including a CO₂-rich population, appear to be a common characteristic feature irrespective of type and age of deposits. Thus, fluids with similar characteristics appear to have been active during quite different stages of the geological evolution. Ore fluids related to epigenetic Cu-Au±Fe oxides display a trend with decreasing salinity, which probably was caused by mixing with meteoric water. Tentatively, this can be linked to different Cu-Au ore paragenesis, including an initial (magnetite)-pyrite-chalcopyrite stage, a main chalcopyrite stage, and a late bornite stage.

Based on the anion composition and the Br/Cl ratio of ore related fluids bittern brines and metaevaporites (including scapolite) seem to be important sources to the high salinity hydrothermal systems generating most of the deposits in Norrbotten. Depending on local conditions and position in the crust these fluids generated a variety of Cu-Au deposits. These include typical IOCG-deposits (Fe-oxides and Cu-Au are part of the same process), IOCG of iron stone type (pre-existing Fe-oxide deposit with later addition of Cu-Au), IOCG of reduced type (lacking Fe-oxides due to local reducing conditions) and vein-style Cu-Au deposits. From a strict genetic point of view, IOCG deposits that formed from fluids of a mainly magmatic origin should be considered to be a different type than those deposits associated with mainly non-magmatic fluids. The former tend to overlap with porphyry systems, whereas those of a mainly non-magmatic origin overlap with sediment hosted Cu-deposits with respect to their origin and character of the ore fluids.

Keywords: Northern Norrbotten ore province, IOCG, Kiruna-type, apatite-iron ore, formation, metallogeny.

1. Introduction

The northern region of the Fennoscandian Shield, involving parts of Finland, Norway and Sweden, is an economically important metallogenic province dominated by Fe oxide and Cu±Au ores showing a large variation in the character of deposits, their mineral composition and host rocks (Martinsson et al., 2001, 2008; Billström et al., 2010). After the pioneering work by Geijer (1910) most studies have been focused on the huge apatite iron ores in the Kiruna area, and these have been very important for the general knowledge of this ore type. More recently the region has been regarded as a typical Fe oxide-Cu-Au (IOCG) province (e.g. Martinsson et al., 2001; Williams et al., 2003). This has been based on the style of Fe and Au-Cu mineralisation and the extensive albite and scapolite alteration. In strictly genetic terms, however, only some of these deposits may be classified as Fe oxide-Cu-Au deposits whilst others only share some of the features characteristic of this ore type (Hitzman et al., 1992; Hitzman, 2000; Groves et al., 2010; Williams et al., 2005). As defined by Hitzman et al. (1992), Fe oxide-Cu-Au deposits constitute a not fully understood, and controversial, deposit type that includes several world class deposits of Cu and Au such as Olympic Dam and Ernest Henry in Australia, Salobo and Cristalino in Brazil and Candelaria in Chile (Porter, 2010).

One of the most fundamental aspects of ore genetic models is the source of fluids and metals, and many different models have been proposed for IOCG type of deposits. These include magmatic and various non-magmatic fluid and metal sources, or a mixture of these sources (Williams et al., 2005, 2010; Porter, 2010). One example of a mixed source model is for Olympic Dam, where early Fe-oxides were deposited from magmatic fluids and subsequently oxidized meteoric fluids introduced Cu-Au (cf. Haynes et al., 1995; Haynes, 2000). A non-magmatic source of evaporitic origin (dissolved evaporates, basinal or surficial brines) has been postulated by Barton and Johnson (1996, 2000), Haynes (2000) and Hitzman (2000). Barton and Johnson (2000) suggested a four-fold grouping of Fe-oxide Cu-Au deposits: 1) magmatic, 2) non-magmatic with coeval brines, 3) non-magmatic with older brines/evaporates, and 4) hybrid or superimposed systems. Group 1) includes porphyry Cu-Au deposits, as well as alkaline magmatic systems, and carbonatites. Group 2) and 3) comprise many of the “typical” Fe-oxide Cu-Au deposits, including Kiirunavaara and Olympic Dam. These deposits have abundant Fe-oxides and sparse sulphides and show a less clear link to intrusive events. Finally, group 4) deposits have mixed characteristics.

In contrast to these non-magmatic models, Pollard (2000, 2001, 2006) strongly advocated a magmatic fluid and metal source related to granitic magmas for Fe-oxide Cu-Au mineralisation in Australia. Based on the occurrence of ore related hypersaline and CO₂-rich fluid inclusions Pollard (2000) argued that these fluids evolved by unmixing of magmatic-derived H₂O-CO₂-salt fluids in contrast to fluids related to porphyry Cu-Au systems where hypersaline brines evolved by boiling of magmatic H₂O-salt fluids. Williams et al. (2005) further suggested that the formation of Fe-oxide Cu-Au deposits may have involved a copper-rich and sulphur-poor fluid phase from CO₂-bearing magmas at >6km depth which could have been channelized by brittle-ductile shear zones and evolved by mineral precipitation, wall rock reactions and mixing with non-magmatic deep-seated fluids. Also for IOCG-deposits in the Coastal Cordillera area of South America a mainly magmatic origin has been proposed with fluids related to oxidized dioritic magmatism (Sillitoe, 2003).

Magmatic versus non-magmatic models for the fluid source have also been tested using halogen ratios (Br/Cl and I/Cl), chlorine and boron isotopes, and noble gas isotopes from a large number of deposits (see Williams et al., 2010 and references therein). Most of these studies have failed to identify a single well defined source for the ore fluids and in many cases mixed sources are indicated. However, the importance of evaporitic components (dissolved halite, basinal brines or bittern brines of surficial origin) has been stressed (e.g. Barton and Johnson, 1996, 2000; Haynes, 2000; Hitzman, 2000; Barton, 2009; Williams et al., 2010) and for some deposits the contribution of magmatic sources to the fluids has been excluded, with the Wernecke Breccias in Yukon as the most convincing example (Hunt et al., 2005, 2010). Based on the possible sources of the ore fluids a slightly different subdivision of IOCG-deposits into three subgroups has been suggested, including magmatic, non-magmatic and hybrid magmatic-non-magmatic types. Eloise and Lightning Creek would be typical examples of the magmatic end-member of IOCG, the Wernecke Breccia and Redbank could typify the non-magmatic end-members, whereas Olympic Dam and Ernest Henry constitute examples of the hybrid magmatic-non-magmatic type of IOCG (Hunt et al., 2007).

Fe oxide-Cu-Au deposits have gained a world wide exploration interest since the discovery of the world class Olympic Dam deposit in Australia in the mid 1970's, and the northern part of the Fennoscandian shield has lately been intensively explored for this type of deposits. However, Northern Norrbotten is probably best known for the presence of large apatite iron ores of the so called Kiruna-type, but it must be noted that several of these Fe oxide deposits also carry some Cu and Au. As a result, in several genetic schemes for the IOCG family of deposits, the Kiruna type apatite iron ores (AIO) has been suggested as an end-member, or at least genetically related to IOCG (Hitzman et al., 1992; Hitzman, 2000; Velasco and Tornos, 2009; Naslund et al., 2002; Porter, 2010; Nold et al., 2014) and in Norrbotten a genetic relation between apatite iron deposits and deposits featuring typical IOCG characteristics has been proposed (Martinsson et al., 2004a). Some genetic models also encompass Cu-Au deposits lacking significant Fe-oxides (e.g. Pollard, 2000; Williams et al., 2005), and in Sweden these could be exemplified by the Pahtohavare Cu-Au deposit. The IOCG deposit type has also been proposed to overlap in style and genesis with other ore systems like carbonatites (Groves and Vielreicher, 2001; Johnston and Pyle, 2005), porphyry Cu-Au in both calc-alkaline and alkaline settings (cf. Pollard, 2000; Sillitoe, 2002; Tornos et al., 2010; Richards and Mumin, 2013), epithermal systems (Kreiner, 2011), and also to possibly include exhalative deposits (Barton, 2009; Nold et al., 2014).

Although spatially related in several districts, the speculative genetic connection between apatite iron ores and iron oxide-copper-gold deposits (Hitzman et al., 1992; Hitzman, 2000) is largely untested, and Groves et al. (2010) have pointed out the need for a more strict definition of IOCG deposits. Another important unresolved problem relates to the mechanism of formation of the apatite iron ores. Some authors (e.g. Nyström and Henríquez, 1994; Naslund et al., 2002; Nyström et al., 2008) have argued that many such ores are strictly orthomagmatic (melt consolidation products reflecting the existence of oxide-silicate melt immiscibility in magma systems) whereas others suggest they are magmatic hydrothermal replacements (e.g. Hilderband, 1986; Hitzman et al., 1992; Rhodes et al., 1999; Sillitoe and Burrows, 2002; Yu et al., 2011) or a combination of orthomagmatic and magmatic hydrothermal processes (Broman et al., 1999; Moore and Modabberi, 2003; Martinsson, 2004a, 2015; Chen et al., 2010). In some genetic models this type of deposits are also suggested to have formed from fluids that underwent little or no interaction with magmas (Barton and Johnson, 1996; Rhodes and Oreskes, 1999).

The current paper will present in more detail the characteristics of selected deposits, and aims to put the global concepts of Fe-oxide Cu-Au mineralisations into a Northern Norrbotten context. The focus will be on iron deposits and various types of deposits containing Fe-oxides and Cu-sulphides in different proportions and that mostly have some characteristics of IOCG. In particular, the origin of fluids (magmatic versus non-magmatic) will be discussed based on fluid characteristics, and new geochronological data are used to link the ore-forming processes to the overall crustal evolution. The geochronological work is based on the occurrence of characteristic hydrothermal alterations (comprising e.g. K feldspar, scapolite, biotite, actinolite, and tourmaline) related to the mineralised zones and the observation that certain minerals, like titanite, form part of these alteration assemblages. Thus, by dating titanite and other phases which, on a textural basis, co-precipitated with the ore minerals or the alteration mineral assemblage, also the ore-forming events are dated. Such information, coupled with stratigraphic position and fluid inclusion investigations on gangue minerals are combined to generate a metallogenetic model for Northern Norrbotten.

2. Regional geology

The large-scale features of the geological evolution of the Fennoscandian shield and the deposit-types in its northern part have been discussed by e.g. Weihed et al. (2005); Billström et al. (2010). It is one of the most important mining areas in Europe, and the northern part, including Sweden, Finland and Russia (Figs. 1 and 2) is intensely mineralised. Unlike most other shield areas, the Fennoscandian shield is more mineralised in the Palaeoproterozoic than in the Archaean. Mineral deposit types, as shown in Fig. 1, include VMS, greenstone-hosted stratiform Fe-Cu-Zn mineralisation, iron formations, Kiruna-type apatite-iron ores, epigenetic Cu-Au ore including porphyry-type Cu-Au mineralisation, orogenic gold deposits and mafic and ultramafic-hosted Ni-(Cu), PGE, Cr and Ti-V mineralisation.

The oldest preserved continental crust in the Fennoscandian Shield was generated during the Saamian Orogeny (3.1–2.9 Ga). Rift-related greenstones, subduction generated calc-alkaline volcanic rocks and tonalitic-trondhjemitic igneous rocks were formed during the Lopian Orogeny (2.9–2.6 Ga). Only a few Archaean economic to sub-economic mineral deposits have been found in the shield, including BIF's, granite hosted Mo-occurrences, and komatiite related Ni-Cu deposits (Gaál, 1986; Eilu, 2012), as well as greenstone-hosted Au deposits (Eilu et al., 2003). The following Palaeoproterozoic units (Fig. 2) were related to several events of rifting and subduction and include Karelian greenstones (2.5–2.0 Ga) and Svecofennian volcanic and sedimentary rocks (1.9 Ga).

Figure 1

Figure 2

2.1. Palaeoproterozoic, 2.5–2.0 Ga Karelian greenstone belts

The Karelian greenstones overlie much of the Archaean craton, and constitute the largest coherent greenstone terrain exposed in the Fennoscandian Shield (Pharaoh, 1985). It extends over 1000 km from the Norwegian coast through Swedish and Finnish Lapland and into adjacent Russian Karelia, and includes the Central Lapland greenstone belt in Finland, the Kiruna greenstones in Sweden and the Kautokeino greenstones in northern Norway. The Karelian rocks have been divided into the Sumi-Sariolan (2.5–2.3 Ga) and Jatulian (2.3–2.0 Ga) subunits (Gorbatshev and Bogdanova, 1993). In northern Sweden the Palaeoproterozoic succession of Karelian greenstones and the following Svecofennian units rest unconformably on a deformed 2.8–2.6 Ga Archaean basement. Stratigraphically lowest is the Kovo Group, ~2.5–2.3 Ga in age (Sumi-Sariolan), which is overlain by the 2.3–2.0 Ga (Jatulian and Ludikowian) Kiruna Greenstone Group. The upper contacts of the Kovo Group and the Kiruna Greenstone Group are characterised by minor unconformities and clasts from these sequences are found in basal conglomerates in overlying units (Martinsson, 1997). Mafic dykes are locally abundant in the Archaean basement and some of these might represent feeders to the extrusive rocks within the Karelian units (Martinsson, 1997, 1999a).

2.1.1. The Kovo Group

The lower unit of the Kovo Group includes conglomerate, quartzite, basaltic lava, and albite-diabase that are found as up to 200 m thick sills. An albite diabase, intruding the lower part of the Kovo Group, has been dated at 2.18 Ga (Skiöld, 1986) and sets a minimum depositional age for the Kovo Group. The conglomerate rests on a weathered surface of Archaean rocks and the clasts are mostly well rounded and dominated by granitoids (Martinsson, 1999b). The overlying unit forms a 1–2 km thick sequence of volcanogenic greywacke with minor intercalations of carbonate rocks and tholeiitic lava having a within-plate basalts (WPB) character. Calc-alkaline basaltic to andesitic volcanic rocks occur locally as a major facies, which represent volcanic centres from where the volcanogenic greywackes derived (Martinsson, 1999b).

2.1.2. The Kiruna Greenstone Group

At Kiruna a thick pile of mafic to ultramafic volcanic rocks overlie the Kovo Group and these rocks constitute the Kiruna Greenstone Group (Martinsson, 1997), and are commonly referred to as Karelian in age (approximately 2.3–2.0 Ga). This sequence is composed of mainly basaltic amygdaloidal lava in the lower part and pillow lava in the upper part. Less important are komatiite, tholeiitic tuff, andesitic to dacitic tuffaceous rocks, and minor black schists and carbonate rocks, which occur in the volcanoclastic units (Fig. 2). Based on petrographical and geochemical criteria, the Kiruna Greenstone Group is divided into six formations, which from the stratigraphically lowest are: Sårkevaratjah Formation, Ädnamvare Formation, Pikse Formation, Viscaria Formation, Peuravaara Formation and Linkaluoppal Formation, with a combined thickness of 3 to 4 km. Albitization and a strong enrichment of Cl and Br in picrite, basalt and conglomerate occurring at the base of the Sårkevaratjah Formation, and silicified carbonate rocks with a nodular structure are suggested to be evidence of former evaporites (Martinsson, 1997). The upper part consists of scapolite altered basaltic lava of WPB-character. The Sårkevaratjah Formation is overlain by peridotitic to basaltic komatiites with a maximum content

of 30% MgO, which constitute the Ädnamvare Formation. In the following Pikse Formation tholeiites form 2 to 20 m thick flows of massive to amygdaloidal lava (Martinsson, 1997).

Volcaniclastic rocks with associated chemical and carbonaceous sediments in the overlying, ore-associated Viscaria Formation record a change in depositional environment, from terrestrial or shallow water, to prevailing subaqueous conditions and exclusively pyroclastic volcanism with MORB-type chemistry and minor intermediate to felsic units. The explosive volcanism generating the Viscaria Formation was succeeded by voluminous eruptions of chemically similar MORB-type pillow lavas in the overlying Peuravaara Formation.

Basin shoaling and a return to pyroclastic volcanism are recorded by the volcaniclastic Linkaluoppal Formation that includes minor graphitic units, iron formations and a carbonate unit at the top. The chemical composition of these volcanic rocks is transitional between MORB and WPB (Martinsson, 1997).

East of a NNE-directed major shear zone, known as the Karesuando-Arjeplog Deformation Zone (KADZ) (Fig. 1), the lithostratigraphy of the greenstones is distinctly different to that of the Kiruna area. These rocks comprise the Veikkavaara greenstones, which represents a rift margin facies of the Karelian greenstones in Northern Norrbotten whilst the Kiruna greenstones have formed within an aborted rift (Martinsson, 2004a,b). Most conspicuous for the Veikkavaara greenstones is the lack of komatiites and basaltic pillow lavas. The Vinsa Formation is a characteristic unit in the uppermost part of the Veikkavaara greenstones and includes graphitic schists and thick units of dolomite and BIF that follows a volcaniclastic unit of basaltic to picritic composition (Martinsson and Wanhainen, 2013).

2.2. Svecofennian, 1.9-1.8 Ga complexes

Svecofennian rocks involve both strong reworking of older crust within the Karelian craton and, importantly, subduction and the accretion of several volcanic arc complexes from the SW towards NE (Weihed et al., 1992; Nironen, 1997; Lahtinen et al., 2003, 2005, 2011). In northern Sweden, the Palaeoproterozoic greenstones are overlain by volcanic and sedimentary rocks comprising several different and partly stratigraphically correlated Svecofennian units. Regionally, most of them exhibit considerable variation in lithologic composition due to partly rapid changes from volcanic to sedimentary dominated facies.

2.2.1. Lower Svecofennian units: Porphyrite Group, Kurravaara Conglomerate, Sammakkovaara Group and the Muorjevaara Group

The Porphyrite Group is restricted to the Kiruna area and is mainly andesitic in composition and commonly porphyritic with abundant 3 to 30 mm phenocrysts of plagioclase. The top and bottom of the Porphyrite Group have not been identified in a single continuous stratigraphic section, hence the total thickness is unknown, but is estimated to a minimum of 2 km in most areas.

The Kurravaara Conglomerate represents a sedimentary dominated facies in the central Kiruna area which is composed of alternating 50 to 300 m thick, ruditic and arenitic beds with minor intermediate to felsic volcanic rocks intercalated in the upper part. Observed sedimentary structures include cross bedding and small channels suggesting a fluvial environment that may represent a fan delta (Kumpulainen, 2002).

East of the Kiruna area the Sammakovaara Group was formed by contemporaneous andesitic volcanism and clastic sedimentation. Rapid facies changes are recorded by different proportions of volcanic and sedimentary rocks and by variations in grain size and composition of the sediments. Intercalations of carbonate rocks, graphitic schists and tourmalinites are found in the Tärendö area and may represent products of temporary anoxic conditions and hydrothermal activity in low energy areas (Martinsson, 2004a,b). The Muorjevaara Group in the Gällivare area has characteristics similar to that of the Sammakovaara Group and may represent an equivalent unit.

Volcanic rocks from the Porphyrite, Sammakovaara and Muorjevaara Groups are calc-alkaline and show a range in SiO₂ content between 52 and 63%. They are probably part of an early Svecofennian continental arc magmatism of mainly intermediate composition, formed during north-east directed subduction at the margin of the Archaean craton (Martinsson and Perdahl, 1995; Martinsson, 2004a).

2.2.2. Middle Svecofennian units: Kiirunavaara Group

The best-documented section of the Kiirunavaara Group is from the central Kiruna area where a 1-4 km thick unit of mainly volcanic rocks are divided into the Hopukka, Luossavaara, and Matojärvi formations (Martinsson, 2004a). From this area a belt of Kiirunavaara Group volcanic rocks extends towards southwest for a distance of at least 100 km (Fig. 1), and these rocks also occur in the Gällivare area. Regional variations within the Kiirunavaara Group are mainly defined by variability in stratigraphic thickness and the proportions of mafic-, intermediate- and felsic volcanic rocks.

The Kiirunavaara Group in the central Kiruna area is dominated by intermediate to felsic volcanic rocks that are overlain by epiclastic sediments (Fig. 2). The Hopukka Formation which traditionally has been called “the syenite porphyry” (Geijer, 1910) is stratigraphically lowest, and is dominated by mostly amygdaloidal andesitic to trachyandesitic lava flows. Felsic porphyries, with abundant 2-5 mm feldspar phenocrysts, and a thickness of up to 800 m constitute the overlying Luossavaara Formation. They have a rhyodacitic composition and Geijer (1919) termed the rock “the quartz-bearing porphyry” due to their quartz-rich groundmass. These rocks are mainly pyroclastic in origin and are dominated by porphyritic tuff-lapillituff. The Matojärvi Formation constitutes a mixed volcanic-sedimentary unit at the top of the Kiirunavaara Group. The lowest part consists of a strongly K feldspar altered and silicified felsic tuff called “the Rektor porphyry” (Geijer, 1910, 1950). The “Rektor porphyry” is overlain by basalts including amygdaloidal lava, agglomerate and reworked volcanoclastics of local origin. At the base they include conglomerate with abundant clasts of iron ore followed by greywacke and pelitic rocks at the top.

Southwest-west of Kiruna, amygdaloidal basaltic lava forms a ~4 km thick pile in the stratigraphically lowest part of the Kiirunavaara Group, which is followed upwards in the stratigraphy by a mainly felsic unit (Martinsson, 2004a). A thin unit of intermediate volcanic rocks, occasionally together with clastic sediments of local origin, usually separates the lower mafic part and the upper felsic dominated part (Martinsson, 2004a). The upper felsic unit is 3 to 5 km thick and is dominated by porphyritic dacitic to rhyodacitic rocks of pyroclastic origin (Offerberg, 1967).

Although the volcanic rocks exhibit a large variation in silica content from ~47 to 76 wt.% they are clearly bimodal in composition with two maxima at ~50 and 70 wt.% SiO₂. The mafic rocks have some

similarities with WPB and low-Ti flood basalts and the felsic units are characterized by high contents of Zr and of most other incompatible elements, which may indicate an intra-plate setting (Martinsson and Perdahl, 1995). Based on the occurrence of a thick basaltic lower unit, the bimodality, and the tendency for a within-plate chemical signature of the volcanic rocks, these rocks have been suggested to correspond to a flood-basalt sequence. However, the tectonic setting of the Kiirunavaara Group is not unequivocally constrained by the chemical characteristics of the volcanic rocks and an alternative interpretation is that the rocks are due to extension of a mature continental arc (Martinsson, 2004a).

2.2.3. Upper Svecofennian units: Hauki and Maattavaara quartzites

The Hauki quartzite, and the similar Maattavaara quartzite located further east, mark a major change in geological evolution in the Kiruna area from a volcanic dominated environment to exclusively epiclastic sedimentation. It is dominated by a quartz-feldspar arenite with planar cross bedding at dm-scale and heavy mineral layers but includes also conglomerate beds. A basal conglomerate is dominated by rounded pebbles of reddish porphyry derived from the underlying Kiirunavaara Group. Although the layering within the Hauki and Maattavaara quartzites is mainly conformable to that of the underlying volcanic units, their lower contacts may be clearly discordant on a local scale. Witschard (1984) suggested that these quartzites formed in small and tectonically active grabens.

2.3. Major Palaeoproterozoic, 1.94-1.76 Ga intrusive rock suites

The major part of the bedrock in the northernmost part of Sweden is composed of intrusive rocks, mainly various types of granitoids but also mafic to ultramafic plutons and dykes. The major suites are: 1) Haparanda Suite (1.94–1.87 Ga) I-type calc-alkaline, granite-granodiorite-tonalite-diorite-gabbro, 2) Perthite-Monzonite Suite (1.88–1.86 Ga) I-type to A-type alkali-calcic, granite-monzonite-monzodiorite-monzogabbro-gabbro-peridotite, 3) Jyryjoki Suite (1.86-1.84 Ga) S-type granite-granodiorite, 4) Lina Suite (1.81–1.76 Ga) S-type, minimum melt and migmatite-associated, granite-pegmatite, and 5) TIB-1 (1.80–1.77 Ga) and TIB-2 (1.72-1.70 Ga) A- to I-type intrusions, alkali-calcic, granite-monzonite-monzodiorite-gabbro and several generations of mafic dykes (Ahl et al., 2001; Bergman et al., 2001).

2.3.2. Calc-alkaline (1.94–1.87 Ga) Haparanda Suite

The oldest rocks related to the Svecokarelian orogen north of the Palaeoproterozoic-Archaean boundary are the Norvijaur type intrusion, dated at $1926 \pm 13/-11$ Ma (Sköld et al., 1993), and a tonalite in the Rombak area which yielded 1940 ± 26 Ma (Romer et al., 1992). These two granitoids show chemical similarities to Haparanda type intrusions and are therefore incorporated into the Haparanda suite, although they are substantially older than other dated intrusions of this suite in northern Sweden. However, similar old ages are recorded in Finland for early tonalitic arc magmatism (Vaasjoki and Sakko, 1988).

The name Haparanda Suite is assigned to ~1.89 and 1.87 Ga intrusions, ranging from gabbro to granite but dominated by diorite-tonalite-granodiorite (Ödman et al., 1949; Hiltunen, 1982). Rocks are typically deformed, and vary in silica content from 56 to 69 wt.% (Martinsson, 2004a). The compositional range and geochemistry, and also age characteristics, of the Haparanda Suite are very similar to that of the arc related volcanic rocks within the Porphyrite-, Sammakovaara- and Muorjevaara Groups. Thus, the Haparanda Suite intrusions are regarded as comagmatic with extrusive phases of the early Svecofennian arc magmatism (Martinsson, 2004a).

2.3.3. Alkali-calcic (1.88–1.86 Ga) Perthite Monzonite Suite

The Perthite Monzonite Suite (PMS) is represented by large peraluminous to metaluminous intrusions in the northwestern part of Norrbotten, but such rocks are rare in the eastern part (Geijer, 1931; Witschard, 1984; Bergman et al., 2001). The plutons are generally undeformed and the silica content shows three major peaks at 38–52, 57–66 and 70–76 wt.%, respectively (Martinsson, 2004a). Monzonite is the dominant rock type and occurs together with varying amounts of gabbro, monzogabbro, monzodiorite, quartzmonzonite, and granite (Bergman et al., 2001). Although only few age determinations exist, the main magmatic event probably occurred at 1.88–1.87 Ga with emplacement of composite monzonitic-granitic intrusions (Skiöld and Öhlander, 1989; Martinsson et al., 1999), whilst some late granites formed at ~1.86 Ga (Skiöld, 1981; Skiöld and Öhlander, 1989). Several ultramafic to mafic complexes occur northwest of Kiruna and include peridotite, troctolite and gabbro (Kathol and Martinsson, 1999).

Intrusions of the Perthite Monzonite Suite are suggested to be co-magmatic with the Kiirunavaara Group volcanic rocks (Witschard, 1984; Martinsson, 2004a). Both display a compositional variation from mafic to felsic combined with a relatively high content of alkali and HFS-elements. Another feature of this intrusive suite that supports such a connection, and a mantle plume origin, is the abundant occurrence of ultramafic-mafic complexes northwest of Kiruna whose existence possibly indicates the location of the plume centre (Martinsson, 2004a). It cannot be excluded, however, that these rocks formed in a back-arc setting related to continuous northeastward subduction.

2.3.4. S-type (1.86-1.84 Ga) Jyryjoki Suite and (1.82–1.76 Ga) Lina Suite

In the eastern part of the study area there are granitoids with U-Pb zircon ages of c. 1.85 Ga. They are mainly represented by granitoids of the Jyryjoki type. These granitoids are associated with pegmatite, and in many places they contain partly assimilated remnants of older rocks (Bergman et al., 2001).

Intrusions of the Lina Suite are extensively developed in Northern Norrbotten (Fig. 1) where they typically occur as granite, pegmatite and aplite of mainly minimum melt composition. Undeformed pegmatite dykes can be seen to cut folded pegmatites demonstrating the existence of different magma generations separated by a deformation event (Martinsson, 2004a). Age determinations indicate an emplacement age between 1.82 and 1.76 Ga for the Lina Suite (Skiöld et al., 1988; Wikström and Persson, 1997; Bergman et al., 2002; Wanhainen et al., 2005), with the younger ages representing pegmatitic phases.

The chemistry of the Lina Suite typically shows a high content of Th and Rb and a depletion of Eu. Even more characteristic is its restricted SiO₂ range between 72 and 76 wt%, which most likely is reflecting an origin as crustal minimum melts (Martinsson, 2004a).

2.3.5. A-, I-type TIB-1 (1.80–1.77 Ga) and TIB-2 (1.72–1.70 Ga) intrusions

Rocks related to the Transscandinavian Igneous Belt (TIB) are coarse- to medium-grained and undeformed. These rocks normally show quartz-poor monzonitic trends, and gabbroic-dioritic-granitic components are relatively common (Romer et al., 1992, 1994; Öhlander and Skiöld, 1994). It is possible to distinguish two generations (~1.8 and 1.7 Ga) of TIB-type intrusions in northern Sweden and adjacent areas of Norway with the younger intrusions occurring in the western part (Romer et al., 1992). In Northern Norrbotten, a local occurrence of 1.79 Ga syenite is known from the Kiruna area (Romer et al., 1994) and further east 1.8 Ga old intrusions of monzonitic to syenitic composition and possibly related gabbroic to anorthositic units exist (Bergman et al., 2001). Many of them have the character of ring dykes and locally they form concentric composite intrusions with a mafic to intermediate composition (Martinsson, 2004a).

The Transscandinavian Igneous Belt has been suggested to have formed in response to eastward subduction (Wilson, 1980; Nyström, 1982; Andersson, 1991; Romer et al., 1992; Weihed et al. 2002), possibly during a period of extensional conditions (Wilson et al., 1986; Åhäll and Larsson, 2000).

3. Metamorphism, deformation and regional alteration

Several ductile and brittle deformation events, spanning between c. 1.88–1.74 Ga, have affected the Norrbotten area during the Svecokarelian orogen. Radiometric data with relevance for the metamorphic evolution are relatively sparse and the spatial and temporal distribution of discrete events is poorly understood. Evidence for an early event of regional metamorphism and deformation at ~1.88 Ga in Northern Norrbotten are given by the difference in the intensity of deformation shown by intrusions of known age of the Haparanda Suite and the Perthite Monzonite Suite (Bergman et al., 2001). This is supported by the occurrence of 1.88 Ga titanite of metamorphic origin in andesite from the Sammakovaara Group (Martinsson, 2004a), and other age data from the Kiruna area (Bergman et al., 2001). A second, possibly more local, metamorphic event at 1.86–1.85 Ga was demonstrated by zircon and monazite data in the Pajala area (Bergman et al., 2006), and an even younger phase at around 1.81–1.78 Ga is recorded by other chronological data from the same area (Bergman et al., 2006; Martinsson, 2004a; Niiranen et al., 2007). The latter event appears to have a regional extent and is clearly temporally related to the 1.8 Ga Lina Suite. A mainly brittle type of deformation and low grade metamorphism at around 1.75–1.74 Ga is evidenced by age results from the Pajala and Gällivare regions (Bergman et al., 2006). Limited P-T determinations indicate varying metamorphic conditions and metamorphic jumps across major shear zones. The metamorphic grade varies generally from upper greenschist to upper amphibolite facies and is mainly of low to intermediate pressure type (Bergman et al., 2001).

The ductile deformation includes at least two phases of folding and also involves the formation of major crustal scale shear zones. The intensity of deformation varies from a strong penetrative foliation to texturally and

structurally well preserved rocks which occur both on a regional and local scale. Axial surface traces of the folds mainly trend in a SE or a SSW direction (Bergman et al., 2001). The age relationship between these two folding events is not well constrained but the SE-directed folds have been interpreted to be the oldest (Bergman et al., 2001; Martinsson, 2004a). Major ductile shear zones are represented by the NNE-directed Karesuando-Arjeplog deformation zone, the N to NNE-directed Pajala shear zone and the NNW-directed Nautanen deformation zone (Fig. 1). Radiometric evidence suggests that these major shear zones were active at ~1.8 Ga, but the zones may have developed prior to 1.8 Ga and at least some of them are supposed to represent reactivated major crustal structures related to the Karelian rift event (Martinsson, 2004a). In general the shear zones in the western part show a western-side-up movement, whilst an eastern-side-up movement is typical for the eastern part (Bergman et al., 2001).

Several types of regional hydrothermal alteration have affected the Precambrian bedrock in Northern Norrbotten and a spatial relation to major shear zones has been noticed for several of them. Most characteristic for the area is scapolite, albite, biotite and carbonate alteration. Other types of alteration involve skarn, K feldspar, sericite, tourmaline, epidote and chlorite (Bergman et al., 2001). There are also more local, ore-associated types of alteration which are characterised by similar mineral assemblages. This type is further discussed in the ore descriptive section below.

Scapolite is widely distributed on a regional-scale and occurs in most intrusive and supracrustal rocks, except for the Archaean domain (Bergman et al., 2001). Intense scapolite alteration can mostly be related to shear zones, contacts of intrusive rocks, and mineral deposits. Strong scapolitisation in certain stratigraphical units of the Kiruna Greenstone Group is suggested to represent the metamorphic expression of former evaporite beds (Martinsson, 1997; 2004a). Regardless of host rock, character and style of scapolite growth the composition is mostly marialitic (Frietsch et al., 1997).

Albite occurs as an alteration mineral in several types of rocks. Some granites belonging to the Perthite Monzonite Suite have an albite-rich rim and mafic sills within the Kovo Group may have an albite-rich central part. Within the Kiruna Greenstone Group the basalts generally have a spilitic character and more intense albitisation have developed within volcanoclastic rocks along the contacts of mafic sills (Martinsson, 1997; Bergman et al., 2001). Extensive albite-carbonate alteration is recognised at many places along the major deformation zone (KADZ) running from Karesuando to Svappavaara and further towards SSW. It affects greenstones, porphyries and in some cases also intrusive rocks, and a number of sulphide occurrences exist in close relation to the shear zones (Ödman, 1939, Eriksson and Hallgren, 1975).

4. Analytical procedures

The geochronological work is based mainly on titanite occurring as part of hydrothermal alteration assemblages related to mineralisation. Titanites were separated from drill cores and hand specimens containing relatively coarse-grained (0.5-4 mm) titanite, which had been macroscopically identified in ore-related parageneses. In most cases it was possible to hand-pick fragments of titanite crystals from coarsely crushed material, but from some samples titanite, and also rutile, were extracted by conventional heavy-liquid and magnetic techniques. The minerals selected for age determinations were basically treated in the same way; washed in acetone, weak HNO₃, and finally rinsed in distilled water. A ²³³⁻²³⁶U/²⁰⁵Pb tracer solution was added to

each sample, which was subsequently dissolved in Teflon bombs at c. 200°C for 3–5 days, using a mixture of concentrated HF and HNO₃ (Krogh, 1973). Eventually, concentrated HCl was also added to the solution for complete dissolution. After evaporation and uptake in HBr, an initial ion exchange step was carried out from which purified Pb aliquots resulted. The uranium fractions went through a second ion exchange procedure in HCl where the remaining Fe was eventually removed. Finally, Pb samples were loaded on single filaments, whilst the uranium samples were loaded using a double-filament arrangement, and the appropriate isotopic ratios were measured on an upgraded Finnigan MAT 261 (TIMS) spectrometer. The chemical procedures and mass spectrometry were carried out at the Department of Geological Sciences at the Swedish Museum of Natural History (NRM) in Stockholm. A software package from Ludwig (1991a,b) was used to calculate and plot relevant ages and associated errors.

Fluid inclusions have been studied in quartz and tourmaline from several deposits, and the analyses were accomplished at the Department of Geological Sciences, Stockholm University. Microthermometric measurements were made using Linkam THM and Chaixmeca heating/freezing stages (-180° to +600°C) on doubly polished 150 µm thick sections of ore samples. Calibration was achieved by measurements on synthetic fluid inclusions (SynFlinC) and high-purity melting-point standards (Merck). To identify solid phases in fluid inclusions and to verify gas compositions as indicated by microthermometry, a Dilor XY laser Raman spectrometer was used. The Raman instrument, which utilized the 514.5 nm line from an Innova 70 argon laser as the excitation source, was wave-number-calibrated with a neon lamp and a silicon standard.

5. Isotope and fluid inclusion results

5.1. U-Pb results

Titanites, and occasionally also epidote and rutile, have been selected from various ore and alteration parageneses (Table 1). Titanites from epigenetic Cu±Au deposits are typically reddish brown, whereas those from samples from Kiruna-type apatite-Fe ores and alteration assemblages tend to be more honey-coloured or have a light-brown colour. Obtained U-Pb data are found in Table 2 and displayed in Figure 3A-D. U-Pb data are relatively concordant (with some exceptions) and define a large range in ²⁰⁷Pb/²⁰⁶Pb ages between c. 1910 and 1740 Ma. Whenever there is a tendency for discordance, it is implied that listed Pb/U and Pb-Pb ages are minimum estimates. Many previously obtained U-Pb titanite and zircon ages from the Swedish Early Proterozoic bedrock have given discordant results and their calculated lower intercept ages are typically in the interval 200–400 Ma (cf. Larson and Tullborg, 1998), suggesting a disturbance resulting in a radiogenic Pb loss at that time. Mobility of lead in this time interval is further reinforced by other isotopic studies (e.g. Romer, 1990; Romer and Wright, 1993). By forcing discordant data points through a fixed lower intercept at 300±150 Ma, it is possible to further constrain the U-Pb ages, see Table 2. It can also be noted that the common Pb contents in the analysed titanites are variable and commonly ²⁰⁶Pb/²⁰⁴Pb ratios are less than 1000, in agreement with many other titanite studies. We believe that the common Pb component was mainly introduced at the time of titanite crystallisation

(and the common Pb correction is done according to the indicated upper intercept age), although mobilization of lead into cracks at some time post-dating the primary crystallization cannot be excluded.

Tab. 1

It is clear from Figure 3 that data from the same locality occasionally comprise significantly different ages, and it is challenging to actually relate a certain age to a specific event. Because of the age complexity, as also shown by variable Re-Os data from a specific deposit (see below), and the likeliness that similar processes operated over extended periods of time, it is not always relevant to speak about specific ore formation dates. Rather, the timing is better suited to discuss with respect to age intervals. Regardless of ore type, the chronological data with bearing on ore formation comprise an essentially bimodal pattern with ages in the 1.88-1.86 Ga and 1.79-1.76 (1.74) Ga intervals, respectively. For instance, it appears that, although there are relatively few titanite data obtained from apatite iron ores (AIO), two age modes are indicated (Fig. 3A); 1.88-1.87 Ga (KUJ-278: 1882 ± 5 Ma and Mert-5: 1882 ± 4 Ma) and 1.74 Ga (Saivo, regression of two points: 1741 ± 4 Ma), respectively. Data for Cu (-Au) mineralisations have been split with respect to the character of the host rock; those hosted by Svecofennian rocks (Fig. 3B) provide a strong tendency for the existence of two age groups. The older group is defined by U-Pb titanite data from Pikkujärvi and Rakkurijärvi, whereas titanite data from Lieteksavo and Gruvberget comprise the younger age group. A similar division is indicated for the greenstone-hosted mineralisation (Fig. 3C), with age data from Veikavaara, Särkivaara, and Jälketurkkio falling between ca 1800-1770 Ma, whereas one analysis from Ö. Äijijärvi fits the older age interval. Finally, data for titanite samples separated from paragenesis interpreted as reflecting alteration and metamorphic assemblages (Fig. 3D) display the same general pattern as shown by data from ore-associated samples; i.e. in this case a spread between ca 1.9 and 1.72 Ga.

Tab. 2

Figure 3

5.2. Fluid inclusion results

The IOCG deposits in Norrbotten contain a variety of different fluid inclusion types (see Table 3 and references therein), but a common characteristic of all of them is populations of aqueous inclusions with high to medium concentrations of salts and of CO₂ inclusions. Based on phases present at room temperature a classification of the inclusions can be made into four main types: L+V+mS, L+V+halite, L+V and CO₂. Fluid inclusion investigations have previously been made on a number of deposits in Norrbotten (Broman and Lindblom, 1995; Lindblom et al., 1996; Martinsson et al., 1997a, 1997b; Harlov et al., 2002; Wanhainen et al., 2003; Edfelt et al., 2007; Gleeson and Smith, 2009) and important results from these locations are summarized in Table 3, together with new data from the deposits at Pikkujärvi, Gruvberget-Cu and Lieteksavo. From the Pikkujärvi deposit fluid inclusions were studied in quartz and tourmaline from chalcopyrite- and pyrite mineralised quartz veins, from the Gruvberget-Cu deposit in chalcopyrite-bearing quartz veins and from the Lieteksavo deposit in quartz from chalcopyrite mineralised quartz-tourmaline veins.

Microthermometry on all types of the salt-rich aqueous inclusions points to a complex composition of the liquid phase. First melting of the aqueous phase occurred in the range -50°C to -70°C which is similar to the

stable eutectic temperature of the NaCl-CaCl₂-H₂O system (-52°C) or close to the metastable eutectic temperature (~-70°C) reported for the same system (Davis et al., 1990). Even though it is difficult to interpret the first melting temperature unambiguously, it is inferred that the most likely composition is an aqueous solution dominated by NaCl and CaCl₂. The salinity of the halite-bearing inclusions is estimated from the halite melting temperature (Sterner et al., 1988) and for halite-free inclusions from the temperature of final ice melting using the data for CaCl₂-rich CaCl₂+NaCl compositions (Oakes et al., 1990).

Tab. 3

Type L+V+mS inclusions are interpreted to be an early population and to represent an initial stage of pyrite-chalcopryrite mineralisation. This inclusion type is found in samples from Pahtohavare, Pikkujärvi, Kallosalmi, Tjårrojåkka-Fe and in late sulphide and hematite-bearing quartz veins in apatite-Fe-oxide deposits (Table 3). The inclusion type comprises multisolid aqueous fluid inclusions which contain a liquid, a vapour and more than two solid phases (Fig. 4). The vapour phase occupies about 2 vol.% of the inclusion volume. Two of the solid phases are identified as halite and sylvite from their cubic shapes, translucency, non-birefringence and behaviour upon heating. A round birefringent solid with a high relief is identified as calcite and an orange coloured solid as hematite from their characteristic Raman spectra. Small unidentified solid phases are occasionally present. Halite and calcite are always the largest of the solids. The inclusions have a primary random appearance in the host minerals with a size that can reach 35 µm, but the typical size is less than 10 µm.

For the Pikkujärvi samples a partial homogenisation of the liquid and vapour to the liquid state took place between 145° and 200°C. Commonly on heating, inclusions decrepitated before complete dissolution of sylvite and halite was reached. For such inclusions an approximate salinity has to be estimated from the phase volumes of the salts (Roedder, 1984). In those inclusions where dissolution of sylvite and halite could be observed, sylvite dissolved in the range 131° to 170°C, and halite between 312° and 450°C. These temperatures indicate salinities of 39 to 53 wt.% NaCl eq.

Figure 4

Type L+V+halite inclusions are always associated with chalcopryrite and indicate that this type represents the fluid during the main stage of copper deposition. This type of inclusions is present in samples from Viscaria, Pahtohavare, Kiskamavaara, Nautanen, Kallosalmi Aitik, Tjårrojåkka, Lieteksavo, Gruvberget and in late stage sulphide- and hematite-bearing quartz veins post-dating apatite Fe-oxide deposits (Table 3). The inclusions consist of an aqueous liquid, a vapour bubble together with a cubic solid phase and sometimes an additional solid with a round shape is also present. The solid with a cubic outline, is translucent and non-birefringent and is identified as halite, whilst the round solid is birefringent and identified as calcite by its characteristic Raman spectrum. Primary inclusions with uniform phase ratios appear randomly scattered in quartz, other inclusions with varying phase proportions occur as secondary in trails along healed microfractures. The size of the inclusions is generally less than 10 µm, but some may be up to 25 µm.

On heating the inclusions from Gruvberget-Cu, a partial homogenisation of the liquid and vapour to the liquid state was measured between 109° and 240°C. Two groups may be distinguished, a first group between 109° and 131°C and a second between 150° and 240°C. The inclusions from Lieteksavo displayed liquid-vapour homogenisation between 143° and 241°C. Dissolution of halite in inclusions from Gruvberget-Cu, in the presence of calcite, took place in the range 224° to 415°C indicating a salinity of 33 to 47 wt.% NaCl eq.

Similarly, for the inclusions from Lieteksavo this occurred between 350° and 373°C giving a salinity of 41.7 to 43.6 wt.% NaCl eq. The dissolution temperature of halite in secondary type 2 inclusions from Lieteksavo form two groups; one where halite dissolved in the range from 134° to 206°C, corresponding to salinities of 29–32 wt.% NaCl eq, and another one where the inclusions decrepitated before halite dissolution was reached. However, visual estimates from the size of the halite in the latter, suggest a salinity of at least 50 wt.% NaCl eq.

Type L+V inclusions are normally associated with bornite and consist of an aqueous liquid and a vapour phase. This type of inclusions occurs in samples from Pahtohavare, Kallosalmi, Lieteksavo, Aitik, Gruvberget and Tjärrojjäcka (Table 3 and references therein). The inclusions have a secondary appearance along well-defined healed microfractures in quartz. Most of the inclusions are below 20 µm in size.

Homogenisation to the liquid state was measured between 109° and 215°C for inclusions from Gruvberget-Cu and between 139° to 226°C for inclusions from Lieteksavo. After freezing the inclusions from Gruvberget-Cu showed final ice melting between -19.9° and -23.9°C, corresponding to a salinity of 20.5 to 22.4 wt.% CaCl₂+NaCl eq. The inclusions from Lieteksavo yielded final ice melting between -22.2° and -25.4°C, corresponding to a salinity of 21.6 to 23.0 wt.% CaCl₂+NaCl eq.

Type CO₂ comprises carbon dioxide inclusions. This type of inclusions is present in samples from Pahtohavare, Lieteksavo, Aitik, Nautanaen and Gruvberget (Table 3). A CO₂-phase has also been observed in some copper-mineralised quartz veins which post-date apatite-Fe-oxide ores in the Per Geijer deposits (Broman and Lindblom, 1995; Gleeson and Smith, 2009). The inclusions appear as clusters, in many places connected to grain boundaries, or along healed microfractures. The CO₂ phase in the clusters is always CO₂ liquid, whereas in healed microfractures the CO₂ consists of two phases, CO₂ liquid plus gas. The majority of the inclusions are less than 10 µm in size; individuals may be up to 25 µm.

After freezing, the inclusions from the Gruvberget-Cu deposit displayed melting of the CO₂ in the range -56.8° to -57.2°C, and homogenisation of the CO₂-phases between -26.5° and +27.4°C. The temperatures give a mean composition of CO₂ with <4 mole% CH₄. Melting of the CO₂ in samples from the Lieteksavo deposit was observed at -56.6°C for inclusions in clusters and at -58.0°C for inclusions in healed microfractures. Homogenisation of the CO₂-phases occurred in the range -2.1° to +16.8°C for the inclusions which appear in clusters and between +15.5° and +24.1°C for inclusions in the healed microfractures. The temperatures indicate two compositional groups; the inclusions in clusters are composed of pure CO₂, whereas the inclusions in the healed microfractures are composed of 90 mole% CO₂ and 10 mole% CH₄.

6. Fe oxide and Cu±Au±Fe oxide deposit types in Northern Norrbotten

Reviews of Fe-oxide Cu-Au deposits in the Northern Norrbotten Ore District in northern Sweden have been published by Carlon (2000) and Billström et al. (2010). The economic deposits are quite variable in character and have traditionally been divided into four major styles of economic mineralisation: stratiform sulphide ores, skarn-rich iron ores, Kiruna type Fe oxide deposits (apatite iron ores) and epigenetic style Cu±Au deposits, as well as subeconomic banded iron formations (BIF). Most of the stratiform-stratabound sulphide occurrences are dominated by Cu, but Zn and Pb are the most important metals in a few cases. Below, the focus

will be on other ore types than the stratiform-stratabound sulphide-rich type (e.g. the Viscaria Cu deposit), and only a few general features will be reiterated in the discussion. Also the epigenetic sulphide deposits are generally Cu-dominated with Au as an additional economic metal in some cases. Less common are high contents of Co or Mo. The skarn-rich iron ores, frequently associated with typical BIF, have traditionally been called skarn iron ores (e.g. Grip and Frietsch, 1973) although they are suggested to represent metamorphosed syngenetic iron deposits. To avoid confusion about their origin (cf. Meinert et al. 2005) they are here named Mg-rich iron formations as they are characterised by magnetite with a high Mg-content (up to 6.2% MgO) and have Mg-rich silicates as typical gangue minerals.

In the following sections the focus will be on deposits containing Fe-oxides and Cu-Au in different proportions including iron formations, apatite iron ores and epigenetic Cu-Au deposits. All of these can display characteristics of IOCG style of mineralization whilst the stratiform base metal deposits are not included in this context. The syngenetic deposits, with a few exceptions, are hosted by the Karelian greenstones whereas the Kiruna type magnetite-apatite deposits occur within the Svecofennian rocks. The epigenetic Cu±Au occurrences include a variety of deposits regarding mineralisation styles and host rocks and a sub-division based on host rock (Karelian or Svecofennian age) will be used. In order to keep the descriptive part relatively condensed only certain major deposits and others of considerable genetic interest are described in some detail below. Further information of the discussed occurrences and their characteristic features are listed in Table 4 and illustrated in Figs. 5 and 6, and their locations are shown in Fig. 1.

Table 4

Figures 5 and 6

6.1. Iron formations

Iron formations occur mainly in the Karelian greenstones but a few examples are also known from Svecofennian units (Table 4 and Fig. 1). In both settings these deposits are mostly low in iron content (20–45%) and relatively small in size (< 100 Mt). In the Karelian rocks, iron formations are extensively developed as up to 200 m thick units in the upper part of the greenstone successions. Based on mineralogy and internal structure the Karelian deposits are divided into BIF (banded iron formation) of dominantly silicate facies and Mg-rich iron formations.

6.1.1 BIF and transitional BIF - Mg-rich iron formations

BIF occur as tabular and layered rock units composed of iron rich silicates or iron oxide and alternating layers of chert. They contain 15–30% Fe and many of them are manganiferous with up to 4% MnO (Martinsson, 2004a). Among this type, the greenstone-hosted *Käymäjärvi* occurrence (Fig. 1) is the largest with a thickness of 150–200 m and a length of at least 18 km. At some places, typical Mg-rich iron formations grade into BIF towards the hanging wall and/or laterally, suggesting that these two types of iron formations are genetically related (Grip and Frietsch, 1973). One of the best examples is the *Tornefors* deposit at Junosuando in northeastern Norrbotten (Figs. 1 and 7) where Mg-rich iron formation containing 25–35% Fe is overlain by typical BIF mineralisation with 15–25% Fe. Other deposits exhibiting a similar transition from Mg-rich iron

formation to BIF include *Marjajärvi* in the Pajala area (Fig. 1) and occurrences in the Masugnsbyn area (Martinsson and Wanhainen, 2013).

Figure 7

Iron formations in Svecofennian units are mainly found in western Norrbotten, most notably in the Jokkmokk and Arjeplog areas, but also southwest of Gällivare, e.g. the *Peltovaara* deposit (Grip and Frietsch, 1973). Ores are associated with felsic to intermediate volcanoclastic and sedimentary rocks.

In the Jokkmokk area, magnetite±hematite mineralisation occurs interlayered with quartz-rich bands and less commonly garnet, epidote, pyroxene and amphibole representing strongly metamorphosed BIF. The host rocks are highly metamorphosed felsic to intermediate volcanoclastic rocks locally containing intercalations of carbonate rocks and calc-silicates. The iron content varies between 25–40%, and the largest occurrence is *Kallak*, containing 92 Mt with 35–38% Fe, 0.04% P, and 0.4% Mn (Frietsch, 1997).

6.1.2. Mg-rich iron formations

The most common ore type in the Karelian greenstones is the Mg-rich iron formations. These form stratiform-stratabound lenses with a length of 600–1300 m, a thickness of 20–150 m and typical sizes between 5–60 Mt. Most of them occur in the stratigraphically upper part of the Karelian greenstones within the Linkaluoppal and Vinsa formations in association with mafic tuff-tuffite, black schist, BIF and dolomite. Some Mg-rich iron formations have a pronounced banded or layered character with the ore minerals occurring, in part, as thin layers (e.g. *Sautusvaara* and *Vathanvaara*), whereas a less pronounced mineralogical banding is common for most of the deposits. The average composition of most Mg-rich iron formations is 30–45% Fe, 1–3% S, 0.02–0.1% P and 0.02–0.08% Cu (Grip and Frietsch, 1973). Magnetite, serpentine, diopside, and tremolite are the main constituents of these deposits and magnetite occurring in association with serpentine is generally rich in Mg with 2 to 6% MgO (Frietsch, 1985; Wanka, 2009; Hasa et al., 2013). Pyrite and pyrrhotite are common minor components, occurring disseminated and in veinlets together with accessory amounts of chalcopyrite. Alteration is mainly developed in the footwall, with the formation of biotite or chlorite and locally scapolite (Bergman et al., 2001).

The *Sahavaara* deposit, situated northwest of Pajala (Fig. 1), comprises three lenses (Stora Sahavaara, Södra Sahavaara and Östra Sahavaara) of Mg-rich iron formation. This deposit occurs in the upper part of the Veikavaara greenstones at the contact between volcanoclastic rocks in the footwall and Svecofennian epiclastic to volcanoclastic rocks in the hanging wall. The footwall tuffites are mainly basaltic in character and have thick intercalations of graphitic schist. The sedimentary rocks in the hanging wall have a rather uniform character and are likely to be predominantly of an andesitic volcanoclastic origin with minor intercalations of quartz arenites (Martinsson, 1995; Bergman et al., 2001).

The ore zone at Stora Sahavaara is up to 80 m thick and 1300 m in length and contains reserves of 80.3 Mt @ 42.7% Fe (Baker et al., 2011). It consists of serpentine-rich magnetite ore with pyrite, pyrrhotite and some chalcopyrite occurring disseminated as minor constituents in the ore (Lundberg, 1967). A Ca-Mg silicate unit usually caps the ore, and displays the sequence serpentine-diopside-tremolite, and locally chert, stratigraphically upwards. The highest content of Fe and S is found in the central part of the deposit, and this area is also slightly

enriched in Cu and Co. Thus, the Sahavaara deposit exhibits a pattern similar to many other Mg-rich iron formations of decreasing contents of Fe, Cu, and Co combined with an increase of Ca, Si, P, and Mn towards the top and peripheral parts of the ore (Martinsson, 1995).

An alteration zone containing scapolite and biotite extends to about 200 m below the Stora Sahavaara ore and it encloses the smaller Östra Sahavaara zone. Noting that the presence of scapolite-rich graphite schists is a regional feature at this stratigraphic position scapolite may not be a product of ore related alteration. The stratigraphically lower part of the alteration zone contains variable amounts of biotite, resulting in enrichment of Mg, K, Ba, and Rb, whilst Na, Ca, and Sr are slightly depleted (Martinsson, 1995).

The *Vathanvaara* deposit in the Vittangi area (Fig. 1) differs in part from the general characteristics of Mg-rich iron formations. In its central part it changes from a laminated and almost massive pyrite-pyrrhotite sulphide facies at the footwall to normal Mg-rich iron formations towards the hanging wall and at its peripheral parts. The *Tapuli* deposit in the Pajala area (Fig. 1) is exceptional in its irregular character. It is hosted by a thick dolomite at the top of the Karelian greenstones with magnetite forming irregular bands, veins and patches within an up to 350 m wide Ca-Mg silicate zone. Locally, almost massive pyrite-pyrrhotite-chalcopryrite veins occur with high Cu-grades (up to 5% Cu) considerably higher than that of the surrounding magnetite ore (less than 0.03% Cu; Denisová et al., 2013). *Tervaskoski* is the largest example of Mg-rich iron formation being enriched in Cu, carrying 30 Mt with 48% Fe, 3% S and 0.1% Cu. A few deposits (*Tjavelk* and *Vieto*, Fig. 1) are anomalous by their high phosphorus content (Grip and Frietsch, 1973).

6.2. Kiruna-type apatite iron ores

Kiruna-type apatite iron ores occur only in Svecofennian rocks and are geochemically different from iron formations. This is clearly seen in Fig. 8 in which the Fe-P-S distribution and the Cu and Au grades are compared showing the apatite iron ores to have higher contents of P whilst the content of S and Cu is lower. Kiruna is the type area for apatite iron ores and the *Kiirunavaara* deposit is the largest and most well known example. In total, about 40 apatite iron occurrences are known from the Northern Norrbotten Ore District, and approximately 2330 Mt of ore have been mined from 11 deposits during the period 1888-2014. Individual deposits have an average content of Fe and P varying between 30–68% and 0.05–5%, respectively. Characteristic is the high contents of V (300–3000 ppm) in the iron oxides in contrast to the relatively low content of Ti (100–600 ppm). Co and Ni are mostly in the range of 20–200 ppm and 50–300 ppm, respectively, and the Cr content is generally less than 100 ppm (Frietsch, 1970). The REE are usually strongly enriched in the ores and occur in apatite, monazite and allanite (Parak, 1973; Harlov et al., 2002; Martinsson et al., 2012). Generally there is a pronounced LREE enrichment and a strong negative Eu-anomaly (Frietsch and Perdahl, 1995). Besides REE, V and Ti, also several other trace elements (Th, Nb, Zr) are slightly enriched in the ores (Martinsson, 2003). The contents of Cu and Au are mostly very low (Cu < 100 ppm and Au < 10 ppb), except for local occurrence of sulphide bearing veins containing chalcopryrite, bornite and pyrite.

Figure 8

The apatite iron ores exhibit a considerable variation in mineralisation style, host rock lithology, alteration, P-content and associated minor components (Table 4). It is possible to distinguish two rather distinct groups of deposits, (1) a breccia type and (2) a tabular, stratiform-strata-bound massive type. A third and less distinct group of deposits is intermediate in character (Martinsson, 1994; Bergman et al., 2001). In addition to these three types there are a few other examples, which fall outside this three-fold division. In most areas the apatite iron ores are not strongly affected by deformation and metamorphic overprinting, other than the Leveäniemi and Malmberget deposits, both of which occur close to larger intrusions of Lina granite.

6.2.1 Breccia-style apatite iron ores

These ores are mainly hosted by intermediate to mafic volcanic rocks, in a stratigraphically low or middle position of the Kiirunavaara Group, or within the underlying Porphyrite Group. Amphibole is always present as a minor component and accessory amounts of pyrite, chalcopyrite, and titanite may be present. Alteration of the host rocks is mainly expressed by the presence of albite, actinolite and scapolite. Characteristic for the breccia-type deposits is a low P-content (usually in the range 0.05–0.3 %), average iron content of only ~30 %, and mostly with magnetite as the only iron oxide.

The *Mertainen* deposit (Fig. 1) is hosted by trachyandesitic lava and is calculated to contain 166 Mt with 35% Fe and 0.05% P (Grip and Frietsch, 1973; Lundberg and Smellie, 1979). It has the character of a large breccia, containing massive lenses or veins of magnetite in its central part that is surrounded by zones with successively lower magnetite content (Fig. 9). Magnetite is accompanied by actinolite, and locally some apatite and calcite. The host rock of the ore is sodic in character with albite and scapolite as the most important alteration minerals. Outside the mineralized breccia zone the volcanic rocks are generally potassic in character with K feldspar as the main feldspar mineral (Martinsson and Virkkunen, 2004). The *Renhagen* deposit south of Kiruna is similar in character, with magnetite occurring as breccia infill, but also disseminated and as amygdule filling in albite altered trachyandesitic lava.

Figure 9

6.2.2. Stratiform-stratabound style apatite iron ores

The other main group of deposits form stratiform-stratabound, tabular lenses at stratigraphically high positions within the Kiirunavaara Group. Hematite is a major iron oxide in this ore type together with varying amounts of magnetite and apatite (the P-content varies from 1 to 5%). However, in this deposit type, hematite is generally formed from magnetite and is only rarely present as a primary component. Amphibole is generally absent, and the gangue minerals are mainly apatite, quartz, and carbonate. Host rock alteration is common including sericite, chlorite, biotite, tourmaline, and carbonate as the most typical products and the hanging wall rocks may also be strongly silicified and K feldspar altered. Sulphides are rare and mainly found in the altered footwall and in crosscutting quartz veins within the ores. Ore of breccia-style is lacking or mainly restricted to the footwall in this type of deposit. Included in this group are the *Per Geijer* ores (i.e. Nukutus, Henry, Lappmalmen, and Rektorn) in the central Kiruna area.

The *Rektorn* ore (Fig. 1) consists of magnetite and hematite in different proportions. Apatite occurs disseminated in the ore, enriched in bands or as crosscutting veins and breccia infill. Quartz and carbonate are, in

places, important constituents of the apatite-rich ore and occur interstitial to apatite or as small patches (Geijer, 1950). The ore shows large variation in textures and grain sizes and mineral composition including skeletal ore and apatite-carbonate rich segregations.

The hanging wall rocks are heterogeneous in character, strongly sheared and belong to the Matojärvi Formation of the Kiirunavaara Group. The Rektor Pophyry, occurring in the lowest part, is a strongly altered rhyolitic unit that varies from a massive K feldspar rock to a strongly silicified rock containing spherulitic aggregates of K feldspar and disseminated hematite. Sericite commonly occurs as a late-stage pervasive alteration phase in association with shear zones, but also occurs as patches or lenses of massive sericite containing radiating aggregates of tourmaline. Intercalations of siliceous, apatite-poor hematite ore with a fragmental character are common in the Rektorn porphyry and may reach several meters in thickness and may contain some barite (Geijer, 1910). Several carbonate-bearing quartz veins which formed in response to shearing crosscut the ore at a high angle. These veins may contain minor amounts of hematite, allanite, pyrite, chalcopyrite and bornite.

6.2.3. Intermediate-style apatite iron ores

Some deposits (e.g. Kiirunavaara, Tjårojkka-Fe, Gruvberget-Fe, Malmberget, and Leveäniemi, see below) share features with both of the main groups of apatite iron ores described above. These are mainly stratabound in character but also comprise ore breccias developed along the wallrock contacts. Magnetite is the dominant, or only, iron oxide. Amphibole is a characteristic minor component and titanite may be present in accessory amounts. The Fe content is high (55–68%), and the average P-content is rather low, although it can vary considerably (from 0.02 to >5% P) within individual deposits. The alteration of host rocks includes the formation of amphibole, albite, biotite, and locally scapolite proximal to ore deposits with more potassic alteration occurring more distal in some occurrences.

The huge *Kiirunavaara* deposit (Figs. 1 and 10) was found in outcrop 1696, but regular mining did not start until 1900 when a railway was built from the Norwegian coast to the town of Kiruna (Grip and Frietsch, 1973). At the end of 2014 the total production of crude ore from open pit and underground mining was 1539 Mt with ore reserves of 656 Mt@47% Fe (LKAB, 2014). The deposit defines a tabular ore body which is ~6 km long, up to 180 m thick, and it extends down to at least 1500 m below the surface. It is situated in the middle part of the Kiirunavaara Group and follows the contact between a thick pile of trachyandesitic lava (Hopukka Fm.) and overlying porphyritic rhyodacite (Luossavaara Fm.). The much smaller Luossavaara ore is situated 1 km further north in a similar stratigraphic position (Fig. 10).

Figure 10

Trachyandesite lava occurs as numerous lava flows, which are rich in amygdules close to the flow tops. A thick sill varying in composition from gabbro to monzonite has intruded the lavas ~1000 m stratigraphically below the ore. Several dykes of granophyric to granitic character cut the ore and a larger body of K-rich granite is found at deeper levels in the mine on the footwall side of the ore. Rhyodacites in the hanging wall are pyroclastic in character and include vesicular lapilli tuff (Martinsson and Hansson, 2004).

Magnetite-actinolite breccias are developed both in the footwall and hanging wall along the contacts of the main Kiirunavaara ore body. In the footwall larger breccia zones may show a change from veined

trachyandesite to breccia with angular to rounded fragments of the wallrock cemented by magnetite±actinolite. Close to the hanging wall contact the ore is often rich in angular to subrounded clasts of rhyodacitic tuff. Veins of magnetite and actinolite may extend tens of meters into the hanging wall and locally form rich ore breccia or smaller lenses of massive ore (Martinsson and Hansson, 2004).

The phosphorus content of the ore exhibits a pronounced bimodal distribution with either < 0.05% P or > 1.0% P. Most of the apatite-poor ore (B-ore) forms slightly irregular and branching bodies of massive and fine-grained (<0.3 mm) magnetite ore. These bodies are usually 40–70 m thick and contain up to 15% of disseminated actinolite in 5–20 m wide zones along the borders, except when in contact with apatite-rich ore. Larger lenses of the B-ore can have a slightly more coarse grained central part enriched in carbonate and small amounts of pyrite (Martinsson and Hansson, 2004). Apatite-rich ore (D-ore) is mainly found towards the hanging wall and in the peripheral parts of the ore body, but also occurs in varying amounts at the footwall contact. Lenses of D-ore within and between B-ore bodies may contain clasts of the wall rocks. The age relationship between B- and D-ore is ambiguous, and both ore types can, at a small scale, be seen cutting each other. However, the lack of actinolite in B-ore in contact with D-ore suggests the latter ore type existed prior to the emplacement of B-ore, which prohibited silica assimilation from the local wall rocks and the formation of silicates. In addition, in some places B-ore that is lacking wall rock clasts are enveloping lenses of D-ore which contain clasts, supporting an earlier formation of the D-ore (Martinsson and Hansson, 2004). Columnar and dendritic magnetite is seen as locally developed textures, which suggest rapid crystallization in a supercooled magma, (Geijer, 1910; Nyström, 1985; Nyström and Henriquez, 1989). Minor constituents of the ore include disseminated anhydrite, monazite, allanite and titanite (Martinsson et al., 2012). Veins of anhydrite, anhydrite-pyrite-magnetite and coarse-grained pyrite are locally encountered in the ore and wall rocks, whilst gypsum may occur as narrow fissure fillings in the ore (Martinsson and Hansson, 2004; Nordstrand and Andersson, 2013).

Strong and variable alteration has affected the host rocks to the Kiirunavaara ore body and some of these might be of more regional character whereas others are clearly related to the ore formation. Extensive albitisation is developed in the footwall to the Kiirunavaara deposit and the most intense albitisation surrounds a gabbroic to monzonitic sill, which itself is strongly altered. Especially the amygdaloidal parts of the lava flows are strongly altered and the albitisation is accompanied by secondary magnetite, actinolite, titanite, and locally some tourmaline (Geijer, 1910). A U-Pb age of 1876 ± 9 Ma was obtained from titanite as amygdale infill in albitised trachyandesite lava, which is considered to be the age of albitisation (Romer et al., 1994; Westhues et al., 2013). Actinolite is a common alteration mineral at both the footwall and hanging wall contacts and it may occur as up to m-wide massive zones bordering the ore and containing some titanite and magnetite. Actinolite also replaces, partly or completely, clasts of wallrocks in the ore and in the ore breccia. Besides actinolite and magnetite veining close to the ore, the hanging wall is, in some areas, affected by biotite-chlorite alteration, which is commonly accompanied by disseminated pyrite and a weak enrichment of Cu, Co, and Mo. Sericite-schist with some tourmaline is locally found in the footwall contact of the ore (Geijer, 1910), and may be related to post-ore shearing.

The *Tjårojokka-Fe* deposit, located some 45 km SW of Kiruna (Fig. 1), is hosted by andesite belonging to the Porphyrite Group (Edfelt et al., 2006). Alteration is extensively developed in the wall rocks of the ore and includes early albitisation and scapolitisation overprinted by Ba-rich K feldspar alteration (Edfelt et al., 2005). The ore is calculated to contain 52.6 Mt of iron ore with 52% Fe and 0.5–1.3% P and consists of a massive

magnetite core surrounded by a breccia ore with 25-40% Fe and 0.4-3% P. The breccia consists of magnetite, carbonate, actinolite and apatite occurring disseminated and as breccia infill in the host rock and contains disseminated chalcopyrite with grades of up to 1% Cu (Martinsson, 1995; Edfelt, 2007). More extensive Cu mineralisation occurs 1 km WNW of Tjårrojåkka-Fe (see Tjårrojåkka-Cu)

The *Gruvberget* apatite iron ore (Fig. 1) is calculated to contain 64.1 Mt with 56.9% Fe and 0.87% P to a depth of ~300 m (Frietsch, 1966) and is since 2010 mined in an open pit. The bedrock consists of metavolcanic rocks of mafic to intermediate composition, probably belonging to the Porphyrite Group. The rocks have a granoblastic texture and they mostly lack primary structures and are strongly altered by albite, scapolite, and K feldspar (Martinsson and Virkkunen, 2004). A N-S directed shear zone is situated in the footwall to the ore and deformed mafic dykes cutting the ore indicate a sinistral movement during shearing (Lindskog, 2001). The apatite iron ore is mostly massive and consists of magnetite in the northern part and hematite in the middle and southern parts of the deposit (Fig. 11). The transition zone between them is a few meters wide, across which magnetite is gradually replaced by hematite. An extensive ore breccia with veins and schlieren of magnetite, hematite, apatite, and amphibole is developed in the footwall at the middle part of the deposit (Frietsch, 1966).

Figure 11

The *Malmberget* deposit (Figs. 1 and 12) consists of several ore lenses with 51–61% Fe and varying contents of phosphorus (Grip and Frietsch, 1973). As of the end of 2014, approximately 674 Mt of crude ore have been produced from open pits and underground workings and the reserves were estimated to 297 Mt with 42% Fe (LKAB, 2014). The ores are hosted by strongly metamorphosed and deformed volcanic rocks of felsic to mafic composition earlier referred to as “leptites” (Geijer, 1930). A porphyritic texture is locally preserved in the felsic rocks and amygdules are occasionally encountered in more mafic rocks, suggesting a mainly extrusive origin and a character similar to that of rocks of the less metamorphosed Kiirunavaara Group (Martinsson et al., 2013). A large intrusion of Lina granite exists northwest of the deposit and the recrystallization of the host rocks increases towards the granite contact. Dykes of granite, aplite and pegmatite are frequently found in the ores and their host rocks.

Brecciation at different scales is developed in the wall rocks of the ores and especially in the eastern part of the Malmberget area. Magnetite, apatite, pyroxene and amphibole are the main constituents of mm- to m wide veins that may develop into networks. Breccias with a high Fe content are mainly found adjacent to the massive iron ores, whilst breccias dominated by pyroxene-amphibole are more distal. Locally anhydrite is abundant together with various amounts of actinolite, magnetite and some sulphides. Magnetite composition and geochemistry of massive ore and ore breccia suggest different origins with crystallisation from a magmatic melt for the former and a hydrothermal origin of the latter (Lund and Martinsson 2013).

Figure 12

The Malmberget deposit is strongly affected by metamorphic recrystallization and ductile deformation, including at least two phases of folding. The present shape of individual ore bodies is mainly a result of stretching parallel to a fold axis dipping 40 to 50° towards SSW (Martinsson et al., 2013). The metamorphic

recrystallization of the ore has caused an increased grain size and the development of different oxidation textures including hematite porphyroblasts, ilmenite exsolution lamellas in magnetite and element redistribution between Fe-Ti oxide minerals (Lund and Martinsson, 2013).

Leveäniemi is the third largest apatite iron ore in Norrbotten and is calculated to contain 204 Mt of high-grade ore with approximately 64% Fe and 104 Mt of low-grade ore with 26% Fe. Both ore types have a P-content varying from 0.02 to 1.1% (Grip and Frietsch 1973). The ore is dominated by massive magnetite, but large volumes of ore breccia occur in the surrounding rocks in up to 100 m wide zones. In the middle part of the deposit, magnetite is largely altered to hematite. The bedrock consists of strongly metamorphosed rocks of volcanic and sedimentary origin, which are intruded by pegmatite and granite, and the immediate host rock is occasionally rich in biotite or scapolite, together with lesser amounts of muscovite, albite and actinolite (Bergman et al., 2001).

6.2.4. Unusual style of apatite iron ores

A few occurrences do not fit into the previously described groups of apatite iron ores. One example is the *Painirova* deposit (Fig. 1) which has the character of a breccia but is different from e.g. Mertainen by the partly coarse grained character, with up to 10 cm apatite crystals that have grown perpendicular to the walls of the magnetite-apatite veins (Lundberg and Smellie, 1979). This texture partly mirrors that of the *Saivo* deposit although its mineralogy and morphology is different. The small *Saivo* Fe-mineralisation (Figs. 1 and 13) is hosted by a coarse-grained pyroxene rock, which forms a 250 x 60 m lens with sharp contacts to surrounding aplitic monzonite and gabbro. The gabbro is partly affected by strong scapolite alteration and it is brecciated by the monzonite which itself is brecciated by the pyroxene rock. Magnetite mostly forms dendritic aggregates up to several 10's of cm that occur intergrown with variable amounts of titanite and these minerals, together with pyroxene, have grown perpendicular to the contact of the monzonite (Lehto, 1972; Martinsson and Wanhainen, 2000).

The *Narken* iron oxide deposits (Fig. 1), hosted by sedimentary and volcanic rocks equivalent to the Sammakovaara Group, form isolated apatite iron ore occurrences in the eastern part of Northern Norrbotten (Frietsch, 1972). They are small in size and have the character of breccia bodies or veins associated with extensive hydrothermal alteration (silicification, epidotisation and chloritisation) of the wall rocks. A typical feature of these deposits is euhedral, but abraded and partly broken, crystals of magnetite, apatite, and pyrite, occurring as breccia infill together with tabular hematite, epidote, chlorite and some quartz and chalcopyrite. Magnetite is generally strongly altered to hematite. These deposits are enriched in REE, and some of them contain up to 0.5% Cu but are low in Au. Magnetite and apatite in the breccia infill have a character similar to that of xenocrysts in magmatic rocks and may initially have formed as magmatic products in an apatite iron ore system that has developed into an explosive hydrothermal phase, bringing these crystals higher up in the crust, and forming breccia bodies cemented by hydrothermally precipitated hematite and epidote with local enrichment of Cu (Martinsson, 2009).

Figure 13

Several small occurrences of apatite iron ores, with a suggested metasomatic origin, occur southwest of Kiruna (Frietsch, 1997). These are generally related to shear zones and contain magnetite, or occasionally hematite, as dissemination, veins or breccia infill. Quartz may be an important gangue mineral, whereas amphibole, apatite and fluorite are less abundant. Related hydrothermal alteration includes the formation of albite, sericite, scapolite, and epidote (Frietsch, 1997).

6.3. Epigenetic Cu-Au±Fe oxide deposits

There is a large diversity in character, alteration, and host rock exhibited by the epigenetic Cu-Au±Fe oxide deposits in the Northern Norrbotten Ore District (Martinsson et al., 2008; Billström et al., 2010). They are mainly hosted by Karelian greenstones or by Svecofennian volcanic rocks, but occurrences are also found in sedimentary units. Some are spatially related to major shear zones, whilst others show a close spatial relation to major apatite iron ores, or are transitional in character to them. There are also cases where mineralisation is related to intrusives with major or minor parts hosted by felsic to intermediate plutons. Although generally not directly genetically related, mafic dykes are commonly spatially related to this type of mineralisation. The mineralisation styles include dissemination, breccia infill, veinlet, vein, and replacement, commonly with several styles present in a single deposit. Alteration is partly related to host rock character, with albite mainly developed in greenstone units, whilst K feldspar, sericite and tourmaline occur in Svecofennian rocks. Other alteration phases, such as scapolite, biotite and carbonate are not clearly constrained to certain rock units. Chalcopyrite and pyrite are the most common ore minerals, with magnetite occurring as a minor or major constituent. Pyrrhotite is less common, and occurs mainly in deposits where graphite schist is present in the host rock successions. Hematite, bornite, chalcocite, and molybdenite may be minor, or locally major, constituents of the ores, and are mainly found in deposits hosted by Svecofennian rocks and in vein-style mineralisations.

Given the large diversity among the occurrences of epigenetic sulphide mineralisation, these are not easily put into classification systems. We have simply chosen the type of host rock as a first-order discrimination factor; i.e. ores are described below as either Greenstone-hosted or Svecofennian-hosted.

6.3.1. Greenstone-hosted Cu±Au deposits

Among the deposits hosted by the Kiruna Greenstone Group, one may distinguish two types; carbonate-quartz vein type (e.g. Kovo) and the Pahtohavare type (e.g. Pahtohavare). *Kovo* is a small historic Cu-mine located 23 km NNE of Kiruna (Fig. 1). The mineralisation is related to quartz veins in an albite-diabase along a mylonitic shear zone. The quartz veins intersect the albite-diabase in a ~10 m wide zone and are partly asymmetrically folded demonstrating a syntectonic origin. A weak mineralisation of chalcopyrite and pyrite is unevenly distributed together with some carbonate and barite in the quartz veins. No distinct alteration of the albite-diabase is identified around the mineralised veins. A bulk sample of the mineralised veins contained 0.15% Cu, 0.024 ppm Au and 5% Ba.

The Pahtohavare type of Cu±Au deposits represents one of the economically most important and best studied within the Palaeoproterozoic greenstone belts in the northern Fennoscandian shield. The *Pahtohavare*

deposit (Martinsson et al., 1997b) is similar in character to the *Bidjovagge* (Bjørlykke et al., 1987) and *Saattopora* (Korvuo, 1991) Cu-Au deposits situated in northern Norway and Finland, respectively. Characteristic is the strong albitisation of the host rocks comprised by graphite schist, mafic volcanoclastic rocks and mafic sills. Biotite-scapolite alteration typically envelopes the mineralised albite zone that contains chalcopyrite and pyrite as dissemination or breccia filling and veins together with carbonate, quartz, albite, and occasionally scapolite. *Suurikuusiko* in Central Finnish Lapland is slightly different in character and has gold as the only commodity.

The *Pahtohavare* Cu-Au deposit, situated in the Kiruna Greenstone Group about 10 km south of Kiruna (Fig. 1 and 14), is the only economic deposit of this type in northern Sweden. It produced 1.7 Mt of ore with 1.9% Cu and 0.9 ppm Au, by underground and open pit mining of two ore lenses during 1990–1997 (Martinsson et al., 1997b).

The mineralisation at Pahtohavare is complex and exhibits a number of different ore styles that also include the stratiform Eastern Pahtohavare, which is similar to the Viscaria deposit (Martinsson et al., 1997b). These deposits are hosted by the volcanoclastic Viscaria Formation, which forms the core of an anticline enveloped by pillow lava of the Peuravaara Formation. The SE-plunging anticline is truncated by a WNW-directed shear zone on its southern limb. The Southern and Central ores occur in a similar stratigraphic position and are spatially associated with a thick black schist zone. The Southern Pahtohavare deposit is mainly stratabound in character, whilst the Southeastern Pahtohavare deposit has a more complex shape with a narrow discordant part (Fig. 14). The discordant ore at SE Pahtohavare is of vein-type, which partly has a mylonitic to cataclastic texture, and is accompanied by albite-carbonate alteration. This part of the ore is mostly high-grade and consists of coarse-grained ferro-dolomite, quartz, scapolite, albite, chalcopyrite, and pyrite in varying proportions (Martinsson et al., 1997b).

Figure 14

In the stratabound deposits ore-grades of Cu are mainly developed in albite felsite, which is a fine grained albite rich rock formed by hydrothermal alteration of graphite schist. Chalcopyrite, pyrite and, more rarely pyrrhotite, are the dominant ore minerals and occur as disseminations, veinlets, and breccia-fillings in the albite felsite together with ferro-dolomite and some quartz (Martinsson et al., 1997b). Native gold and tellurides are found as inclusions in chalcopyrite (Hålenius, 1986, 1988) and the distributions of Cu and Au grades within the ores exhibit similar patterns suggesting a co-precipitation. The ore is partly also enriched in U, REE and Co (Martinsson et al., 1997b).

Several types of alteration are recognised in Pahtohavare, with albite, marialitic scapolite, biotite, actinolite, and ferro-dolomite as characteristic minerals, the latter also occurring as coarse-grained veins. Carbonate veins rich in sulphides may have a cataclastic texture, otherwise most carbonate veins show little evidence of deformation. Accessory minerals in these veins are brannerite, rutile, allanite, xenotime, and minute grains of thortveitite. Biotite-scapolite alteration is most extensive and surrounds ore-bearing albite felsites. The contacts between ore-related albite felsite and surrounding tuffite or black schist are usually interfingering, highly irregular and discordant to the primary bedding. Scapolite forms characteristic porphyroblasts and

networks outside the albite-altered host rock. Scapolite veinlets are commonly deformed by folding and shearing, and are in places mylonitic (Martinsson et al., 1997b).

Jälketurkkio is a small Cu-occurrence in the Vittangi area (Fig. 1) that is hosted by a fine grained albite felsite occurring adjacent to graphite schist and metadiabase in the middle part of the Kiruna Greenstone Group. The rocks are deformed and metamorphosed in middle amphibolite facies, while the albite felsite and the mineralisation are mainly undeformed and lack evidence of metamorphic recrystallization. The albite felsite is brecciated with amphibole, magnetite and pyrite forming rich disseminations in the calcite-dominated breccia infill. Chalcopyrite and chlorite are minor constituents and titanite is an accessory component. A bulk sample of the mineralised breccia contained 0.25% Cu, 0.062 ppm Au, 474 ppm Co and a weak enrichment of Ni and U (Martinsson, 1994).

During the 18th century some Cu was produced from *the Särkivaara* Cu deposit in the Svapavaara area (Fig. 1). The ore is situated in scapolite-rich mafic tuffites with intercalations of carbonate rocks and garnet-pyroxene-scapolite skarn in a sequence belonging to the upper part of the Kiruna Greenstone Group. Pyrite, pyrrhotite, chalcopyrite, and some molybdenite occur disseminated in the garnet-pyroxene skarn and constitute the main ore type (Frietsch, 1966; Martinsson, 1994). Besides Cu and Mo, the deposit is slightly enriched also in LREE, Co, Ni and Au (Martinsson, 1994).

Veikkavaara is a small Cu-prospect (Fig. 1) where mineralisation occurs at the border of a 30 to 40 m thick mafic sill, which has intruded mafic volcanoclastic rocks belonging to the upper part of the Veikkavaara Greenstones. These rocks are metamorphosed at middle to upper amphibolite facies and have a more or less well developed mineral lineation or foliation. Sulphide mineralisation is restricted to veins or lenses of skarn close to the western border of the sill. The skarn consists of coarse grained scapolite, pyroxene, amphibole, some biotite, locally garnet, and accessory titanite. Sulphides are dominated by pyrrhotite occurring together with some chalcopyrite as irregular dissemination and patches. A bulk sample of rich sulphide mineralisation from the Veikkavaara deposit contained 0.91% Cu and 0.008 ppm Au and was slightly enriched in Co and LREE.

Östra Äijäjärvi is a Cu deposit (Fig. 1) that consists of several lenses hosted by mafic tuffites belonging to the Kiruna Greenstone Group. The host rock tuffite occurs as a large inclusion in Perthite Monzonite granite that forms a 2x1 km satellite intrusion at the northern border of a larger granite pluton. The ore-associated granite has a poorly constrained U-Pb zircon age of ~1.86 Ga (Skiöld, 1981). Fine grained gabbro to diorite occurs locally in the granite and exhibits mingling structures with the granite. The tuffite is mostly banded at mm- to cm scale, and more or less albitised and locally scapolite is abundant. Skarn-rich patches, veins or breccias with amphibole, pyroxene, and occasionally epidote, are common. The mineralisation is dominated by pyrrhotite and pyrite, whilst chalcopyrite is a minor constituent and magnetite is only locally encountered. Two mineralisation styles are recognized: 1) dissemination of fine grained pyrrhotite in gabbro together with minor disseminated pyrite and chalcopyrite, 2) pyrrhotite, some pyrite and minor chalcopyrite in tuffite, with sulphides occurring as irregular dissemination, veinlets and massive veins or breccia infill. Elevated contents of Ni and Co support a relationship between sulphide mineralisation and the gabbro, whereas locally anomalous contents of Mo indicate a contribution also from the granite.

6.3.2. Svecofennian-hosted Cu-Au±Fe oxide deposits

Noting their diversity, and leaving out the earlier described apatite-iron type, the Svecofennian-hosted, epigenetic deposits may tentatively be further subdivided into the following styles of mineralisation; intrusion-related, shear zone-related, and vein-style. All deposits representing these styles, with the major exception of Aitik, are sub-economic. These deposits, occurring in the Porphyry Group and the Kiirunavaara Group (e.g. Aitik, Tjärrojäkka, Kiskamavaara, Lieteksavo), are characterised by an alteration comprising K feldspar, scapolite, biotite, and some tourmaline and sericite. Extensive albitisation is only locally developed and then mainly in association with intermediate to felsic intrusions. The main ore minerals are pyrite, occasionally rich in Co, and chalcopyrite, with magnetite as a minor to major constituent in most occurrences. Occasionally, there is field evidence suggesting that an early stage of Fe oxide mineralisation developed prior to the deposition of copper rich zones (the situation at e.g. Tjärrojäkka). Several deposits also contain bornite and minor amounts of hematite and molybdenite. The ore minerals occur disseminated, as breccia infill, in veinlets or in quartz-tourmaline veins. Generally the occurrences are structurally (shear zone) controlled and they occur mainly within areas dominated by K feldspar alteration, whilst scapolite-biotite alteration may be more important outside of the mineralised area.

6.3.2.1. Intrusion-related Cu-Au±Fe oxide deposits

Basically, all the intrusive-hosted mineralisation discussed in this section show similarities with porphyry style mineralisation: a) they are hosted by, or are spatially associated with, intrusive rocks, b) ore minerals occur as disseminations, veinlets, and locally as stockworks, c) age determinations confirm a temporal relationship between intrusions and hydrothermal activity, d) alteration style and zoning may have similarities to porphyry style mineralisation. Aitik is the only occurrence that has been interpreted as a typical porphyry deposit (Wanhainen, 2005), while most others have been suggested to be related to the Fe oxide-Cu-Au (IOCG) family of deposits (Williams et al., 2003). However, we have chosen to use the non-genetic name “Intrusion-related Cu-Au±Fe oxide deposits” for this mineralisation style.

The Aitik Cu-Au-Ag deposit is Sweden's largest sulphide mine and the third largest copper mine in Europe. The mine is situated 15 km southeast of Gällivare (Figs. 1 and 15) and close to the regionally important, NW-SE striking Nautanen Deformation Zone. In 2014, the mine production was 39.1 Mt of ore with 0.20% Cu, 0.09 ppm Au and 2.1 ppm Ag. The total production 1968-2014 is 672 Mt and reserves are estimated at 1126 Mt with 0.22% Cu, 0.14 ppm Au, 1.7 ppm Ag (New Boliden, 2014). Possibly also molybdenum will be extracted in the future (Nordin, 2008).

Figure 15

The geology of the Aitik deposit comprises Palaeoproterozoic schists and gneisses of volcanoclastic origin belonging to the Muorjevaara Group (Wanhainen and Martinsson, 1999), intermediate to mafic intrusions of the Haparanda Suite, and younger dykes. A 1.89 Ga subvolcanic quartz monzodioritic intrusion occurs in the footwall of the deposit and is weakly mineralised with chalcopyrite, pyrite, molybdenite, and magnetite occurring as veins and dissemination (Monro, 1988; Wanhainen et al., 2006). Undeformed pegmatite dykes of the Lina Suite crosscutting the ore zone and hanging wall units are 1.76 to 1.73 Ga in age (Wanhainen et al.,

2005). The ore is hosted by garnet-biotite schist and quartz-muscovite-sericite schist formed by hydrothermal alteration and metamorphic recrystallization of the volcanoclastic rocks (Zweifel, 1976; Wanhainen and Martinsson, 1999). The footwall also comprises feldspar-biotite-amphibole gneiss with sub-economic mineralisation. The hanging wall, which is separated from the ore zone by a major thrust, comprises feldspar-biotite-amphibole gneiss, which is barren of mineralisation.

Chalcopyrite and pyrite are the main ore minerals in Aitik, whilst magnetite, pyrrhotite, bornite, chalcocite and molybdenite constitute minor components. Ore minerals occur mainly disseminated, as veinlets, as patches and clots, and also in several types of veins together with varying amounts of other minerals such as quartz, biotite, amphibole, garnet, tourmaline, barite, zeolites (stilbite and chabazite) and thaumasite, and in pegmatite dykes (Wanhainen and Martinsson, 2003). Alteration within the ore zone is extensive and involves biotite alteration, commonly accompanied by garnet porphyroblasts of spessartine-almandine composition (Monro, 1988), and sericite-pyrite alteration. K feldspar-epidote alteration is mainly restricted to the footwall and hanging wall contacts. U-Pb dating of titanite from these alteration mineral associations defines a range in ages between 1.81 and 1.75 Ga with a peak around 1.78 Ga (Wanhainen et al., 2005).

The *Pikkujärvi* deposit is situated at the NW border of a quartz-monzonitic intrusion (Fig. 1) belonging to the Perthite Monzonite suite, which has intruded amygdaloidal basalts and felsic porphyries belonging to the lower part of the Kiirunavaara Group. The volcanic rocks are intruded by felsic to mafic dykes, aplites, small satellite intrusions of monzonitic rocks (Fig. 16).

Chalcopyrite, pyrite and small amounts of bornite, molybdenite, and magnetite form several less well-defined ore lenses. The ore minerals occur irregularly disseminated or as veinlets in altered basalt and in the felsic intrusions (Martinsson, 1995; Bergman et al., 2001).

Several types of alteration have affected the volcanic rocks and most extensive is biotite-scapolite alteration, which frequently is also developed in the mafic dykes. K feldspar alteration is more restricted, and mainly spatially associated with Cu-mineralisation. In many cases the alteration has completely obliterated the primary character of the basalt and changed it into a reddish K-feldspar rock containing locally preserved amygdules, or occasionally into scapolite-rich rocks. Tourmaline and epidote are common in veinlets and as small patches, whilst zeolites occur as late phases of the mineralisation event (Martinsson, 1995; Bergman et al., 2001).

Figure 16

6.3.2.2. Shear zone-related deposits

The Nautanen region (Fig. 1) forms a type area for shear zone-related deposits. Here, several deposits (e.g. *Nautanen*) occur in intensely altered and deformed rocks within a 300 m wide and 5 km long part of the Nautanen Deformation Zone. Less deformed clastic metasedimentary rocks, belonging to the Muorjevaara Group, are found on both sides of the shear zone. Ore minerals including magnetite, pyrite, chalcopyrite, bornite and small amounts of chalcocite and molybdenite occur in several different styles of mineralisation that also might represent different generations. Most of the mineralised shear zone is affected by potassic alteration that in the eastern part overprints scapolite alteration. Several other alteration minerals occur more locally and are usually spatially related to different style of mineralisation (Martinsson and Wanhainen, 2004).

Massive magnetite with disseminated pyrite and chalcopyrite occur as pods or folded and disrupted vein-style mineralisation bordered by garnet-biotite alteration in the eastern part of the shear zone. Mineralisation rich in magnetite may also contain carbonates, amphibole, quartz, garnet, barite and apatite. In the western part mineralization is mainly parallel to the shear zone with disseminated chalcopyrite, pyrite and magnetite related to strong tourmaline-quartz alteration and also with disseminated pyrite, chalcopyrite, bornite and magnetite occurring in strongly sheared and sericite altered rocks. Minor quartz veins with bornite and chalcocite that may contain stilbite and chabazite in druses probably represent the latest style of mineralisation in this area. The occurrence of rotated garnet porphyroblasts as well as folded and boudinaged vein-style mineralisation indicate that at least part of the mineralisation and alteration is pre- to syn-deformation (Martinsson and Wanhainen, 2004).

The deposits at Kiskamavaara and Ahmavuoma (Fig. 1) are different from the ores in the Nautanen area with respect to mineralogy and alteration. The Kiskamavaara Cu-Co-Au deposit (Fig. 17) occurs within a major tectonic zone (KADZ) trending from Karesuando to Svappavaara and further SSW towards Arjeplog (Fig. 1). The host rock to this deposit is a hydrothermal breccia within the Porphyrite Group, with strongly K feldspar altered and subrounded clasts of intermediate volcanic rocks. Several types of alteration, including scapolite, albite, carbonate and biotite have affected the surrounding rocks (Bergman et al., 2001; Martinsson, 2011).

The deposit is regarded to be of typical IOCG-style (Martinsson, 2011) and consists of Co-bearing pyrite occurring disseminated in the breccia groundmass together with magnetite and some chalcopyrite, whereas calcite and (locally) quartz occur as gangue minerals. The composition of the groundmass changes from almost massive pyrite in the central part of the richer ore lenses to disseminated magnetite-pyrite in the peripheral parts and hematite-magnetite outside the sulphide mineralisation. The mineralisation may be part of a symmetrically zoned system, as indicated by a core rich in magnetite with traces of molybdenite that is surrounded by an inner magnetite-pyrite-chalcopyrite assemblage and an outer and more extensive zone of hematite (Martinsson, 2011).

Figure 17

At Ahmavuoma (Fig. 1), there are andesitic volcanic rocks belonging to the Porphyrite Group with mineralisation of magnetite and pyrite. The most important mineralisation is a subeconomic pyrite lens containing Cu, Co, and Au. It contains K feldspar-altered fragments of andesite and is surrounded by a wide zone of strong biotite alteration, whilst scapolite and garnet are locally occurring constituents. Magnetite is occurring as veinlets, networks and breccia infill in andesite in other parts of the area. The breccia-type has the highest Fe content (25-50 % Fe), and is usually slightly enriched in REE.

The Tjårrojåkka-Fe deposit was described above as a Kiruna-type magnetite-apatite ore, but in addition several occurrences of Cu are found in the vicinity of the apatite iron ore. The largest, Tjårrojåkka-Cu, is situated 1 km WNW of the Tjårrojåkka-Fe deposit (Fig. 1). Disseminated pyrite, chalcopyrite, and bornite occur together with apatite-magnetite veins in a 700 m long and up to 30 m wide zone (Edfelt et al., 2005, 2006). A weak enrichment of Mo is locally found and a Re-Os age of 1.78-1.76 Ga was given by molybdenite (Stein, 2004). Due to a close spatial association and mineralogical similarities to Tjårrojåkka-Fe, a genetic relationship has been suggested for the apatite iron ore and the Cu-mineralisation at Tjårrojåkka (Martinsson, 1995; Edfelt, 2007). Several types of alteration are recognised at Tjårrojåkka-Cu including, K-feldspar, biotite, scapolite and

albite. K feldspar alteration is closely connected to the Cu-mineralisation, while albitisation is mainly restricted to the apatite-magnetite veined footwall to the Cu-mineralisation. Scapolite, frequently accompanied by biotite, forms an extensive alteration in the hanging wall to the Cu-mineralisation and in the ore zone (Martinsson, 1995). Characteristic for the Cu-mineralised area is the occurrence of accessory barite, and enrichments of sulphate in scapolite (Edfelt et al., 2005).

The *Gruvberget* Cu deposit (Fig. 1) is a historic mine situated close to the *Gruvberget* apatite Fe-ore (see above). Cu-sulphides are scattered throughout the area, but higher grade mineralisation is mainly developed in the footwall to the AIO along the transition zone between hematite ore to the south and magnetite ore to the north (Fig. 11). Chalcopyrite and some bornite are the main ore minerals, occurring disseminated together with magnetite, or as rich ore shoots at the contact with the iron ore. Druses with epidote, magnetite, pyrite, Cu-sulphides, stilbite, and accessory molybdenite are common within the bornite-bearing sulphide mineralisation (Frietsch, 1966; Martinsson and Virkkunen, 2004). Several small mine shafts are found close to metadiabases and Cu-mineralisation seems to be controlled by the same structures as the dykes and the redox boundary outlined by the hematite-magnetite transition. As the dykes cut the iron ore, Cu-mineralisation is probably younger and genetically unrelated to the apatite iron ore (Lindskog, 2001). However, the competent iron ore probably has acted both as a tectonic and chemical trap for later Cu-bearing fluids (Martinsson and Virkkunen, 2004). Close to the Cu-occurrences the bedrock is mainly affected by biotite-scapolite alteration with locally intense and overprinting K feldspar alteration replacing earlier scapolite. This is especially developed in association with mineralisation of bornite in late structures on the footwall side (Martinsson and Virkkunen, 2004).

At *Rakkurijärvi*, two occurrences of Cu mineralisation (*Rakkurijärvi* North and *Rakkurijärvi* Discovery Zone) are found within Svecofennian rocks immediately east of the greenstone-hosted *Pahtohavare* deposit (Fig. 1). *Rakkurijärvi* North is a low grade Cu-mineralisation hosted by andesites belonging to the Porphyrite Group. The ore minerals are pyrite, chalcopyrite and magnetite, which occur disseminated and in carbonate veinlets in andesite that might be strongly albite altered. Coarse-grained dolomite veins, resembling those at *Pahtohavare*, occur locally and may contain chalcopyrite and pyrite in significant amounts.

The *Rakkurijärvi* Discovery Zone is a recently discovered Cu-Au deposit suggested to be of IOCG-style (Smith et al., 2007). It is located 500 m southeast of *Rakkurijärvi* North, close to an ENE-trending shear zone. The host rock is affected by early albitisation followed by brecciation and mineralisation. Breccias range from jigsaw breccia distal to the main mineralisation to lithic breccia close to the ore zone, which is dominated by magnetite breccia. The breccia infill consists of albite, scapolite, actinolite and magnetite together with chalcopyrite, pyrite and accessory molybdenite. An early albite alteration is overprinted by magnetite-actinolite alteration followed by K feldspar, scapolite and biotite alteration, which partly develops into biotite-scapolite schist. According to Re-Os data of molybdenite the mineralisation has an age of 1.86–1.85 Ga (Smith et al., 2007).

6.3.2.3. Vein-style Cu±Au deposits

This category of deposits is mostly related to deformation zones but occur outside the main shear zones along secondary structures. The *Lieteksavo* Cu-Au-(Mo-W) deposit is a small, but rich, Cu-Au mineralisation (Fig. 1), with significant contents of Mo and W and also some Bi. The bedrock in the area comprises

amygdaloidal lava of basaltic to andesitic composition belonging to the upper part of the Porphyrite Group and the lower part of the overlying Kiirunavaara Group (Bergman et al., 2001). The deposit consists of quartz-tourmaline veins, up to 5 m wide, with bornite, chalcopryrite and some molybdenite that are hosted by a metadiabase occurring along a shear zone. The quartz-tourmaline veins are generally symmetrically surrounded by scapolite-biotite-tourmaline alteration at a distance of up to 7 m from the vein contacts. Close to the veins, coarse grained and almost massive scapolite may occur together with some tourmaline and accessory biotite, epidote and titanite. Further away, scapolite forms gradually less frequent porphyroblasts in a biotite-rich groundmass (Bergman et al., 2001).

The Ferrum and Fridhem deposits were worked on a small scale in the early 1900s. They are situated east of Nautanen (Fig. 1), in an area dominated by volcanogenic sandstones and pelites of the Muorjevare Formation, which have been metamorphosed in middle-upper amphibolite facies. Veins of quartz, quartz-tourmaline and quartz-microcline-pyroxene-amphibole-calcite contain chalcopryrite and bornite in varying proportions. Molybdenite and bismuthinite are accessory minerals and stilbite is locally found as a late phase. The quartz veins are mainly unaffected by deformation, whereas veins containing microcline-pyroxene-amphibole may show evidence of ductile deformation (Ros, 1980). Host rocks are affected by scapolitisation and many veins are enveloped by a massive scapolite zone containing some tourmaline (Geijer, 1918). At Fridhem, worked for gold only, a set of parallel quartz veins up to 2 m wide, contain small amounts of bornite and chalcocite. The richest Au mineralisation was found occurring in narrow and drusy quartz veins, containing chalcocite, native gold, calcite and druse fillings of hematite, stilbite, and chabazite (Geijer, 1918).

7. Discussion

Ore-forming processes are generally related to geological systems that are controlled by plate tectonic activity, with certain ore types spatially and temporally related to specific stages of the geodynamic evolution. An important aspect of the Northern Norrbotten metallogenic province is the long and complex geological history that paved the way for the deposition of a diverse range of ore deposits at different stages of the regional evolution (Billström et al., 2010). Palaeoproterozoic rock-forming processes span over more than 700 Ma years and are connected to a series of tectonic settings starting with continental rifts, followed by passive margins and ending with subduction and collisional orogens (Martinsson, 2004a; Weihed et al., 2005; Billström et al., 2010). Among the ore types identified in this province (stratiform-stratabound Cu±Zn±Pb sulphide deposits, iron formations, apatite iron ores and epigenetic sulphide deposits) several are restricted to certain stratigraphic units implying some kind of genetic connection to these rock forming events. Perhaps the most obvious observation is that both stratiform sulphide deposits and the iron formations are only found in submarine volcanoclastic units that also include carbonate and carboniferous rocks. Iron formations are almost exclusively situated in the uppermost part of the rift related Karelian greenstones (Vinsa Formation in eastern Norrbotten and Linkaluoppal Formation in the Kiruna area), whereas the stratiform sulphide deposits are either found at the top of the Viscaria Formation in the Kiruna area or at the top of the Vinsa Formation in eastern Norrbotten (Martinsson, 2004a). This suggests a connection to specific stages of rift evolution and rift environments during the development of the rift basin. The apatite iron ores have a less distinct stratigraphic position as they are hosted by rocks of both the Svecofennian Porphyrite and Kiirunavaara Groups. However, major apatite iron ores are clearly restricted to

areas occupied by the Kiirunavaara Group which supports a relation to that specific stage of the evolution (Martinsson, 2004a). The epigenetic sulphide deposits occur in most rock units and their formation seems to be mainly structurally controlled. Yet, the stratigraphic position and type of host rocks exert some control over the ore mineral compositions and ore alteration assemblages.

Another conspicuous feature of the Northern Norrbotten Ore Province is the high salinity of ore forming fluids (Broman and Martinsson, 2000; Billström et al., 2010) and the regional occurrence of scapolite (Frietsch et al., 1997). Both of these features might be the expression of evaporites being mobilised from the Karelian rift units (Frietsch et al., 1997; Martinsson et al., 1997a; Kyläksoski, 2009). Most of the ore deposits in this region are probably of hydrothermal origin with the apatite iron ores as a major exception given that these are interpreted to be at least partly formed from iron oxide melts (Frietsch, 1978; Martinsson, 2004a; Nyström et al., 2008; Billström et al., 2010). It is generally suggested that each of the different ore types which formed by hydrothermal processes had rather specific fluid characteristics that led to their individual types of metal associations, ore minerals, alteration minerals, and zonation patterns. However, the large variation in geological characteristics of the studied epigenetic sulphide deposits may also be attributed to the interplay between fluids of similar character (origin) with varying host rocks (Billström et al., 2010). To lay the foundation for further discussions related to ore formation, constraints on the crustal evolution in the study area, during the time period of concern (~ 2.5-1.75 Ga), are presented below.

Karelian rift events: The Karelian evolution includes several rift events starting with the emplacement of 2.50-2.44 Ga plume related mafic layered intrusions in the northeastern part of the Fennoscandian shield and the eruption of felsic volcanic rocks of the Sumi-Sariolan Group (Amelin et al., 1995; Manninen et al., 2001; Räsänen and Huhma, 2001; Iljina and Hanski, 2005). This was followed by episodes of erosion, sedimentation, sporadic volcanism and dyke injections at 2.22 Ga, before a stage of extensive mafic volcanic activity and continental break up at c. 2.15-2.05 Ga (Sokolov and Heiskanen, 1985; Vuollo, 1994; Martinsson, 1997; Hanski et al., 2001). A later rift event along the craton margin is recorded by the 1.95 Ga Jormua ophiolite (Peltonen, 2005).

The Palaeoproterozoic Kovo Group (~ 2.5-2.3 Ga) and the Kiruna Greenstone Group (~2.3–2.0 Ga) are part of a volcano-sedimentary unit deposited on an Archaean granitoid-gneiss basement, demonstrating a continental setting. The mixed chemical character of the volcanic rocks in the Kovo Group including both WPB and calc-alkaline affinities make interpretations of tectonic settings problematic. However, stratigraphically equivalent volcanic and sedimentary units in other parts of the Fennoscandian Shield are suggested to have formed in a continental rift environment (Amelin et al., 1995; Buiko et al., 1995). This may also be the case for the Kovo Group with the calc-alkaline volcanic rocks formed from crustal melts, whilst lava of WPB type is likely to have a mantle origin.

Characteristics for the Kiruna Greenstone Group include the mainly tholeiitic composition of the volcanic rocks, the occurrence of vesiculated terrestrial flows in the lower part and sub-aquatic pillowed flows in the upper part. The temporal evolution in composition and character of the volcanism suggests a rift environment, and age determinations of Palaeoproterozoic greenstones in northern Fennoscandia are consistent with a main magmatic and rifting event at ~2.1 Ga. In particular, the spatial association of WPB-basalts, coarse-grained clastic sediments, and possible evaporites in the lower part of the greenstones is indicative of an intracontinental rift setting generated by a mantle plume. The following units record a change in magma composition and

depositional environment that includes komatiites and tholeiites of LKT and MORB affinity and a change to subaqueous conditions higher up in the sequence. The generation of MORB-type magma indicates strong crustal attenuation and decompressional mantle melting (Bailey, 1983) and this change from early amygdaloidal WPB lava to late MORB-type pillow lava monitors the magmatic evolution in many flood basalt provinces (Marsh, 1987). Rifting of the Archaean craton, in a NW-direction from Ladoga to Lofoten, was accompanied by injection of ~2.1 Ga NW-trending dyke swarms (Vuollo, 1994). The Kiruna Greenstone Group and dyke swarms north of Kiruna, outline a NNE-trending magmatic belt extending into northern Norway. This belt is almost perpendicular to the major rift, and may represent a failed rift arm related to a triple junction south of Kiruna. The rapid basin subsidence, accompanied by eruption of MORB-type pillow lava in the Peuravaara Formation, is suggested to be expressions of the failed rifting event (Martinsson, 1997).

The lack of MORB-type pillow lava in eastern Norrbotten indicates a deposition of the Veikavaara greenstones units outside the failed rift basin in a more shallow water and evaporitic environment with the development of thick units of BIF and dolomite at the top. Characteristic is the enrichment of Cl, Br, Mn and Ba within graphitic tuffite and BIF within the Vinsa Formation (Martinsson, 2004b). After successful rifting of the Archaean craton a passive margin was developed along a line from Lofoten in the northwest to Ladoga in the southeast (Martinsson, 1997; Wanke and Melezhik, 2005). Along this margin a thick succession of dolostones, arenites, volcanoclastic and mafic volcanic rocks was deposited in a variable depositional environment associated with transition from a marine-influenced, rifted margin to a rimmed carbonate shelf. This was followed by a more than 2000m thick unit of deep-water shales, deposited on the drowned carbonate platform in response to tectonically enhanced subsidence along a passive margin (Wanke and Melezhik, 2005).

Svecofennian subduction: The Palaeoproterozoic tectonic evolution between 1.95 and 1.80 Ga north of the Archaean-Proterozoic Palaeoboundary is rather poorly understood in detail. Intrusions with relatively imprecise ages between ~1.95-1.93 Ga located in the Rombak window (Romer et al., 1992) and at Norvijaur (Skiöld et al., 1993) probably mark the onset of the Svecokarelian orogeny and a transition from a passive margin to a convergent setting. The first main expression of the Svecokarelian orogeny within Northern Norrbotten is the presence of rocks of the Porphyrite Group. Based on the present knowledge of ages and petrochemistry, these rocks are attributed to an event of convergent continental margin magmatism within a terrestrial to shallow marine environment with rapid facies changes from volcanic to sedimentary dominated rocks outwards from central volcanoes (Martinsson, 2004a). This period of convergence was manifested by the numerous intrusions of Haparanda-type calc-alkaline plutons. Within a few m.y. this period of convergent margin magmatism was followed by more alkaline and terrestrial volcanism producing the Kiirunavaara Group, which includes tholeiitic flood basalt, minor trachyandesite and extensive felsic pyroclastic units. Related plutonism generated the Perthite-Monzonite Suite (PMS) and this magmatic event took roughly place between 1.88 and 1.86 Ga, with the main volcanic episode lasting probably less than 10 m.y. The PMS intrusions, including mafic-ultramafic layered intrusions, could be related to mature continental arc magmatism developed in older continental crust or related to underplating and crustal extension caused by a mantle plume (Martinsson, 2004a). The magmatic evolution after ~1.86 Ga is mainly recorded in an extensive S-type magmatism (~1.85 Ga Jyryjoki and 1.81-1.78 Ga Lina-type) derived from anatectic melts in the middle crust. In the western part of the shield extensive I- to A-type magmatism (1.80-1.70 Ga) formed roughly N-trending batholiths (the Transscandinavian Igneous Belt) coeval with the S-type magmatism. Scattered TIB intrusions also occur further

east. The period at 1.80 Ga probably also involved a shift in orogenic vergence from NE to ~E-W in the northern part of the shield (Weiherd et al., 2002).

Svecokarelian orogen: High T - low P regional Svecokarelian metamorphism characterise large regions of Paleoproterozoic terrains of Sweden, but as pointed out by Bergman et al. (2001), the estimated pressures and temperatures are not constrained in time and could be related to different metamorphic events. Calculations of pressure (P) and temperature (T) conditions of metamorphism in Northern Norrbotten were performed on regional metamorphic assemblages in metaargillites and mafic metavolcanic rocks, interpreted to be of Svecofennian age. Generally, it is indicated that the metamorphic conditions show geographic variations. This led Bergman et al. (2001) to divide the pre-1.88 Ga rocks in northernmost Sweden into low- medium- and high-grade following the definitions of Winkler (1979). This is reflected by P estimates of 2–4 and 6–7.5 kbar, with the corresponding T estimates being 510–570°C and 615–805°C, respectively. It is interesting to note that most low-grade areas (i.e. Kiruna) are located in the westernmost part of Norrbotten whilst the majority of medium- to high-grade metamorphic rocks are located in the central to eastern part where also the vast majority of the Lina type granites (~1.81 to 1.78 Ga) are situated. The strong spatial relationship between the higher-grade metamorphic rocks and the S-type granites is probably the result of deeper erosional level of the crust exposed in these areas.

Moreover, radiometric evidence sheds further light on the timing of metamorphic events in northern Sweden, but data are generally too sparse to allow a detailed picture to be built up. Evidence for an early metamorphic event at ~1.93 Ga come from a Sm-Nd isochron of greenschist facies metamorphic minerals from a mafic sill within the Kiruna Greenstone Group (Skiöld and Cliff, 1984), and this is temporally related to the onset of Svecofennian magmatism. Subsequently, as mentioned previously, several lines of evidence call for a more regional metamorphic event at around 1.88 Ga ago, which, at least, affected a relatively large region around Kiruna. Following this, Bergman et al. (2006) argued for a possibly local magmatic-metamorphic event at ~1.86–1.85 Ga, based on radiometric data available from the area north of Pajala in the northeastern part of Norrbotten. A second metamorphic episode of regional significance occurred at around 1.8 Ga, and this was followed by more brittle deformation in local areas. It is also noteworthy that all suggested (ductile) metamorphic events generally coincide in space and time with known episodes of magmatism in this part of Sweden. Possibly later and low grade metamorphic events are indicated by a few age determinations giving U-Pb ages at ~1.62 and 1.73 Ga (Romer, 1996; Wanhainen et al., 2005; Westhues et al., 2013).

7.1 Genetic aspects and timing of deposits

The discussion below is focused on the detailed metallogenetic evolution in northernmost Sweden and for each deposit-type the timing of ore-forming events and the character of ore-forming fluids and environment of ore deposition are considered. Following this an attempt is made to integrate all available information, including a range of analytical data, metal commodities, stratigraphic relationships and alterations into dynamic genetic models and relate these to the IOCG concept.

There are three sets of radiometric analyses with relevance for ore formation in Northern Norrbotten, each carried out by means of different techniques; (i) Re-Os data on molybdenites, (ii) laser ablation (LA) ICP-MS, U-Pb in-situ analyses of chiefly titanite, but also to some extent of allanite, rutile and apatite and (iii)

conventional U-Pb analyses (this study). These three sets of data give in some cases different results for the timing of mineralisation. This could be explained by difficulties to relate the dated mineral to a specific hydrothermal alteration and mineralisation, since multiple mineralising and remobilization events could easily cause confusion but possibly also be related to the different analytical techniques.

The Re-Os data are, in a way, the simplest to interpret as such data are believed to be very robust (e.g. neither inheritance, nor later resetting is known to affect Re-Os molybdenite data (Stein et al., 1998). However, it may be more complicated to relate the obtained Re-Os age to a specific event (e.g. ore formation, metamorphism and alteration), but this problem is also shared by workers interpreting other type of radiometric data. A brief look at existing Re-Os data from northern Sweden reveals that the age spread is quite large between c. 1900 and 1730 Ma (Stein, 2004) and, generally speaking, a similar spread was found also in the present study and from earlier (LA) ICP-MS analyses. These data tell us that there exist a number of significant events, clearly separated in time, capable of promoting growth of molybdenite. The LA-ICP-MS results, on the other hand, are complicated to interpret. Storey et al. (2007) could demonstrate that titanite grains from certain apatite-iron ores (AIO) show complex internal structures. From BSE (back-scattered electron) imaging and REE analysis, core and rim zones were identified and a two-stage model was suggested. Certain core parts had LA-ICP-MS ages of c. 2050 Ma and were thought to represent an early alteration/metamorphism of the host rocks belonging to the Porphyry Group, whereas the rims are much younger with ages below c. 1900 Ma. Generally, the younger ages were interpreted either as due to new growth or to isotopic resetting of earlier formed grains coupled to ore formation or metamorphic events. However, unlike the typical core-rim development in e.g. zircon, the BSE dark (“cores”) and bright (“rims”) parts in titanites are unevenly distributed, and old ages were not always obtained from dark areas (cf. data from Gruvberget). Moreover, age data occasionally straddles a more or less continuous part of the concordia (cf. data from Malmberget) which renders it difficult to obtain undisputable and precise ages. The results also contradict existing conventional U-Pb zircon ages of these rocks. Smith et al. (2009) applied the LA-ICP-MS technique on samples from IOCG-style mineralisations. These authors were not able to demonstrate clear signs of any old 2050 Ma event but the obtained data are relevant for discussing the timing of ore-forming events. Obviously, these papers are very important in pointing out the age complexities noted especially for titanite, which only can be observed by in-situ studies. Also these results cast doubt on the relevance of conventional (TIMS) U-Pb data as a specific titanite age could readily be interpreted as corresponding to a geologically insignificant mean value of internal domains, each representing a unique age. It is beyond the scope of this paper to discuss these LA-ICP-MS results in further detail, although a few additional comments are to be found in the coming paragraphs. However, in a separate paper (Billström et al., in prep.) it will be proposed that (1) the outlined age complexities appear to selectively affect certain AIO type of ores which have suffered strong, multi-stage over-printing, but not Cu (-Au) mineralisations, (2) the inferred old (c. 2050 Ma) alteration effect may be given alternative explanations, and previously reported U-Pb zircon ages of the Porphyry group (≤ 1.9 Ga) are favoured. Following this, the conventional U-Pb data presented in this paper (Table 2) are considered to be geologically significant, a view principally supported by robust Re-Os data.

7.1.1 BIF and Mg-rich iron formations

These deposit types are grouped together since they occur as stratiform formations in similar (Greenstone-hosted) settings and are interpreted to have a mainly syngenetic origin. Basically, no radiometric

ages with relevance for primary mineralisation events exist for these deposits, and therefore other isotopic evidence and field relationships have to be used to place age and genetic constraints. BIF's and Mg-rich iron formations are located at stratigraphically high positions within the Karelian greenstones, which suggests an approximate formation age between c. 2.1 to 2.0 Ga. The pre-Svecofennian age for the Mg-rich iron formations is supported by their local occurrence as xenolites in Svecofennian granitoids (Frietsch, 1967) and cross-cutting porphyry dykes in some deposits. More rarely, there are also iron formations (e.g. Peltovaara) found in Svecofennian volcanic and sedimentary units with an approximate age of 1.9 Ga.

Most of the Mg-rich iron formations occur in the uppermost part of the greenstone succession (Vinsa Formation), partly in close association to typical BIF but also within dolomite occurring at the top of the Karelian greenstones (e.g. Tapuli). A few Mg-rich iron deposits occur within the underlying Viscaria Formation. These mineralisations consist of lenses of Ca-Mg silicates (mainly serpentine or Mg-rich pyroxene and amphibole $\pm$ phlogopite) containing magnetite and with minor disseminated sulphides (Grip and Frietsch, 1973). Ca-Mg silicate dominated deposits mostly have a more or less well-developed layering with alternating mm to dm thick bands of magnetite and pyroxene-amphibole. In a few deposits chert bands also occur. Deposits containing mainly serpentine as gangue minerals generally lack well developed layering, but a more diffuse banding in decimeter- to meter scale may occur.

Certain Mg-rich iron mineralisations possess some specific mineralogical, textural or structural characteristics indicating partly or totally different mineralizing processes compared to most of them. The Vathanvaara deposit has a banded to laminated character and is, in part, very rich in iron sulphides, resulting in a change from sulphide facies to oxide facies, whilst the Tjavelk deposit is unusually rich in apatite. Tapuli is situated at the same stratigraphic position as Sahavaara but is comprised of irregular lenses of magnetite-rich serpentine-pyroxene-amphibole within a thick dolomite. The Puoltsa deposit in the Kiruna area occupies a stratigraphically low position in the Kiruna Greenstone Group and, compared to other greenstone-hosted deposits, it also differs by its coarse grained character of pyroxene-amphibole skarn and the very irregular distribution of magnetite within the skarn body. A few deposits have slightly enhanced average Cu-contents (e.g. $\geq 0.1\%$ at Tervaskoski). These deposits usually occur close to granitoid intrusions and chalcopyrite is mainly occurring in veinlets suggesting a later introduction.

The prevalent view is that the source for iron in banded iron formations is mainly of hydrothermal origin (Ohmoto et al., 2006; Bekker et al., 2010). The generation of extensive BIF is proposed to be related to major magmatic events causing large scale addition of iron to sea water by seafloor hydrothermal discharge (Bekker et al., 2010). In contrast to the traditional model for BIF formed by upwelling of reduced iron-rich seawater to oxygenated coastal areas (Garrels et al., 1973; Holland, 1973) an exhalative model was presented by Ohmoto et al. (2006) involving the precipitation of chemical sediments from iron and silica-rich brine pools at the sea floor in land-locked seas. The subeconomic stratiform iron formations (BIF's) and the Mg-rich iron formations in Norrbotten is considered to be such iron rich exhalites which formed at a late stage of magmatism and later became variably metamorphosed.

The iron and the silica appear to have been deposited synchronously with carbonate-rich sedimentation in most of the Mg-rich iron formations which, during subsequent alteration reactions associated with regional ~ 1.9 Ga Svecokarelian amphibolite-grade metamorphism, got transformed into banded to laminated magnetite-amphibole-pyroxene rocks. In some deposits serpentine is the main gangue mineral and it may have formed

either as an alteration product of dolomite or as the metamorphic expression of syngenetic Mg-silicates. Several different Mg-silicates (sepiolite, stevensite, kerolite) may form at low temperatures in evaporitic basins instead of dolomite if silica activity is increased by hydrothermal or clastic input (Helgeson et al., 1969; Stoessel and Hay, 1978; Stoessel, 1988; Gehring et al., 1994). Mg-silicates, including talc, may also form from moderate to high temperature (140-380°C) hydrothermal discharge at the seafloor through interaction of hydrothermal silica and seawater magnesium (Lonsdale et al., 1980; Koski et al., 1985; Buatier et al., 1995; Dekov et al., 2008). Monomineralic beds of talc associated with the Archaean Matasgami VMS-deposit (Costa et al., 1980) are similarly suggested to represent seafloor precipitates.

The observed gradation in some deposits in Norrbotten from Mg-rich iron formations to chert-banded iron-silicate deposits suggests a genetic relationship and a similar exhalative origin. The lens-shaped and sulphide-bearing Mg-rich type of iron formation containing significant magnetite may represent more proximal deposits, whereas the chert-banded type of mainly silicate facies BIF might correspond to more distal and low temperature chemical sediments mainly lacking sulphides and having lower iron content. In our model, BIF's formed as extensive sheets of chemical sediments in a reducing environment due to venting of low to moderate temperature hydrothermal fluids adding mainly iron, silica, manganese and barium into a restricted basin with accumulations of organic carbon in a shallow water evaporitic environment. The initial BIF-forming stage was followed by extensive deposition of dolomite in the evaporitic basin, and continued hydrothermal discharge caused local addition of silica and the deposition of Mg-silicates (instead of dolomite) together with iron oxides and smaller amounts of sulphides. The present mineralogical and textural characteristics of these lens-shaped Mg-rich iron formations are mainly the result of metamorphic reworking of chemical precipitates that originally formed at or close to the seafloor (e.g. Sahavaara) or as sub-seafloor replacement (e.g. Tapuli).

Sulphur isotope composition of sulphides is near zero per mil for most of the Mg-rich iron formations (Frietsch et al., 1995; Martinsson, unpubl. res.) supporting that sulphur (and iron and silica) was leached from the volcanic pile. A similar origin of sulphur has been implied also for the volcanic associated banded iron formations of Algoma type (Bowins and Crocket, 1994). In contrast, sulphides from more typical BIF deposits in the Vinsa Formation show a large variation in sulphur isotope composition (Frietsch et al., 1995; Martinsson, unpubl. res.) compatible with biogenic processes interacting within low temperature chemical precipitates in a reduced basin.

7.1.2 Kiruna-type apatite iron ores

The origin of apatite iron ores (AIO) has been debated for more than a century and evidence have been put forward to support either of two dominant genetic models; a magmatic and a hydrothermal model. Geochemical studies (Naslund et al., 2003) show that an iron-rich magma may be generated by immiscibility from a silicate-magma. In a model where this ore type forms from a cooling Fe-oxide rich magma it would be expected that magnetite-apatite ores are emplaced as lava domes, dykes (sometimes forming intrusive breccia), lava flows and pyroclastic deposits, which all are ore styles described from El Laco in Chile and Cerro de Mercado in Mexico (Park, 1961; Henriquez and Martin, 1978; Henriquez and Nyström, 1998; Nyström and Henriquez 1994; Lyons, 1988). Similar evidence for a magmatic origin is found in the Gushan apatite iron ore in China with gas bubbles as well as tube-, miarolitic-, amygdaloidal- and flow structures in the ores (Hou et al., 2011). Evidences for a magmatic origin involving a high content of volatile components has also been demonstrated from the Nukutus deposit at Kiruna. The existence of volatiles is expressed by vesicles or partly unfilled spaces between magnetite laths in skeletal ore. Blebs of apatite-carbonate-quartz occurring in the ore are suggested to represent globules formed by immiscibility of volatile rich and iron-poor magma that in some areas have accumulated to larger masses forming the apatite-carbonate-quartz rocks. Contact relations of these rocks to iron ore vary from cutting to mingling. Apatite and carbonate are also occurring disseminated and as breccia infill in altered hanging wall rocks demonstrating that volatile components were expelled during formation of the apatite iron ores generating hydrothermal brecciation and alteration in the wall rocks (Martinsson, 2015).

The generally sharp contacts to the wallrocks found both at El Laco and Kiirunavaara and the primary ore textures, such as columnar and dendritic magnetite (Geijer, 1910; Nyström, 1985; Nyström and Henriquez 1989) suggest a rapid crystallization from an iron-rich magma. On the contrary, deposits formed at hydrothermal conditions (cf. the model of Rhodes and Oreskes, 1999), would be expected to show more gradational contacts and evidence of more extensive replacement of host rocks than are seen in most Norrbotten deposits. Besides, a selective replacement origin as the dominant mode of formation appears to be excluded on the basis of the lack of primary textures in the wall rocks mimicking those found in the magnetite ore (e.g. dendritic and columnar magnetite). Although a magmatic model is preferred for most of the AOI deposits in Norrbotten all their features can not be explained by magmatic processes, especially not the extensive alteration zones related to most of them (Bergman et al., 2001; Martinsson, 2003).

7.1.2.1 Deposits with magmatic-hydrothermal transitional characters

We believe that there is compelling evidence supporting a mainly primary magmatic origin for Kiruna-type ores based on specific textures of magnetite, the massive character of most deposits and their relation to the host rocks. However, the large variation in mineralisation style (massive, disseminated and breccia-type), both within and between individual Kiruna-type deposits, alteration of host rock and late veins, can not be explained by a single process. Instead, a combined activity of magmatic and subsequent hydrothermal processes has to be invoked (Martinsson, 2003, 2004a). For instance, whereas pyroxene replaces andesite xenoliths within massive ore at El Laco (Naslund et al., 2002), actinolite forms a narrow border zone at both the hanging- and footwall contacts and around porphyry xenoliths within the ore at Kiruna. Such relationships may be the result of a chemical exchange between an iron oxide melt and surrounding silicate rocks. It is also possible that the breccia style mineralisation at Kiruna could be partly, or totally, of hydrothermal origin similar to what is found at El Laco (Naranjo et al., 2010).

Further evidence for a hydrothermal over-print includes K feldspar-silica-biotite-chlorite-carbonate-sericite alteration together with minor pyrite, chalcopyrite, molybdenite mineralisation, and oxidation of magnetite to hematite. In the hanging wall to the Per Geijer ores at Kiruna extensive K feldspar alteration and silicification also includes hematite mineralisation with disseminated barite and low phosphorus content (Geijer, 1910; Frietsch, 1979) that most probably represents hydrothermal products. Breccia style AIO deposits are generally hosted by rocks affected by strong sodic alteration that, at distance, changes into potassic alteration (e.g. Mertainen) illustrating the generation of large hydrothermal alteration zones during their deposition.

The hydrothermal fluids were probably released during crystallisation of the volatile rich Fe-oxide magma (Broman et al., 1999; Nyström et al., 2008) in a process similar to the one producing highly vesiculated flows of iron oxide at El Laco (Naslund et al., 2002). At Kiruna lenses of massive magnetite ore (B-ore) may show a transition from very fine grained outer parts, with a low content of gangue minerals, to slightly more coarse grained internal parts, with increasing amounts of calcite together with small amounts of anhydrite and pyrite. This may indicate an enrichment of volatiles during the final stage of solidification, and an analogous process could also explain local patches of coarse-grained and skeletal magnetite within massive ore. Veins of anhydrite±pyrite±magnetite occurring in the ore and surrounding rocks may also be expressions of hydrothermal precipitation of components released from the crystallization of an iron oxide melt. A similar process may also be responsible for the replacement of magnetite by hematite in many of the AIO deposits in Norrbotten in accordance to what is found at El Laco (Nyström et al., 2008).

A hydrothermal imprint can also be suggested on the basis of observations from other AIO deposits in Northern Norrbotten. At Malmberget, south of the Kiruna area, overprinting metamorphism in upper amphibolite facies, strong deformation and injection of granitic, aplitic and pegmatitic dykes partly obscure the primary features of the ores (Martinsson and Virkkunen, 2004). However, in general the ore lenses are surrounded by extensive amphibole-pyroxene-feldspar alteration zones that may change into amphibole-pyroxene skarn closer to the ores. At the Fabian ore body, the massive magnetite ore is bordered by a wide zone of skarn breccias including massive and disseminated pyroxene-amphibole with varying amounts of anhydrite and smaller amounts of magnetite, pyrite and chalcopyrite. Based on these observations, and on the lower content of trace elements in magnetite from skarn breccia compared to what is found in massive ore, the latter was suggested to have formed from an iron oxide melt whereas the surrounding breccia was generated by a related hydrothermal process (Lund and Martinsson, 2013). A transition from magmatic to hydrothermal conditions has also been suggested for the Narken deposit in eastern Norrbotten. Here magmatically formed euhedral magnetite and apatite crystallized at depth and were brought to higher positions in the crust by hydrothermal fluids, where they were re-deposited as xenocrysts together with hydrothermally formed platy hematite and epidote forming a breccia infill (Martinsson, 2009). The Saivo mineralisation, located south-east of Kiruna, may also have formed at magmatic-hydrothermal transitional conditions similar to those producing pegmatites. Large crystals of diopside and minor magnetite and titanite show an inward growth with magnetite and titanite occurring in dendritic aggregates at the outer border of the lens-shaped ore body. The surrounding fine-grained syenite is brecciated and veined by pyroxene and amphibole (Martinsson and Virkkunen, 2004). It locally exhibits textures indicating simultaneous crystallization of hydrothermal pyroxene-amphibole with a co-existing magma producing a feldspar-rich rock. A similar relationship between magmatic and hydrothermal products has been described from the Roxmere pluton within the Williams batholith, Cloncurry district (Mark and Foster, 2000).

7.1.2.2 Fluid inclusion and radiometric evidence

At the Pliocene and unmetamorphosed El Laco deposit ore textures, oxygen isotopes and fluid inclusions have been used for genetic interpretations. Based on textures and oxygen isotopes of magnetite, ore formation is suggested to mainly be the product of a cooling magma at c. 1000°C (Nyström et al., 2008). In a study of fluid inclusions from El Laco, Broman et al. (1999) and Sheets (1997) found hydrosaline melt inclusions in pyroxene formed at temperature >800°C. This marks a transition from late-stage magmatic melts, represented by hydrous saline Na-K-Cl melt inclusions with anhydrite daughter crystals, to hydrothermal stage fluids of successively lower salinities (from 60 to 0 wt.% NaCl eq.). In Norrbotten, similar salt inclusions have been suggested to occur in magnetite from Kiirunavaara based on geochemical mass balance calculations (Lundkvist, 1998). This scenario is similar to what has been observed in the Stillwater Complex in USA, with magmatic halite melt inclusions occurring in olivine followed by high temperature (700-715°C) brine-carbonic fluids in felsic differentiates that during further cooling unmixed into CO₂-dominated and NaCl-H₂O dominated fluids observed as inclusions in quartz that records progressively decreasing salinity (Hanley et al., 2008). Fluid inclusion data from AIO-deposits in the Kiruna area have mainly been obtained from post-ore quartz (hematite±magnetite±sulphides) veins related to shearing that cuts the ores (e.g. the Kiirunavaara, Loussavaara, Henry, Rektorn and Nukutusvaara deposits). They have high salinity fluid inclusions (31 to 60 wt.% NaCl) (Broman and Lindblom, 1995; Harlov et al., 2002; Gleeson and Smith, 2009) showing, in general, the same characteristics as those from Cu-Au deposits in the region. However, fluid inclusion data from Tjärrojjåkka-Fe shows a continuum from moderate saline (15-17 wt.% NaCl eq. and Ca-Na dominated) high temperature (400-500°C) fluid inclusions in apatite to more saline fluid inclusions in quartz and calcite related to Cu-mineralisation within ore breccia. These fluid inclusions are identical with those found at Tjärrojjåkka-Cu and suggest these two deposits to be part of the same ore-forming system showing a progression from a high temperature apatite iron ore system to a typical IOCG system (Edfelt et al., 2007).

Published and new geochronological data help to set constraints on the timing of formation of the AIO deposits. Most ages are obtained from the Kiruna area and are based on titanite that exists together with magnetite or with actinolite, or zircon ages from cross-cutting felsic dykes. At Loussavaara coarse-grained magnetite dykes occur within biotite-chlorite altered trachyandesite in the footwall. Large platy crystals of titanite from these veins have given an U-Pb age of 1888±6 Ma (Romer et al., 1994), whilst granophyric to granitic dykes cross-cutting the stratabound type AIO at Kiirunavaara have an U-Pb zircon age of 1880±3 Ma (Cliff et al., 1990). Taken together these two ages indicate a deposition of the ore around 1.88 Ga, which is strengthened by the KUJ 278 titanite from a magnetite-bearing actinolite skarn assemblage at the contact to the ore which yielded a 1882±11 Ma age (Table 2). A similar age of 1880±3 Ma was obtained from titanite in a massive magnetite dyke at the Mertainen AIO (Fig. 3A). These ages, which are also compatible with TIMS and LA-ICP-MS data of titanite from amygdals within the footwall to the main apatite iron mineralisation at Kiruna (Romer et al., 1994; Smith et al., 2009; Westhues et al., 2013), suggest the Kiruna-type apatite iron ores to have formed between 1.89 and 1.88 Ga and argue for similar ages for both stratabound and breccia-style of AIO deposits. Based on the zircon ages of the hanging wall rhyodacite at Kiirunavaara (1882±24 Ma; Welin, 1987 and 1886±8.6 Ma; Westhues et al., 2013), the Kiruna type deposits seem to be almost coeval with their host rocks. Thus, regardless of the preferred mode of origin for these deposits, there is both a spatial and temporal correlation with rocks of the Kiirunavaara Group. There is also a temporal correlation to the older intrusions of the Perthite-Monzonite Suite, which is considered to be co-magmatic with the Kiirunavaara Group. Significant

for the age relations is also the occurrence of AIO as boulders in conglomerates in the uppermost part of the Kiirunavaara Group demonstrating that they had been exposed to erosion by that time.

Contrary to c. 1.89-1.88 Ga ages obtained from the Kiruna area, a significantly younger age is evidenced by the Fe-Ti deposit at Saivo for which U-Pb titanite data suggest a mineralisation event close to 1740 Ma (Fig. 3A). This AIO deposit is unusual in that it forms a pyroxene-rich lens within a body of monzonite and gabbro, probably belonging to a late 1.8 Ga TIB-1 generation, and the mineralisation is poor in apatite. Thus, the Saivo deposit may represent an event concentrating iron which is temporarily different from that giving rise to more typical AIO mineralisations. Moreover, U-Pb titanite and Sm-Nd data from the Tjårojåkka-Fe ore, give ages in the range 1.75-1.70 Ga (Edfelt et al., 2007). This iron ore is a representative example of a Kiruna-type apatite iron ore, and these results suggest that this ore type may have developed at two different stages, separated in time by about 100 m.y.

7.1.3. Epigenetic vein-, shear zone-style and intrusion related Cu-Au deposits

The epigenetic Cu-Au deposits in Northern Norrbotten, which are hosted either by Karelian greenstones or by Svecofennian rocks, display a complex pattern of mineralisation. Many of these deposits are located in shear zone environments, where large volumes of highly saline aqueous fluids were focused, whereas others are spatially (and genetically) related to intrusions of intermediate to felsic composition and occur at the borders of Haparanda and PMS plutons. Differences between deposits can be noted with respect to host rocks, mineralisation style, ore mineralogy, alterations and metal associations (cf. Table 4). Some features are mainly controlled by stratigraphic position and type of host rock whilst the Au/Cu ratio, for example, may partly be controlled by large scale regional variations in fluid composition, as indicated by increasingly more Au-dominated deposits and less abundant scapolite in ore-related alteration when going eastwards from the Kiruna area. In Karelian greenstone hosted deposits chalcopyrite and pyrite are abundant, with pyrrhotite and magnetite occurring in varying amounts. However, when hosted by graphitic sediments the deposits may be lacking iron oxides (cf. the Pahtohavare deposit) and pyrrhotite may become the most common iron-bearing mineral. Deposits hosted by the overlying Svecofennian units may have a high content of iron oxides together with chalcopyrite and pyrite, whereas pyrrhotite is rare or absent. These latter deposits may also contain bornite, chalcocite and molybdenite.

7.1.3.1 Fluid inclusion constraints

Fluid inclusion studies of epigenetic Cu-Au deposits in Northern Norrbotten display certain distinct features, which suggest a multi-stage fluid evolution, involving one or several fluids, where salinity progressive decreases during ore formation. This pattern is present in deposits of both the older (1.88-1.86 Ga) and the younger (1.79-1.76 Ga) group. The fluid inclusions present in the studied samples are typical for IOCG deposits worldwide; aqueous inclusions with high salinities and paragenetically later carbon dioxide inclusions. The fluid inclusions are characteristically Ca-rich, with halite, sylvite, calcite, pyrosmalite and hematite as solid inclusions. For the Norrbotten deposits the below overall model is built on an integration of data since there is a variation in fluid inclusion types between deposits; some of the deposits contain all of the observed types of fluid inclusions whereas others are dominated by a certain type.

Primary multi-solid aqueous inclusions are interpreted to represent an early stage of ore formation. These inclusions were trapped at temperatures that may have reached above 500°C. These inclusions consist of a CaCl_2 -NaCl-rich liquid, which is supersaline, with a salinity of 33 to more than 60 wt.% NaCl eq., and contain solid phases of halite, sylvite, calcite and hematite. Fluid inclusions from the Gruvberget-Cu and the Lieteksavo deposits have homogenisation temperatures between 210° and about 500°C. These temperatures extend the upper range established by similar fluid inclusions in the previously studied Cu-rich deposits in Norrbotten where the recorded total homogenisation temperatures were between 125° and 373°C (Table 3).

In most deposits (e.g. Pikkujärvi, Pahtohavare, Tjärrojåkka) these highly saline hydrothermal fluids caused extensive biotite-scapolite alteration. Sulphides are generally rare in these alteration zones, but magnetite may be a common accessory mineral formed by alteration of iron-bearing silicates. The wide zones of biotite-scapolite alteration are generally overprinted by more localised K feldspar- and/or albite alteration in their innermost parts. Primary aqueous fluid inclusions with halite $\pm$ calcite are typical for the main phase of chalcopyrite precipitation and indicate that the salinity of the fluid has become slightly lower, in the range 30 to 55 wt. % NaCl eq. The decrease in salinity may best be explained by mixing with an infiltrating, less saline fluid or by consumption of Cl and Na during formation of scapolite. Metal precipitation is most likely related to mixing with less saline fluids, but chemical changes caused by fluid reactions with the host rock may also be of importance. This is especially obvious when mineralisation is hosted by graphite schist, and at Pahtohavare mineralisation is related to extensive graphite decomposition and generation of CO_2 , which is also reflected by negative $\delta^{13}\text{C}$ of ferrodolomite occurring disseminated and as veins in the altered rocks (Martinsson, 1997b).

An overlapping stage that involved a fluid trapped in secondary inclusions with a less saline, but still Ca-Na-Cl-rich, composition created a final bornite-dominated Cu-mineralisation at the Gruvberget-Cu and Lieteksavo deposits. The lower salinity (between 20 to 23 wt.% NaCl+ CaCl_2 eq. dissolved salts) and deposition of bornite instead of chalcopyrite was probably a result of further mixing with the infiltrating water. Fluid inclusions related to this stage display homogenisation temperatures in the range 109° to 226°C, which are minimum estimates for the bornite-dominated stage. This type of inclusions, regarded as representing the final evolutionary stage of the NaCl- CaCl_2 fluid, has also been involved in bornite mineralisation in the Aitik (Wanhainen et al., 2003) and Tjärrojåkka (Edfelt et al., 2007) deposits (Table 3). Furthermore, such inclusions are commonly reported as the typical saline aqueous liquid-vapour type responsible for an overprinting sulphide stage in the genesis of many IOCG-deposits (Billström et al., 2010).

The significance of the late CO_2 phase has not yet been completely established. Typically, in many IOCG deposits CO_2 inclusions coexist with the high salinity inclusion types, and this has been interpreted as the result of unmixing of the carbonic phase from an original CO_2 -bearing magmatic hypersaline brine (Pollard, 2000, 2001). Mixing of two fluids is an alternative process where an inflowing CO_2 component is suggested to be a metamorphic volatilisation product (Fisher and Kendrick, 2008) that got added to the high salinity fluids. The carbonic inclusions from the Gruvberget-Cu and the Lieteksavo deposits (Table 3) are composed of almost pure CO_2 with less than 10 mole% CH_4 and they typically occur as primary inclusions in clusters or as secondary inclusions along healed microfractures that cut quartz-containing primary salt-rich inclusions. The large range of CO_2 homogenisation temperatures, also observed in other deposits in Norrbotten (Table 3), and frequently reported for CO_2 inclusions in IOCG deposits, with nearly unvaried temperatures for the CO_2 melting, indicates trapping at varying pressure conditions. The inclusions in clusters always have the lowest homogenisation

temperatures implying the highest pressures and the inclusions in healed microfractures were trapped after a stage of fracturing and rapid decompression. The connection of CO₂-rich fluid inclusions to a specific stage of mineralisation is less apparent, however, enhanced contents of gold in some deposits may tentatively be correlated with the CO₂ fluid as was put forward by Lindblom et al. (1996), Broman and Martinsson (2000) and Sammelin et al. (2011).

7.1.3.2 Radiometric constraints

As outlined previously, two main age peaks of titanite (rutile) growth/ore formation can be defined in the Northern Norrbotten Ore Province (Fig. 3A-C); one at c. 1.88–1.86 Ga and the other in the 1.79–1.76 Ga interval. Examples of the “older” age group (1.88–1.86 Ga) include the deposits at Ö. Äijijärvi (titanite; magmatic origin from suggested ore related granite), Rakkurijärvi (rutile; together with chalcopyrite and pyrite, in a carbonate vein), and Pikkurijärvi (titanite; in various hydrothermal assemblages; cf. Table 1). The significance of the rutile age (upper intercept age at 1868 Ma), sampled from the low-grade Rakkurijärvi Cu-deposit, finds support from Re-Os analyses of two separates of molybdenite intergrown with magnetite from the nearby IOCG-type of mineralisation at Rakkurijärvi (Discovery Zone), yielding ages in the 1.86–1.85 Ga interval (Smith et al., 2007). The significance of this time interval is also confirmed by LA-ICP-MS data on (mainly) allanite obtained from the Discovery Zone (Smith et al., 2009). Pahtohavara is situated close to the Rakkurijärvi deposits and might be of similar age. This suggestion is supported by the 1885 ± 18 Ma age of davidite, which is related to the gold mineralization at the genetically similar Bidjovagge Au-Cu deposit in northern Norway (Bjørlykke et al., 1990). Titanite data from Pikkurijärvi, including samples from the hydrothermally altered host rock and from ore related quartz monzonite, display a relatively consistent age pattern with ages between 1880–1860 Ma. No ages have been recorded for the Kiskamavaara deposit, which has all of the typical characteristics of IOCG deposits, and is situated within a major crustal shear zone that includes extensive albite carbonate alteration (Martinsson, 2011). Breccia infill of amphibole is locally developed within the altered rocks, from which titanite ages of 1.86 Ga were obtained, demonstrating hydrothermal activity within this shear zone at the same time as the generation of the deposits of the older group. However, the Kiskamavaara deposit itself could have a different age as several events of hydrothermal activity might have occurred.

Deposits giving conventional titanite U-Pb age data within the “younger” age group (1.79–1.76 Ga) include Jälketurkkio, Särkivaara, Veikavaara and Lieteksavo. However, independent radiometric evidence suggesting “young” mineralisation events exists only for Särkivaara where three molybdenites give Re-Os ages in the interval c. 1.78–1.77 Ga (Stein, 2004), which fits with the single conventional titanite analysis from this locality. Probably, the slightly older 1.82 Ga titanite age from a hydrothermal amphibole vein in a diorite at Ahmavuoma belongs to this age category as well. In addition, a titanite age (1.74 Ga) at the low end of the spectrum was obtained for a local carbonate vein at Mertainen, that contains some bornite and molybdenite and which is overprinting the apatite iron ore. The mineralisation at Tjärrojåkka can also be considered part of this group based on Re-Os molybdenite ages of 1.78–1.76 Ga (Stein et al., 2004) and U-Pb titanite ages of 1.78 Ga (Edfelt, 2007). However, a comparison between different data from the Gruvberget Cu occurrence reveals an age discrepancy. Four conventional titanite (sampled from feldspar-bearing quartz veins and K feldspar alteration related to bornite) analyses yield consistent ages between 1.77 and 1.78 Ga. The single molybdenite age from Gruvberget, taken from the same alteration zone, yields a 1895 Ma Re-Os age (Stein, 2004), and finally LA-ICP-

MS titanite data from a scapolite altered metavolcanic rock form an age cluster between c. 1900 and 1800 Ma (Storey et al., 2007). The reason for the apparent age discrepancy is not known but the large LA-ICP-MS titanite data spread is probably due to a disturbance created by a ≤ 1.8 Ga hydrothermal event, which may correspond to the obtained conventional titanite age. Radiometric data from Gruvberget may also be explained by a multi-phase mineralisation process with early formed AIO and later copper (Lindskog, 2001), and, if this hypothesis is correct, the 1.78 Ga age may date the stage of bornite precipitation. Finally, included in this younger group, but discussed separately (Martinsson et al., in prep.), are a number of vein-style deposits situated within the Nautanen deformation zone; e.g. Ferrum and Nautanen. The latter have Re-Os molybdenite ages of 1.77-1.75 Ga (Stein, 2004). At least some of these younger Cu $\pm$ Au deposits might have formed by remobilization of older deposits similar to what is found at e.g. Aitik (Wanhainen et al., 2005).

Some titanites taken from regional alteration and metamorphic paragenesis, typified by skarn and scapolite alteration, and with no obvious relationship to a nearby mineralisation, were also included (Fig. 3D). The inferred overall age range between ca 1900-1760 Ma for multi-stage metamorphism and hydrothermal alteration is similar to the data distribution from ore-associated mineral assemblages. At Nunasvaara, situated in the Vittangi area, an intensely scapolite-altered Haparanda type diorite intrudes the lower part of the Kiruna Greenstone Group. Titanites from this intrusion indicate two events at c. ~ 1.91 and 1.86 Ga, respectively. A 1903 Ma LA-ICP-MS titanite age (Smith et al., 2009) corroborates the significance of the older event. Hypothetically, the ~ 1.91 Ga event was the time of an early phase of regional scapolitisation related to Haparanda Suite magmatism mobilizing evaporites at the base of the greenstones and the younger age is representing a later metamorphic overprint. The Haparanda Suite magmatism could locally have commenced even earlier, cf. the Norvijaur type intrusions (Skiöld et al., 1993), and may have been contemporaneous with a suggested 1.93 Ga metamorphic event detected at Kiruna (Skiöld and Cliff, 1984). A later hydrothermal episode at Nunasvaara is suggested by c. ≥ 1.81 Ga old titanite from an amphibole-scapolite breccia at the border of the intrusion. The other alteration-related titanite ages, shown in Fig. 3D, could collectively be explained by mineral growth during metamorphic events recognized from the Norrbotten area. Such events include a 1.86-1.85 Ga episode based on data from Pajala (Bergman et al., 2006), and the major (regional) metamorphism close to 1.8 Ga, and a more, brittle metamorphic event between 1.75-1.73 Ga (Romer, 1996).

7.2 A regional metallogenetic frame-work for Fe-oxide, Cu-Au deposits in Northern Norrbotten

Fe-oxide Cu-Au mineralisation in Norrbotten is characterized by a large variety of mineral deposits containing various proportions of Fe-oxides and sulphide minerals. However, when deposits are put in stratigraphic schemes or observed with reference to their inferred depth of formation and timing, it is possible to establish some general patterns regarding deposit character, element association, compositional control and type of ore minerals and alterations that are useful for the metallogenetic interpretations. Also the character and origin of the ore fluids have important implications for the genetic interpretations in general and the IOCG-concept in

particular. Before addressing these aspects of ore formation, a brief over-view of the geological history and its couplings to ore-forming stages is given below.

7.2.1. Geological context and timing of mineralisation

It is useful in ore genetic modelling to aim at unravelling possible temporal relationships between ore formation (dated by ore-associated minerals such as titanite) and known crustal events (dated mainly by zircon). In Norrbotten the formation of iron oxide±Cu±Au deposits had a duration of approximately 300 m.y., from about ≤ 2.1 Ga to 1.75 Ga, with major peaks related to specific geological events (Fig. 18). The economically most important stage of ore formation (leading e.g. to the huge deposits at Kiirunavaara and Aitik) took place ≤ 1.89 Ga. This was a period of continental margin magmatism, reworking of older rift-related sedimentary sequences (including evaporites), and focused fluid flow in major crustal-scale shear zones linked to syn- and post-peak metamorphism. Several of the major crustal scale shear zones are associated with abrupt changes in metamorphic grade, indicating that these zones have been active after the 1.8 Ga peak metamorphism, and several of the epigenetic Cu-Au deposits show a strong spatial relationship with these shear zones. Thus, geochronological and structural evidence indicates a linkage to late- to post-metamorphic conditions for many of these mineralisations and it is also evident that deposits which display a structural control on a deposit-scale are also spatially associated with structures on a regional scale.

During the Karelian stage of evolution (≥ 2.0 Ga) stratiform sulphide mineralisation (Viscaria type) was associated with rapid basin subsidence and submarine basaltic eruptions (Martinsson, 2004a), whilst iron formations mainly were formed at a late Karelian stage in shallow evaporitic basins in relation to widespread hydrothermal discharge. The Svecofennian stage includes early (≤ 1.93 Ga), juvenile calc-alkaline I-type magmas, which were emplaced in a continental arc with submarine to subareal volcanism. These processes led to the formation of intrusion related deposits (e.g. Aitik), associated with the 1.93-1.89 Ga Haparanda intrusive suites, which are coeval and co-genetic with the Porphyrite Group volcanic rocks. After a short event of metamorphism and deformation, chemically more evolved magmatic rocks formed in a mature arc, or possibly due to introduction of mantle melts. These later c. 1.88-1.86 Ga intrusive and extrusive rocks include mafic to felsic and more alkali-rich magmas of the Perthite Monzonite Suite (PMS) and the Kiirunavaara Group volcanic rocks. During this stage Kiruna-type apatite-iron ores were formed by mainly magmatic processes. Simultaneously, hydrothermal deposits of diverse character, including typical IOCG-style of mineralisation (e.g. Pikkujärvi), were formed. Subsequently, during the late Svecofennian stage (c. 1.82 to 1.79 Ga), a second major input of magmas comprising large volumes of S-type granites (Lina-type) and subduction-related A-/I-type granitoids (TIB-type) occurred. These major plutons contributed with large volumes of magmas and heat transfer to the crust, resulting in greenschist to amphibolite facies metamorphism (1.81-1.78 Ga), deformation and the generation of vast amounts of fluids that were largely focused into crustal-scale shear zones in the middle crust. This was followed by low grade metamorphism and mainly brittle deformation along reactivated major shear zones until 1.74 Ga. Simultaneously with these magmatic-metamorphic processes, epigenetic and mainly vein-style Cu±Au deposits (1.82-1.74 Ga) formed at structurally favourable sites (e.g. in the Nautanen area). As a result of these processes metals from older mineral deposits (e.g. at Aitik) were also locally remobilized. The local occurrence of bornite and molybdenite in Lina-type pegmatites in the Norrbotten region (Geijer, 1924)

might represent more large scale remobilization of metals. During this stage a younger generation of apatite iron ores may also have formed.

Figure 18

7.2.2 Ore fluid sources

Most ore deposits in the Northern Norrbotten Ore Province are probably of hydrothermal origin, with the exception of the apatite ores being mainly formed from iron oxide melts (Martinsson, 2004a; Nyström et al., 2008; Billström et al., 2010). These fluids have partly high contents of Cu, but also Fe, Zn and Pb according to LA-ICP-MS and leachate analysis of fluid inclusions (Smith et al., 2013). At Nautanen the fluids also contain significant amounts of sulphate (Smith et al., 2013), which is reflected in a rather high content of sulphate in ore related scapolite (Frietsch et al., 1997), suggesting a rather oxidised character for the fluids. Ore types formed by hydrothermal processes generally have rather specific sources of fluids and specific fluid characteristics that give the deposit their signature including metal associations, ore minerals, alteration minerals, and zonation patterns. However, in Norrbotten the geological characteristics of the epigenetic sulphide deposits exhibit large variations that may also be partly attributed to the interplay between fluids having similar characteristics and their varying host rocks (Billström et al., 2010).

Characteristic for the Northern Norrbotten Ore Province is the high salinity of ore forming fluids (Broman and Martinsson, 2000; Billström et al., 2010) and a complex chemical composition (Smith et al., 2013). The fluids are typically Ca-rich and oxidized, with halite, sylvite, calcite, pyrosmalite and hematite as daughter minerals in fluid inclusions. These features compare well with characteristics for IOCG deposits worldwide, which typically feature aqueous inclusions with high salinities and carbon dioxide inclusions (Billström et al., 2010). In addition to those occurring in IOCG deposits, highly saline ore fluids are involved in many different types of ore deposits including porphyry copper, skarn, SEDEX, MVT and sediment hosted copper deposits, with the first and last mentioned showing the most similarities to IOCG. Ore genetic models are commonly based on a mainly magmatic source for fluids in the porphyry copper and skarn deposits and a non-magmatic fluid source in the others. These conceptual models have been tested by different techniques including stable isotope (S, O and H) ratios of halogens and noble gases. Another conspicuous parameter is the chemistry of the ore fluids. For instance, a Na-Ca dominated composition is typically attributed to fluids suggested to be of a non-magmatic (basinal, or bittern brine or evaporitic) origin, whereas such compositions are uncommon for magmatically derived fluids in porphyry systems (Wanhainen et al., 2003). However, as pointed out by Yardely et al. (2000) most anions and isotopic tracers as $\delta^{18}\text{O}$ mainly retain memory of the immediate environment of final emplacement or represent a mixed memory of the flow path and may not show the characteristics of the original fluid source.

7.2.2.1 Halogen ratios

High salinity brines may have different origins and their origin can be distinguished on the basis of conservative halogen tracers such as the ratio of Br/Cl and the chlorine isotope composition (Yardely et al., 2000). Such analyses can also indicate whether or not inferred evaporate-bearing strata have a metallogenic importance in the Norrbotten ore district. It may be recalled that primary evaporite Ca-Na-Mg-K brines (bittern

brines) which have evolved beyond halite saturation have salinities of ca 28-39 wt.% and a 1000xBr/Cl molar ratio from 1.54 (seawater ratio) and up to 10.7 (Fig. 19). Dissolution of evaporite halite generates secondary brines dominated by Na and with a 1000xBr/Cl molar ratio <1.54 (mostly 0.3-0.4). Magmatic, high temperature (>600°C), fluids generating hydrothermal mineral deposits are typically Na-K dominated and rich in Fe. Before brine-vapour unmixing such fluids are moderately saline (2-17 wt.%), and CO₂ bearing, with a 1000xBr/Cl molar ratio of 0.6-2.5, which is close to the suggested magmatic ratio of 1-2 (Kendrick et al., 2001a,b; Rusk et al., 2011). If fluid separation occurs, this produces a CO₂ rich phase and a highly saline Na-K-(Fe) brine with salinities up to 65 wt.% (Kendrick et al., 2001a,b). These fluids might change composition by fluid-rock reactions and evolve into more Ca-rich composition by albitisation (Yardely et al., 2000) or during calc-silicate formation in skarn deposits (Kwak, 1986; Layne and Spooner, 1991). Recent studies of ore fluids related to IOCG-deposits using Br/Cl ratios have revealed a complex scenario and interpretations either supporting a magmatic origin, a mixture of magmatic and non-magmatic fluids or a mainly non-magmatic origin (e.g. Kendrick et al., 2006, 2008; Chiaradia et al., 2006; Baker et al., 2008; Fisher and Kendrick, 2008; Williams et al., 2010; Marschick et al., 2011; Smith et al., 2013).

Analyses of the Br/Cl ratio in fluid inclusions from different ore deposits in Northern Norrbotten have been done by a crush-leaching technique (Gleeson and Smith, 2009, Smith et al., 2013). Additional PIXE (Williams et al., 2010) and LA-ICP-MS data (Smith et al., 2013) exist from some of the same deposits. Although there might be certain differences between PIXE, LA-ICP-MS and crush-leaching analytical data, with lower values typically recorded by the latter method (where all existing inclusions in the sample are released), a rather consistent pattern arises with 1000xBr/Cl molar ratios varying between 0.08-1.4 for Cu-Au deposits in Norrbotten. A similar range of values (0.2-1.1) is recorded for post-ore vein type samples associated with AIO deposits (Fig. 19). These values overlap with the lower range of typical magmatic fluids and are also consistent with halite dissolution brines. Aitik is the only exception among the analysed deposits, with a 1000xBr/Cl molar ratio of 4.4 (Fig. 19), which is consistent with an origin from bittern brines. A common feature in the Northern Norrbotten Ore Province is the Na-Ca rich character, which, together with the Br/Cl characteristics, precludes a juvenile magmatic origin of ore fluids. Although the Ca-rich character could be a secondary feature caused by extensive fluid-rock interaction connected with albitisation, masking a juvenile magmatic origin, this process should not significantly have affected the Br/Cl ratio of the fluids (Yardely et al., 2000).

Figure 19

The common occurrence of scapolite in the Northern Norrbotten Ore Province is suggested to reflect the occurrence of former evaporites within the stratigraphic record of the Kiruna Greenstones (Frietsch et al., 1997), in agreement with other studies suggesting that scapolite-rich metasedimentary units are commonly reflecting former evaporite-bearing strata (Ramsay and Davidson, 1970; Kwak, 1977; Ortega-Gutierrez, 1984). Strong, regional scapolite alteration, with no obvious relation to ores, is developed at two stratigraphic levels within the Kiruna Greenstones (Martinsson, 1997; 2004a,b). The lower position is at the base of the greenstone pile in association with clastic sediments and carbonate rocks. Scapolite is extensively developed in the overlying basalts, whereas parts of the clastic sediments show strong enrichment in Na and Mg. The upper position is close to the top of the greenstones and is developed as a scapolite and calc-silicate rich tuffitic unit of metaevaporite

character. It is partly graphitic and situated between underlying silicate facies BIF and overlying dolomite. It is suggested that whole rock Br and Cl data from these scapolite-rich units are mimicking the original Br/Cl ratios of the scapolite-forming fluids, as is also indicated from other studies (Pan and Dong, 2003). The lower scapolite unit has 1000xBr/Cl molar ratios of 1.11-1.64 whereas the upper unit has values of 1.82-4.44 (Fig. 19). This suggests that scapolite in the lower unit has formed from brines generated by dissolution of halite and possibly mixed with seawater. In contrast, scapolite in the upper unit probably formed from brines generated by evaporation of seawater beyond the point of halite saturation.

The previously mentioned greenstone-hosted Viscaria stratiform Cu deposit carries scapolite as a metamorphic, ore-associated mineral in the footwall alteration zone, and also within tuffitic units at the hanging wall contact. Scapolite-bearing rocks at both stratigraphic levels have a 1000xBr/Cl molar ratio of 0.97-1.25, which is comparable to the range within the regionally developed lower scapolite unit. We interpret this feature as pointing to an ore fluid brine which was generated by convecting seawater dissolving evaporites at the base of the greenstones, leaching metals from the volcanic pile, and generating stratiform sulphide deposits higher up in the sequence. A similar origin is suggested for the Iron formations with the Mg-rich iron formations formed at proximal sites from low to medium temperature discharge, whereas typical BIF was generated as extensive exhalite sediments in the surrounding areas.

During the Svecofennian period, the Karelian evaporites might have been transformed into metaevaporites, including scapolite, and this mineral has been suggested as a possible source for chlorine in hydrothermal fluids (Fisher and Kendrick, 2008; Kendrick et al., 2008). Apart from Aitik, with a 1000xBr/Cl molar ratio of 4.4 (Fig. 19), which is consistent with an origin from bittern brines, the epigenetic Cu-deposits in Northern Norrbotten have mostly 1000xBr/Cl molar ratios of 0.08-1.0. Such ratios are compatible with an origin mainly from halite or scapolite dissolution from the lower scapolite unit (Fig. 19). Similar ratios (0.2-1.1) are also found in post-ore veins related to AIO deposits suggesting that this type of fluid is of regional metallogenetic importance. However, earlier studies, partly based on oxygen isotopes, have favoured a mainly magmatic source (Lindblom et al., 1996; Edfelt et al., 2007; Gleeson and Smith, 2009) or a mixed magmatic and non-magmatic source (Wanhainen et al., 2003; Williams et al., 2003) for the fluids.

Gleeson and Smith (2009) also presented chlorine isotope data for fluid inclusions from AIO and epigenetic Cu-deposits. In general there is a rather small variation in $\delta^{37}\text{Cl}$ for AIO-deposits (-3.1 to -1.0‰), whereas larger variations are recorded for epigenetic Cu-deposits (-6.6 to 0.5 ‰). A magmatic origin for chlorine was favoured by Gleeson and Smith (2009) as data largely overlap with mantle signatures of -3 to 0 ‰. Strongly negative $\delta^{37}\text{Cl}$ values as low as -7.8 have only been reported for subduction zone pore fluids and the more negative values in Norrbotten could possibly be related to fluid-mineral fractionation (Gleeson and Smith, 2009).

7.2.3 Ore style, alterations and structural control

The character of the studied ores appears, at least partly, to be related to their stratigraphic positions. This concept is further developed in Fig. 20, where a division is made between iron ore of Kiruna-type (AIO) and two age generations of epigenetic Cu-Au mineralisations. AIO deposits are diverse in style, comprising mainly breccia-style mineralisation at larger depths (Porphyrite Group or lower part of the Kiirunavaara Group) and more stratabound massive lenses at higher positions (middle to upper parts of the Kiirunavaara Group). The

former mainly consist of magnetite with apatite and actinolite as most important gangue minerals, whereas those at higher stratigraphic positions might be dominated by hematite and carry apatite, carbonate and quartz as important gangue minerals. More rarely, shear-zone related (1.8 Ga?) AIO types (tentatively occurring at e.g. Tjårrojåkka-Fe) formed at various depths under magmatic to hydrothermal conditions. Stratigraphically highest are the epiclastic units, which locally contain significant amounts of erosional products from apatite iron ores. A special type of deposits has features indicating a formation at the magmatic-hydrothermal transition and is represented by AIO style mineralisations at Saivo and Narken. Saivo comprises “pegmatitic” structures in syenite with an amphibole-pyroxene assemblage mimicking similar magmatic-hydrothermal structures in Australia (Mark and Foster, 2000) and also described by Velasco and Toros (2009) for apatite iron ores in the costal belt of Chile, whilst the Narken deposit has the character of a hydrothermal breccia (Martinsson, 2009).

Figure 20

When the epigenetic Cu-Au type is inspected (Fig. 20), several styles of mineralisation could be identified. Disseminated and vein-style Cu-Au mineralisation could cap c. 1.88 Ga intrusives crystallizing at intermediate crustal depths and occurring together with shear zone related deposits of various characteristics. However, shear-related and vein-style deposits become more abundant with the younger generation of ore deposition linked to $\sim$ 1.8 Ga deformation. Structural observations would suggest that late-stage deformation took place both in ductile and ductile-brittle transitional environments.

The main features with relevance for ore-related alterations and metal associations are given in Fig. 21. Scapolite occurs both as a regional alteration phase (probably representing metaevaporites and is due to large-scale mobilisation of evaporite-fluids) as well as spatially connected to different ore types. Other typical alteration minerals include albite, K feldspar, actinolite, sericite, carbonate, biotite, quartz and tourmaline. The intensity and type of alteration appear to follow patterns which show both temporal and depth relationships, and, for instance, K feldspar, quartz and sericite are alteration phases developed at shallower crustal levels for both Kiruna-type and epigenetic Cu-Au deposits. The ore mineralogy also shows a similar pattern in that, for example, hematite partly replaces magnetite (Kiruna-type), and bornite succeeds chalcopyrite (epigenetic Cu-Au) at shallower crustal levels. In addition, certain elements, like Ba, substituting for potassium in biotite and K feldspar, or more rarely occurring as barite, is typical for several types of deposits (AIO and epigenetic Cu-Au). All these vertical zoning patterns comprising ore, gangue and alteration minerals are rather similar for AIO and Cu-Au deposits. They also have many common features with the IOCG model (Hitzman et al., 1992) and probably reflect an evolution of the mineralising fluids as they move towards more shallow levels in the crust. This evolution includes a decrease in temperature, salinity and pH as a result of mineral precipitation, wall rock interaction, and mixing with meteoric fluids.

Figure 21

7.2.4 A remarkable concentration of AIO type of deposits in the Kiruna-Gällivare area

The spatial distribution of Kiruna type iron ores in Norrbotten is mainly restricted to the Kiruna-Gällivare area, which contains more than 2000 Mt of iron. Only a few and uneconomic occurrences are known in other

areas (Frietsch, 1975). Importantly, the spatial distribution of strong regional scapolitisation, rocks of the Perthite Monzonite Suite, and thick basaltic units of the Kiirunavaara Group coincide. Thus, the main occurrences of apatite iron ores occur in a limited area combining these specific geological features, which also involves a deep crustal structure inferred to have formed as a failed rift at 2.1 Ga. This geological situation is similar to the setting of the Coastal Iron Belt in Chile, found within a back-arc basin, including evaporites, shoshonitic basalt, and basaltic andesite of flood basalt character (Oyarzon et al., 2003). We believe that such geological circumstances were crucial for the development of large AIO of a presumable primary magmatic origin in these two areas. A similar origin, occasionally superimposed by hydrothermal processes, may be advocated for other deposits, including those at El Lago (Park, 1961; Henriquez and Martin, 1978; Henriquez and Nyström, 1998; Nyström and Henriquez 1994, Naslund et al., 2002), Magnetita Pedernales (Grez et al., 1991), Marcona (Chen et al., 2010), Cerro de Mercado (Lyons, 1988), La Perla (Corona-Esquivel et al., 2010, Pea Ridge (Nold et al., 2013), and in the Bafq (Förster and Jafarzadeh, 1994; Moore and Modabberi, 2003) and Gushan areas (Hou et al., 2011). Geochemical studies (Naslund et al., 2003) show that iron-rich magma is most likely generated by immiscibility from a silicate-magma and melt inclusions from El Lago indicates that a high sulphate content is important to generate Ti-poor Fe-oxide melts from basaltic magmas (Naslund et al., 2009). Experimental data suggests that the miscibility gap in the silicate-iron oxide system is enlarged by increased oxygen fugacity (Vicenzi et al., 1994) and by high contents of F⁻, Cl⁻, CO₂²⁻ and BO₃³⁻ (Veksler, 2004). An increased content of SO₄²⁻, Cl⁻, CO₂²⁻, and BO₃³⁻ may be governed by crustal assimilation, including metaevaporites. The importance of evaporites in the generation of AIO has also been stressed by Barton and Johnson (1996) and Marchik et al. (2008). An oxidized character of mineralizing systems in the Kiruna area is, indeed, indicated by local hematite alteration of magnetite and disseminated anhydrite in the Kiirunavaara ore, extensive hematite alteration and the occurrence of barite in the Per Geijer ores, and abundant anhydrite in skarn breccia at the Fabian ore body in Malmberget. Furthermore, the tendency for unmixing of a Fe-rich magma is favoured by a high Fe content in the silicate starting magma and this, in turn, is governed by fractionation of plagioclase and olivine (Veksler, 2009). The presence of abundant mafic to ultramafic intrusions belonging to PMS in the Kiruna area seems to be consistent with such fractionation processes.

Briefly, based on the regional geological context in Northern Norrbotten and the ideas of Naslund et al. (2009), a hypothetical ore forming scenario is as follows. Tholeiitic mantle magma was injected into the crust along old zones of weakness and interacted with evaporitic units in mid-crust settings, adding sodium, calcium, chlorine and sulphate. This interaction with evaporites is supported by the sulphur isotopic composition of anhydrite at Kiirunavaara that is similar to Palaeoproterozoic seawater (O'Farrelly, 1990). Thus, the magmas gained components favouring the unmixing of an iron-rich and volatile-rich melt from the fractionating silicate magma that led to the generation of AIO deposits higher up in the crust.

7.2.5 A linkage between Kiruna type (AIO) and IOCG deposits?

Following the definition of Porter (2010) and Williams et al. (2005) only some of the Cu-Au deposits in Norrbotten may be classified as IOCG *sensu stricto*; including Rakkurijärvi (Discovery zone), Kiskamavaara, Nautanen, Tjärrojjåkka-Cu, Gruvberget-Cu, Jälketkurkkio and Pahtohavare, with the last mentioned deposit mainly lacking iron oxides due to a reducing host rock. Others may be regarded as vein-type (Kovo, Lieteksavo,

Fridhem, and Ferrum), intrusion related (Pikkujärvi and Östra Äjijärvi), or Mg-rich iron formations with anomalous Cu-Au (e.g. Sahavaara and Tjavelk) or Cu-Au anomalous AIO (Tjärrojåkka-Fe, Fabian, and Narken). In addition, a few deposits (e.g. Tervaskoski) may be characterized as Mg-rich iron formations with overprinting Cu-mineralisation.

The issue of whether or not a continuum exists between apatite-Fe and Cu-Au deposits is controversial and hard to resolve. However, two excellent examples illustrating the different relationships between AIO and IOCG (Gruvberget and Tjärrojåkka) are known in Northern Norrbotten. At Tjärrojåkka magnetite-rich (AIO) and Cu±Au rich deposits occur closely related. The Cu-mineralisation mainly occurs within a breccia-style AIO, or as a separate body at a distance of less than 1 km from the AIO. It is tentatively inferred that Fe- and Cu mineralisations are genetically related and formed close in time. The Fe-oxide (AIO) system, being slightly older, was associated with proximal albite-scapolite alteration formed at a high temperature. Potassic alteration occurs distal to the AIO, representing a lower temperature stage, with Cu-mineralisation occurring mainly in the transition from sodic to potassic zones. Minor Cu-mineralisation is also associated with K feldspar alteration overprinting albite-scapolite alteration close to the AIO. This is corroborated by similar fluid inclusion and stable isotope systematics for the two types of ore mineralisation at Tjärrojokka (Edfelt et al., 2007). In contrast, the IOCG style mineralisation at Gruvberget is younger (c. 1.78 Ga), clearly separated in time from the AIO, and related to overprinting K feldspar alteration (Lindskog, 2001; Martinsson and Virkkunen, 2004). The Cu-mineralisation occurs mainly in the transition zone between hematite and magnetite dominated apatite iron ore, suggesting deposition at a redox front by oxidised fluids, with magnetite acting as a reductant. A similar late overprint of Cu-mineralisation is recorded by bornite in calcite veins (1.74-1.73 Ga) cutting the apatite iron ore (1.88 Ga) at Mertainen.

The transition from magmatic to hydrothermal mineralisation processes during crystallization of volatile-rich iron-oxide magma is high-lighted in this paper and also suggested to be an important part of the development of other AIO deposits (Broman et al., 1999; Moore and Modabberi, 2003; Nyström et al., 2008; Naranjo et al., 2010; Chen et al., 2010, 2011, Nold et al., 2013, 2014; Chai et al., 2014; Sabeth-Mobarhan-Talab et al., 2015). Examples illustrating that AIO systems could develop hydrothermal mineralisation of typical IOCG-style in Norrbotten may be given from Tjärrojåkka-Fe, Kiirunavaara, the Fabian iron ore body at Malmberget and the Narken iron ore deposits, as these deposits also include minor sulphides (pyrite±Cu±Co±Au±Mo). However, most IOCG-systems seem to be products of metamorphic or magmatically driven large-scale hydrothermal systems involving significant components of non-magmatic origin, or to totally be a product of non-magmatic fluids similar to that of the Wernecke Breccia system (Hunt et al., 2007).

In the Norrbotten ore district the diverse character of Cu-Au deposits seems to be the product of two major events of mineralization. Each of these is related to regional magmatic and metamorphic processes involving fluids having many similar characteristics, but with different proportions of magmatic and non-magmatic components, and is schematically illustrated in Fig. 22. Several of the deposits would be classified as IOCG sensu stricto according to Williams et al., (2005), whilst others are better explained as previously existing Fe-oxide deposits overprinted by later Cu-Au mineralization, or intrusion related to vein style deposits without significant Fe-oxides.

Figure 22

8 Conclusions

- (1) Epigenetic Cu-Au $\pm$ iron oxides formed at two stages ($\approx$ 1.9 and 1.8 Ga, respectively) of the Svecofennian evolution in association with magmatic and metamorphic events, with fluids generated from magmatic sources and by metamorphic dewatering. A non-magmatic origin for the fluids may dominate, as indicated by cation compositions and Br/Cl ratios of the ore fluids with bittern brines, or, more commonly, dissolution of halite or scapolite, contributing to their high salinity. For intrusion-related deposits metals and sulphur are believed to have a mainly magmatic-hydrothermal source, but for other type of deposits most components were probably leached from the middle crust. In most cases a close spatial relation to intrusive rocks is lacking and, depending on local conditions and position in the crust, ore fluids generated a variety of Cu-Au deposits that includes typical IOCG-deposits (Fe-oxides and Cu-Au formed from the same process), IOCG of iron stone type (pre-existing Fe-oxide deposit with later addition of Cu-Au), IOCG of reduced type (lacking Fe-oxides due to locally reducing conditions) and vein-style Cu-Au deposits. From a strictly genetic point of view deposits formed from fluids of a mainly magmatic origin should be considered as a different type of deposit compared to those of a mainly non-magmatic origin. In that perspective, IOCG-style deposits of a mainly magmatic origin overlap with porphyry systems, whilst those of a mainly non-magmatic origin overlap with sediment hosted Cu-deposit regarding origin and character of the ore fluids.
- (2) AIO $\pm$ anomalous Cu-Au-Co-Mo deposits were generated from volatile-rich Fe-oxide melts that, during crystallisation, expelled hydrothermal fluids generating minor IOCG-type deposits. Fe-oxide magmas were generated by immiscibility of tholeiitic magmas in a mature continental arc or a plume related rift setting due to fractionation and contamination of evaporitic deposits in middle crust.
- (3) Mg-rich iron formations $\pm$ anomalous Cu-Au-Co were generated as proximal exhalative deposits formed from convecting seawater with increased salinity due to transformation into bittern brines in evaporitic shallow basins, or by dissolving halite from evaporites at the base of the greenstone pile. Metals and sulphur were leached from the volcanic pile during the vaning stage of rift related volcanism. Later overprinting by Cu-Au mineralisation might have generated the IOCG of iron stone type.

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Figure captions

Figure 1. Simplified geological map of northern Norrbotten with selected mineral deposits (modified from Bergman et al., 2001). Inset shows the location of northern Norrbotten in Scandinavia.

Figure 2. Lithostratigraphy of Archaean and Palaeoproterozoic units in the Kiruna area (from Martinsson et al., 1999).

Figure 3. U-Pb age data of minerals selected to date ore-forming episodes in northern Sweden, and possibly related metamorphic events. A) AIO mineralization; B) Svecofennian-hosted Cu(-Au) mineralization; C) Greenstone-hosted Cu(-Au) mineralization; D) non-ore related events.

Figure 4. Images (a) and (b) show type L+V+mS fluid inclusions from Pikkujärvi. These are multisolid inclusions with an aqueous liquid, a vapour bubble and more than two solid phases. (a) Fluid inclusion in tourmaline, (b) fluid inclusions in quartz. The photographs (c) to (f) show typical fluid inclusions in quartz from Lieteksavo and Gruvberget-Cu. (c) Type L+V+halite inclusions with an aqueous liquid, a vapour bubble and a cubic halite crystal. d) Type L+V+halite inclusions with an additional rounded solid identified as calcite. e) Type L+V inclusions with an aqueous liquid and a vapour bubble in a healed microfracture. f) Type CO₂ inclusions along a healed microfracture. Type L+V+mS inclusions from Pikkujärvi.

Figure 5. Photographs showing representative examples of ore types in northern Norrbotten. a) Sautusvaara, skarn-rich iron formation with alternating silicate and magnetite layers. b) Narken, breccia infill of epidote and hematite altered magnetite. Fragments of wall rocks chlorite-epidote altered (brownish) and silicified (white). c) Kiirunavaara, massive apatite iron ore varying from apatite rich (left) to apatite poor (right). d) Renhagen, breccia style apatite iron ore. e) Malmberget, coarse grained apatite ore with magnetite and greenish apatite. f) Nautanen, folded vein with magnetite, pyrite and chalcopyrite (at scale) bordered by scapolite alteration. g) Pahtohavare, brecciated albite felsite with infill of chalcopyrite and ferrodolomite. h) Kiskamavaara, hydrothermal breccia with K-feldspar altered clasts and infill of pyrite and magnetite.

Figure 6. Photographs showing alteration types associated with mineralization in northern Sweden. A) Rektorn, silicified and K-feldspar altered felsic volcanic rock. b) Saivo, albite-rich felsic intrusion exhibit mingling structures with hydrothermally formed amphibole-pyroxene. c) Mertainen, albite altered and brecciated trachyandesite with infill of scapolite. d) Nautanen, tourmaline altered intermediate volcanic rock veined by quartz. e) Pahtohavare, albite altered graphite schist with partly preserved graphite in lower left. f) Pahtohavare, ferrodolomite alteration associated with quartz veining. g) Pahtohavare, brecciated albite felsite with infill of ferrodolomite. h) Pahtohavare, biotite altered mafic sill with veinlets and porphyroblasts of scapolite.

Figure 7. Geological map of the Tornefors iron formation deposit (from Martinsson, 1995).

Figure 8. Chemical characteristics of apatite iron ores, skarn iron ores, and epigenetic Cu-Au ores. a) Fe-S-Cu b) Fe-P-S.

Figure 9. The Mertainen apatite-iron ore deposit. Top: Geology and Fe contents, Bottom left: Fe content in drill cores from profile A, Bottom right: Magnetic and gravimetric anomalies at Mertainen (modified from Lundberg and Smellie, 1979).

Figure 10. Geology and apatite-iron ores in the central Kiruna area (from Wanhainen and Martinsson, 2000).

Figure 11. Geology of the Gruvberget apatite iron ore deposit (modified from Frietsch, 1966).

Figure 12. Geology of the Malmberget apatite iron ore deposit (modified from Geijer 1930).

Figure 13. Geological map of the Saivo iron ore deposit (modified from Lehto, 1972).

Figure 14. Geological map of the Pahtohavare area (top) and the Southern Pahtohavare Greenstone-hosted Cu-Au deposit (bottom) (from Martinsson, 1997).

Figure 15. Geological map and profile (looking north) of the Aitik copper mine area, showing the main rock types and the outlines of the Aitik main open pit (the northern pit) and the Salmijärvi open pit (the southern pit)(modified from Nordin et al., 2013).

Figure 16. Geological map of the Pikkujärvi area (top) and the Pikkujärvi copper deposit (bottom).

Figure 17. Geology of the Kiskamavaara area showing the outline of Cu-Co-Au ore bodies (Martinsson and Wanhainen, 2000).

Figure 18. Ore forming-episodes integrated into the overall crustal evolution in northern Norrbotten.

Figure 19. Scatter plot (Log scale) of Br/Cl ratios (red dots) in fluid inclusions from AIO (quartz vein samples associated with apatite iron ores) and IOCG (samples from Fe oxide-Cu-Au ores) in the Norrbotten ore province after data in [1] Williams et al. 2010; [2] Gleeson and Smith 2009; [3] Williams unpublished, and of Br/Cl ratios in scapolite-rich rocks in ore-related alteration zones and within the Kiruna Greenstone Group, KGG [This study]. Salinities of fluid inclusions in AIO and IOCG samples vary between 30 and 60 wt.% $\text{CaCl}_2\text{-NaCl}$ eq. Black line between red dots stands for a range of data. The seawater Br/Cl ratio (1.54) is marked with a black star and the magmatic fluid range as a yellow field after data in [4] Kendrick et al. 2001a, 2001b and [6] Böhlke and Irwin 1992. The range of Br/Cl ratios that indicates fluid interaction with halite-bearing evaporites or scapolite is from [6] Böhlke and Irwin 1992 and [1] Williams et al. 2010. Typical ranges of Br/Cl ratios in fluid inclusions from IOGC (blue field) and PCD (porphyry copper related fluids; grey field) are compiled after data in [1] Williams et al. 2010; [4] Kendrick et al. 2001a, 2001b; [5] Baker et al. 2008.

Figure 20. Relationships between stratigraphic position and character of 1.8 and 1.9 Ga mineral deposits of AIO and IOCG type in northern Norrbotten.

Figure 21. Variations in ore mineralogy and alteration types in relation to stratigraphic position for 1.8 and 1.9 Ga mineral deposits of AIO and IOCG type in northern Norrbotten.

Figure 22. Cartoon illustrating hypothetical fluid sources involved in the formation of Fe oxide, Cu-Au deposits in northern Sweden. Fluid of hybrid origin may form by mixing of fluids of magmatic and evaporate origin, respectively or by magmatic fluids modified by fluid-rock reactions during ascent. Saline fluids may also form during solidification of Fe-oxide melts.

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Table 1 Description of samples for U-Pb age determinations. Specimens were selected to include parageneses representing three categories; ore, alteration and metamorphism.

Locality; sample #	Type of deposit	Host to sample ¹	Paragenesis ²	Morphology of dated phase ³
<i>Ore deposit</i>				
Kiirunavaara; KUJ-278	AI O	Skarn at contact to ore: KG	Act with some mt and titanite	Subhedral, brown, ≤ 3 mm
Mertainen; Mert-5	AI O	Actinolite skarn in magnetite ore: KG	Act, mt with some cal, qz and titanite	Sub-anhedral, brown-orange, ≤ 2 mm
Mertainen; Mert-6	AI O	Calcite vein in magnetite ore: KG	Cal, act, mt, titanite and accessory cp	Subhedral, light brown, 3 mm
Saivo; 29KOM58	AI O?	Skarn in monzonite: TIB?	Px, act, mt, and titanite	An-subhedral, brown-dark brown, ≤ 50 mm
Ahmavuo ma; AHM- 1:158,0	IO CG	Amphibole vein in metadiorite: PG	Act, some py, titanite	Sub-anhedral, brown, 0.5-1.0 mm
Gruvberg et; Gru-14B	IO CG	Quartz vein in metaandesite:PG	Qz, K-fsp, hbl, some mt, cp, bn, and titanite	Euhedral, brown, 0.5-1 mm
Gruvberg et; Gru-3A	IO CG	Bornite in K-feldspar alteration within scapolite-altered andesite:PG	Scp, K-fsp, ep, titanite, mt, bn, mo, cal, stl	Subhedral, brown, 1-2 mm
Gruvberg et; Gru-17	IO CG	Bornite in K-feldspar alteration within scapolite-altered andesite:PG	Scp, K-fsp, ep, titanite, mt, bn, cal,	Subhedral, brown, 1-2 mm
Jälketkurkio; Jäl-1	IO CG	Carbonate breccia in albite-altered greenstone	Cal, mt, py, amp, accessory cp, titanite	Sub-anhedral, brown, 0.5 mm

Pikkujärvi; Pick-32A:19.7	IO CG	Qz- monzonite: PMS	Fsp, qz, bt and accessory mt and titanite	Subhedral, dark- brown, 1-4 mm
Pikkujärvi; Pick-32B:23.3	IO CG	Qz- monzonite: PMS	Fsp, qz, bt and accessory mt, titanite	Eu-subhedral, dark brown, 1-4 mm
Pikkujärvi; Pick18:15.3	IO CG	Qz- monzonite: PMS	Fsp, qz, act, mt and accessory titanite	Eu-subhedral, dark brown 1-3 mm
Pikkujärvi; PIK18:100.0	IO CG	Scapolite altered metabasalt: KG	Scp, some ep, bt, titanite and chz in druse	Subhedral, brown , 0.5 mm
Lieteksav o; Lie4:85,4 5	Epi -Cu	Quartz vein in metadiabase in KG	Qz, tur, some chl, titanite	Euhedral, light brown, 2-5 mm
Rakkurijä rvi; 73001, 72.3	Epi -Cu	Carbonate vein in metadiabase:PG	Fe-dol, some py, cp, accessory rt	Rutile, subhedral, dark brown, 0.6-1.0 mm
Särkivaar a; Sär-1A	Epi -Cu	Scapolite skarn in greenstone	Scp, some px, amp, ep, titanite, py, cp, mo	An-euhedral, brown, ≤ 2 mm
Veikkava ara; 28LOM1 3C	Epi -Cu	Scapolite skarn in metadiabase within greenstone	Scp, px, some cp, po, titanite	Anhedral, brown, ≤ 1mm
Östra Äijäjärv; ÖÄ,6800 3:11.3	Epi -Cu	Granite:PS M	Qz, fsp, some amp, bt, chl, ep, accessory py, titanite	Euhedral, brown, 0.3-2.0 mm
Alteration				
Nunasvaa ra; 29KOM2 81A+B	Sca politization	Metadiorite: HS	Scp, some px and disseminated py, titanite.	Sub-anhedral, light brown, ≤ 1 mm

Nunasvaara; 29KOM6 2-F	So dic-calcic alteration	Amphibole breccia in greenstone	Ab-altered scp, amp, accessory titanite.	Anhedral, light brown, ≤ 0.4 mm
Paurankivaa; 29KOM2 98W	So dic-calcic alteration	Amphibole skarn in albite- carbonate alteration zone:PG	Act, some disseminated mt, cp, cal, titanite	Anhedral, brown- grayish, brown, 1-4 mm
Metamorphism				
Annaniva; 29MOM4 0A+B	Me tamor- phism?	Pyroxene skarn in greenstone	Px, some titanite	Sub-anhedral, brown, ≤ 10 mm.
Dundret; 28KDun- 1	Me tamor- phism	Pyroxene vein in amphibolite: PG?	Px, amp, ep, fsp some titanite.	Anhedral- subhedral, reddish brown, 0.2-1 mm
Junosuan do OM95026	Me tamor- phism	Metadacite: PG	Fsp, qtz, amp, ep with accessory titanite	Anhedral- subhedral, reddish brown, 0.1-0.3 mm
Nunisvaara; 28KOM7 0	Me tamor- phism	Metaandesit e: PG	Fsp, amp with accessory titanite	Anhedral, brown, 0.4-0.8 mm
Tiankijoki; Tia9:68,7	Me tamor- phism	Pyroxene skarn in metasediment: PG	Px, some amp, cal, po, accessory titanite	Eu-subhedral, dark brown, ≤ 0.8 mm

- 1) abbreviations used for rock sequences; HS=Haparanda Suite, KG=Kiirunavaara Group, PG=Porphyrite Group, PMS=Perthite-Monzonite Suite, TIB=Transcandinavian Igneous Belt
- 2) mineral abbreviations: ab=albite, act=actinolite, amp=amphibole, bn=bornite, bt=biotite, cal=calcite, cp=chalcopryrite, chz=chabazite, chl=chlorite, ep=epidote, Fe-dol=ferrodolomite, fsp=feldspar, hbl=hornblende, K-fsp=K-feldspar, mo=molybdenite, mt=magnetite, po=pyrrhothite, px=pyroxene, py=pyrite, qz=quartz, rt=rutile, scp=scapolite, stl=stilbite
- 3) dated mineral is titanite, with the exception from Rakkurijärvi from where rutile was utilized

Table 2 U-Pb TIMS data for titanites and rutile from Fe oxide, Cu-Au ore deposits and non-ore settings in N. Norrbotten.

Sa	V		Pb tot (ppm)	²⁰⁶ Pb me	Cor (p	²⁰⁶ Pb - ²⁰⁷ Pb radiog. Pb	²⁰⁶ Pb	²⁰⁷ Pb	²⁰⁷ Pb/ ²⁰⁶ Pb age (Ma)
Apatite-Fe ores									
Kl	1		35.8	212	0.3	33.4 – 3.8	0.3	5.2	1878±4
M	2		12.8	482	0.9	58.2 – 6.7	0.3	5.3	1880±3
M	2		15.0	765	0.8	63.1 – 6.7	0.2	4.3	1727±3
Sa	1		54.7	214	1.3	80.7 – 8.6	0.2	4.1	1774±4
Sa	7		59.6	302	1.0	80.4 – 8.5	0.3	4.4	1740±2
Epigenetic-style Cu-Au ores									
Svecofennian									
A	90	44.1	21.9	878	0.9	60.9 – 6.8	0.3	5.130±42	1814±7
G	15	117	39.7	2430	0.8	77.9 – 8.5	0.3	4.518±33	1779±2
G	21	133	43.5	789	3.1	82.7 – 10	0.3	4.607±21	1772±4
G	21	103	33.3	543	3.3	81.0 – 10	0.2	4.430±21	1771±6
G	68	57.1	17.8	1400	0.8	87.1 – 10	0.3	4.656±39	1773±7
K	70	64.6	20.0	1062	1.0	88.6 – 9.5	0.3	4.481±16	1755±5
P	7		59.6	182	1.1	52.2 – 6.4	0.3	4.8	1880±4
P	12		57.1	316	8.0	66.0 – 10	0.3	5.2	1863±11
P	44		168	141	5.4	67.7 – 8.4	0.3	5.3	1879±4
R	63	89.1	29.6	3810	0.3	83.1 – 9.5	0.3	4.971±11	1859±2
Greenstone-belt									
J	15	136	49.5	932	2.1	77.1 – 8.4	0.3	4.684±32	1776±4
S	18	180	67.8	339	9.6	83.2 – 9.0	0.3	4.673±16	1777±3
V	90	98.8	31.3	1944	0.9	83.1 – 9.7	0.3	4.606±27	1800±3
C	56	85.5	31.5	555	2.8	84.0 – 9.6	0.3	5.122±22	1856±5
Alteration and metamorphism									
Ar	1	102	32.2	605	3.0	89.4 – 9.4	0.2	4.356±11	1724±2
Ar	1	38.5	11.1	623	0.8	71.0 – 7.4	0.2	3.133±10	1688±3
Du	1	534	303	476	20	49.0 – 5.3	0.3	4.485±30	1763±2
Ju	9	94.1	45.0	2707	0.6	57.3 – 6.3	0.3	4.713±14	1783±2
Nu	1	7.8	12.8	338	0.4	17.5 – 2.0	0.3	5.024±28	1851±6
Nu	2	14.9	11.4	482	0.4	21.7 – 4.1	0.3	4.986±25	1850±7
Nu	1		25.0	1267	0.7	49.4 – 5.5	0.3	4.839±46	1813±6
Nu	8	53.6	52.9	1047	0.9	27.1 – 2.9	0.3	4.530±29	1769±6
Pa	1	62.2	48.7	2406	0.5	37.6 – 4.2	0.3	5.263±53	1853±7
Ti	1	114	37.4	1633	1.2	82.7 – 8.9	0.3	4.517±27	1760±5

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a: sample abbreviations (complete sample numbers are found in Table 1); Ahm=Ahmavuoma , Gru=Gruvberget, Jäl=Jälketkurrkio, K-Lie=Lieteksavo, KUJ=Kiirunavaara, Mert=Mertainen, Nun=Nunasvaara, Paur =Paurankivaara, Pick=Pickujärvi, Rak=Rakurrijärvi, Sär=Särkivaara, Tia=Tiankijokki, Veik=Veikavaara and ÖÄ=Östra Äijijärvi.

b: corrected for mass fractionation (0.10 ‰ per a.m.u.) and spike

c: corrected for mass fractionation, blank, spike and common Pb.

Errors in the isotope ratios are given at the 95 % confidence level.

Ore deposit	Type of inclusions	T h °C T o liquid P artial* or total	T h °C B y melting of halite P artial* or total	S alinity wt % NaCl eq.	T hCO ₂ °C	T mCO ₂ °C	Composit ion CO ₂ phase	Reference
1) Stratiform Cu								
2) Epigenetic Cu- Au								
3) Apatite-Fe- oxide (late quartz veins)								
1) Viscaria	L+V+ halite	8 2 to 138*	1 12 to 214	2 8 to 33	--- ---	--- ---	---	Martinsson et al., 1997
2) Aitik	L+V+ halite±cc L+V CO ₂	1 01 to 228* 1 00 to 222	1 40 to 373	3 0 to 44 1 8 to 27	- 17.0 to +30.0	- 56.6 to - 56.7	CO ₂ + <1 mole % CH ₄	Wanhainen et al., 2003
2) Gruvberget-Cu	L+V+ halite±cc L+V CO ₂	1 09 to 240* 1 09 to 215	2 24 to 415*	3 3 to 50 2 0 to 22	- 26.5 to +27.4	- 56.6 to - 57.2	CO ₂ + <4 mole % CH ₄	This study Gleeson and Smith, 2009
2) Kallosalmi	L+V+ mS L+V+ halite L+V	1 11 to 154* 9 0 to 107* 1 10 to 138	2 20 to 510* 2 00 to 220	3 3 to 61 3 2 to 33 2 4 to 25	--- ---	--- ---	---	Gleeson and Smith, 2009
2) Kiskamavaara	L+V+ halite	1 03 to 150*	1 90 to 285	3 1 to 37	--- ---	--- ---	---	Gleeson and Smith, 2009
2) Lieteksavo	L+V+ halite L+V CO ₂	1 43 to 241* 1 39 to 226	2 10 to ~500	3 2 to 60 2 1 to 23	- 2.1 to +24.1	- 56.6 to - 58.0	CO ₂ + <10 mole % CH ₄	This study
2) Nautanen	L+V+ halite L+V± CO ₂	1 12 to 285* 1 06 to 300	~ 100 to 335	2 8 to 41 4 to 28	? ---	- 56.6	Pure CO ₂	Gleeson and Smith, 2009
2) Pahtohavare	L+V+ mS L+V+ halite L+V	1 13 to 179* 7 0 to 126*	2 21 to >500* 1 25 to 170	3 3 to >60 2 9 to 30 0	- 4.2 to	- 57.7 to -	CO ₂ + < 5 mole % CH ₄ ± N ₂	Lindblom et al., 1996 Gleeson and Smith, 2009

	CO ₂	8 0 to 340		.5 to 25	+25.8	57.9		
2) Pikkujärvi	L+V+ mS	1 45 to 200*	3 12 to 450*	3 9 to 53	---	---	---	This study
2) Tjärrojäkka-Cu	L+V+ halite L+V	1 55 to 209* 1 11 to 191	1 96 to 292	3 1 to 37 2 3 to 27	---	---	---	Edfelt et al., 2007
3) Kiirunavaara – Loussavaara	L+V+ mS L+V	9 4 to 156* 1 04 to 277	~ 190 to 340*	3 1 to 41 1 3 to 22	---	---	---	Broman and Lindblom, 1995 Harlov et al., 2002 Gleeson and Smith, 2009
3) Mertainen	L+V+ mS	1 05 to 164*	3 20 to 500*	4 0 to 60	---	---	---	Gleeson and Smith, 2009
3) Per Geijer ores	L+CO ₂ +mS L+V+ halite L+V	9 6 to 242* 8 1 to 202* 7 0 to 269	2 21 to 300* 1 85 to 344	3 3 to 38 3 1 to 41 1 3 to 15	+2 6.9 to +28.5	- 57.0	CO ₂ + 3 mole % CH ₄	Broman and Lindblom, 1995 Harlov et al., 2002 Gleeson and Smith, 2009
3) Tjärrojäkka-Fe	L+V+ mS L+V+ halite L+V	~ 150* 1 15 to 135* 1 13 to 270	3 00 to >500* 1 93 to 282	3 8 to >60 3 1 to 37 1 8 to 28	---	---	---	Edfelt et al., 2007

Table 3. Summary of fluid inclusion microthermometric data from northern Norrbotten IOCG deposits.

L = aqueous liquid, V = vapour, mS = multisolid, cc = calcite, Th = homogenisation temperature, * = partial Th = L+V+halite±mS → L+halite±mS → L+mS

total Th = L+V → L or L+halite → L, ThCO₂ = homogenisation of CO₂(l) + CO₂(g) → CO₂(l), TmCO₂ = melting of CO₂

Table 4. Deposit characteristics for the studied ore types in northern Norrbotten (for references, see text).
A division is made by grouping individual deposits into main ore types and host rock units.

Ore type and host rock unit for selected mineralisations	Associated rocks	Alteration minerals	Ore/gangue minerals	Element association	Ore character
Iron formation – hosted by the Kiruna Greenstone Group					
Käymäjärvi: hosted by Vinsa Formation	Picritic tuff, black schist, dolomite	scp, bt (in footwall)	mt/ qz, gru, act, fa	Fe, (Mn, Ba)	BIF. Stratiform, oxide- and silicate facies, banded, laminated
Sahavaara: hosted by Vinsa Formation	Tuffite, black schist, arenite	scp, phl (in footwall)	mt, py, po/ srp, di, trm	Fe, (Cu, Co)	Mg-rich iron formation. Stratiform lens, massive, banded, disseminated
Sautusvaara: hosted by Linkaluoppal Formation	Tuffite, black schist, conglomerate	bt, scp (in footwall)	mt, py, po/ di, trm, act	Fe, (Cu, Co)	Mg-rich iron formation. Stratiform lens, massive, banded, laminated
Tapuli: hosted by Vinsa Formation	Dolomite, tuffite, arenite	phl, scp (in footwall)	mt, (py, po)/ srp, di, trm, act, bt	Fe	Mg-rich iron formation. Stratabound lenses, veins massive, disseminated,
Tervaskoski: unspecified host unit	Tuffite	bt	mt, py, po, cp/ trm, act	Fe, (Cu, Co)	Mg-rich iron formation. Stratiform lens, massive, disseminated, sulphides in veinlets and disseminated
Tjavelk: hosted by Linkaluoppal Formation	Tuffite, black schist	phl, scp	mt, py, po, cp/ srp, trm, act, ap	Fe, (Cu, Co, Ni, V, LREE, P)	Mg-rich iron formation. Stratiform lens, massive, banded, disseminated
Tornefors: hosted by Vinsa Formation	Basaltic tuff, tuffite	bt (in footwall)	mt, py, po/ qz, srp, gru, di, ep	Fe	Transitional Mg-rich iron formation/BIF. Stratiform lens, banded, laminated
Vathanvaara: hosted by Linkaluoppal Formation	Tuffite, black schist	bt, scp (in footwall)	mt, py, po/ trm,	Fe, (Co, Cu,	Mg-rich iron formation. Stratiform

		footwall)	act	LREE)	lens, massive, banded, laminated
Vieto: <i>unspecified host unit</i>	Tuffite, dolomite, mafic dykes	phl, scp	mt, py, po, cp/trm, act	Fe, (Cu, Co, Ni, V, LREE, P)	Mg-rich iron formation. Stratiform lens, massive, banded, disseminated
Iron formation – hosted by Svecofennian units					
Peltovaara: <i>unspecified host unit</i>	Mica schist, volcanics	bt	mt/di, trm, act, cal	Fe	Stratiform-stratabound, banded, disseminated
Kiruna-type (apatite iron ore) - hosted by the Kiirunavaara Group					
Henry: <i>at contact between Luossavara and Matojärvi Formation</i>	Rhyodacite and rhyolite pyroclastics	K-fsp, ser, chl, cb	ht, mt/ap, cal	Fe, (P, V, LREE)	Stratabound lens with discordant extension into footwall, massive, banded
Kiirunavaara: <i>at contact between Hopukka and Luossavaara Formation</i>	Trachyandesite lava and rhyodacite pyroclastics	act, ab (footwall) bt (hangingwall)	mt/ap, act, bt, cal	Fe, (P, V, LREE)	Tabular, massive, minor breccia infill
Lappmalmen: <i>at contact between Luossavara and Matojärvi Formation</i>	Rhyodacite and rhyolite pyroclastics	K-fsp, ser, chl	mt, ht/ap, cal, bt, qz	Fe, (P, V, LREE)	Stratabound lens, massive, disseminated
Malmberget: <i>unspecified host unit</i>	Felsic-mafic volcanic rocks, aplite and granite	ab, bt, scp, act, di, K-fsp, anh	mt, ht/ap, act, bt	Fe, (P, V, LREE)	Massive to banded lenses, stocks, minor breccia infill
Mertainen: <i>unspecified host unit</i>	Trachyandesite lava	ab, scp, act, bt, ser, cb	mt/act, bt, cal	Fe, (V)	Breccia infill, massive lenses and veins
Nukutus: <i>at contact between Luossavara and Matojärvi Formation</i>	Rhyodacite and rhyolite pyroclastics	K-fsp, ser, chl, cb	mt, ht/ap, cal, qz	Fe, (P, V, LREE)	Stratabound lens, massive
Painirova: <i>unspecified host unit</i>	Trachyandesite lava	ab, act, scp	mt/ap, act	Fe, (P, V)	Breccia infill and veins
Pattok: <i>unspecified</i>	Trachy	act,	ht,	Fe	Lens, massive,

<i>host unit</i>	andesite lava	ab, scp, ser, chl, ep, tur	(mt)/ ap, cal, act,	(P, V)	breccia infill
Rektorn: <i>at contact between Luossavara and Matojärvi Formation</i>	Rhyod acite and rhyolite pyroclastics	K- fsp, ser, qz, chl, cb, tur	mt, ht/ ap, cal, qz	Fe, (P, V, LREE)	Stratabound lens, massive and partly banded
D:o - <i>hosted by the Porphyrite Group</i>					
Gruvberget-Fe: <i>unspecified host unit</i>	Mafic to intermediate volcanic rocks, mafic dykes	ab, scp, K-fsp, ep, grt, act, ser	mt, Fe, ht/ ap, cal, grt (P, V)		Massive lens, minor breccia infill
Leveäniemi: <i>unspecified host unit</i>	Felsic- mafic volcanic rocks	act, ab, bt, ser, scp	mt, ht/ ap, act, bt, cal	Fe, (V, LREE)	Massive lens, breccia infill
Narken: <i>unspecified host unit</i>	Interm ediate volcanic rocks, arenitic sediments	ep, chl, qz	mt, ht, (py, cp)/ep, chl, (qz)	Fe, (P, Cu, Au, Th, LREE)	Breccia infill
Tjårojokka-Fe: <i>unspecified host unit</i>	Andesi te, diorite and porphyry dikes	act, ab, scp, K-fsp	mt, (cp)/ ap, act, cal	Fe, (P, V, Cu, LREE)	Massive lens, breccia infill
D:o - <i>hosted by intrusive rocks</i>					
Saivo: <i>hosted by the TIB 1 Suite</i>	Monzo nite and gabbro	scp, bt	mt/ di, ttn, act	Fe, Ti, (Nb)	Dendritic aggregates of magnetite in pyroxene lens
Karelian-hosted Cu-Au					
Jälketkurkkio: <i>hosted by the Kiruna Greenstone Group</i>	Black schist, tuffite, mafic sill	ab, scp, bt	cp, mt, py/ cal, act, chl	Cu, (Co, Ni, Au, Fe)	Breccia infill, veinlets
Pahtohavare: <i>hosted by the Kiruna Greenstone Group</i>	Tuffite, black schist, mafic sill, pillow lava	ab, bt, scp, cb	cp, py, (po)/ dol, ab, scp, qz, bt	Cu, Au, (Co, LREE, U, Mo)	Breccia infill, veins, veinlets, disseminated
Särkivaara: <i>hosted by the Kiruna Greenstone Group</i>	Tuffite, diabase, carbonate rock	scp, bt	py, po, cp, mo/ scp, cpx, ep, grt	Cu, Mo, (Au, Ni, Co, Zn, U, LREE)	Disseminated, patches
Veikavaara: <i>hosted</i>	Mafic	scp,	cp,	Cu,	Disseminated

<i>by the Veikavaara Greenstone Group</i>	sill, tuffite	cpx, act	po/ scp, cpx, act, grt	(Co)	
<i>Ö.Äijärvi: hosted by the Kiruna Greenstone Group</i>	Granite, gabbro and tuffite	ab, scp, act, di, epi	cp, po, py/ /ab, act, scp, di	Cu, (Ni, Co, V, Zn, Mo)	Disseminated, veinlets
Svecofennian-hosted; intrusion related Cu-Au					
<i>Aitik: hosted by rocks of the Haparanda Suite and Muorjevare Group</i>	Diorite, andesitic volcanics	bt, ser, grt, K-fsp, scp, ep, tur	cp, py, po, (mt, mo)/ qz, bt, ser, grt, act, brt	Cu, Au, Ag, (Mo)	Disseminated, veinlets, patches
<i>Pikkujärvi: hosted by rocks of the Perthite-Monzonite Suite and Kiirunavaara Group</i>	Quartz-monzonite, monzonite, basalt	K-fsp, bt, scp, tur, ep	cp, py, mt (mo)/ K-fsp, scp, act, ep	Cu, (Au, Co)	Disseminated, veinlets
Svecofennian-hosted; shear-zone related Cu-Au:					
<i>Ahmavuo: hosted by the Porphyrite Group</i>	Andesite	bt, scp, K-feld, ab, ep	py, cp, mt, (mo)/ fsp, act	Cu, Co, Au, (Fe, Mo)	Breccia infill, disseminated, veins
<i>Gruvberget-Cu: hosted by the Porphyrite Group</i>	Mafic to intermediate volcanics	K-fsp, scp, ep	cp, bn, mt, (py, mo)/K-fsp, ep, qz, scp, cal	Cu, (Au, Mo)	Disseminated, veinlets, patches
<i>Kiskamavaara: hosted by the Porphyrite Group</i>	Andesite breccia	K-fsp, ser, scp, tur, chl	cp, py, mt, ht, (mo)/ cal, qz, chl	Cu, Co, Au, (Fe, Mo)	Breccia infill, disseminated
<i>Nautanen: hosted by the Muorjevaara Group</i>	Mafic to intermediate volcanics	K-fsp, grt, bt, scp, act, tur, ser	cp, mt, py, bn/ act, bt, grt, cpx, tur, qz, ser, ap, cal	Cu, Au, (Ag, Fe)	Disseminated, veins, veinlets
<i>Rakkurijärvi-Discovery zone: hosted by rocks of the Porphyrite and Kiirunavaara Groups</i>	Intermediate porphyres	ab, bt, scp, act, K-fsp, cb	cp, mt, py, (mo)/ act, scp, ab, cal, chl	Cu, Au, Fe, (Mo, LREE, Co, Ni)	Breccia infill, disseminated, veinlets

Tjårojokka-Cu: <i>hosted by the Porphyrite Group</i>	Andesite	K- fsp, scp, bt, ab, chl	cp, bn mt, (py)/ ap, act, scp	Cu, (Fe, P, LREE)	Disseminated and veinlets
Karelian-hosted; epigenetic vein-style Cu-Au					
Kovo: <i>hosted by the Kovo Group</i>	Mafic sill	chl, ep	cp, py (mt, ht)/ qz, brt, cal	Cu, Ba	Quartz-barite veins
Svecofennian-hosted; epigenetic vein-style Cu-Au					
Ferrum: <i>hosted by the Muorjevara Group</i>	Andesitic volcaniclastics	scp, tur, bt	cp, bn, (py, mt, mo)/qz, tur, cpx, act, cal	Cu, Au, (Ag)	Quartz-tourmaline veins
Fridhem: <i>hosted by the Muorjevara Group</i>	Andesitic volcaniclastics	tur, scp, bt, chl, ep	cc, (bn, gld, ht)/ qz	Au, (Cu)	Quartz veins
Lieteksavo: <i>hosted by rocks of the Porhyrite and Kiirunavaara Groups</i>	Mafic sill and basaltic to rhyodacitic volcanic rocks	scp, bt, tur, K-fsp	cp, bn, mo/ qz, tur	Cu, Au, Ag, (W, Mo)	Quartz-tourmaline veins
Pahtavaara: <i>hosted by the Porhyrite Group</i>	Andesitic lava	ab, bt, chl, cb	cp, py, mt/ cb, (qz)	Cu, (Fe, Au, Sc)	Carbonate veins, breccia infill

Mineral abbreviations: act=actinolite, ab=albite, anh=anhydrite, ap=apatite, bn=bornite, brt=barite, bt=biotite, cal=calcite, cb=carbonate, cc=chalcocite, chl=chlorite, chn=chondrodite, cp=chalcopyrite, cpx=clinopyroxene, di=diopside, dol=dolomite, ep=epidote, fa=fayalite, fsp=feldspar, gld=gold, grt=garnet, gru=grunerite, ht=hematite, K-fsp=K feldspar, mo=molybdenite, ms=moscovite, mt=magnetite, phl=phlogopite, po=pyrrhotite, px=pyroxene, py=pyrite, qz=quartz, scp=scapolite, ser=sericite, srp=serpentine, skn=skarn, ttn=titanite, tur=tourmaline, trm=tremolite.