

THE GEOLOGY OF THE PROTEROZOIC HAVERI AU-CU DEPOSIT, SOUTHERN FINLAND

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TOBY A L STRAUSS

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ABSTRACT

The Haveri Au-Cu deposit is located in southern Finland about 175 km north of Helsinki. It occurs on the northern edge of the continental island arc-type, volcano-sedimentary Tampere Schist Belt (TSB) within the Palaeoproterozoic Svecofennian Domain (2.0 – 1.75 Ga) of the Fennoscandian Shield. The 1.99 Ga Haveri Formation forms the base of the supracrustal stratigraphy consisting of metavolcanic pillow lavas and breccias passing upwards into intercalated metatuffs and metatuffites. There is a continuous gradation upwards from the predominantly volcanoclastic Haveri Formation into the overlying epiclastic meta-greywackes of the Osara Formation. The Haveri deposit is hosted in this contact zone. This supracrustal sequence has been intruded concordantly by quartz-feldspar porphyries.

Approximately 1.89 Ga ago, high crustal heat flow led to the generation and emplacement of voluminous synkinematic, I-type, magnetite-series granitoids of the Central Finland Granitoid Complex (CFGK), resulting in coeval high-T/low-P metamorphism (hornfelsic textures), and D₁ deformation. During the crystallisation and cooling of the granitoids, a magmatic-dominated hydrothermal system caused extensive hydrothermal alteration and Cu-Au mineralisation through the late-D₁ to early-D₂ deformation. Initially, a pre-ore Na-Ca alteration phase caused albitisation of the host rock. This was closely followed by strong Ca-Fe alteration, responsible for widespread amphibolitisation and quartz veining and associated with abundant pyrrhotite, magnetite, chalcopyrite and gold mineralisation. More localised calcic-skarn alteration is also present as zoned garnet-pyroxene-epidote skarn assemblages with associated pyrrhotite and minor sphalerite, centred on quartz-calcite±scapolite veinlets. Post-ore alteration includes an evolution to more K-rich alteration (biotitisation). Late D₂-retrograde chlorite began to replace the earlier high-T assemblage. Late emanations (post-D₂ and pre-D₃) from the cooling granitoids, under lower temperatures and oxidising conditions, are represented by carbonate-barite veins and epidote veinlets. Later, narrow dolerite dykes were emplaced followed by a weak D₃ deformation, resulting in shearing and structural reactivation along the carbonate-barite bands. This phase was accompanied by pyrite deposition.

Both sulphides and oxides are common at Haveri, with ore types varying from massive sulphide and/or magnetite, to networks of veinlets and disseminations of oxides and/or sulphides. Cataclastites, consisting of deformed, brecciated bands of sulphide, with rounded and angular clasts of quartz vein material and altered host-rock are an economically important ore type. Ore minerals are principally pyrrhotite, magnetite and chalcopyrite with lesser amounts of pyrite, molybdenite and sphalerite. There is a general progression from early magnetite, through pyrrhotite to pyrite indicating increasing sulphidation with time. Gold is typically found as free gold within quartz veins and within intense zones of amphibolitisation. Considerable gold is also found in the cataclastite ore type either as invisible gold within the sulphides and/or as free gold within the breccia fragments.

The unaltered amphibolites of the Haveri Formation can be classified as medium-K basalts of the tholeiitic trend. Trace and REE support an interpretation of formation in a back-arc basin setting. The

unaltered porphyritic rocks are calc-alkaline dacites, and are interpreted, along with the granitoids as having an arc-type origin. This is consistent with the evolution from an initial back-arc basin, through a period of passive margin and/or fore-arc deposition represented by the Osara Formation greywackes and the basal stratigraphy of the TSB, prior to the onset of arc-related volcanic activity characteristic of the TSB and the Svecofennian proper.

Using a combination of petrogenetic grids, mineral compositions (garnet-biotite and hornblende-plagioclase thermometers) and oxygen isotope thermometry, peak metamorphism can be constrained to a maximum of approximately 600 °C and 1.5 kbars pressure. Furthermore, the petrogenetic grids indicate that the REDOX conditions can be constrained at 600 °C to $\log f(\text{O}_2)$ values of approximately -21.0 to -26.0 and -14.5 to -17.5 for the metasedimentary rocks and mafic metavolcanic rocks respectively, thus indicating the presence of a significant REDOX boundary.

Amphibole compositions from the Ca-Fe alteration phase (amphibolitisation) indicate iron enrichment with increasing alteration corresponding to higher temperatures of formation. Oxygen isotope studies combined with limited fluid inclusion studies indicate that the Ca-Fe alteration and associated quartz veins formed at high temperatures (530 – 610 °C) from low CO_2 , low- to moderately saline (<10 eq. wt% NaCl), magmatic-dominated fluids. Fluid inclusion decrepitation textures in the quartz veins suggest isobaric decompression. This is compatible with formation in high-T/low-P environments such as contact aureoles and island arcs.

The calcic-skarn assemblage, combined with phase equilibria and sphalerite geothermometry, are indicative of formation at high temperatures (500 – 600 °C) from fluids with higher CO_2 contents and more saline compositions than those responsible for the Fe-Ca alteration. Limited fluid inclusion studies have identified hypersaline inclusions in secondary inclusion trails within quartz. The presence of calcite and scapolite also support formation from CO_2 -rich saline fluids. It is suggested that the calcic-skarn alteration and the amphibolitisation evolved from the same fluids, and that P-T changes led to fluid unmixing resulting in two fluid types responsible for the observed alteration variations. Chlorite geothermometry on retrograde chlorite indicates temperatures of 309 – 368 °C. As chlorite represents the latest hydrothermal event, this can be taken as a lower temperature limit for hydrothermal alteration and mineralisation at Haveri.

The gold mineralisation at Haveri is related primarily to the Ca-Fe alteration. Under such *P-T-X* conditions gold was transported as chloride complexes. Ore was localised by a combination of structural controls (shears and folds) and REDOX reactions along the boundary between the oxidised metavolcanics and the reduced metasediments. In addition, fluid unmixing caused an increase in pH, and thus further augmented the precipitation of Cu and Au. During the late D_2 -event, temperatures fell below 400 °C, and fluids may have remobilised Au and Cu as bisulphide complexes into the shear-controlled cataclastites and massive sulphides.

The Haveri deposit has many similarities with ore deposit models that include orogenic lode-gold deposits, certain Au-skarn deposits and Fe-oxide Cu-Au deposits. However, many characteristics of the Haveri deposit, including tectonic setting, host lithologies, alteration types, proximity to I-type granitoids and *P-T-X* conditions of formation, compare favourably with other Early Proterozoic deposits within the TSB and Fennoscandia, as well as many of the deposits in the Cloncurry district of Australia. Consequently, the Haveri deposit can be seen to represent a high-T, Ca-rich member of the recently recognised Fe-oxide Cu-Au group of deposits.

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1. INTRODUCTION AND REGIONAL GEOLOGY

The study area (Figure 1.1) is centred on the Haveri Au-Cu deposit, located in Southern Finland about 175 km north of Helsinki, and 35 km north west of Tampere, Finland's second biggest city. The nearest village is Viljakkala, located 1 km east of the mine. The setting is rural, with farms, forests and lakes.

The Haveri deposit has been the site of periodic mining for over 250 years. In the most active part of its history (1942 to 1960), the main commodities were copper and gold, which were extracted from open pit and shallow underground workings. Since 1962 the property has been inactive, except for limited base metal exploration in the immediate area, and the more recent Au exploration carried out by Glenmore Highlands Inc. in the late 1990's.

The Haveri Au-Cu deposit is situated on the northern edge of the Tampere Schist Belt within the Palaeoproterozoic Svecofennian Domain of the Fennoscandian Shield.

1.1 OVERVIEW OF THE FENNOSCANDIAN SHIELD

The East European Craton is composed of three crustal segments (Figure 1.2): Sarmatia, Volgo-Uralia and Fennoscandia. Fennoscandia is the northern crustal segment, comprising the Fennoscandian Shield and its buried continuation (Gorbatshev and Bogdanova, 1993). The crustal segments are separated by junctures, that since the Mesoproterozoic have been the approximate sites of the rifting that eventually created the Volhyn-Orsha, Central Russian, and Pachelma systems of palaeorifts, aulacogens and depressions (Gorbatshev and Bogdanova, 1993). The south-west margin of the craton is defined by the Tornquist Line / Transeuropean Fault system which separates the older and thicker crust of north-eastern Europe from the hotter, thinner crust of the Phanerozoic mobile belts in central and western Europe (Gorbatshev and Bogdanova, 1993).

The Fennoscandian crustal segment consists of the Archaean Domain in the northeast, the Palaeoproterozoic Svecofennian Domain in the centre and the Southwest Scandinavian Domain in the west. This geochronological zoning with younging from the northeast to the southwest reflects the formation of the continental crust during three major consecutive orogenic events: Lopian orogeny (2.9 - 2.6 Ga), Svecofennian orogeny (2.0 – 1.75 Ga) and the Gothian orogeny (1.75 – 1.5 Ga) (Gaál and Gorbatshev, 1987; Gorbatshev and Bogdanova, 1993).

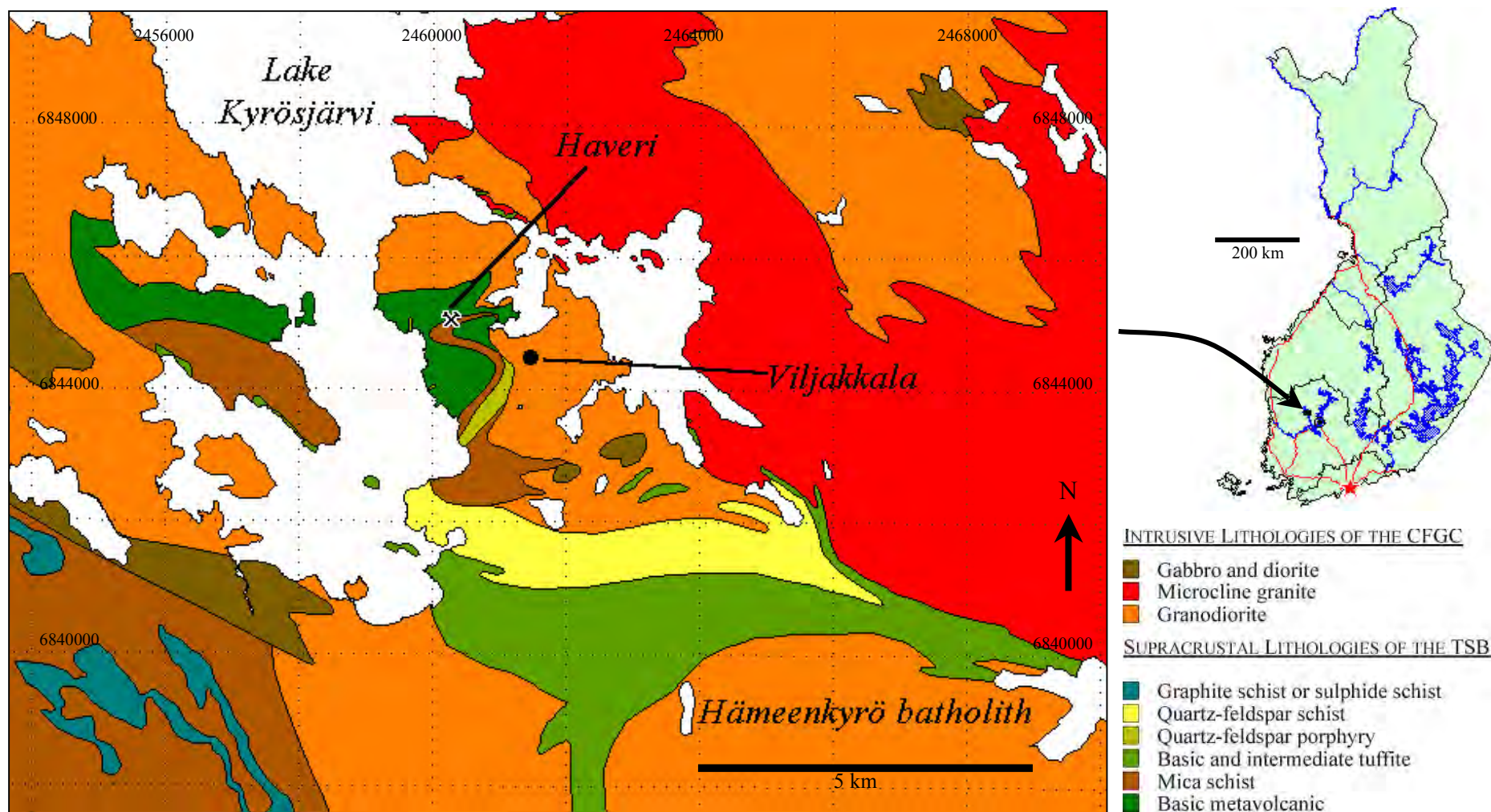


Figure 1.1: The location and the regional geology of the study area. Coordinates in the Finnish KKKJ-2 coordinate system.

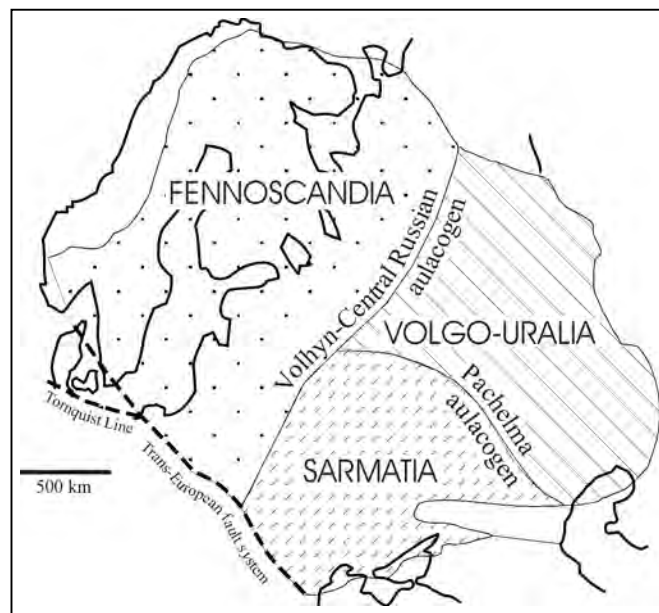


Figure 1.2: Crustal segments of the East European Craton (Gorbatshev and Bogdanova, 1993).

1.1.1 The Archaean Domain

The Archaean Domain of the Fennoscandian Shield has been subdivided into three crustal provinces, comprising the granite-greenstone terrain of the Karelian Province and the high-grade gneiss terrains of the Belomorian and Kola Peninsula Provinces (Figure 1.3).

The Karelian Province contains more than 20 major and numerous minor greenstone belts enveloped by Archaean tonalites. The greenstone belts have a pronounced linear alignment forming a pattern of NNW to NW and NS-trending occurrences (Gaál and Gorbatshev, 1987). The greenstone belts in the Karelian Province have an abundance of volcanic rocks, and only some of the belts exhibit the “classic” stratigraphy of basal komatiites and tholeiitic basalts followed by calc-alkaline volcanics and sediments. Other greenstone belts have calc-alkaline volcanics, basal pelites or metaturbidites as the lowermost units (Gaál and Gorbatshev, 1987). All of the greenstone belts have undergone metamorphism under conditions of low pressure and low- to medium-temperatures. Most of the radiometric dates fall into the range 2.9 – 2.6 Ga, and thus can be attributed to the Lopian orogeny.

The Belomorian Province, which lies to the northeast of the Karelian Province (Figure 1.3), is a high-grade gneiss region characterised by penetrative folding and metamorphism at elevated pressures. The protoliths were predominantly supracrustals and tonalitic-granodioritic plutonics. U-Pb dating of zircons indicates that the Belomorian Supracrustal Complex was deposited 2.9 – 2.7 Ga ago, with an intense magmatic-metamorphic event 2.8 – 2.7 Ga ago, producing the tonalitic-granodioritic rocks (Tugarinov and Bibikova 1980, in Gorbatshev and Bogdanova, 1993; Mitrofanov and Pozhilenko

1991, in Gorbatshev and Bogdanova, 1993). The dominant structure of the Belomorian sequence is a NW-trending linear synformal belt. Deformation and metamorphism have repeatedly affected all the rocks in the Belomorian Province.

The Kola Peninsula Province, which lies to the northeast of the Belomorian Province (Figure 1.3), contains another metasedimentary belt occurring in the linear synformal zone of the Keivy ridge. U-Pb dating of zircons indicates that the depositional age of the Kola Peninsula Gneisses is 2.9 – 2.7 Ga (Tugarinov and Bibikova 1980, in Gorbatshev and Bogdanova, 1993). The rocks in the Kola Peninsula Province exhibit complex polyphase deformation associated with early granulite facies metamorphism and later retrograde amphibolite facies metamorphism (Dobrzhinetskaya 1978, in Gaál and Gorbatshev, 1987).

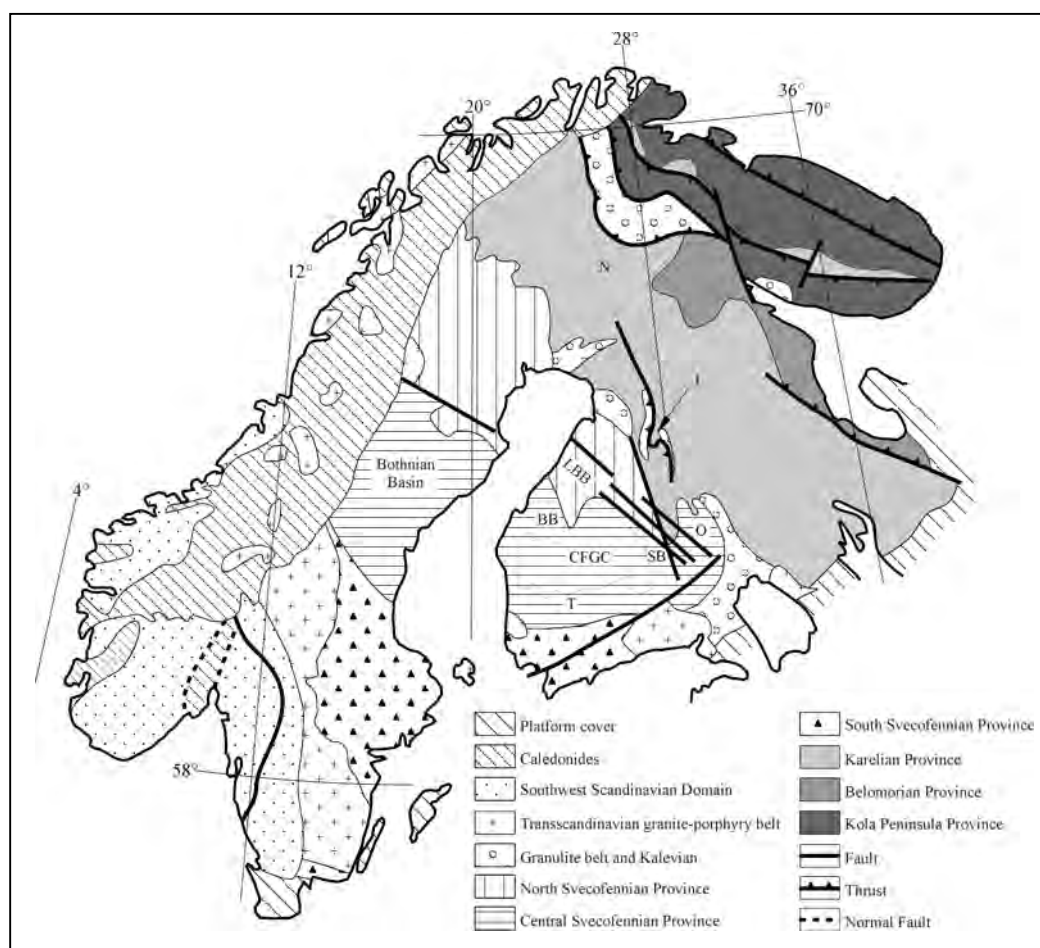


Figure 1.3: The major geological subdivisions of the Fennoscandian Shield. Ladoga – Bothnian Bay tectonic zone (LBB), Central Finland Granitoid Complex (CFGC), Tampere Schist Belt (T), Bothnia Belt (BB), Savo Belt (SB), Outokumpu Ophiolite (O), Jormua Ophiolite (J), Nuttio Serpentinities (N). Modified after Gaál and Gorbatshev (1987), Weihed et al. (1992) and Nironen (1997).

In summary, the development of the rocks within the Archaean Domain occurred in two main periods, firstly, the development of the tonalitic-trondhjemitic basement during the Saamian orogeny (> 3.0 Ga). This was followed by the Lopian orogeny (2.9 – 2.6 Ga) wherein the supracrustals were deposited onto Saamian basement (2.9 – 2.7 Ga ago), and were subsequently intruded by granitic rocks, deformed and metamorphosed (2.7 – 2.6 Ga ago). Gaál and Gorbatshev (1987) have suggested a plate-tectonic model for the Lopian orogeny, in which they interpret the high proportion of greywackes and abundant tonalitic-granodioritic plutonics in the Belomorian Province as partial evidence for crustal stacking. They therefore interpret the high-grade gneisses of the Belomorian Province (and Kola Peninsula Province) as representing a mobile-belt region, wherein the Belomorian has been subducted beneath the Saamian craton in the west, with overthrusting at higher crustal levels. In this model, the Karelian Province represents the rifted foreland complex of the Lopian orogeny.

1.1.2 Early Proterozoic: Svecokareliides

The Early Proterozoic Svecokareliides consist of two coeval but distinct parts: the Karelian schist belt (Kareliides) and the Svecofennian complex (Svecofennides) (Park, 1985). The Karelian (or Kareliides) refers to the Proterozoic rocks within the Archaean Domain. These are craton-covering rocks underlain by Archaean continental crust, forming a marginal zone along the southwest margin of the Archaean basement. They were deposited during the break-up of the Archaean continent in an extensional tectonic regime. The Svecofennian (or Svecofennides) refers to rocks from the Svecofennian Domain that have no (known) Archaean basement, and form juvenile Early Proterozoic crust. The orogenic rocks of the Svecofennides were formed in a compressional/transpressional tectonic regime, with the accretion of a succession of allochthonous volcanic arcs onto the Karelian continent along the NW-SE trending Lake Lagoda-Bothnian Bay lineament zone.

1.1.2.1 Early Proterozoic: Anorogenic Karelian Formations

At the end of the Lopian orogeny, tectonic activity within the Archaean Domain appears to have ceased and the Archaean Domain was converted into a cratonic realm. There then followed a period of anorogenic supracrustal deposition that lasted throughout the Late Archaean and Early Proterozoic (Gaál and Gorbatshev, 1987; Lehtonen et al., 1998). The oldest rocks covering the eroded Archaean Basement are the Salla and Onkamo Groups together consisting of a 200-metre thick basal unit of ultramafic pyroclastic rocks and lava-flows. The Salla and Onkamo Groups are interpreted to have been deposited during the formation of an intracratonic rift (Lehtonen et al., 1998). A lower age limit for the Salla Group is provided by a crosscutting 2.44 Ga old gabbro (Alapieti, 1982). Layered gabbro complexes of this age are common in northern Finland, and occur in a roughly 300 km-long belt.

The Salla and Onkamo Groups can be correlated with the Sumian-Sariolan Group, which occurs further south in Finland and in Russia (Lehtonen et al., 1998). The Sumian-Sariolan Group was deposited onto faulted Archaean basement of considerable topographic relief, and consists of immature clastic sediments with interbeds and intrusions of mafic and felsic eruptives. Marmo and Ojakangas (1984) have interpreted the diamictites as glaciogenic deposits, whereas the coarse-grained sediments and talus breccias are interpreted as molasse deposits from the end of the Lopian orogeny (Sokolov and Heiskanen, 1985). The deposition of the Sumian-Sariolan rocks was restricted to linear fault-controlled troughs trending NW to NNW (Gaál and Gorbatshev, 1987).

The abundant volcanic activity of the Salla and Onkamo Groups was followed by a more tranquil period during which thick and widespread epiclastic sediments were deposited in either a cratonic setting or at a cratonic margin (Lehtonen et al., 1998). These sediments are represented by the arkosic and micaceous metasediments, metapelites and quartzites of the Sodankylä Group, which can be constrained to a minimum age of 2.2 Ga by intrusive sills (Hanski, 1987; Lehtonen et al., 1998).

With gradual deepening of the depositional basin, the Sodankylä Group sediments give way to the black schists and mafic and ultramafic volcanics of the Savukoski Group followed by the mainly volcanic sequence with intercalated sediments represented by the Kittilä Group. The Kittilä Group volcanics give way to mica schists and arkosic greywackes, that are interpreted to represent the final stages in the opening and evolution of an oceanic basin (Lehtonen et al., 1998). A minimum age for the Savukoski and Kittilä Groups is provided by 2.05 Ga old diabase dykes and ~2.0 Ga old felsic and basic metavolcanic rocks respectively (Lehtonen et al., 1998).

The development of the overlying Lainio Group commenced with calc-alkaline volcanic activity, and contains meta-arkoses, quartzites and conglomerates. These rocks are interpreted to have formed in an island-arc environment ~ 1.9 Ga ago (Lehtonen et al., 1998). The < 1.88 Ga Kumpu Group contains molasse-type meta-arkoses and conglomerates that are interpreted as having formed in a foreland basin environment (Lehtonen et al., 1998). Consequently, the Lainio and Kumpu Groups represent the initial onset of Svecofennian collisional tectonism after the development of a passive continental margin, and stabilisation of the Archaean crust. Prior to the deposition of the marine sequence of the Kalevian Group, was a period of intense faulting, accompanied by the intrusion of predominately NW-trending dolerite dyke swarms into the Archaean basement and its cover. The faulting and dyke intrusions along the western margin of the Karelian Province signify an extensional tectonic regime, and mark the onset of rifting, and the break-up of the Archaean craton (Gaál and Gorbatshev, 1987). The dolerite dykes have been dated at between 2.25 and 2.0 Ga (Sakko, 1971, in Gaál and Gorbatshev, 1987).

Prior to the main period of Svecofennian compressional tectonics, there existed a pre-orogenic phase of extensional tectonism, represented by the Kalevian sediments and the formation of oceanic crust. These rocks have commonly been grouped into the Svecofennian (Gaál and Gorbatshev, 1987). Consequently the Svecofennian can be divided into a pre-orogenic extensional phase, representing the

culmination of the break-up of the Archaean craton, and the main orogenic phase of compressional tectonics represented by the Svecofennian proper (Gaál and Gorbatshev, 1987).

The Kalevian Group consists of turbiditic sediments intercalated with tholeiitic volcanics, which developed on the passive continental margins in a series of N-S to NNW-trending, sub-parallel intracratonic troughs (Ward, 1987). The lower part of the Kalevian Sequence consists of an association of dolomites, Lake Superior-type iron formations, fluxo-turbidites, mid-fan turbidites and conglomerates, reflecting shelf to near-shelf environments close to a continental margin. The upper part of the Sequence contains distal turbidites and slices of ophiolites (Gaál and Gorbatshev, 1987; Kohonen, 1995).

The 1.97 – 1.95 Ga old ophiolite-like complexes at Outokumpu (Koistinen, 1981), Jormua (Kontinen, 1987; Peltonen et al., 1996) and the Nuttio serpentinite belt (Hanski, 1997) represent remnants of oceanic crust formed by rifting in an extensional tectonic regime (Nironen, 1997). Although the depositional age of the Kalevian sediments is not fully constrained, analyses of detrital zircons in two “Upper Kalevian” greywackes indicates that a major area of 2.1 – 1.9 Ga old crust was in an erosional position approximately 1.9 Ga ago (Claesson et al., 1993). A maximum age for the sediments is provided by the 1.96 Ga Outokumpu and Jormua ophiolites.

1.1.2.2 Early Proterozoic: Orogenic Svecofennian Domain

The Svecofennian is generally regarded to refer to the rocks formed between 2.0 and 1.75 Ga ago (Gaál and Gorbatshev, 1987) during the “Svecofennian Orogen” in a compressional tectonic regime (Kähkönen et al., 1989). At some stage after the formation of oceanic crust in an extensional environment 1.95 Ga ago, the tectonic regime changed to one of compression, and the main phase of Svecofennian crustal formation commenced. Although granitoids are now the principal component (Front and Nurmi, 1987), Gaál (1987) has divided the Svecofennian Domain into three provinces on the basis of supracrustal lithologies. These are the volcanic Northern and Southern Svecofennian Provinces and the intervening sedimentary Central Svecofennian Province (Figure 1.3). Stratigraphic relationships between the three provinces have been largely obscured by the voluminous granitoid intrusions that occupy the Svecofennian Domain.

1.1.2.2.1 *Svecofennian Volcanic Belts and the Bothnian Basin*

The two volcanic belts have many characteristics in common, including, at lower stratigraphic levels, a predominance of calc-alkaline volcanic suites (mainly rhyolites and dacites) interbedded with carbonate rocks and minor epiclastic sediments. Higher in the stratigraphy, tholeiitic basalts, occasional ultramafic volcanics, conglomerates, metapelites and greywackes dominate (Gaál and Gorbatshev, 1987). In both the Northern and Southern Svecofennian Provinces, thick basal greywackes suggest continental margin environments (Gaál and Gorbatshev, 1987). The Central Svecofennian Province is made up chiefly of metagreywackes and metapelites, with subordinate, mostly mafic metavolcanic rocks. These rocks were deposited in a marine basin, commonly referred to as the Bothnian Basin

(Gaál and Gorbatshev, 1987; Hietanen, 1975). In the centre of this basin, the metagreywackes attain thicknesses of 10 kilometres (Gaál and Gorbatshev, 1987; Lundqvist, 1987).

1.1.2.2.2 *Svecofennian Granitoids*

Granitoids are the predominant component of the Svecokarelian crust (Front and Nurmi, 1987). The Proterozoic granitoids have traditionally been divided into three principal groups: early-, late- and post-orogenic granites, in terms of their relationship to the “Svecofennian Orogeny” (Andersson, 1991; Gaál and Gorbatshev, 1987).

The early orogenic, synkinematic granitoids make up the bulk of the Svecofennian crust. They have U-Pb zircon ages from about 1900 to 1860 Ma (with a peak between 1890 and 1870 Ma), being roughly coeval with the Svecokarelian metavolcanic and mafic plutonic rocks (Aho, 1979; Simonen, 1980; Neuvonen et al., 1981; Huhma, 1986; Patchett and Kouvo, 1986; Front and Nurmi, 1987). The synkinematic granitoids occur in schist belts as intrusions of varying size, and form the granitoid complex of central Finland, which occupies an area of more than 35,000 km² (Nurmi and Haapala, 1986). The granitoids in schist belts form concordant circular or elongate intrusions of variable size. (Nurmi and Haapala, 1986; Front and Nurmi, 1987). The granitoids generally crystallised after the main deformation phase, and soon after the peak of metamorphism, but before the younger deformation phases, which are locally visible as discordant foliations or shear zones (Front and Nurmi, 1987). The Central Finland Granitoid Complex (CFGC) (Figure 1.3) is made up of a vast number of plutons varying considerably in size. The contacts between the various granitoids are intrusive, tectonic, or else gradational over metres to tens of metres. In spite of distinct tectonic differences, U-Pb zircon ages from different granitoids are practically indistinguishable (Nurmi and Haapala, 1986; Front and Nurmi, 1987). Supracrustal rocks occur within the CFGC as minor schist belts, and commonly also as xenoliths of medium and high-grade mafic to felsic gneiss (Front and Nurmi, 1987). Chemically, the early, synkinematic granitoids are a differentiated suite of calcic and calc-alkaline, I-type rocks ranging from gabbros and diorites through to more dominant tonalites-trondhjemites and granodiorites-granites (Front and Nurmi, 1987; Gaál and Gorbatshev, 1987). Isotopic and geochemical data suggest that they derive predominantly from material newly segregated from the mantle during the early Proterozoic (Huhma, 1986; Patchett and Kouvo, 1986; Gaál and Gorbatshev, 1987; Patchett et al., 1987). By about 1.87 Ga ago, most of the Svecofennian crust had been sufficiently consolidated to allow rifting and the formation of several swarms of 1.87 – 1.83 Ga dolerite dykes (Gaál and Gorbatshev, 1987).

The late-orogenic Svecofennian granites have U-Pb ages of 1.83 – 1.77 Ga (Andersson, 1991; Gaál and Gorbatshev, 1987; Patchett et al., 1987). These granitoids are generally crustal, anatectic, mostly S-type undifferentiated granites with only a very minor granodioritic component (Gaál and Gorbatshev, 1987). The late-kinematic granites are closely associated with high-grade metamorphosed Svecokarelian mica gneisses and schists, migmatites and large amounts of pegmatite, and are especially widespread in the area south of the CFGC (Gaál and Gorbatshev, 1987; Nurmi and Haapala, 1986). In the two volcanic belt provinces, the Late Svecofennian granites form numerous minor massifs and dyke networks, whereas they form large masses in the old major sedimentation basins, such as the western part of the Central Svecofennian Province (Gaál and Gorbatshev, 1987). The late-kinematic granites define an interval of extensive reworking of the pre-existing crust, associated with minor juvenile additions (Andersson, 1991).

The late-orogenic Svecofennian granites include the “post-kinematic” granitoids of Front and Nurmi (1987) and Nurmi and Haapala (1986), which are present as small, roundish, composite stocks and ring dykes, intruded around 1800 Ma ago. The rock types vary from quartz monzodiorite and tonalite to quartz monzonite, granodiorite and granite. These granitoids clearly crosscut the late-kinematic Svecofennian granites (Claesson and Lundqvist, 1995; Nurmi and Haapala, 1986), and also differ in terms of geochemistry and magma type, with the “post-kinematic” granitoids being formed from volatile-poor, high-temperature melts (Claesson and Lundqvist, 1995). However, the synchronous crystallisation of the 1.83 – 1.77 Ga old “late-kinematic” and the 1.80 Ga old “post-kinematic” granitoids is clearly confusing, and this terminology to differentiate these two groups should be avoided (Andersson, 1991). Consequently any granitoids from the second episode of plutonism within the Svecofennian Domain, dated between 1.83 – 1.77 Ga, are all referred to as late-orogenic Svecofennian.

The anorogenic granites (forming the Trans-Scandinavian granite-porphyry belt) are the third major episode of granitoid generation in the Svecofennian, covering a period between 1.70 – 1.54 Ga (Vaasjoki, 1977; Front and Nurmi, 1987; Gaál and Gorbatshev, 1987; Nurmi and Haapala, 1986; Andersson, 1991). These are predominately large multiple batholiths and stocks of rapakivi granites with anorogenic (A-type) characteristics, and minor coeval gabbros, anorthosites and diabases (Andersson, 1991; Front and Nurmi, 1987; Gaál and Gorbatshev, 1987; Nurmi and Haapala, 1986). These were generated by crustal melting associated with the injection of mantle-derived magmas in a tensional intracratonic environment (Andersson, 1991).

1.1.2.3 Tectonic models for the evolution of the “Svecokarelian Orogen”

The first attempt at describing the Svecofennian in terms of plate tectonics was by Hietanen (1975). This model proposed a Cordilleran-style tectonic regime whereby an Early Proterozoic oceanic plate was subducted to the northeast under the Karelian continent, producing an island arc and an interarc basin (the greywacke-rich Bothnian Basin) (Figure 1.4). Since this initial work a number of increasingly complex models have been proposed for the tectonic evolution of the Early Proterozoic.

With the exception of the model proposed by Welin (1987), all of the models for the evolution of the Svecofennian are based on plate tectonics. Although most of these models agree both on the Circum-Pacific-type collision model of Hietanen (1975), and on the accretion of one or more island-arcs onto the craton edge, there are considerable differences between the models in terms of the number of arcs involved, the location of their formation and the directions of subduction.

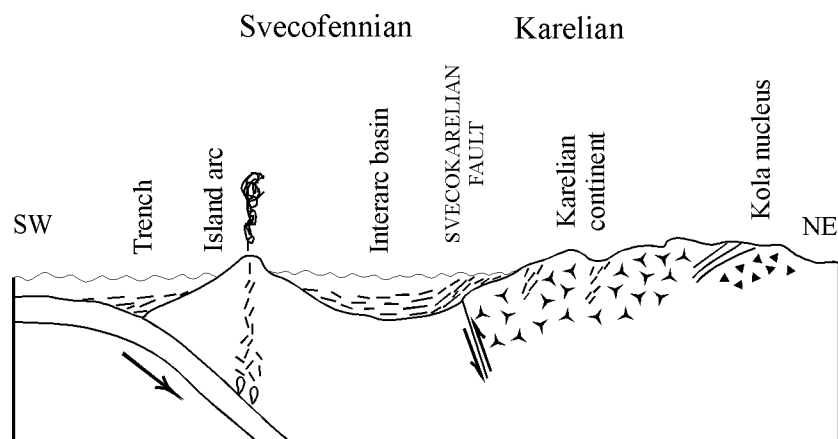


Figure 1.4: Cross-section from southwest to northeast through the Fennoscandian Shield showing the possible Svionian island arc 2.5 to 2.0 Ga ago. Deposition of quartz sand, dolomite and pelite on the Karelian continental shelf, deposition of greywacke-type sediments in the Bothnian interarc basin, and volcanism in the Svionian island arc were coeval. The oldest nucleus is in the Kola Peninsula at right, and the subduction zone is marked by a trench on the surface at left. After Hietanen (1975).

Some of the earlier models (Gaál, 1986; Gaál and Gorbatshev, 1987; Park, 1985; Park et al., 1984; Pharaoh and Pearce, 1984) had an initial eastward subduction of oceanic crust, creating a marginal basin and the formation of oceanic crust (Jormua, Outokumpu and Nuttio “ophiolite” assemblages) in a back-arc environment. However, subsequent studies (Kontinen, 1987; Peltonen et al., 1996) suggest that the 1.97 Ga old Jormua ophiolite formed as an incipient ocean rather than as a back-arc basin. This has led to more recent models of an earlier phase of divergence with marginal rifts and aulacogens (Gaál, 1990; Nironen, 1997; Lahtinen, 1994).

Gaál (1990) proposed that the Early Proterozoic continental margin split about 2.0 Ga ago, with the development of NW-SE-trending marginal basins parallel to the continental margin. This was followed by westward subduction and the development of oceanic island arcs between 1.93 – 1.90 Ga ago. The oceanic arcs and the contents of the marginal basin were subsequently accreted onto the Archaean craton. Further compression caused a reversal of the subduction to the east under the newly formed Early Proterozoic crust, giving rise to extensional continental island arcs 1.88 Ga ago. Retreat of the subduction zone to the west and magmatic underplating led to the generation of large volumes of granite and to westward accretion of continental crust 1.81 – 1.7 Ga ago.

Park et al. (1984) and Park (1985, 1991) recognised that many of the Svecofennian belts have east-west trends that differ from the NW-SE trends of the pre-Svecofennian rocks in the Karelides. This difference has been interpreted as the northward movement of oceanic lithosphere, where successively formed early Proterozoic, E-W-trending oceanic island-arcs moved northwards and were accreted onto the stable craton (Figure 1.5). In northern Sweden, subduction occurred directly beneath the craton producing the contemporaneous Skellefteå arc. Transcurrent movements along the NW-SE-trending Raahe – Ladoga lineament, blocked the Skellefteå subduction zone. By ca. 1.90 Ga, the site of subduction had shifted southwards from the Skellefteå belt to firstly the intraoceanic Tampere Belt, and then subsequently the Skåldö arc and then to the Örebro arc. As the Svecofennides accreted onto the Karelides and their Archaean basement, stresses were accommodated along the Raahe – Ladoga lineament, initially by thrusts and NW-trending dextral wrench faults, and then later by wrench faults alone, which remained sites of movement and granitoid emplacement for a further 100 Ma. The resultant “Svecofennian Orogen”, consisting of a stable craton, obducted nappes, exotic terranes, transcurrent faults, and island arcs that moved obliquely to the margin of the craton, therefore shows many similarities to the western cordillera of N America.

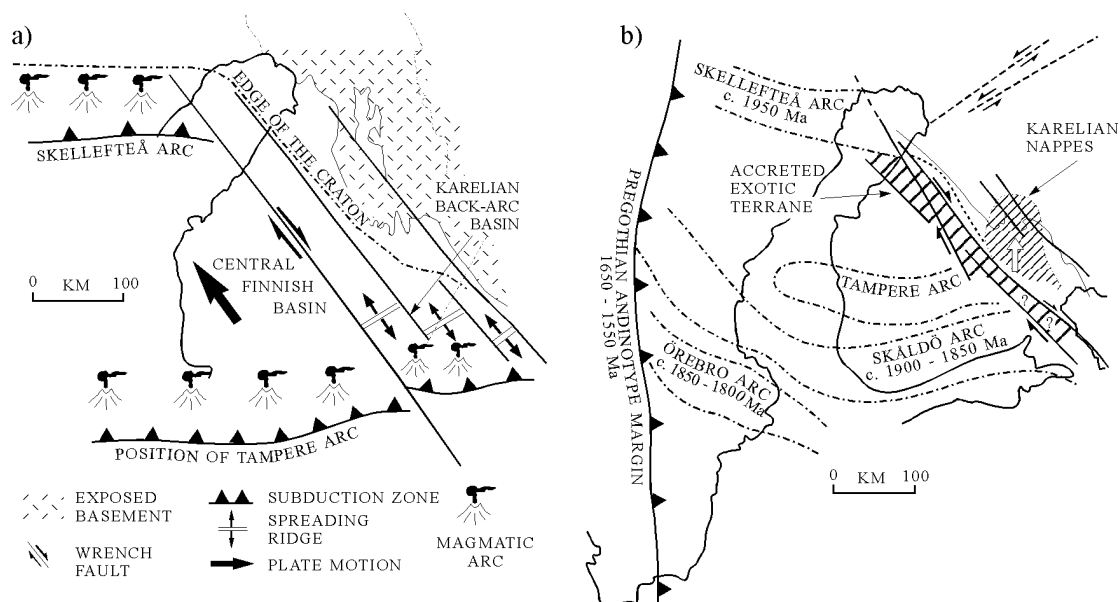


Figure 1.5: a) Schematic reconstruction of Sarmatian margin ca. 2.0 – 1.95 Ga, at time of active Skellefteå arc and Karelian back-arc basin and arc; eventual position of younger Tampere arc is shown. b) Major geochronological provinces in Svecokarelides and pregothides and proposed arc terranes; position of allochthonous exotic terranes in Karelia and Pohjanmaa is indicated. After Park (1985)

According to this model, the arcs in the south are younger than those in the north. However, there is no evidence of age zoning of this kind (Patchett and Kouvo, 1986; Skiöld, 1987; Welin, 1987; Vaasjoki and Sakko, 1988; Kähkönen et al., 1989), and the abundant data available on the plutonic rocks indicate that the formation of Svecofennian crust 1.90 – 1.87 Ga ago occurred simultaneously over a wide area. This process was complex and may have involved several coexisting spreading-subduction

systems with different subduction directions and polarities (Kähkönen et al., 1989). Such models for the evolution of the Svecofennian involving up to three successive orogenic terranes and collision events have been proposed by Gaál (1990), Ekdahl (1993), Lahtinen (1994, 1996, 2000), Lahtinen and Huhma (1997) and Nironen (1997). These models differ principally in the number and direction of subduction events. However, there are many similarities between the models of Lahtinen (1994, 1996, 2000), Lahtinen and Huhma (1997) and Nironen (1997), and consequently, there is at present a broad consensus for the development of the Svecofennian, which will be outlined below according to Nironen (1997).

The model detailed by Nironen (1997) proposes an early extensional phase, leading to the development of the Outokumpu, Jormua and Nuttio “ophiolites” 1.95 Ga ago. These were formed either in marginal rifts due to intracratonic rifting and perhaps mantle plumes or, alternatively, on a passive margin in developing aulacogens, with as yet unrecognised megablocks separating from the Archaean craton. Regardless of the exact setting for the extensional phase, an arc system consisting of several arcs, basins and subduction zones had developed to the (present) southwest of the Archaean craton over an inferred SW-dipping subduction zone. The arcs had amalgamated to form new felsic crust. Island-arc basalts and andesites and their infracrustal equivalents (e.g. the 1.93 – 1.92 Ga volcanics and plutonics of the Savo Belt) had been formed at the active margin of an older Palaeoproterozoic nucleus (Central Finland Granitoid Complex). There was a large ocean basin (the Bothnian Basin) between this arc complex and another arc complex (including southern Finland and south-central Sweden) further away.

This arc complex was accreted to the continent about 1.91 Ga ago. At the same time, an ocean basin was closing to the north (Figure 1.6a). During this phase the Archaean crust started to thicken tectonically, and the craton margin ophiolites, as well as early Proterozoic cover sequences (Karelian Formations) were thrust NE onto the Archaean crust.

After this initial accretion, there was either subduction reversal and rotation of plate motion direction, or alternatively another (micro)plate approached from the south, and plate convergence was accommodated by an E-W trending subduction zone (Figure 1.6b). The Tampere arc developed 1.90 Ga ago at the southern margin of the arc complex. The Skellefteå, Bothnia and Tampere volcanics comprise volcanic arcs with different tectonic settings: Tampere represents active continental margin volcanism, and Bothnia and Skellefteå island-arc volcanism. Since plate convergence at this stage was oblique with respect to the continental margin, it led to lateral displacement at the margin, and cratonic megablocks were translated northwards. This lateral displacement marks the initial development of the Lake Ladoga – Bothnian Bay zone (Raahe – Ladoga lineament). The subduction zone was pinned in the east (below the Tampere arc) and the western part of the zone started to curve while migrating northwards. Eventually, a dextral transform fault developed between two curving subduction zones.

Simultaneously, to the south another arc complex including the Hame and Bergslagen volcanics with an older Palaeoproterozoic nucleus was developing 1.89 Ga ago. This arc complex approached from the south overriding a retreating southerly dipping subduction zone.

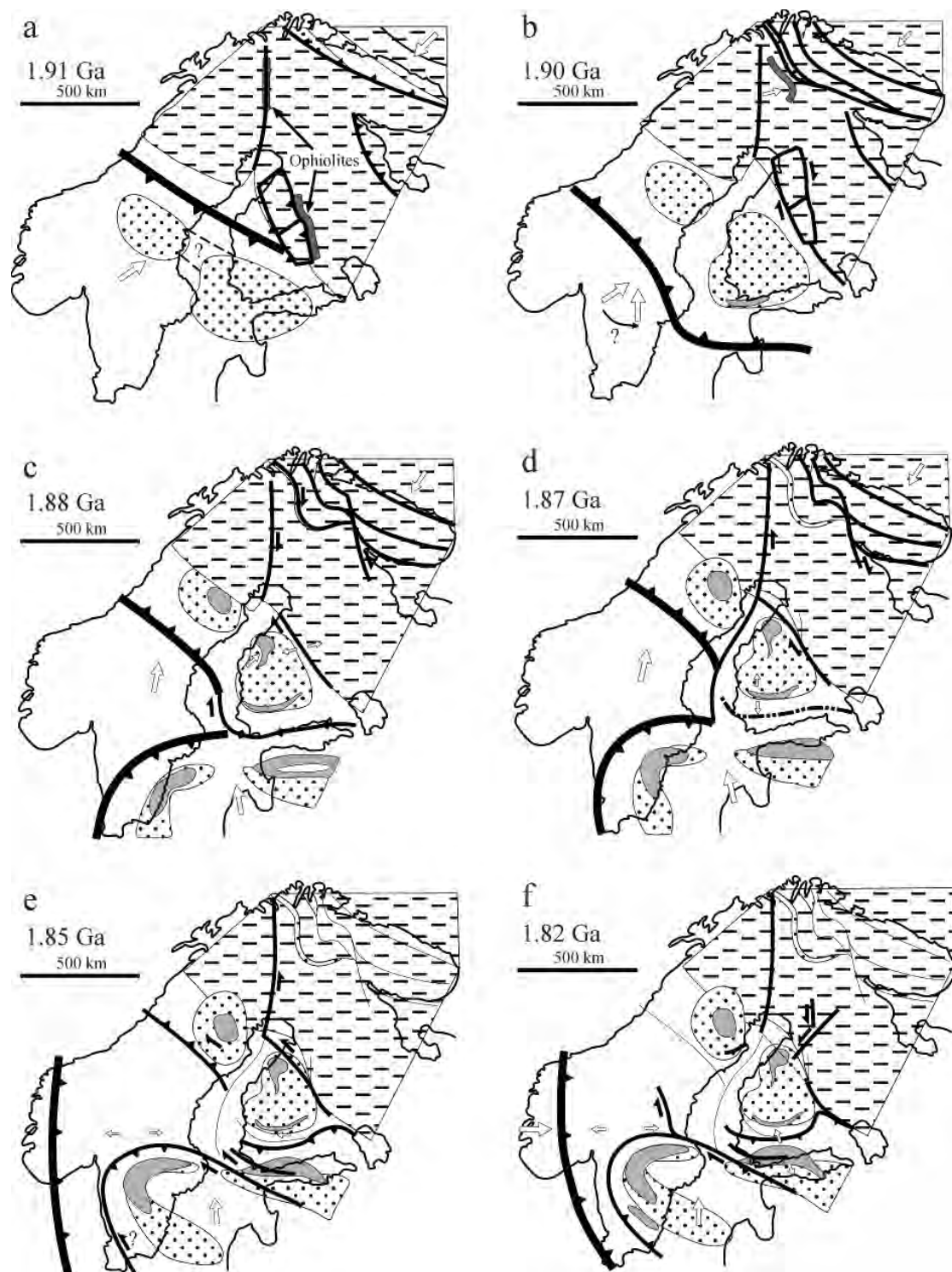


Figure 1.6: A model for the development of the Svecofennian Orogen between 1.91 and 1.82 Ga. The Proterozoic cover on the Archaean craton (broken lines) as well as syn- and post-collisional plutonism and continental volcanism are omitted for simplicity. The area without overprint is oceanic crust and sedimentary rocks, the dotted area is the arc complex, and the shaded area is volcanic arc. Subduction zones are marked by thick black lines (filled triangles on the subducting side), thrust/shear zones are marked by thinner lines (thick lines are active zones, thin lines are inactive). The big arrow shows the direction of relative plate motion, the small ones show shear/extensional senses of movement. (a) Initial accretion of an island arc complex to the continent. An ocean basin

was closing in the north. (b) Subduction reversal and rotation of plate motion direction, or convergence of another (micro)plate. (c) Collision of another arc complex against the accreted one in the east. Major dextral shear zones developed within the continent. (d) Subduction and sediment accumulation continued in the west while the crust was shortening in the east. A sinistral megashear extended from the arc system into the continent. (e) A transpressional zone developed in the east during continued convergence. Subduction from the west caused extension at the active margin and compression further east. (f) Transpression shifted westward and extensional collapse occurred in the east. After Nironen (1997).

The Bothnian Basin closed in the east 1.89 Ga ago when the southern arc complex collided with the northern one and the two accretionary prisms merged (Figure 1.6c). This is similar to the present Mollucca sea in Indonesia (e.g. McCaffrey and Abers, 1991). Subduction below the Skellefteå arc and deposition of turbiditic sediments in the western Bothnian Basin continued while the eastern part of the same basin was shortening.

Collision of the two arc complexes, and simultaneous continental collision in the north led to the development of dextral megashears in the north and to lithospheric thickening and peak metamorphism at 1.89 – 1.88 Ga in the east. A transtensional shear system developed at the continental margin (Lagoda – Bothnian Bay Zone). By 1.87 Ga the eastern part of the Svecofennian crust was largely stabilised and deformation was partitioned into ductile shear zones (Figure 1.6d). Continued convergence caused the transform fault to develop into a sinistral megashear that extended into the continent.

Subduction ceased beneath the Skellefteå arc 1.87 – 1.86 Ga ago. Curving of the southern arc complex continued until the complex doubled, and caused lithospheric thickening in this area (Figure 1.6e). Intracontinental transpression commenced approximately 1.86 Ga ago in the area of the merged accretionary prisms. Transpression resulted in partial thrusting of the southern accretionary prism onto the northern one and in the development of a shear zone system that spread from southern Finland to Sweden. Subduction below the continental margin in the west caused the development of a N-S rift at the margin and compression further east. Subduction in the west continued to 1.82 Ga ago when the transpressional area in the east was extending (Figure 1.6f).

The voluminous 1.89 – 1.88 Ga granitoids of the CFGC reflect a period of high crustal heat flow. Nironen (1997) describes a process whereby accretion from 1.91 Ga onwards resulted in thickening of the crust as well as the underlying mantle lithosphere. Despite subduction-induced heating in the mantle wedge between 1.90 and 1.89 Ga, the whole base of the lithosphere became cold, dense and gravitationally unstable with respect to the surrounding asthenosphere. Approximately 1.89 Ga ago, the base of the lithosphere delaminated by convective removal, and was replaced by hot asthenosphere (magmatic underplating). This remelted the lower crust and generated a variety of magmas by mixing and mingling of crustal and mantle magmas. High heat flow associated with crustal underplating and intraplate magmatism generated high-temperature and low-pressure metamorphism

that culminated coevally with the emplacement of magmas (Nironen, 1997; Lahtinen 1994, 1996, 2000; Lahtinen and Huhma, 1997).

The post-1.85 Ga granitoids were generated as the convergent tectonic regime was modified 1.86 Ga ago into transpressional ductile shear zones, in an E-W zone from southern Finland to central Sweden. Partial remelting of sedimentary rocks caused migmatisation and emplacement of anatectic granites 1.84–1.82 Ga ago within the transpressional zone as the result of extensional collapse (Nironen, 1997).

1.1.3 The Southwest Scandinavian Domain

The Southwest Scandinavian Domain (Figure 1.3), consists mainly of 1.75 – 1.50 Ga Gothian-aged rocks that were accreted to the Fennoscandian Shield during the Middle Proterozoic. The Southwest Scandinavian Domain is separated from the rest of the Fennoscandian Shield by the nearly 1600 km long Transscandinavian Granite – Porphyry Belt, and the “Protogine Zone”, a major belt of shearing and faulting accompanied by aligned intrusions of syenite, granite and mafic dykes (Gaál and Gorbatshev, 1987).

The Transscandinavian Belt is a 20 – 150 km wide zone of predominantly felsic volcanics and differentiated but mostly monzodioritic, monzonitic and granitic plutonic rocks extending along the western edge of the Svecofennian Domain in southern and central Scandinavia but cutting into the Archaean Domain in the north. Numerous fragments and enclaves of older rocks demonstrate intrusion into pre-existing continental crust (Gaál and Gorbatshev, 1987). The “Protogine Zone” constitutes a sharp boundary between the strongly foliated rocks of the Gothian Southwest Scandinavian Domain and the considerably better preserved rocks of the Svecofennian Domain.

The Southwest Scandinavian Domain itself exhibits a marked lithological zonation (Gaál and Gorbatshev, 1987). Its crust is more felsic in the east, with alkali – calcic and alkaline plutonic rocks. Further west, similarly aged Gothian plutonic rocks are more calcic/calc-alkaline in character. The supracrustal rocks also display a spatial variation, being shallow-water or continental in the east and marine in the west. Gorbatshev (1980) has interpreted the lithology as representing island-arc type volcanic belts with intervening basinal facies. The Southwest Scandinavian Domain can be correlated with various Grenvillian and pre-Grenvillian terrains in North America, thus indicating that they form part of a major, intercontinental pattern of Proterozoic crustal accretion (Gorbatshev, 1980; Gaál and Gorbatshev, 1987).

After the formation of the Southwest Scandinavian Domain, the crust was reworked by the 1.5 – 1.4 Ga old Hallandian “orogeny”, the 1.25 – 0.9 Ga Sveconorwegian – Grenvillian orogeny, and by the 0.6 – 0.4 Caledonian orogeny. However, despite the substantial volumes of Sveconorwegian granites and Caledonian nappes, post-Gothian additions to the continental crust of the Fennoscandian Shield are relatively minor, and confined to the western margins in Sweden and Norway.

1.2 METALLOGENESIS OF THE FENNOSCANDIAN SHIELD

Metal mining in the Fennoscandian Shield is of great economic importance for northern Europe. By far the most important deposits of the region are of Early Proterozoic age (Gaál, 1990). The association of various types of ore deposits with different tectonic environments is well recognised. This association has permitted Gaál (1990) to outline the main metallogenic districts within the Fennoscandian, according to plate tectonic processes (Figure 1.7). The plate tectonic evolution of the Fennoscandian has been discussed in Section 1.1.2.3. A brief account of the associated metallogenesis, based on the paper by Gaál (1990) will be outlined below.

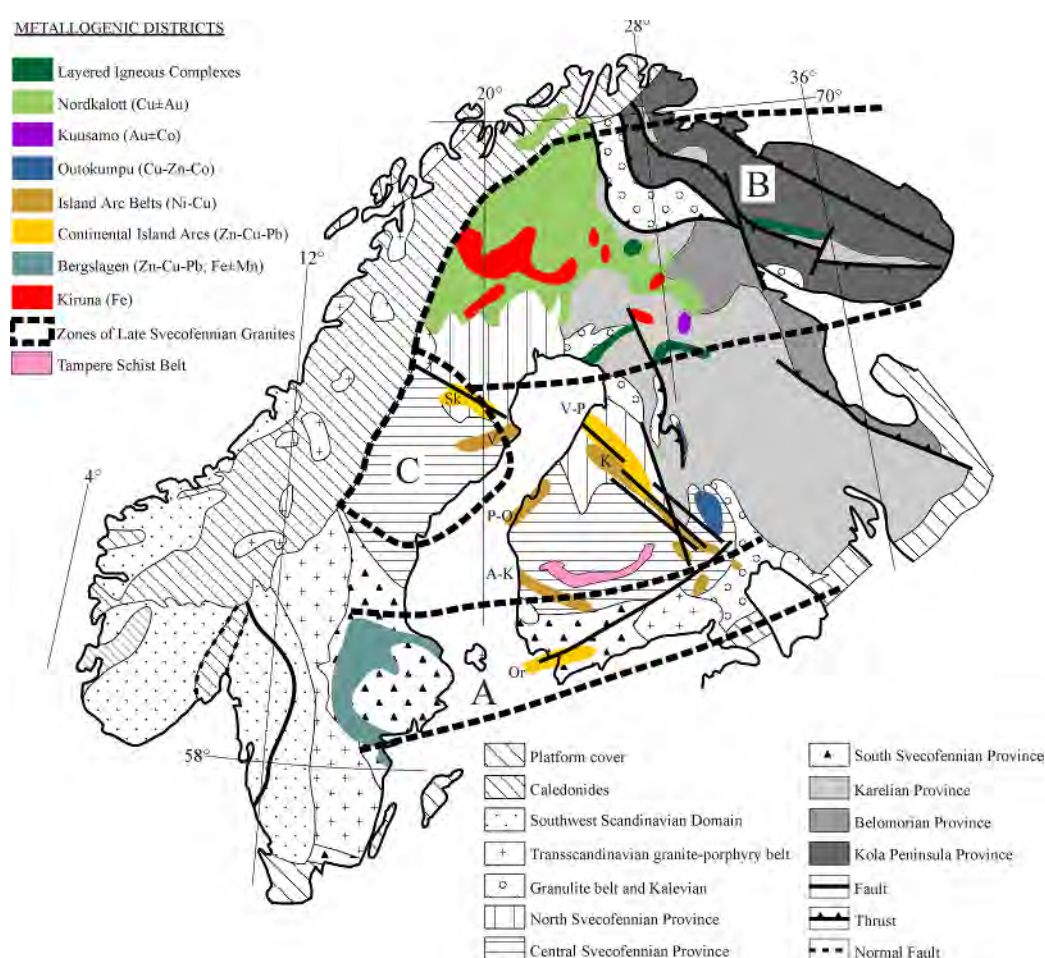


Figure 1.7: Metallogenic districts of the Fennoscandian Shield. Västerbotten (V); Kotalahti (K); Petolahti-Oravainen (P-O); Ahlainen-Kylmäkoski (A-K); Skellefteå (Sk); Vihanti-Pyhäsalmi (V-P); Orijärvi (Or). Zones of Late Svecofennian Granites: (A) Southern Zone; (B) Northern Zone; (C) Revsund granites. Modified after Gaál and Gorbatshev (1987), Gaál (1990), Weihed et al. (1992) and Nironen (1997).

1.2.1 Karelian Metallogenesis

With the consolidation of the Archaean crust 2.5 Ga ago, and the onset of the Karelian stage of crustal development, metallogenesis was largely connected with the abundant tholeiitic magmatism of this period (Gaál, 1990). The mafic-ultramafic layered igneous complexes of the roughly 300 km-long Kemi-Koillismaa belt in the northern part of the shield host a number of deposits, including: the chromite deposits at Kemi; PGE mineralisation at Penikat; Ni-Cu-PGE mineralisation and also Fe-V-Ti deposits in Koillismaa. Alapieti (1982) suggested that the individual layered magmatic bodies of this belt originally formed either a single large sheet, comparable in size to the Bushveld Complex, or several large lopoliths disrupted and fragmented by subsequent faulting (Gaál, 1990).

The tholeiitic volcanics of the numerous greenstone belts of the Lapponian Supergroup, are host to over twenty economically significant Cu±Au deposits (and numerous occurrences) of the “Nordkalott Tholeiitic Province” (Pharaoh et al., 1987). Some of the more significant mines include Aitik, Bidjovagge, Saattopora, Suurikuusikko and Pahtohavare which have gold tenors ranging from 0.3 to approximately 6 g/t (Gaál, 1990;). The most characteristic feature of these deposits is the hydrothermal alteration of the country rocks, which includes albitisation, sericitisation, carbonatisation, silicification, tourmalinisation and scapolitisation (Bjørlykke et al., 1987; Gaál, 1990; Lindblom et al., 1996; Frietsch et al., 1997; Eilu, 2003). The deposits themselves are typically hosted by altered metasediments, phyllites and mica schists, whereas quartzites, and tholeiitic rocks or their metamorphic equivalents, occur as wall rocks, invariably on the footwall side (Gaál, 1990). Ore genetic models have varied from syngenetic Besshi-type volcanogenic-exhalative (Gaál, 1990) to mesothermal (Bjørlykke et al., 1987; Lindblom et al., 1996; Eilu, 2003), to porphyry-Cu style for Aitik (Monro, 1988). Frietsch et al. (1997) noted many similarities between these deposits and the Fe-oxide Cu-Au deposits common in the Australian Proterozoic. Furthermore, Frietsch et al. (1997) recognised that magnetite is an essential component of the ores, although only as a minor component in many of the Fennoscandian deposits compared to the Australian deposits. The alteration associated with these deposits occurred mainly around 1.9 Ga at the peak of the main regional metamorphism and the intrusion of granitoids through to around 1.8 Ga (Frietsch et al., 1997). These deposits despite being hosted by Karelian rocks, were formed during the Svecofennian, and have been therefore been misclassified by Gaál (1990) as Karelian metallogenic events. Hydrothermally altered tholeiitic rocks in the Kuusamo district are also host to polymetallic Au-Co-Cu deposits.

1.2.2 Svecofennian Metallogenesis

During the pre-orogenic phase of extensional tectonism marked by the Kalevian sediments, thinning of the continental crust was accompanied by the opening of marginal basins and the formation of oceanic crust. Remnants of the 1.97 – 1.95 Ga old oceanic crust formed in these basins, have been preserved as ophiolite-like complexes at Outokumpu, Jormua and the Nuttio serpentinite belt (Section 1.1.2.1). The ophiolite-like complex in the Outokumpu district hosted eight massive Cu-Zn-Co sulphide deposits containing some 50×10^6 tons of ore with an average grade of 2.7% Cu, 1% Zn, 0.2% Co, 0.1% Ni, 26% Fe and 21% S (Gaál, 1990).

With the change to a compressional (and in places transpressional) tectonic regime, oceanic island arcs formed 1.93 – 1.86 Ga ago to the south-west of the Archaean craton. These oceanic island arcs are now represented by linear belts containing mafic and ultramafic rocks hosting economically important Ni-Cu deposits. These belts include: the Västerbotten Belt (containing seven deposits); the Kotalahti Belt (containing 34 deposits, prospects and showings); the Kylmäkoski Belt (containing two deposits and seven prospects); and the Oravainen Belt (containing four deposits) (Gaál, 1990).

Continental island arcs (1.9 – 1.87 Ga) also formed, generating a large variety of volcanic rocks of felsic, intermediate and mafic composition, and of calc-alkaline and tholeiitic affinity, together with epiclastic sediments and various plutonic rocks. The continental island arcs were accreted to the Archaean craton (Section 1.1.2.3) and are now represented by several individual volcanic belts, which from north to south, are: Kiruna, Skellefte, Vihanti-Pyhäsalmi, Haukivesi, Tampere, Hämeenlinna, Aijala-Orijärvi, and Bergslagen. These volcanic belts are the sites of the economically most significant mineralisations of the Fennoscandian shield, containing volcanogenic, massive Zn-Cu-Pb-Ag-Au sulphide deposits, apatite-bearing iron deposits, quartz-banded iron ores, and skarn iron ores with or without manganese (Gaál, 1990).

In the Kiruna district, a number of apatite-bearing iron ore deposits of the Kiruna-type are associated with Svecofennian porphyries. The major deposit groups have total iron ore reserves of 3.4×10^9 tons of ore grading 49-63% Fe (Grip, 1978; in Gaál, 1990). Sodic alteration of the lower part of the volcanic pile and potassic alteration of uppermost part, is a characteristic of these deposits suggesting a genetic link to Fe-oxide Cu-Au deposits (Williams, 1999b). There has been much controversy over the formation of the Kiruna-type deposits, leading essentially to two different models of either sedimentary exhalative or magmatic systems (e.g. Blake, 1999 and references therein).

The Skellefteå district hosts some 30 massive Zn-Cu-Pb-Ag-Au sulphide deposits, as well as other types of deposits including Ni mineralisation (in mafic to ultramafic sills and dykes), gold-rich porphyry-type deposits and minor W, U, Mo and Li mineralisations (Gaál, 1990; Weihed et al., 1992). The VMS deposits and gold-rich porphyries were formed ~1.89 Ga in an area resembling a modern island arc environment. Gold was emplaced 20-60 Ma later during peak metamorphism at the end of the Svecofennian orogeny. Finally, at ~1.80 Ga, after accretion of the island arc to the Archaean continent in the northeast, large granitoids intruded. Some of these have associated W, U, Mo and Li mineralisations, taken as evidence for crustal or cratonic influence (Weihed et al., 1992).

The NW-SE trending Vihanti-Pyhäsalmi district (hosting 19 Zn-Cu VMS deposits) and located along the Ladoga-Bothnian Bay tectonic zone is actually the continuation of the Skellefteå district on the eastern side of the Bothnian Bay. However, there are a number of differences between the two districts. Compared with the Skellefteå ores, the ores of the Vihanti-Pyhäsalmi district are rich in iron but poor in precious metals and Pb. The Au content, for instance, is one-tenth that of Skellefteå (Gaál, 1990). In

addition, the grade of metamorphism in the Skellefteå district increases from greenschist facies in the west to lower-amphibolite facies in the east. In the Vihanti-Pyhäsalmi district, the northern part was subjected to low to high amphibolite-facies metamorphism, whereas the southern part was metamorphosed 1.88 Ga ago under low-pressure high-temperature conditions of the granulite facies (Korsman et al., 1984; in Gaál, 1990).

The Orijärvi district in southern Finland hosted four small Zn-Cu-Pb-Ag sulphide deposits containing a total of 7.8×10^6 tons of ore, grading on average 2.1% Zn, 0.4% Cu, 0.9% Pb and 33 g/t Ag (Gaál, 1990). The deposits of the Tampere belt will be discussed in detail in Section 1.3.5.

The Bergslagen district in southern Sweden contains some thirty sulphide deposits that have been mined during the ca. 1000 years of mining in the district. The district is estimated to have contained 4.1×10^6 tons of Zn, 1.3×10^6 tons of Pb, 0.45×10^6 tons of Cu, 30 tons of Au and 3,450 tons of Ag (Grip, 1978; in Gaál, 1990). Compared with the Skellefteå and Vihanti-Pyhäsalmi districts, the Bergslagen district has relatively high values of Zn and Pb, which correlates with the higher abundance of carbonate rocks hosting the Zn-Pb-Cu-Au deposits of the Bergslagen district (Gaál, 1990). In addition to the massive sulphide deposits, the Bergslagen district also has numerous iron deposits, of which the economically most important are the Kiruna-type apatite-bearing iron ores, although banded iron formations (BIF's), manganiferous iron ores and skarn iron ores are also present (Gaál, 1990). The Bergslagen district is characterised by a thick pile of felsic volcanics, overlain by greywackes (with intercalations of mafic volcanics) and conglomerates. All of the Zn-Pb-Cu-Ag deposits are located in the uppermost part of the felsic volcanic units, close to the overlying sedimentary units. As seen in the Kiruna district, sodic alteration of the lower part of the volcanic pile and potassic alteration of the uppermost parts (Gaál, 1990) is a feature that perhaps links the Bergslagen to Fe-oxide Cu-Au deposits (Williams, 1999b).

Svecofennian granitoids (Section 1.1.2.2.2) are the predominant component of the Svecokarelian crust, so it is no surprise to see a number of deposits, prospects and showings related to granitoid intrusions. The early-orogenic Svecofennian granitoids (1.89 – 1.87 Ga) are typically I-type granitoids occurring either as part of the CFGC or as rounded batholiths enveloped by Svecofennian supracrustal rocks (Front and Nurmi, 1987). These I-type granitoids are typically associated with low-grade Cu-Au, Mo, Mo-Cu and Cu-W mineralisations of porphyry type. Of these, only the Paroinen deposit in the Tampere Schist Belt has been mined (Gaál, 1990) and will be discussed in detail in Section 1.3.5.2. The late-orogenic Svecofennian granitoids (1.84 – 1.75 Ga) on the other hand are typically S-type granitoids. Mineralisations associated with such granitoids contain Mo, W, U and elements enriched in complex pegmatites (Sn, Be, Li, Nb and Ta) (Gaál, 1990). The late-orogenic granites are found in three zones: the northern zone composed of the central Lapland granite complex, lying between Kuusamo and north of Skellefte; the southern zone extending from Lake Ladoga into the Bergslagen Province; and the Revsund granites intruded into the Svecofennian metaturbidites in the centre of the Bothnian Basin.

1.3 TAMPERE SCHIST BELT

The Svecofennian Domain in Finland is dominated by the extensive Central Finland Granitoid Complex surrounded by a number of volcanic-sedimentary belts: the Bothnia Belt in the west, the Tampere Schist Belt in the south and the Savo Belt in the east (Lahtinen and Huhma, 1997). The Haveri mine is located on the northern edge of the east-west trending Tampere Schist Belt close to the contact with the granitoids of the Central Finland Granitoid Complex (Figure 1.8).

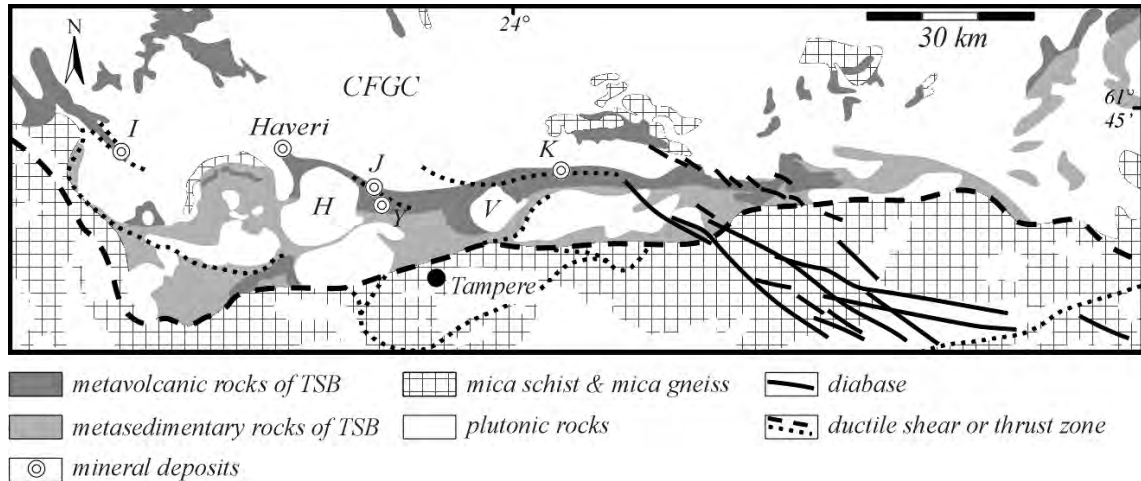


Figure 1.8: Simplified geological map of the Tampere Schist Belt (TSB). CFGC: Central Finland Granitoid Complex; H: Hämeenkyrö Batholith; V: Värmälä Stock. Also shown are the main mineral deposits within the TSB: I: Isovesi; J: Järvenpää; Y: Ylöjärvi; K: Kutemajärvi (modified after Poutiainen and Grönholm, 1996).

1.3.1 Stratigraphy and tectonic setting

The Tampere Schist Belt (TSB) lies about 50 km north of the suggested boundary between the southern and central Svecofennian provinces of Gaál and Gorbatshev (1987). It is a narrow, discontinuous, E-W-trending belt about 200 km long and about 20 km across at its widest. To the north the TSB is bordered by the 1.889 to 1.879 Ga old Central Finland Granitoid Complex (Huhma, 1986). To the south, the schists change abruptly to gneisses and migmatites with associated intrusive rocks (Kähkönen, 1989; Nironen, 1989).

In the conventional stratigraphic scheme of the TSB (Simonen, 1980), greywackes, siltstones and mudstones comprise the lowermost units. They are overlain by felsic schists, mafic and intermediate volcanics, conglomerates and related sedimentary rocks and, finally, by mafic volcanics (Figure 1.9). Revisions to this generalised scheme have been made by Mäkelä (1980) and Kähkönen (1989), which place the pillow basalt-bearing Haveri Formation below the greywackes-mudstones, and by recognising volcanic units within the lowest greywackes near Lake Näsijärvi.

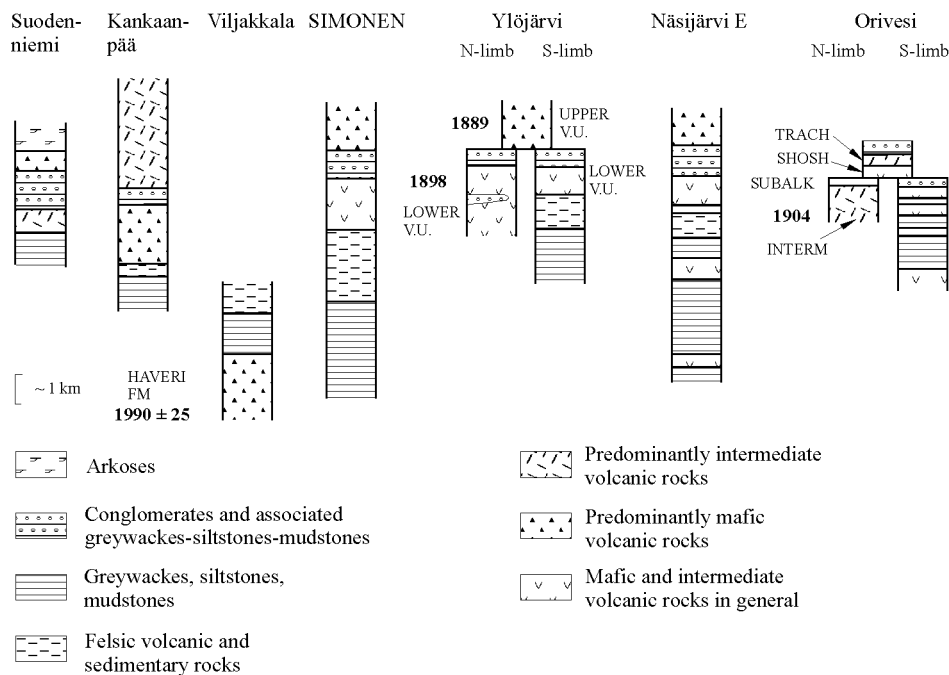


Figure 1.9: Stratigraphic sections of the Tampere Schist Belt according to Mäkelä (1980), Simonen (1980), Kähkönen (1989) and the references therein. The Kankaanpää area is 70-80 km west-northwest, and the Suodenniemi area 50 km west of Tampere. SIMONEN = generalised stratigraphy after Simonen (1980); INTERM = "Intermediate Unit" at Orivesi; SUBALK = Subalkaline Basaltic to Rhyolitic Unit at Orivesi; SHOSH = Shoshonitic Unit at Orivesi; TRACH = Trachytic Unit at Orivesi; LOWER V.U. = Lower Volcanic Unit at Ylöjärvi; UPPER V.U. = Upper Volcanic Unit at Ylöjärvi; HAVERI FM. = Haveri Formation. The figures 1904, 1898, and 1889 give U-Pb zircon ages in Ma (Kähkönen et al., 1989). 2-sigma error limits are 4-5 Ma. The Pb-Pb whole-rock age of the mafic volcanic rocks of the Haveri Formation is 1990 ± 25 Ma (Vaasjoki and Huhma, 1987).

The observed stratigraphy of the TSB is consistent with an arc-related geotectonic environment. Mäkelä (1980) suggested that the metabasalts in Haveri, approx 20 km NW of Ylöjärvi and belonging to the TSB, were generated at the initial stages of an island arc or back-arc, followed by calc-alkaline volcanism. According to Kähkönen (1987), the metavolcanic rocks in the TSB resemble those of mature island arcs or volcanic arcs at active continental margins. A volcanic-arc origin is also probable considering the conclusions of Ojakangas (1986) who assigns the metaturbidites of the TSB to a fore-arc basin environment on sedimentological criteria. The fore-arc would have been either within an island arc or within an active margin of somewhat older Proterozoic crust (Nironen, 1989).

1.3.2 Radiometric dating

Radiometric dating is consistent with the stratigraphic interpretation of the Tampere Schist Belt in Figure 1.7. U-Pb zircon ages from volcanic units have been provided by Kähkönen et al. (1989). The oldest age, 1904 ± 4 Ma, is from the “Intermediate Unit” at Orivesi, which is dominated by calc-alkaline, high-K dacites and andesites. This unit is the lowest in the northern limb of the syncline at Orivesi. At Ylöjärvi, a sample from the Lower Volcanic Unit has an age of 1898 ± 4 Ma, while a sample from the Upper Volcanic Unit has an age of 1889 ± 5 Ma. Lead isotopic results from the metabasalts of the Haveri Formation yield a Pb-Pb model age of 1990 ± 25 Ma (Vaasjoki and Huhma, 1987), supporting the stratigraphic scheme placing the Haveri Formation at the base of the Tampere Schist Belt. The considerable age difference between the Haveri Formation and the rest of the Tampere Schist Belt will be discussed later (Section 3.1.4) following a detailed appraisal of the Haveri Formation in terms of its geochemistry and interpreted geotectonic provenance.

The TSB is intruded by the Hämeenkyrö batholith and the Värmälä stock. The Hämeenkyrö “batholith” is actually a composite stock. The major units of the complex are granodiorite, a marginal tonalite, and a central monzogranite. All are hornblende- and biotite-bearing, metaluminous, I-type granitoids, exhibiting extensive evidence of magma mixing. Ilmenite is widespread as an accessory mineral, but the assemblage magnetite-sphene in many units implies high magmatic $f(\text{O}_2)$ conditions (Clark 1997). Nironen (1989) demonstrated the syntectonic nature of the pluton, which was emplaced at 1882 ± 6 Ma (Patchett and Kouvo, 1986) during regional metamorphism and the major D_1 deformational event in this segment of the TSB. Intrusion was initially passive, but was forceful in the later stages. The pluton became rigid towards the close of the D_1 deformation, and was affected by D_2 foliation at *ca.* 1860 Ma, during the waning stages of the thermal evolution of the Svecofennide volcano-plutonic arc (Clark, 1997).

The Värmälä stock is mainly composed of granodiorites and monzogranites, with I-type characteristics (Nurmi and Haapala, 1986). The Värmälä stock has a U-Pb zircon age of 1901 ± 28 Ma (Kähkönen et al., 1989). This relatively high age is not considered to be contradictory by Kähkönen et al. (1989), due to a combination of high analytical errors, and minor inherited zircon.

1.3.3 Metamorphic conditions

The rocks of the TSB recrystallised under low-pressure metamorphic conditions near the greenschist-amphibolite facies transition. Campbell (1978) estimated conditions between 500 °C, 1.5 kbar and 600 °C, 3 kbar for the peak of metamorphism in the Takamaa-Siivikkala area (to the east of the Hämeenkyrö Batholith, west of Lake Näsijärvi). On the basis of mineral parageneses in the Haveri Formation, Mäkelä (1980) has estimated metamorphic conditions to be approximately 550 °C and 2.5 kbar pressure.

1.3.4 Structure

Although Campbell (1978) recognised a structural sequence with five deformational phases in the Takamaa – Siivikkala area of the TSB, evidence of D_1 and D_3 deformational events are scarce. Therefore, the structural sequence as outlined by Nironen (1989) is more commonly taken to represent the structural evolution of the TSB.

The earliest structures recognised in the TSB include convolute bedding and load structures in metaturbidites and volcanoclastics. These features have been interpreted as syndepositional or early D_1 (Nironen, 1989). The regionally dominant tectonic fabric is formed by F_1 (equivalent to F_2 of Campbell, 1978). F_1 axial surfaces strike generally E-W and dip subvertically, while F_1 fold axes usually plunge gently east or west. F_1 folds are upright or slightly overturned and vary from open to isoclinal. The wavelength of F_1 folds is extremely variable ranging from less than 1 m to greater than 1 km. The style of folding varies from similar to parallel in more competent units. A penetrative S_1 schistosity is axial planar to F_1 folds, and in metaturbidites is expressed as a preferred orientation of micas, and as a preferred orientation of MS_1 hornblende and biotite in stratified volcanics. A steeply to vertically plunging mineral lineation lies within the plane of S_1 , and has been interpreted by Nironen (1989) as an L_1^1 stretching lineation. According to both Campbell (1978) and Nironen (1989), the peak of metamorphism occurred contemporaneously with D_1 , but outlasted the peak of deformation.

D_2 structures (equivalent to D_4 of Campbell, 1978) are found throughout the TSB, although they are well developed only in localized zones, and especially in micaceous rocks (Nironen, 1989). F_2 axial surfaces are subvertical and range from E-W to NE-SW in strike, while F_2 fold axes plunge steeply to the east. F_2 folds are open to tight and similar in style with hinge zones varying from round to angular. They are mostly asymmetric and both F_2 folds and S_2 have dextral vergence with respect to S_1 . The wavelength of the folds varies usually from a few centimetres to tens of centimetres but may reach several metres. S_2 is axial planar to F_2 folds and occurs both as microfolding and as discrete crenulation cleavage. In mica-rich lithologies a distinct L_2^1 intersection lineation is subparallel to L_1^1 . MS_2 - MP_2 biotite flakes are subparallel to S_2 , but MP_2 biotite, muscovite and chlorite flakes which have overgrown the F_2 microfolded fabric are randomly orientated (Nironen, 1989), indicating that the retrograde metamorphism continued beyond D_2 deformation.

D_3 structures (equivalent to D_5 of Campbell, 1978) are rarely observed, and the effect of D_3 deformation has been minor (Nironen, 1989). F_3 kink folds, which deform earlier tectonic fabrics in narrow zones, are usually open, although tight closures also occur. Their wavelength ranges from a few millimetres to some tens of centimetres. The fold axial planes, striking NW-SE to NNW-SSE and dipping steeply, have sinistral vergence with respect to S_1 , but NNW-SSE to N-S trending planes with dextral vergence have also been recorded. Small-scale extensional fracturing is associated with F_3 kinking, and the fillings are commonly K-feldspar or carbonate. Biotite is altered to chlorite, and MP_2 chlorite flakes are strained within D_3 deformed zones (Nironen, 1989).

1.3.5 Mineralisation

The TSB is reasonably well represented in terms of gold mineralisation, and has hosted a number of mines and deposits (Figure 1.8). At present only the Kutemajärvi (*aka.* Orivesi) gold mine is in active production, although historically there has been gold, copper and iron mining at Haveri and copper, tungsten ± gold mining at Paroinen (*aka.* Ylöjärvi). However, there are a number of other gold-bearing mineral showings within the TSB, including: Ahvenlammi, Järvenpää and Isovesi (Luukkonen, 1994; Clark, 1997).

1.3.5.1 Kutemajärvi (Orivesi)

The Kutemajärvi (Orivesi) gold deposit is owned by Outokumpu Oy.. Mining began in 1994, and the estimated reserves are approximately 0.4 Mt averaging 8-9 g/t Au. The Kutemajärvi deposit is in a sequence comprising felsic, intermediate and mafic metavolcanic (calc-alkaline) and volcanogenic metasedimentary rocks of the 1904 ± 4 Ma Koskuenjärvi Formation of the TSB. The host rocks are syn- to late metamorphic (ca. 1.89 Ga), strongly hydrothermally altered, intermediate metavolcanic rocks. The deposit is situated on the northern margin of the TSB. Less than 250 metres to the north of the deposit, is the Pukala Granitoid; a hypabyssal tonalitic to monzogranitic, porphyry intrusion, in contact with the alteration halo around the deposit. The surface extent of the Pukala Granitoid is 1-2 km x 15 km (Luukkonen, 1994; Poutiainen and Grönholm, 1996; Clark 1997; Eilu, 2003). Peak metamorphism occurred during an intense period of magmatism at 1.88 ± 0.01 Ga at 470-570°C and 3-4 kbar. This corresponds to a low-pressure domain transitional from greenschist to amphibolite facies. (Eilu, 2003).

The alteration at Kutemajärvi consists of an outer alteration zone characterised by sericitisation, chloritisation, silicification and pyritisation. Andalusite and phlogopite are common, and show post-tectonic crystallization textures. Also present are boudinaged and chloritised magnetite-bearing amphibole schist interlayers, representing either sills intruded during extensional stages of deformation (Grönholm, 1992, in Poutiainen and Grönholm, 1996) or altered metatuffites (Clark, 1997). The inner zone of alteration is characterised by sericitisation and silicification. It is dominated by sericite-quartz schist containing variable amounts of topaz, fluorite, andalusite, and occasionally pyrophyllite, kaolinite and sulphides (Poutiainen and Grönholm, 1996).

Gold is concentrated in at least five distinct vertical ore pipes, spatially associated with sericite-quartz schist and siliceous rocks (Eilu, 2003). In plan-view section the ore pipes are small lensoids (50 – 400 m²), extending to at least 820 metres below the surface (Poutiainen and Grönholm, 1996; Eilu, 2003). Enhanced gold grades are related to zones where the rocks are intensively deformed and silicified (i.e. sheared, boudinaged, folded, and quartz veined) during syn- to late-stage deformation (Poutiainen and Grönholm, 1996), suggesting remobilisation of the metals. The major part of the gold is carried as very

fine (generally < 60 µm) native gold, occurring mainly along quartz-grain boundaries and later fractures. Gold is locally present as Au-Te tellurides, also occurring at quartz-grain boundaries and in fracture fillings in quartz and tourmaline. Sulphides are common in the hydrothermally altered area. The outer alteration zone of chlorite-sericite schist, is characterised by pyrite dissemination. The inner alteration zone of sericite-quartz schist and quartz bodies contains very few sulphides. In the rocks of the ore pipes, tellurides are often more abundant than sulphides (mainly pyrrhotite, pyrite, chalcopyrite and sphalerite) which occur as disseminations, accumulations, small veinlets and vertically elongated massive pockets. Minor amounts of sulphosalts are associated with the sulphides (Poutiainen and Grönholm, 1996).

Five main mineralisation stages have been recognised on the basis of the ore mineral paragenesis at Kutemajärvi (Grönholm, 1992; Luukkonen, 1994; Poutiainen and Grönholm, 1996). The first stage was the initial host-rock alteration (sericite-quartz alteration) and sulphide mineralisation (cassiterite, ilmenite, magnetite and pyrite). This was followed by the second, main sulphide stage, with sulphides, gold mineralisation and silicification. Sulphosalts are also present in this phase. The third stage is characterised by gold tellurides, sulphides and silicification. The final hypogene stage is a “residual” stage with remobilization of metals and sulphides, including gold, copper, lead, antimony and arsenopyrite. The fifth and final stage is characterised by supergene mineralisation.

Fluid-inclusion studies (Poutiainen and Grönholm, 1996) indicate that there was an overall fluid evolution at Kutemajärvi. The initial H₂O-CO₂ fluids associated with the first, oxide-dominated alteration system, are interpreted as being generated during regional metamorphism by devolatilisation of the lower crust, at deeper crustal levels than are currently exposed. Fluid generation was greatest in the vicinity of crustal shear zones, where enhanced heat flow and granitoids emplacement created extensive thermal anomalies. The P-T conditions for these fluids (and hence peak metamorphism) are interpreted to have been in the range 470 – 570 °C, at 3 – 4 kbar pressure.

Fluids expelled from the volcano-sedimentary and granitic rocks (fluorine and boron enrichment) migrated upward into favourable dilatant structures (shear zones) which continuously acted as conduits for the fluids (fluid mixing) during Svecofennian deformation. As the fluids ascended along the shear zones to lower P-T conditions, they evolved (CH₄ enrichment), and reacted with wall rocks, becoming enriched in sulphides ± gold. The main sulphide stage (sulphide ± gold + silicification) took place in a P-T range of 380 – 320 °C, at 2.8 – 2.0 kbar pressure (compatible with the greenschist to lower amphibolite facies transition), in the presence of low salinity H₂O-CO₂-CH₄ fluid (Poutiainen and Grönholm, 1996).

The telluride stage took place in the P-T range of 330 – 270 °C, at 1.6 – 0.7 kbar pressure (at or below the brittle-ductile transition), in the presence of low-salinity CO₂-CH₄ fluids. Phase separation (effervescence) due to local pressure fluctuations (hydraulic fracturing and uplift of the area) may have caused the tellurium enrichment, by fractionation into a carbonic vapour phase.

Secondary enrichment and remobilization of earlier gold and some base metals probably occurred in relation to a later thermal event (i.e. potassium granite migmatization ca. 1.83 Ga) at P-T conditions of < 380 °C at 1kbar. This was associated with brittle-ductile deformation, and interaction with deeply circulating meteoric fluids, causing cooling and dilution of ore fluids. The deposit was also affected by late CO₂-CH₄ ± N₂ fluids, which could have been involved in the earlier mineralisation event, or which could have caused late epithermal telluride deposition (Poutiainen and Grönholm, 1996).

Eilu et al., (2003) have suggested that the alteration zonation and geochemical features at Kutemajärvi are best explained by premetamorphic, high-sulphidation, epithermal-style alteration, with metamorphic recrystallisation of the altered rocks. In such a model the original alteration zones (from proximal to distal) of a quartz rich core, through an advanced argillic zone into a distal propylitic zone, are after metamorphism at lower-amphibolite facies, now represented by quartz, quartz-pyrophyllite-andalusite, and sericite-quartz-chlorite respectively.

In Kutemajärvi, the close spatial relationship between the gold deposit and plutonic-metamorphic rocks, as well as the F-rich minerals suggests a genetic link with local magmatic and metamorphic-deformational processes that led to mineralisation associated with strong hydrothermal alteration in a continental island-arc or in an active continental margin environment (Poutiainen and Grönholm, 1996). The preservation potential of an epithermal deposit in such a terrane would be very low, unless associated with rapid subsidence, such as during arc extension (Eilu et al., 2003).

1.3.5.2 Paroinen (Ylöjärvi)

The Paroinen (Ylöjärvi) Cu-W deposit was mined in the period 1942 – 1966, and closed due to the exhaustion of reserves. During the life of the mine some 128,343 t of Cu, 49.5 t of Ag and only 270 kg of Au was removed from the deposit. The main Cu-W ore phase is related to porphyry copper-type tourmaline breccias (Clark, 1997).

The Paroinen deposit occurs within the largest of a swarm of tourmaline-cemented breccia bodies hosted by predominantly metabasaltic volcanic rocks within 1 km of the eastern contact of the Hämeenkyrö batholith. The eastern, 1 km wide, marginal zone of the pluton is affected by widespread hydrothermal alteration (Gaál et al., 1981). Tourmalinisation is the most extensive expression of alteration, but a ca. 3.5 km² area 2 – 3 km SSW of Paroinen, exhibits locally intense scapolitisation and carbonatisation (Clark, 1997). The eastern contact zone of the intrusion is also a nucleus for the emplacement of a complex of porphyritic dykes, largely of granodioritic (dacitic) and monzogranitic composition, which Clark (1997) suggests are identical to plagioclase porphyries in the immediate vicinity of the Paroinen deposit, and thus are suggestive of a broad genetic relationship between the Hämeenkyrö pluton and the Paroinen deposit.

Clark (1964, 1997) has identified three main stages of mineralisation $\pm$ alteration at Paroinen. The first stage (Stage I) of alteration-mineralisation was initiated by intense Fe (-Ca) metasomatism of mafic and intermediate rocks in the vicinity of the evolving D₂ shear-zone, generating wide, skarn-like, zones rich in magnetite and ferrotschermakite, and less abundant bytownite, apatite, Fe-Mn garnet, scapolite and several Sn minerals. Sulphides (and in particular chalcopyrite, scheelite and arsenopyrite) are noticeably absent from this mineral assemblage. Radical depletion in Na is reflected in the extremely calcic compositions of the amphibole and plagioclase. On the basis of amphibole-plagioclase thermometry, fluid inclusion studies and mineral compositions, Clark (1997) suggests that this intense hydrothermal stage took place at temperatures of at least 600 °C, and involved saline fluids (20 – 27 equiv. wt.% NaCl from secondary inclusions in apatite). The elevated temperatures, and the salinities have been used by Clark (1997) to infer that source of the heat and the fluids are of magmatic origin (rather than metamorphic) and that the most likely correlation is with the porphyry dyke complexes. The Stage I assemblages are sulphide-poor but enriched in both Sn and Au.

The major locus of the magnetite-rich assemblages is in the strongly sheared metabasaltic tuffs, where the alteration forms a series of crudely tabular zones, broadly paralleling the trend of the shear zone, with an aggregate width of at least 25 m, paralleling the SE margin of the breccia complex (Stage II), and separated by septa of less altered metabasalt (Clark, 1997). The alteration is developed over the entire vertical interval of the mine, and is intersected in several drill-holes at depths greater than the lower extremity of the main breccia bodies at 380 m (Himmi et al., 1979). Clark (1997) also describes Fe-rich biotite and chlorite intimately associated with the magnetite-hornblende assemblages, forming monomineralic septa. Although the paragenetic sequence is “not entirely clear”, biotite locally pseudomorphs amphibole, and Fe-rich chlorite forms felted masses replacing coarse biotite flakes. Clark (1997) has thus interpreted an overall sequence from early amphibole, through biotite to chlorite, and thus a model in which early Fe-Ca metasomatism was followed by strong K-metasomatism, with later retrograde replacement of biotite by chlorite.

The development of Stage I alteration-mineralisation overlapped temporally with the initiation of true brecciation and the early part of Stage II development, and the main phase of Cu, W and As mineralisation. Stage II mineralisation occurs in two main forms, as the dominant tourmaline-cemented breccias and as emplacements into sheeted D₂ shear fractures (hosting about 40% of the mineralisation) in metatuffites (Clark, 1997).

Stage II mineralisation commenced with early scheelite crystallization, in association with arsenopyrite and minor cassiterite, magnetite, ilmenite and rutile, prior to the emplacement of the tourmaline cement. Au-rich electrum is also associated with this early part of Stage II mineralisation, as is native gold which has been observed in close association with arsenopyrite. The main part of Stage II mineralisation is the chalcopyrite ($\pm$ apatite-pyrrhotite-sphalerite) assemblage associated with the emplacement of the tourmaline cement. This main Cu ore-forming sub-stage is Au-poor. The final

Stage II assemblages are similarly Au-poor but are enriched in Ag, occurring as the native metal, acanthite, sternbergite, hessite and several sulphosalt minerals (Clark, 1997).

Intense silicification (bleaching) has been observed on the margins of breccia fragments, and is associated with sporadic fine-grained muscovite (Himmi et al., 1979; Clark 1997). Clark (1997) also states that muscovite is a widespread (although megascopically inconspicuous) constituent of the tourmaline-sulphide matrix. Therefore Stage II mineralisation can be characterised as being associated with quartz >> sericite (muscovite) alteration, indicative of strong K and Al leaching, and thus that the hydrothermal fluid was markedly acidic (Clark, 1997).

Stage II commenced at temperatures in excess of 489 °C, when non-boiling, moderately saline (≤ 18 equiv. wt.% NaCl) fluids penetrated the shear zone, depositing considerable scheelite and arsenopyrite. Subsequent deposition of the main Cu ore-forming sub-stage may have taken place from boiling, moderately-saline (*ca.* 12 % equiv. wt.% NaCl) fluids at *ca.* 455 °C and at a pressure of 400 – 500 bars (from fluid inclusions in apatite). Independent pressure estimates of co-existing sphalerite and hexagonal pyrrhotite suggest pressures of 1250 ± 250 bars (Clark, 1997). The final Ag-rich and sulphosalt-bearing phases of Stage II are believed to have formed in the terminal stages of breccia consolidation, as temperatures decreased to below 145 °C (Clark, 1997).

The terminal stage (Stage III) in the evolution of the Parainen hydrothermal system is volumetrically minor. Stage III mineralisation is characterised (Clark, 1997) by discontinuous, steeply-dipping, carbonate-bearing veins, which cross-cut both the tourmaline breccias and the sulphosalt-rich later stages of the breccia cement. The carbonate-quartz veinlets are dominated by ankerite and calcite, and contain galena, sphalerite, minor chalcopyrite and pyrite. Joints filled with gypsum, pyrrhotite, and pyrite are also present. These late-stage veins also include very low-sulphidation sulphide associations, and are Au-bearing. The dominant sulphide minerals are troilite, Fe-rich hexagonal pyrrhotite and mackinawite. Pyrite and marcasite are less abundant and do not co-exist with the troilite or mackinawite. Tennantite and Fe-poor sphalerite occur in direct association with pyrite. Au-rich electrum has been observed in contact with troilite and hexagonal pyrrhotite (Clark, 1997).

Another feature described by Clark (1997) and apparently contemporaneous with the carbonate-bearing veins, is the presence of widespread, sinuous shear planes, termed *harniska pinta* by the miners, which contain strongly sheared accumulations of chlorite, quartz, ankerite and calcite. Despite containing no sulphides, the *harniska pinta*, contain significant Au, and are interpreted as recording transcurrent reactivation of the S₂ structures.

In terms of an ore genetic model for Parainen, Clark (1997) has appraised the data in comparison to porphyry copper deposits and allied magmatic-hydrothermal deposits, and “hypozonal” mesothermal lode-gold deposits. Despite many similarities to the “hypozonal” clan of mesothermal Au lodes, a magmatic genesis for Parainen is supported by *P-T-X* conditions (the high-temperature, saline and

CO₂-poor fluid inclusions) in the earlier stages of mineralisation. Another fact in support of a porphyry (or allied magmatic-hydrothermal) model is the widespread occurrence of magnetite-rich alteration-mineralisation (Fe-metasomatism), which is recognised globally as an important constituent phase in magmatic (e.g. porphyry copper) and magmatic-hydrothermal (e.g. Fe-oxide Cu-Au deposits), whereas it is generally undocumented in mesothermal deposits. Finally, in an analysis of porphyry copper deposits, Clark (1997) reveals that Au is widely enriched in early (“A” or “M”) or late (“D”) vein systems, but only very rarely in the transitional, “B”, veins and associated tourmaline breccias. Therefore, despite failing to unequivocally identify the actual magmatic source rocks, Clark (1997) clearly favours a proximal magmatic-hydrothermal model for the generation of the Paroinen deposit, with the hydrothermal activity being tentatively assigned to the post-D₁ porphyry dykes.

1.3.5.3 Ahvenlammi

The Ahvenlammi tungsten-polymetallic prospect is located on the southern fringe of the TSB where the schists have been intruded by the Värmälä granodiorite stock in the NW and by the Siitama granite batholith in the SW. The TSB is bound on the south by banded mica gneiss complex (Luukkonen, 1994). The dominant rocks in the Ahvenlammi area are metaurbidites that have mafic, intermediate and felsic metavolcanic interlayers. Mineralisation is hosted largely by metagreywackes ca. 500 m from the contact with the Siitama granite. Scheelite is concentrated in quartz veins and disseminations in a vertical pipe (about 100 m wide) extending down to at least 125 m (Luukkonen, 1994; Clark 1997).

The paragenetic sequence of mineralisation (Luukkonen, 1994; Clark 1997) commenced with the deposition of scheelite, calcite and arsenopyrite, and the formation of quartz veins. This was followed by sulphide base and minor precious metal precipitations: the Cu-Sn-Zn stage, with chalcopyrite, rutile, ilmenite, sphalerite, scheelite (minor), pyrite, marcasite, arsenopyrite, galena, stibnite and proustite; and the Bi-Ag stage, with pyrite, marcasite, pyrrhotite, herzenbergite, bismuth and electrum. Mineralising events have been followed by secondary alteration processes under supergene conditions.

Hydrothermal activity was channeled by ductile shearing associated with a major shear zone defining the interface between the metagreywackes and higher-grade micaceous gneisses. Late-stage brittle fractures host quartz veins, but these are apparently barren. Quartz veining may have begun during D₁ shearing, but metal emplacement is inferred to have occurred during the development of D₂ fracture cleavage and continued shearing (Luukkonen, 1994; Clark 1997).

The abundance of scheelite and arsenopyrite at Ahvenlammi is strongly suggestive of parallels with Paroinen, as is their predominance in the earliest stage of ore deposition. Sn is also enriched in both deposits, although at Ahvenlammi it occurs with the main Cu-Zn sulphide and late-stage, Bi-rich, sub-stages. Electrum deposition apparently took place only in the latest hypogene stages, and was not

associated with scheelite or arsenopyrite as at Paroinen. Tourmaline has not been observed at Ahvenlammi, and there is no record of hydrothermal alteration effects (Clark, 1997).

1.3.5.4 Järvenpää

The Järvenpää Au-Zn-Sb prospect is located 4 km north of the Paroinen deposit, and is situated on the northern boundary of the TSB close to the contact with the CFGC, and 1.7 km east of the Hämeenkyrö batholith. The mineralisation is hosted by sericite-quartz schists (as at Kutemajärvi), forming a lens-shaped body enclosed largely by calc-alkaline intermediate to mafic metavolcanic rocks of pyroclastic origin. Outcrops are poor, but Luukkonen (1994) interprets the host-rocks as representing variably sericitised, silicified and pyritised zones, with probable intermediate protoliths.

Luukkonen (1994) proposes the following sequence of ore mineral formation for Järvenpää: The first stage was metasomatic silicification of the intermediate volcanic rocks, and the deposition of disseminated pyrite, rutile, pyrrhotite, chalcopyrite and arsenopyrite, possibly due to boiling of fluids related to a decrease in pressure. The second stage was the formation of compact lenses and pockets of semi-massive sulphide (pyrite, chalcopyrite, pyrrhotite and sphalerite) and the formation of quartz veins. This was followed by a period of fracturing and remobilisation of quartz. The main stage of ore formation, and the deposition of Sb, Pb, Au and Ag minerals, has been subdivided into an earlier phase with silver-rich electrum, and a later phase with Au-rich electrum. The Ag-rich phase contains ullmannite, chalcopyrite, electrum, tetrahedrite and pyrite. The Au-rich phase includes tetrahedrite, bournonite, jamesonite, boulangerite, galena, pyrrhotite, arsenopyrite, gudmundite, hessite, altaite, antimony, electrum and aurostibite. Luukkonen (1994) suggests that the sericitisation and pyritisation occurred after the culmination of D₁ and metamorphism, while the main mineralizing event occurred either syn- or post-D₂ deformation.

The Järvenpää assemblages appear to correspond closely to those of the main Paroinen Stage II. There is, however, no evidence of the occurrence of associations corresponding to stages I and III of the Paroinen deposit. In addition the late sulphosalt assemblage is apparently relatively more abundant at Järvenpää, and is enriched in both Te and Au relative to the corresponding sub-stage at Paroinen. One of the major differences between Järvenpää and Paroinen, is the absence of tourmaline and associated breccias at Järvenpää (Clark, 1997). There are in many respects more similarities between Järvenpää and Kutemajärvi, including early silicification (and the deposit hosted by quartz-sericite schists), the presence of tellurides (and later sulphosalts), the absence of tourmaline, and perhaps not least of all, a geological setting close to the CFGC. Noticeably absent from the described assemblages at the Järvenpää prospect is any evidence of (early) hydrothermal magnetite.

1.3.5.5 Isovesi

The Isovesi Au prospect is situated at the SE end of the Kankaanpää schist belt, surrounded by granitoid rocks of the CFGC. The Kankaanpää schist belt can be considered as a western extension of the TSB. The host rocks of the Isovesi mineralisation are variably altered intermediate metavolcanic rocks (metatuffs). Plagioclase and uraltite porphyries are also locally present. The prospect lies in the immediate envelope of a moderately-foliated granite pluton, and is contiguous with a major, regional shear-zone.

Hydrothermal alteration has been intensive in the intermediate metatuff, especially in the vicinity of shear zones, and the metatuff is rich in epidote, clinozoisite, hedenbergite and carbonate. The alteration of amphibole leads to the formation of biotite and chlorite. The “skarn” alteration is also reportedly developed in the uraltite porphyry. Retrograde sericitisation of plagioclase is also a typical feature of the intermediate metatuffs (Luukkonen, 1994). Skarn-type alteration involved enrichment in Ca and Mg, retention of Fe, and depletion of Si, Na and K (Clark, 1997).

Luukkonen (1994) defines the following sequence of hydrothermal episodes: firstly, upward percolating fluids in the shear zones, caused intensive fracturing and skarn reactions, and the precipitation of pyrrhotite, arsenopyrite and löllingite. This was followed by weak silicification and the emplacement in fractures of pyrrhotite, pyrite, chalcopyrite, arsenopyrite, sphalerite and sphene associations. Finally, there was a late Au-Bi mineralisation stage, comprising pyrite, pyrrhotite, arsenopyrite, sphalerite, electrum, gold, pilsenite (a Bi_2Te mineral), galena and hematite (Luukkonen, 1994; Clark, 1997).

Arsenopyrite and pyrrhotite were apparently deposited throughout the hypogene evolution, and Au is strongly correlated with late Bi and Sb-rich sulphosalt-native metal associations, and with Te minerals, rather than with early formed arsenopyrite. The maximum temperature of mineralisation is defined as less than 440 °C by the equilibrium assemblage of arsenopyrite – löllingite (Luukkonen, 1994). Noticeably absent from the assemblages are magnetite and andradite, and this combined with the presence of hedenbergite, has led Luukkonen (1994) to infer low oxygen activities during the skarn reactions. Although there are some differences in assemblages, Clark (1997) has drawn analogies between Isovesi and Paroinen, with the early Ca-enrichment associated with the skarn reactions being analogous to the Paroinen Stage I, whereas the main phases of mineralisation is analogous to the Paroinen Stage II mineralisation (the main Cu (-W, As) ore forming episode).

1.4 HISTORY OF THE HAVERI MINE

Mining for iron was started at Haveri as early as 1730 and continued periodically until the end of the 1800's. Attempts to make the operations viable were stymied by the presence of sulphides in the magnetite iron ores. In 1935, investigations into Haveri's gold and copper potential commenced, prompted by reports on the presence of gold and copper in the old iron pits. Soon, sufficient indications had been gathered to warrant test mining, and in 1939, mining began with the onset of the Winter War with Russia (Lupander and Räisänen, 1954). The mine operated fairly inefficiently until 1949, when fire damage allowed the construction of a more efficient mill, complete with a cyanide circuit for recovery of gold. This increased production from 50,000 tons to 100,000 tons / year (Lupander and Räisänen, 1954).

The ore was hoisted from an open pit (about 100 m long and 65 m deep) and from underground workings developed on the 50 and 96 m levels. The underground workings were serviced by a small shaft with a depth of 120 m. Around 1953, a 3-compartment shaft, with a concrete collar, was sunk to a depth of about 120 m, north-west of the original shaft. This new shaft, however, was never used. In the latter stages of the mine's life, a decline was sunk at the base of the old shaft and, a second shaft was sunk directly below the original to a depth of 230m, leaving about 10 m of rock between the two shafts (Lupander and Räisänen, 1954; Karvinen, 1996; Strauss, 1998).

By the mid-1950's mining appears to have overtaken exploration efforts to find and develop more ore (there was no geologist employed). In the light of poor gold prices, diverted interests into iron elsewhere in Finland, and changes in majority shareholders in the company, a decision was made in 1959 to close the mine. The last mining engineer at the mine, convinced of the potential, mounted a last valiant effort to find more ore. It was in this period that the decline was sunk and the shaft deepened. This was accompanied by a significant program of diamond drilling both from surface and underground. This final phase of activity at Haveri located: (a) the SW copper zone, supposedly rich in copper and gold, which was partially (?) developed; and (b) an area at Peltosaari, about 700 m east of the mine, which was never developed. However, in 1962 the mine was finally closed (Karvinen, 1996; Strauss, 1998).

In total, 1.5 million tons of ore, grading an average of approximately 2.8 g/t gold and 0.37% copper, is the official production from the Haveri Mine during the period 1942 to 1962 (Isokangas, 1978). All reports on the mine state that the gold was "invisible". However, several spectacular samples of native gold from the Haveri Mine can be found at museums around the country, as well as in the collections of some of the old local miners (Strauss, 1998).

The release of low-level airborne geophysical data by the government attracted Outokumpu to carry out base metal exploration in the area in 1976 – 1977: During this time geological mapping, magnetic

surveys, EM measurements using Slingram, gravity measurements, surface sampling and diamond drilling were carried out. This work was mostly done to the south of the mine at Ansomäki. A number of EM anomalies were delineated and appear to have guided some, if not all, of the 12 drill holes known which totalled approximately 1670 m. (A 13th hole was drilled, but the location is unknown). At Peltosaari, only surface sampling of a sulphide zone with a hand-held drill and a Slingram EM survey were carried out (Strauss, 1998).

From 1980 to 1983, Outokumpu did auger sampling of the tailings, which is estimated to be about 1 million tons. A total of 1201 samples at 1 m spacing, from 165 holes were taken and analysed for a number of elements including gold and copper.

From 1996 until 2000, extensive gold exploration was carried out in the Haveri region by Glenmore Highlands Inc. During this period, exploration activities included: clearing overburden, trenching, numerous geophysical surveys, RC drilling (1047 metres), diamond drilling (7952 metres), and finally, dewatering and partial rehabilitation of the old mine to the 100 metre level. In addition, during the course of the exploration, considerable archival material (including borehole assays for over 18500 metres of drilling) was uncovered from the period of mining (Strauss, 1998).

1.5 PREVIOUS WORK IN THE HAVERI AREA

A number of studies have been completed at Haveri. With the added benefit of observations undertaken during mining at Haveri, Stigzelius (1944) produced a thorough and complete description of the mine, and its geology. This study described the mineral assemblages and petrology of the rocks in the Haveri area, as well as the styles and distribution of Au-Cu mineralisation in the mine. From this, Stigzelius (1944) developed an epigenetic theory for the ore genesis, related to the nearby granitoids, although at a later stage he apparently favoured a syngenetic model (Stigzelius, 1976). Paarma (1947) studied the sulphide mineral assemblage at Haveri, and recognised a pyrrhotite-dominated assemblage, rich in magnetite and chalcopyrite. In certain areas of the old mine, Paarma (1947) identified considerable quantities of cobaltite and glaucodot with the massive sulphide. Pyrite was variably present in the mineral assemblage, and was noticeably absent from cobalt-rich massive sulphide. Minor pre-ore pyrite was observed (fractured by pyrrhotite) but was principally a later retrograde product associated with chlorite. In the copper-rich portions of the mine, cubanite lamellae were identified in the chalcopyrite, as well as trace amounts of valleriite. Other minor minerals include gersdorffite, arsenopyrite, molybdenite, scheelite, and sphalerite. Gold was associated with sulphides (especially cobalt-bearing phases) in the massive sulphide, and also occurred as free gold. Paarma (1947) also reports the occurrences of tellurides (bismuth telluride and hessite) again associated with the cobalt-rich phases in the massive sulphide, but mentions that there was no evidence of gold tellurides at Haveri. Paarma

(1947) suggested an epigenetic origin for the ore. A summary of the Haveri deposit and the mining methods was provided by Lupander and Räisänen (1954).

Anttonen (1968) studied the paragenesis of the chalcopyrite at Haveri, and made comparisons to the chalcopyrite occurrences at Outokumpu and Orijärvi. On the basis of these studies Anttonen (1968) concluded a pneumatolytic origin for the mineralisation at Haveri. Ollila (1977) described the rocks in the Haveri region as part of an MSc thesis, and largely on the basis of the ore mineral descriptions of Paarma (1947) concluded that the ore was pneumatolytic in origin, but was unable to determine whether the cause was the granodiorites or the porphyries. Mäkelä (1980) completed a study on the Haveri mine area, as a result of Outokumpu's base metal exploration, and concluded that the ore was syngenetic in origin, and could be compared to Cyprus-style massive sulphide mineralisation. Kähkönen et al. (1981), on the basis of the REE geochemistry of the basalts concluded that they were formed in a marginal basin environment, and supported the syngenetic argument for ore genesis. A lead isotopic study of the Haveri Formation, led Vaasjoki and Huhma (1987) to conclude an oceanic mantle origin for the volcanics, and an implied temporal relationship with the ophiolite complexes at Outokumpu and Jormua. Detailed geochemical and structural studies of the Haveri area were completed by Kähkönen and Nironen (1994) and Nironen (1994). These studies did not dispute the syngenetic origin for the Au-Cu mineralisation, although Nironen (1994) suggested that the ore was mobilised during both D₁ and D₂ deformation.

2 GEOLOGY OF THE HAVERI AREA

2.1 STRATIGRAPHY

The rocks in the study area comprise a supracrustal succession that has been intruded by several intrusive rock types. The supracrustal succession has been divided into three Formations (Figure 2.1): the basal Haveri Formation (Kähkönen and Nironen, 1994; Mäkelä, 1980) is composed predominantly of metamorphosed mafic volcanics, with pillow structures and flow breccias. This passes upwards into an intercalated zone of layered metavolcanic rocks and metasedimentary rocks, interpreted as metamorphosed mafic tuffs and tuffites. There is a continuous gradation upwards from the metalavas and the metatuffites of the Haveri Formation, being dominated by volcanoclastic material, into metasedimentary lithologies of an epiclastic nature. The metasedimentary rock unit, consisting of metagreywackes, has been termed the Osara Formation (Kähkönen and Nironen, 1994). To the south of the Osara Formation (outside of the study area) occur various metavolcanic rocks (streaky/banded tuffs and tuffites, tuff breccias and lapilli tuffs) of mostly intermediate compositions. These have been termed the Harhala Formation (Kähkönen and Nironen, 1994). This supracrustal sequence has been intruded by concordant quartz-feldspar porphyries, plutonic granites and granodiorites, and late stage discordant mafic dykes.

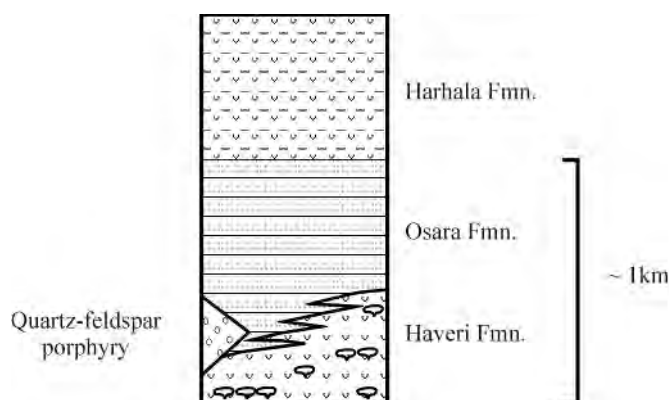
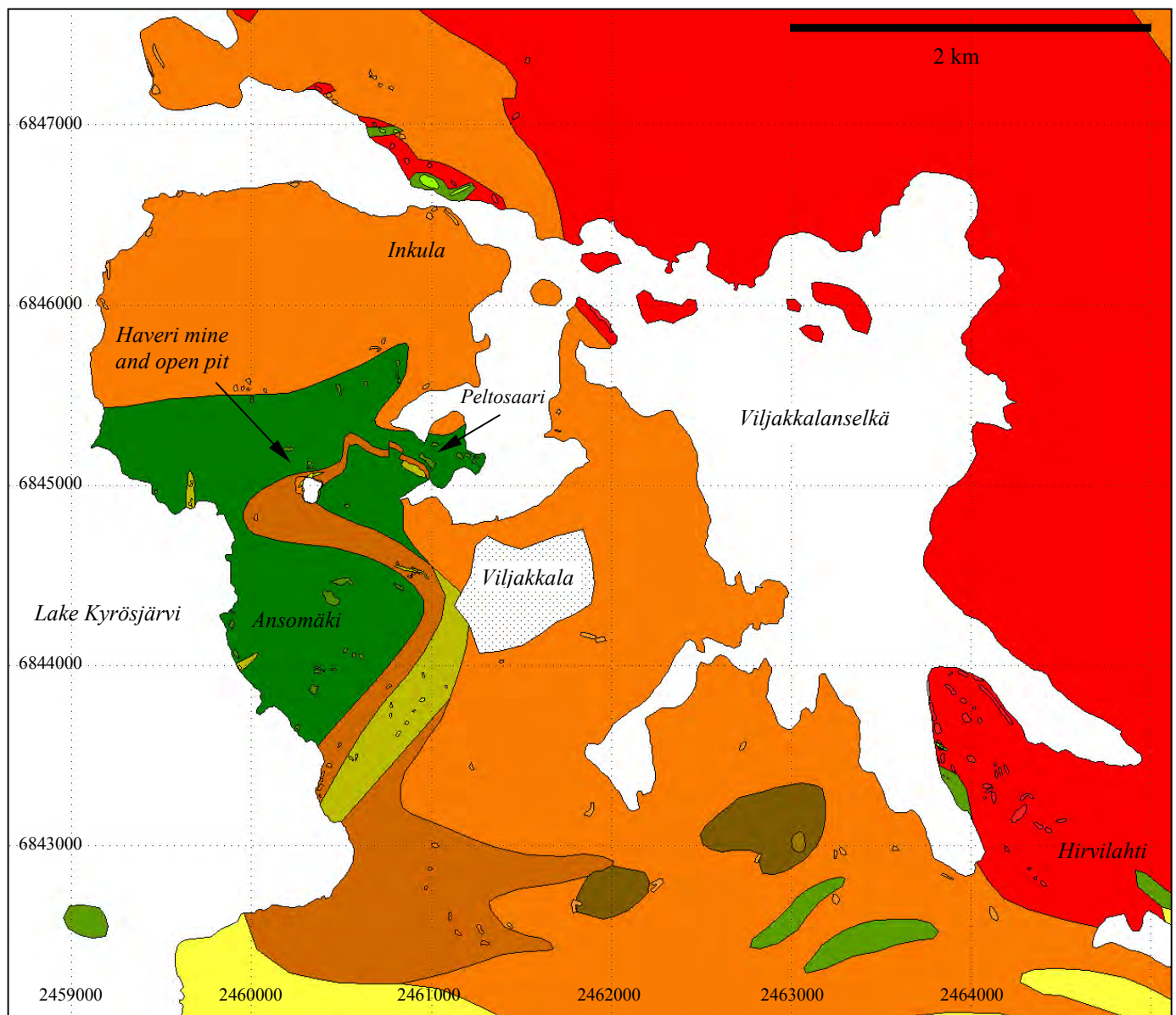


Figure 2.1: Stratigraphy of the supracrustal rocks in the Haveri area. Modified from Mäkelä (1980) and Kähkönen and Nironen (1994)

The bedrock geology of the study area (Figure 2.2) is largely obscured by Quaternary glacial cover. Outcrops of bedrock are sparsely distributed. The glacial cover in the region is typically of 2 – 3 metres in depth and is composed predominantly of unsorted diamictites, interpreted as basal till deposits. Glaciofluvial deposits are not widely represented within the study area, but where present are represented by fragments of eskers. These form ridges composed of rounded cobbles and boulders, gravel and sand. The bedrock surface beneath the glacial cover is typically fresh and relatively unweathered, with only minor oxidation of ferruginous minerals. Striations on outcrops (mainly along the shores of Lake Kyrösjärvi), and on bedrock surfaces from which the cover has been removed, indicate a general ice direction from the N to NNW. This is consistent with formation during the youngest glacial stage (Salpausselkä stage) which ended ca. 9,000 years ago (Koljonen & Tanskanen, 1992).



INTRUSIVE LITHOLOGIES OF THE CFGC

- Gabbro and diorite
- Microcline granite
- Granodiorite
- Quartz-feldspar porphyry

SUPRACRUSTAL LITHOLOGIES OF THE TSB

- Quartz-feldspar schist
- Basic and intermediate tuffite
- Mica schist
- Basic metavolcanic rock

Figure 2.2: Simplified bedrock geology of the study area, displaying the limited distribution of outcrops (irregular polygons on map). The town of Viljakkala is also displayed. Modified from the 1:100,000 map sheets produced by the Geological Survey of Finland (Sheet numbers 2122 & 2124).

2.2 LITHOLOGICAL DESCRIPTIONS AND PETROGRAPHY

2.2.1 Haveri Formation

2.2.1.1 Metavolcanic Rocks

The metavolcanic rocks of the Haveri Formation typically form areas of high relief relative to the metatuffites and metasedimentary rocks. Outcrops of the mafic metavolcanic rocks are usually extensive, and covered in a layer of moss which can be pulled back, to reveal a clean, fresh, unweathered surface of pale grey – green coloured amphibolite, in which textures can easily be discerned.

The metavolcanic rocks are generally fairly massive rock types, representing amphibolitised mafic lava flows. Locally, at Peltosaari (800 metres to the east of the Haveri mine) and at Ansomäki (800 metres to the south of the Haveri mine), pillow structures have been identified, indicating sub-aqueous deposition. The pillows vary in size from 20 cm in diameter to elongated pillows up to about 1 metre in length. Amygdules are up to 1 cm in size, and are found on the inside margins of pillows. The spaces between pillows are commonly filled with epidote. Although many of the pillows have been deformed, at Peltosaari, on the basis of “tails” of the pillows, and amygdules on the opposite margin to the “tails”, the depositional younging direction has been determined to be to the west, towards the old mine. At Ansomäki, although the structures are clearly discernible as pillows, the deformation prevents an unambiguous interpretation of the “way up”. Pillow structures have also been reported from the lateral extension to the volcanics west of Lake Kyrösjärvi (Mäkelä, 1980), indicating a depositional younging direction to the south.

Other structures also present locally in the mafic metavolcanic rocks are breccia structures with angular clasts up to 30 cm long composed of autoclastic mafic volcanic, pillow fragments, and inter-pillow hyaloclastic and epidote rich breccias (Figures 2.3 – 2.5).

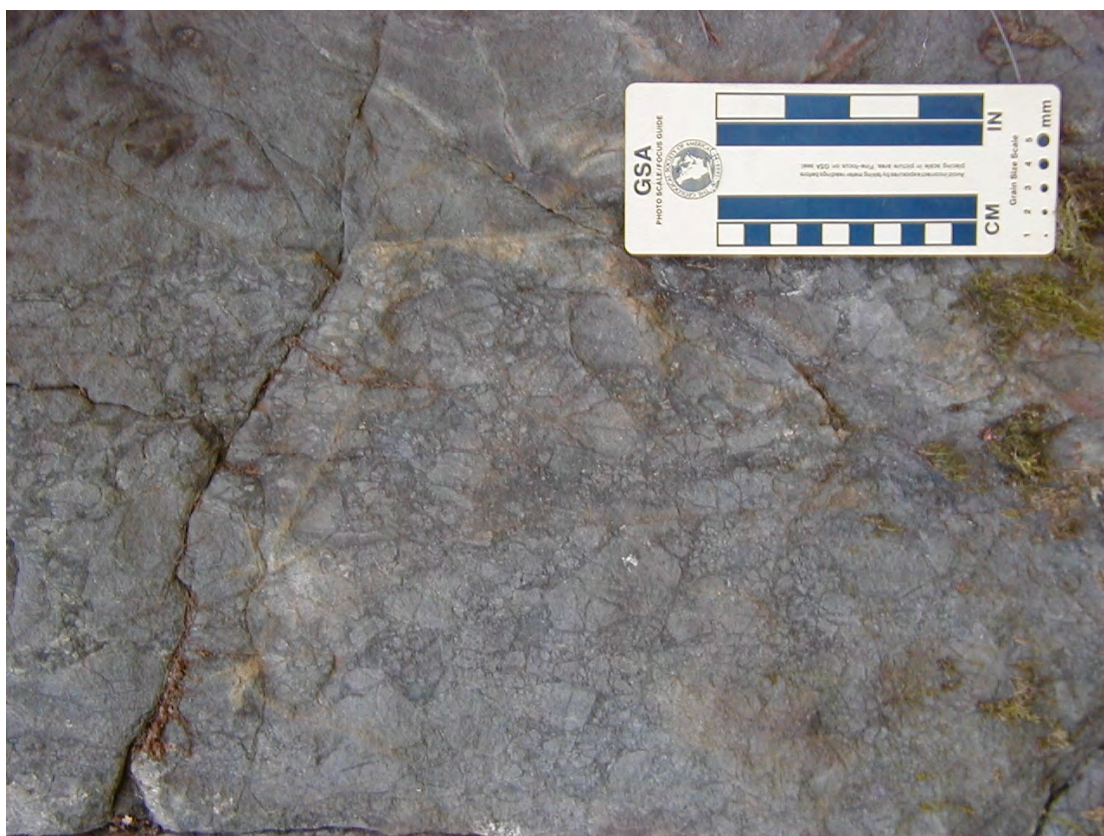


Figure 2.3: Brecciated clasts of mafic metavolcanic rock at Ansomäki. The matrix is composed of amphibole.



Figure 2.4: Amphibole veins brecciating the mafic metavolcanic rock at Ansomäki.



Figure 2.5: Brecciated mafic metavolcanic rock with epidote-rich clasts of interpillow and hyaloclastic material.

In thin section (Figure 2.6), the mafic metavolcanic rocks are typical amphibolites, composed predominantly of granoblastic hornblende and plagioclase, with lesser epidote, sphene and opaques (Table 2.1). The grain sizes of the amphiboles and the plagioclases are on average about 100 – 200 μm , with a maximum size of 350 μm . The grains are generally unaligned, although a weak orientation of the amphiboles is occasionally observed. Epidote, sphene and opaque phases are generally less than 50 μm in size. The opaque phases present in the less altered volcanics consist of intergrown grains of magnetite and ilmenite and fine disseminations of pyrrhotite, chalcopyrite and pyrite. However, it is not clear when these minerals were introduced in relation to peak metamorphism (Figure 2.7).

Mineral	Grain Size	Content (%)
Plagioclase	50 – 350 μm (avg. 100)	30 – 45
Amphibole	50 – 350 μm	50 – 65
Epidote	50 – 100 μm	0 – 2
Sphene	< 60 μm	0 – 2
Opaque	< 50 μm	0 – 2
Biotite	< 150 μm	0 – 5
Apatite	< 50 μm	Trace

Table 2.1: Typical components of the metavolcanic rocks of the Haveri Formation

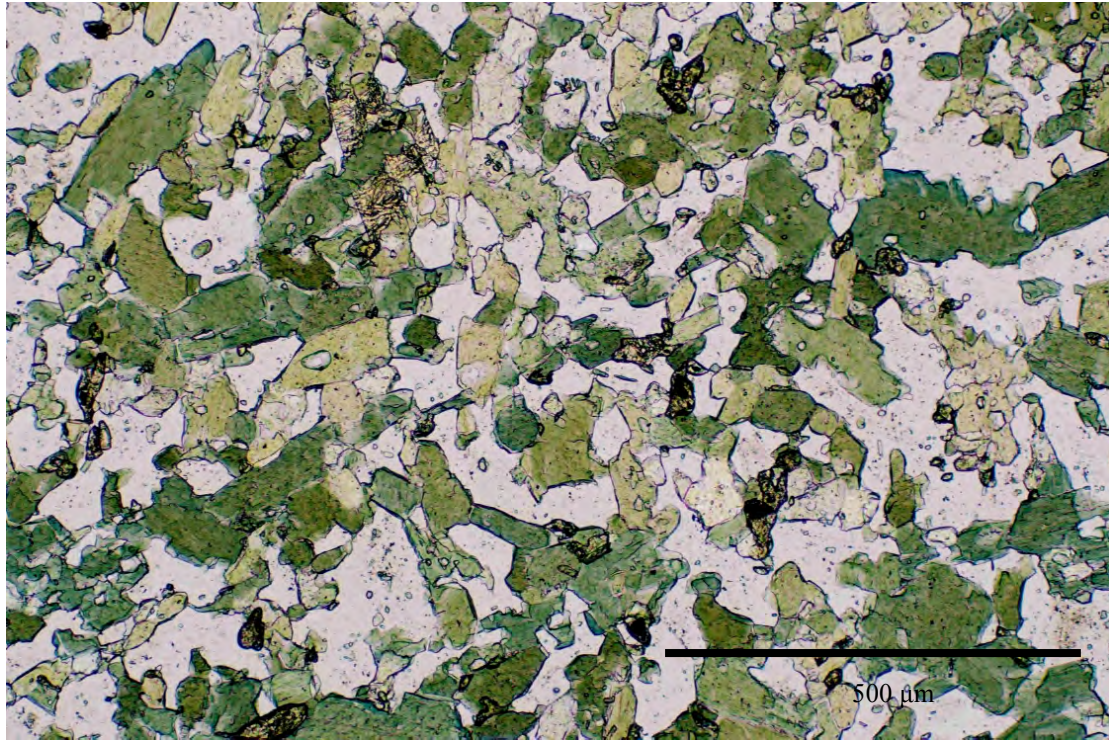


Figure 2.6: Photomicrograph of a typical amphibolite from the Haveri Formation. In addition to the hornblende and plagioclase, there is also epidote and sphene forming a rim around an opaque phase (ilmenite). (Sample P2-89).

The metavolcanic rocks have been affected by a number of later events including a major amphibolitisation event, a calcic-skarn event, and a number of phases of quartz veining and chlorite veining (Section 2.3).

Amphibolites that are relatively free of hydrothermal alteration, commonly display pale spots of 0.5 – 1.0 mm diameter. These consist of partially saussuritised plagioclase “megacrysts” or symplectic textured epidotes or zoisites. These are likely to represent metamorphic replacement of earlier coarse-grained plagioclase amygdales. In addition to discrete grains of sphene, in the least altered mafic-volcanics, sphene commonly forms as a rim around a core of ilmenite (Figure 2.6). Spear (1981) has interpreted this texture as a reaction relationship between sphene and ilmenite with decreasing temperature. It is therefore likely that at least some sphene (that displays this texture) formed later than ilmenite and is a retrograde mineral and not part of the main prograde metamorphic assemblage.

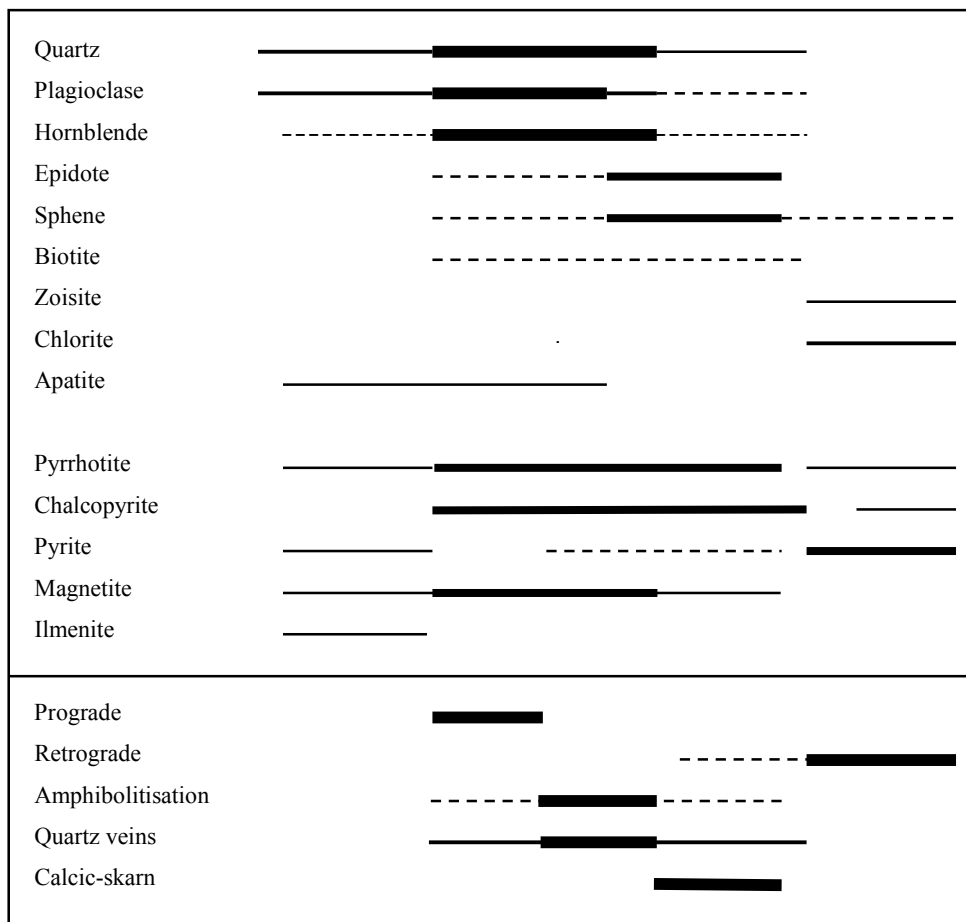


Figure 2.7: *Simplified paragenetic sequence for the volcanogenic rocks (meta-lavas and meta-tuffites) of the Haveri Formation. The length and thickness of the line indicates the duration and relative importance of the minerals formation relative to the main metamorphic and alteration / veining events (lower box).*

2.2.1.2 Meta-tuffites

The meta-tuffites are an intermediate zone between the underlying mafic lavas and breccias, and the overlying metasedimentary rocks of the Osara Formation. The meta-tuffites are characterised by alternating bands of intercalated amphibole-dominated volcanoclastic and biotite-dominated epiclastic material, with a high sulphide content. Outcrops of the meta-tuffites are generally in areas of a low topographic relief. The outcrops are usually limited in extent, with heavily weathered surfaces and poor exposure. Where visible in outcrop, the meta-tuffites display a rhythmic layering (Figure 2.8), visually defined by extensive weathering and development of iron hydroxides, as well as variations in the biotite content affecting the friability of the rock. In places the banding is highly deformed with very chaotic textures (Figure 2.9). This is interpreted as a combination of post-depositional (soft sediment) deformation overprinted by later tectonic deformation. Primary volcanic textures are absent, and sedimentary breccias with volcanic/volcanogenic clasts in a sedimentary matrix have been observed.

The metatuffites represent a gradational zone whereby the volcanoclastic and epiclastic bands vary both in thickness and frequency upwards through the sequence. The main part of the tuffites is characterised by a regular compositional banding (Figure 2.8), with individual bands approximately 1 cm in thickness. At the base of the zone, the metavolcanic rock component predominates with only occasional 5 – 10 cm bands of metasedimentary material. Many of these metasedimentary units are biotite-rich and contain porphyroblasts of garnet and cordierite. At the top of the zone, the volcanoclastic units are subordinate to the sediments, and decrease to nothing as the rock grades into the overlying turbiditic sediments of the Osara Formation. The total thickness of this zone varies between 50 and 200 metres. However due to lack of continuity in exposures, it is possible that the true thickness of this zone is thinner, and that there has been considerable structural thickening.

The composition of the bands within the metatuffite is variable, having characteristics reflecting the source material of the sediment from which they were formed. Thus, as the provenance of the sediments becomes more volcanogenic, the assemblage becomes increasingly dominated by amphibole, and the biotite content decreases. Conversely, as the volcanogenic input of the sediment decreases so does the amphibole content and the assemblage becomes increasingly dominated by biotite.

In thin section, the volcanogenic and volcanic-rich epiclastic bands have characteristics common to the metavolcanic rocks, being dominated by typical amphibolite assemblages. The main minerals are amphibole and plagioclase, with lesser amounts of epidote, sphene and opaques. There is no evidence of textures indicating a primary volcanic genesis, such as vesicle replacement. The opaque content is generally higher than in the more massive meta-lavas lower in the sequence, although individual bands may themselves have low opaque contents. The opaque assemblage consists of fine (ca. 50 μm) disseminations, blebs and stringers of pyrrhotite, chalcopyrite, pyrite, magnetite and ilmenite. As in the metavolcanic rocks, ilmenites are commonly rimmed by sphene. There is also minor replacement of amphibole by biotite.

The epiclastic bands within the meta-tuffites have characteristics similar to the metasedimentary rocks of the Osara Formation, with a mineral assemblage dominated by biotite. As the source of the sediment becomes more argillaceous in nature, the amphibole content decreases and the biotite content increases. These pelitic assemblages are essentially the same as the argillaceous portions of the Osara Formation, with metaluminous minerals such as garnet and cordierite (and their break-down products) being present. The main difference between the epiclastic bands in the metatuffite and the meta-greywackes of the Osara Formation (Table 2.2) is in the opaque content, which is generally higher within the metatuffite, although individual bands may have low contents of opaques. The opaques in the epiclastic portions of the metatuffites consist of disseminations, blebs and stringers of pyrrhotite, chalcopyrite, pyrite, magnetite and ilmenite.

The opaque content of the meta-tuffite unit varies greatly between individual bands, although individual bands may contain in excess of 20 % opaque phases, either disseminated homogeneously within a band, or as blebs and stringers. The disseminated opaques are typically dominated by pyrrhotite and/or ilmenite, with lesser amounts of chalcopyrite, magnetite and pyrite. The heterogeneous distribution of the disseminated opaques *between bands* may either be a primary compositional feature, with the oxides and pyrite being enriched in certain heavy mineral layers, or alternatively may be a post-depositional addition to the rock, controlled by porosity variations between the bands, or a combination of both.

The blebs and stringers are dominated by anhedral pyrrhotite and chalcopyrite with both skeletal and subhedral to euhedral pyrite. Blebs of magnetite and ilmenite are closely associated with the pyrrhotite, and are commonly found as inclusions within pyrrhotite. This suggests that most magnetite and ilmenite predate pyrrhotite, although contemporaneous formation is also likely to have occurred. Chalcopyrite is found as minor inclusions within the pyrrhotite, as fracture fillings, and as disseminations close to and at the apices of fractures, especially within amphibolised rock, and associated with quartz and amphibole veinlets. This suggests that the chalcopyrite formed contemporaneously, but outlasted, pyrrhotite formation, and has been susceptible to remobilisation. The difference in form of the pyrite suggests that there are at least two phases of pyrite formation, with the early phase now remaining as skeletal pyrite, and a more recent subhedral to euhedral phase of pyrite formation. The blebs, stringers and fracture fillings are all interpreted as post-diagenetic events, due to hydrothermal processes and / or structural remobilisation.

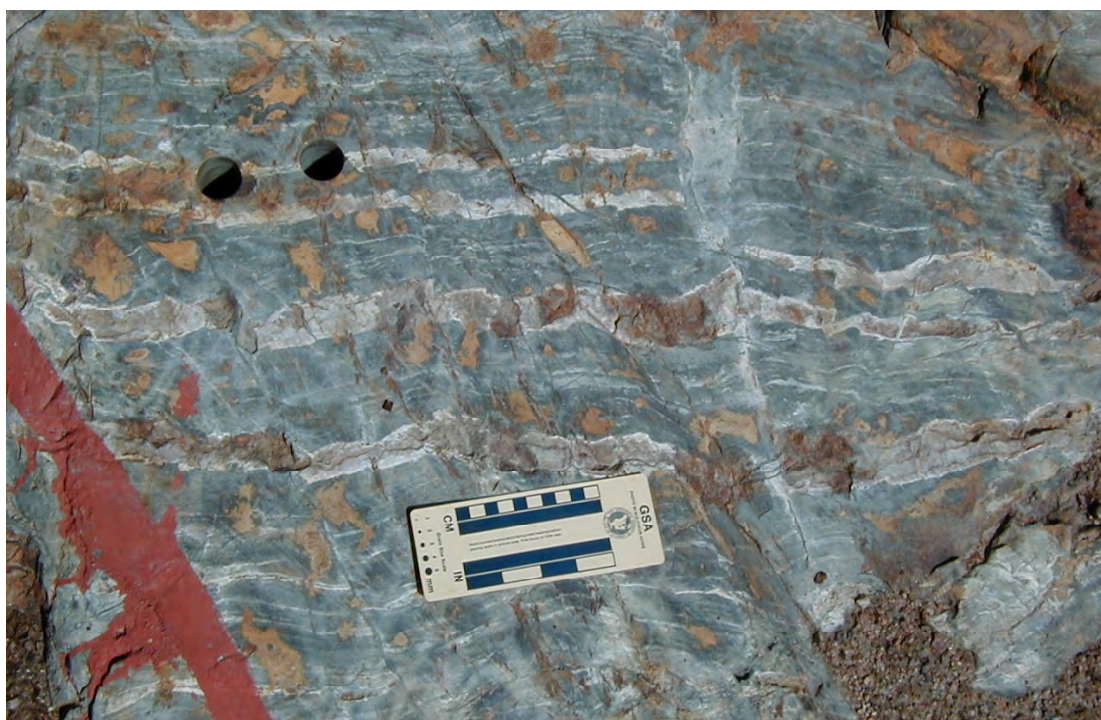


Figure 2.8: Compositional banding between the green volcaniclastic material and the paler (weathered) epiclastic material in the metatuff close to the Haveri open pit. The bands are cut by an S_2 cleavage.

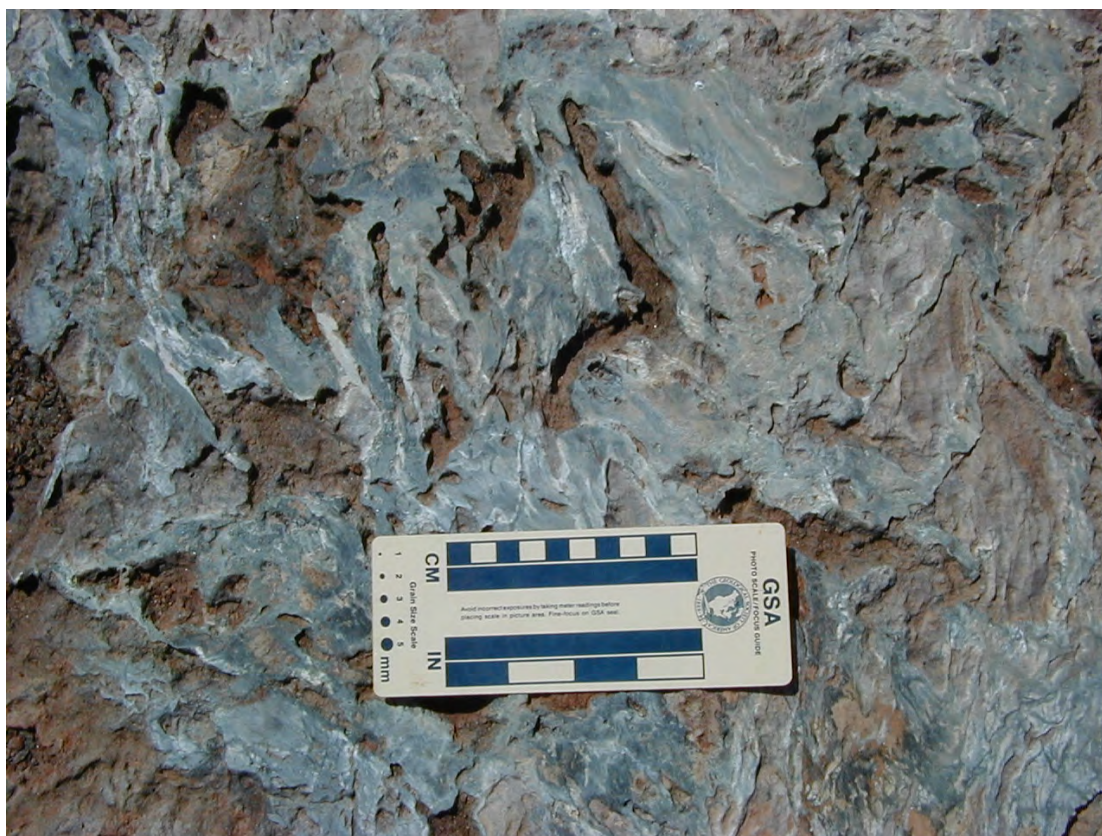


Figure 2.9: *Highly deformed chaotic structures in the tuffites, interpreted as a combination of soft-sediment deformation overprinted by tectonic deformation. Close to Haveri open pit.*

2.2.2 Osara Formation

The metasedimentary rocks of the Osara Formation typically form areas of low topographic relief. The outcrops themselves are generally flat, showing very little relief except along fracture surfaces, making structural measurements more difficult to obtain. The outcrop surfaces are slightly more susceptible to weathering than the volcanics, but more resistant than the tuffites. The best exposures of the Osara Formation are along road cuttings or along the eastern shore of Lake Kyrösjärvi, where the surfaces are relatively clean and internal structures easily discernible.

From clean exposures and from diamond drilling, the Osara Formation can be characterised as grey coloured meta-turbidites with bed thicknesses up to 2 metres. The beds are predominantly arenaceous grading into a more biotite-rich upper portion a few centimetres thick. The biotite-rich upper portions represent metamorphosed finer grained, muddier portions of each turbidite unit. The more voluminous sandy portions of the beds are fairly homogenous and massive in texture. The more biotite-rich upper portions of the beds commonly contain pale spots up to 1 cm in diameter. The pale spots include garnet and cordierite porphyroblasts, or their retrograde breakdown products. On the basis of graded bedding visible on the east shore of Lake Kyrösjärvi, the younging direction is inferred to be to the south, which is in agreement with Mäkelä (1980) and Kähkönen and Nironen (1994).

The present mineralogy of the metasedimentary rocks is dependent on both their initial composition, and also on subsequent metamorphic and metasomatic events. The arenaceous portions of the turbidites are now represented by a mineral assemblage consisting mainly of quartz, plagioclase, and biotite (Table 2.2). In addition there are lesser amounts of amphibole, K-feldspar, fluorite and opaque phases. Most of the plagioclase is cloudy and has only very poorly developed twinning (Figure 2.10) thus making it difficult to differentiate from the K-feldspar. However, it is clear that there is a considerable quantity of plagioclase within the assemblage, suggesting that the sediments were relatively immature, and thus consistent with a turbiditic interpretation.

The hornblende content in the metasedimentary rocks is invariably low, and is most easily explained as minor volcanoclastic contamination of a mainly epiclastic sediment. An alternative explanation for the presence of amphibole is as a manifestation of distal amphibolitisation. However, the amphibole is distributed evenly throughout the rock, and there is no evidence of veining, or other textures that would suggest that the amphiboles formed later than the other metamorphic minerals. In some of the arenaceous units with considerable amphibole and biotite, there are also accessory grains of fluorite (Figure 2.11). This indicates, at the very least, that hydrothermal fluids passed through such fluorite-bearing units, and may even have been more widespread.

Mineral	Argillaceous		Arenaceous		Texture	Genesis
	Grain size	Content (%)	Grain size	Content (%)		
Quartz	< 300 μm	5 – 40	< 200 μm	35 – 45		Metamorphic
Plagioclase	< 350 μm	0 – 10	< 150 μm	35		Metamorphic
Biotite	< 800 μm	45 – 85	< 150 μm	25	Foliated	Metamorphic
Amphibole		< 5		< 5		Metamorphic
K-feldspar						Metamorphic
Andalusite	< 100 μm	< 1				Metamorphic
Garnet	< 10 mm	0 – 15			Porphyroblast	Metamorphic
Cordierite	< 10 mm	0 – 15			Poikiloblast	Metamorphic
Muscovite	< 150 μm	0 – 15		0 – 45	Replacement	Retrograde
Chlorite	< 250 μm	0 – 20			Replacement	Retrograde
Epidote	< 200 μm	< 5			Replacement	Retrograde
Zoisite						Retrograde
Apatite	< 250 μm	< 2				Primary
Allanite (?)	< 25 μm	< 1				Primary
Spinel	< 100 μm	< 1				Metamorphic
Sphene			< 50 μm	0 – 5		Metamorphic
Opaque	< 300 μm	5 – 10	< 50 μm	5 – 10		(Primary)/ Met

Table 2.2: Typical components of Osara Formation meta-turbidites

The argillaceous upper portions of the turbidites are represented by a pelitic, biotite-rich assemblage, which due to metamorphic recrystallisation, is coarser grained than the sandier portions (Table 2.2). These pelitic portions are composed primarily of biotite and opaques, with minor amounts of quartz, apatite, andalusite, plagioclase, K-feldspar and spinel (Figure 2.12). In addition, the peraluminous pelitic portions commonly contain porphyroblasts and poikiloblasts of garnet and cordierite or their breakdown products such as muscovite and chlorite. The biotites in all of the metasedimentary rocks contain numerous pleochroic haloes, surrounding a radioactive mineral. These grains are too small to determine optically, but may possibly be allanite. However, electron microprobe analyses are required to enable accurate identification. Crenulation cleavages are commonly developed and are easily discernible in the biotite-rich units (e.g. Figures 2.55 – 2.56).

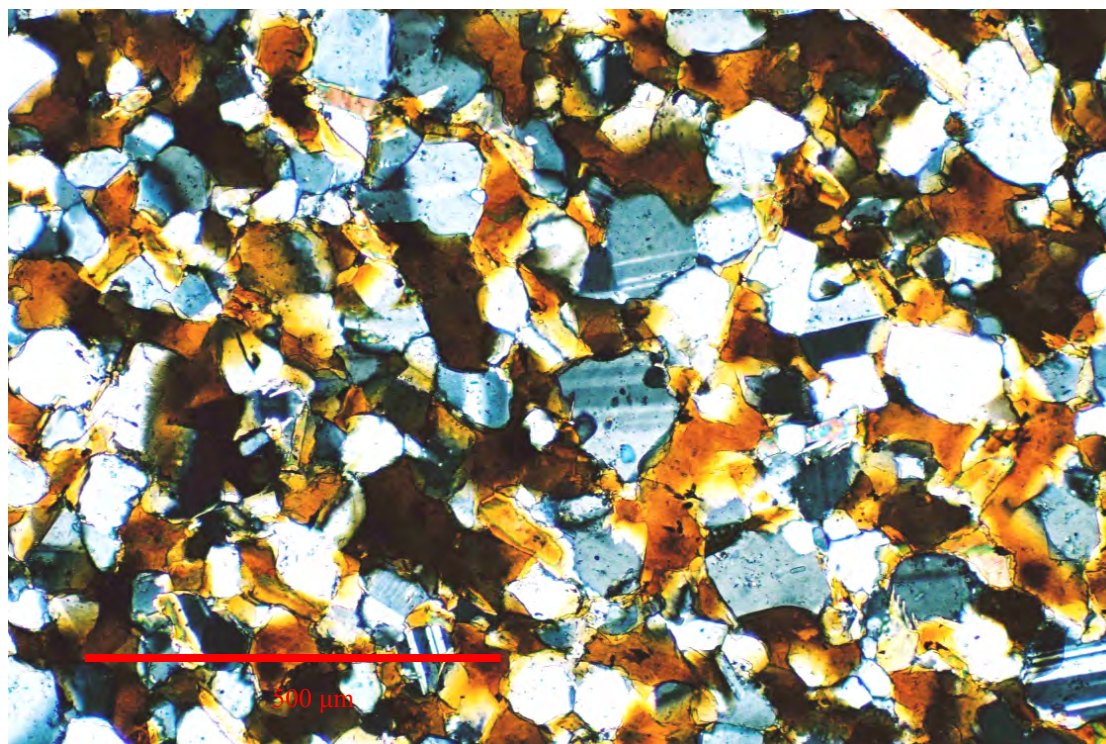


Figure 2.10: A typical arenaceous metasedimentary rock, primarily composed of biotite, quartz and poorly twinned plagioclase. Pleochroic haloes are evident in the biotite. (Sample P9-27).

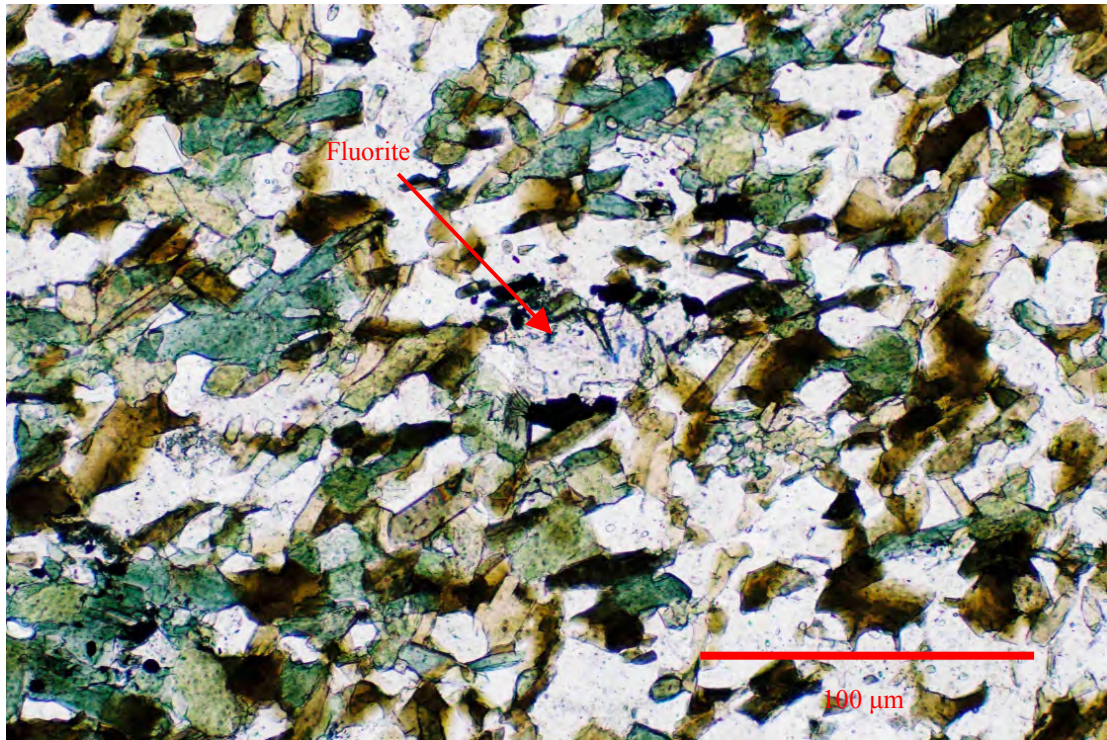


Figure 2.11: A fluorite grain (centre field) in a mafic tuff (Sample TS-37).

The garnet porphyroblasts are usually less than 1 cm in diameter and are euhedral (Figure 2.13). They contain inclusions of biotite, quartz/plagioclase, spinel and opaques. The euhedral faces of the garnet at times truncate the biotites, and at other times the biotite “flows” around the garnet. This suggests that the biotites and the garnet porphyroblasts formed contemporaneously, and most likely record the peak metamorphic event. The opaque minerals are principally magnetite and ilmenite, and occur both interstitially and as inclusions in the prograde minerals, biotite, garnet and cordierite. In addition to magnetite and ilmenite, other opaque phases include pyrrhotite and chalcopyrite, occurring as later phases.

Cordierite porphyroblasts and poikiloblasts are usually less than 1 cm in diameter, and are ovoid in shape. They are typically sieve textured with inclusions of biotite and opaques, and commonly display well developed sector zoning (Figure 2.14). Another characteristic of the cordierite porphyroblasts is the presence of yellow pleochroic haloes indicating the onset of pinitisation. The porphyroblasts of garnet and cordierite are commonly replaced by retrograde products. The garnets are typically replaced by zoisite, epidote and opaques, perhaps with the skeletal remains of the original garnet. Pale blotches with a rim of large muscovite laths and a core composed of mainly muscovite and chlorite, and lesser biotite, plagioclase and opaques, are interpreted as pseudomorphed porphyroblasts after cordierite (Figure 2.15), although Kähkönen and Nironen (1994) suggested that they are pseudomorphs of andalusite or staurolite.

Most of the metasedimentary rocks in the study area show the prograde assemblages described above. However, there is commonly a certain amount of retrograde alteration overprinting the prograde mineral assemblages. One of the most common manifestations of retrograde alteration is the replacement of biotite by chlorite (Figure 2.16). This is commonly accompanied by needle-like opaques occurring along cleavage. This is interpreted as the inability of the chlorite to accommodate titanium from the original biotite, and thus it is exsolved as needles of a titanium-rich phase, probably rutile. Another texture interpreted as a retrograde alteration product of biotite, is the development of wavy, colourless, 4 mm long “flasers” of epidote within the biotite (Figure 2.17). These occur in the most argillaceous metasedimentary units, composed almost entirely of coarse-grained biotite. In hand specimen, the “flasers” look like paler biotites.

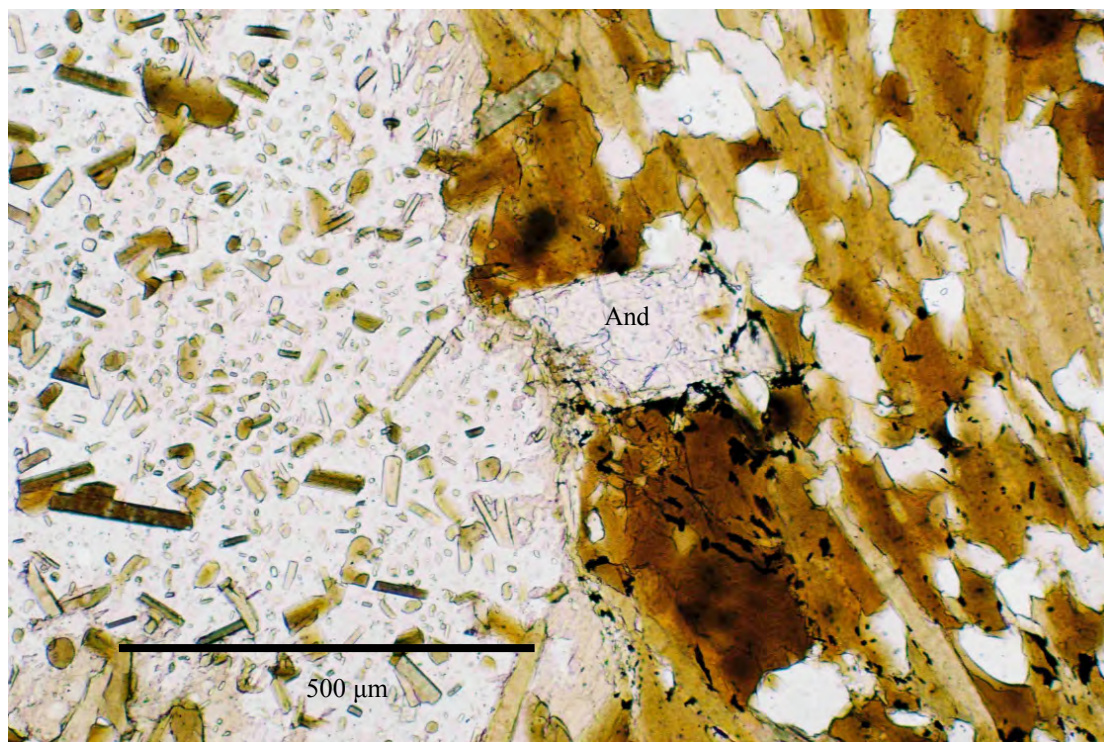


Figure 2.12: *Andalusite (And) close to the contact between a sieve-textured cordierite (left) and a biotite-quartz-feldspar argillaceous metasedimentary rock. (Sample P9-27).*

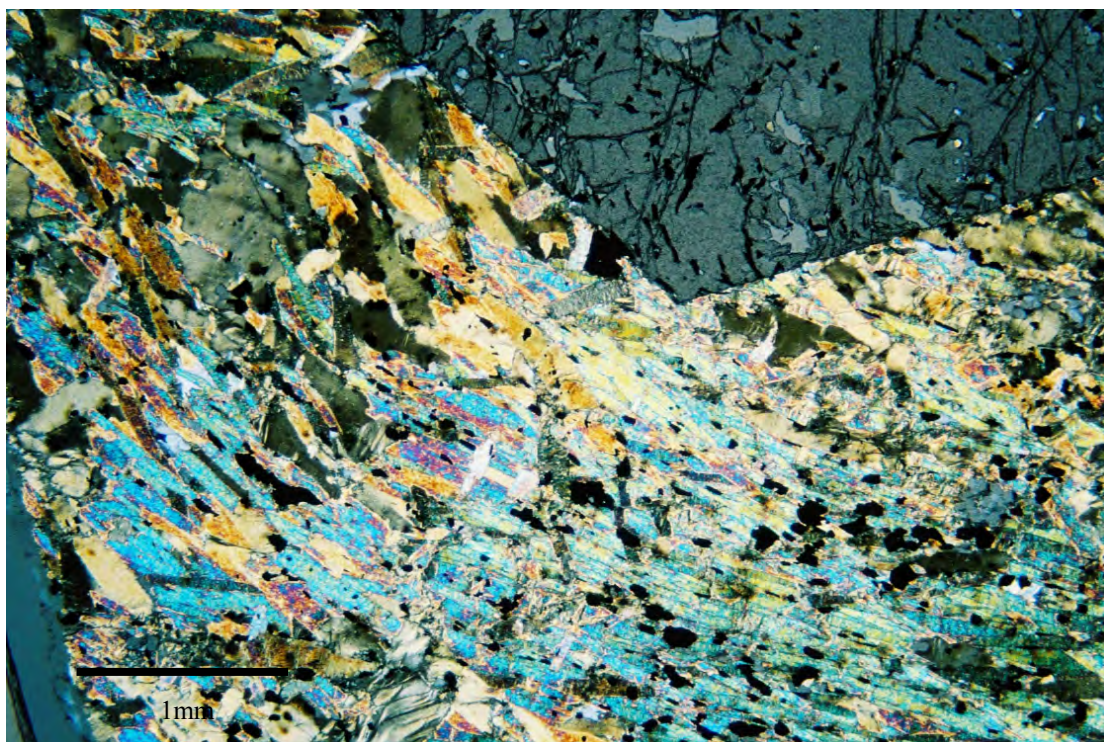


Figure 2.13: Biotite flowing around, and being truncated by, an euhedral garnet porphyroblast in an argillaceous unit (the polarisers are slightly offset) (Sample P5-16).

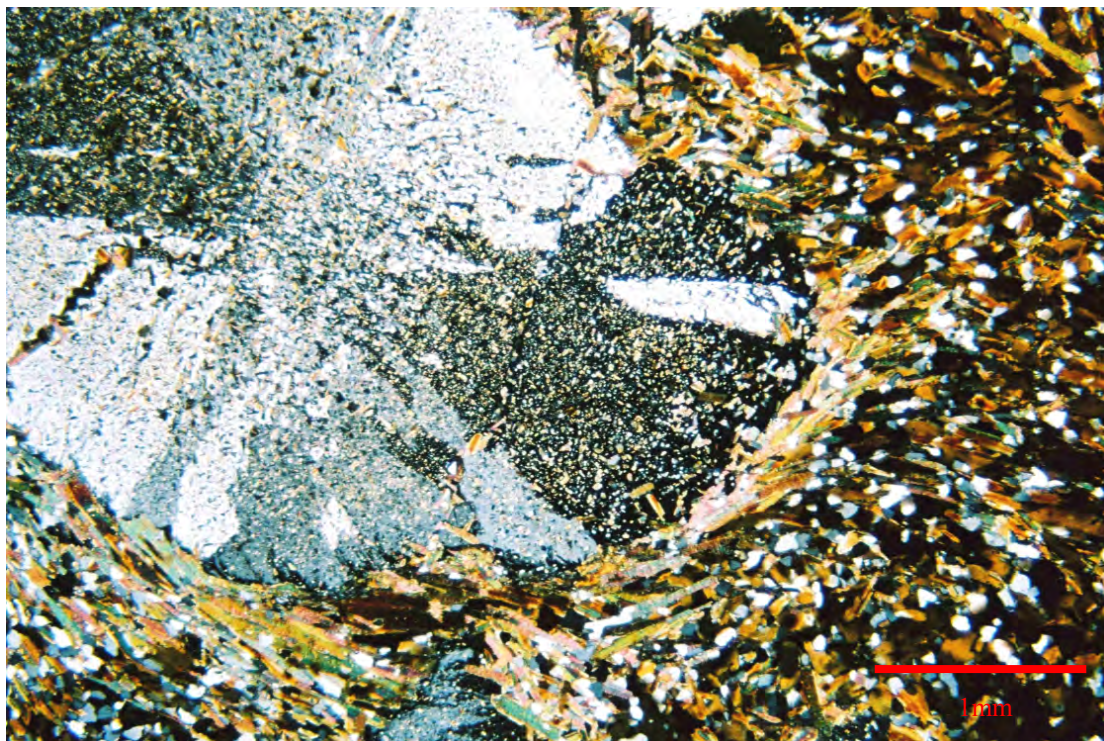


Figure 2.14: Sector zoning in a sieve-textured cordierite porphyroblast, with MS1 biotite “flowing” around the cordierite (Sample P9-27).

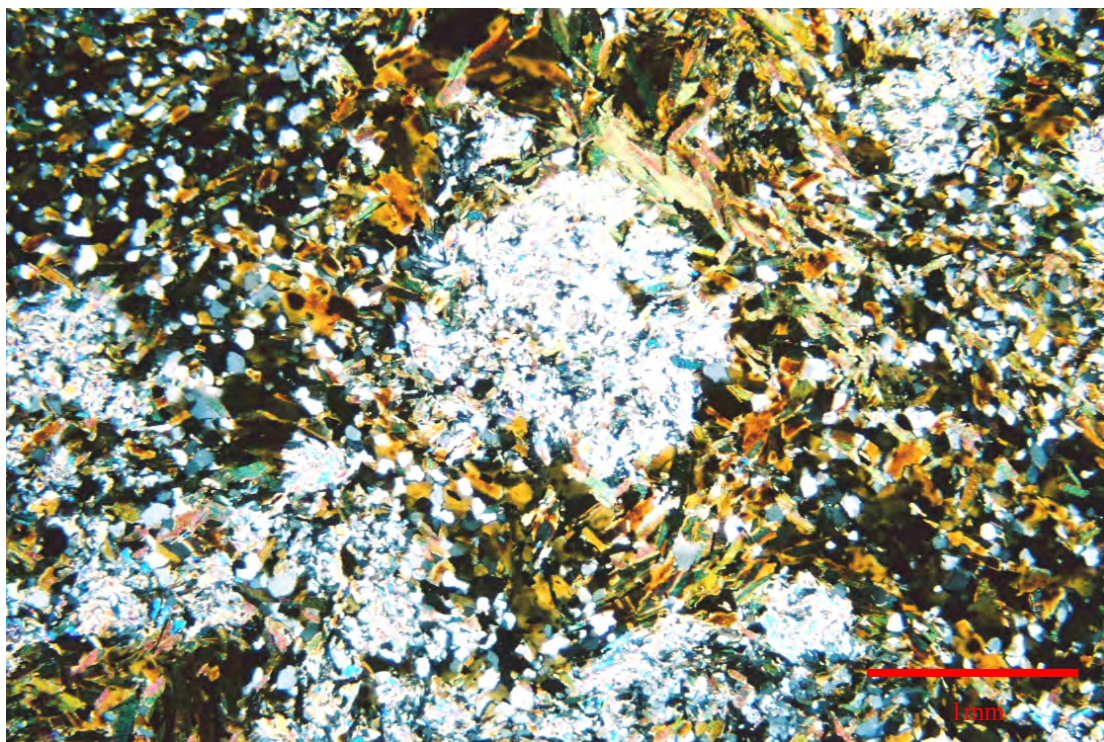


Figure 2.15: Retrograde muscovite pseudomorphing a cordierite porphyroblast. (Sample P9-43).

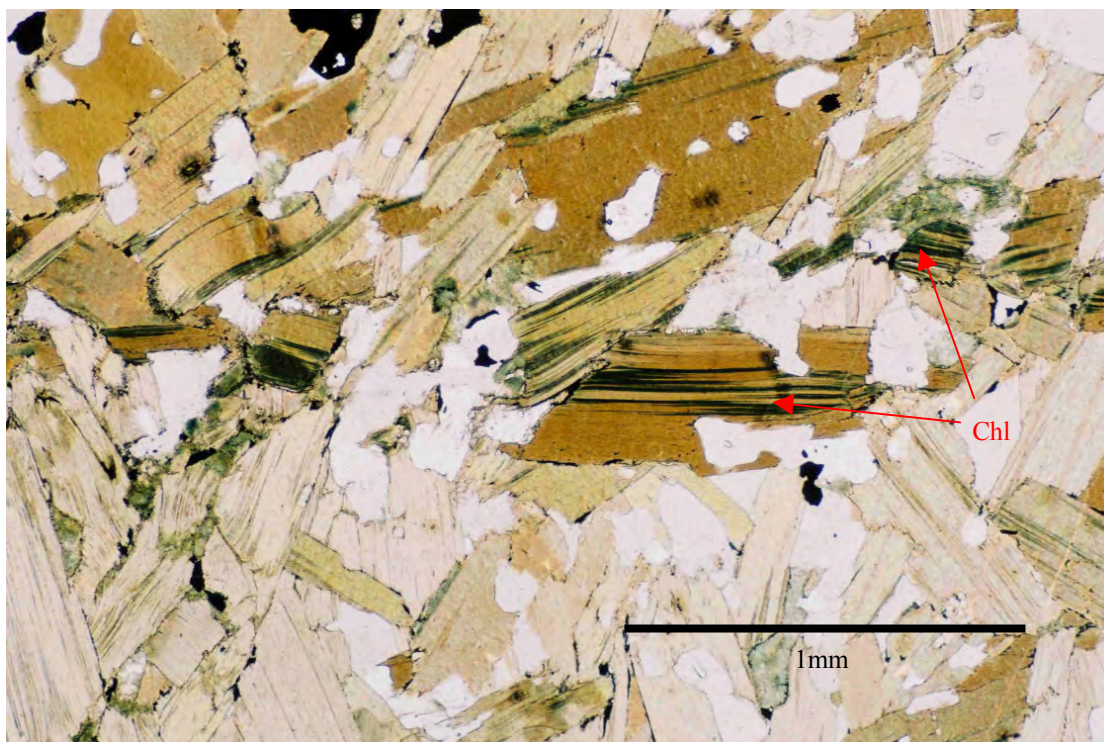


Figure 2.16: Retrograde chlorite replacing biotite along cleavage. (Sample P13-28).

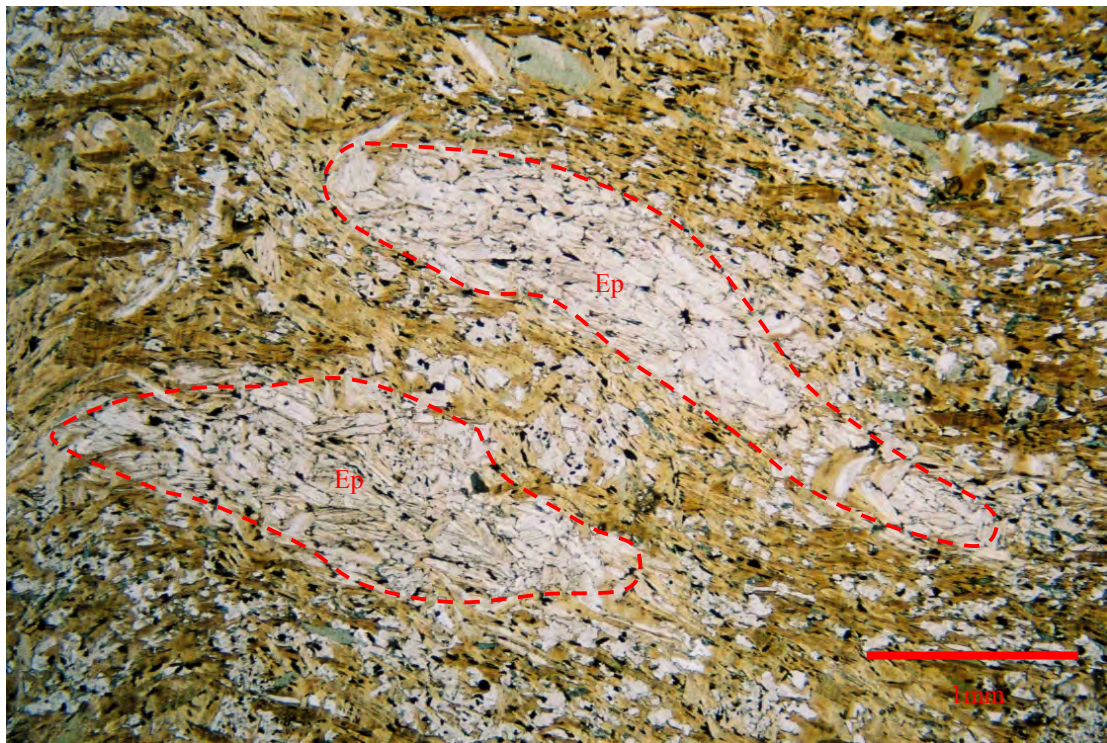


Figure 2.17: Epidote flasers in biotite, interpreted as a retrograde product. (Sample R7-104b).

In the meta-pelitic units, the opaque phases occur both as inclusions within the metamorphic biotite grains (i.e. poikilitic texture), as well as interstitial to the biotite grains, being trapped between grain boundaries. This indicates that many of the opaques (i.e. interstitial minerals) were formed, or at least recrystallised, during metamorphism, whereas others (i.e. inclusion in poikilitic biotite), predate metamorphism. A simplified paragenetic sequence is shown in Figure 2.18. The overall opaque content of the metasedimentary units varies considerably between thin sections, although on a smaller scale it is apparent that the opaques are distributed in layers. This suggests that the distribution of opaques in the metasedimentary rocks is a result of a primary compositional layering, or alternatively, are secondary and reflect porosity variations in the rock.

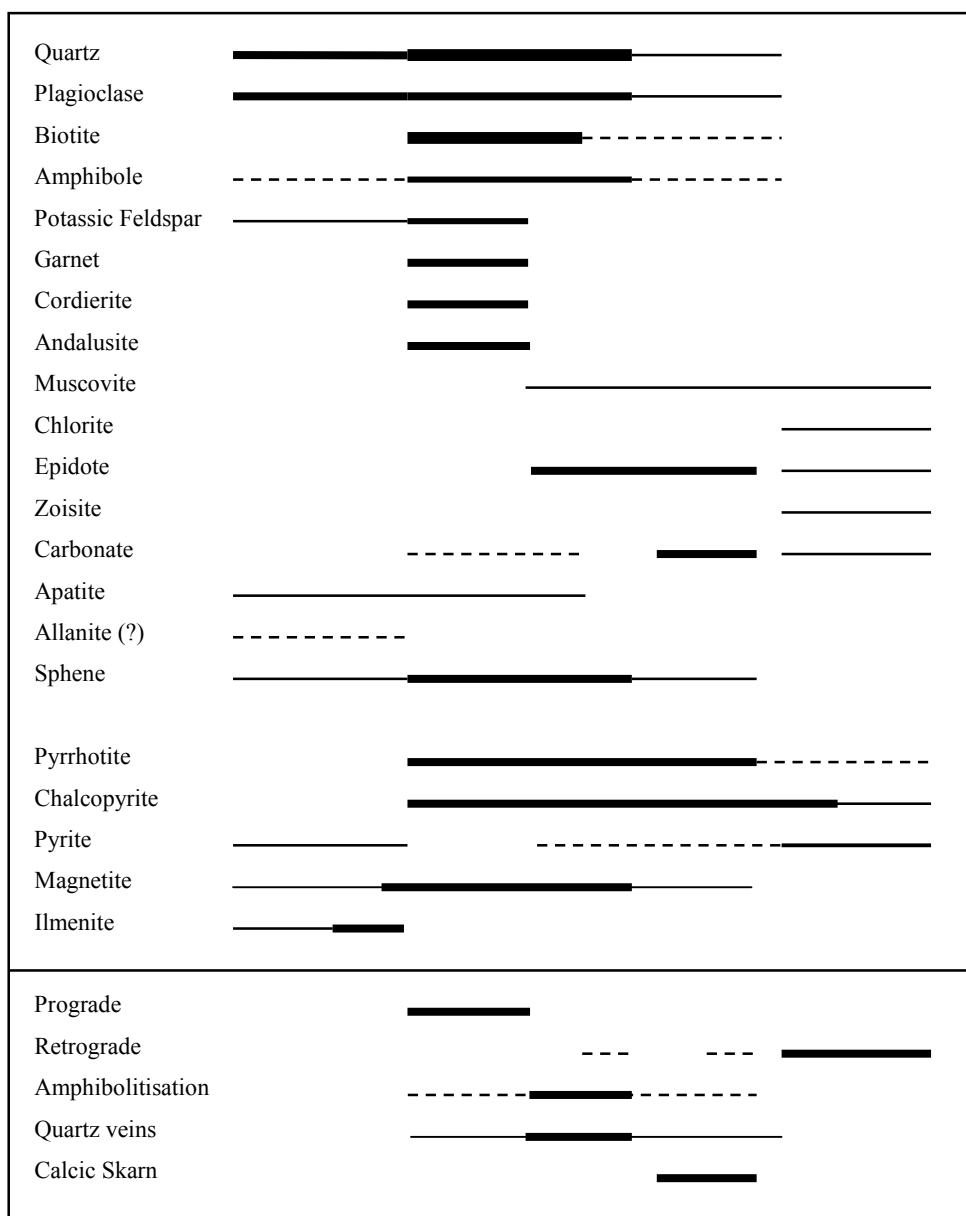


Figure 2.18: Simplified paragenetic sequence for the epiclastic units in the metatuffites and the metaturbidites of the Osara Formation. The length and thickness of the line indicates the duration and relative importance of the minerals formation relative to the main metamorphic and alteration / veining events (lower box).

2.2.3 Intrusives

A number of intrusive lithologies are found in, and surrounding the study area. These include the concordant porphyritic rocks of the Haveri Formation (Mäkelä, 1980), the regional granites and granodiorites of the Central Finland Granitoid Complex, and minor late-stage mafic dykes.

2.2.3.1 Quartz-Feldspar Porphyries

The quartz-feldspar porphyries are fairly resistant in relation to the metasedimentary rocks and tuffites, and thus form minor topographic highs when located within these rock types. The outcrops of porphyry, like the volcanics, are usually covered by a layer of moss that can be pulled back to reveal an unweathered and fresh surface, enabling the quartz and feldspar phenocrysts to be easily identified.

The quartz-feldspar porphyry has previously been included within the supracrustal sequence of the Haveri Formation (Mäkelä, 1980). Kähkönen and Nironen (1994) suggest that these felsic and intermediate dyke rocks are subvolcanic in origin and cut the supracrustal rocks. From outcrop mapping combined with drill hole data, it is clear that the quartz-feldspar porphyries are generally concordant to the Haveri Formation on a regional scale, and most have been intruded into the tuffites close to, or at the contact of, the overlying Osara Formation. On outcrop scale the porphyries can be seen to be locally discordant with respect to the supracrustal sequence. Less extensive intrusions of quartz-feldspar porphyries (less than 2 metres thick) have been intruded into the volcanics lower in the stratigraphy. The maximum thickness of the porphyry is approximately 40 – 50 metres, although, as is the case in the outcrop west of the Haveri open pit, a certain amount of this may be accounted for by structural thickening. In boreholes, 10 – 20 metres is a more normal intersection. It is not clear whether there are several quartz-feldspar porphyries present, or whether the smaller ones are apophyses of a single porphyry. There is no evidence of chilled margins on the porphyries, indicating that they were emplaced into relatively hot country rocks.

In outcrop the typical porphyry is a massive, fine-grained pale grey rock with 0.5 – 4.0 mm-sized plagioclase and quartz phenocrysts. Quartz phenocrysts are not always present. Some occurrences of the porphyry also have amphibole phenocrysts and coarse clusters of amphibole; these porphyries are very distinct in hand specimen being darker than the “normal” porphyry. These latter porphyries are commonly mapped in Finland as “uralite porphyry”. At Haveri, they are almost always observed in gradational contact with the normal quartz-feldspar porphyry, and hence are interpreted as an alteration product, rather than as a discrete porphyritic intrusion.

In thin section, the porphyry consists of megacrysts of quartz, plagioclase and amphibole, with a groundmass of mainly quartz, plagioclase, amphibole with or without biotite. There are also minor amounts of epidote, sphene, apatite, carbonate, and possibly monazite. Typical mineral phases present

in the porphyries are shown in Table 2.3. The megacrysts generally account for 15 – 20 % of the rock mass. The average grain size in the groundmass is 100 μm (although maximum sizes are somewhat higher).

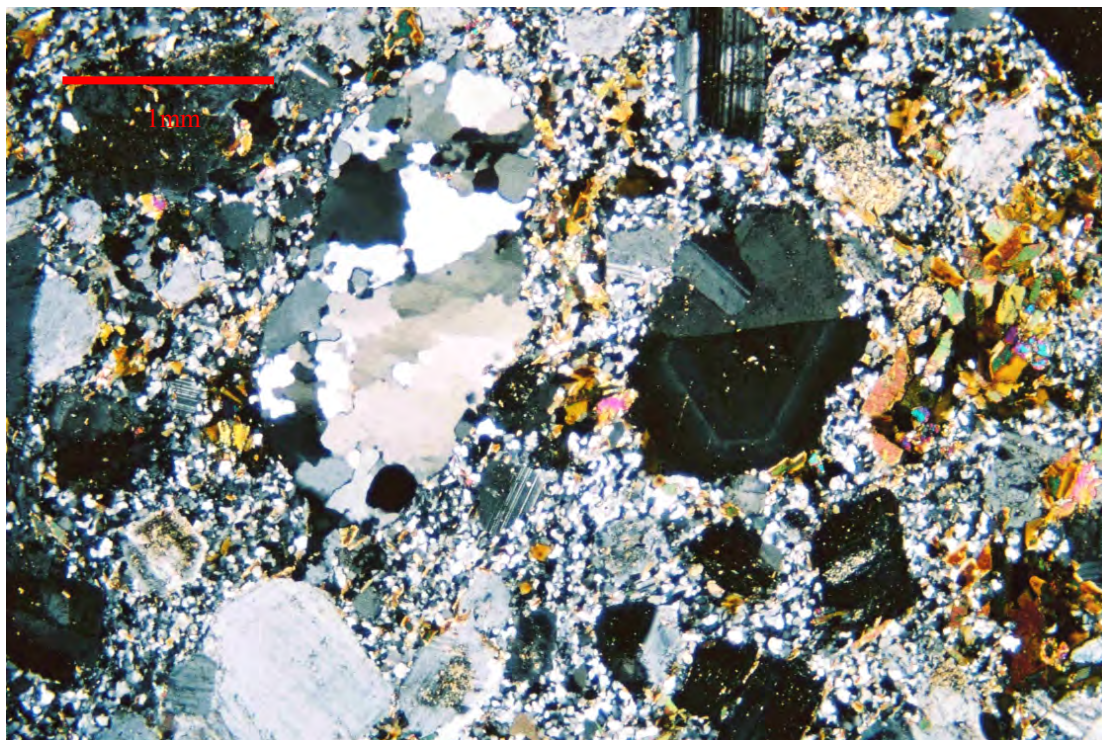


Figure 2.19: Photomicrograph of a typical quartz-feldspar porphyry (Sample P11-22).

Mineral	Grain size	Content (%)	Texture
Quartz	0.5 – 4 mm	0 – 5	Megacryst
Plagioclase	0.5 – 3.5 mm	5 – 20	Phenocryst
Amphibole	0.5 – 2 mm	0 – 5	Megacryst
Quartz	100 μm	25	Groundmass
Plagioclase	100 μm	25 – 35	Groundmass
Amphibole	100 μm	0 – 35	Groundmass
Biotite	< 300 μm	0 – 40	Groundmass
Epidote	< 150 μm	< 1	Groundmass
Sphene	< 250 μm	< 2	Groundmass
Apatite	< 200 μm	< 1	Groundmass
Monazite (?)	< 60 μm	Trace	Groundmass
Opaque	< 150 μm	0 – 2	Groundmass

Table 2.3: Typical components of the quartz-feldspar porphyry

The quartz megacrysts, when present, vary in size between 0.5 mm and 4.0 mm. They are composed of polycrystalline quartz with well sutured grain boundaries and extensive sub-grain development. These textures indicate that the quartz megacrysts have been subject to high stress and deformation. Occasionally the quartz megacrysts contain minor inclusions of biotite, epidote and carbonate.

The plagioclase phenocrysts are the dominant megacrystic phase in the porphyries and vary in size between 0.5 and 3.5 mm. The plagioclase phenocrysts are commonly zoned with an inclusion-rich core and a more homogeneous rim. The inclusions within the feldspar phenocrysts are mainly calcic minerals such as amphibole with minor epidote, sphene and carbonate. Biotite inclusions reflect the presence of potassium. The calcic-inclusion-rich core indicates a more calcic-rich composition relative to a more sodic rim. The phenocrysts, although containing inclusions, are generally not saussuritised, except close to skarn veins and veinlets. This implies that any retrograde effects are only local in distribution, and are related to the skarn veins rather than to a pervasive regional event.

The amphibole megacrysts in the “uralite porphyry” commonly occur in two forms. One type is as coarse (< 1.75mm) isolated grains full of quartz inclusions. Roberts et al. (2000) interpret this texture as a retrograde uralitisation, pseudomorphing pyroxene. The second type of amphibole megacrysts is as clusters (< 2mm) of densely packed amphiboles (< 300 μm), with minor quartz, epidote and opaque phases associated with them (Figure 2.20). Many of the amphiboles display pleochroic haloes, indicating the presence of inclusions of radioactive accessory minerals. The “uralite porphyry” as well as having amphibole megacrysts also has a higher content of amphibole in the groundmass, causing the darker appearance.

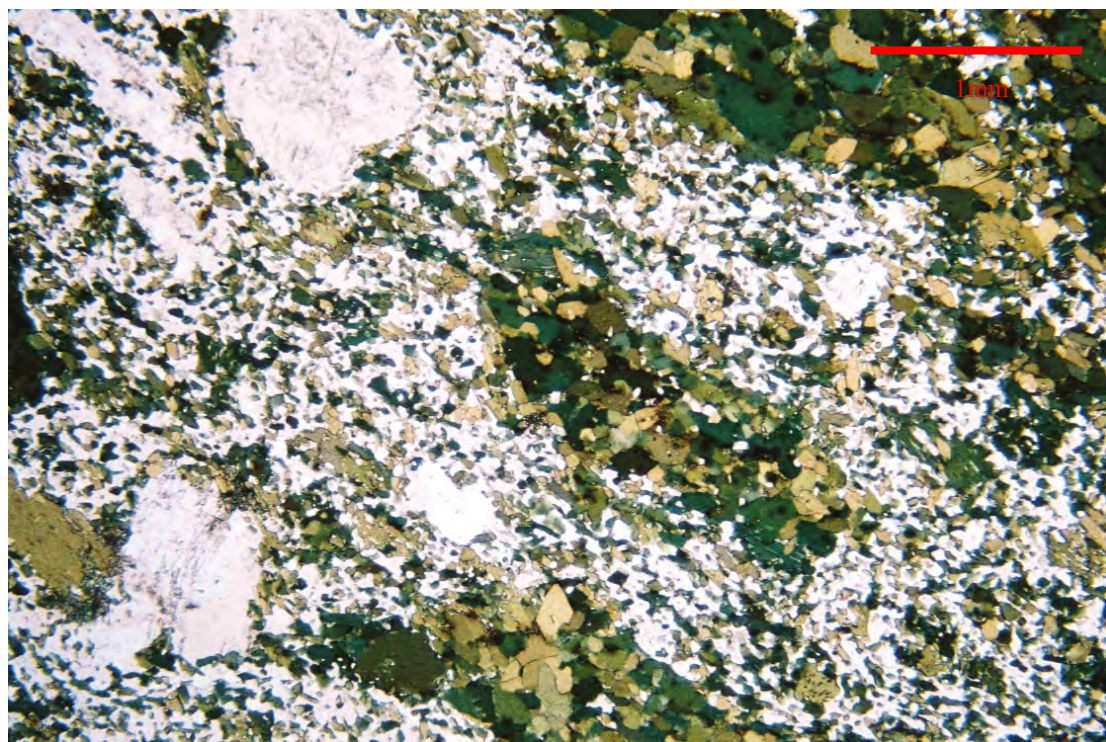


Figure 2.20: Amphibole megacrysts in the uralite porphyry (Sample P13-98).

The main mineral phases in the groundmass are plagioclase, quartz and amphibole. Groundmass quartz and feldspar together make up approximately 40 – 50 % of the rock mass. Much of the plagioclase has only very poorly developed twinning. One noticeable feature of the porphyries is the varying presence of biotite. Some sections are totally devoid of biotite, whereas in others biotite is a major constituent. Even where biotite is present within a section, the distribution is heterogeneous. Where both biotite and amphibole are present the textures suggest that biotite is replacing the amphibole.

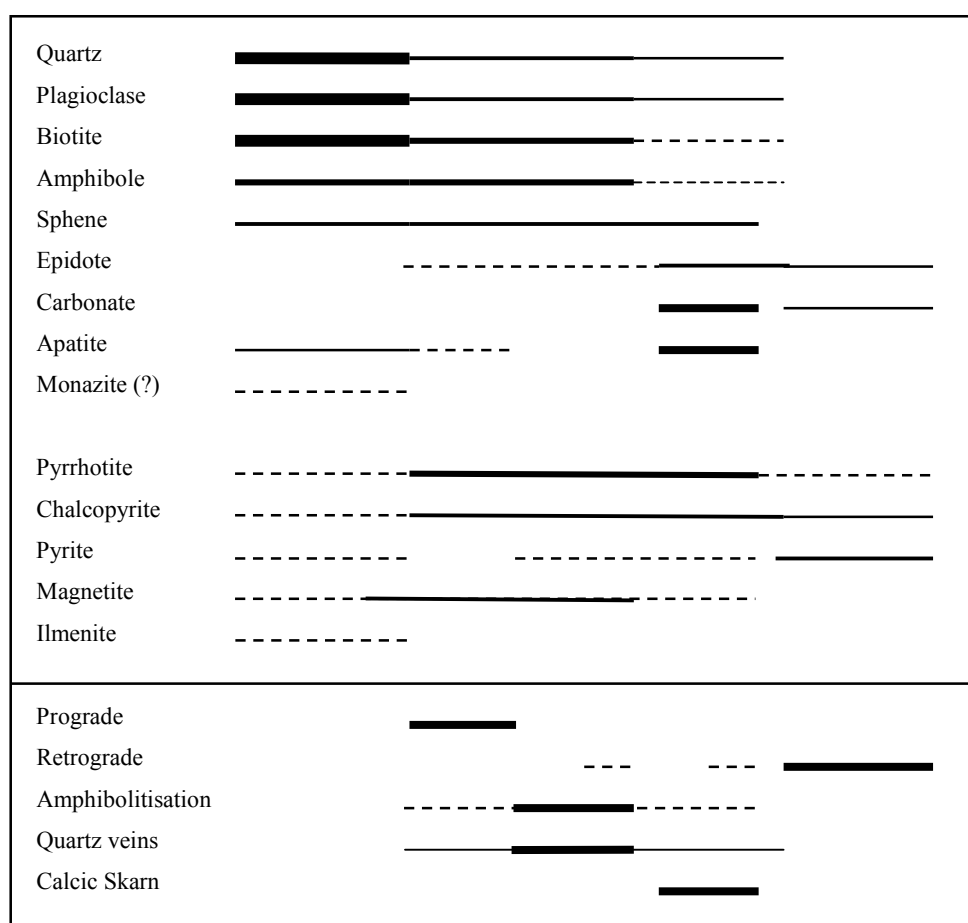


Figure 2.21: Typical paragenetic sequence for the quartz-feldspar porphyry. The length and thickness of the line indicates the duration and relative importance of the minerals formation relative to the main metamorphic and alteration / veining events (lower box).

Primary, non-hydrothermal, opaque phases in the porphyries are only a minor constituent of the mineral assemblage. When present they rarely account for more than 2 % of the assemblage, and occur as disseminated grains of pyrrhotite, with lesser amounts of chalcopyrite, pyrite and magnetite.

Radioactive phases are most commonly seen within the biotites and amphiboles, and are recognisable as minute grains surrounded by a pleochroic halo. These grains are too small to be determined optically. Only one section examined contained a radioactive grain of sufficient size to enable optical examination. This grain is approximately 100 µm in diameter, and is contained in a feldspar phenocryst. This is possibly monazite.

2.2.3.2 Granites and Granodiorites

Granitoids are the predominant rock types of the Svecofennian of Finland, and are thus commonly closely associated with Svecofennian supracrustal successions (e.g. Lahtinen, 1994). The southern margin of the Central Finland Granitoid Complex lies approximately 500 metres to the north of the Haveri mine, and approximately 1 km to the east (Figures 1.1; 1.7; 2.2). Drilling below Lake Kyrösjärvi combined with geophysics suggests that granitoids are also present approximately 800 metres to the west of the Haveri mine. Five kilometres to the south of Haveri lies the 1.88 Ga Hämeenkyrö Batholith.

In the immediate vicinity of the Haveri mine, there are two main types of granitoid intrusion (Figure 2.2): a microcline granite and a granodiorite. The granodiorite is spatially the closest granitoid intrusion being located to the north, east and west of the mine. Within the old mine itself, there is a 15 cm thick horizontal dyke of granodiorite. The microcline granite covers a large area but is found further from the mine site (Figure 2.2). The nature of the contact between the two rock types has not been observed.

Outcrops of both the granite and granodiorite are typically positive topographical features in the landscape. The outcrops are either covered by a layer of moss, which can be pulled back to reveal fresh exposures, or alternatively, are bare rock covered by extensive growth of lichen, obscuring the nature of the rock type. Both the granite and the granodiorite are massive, coarse-grained rocks.

The granodiorite is typically pale grey coloured, with easily identifiable grains of quartz, feldspar, biotite and amphibole. The granodiorite has a porphyritic texture, with the average grain size of the quartz and feldspar phenocrysts being about 2 – 3 mm. The granodiorite is commonly cut by green epidote and carbonate veinlets, as well as by pink veinlets of microperthitic potassic feldspar.

In thin section the granodiorite is composed of primarily quartz, plagioclase, K-feldspar, biotite and amphibole, with lesser amounts of epidote, sphene, carbonate, muscovite, apatite, and opaques (Table 2.4). Many of the plagioclase phenocrysts display compositional zoning, with cores containing abundant epidote, carbonate and muscovite inclusions, indicating a more calcic core relative to a sodic rim. In addition, the potassic feldspar and the plagioclase display considerable sericitisation and saussuritisation respectively, although the degree of alteration varies from grain to grain. The quartz grains show a reasonably well developed sub-grain development, suggesting post-crystallisation deformation. Biotite is ubiquitous in the granodiorite, and typically contains pleochroic haloes surrounding small, euhedral crystals of zircon. Biotite is interstitial to the quartz and feldspars, and is associated with epidote, sphene, and when present, also hornblende. The hornblende is present in the granodiorites to the north and east of Haveri mine, but is noticeably absent from the granodiorite intercepted in drill holes under Lake Kyrösjärvi to the west of the mine. It is unclear whether this occurrence represents a zonation in the granodioritic intrusion (the sample being taken close to the

contact with the amphibolite) or whether it is a separate intrusion altogether. The opaque phases consist primarily of magnetite, with minor pyrite and very occasionally some chalcopyrite.

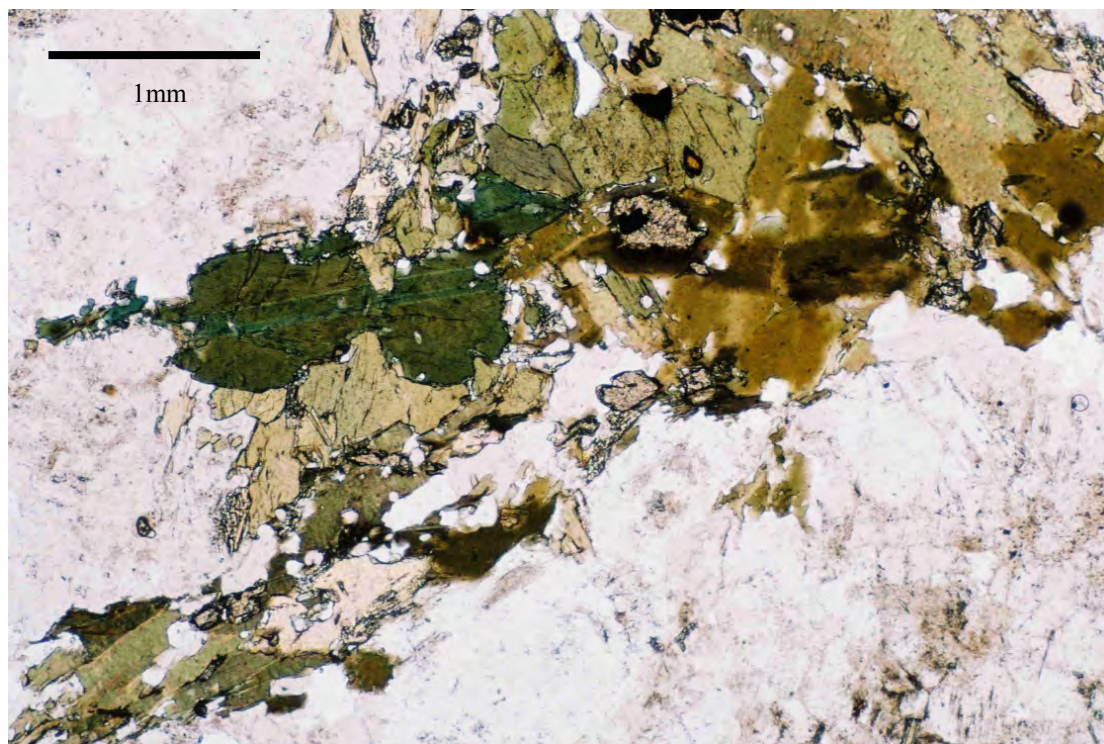


Figure 2.22: Photomicrograph of a typical sample of granodiorite. The high proportions of calcic amphibole and sphene, are indicative of a Ca-rich chemistry (Sample TS-19).

The presence of biotite and biotite-hornblende assemblages and the lack of magmatic muscovite, allow the granodiorite to be classified as an I-type granodiorite with an igneous source (Chappell and White, 1974). The presence of magnetite enables the granodiorite to be classified as a magnetite-series granodiorite (Ishihara, 1981).

The microcline granite is a pink-coloured rock dominated by large crystals of microcline feldspar, in a variably dark to light coloured matrix. Near Inkula, 1 km north of the Haveri mine, the matrix is typically dark with a higher proportion of hornblende and biotite, whereas 2 km ESE of Haveri the matrix is composed primarily of feldspar and quartz, with a much lower ferromagnesian content. This variation may be due to zoning within a single intrusion or alternatively may be due to separate intrusions.

In thin section the microcline granite (Table 2.4) is composed primarily of large K-feldspar phenocrysts showing distinctive crosshatched twinning and commonly a microperthitic texture. The K-feldspar shows considerable alteration, with many of the smaller grains (and the plagioclase) being almost completely altered to sericite. Only patches of relict multiple twinning can be recognised to identify the original mineral as plagioclase. The quartz typically has a polycrystalline nature with well sutured grain boundaries, but only very minor sub-grain development. The ferromagnesian minerals are generally

found together, occurring interstitially to the feldspars and quartz. Biotite and epidote commonly form a rim around amphiboles, and are also distributed in fractures through the amphibole, where they are associated with magnetite, sphene and quartz. In addition there is partial replacement of the amphibole by biotite. All of these textures suggest that biotite, epidote, magnetite and sphene formed after amphibole crystallisation. Zircon and apatite crystals are also present in the rock as accessory minerals.

	Granodiorite		Microcline Granite	
Mineral	Grain size	Content (%)	Grain size	Content (%)
Quartz	< 500 μm	25 – 30	< 2 mm	25 – 30
Plagioclase	< 3 mm	40 – 45	< 3 mm	5
K-feldspar	< 3 mm	10 – 15	< 10 mm	50 – 55
Amphibole	< 1mm	1 – 10	< 5 mm	5 – 10
Biotite	< 500 μm	10	< 500 μm	2 – 4
Muscovite	< 300 μm	< 1		
Epidote	< 300 μm	1 – 3	< 500 μm	1 – 4
Sphene	< 150 μm	1 – 2	< 200 μm	1
Carbonate	< 150 μm	< 1		
Magnetite	< 250 μm	1	< 200 μm	1
Pyrite	< 150 μm	< 1		
Apatite	< 80 μm	Trace	< 100 μm	< 1
Zircon	< 30 μm	Trace	< 100 μm	Trace

Table 2.4: Typical components of the granitoids near Haveri

The presence of hornblende, biotite, sphene and magnetite enables the microcline granite to be classified as an I-type granite under the classification scheme of Chappell and White (1974). In terms of the classification proposed by Isihara (1981), the microcline granite can be classified as a magnetite-series granite.

2.2.3.3 Mafic Dykes

Mafic dykes have been recognised at the Haveri mine site, at Peltosaari and also from diamond drill holes. From outcrop observations they are the youngest rock types seen at Haveri, cross-cutting, and hence post-dating, the supracrustal Haveri Formation, the quartz-feldspar porphyries (Figure 2.23), the carbonate skarn veins, and the granodiorite north of Haveri. The dykes display no evidence of metamorphic recrystallisation, and are commonly displaced by pyrite-filled shear zones (Figure 2.24). In addition, minor open folding, with axial planar quartz veinlets and fractures in the fold hinges can be recognised in the mafic dykes. This indicates that they were emplaced after peak metamorphism, and prior to the final phase of deformation.

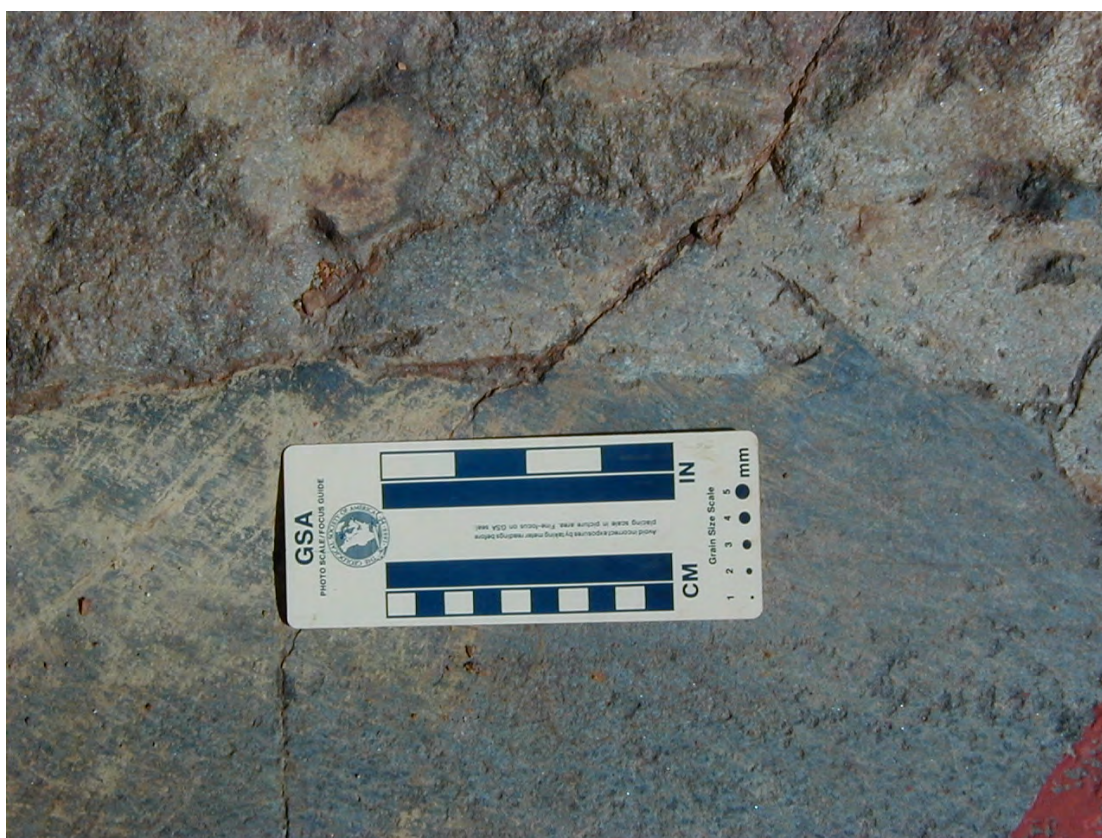


Figure 2.23: Mafic dyke (lower part of photograph) intruded into a quartz-feldspar porphyry. The chilled margin to the dyke is clearly visible. Close to the Haveri open pit.

The mafic dykes are less than a metre thick and are a dark grey-green in colour. They have well developed chilled margins, indicating emplacement into relatively cold country rocks (Figure 2.23). The cores of the dykes contain large (2 mm) dark green grains of amphibole, in a paler grey-green matrix. All of the dykes seen in outcrop trend approximately east – west, and dip fairly steeply to the north.



Figure 2.24: A sheared and displaced mafic dyke. The shear zone is filled with pyrite.

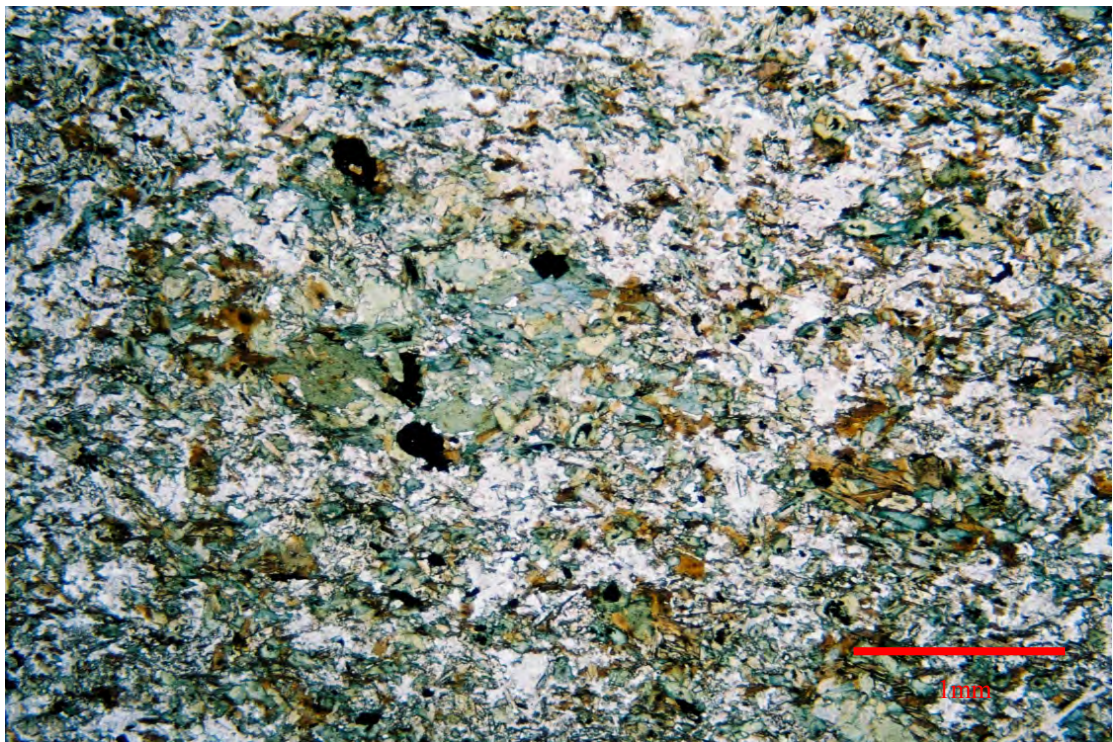


Figure 2.25: Photomicrograph of a typical mafic dyke (Sample TS-10).

In thin section (Figure 2.25) the dykes have a porphyritic texture with phenocrysts of amphibole and plagioclase, and a groundmass of amphibole, biotite, plagioclase and quartz with minor amounts of epidote and opaques (Table 2.5). Most of the green spots of amphibole visible in hand specimen are composed of clusters of hornblende and/or biotite with minor quartz. The remaining green amphibole spots are composed of large (0.5 – 2 mm) megacrysts of hornblende with numerous bleb-like quartz inclusions. This texture has been interpreted by Roberts et al. (2000) as a replacement texture after pyroxene. Some of the hornblende megacrysts have inclusions of ilmenite along the cleavages. This is also a replacement texture whereby the new mineral phase (hornblende) is unable to accommodate the titanium present in the original clinopyroxene, and thus exsolves ilmenite along its cleavage. The plagioclase laths, when present, are 0.5 – 2 mm in length and have numerous inclusions of biotite and occasionally hornblende. The groundmass is composed of randomly orientated fine-grained (approximately 100 μm) crystals of biotite, plagioclase, quartz and hornblende with minor epidote, opaques and rare apatite and sphene. The biotites occasionally form clusters up to 2 mm in size with coarser crystal sizes up to 500 μm in size. Pleochroic haloes are common in the biotites, indicating the presence of a radioactive mineral. The opaque phases are mainly ilmenite with minor amounts of pyrite and pyrrhotite. In the vicinity of narrow quartz and carbonate veinlets, plagioclase is extensively saussuritised to sericite. The paucity of sulphide or magnetite mineralisation, and the absence of amphibolitisation and calcic-skarn alteration indicates that the mafic dykes were emplaced post-mineralisation.

Mineral	Grain size	Content (%)	Texture
Hornblende	0.5 – 2 mm	10 – 15	Megacryst
Biotite	< 500 μm	0 – 10	Megacryst
Plagioclase	0.5 – 2 mm	0 – 15	Phenocryst
Quartz	< 100 μm	15	Groundmass
Hornblende	< 150 μm	10 – 20	Groundmass
Plagioclase	< 150 μm	20 – 25	Groundmass
Biotite	< 300 μm	15 – 20	Groundmass
Epidote	< 150 μm	0 – 5	Groundmass
Ilmenite	< 200 μm	1 – 2	Replacement
Apatite	< 200 μm	< 1	
Sphene	< 100 μm	Trace	
Pyrrhotite	< 100 μm	< 1	
Pyrite	< 200 μm	< 1	

Table 2.5: Typical components of the late-stage mafic dykes

2.3 ALTERATION TYPES AND VEINING

A number of different alteration assemblages can be recognised at Haveri. These include the prolific amphibolitisation, the calcic-skarn assemblages, as well as post-metamorphic retrograde alteration and chlorite veining, and minor biotite alteration. Quartz veins and veinlets are a prominent feature of the Haveri geology. Minor carbonate, epidote and potassic feldspar veins and veinlets are also present.

2.3.1 Amphibolitisation

Amphibolitisation is the most widespread of the alteration types seen at Haveri. This type of alteration is characterised by fine, dark-green amphibole veinlets which, with increasing degree of alteration, develop into an increasingly dense network of veins, finally culminating in total replacement of the host rock as an intense dark green amphibole skarn. Fine amphibole veinlets commonly pass laterally into and out of quartz $\pm$ plagioclase veinlets (Figure 2.26). The initial texture of the host rock appears to have some degree of control on the development of the amphibole skarn. In the meta-lavas, vesicles are commonly replaced by amphibole, forming a spotted amphibolite (Figure 2.27). In the more massive metavolcanic rocks and volcanogenic tuffites, network veining of amphibole is the dominant texture, whereas in the metasedimentary rocks, amphibolitisation is a minor alteration phase, characterised by rare, localised veins of amphibole (Figure 2.28), and a weak replacement of biotite by amphibole. Amphibole alteration of the porphyries forms the darker looking “uralite porphyries”.

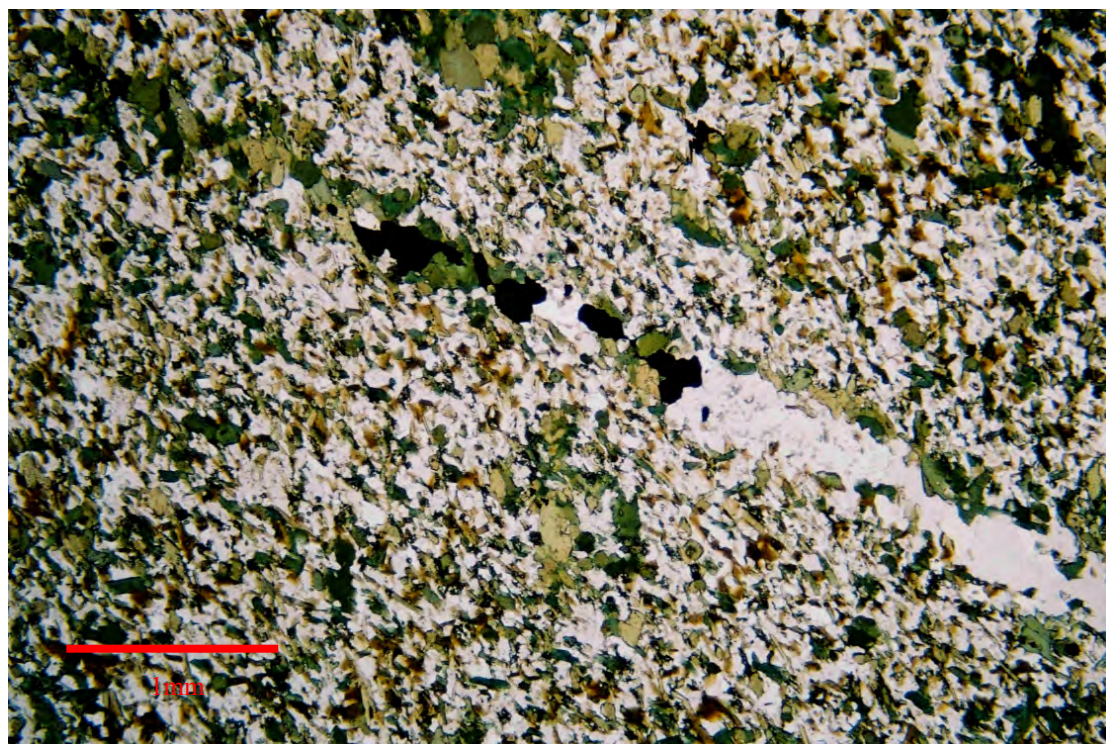


Figure 2.26: *Amphibole-quartz veinlet with pyrrhotite, cutting a biotitic amphibolite (metatuff) (Sample H5-139).*

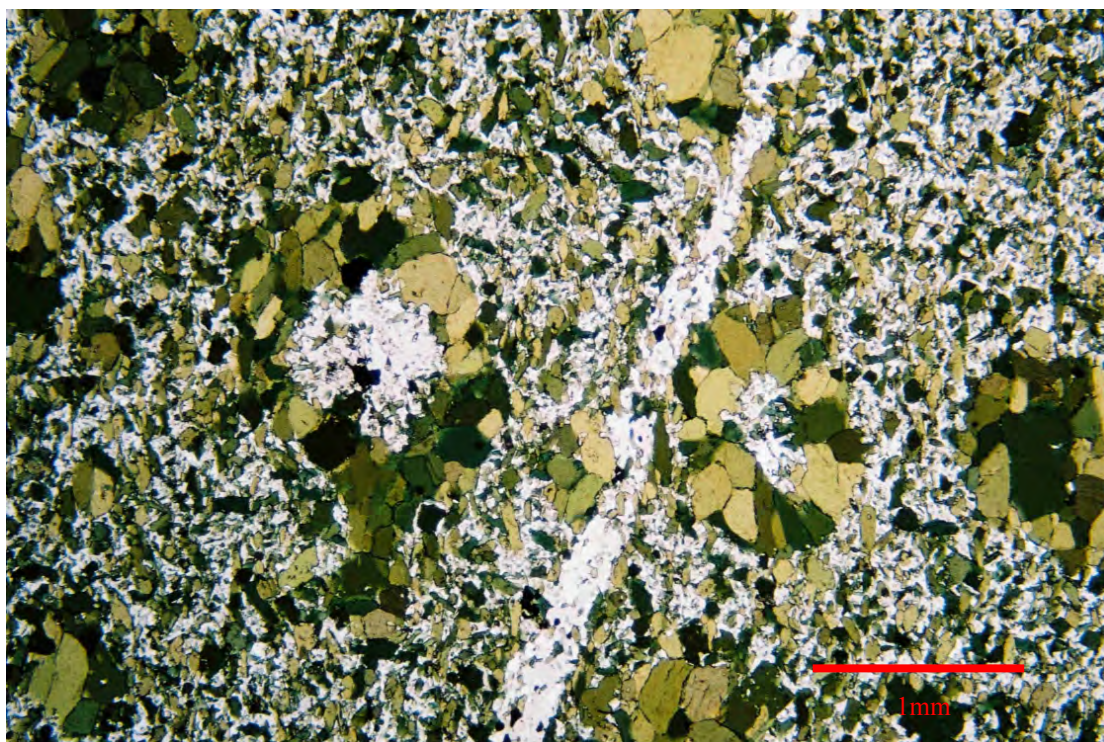


Figure 2.27: Spotted amphibolite cut by a quartz veinlet. The cores of the amphibole “megacrysts” are composed mainly of plagioclase, with minor epidote, sphene and opaques (magnetite, pyrrhotite and chalcopyrite) (Sample H5-164a).

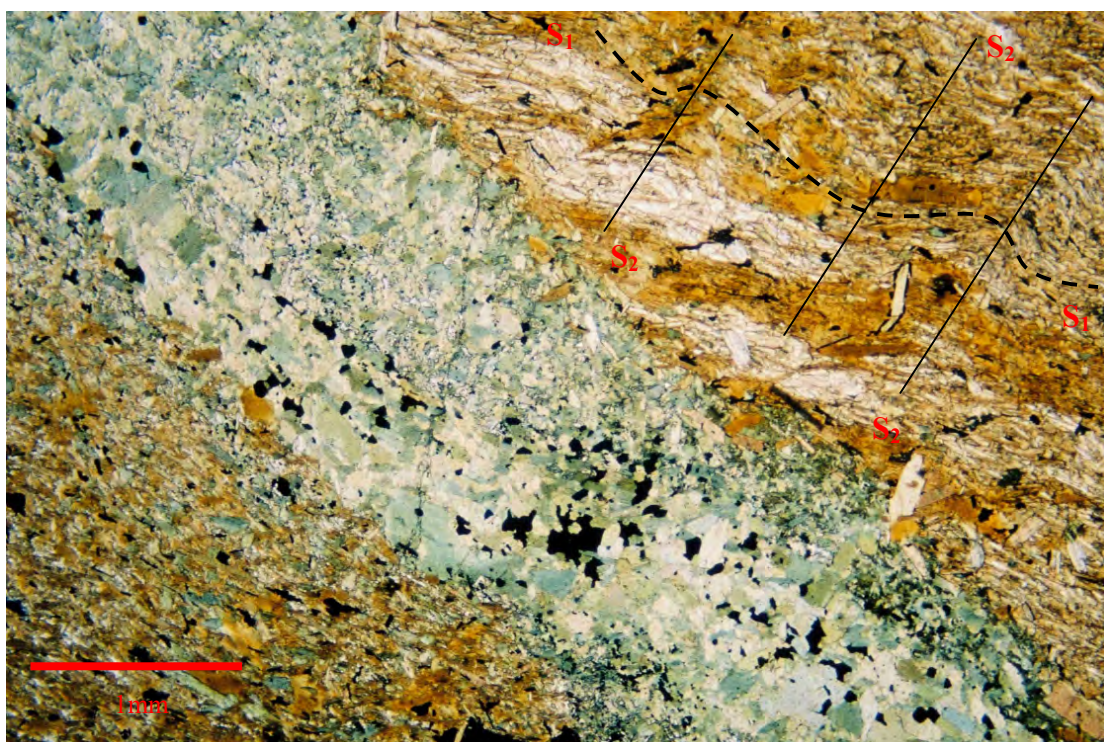


Figure 2.28: Amphibole veinlet cutting a biotite-rich metasedimentary rock. To the right of the vein, flasers of epidote are present, as well as a weak crenulation cleavage developed in the biotites. (Sample H5-240).

In a number of the amphibole veins, there is a core of pyroxene, surrounded by a rim of dark green amphibole. However, more commonly, the veins (and the intense amphibole alteration) are represented by amphiboles of different composition (Section 3.2.2.1). The outermost phase (Figure 2.29), which forms a narrow border with the host rock, is composed of predominately granoblastic, coarse-grained amphibole, responsible for the dark green appearance in hand specimen. This has been termed “rim-type” amphibole, and has the same appearance as the amphibole in the narrow veinlets, and in the pervasive dark-green amphibole. The core of the amphibole vein is of slightly finer grained amphibole that is intergrown with bleb-like quartz. This has been termed the “core-type” amphibole. This texture has been suggested by Roberts et al. (2000) to indicate the replacement of pyroxene by amphibole. Within this “core-type” amphibole, are small patches of washed-out, pale-green amphiboles, with actinolitic compositions (Section 3.2.2.1). These patches are interpreted as later, lower-temperature, retrograde amphiboles, that have replaced the initial hydrothermal pyroxenes, as the hydrothermal fluids further cooled.

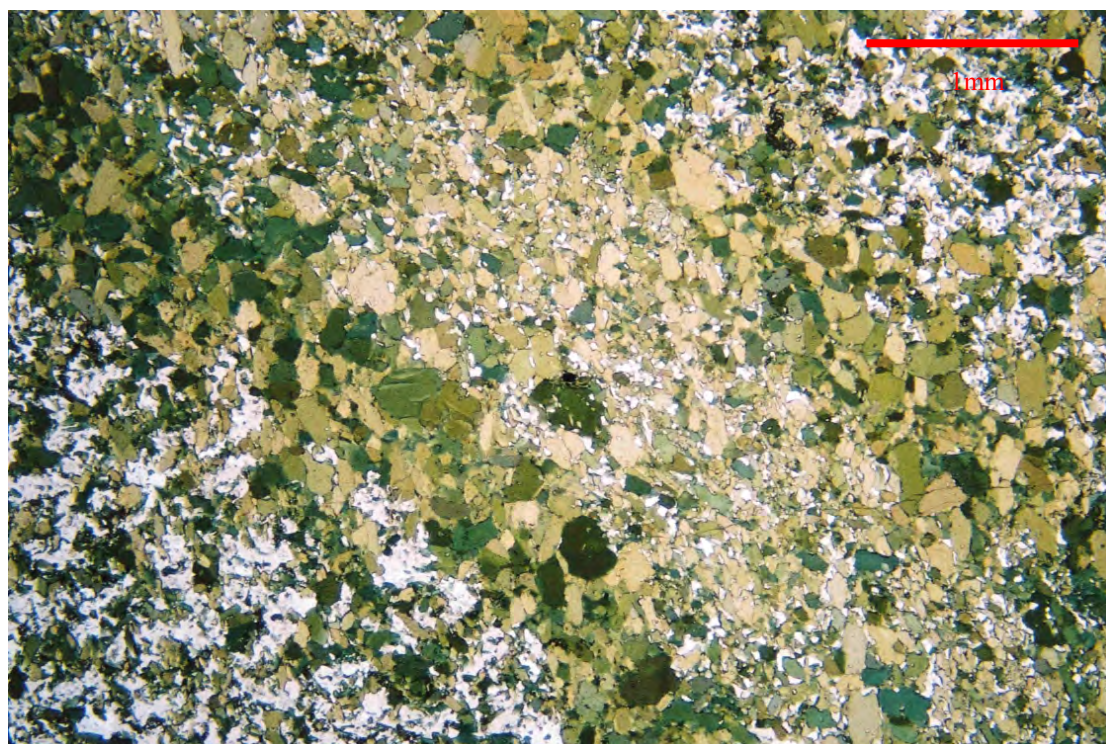


Figure 2.29: Amphibole vein cutting amphibolite. Clearly visible are the rim and the core of the amphibole veins. Also visible is the bleb-like quartz in the vein core (Sample H5-182).

In the metavolcanic rocks, the amphibole skarn is occasionally accompanied by a concomitant de-amphibolitisation / albitization of the metamorphic amphibolite host rock. In hand specimen, the metavolcanic rock takes on a brittle brecciated look, with pale grey to pink, angular clasts, of de-amphibolitised host rock (Figure 2.30). The matrix to these brittle clasts is the dark green amphibole veins described above. In thin section (Figures 2.31 – 2.32), the pale clasts consist of fine-grained plagioclase with only a low amphibole content. The pink colour is caused by a high sphene content in the clasts. The sphene commonly occurs as fine strings or veinlets cutting the clasts.

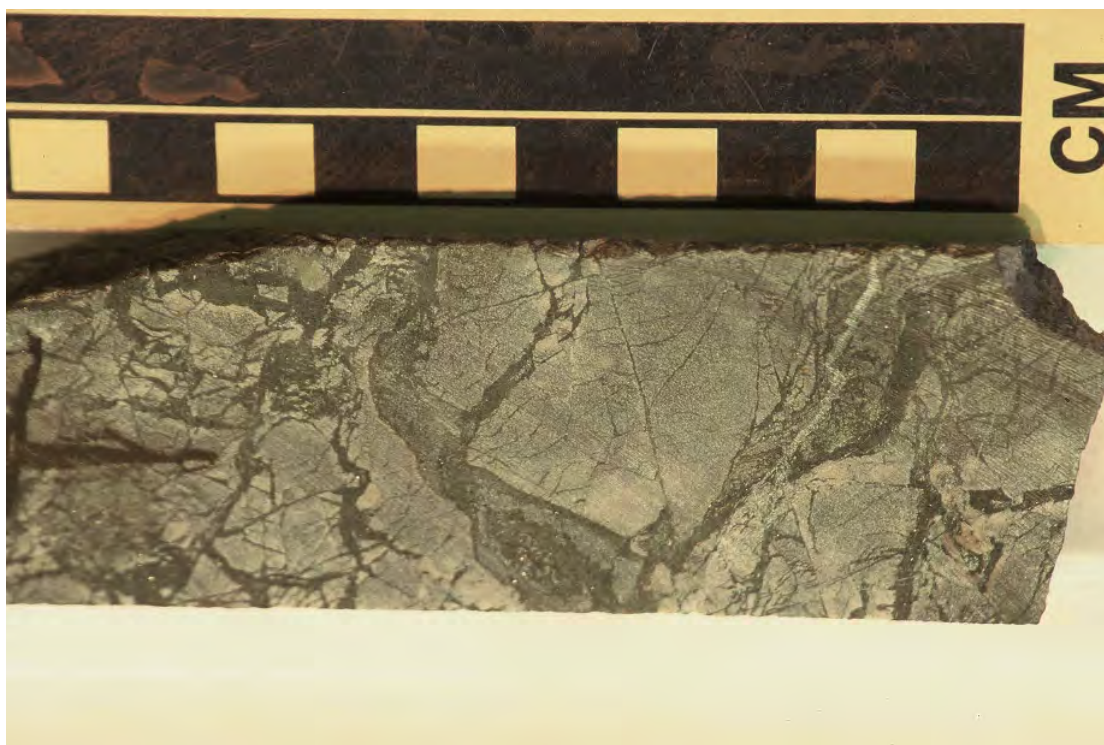


Figure 2.30: *Amphibole brittle-brecciation of mafic metavolcanic rocks and deamphibolitisation and/or albitisation of the metavolcanic clasts (Sample R8-241).*

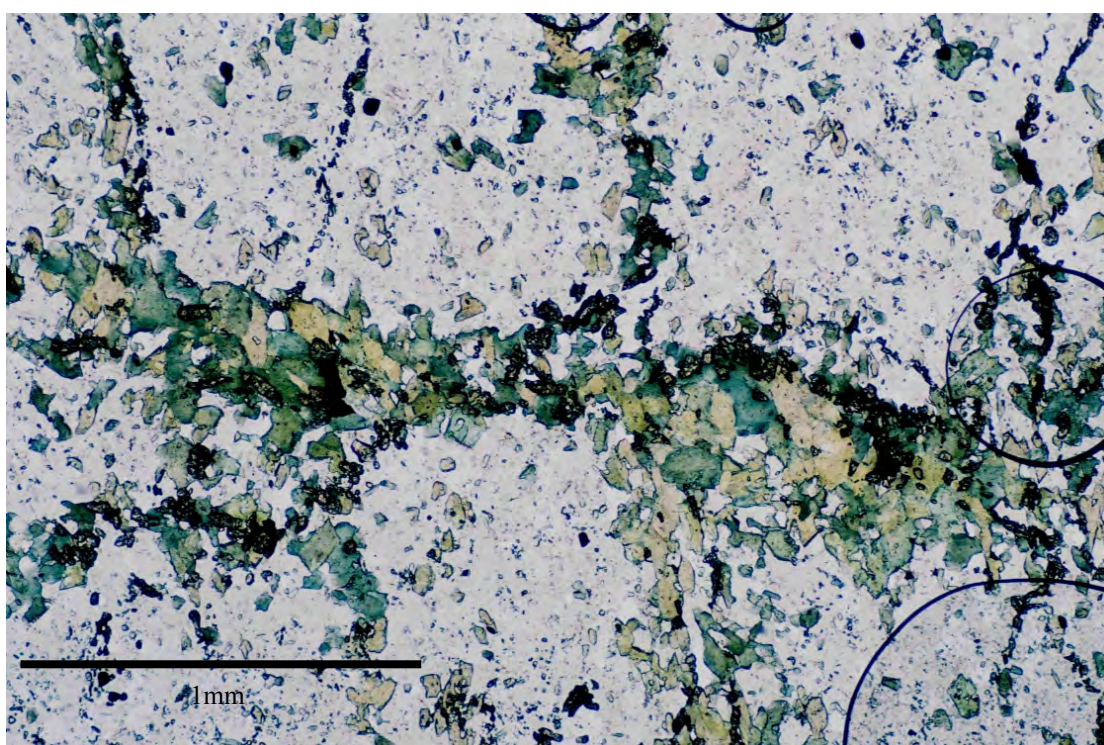


Figure 2.31: *Deamphibolitised (and albitised) mafic metavolcanic rock. Veins of amphibole (green) and sphene (dark and high relief) brecciate the pale grey clasts of “leached” metavolcanic material (Sample R8-225f).*

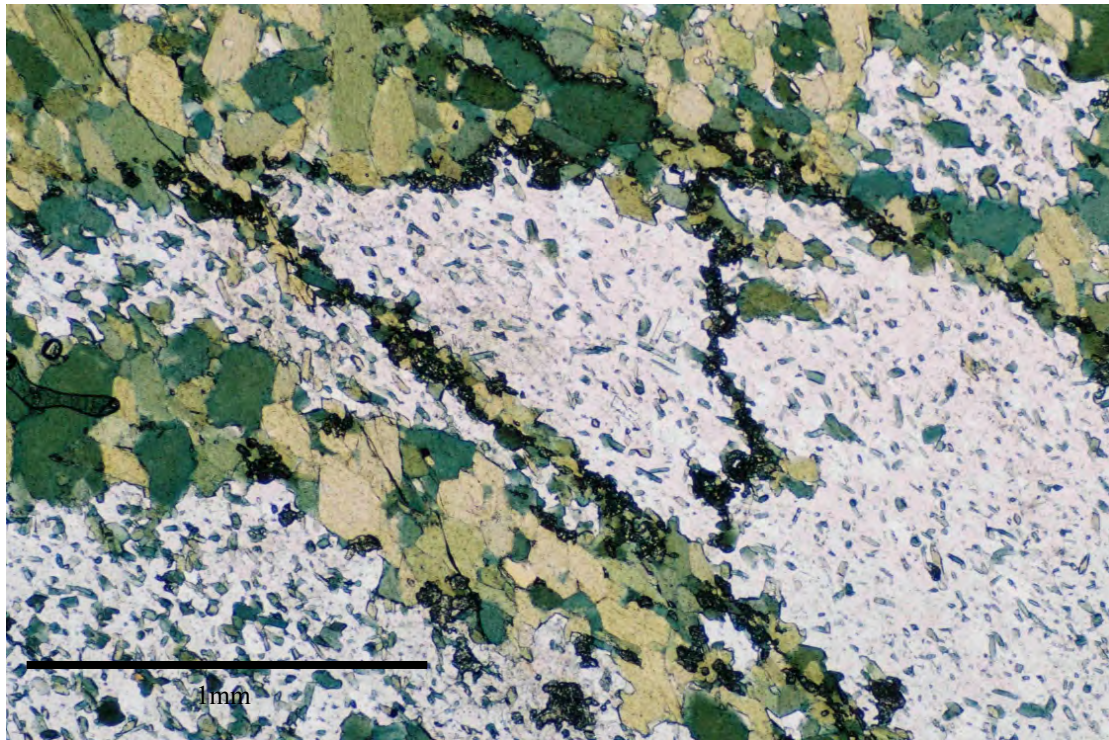


Figure 2.32: Deamphibolitisation and albitisation of the mafic metavolcanic rocks, with brecciation of clasts by an amphibole-sphene matrix (Sample H5-130a).

This texture is interpreted as a hydrothermal brecciation, with the preferential removal of Ca, Fe and Mg from the rock and its integration into the amphibole veining. The sphene and plagioclase have been left behind causing an apparent enrichment in Si and Ti. Any remaining Ca is fixed in sphene. In addition, plagioclase compositions are more sodic (Section 3.2.2.4), suggesting partial albitization. The resultant brittle, brecciated texture is important in terms of gold mineralisation, which will be discussed later (Chapter 5).

Amphibolitisation is closely associated with both sulphide and oxide mineralisation, and amphibolitised rock is also the most important host for gold mineralisation. Pyrrhotite is the dominant sulphide mineral with lesser chalcopyrite and minor pyrite, and occurs as massive zones, veins and disseminations. Magnetite is also a widespread and important hydrothermal phase, forming intergrowths with pyrrhotite and chalcopyrite as well as forming patches of massive magnetite. The many styles of sulphide and oxide mineralisation are discussed in greater detail in Section 2.4. A paragenetic sequence of events for the amphibolitisation is depicted in Figure 2.33.

The timing of the amphibolitisation in relation to metamorphism is clearly important. One possibility is that the alteration occurred pre-metamorphism as sub-seafloor spilitisation. In such a scenario, the chlorite-dominated alteration (spilitisation) was metamorphosed into the amphibole-dominated assemblage seen today. The alternative model is that the amphibolitisation is syn- to post-metamorphic. The degree of amphibolitisation of the rock is variable ranging from fine amphibole veinlets, to an intense and pervasive total replacement of the entire rock. In areas where the degree of alteration is

fairly minor and the alteration is represented by amphibole veinlets, it is possible to see the amphibole veins cutting individual metamorphic amphiboles of the host rock, this clearly indicates a post-metamorphic genesis for the amphibolitisation. This is further supported by the progressive amphibolitisation of the quartz-feldspar porphyries, resulting at its most extreme in the “uralite porphyries”. The amphibolitisation must therefore be post-porphyry emplacement. The weight of evidence thus clearly supports a syn- to post-metamorphic, metasomatic origin for the amphibolitisation.

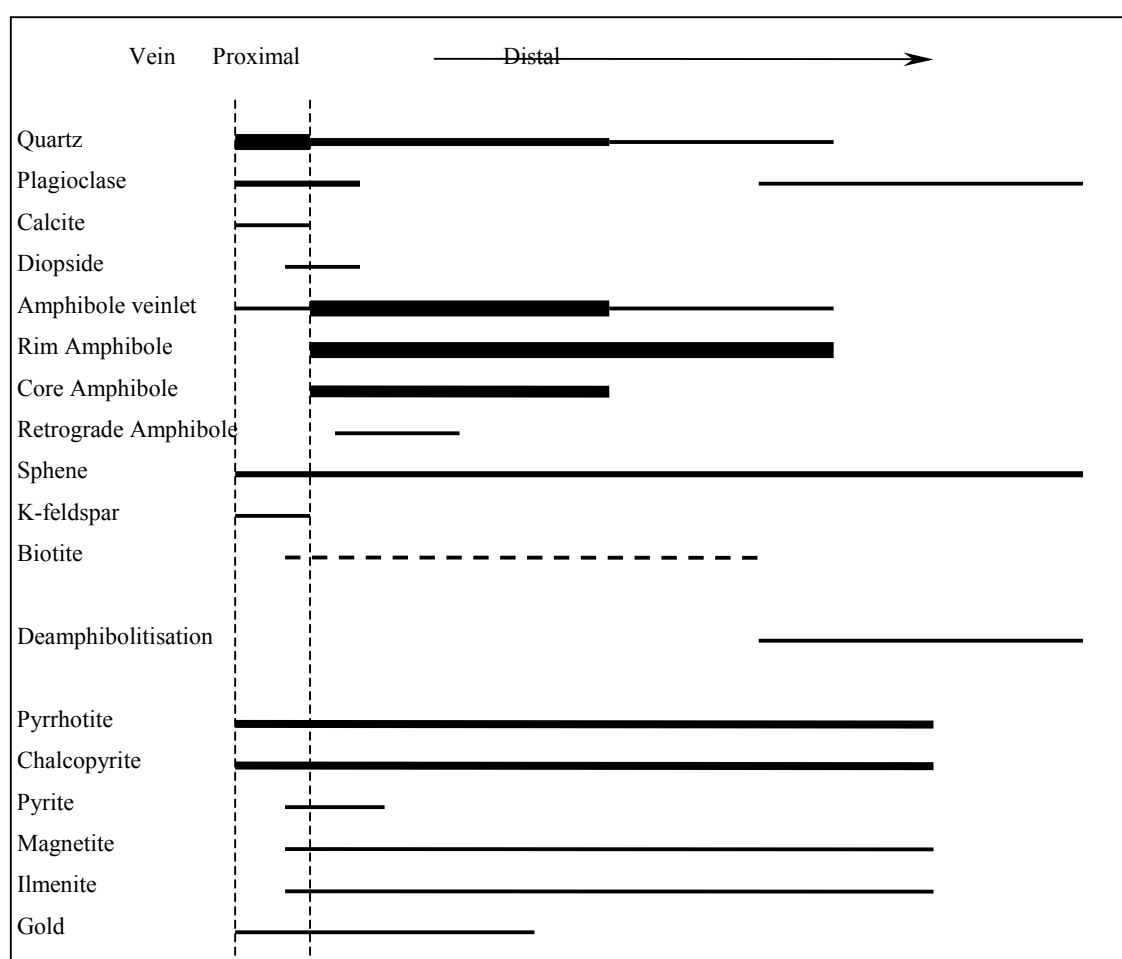


Figure 2.33: Schematic summary of the paragenetic alteration sequence associated with the amphibolitisation. Proximity to the core of the alteration system increases to the left.

2.3.2 Calcic-skarn alteration

The calcic-skarn alteration affects both the supracrustal suite and the porphyries, but has not been observed in either the granitoids or the mafic dykes. The calcic-skarn is most commonly expressed as discrete, spatially confined, zoned mineral assemblages centred on calcite veins. The mineral zonation around the veins is commonly complicated by a retrograde overprint, representing the final phases of skarn formation, whereby the hydrothermal system has contracted and become cooler, resulting in the destruction and replacement of the earlier skarn minerals by lower temperature, hydrous minerals such as amphibole and epidote.

In the calcic-skarn, the proximal mineral assemblages are commonly zoned around carbonate veins and veinlets. The best examples of zonation are around fine, millimetre-scale, carbonate-quartz veinlets. These commonly have, from proximal to distal: a core zone of garnet $\pm$ diopside; a diopside-epidote zone; an epidote-amphibole zone; and a distal amphibole assemblage (Figure 2.34). In plagioclase-rich rocks, the distal zone is surrounded by a halo of saussuritisation. The zonation described above is the dominant mineral zonation; however, variations are common with a number of other minerals commonly being present, including sphene, and more rarely scapolite (Figure 2.35) and serpentine/talc, as a late replacement. As mentioned previously, this assemblage commonly displays a late retrograde overprint whereby distal mineral phases and/or lower-temperature hydrous mineral phases overprint the more proximal minerals. Examples of this include epidote and amphibole replacing garnet and pyroxene.



Figure 2.34: A zoned calcic-skarn vein in amphibolite. A fragment of a discontinuous quartz-carbonate vein is surrounded by a rough alteration halo with a core zone of garnet, passing into a diopside-epidote zone and finally into an amphibole zone. Large blebs of pyrrhotite and chalcopyrite are also present (Sample R8-169).

Meinert (1992) has summarised the formation of skarns, and identifies three main stages related to and correlated with the late magmatic and hydrothermal events of pluton emplacement. Thus the formation of skarns involves stages of (1) isochemical metamorphism, with hornfelsing of non-carbonate units, and the development of reaction skarns in carbonate rocks. This is followed by (2) a metasomatic stage with development of exo- and endo-skarns. Towards the final phases of cooling, (3) a retrograde stage occurs, during which the early skarn assemblages are partially destroyed and replaced by hydrous minerals.

The calcic-skarn alteration has a number of opaque phases associated with it, including pyrrhotite, pyrite, chalcopyrite, magnetite and less commonly sphalerite. The sphalerite is associated with an intergrowth of anhedral pyrrhotite, chalcopyrite and magnetite with inclusions of subhedral pyrite. In addition there is a late phase of fracture-filling pyrite. Results from the gold exploration programme indicated that the calcic-skarn veins are unmineralised with respect to gold. Microscopic examinations have also failed to identify gold grains within the calcic-skarn. A simplified paragenetic sequence for the calcic-skarn alteration is shown in Figure 2.36.

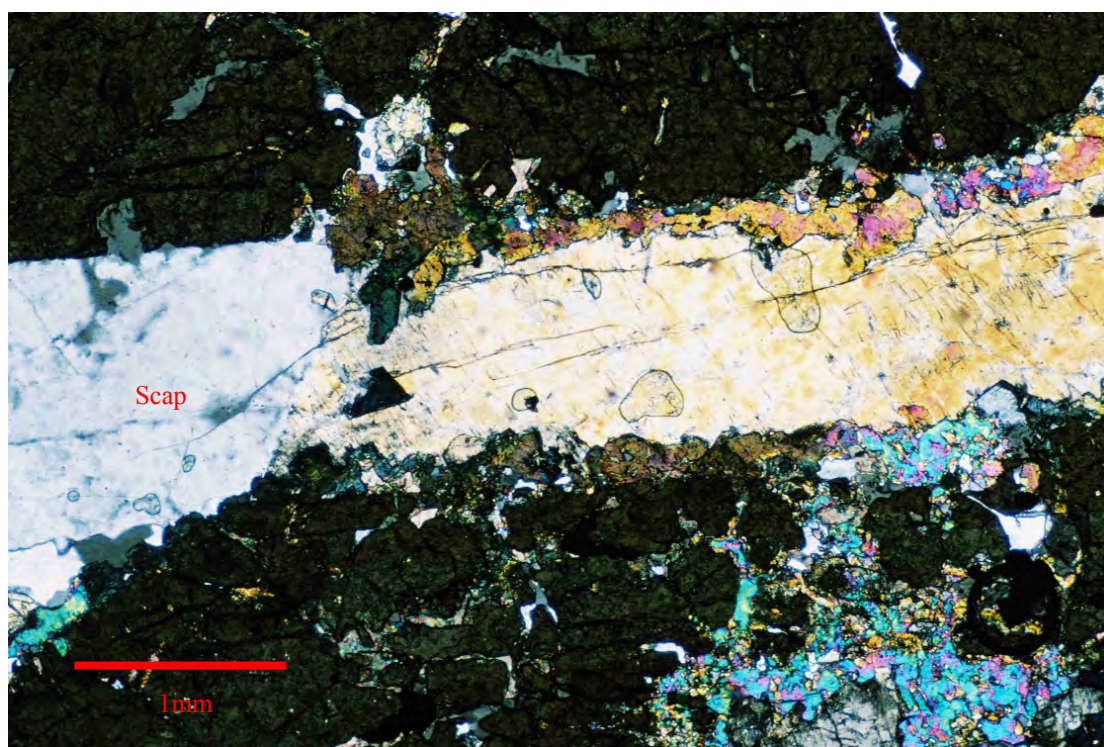


Figure 2.35: Calcic-skarn vein of scapolite (scap) and quartz (slightly thick section), surrounded by garnet and retrograde epidote. Photographed using offset polarisers (Sample R8-169a).

The relationship between the calcic-skarn event and the amphibolitisation is not totally clear. No amphibole veins cut the calcic-skarn/carbonate zones, indicating at least that the amphibolitisation is not later than the calcic-skarn. However, no observations have been made of the calcic-skarn unambiguously cutting amphibole veinlets. There are therefore several possibilities remaining. Either

the calcic-skarn is a later, separate event, to the amphibolitisation, or, alternatively, the calcic-skarn is contemporaneous with the amphibolitisation, and that it is simply a proximal phase of the same alteration event. It is also possible that the calcic-skarn, although being later than the amphibolitisation is in fact simply a later, more evolved phase, of the same event.

Despite the lack of an unambiguous cross-cutting relationship, evidence tends to support the argument that the two events are separate, and that the calcic-skarn is indeed a later event. One of the main differences is that the amphibole alteration is always devoid of epidote (Section 2.3.1), whereas epidote is ubiquitous to the calcic-skarn (Figure 2.36). Another difference is that the amphibolitisation has an alteration halo of deamphibolitisation, with a consequent enrichment in plagioclase content, whereas the calcic-skarn has an alteration halo with saussuritisation, and the presence of hydrous lower-temperature minerals such as zoisites. Finally, the amphibole alteration is commonly associated with gold, whereas the calcic-skarn is invariably devoid of gold mineralisation. Therefore, on balance, it appears that the calcic-skarn is a later hydrothermal event, although it may also represent a lower-temperature evolution of the main amphibolitisation event. This relationship is discussed later in Section 5.2.2.

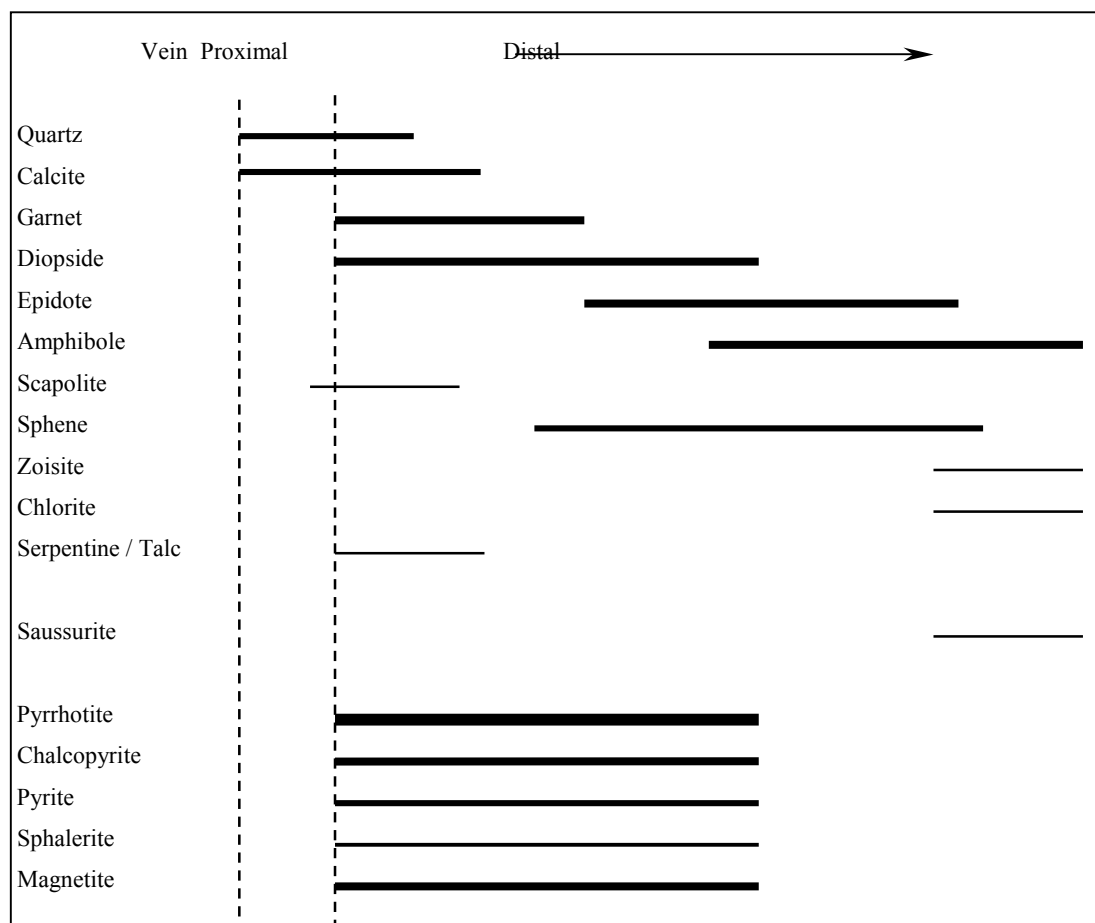


Figure 2.36: Schematic summary of the paragenetic alteration sequence associated with the calcic-skarn alteration. Proximity to the core of the alteration system increases to the left.

2.3.3 Carbonate Veining

Carbonate veins up to 1 metre thick have been observed in outcrop, in the old mine and in boreholes. In the immediate vicinity of the old mine there is a zone with several carbonate bands that can be traced at least to the old 100 metre level. These bands have previously been interpreted as primary sedimentary limestone bands (Kähkönen and Nironen, 1994; Nironen, 1994). However, in the old open pit these carbonate bands have been observed cross-cutting a quartz-feldspar porphyry, and thus must be of hydrothermal origin. This is also supported by carbon isotope analyses (Section 3.3.7). To the west of the Haveri open pit, one of the carbonate veins is cross-cut by a narrow mafic dyke, indicating that they were emplaced post-porphyry and pre-mafic dyke.

The veins themselves are composed primarily of calcite, barite and chlorite. These thicker, calcite-barite dominated veins generally do not display the proximal minerals seen associated with the calcic-skarn, such as garnet and scapolite, but rather are characterised by a greater degree of late-stage retrograde alteration, with a pale green halo of epidote, amphibole and chlorite. However, diopside is also commonly present, and diamond drilling has intersected garnet-rich portions within the carbonate veins. These are indicative of higher temperature conditions. The main opaque phase is pyrite with lesser amounts of pyrrhotite and sphalerite.

It is possible that the carbonate veins are genetically related to the calcic-skarn alteration seen around the finer carbonate veinlets. However, there are several major differences. Firstly, the calcic-skarn is centred on fine carbonate veinlets, with no barite in the veinlets, whereas there is considerable barite in the carbonate veins. Secondly, the calc-silicate mineralogy which is so dominant in the calcic-skarn, is but a minor component of the calcite-barite veins. Another point of difference is that whereas the calcic-skarn contains considerable pyrrhotite, the carbonate veins contain mainly pyrite. Finally, carbonate isotope studies (Section 3.3.7) show broadly similar isotopic compositions for carbonates from both the carbonate-barite veins and the fine carbonate veins associated with the calcic-skarn, indicating that both are of hydrothermal (or magmatic origin). However, the carbonates from the thick carbonate veins appear to have slightly lighter $\delta^{13}\text{C}$ values than the calcic-skarn carbonates. Therefore, it is likely that the carbonate veins are either genetically unrelated to the calcic-skarn, or that they represent a later phase of the same event, perhaps being formed at lower temperatures, under more oxidising conditions (hence barite and pyrite).

2.3.4 Quartz Veining

Quartz veining is ubiquitous to all the supracrustal lithologies and ranges in thickness from the millimetre scale to tens of centimetres. In outcrop, most of the veins encountered are pale grey. The quartz veins and veinlets display a number of structural styles including: intensely folded veins (Figure 2.37), boudinaged veins (Figure 2.38), and laminated veins with slivers of amphibolitised wall rocks

(Figure 2.39). These structural styles are indicative of ductile deformation (Eilu et al., 1997, 1999). In addition, where the host rock is strongly foliated and amphibolitised, parallel to sub-parallel quartz-rich bands are present (Figure 2.39), which have been interpreted as recrystallised, strongly deformed quartz veins in a banded brittle-ductile shear zone (Eilu et al., 1997, 1999). In thin section the veins and veinlets are quartz-dominated, with numerous inclusions of amphibole and sulphides, and occasionally minor amounts of calcite and/or plagioclase.

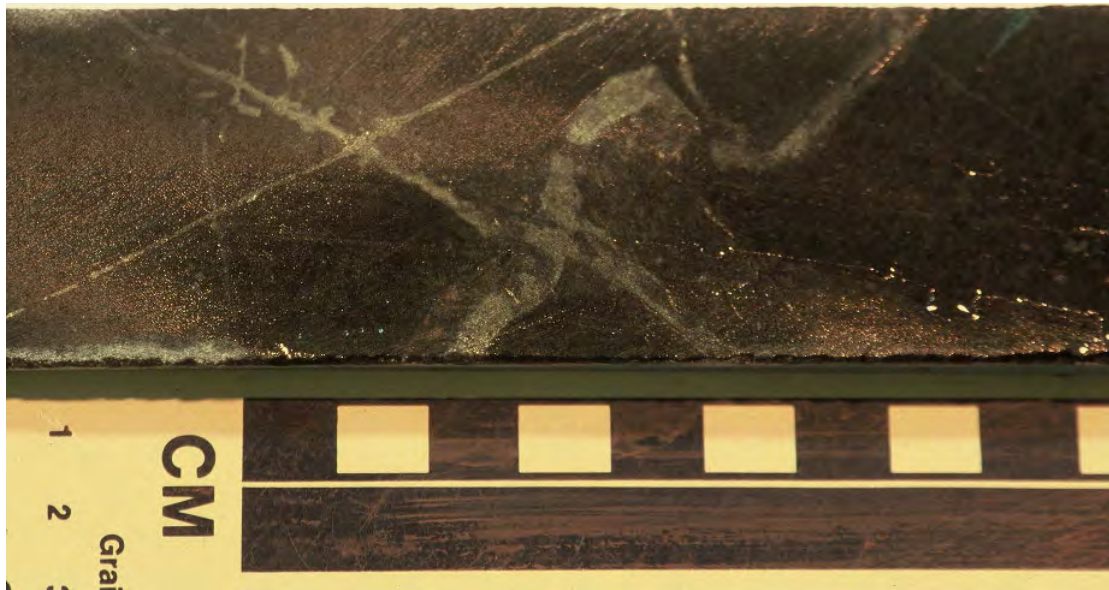


Figure 2.37: *Folded quartz vein, and evidence of several phases of quartz veining (Sample R8-188).*



Figure 2.38: *Boudinaged quartz vein, with a rim of amphibolitisation. Sample from underground in the Haveri mine.*

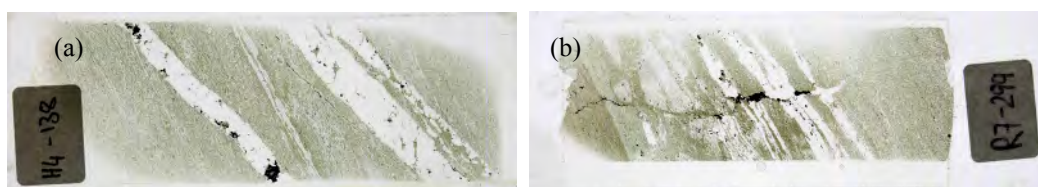


Figure 2.39: (a) Quartz veins subparallel to foliation, with slivers of amphibolitised wall rock. (Sample H4-138). (b) Strongly foliated zone of quartz-rich bands in amphibolite. (Sample R7-299). The main foliated fabric is cross-cut by a vein of amphibole-quartz-pyrrhotite.

The relative timing of vein formation to the host rocks and other geological events is highly variable. The majority of the quartz veinlets are either contemporaneous with, or slightly later than the amphibole veinlets (Figure 2.40), although amphibole veinlets do also cut quartz veinlets which are typically highly deformed and folded, and thus are likely to be emplaced syn-D₁. This indicates that quartz veining has continued at Haveri since pre-F₁ right through until at least D₂ deformation and possibly later. The majority of the quartz veinlets are pre-F₂ and probably syn-S₁, as they are parallel to the S₁ biotite schistosity, and have been folded by the F₂ crenulations in the biotite-rich rock (Figure 2.55 – 2.56).

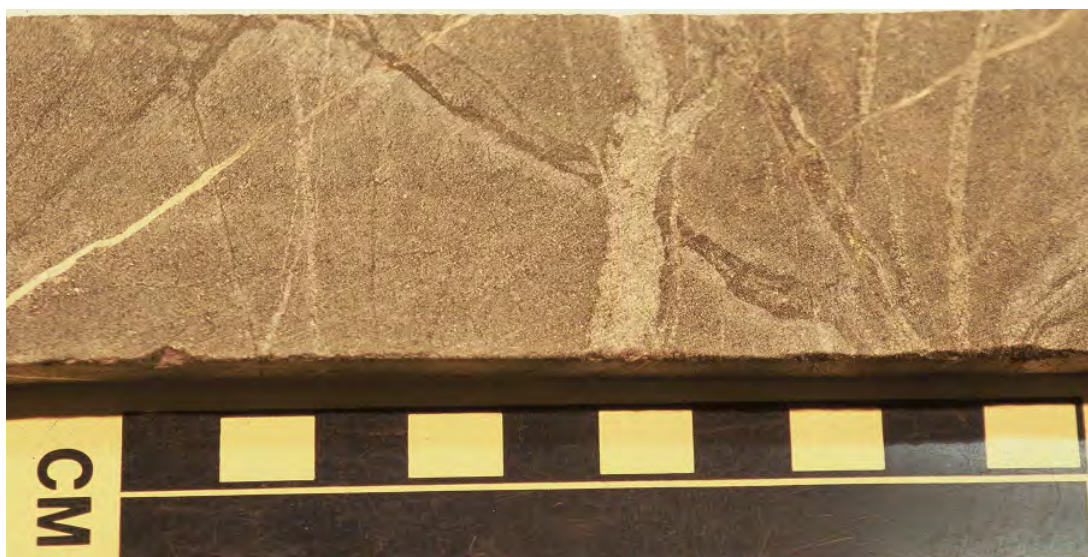


Figure 2.40: Quartz veinlet cutting both the clasts and the matrix of a brecciated volcanic, showing signs of minor deamphibolitisation along clast margins. (Sample R7-297).

The quartz veinlets are considered to be one of the three most important hosts for gold mineralisation at Haveri in addition to the intense amphibolitisation and the massive sulphides (Section 2.4.2). Some of the more spectacular samples containing visible gold have been recovered from the quartz veinlets.

Milky white quartz veins have occasionally been encountered at Haveri. These are generally pure quartz with few mineral inclusions and no sulphides. Under the microscope they show no evidence of sub-grain development and thus are relatively undeformed, and must be late in the overall geological history at Haveri.

2.3.5 Retrograde Alteration

Evidence of retrograde alteration is present in the form of chlorite veins and veinlets cutting the rock, and the local chloritisation of the host rock. Chlorite veinlets are demonstrably a late event, cross-cutting the calcic-skarn assemblages, the amphibolitisation, as well as the quartz veinlets. The chlorite veinlets commonly contain a high proportion of opaque minerals (e.g. Sample P13-28). Chloritisation of the metasedimentary rocks is typically observed where metamorphic biotite has been pseudomorphed by chlorite, with titanium-rich opaques (most likely needles of rutile) being exsolved along the cleavage planes. Also associated with the chloritisation, is the development of zoisite/clinozoisite, and the breakdown of plagioclase.

2.3.6 Biotite Alteration

Biotitisation is another alteration texture that has been observed in thin section. This is seen as the partial replacement of amphiboles by biotite (Figure 2.41), and, when present, is most noticeable within the metavolcanic rocks and the quartz-feldspar porphyry. The extent of this alteration within other lithologies is unclear, due to the presence of biotite in the host rock assemblages. The timing of the biotite alteration is also unclear. However, it is likely to be contemporaneous with, or later than, the amphibolitisation and peak metamorphism, as the biotite is observed replacing the amphiboles, although it has not actually been conclusively observed replacing amphibole associated with amphibolitisation. Amphibole replacement of biotite has also not been observed. It is possible that the biotite alteration is more extensive than what has been observed, especially within the meta-tuffites, where biotite and amphibole are part of the same mineral assemblage.

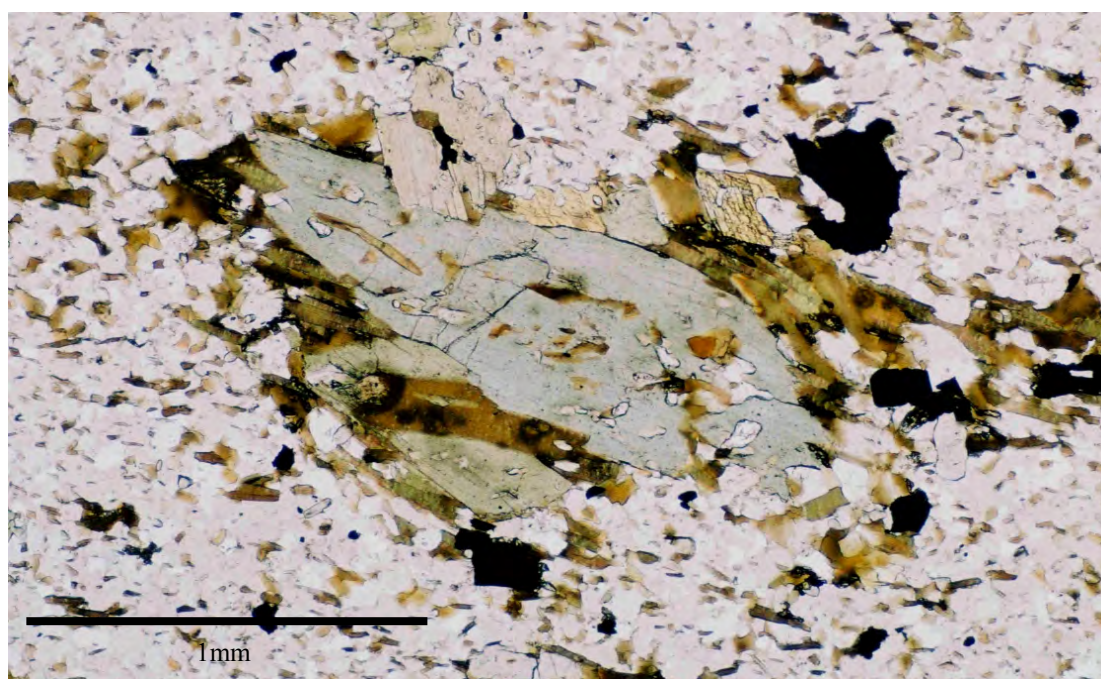


Figure 2.41: Biotite replacing amphibole. (Sample H5-70).

2.4 MINERALISATION

Opaque phases are ubiquitous in the rocks at Haveri, and it is possible to distinguish a number of different textures and assemblages related to a range of geological processes. Results from the recent exploration programme, supplemented by microscopic examinations during this study, show that the gold mineralisation is less extensive than the sulphide mineralisation, and is constrained to a few main ore types.

The main ore minerals encountered at Haveri are the sulphides pyrrhotite, chalcopyrite and pyrite, and the oxides magnetite and ilmenite. In addition there are minor amounts of molybdenite, sphalerite and gold. Also reported from Haveri, although not encountered in the present study, are galena (Paarma, 1947; Vaasjoki and Huhma, 1987), arsenopyrite, cobaltite, glaucodot and scheelite, as well as trace occurrences of tellurides (Paarma, 1947). The ore minerals have been introduced at various stages during the geological evolution of the Haveri deposit, and consequently a number of models have been proposed, or may be utilised in explaining the gold mineralisation at Haveri. These models will be examined in depth at a later stage within this study.

2.4.1 Sulphide and Oxide Mineralisation

The sulphides, oxides and gold occur in a number of different textural styles and mineral assemblages, many of which contain insignificant quantities of gold. However, all of these styles are important in constructing a model for the geological evolution of the Haveri deposit. Many of these styles may overlap both spatially and/or in terms of their genesis. The main styles of mineralisation observed at Haveri are:

- I. Massive sulphide / “cataclastite” / network sulphide and oxide ore
- II. Disseminated sulphides and oxides
- III. Quartz-vein related sulphides
- IV. Calcic-skarn related sulphides and oxides

2.4.1.1 Massive sulphide / “cataclastite” / network sulphide and oxide ore

The massive sulphide type ore is the principal ore type at Haveri in terms of gold mineralisation. The recent exploration programme indicates that this ore type occurs at various scales ranging in size from centimetre thick veins, to semi-continuous zones up to 10 metres thick and up to 50 metres long.

The massive sulphide type ore consists of patches and “veins” of massive sulphide, within a network of sulphide stringers. The zones of massive sulphide commonly display breccia textures, with angular, fractured clasts of biotite-rich rock (Figure 2.43), amphibolitised host rock, or rounded quartz-vein material (Figure 2.45) in a highly deformed pyrrhotite-dominated matrix, with monomineralic bands of pyrrhotite, cobaltite and chalcopyrite. This cataclastic texture with ductile flow of the sulphides has been termed by Vokes (1969) as “durchbewegung” and is indicative of remobilisation of the sulphides with deformation. Within the massive-sulphide ore type are patches of massive magnetite, with sulphide inclusions of predominately chalcopyrite, and minor pyrrhotite. Magnetite is found both as inclusions in pyrrhotite (and chalcopyrite), and also containing inclusions of pyrrhotite, thus indicating contemporaneous formation, perhaps during deformation. However, textural evidence indicating a slightly earlier formation of the magnetite is more common, including sulphide veinlets cutting the massive magnetite (Figure 2.44). It is therefore likely that magnetite precipitation commenced prior to sulphide precipitation. The pyrrhotite-dominated massive sulphide commonly contains euhedral to subhedral grains of pyrite (Figure 2.46), as well as a more anhedral looking mineral identified as molybdenite.



Figure 2.42: Massive sulphide sample from Peltosaari. Clasts of coarse biotite-rich rock. The principal sulphide present is pyrrhotite with lesser amounts of chalcopyrite, molybdenite and pyrite (Diameter of coin is 26 mm).



Figure 2.43: Massive sulphide sample with clasts of amphibolite containing quartz vein material. (Sample R7-22).



Figure 2.44: A thin vein of sulphide cataclastite intruded through massive magnetite containing sulphide stringers. (Sample R7-191).



Figure 2.45: Massive sulphide / cataclastite with rounded clasts of vein quartz. Photograph taken underground at the Haveri mine.

The network of sulphide stringers and massive sulphide is commonly closely associated with intense amphibolitisation of the host rock. The amphibolitisation with sulphides can be so intense and pervasive that the initial host rock is unidentifiable.

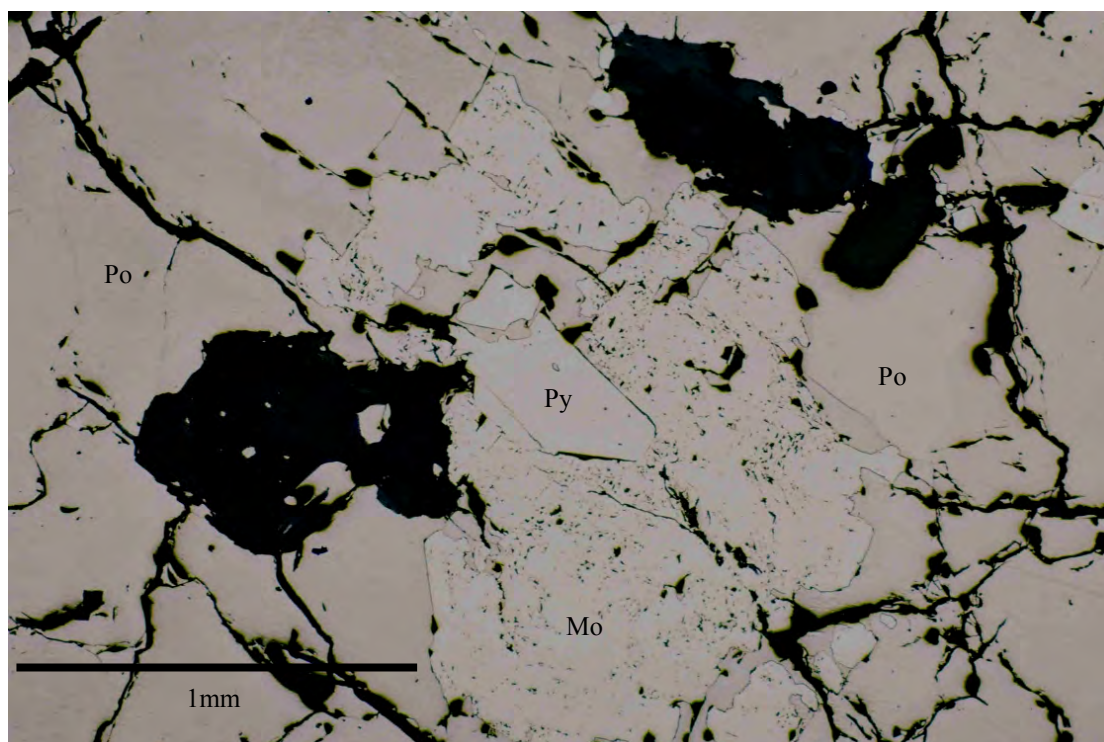


Figure 2.46: Reflected light photomicrograph of massive sulphide dominated by pyrrhotite (Po), with subhedral crystals of pyrite (Py) and molybdenite (Mo). (Sample R7-174a).

There appears to be both a lithological and a structural control on the distribution of the massive sulphide type ore. Lithologically, the massive sulphide ore occurs in the transition zone of the meta-tuffites between the meta-lavas and the metasedimentary rocks. Additionally, the proximity of the quartz feldspar porphyry is also worthy of note. Structurally, the main concentrations of massive sulphide ore appear to be located (at the deposit scale) within the hinges of folds (as suggested by Nironen, 1989), and where they have been observed in outcrop or underground, occur along N-trending fractures and veinlets (Figure 2.45). The massive sulphide ore also occurs in areas of competency contrast, such as along the contact of the porphyry, and even penetrating into fractures on the edges of the porphyry. However, no brecciated clasts of porphyry have been observed within the massive sulphide ore.

The genesis of the massive sulphide ore type is unclear. The cataclastic textures associated with many of the sulphides indicate that at least some of the sulphides have clearly been deformed and remobilised. In addition, there is a close spatial relationship with amphibolitisation of the host rocks. However, whether there is a genetic relationship between the amphibolitisation and the massive sulphide is unclear. A number of possibilities exist. It is possible that the sulphides are syngenetic with respect to the host rocks, and were emplaced before (or pene-contemporaneously with) the porphyry, and have subsequently been deformed and remobilised, perhaps during the amphibolitisation event. Alternatively, the sulphides within the massive sulphide ore type may be epigenetic, and introduced as a post-porphyry event, contemporaneously with the amphibole alteration and/or the calcic-skarn, either pre- or syn- deformation, and that the competency contrast afforded by the porphyry and the host rocks

controlled the local stress fields to the extent that the massive sulphide was emplaced in the low-stress zone “shadow” of the porphyry. An epigenetic interpretation for the cataclastites is clearly supported by underground observations indicating that the cataclastite bands crosscut the stratigraphy. This interpretation is further supported by structural measurements indicating that the cataclastites are in D₂ structures (Section 2.5).

2.4.1.2 Disseminated sulphides and oxides

Disseminated sulphides and oxides are present in all rock types as a primary or metamorphic assemblage. There are areas, however, especially within the volcanics and tuffites, where the disseminations are dominated by sulphides, accounting for up to approximately 20% of the rock mass. When these disseminations are associated with amphibolitisation of the host rock, it is easy to infer a hydrothermal origin. In these cases it is likely that many of the disseminations are a hydrothermal feature. Grain sizes of disseminated opaques are generally fine, being less than 50 µm in diameter. The opaque phases with disseminated textures are pyrrhotite, chalcopyrite, pyrite, magnetite and ilmenite. This is the typical opaque phase assemblage seen at Haveri, and consequently identifying the origins of the disseminations is problematic.

In the volcanics and tuffites, sulphide and oxide disseminations are common, and are visible as a very fine “dusting” of sulphides in the rock, producing a typically rusty weathering surface. The disseminated sulphides and oxides can in places be seen to define a brecciated texture. In this case the disseminations are primarily pyrrhotite, with lesser amounts of chalcopyrite, ilmenite and pyrite. The sulphides are generally subhedral, whereas the oxides have more rounded edges. A brecciated texture can be seen with a matrix of pervasively altered rock consisting almost entirely of amphibole and up to 20% of disseminated opaques. The “clasts”, on the other hand, have a lower sulphide/oxide content than the “matrix”. And thus relatively, a higher amphibole content. This texture can be generated in a number of different ways:

- a) The breccia may be a primary autoclastic breccia created prior to lithification, whereby the sulphides and oxides are also primary and reflect the sulphide and oxide contents of the initial volcanic material.
- b) Both the brecciated texture of the rock and the disseminated opaques may be post-lithification alteration features, caused by metamorphic or hydrothermal alteration.
- c) The breccia itself may be a primary breccia, with the sulphides and oxides being introduced at a later time by metamorphic or hydrothermal fluids. The heterogeneous distribution of the opaque phases may reflect porosity variations of the initial breccia. It is also possible that the variation in amphibole content may, to a certain extent, be a hydrothermal feature, also reflecting the porosity variations created by the initial pre-lithification brecciation.

Metavolcanic rocks which display little evidence of hydrothermal alteration also contain a certain proportion of disseminated opaques, which are mainly composed of rounded grains of ilmenite (with a rim of sphene), as well as minor pyrrhotite, chalcopyrite, pyrite and magnetite. It is therefore likely that at least the ilmenite is either primary or metamorphic. However, it is not possible to determine if the sulphides and the magnetite are primary or secondary, although a broad spatial association of the disseminations and amphibolitisation of the host rock suggests a likely genetic relationship with amphibolitisation.

In the metasedimentary rocks, the main style of sulphide and oxide mineralisation is as disseminated opaques, with only relatively minor sulphide veining. The opaques in the metasedimentary rocks are principally pyrrhotite, with lesser amounts of chalcopyrite and pyrite. The presence of magnetite as disseminations within the metasedimentary rocks is variable, with some units containing high concentrations, but with the majority of the metasedimentary rocks being devoid of magnetite, or containing only minor concentrations. The magnetite within the metasedimentary rocks, and especially the units containing high concentrations, is most likely to be a primary feature (rather than hydrothermal) reflecting sedimentary heavy mineral layers. A secondary (hydrothermal) origin, related to variations in the porosity of the rock, is considered less likely due to the REDOX conditions of the metasedimentary rocks. As will be discussed in later sections, the metasedimentary rocks represent a reducing environment (relative to the metavolcanic rocks), that favour the precipitation of pyrrhotite rather than magnetite from hydrothermal fluids (although this is also controlled by the availability of sulphur). The sulphides within the metasedimentary rocks are likely to be a combination of both primary and hydrothermal (metamorphic) in origin.

2.4.1.3 Quartz-vein related sulphides

The quartz veins and veinlets do not generally have a high content of sulphides or oxides associated with them. Within the veins themselves, sulphide contents generally account for less than 10% of the total mineral assemblage. The main sulphide phase is pyrrhotite with lesser amounts of chalcopyrite and pyrite. Both the pyrrhotite and the chalcopyrite occur as anhedral blebs, whereas the pyrite commonly has a skeletal habit with well developed crystal faces. The quartz veins and veinlets usually contain small amphibole crystals, and slivers of amphibolitised wall rock.

As mentioned previously (Section 2.3.4), the quartz veins and veinlets are commonly intensely deformed, with folding of the veins and sub-grain development within the quartz grains (Figures 2.37 – 2.38). It is not clear whether the sulphides were primary with respect to the formation of the quartz veins (pre-deformation), or whether the sulphides were introduced during deformation. However, the complex relationships of the quartz veins with respect to deformation and the alteration, indicates that quartz veins were being formed throughout much of the active geological history at Haveri. Consequently, many of the sulphides in the quartz veins are likely to be pre-deformation, whereas others may be syn-deformation.

Sulphides are also found beyond the quartz vein margins, occurring both along the contact and within the host rock itself. These sulphides are primarily pyrrhotite with lesser amounts of chalcopyrite, pyrite and molybdenite and trace amounts of magnetite. The sulphides are intergrown, with anhedral pyrrhotite commonly enveloping subhedral pyrite (commonly with inclusions of pyrrhotite), and molybdenite intergrown with pyrrhotite and pyrite with a laminated texture. Chalcopyrite is found both as inclusions within pyrrhotite, and containing inclusions of pyrrhotite. In addition chalcopyrite is commonly observed as fracture fillings within pyrrhotite, indicating that chalcopyrite continued to be introduced or mobilised after the crystallisation of pyrrhotite. These textures suggest that all of the above-mentioned sulphides formed pene-contemporaneously, with the formation of pyrite occurring at a slightly later stage than the bulk of the ore assemblage. Chalcopyrite formation occurred throughout the ore-forming stage, and continued later. This is due to the relative ease of chalcopyrite remobilisation.

2.4.1.4 Calcic-skarn related sulphides and oxides

The calcic-skarn zones contain an ore mineral assemblage which is for the main part the same as in other styles of mineralisation. The main minerals are pyrrhotite, pyrite, chalcopyrite and magnetite. The calcic-skarn assemblage differs from the quartz-vein related sulphides (and amphibole-hosted sulphides) in that minor sphalerite has been observed in the calcic-skarn assemblage (Figure 2.47), and gold is absent from the calcic-skarn assemblage. The gangue mineralogy (Section 2.3.2), with well zoned mineral assemblages, is perhaps the most obvious way of recognising this style of mineralisation. The sulphides and oxides within the calcic-skarn zones are intergrown with the gangue minerals, and are thus likely to have been formed contemporaneously. However, many of the gangue minerals, especially in the outer zones, show evidence of retrograde replacement, and thus it is possible that in such zones the sulphides and oxides crystallised later from the cooling hydrothermal fluids as the skarn system contracted.

2.4.1.5 Discussion

On the basis of the preceeding sections on sulphide and oxide mineralogy, as well as the following section on gold mineralisation (Section 2.4.2), it is possible to construct a paragenetic sequence for the ore mineralogy at Haveri (Figure 2.48). The evolution of the ore phases from the different styles of mineralisation reflect an overall increase in the degree of sulphidation throughout time. This is represented by the progression from an early oxide (FeO) phase, dominated by magnetite $\pm$ ilmenite; followed by pyrrhotite + chalcopyrite + molybdenite + sphalerite (an FeS phase); followed by late pyrite (an FeS₂ phase). This pattern of increasing sulphidation has been described by Clark (1997) for the Paroinen deposit, where sulphosalts are present in the final stages of ore formation. This will be discussed further in Sections 5 and 6.

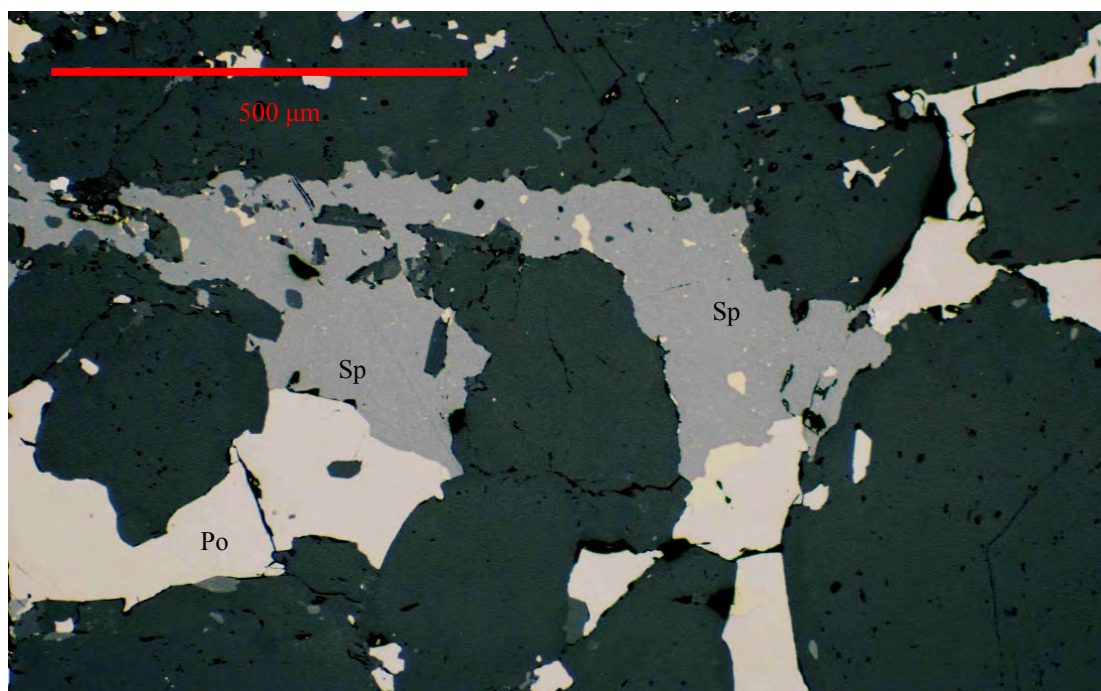


Figure 2.47: Sphalerite (Sp) and pyrrhotite(Po) in a calcic-skarn assemblage. (Sample R12-51).

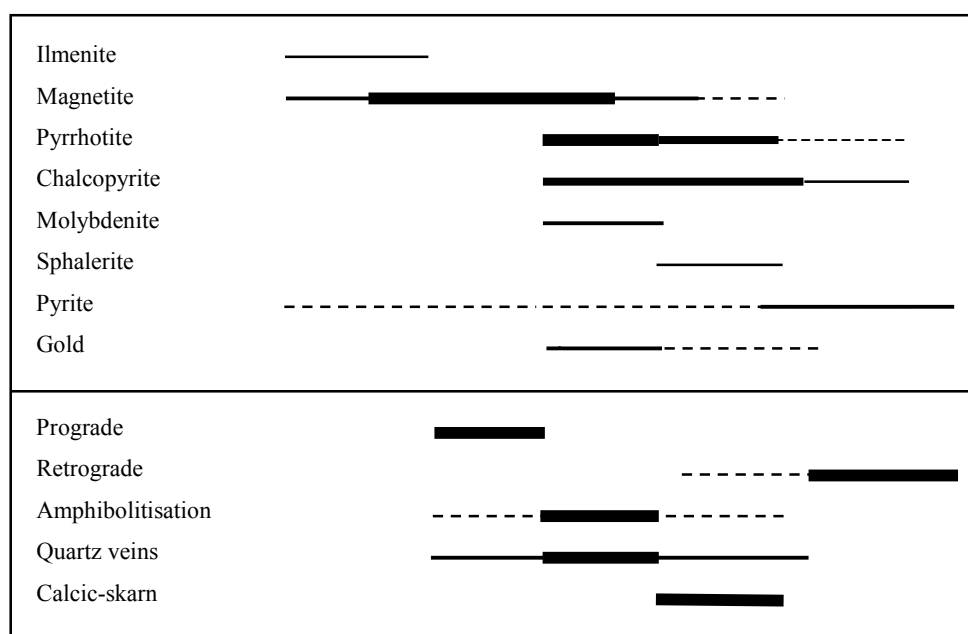


Figure 2.48: Schematic paragenetic sequence for the ore mineralogy at Haveri constructed from the ore petrography. The length and thickness of the line indicates the duration and relative importance of the minerals formation relative to the main metamorphic and alteration / veining events (lower box).

2.4.2 Gold Mineralisation

On a broad scale, the gold mineralisation at Haveri is lithologically constrained to the transitional zone of meta-tuffites between the metavolcanic rocks of the Haveri Formation and the metasedimentary rocks of the Osara Formation. Within this transitional zone the gold mineralisation can be seen to occur in several different ore types. Gold fineness has been determined by electron microscope analysis and is discussed in section 3.2.2.11.

2.4.2.1 Amphibole-hosted gold

The intense and pervasive amphibolitisation of the meta-tuffites and mafic metavolcanic rocks of the Haveri Formation, is responsible for the creation of one of the main host-rocks for the observed gold mineralisation. Gold is typically observed as easily identifiable specks, and even spectacularly concentrated grains, in the intensely amphibolitised tuffites and volcanics. Ore minerals, when present, commonly occur as veinlets, blebs and disseminations within the amphibolitised rock, and consist of pyrrhotite, chalcopyrite, pyrite, and magnetite. Gold is most commonly associated with amphibole, both as inclusions and along grain boundaries (Figures 2.49 – 2.50). The amphibolitised rock is typically dark green, although there are patches of paler, retrograded amphibole. Gold occurs in both of these, as well as less commonly along the grain boundaries of some of the associated sulphides, and in particular pyrrhotite. Gold is also occasionally observed associated with other gangue minerals, such as plagioclase, and ore minerals (usually pyrrhotite or chalcopyrite), where it occurs along grain boundaries and rarely as inclusions (Figure 2.51). Gold grains are anhedral and up to 150 μm in size.

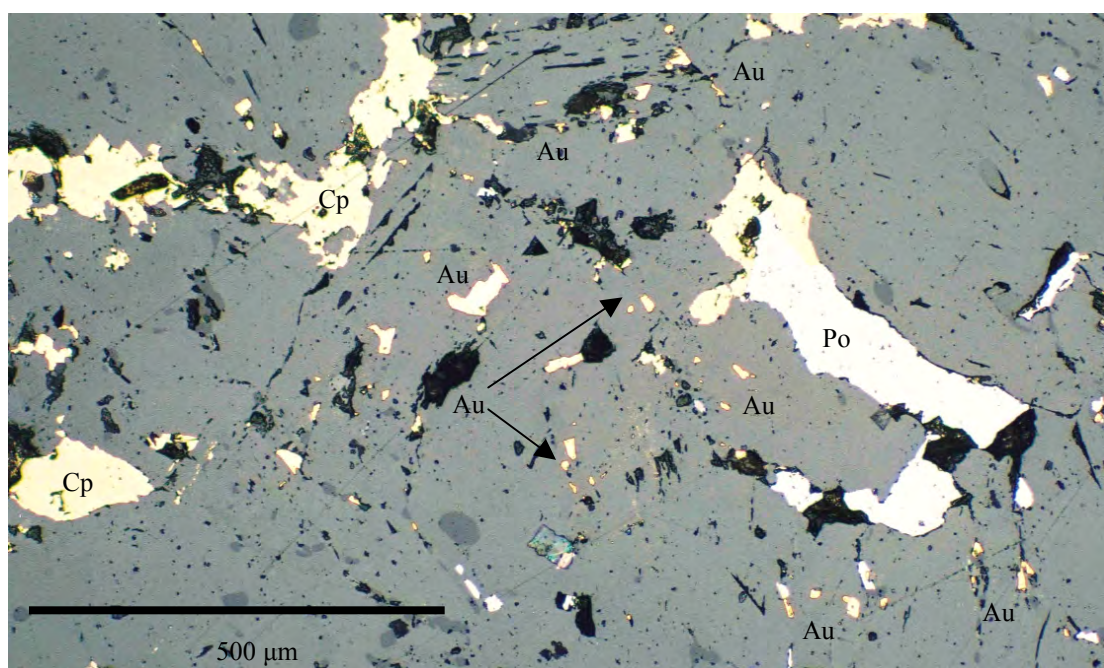


Figure 2.49: Discrete gold grains (Au) in an intensely amphibolitised mafic metavolcanic rock. Also present are grains of pyrrhotite (Po) and chalcopyrite (Cp). (Sample P22-72b).

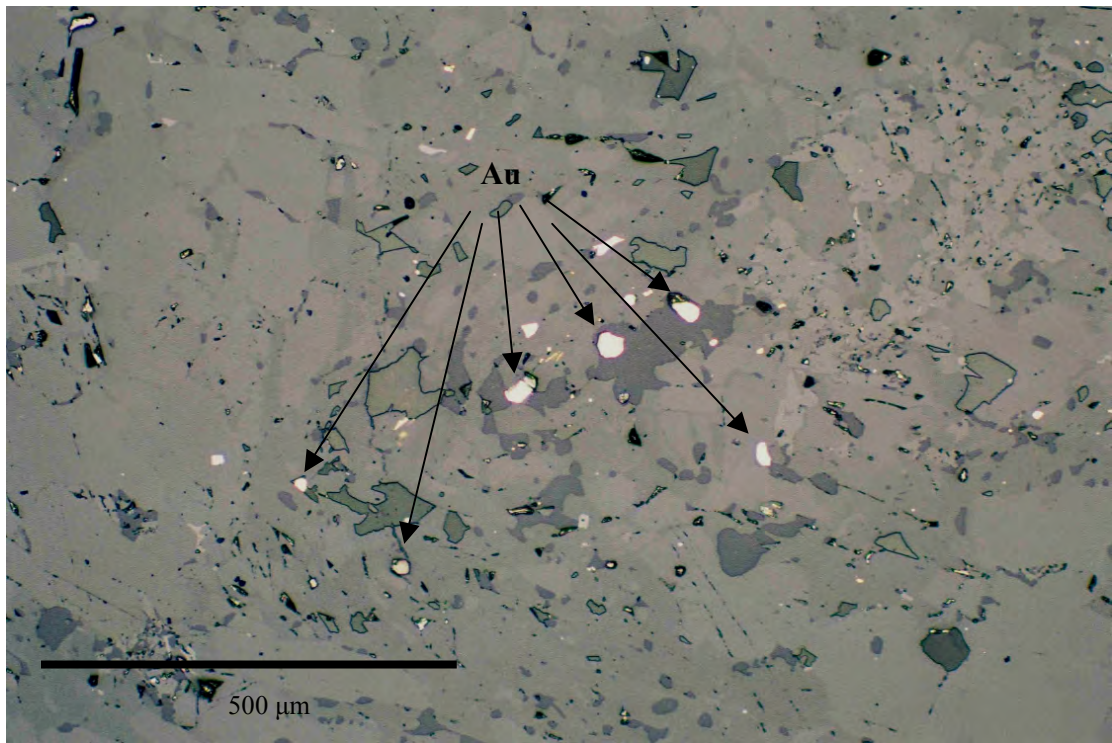


Figure 2.50: Discrete gold grains (Au) in intense and pervasively amphibolitised mafic metavolcanic rock with no associated sulphides. (Sample P22-70). Note that photograph is overexposed accounting for poor colour.

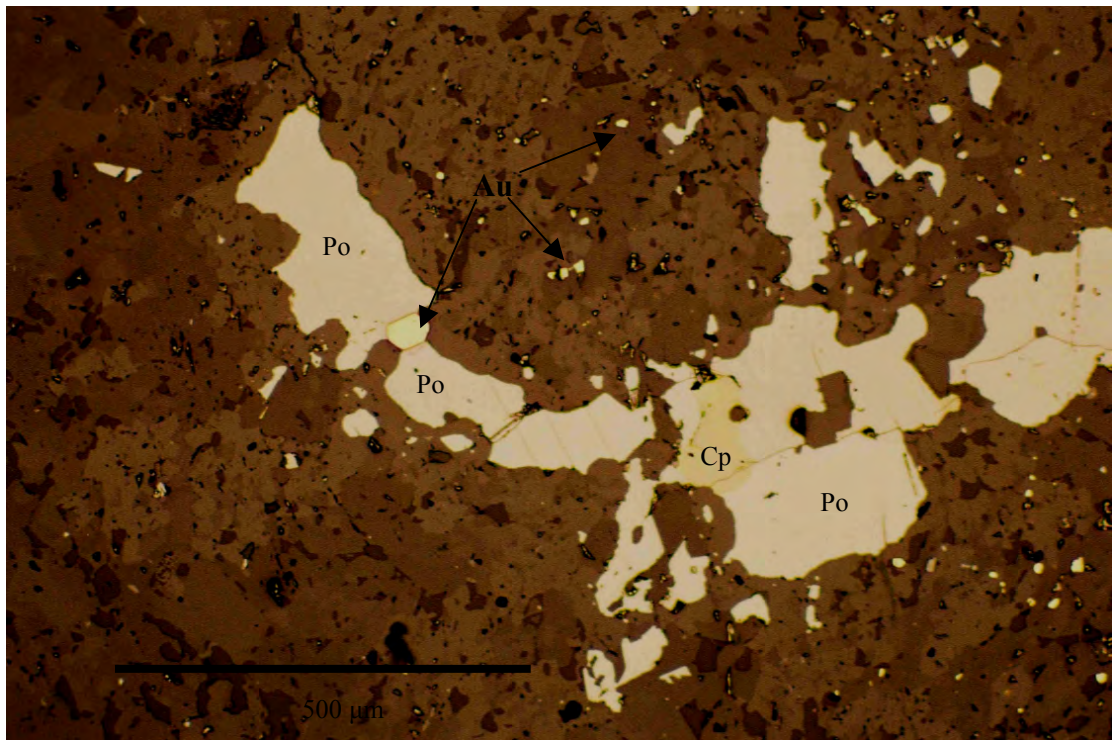


Figure 2.51: Gold associated with pyrrhotite (Po), as well as being present as discrete gold grains in amphibolite. Chalcopyrite (Cp) is also present. (Sample P22-70).

2.4.2.2 Quartz-vein related gold

Gold is commonly observed in and related to quartz veins and veinlets. The texture of the quartz veins is variable. Sometimes the veinlets are discrete, with well defined borders with the volcanic or tuffite wall rock. At other times the quartz veins have a gradational relationship with amphibolitised or sulphide-rich wall rock. The most common ore minerals associated with the quartz veins, are pyrrhotite, chalcopyrite and magnetite. These occur both within the quartz veins and adjacent to them as disseminations within the wall rock. The quartz veins contain other gangue minerals including mainly amphibole, but also occasionally plagioclase, biotite and calcite. The gold itself occurs both in the quartz veinlets (Figure 2.52), and as disseminations within and along the contact of the altered country rock (Figure 2.53). The gold is most commonly present as discrete grains within quartz grains and along grain boundaries. Less commonly it is associated directly with a sulphide phase such as pyrrhotite or chalcopyrite, again occurring either along grain boundaries or as inclusions within the sulphide.

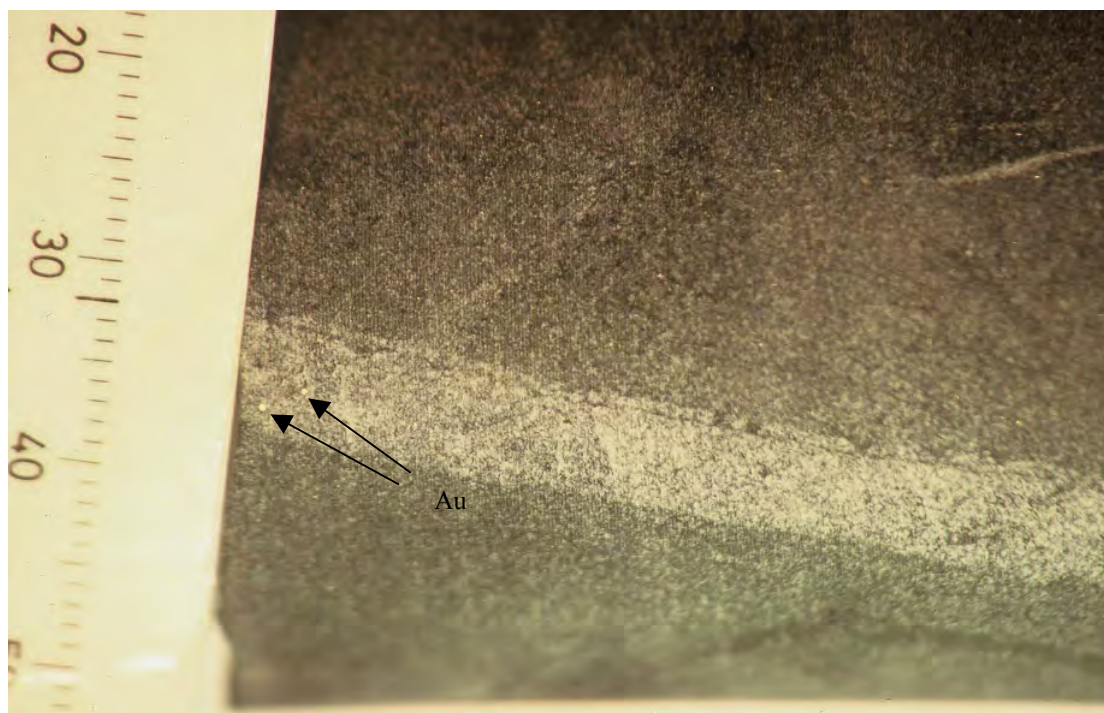


Figure 2.52: Gold grains in quartz veinlet (Scale units in millimetres).(Sample R8-241).

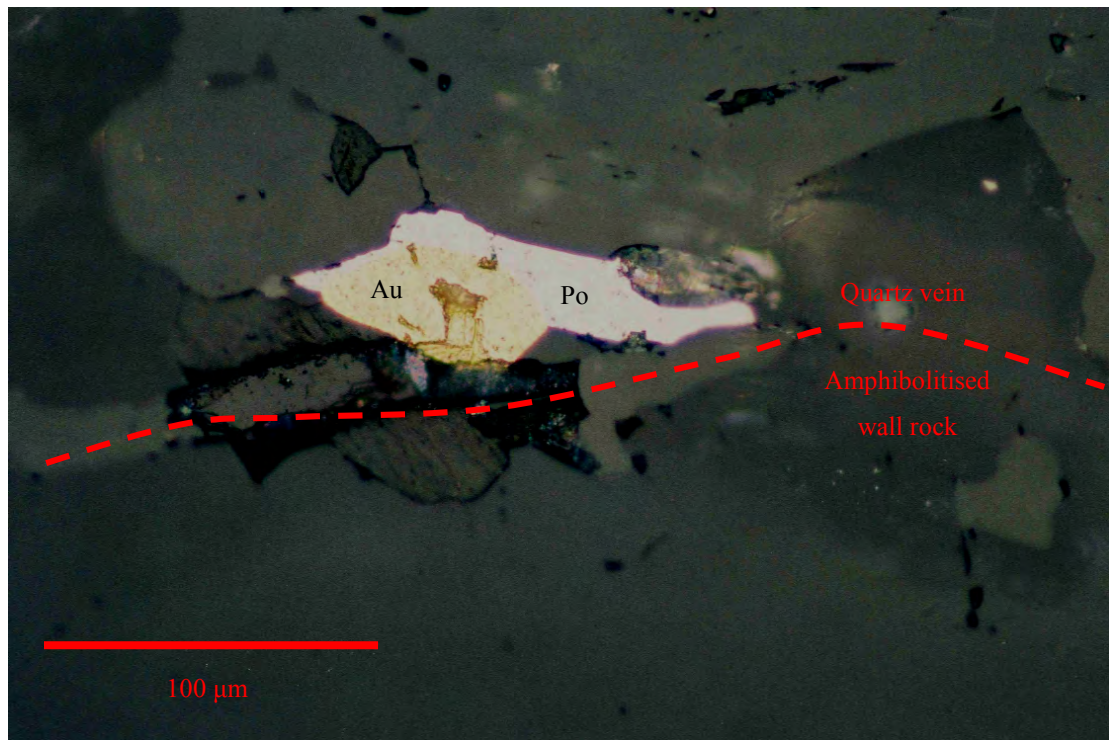


Figure 2.53: Gold associated with pyrrhotite (Po), along the contact of a quartz vein and the amphibolised wall rock. (Sample R8-219).

2.4.2.3 Sulphide-related Gold

Gold mineralisation within the massive sulphide is typically in the order of 6 – 12 g/t Au over the width of the zone, although individual samples of approximately a metre in length have Au values ranging between 1.5 g/t and 20 g/t Au. However, the location of the gold is somewhat problematic, with microscopic examinations consistently failing to identify any visible gold within the massive sulphide itself. Examinations of the individual samples with a high gold grade (>6 g/t Au) usually identify visible gold present in fine quartz veinlets and/or amphibole alteration zones. Where the gold grade is greater than about 2 g/t Au, fine quartz veinlets, fragments of quartz veinlets or amphibole alteration can almost always be identified in the assemblage, and may therefore be the host to the gold, rather than the massive sulphide. On the other hand, samples with disseminated sulphides, sulphide stringers and massive-sulphide ore, and without quartz veinlets and significant amphibole alteration, generally have elevated gold values although these are usually in the order of 0.5 – 1.5 g/t Au. This suggests that there is likely to be at least some gold hosted by the sulphides. However, this has failed to be identified under the optical microscope. Therefore, it must be considered that some of the gold at Haveri is “invisible gold”, either too fine to be discernible by optical microscopy, or alternatively that it is effectively in solid solution with a sulphide phase, most likely pyrrhotite or chalcopyrite.

2.5 *STRUCTURE*

The low density of outcrops within the study area means that any structural interpretations will be by necessity highly simplified, and heavily biased towards those areas where structural measurements can be readily obtained. The majority of structural measurements have been taken from the large area of recently exposed bedrock around the Haveri open pit (Figure 2.54), and from the old mine itself, with a relatively minor input from outcrops elsewhere in the study area. Consequently, a structural interpretation of Haveri is going to be strongly influenced by the structures observed in and close to the Haveri mine.

The earliest fabrics visible at Haveri are the primary (S_0) structures in the supracrustal sequence. Primary S_0 structures are not seen in the metalavas, but would include the alignment of pillow structures. S_0 is most commonly seen as a broad compositional layering in the metatuffites and in the metasedimentary rocks of the Osara Formation. Evidence of syndepositional deformation can be seen within the banded tuffites. This is most clearly seen as chaotic breccia structures within an otherwise well layered series of tuffites (Figure 2.9). Early faulting can also be seen in the banded tuffites where the more competent volcanoclastic layers are faulted while the epiclastic layers are plastically deformed. Both of these structures are pre-lithification, and hence pre- D_1 .

A penetrative schistosity (S_1) is developed parallel or sub-parallel to the primary layering. In the volcanoclastic bands of the metatuffites, S_1 is expressed by a weak alignment of hornblende, whereas in the epiclastic bands and in the metasedimentary rocks, S_1 is defined by the alignment of biotite. In the argillaceous units, MS_1 biotite wraps around porphyroblasts of garnet and cordierite (Figure 2.13), indicating that the porphyroblasts grew before the termination of D_1 , and that D_1 deformation occurred coevally with, and outlasted, the peak metamorphism.

In the biotite-rich argillaceous units within the tuffites and the metasedimentary rocks, the S_1 foliation has been crenulated by D_2 folding (Figures 2.55 – 2.56). Kähkönen and Nironen (1994) have attributed D_2 deformation with the retrograde metamorphism on the basis of chlorite flakes parallel to S_2 overgrowing biotite. However, chlorite is pseudomorphing the biotite, and hence it is the biotite that has formed parallel to S_2 , and is therefore MS_2 . In addition, there is only a limited amount of MS_2 biotite/chlorite parallel to S_2 , with most biotite being parallel to the crenulated S_1 (Figure 2.28). It is therefore considered likely that the retrograde metamorphism is a post- D_2 event. F_2 folding of S_1 can be observed in outcrop (Figure 2.57). Structural measurements of the S_0/S_1 foliations at Haveri have been plotted as poles to planes on a stereoplot (Figure 2.58). This clearly shows that the foliations have been refolded by F_2 with a (beta) fold axis plunging quite steeply to the southwest (62° - 225°). Within the metasedimentary rocks, it is commonly possible to observe the axial planes of minor F_2 folds, rodding lineations and intersection (S_2 on S_0/S_1) lineations L_2 . These lineations (Figure 2.58) have a vector mean plunging quite steeply to the southwest (57° - 234°), supporting the F_2 (beta) fold axis calculated from the foliations.

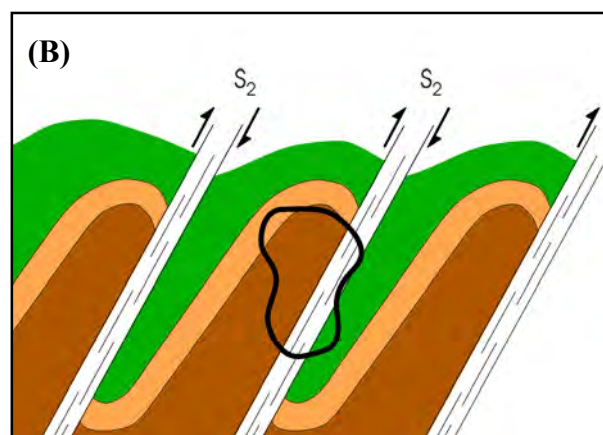
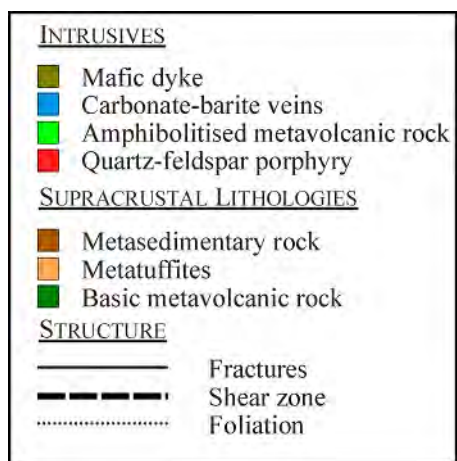
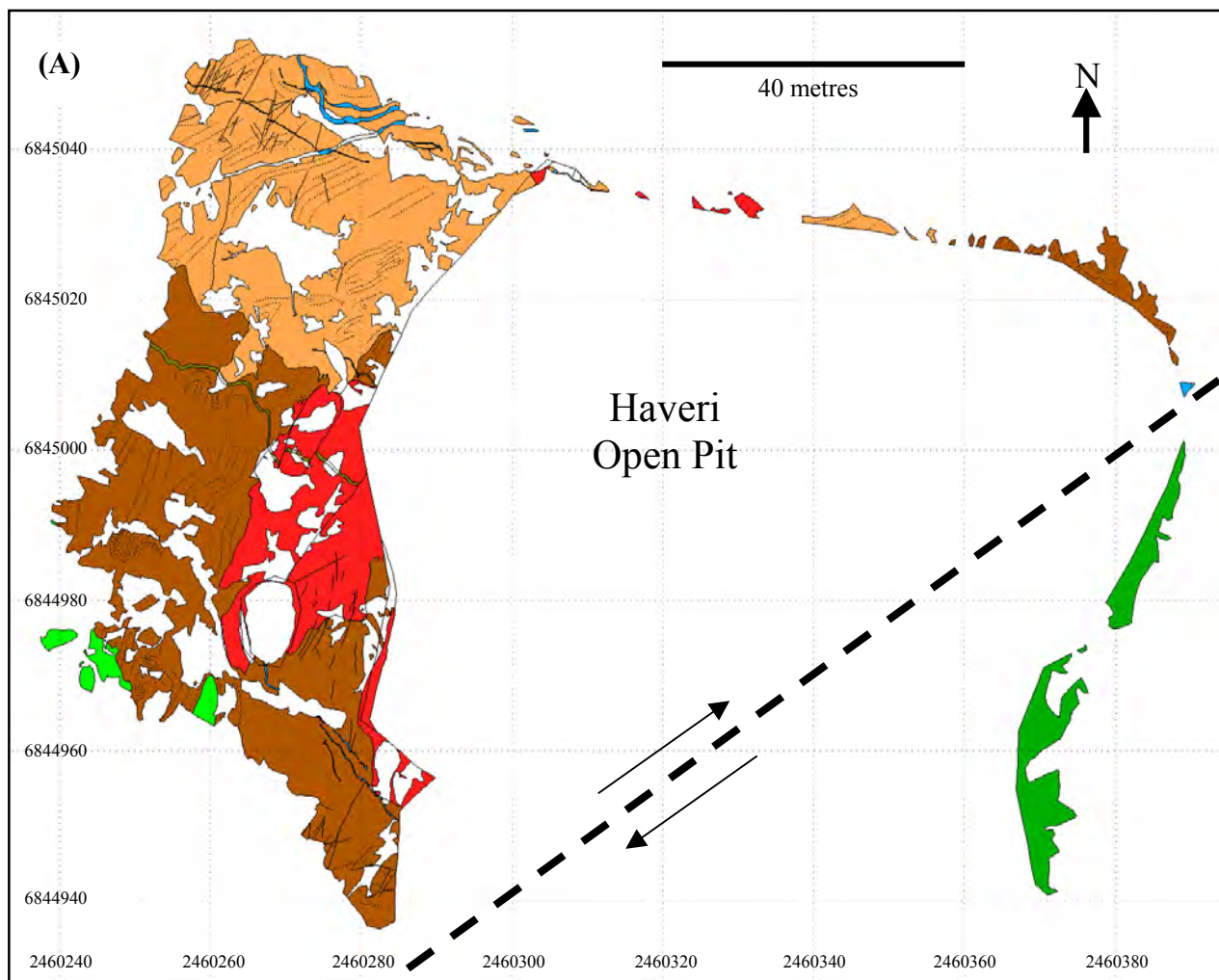


Figure 2.54: (A) Simplified geological map of the outcrop around the Haveri open pit. (B) Schematic sketch of the interpreted structure in the open pit area. The outline of the open pit is shown.

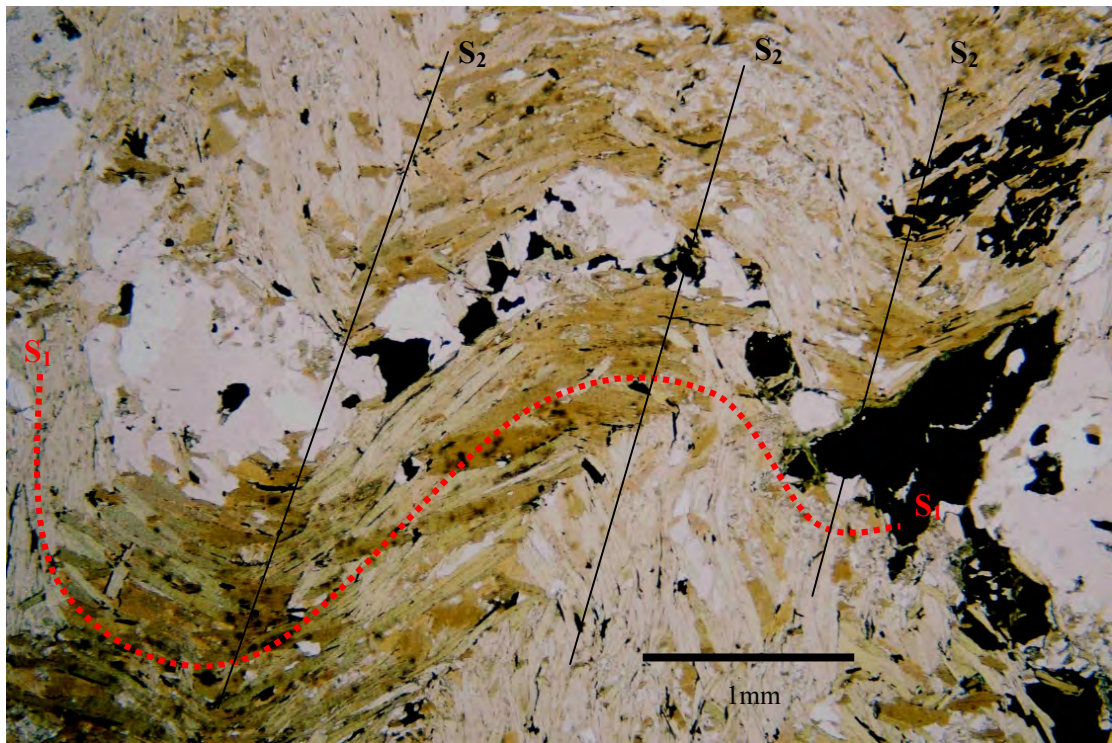


Figure 2.55: MS_1 biotite in argillaceous unit crenulated by F_2 . The quartz vein and sulphides have also been crenulated by F_2 folding. (Sample H5-291a).

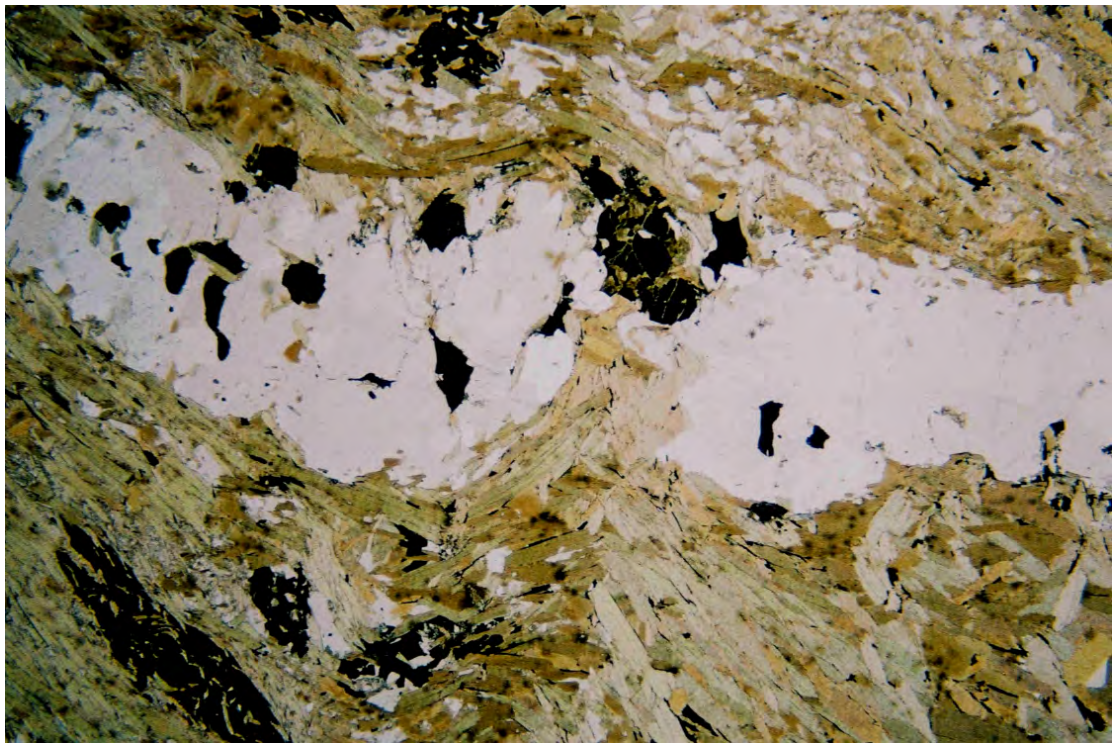


Figure 2.56: MS_1 biotite crenulated by F_2 . The pre- D_2 quartz vein has been boudinaged by the S_2 cleavage with the precipitation of sulphides, along S_2 , in an area of low strain. (Sample H5-291a).

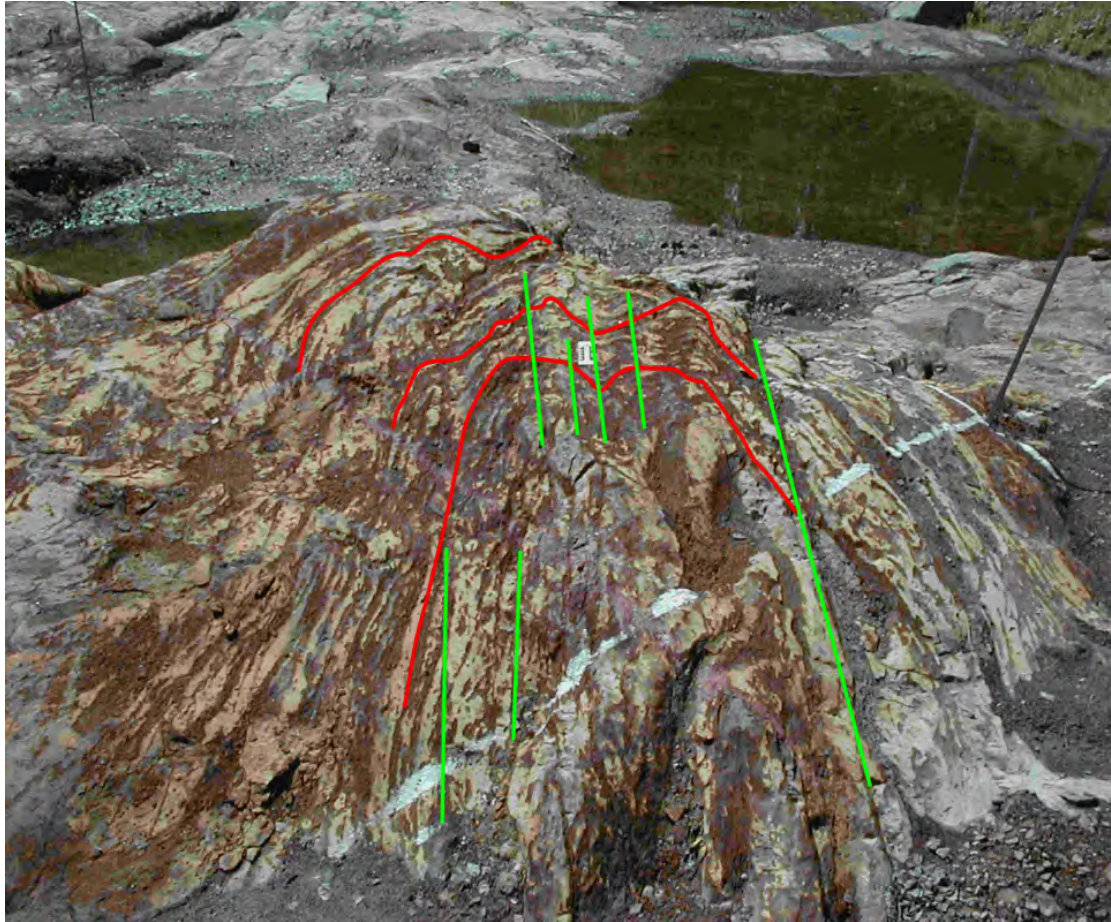


Figure 2.57: Photograph of an F_2 fold in the metasedimentary rocks to the SW of the Haveri open pit. The red lines trace the S_0/S_1 schistosity; the green lines trace the S_2 cleavage. The painted lines on the outcrop (red and white) are on a two metre grid.

Both in outcrop, and underground in the old mine, it is possible to see numerous fractures and sulphide veins and breccias (cataclastites) that are clearly post- D_1 . The sulphide bands are generally sub-parallel to the primary foliation (S_1) which (except in fold hinges) is itself commonly sub-parallel to the S_2 foliation. A stereoplot of the poles to the sulphide veins and breccias (Figure 2.59), clearly shows that they have a strong preferred orientation, striking between NE-SW and N-S, and dipping quite steeply (between 35° and 85°) to the west. The average dip and dip direction of the sulphide bands is 62° towards 281. The sulphide bands do appear to be slightly folded, and have a calculated (beta) fold axis plunging quite steeply westwards (56° -245).

Whereas the sulphide veins and breccias are generally all dipping roughly westwards, the fractures (Figure 2.60) are dipping steeply both to WNW and to ESE. However, what is clear, is that the sulphide bands and the fractures (Figure 2.61) are all striking roughly NNE-SSW, which is subparallel to the F_2 fold axial plane. It is therefore likely that the sulphide bands are an S_2 feature, with a minor F_3 overprint caused by a subparallel later deformation event (D_3). This D_3 event is likely to be responsible for the fractures (as discussed below).

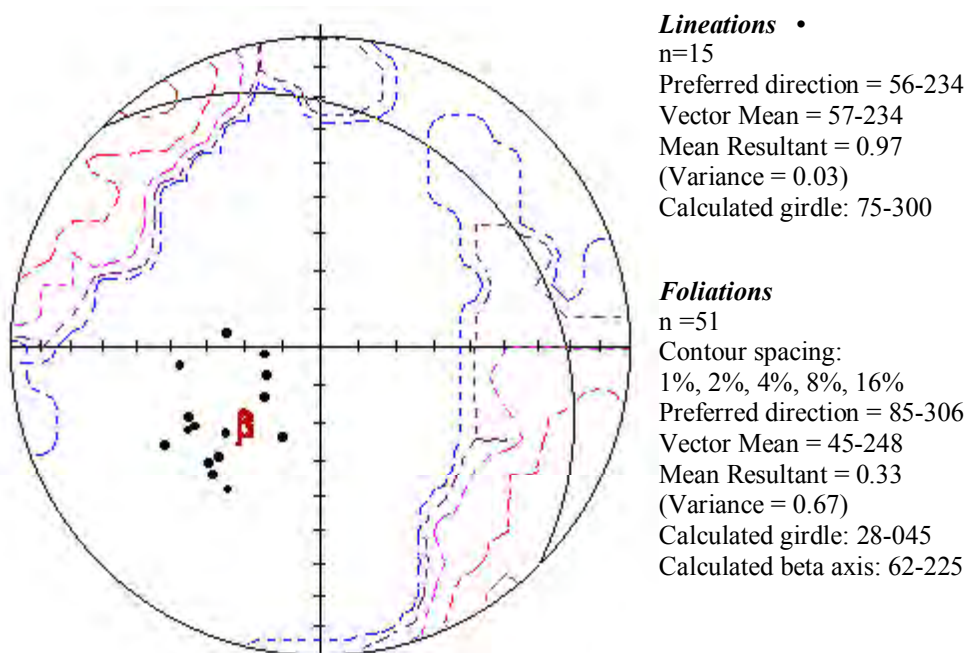
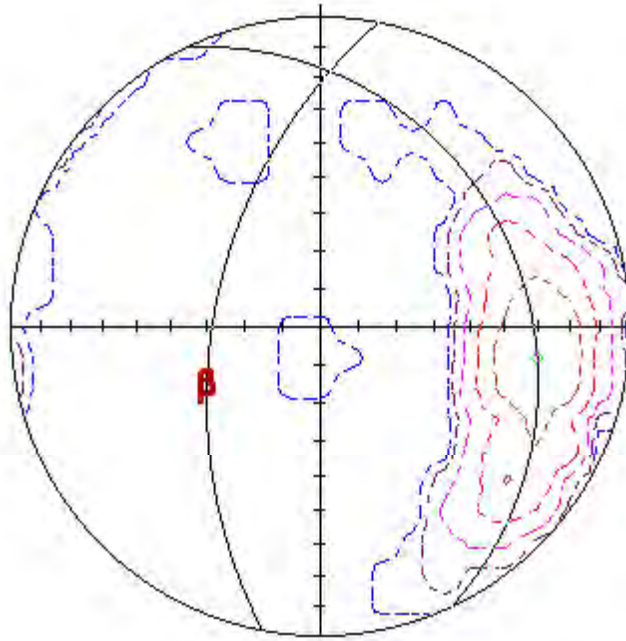


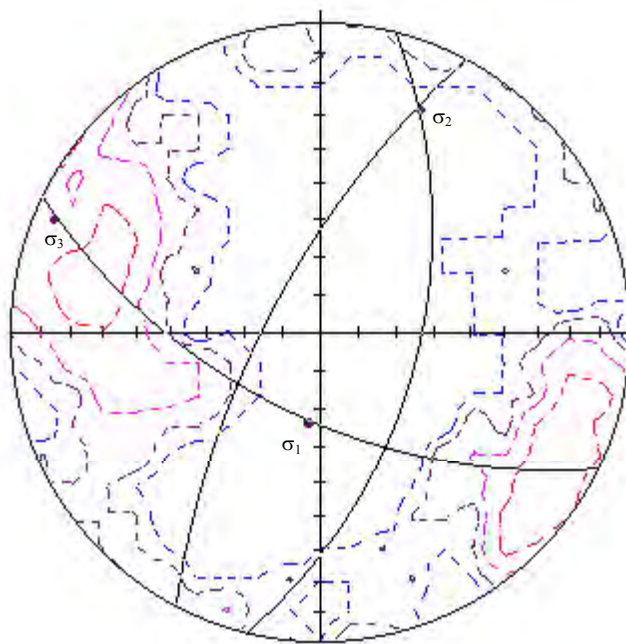
Figure 2.58: Lower hemisphere, equal-area stereoplot that shows contoured poles to S_0/S_1 foliations, folded around an F_2 fold axis (beta). Also shown are intersection (S_2 on S_1) lineations L_2 .



Sulphide veins and breccias
n = 66

Contour spacing:
1%, 2%, 4%, 8%, 16%, 32%
Preferred direction = 62-281
Vector Mean = 59-280
Mean Resultant = 0.86
(Variance = 0.14)
Calculated girdle: 34-065
Calculated beta axis: 56-245

Figure 2.59: Lower hemisphere, equal-area stereoplot that shows contoured poles to sulphide veins and breccias. Also shown is the calculated (beta) fold axis.



Fractures

n = 96
Contour spacing:
1%, 2%, 4%, 8%, 16%
Preferred direction = 84-110
Vector Mean = 18-019
Mean Resultant = 0.35
(Variance = 0.65)
Calculated girdle: 32-210
Calculated beta axis: 58-030
Girdle 1: 64-104 [axis: 26-284]
Girdle 2: 76-298 [axis: 14-118]
Girdle 3: 68-206 [axis: 22-026]

σ_1 = 66-188
 σ_2 = 22-024
 σ_3 = 7-293

Figure 2.60: Lower hemisphere, equal-area stereoplot that shows contoured poles to fracture planes. Also shown are the principal stress axes if the bimodality of the data is assumed to represent a conjugate fracture set.

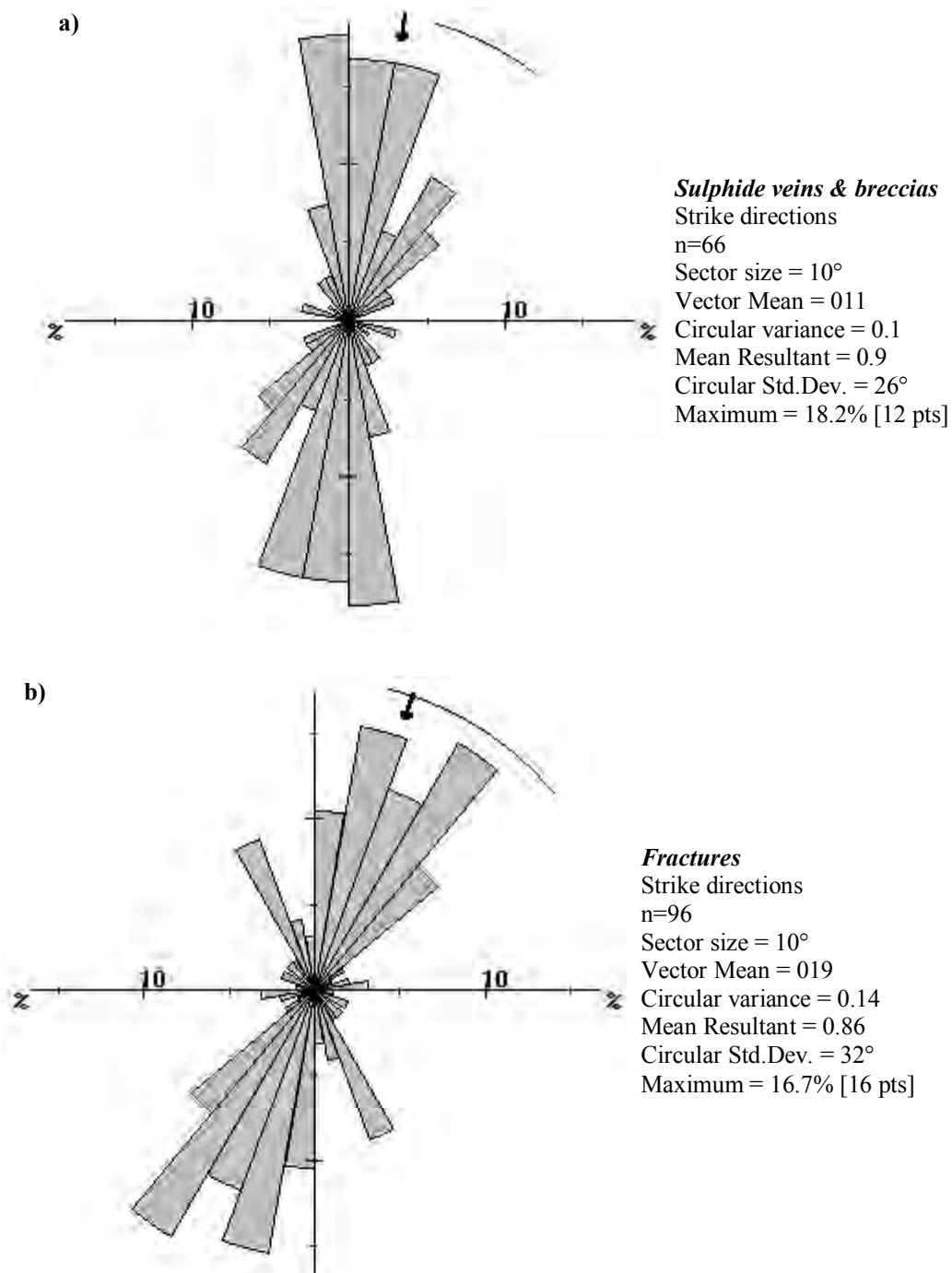


Figure 2.61: Strike directions of a) sulphide veins and breccias and b) fractures.

Evidence for D₃ deformation is the shearing of the mafic dykes, granodiorites and porphyries and the minor folding of the late stage mafic dykes. D₃ shears are pyrite-rich and trend NE – SW. On the basis that the pyrrhotite-rich sulphide bands appear to be associated with amphibolitisation and gold mineralisation (Section 2.4), which predates the mafic dykes, it is likely that the pyrrhotite-rich sulphide bands are syn-S₂. The fractures, on the other hand, do cross-cut the dykes, and are thus syn-D₃.

If the observed bimodality in the dips of the fractures is assumed to represent conjugate fracture sets, then it is possible to calculate the principal stress axes for D₃ deformation (Figure 2.60). In this case, with a relatively steep maximum stress axis ($\sigma_1 = 66^\circ\text{-}188$) and relatively flat intermediate and minimum ($\sigma_2 = 22^\circ\text{-}024$ and $\sigma_3 = 7^\circ\text{-}293$ respectively) stress axes, the fractures are related to mainly normal faulting with a minor strike-slip component.

Two sets of visibly conjugate fractures from the outcrop west of the Haveri open pit have been plotted in Figure 2.62 and the average principal stress axes calculated. The intermediate stress axis for the average of the two conjugate sets is steep ($\sigma_2 = 74^\circ\text{-}249$), whereas the maximum and minimum stress axes are flat ($\sigma_1 = 10^\circ\text{-}025$ and $\sigma_3 = 10^\circ\text{-}117$ respectively). This orientation of the stress axes is indicative of a stress system dominated by strike-slip movement.

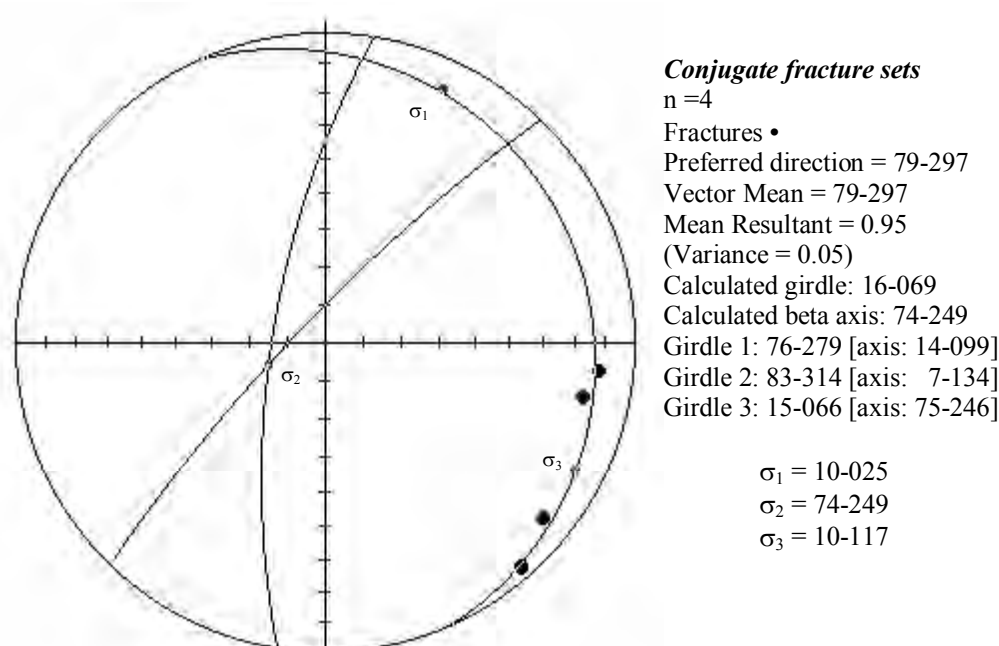


Figure 2.62: Lower hemisphere, equal-area stereoplots that shows the principal stress axes calculated for the average of two conjugate fracture sets, from the outcrop west of the Haveri open pit.

The deformation event D₃ is therefore likely to have been a mainly strike-slip dominated event, with perhaps an element of normal faulting in an extensional regime. This event is responsible for the brittle-ductile shears and brittle fractures observed around the Haveri mine area.

The relative timing of the granitoid emplacement is not totally clear. However, thin horizontal granodiorite dykes seen underground at the Haveri mine display little evidence of major folding, but have been faulted by sulphide-rich shears. These granitoid dykes are therefore likely to be post-D₁, and pre-D₃. This is consistent with syn-D₂ emplacement, as suggested by Kähkönen and Nironen (1994).

2.6 DISCUSSION

It is possible to construct a chronological sequence of events for Haveri (Figure 2.63) based on the field relationships, observed mineral textures, and structural interpretations discussed throughout this chapter (Sections 2.1 – 2.5). The earliest event at Haveri was the deposition of the supracrustal sequence (with the S_0 bedding of the metatuffites and metasedimentary rocks). The quartz-feldspar porphyries have been deformed by F_1 folds, and were thus emplaced into the supracrustal sequence either pre- or syn- D_1 , which also generated the S_1 penetrative schistosity. In the pelitic units the S_1 schistosity is defined by MS_1 biotite, which wraps around the peak metamorphic garnet and cordierite porphyroblasts. This suggests that D_1 deformation was coeval with, and outlasted the peak metamorphism. On the basis of the chlorite/biotite relationships in the pelitic rocks displaying crenulation cleavages it is possible to infer that the D_2 crenulations occurred prior to the retrograde alteration. It is therefore considered likely that the granitoids were emplaced during D_2 , and the retrograde alteration relates to the crystallisation and cooling of these large intrusive bodies.

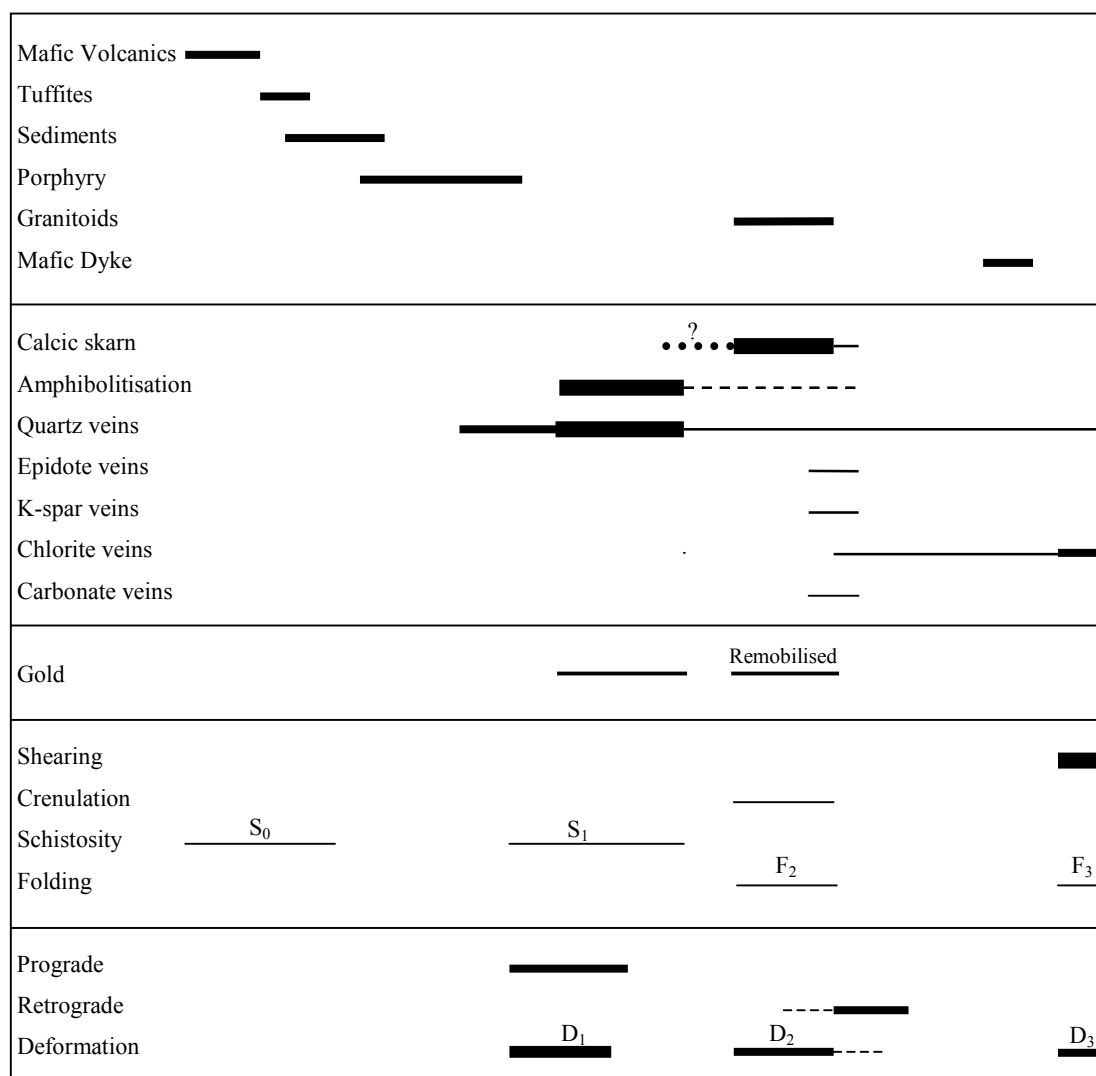


Figure 2.63: Chronological sequence of events at Haveri.

On the basis of the field relationships and lack of metamorphic recrystallisation and alteration, it is apparent that the mafic dykes represent the youngest rock type at Haveri. However, they have been affected by later D₃ deformation associated with pyrite mineralisation (Section 2.2.3.3).

The timing of the alteration phases, veining and gold and sulphide mineralisation can also be constrained relative to the main phases of deformation. The amphibolitisation of the rocks can be clearly seen to affect the pre- to syn-D₁ porphyries, creating at its most pervasive, the “uralite porphyries”. This therefore excludes the possibility of the “amphibole alteration” being interpreted as pre-metamorphic sub-seafloor spilitisation that was subsequently metamorphosed, converting the chlorite into amphibole. No evidence has been observed of the granitoids being affected by the amphibolitisation event, and thus the amphibolitisation can be constrained to syn-D₁ and pre-granitoid. The relationship between the calcic-skarn event and the amphibolitisation is much less clear, although it is likely that the calcic-skarn event is either contemporaneous with the latter phases of amphibolitisation, or is slightly later. This will be discussed in greater detail in Chapter 5.

Quartz veining is ubiquitous in the rocks at Haveri, although the majority of the quartz veins are contemporaneous with or slightly post-date the amphibole veins associated with amphibolitisation. However, intensely deformed quartz veins are also present and suggest that quartz veining has continued at Haveri since pre-D₁ right through until at least D₂ deformation and possibly later (Section 2.3.4).

Of the four main styles of mineralisation that have been recognised at Haveri (Section 2.4), two are directly related to the veins that host them (quartz veins and calcic-skarn veins). The remaining, and most important mineralising styles in terms of gold mineralisation, are those hosted in amphibolitised rock, and the cataclastite/massive sulphide veins. Field observations and petrographical studies clearly reveal a very close association between gold and the intense amphibolitisation of the rock, as well as between amphibole veining and sulphides and oxides. It is therefore considered likely that the gold mineralising event was broadly accompanied by sulphides and oxides contemporaneously with the amphibolitisation, and was therefore syn-D₁. However, it is possible that there were also *minor* sulphides and oxides present in the pre-metamorphic protoliths. The genesis of the gold mineralisation will be discussed in greater detail in Chapter 5.

Structural measurements (Section 2.5) of the cataclastite/massive sulphide bands clearly indicate that they are hosted in D₂ shear zones, and can be seen underground to crosscut stratigraphy. In addition the breccia fragments within the cataclastites consist of highly deformed fragments of quartz veinlets, as well as amphibolitised rock. All of these features clearly indicate that they are post-D₁, and therefore syn-D₂. This totally excludes the possibility of these sulphide bodies representing syngenetic massive sulphide.

3 GEOCHEMICAL STUDIES AT HAVERI

3.1 WHOLE ROCK GEOCHEMISTRY

A representative suite of 38 samples was taken from across the study area for whole rock geochemistry. This suite contains samples of all the principal rock types, displaying varying degrees of alteration from relatively unaltered specimens to intensely altered specimens.

The objectives of the whole rock geochemistry were to deduce:

- a) The primary protoliths of the various lithologies, their compositions, and to make some inferences regarding their genesis, both in terms of their geotectonic environment as well as their relationship with other lithologies.
- b) The nature and extent of the hydrothermal alteration, in terms of compositional changes and element mobility.

The whole rock geochemistry of the Haveri area has been examined as part of a broader study of the Tampere Schist Belt by Kähkönen (1987, 1989), Kähkönen and Leveinen (1994) and Lahtinen (1996). More detailed studies of the Haveri area, that have incorporated whole rock geochemistry, have been completed by Ollila (1977), Mäkelä (1980), Kähkönen et al. (1981) and, the most comprehensive study of all, Kähkönen and Nironen (1994).

3.1.1 Analytical Techniques

The samples for bulk geochemical analysis consisted of single specimens, taken from either outcrop or diamond drill core, that were fresh with respect to surface weathering features, and chosen to be representative of the rock whether altered or unaltered. Analyses were done by the commercial company, Activation Laboratories Ltd. (ACTLABS) of Ontario, Canada. All analyses were conducted on fusion beads of the samples. The major oxides, loss on ignition (LOI) and the trace elements Sc and Be were analysed using the ICP-AES. The remaining trace elements were analysed using ICP-MS on the dissolved fusion beads. In addition the FeO/Fe₂O₃ ratios were determined by titration. The analytical methods and detection limits of the various oxides and elements analysed are listed in Table 3.1.

Major Oxides								
SiO₂	ICP-AES	0.01%	MgO	ICP-AES	0.01%	Na₂O	ICP-AES	0.01%
Al₂O₃	ICP-AES	0.01%	MnO	ICP-AES	0.01%	K₂O	ICP-AES	0.01%
FeO	Titration	0.1%	CaO	ICP-AES	0.01%	P₂O₅	ICP-AES	0.01%
Fe₂O₃	ICP-AES	0.01%	TiO₂	ICP-AES	0.01%	LOI	ICP-AES	0.01%
Trace Elements (Detection Limits in ppm)								
Ag	ICP-MS	0.5	Ni	ICP-MS	5	La	ICP-MS	0.01
As	ICP-MS	5	Pb	ICP-MS	5	Ce	ICP-MS	0.01
Ba	ICP-MS	0.01	Rb	ICP-MS	0.01	Pr	ICP-MS	0.005
Bi	ICP-MS	0.05	Sb	ICP-MS	0.01	Nd	ICP-MS	0.01
Co	ICP-MS	0.1	Sn	ICP-MS	0.2	Sm	ICP-MS	0.01
Cr	ICP-MS	10	Sr	ICP-MS	0.01	Eu*	ICP-MS	0.005
Cs	ICP-MS	0.01	Ta	ICP-MS	0.005	Gd	ICP-MS	0.01
Cu	ICP-MS	5	Th	ICP-MS	0.005	Tb	ICP-MS	0.01
Ga	ICP-MS	1	Tl	ICP-MS	0.01	Dy	ICP-MS	0.01
Ge	ICP-MS	0.1	U	ICP-MS	0.005	Ho	ICP-MS	0.01
Hf	ICP-MS	0.05	V	ICP-MS	5	Er	ICP-MS	0.01
In	ICP-MS	0.1	W	ICP-MS	0.01	Tm	ICP-MS	0.005
Mo	ICP-MS	0.01	Y	ICP-MS	0.1	Yb	ICP-MS	0.01
Nb	ICP-MS	0.01	Zn	ICP-MS	2	Lu	ICP-MS	0.002
Zr	ICP-MS	0.01	Sc	ICP-AES	2 ppm	Be	ICP-AES	1 ppm
* Eu determinations are semiquantative in samples having extremely high Ba concentrations								

Table 3.1: Analytical methods and detection limits (as supplied by Activation Laboratories Ltd.) for the various oxides and elements being analysed. At the detection limit the analytical accuracy is $\pm 100\%$, at 10 times the detection limit the analytical accuracy is $\pm 25\%$, at 100 times the detection limit the analytical accuracy is $\pm 5-6\%$.

The sampling sites and short descriptions of the analysed samples are given in Appendix B, along with the full analytical results for each sample.

3.1.2 Rock Classifications

Rocks can be classified on the basis of many factors, such as geotectonic environment, mineral assemblages and geochemistry. Many of these classification schemes are most applicable to modern rocks, and become less definitive with older rocks as environments of formation become less clear, and

secondary processes such as metamorphism and hydrothermal alteration become more prevalent. Most geochemical classification schemes are based on modern analogues unaffected by secondary processes, and consequently caution must be applied when attempting to classify older rocks. However, due to behavioural variations amongst the major, trace and rare earth elements, it is possible to select classification schemes based on carefully chosen elements, that are suitable to the rock suite in question, which have the potential to reflect the pre-metamorphic or pre-alteration protoliths.

3.1.2.1 Haveri Amphibolites

The igneous rocks can be chemically classified by a number of different schemes. One of the more useful schemes is by means of plotting the total alkalis vs. silica (TAS) diagram (Le Maitre et al., 1989; Rollinson, 1993). Although this scheme is intended for fresh volcanic rocks (i.e. neither altered nor metamorphosed), it can still be useful when looking at the range of igneous rocks at Haveri (Figure 3.1).

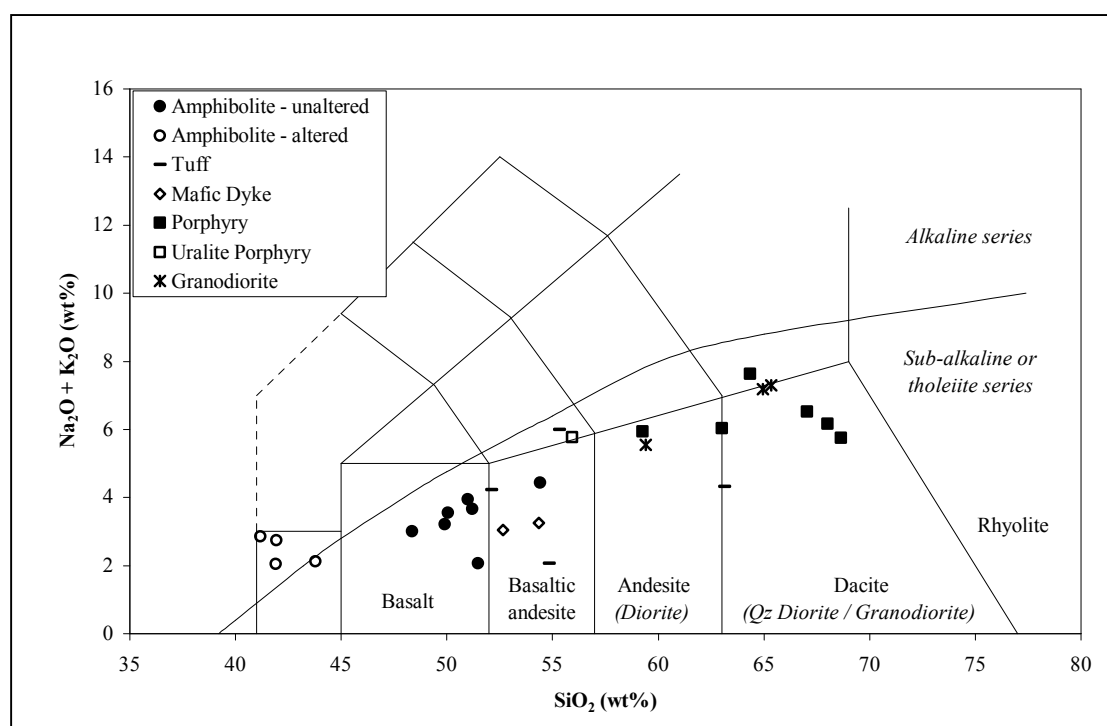


Figure 3.1: Total alkalis vs. silica (TAS) diagram (after Le Maitre et al. (1989)). The division between the alkalic and sub-alkalic magma series has been taken from Irvine and Baragar (1971).

The Haveri amphibolites plot in two major clusters. The “unaltered” amphibolites consist of samples of vesiculated amphibolite and pillowed amphibolite, as well as some samples displaying a weak amphibole alteration, with minor network amphibolitisation, and amphibole spots. These samples form a cluster plotting mainly in the basalt field on the diagram within the subalkalic magma series, whereas

the more heavily altered amphibolites plot in the picro-basalt field, and straddle the alkalic – sub-alkalic boundary. These more altered amphibolites are intensely amphibolitised, and display a relative depletion in both silica and alkali contents. The subalkalic series has been subdivided by Le Maitre et al. (1989). The unaltered Haveri amphibolites plot in the medium-K field close to the low-K field boundary (Figure 3.2), whereas the altered samples plot outside of the defined fields. Under the classification scheme of Rickwood (1989), the unaltered amphibolites are classified as calc-alkaline series. On the AFM diagram of Irvine and Baragar (1971), however, the Haveri amphibolites plot within the tholeiite series field (Figure 3.3).

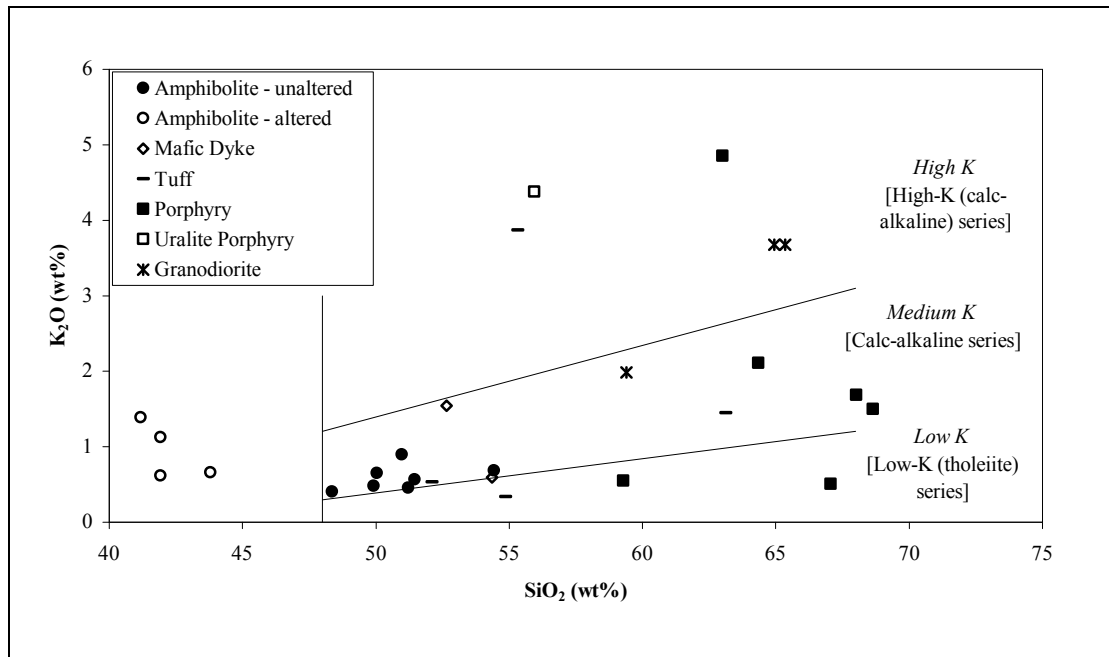


Figure 3.2: K_2O vs. SiO_2 diagram showing the subdivision of subalkalic rocks after Le Maitre et al. (1989a) (nomenclature in italics). These fields broadly coincide with those of Rickwood (1989c) (nomenclature in parentheses).

The samples of tuff show a wide variation on the TAS diagram, reflecting the heterogeneous nature of the samples. Due to the finely banded texture of the tuffs, it is difficult to selectively sample the volcanoclastic bands without a certain degree of contamination with the epiclastic bands. Most of the tuff samples plot either in, or close to, the basaltic andesite field and are thus consistent with the results of Kähkönen and Nironen (1994). The high SiO_2 tuff sample is of a rusty band, interpreted as an epiclastic unit within the tuffs.

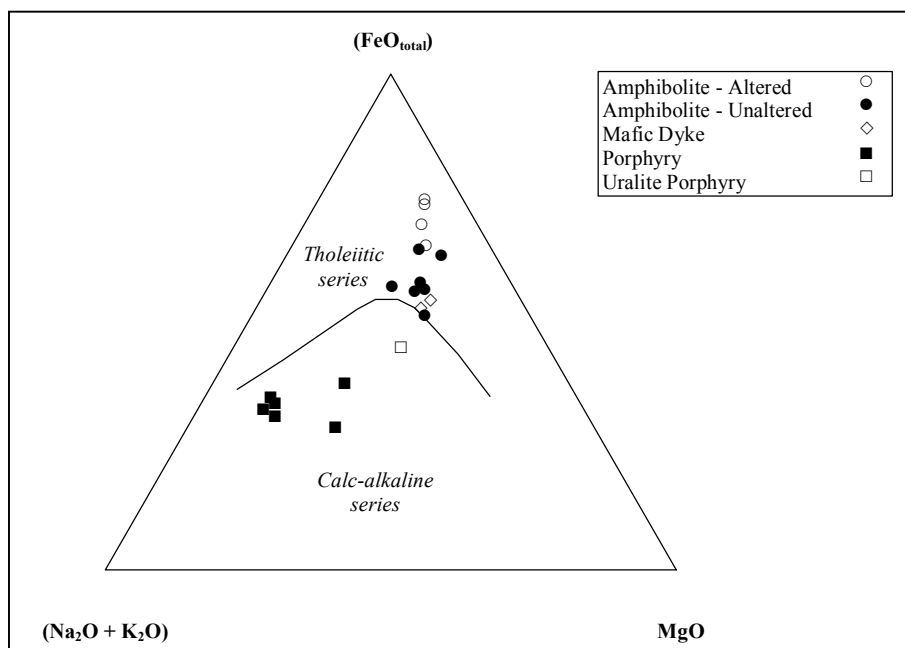


Figure 3.3: An AFM diagram of the Haveri volcanic rocks. The boundary between the tholeiitic and the calc-alkaline series is from (Irvine and Baragar, 1971)

Tectono-magmatic discrimination diagrams are commonly used for classifying igneous rocks, especially of basaltic to andesitic compositions, in terms of their tectonic settings during emplacement. However, in using such discrimination diagrams, it is necessary to both understand the basis of their construction and to be aware of their limitations and inherent problems. Discrimination diagrams are based on the assumption that different tectonic environments generate magmas that can be chemically distinguished. These chemical fingerprints are assumed to be caused by a combination of factors which vary between tectonic environments, such as variations in the source area for the magma, the degree and method of partial melting that produces the magma, and subsequent modification of the magmas, by magma mixing, fractional crystallisation and contamination. As the body of geological knowledge has grown, it has become apparent that magmas from different tectonic environments are in fact more heterogeneous than earlier assumed (Rollinson, 1993), and hence a spread of data is to be expected when plotted in discrimination diagrams. Most tectono-magmatic discrimination diagrams have been empirically derived from suites of recent, relatively unaltered igneous rocks. Consequently, as one moves further from the “control” rocks used to derive the diagram, either in terms of age or degrees of alteration, one can expect greater deviations from the “true” tectonic environment of formation.

Some of the more useful discrimination diagrams use high-field-strength elements (HFSE), such as Ti, Zr, Y, Nb, and P, which are thought to be relatively immobile in aqueous fluids and thus stable under conditions of sea-floor alteration, hydrothermal alteration and metamorphic grades up to mid-amphibolite facies (Rollinson, 1993). Consequently, discrimination diagrams can still be of some use when applied to metamorphic and hydrothermally altered rocks (such as those at Haveri), provided that

the results are treated with caution, and that they are utilised in conjunction with other sources of geological information.

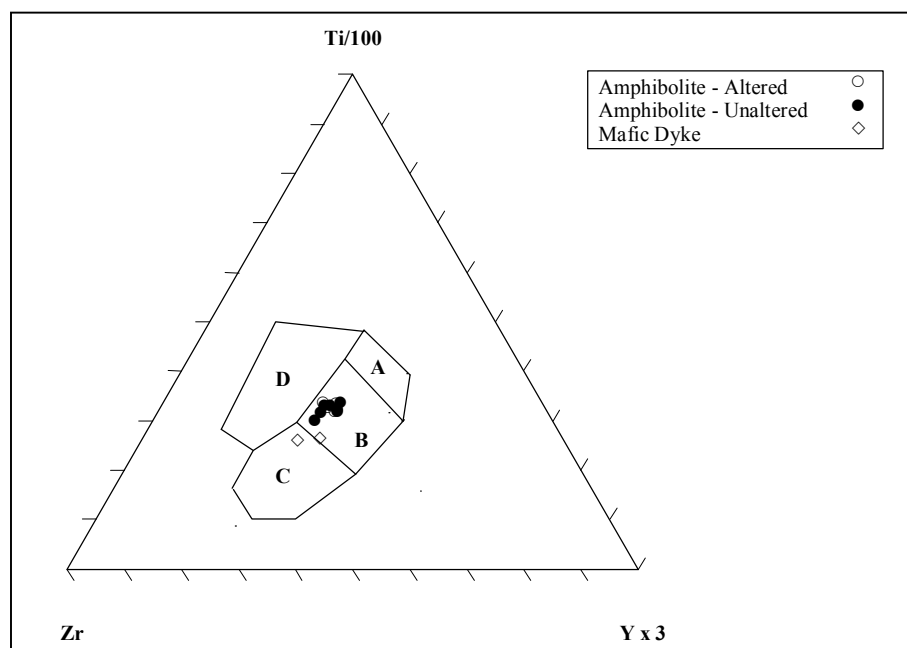


Figure 3.4: The Haveri mafic rocks plotted on the Ti-Zr-Y tectonomagmatic discrimination diagram (after Pearce and Cann, 1973). The fields are as follows: A, island-arc tholeiites; B, MORB, island-arc tholeiites and calc-alkali basalts; C, calc-alkali basalts; D, within-plate basalts.

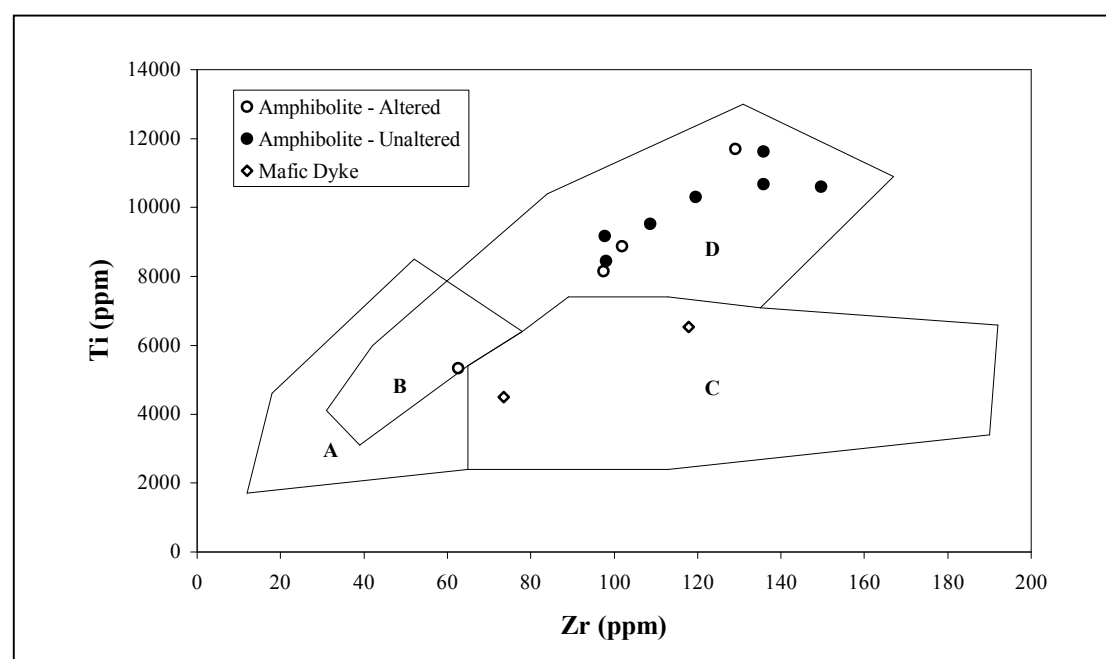


Figure 3.5: Haveri mafic rocks plotted on the Ti-Zr discrimination diagram for basalts (after Pearce and Cann, 1973). The fields are as follows: A, island-arc tholeiites; B, MORB, calc-alkali basalts and island-arc tholeiites; C, calc-alkali basalts; D, MORB.

The Ti-Zr-Y and Ti-Zr diagrams of Pearce and Cann (1973) have been widely used in determining the tectonic setting of basalts. Figures 3.4 – 3.5 suggest that the amphibolites of the Haveri Formation were basalts with mid-ocean ridge (MORB) affinities, although the diagrams are unable to determine which of the various types of MORB the Haveri amphibolites represent. It is therefore necessary to look at other discrimination diagrams that have been created to distinguish the various types of MORB (Figures 3.6 – 3.8). It is clear that none of these are able to satisfactorily classify the Haveri samples, and indeed there is considerable contradiction between the various schemes (Table 3.2). This perhaps highlights the dangers inherent in utilising discrimination diagrams without due caution.

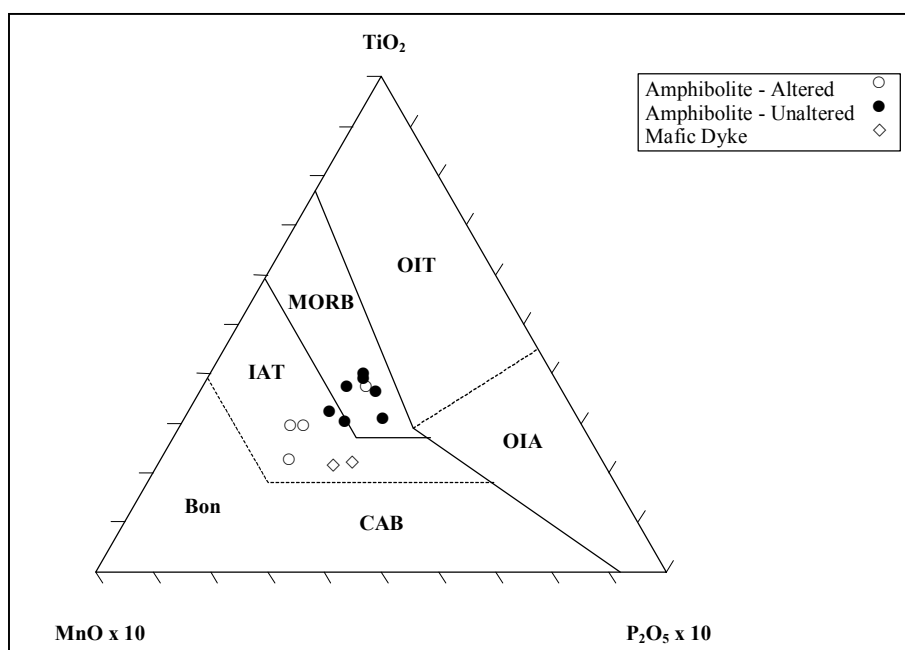


Figure 3.6: The Haveri mafic rocks plotted on the $\text{MnO-TiO}_2\text{-P}_2\text{O}_5$ discrimination diagram for basalts and basaltic andesites (after Mullen, 1983). The fields are MORB; OIT, ocean-island tholeiite; OIA, ocean-island alkali basalt; CAB, island-arc calc-alkaline basalt; IAT, island-arc tholeiite; Bon, boninite.

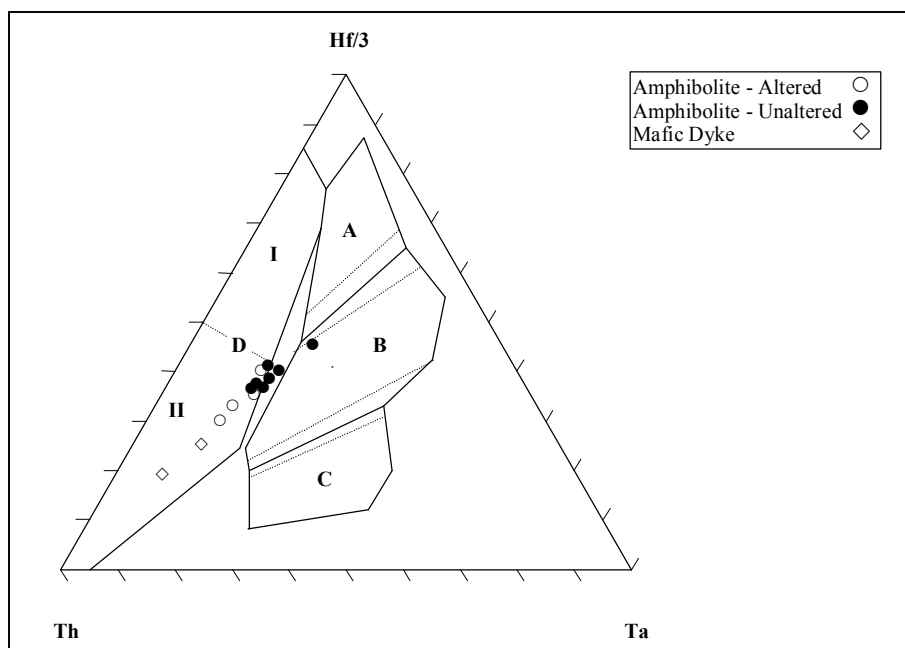


Figure 3.7: The Haveri mafic rocks plotted on the Th-Hf-Ta discrimination diagram for basalts (after Wood, 1980). The fields are: A, N-type MORB; B, E-type MORB and within-plate tholeiites; C, alkaline within-plate basalts; D, volcanic-arc basalts; DI, island-arc tholeiites; DII, calc-alkaline basalts. The broken lines indicate transitional zones between basalt types.

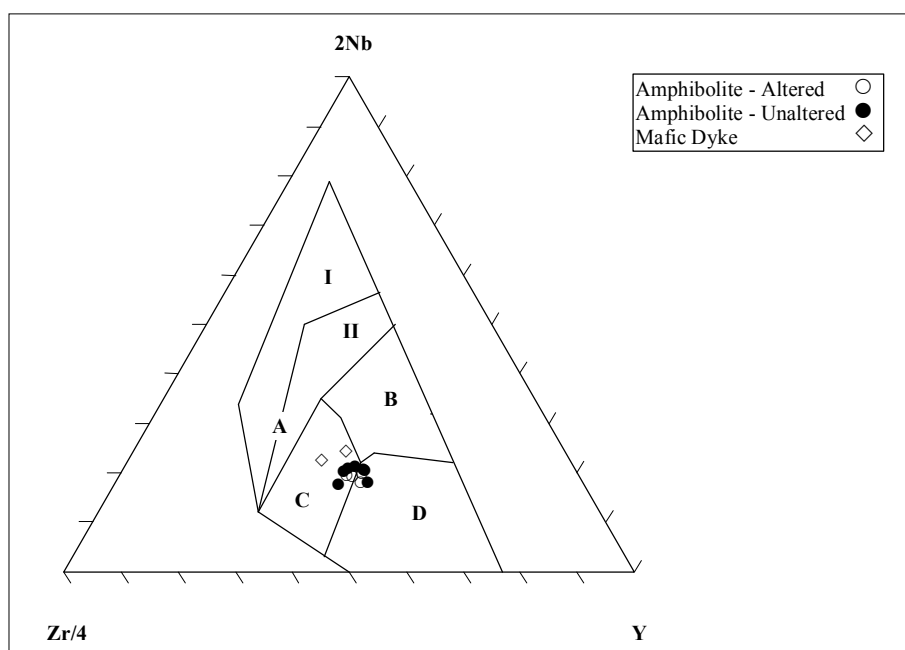


Figure 3.8: The Haveri mafic rocks plotted on the Zr-Nb-Y discrimination diagram for basalts (after Meschede, 1986). The fields are: AI, within-plate alkali basalts; AII, within-plate alkali basalts and within-plate tholeiites; B, E-type MORB; C, within-plate tholeiites and volcanic-arc basalts; D, N-type MORB and volcanic-arc basalts.

The geochemical and geotectonic classification of the Haveri amphibolites, although far from definitive, provides sufficient information to state the following: the Haveri amphibolites are medium-K basalts, of either the calc-alkaline (Figure 3.2) or tholeiitic (Figure 3.3) series. However, on both of these plots, the unaltered samples lie close to the boundary with either the tholeiitic or calc-alkaline fields. Kähkönen and Nironen (1994) identified the Haveri metalavas as low-K/medium-K basalts with tholeiitic affinities. In terms of the geotectonic environment of formation, the Haveri amphibolites consistently show certain MORB characteristics, although they commonly overlap the island-arc and calc-alkaline basalt fields also. A more detailed assessment is not possible without additional geological information.

<i>Classification scheme:</i>	<i>Classified as:</i>
TAS	Basalt
K ₂ O vs. SiO ₂	Medium-K; calc-alkaline series
AFM	Tholeiitic series
Zr-Ti-Y	MORB, island-arc tholeiites, calc-alkali basalts
Ti-Zr	MORB
MnO-TiO ₂ -P ₂ O ₅	MORB, island-arc tholeiite
Th-Hf-Ta	E-type MORB, within-plate tholeiites, calc-alkaline basalts
Zr-Nb-Y	N-type MORB, volcanic-arc basalts, within plate tholeiites

Table 3.2: Summary of classifications of unaltered Haveri amphibolites.

The rare earth elements (REE) are regarded as amongst the least soluble trace elements and are relatively immobile during weathering, low-grade metamorphism and hydrothermal alteration (Rollinson, 1993; Wilson, 1989). However, the REE are not totally immobile and there is no simple relationship between the degree of mobility of the REE and rock type or metamorphic grade. Rather, REE mobility is controlled by mineralogical and fluid controls (Humphries, 1984; Rollinson, 1993). Consequently caution must be exercised when interpreting the REE patterns of heavily altered or highly metamorphosed rocks. Nevertheless, REE patterns even in slightly altered rocks can faithfully represent the original composition of the unaltered parent (Rollinson, 1993).

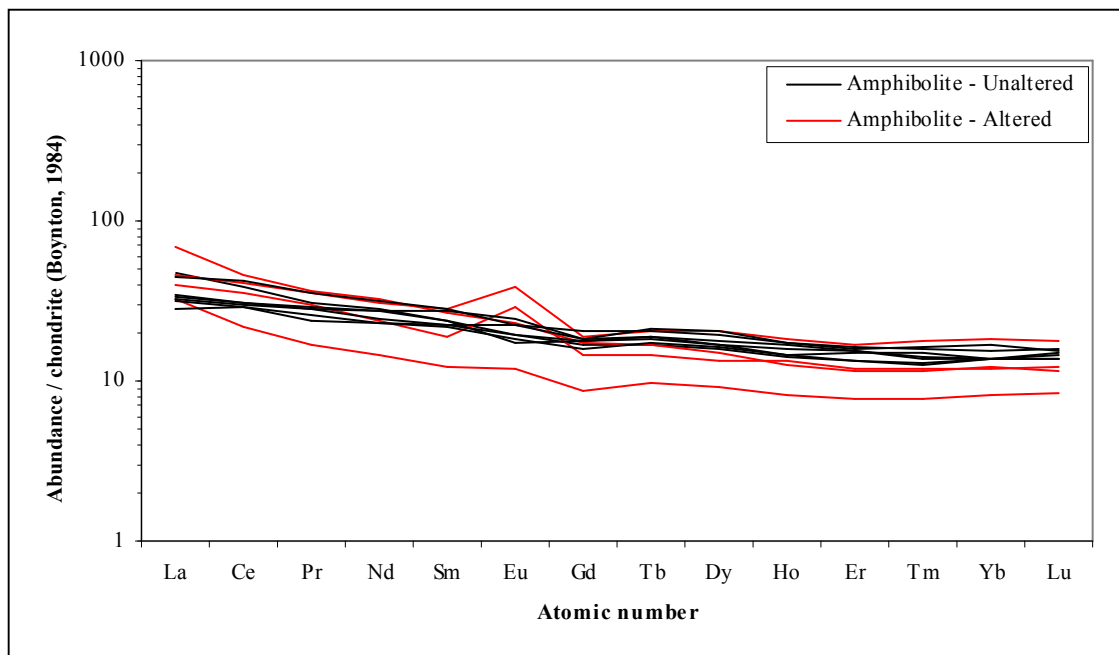


Figure 3.9: Chondrite-normalised REE patterns of the Haveri amphibolites. Chondrite values taken from Boynton (1984).

The relatively unaltered amphibolites from Haveri (Figure 3.9) display flat, gently sloping patterns with the light REE (LREE) slightly enriched relative to the heavy REE (HREE). The patterns of the altered samples are discussed later (Section 3.1.3). All seven samples of unaltered amphibolite analysed have closely matching patterns which suggests that any subsequent mobilisation of REE (if any) has happened equally to all the samples. The four patterns (Figure 3.9) of the altered amphibolites (all samples altered by amphibolitisation) display marked deviations from the close patterns of the unaltered samples, and in particular an obvious enrichment of europium. This suggests that the REE have indeed been somewhat mobilised during amphibolitisation. Consequently, the REE patterns of the unaltered samples can be assumed to represent either the original composition of the host rocks or a regional, and remarkably consistent metamorphic mobilisation event. It is considered likely that the REE patterns of the unaltered amphibolites are fair representations of the original compositions of the igneous host rock.

Three major factors control the distribution of the REE in igneous rocks.

- 1) Composition of the source
- 2) Degree and conditions of partial melting of the source
- 3) Fractional crystallisation

The highly incompatible nature of the REE means that with a high degree (15 – 30 %) of partial melting (Green and Ringwood, 1967; Sun et al., 1979; Wilson 1989), as is the case with tholeiitic basalts, there is insignificant fractionation of the REE and that the REE composition of the melt phase is very similar to that of the starting material (Haskin, 1984). Therefore, in the case of the Haveri

amphibolites, partial melting can effectively be ruled out as a controlling factor in the distribution of the REE.

Fractional crystallisation of the melt phase can also be excluded as a major factor in producing the observed REE patterns. With increasing crystal fractionation of olivine, plagioclase and clinopyroxene, the remaining composition of the melt has a REE pattern scarcely changed from the starting composition but with increasing overall abundances. This continues until about 80% of the original melt has crystallised, at which point the overall abundances of the REE in the remaining melt become greatly enriched (relative to both chondrites and the original starting composition), and a negative europium anomaly becomes increasingly obvious (Haskin, 1984). In the case of the Haveri amphibolites, the overall abundances are only moderately enriched relative to chondrite (28 – 48 times for La; 13 – 16 times for Lu), and have no significant europium anomalies¹ ($[Eu/Eu^*]_N$ of 0.8 – 1.1). Therefore, it can be concluded that the REE patterns observed for the Haveri amphibolites reflect the original composition of the source.

The REE patterns of the unaltered Haveri amphibolites (Figure 3.9), and hence of the source area are slightly enriched in the LREE ($[La/Yb]_N$ of 2.0 – 3.4) relative to the HREE, with smooth gently sloping patterns. The LREE display a mildly greater degree of fractionation ($[La/Sm]_N$ of 1.3 – 2.0) relative to the HREE ($[Gd/Yb]_N$ of 1.1 to 1.4). These patterns correspond closely to the data presented by Kähkönen et al. (1981) and Kähkönen & Nironen (1994). Although there is a continuous spectrum of magma type between N-type and E-type MORB, N-type MORB tends to show a relative depletion in the LREE ($[La/Sm]_N < 1$), whereas E-type MORB in many cases shows a tendency for LREE enrichment ($[La/Sm]_N > 1$) (Wilson, 1989). Flat REE patterns weakly enriched in LREE (6 – 30 times chondrite) are also produced by tholeiitic island-arc basalts as well as in certain back-arc basin basalts. However, the LREE-enriched island-arc basalts from the South Sandwich Islands display a europium anomaly that is absent from the LREE-enriched back-arc basin basalts from the East Scotia Sea (Hawkesworth et al., 1977; Wilson, 1989). Therefore on the basis of the REE patterns, the Haveri amphibolites appear to have a greater affinity for an E-type MORB or a back-arc basin setting, than for an N-type MORB.

The unaltered Haveri amphibolites have been plotted on the MORB-normalised trace element variation diagram (Figure 3.10) devised by Pearce (1983). The elements are ordered in the diagram according to their mobility in an aqueous fluid, and their relative incompatibility. The large ion lithophile (LIL) elements (Sr, K, Rb and Ba) are classed as mobile and plot at the left end of the pattern, whereas the high field strength (HFS) elements (Th to Yb) are generally immobile, and are plotted at the right of the pattern. The elements are arranged so that the incompatibilities of both mobile and immobile elements increase from the margins to the centre of the pattern. The unaltered Haveri volcanics display a “humped” pattern (Figure 3.10) whereby the more incompatible elements are enriched relative to N-

¹ $[Eu/Eu^*]_N = Eu_N / \sqrt{[(Sm_N) \cdot (Gd_N)]}$ from Wilson (1989)

type MORB. The mobile LIL elements are enriched relative to N-type MORB by a factor of <10 . The more “immobile” HFS elements and other alteration resistant elements define a gently sloping pattern from Th to Yb, with minor Ta, Nb and P depletions and a minor Ce enrichment. This pattern is broadly comparable to the results of Kähkönen and Nironen (1994). One sample displays a major enrichment in Ba, which is likely to be caused by hydrothermal alteration associated with carbonate veining.

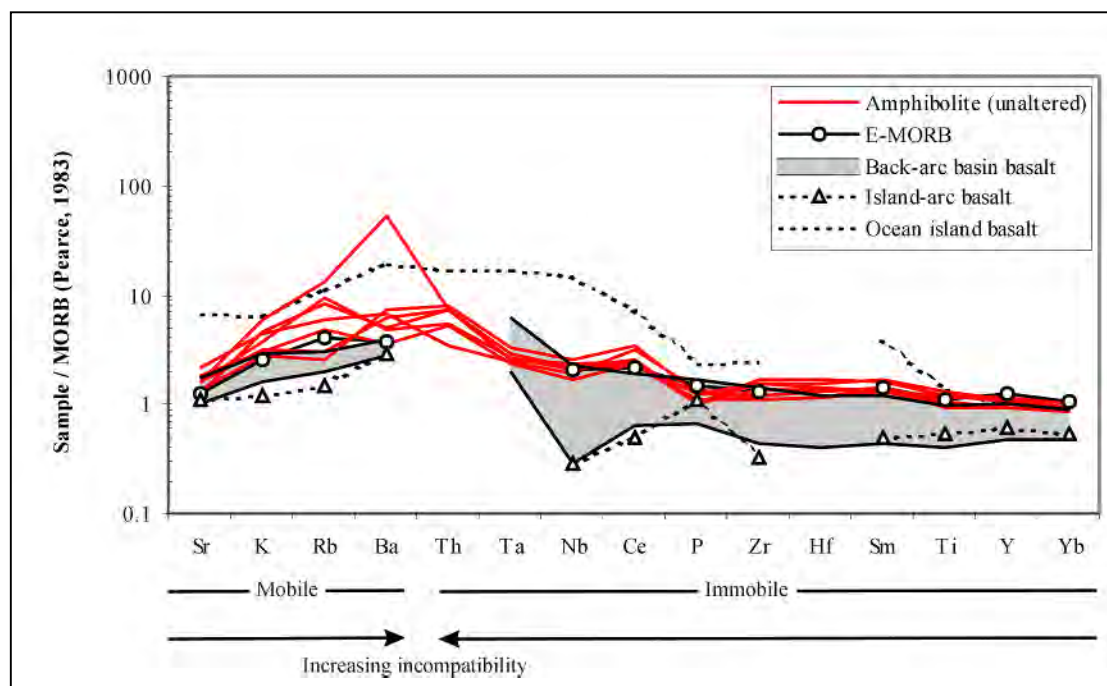


Figure 3.10: Normal mid-ocean ridge basalt (N-MORB) normalised trace element variation diagram of the unaltered Haveri amphibolites compared to a range of tholeiites from different tectonic environments. N-MORB values taken from Pearce (1983). E-MORB values taken from Humphries et al. (1985). The range of back-arc basin values is taken from the East Scotia Sea (Saunders and Tarney, 1979; Wilson 1989). The values for the island-arc basalts is taken from the South Sandwich island-arc (Luff, 1982, in Wilson 1989). The values for the ocean island basalt are average OIB values from Sun (1980).

Tholeiitic basalts are assumed to be the result of high degrees (15 – 30 %) of partial melt (Green and Ringwood, 1967; Sun et al., 1979; Wilson 1989), which results in incompatible element abundances being similar to their source (Wilson, 1989). Thus the very obvious enrichments in the most incompatible elements with respect to N-type MORB must reflect a similar enrichment in the source area. MORB normalised multi-element plots with a pattern of incompatible enrichment are distinctive of tholeiites from ocean islands, plume-ridge segments of MORB (P-type or E-type MORB) and some tholeiites associated with back-arc basins (Wilson, 1989).

N-type and E-type MORB represent the two end-member compositions of a continuous spectrum of magmas with transitional types occurring between them. N-type MORB is considered to be derived from a fairly homogeneous, well mixed, asthenospheric upper-mantle reservoir depleted in the highly

incompatible elements Rb, Ba, K, Th, La, Ce, Nb and Ta (Saunders and Tarney, 1984). It is generally accepted that the low abundances of these elements, relative to other less incompatible species (e.g. Zr, Hf and HREE), implies derivation of parental magmas from a source which has undergone previous melt extractions (Saunders and Tarney, 1984). E-type MORB is derived from sources containing variable amounts of a hotspot component from a lower, heterogeneous mantle reservoir (which is also the source of oceanic island basalts (OIB)), mixed with the depleted N-type MORB source (Wilson, 1989). E-type MORB is strongly enriched in LIL elements and contains high abundances of K, Rb, Th, Ba and LREE, and enhanced abundances of Nb and Ta (Saunders and Tarney, 1984).

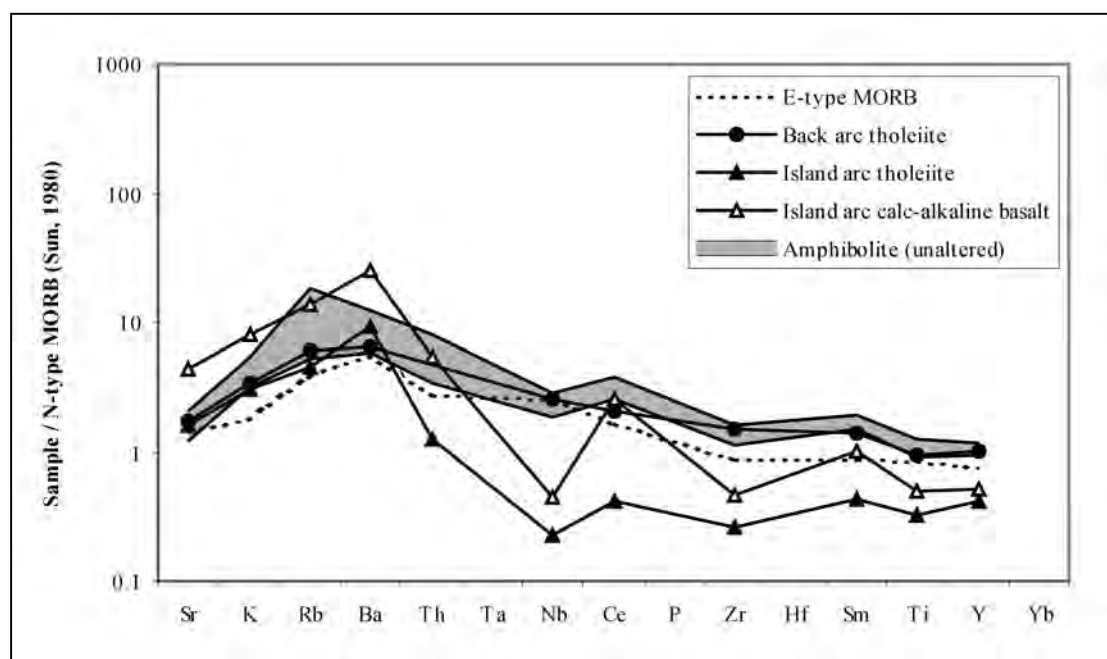


Figure 3.11: N-type MORB normalised trace element patterns for a variety of basaltic rocks. Also included is the range of data for the unaltered Haveri amphibolites. The high Ba sample (TS22) has been removed for clarity. Data is from Sun (1980, Wilson 1989) and Hawkesworth et al (1977, Wilson 1989). After Pearce (1983).

Island-arc magmas have a characteristic geochemistry, being enriched in the LIL elements and depleted in the immobile HFS elements relative to MORB. Furthermore, island-arc volcanic rock suites commonly contain a variety of magma series, with distinct variations in their geochemistry. Low-K island-arc tholeiites are characterised by a MORB-normalised pattern (Figure 3.11) of LIL enrichment, and an overall depletion in the HFS elements, resulting in the Ta to Yb portion of the pattern lying parallel to MORB but at a lower level (Pearce, 1983). The calc-alkaline island-arc basalts have a more “spiked” pattern on a MORB-normalised trace element diagram (Figure 3.11), with the LIL elements displaying even greater enrichment. In addition Ce, P and Sm are also enriched. However, the immobile elements Ta, Nb, Zr, Hf, Ti, Y and Yb still define a relatively flat trend parallel to the MORB pattern (Wilson, 1989), although Ta and Nb show a more pronounced negative anomaly, due to the adjacent enrichment of Th and Ce.

The generation of arc magmas, (with LIL enrichment and HFS depletion relative to N-type MORB) is apparently not possible through partial melting and fractionation of N-type MORB. They are considered most likely to be caused by the enrichment or metasomatism of the mantle source area by LIL-enriched fluids derived from the dehydrated subducted oceanic crust, leaving the HFS elements bonded within the residual crystal lattices (Saunders and Tarney, 1984). Thus the mantle wedge becomes enriched in LIL elements, H₂O and other slab-derived volatiles, producing the characteristic high LIL / HFS ratios seen in arc basalts. Repeated melt extraction from the mantle wedge depletes the wedge in both LIL elements and HFS incompatible elements, but the LIL elements are selectively replenished from dehydration of the subducted slab, thereby decreasing the abundances of the HFS elements more rapidly than the LIL elements.

The Ta and Nb depletion often seen in MORB-normalised plots of island arc basalts may, as mentioned simply be a function of the element arrangement in the diagram, and be a result of relative Th and Ce enrichments (being located either side of Ta and Nb on the plot). However, a number of alternative explanations have been proposed for these depletions. Under conditions of high P_{H₂O} and P_{O₂}, some HFS elements will be stabilised in minor mineral phases (e.g Zr in zircon, and Ta and Nb in rutile, ilmenite or sphene). These may then be retained in the source during melting (Saunders et al., 1980; Saunders and Tarney, 1984; Wilson 1989). Alternatively, it has been suggested that the upper mantle as a whole is subject to re-enrichment of the LIL elements and Ta and Nb (creating E-type MORB and OIB). In the case of an island-arc, the re-enrichment of the mantle wedge is physically prevented by the presence of the subducted slab (Saunders and Tarney, 1984).

Tholeiitic basalts from back-arc basins also show a wide range of compositions, and can be considered to contain a mixture of MORB-like and arc-like characteristics (Wilson, 1989). This is to be expected, as the back-arc basin environment is one in which fluids from the subducting slab are likely to become involved in the magma generation process, producing basalts with geochemical characteristics transitional to arc basalts (Wilson, 1989). Tarney et al. (1977) have suggested that the composition of the back-arc basin tholeiite is likely to have greater arc-like affinities early in the basin's evolution, and then diminish as the basin widens, although this will depend on the geometry of the particular subduction system, and be influenced by such factors as marginal basin width, spreading rates, maturity of the subduction zone and the angle of subduction (Saunders and Tarney, 1984).

The Haveri amphibolites (Figure 3.10) display a MORB-normalised trace element pattern that closely corresponds to both E-type MORB and to certain back-arc basin compositions. However, caution must be exercised when comparing absolute trace element abundances in this way, as even relatively small variations in the degrees of partial melting will produce variations in the trace element abundances (Wilson, 1989). Consequently, Wilson (1989) has suggested that the use of trace element ratios are perhaps more reliable petrogenetic indicators for mantle source compositions.

	Haveri amphibolites (unaltered)	N-type MORB (depleted)	E-type MORB (enriched)	East Scotia Sea back-arc basin basalts	Oceanic- island tholeiite
K/Rb	256 – 673	1046	414	400 – 800	400
K/Ba	7 – 56	109	34	40 – 60	25 – 40
Rb/Sr	0.031 – 0.135	0.0079	0.043	0.025 – 0.04	0.01 – 0.05
Zr/Rb	6 – 26	37	6	16 – 54	6 – 15

Table 3.3: Trace element characteristics of the unaltered Haveri amphibolites compared to back-arc basin basalts from the East Scotia Sea, N-type and E-type MORB and oceanic island basalts (OIB). After Wilson, (1989); data from Saunders & Tarney (1979) and Basaltic Volcanism Study Project (1981).

Table 3.3 shows trace element ratios of the unaltered Haveri amphibolites compared to tholeiites from other environments. Despite the wide spread of data for the Haveri amphibolites, the data suggest an affinity closest to the back-arc basins. However, the elements used are particularly susceptible to mobilisation, and consequently caution must again be exercised when interpreting the data. Low-temperature interaction between seawater and basalts has been recognised as causing LIL element enrichment, and in extreme cases, even LREE enrichment. Saunders and Tarney (1984) showed that low-grade alteration of basalts causes a strong enrichment in K and Rb (no data for Ba), and a lesser enrichment in Sr. The HFS elements (Nb to Yb) and Th were essentially immobile. The relative immobility of Th under low-grade conditions has been used to aid recognition of secondary alteration, whereby an altered basalt with LIL element enrichment displays a negative Th anomaly, relative to the general N-type MORB normalised pattern. On the basis of the absence of a negative Th anomaly for the Haveri basalts (Figure 3.10), it is assumed that the trace element patterns observed are not the result of seawater alteration. However, the mobility of the LIL elements and Th is uncertain in meta-igneous suites affected by higher grades of metamorphism, and consequently there is a possibility that the trace element patterns have been affected by metamorphism.

A back-arc basin interpretation for the geotectonic environment of formation of the Haveri Formation amphibolites is consistent with the regional geological model for the development of the Tampere Schist Belt within the Svecofennian (Section 1.3). Lahtinen and Huhma (1997) envisage a subcontinental lithospheric mantle and thermal boundary layer beneath a partially evolved continental nucleus surrounded by more juvenile island-arc accretions (i.e the CFGC and TSB) before 1.91 Ga. These acted as traps for fluids and small amounts of melts derived from fluid release and melting of the slab. The enriched reservoir that was formed was the source for the Haveri metalavas generated in an initial rift phase in a marginal basin. The location of Haveri Formation on the northern margin of the TSB is also consistent with a marginal basin setting in an evolving arc on a continental margin.

3.1.2.2 Quartz-Feldspar Porphyry

The quartz-feldspar porphyries show a spread between andesite and dacite when plotted on the total-alkalis–silica (TAS) plot (Figure 3.1). However, there is a cluster of three points within the dacite field, which correspond to fresh, relatively unaltered samples taken from fairly thick (>5m), massive porphyries. The two samples which plot in the andesite – dacite and andesite areas of the diagram, are both samples from thin (<2 m) porphyry bodies. Consequently these samples are likely to be more susceptible to hydrothermal alteration, and indeed, like the altered amphibolites, they show a relative silica depletion. It is interesting to note that the sample of uraltite porphyry also plots along this trend of alteration, and may be just a more intensely altered quartz-feldspar porphyry, rather than a separate rock type. The porphyry sample plotting with a high alkali content was a sample of porphyry that had visible potassic alteration.

The unaltered quartz-feldspar porphyries are dacitic in composition and lie within the field of sub-alkalic or tholeiitic series in Figure 3.1. On the AFM diagram (Figure 3.3), the porphyries can be seen to be of the calc-alkaline series, and of a different trend than the Haveri amphibolites, suggesting that the quartz feldspar porphyries are not a fractionation product related to the amphibolites.

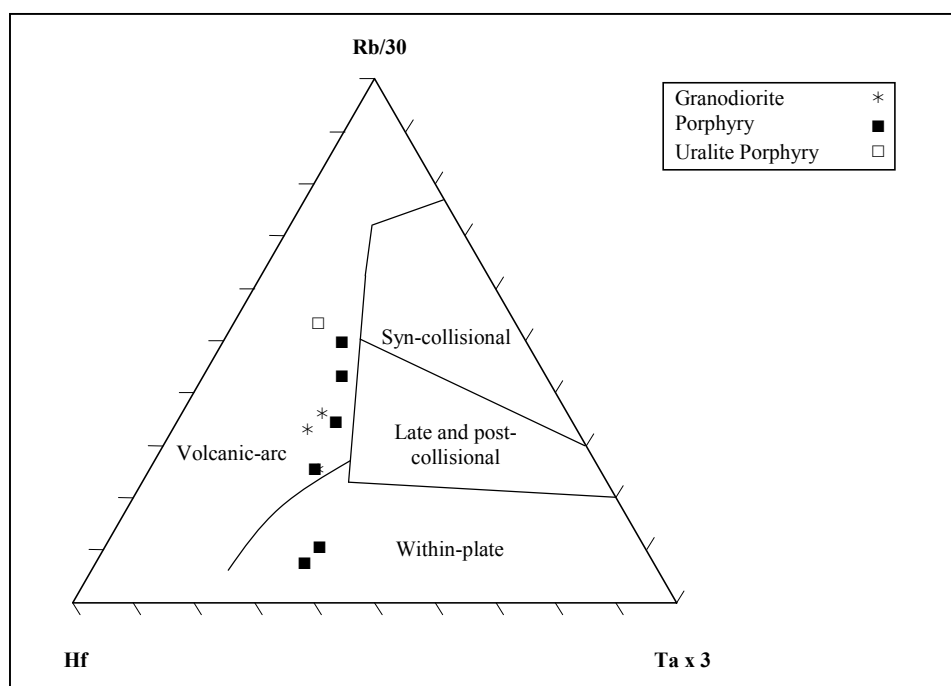


Figure 3.12: The Haveri felsic igneous rocks plotted on the Hf-Rb-Ta discrimination diagram for granites after Harris et al. (1986)

When plotted on the trace element discrimination diagrams (Pearce et al., 1984; Harris et al., 1986), it is clear that the porphyries have tectonic affinities with volcanic-arc granites (Figures 3.12 – 3.14). Although the discrimination diagrams were developed for plutonic granitic rocks, Twist and Harmer (1987) applied them to the Proterozoic Bushveld granites and felsitic lavas, and demonstrated that the diagrams are applicable to felsitic lavas, provided that they are uncontaminated (Rollinson, 1993).

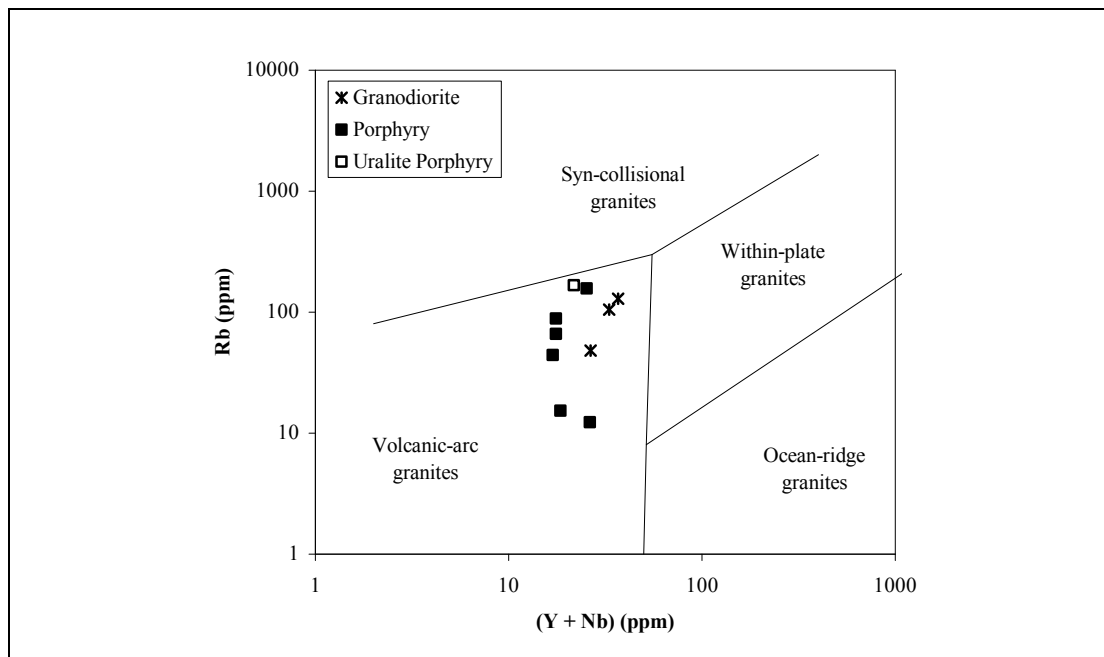


Figure 3.13: The Haveri felsic rocks plotted on the Rb-(Y + Nb) discrimination diagram of Pearce et al. (1984)

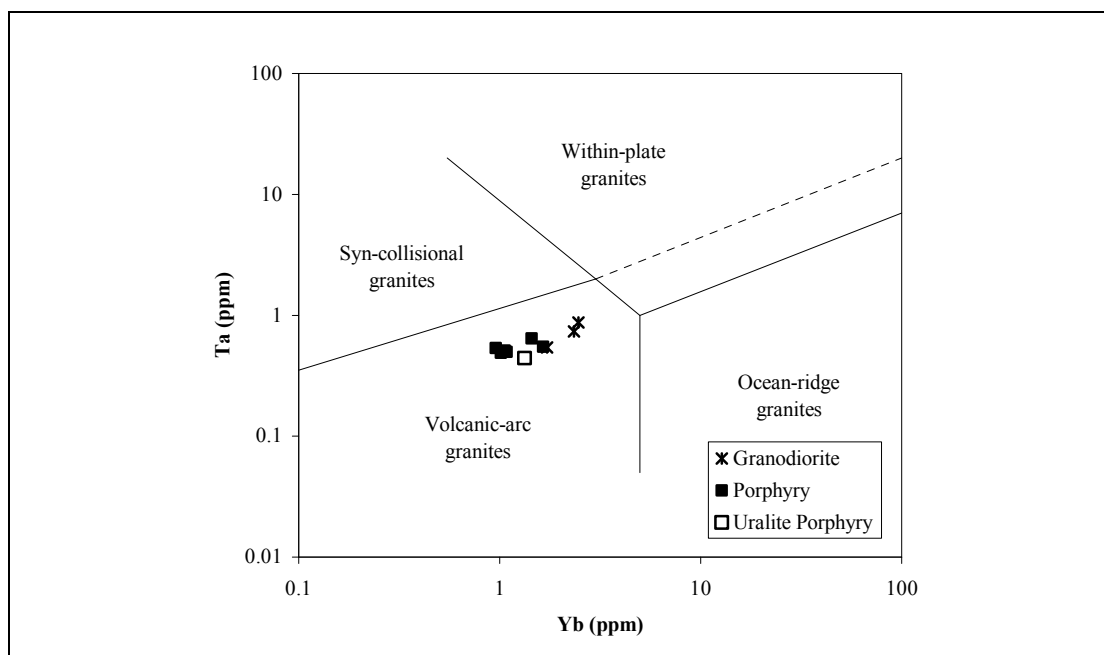


Figure 3.14: The Haveri felsic rocks plotted on the Ta-Yb discrimination diagram of Pearce et al. (1984)

Discrimination diagrams for granitic rocks, like other discrimination diagrams, are susceptible to errors caused by element mobility and crystal fractionation (Rollinson, 1993). Diagrams which use the mobile LIL elements (e.g. Rb) as discriminating elements must be treated with caution, and the results from

such diagrams should only be used in conjunction with other indicators (such as diagrams utilising immobile elements) when interpreting the geotectonic environment of formation for the rocks. Pearce et al. (1984) also identified crystal accumulation of plagioclase as responsible for shifting within-plate or ocean-ridge granites into the volcanic-arc field, and the accumulation of ferromagnesian minerals and minor phases for shifting syn-collisional or volcanic-arc granites into the within-plate or oceanic granite field. Consequently, discrimination diagrams are a tool that must be used with caution, and ideally as a supplement to other sources of geological information.

Models to explain the petrogenesis of basic and ultrabasic rocks are simple compared to those required to explain the petrogenesis of intermediate and silicic rocks. Andesites, dacites and granitic rocks can be derived from a multitude of sources, and by a variety of crystallisation paths (Cullers and Graf, 1984). Simple one-stage melting or crystallisation models are unable to satisfactorily explain the trace element data observed in these more evolved rocks, and it becomes necessary to invoke at least two-stage models including combinations of melting, crystallisation, magma mixing, assimilation and volatile transport (Cullers and Graf, 1984). Furthermore, as magmas evolve and become more silicic, the partition coefficients (K_D values) change, and small amounts of accessory minerals concentrate the REE and other trace elements, severely limiting the use of these elements in interpretation of igneous processes (Cullers and Graf, 1984). Although the melt composition is the most important single factor controlling mineral/melt partition coefficients, other factors such as temperature, pressure and oxygen fugacity can all significantly change the partition coefficients (Rollinson, 1993). Consequently, trace element modelling and interpretation of trace element data become increasingly difficult.

The porphyritic texture of the quartz-feldspar porphyries at Haveri indicates the presence of a high-level magma chamber and the likely occurrence of crystal fractionation (and perhaps other processes) of a more basic melt, producing the observed andesitic to dacitic compositions. Therefore, in contrast to the REE patterns of the Haveri basalts, the REE patterns of the porphyries are unlikely to represent the compositions of the source. In rocks of the calc-alkaline series, as the SiO_2 content increases, the LREE behave as incompatibles and become increasingly enriched, whereas the HREE remain constant or decrease with fractionation as they become accommodated within amphibole, biotite and zircon.

The chondrite-normalised REE patterns produced by the thicker, unaltered porphyries (Figure 3.15), are moderately enriched in the LREE ($[\text{La/Yb}]_N$ of 7.0 – 8.4) relative to the HREE and display a weak negative europium anomaly ($[\text{Eu/Eu}^*]_N$ of 0.8 – 0.9). The LREE display a greater degree of fractionation ($[\text{La/Sm}]_N$ of 3.5 – 4.2) relative to the HREE ($[\text{Gd/Yb}]_N$ of 1.2 – 1.4). Indeed the heaviest of the HREE display a positive anomaly between Ho and Lu ($[\text{Ho/Lu}]_N$ of 0.8). The weak negative europium anomaly is most likely caused by the fractionation, and removal from the melt, of minor amounts of plagioclase. The relative depletion and “dish” shape seen within the HREE most likely reflects hornblende (and to a lesser extent clinopyroxene) fractionation. In felsic and intermediate liquids, the highest partition coefficients for hornblende (and to a lesser extent clinopyroxene) are between Dy and Er. Consequently, the fractionation and removal of hornblende would preferentially

deplete the remaining melt in these elements, producing the observed pattern (Rollinson, 1993). Hornblende fractionation also produces a positive europium anomaly, thus reducing the effect of a negative anomaly caused by plagioclase fractionation (Rollinson, 1993).

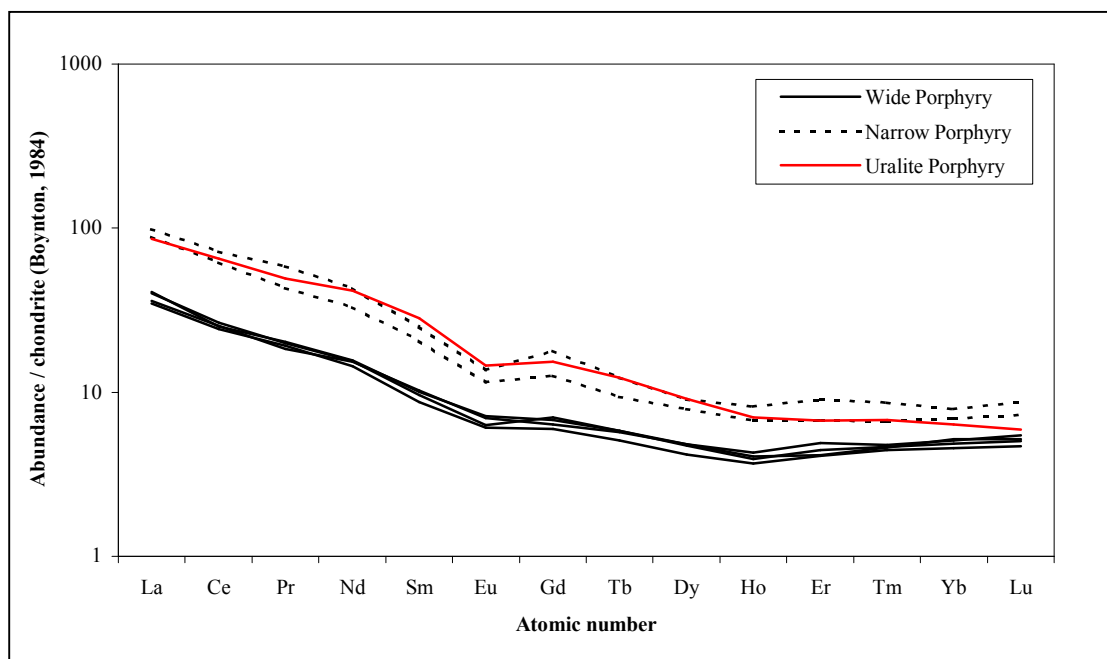


Figure 3.15: Chondrite-normalised REE patterns of the Haveri porphyritic rocks. Chondrite values taken from Boynton (1984)

The chondrite-normalised REE plot of the porphyries (Figure 3.15) shows that there are two distinct REE patterns, whereby one group has higher overall abundances than the other. These two patterns show the same differentiation between samples as in the TAS diagram (Figure 3.1), whereby the three samples from thicker porphyries correspond closely with each other, and the two samples from narrow porphyry bodies have REE patterns similar to each other and to the uralite porphyry (Table 3.4). This supports the notion that the wider porphyries are unaltered, whereas the narrower porphyries have been affected by hydrothermal alteration. Furthermore, the similar patterns displayed by the narrow porphyries and the uralite porphyry, imply a genetic relationship between the uralite porphyry and the quartz-feldspar porphyries.

	Wide porphyries	Narrow porphyries	Uralite porphyry	Granodiorite
[La/Yb] _N	7.0 – 8.4	12.5 – 12.9	13.6	10.0 – 10.7
[Eu/Eu*] _N	0.9 – 0.9	0.6 – 0.7	0.7	0.6 – 0.7
[La/Sm] _N	3.5 – 4.2	3.9 – 4.3	3.0	3.1 – 4.2
[Gd/Yb] _N	1.2 – 1.4	1.8 – 2.3	2.4	1.5 – 1.9
[Ho/Lu] _N	0.8	0.9	1.2	1.0 – 1.1

Table 3.4: A comparison of the rare earth element (REE) ratios for the various felsic rock types encountered at and around Haveri.

Both the narrow porphyries and the uraltite porphyry show a greater degree of REE fractionation than the wide, unaltered quartz-feldspar porphyry, and both have larger europium anomalies, suggesting the fractionation of feldspar. However, the narrow porphyries exhibit a far greater fractionation of the LREE than the HREE, which is much less pronounced in the uraltite porphyry. Although there are many fractionating phases that could generate these patterns, likely mineral phases include pyroxenes, hornblende and garnet.

The Haveri porphyritic rocks are of the calc-alkaline series (Figure 3.3), and consequently have different geochemical characteristics compared to the tholeiitic Haveri amphibolites. The trace element compositions of calc-alkaline basalts show a distinctive pattern of not only alkali and alkaline earth element enrichment, but also an enrichment in Th, Ce, P and Sm, whilst the other incompatible elements Ta, Nb, Zr, Hf, Ti, Y and Yb remain at low abundances (Pearce, 1982). Furthermore, as fractional crystallisation progresses to produce the more evolved compositions of andesites and dacites, the elements Rb, Ba, Th, Nb, Zr and the LREE all behave as incompatibles and become enriched in the melt (Figure 10 in Pearce, 1982), whereas Y and the HREE remain relatively constant or even decrease as they become accommodated within amphibole, biotite and zircon (Pearce, 1982).

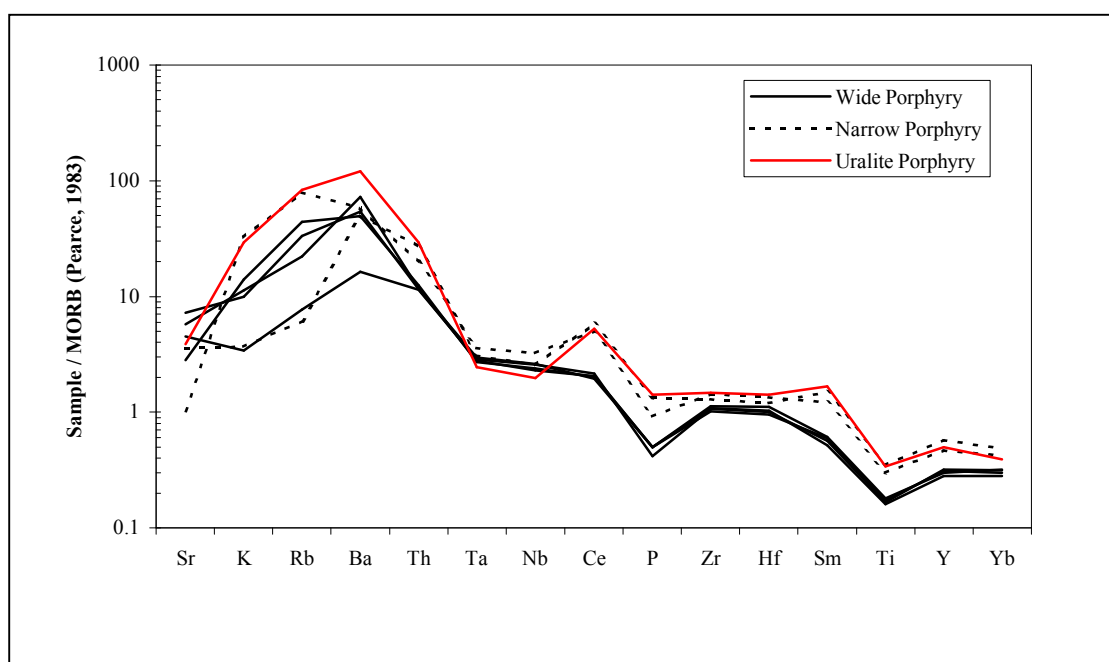


Figure 3.16: MORB-normalised trace element variation diagram of the Haveri porphyritic rocks. Normalising values are for N-type MORB (Pearce, 1983)

The LIL enrichment, and steep slopes from Th to Ta and Nb on the MORB-normalised trace element plot (Figure 3.16) are typical characteristics of arc-generated magmas (Pearce, 1982; Kähkönen and Nironen, 1994), and, combined with the calc-alkaline character of the magma and the geotectonic plots, strongly indicate an arc-type origin.

The MORB-normalised trace element plot for Haveri porphyries (Figure 3.16) again shows a differentiation between the three samples from the wide unaltered porphyries and the samples from the

two narrow porphyries. The patterns for the three wide samples show many of the characteristics associated with calc-alkaline suites described above. The mobile LIL elements (K, Rb and Ba) and Th all show distinct enrichments, whereas P and Ti both display considerable depletions. Ce and Sm display no obvious enrichment or depletion. The Th enrichment is considered to represent both the natural Th enrichment of the calc-alkaline suite combined with a Th enrichment associated with fractional crystallisation. The P enrichment expected in the calc-alkaline suite has been offset by the fractional crystallisation of apatite, causing an overall depletion in P. Similarly the Ti depletion is caused by the fractional crystallisation of magnetite. The reasons for the absence of any Ce or Sm enrichment expected in the calc-alkaline suite are unclear.

The uraltite porphyry and the two samples from narrow porphyries display higher abundances of the mobile LIL elements, as well as overall higher abundances in the immobile elements. Furthermore, these samples also display the expected enrichments in Th, Ce and Sm, and possibly in P (when offset against apatite crystallisation). A number of processes are possible for increasing the overall abundances observed in both the trace elements and in the REE plots, including fractional crystallisation and magma mixing with more evolved melts. However, fractional crystallisation alone is incompatible with the dark matrix, lower SiO₂ contents, and higher amphibole contents associated with these samples. It is considered likely that the geochemistry of the uraltite porphyry and the narrow porphyries is a result of an enrichment process that has subsequently been modified by hydrothermal alteration.

The reasons why the wide, unaltered porphyry samples don't display any Ce or Sm enrichment as expected in the calc-alkaline suite, whereas the uraltite porphyry and the narrow porphyry samples do, are unclear, as the cause for the selective enrichments of these elements in calc-alkaline suites is as yet not understood (Pearce, 1982).

3.1.2.3 Granites and Granodiorites

The plutonic rocks in the vicinity of Haveri can be classified on the TAS diagram (Figure 3.1) as diorites and quartz diorites / granodiorites (Wilson, 1989). The granodiorites are high-K calc-alkaline in character, whereas the diorite is medium-K calc-alkaline (Figure 3.2). The sample of dioritic composition is taken from an outcrop close to the contact with the amphibolites, 500 m northwest of the Haveri mine, and may represent a separate phase within a pluton of granodioritic composition (Kähkönen and Nironen, 1994). The fresh samples of granodiorite were taken from boreholes to the east and south east of the Haveri mine. The granodiorites represent the main plutonic phase in proximity to the study area, and veins of granodiorite have been observed cutting the supracrustal succession within the old mine. The microcline granite that is encountered beyond the granodiorites further into the Central Finland Granitoid Complex has not been analysed in this study.

The granodiorites and the diorite all plot within the volcanic-arc field on the tectonomagmatic discrimination diagrams (Figures 3.12 – 3.14). The field of volcanic-arc granites includes granites from both oceanic arcs dominated by tholeiitic or calc-alkali basalts as well as granites from active continental margins (Pearce et al., 1984).

The use of trace element and rare earth element geochemistry in deciphering the petrogenesis of plutonic rocks is considerably more complex than in the case of basaltic rocks. Granitic melts can be produced by melting a variety of sources, and the resultant melts can be modified by variable combinations of crystal fractionation, magma mixing, liquid immiscibility, assimilation of wall rocks and evolution of volatiles, all of which can drastically alter the trace element and REE contents of the melt (Cullers and Graf, 1984). Furthermore, as the melts evolve and physico-chemical conditions change, different mineral parageneses occur, commonly with huge (and sometimes unknown) variations in the distribution coefficients (D values) for the REE in particular. However, in spite of these problems, REE studies of granitoids, especially when used in conjunction with field relations integrated with petrographic and major element studies, can be of use in limiting the petrogenetic possibilities (Cullers and Graf, 1984).

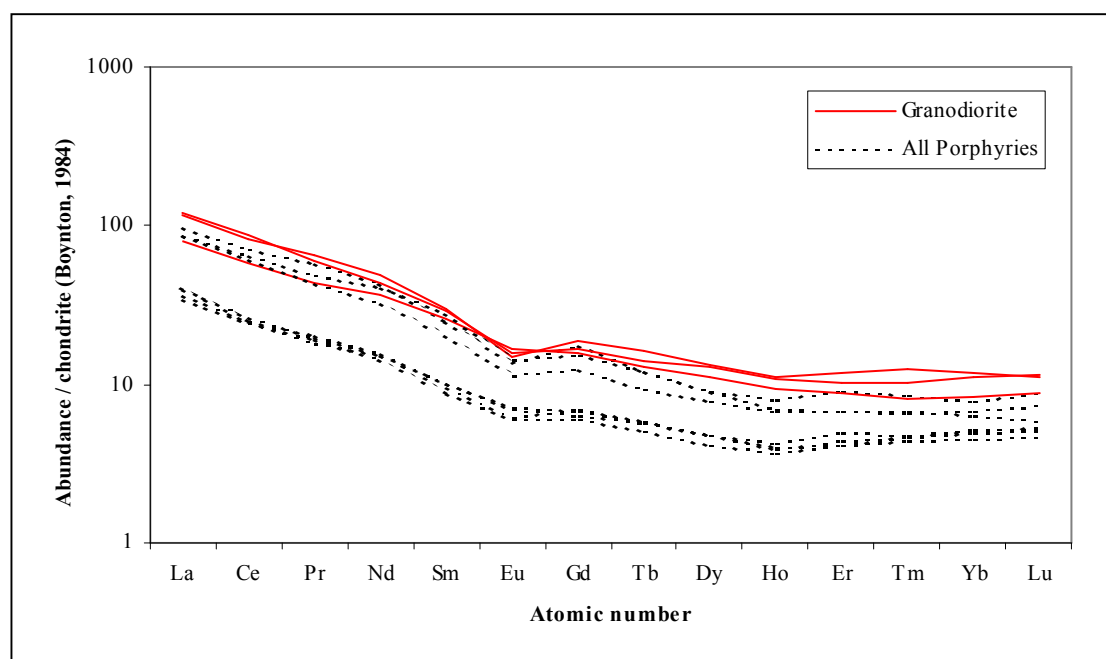


Figure 3.17: Chondrite-normalised REE patterns of the Haveri granodiorites and porphyritic rocks. Chondrite values taken from Boynton (1984).

The samples of granodiorite and diorite at Haveri (Figure 3.17 & Table 3.2) define a REE pattern which is considerably enriched in the LREE relative to the HREE ($[La/Yb]_N$ of 10.0 – 10.7).

Furthermore, the LREE have been fractionated to a far greater degree than the HREE, and there is a moderately sized negative europium anomaly.

Negative europium anomalies in granitoids can be produced by either partial melting of a plagioclase-rich source with residual plagioclase, or fractional crystallisation of plagioclase, or a combination of both (Cullers and Graf, 1984). Potassic alteration with the destruction of plagioclase can also produce a negative europium anomaly (Cullers and Graf, 1984). Granodiorites (with no evidence of potassic alteration) containing negative europium anomalies and significant REE contents and LREE/HREE ratios require sources with abundant residual plagioclase (to produce the Eu anomaly) and minor amounts of garnet, amphibole or pyroxene (to reduce the HREE relative to the LREE). Likely sources are garnet-bearing siliceous granulite, hornblende-garnet-clinopyroxene-plagioclase basic rock or meta-greywacke (Cullers and Graf, 1984). The fractional crystallisation of plagioclase and magnetite is likely to have occurred in the Haveri granodiorites, producing the negative Eu (Figure 3.17) and Ti anomalies (Figure 3.18), respectively. The negative Ti anomaly (Figure 3.18) is consistent with the I-type (magnetite series) of the granodiorites, and implies melting from a subducted slab in an arc environment (Chappell and White, 1974). The LIL enrichment, and steep slopes from Th to Ta and Nb on the MORB-normalised trace element plot (Figure 3.18) are also consistent with arc-generated magmas (Pearce, 1982; Kähkönen and Nironen, 1994).

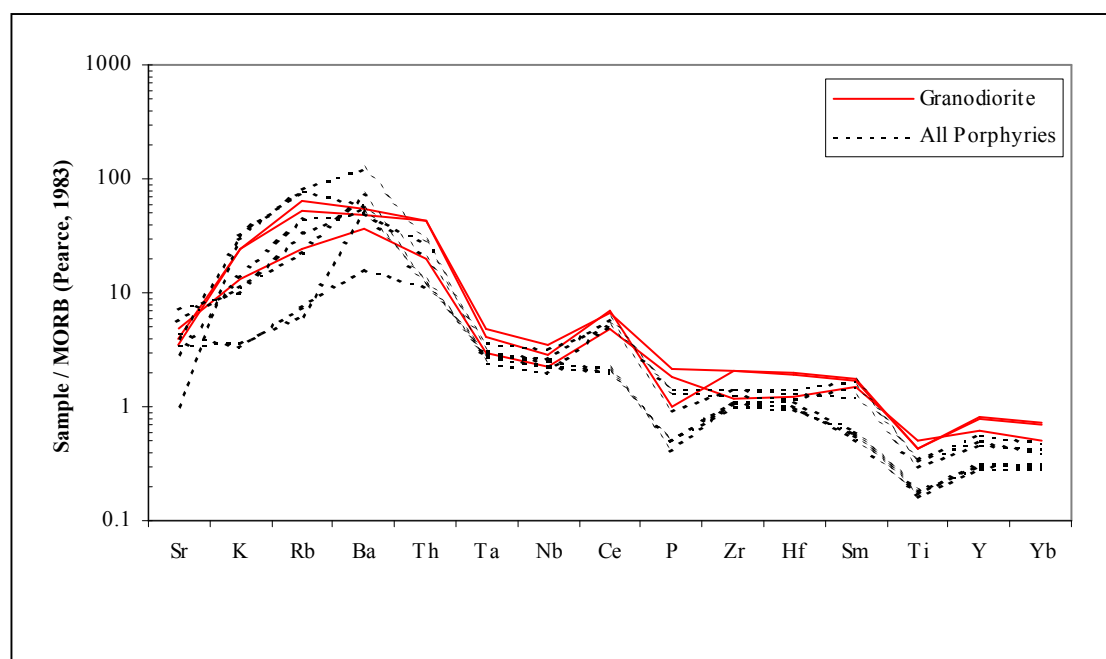


Figure 3.18: MORB-normalised trace element variation diagram of the Haveri granodiorites and porphyritic rocks. Normalising values are for N-type MORB (Pearce, 1983)

The petrogenesis of the granitoids close to Haveri is a study in itself, and beyond the scope of this study. However, detailed studies of the field relationships of the plutons both to geographically close volcanic rocks, as well as zonations within the granitoids themselves, combined with detailed

petrographic and geochemical studies, including isotopic analyses, would provide useful information on the genesis and evolution of the granitoids.

3.1.2.4 Mafic Dykes

The mafic dykes are minor components of the geology at Haveri, and have been intruded relatively late in the geological history of the region. Field relations alone are sufficient to show that there is no genetic relationship between the mafic dykes and the other rock types at Haveri. They have been observed cross-cutting the Haveri amphibolites and metasedimentary rocks, the porphyries, and even the granitoids and some carbonate veins. The dykes are medium-K basaltic andesite (calc-alkaline series) in composition (Figures 3.1 – 3.2) and are close to the tholeiite – calc-alkaline boundary of Irvine and Baragar (1971) (Figure 3.3).

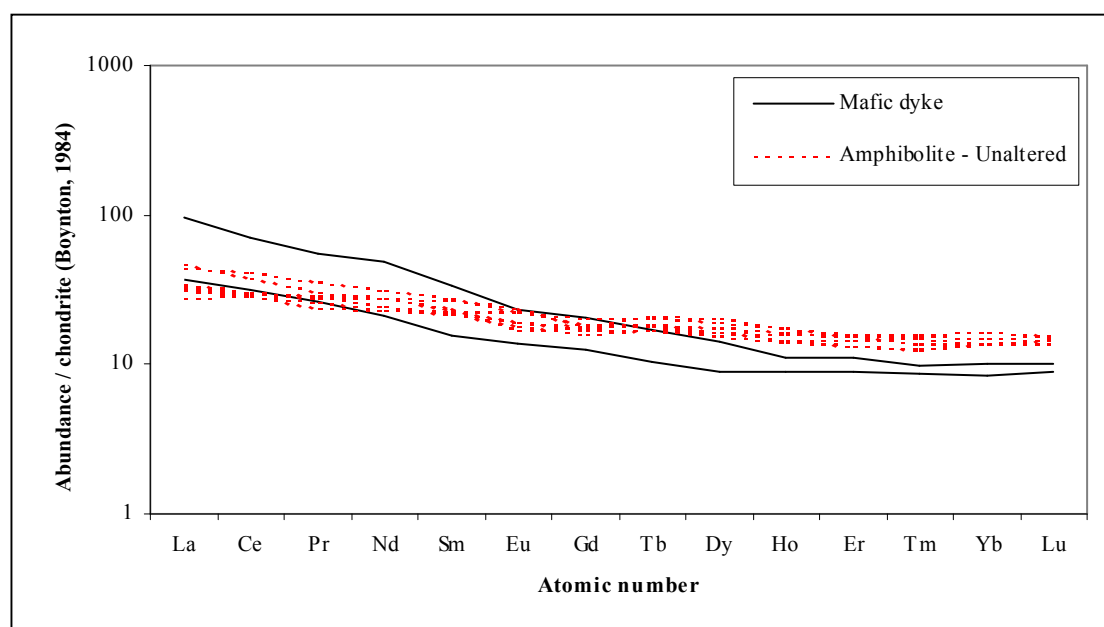


Figure 3.19: Chondrite-normalised rare earth element diagram of the mafic dykes compared to the unaltered Haveri amphibolites. Chondrite values taken from Boynton (1984).

The chondrite-normalised REE patterns of the mafic dykes (Figure 3.19) show considerable variations between the two samples, especially in terms of overall REE abundances. However, there are a number of similarities that can be observed in the REE patterns of the mafic dykes. Both samples show an enrichment in the LREE over the HREE ($[La/Yb]_N$ of 4.5 – 9.4), with a greater degree of LREE fractionation ($[La/Sm]_N$ of 2.3 – 2.9), relative to the HREE ($[Gd/Yb]_N$ of 1.5 – 2.0). There are no discernible europium anomalies ($[Eu/Eu^*]_N$ of 0.9 – 1.0). The variations observed between the two samples could be caused either by differences in the degrees of partial melting, or by a greater degree of fractional crystallisation causing the REE enrichments observed, or by a combination of these two processes.

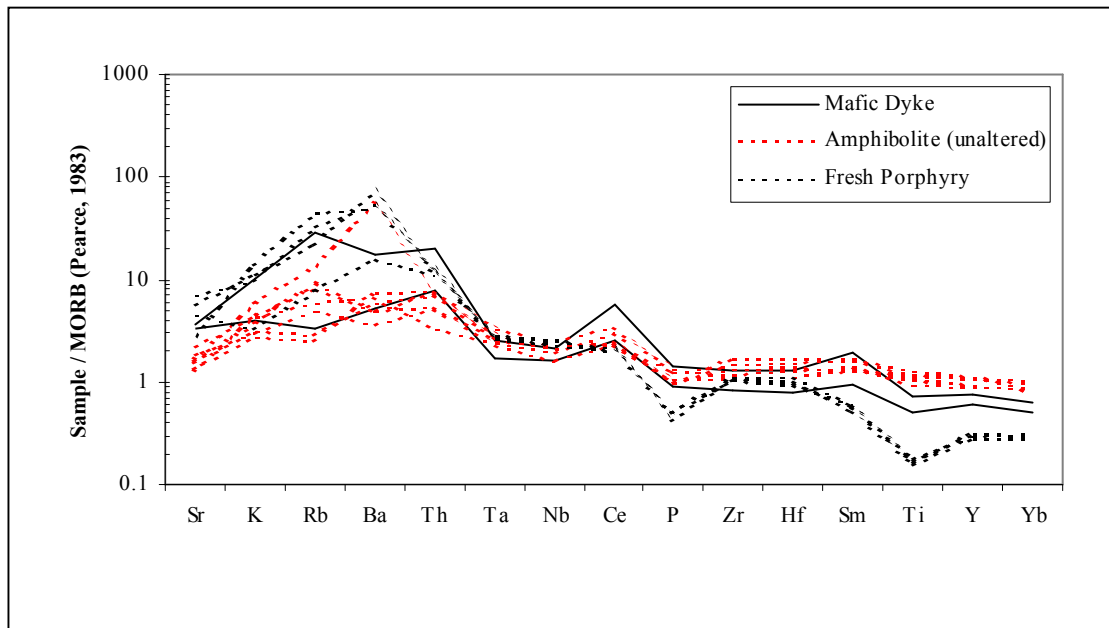


Figure 3.20: MORB-normalised trace element variation diagram of the mafic dykes compared to the unaltered Haveri amphibolites. Normalising values are for N-type MORB (Pearce, 1983)

The MORB-normalised trace element plot for the mafic dykes (Figure 3.20), displays the distinctive calc-alkaline patterns of LIL enrichment, steep slopes from Th to Ta and Nb, and enrichment in Th, Ce and Sm (Pearce, 1982). The absence of any discernible enrichment in P is likely to be caused by the fractionation of apatite. The geochemistry of the mafic dykes confirms the observed field relations, that the mafic dykes are genetically unrelated to the Haveri amphibolites.

3.1.2.5 Metasedimentary rocks

The metasedimentary rocks that have been sampled at Haveri have all been taken from the base of the Osara Formation and from sedimentary intercalations with the Haveri Formation amphibolites and tuffs. The exact stratigraphic positions of the samples are in many cases unclear, due to folding and faulting. However, all of the analysed samples have been taken from close to the contact between the Haveri Formation and the Osara Formation. A consequence of this is that many of the “metasedimentary rocks” are in fact reworked tuffs, and have similar geochemical characteristics to the tuffs.

Kähkönen and Leveinen (1994), in their study of the metasedimentary rocks of the Tampere Schist Belt (TSB), observed that the chemical compositions of the Osara Formation greywackes are similar in chemical composition to the Myllyniemi greywackes, located further east within the TSB. However, the REE pattern for the Osara sample deviates from most of the TSB sedimentary rocks in that it has a relatively high LREE enrichment and a pronounced negative europium anomaly.

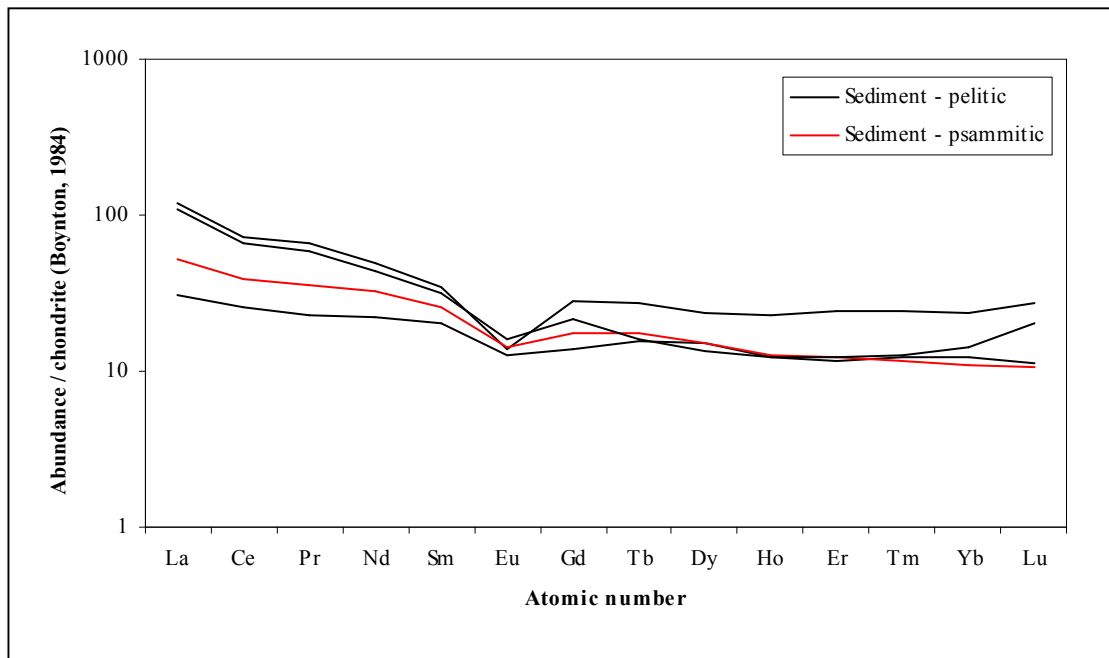


Figure 3.21: Chondrite-normalised rare earth element plot of selected metasedimentary samples from close to the Haveri Formation – Osara Formation contact. Chondrite values taken from Boynton (1984; in Rollinson, 1993)

A number of metasedimentary samples have been selected that are most clearly epiclastic in origin rather than tuffs, and are relatively free of hydrothermal alteration. These samples have been plotted on a chondrite-normalised REE plot (Figure 3.21). The REE patterns for the metasedimentary rocks are enriched in the LREE relative to the HREE ($[La/Yb]_N$ of 2.5 – 7.6), and have a greater degree of LREE enrichment ($[La/Sm]_N$ of 1.5 – 3.4) than HREE enrichment ($[Gd/Yb]_N$ of 1.1 – 1.6). In addition the REE patterns have pronounced negative europium anomalies ($[Eu/Eu^*]_N$ of 0.4 – 0.7). These data therefore confirm the results of Kähkönen and Leveinen (1994). Two of the pelitic samples in Figure 3.21 show a flat to positive slope for the heavier REE, that is likely to be caused by the presence of HREE-enriched minerals such as zircon, garnet or hornblende (which would cause the dish-shaped HREE pattern).

The REE patterns of siliciclastic sedimentary rocks reflect the REE patterns of the source area. Consequently a significant enrichment in LREE relative to HREE, indicates a similar enrichment in the source area reflecting a source area composed of evolved rocks rather than primitive rocks (McLennan and Taylor, 1991). A negative europium anomaly in the metasedimentary rocks either reflects strong weathering of the metasedimentary material, and in particular of plagioclase, or alternatively reflects a europium-depleted source region, with the fractionation of feldspars during crustal formation in the source region. Such a process occurs during cratonisation and intracrustal differentiation (Kähkönen and Leveinen, 1994). As the metasedimentary rocks of the Osara Formation do not display evidence of significant weathering, and (in the psammitic units at least) contain considerable plagioclase, it is likely

that the negative europium anomaly reflects the source region of the sediments. In view of the absence of other evidence for the presence of a nearby craton (Kähkönen and Leveinen, 1994), it is considered likely that the source region for the Osara Formation is from a nearby volcanic arc, mature enough to initiate intracrustal differentiation. The data from the metasedimentary rocks of the Osara Formation have been plotted on the discrimination diagrams for greywackes (Figure 3.22) of Bhatia and Crook (1986). Data from the tuffs have also been plotted, to indicate the effects of an increased volcanoclastic component on the distribution of the data points. Although the data do not fall neatly within the fields outlined by Bhatia and Crook (1986), it does appear that in general, the data points plot closest to the fields for oceanic island-arcs and continental island-arcs. Despite being inconclusive, the geochemistry of the Osara Formation does tend to support a source region consisting of a mature volcanic arc, rather than a cratonic source.

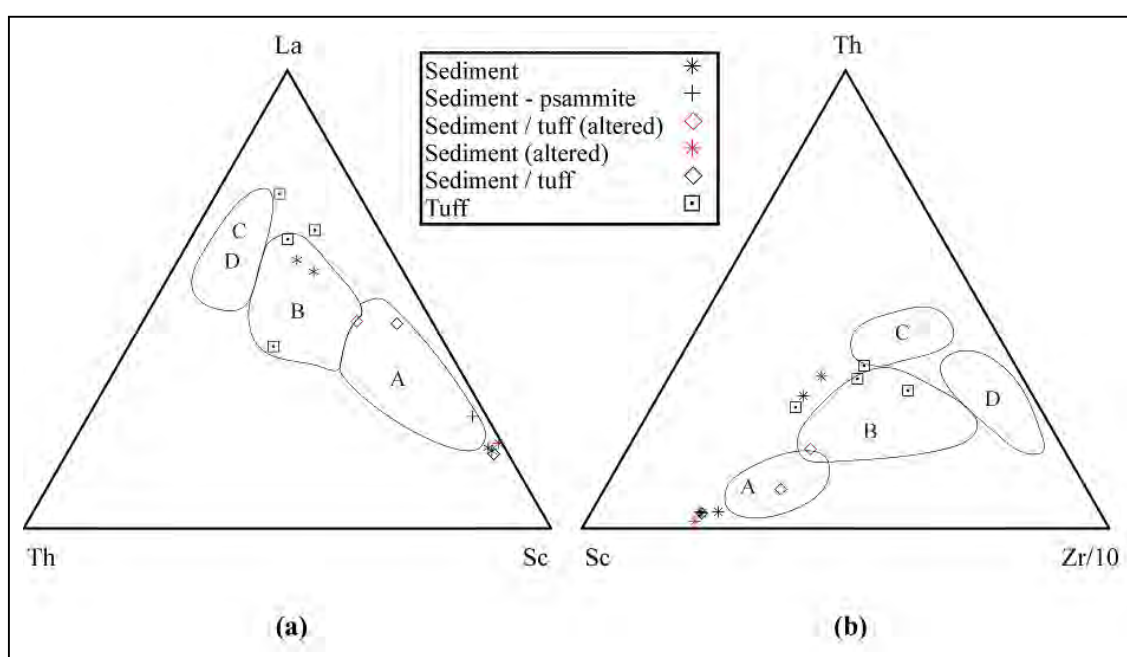


Figure 3.22: (a) La-Th-Sc discrimination diagram for greywackes; (b) Th-Sc-Zr/10 discrimination diagram for greywackes. The fields are: A, oceanic island-arc; B, continental island-arc; C, active continental margin; D, passive margin (after Bhatia and Crook, 1986).

3.1.3 Hydrothermal Alteration

Amphibolitisation is the dominant form of hydrothermal alteration at Haveri, and is principally recognised by the development of calcic amphibole. In the proximal and more intensely altered zones other minerals such as clinopyroxene also become apparent. Figure 3.23 is an enrichment – depletion diagram showing the relative mobility of the major element oxides of the altered amphibolites compared to an average of the unaltered amphibolites. An “average” value for the unaltered amphibolites has been used rather than a single unaltered sample, because this reduces the effect of

contamination caused by the minor amounts of the various alteration types present in most samples. It is apparent that the main enrichments, relative to the unaltered samples, are in Fe (Fe^{2+} and Fe^{3+}), Mn, Ca and K, and the main depletions being Si, Al, Na, Ti and P. The enrichments in Fe^{2+} , Mn and Ca are unsurprising as they are the main elements in calcic amphiboles. The enrichment in Fe^{3+} is most likely controlled by the presence of magnetite in the alteration assemblage, although a minor amount of Fe^{3+} may substitute for Al^{3+} in the amphiboles. The K enrichment is in the more distal (and therefore less intensely altered) alteration zones without pyroxene, and is effectively absent from the proximal-zone alteration type. This is perhaps a manifestation of the biotite alteration as a distal equivalent of amphibolitisation.

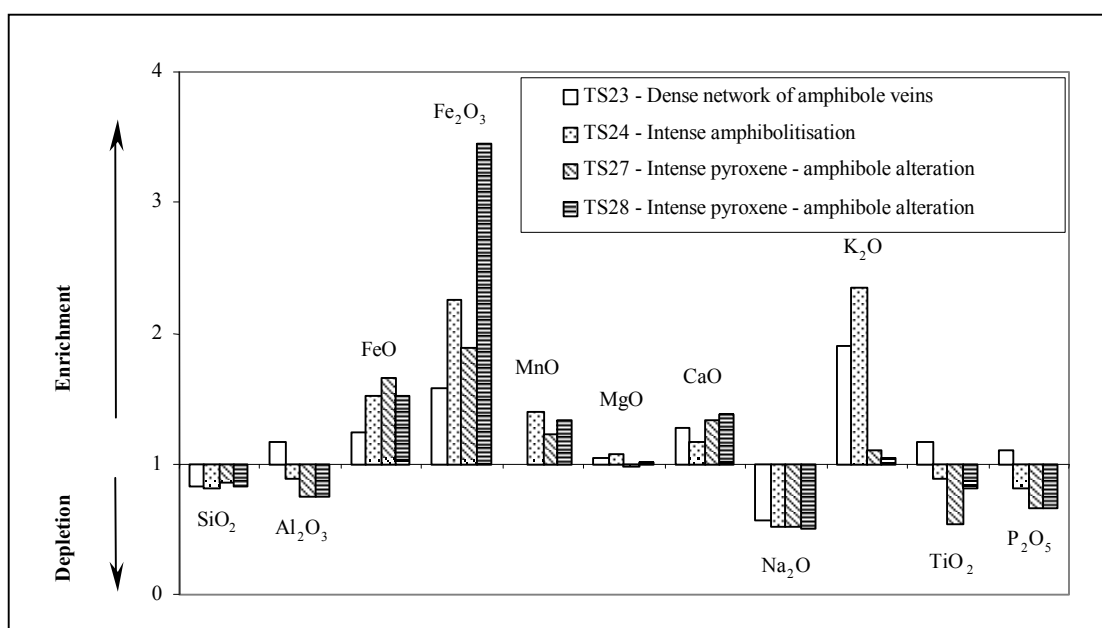


Figure 3.23: Enrichment – depletion diagram for the major element oxides of the amphibolitised Haveri amphibolites. The diagram compares the oxide abundances of the amphibolitised samples to the averages of the unaltered amphibolites.

A similar enrichment – depletion diagram for trace elements and REE is not suitable, due to the wide variation of element enrichments and depletions, and the proximity to detection limits for many of the elements in a number of the base level “unaltered” samples. Consequently it is easier to observe the effects of alteration using the REE plots and MORB-normalised trace element plots (Figure 3.9 and Figure 3.24 respectively).

The chondrite-normalised REE patterns (Figure 3.9) of the altered amphibolites display a greater variation between samples than the unaltered amphibolites, and so caution must be applied when making generalisations about the effects of amphibolitisation. However, a number of observations can be made. The LREE have been slightly enriched relative to the HREE ($[\text{La}/\text{Yb}]_N$ of 2.5 – 5.6), with a greater degree of fractionation within the LREE themselves ($[\text{La}/\text{Sm}]_N$ of 1.7 – 2.7). The fractionation of the HREE essentially remains unchanged ($[\text{Gd}/\text{Yb}]_N$ of 1.1 – 1.4) although there is a perceptible

overall decrease in HREE abundances in three of the four altered samples. Perhaps the most obvious effect of the amphibolitisation is the positive europium anomalies ($[Eu/Eu^*]_N$ of 1.1 – 1.7). Positive europium anomalies are commonly caused by the addition of feldspar, although they can also be caused by the removal of hornblende, pyroxene, garnet or sphene. However, since the alteration assemblage is characterised by the *addition* of calcic amphiboles and pyroxene, it can be concluded that the positive europium anomaly associated with amphibolitisation is most likely caused by the addition of feldspar. This presents a problem, as the whole-rock alteration patterns (Figures 3.23 – 3.24) display Na and Sr depletions, a feature characteristic of feldspar removal. The most likely explanation for this is that the amphibolitisation is accompanied by albitisation, but, due to the predominance of calcic amphibole (containing no Na), the bulk composition of the rock actually reflects a Na decrease with sodic alteration, as there is less plagioclase in the rock (i.e. dilution). This is supported by the plagioclase mineral chemistry (Section 3.2.2.4) which indicates that the plagioclase compositions become more sodic with alteration. In addition the pale, brittle, and brecciated metavolcanic rock (e.g. Figure 2.30), commonly associated with amphibolitisation, is a typical feature of albitisation. Albitisation is a common feature of alteration zones of many deposit types, including mesothermal lode-gold deposits, porphyry deposits, and Fe-oxide Cu-Au deposits. (Eilu et al., 1997, 1999; Pirajno, 1992; Williams, 1999a).

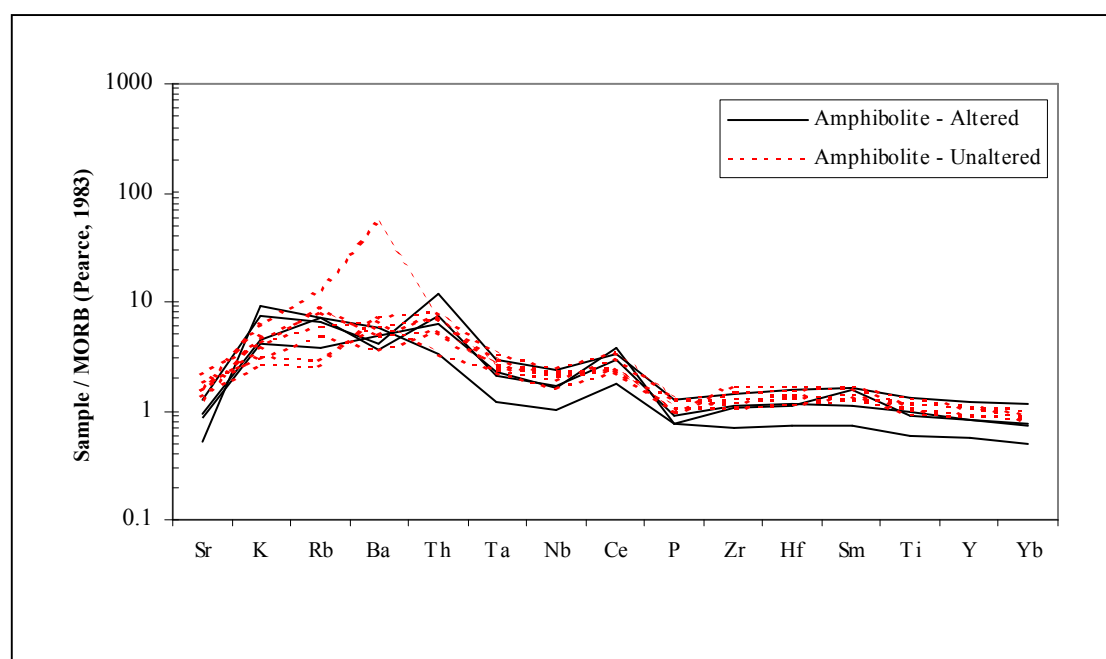


Figure 3.24: MORB-normalised trace element variation diagram of the Haveri amphibolites, showing the effects of amphibolitisation. Normalising values are for N-type MORB (Pearce, 1983)

There are few significant changes in the incompatible trace element geochemistry of the amphibolites with amphibolitisation (Figure 3.24), with the MORB-normalised patterns simply showing a greater variability, and a general enhancement in peaks and troughs in the altered amphibolites than in the unaltered. The most consistent changes are that there is a depletion in Sr, and an enrichment in K (as

previously observed). Although not applicable to all of the altered samples, it is possible to recognise a general depletion in the immobile elements between P and Yb, as well as general enrichments in Th and Ce.

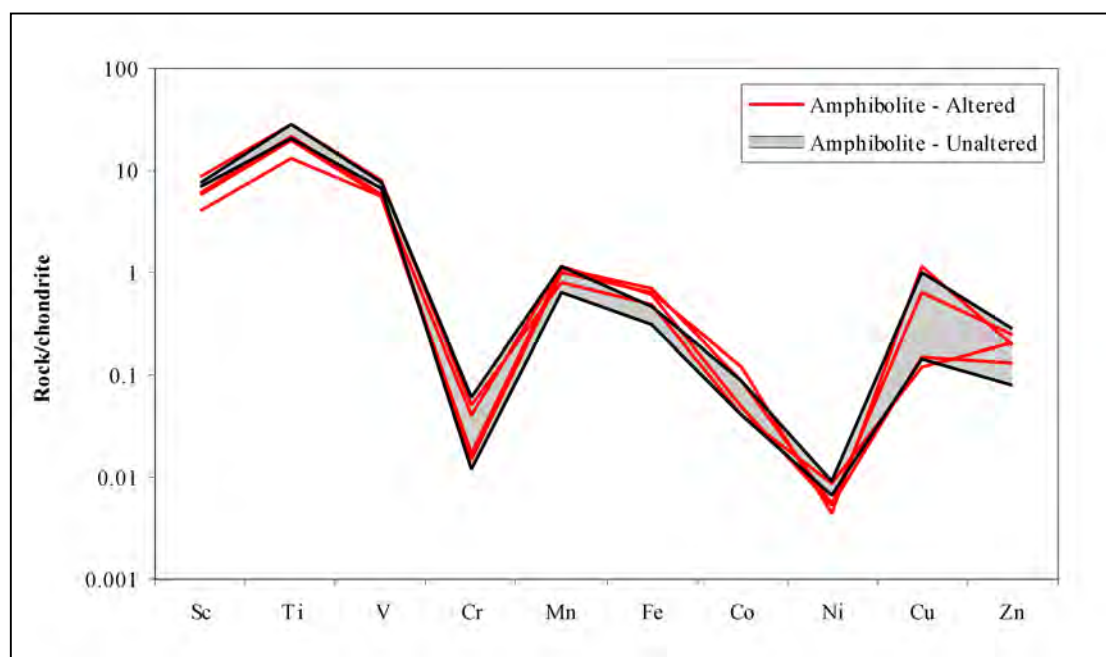


Figure 3.25: Chondrite normalised transition metal concentrations for both altered and unaltered Haveri amphibolites. The normalising values are taken from Langmuir et al. (1977) with the exception of Fe, which was taken from Bougault et al. (1980).

Variations in the concentrations of the transition metals between the altered and unaltered amphibolites are shown in Figure 3.25. The wide degree of variation observed in Cu and Zn for both the altered and unaltered samples is not surprising, as these elements are particularly mobile during metamorphism and alteration. Iron is the only element in which the altered samples show any consistent variation from the unaltered samples. However, even this variation is only a minor enrichment, although this perhaps reflects the sampling regime, which focussed on the silicate alteration, and avoided altered samples with obvious magnetite and sulphide enrichment.

An enrichment – depletion diagram (Figure 3.26) for the altered porphyries, compared to the average of the unaltered porphyries, shows that the main major element enrichments are Fe^{2+} , Mn, Mg, Ca, K, Ti, and P, whereas the main depletions are Fe^{3+} and Na. The enrichments in Fe^{2+} , Mn, Mg, Ca, K and Ti are likely to be controlled by the addition of amphibole and biotite, although Ca and Ti may also be controlled by the addition of sphene. Enrichments in P are likely to be caused by apatite additions. The depletion in Na is most likely caused by the removal of plagioclase and its replacement in the matrix by the darker ferromagnesian minerals. The depletion in Fe^{3+} is probably caused by the removal of magnetite. The Na depletion and K enrichment is not evident in sample TS14, which actually shows a minor enrichment in Na and a K depletion. This sample is from a narrow porphyry in the “R7 trench”

north of the Haveri mine, which does not display a dark matrix, but rather still has a high plagioclase content.

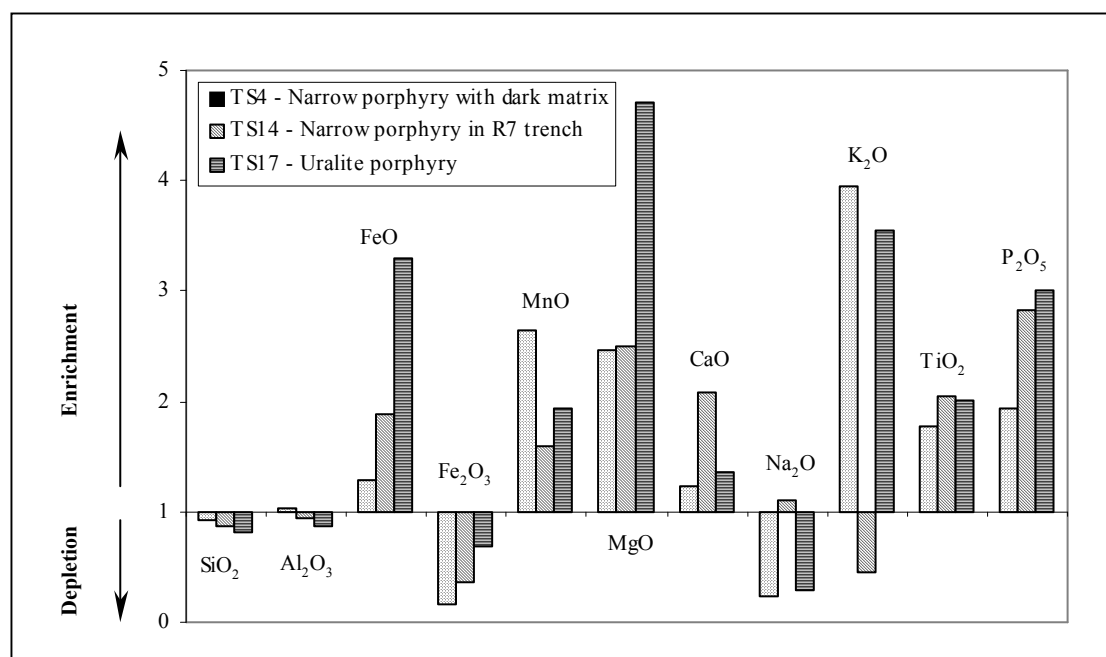


Figure 3.26: Enrichment – depletion diagram for the major element oxides of the narrow porphyries and the uralite porphyry. The diagram compares the oxide abundances of the “altered” samples to the average of the unaltered porphyries.

The main differences in the effect of the alteration on the major element geochemistry between the amphibolites and the porphyry is that Mg and K, and to a certain extent Fe^{2+} , are more greatly enriched in the porphyries than the amphibolites (reflecting a higher biotite content); there is an enrichment in Ti (also reflecting a higher biotite content) and P in the porphyries as opposed to a depletion in the amphibolites; and that Fe^{3+} is depleted in the porphyries, whereas it is enriched in the amphibolites.

As mentioned previously, the altered porphyries have higher overall abundances of the REE and the trace elements compared to the fresh, unaltered porphyries (Figures 3.15 – 3.16; Table 3.4). This is believed to be caused by a combination of fractional crystallisation and subsequent alteration rather than representing a distinct and separate phase of porphyry intrusion. An enrichment in REE is commonly associated with alteration in many deposit types including Fe-oxide Cu-Au deposits (Hitzman et al., 1992).

3.1.4 Discussion

On the basis of the whole rock geochemistry it is possible to construct a likely geological model for the main lithologies at Haveri. The oldest rocks encountered in the study are the amphibolites of the Haveri Formation. These represent metamorphosed submarine pillow basalts most likely formed in a back-arc basin environment formed by the splitting of an oceanic volcanic arc (e.g. Kähkönen and Nironen, 1994). Overlying the main bulk of the Haveri pillow basalts, and intercalated near the top of the volcanics, are banded volcanoclastic tuffites. The likely source for these volcanoclastics is the volcanic arc itself, which typically is volumetrically the most important source in back-arc basins (Carey and Sigurdsson, 1984). The volcanoclastics may be produced by explosive subaerial and/or subaqueous and secondary erosion of the arc complex. Transport of the material to the deeper parts of the marginal basins occurs by passive settling through the water column or by a variety of sediment gravity flows (Carey and Sigurdsson, 1984). The back-arc spreading centre also provides small volumes of hyaloclastites and hydrothermally derived, iron sulphide-rich, volcanogenic sediments that are incorporated into the basal parts of the overlying sedimentary sequence (Carey and Sigurdsson, 1984). The volcanoclastic tuffites grade upwards into the epiclastic turbiditic sequence of the Osara Formation, with intercalated tuffite layers declining upwards from the base. The source of the epiclastic turbidites is likely to be from erosion of the subaerial and/or subaqueous arc complex. In the later stages of basin development, back-arc spreading declines, perhaps due to the migration of the spreading centre away from the region of tensional stress induced by the subduction process, and pelagic sedimentation becomes the dominant process (Carey and Sigurdsson, 1984).

The formation of the Haveri Formation in a back-arc basin, differentiates it significantly from the main part of the Tampere Schist Belt (TSB), which formed in an active volcanic arc (e.g. Nironen, 1997). In addition the metavolcanic Haveri Formation is separated from the arc-volcanics of the TSB by a thick sequence of greywackes represented by the Osara Formation and the basal part of the TSB stratigraphy. This hiatus in arc activity represents a considerable period of passive margin and/or fore-arc deposition, prior to the onset of volcanic activity typical of an active arc environment. Therefore, although the Haveri Formation forms the stratigraphic base of the TSB, it actually represents an earlier geotectonic period, and as such is not related to the genesis of the TSB. Kähkönen and Nironen (1994) suggested that the Haveri Formation represents an exotic remnant of an older ocean basin within an arc complex. This accounts for the significant age differences (~100 Ma) observed between the Pb-Pb model age of the Haveri Formation (Vaasjoki and Huhma, 1987) and the dates obtained by Kähkönen et al. (1989) from considerably higher in the stratigraphy of the TSB (Section 1.3).

The post-depositional evolution of the geology at Haveri, including metamorphism, hydrothermal alteration and mineralisation is discussed in detail in later sections. Suffice to say, the most widespread alteration phase at Haveri, therefore, includes mainly amphibolitisation accompanied by albitisation and iron enrichment and, thus, may be interpreted to represent a Ca-Na-Fe alteration phase.

3.2 MINERAL CHEMISTRY

3.2.1 Analytical Techniques

The polished sections were studied under a polarising microscope with both transmitted and reflected light. The mineral analyses were carried out at Rhodes University using a JEOL-733 Superprobe electron microprobe, equipped with four wavelength dispersive spectrometers (WDS). The analytical conditions used for silicate analyses were 15kV and 20 nA, and 20kV and 30 nA for sulphides and gold. Standards and X-ray lines used for silicate, sulphide and gold analyses are displayed in Table 3.5.

Silicates			Sulphides and gold		
<i>Element</i>	<i>Line</i>	<i>Standard</i>	<i>Element</i>	<i>Line</i>	<i>Standard</i>
Si	K _α	Wollastonite	Fe	K _α	Chalcopyrite
Al	K _α	Plagioclase	Cu	K _α	Chalcopyrite
Fe	K _α	Olivine	Zn	K _α	Sphalerite
Mg	K _α	Olivine	S	K _α	Chalcopyrite
Mn	K _α	Willemite	Au	L _α	Gold metal
Cr	K _α	Chromite	Ag	L _α	Silver metal
Ti	K _α	Rutile			
Ca	K _α	Wollastonite			
Na	K _α	Plagioclase			
K	K _α	Orthoclase			

Table 3.5: Elements analysed on studied silicates, sulphides and gold. Used X-ray lines and standards are listed after each element.

3.2.2 Mineral Compositions

3.2.2.1 Amphibole

Amphibole is one of the most widespread silicates at Haveri, occurring in all the rock types, and in the igneous, metamorphic and hydrothermal mineral assemblages. Amphiboles are double-chain silicate minerals and are classified on the basis of the chemical contents of the standard amphibole formula $A_{0-1}B_2C_5^{VI}T_8^{IV}O_{22}(OH)_2$ where in structural terms the *A*, *B*, *C* and *T* atoms are respectively in *A* sites; *M4*; *M1* + *M2* + *M3*; and *T* (tetrahedral) (Deer et al., 1992). The most common cations that can enter each structural site are (Deer et al., 1992):

A Na, K

B Na, Ca, Mg, Fe²⁺ but also Mn, Li

C Mg, Fe²⁺, Al, Fe³⁺ but also Mn, Zn, Cr, Ti and Li

T Si, Al

Hydroxyl ions can be replaced by Cl as well as by F, and sometimes by oxygen

The amphiboles are classified primarily into four groups depending on the occupancy of the *B* sites (IMA, 1997):

- 1) *Mg-Fe-Mn-Li amphiboles*: $(Ca + Na)_B < 1.00$ and L type ions $(Mg, Fe, Mn, Li)_B \geq 1.00$
- 2) *Calcic amphiboles*: $(Ca + Na)_B \geq 1.00$ and $Na_B < 0.50$. Usually, but not always, $Ca_B > 1.50$
- 3) *Sodic-calcic amphiboles*: $(Ca + Na)_B \geq 1.00$ and Na_B 0.50 to 1.50
- 4) *Sodic amphiboles*: $Na_B \geq 1.50$

Analytical data from electron microprobe analyses are unable to assign the elements to site occupancies in the crystal lattice, and it is necessary to recalculate the analyses to provide the chemical formula, and hence the name of the amphibole. However, these formulae amount to arithmetic “best-fits”, and cannot be confirmed without direct structural evidence (IMA, 1997). The electron microprobe data are also unable to provide analyses for H₂O, and consequently the amphibole formula has been calculated on the basis of 23(O) with 2(OH, F, Cl) assumed. Another common uncertainty in calculating an amphibole formula is caused by the inability of an electron microprobe to distinguish the valence states of elements, and in particular, Fe²⁺ and Fe³⁺, which can influence the resulting name of the mineral (IMA, 1997). The stoichiometric calculations for amphibole formulae and the estimations of the ferric iron content have been done according to the method set out by J. C. Schumacher (IMA, 1997), although there are a number of alternative methods of estimating ferric iron contents (e.g. Droop, 1987). The Schumacher method of ferric iron determination yields a maximum and minimum ferric iron content based upon the permissible and usual site occupancies, defining a range of possible formulae and possible names. Selecting a single structural formula from the range of possibilities requires some other constraint or assumption. For this study the mean value between the maximum and the minimum has been used for the mineral formulae and the names of the amphibole species. However, it is recognised that the mean ferric iron content provides a formula that is no more or no less valid than a formula provided by the maximum or minimum ferric iron contents, or any ferric iron content that lies in between (Schumacher in IMA, 1997).

Petrographically, amphiboles are the most prevalent mineral at Haveri, occurring in practically all the mineral assemblages. In total some 182 acceptable amphibole microprobe analyses were completed from 19 samples. The average amphibole compositions have been summarised in Table 3.6, and the full dataset is supplied in Appendix C. After recalculation and ferric iron estimation, it can be seen that all of the analysed amphiboles are calcic amphiboles, the majority of which can be classified as magnesiohornblendes and actinolites in composition (Figure 3.27).

The most common amphiboles at Haveri are those associated with the regional metamorphic assemblage, and those associated with the amphibolitisation event. As mentioned earlier (Section 2.3.1) the amphibolitisation occurs both as a pervasive, dark green amphibole representing an intense amphibolitisation of the host rock, as well as in amphibole veinlets cutting the host rock. The veinlets themselves are either composed of a homogenous amphibole or alternatively are zoned, commonly consisting of a dark green rim to the vein, and a paler core, which sometimes displays a “washed-out” retrograde appearance. The metamorphic amphiboles are recognised as those amphiboles that are associated with plagioclase in the main rock mass, with no obvious hydrothermal overprint or direct association with amphibole veining.

When plotted on the diagram for classifying calcic amphiboles (Figure 3.28), it is clear that they define a distinct trend from ferrotschermakite to actinolite. The metamorphic amphiboles plot primarily in the magnesiohornblende field of the diagram, as do the homogenous amphibole veinlets and the rims of the zoned amphibole veinlets. The cores of the veinlets straddle the actinolite – magnesiohornblende fields, and the “washed-out”, retrograde cores plot in the actinolite field. The amphiboles associated with the intense amphibolitisation plot in the ferrotschermakite field.

This pattern is interpreted as a function of the changing P - T - X conditions over both distance and time from the focal point of the alteration. In the most intensely altered areas, the P - T - X conditions of the hydrothermal fluids are at their most extreme, resulting in the almost total replacement of the host rock by amphibole. Further out from this point, the nature of the alteration changes from a pervasive replacement, to one of network veining of the host rock. The smaller veins and veinlets rapidly choke their conduits, preventing any further reaction with the hydrothermal fluids, and thereby retain a homogenous character. The larger veins, would remain open longer, thereby allowing the centre of the vein to be modified according to the changing P - T - X conditions of the fluids. The “washed out” retrograde actinolites in the centers of the vein, are interpreted as later retrograde replacements of initial pyroxene (Roberts et al., 2000). The similarities between the metamorphic amphiboles and the veinlets and rim-type amphiboles, suggests that they formed at similar P - T - X conditions, and supports the theory that the amphibolitisation is syn-metamorphism.

Class	Metamorphic amphibole	Intense amphibole	Core of amphibole veins	Rim of amphibole veins	Retrograde amphibole in cores	Amphibole veinlets	Amphibole in quartz vein	Amphibole on rim of quartz vein	Amphibole near quartz vein	Associated with pyroxene	Zoned with Gt and Pyx	Associated with carbonate	Associated with epidote	Associated with sulphide	Associated with cataclastite	Associated with gold
n	36	6	14	10	11	8	6	7	2	22	7	27	2	11	3	10
Name	MH	FT	MH	MH	ACT	MH	MH	MH	MH	MH	MH	MH/FH	MH	MH	MH/FT	FT
SiO ₂	46.35	40.89	49.65	47.09	51.47	46.75	48.77	46.15	48.79	47.39	47.53	51.49	50.89	46.03	42.88	40.85
TiO ₂	0.80	0.83	0.34	0.76	0.07	0.77	0.40	0.78	0.71	0.38	0.39	0.26	0.12	0.54	0.83	0.80
Al ₂ O ₃	7.70	10.18	4.89	6.77	2.78	7.46	5.70	7.71	6.41	7.05	7.78	5.17	4.28	8.08	9.25	11.16
Cr ₂ O ₃	0.04	0.01	0.00	0.01	0.00	0.02	0.00	0.00	0.05	0.01	0.02	0.01	0.06	0.03	0.00	0.01
FeO	18.06	23.60	17.66	19.09	17.45	17.85	16.78	18.62	17.12	15.72	19.19	10.22	15.54	18.77	20.52	23.21
MnO	0.25	0.40	0.23	0.22	0.30	0.24	0.21	0.28	0.17	0.62	0.38	0.30	1.32	0.67	0.28	0.28
MgO	10.58	7.09	11.88	10.52	13.12	10.90	12.74	10.23	12.37	12.55	9.53	16.37	13.58	10.61	9.53	6.76
CaO	11.64	11.67	11.47	11.52	11.74	11.63	11.95	11.96	11.67	12.33	12.04	12.76	12.40	11.79	11.78	11.39
Na ₂ O	0.98	1.02	0.68	0.91	0.24	0.84	0.65	0.90	0.77	0.68	0.67	0.51	0.36	0.84	0.87	0.96
K ₂ O	0.41	1.10	0.10	0.27	0.09	0.47	0.20	0.46	0.04	0.67	0.31	0.19	0.04	0.13	1.02	1.37
Cl	0.02	0.72	0.28	0.19	0.12	0.10	0.15	0.00	0.15	0.20		0.13	0.00	0.03	0.96	0.80
F	0.01	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00		0.00	0.00	0.00	0.02	0.04
Total	96.81	96.78	96.90	97.15	97.26	96.94	97.40	97.09	98.10	97.41	97.85	97.28	98.59	97.49	96.97	96.79
-O=Cl,F	0.01	0.16	0.06	0.04	0.03	0.01	0.04	0.00	0.03	0.02	0.00	0.01	0.00	0.01	0.23	0.20
Total	96.80	96.61	96.84	97.11	97.23	96.93	97.36	97.09	98.06	97.39	97.85	97.27	98.59	97.48	96.74	96.59
AVERAGE																
Si	6.95	6.33	7.35	7.04	7.57	6.98	7.16	6.92	7.11	6.97	7.07	7.36	7.34	6.83	6.50	6.32
Al ^{iv}	1.05	1.67	0.65	0.96	0.43	1.02	0.84	1.08	0.89	1.03	0.93	0.64	0.66	1.17	1.50	1.68
sum T	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
Al ^{vi}	0.31	0.19	0.21	0.23	0.06	0.29	0.16	0.29	0.21	0.22	0.44	0.24	0.07	0.24	0.15	0.36
Ti	0.09	0.10	0.04	0.09	0.01	0.09	0.04	0.09	0.08	0.04	0.04	0.03	0.01	0.06	0.09	0.09
Fe ³⁺	0.34	0.83	0.34	0.42	0.33	0.37	0.45	0.34	0.46	0.41	0.23	0.21	0.52	0.67	0.79	0.69
Cr	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.01	0.00	0.00	0.00
Mg	2.36	1.64	2.62	2.34	2.87	2.42	2.78	2.28	2.69	2.74	2.11	3.48	2.92	2.35	2.15	1.56
Fe ²⁺	1.88	2.22	1.79	1.92	1.73	1.81	1.54	1.98	1.55	1.55	2.14	1.02	1.36	1.64	1.79	2.27
Mn	0.02	0.03	0.01	0.01	0.00	0.01	0.01	0.02	0.01	0.04	0.03	0.02	0.11	0.05	0.02	0.02
sum C	5.00	5.00	5.00	5.00	5.00	5.00	4.99	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00
Mg	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Fe ²⁺	0.04	0.01	0.06	0.05	0.09	0.05	0.07	0.02	0.08	0.00	0.02	0.01	0.00	0.02	0.02	0.04
Mn	0.02	0.02	0.02	0.02	0.03	0.02	0.01	0.02	0.02	0.04	0.02	0.01	0.05	0.04	0.02	0.02
Ca	1.87	1.94	1.82	1.84	1.85	1.86	1.88	1.92	1.82	1.95	1.92	1.96	1.92	1.87	1.91	1.89
Na	0.07	0.03	0.09	0.08	0.03	0.08	0.05	0.04	0.09	0.01	0.04	0.02	0.04	0.06	0.05	0.06
sum B	2.00	2.00	2.00	2.00	2.01	2.00	2.03	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00
Na	0.21	0.27	0.10	0.18	0.04	0.17	0.13	0.22	0.13	0.19	0.15	0.12	0.06	0.18	0.21	0.23
K	0.08	0.22	0.02	0.05	0.02	0.09	0.04	0.09	0.01	0.13	0.06	0.03	0.01	0.02	0.20	0.27
sum A	0.29	0.49	0.12	0.23	0.05	0.26	0.17	0.31	0.14	0.32	0.21	0.15	0.07	0.20	0.41	0.50
Total	15.29	15.49	15.12	15.23	15.06	15.26	15.18	15.31	15.14	15.32	15.21	15.16	15.07	15.20	15.41	15.50
(Na + K) _A	0.29	0.49	0.12	0.23	0.05	0.26	0.17	0.31	0.14	0.32	0.21	0.15	0.07	0.20	0.41	0.50
Mg/(Mg + Fe ²⁺)	0.55	0.42	0.59	0.54	0.61	0.57	0.63	0.53	0.62	0.63	0.50	0.77	0.68	0.59	0.54	0.40
Ca _B	1.87	1.94	1.82	1.84	1.85	1.86	1.88	1.92	1.82	1.95	1.92	1.96	1.92	1.87	1.91	1.89
Si	6.95	6.33	7.35	7.04	7.57	6.98	7.16	6.92	7.11	6.97	7.07	7.36	7.34	6.83	6.50	6.32

Table 3.6: Summary table of average amphibole compositions, for various category of amphibole encountered at Haveri. The stoichiometry uses the average ferric iron content calculated according to Schumacher (IMA, 1997). MH = Magnesiohornblende, FT = Ferrotschermakite, ACT = Actinolite, FH = Ferrohornblende.

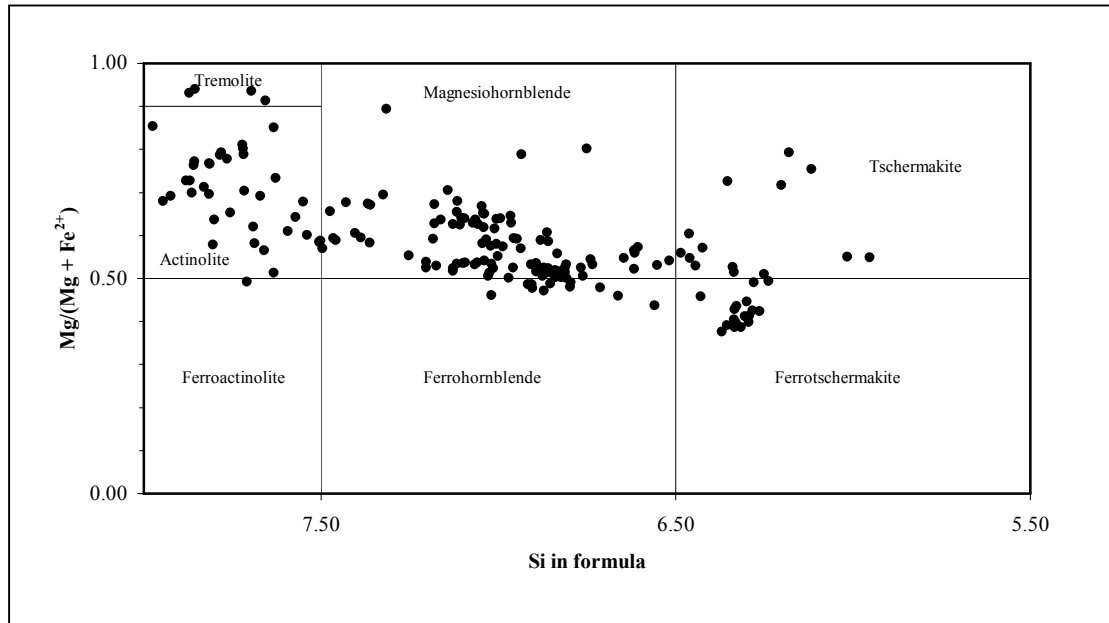


Figure 3.27: Classification of all the Haveri amphibole analyses on the calcic-amphibole plot (IMA, 1997)

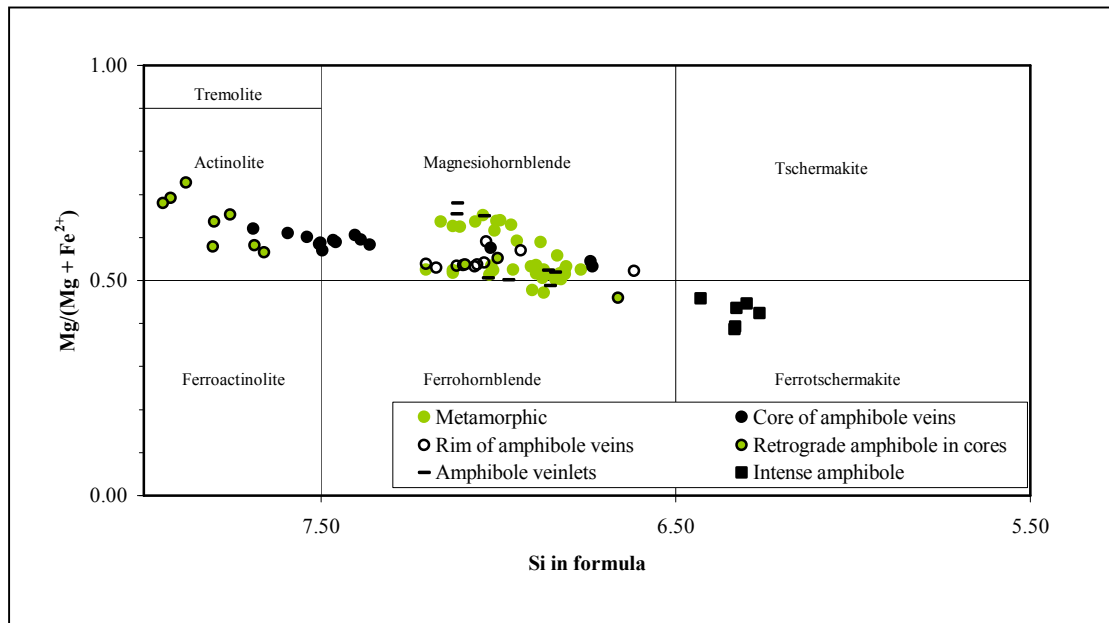


Figure 3.28: Electron microprobe analyses of amphiboles associated with metamorphism and amphibolitisation at Haveri (after IMA, 1997).

The compositions of the amphiboles associated with alteration also display a pattern of iron enrichment with increasing alteration. Thus the mineral chemistry supports the whole-rock data (Section 3.1) of increasing iron enrichment with increasingly pervasive alteration.

The amphiboles related to calcic-skarn mineral assemblages can be classified on the basis of their mineral associations (Figure 3.29). Many of these amphiboles display retrograde textures, whereby the amphiboles are retrograde products formed either by the breakdown of pyroxene, or alternatively by the breakdown of an earlier amphibole species. Consequently, each class is likely to contain a series of analyses which broadly define the spectrum of compositions from earlier proximal amphiboles to later distal amphiboles. The analyses for the amphiboles associated with the zoned garnet – pyroxene assemblages, as well as those associated with pyroxene, both define series which are similar to those displayed by the amphibolitisation (Figure 3.28), except there is no ferrotschermakite. The amphiboles associated with carbonates are more Mg-rich, and define a series from tschermakite to tremolite-actinolite. The calcic-skarn displays a mineral assemblage of carbonate to pyroxene to garnet, that reflects an increase in iron with increasing skarnification. This general iron enrichment with increasing skarnification is also displayed in the amphibole compositions associated with the calcic-skarn (Figure 3.29), thus showing a similar trend to the compositions associated with amphibolitisation.

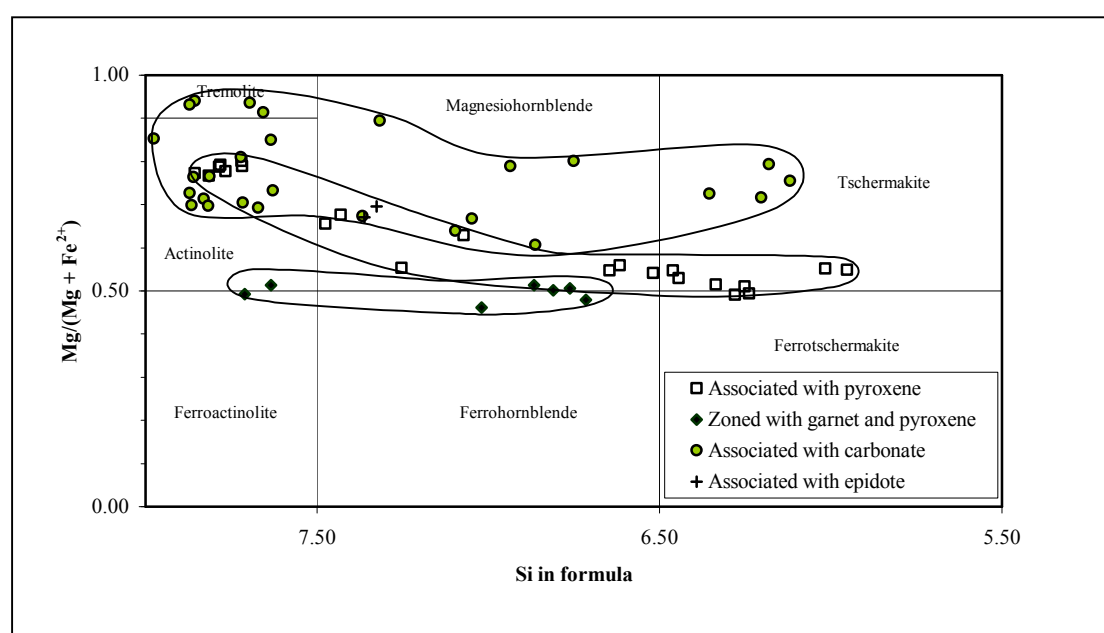


Figure 3.29: Electron microprobe analyses of amphiboles related to the calcic-skarn alteration.

Amphiboles are also present in and around the quartz veins and veinlets (Figure 3.30). These are primarily magnesiohornblendes and have compositions similar to both the metamorphic amphiboles and the homogenous amphibole veinlets. The amphibole analyses from within the quartz veins themselves define a similar trend to that defined by the amphibolitisation, with compositions ranging from actinolite, through magnesiohornblende, into the ferrohornblende field. This similarity may be indicative of a genetic relationship between the quartz veins and the amphibolitisation.

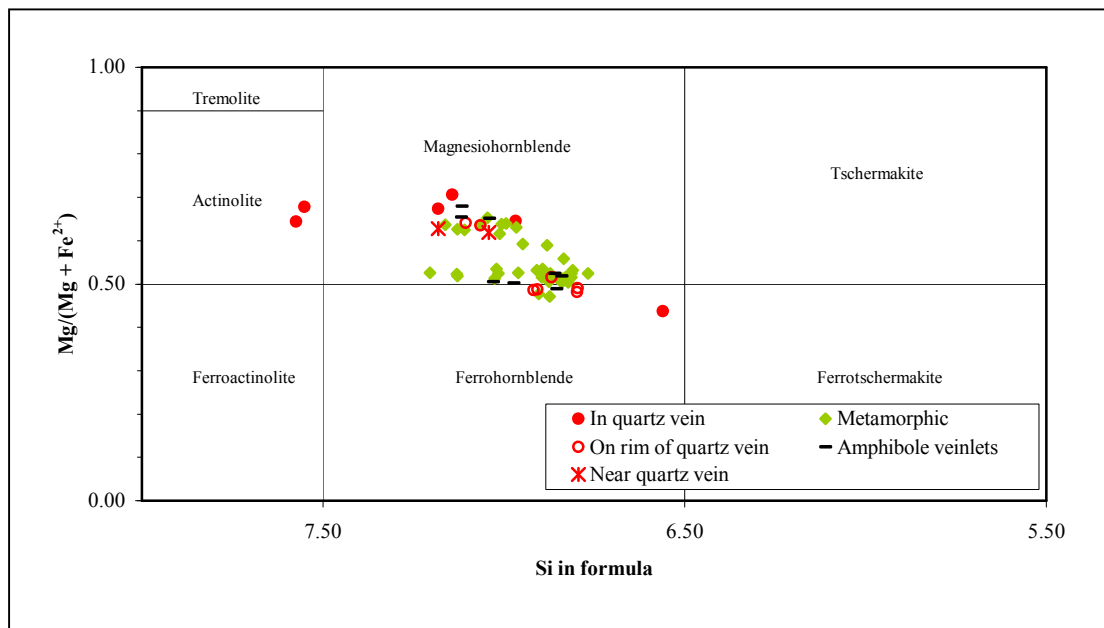


Figure 3.30: Electron microprobe analyses of amphiboles relating to quartz veins.

Amphiboles are also commonly associated with sulphides and gold (Figure 3.31). Gold is commonly observed as coarse grains within a dark green amphibole mass. The electron microprobe data indicate that the amphiboles associated with gold are ferrotschermakites with similar compositions to the intense amphibole alteration. These amphiboles associated with gold are more iron-rich than those associated with sulphides and cataclastites, which are primarily magnesiohornblendes with a few analyses in the tschermakite field. This trend of iron-enrichment and Si depletion is similar to that of the amphibolitisation and the calcic-skarn, rather than the carbonate veining (as expected). The obvious conclusion to draw from this is that the gold mineralisation is related to the intense amphibole alteration phase, which is characterised by the most iron-enriched amphiboles.

It is apparent that there is a similar trend of Fe enrichment and Si depletion in amphiboles which represent more intense and/or proximal alteration, from the various assemblages analysed (i.e. amphibolitisation, calcic-skarn assemblages, quartz-vein assemblages and the mineralisation assemblages). These similarities observed in the trends of amphibole compositions defined by the various assemblages can be caused by several processes. It may well be that the amphiboles from these different assemblages did in fact form contemporaneously, and are effectively co-genetic. However, it is also possible that *P-T-X* conditions were similar at different stages of the geological evolution of the Haveri area, and hence amphiboles of similar composition may have been produced independently of one another.

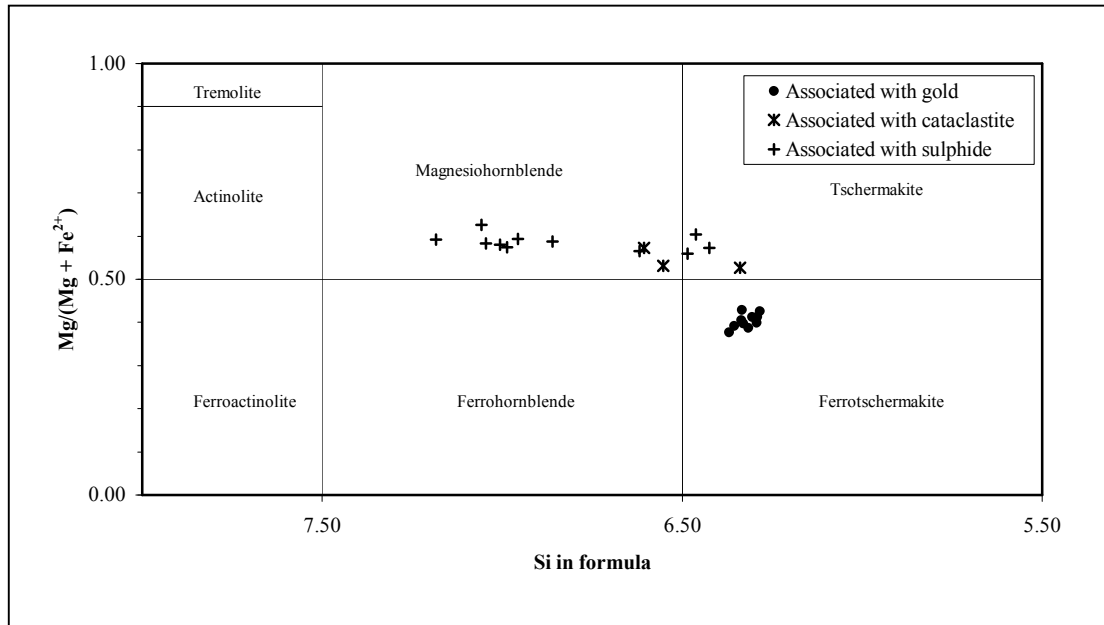
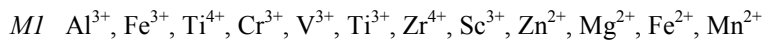
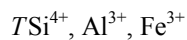


Figure 3.31: Electron microprobe analyses of amphiboles related to mineralisation.

3.2.2.2 Pyroxene

The pyroxenes at Haveri are now only part of the hydrothermal mineral assemblage, although they were perhaps also part of the pre-metamorphic mineral assemblages. Pyroxenes are single-chain silicates that are classified on the basis of the chemical contents of the standard pyroxene formula $[(M2)(M1)T_2O_6]$ where $M2$ and $M1$ refer to cations in irregular and regular octahedral coordination respectively, and T to tetrahedrally coordinated cations (Deer et al., 1992). The most common cations that can enter each structural site are, in terms of ideal site occupancy, and in the order of assigning atoms to each site (Deer et al., 1992):



The pyroxenes are divided broadly into three main subgroups (Deer et al., 1992):

- 1) *Mg-Fe pyroxenes*: Magnesium and iron occupy more than 90% of the $M1$ and $M2$ sites.
- 2) *Calcium pyroxenes*: Calcium occupies more than two third of the $M2$ sites.
- 3) *Sodium pyroxenes*: The $M2$ site is largely occupied by Na and the $M1$ sites by Al, Fe^{3+} or Cr.

In addition there are two minor subgroups: the calcium-sodium pyroxenes and the lithium pyroxenes.

As mentioned previously (Section 3.2.1), electron microprobe analyses are unable to identify the structural sites and the valence states of the elements (in particular iron). Consequently it is necessary to recalculate stoichiometrically on the basis of six oxygens to provide the chemical formula. Ferric iron estimations have been determined according to Droop (1987).

In total, 50 pyroxenes were analysed from 10 separate samples. The averages of these samples have been summarised in Table 3.7, and the full dataset are supplied in Appendix C. The electron microprobe analyses of pyroxenes from the Haveri area clearly plot within the calcium pyroxene fields of diopside ($\text{CaMgSi}_2\text{O}_6$) and hedenbergite ($\text{CaFeSi}_2\text{O}_6$) when plotted on the Ca-Mg-Fe pyroxene classification diagram (Figure 3.32)

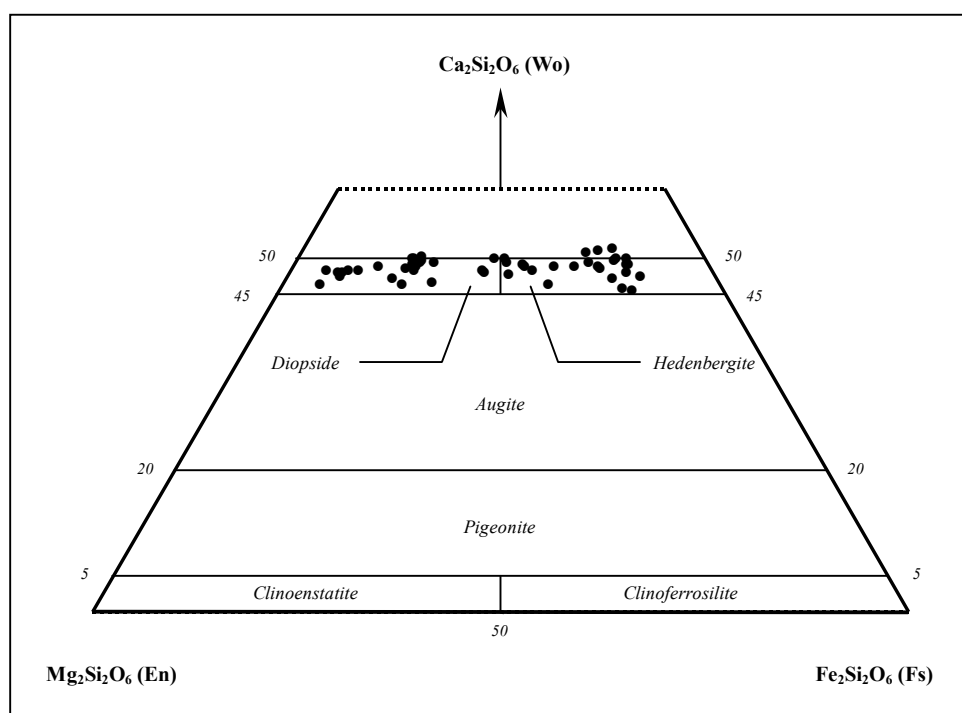


Figure 3.32: Composition ranges and the nomenclature of the Haveri Ca-Mg-Fe clinopyroxenes (after Deer et al., 1992)

The clinopyroxenes analysed at Haveri plot in two main fields: diopside and hedenbergite. When the data are further differentiated and plotted on an Mg-Fe-Mn diagram (Figure 3.33), it is apparent that these two fields distinguish pyroxenes according to their associated alteration assemblage. The more Mg-rich diopsides are all from the amphibolitisation assemblages, where they are associated with amphibole, but devoid of garnet. The more Fe-rich hedenbergites are from the calcic-skarn assemblage where they are associated with zoned garnet-pyroxene ($\pm$ amphibole) mineral assemblages.

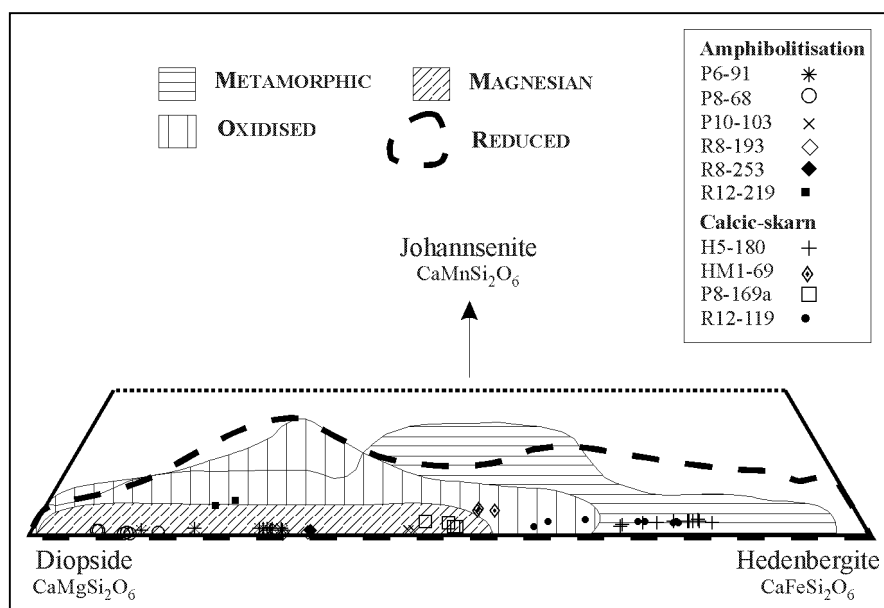


Figure 3.33: Compositional range of pyroxenes at Haveri. Also shown are the fields of pyroxene compositions from reduced, oxidised, magnesian and metamorphic Au skarns (after Meinert, 1998).

It is possible to interpret this compositional difference in two ways: either the amphibolitisation and the calcic-skarn alteration are two separate and distinct alteration events. Or alternatively, the calcic-skarn event represents a more proximal manifestation of the amphibolitisation, and that the pyroxene chemistry displays a gradational variation from the more distal amphibolitisation into the proximal calcic-skarn alteration. This relationship between the amphibolitisation and the calcic-skarn will be discussed further in Section 5.2.1.2.

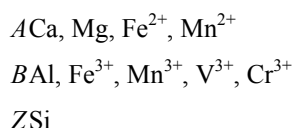
Figure 3.33 also shows the fields of pyroxene compositions from a variety of Au skarn types as outlined by Meinert (1998). The compositions of the pyroxenes from Haveri indicate that both the amphibolitisation and the calcic-skarn are consistent with being formed as part of either a metamorphic or a reduced Au-skarn system. Ore genetic models, and the implications of mineral chemistry and other geochemical parameters will be discussed in detail in Section 5.

Sample Alteration n	R12-119 C-S 8		P8-68 C-S 6		H5-180 C-S 11		P8-169a C-S 4		HM1-69 C-S 3		P6-91 Amph. 9		R8-193 Amph. 3		P10-103 Amph. 2		R12-219 Amph. 3		R8-253 Amph. 1
	Average	Std. Dev.	Average	Std. Dev.	Average	Std. Dev.	Average	Std. Dev.	Average	Std. Dev.	Average	Std. Dev.	Average	Std. Dev.	Average	Std. Dev.	Average	Std. Dev.	Average
SiO ₂	50.53	0.83	53.91	0.70	48.87	0.37	50.96	0.35	51.00	0.02	52.18	0.56	52.23	0.29	50.94	0.15	53.32	0.61	51.83
TiO ₂	0.08	0.04	0.10	0.09	0.04	0.03	0.01	0.02	0.00	0.00	0.05	0.04	0.03	0.02	0.02	0.03	0.03	0.03	0.00
Al ₂ O ₃	1.04	0.09	0.99	0.62	0.94	0.31	0.75	0.12	0.87	0.14	0.87	0.43	0.50	0.04	0.77	0.17	0.36	0.12	0.52
Cr ₂ O ₃	0.04	0.04	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.03	0.04	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Fe ₂ O ₃	0.36	0.75	0.95	0.81	1.72	1.26	1.95	0.78	1.02	0.62	1.88	0.87	3.29	0.44	3.17	0.20	0.79	1.37	2.07
FeO	20.39	1.54	2.80	1.25	21.39	1.79	13.38	1.20	15.26	0.88	6.44	2.32	6.85	0.42	11.54	0.32	6.44	1.15	8.93
MnO	0.60	0.08	0.25	0.07	0.60	0.08	0.51	0.10	1.12	0.04	0.34	0.05	0.25	0.06	0.29	0.01	1.63	0.13	0.32
MgO	4.79	1.33	16.89	0.74	3.83	0.65	8.60	0.27	7.66	0.19	13.20	1.35	12.46	0.41	9.67	0.26	13.96	0.33	12.12
CaO	22.87	0.84	24.39	0.49	22.17	1.20	23.63	0.81	23.21	0.35	24.40	0.33	24.39	0.48	23.16	0.60	23.85	0.86	22.44
Na ₂ O	0.24	0.03	0.01	0.02	0.27	0.04	0.31	0.02	0.26	0.02	0.19	0.06	0.41	0.02	0.47	0.30	0.02	0.03	0.51
K ₂ O	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Total	100.95	0.95	100.29	1.16	99.83	1.01	100.11	1.02	100.41	0.37	99.57	0.83	100.40	1.04	100.04	0.64	100.41	0.21	98.73
Si	1.98	0.01	1.97	0.02	1.97	0.01	1.98	0.01	1.98	0.01	1.97	0.01	1.97	0.00	1.97	0.00	1.99	0.01	1.99
Al	0.02	0.01	0.03	0.02	0.03	0.01	0.02	0.01	0.02	0.01	0.03	0.01	0.02	0.00	0.03	0.00	0.01	0.01	0.01
Al	0.03	0.01	0.01	0.01	0.02	0.01	0.01	0.01	0.02	0.01	0.01	0.01	0.00	0.00	0.01	0.01	0.01	0.01	0.01
Fe ³⁺	0.01	0.02	0.03	0.02	0.05	0.04	0.06	0.02	0.03	0.02	0.05	0.02	0.09	0.01	0.09	0.01	0.02	0.04	0.06
Cr	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Ti	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Fe ²⁺	0.67	0.05	0.09	0.04	0.72	0.07	0.43	0.04	0.49	0.03	0.20	0.07	0.21	0.01	0.37	0.01	0.20	0.04	0.29
Mn	0.02	0.00	0.01	0.00	0.02	0.00	0.02	0.00	0.04	0.00	0.01	0.00	0.01	0.00	0.01	0.00	0.05	0.00	0.01
Mg	0.28	0.08	0.92	0.04	0.23	0.04	0.50	0.01	0.44	0.01	0.74	0.08	0.70	0.03	0.56	0.01	0.78	0.01	0.69
Ca	0.96	0.04	0.95	0.02	0.96	0.05	0.98	0.03	0.97	0.01	0.99	0.01	0.99	0.01	0.96	0.02	0.95	0.04	0.92
Na	0.02	0.00	0.00	0.00	0.02	0.00	0.02	0.00	0.02	0.00	0.01	0.00	0.03	0.00	0.04	0.02	0.00	0.00	0.04
K	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
TOTAL	4.00	0.01	4.01	0.01	4.01	0.01	4.02	0.01	4.01	0.01	4.02	0.01	4.03	0.00	4.03	0.00	4.01	0.01	4.02
En	14.30	3.94	46.12	1.87	11.50	1.90	24.75	0.58	22.28	0.40	36.92	3.34	34.50	1.44	27.55	0.45	38.72	1.21	34.50
Fs	35.72	2.72	5.96	1.35	39.57	2.65	25.19	1.15	28.21	0.83	13.28	2.76	15.50	0.91	23.30	0.01	13.68	0.83	17.67
Wo	49.04	1.59	47.87	0.74	47.88	1.73	48.90	0.99	48.53	0.36	49.10	0.58	48.52	0.58	47.41	0.71	47.53	1.18	45.93
Ac	0.93	0.12	0.05	0.08	1.05	0.17	1.16	0.08	0.97	0.07	0.69	0.23	1.49	0.08	1.75	1.14	0.07	0.11	1.89

Table 3.7: Summary table of average pyroxene compositions from Haveri with standard deviations. The samples from the calcic-skarn (C-S) alteration assemblage are zoned with associated garnet. The samples from the amphibolitisation (Amph.) alteration assemblage have no associated garnet. The values shown have been recalculated for ferric iron content according to Droop (1987).

3.2.2.3 Garnet

Garnets are common at Haveri in the metasedimentary lithologies, as well as in the proximal part of the hydrothermal mineral assemblage. Garnets form an isomorphous series represented by end-members with the general formula $X_3Y_2Z_3O_{12}$. The most common cations that can enter each structural site are (Deer et al., 1992):



In hydrogarnets there is replacement of SiO_2 by $2\text{H}_2\text{O}$, with vacant Si spaces in the structure

The end-members of the garnet group are as follows:

Pyrope	$\text{Mg}_3\text{Al}_2\text{Si}_3\text{O}_{12}$
Almandine	$\text{Fe}_3^{2+}\text{Al}_2\text{Si}_3\text{O}_{12}$
Spessartine	$\text{Mn}_3\text{Al}_2\text{Si}_3\text{O}_{12}$
Grossular	$\text{Ca}_3\text{Al}_2\text{Si}_3\text{O}_{12}$
Andradite	$\text{Ca}_3(\text{Fe}^{3+}, \text{Ti})_2\text{Si}_3\text{O}_{12}$
Uvarovite	$\text{Ca}_3\text{Cr}_2\text{Si}_3\text{O}_{12}$
Hydrogrossular	$\text{Ca}_3\text{Al}_2\text{Si}_2\text{O}_8(\text{SiO}_4)_{1-m}(\text{OH})_{4m}$

The name assigned to a garnet is dependent on the dominant ‘molecular’ species present. The garnets may be divided into two series: pyrospite (*pyrope*, *almandine*, *spessartine*) and ugrandite (*uvarovite*, *grossular*, *andradite*). Fairly complete and continuous variation in composition occurs within these two series, but there appears to be no continuous variation between pyrospite and ugrandite (Deer et al., 1992).

Electron microprobe analyses have been recalculated on the basis of 24(O) to give the stoichiometric formulae, and also to provide the molecular proportions of their end-member components. Ferric iron has been estimated according to Droop (1987). In total, 66 garnet analyses were completed, from 6 separate samples. This includes a traverse of 15 analyses across one garnet to determine its compositional variation (Figure 3.35). The averages of these samples have been summarised in Table 3.8, and the full dataset are supplied in Appendix C.

n	Metasomatic		Metamorphic		Metamorphic		Metamorphic	
	27		All		Rim		Core	
	Average	Std. Dev.	Average	Std. Dev.	Average	Std. Dev.	Average	Std. Dev.
SiO₂	36.64	1.02	37.88	0.45	37.79	0.22	38.07	0.45
TiO₂	0.52	0.23	0.02	0.02	0.02	0.02	0.01	0.02
Al₂O₃	6.36	2.39	21.06	0.63	21.20	0.31	21.15	0.64
Fe₂O₃	21.05	3.93	0.26	0.46	0.18	0.26	0.27	0.45
FeO	2.63	1.84	32.20	1.10	33.24	0.73	31.39	0.79
MnO	0.59	0.25	2.09	0.50	1.99	0.38	2.10	0.56
MgO	0.04	0.02	4.36	0.75	3.83	0.51	4.82	0.72
CaO	31.98	1.15	2.09	0.20	2.06	0.17	2.10	0.12
Cr₂O₃	0.02	0.02	0.02	0.02	0.01	0.02	0.01	0.02
Total	99.83	1.15	99.97	0.72	100.30	0.57	99.92	0.61
Si	6.36	0.06	6.03	0.02	6.02	0.02	6.04	0.02
Al^{iv}	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Al^{vi}	1.29	0.42	3.95	0.08	3.98	0.04	3.95	0.08
Fe³⁺	2.61	0.51	0.03	0.06	0.02	0.03	0.03	0.05
Cr	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Ti	0.07	0.03	0.00	0.00	0.00	0.00	0.00	0.00
Fe²⁺	0.36	0.24	4.29	0.16	4.43	0.10	4.17	0.13
Mn	0.09	0.04	0.28	0.07	0.27	0.05	0.28	0.08
Mg	0.01	0.00	1.03	0.17	0.91	0.12	1.14	0.16
Ca	5.96	0.25	0.36	0.04	0.35	0.03	0.36	0.02
Total	16.75	0.14	15.99	0.03	15.99	0.02	15.98	0.03
pyrope	0.18	0.08	17.34	2.93	15.25	2.06	19.15	2.84
almandine	5.61	3.82	71.95	2.31	74.36	1.25	70.11	1.59
spessartine	1.35	0.61	4.72	1.15	4.51	0.85	4.75	1.28
uvarovite	0.06	0.07	0.05	0.05	0.03	0.06	0.05	0.05
grossular	27.32	9.78	5.16	0.96	5.33	0.41	5.12	1.11
andradite	65.49	11.85	0.78	1.42	0.53	0.78	0.82	1.37
pyralspite	7.14	3.82	94.01	0.60	94.12	0.45	94.01	0.35

Table 3.8: Summary table of average garnet compositions from Haveri. The values shown have been recalculated for ferric iron content according to Droop (1987).

The compositions of the garnets analysed at Haveri (Figure 3.34) clearly indicate that both the pyralspite and the ugrandite series are represented. The pyralspite garnets are from metasedimentary rocks (samples P5-16 and P6-43), and therefore represent the metamorphic garnets, that are variably associated with biotite and/or cordierite. The ugrandite series garnets are those observed in the calcic-skarn assemblages, usually adjacent to pyroxene. These are thus clearly metasomatic garnets, and are andradite-rich garnets.

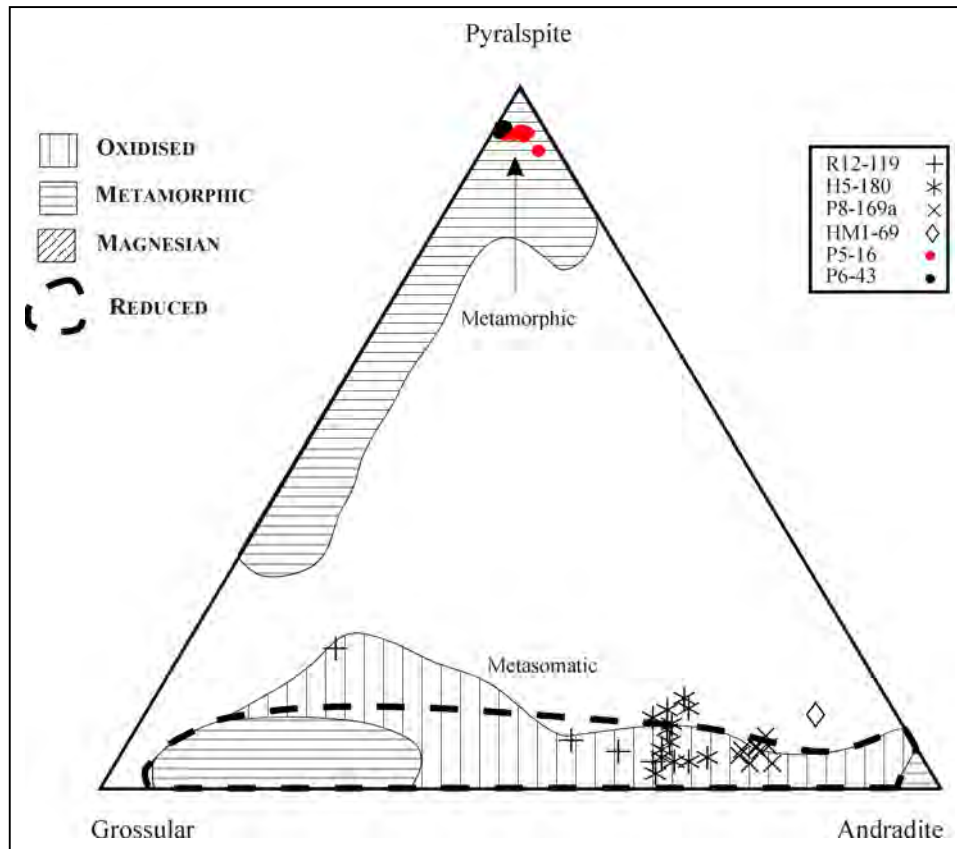


Figure 3.34: Compositional range of garnets at Haveri. Also shown are the fields of garnet compositions from reduced, oxidised, magnesian and metamorphic Au skarns (after Meinert, 1998).

The garnet-biotite exchange thermometer is based on the temperature dependent exchange of Mg and Fe between coexisting biotite and garnet, and is thus of particular application to determining P - T - X conditions of pelitic rocks. The garnet-biotite geothermometer, and the phase relations of garnet with biotite and cordierite are discussed in a later section on the metamorphic P - T - X conditions of pelites (Section 4.1.1). The metamorphic garnets in the Haveri pelites are dominated by the almandine molecular species, with lesser amounts of the pyrope and spessartine molecules (Table 3.8). However, there is a compositional zonation across the garnet crystal (Figure 3.35), with the rim (outermost 0.5 mm) being enriched in iron (pyrope) relative to magnesium (almandine), and the spessartine content remaining constant.

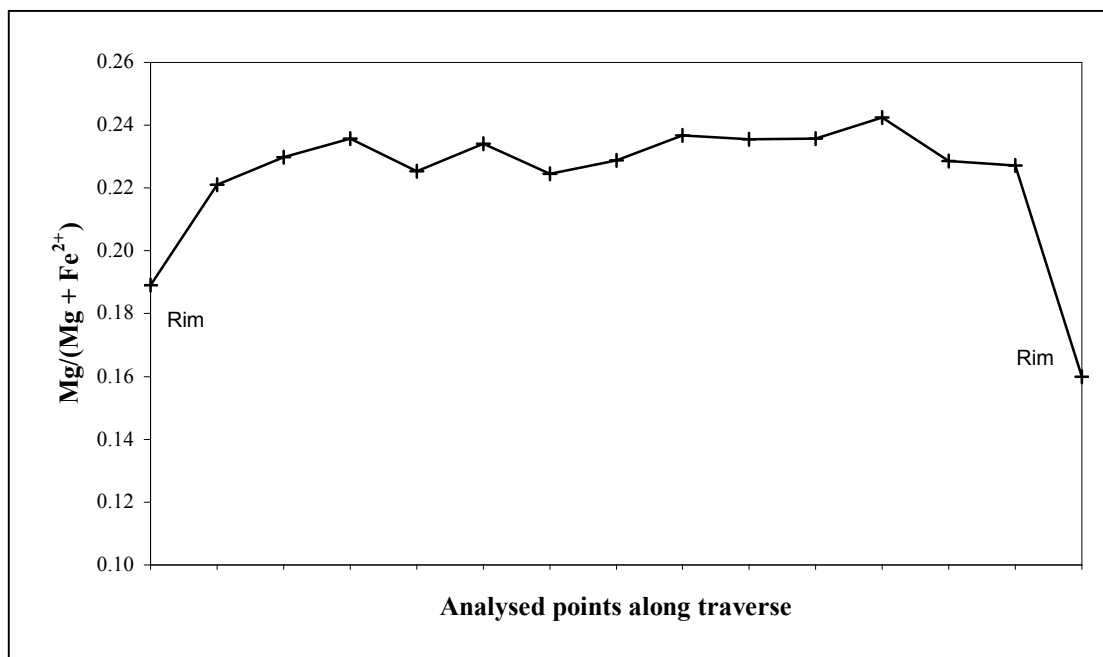


Figure 3.35: Compositional variation across a 5mm diameter metamorphic garnet (sample P6-43).

The metasomatic garnets are dominated by andradite (average = 65%) and grossular (average = 27%) molecular components, with minor amounts of the pyrospite (average = 7%) components. Grandite series garnets are characteristic of calcic-skarn alteration, and the variation in their chemistry has been used to aid in the classification of skarns. Meinert (1998) has identified the compositional fields of garnets associated with various types of gold skarns (Figure 3.34). It is apparent, that the majority of the data for the metasomatic garnets associated with the calcic-skarn alteration falls within the fields of reduced and oxidised Au skarns, and that the metamorphic garnets, not unsurprisingly, fall within the field of metamorphic Au skarns. The chemistry of the pyroxenes associated with the garnet in the calcic-skarn assemblage indicated that the likely skarn type was either a reduced or metamorphic Au skarn. On the basis of the combined garnet (metasomatic) and pyroxene chemistries, it is considered most likely, that the calcic-skarn alteration observed at Haveri formed under conditions similar to those that Meinert (1998) has recognised for the generation of reduced Au skarns. These conditions, and the implications for an ore genetic model for Haveri, will be discussed further in a later section.

3.2.2.4 Plagioclase

Plagioclase is one of the most abundant minerals at Haveri and is an integral part of most mineral assemblages. Plagioclase is a member of the framework silicate feldspar group, a ternary system with end members: albite (Ab), anorthite (An) and KAlSi_3O_8 (K-feldspar, Or). Compositions between $\text{NaAlSi}_3\text{O}_8$ and KAlSi_3O_8 are referred to as alkali feldspars, and those between $\text{NaAlSi}_3\text{O}_8$ and $\text{CaAl}_2\text{Si}_2\text{O}_8$ as plagioclase feldspars. The alkali feldspars generally contain less than 5 – 10 % of the

calcium feldspar in solid solution; although this increases with sodium-rich K-feldspars. Similarly plagioclase feldspars generally contain less than 5 to 10 per cent of the potassium component. Other ions which may be present in very limited amounts include Ti, Fe^{3+} , Fe^{2+} , Mn, Mg, Ba and Sr (Deer et al., 1992). Electron microprobe analyses are recalculated stoichiometrically on the basis of 32(O) in the unit cell, to give the formulae, and the analyses are re-stated in terms of the molecular percentage of their end-member components, albite, anorthite and orthoclase. Most of the iron reported in feldspar analyses is Fe^{3+} , and is considered to replace Al in the structure. Plagioclase compositions are usually expressed in terms of the two end-members albite and anorthite $\text{Ab}_x\text{An}_{100-x}$. The series is divided arbitrarily into albite, oligoclase, andesine, labradorite, bytownite and anorthite at An mole percentages of 0 – 10, 10 – 30, 30 – 50, 50 – 70, 70 – 90, 90 – 100 respectively.

In total, 58 feldspar analyses were completed, from 9 separate samples. This includes 54 plagioclase analyses and 4 k-feldspar analyses (Figure 3.36). The averages of these samples have been summarised in Table 3.8, and the full dataset are supplied in Appendix C.

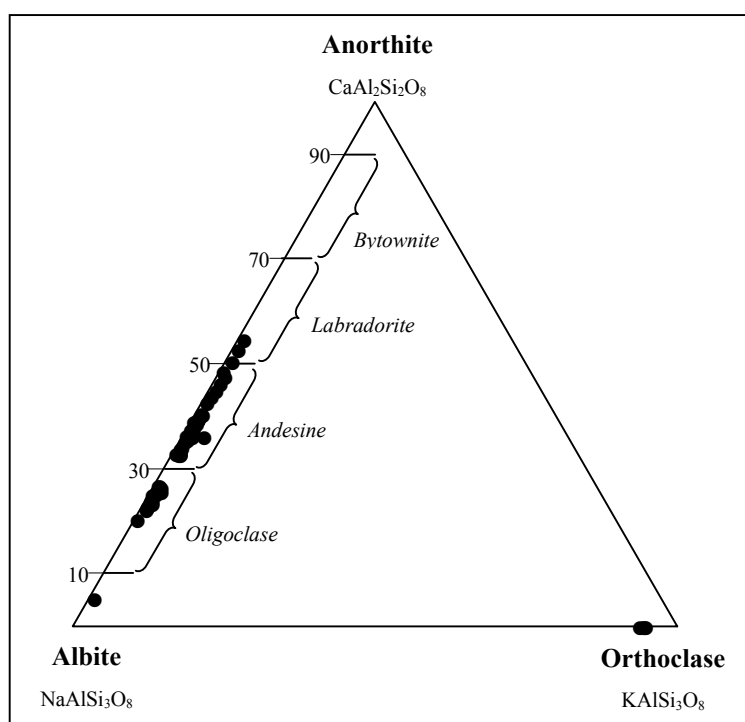


Figure 3.36: Compositional range of homogeneous high-temperature disordered feldspars (after Deer et al., 1992)

The electron microprobe analyses of the feldspars at Haveri (Figure 3.36), show that the plagioclases have compositions between $\text{An}_{5.4}$ and $\text{An}_{54.9}$, and the K-feldspars have compositions of approximately Or_{94} , corresponding to sanidine (Deer et al., 1982). The range of data exhibited for the compositions of plagioclase reflect both metamorphic and hydrothermal compositions.

Sample	In rock mass				Amphibole veinlet				Amphibole alteration				By QV	with scapolite		Plag. Porphyry		K-spar Sediment	
	Metamorphic		Rim		Far		Near		Pervasive		Intense								
Alteration					Amphibolitisation		Amphibolitisation		Amphibolitisation		Amph.	Amph.	Amph.	Calcic Skarn					
n	17		3		3		8		17		1	1		2		2		4	
	Average	Std. Dev.	Average	Std. Dev.	Average	Std. Dev.	Average	Std. Dev.	Average	Std. Dev.	Average	Average		Average	Std. Dev.	Average	Std. Dev.	Average	Std. Dev.
SiO ₂	56.91	1.58	53.06	0.51	57.64	0.70	60.45	3.42	61.21	0.99	64.74	62.15		59.94	0.12	58.23	0.13	62.92	0.73
TiO ₂	0.02	0.02	0.04	0.01	0.03	0.01	0.00	0.01	0.07	0.20	0.00	0.00		0.00	0.00	0.00	0.00	0.06	0.05
Al ₂ O ₃	26.91	0.79	29.33	0.13	26.69	0.41	25.38	1.66	23.94	0.66	20.95	23.65		25.20	0.26	25.80	0.29	18.33	0.09
Fe ₂ O ₃	0.19	0.17	0.15	0.04	0.14	0.05	0.17	0.11	0.24	0.14	0.09	0.42		0.18	0.01	0.12	0.04	0.09	0.07
MgO	0.02	0.07	0.00	0.00	0.00	0.00	0.00	0.01	0.01	0.04	0.00	0.00		0.00	0.00	0.00	0.00	0.00	0.00
CaO	8.45	0.90	11.08	0.53	8.24	0.57	6.96	0.89	5.20	0.28	1.20	4.82		6.33	0.00	7.46	0.28	0.00	0.00
Na ₂ O	6.87	0.69	5.41	0.22	7.10	0.01	7.03	0.69	8.72	0.40	11.59	7.90		7.03	0.09	6.62	0.03	0.72	0.06
K ₂ O	0.07	0.02	0.07	0.01	0.25	0.25	0.10	0.05	0.09	0.04	0.07	0.00		0.04	0.06	0.00	0.00	16.68	0.19
Total	99.43	0.82	99.14	0.77	100.08	0.13	100.09	0.83	99.47	0.79	98.64	98.96		98.72	0.09	98.23	0.13	98.80	0.57
Si	10.26	0.20	9.68	0.04	10.33	0.12	10.72	0.44	10.93	0.11	11.56	11.08		10.76	0.03	10.55	0.02	11.86	0.05
Al	5.72	0.20	6.31	0.03	5.64	0.09	5.31	0.41	5.04	0.13	4.41	4.97		5.33	0.05	5.51	0.06	4.07	0.04
Fe ³⁺	0.03	0.02	0.02	0.01	0.02	0.01	0.02	0.01	0.03	0.02	0.01	0.06		0.02	0.00	0.02	0.00	0.01	0.01
Ti	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.01	0.03	0.00	0.00		0.00	0.00	0.00	0.00	0.01	0.01
Mg	0.01	0.02	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.00		0.00	0.00	0.00	0.00	0.00	0.00
Ca	1.63	0.19	2.17	0.09	1.58	0.11	1.33	0.19	1.00	0.06	0.23	0.92		1.22	0.00	1.45	0.05	0.00	0.00
Na	2.40	0.23	1.92	0.09	2.47	0.00	2.42	0.26	3.02	0.15	4.01	2.73		2.45	0.03	2.33	0.01	0.26	0.02
K	0.02	0.01	0.02	0.00	0.06	0.06	0.02	0.01	0.02	0.01	0.02	0.00		0.01	0.01	0.00	0.00	4.01	0.07
TOTAL	20.07	0.09	20.11	0.03	20.10	0.05	19.83	0.37	20.05	0.12	20.24	19.77		19.79	0.00	19.85	0.00	20.23	0.08
Ab	59.24	4.86	46.75	2.18	60.11	1.05	64.28	1.59	74.80	1.49	94.21	74.78		66.61	0.52	61.63	1.02	6.19	0.42
An	40.39	4.85	52.88	2.18	38.52	2.14	35.11	1.66	24.71	1.46	5.40	25.22		33.13	0.16	38.37	1.02	0.00	0.00
Or	0.38	0.14	0.37	0.06	1.38	1.43	0.61	0.28	0.50	0.24	0.39	0.00		0.26	0.36	0.00	0.00	93.81	0.42
Max An	48.70		54.90		40.40		36.80		26.90					33.20		39.10			
Min An	32.90		50.60		36.20		32.90		22.40					33.00		37.60			

Table 3.9: Summary table of average feldspar compositions from Haveri. The plagioclases described as in the rock mass, have no obvious and visible alteration associated with them, and are interpreted as metamorphic. The rim plagioclases are analyses taken from the rims of the metamorphic plagioclases. The plagioclases associated with amphibolitisation are arranged with intensity of alteration increasing to the right. The sample close to the quartz vein has amphibolitisation associated with it. The sample from the calcic-skarn assemblage is associated with scapolite. The samples from the porphyry and the metasediment have no associated alteration.

The metamorphic plagioclases are compositionally defined as andesines, with an average composition of An₄₀ (the range of compositions are approximately An₃₃ to An₄₉). These are typical values for plagioclase formed during low-pressure amphibolite-facies metamorphism (Liou et al., 1974). The wide range of values may be due to a number of reasons. The plagioclases with lower anorthite contents may have been formed at lower temperatures than the plagioclases with higher anorthite contents, perhaps further down the thermal gradient around the metamorphic heat source (e.g. granitoid intrusions). Another possibility is that they have been affected by albitisation accompanying hydrothermal alteration. It is apparent that with increasing amphibolitisation there is a concomitant decrease in the anorthite content of the plagioclase (Table 3.9). This can be detected even in the vicinity of thin (< 1mm) amphibole veinlets, where the plagioclase in the surrounding rock displays a reduction in the anorthite content to an average of An₃₅ (the range of compositions are approximately An₃₃ to An₃₇). As the amphibolitisation becomes more pervasive (amphibole veinlets and spotted amphibolites) the anorthite contents of the plagioclases drop to an average of An₂₅, with a range between An₂₂ and An₂₇. At its extreme, in a gold-bearing sample, where there is intense amphibolitisation surrounding a quartz vein, the plagioclase is albitic (An_{5.4}) in composition. It is therefore considered likely that the more sodic andesines from the metamorphic class of plagioclase, have actually been affected by weak albitisation, associated with amphibolitisation.

Although it is apparent that there has been albitisation of the plagioclases, associated with the amphibolitisation at Haveri, there is also evidence of some Ca enrichment. Three plagioclase analyses taken from the rims of the metamorphic plagioclases, are clearly more calcic (An₅₁₋₅₅) than the typical “core” analyses. These rim analyses are enriched in Ca and Al, and depleted in Si. Furthermore there is no significant enrichment in Fe and/or Mg. This suggests that this is a true mineral zonation, rather than a bad analysis partially analysing the surrounding amphibole. A plagioclase zonation with a Ca enriched rim is likely to reflect either an increase in temperature during formation or alteration by a Ca-rich hydrothermal fluid.

The coexistence of hornblende and plagioclase has been widely used in geothermobarometry (e.g. Blundy and Holland, 1990; Holland and Blundy, 1994), and is thus especially applicable to mafic rocks metamorphosed to amphibolite facies. The application of hornblende-plagioclase thermobarometry is examined in detail in section 4.1.2.1.

3.2.2.5 Epidote

Epidote is present at Haveri, both in the metamorphic mineral assemblage as well as the hydrothermal alteration mineral assemblage. Epidote has the chemical formula $\text{Ca}_2\text{Al}_2\text{O}(\text{Al},\text{Fe}^{3+})\text{OH}[\text{Si}_2\text{O}_7][\text{SiO}_4]$, with the main replacement being between Fe^{3+} and Al. The two end-members of this substitution are $\text{Ca}_2\text{Al}_3\text{Si}_3\text{O}_{12}(\text{OH})$ (clinozoisite) and $\text{Ca}_2\text{Fe}^{3+}\text{Al}_2\text{Si}_3\text{O}_{12}(\text{OH})$ (pistacite, Ps). Electron microprobe analyses are recalculated stoichiometrically on the basis of 12.5(O) in the unit cell, to give the formulae, and the analyses are re-stated in terms of the molecular percentage of the pistacite molecule, that is a measure of the $\text{Fe}^{3+}:\text{Al}$ ratio.

In total, 20 epidote analyses were completed, from 6 separate samples. This includes epidotes from the calcic-skarn alteration assemblage, as well as epidotes formed as part of a retrograde mineral assemblage. The averages of these samples have been summarised in Table 3.10, and the full dataset are supplied in Appendix C.

Section	R12-19 Associated with carbonate		R12-119 C-S (amp-pyx-gt zoned)		R12-27 With amphibole + carbonate		R8-169a C-S (amp-pyx-gt- scap zoned)		HM1-69 C-S (amp-pyx-gt zoned)		R7-104b Retrograde	
n	3		4		5		3		2		3	
	Average	Std. Dev	Average	Std. Dev	Average	Std. Dev	Average	Std. Dev	Average	Std. Dev	Average	Std. Dev
SiO₂	39.45	1.68	38.80	0.44	37.610	1.165	44.00	11.25	37.98	0.55	43.03	0.36
TiO₂	0.03	0.04	0.05	0.06	0.210	0.250	0.08	0.02	0.05	0.00	0.32	0.15
Al₂O₃	25.62	4.21	25.72	1.32	21.130	1.275	19.90	4.75	23.63	1.08	23.01	0.30
Fe₂O₃	8.95	2.24	10.73	1.94	14.561	0.522	12.09	1.99	12.13	1.93	2.12	0.11
MnO	0.14	0.08	0.09	0.02	0.437	0.107	0.10	0.10	0.21	0.04	0.00	0.00
MgO	1.06	1.83	0.02	0.03	0.447	1.000	0.02	0.03	0.00	0.00	0.20	0.08
CaO	23.19	0.97	23.62	0.21	22.551	0.936	21.05	4.62	24.15	0.57	28.05	0.33
Na₂O	0.01	0.02	0.00	0.00	0.018	0.024	0.00	0.00	0.00	0.00	0.00	0.00
K₂O	0.00	0.00	0.00	0.00	0.000	0.000	0.00	0.00	0.00	0.00	0.00	0.00
Total	98.46	1.06	99.03	0.98	96.963	0.692	97.23	0.08	98.14	0.31	96.72	0.64
Si	3.06	0.12	3.02	0.01	3.042	0.074	3.43	0.71	3.01	0.02	3.36	0.00
Al	0.00	0.01	0.00	0.00	0.003	0.007	0.00	0.00	0.00	0.00	0.00	0.00
Al	2.34	0.38	2.36	0.12	2.012	0.130	1.85	0.51	2.21	0.08	2.12	0.02
Fe³⁺	0.52	0.13	0.63	0.11	0.886	0.033	0.72	0.14	0.72	0.12	0.12	0.01
Ti	0.00	0.00	0.00	0.00	0.013	0.015	0.00	0.00	0.00	0.00	0.02	0.01
Mn	0.01	0.00	0.01	0.00	0.030	0.007	0.01	0.01	0.01	0.00	0.00	0.00
Mg	0.12	0.21	0.00	0.00	0.053	0.119	0.00	0.00	0.00	0.00	0.02	0.01
Ca	1.93	0.08	1.97	0.01	1.955	0.087	1.78	0.45	2.05	0.03	2.35	0.04
Na	0.00	0.00	0.00	0.00	0.003	0.004	0.00	0.00	0.00	0.00	0.00	0.00
K	0.00	0.00	0.00	0.00	0.000	0.000	0.00	0.00	0.00	0.00	0.00	0.00
TOTAL	8.00	0.01	7.99	0.01	7.996	0.019	7.79	0.39	8.02	0.01	8.00	0.01
Ps_{avg}	18.6	6.5	21.0	3.9	30.604	1.931	28.2	1.8	24.7	3.8	5.5	0.3
Ps_{max}	26.2		26.8		33.4		30.3		27.4		5.8	
Ps_{min}	14.7		18.7		28.5		27.1		22.0		5.2	

Table 3.10: Summary table of average epidote compositions from Haveri.

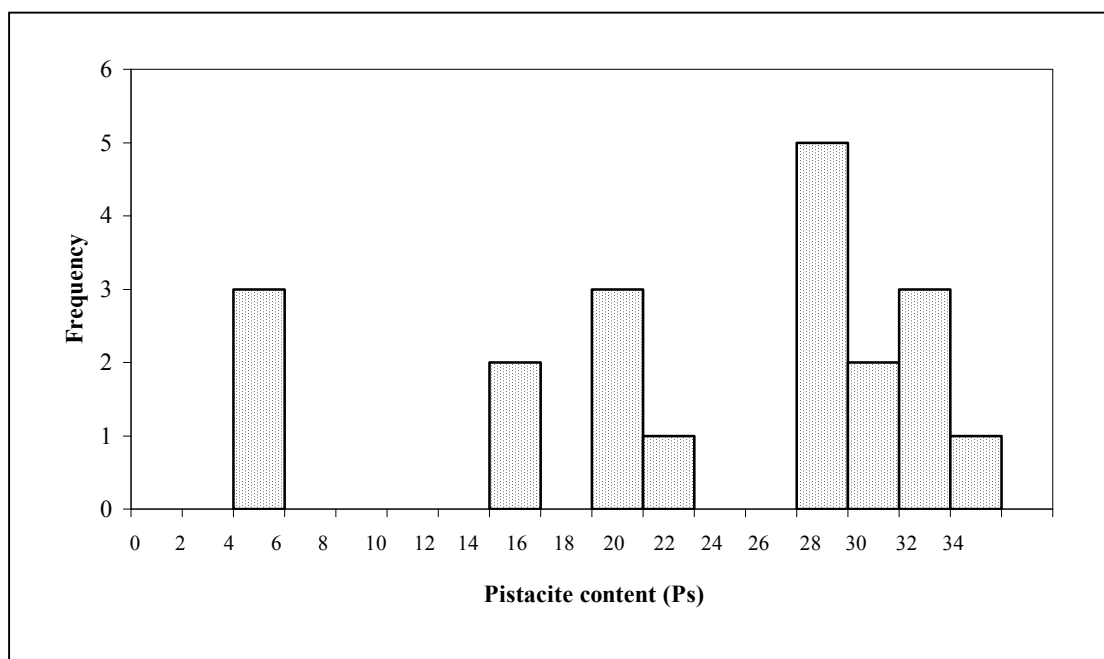


Figure 3.37: Histogram displaying the variation in the pistacite content (Ps) of epidotes from the Haveri area analysed by electron microprobe.

The variation in pistacite contents of the epidotes from Haveri is considerable (Figure 3.37), ranging from $Ps_{5.2}$ to $Ps_{33.4}$. The “flaser” textured epidotes from sample R7-104b are in a retrograde mineral assemblage associated with biotite and the break-down products of cordierite or garnet. These retrograde epidotes have pistacite contents of $Ps_{5.2}$ to $Ps_{8.5}$. The remaining epidotes are all hydrothermal and are associated with either carbonate, amphibole ($\pm$ carbonate), or the zoned calcic-skarn assemblage. All of these hydrothermal epidotes have higher pistacite contents (up to $Ps_{33.4}$), and thus, like the mineral chemistry of the amphiboles, demonstrate iron-enrichment associated with hydrothermal alteration.

3.2.2.6 Scapolite

Scapolite has only been observed at Haveri in one thin section, and is associated with the proximal veinlet of the calcic-skarn hydrothermal alteration assemblage. The presence of scapolite at Haveri, may however be more widespread and largely unrecognised, due to its similarities in appearance to plagioclase in hand specimen. The composition of scapolite may be expressed by the general formula $W_4Z_{12}O_{24}.R$ where:

$$W = \text{Ca, Na, K}$$

$$Z = \text{Si, Al}$$

$$R = \text{Cl, CO}_3, \text{SO}_4$$

Compositions of the two main end-members of the scapolite series may be expressed as $3\text{NaAlSi}_3\text{O}_8 \cdot \text{NaCl}$ (marialite, Ma), and $3\text{CaAl}_2\text{Si}_2\text{O}_8 \cdot \text{CaCO}_3$ (meionite, Me). Three substitutions occur in members of the series: $\text{Na}^+ \leftrightarrow \text{Ca}^{2+}$, $\text{Al}^{3+} \leftrightarrow \text{Si}^{4+}$, $\text{Cl}^- \leftrightarrow \text{CO}_3^{2-}$; all are active in the more sodic ($\text{Ca}/(\text{Ca} + \text{Na}) < 0.75$) part of the series. In the more calcic ($\text{Ca}/(\text{Ca} + \text{Na}) > 0.75$) part, the minerals are fully carbonated and the chemical variation is governed by the coupled substitution $\text{NaSi} \leftrightarrow \text{CaAl}$. The compositional variation of the common scapolites can thus be considered to be due to the presence within marialite – meionite of two solid-solution series:

$\text{Me}_0 - \text{Me}_{75}$ governed by the coupled replacement $[\text{Na}_4\text{Cl}]\text{Si}_2 \leftrightarrow [\text{NaCa}_3.\text{CO}_3]\text{Al}_2$

$\text{Me}_{75} - \text{Me}_{100}$ characterised by the substitution $[\text{NaCa}_3.\text{CO}_3]\text{Si} \leftrightarrow [\text{Ca}_4.\text{CO}_3]\text{Al}$

Natural pure end-members are unknown and the majority of scapolite compositions range from Me_{20} to Me_{80} (Deer et al., 1992).

Electron microprobe analyses of scapolite are unable to provide information on CO_3 and SO_4 , and so formula recalculations are based on $12(\text{Si},\text{Al})$ in the unit cell. From this, the analyses can be restated in terms of the molecular percentage of the meionite molecule.

Six electron microprobe analyses were made from two grains of scapolite in a zoned calcic-skarn veinlet in sample R8-169a (Table 3.11). These analyses indicate that the few scapolites encountered at Haveri are dominated by the meionite component, with values of $\text{Me}_{51.6}$ to $\text{Me}_{61.0}$ and chlorine contents of 0.76 – 1.38 wt% Cl.

Scapolite is most commonly found in metamorphic and metasomatic rocks, particularly in skarns, where it is clearly related to metasomatic processes accompanying the emplacement, crystallisation, and cooling of intrusions (Einaudi et al., 1981). Although a wide range of compositions has been reported for scapolite over the entire spectrum of skarn deposits, marialitic scapolite is characteristic of most shallow skarn deposits, whereas the meionitic variety is commonly found in those from medium- to high-grade deposits (Pan, 1998). In addition to being found in skarn deposits, scapolite is regionally distributed throughout northern Fennoscandia, leading Frietsch et al. (1997) to suggest regional Na-Cl metasomatism related to igneous activity. Such alteration, and the presence of scapolite, is commonly associated with Fe-oxide Cu-Au deposits (Williams, 1999a).

Section	R8-169a Scapolite 1.1	R8-169a Scapolite 1.2	R8-169a Scapolite 1.3	R8-169a Scapolite 1.4	R8-169a Scapolite 1.5	R8-169a Scapolite 2.1
Al₂O₃	25.01	25.93	26.18	26.15	25.31	25.59
SiO₂	49.02	49.04	49.14	48.54	48.54	49.83
CaO	13.11	14.14	13.61	14.48	14.30	11.71
MnO	0.02	0.00	0.02	0.03	0.02	0.01
Na₂O	5.24	4.88	5.21	5.07	4.98	5.93
K₂O	0.19	0.20	0.23	0.26	0.26	0.29
MgO	0.00	0.00	0.00	0.02	0.00	0.02
FeO	0.06	0.07	0.06	0.05	0.06	0.06
TiO₂	0.00	0.00	0.01	0.00	0.03	0.00
Cr₂O₃	0.00	0.02	0.00	0.00	0.03	0.00
Cl	1.14	0.76	1.21	1.08	1.37	1.38
F	0.00	0.03	0.00	0.00	0.00	0.01
Total	93.79	95.07	95.67	95.69	94.90	94.82
Al	4.51	4.61	4.63	4.66	4.57	4.52
Si	7.49	7.39	7.37	7.34	7.43	7.48
Ca	2.15	2.28	2.19	2.35	2.35	1.88
Mn	0.00	0.00	0.00	0.00	0.00	0.00
Na	1.55	1.43	1.51	1.49	1.48	1.72
K	0.04	0.04	0.04	0.05	0.05	0.06
Mg	0.00	0.00	0.00	0.00	0.00	0.00
Fe²⁺	0.01	0.01	0.01	0.01	0.01	0.01
Ti	0.00	0.00	0.00	0.00	0.00	0.00
Cr	0.00	0.00	0.00	0.00	0.00	0.00
Cl	0.29	0.20	0.31	0.28	0.35	0.35
F	0.00	0.01	0.00	0.00	0.00	0.00
Total (Si, Al)	12.00	12.00	12.00	12.00	12.00	12.00
Me	57.56	60.98	58.52	60.57	60.67	51.57
Ma	42.44	39.02	41.48	39.43	39.33	48.43

Table 3.11: The six electron microprobe analyses of the two scapolite grains encountered in the core of the zoned garnet-pyroxene-amphibole calcic-skarn veinlet.

3.2.2.7 Mica

Biotite is a common mineral at Haveri occurring both as a metamorphic mineral and as a hydrothermal mineral perhaps as a distal equivalent of the amphibolitisation. Muscovite is also present, principally as a retrograde breakdown product of garnet and cordierite within the metasedimentary rocks.

Micas are sheet silicates in which a layer of octahedral coordinated Y cations is sandwiched between two identical layers of linked $(\text{Si,Al})\text{O}_4$ tetrahedra. The chemical composition of micas may be expressed by the general formula $X_2Y_{4-6}Z_8\text{O}_{20}(\text{OH},\text{F})_4$ where:

X is mainly K, Na or Ca but also Ba, Rb, Cs, etc.

Y is mainly Al, Mg or Fe but also Mn, Cr, Ti, Li, etc.

Z is mainly Si or Al but perhaps also Fe^{3+} and Ti

The micas can be subdivided into di-octahedral and tri-octahedral classes in which the number of Y ions is 4 and 6 respectively. A further subdivision can be made according to the nature of the principal X constituent. In the common micas X is largely K or Na, but in the so-called “brittle micas” X is largely Ca; further subdivisions of the common micas are made according to the principal constituents in the categories X , Y , and Z . A chemical feature which most micas have in common is their water content; analyses, except for those with high fluorine content, show approximately 4 – 5 percent H_2O^+ (Deer et al., 1992).

The electron microprobe is unable to measure the (OH) content, or to identify the structural sites for the ions, so it is necessary to recalculate the data stoichiometrically. The electron microprobe analyses have been recalculated on the basis of 22(O) to provide the chemical formulae. In the phlogopite – biotite series, phlogopite is the name for compositions near the Mg-rich end-member. There is no agreed demarcation between phlogopite and biotite, although the Fe-rich analogue of phlogopite is annite.

In total 26 electron microprobe analyses were made of biotites from six separate samples. The average compositions of these biotites are shown in Table 3.12. The full dataset is supplied in Appendix C. The $\text{Mg}/(\text{Mg} + \text{Fe}^{2+})$ of the analysed biotites at Haveri range between 0.38 and 0.61. The composition of biotite in equilibrium with other ferromagnesian minerals is important in terms of phase relations, and geothermometry. The phase relationships of biotite and coexisting minerals in pelites, as well as the application of the biotite – garnet thermometer are discussed in greater detail in Section 4.1.1.

Section	R6-284		P13-39		P13-67a	P5-16		R7-104b	P6-43	
n	2		5		1	12		1	5	
	Average	Std. Dev	Average	Std. Dev		Average	Std. Dev		Average	Std. Dev
SiO ₂	35.35	0.04	35.33	0.89	34.68	35.70	0.44	30.27	37.21	0.27
TiO ₂	2.46	0.05	0.94	0.13	2.55	1.54	0.05	1.32	1.20	0.07
Al ₂ O ₃	15.91	0.06	16.66	0.48	14.93	18.32	0.38	16.00	18.29	0.16
FeO	21.98	0.28	24.60	0.42	21.35	19.11	0.77	28.01	16.23	0.26
MnO	0.09	0.00	0.30	0.03	0.23	0.02	0.04	0.16	0.05	0.03
MgO	9.72	0.16	8.79	0.07	10.98	11.55	0.20	10.06	13.72	0.15
CaO	0.05	0.00	0.03	0.04	0.00	0.02	0.03	0.10	0.01	0.01
Na ₂ O	0.11	0.00	0.08	0.03	0.07	0.29	0.05	0.07	0.52	0.02
K ₂ O	10.11	0.24	8.33	1.03	9.34	8.75	0.70	4.38	7.28	0.22
Total	95.78	0.64	95.04	0.52	94.13	95.29	0.93	90.36	94.52	0.42
Si	5.47	0.03	5.52	0.08	5.45	5.41	0.04	5.03	5.54	0.03
Al	2.53	0.03	2.48	0.08	2.55	2.59	0.04	2.97	2.46	0.03
Total Z	8.00		8.00		8.00	8.00		8.00	8.00	
Al	0.38	0.03	0.58	0.11	0.21	0.68	0.07	0.17	0.75	0.01
Ti	0.29	0.01	0.11	0.02	0.30	0.18	0.01	0.16	0.13	0.01
Fe ²⁺	2.85	0.02	3.21	0.08	2.80	2.42	0.11	3.89	2.02	0.03
Mn	0.01	0.00	0.04	0.00	0.03	0.00	0.00	0.02	0.01	0.00
Mg	2.24	0.03	2.05	0.03	2.57	2.61	0.03	2.49	3.05	0.03
Total Y	5.76		5.99		5.92	5.89		6.74	5.95	
Ca	0.01	0.00	0.00	0.01	0.00	0.00	0.00	0.02	0.00	0.00
Na	0.03	0.00	0.02	0.01	0.02	0.08	0.01	0.02	0.15	0.01
K	2.00	0.04	1.66	0.22	1.87	1.69	0.13	0.93	1.38	0.04
Total X	2.04		1.69		1.89	1.78		0.97	1.54	
TOTAL	15.80		15.68		15.81	15.67		15.71	15.49	
Mg/(Mg+Fe ²⁺)										
Average	0.44		0.39		0.48	0.52		0.39	0.60	
Max	0.44		0.48			0.53			0.61	
Min	0.44		0.38			0.49			0.60	

Table 3.12: Summary table of average biotite compositions from Haveri.

3.2.2.8 Cordierite

Cordierite is found within the prograde metamorphic mineral assemblage in the Osara Formation metasedimentary rocks. The formula for cordierite is (Mg,Fe)₂[Si₅Al₄O₁₈].nH₂O. In most cordierites the octahedrally coordinated positions are occupied predominantly by magnesium, and cordierites containing more than one Fe²⁺ atom per formula unit are rare. Compared with other associated ferromagnesian minerals cordierites are preferentially enriched in magnesium relative to iron. Electron microprobe analyses of cordierites are stoichiometrically recalculated on the basis of 18(O) to provide the chemical formula.

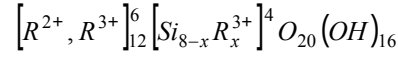
In total five electron microprobe analyses of cordierite were made from two sections. The compositions of these are summarised in Table 3.13 (as well as in Appendix C). The cordierites analysed from Haveri have a range of Mg* of between 59.8 and 75.5 (where $Mg^* = 100 (Mg / (Mg + Fe))$). As can be seen from Table 3.13, the cordierites from the two sections have Mg* values that cluster around 60 and 75. The compositions of the cordierites are controlled by the bulk compositions of the pelites in which they are found, and thus also by the compositions of associated phases (e.g. biotite, garnet, andalusite etc.). This is discussed in more detail in section 4.1.1

Section:	P9-27	P9-27	P9-27	P6-43	P6-43
	Cordierite 1.1	Cordierite 1.2	Cordierite 1.3	Cordierite 1.1	Cordierite 1.2
SiO₂	48.3	48.71	47.64	48.77	48.91
TiO₂	0.01	0.00	0.03	0.01	0.02
Al₂O₃	32.82	33.16	32.98	32.69	33.21
FeO	9.32	9.06	9.40	5.81	5.82
MnO	0.28	0.26	0.22	0.03	0.16
MgO	7.95	7.93	7.83	9.72	10.07
CaO	0.01	0.01	0.02	0.00	0.03
Na₂O	0.08	0.10	0.13	0.25	0.09
K₂O	0.00	0.00	0.00	0.00	0.00
Total	98.78	99.22	98.25	97.31	98.30
Si	4.98	4.99	4.94	5.02	4.98
Ti	0.00	0.00	0.00	0.00	0.00
Al	3.99	4.00	4.03	3.96	3.99
Fe²⁺	0.80	0.78	0.82	0.50	0.50
Mn	0.02	0.02	0.02	0.00	0.01
Mg	1.22	1.21	1.21	1.49	1.53
Ca	0.00	0.00	0.00	0.00	0.00
Na	0.02	0.02	0.03	0.05	0.02
K	0.00	0.00	0.00	0.00	0.00
TOTAL	11.03	11.02	11.05	11.03	11.03
<u>Mg* = 100(Mg/Mg+Fe)</u>					
Mg*	60.3	61.0	59.8	74.9	75.5

Table 3.13: The five electron microprobe analyses of two cordierite.

3.2.2.9 Chlorite

Chlorite is a widespread retrograde mineral, present in late veins and as a replacement of biotite and amphibole. Most chlorite compositions can be represented by the formula:



where R^{2+} = Mg, Fe, Mn, Ni, Zn

R^{3+} = Al, Fe, Cr

And x (generally) $\sim 1 - 3$

In most common chlorites there are 12.0 octahedral cations per $O_{20}(OH)_{16}$, and approximately equivalent amounts of aluminium in tetrahedral and octahedral sites. Such minerals are referred to as tri-octahedral chlorites. Mg $\leftrightarrow$ Fe substitution occurs over a very wide range and natural tri-octahedral chlorites can as a first approximation be expressed as binary combinations of the $(Mg_{10}Al_2)(Si_6Al_2O_{20})(OH)_{16}$ clinochlore and $(Fe_{10}^{2+}Al_2)(Si_6Al_2O_{20})(OH)_{16}$ chamosite end-members (Deer et al., 1992).

In total 14 chlorites from two samples have been analysed by electron microprobe and recalculated stoichiometrically on the basis of 28(O) and an assumed 16(OH) to provide the structural formula. The average compositions of the two samples are supplied in Table 3.14, and the whole dataset is provided in Appendix C.

The analyses from sample P13-39, which are chlorites replacing biotite have a Mg* (where $Mg^* = 100 (Mg / (Mg + Fe^{2+}))$) of approximately 25, indicating that they are composed mainly of the Fe-rich chamosite end-member. The analyses from sample R7-104b, taken from retrograded cordierites (which are now mainly chlorite), are more Mg-rich. These chlorites have a range of Mg* between 45.5 and 47.3, and are thus intermediate between chamosite and clinochlore.

A number of the chemical variables in chlorite have been shown to be temperature dependent and thus chlorite has been used as a geothermometer (Cathelineau, 1988; Cathelineau and Nieva, 1985; McDowell and Elders, 1980). The chlorite geothermometry is examined in greater detail in Section 4.4.

Section n	P13-39		R7-104b	
	Replacing biotite 4		Retrograde cordierite 10	
	Average	Std. Dev	Average	Std. Dev
SiO₂	24.37	0.17	25.09	0.37
TiO₂	0.21	0.05	0.10	0.03
Al₂O₃	14.89	0.30	20.69	0.54
Cr₂O₃	0.04	0.03	0.01	0.01
FeO	40.12	0.20	27.85	0.31
MnO	0.17	0.02	0.21	0.04
MgO	7.41	0.10	13.56	0.35
CaO	0.06	0.03	0.00	0.00
Na₂O	0.03	0.03	0.00	0.00
K₂O	0.00	0.00	0.00	0.00
Total	87.29	0.46	87.51	1.04
Si	5.68	0.03	5.38	0.05
Al^{IV}	2.32	0.03	2.62	0.05
Al^{VI}	1.76	0.04	2.61	0.05
Cr	0.01	0.01	0.00	0.00
Ti	0.04	0.01	0.02	0.01
Fe²⁺	7.81	0.09	4.99	0.10
Mn	0.03	0.00	0.04	0.01
Mg	2.57	0.02	4.33	0.07
Ca	0.02	0.01	0.00	0.00
Na	0.01	0.01	0.00	0.00
K	0.00	0.00	0.00	0.00
TOTAL	12.25		11.99	
OH	16.00		16.00	
Mg*=(Mg/Mg+Fe)	24.78	0.34	46.47	0.81

Table 3.14: Summary table of average chlorite compositions from Haveri, determined by electron microprobe analysis.

3.2.2.10 Sphalerite

Sphalerite occurs within the calcic-skarn hydrothermal mineralisation, and is a relatively uncommon mineral at Haveri. Sphalerite has the simple chemical formula of ZnS. The principal substitute for zinc in the sphalerite structure is iron, with up to 26 wt% Fe being reported (Deer et al., 1992). Cadmium, copper, silver and tin are also commonly reported in sphalerite analyses but some of these elements may be present in small inclusions of other minerals.

Section Description	R12-51		R12-51	R12-51
	Average	Std.Dev.	Max FeS (mol%)	Min FeS (mol%)
n	18			
Zn (wt%)	57.83	1.16	57.10	60.90
S (wt%)	34.30	0.48	34.69	33.75
Fe (wt%)	7.48	0.66	8.48	6.12
Total	99.61	0.93	100.27	100.78
FeS (mol%)	13.16	1.2	14.81	10.52
ZnS (mol%)	86.84	1.2	85.19	89.48

Table 3.15: Summary table of sphalerite compositions determined by electron microprobe.

In total, 18 electron microprobe analyses of sphalerite were completed from one section. The average composition (as well as the most and least Fe-rich) are shown in Table 3.15. The complete dataset is provided in Appendix C. The analyses have been recalculated to show the mole percent FeS content of the sphalerites (Figure 3.38), which is a measure of the amount of iron that has substituted for zinc, being controlled by pressure, temperature and associated iron sulphides. The range of values is between 10.5 and 14.8 mole % FeS.

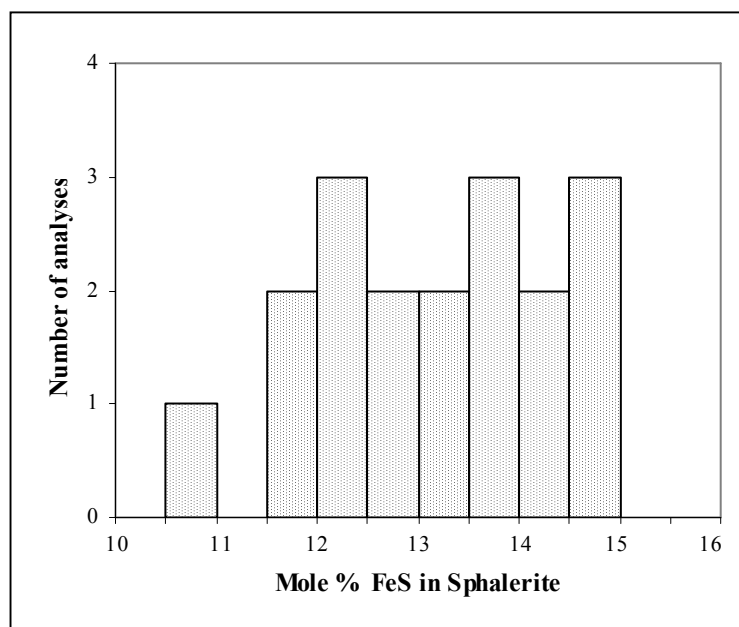


Figure 3.38: Electron microprobe analyses of sphalerites from Haveri showing the replacement of iron for zinc as mole % FeS in sphalerite.

The iron content in sphalerite is sensitive to pressure, temperature and FeS activity, and can be useful as a geobarometer (Barton and Toulmin, 1966; Lusk and Ford, 1978; Scott, 1973, 1976; Scott and Barnes, 1971). This is examined in greater detail in Section 4.3.2.

3.2.2.11 Gold

Gold at Haveri occurs in a number of associations, including the intense zones of amphibole alteration, quartz veins, and in association with sulphides. However, visible gold is rarely observed in the sulphides, usually only being identified from assay results. Visible gold has been analysed by the electron microprobe to determine the fineness of the gold (i.e. the Au:Ag ratio expressed as parts per thousand)

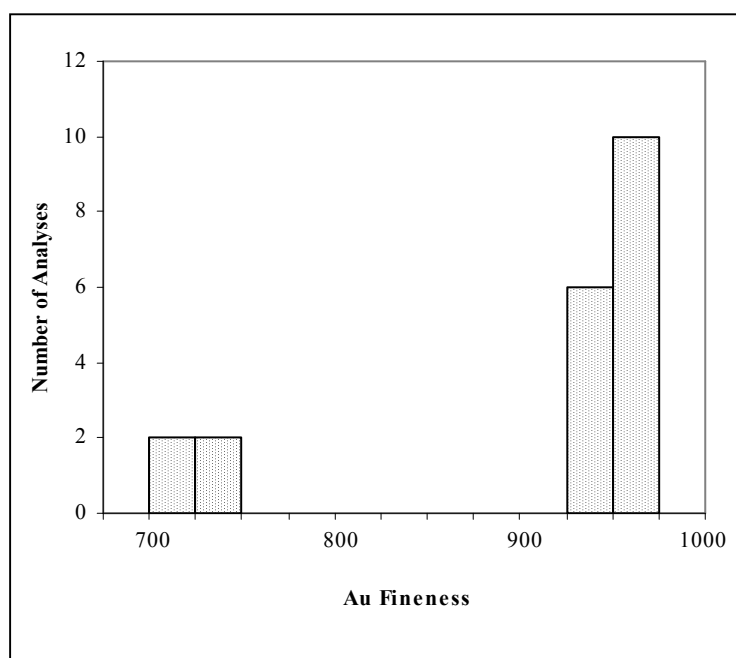


Figure 3.39: Electron microprobe analyses of gold grains at Haveri showing the fineness of the gold where fineness = 1000 ($Au/(Au + Ag)$).

Two populations of gold fineness can be clearly identified (Figure 3.39). The bulk of the analyses have a very high fineness of between 927 and 972. These analyses are from discrete, individual, gold grains up to 100 μm in size. The gold grains are situated in either quartz veins or in intensely amphibolitised rock, both with and without associated sulphides. The sulphides, when present, are pyrrhotite and chalcopyrite, which are not actually in contact with the gold. The second population of analyses has a much higher silver content, and has a range of fineness between 704 and 740. This population is from a sample where gold is visible as discrete inclusions within pyrrhotite grains, which themselves are in amphibolitised volcanics.

3.3 STABLE ISOTOPES

The stable isotopes of hydrogen, carbon, oxygen and sulphur are powerful tracers of fluid sources, sources of carbon and sulphur, and of the processes of fluid-rock interaction in a wide variety of hydrothermal environments. The utility of these isotopes rests in part on the fact that many of the major crustal and upper-mantle reservoirs of rock and water are isotopically distinct from one another in one or more of H, C, O and S. Thus fluids equilibrated with one reservoir commonly will undergo significant isotopic modification upon interaction with other crustal (rock or fluid) reservoirs. Further, significant changes in isotopic ratios of these elements result from many geological and geochemical processes pertinent to hydrothermal systems including decarbonation reactions, precipitation, fluid-rock interaction, and oxidation/reduction reactions. In principle then, the isotopic compositions of these elements in hydrothermal systems, and their variation in space and time within the system, can tell us much about the sources of fluids and dissolved components, and about the physico-chemical processes resulting in the hydrothermal alteration assemblages (Bowman, 1998b).

3.3.1 Analytical Techniques

Isotopic analyses were carried out for O, H, C, and S on a range of whole rock samples and mineral separates. CO₂ was extracted from 8 carbonate samples (6 calcite and 2 dolomite) by the author at the University of Cape Town (UCT), aided by Dr. Ron Sheets. The carbon isotope analyses as well as the oxygen isotope analyses of the CO₂ extracted from the carbonates were completed by the staff of UCT (Table 3.17), as well as 12 whole rock oxygen isotope analyses and 7 oxygen isotope analyses on quartz from quartz veins, were undertaken by UCT staff (Table 3.16). A further 16 oxygen isotope analyses from mineral separates (including a duplicate of a quartz sample analysed at UCT) and 6 hydrogen isotope analyses on the hydrous minerals, were analysed at the Department of Geological Sciences, Queens University, Canada (Table 3.16). Sulphur isotope analyses were carried out on 15 pyrrhotite separates (including one duplicate analysis) at the Central Science Laboratory, University of Tasmania (UTA), Australia (Table 3.17).

The 12 whole rock samples for analysis at UCT were crushed and sieved, and the –120 mesh sample was milled to produce a sample ready for oxygen extraction. The quartz vein material analysed at UCT, was prepared by reacting the crushed quartz vein material with aqua regia, and then hand picking the clean quartz grains using a binocular microscope. The mineral separates sent to Queens University and the University of Tasmania, were prepared by crushing the sample material, and then washing the sample through a series of sieves with distilled water. The medium-grained portion of the sample material was then further segregated by running the sample through a magnetic separator several times and at different settings, according to the particular mineral separate required. The final separation, was performed by hand picking the purest mineral fragments using a binocular microscope.

Oxygen (O_2) was liberated from the silicate minerals with bromine pentafluoride (BrF_5) on a conventional extraction line. Hydrogen was liberated using an RF generator to release the water and then converting it to hydrogen (H_2) by reaction with depleted uranium at high temperature. The pyrrhotite samples were reacted with V_2O_5 at high temperature to produce SO_2 .

The carbonate samples were crushed and sieved, and the -120 mesh sample was milled. A portion of this was analysed using an X-Ray Diffractometer at Rhodes University, to determine the approximate total carbonate content of the samples, and the relative proportions of carbonate types. The samples were dried in an oven at $50\text{ }^\circ\text{C}$ overnight. The calcite samples were reacted with phosphoric acid (H_3PO_4) at $25\text{ }^\circ\text{C}$ for at least two hours, and the CO_2 extracted. Samples with dolomite in were further reacted at $50\text{ }^\circ\text{C}$ to prevent mixing of CO_2 from dolomite and any remaining calcite that had not reacted fully with the phosphoric acid. The sample was then reacted at $50\text{ }^\circ\text{C}$ for at least eight hours, and the CO_2 released from the dolomite extracted for analysis.

The isotopic compositions of the gas extracts were measured using mass spectrometers. The final isotopic data is then reported in terms of isotope ratios relative to a standard, and are expressed in parts per thousand, i.e. parts per mil (‰). The isotope ratio is expressed as a δ value. The standard used for oxygen ($^{18}\text{O}/^{16}\text{O}$) and hydrogen (D/H) is Standard Mean Ocean Water (SMOW). Carbon isotope compositions are reported as a measure of the ratio of $^{13}\text{C}/^{12}\text{C}$ relative to a standard belemnite sample from the Cretaceous Peedee Formation (PDB). Sulphur isotope compositions are reported as a measure of the ratio of $^{34}\text{S}/^{32}\text{S}$ relative to the reference standard of troilite (FeS) from the Canon Diablo iron meteorite (CDT) (Rollinson, 1993).

The analytical precision for the reported isotopic compositions are approximately $\pm 0.2\text{ }‰$ for $\delta^{18}\text{O}$ values; $\pm 5\text{ }‰$ for δD values; $\pm 0.1\text{ }‰$ for $\delta^{13}\text{C}$ values; and $\pm 0.2\text{ }‰$ for $\delta^{34}\text{S}$ values.

During this study only two duplicate samples were analysed: a sample of quartz from a quartz vein (P11-79) was analysed both at UCT and at Queens; and a sample of pyrrhotite from a cataclastite (R7-191) was analysed twice (with different laboratory numbers) at the University of Tasmania. The duplicate quartz sample (P11-79) has a difference in $\delta^{18}\text{O}$ values of $0.65\text{ }‰$ (Table 3.16), which suggests that there is a minor variation between laboratories (beyond analytical precision). However, both of these analyses are anomalously high relative to the other quartz vein samples. The sample of pyrrhotite analysed twice has a difference in $\delta^{34}\text{S}$ values of $0.19\text{ }‰$ (Table 3.17), which is within the expected analytical error ($\pm 0.2\text{ }‰$ for $\delta^{34}\text{S}$ values).

<i>Sample</i>	<i>Description</i>	<i>Mineral etc.</i>	δ^8O	δD	<i>Laboratory</i>
TS-2	Granodiorite (NW of Haveri)	Whole Rock	7.94		UCT
TS18	Granodiorite (R13)	Whole Rock	8.82		UCT
TS18	Granodiorite (R13)	Quartz	10.3		Queens
TS18	Granodiorite (R13)	Plagioclase	7.8		Queens
TS18	Granodiorite (R13)	Biotite	4.3	-86	Queens
TS15	Porphyry (HDD6)	Whole Rock	10.26		UCT
TS15	Porphyry (HDD6)	Quartz	11.1		Queens
TS15	Porphyry (HDD6)	Plagioclase	9.2		Queens
TS15	Porphyry (HDD6)	Biotite	4.4	-77	Queens
TS-9	Porphyry (west of pit)	Whole Rock	10.15		UCT
P8-24	Porphyry (skarned)	Whole Rock	9.67		UCT
P13-97	Porphyry (dark matrix)	Whole Rock	8.44		UCT
P13-98	Uralite porphyry	Whole Rock	7.83		UCT
TS-5	Pillowed amphibolite	Whole Rock	7.56		UCT
P7-105	Vesiculated amphibolite	Whole Rock	7.91		UCT
P7-105	Vesiculated amphibolite	Amphibole	6.8	-91	Queens
P7-105	Vesiculated amphibolite	Plagioclase	10.0		Queens
P5-79	Intense amphibolitisation	Whole rock	7.00		UCT
P7-82	Intense amphibolitisation	Magnetite	1.8		Queens
P7-82	Intense amphibolitisation	Amphibole	6.5	-105	Queens
R11-137	Biotite Hornfels	Whole Rock	8.99		UCT
H5-240	Biotite Hornfels (skarned)	Whole Rock	7.34		UCT
P6-43	Pelite	Biotite	5.8	-74	Queens
R8-168a	Skarned quartz-magnetite vein	Magnetite	2.6		Queens
R8-168a	Skarned quartz-magnetite vein	Quartz	10.1		Queens
R8-168a	Skarned quartz-magnetite vein	Plagioclase	9.0		Queens
R7-331	Quartz vein in amphibolitisation	Quartz	10.47		UCT
R7-331	Quartz vein in amphibolitisation	Amphibole	5.7	-90	Queens
H5-128	Quartz-carbonate vein	Quartz	10.50		UCT
H5-280a	Quartz vein with Au	Quartz	11.25		UCT
H5-291	Quartz vein	Quartz	9.95		UCT
P11-79	Quartz vein	Quartz	14.15		UCT
P11-79	Quartz vein	Quartz	13.5		Queens
R7-309	Quartz vein	Quartz	10.08		UCT
R7-312	Quartz vein	Quartz	9.80		UCT

Table 3.16: δ^8O and δD values for whole rock and mineral analyses from Haveri

<i>Sample</i>	<i>Description</i>	<i>Mineral etc.</i>	$\delta^{18}O$	$\delta^{13}C$	$\delta^{34}S$	<i>Laboratory</i>
P22-62b	Cataclastite	Pyrrhotite			5.99	UTA
H5-291	Cataclastite	Pyrrhotite			2.48	UTA
R7-191	Cataclastite	Pyrrhotite			2.89	UTA
P22-63b	Sulphide & Au	Pyrrhotite			5.47	UTA
P22-70	Sulphide & Au	Pyrrhotite			11.35	UTA
HM1-14	Disseminated in meta-tuff	Pyrrhotite			-0.27	UTA
R12-51	Calcic skarn	Pyrrhotite			4.25	UTA
R7-125	Quartz vein related	Pyrrhotite			3.32	UTA
R7-174a	Quartz vein related	Pyrrhotite			3.17	UTA
R7-286	Quartz vein related	Pyrrhotite			2.79	UTA
H5-280	Quartz vein related	Pyrrhotite			3.65	UTA
R11-53	From sediment	Pyrrhotite			4.03	UTA
P13-39	From porphyry	Pyrrhotite			4.89	UTA
TS-29	From sediment	Pyrrhotite			-0.28	UTA
R7-191	Cataclastite (duplicate)	Pyrrhotite			3.08	UTA
TS-12	Quartz-carbonate vein	Calcite	13.43	-5.08		UCT
TS-13	Quartz-carbonate vein	Calcite	12.11	-5.51		UCT
TS-13	Quartz-carbonate vein	Dolomite	9.51	-5.20		UCT
H5-280e	Quartz-carbonate vein	Calcite	13.74	-6.31		UCT
H5-280e	Quartz-carbonate vein	Dolomite	11.11	-6.04		UCT
H5-237	Quartz-carbonate vein	Calcite	12.29	-7.09		UCT
R11-232	Quartz-carbonate vein	Calcite	14.12	-7.47		UCT
H5-128	Quartz-carbonate vein	Calcite	14.40	-7.80		UCT

Table 3.17: $\delta^{18}O$, $\delta^{13}C$ and $\delta^{34}S$ values for carbonates and sulphides from Haveri.

3.3.2 Whole-rock $\delta^{18}O$

The whole-rock $\delta^{18}O$ value is a function of the $\delta^{18}O$ values of the component minerals and their relative proportions within the rock. The $\delta^{18}O$ value of the rocks is controlled by the initial isotopic composition of the rock (which commonly has a restricted range), and any subsequent interaction with natural fluids. The mantle has a $\delta^{18}O$ value of 5.7 ± 0.3 (Taylor, 1980). Most granites, metamorphic rocks and sediments are enriched in $\delta^{18}O$ relative to the mantle value, whereas seawater and meteoric waters are depleted, thus forming complementary $\delta^{18}O$ reservoirs.

A suite of representative samples from the principal lithologies were analysed for their whole-rock $\delta^{18}O$ values (Table 3.18). This suite includes samples that are relatively unaltered, as well as samples of progressively altered rock. In all cases the alteration described is amphibolitisation.

Sample	Description	Mineral etc.	$\delta^{18}\text{O}$	Laboratory
TS-2	Granodiorite (NW of Haveri)	Whole Rock	7.94	UCT
TS-18	Granodiorite (R13)	Whole Rock	8.82	UCT
TS-15	Porphyry (HDD6)	Whole Rock	10.26	UCT
TS-9	Porphyry (west of pit)	Whole Rock	10.15	UCT
P8-24	Porphyry (skarned)	Whole Rock	9.67	UCT
P13-97	Porphyry (dark matrix)	Whole Rock	8.44	UCT
P13-98	Uralite porphyry	Whole Rock	7.83	UCT
P7-105	Vesiculated amphibolite	Whole Rock	7.91	UCT
TS-5	Pillowed amphibolite	Whole Rock	7.56	UCT
P5-79	Intense amphibolitisation	Whole rock	7.00	UCT
R11-137	Biotite Hornfels	Whole Rock	8.99	UCT
H5-240	Biotite Hornfels (skarned)	Whole Rock	7.34	UCT

Table 3.18: $\delta^{18}\text{O}_{\text{whole rock}}$ values for a suite of variably altered rocks from Haveri. The porphyries, amphibolites and metasedimentary rocks are arranged in order of increasing alteration.

I-type granitoids typically have $\delta^{18}\text{O}$ values between +6 ‰ and +8 ‰, whereas magmatic water from S-type granitoids range from +8 ‰ and + 12 ‰ (Ohmoto, 1986). The two samples of granodiorite at Haveri have $\delta^{18}\text{O}$ values that are slightly high for I-type granitoids, with one sample being more similar to an S-type granitoid. This deviation from the normal value may be due to interaction with metamorphic fluids enriched in $\delta^{18}\text{O}$. Such a metamorphic fluid can be generated from sediments or even volcanic rocks, such as the Haveri amphibolites, which have interacted with seawater at temperatures below 200 °C, and subsequently been subjected to higher temperatures and pressures (Ohmoto, 1986). Alternatively, enrichment in ^{18}O can also be caused by hydrothermal alteration at very low temperatures. (Taylor, 1997).

The porphyries, metavolcanic rocks and the metasedimentary rocks all display $\delta^{18}\text{O}$ depletions with increasing hydrothermal alteration. This is a commonly observed characteristic in many hydrothermal ore deposits which involve meteoric-hydrothermal convection systems (e.g. Ohmoto, 1986).

3.3.3 Oxygen isotope thermometry

Isotopic geothermometers offer a number of advantages over cation-exchange thermometers, including applicability to common mineral assemblages, and the absence of a measurable pressure effect (Valley, 1986). In addition, there are potentially a large number of thermometers available in a single rock, a situation which rarely arises from cation-exchange thermometers. This allows the assumption of isotopic equilibrium to be tested (Rollinson, 1993).

Isotopic disequilibrium can arise by a number of means. In slowly cooled rocks, individual minerals cease to exchange oxygen as they pass through their own blocking temperatures, and it is only the last mineral-pair to exchange oxygen that will be in equilibrium (Giletti, 1986). Consequently, isotopic disequilibrium may be expected to be the norm in slowly cooled rocks, and thus rather than recording peak temperatures, the various mineral-pairs are actually recording a retrograde event, as a function of the oxygen diffusivities of the various minerals. In studies of hydrothermal ore deposits it is more common for isotopic disequilibrium to be caused by isotopic exchange with infiltrating fluids. The nature and the extent of the disequilibrium resulting from differences in fluid composition, temperature and fluid/rock ratios.

Oxygen isotope thermometry is based upon the temperature dependence of isotopic exchange of ^{18}O between equilibrium mineral pairs. This requires a calibration for the temperature dependent fractionation factors for the minerals, according to the general equation:

$$1000\ln \alpha_{\text{mineral1-mineral2}} = A(10^6/T^2) + B$$

where T is in kelvin and A and B are constants, either theoretically, empirically or experimentally determined. There are a large number of different calibrations of oxygen isotope exchange reactions, some of which give conflicting results. However, the more recent polynomial calibrations (Clayton and Keiffer, 1991; Zheng, 1991, 1993a,b; Zheng and Simon, 1991) combining experimental and theoretical studies are considered by Rollinson (1993), to be more reliable than the earlier empirically and experimentally determined calibrations (O'Neil and Taylor, 1967; Bottinga and Javoy, 1973, 1975; Matthews et al., 1983; Chiba et al., 1989).

3.3.3.1 Granodiorite and quartz-feldspar porphyry

Mineral separates of quartz, biotite and plagioclase were extracted from samples of the granodiorite and the quartz-feldspar porphyry. These were analysed for $\delta^{18}\text{O}$ (and δD of biotite) at Queens University, Ontario. Table 3.16 shows the results of the analyses.

A number of potential geothermometers exist for these minerals, as well as numerous calibrations for the fractionation factors. The results of the thermometers with the various calibrations are presented in Table 3.19. It is apparent that there is a considerable variation in calculated temperatures both for the different calibrations of the same thermometers as well as between different thermometers, for both the porphyry (TS-15) and the granodiorite (TS-18). These variations indicate isotopic disequilibrium. As mentioned earlier, the potential causes for isotopic disequilibrium are either slow cooling through mineral blocking temperatures and/or isotopic re-equilibration with infiltrating fluids.

Thermometer Calibration	1000ln α	TS-15 QFP	TS-18 GRD
<u>Quartz-Albite:</u>			
Chiba et al. (1989)	$0.94 (10^6/T^2)$	434 °C ⁺⁸⁹ / ₋₆₄	343 °C ⁺⁵⁶ / ₋₄₄
Bottinga & Javoy (1975)	$0.97 (10^6/T^2)$	445 °C ⁺⁹⁰ / ₋₆₅	353 °C ⁺⁵⁷ / ₋₄₅
Matthews et al. (1983)	$0.5 (10^6/T^2)$	242 °C ⁺⁶⁵ / ₋₄₇	176 °C ⁺⁴¹ / ₋₃₂
Zheng (1993a, b)	$0.15 (10^6/T^2) + 1.39 (10^3/T) - 0.57$	387 °C ⁺¹¹¹ / ₋₈₁	273 °C ⁺⁷⁰ / ₋₅₄
<u>Plagioclase (An₃₀)-H₂O:</u>			
i) Bottinga & Javoy (1973)	$(3.13 - 1.04\text{An}) 10^6/T^2 - 3.70$		
ii) O'Neill & Taylor (1967)	$(2.91 - 0.76\text{An}) 10^6/T^2 - 3.41 - 0.14\text{An}$		
<u>Quartz-H₂O:</u>			
iii) Zheng (1993a)	$4.48 (10^6/T^2) - 4.77 (10^3/T) + 1.71$		
<u>Quartz-Plagioclase (An₃₀):</u>			
iii) – i)	$1.662 (10^6/T^2) - 4.77 (10^3/T) + 5.41$	424 °C ⁺⁰ / ₋₁₅₇	848 °C ⁺²⁹⁸ / ₋₂₈₇
iii) – ii)	$1.798 (10^6/T^2) - 4.77 (10^3/T) + 5.162$	481 °C ⁺⁰ / ₋₀	270 °C ⁺¹⁰⁶ / ₋₄₉
<u>Biotite-H₂O:</u>			
iv) Zheng (1993b)	$3.84 (10^6/T^2) - 8.76 (10^3/T) + 2.46$		
<u>Quartz-Biotite:</u>			
Bottinga & Javoy (1975)	$3.69 (10^6/T^2)$	472 °C ⁺²³ / ₋₂₁	514 °C ⁺²⁸ / ₋₂₅
iii) – iv)	$0.64 (10^6/T^2) + 3.99 (10^3/T) - 0.75$	396 °C ⁺³² / ₋₂₉	453 °C ⁺³⁹ / ₋₃₄
<u>Biotite-Plagioclase (An₃₀):</u>			
iv) – i)	$1.022 (10^6/T^2) - 8.76 (10^3/T) + 6.16$	387 °C ⁺³² / ₋₃₀	498 °C ⁺⁴⁰ / ₋₃₇
iv) – ii)	$1.158 (10^6/T^2) - 8.76 (10^3/T) + 5.912$	381 °C ⁺³⁴ / ₋₃₂	500 °C ⁺⁴³ / ₋₄₀

Table 3.19: Oxygen isotope temperatures calculated from $\delta^{18}\text{O}$ values of minerals in the granodiorite (TS-18) and the quartz-feldspar porphyry (TS-15). Temperatures have been calculated according to a number of calibrations of the fractionation factors.

3.3.3.2 Unaltered Amphibolite

Oxygen isotope analyses on mineral separates of hornblende and plagioclase from an unaltered, vesiculated amphibolite have been used for isotope thermometry (Table 3.20). Three separate calibrations were used for the calculations, two of which were calculated using an estimated plagioclase composition of An_{30} (Section 3.2.2.4). The three different calibrations produce temperatures which are in reasonably close agreement, with an average temperature of approximately 508 °C (standard deviation = 11 °C). With only two co-existing minerals analysed for oxygen isotopes, it is not possible to confirm categorically whether the minerals are in equilibrium or not. However, the calculated temperature is considerably lower than that calculated by the amphibole-plagioclase cation exchange thermometer (Section 4.1.2.1) and the peak metamorphic temperature determined by the pelitic mineral assemblage (Section 4.1.1). This does suggest that the minerals may not be in isotopic equilibrium, and that isotopic re-equilibration may therefore have occurred.

Thermometer	Calibration	$1000\ln\alpha$	P7-105
<u>Quartz-albite:</u>			
i)	Bottinga & Javoy (1975)	$0.97 (10^6/T^2)$	
<u>Quartz-Hornblende:</u>			
ii)	Bottinga & Javoy (1975)	$3.15 (10^6/T^2) - 0.3$	
<u>Plagioclase (An_{30})-H_2O:</u>			
iii)	Bottinga & Javoy (1973)	$(3.13 - 1.04An) 10^6/T^2 - 3.70$	
iv)	O'Neill & Taylor (1967)	$(2.91 - 0.76An) 10^6/T^2 - 3.41 - 0.14An$	
<u>Hornblende-H_2O:</u>			
v)	Zheng (1993b)	$3.89 (10^6/T^2) - 8.56 (10^3/T) + 2.43$	
<u>Feldspar-Hornblende:</u>			
	ii) – i)	$2.18 (10^6/T^2) - 0.3$	$519\text{ °C}^{+50} / -_{42}$
	iii) – v) at An_{30}	$-1.072 (10^6/T^2) + 8.56 (10^3/T) - 6.13$	$497\text{ °C}^{+42} / -_{39}$
	iv) – v) at An_{30}	$-1.208 (10^6/T^2) + 8.56 (10^3/T) - 5.798$	$509\text{ °C}^{+46} / -_{43}$

Table 3.20: Oxygen isotope temperatures calculated from $\delta^{18}O$ values of plagioclase and hornblende in an unaltered vesiculated amphibolite. Temperatures have been calculated according to a number of calibrations of the fractionation factors.

Isotopic re-equilibration is to be expected at Haveri, as $\delta^{18}O$ depletions are common not only in hydrothermal systems (as mentioned earlier) but also within contact metamorphic aureoles. Forester and Taylor (1977) have shown that significant $\delta^{18}O$ depletions can occur in basalts over 3 km from the intrusion contact, and are still detectable up to 5 km from the intrusion contact, although Valley (1986) states that this distance can be much less. The location of Haveri, less than 1 km from the Central Finland Granitoid Complex, means that it is impossible to totally avoid such isotopic re-equilibration,

and that the degree of $\delta^{18}\text{O}$ depletion will be a function of the degree of metamorphism (i.e. proximity to intrusion) and the extent of hydrothermal alteration.

3.3.3.3 Amphibolitisation

Hydrothermal alteration at Haveri is dominated by a pervasive amphibolitisation. In the areas of most intense alteration, the rock consists almost entirely of hornblende with or without sulphides, magnetite and gold. In terms of oxygen isotope thermometry, the only mineral-pairs with oxygen are hornblende and magnetite. Mineral separates of hornblende and magnetite were taken from a sample of intense amphibolitisation (P7-82), with large patches of massive magnetite. The results of the oxygen isotope thermometry are displayed in Table 3.21.

Thermometer	Calibration	$1000\ln\alpha$	P7-82
<u>Magnetite-H_2O:</u>			
i)	Zheng (1991)	$3.02 (10^6/\text{T}^2) - 12.0 (10^3/\text{T}) + 3.31$	
<u>Hornblende-H_2O:</u>			
ii)	Zheng (1993b)	$3.89 (10^6/\text{T}^2) - 8.56 (10^3/\text{T}) + 2.43$	
<u>Quartz-Hornblende:</u>			
iii)	Bottinga & Javoy (1975)	$3.15 (10^6/\text{T}^2) - 0.3$	
<u>Quartz-Magnetite:</u>			
iv)	Chiba et al. (1989)	$6.29 (10^6/\text{T}^2)$	
v)	Bottinga & Javoy (1975)	$5.57 (10^6/\text{T}^2)$	
vi)	Matthews et al. (1983)	$6.11 (10^6/\text{T}^2)$	
vii)	Clayton & Keiffer (1991)	$0.012 (10^{18}/\text{T}^6) - 0.322 (10^{12}/\text{T}^4) + 6.442 (10^6/\text{T}^2)$	
iiix)	Zheng (1991)	$1.22 (10^6/\text{T}^2) + 8.22 (10^3/\text{T}) - 4.35$	
<u>Hornblende-Magnetite:</u>			
ii) – i)		$0.87 (10^6/\text{T}^2) + 3.44 (10^3/\text{T}) - 0.88$	$538\text{ }^\circ\text{C}^{+50} /_{-44}$
iv) – iii)		$3.14 (10^6/\text{T}^2) + 0.3$	$573\text{ }^\circ\text{C}^{+41} /_{-36}$
v) – iii)		$2.42 (10^6/\text{T}^2) + 0.3$	$470\text{ }^\circ\text{C}^{+36} /_{-32}$
vi) – iii)		$2.96 (10^6/\text{T}^2) + 0.3$	$549\text{ }^\circ\text{C}^{+40} /_{-35}$
vii) – iii)		$0.012 (10^{18}/\text{T}^6) - 0.322 (10^{12}/\text{T}^4) + 3.292 (10^6/\text{T}^2) + 0.3$	$529\text{ }^\circ\text{C}^{+47} /_{-40}$
iiix) – iii)		$-1.93 (10^6/\text{T}^2) + 8.22 (10^3/\text{T}) - 4.05$	$221\text{ }^\circ\text{C}^{+107} /_{-25}$

Table 3.21: Oxygen isotope temperatures calculated from $\delta^{18}\text{O}$ values of coexisting hornblende and magnetite in intensely amphibolitised volcanic. Temperatures have been calculated according to a number of calibrations of the fractionation factors.

With the exception of one anomalously low calculated temperature, the calibrations produce reasonably consistent temperatures with an average of 532 °C (standard deviation = 38 °C). This is slightly lower than the temperatures calculated for peak metamorphism by cation-exchange thermometry and phase equilibria (Section 4.1).

3.3.3.4 Quartz veins

Amphibolitisation has been shown to be essentially contemporaneous with many of the quartz veins observed at Haveri. Consequently, the geothermometry of the quartz veins should be expected to have similar temperatures to the amphibolitisation (and hence also peak metamorphism). The mineralogy of the quartz veins is variable, being composed primarily of quartz with or without hornblende, plagioclase, carbonate, magnetite and the sulphides pyrrhotite and chalcopyrite. In terms of oxygen isotope geothermometry, two veins with co-existing mineral pairs were analysed. Sample R8-168a from an altered and gold-rich portion of metavolcanic rock 300 metres north of the Haveri mine provided separates of quartz, plagioclase and magnetite. This assemblage permits three oxygen isotope thermometers and numerous calibrations to be used (Table 3.22).

The calculated temperatures (mean = 611 °C; standard deviation = 71 °C) for this quartz vein are, in contrast to the unaltered amphibolite, in good agreement with temperatures determined for peak metamorphism by the amphibole-plagioclase cation exchange thermometer (Section 4.1.2.1) and by the petrogenetic grid for pelitic mineral assemblages (Section 4.1.1). If the assumption that this quartz vein is contemporaneous with the amphibolitisation is correct, then it is possible to state that both the quartz veins and amphibolitisation are also contemporaneous with the peak of metamorphism.

Thermometer	Calibration	1000ln α	R8-168a
<u>Quartz-Magnetite:</u>			
	Chiba et al. (1989)	$6.29 (10^6/T^2)$	$646\text{ }^{\circ}\text{C}^{+26} / -_{24}$
	Bottinga & Javoy (1975)	$5.57 (10^6/T^2)$	$591\text{ }^{\circ}\text{C}^{+24} / -_{22}$
	Matthews et al. (1983)	$6.11 (10^6/T^2)$	$632\text{ }^{\circ}\text{C}^{+25} / -_{23}$
	Clayton & Keiffer (1991)	$0.012 (10^{18}/T^6) - 0.322 (10^{12}/T^4) + 6.442 (10^6/T^2)$	$629\text{ }^{\circ}\text{C}^{+27} / -_{25}$
i)	Zheng (1991)	$1.22 (10^6/T^2) + 8.22 (10^3/T) - 4.35$	$549\text{ }^{\circ}\text{C}^{+25} / -_{23}$
<u>Quartz-Albite:</u>			
	Chiba et al. (1989)	$0.94 (10^6/T^2)$	$656\text{ }^{\circ}\text{C}^{+236} / -_{133}$
	Bottinga & Javoy (1975)	$0.97 (10^6/T^2)$	$670\text{ }^{\circ}\text{C}^{+239} / -_{136}$
	Matthews et al. (1983)	$0.5 (10^6/T^2)$	$404\text{ }^{\circ}\text{C}^{+172} / -_{97}$
ii)	Zheng (1993a)	$0.15 (10^6/T^2) + 1.39 (10^3/T) - 0.57$	$661\text{ }^{\circ}\text{C}^{+265} / -_{164}$
<u>Albite-H₂O:</u>			
iii)	Zheng (1993a)	$4.33 (10^6/T^2) - 6.15 (10^3/T) + 1.98$	
<u>Plagioclase (An₃₀)-H₂O:</u>			
iv)	Bottinga & Javoy (1973)	$(3.13 - 1.04\text{An}) 10^6/T^2 - 3.70$	
v)	O'Neill & Taylor (1967)	$(2.91 - 0.76\text{An}) 10^6/T^2 - 3.41 - 0.14\text{An}$	
<u>Magnetite-H₂O:</u>			
vi)	Zheng (1991)	$3.02 (10^6/T^2) - 12.0 (10^3/T) + 3.31$	
<u>Feldspar-Magnetite:</u>			
	Chiba et al. (1989)	$5.35 (10^6/T^2)$	$644\text{ }^{\circ}\text{C}^{+30} / -_{27}$
	Matthews et al. (1983)	$5.61 (10^6/T^2)$	$666\text{ }^{\circ}\text{C}^{+31} / -_{28}$
	i) – ii)	$1.07 (10^6/T^2) + 6.83 (10^3/T) - 3.78$	$531\text{ }^{\circ}\text{C}^{+28} / -_{26}$
	iii) – vi)	$1.31 (10^6/T^2) + 5.85 (10^3/T) - 1.33$	$668\text{ }^{\circ}\text{C}^{+43} / -_{39}$
	iv) – vi)	$- 0.202 (10^6/T^2) + 12.0 (10^3/T) - 7.01$	$607\text{ }^{\circ}\text{C}^{+28} / -_{26}$
	v) – vi)	$- 0.338 (10^6/T^2) + 12.0 (10^3/T) - 6.762$	$612\text{ }^{\circ}\text{C}^{+29} / -_{27}$

Table 3.22: Oxygen isotope temperatures calculated from $\delta^{18}\text{O}$ values of quartz, magnetite and plagioclase in a vein associated with amphibolitisation. Temperatures have been calculated according to a number of calibrations of the fractionation factors.

A second quartz vein (sample R7-331) located directly below the old Haveri open pit, has provided vein quartz and a hornblende separate from the wallrock adjacent to the vein. Two calibrations of the quartz-hornblende oxygen isotope thermometer (Table 3.23) have yielded an average temperature of 536 °C (standard deviation = 25 °C). This is somewhat lower than the temperatures calculated for peak metamorphism. There are a number of potential explanations for this. It is possible that the minerals are not in isotopic equilibrium, either because the quartz vein and the adjacent hornblende never attained equilibrium, or that there has been isotopic exchange with an externally derived fluid, resulting in either partial or total isotopic reequilibration at lower temperatures. Another possibility is that the samples are simply reflecting a temperature gradient away from the intrusions, and that therefore these are true temperatures. However, the calculated temperatures do represent a minimum temperature for the quartz vein. Therefore the emplacement of the quartz vein and the adjacent amphibole alteration are likely to have been either contemporaneous with metamorphism, or at least soon thereafter, before temperatures decreased below 536 °C.

Thermometer	Calibration	$1000\ln\alpha$	R7-331
<u>Quartz-H₂O:</u>			
i)	Zheng (1993a)	$4.48 (10^6/T^2) - 4.77 (10^3/T) + 1.71$	
<u>Hornblende-H₂O:</u>			
ii)	Zheng (1993b)	$3.89 (10^6/T^2) - 8.56 (10^3/T) + 2.43$	
<u>Quartz-Hornblende:</u>			
	i) – ii)	$0.59 (10^6/T^2) + 3.79 (10^3/T) - 0.72$	553 °C ⁺⁵⁶ / ₋₄₈
	Bottinga & Javoy (1975)	$3.15 (10^6/T^2) - 0.3$	518 °C ⁺³³ / ₋₂₉

Table 3.23: Oxygen isotope temperatures calculated from $\delta^{18}\text{O}$ values of quartz in a quartz vein, and hornblende on the selvedge of the vein. The vein appears to be associated with amphibolitisation. Temperatures have been calculated according to a number of calibrations of the fractionation factors.

3.3.4 Isotopic composition of water in hydrothermal fluids

The origin of ore-forming fluids (e.g. magmatic, meteoric, etc.) and the hydrologic nature of the plumbing systems (e.g. geometry, water/rock ratios) can be estimated by comparing the H and O isotopic compositions of ore-forming fluids with those of other natural waters, such as sea-, meteoric-, magmatic-, connate-, and metamorphic waters; and comparing the H and O isotopic compositions of minerals and rocks surrounding ore deposits with those of “normal” rocks (Ohmoto, 1986).

Direct and indirect methods have been used to estimate the H and O isotopic composition of ore-forming fluids. The direct method consists of measuring $\delta^{18}\text{O}$ and δD from “fossil” water preserved in fluid inclusions in host minerals. This is done by crushing a large amount of a mineral in a vacuum, and

measuring the extracted fluid. However, the extracted fluid may be a mixture of fluid inclusions of different generations, and thus distinguishing the contribution of the different generations of fluids may be problematic (Ohmoto, 1986). Consequently, the indirect method is most commonly used, and is based on the isotopic composition of minerals which were in equilibrium with the fluid (Rollinson, 1993). If it can be assumed that there was a close approach to isotopic equilibrium between the mineral and the hydrothermal fluid, then laboratory calculations of equilibria between rock-forming minerals and water can be used to calculate the isotopic composition of the hydrothermal solution. The calculation requires a knowledge of the temperature of equilibration, which needs to be measured independently by using such techniques as fluid inclusion thermometry, or phase equilibria (Rollinson, 1993).

<u>Biotite-Water:</u>									
Oxygen:		$1000 \ln \alpha = 3.84 (10^6/T^2) - 8.76 (10^3/T) + 2.46$				Zheng, 1993b			
Hydrogen:		$1000 \ln \alpha = -21.3 (10^6/T^2) - 2.80$				Suzuoki & Epstein, 1976			
Sample:			<u>TS-15</u>		<u>TS-18</u>		<u>P6-43</u>		
Description:			Biotite in QFP		Biotite in GRD		Biotite in pelite		
			$\delta^{18}\text{O}_{\text{biotite}}$	$\delta\text{D}_{\text{biotite}}$	$\delta^{18}\text{O}_{\text{biotite}}$	$\delta\text{D}_{\text{biotite}}$	$\delta^{18}\text{O}_{\text{biotite}}$	$\delta\text{D}_{\text{biotite}}$	
1000Lna			4.4	-77	4.3	-86	5.8	-74	
T °C	$\delta^{18}\text{O}$	δD	$\delta^{18}\text{O}_{\text{water}}$	$\delta\text{D}_{\text{water}}$	$\delta^{18}\text{O}_{\text{water}}$	$\delta\text{D}_{\text{water}}$	$\delta^{18}\text{O}_{\text{water}}$	$\delta\text{D}_{\text{water}}$	
550	-2.51	-34.24	6.9	-42.8	6.8	-51.8	8.3	-39.8	
600	-2.54	-30.74	6.9	-46.3	6.8	-55.3	8.3	-43.3	
650	-2.52	-27.79	6.9	-49.2	6.8	-58.2	8.3	-46.2	
800	-2.37	-21.30			6.7	-64.7			
850	-2.30	-19.69			6.6	-66.3			
<u>Hornblende-Water:</u>									
Oxygen:		$1000 \ln \alpha = 3.89 (10^6/T^2) - 8.56 (10^3/T) + 2.43$				Zheng, 1993b			
Hydrogen:		$1000 \ln \alpha = -23.9 (10^6/T^2) + 7.9$				Suzuoki & Epstein, 1976			
Sample:			<u>P7-105</u>		<u>R7-331</u>		<u>P7-82</u>		
Description:			Hb in unaltered amphibolite		Hb selvage on quartz vein		Hb in intense amphibolite alteration		
			$\delta^{18}\text{O}_{\text{hb}}$	$\delta\text{D}_{\text{hb}}$	$\delta^{18}\text{O}_{\text{hb}}$	$\delta\text{D}_{\text{hb}}$	$\delta^{18}\text{O}_{\text{hb}}$	$\delta\text{D}_{\text{hb}}$	
1000Lna			6.8	-91	5.7	-90	6.5	-105	
T °C	$\delta^{18}\text{O}$	δD	$\delta^{18}\text{O}_{\text{water}}$	$\delta\text{D}_{\text{water}}$	$\delta^{18}\text{O}_{\text{water}}$	$\delta\text{D}_{\text{water}}$	$\delta^{18}\text{O}_{\text{water}}$	$\delta\text{D}_{\text{water}}$	
550	-2.23	-27.37	9.0	-63.6	7.9	-62.6	8.7	-77.6	
600	-2.27	-23.45	9.1	-67.6	8.0	-66.6	8.8	-81.6	
650	-2.28	-20.14	9.1	-70.9	8.0	-69.9	8.8	-84.9	

Table 3.24: Calculated $\delta^{18}\text{O}$ and δD values of fluids in equilibrium with biotite and hornblende, at various temperatures. The fractionation factors for biotite- H_2O and hornblende- H_2O are those of Zheng (1993a) and Suzuoki and Epstein (1976), for ^{18}O and D respectively.

Table 3.24 shows the calculated equilibrium fluid compositions for biotites and hornblende from a range of rock types. The biotite-H₂O and hornblende-H₂O fractionation factors used are those of Zheng (1993a) and Suzuoki and Epstein (1976) for oxygen and hydrogen respectively. Equilibrium fluids have been calculated at temperatures of 550 °C, 600 °C and 650 °C, as well as at 800 °C and 850 °C for the biotite from a granodiorite. The results at 600 °C are also plotted on Figure 3.40, along with the fields of δD vs $\delta^{18}O$ values for different water types. The biotite from the granodiorite is also plotted for a temperature of 850 °C, and as expected, plots within the magmatic water field of Taylor (1974). The biotite from porphyry has not been calculated at the higher temperatures as there is sufficient field (and petrographic) data that suggest it was intruded prior to peak metamorphism and hydrothermal alteration.

It is clear from Figure 3.40 that the biotite from the granodiorite (calculated at 850 °C) plots in the field for magmatic water (as expected). The biotites from the pelite and the porphyry have isotopic compositions falling within the field of overlap between magmatic water and metamorphic water. On the basis of field and petrographic evidence, it is likely that these represent metamorphic isotopic signatures, and that the biotites have equilibrated with metamorphic fluids.

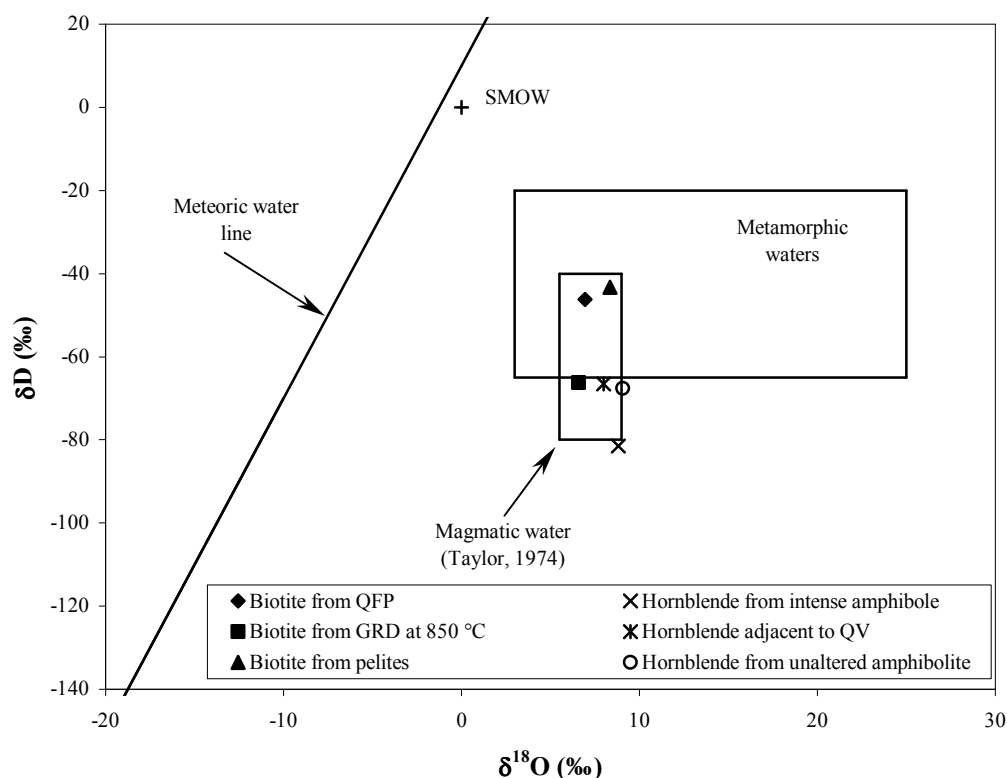


Figure 3.40: Plot of calculated δD vs $\delta^{18}O$ values of water in equilibrium with biotite and hornblende from Haveri at 600 °C. The field for magmatic water is taken from Taylor (1974). The meteoric water line is from Epstein et al. (1965) and Epstein (1970). The metamorphic field combines the values of Taylor (1974) and Sheppard (1981). After Rollinson, (1993).

The hornblende from the unaltered, vesiculated, amphibolite plots on the edge of the magmatic water field, and outside of the metamorphic water field. However, it is unlikely that a metamorphic hornblende will have equilibrated with water from a magmatic source. Rather, it is likely that the isotopic composition of the hornblende displays the effect of some minor re-equilibration, resulting in a depletion in δD , from an originally metamorphic composition. Indeed, this is supported by the even greater degree of D depletion displayed by the hornblende from the intensely altered amphibole. It is therefore considered likely that even the “unaltered” amphibolite is displaying a minor degree of hydrothermal alteration.

The relative effects of hydrothermal alteration on the isotopic composition of the host rocks is controlled by the isotopic composition of the fluids (and hence source) combined with the fluid / rock ratio. Taylor (1974) recognised six types of naturally occurring waters, that can be used as reference compositions to determine the source (or sources) of the fluid. These include seawater (SMOW), meteoric water, geothermal water, formation water, metamorphic water and magmatic water (Rollinson, 1993). Hydrogen is a minor component of most rocks and so, except when there is a very low fluid/rock ratio, the hydrogen isotope composition of rocks and minerals is very sensitive to the isotopic composition of interacting fluids. Oxygen, on the other hand, is a major component in most rocks and minerals, and so is less sensitive to the oxygen isotope ratio of interacting fluids, except at high fluid/rock ratios (Rollinson, 1993). Consequently, the observed decrease in δD and the relative consistency in $\delta^{18}O$ values in hornblende with increasing alteration at Haveri, can be explained as an interaction with a hydrothermal fluid at moderate fluid/rock ratios. In addition, due to the shift in δD away from the metamorphic fluid field with increasing alteration, it is unlikely that the fluids are derived from metamorphic waters, and are most likely to be caused by waters with an isotopic character similar to either meteoric or magmatic-meteoric waters. In view of the proximity to the Central Finland Granitoid Complex, and isotopic compositions similar to Taylor’s magmatic water, a magmatic dominated magmatic-meteoric fluid source is considered most reasonable for the hydrothermal fluids responsible for amphibolitisation.

The isotopic composition of a fluid in equilibrium with the hornblende adjacent to the quartz vein, shows a depletion in ^{18}O relative to the “unaltered” sample. This can perhaps be explained by interaction with a fluid dominated by a magmatic water component with a lower $\delta^{18}O$ value and a similar δD value. This would produce the ^{18}O depletion with only a negligible change in D. The implications of this on the origin of quartz veins are discussed in Section 3.3.6.

3.3.5 Fluid / rock ratios in hydrothermal alteration

One of the major uses for stable isotope geochemistry in hydrothermal or ore deposit systems is to constrain the extent of fluid/rock interaction. Fluid/rock interaction can vary between two extremes. When the water/rock ratio is small and the $\delta^{18}O$ in the rock dominates the system it is the fluid

composition which is changed, as happens in geothermal systems. On the other hand, when the water/rock ratio is large and the $\delta^{18}\text{O}$ of the water dominates, the $\delta^{18}\text{O}$ of the rock is modified (Rollinson, 1993).

Oxygen isotopes are able to provide information regarding the water/rock ratios involved in hydrothermal alteration. Taylor (1974, 1977) derived mass balance equations from which the water/rock ratio may be calculated from $\delta^{18}\text{O}$ values. In a closed system, from which none of the water is lost, the water/rock ratio, integrated over the lifetime of the hydrothermal system, is:

$$W / R_{\text{closed}} = \frac{\delta^{18}\text{O}_{\text{rock}}^{\text{final}} - \delta^{18}\text{O}_{\text{rock}}^{\text{initial}}}{\delta^{18}\text{O}_{\text{fluid}}^{\text{initial}} - \delta^{18}\text{O}_{\text{fluid}}^{\text{final}}}$$

This is the effective water/rock ratio which can differ from the actual water/rock ratio depending upon the efficiency of the exchange reaction. The initial value for the rock is obtained from ‘normal’ values for the particular rock type, or from an unaltered sample of the rock suite being analysed. The final value for the rock is the measured value. The initial value for the fluid is assumed. The composition of the final fluid can be calculated from the mineralogy of the altered rock, and an independent estimate of the temperature (Rollinson, 1993).

The equation for an open system water/rock ratio, through which water makes only a single pass, is given by Taylor (1974, 1977) as:

$$W / R_{\text{open}} = \ln(W / R_{\text{closed}} + 1)$$

It is likely that the behaviour of any given hydrothermal system will be somewhere between the two extremes (Rollinson, 1993).

For the amphibolitisation at Haveri, whole rock analyses of the unaltered and intensely altered amphibolites provide the $\delta^{18}\text{O}$ values for the initial ($\delta^{18}\text{O} = 7.91 \text{ ‰}$) and final ($\delta^{18}\text{O} = 7.00 \text{ ‰}$) rocks respectively. In addition the $\delta^{18}\text{O}$ value for the final composition of the fluid is $\delta^{18}\text{O} = 8.8 \text{ ‰}$. This has been calculated (Table 3.24) from water in equilibrium with alteration hornblende at 600 °C. It is also necessary to make an assumption about the initial composition of the fluid. As mentioned previously it is likely that the fluid is a combination of magmatic water and meteoric water. The composition of magmatic water from I-type granitoids is between +7.0 and +9.0 ‰. The composition of meteoric water can be calculated from the D/H ratio of the alteration assemblage and the meteoric water equation: $\delta\text{D} = 8\delta^{18}\text{O} + 10$ (Rollinson, 1993). From Table 3.24, at 600 °C the δD of the fluid in equilibrium with alteration amphibole is $\delta\text{D} = -81.6 \text{ ‰}$. Consequently, using the meteoric water equation, the $\delta^{18}\text{O}$ of the meteoric water would be -11.45 ‰ . In calculating the fluid/rock ratios, the magmatic water and the meteoric water represent the end members of initial composition of the fluid,

with the actual initial isotopic composition of the fluid having a value between the two depending on the proportions of water types involved.

	<i>Amphibolite</i>	
$\delta^{18}\text{O}_{\text{final rock}}$	7.00 ‰	
$\delta^{18}\text{O}_{\text{initial rock}}$	7.91 ‰	
$\delta^{18}\text{O}_{\text{final fluid}}$	8.8 ‰	
$\delta^{18}\text{O}_{\text{initial fluid}}$	Magmatic	Meteoric
	8.0 ‰	-11.45 ‰
Fluid/Rock _{closed}	1.14	0.045
Fluid/Rock _{open}	0.76	0.044

Table 3.25: *Fluid/rock ratios for the altered amphibolite, calculated for both magmatic and meteoric waters, using the closed and open systems of Taylor (1974, 1977).*

Table 3.25 shows the calculated fluid/rock ratios for the altered amphibolites. There is clearly a wide discrepancy between the end-member results for magmatic and meteoric waters. Typical fluid/rock values averaged over an entire hydrothermal system are seldom greater than 1 (Taylor, 1997). Furthermore, if one defines a plausible length time scale and flow vector, then these fluid/rock ratios of about unity translate into time-integrated H_2O fluxes of hundreds to thousands of kg/cm^2 . These enormous amounts of water are required by simple material balance, and they explain the pervasive “soaking” that typically occurs in such areas of hydrothermal alteration, where the entire bulk volume of rock in an outcrop is depleted in ^{18}O , and not just the rocks adjacent to veins or large fractures (Taylor, 1974). Although, only limited isotopic analyses have been conducted at Haveri, the intense amphibolitisation is indeed pervasive, and widely distributed. Consequently, it is considered likely that the fluid/rock ratios are closer to the magmatic values (as calculated above) rather than the meteoric values, and that the source of the hydrothermal fluids responsible for amphibolitisation is predominately magmatic with only a minor meteoric component.

Buick (1998) has detailed a number of problems with water/rock ratios. The use of water/rock ratios assumes complete equilibration between the fluid and the rock. However, this implies infinite water/rock ratios and infinite time, which is clearly impossible. In addition, the amounts of water are expressed in terms of mass of water to unit mass of rock. These units do not however account for the fact that hydrogeological systems are systems of fluid transport, which use fluid fluxes (over a period of time). The water/rock ratio is also unable to account for incremental changes in the isotopic composition of both rocks and the fluids as they interact along the fluid flow paths. Consequently, a water/rock ratio is relevant only to the particular site of alteration being examined.

3.3.6 Origin of Quartz veins

Veins are a common feature of metamorphic rocks, and are commonly considered to represent fluid pathways for high time-integrated fluid fluxes (Buick, 1998). However, their origin is controversial. The idea that veins record large amounts of fluid flow comes from a consideration of the solubility of silica in aqueous fluids. Silica solubility decreases with decreasing temperature and pressure, so that quartz veins may be precipitated along P-T-t paths of decreasing temperature and pressure. The amount of fluid needed to deposit quartz in veins along long-distance flow paths of decreasing P and T is very large (Buick, 1998). Alternatively, rather than being precipitated during fluid flow, vein systems may be locally derived from adjacent rock units through diffusive mass transport (Gray et al., 1991; Yardley and Bottrell, 1993; Cartwright and Oliver, 1994; Buick, 1998). Stable isotopes can potentially discriminate between veining as a result of interaction with an internally derived fluid, and veins generated through open-system processes (Buick, 1998).

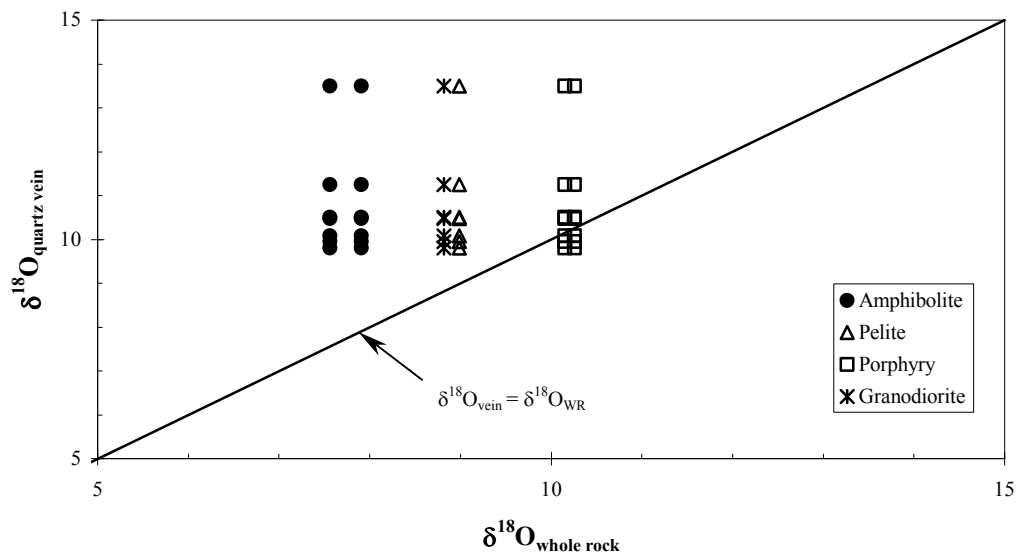


Figure 3.41: Isotopic composition of quartz veins plotted against the isotopic composition (whole rock) of the unaltered lithologies at Haveri, to discriminate the source of the quartz veins (after Buick, 1998).

Figure 3.41 is a plot of the $\delta^{18}\text{O}$ values of quartz veins plotted against the $\delta^{18}\text{O}_{\text{whole rock}}$ values of the lithologies encountered at Haveri. When veins are locally precipitated with material from adjacent interlayered wall rocks of different isotopic composition, the data should spread out along a 45° line where $\delta^{18}\text{O}_{\text{vein}} = \delta^{18}\text{O}_{\text{whole rock}}$. If the isotopic composition of the vein is buffered by a fluid of an external origin, the isotopic composition of the veins is independent of the isotopic composition of the wall rocks, and the data will plot away from the line. With the exception of one high value, the quartz vein data from Haveri plot in a field of approximately uniform isotopic composition away from the 45° line. The isotopic composition of the veins can therefore be interpreted as being buffered by an external

fluid reservoir of uniform isotopic composition (Gray et al., 1991). However, the isotopic compositions of the quartz veins do correlate well with the isotopic composition of the porphyry, and it is therefore possible that the porphyries alone are the source of the fluids for the quartz veins. The high $\delta^{18}\text{O}$ value of one vein compared to the others may indicate that the vein is from a separate generation, although it is possible that the high value may have been caused by later fluid flow and re-equilibration of the isotopic composition of the vein at lower temperature (Taylor, 1997).

3.3.7 Quartz – Carbonate veins

3.3.7.1 Origin of Quartz – Carbonate veins

Quartz $\pm$ carbonate veins are common at Haveri, and include the large (1 metre thick) parallel calcite-barite veins, the quartz-calcite veins of the calcic skarn and the numerous, (commonly) gold-rich, quartz veins and veinlets associated with amphibolitisation. Figure 3.42 shows the ranges of isotopic compositions for carbonates from quartz-carbonate veins at Haveri.

Decarbonation of a marine limestone during metamorphism causes ^{13}C and ^{18}O depletions in calcite relative to the original isotopic compositions of the limestone. The resultant CO_2 is correspondingly enriched in these isotopes (Rollinson, 1993). These depletions in the carbonate result from continuous (Rayleigh distillation) separation of CO_2 from residual rock resulting from decarbonation reactions. Normal or equilibrium decarbonation reactions can produce significant ^{13}C depletion but only limited ^{18}O depletion, due to the “calc-silicate limit” (Bowman, 1998b). This limit exists because the dominant oxygen reservoir will always be in the silicates in the residual metamorphosed rock (Valley, 1986).

It is apparent that the isotopic compositions of the carbonates from Haveri cannot be derived from decarbonation of typical marine limestones, although Baker and Fallick (1989) described a field of 2.0 Ga Proterozoic limestones with isotopic compositions that upon decarbonation could produce the observed isotopic compositions. However, Karhu (1993) has shown that sedimentary carbonates from the Palaeoproterozoic Svecofennian Domain have isotopic compositions similar to normal sedimentary (marine) carbonates. Calcites from the Central Svecofennian Province have a range of compositions between $\delta^{13}\text{C}$ values of 0.57 to 1.47 ‰ and $\delta^{18}\text{O}$ values of between 19.72 and 24.10 ‰ (Karhu, 1993). The results of normal calc-silicate decarbonation (at both 300 °C and 550 °C) on the sedimentary carbonates from the Central Svecofennian Province are shown in Figure 3.42. It is apparent that the calcites from Haveri plot outside the fields of decarbonation from a sediment, and mainly outside the typical “igneous” calcite field. Consequently, on isotopic evidence alone, it is clear that the Haveri carbonates are not of sedimentary or magmatic origin, and must therefore be of hydrothermal origin. The bulk of the calcites plot within the field of skarn calcite observed in the Mines Gaspé (Shelton, 1983, in Bowman, 1998b). The isotopic compositions of the carbonates will be discussed in reference to other hydrothermal deposit types in a later section (Section **Error! Reference source not found.**).

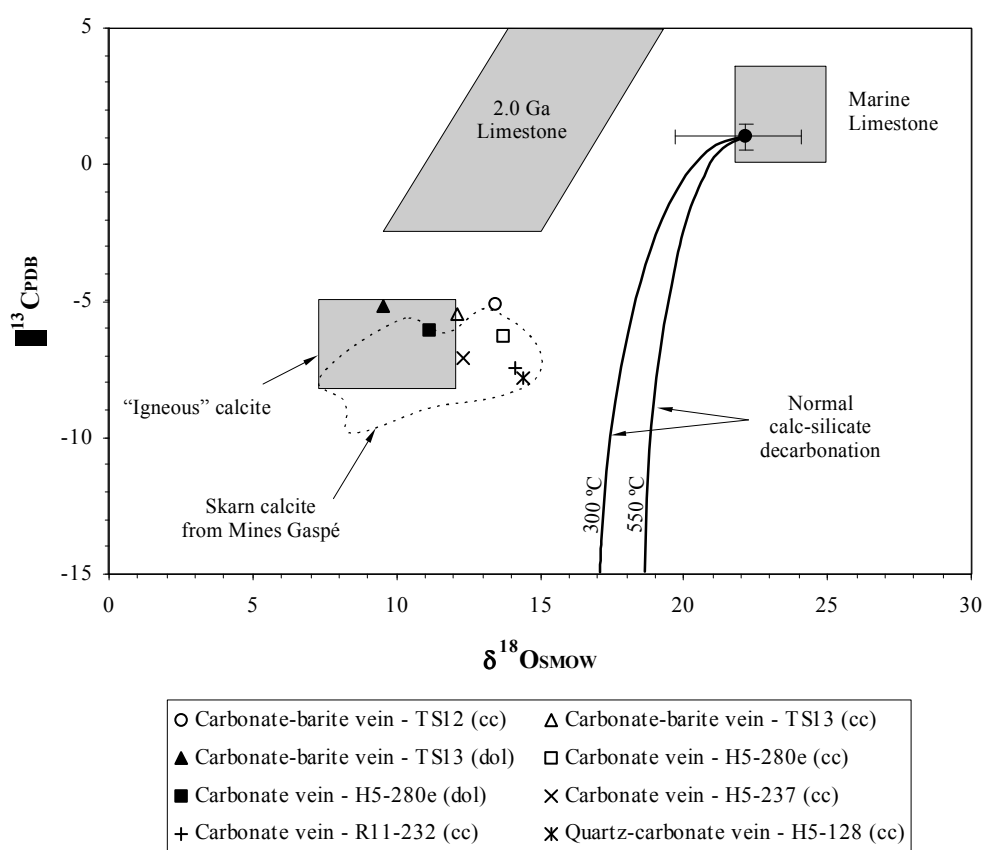


Figure 3.42: $\delta^{18}\text{O}$ vs $\delta^{13}\text{C}$ plot showing the composition of carbonates from carbonate and quartz-carbonate veins. Boxes outlining the $\delta^{13}\text{C}$ - $\delta^{18}\text{O}$ fields for marine limestone, "igneous" calcite (Bowman, 1998b) and 2.0 Ga limestone (Baker and Fallick, 1989). Igneous calcite is calcite in equilibrium with normal magmatic carbon ($\delta^{13}\text{C} = -5$ to -8 per mil). Decarbonation (Rayleigh distillation) trends at 300 and 550 °C for a sedimentary calcite from the Central Svecofennian Province (Karhu, 1993) (after Bowman, 1998b).

The 1 metre thick calcite-barite veins in the vicinity of the old mine and seen underground have previously been interpreted as "chemical sediments" and "limestone interlayers", indicative of a syngenetic origin (Stigzelius, 1944; Lupander and Räisänen, 1954; Anttonen, 1968; Ollila, 1977; Mäkelä, 1980; Kähkönen and Nironen, 1994; Nironen, 1994). These parallel calcareous layers have been used as evidence of primary layering, as well as in formulating models of ore genesis. However, in the course of recent exploration activities, and the subsequent overburden removal and dewatering of the Haveri open pit, more bedrock has been exposed, enabling rock relationships to be observed which have hitherto been hidden. In the walls in the open-pit, the parallel carbonate bands have been observed cutting the quartz-feldspar porphyry, thus demonstrating the epigenetic nature of the calcareous bands. Therefore both the isotopic data (samples TS12 and TS13) and the field relationships are in agreement that the carbonate bands are hydrothermal in origin.

3.3.7.2 Calcite-Dolomite isotope thermometry

Two samples of carbonate from quartz-carbonate veins contained calcite-dolomite pairs, determined by XRD. With respect to $\delta^{13}\text{C}$ the calcite-dolomite pairs are in near equilibrium, whereas the $\delta^{18}\text{O}$ values show strong disequilibrium, with the $\delta^{18}\text{O}$ values of dolomite being lower than those for calcite. Sheppard and Schwarcz (1970) showed that the oxygen isotope shift can be caused by a more rapid exchange of calcite with an aqueous fluid relative to dolomite. Consequently, Karhu (1993) has suggested that this pronounced disequilibrium of oxygen isotope compositions has resulted from retrograde, postmetamorphic reactions with an external oxygen reservoir, such as a retrograde aqueous fluid.

The fractionation of ^{13}C between calcite and dolomite is temperature dependent, and is therefore a potential geothermometer. Sheppard and Schwartz (1970) calibrated the calcite-dolomite isotope thermometer for a temperature range of 100 – 650 °C. However, the fractionation of carbon into calcite and dolomite is very similar, and thus even minor variations in $\delta^{13}\text{C}$ values of either calcite or dolomite will result in large variations in calculated temperatures.

Table 3.26 shows the calculated temperatures for the dolomite-calcite pairs from the Haveri carbonates according to the isotope thermometer of Sheppard and Schwartz (1970). The calculated temperatures are clearly far too high to be credible. This is to be expected because the composition of dolomite has been affected by retrograde reactions, as shown in Figure 3.42. The dolomite-calcite isotope thermometer is therefore unsuitable for determining temperatures of formation of the Haveri carbonates.

$1000\text{Ln}\alpha = 0.18(10^6/T^2) + 0.17 [T \text{ in } ^\circ\text{K}]$ Sheppard & Schwartz (1970)				
Sample	$\delta^{13}\text{C}_{\text{calcite}}$	$\delta^{13}\text{C}_{\text{dolomite}}$	$1000\text{Ln}\alpha$	T (°C)
TS13	-5.51	-5.2	0.310	861
H5-280e	-6.31	-6.04	0.270	1069

Table 3.26: *Calculated temperatures for calcite-dolomite pairs from Haveri according to the isotope thermometer of Sheppard and Schwartz (1970).*

3.3.8 Sulphur isotopes

In most metallic ore deposits the economic metals are fixed as sulphides. Even in deposits such as Haveri, where at least some of the economic metals are in the form of non-sulphides (e.g. native gold), sulphide minerals are typically associated with the ore. The sulphur in such deposits does not necessarily have the same history as the metals, or the fluids that transported the metals to the site of mineral deposition. Understanding the geochemistry of the sulphur (i.e. its source, transport and precipitation mechanisms) is therefore an important part of ore deposit studies (Ohmoto, 1986; Ohmoto and Goldhaber, 1997).

Sulphur occurs in nature in many different valence states: S^{2-} in sulphide species, S^0 in native sulphur, S^{4+} in SO_2 , and S^{6+} in sulphate species. Consequently, the redox conditions, as well as other physico-chemical parameters such as temperature, sulphur activity and pH, are important in controlling which sulphur species are present, and thus the overall isotopic composition of the sulphur, as the isotopic composition varies due to isotopic fractionation with sulphur speciation.

About 10% of the sulphur in the crust + ocean system is as sulphate in seawater; the remaining 90% is found in sedimentary rocks (~ 60%) as recycled seawater-sulphur, and igneous rocks (~ 30%) (Ohmoto, 1986). The isotopic composition of seawater sulphate has varied over time, present-day seawater has a $\delta^{34}S$ value of $+20.0\text{‰} \pm 1.0\text{‰}$. In the Archaean, $\delta^{34}S$ values are likely to have been in the region of +2, which increased (with decreasing ocean temperatures) to about +10‰ about 2.5 Ga ago (Ohmoto and Goldhaber, 1997).

Seawater-sulphur can be recycled and incorporated into sedimentary rocks, and ultimately into sulphides in ore deposits by a number of different mechanisms, with associated isotopic variations (Ohmoto and Goldhaber, 1997). One mechanism for the fixing of seawater-sulphate is by the direct precipitation of gypsum in evaporite sequences, with only a minor isotopic fractionation involved (enriched by about 1.5‰ at 25 °C), creating a crustal sulphur reservoir with average $\delta^{34}S$ values of +22‰. This mechanism accounts for approximately 60% of the amount of sulphur fixed from seawater-sulphate. After burial deep in a sedimentary basin, the gypsum may be converted to anhydrite by diagenesis and/or dissolved in basin water, and converted to H_2S by nonbacterial organic reduction.

Another mechanism (accounting for approximately 30% of the sulphur fixed) is by formation of biogenic pyrite in normal marine sedimentary rocks. Pyrite is formed by reactions between iron-bearing minerals in seawater (in euxinic basins) and/or sediments and H_2S (and/or HS^-) generated by bacterial reduction of seawater sulphate. This produces a crustal reservoir with average $\delta^{34}S$ values of -25‰ (and a range between -50‰ and +10‰).

The remaining 10% of sulphur fixed from seawater forms disseminated gypsum/anhydrite, pyrite and/or pyrrhotite in euxinic basins or in oceanic basalts by high-temperature (150 – 350 °C) reduction of seawater-sulphate during the convective circulation of seawater through the underlying volcanics. The $\delta^{34}S$ values of such sulphides are therefore shifted from the normal igneous value (0‰) to values typically between +5‰ and +8‰, reflecting the seawater-sulphate input.

During later diagenesis and metamorphism, the sulphides and sulphates produced by the above mechanisms can be converted into H_2S by decomposition, permitting replacement of biogenic sulphides or transport by hydrothermal fluids and precipitation as sulphide minerals elsewhere (Ohmoto and Goldhaber, 1997).

The $\delta^{34}\text{S}$ values of most igneous rocks were for years believed to be within the range of $0\text{‰} \pm 5\text{‰}$ (e.g. Ohmoto and Rye, 1979). However, with new data it is apparent that igneous rocks with $\delta^{34}\text{S}$ values outside of this range are quite common (Ohmoto and Goldhaber, 1997). It has become clear that the $\delta^{34}\text{S}$ values of the igneous rocks actually reflect the $\delta^{34}\text{S}$ values of the country rocks through which they are emplaced. There are several mechanisms through which a magma (or an igneous rock) can acquire country rock sulphur: (1) bulk assimilation of country rocks by a magma (i.e. transfer of both volatile and non-volatile elements from country rocks to magma); (2) selective assimilation (volatile transfer) of country rock sulphides by a magma; and (3) subsolidus reaction with fluids that circulated through the country rocks and the intrusive rock (Ohmoto and Rye, 1979).

The dominant sulphur species in magmatic fluids derived from I-type granitoid magmas can be either SO_2 and/or H_2S ; the lower $P_{\text{H}_2\text{O}}$ values (i.e. shallow magmas) favouring SO_2 . Because there exists $\sim 3\text{‰}$ to $\sim 6\text{‰}$ fractionation between SO_2 and H_2S at magmatic temperatures, the $\delta^{34}\text{S}_{\Sigma\text{S}}$ of fluids from I-type magmas can be similar to, or $\sim 5\text{‰}$ heavier than, the $\delta^{34}\text{S}$ values of sulphides in igneous rocks (Ohmoto and Goldhaber, 1997; Ohmoto and Rye, 1979).

3.3.8.1 Origin of sulphur in hydrothermal fluids

Studies summarised by Ohmoto and Rye (1979) show that the fractionation of ^{34}S between the different sulphur-bearing species in hydrothermal fluids is not simply controlled by temperature. Rather it is a function of the physico-chemical conditions of the fluid: these include oxygen activity, sulphur activity, pH and the activity of cations associated with sulphate (Rollinson, 1993). These studies imply firstly that the $\delta^{34}\text{S}$ of a hydrothermal fluid cannot be directly estimated from the $\delta^{34}\text{S}$ of sulphide minerals unless variables such as oxygen activity and pH are also known. Secondly, the $\delta^{34}\text{S}$ of a sulphide mineral cannot be used directly to determine the source of the sulphur since it is a function of many more variables than simply temperature.

In the determination of the sources of sulphur in ore deposits, Ohmoto (1986) states that in most geologic fluids with $T > 200\text{ }^\circ\text{C}$, the $\delta^{34}\text{S}$ values of individual fluid species are likely to be in equilibrium. Furthermore, although the $\delta^{34}\text{S}$ values of individual sulphur species will not indicate their source, the $\delta^{34}\text{S}$ value of total S in the fluid may be diagnostic of its source, and can be calculated for temperatures above $350\text{ }^\circ\text{C}$, by the following mass balance equation (Ohmoto, 1972):

$$\delta^{34}\text{S}_{\Sigma\text{S}} = \delta^{34}\text{S}_{\text{H}_2\text{S}} + \Delta_{\text{SO}_2-\text{H}_2\text{S}} \left(R / (1 + R) \right)$$

where $\Delta_{\text{SO}_2-\text{H}_2\text{S}}$ is the equilibrium fractionation factor between SO_2 and H_2S , and R is the mole ratio of SO_2 and H_2S in solutions

$$R = X_{\text{SO}_2} / X_{\text{H}_2\text{S}}$$

At very high temperatures ($T > 400\text{ }^{\circ}\text{C}$) the dominant sulphur species in hydrothermal systems are H_2S and SO_2 (Rollinson, 1993). Furthermore, as a rule (Ohmoto, 1986), when pyrrhotite occurs as a stable sulphide mineral, $\text{pH} < 6$, and temperatures are less than about $500\text{ }^{\circ}\text{C}$, H_2S is the dominant sulphur species in fluids (i.e. $R < 0.01$). This would imply that

$$\delta^{34}\text{S}_{\text{sulphides}} \approx \delta^{34}\text{S}_{\Sigma\text{S}}$$

In other words, that the $\delta^{34}\text{S}$ values from the sulphides may be diagnostic of their source.

Mäkelä (1980) analysed a suite of pyrrhotites for sulphur isotopes from the Haveri Formation, taking the samples from the metalavas breccias, the metatuffs and the metatuffite (Table 3.27).

Rock type	Samples (n)	Average $\delta^{34}\text{S}$	Maximum $\delta^{34}\text{S}$	Minimum $\delta^{34}\text{S}$
Metalavas	8	+6.3	+8.7	+3.4
Metatuff	4	+4.3	+4.7	+3.4
Metatuffite	2	-0.3	0.0	-0.6
	14	+4.8		

Table 3.27: $\delta^{34}\text{S}$ values of pyrrhotite from various lithologies within the Haveri Formation. Data from Mäkelä (1980)

During the course of this present study a further 14 samples of pyrrhotite were taken from a variety of host rocks and ore types (Table 3.28). These samples have similar $\delta^{34}\text{S}$ values to those obtained by Mäkelä (1980).

Sample	Description	Mineral etc.	$\delta^{34}\text{S}$	Laboratory
P22-62b	Cataclastite	Pyrrhotite	5.99	UTA
H5-291	Cataclastite	Pyrrhotite	2.48	UTA
R7-191	Cataclastite	Pyrrhotite	2.89	UTA
P22-63b	Sulphide & Au	Pyrrhotite	5.47	UTA
P22-70	Sulphide & Au	Pyrrhotite	11.35	UTA
HM1-14	Disseminated in meta-tuff	Pyrrhotite	-0.27	UTA
R12-51	Calcic skarn	Pyrrhotite	4.25	UTA
R7-125	Quartz vein related	Pyrrhotite	3.32	UTA
R7-174a	Quartz vein related	Pyrrhotite	3.17	UTA
R7-286	Quartz vein related	Pyrrhotite	2.79	UTA
H5-280	Quartz vein related	Pyrrhotite	3.65	UTA
R11-53	From sediment	Pyrrhotite	4.03	UTA
P13-39	From porphyry	Pyrrhotite	4.89	UTA
TS-29	From sediment	Pyrrhotite	-0.28	UTA
R7-191	Cataclastite (duplicate)	Pyrrhotite	3.08	UTA

Table 3.28: $\delta^{34}\text{S}$ values of pyrrhotite from various lithologies and ore types at Haveri.

The majority of the analyses are enriched in ^{34}S with values between +2.48‰ and +5.99‰. Two samples (HM1-14 disseminated in meta-tuff; TS-29 from sediment) have $\delta^{34}\text{S}$ values close to zero, which is typically indicative of a magmatic source ($\delta^{34}\text{S} = 0\text{‰}$) for the sulphur. However, as the other sulphide sample from the sediment does not show this relative depletion, it is perhaps more likely that there is a mixed sulphur source, with a minor input from isotopically lighter biogenic pyrite in the sediment and meta-tuff. One sample (P22-70, sulphide + Au) has an anomalously high value of $\delta^{34}\text{S}$ ($\delta^{34}\text{S} = 11.35\text{‰}$) which is considered to be as a result of analytical error, or sample contamination. Excluding sample P22-70, the pyrrhotite samples have a narrow range of $\delta^{34}\text{S}$ values with a mean of +3.25‰. This range of values is consistent with isotopic compositions of a number of magmatic-hydrothermal deposit types (Ohmoto and Goldhaber, 1997), including Cu-Au skarn deposits and Fe-oxide Cu-Au deposits as well as mesothermal lode-Au deposits (e.g. Ohmoto, 1986; Ohmoto and Goldhaber, 1997; Taylor, 1987). These values are also consistent with volcanogenic massive sulphide (VMS) deposits of Proterozoic and Phanerozoic ages ($\delta^{34}\text{S}$ values of $+5\text{‰} \pm 5\text{‰}$), and was used by Mäkelä (1980) as one of the principal reasons for proposing such an ore deposit model for Haveri. The various deposit types potentially comparable to the Haveri deposit are discussed in greater detail (including reference to their typical sulphur isotope compositions) in Section 5.1.

3.3.8.2 Sulphur isotope thermometry

One of the applications of sulphur isotopes is in thermometry. There are a number of theoretical and experimental determinations of the fractionation of $\delta^{34}\text{S}$ between coexisting sulphide phases (sulphide-sulphide and sulphide-sulphate) as a function of temperature. The partitioning of sulphur isotopes between sulphides is not a particularly sensitive thermometer and requires precise isotopic determinations (Rollinson, 1993). The mineral pairs which have equilibrium isotopic fractionation factors most sensitive to temperature are those involving sulphate (e.g. anhydrite and barite) and sulphides (e.g. pyrite, sphalerite). Co-existence of sulphides and sulphates is quite common in a variety of deposits, and has been observed as co-existing barite and pyrite in the parallel carbonate bands to the west of the Haveri open pit. However, the observed fractionation factors between sulphate and sulphide minerals are not the same as the equilibrium values at depositional temperatures. Because the observed fractionation factors are usually smaller than the equilibrium values, sulphide-sulphate mineral pairs are not suitable for obtaining temperatures of mineralisation (Ohmoto, 1986).

3.4 FLUID INCLUSIONS

Fluid inclusions are tiny blebs of fluid trapped in crystals at the time of their growth, or introduced along microcracks and cleavages after crystallisation of the host mineral. They are usually $<50\ \mu\text{m}$ in size, and are found trapped within single crystals of both ore and gangue minerals, although normally only transparent phases are studied. As fluid inclusions are actual samples of long-gone fluids that were present in the rock, they are able to provide direct, quantitative data on the temperature, pressure, density and composition of the hydrothermal fluid in question.

In terms of their origin, three types of fluid inclusions are recognised, namely: primary, secondary and pseudosecondary. Primary inclusions are those that became trapped during the growth of the host mineral, and are therefore associated with crystallisation features – such as growth zones; or they may occur isolated due to imperfections of the crystal during its growth. Secondary inclusions are those that form after the growth of the host mineral is completed. They cut across growth zones and even crystal boundaries, and may represent infilling of microcracks by late fluids that might well be unrelated to the ore-forming event. Pseudosecondary inclusions form during the two stages outlined above, and are characterised by their alignment with microcracks that end against a growth zone. (Pirajno, 1992). These definitions are somewhat simplistic, and it is often difficult to distinguish between different types especially as most natural samples contain numerous generations of fluid inclusions.

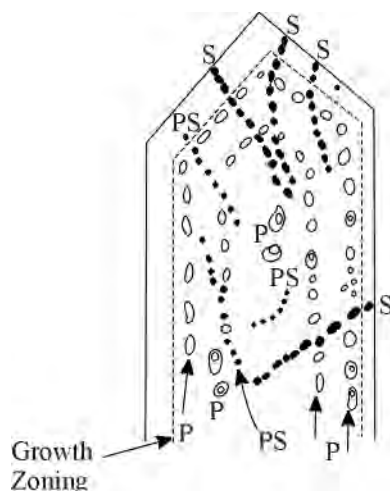


Figure 3.43: Types of fluid inclusions, and their distribution in a quartz crystal. **P** Primary; **S** secondary; **PS** pseudosecondary (After Shepherd et al., 1985).

Inclusions are usually formed by the trapping and hence preservation of a single homogeneous fluid at some elevated temperature (and pressure) during the growth of the crystal. However, on cooling to room temperature a variety of phase changes may occur. The most common of these is the formation of a vapour bubble by differential shrinkage on cooling (the fluid shrinks more than the host crystal), but other phases, including crystals (“daughter” phases) may also form in the fluid (Roedder and Bodnar,

1997). The classification scheme of Shepherd et al. (1985), based on morphology and contents of inclusions, is summarised below, and diagrammatically shown in Figure 3.44.

- I. Monophase inclusions: entirely filled with liquid (L).
- II. Two-phase inclusions: filled with a liquid phase and a small vapour bubble (L + V).
- III. Two-phase inclusions: in which the vapour phase is dominant and occupies more than 50 % of the volume (V + L).
- IV. Monophase vapour inclusions (V): entirely filled with a low density vapour phase (generally mixtures of H₂O, CH₄, CO₂).
- V. Multiphase inclusions containing solids (S + L ± V): contain solid crystalline phases known as daughter minerals. These are commonly halite (NaCl) and sylvite (KCl), but many other minerals may occur including sulphides.
- VI. Immiscible liquid inclusions: contain two liquids, usually one H₂O-rich and the other CO₂-rich (L1 + L2 ± V)

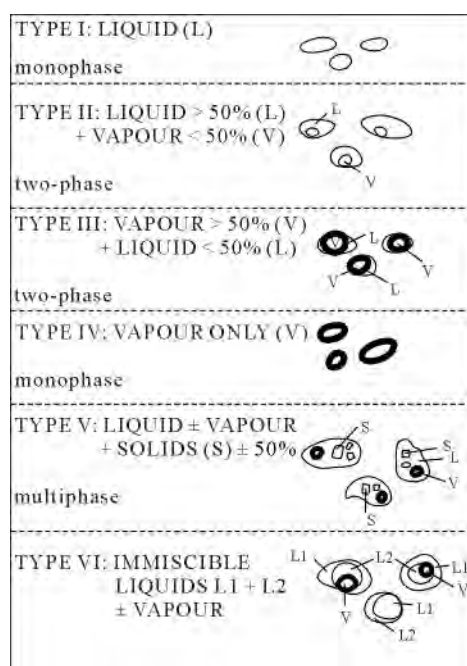


Figure 3.44: Classification of fluid inclusions observed at room temperature (after Shepherd et al., 1985).

There are a number of basic (but important) assumptions that are applicable to all studies of fluid inclusions (Roedder and Bodnar, 1997):

1. A representative sample of homogeneous fluid (possibly the ore-forming fluid in ore-deposit studies) is trapped and sealed during the growth of a crystal.
2. Nothing is added or lost from the inclusion after trapping.

3. The volume of the vacuole (cavity) surrounding the inclusion fluid does not increase or decrease after trapping.
4. The relationships are known between the actual trapping event, both in space and particularly in time, and the geologic process of interest, such as deposition of ore.
5. The effects of pressure are insignificant or are known.

Although many samples do not fulfill all of these assumptions, if the deviations are known (or can be estimated) and are thoughtfully considered, it is still possible to infer useful information from the fluid inclusions.

3.4.1 Analytical Techniques

After accurate petrography has been carried out on the samples, and the overall mineral assemblages have been correlated to particular events, samples can be prepared for description and analysis. Sample preparation involves making a thin section polished on both sides to a high quality of polish. In preparation, it is important to keep temperatures below 100 °C to avoid decrepitation of the inclusions.

The doubly polished thin section (removed from a glass slide) is examined under the optical microscope for fluid inclusions. Observed fluid inclusions are mapped in relation to recognisable features under the microscope, and described in terms of their size, genesis, morphology and contents, as well as relationships to other inclusions (individual, clusters or trails). If the inclusions are two-phase, an estimate of the proportions is made. Once the petrography of the inclusions has been documented, the section is then carefully broken into pieces small enough to fit into the heating-cooling stage (approx. 1 cm in diameter).

The heating-cooling stage used for the analyses is a Linkham THMSG 600, controlled by a Linkham TMS 93 with an attached liquid nitrogen pump. The microscope used was a Nikon E600-Pol with a 50x objective and a 10x ocular, providing 500x magnification.

Having located the inclusions in question under the microscope in the heating-cooling stage, the sample is super-cooled until the whole inclusion has become solid. The temperature is slowly increased, and the temperature of initial melting (T_e) is recorded (if possible). This provides the eutectic melting and the likely composition of the system. Further heating is continued until the temperature of melting (T_m) is attained with the specific phase indicated (e.g. $T_{m \text{ NaCl}}$ or $T_{m \text{ ice}}$). This temperature provides an indication of the salinity in hydrous fluids. When all the phases have melted, heating is continued until phases start to disappear, and the homogenisation temperatures recorded (T_h). In a two-phase inclusion this is usually when the vapour bubble decreases in size until it finally homogenises with the fluid. Homogenisation temperatures are not measured until all of the cooling temperature measurements have been conducted, due to the possibility of stretching or decrepitating the inclusions during heating. In

fluid inclusion studies, all phase changes of all observable phases are recorded with increasing temperature to avoid erroneous temperatures caused by phase metastability.

3.4.2 Observations and Results

At Haveri a number of samples were examined for fluid inclusions. The samples prepared were taken from mineralised and unmineralised quartz veins associated with amphibolitisation, and containing amphibole inclusions, as well as sulphides.

The amphiboles both in the veins and in the host-rock were examined for fluid-inclusions. Unfortunately, although the presence of possible inclusions was occasionally detected, they were so small ($<1\ \mu\text{m}$) that they were beyond the resolution of the microscope, making it impossible to determine if they were indeed fluid inclusions, or minute mineral inclusions.

The quartz veins themselves generally have a high density of fluid inclusions with the majority being in visible trails. Unfortunately, the quartz veins show considerable sub-grain development, indicative of post-vein deformation. Most of these secondary inclusion trails either outline the sub-grain development of the quartz veins, or even cross-cut the sub-grain boundaries. These secondary trails are interpreted as late- to post-deformation, and thus clearly post-date the formation of the quartz veins. The clearest quartz grains have no inclusions, and thus possibly indicate recrystallisation of the quartz.

Rare individual fluid inclusions have been observed in the less deformed grains, commonly close to mineral inclusions within the quartz vein (Figure 3.45). It is possible that the presence of such mineral inclusions (typically sulphides) provided a localised region of lower stress, enabling the inclusions to survive the later deformation. Although these individual inclusions do not appear to be related to visible inclusion planes, they can not be definitively classified as “primary” according to the scheme of Roedder (1984). However, they are the earliest generation of secondary inclusions present in the sample, and may be related to the initial growth of the quartz in the veins. In total only 7 of these small early inclusions have been identified from two separate sections. These rare inclusions are typically small, being less than $5\ \mu\text{m}$ in the longest dimension, and are either oblate spheroids in shape or exhibit negative crystal shapes. Unfortunately, their small size makes quantitative data collection problematic, although it is still possible to extract meaningful information from them. All of these inclusions can be classified as Type II two-phase inclusions with liquid $>$ vapour (at room temperature). None of the inclusions display visible evidence of containing a CO_2 phase, indicating low concentrations of CO_2 . Due to the small size of these inclusions it was only possible to obtain microthermometric measurements on three inclusions (Table 3.29), the others were simply too small to see phase changes. Even for the three inclusions for which measurements were made, the size of the inclusions made it impossible to determine the eutectic temperatures (T_e). Consequently it was necessary to determine salinities from the freezing point depressions ($T_{\text{m ice}}$) in terms of equivalent weight percent NaCl, according to the calculations of Bodnar (1992).

Sample	Dimensions	Shape	Type	$T_{m \text{ ice}}$	wt% NaCl	$T_{h\text{-liq}}$
R7-292 (area 3a)	5 x 4 (μm)	negative crystal	Type II: $L > V$	-3.3 °C	5.4	136 °C
R7-292 (area 5a)	4 x 3 (μm)	negative crystal	Type II: $L > V$	-6.4 °C	9.7	141 °C
R7-292 (area 5b)	4 x 3 (μm)	negative crystal	Type II: $L > V$	-6.5 °C	9.9	172 °C

Table 3.29: Description of early-formed individual fluid inclusions from quartz in quartz vein, which have no visible relationship to inclusion planes or other secondary features. The equivalent wt% NaCl has been determined from freezing point depression tables calculated by Bodnar (1992).

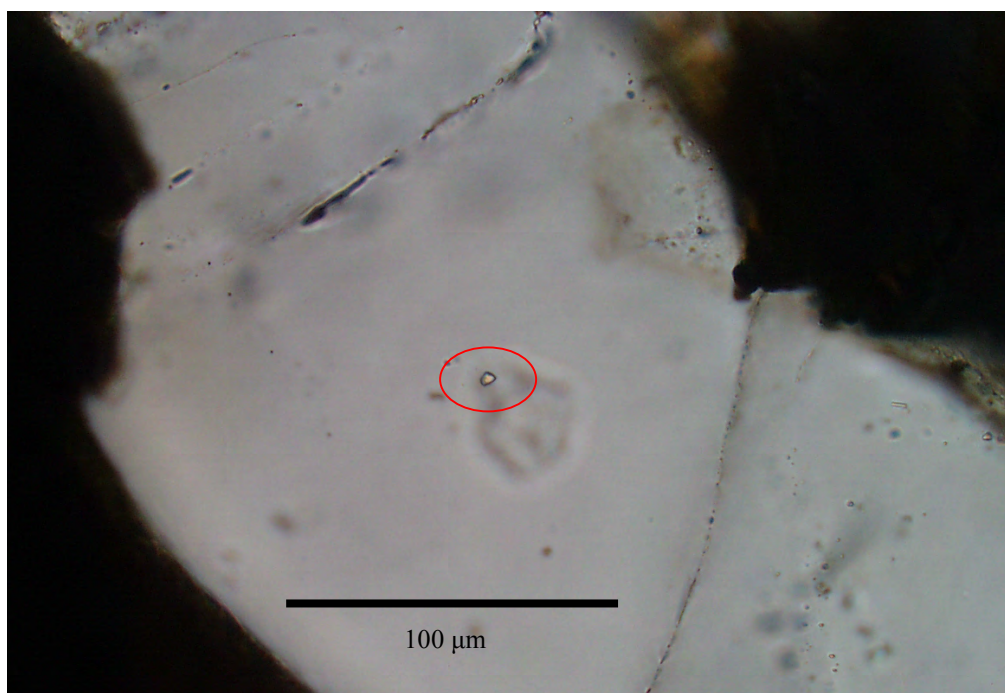


Figure 3.45: Early formed Type II: $L > V$ fluid inclusion in a relatively undeformed quartz vein, close to inclusions of pyrrhotite. There is no obvious relationship to inclusion planes or other secondary features. The inclusion is however spatially close to an amphibole crystal. Area 3a from sample R7-292.

Before the results from these measurements are considered, it is necessary to discuss other fluid inclusion textures observed in the quartz veins. In addition to the rare individual inclusions and the extensive late-stage secondary inclusion trails, there are also inclusion textures indicative of reequilibration.

One of the basic assumptions of fluid inclusion geothermobarometry is that fluids represent constant volume (isochoric) systems. For most inclusions trapped in most minerals from the majority of geological environments this assumption is valid, and the fluid inclusion data provide a reasonable approximation of the P-T conditions of formation (Bodnar et al., 1989). However, if the fluid inclusions follow a P-T path that differs from the inclusion isochore, then the inclusion may reequilibrate with leakage or decrepitation occurring, and the loss of some or all of the inclusion contents. When this occurs the inclusions no longer provide accurate information on the P-T conditions of formation and/or the composition of the fluid from which the mineral grew (Bodnar et al., 1989). However, the textures produced during reequilibration can themselves give information on the P-T path that rock has been subjected to since the fluid inclusions were formed (Bodnar et al., 1989; Sterner and Bodnar, 1989; Vityk and Bodnar, 1995).

Vityk and Bodnar (1995) recognise four general non-isochoric P-T paths which a fluid inclusion might follow after entrapment that leads to reequilibration (Figure 3.46). Each path may involve a variety of geological processes (such as subduction, thrusting, tectonic and erosional denudation, magma intrusion, etc.) that cause perturbation of the crustal thermal structure and/or thickness. Thus crustal thickening by subduction, thrusting and sedimentation may result in a nearly isothermal compressional P-T path (ITC). Crustal thinning accompanied by tectonic uplift, erosion and extension might produce an isothermal decompressional P-T path (ITD). The isobaric cooling path (IBC) usually occurs in crust of near-normal thickness and might be induced by processes such as magma accretion into and under continental crust, extension of normal crust with concomitant magmatic underaccretion, or extension of previously over-thickened crust in collisional or continental arcs. A nearly isobaric heating path (IBH) simulated magmatic underplating. The effect of these various crustal P-T histories on fluid inclusion characteristics depends on many factors including P-T conditions of reequilibration, magnitude of the departure of the inclusion from the original isochore, slope and shape of the P-T trajectory, rates of crustal perturbations, thermal conductivity of rocks, etc. The four types of P-T path can be divided into those that result in internal overpressure (IBH and ITD) and those resulting in internal underpressure (IBC and ITC) depending on whether the pressure inside the inclusion either exceeds or is less than the confining pressure respectively. Reequilibration under both underpressure and overpressure conditions may involve fracturing, stretching, shrinking, dissolution or recrystallisation of the inclusion walls or preferential loss of some fluid components (Vityk & Bodnar, 1995).

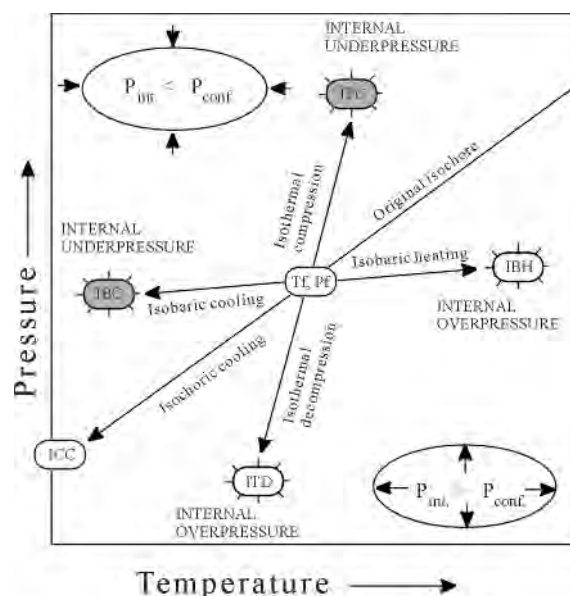


Figure 3.46: Schematic representation of P-T paths which a fluid inclusion might follow after entrapment. T_f , P_f correspond to the initial trapping conditions. After trapping, the inclusion might be subjected to essentially isobaric heating (IBH) or isothermal decomposition (ITD), both of which generate internal pressures (P_{int}) that exceed the confining pressure (P_{conf}). Conversely, the path may approximate isobaric cooling (IBC) or isothermal compression (ITC) and lead to internal underpressure. On rare occasions the path may approximate the fluid inclusion isochore [isochoric cooling (ICC)], resulting in an effective pressure of zero ($P_{internal} = P_{external}$). Each of the different P-T paths described may result in a different and characteristic fluid inclusion re-equilibration texture (after Vityk & Bodnar, 1995).

By experimental reequilibration of synthetic 10 wt% NaCl-H₂O inclusions in natural quartz, Vityk and Bodnar (1995) have shown that characteristic fluid inclusion re-equilibration textures can be produced by the four different P-T paths described above. Specifically, implosion textures generated under conditions of isothermal compression at high temperatures, and implosion textures produced at low temperatures during isobaric cooling, are distinctly different. Similarly explosion textures produced during isothermal decomposition are different from those produced as a result of isobaric heating.

In addition to textural and morphological changes with reequilibration, Vityk and Bodnar (1995) have also shown the effects on the density (and hence homogenisation temperatures) of the inclusions. In the majority of cases (and with some differences in detail), the homogenisation temperatures of the inclusions span the range of temperatures between the original formation conditions, and the homogenisation temperatures expected if the inclusions completely reequilibrated to the new P-T conditions.

Extrapolation of the results from synthetic fluid inclusions to natural inclusions is potentially problematic. However, Vityk and Bodnar (1995) report that the application of laboratory results to natural samples has been remarkably successful in reconstructing the history of naturally deformed rocks.

Fluid inclusion reequilibration textures have been observed in the quartz veins at Haveri. Although they are not plentiful, they occur in similar proportions to the rare individual inclusions described above. Furthermore, the distribution of the re-equilibrated inclusions is similar, being located in isolation within somewhat less altered quartz grains, and having no obvious relationship to secondary planes. The reequilibrated fluid inclusions are recognised as three-dimensional clusters of small secondary inclusions surrounding the parent inclusion (Figure 3.47). These textures have been described by Sterner and Bodnar (1989) and Vityk and Bodnar (1995), and have been ascribed to conditions of internal underpressure with an effective pressure differential of between 1.8 – 2.7 kbar. Vityk and Bodnar (1995) have characterised this texture to a P-T path of isobaric cooling.



Figure 3.47: Photomicrograph of a decrepitated fluid inclusion from sample R8-219. The three-dimensional cluster of small secondary inclusions surrounding the parent inclusion, is an implosion texture resulting from isobaric cooling (Vityk & Bodnar, 1995).

The homogenisation temperatures observed by Vityk and Bodnar (1995) at equivalent levels of internal underpressure (2.4 kbar), and along a P-T path of isobaric cooling, span a range of 192 – 299 °C. This covers the range from slightly higher temperatures than for the original conditions of formation ($T_h = 284$ °C) down to those expected if the inclusions completely reequilibrated to the new P-T conditions ($T_h = 196$ °C). The fact that some inclusions have slightly higher T_h values than the original T_h , suggests that some of the inclusions experienced a small decrease in density. The implications of this to the observed data at Haveri, is that the measured homogenisation temperatures (for the early inclusions) are most likely to be lower than their original values. If microthermometric measurements could be made on sufficient inclusions, the homogenisation temperatures would be expected to define a range of T_h values with the maximum approximating the original T_h values, and the minimum approximating the T_h value of the reequilibration conditions.

In summary, the early-phase fluid inclusions related to quartz veining (and hence amphibolitisation and gold mineralisation) have textures indicative of isobaric cooling with an effective internal underpressure of approximately 2.4 kbar. Consequently, homogenisation temperatures (136 – 172 °C) are likely to reflect reequilibration to lower P-T conditions. The composition of the inclusions suggests that they are < 10 wt% equivalent NaCl-H₂O inclusions with a low concentration of CO₂. Low salinity inclusions (with < 10 wt% equivalent NaCl-H₂O) are characteristic of epithermal, VMS and lode-gold deposits, as well as some skarn deposits (Figure 3.48) (Lattanzi, 1994). In addition, Pollard (1999) reports the presence of low-salinity, aqueous inclusions, in many of the Fe-oxide Cu-Au deposits of the Cloncurry district, in the Proterozoic of Australia. Homogenisation temperatures for these deposits is typically higher than those observed at Haveri. However, allowing for the effects of reequilibration, it is likely that the early-phase inclusions from Haveri, originally plotted within these fields. The significance of the fluid inclusion results in terms of mineralisation are discussed further in Section 5. However, it is sufficient to say that the reequilibration textures support a model of contact metamorphism, with the intrusion of hot, granitoids at shallow levels (low pressure) within the crust, possibly in an arc environment (Section 4).

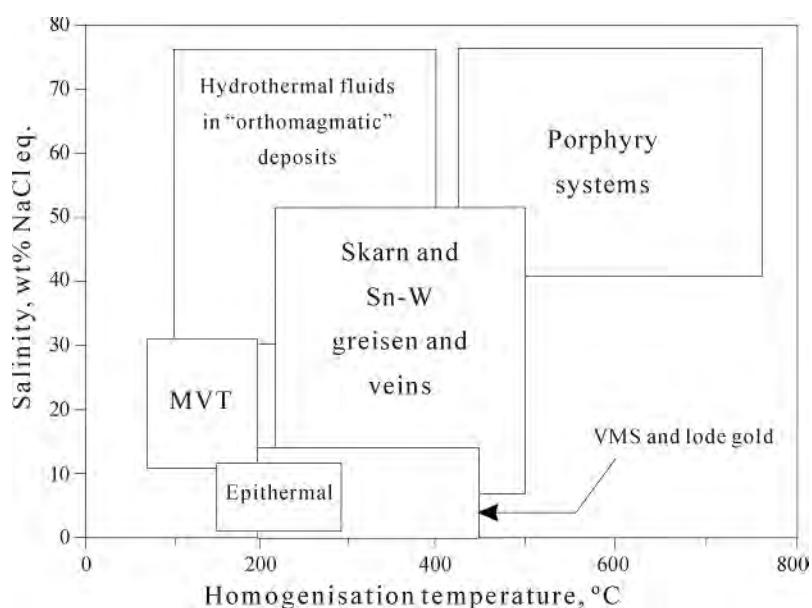


Figure 3.48: Schematic representation of salinity-temperature relationships of fluid inclusions typically associated with some deposit types. The boundaries for each field are somewhat arbitrary, and serve only as a broad approximation. (Modified after Lattanzi, 1994).

The later secondary trails of fluid inclusions are also able to provide information on the P-T conditions that the rocks have been subjected to. The trails themselves are commonly randomly orientated (although parallel planes are also not uncommon) and cut across both grain boundaries and sub-grain boundaries in the quartz (Figure 3.49), indicating that they are relatively late and post-date the deformation. The fluid inclusions within the trails have a variety of shapes ranging from negative crystal shapes through to arcuate shapes. Likewise, the inclusions are a variety of types ranging from Type I (liquid) and Type IV (vapour) monophasic inclusions through Type II and Type III two-phase

inclusions, to Type V multiphase inclusions with a solid phase. The Type V inclusions, have been observed with cubic solid phases (Figures 3.50 – 3.51), which have been assumed to represent a hypersaline fluid. However, the majority of inclusions in the secondary trails are Type II two-phase, L>V inclusions. The sizes of the inclusions varies from below optical resolution (at 500x magnification), up to 15 x 10 μm for an inclusion with a negative crystal shape.

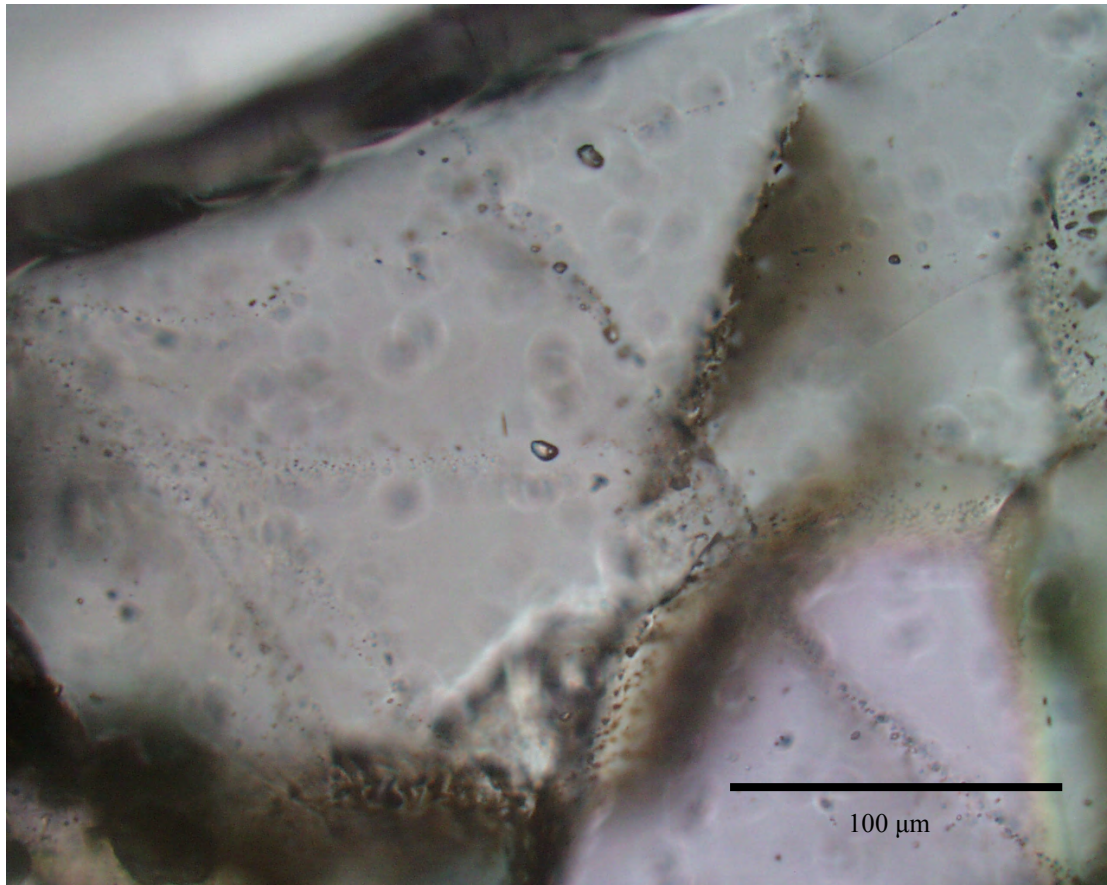


Figure 3.49: Fluid inclusions in quartz vein material in sample R7-292. Secondary inclusion trails are clearly visible, cutting grain and sub-grain boundaries. Two Type II: L > V inclusions, are visible that are related to trails.

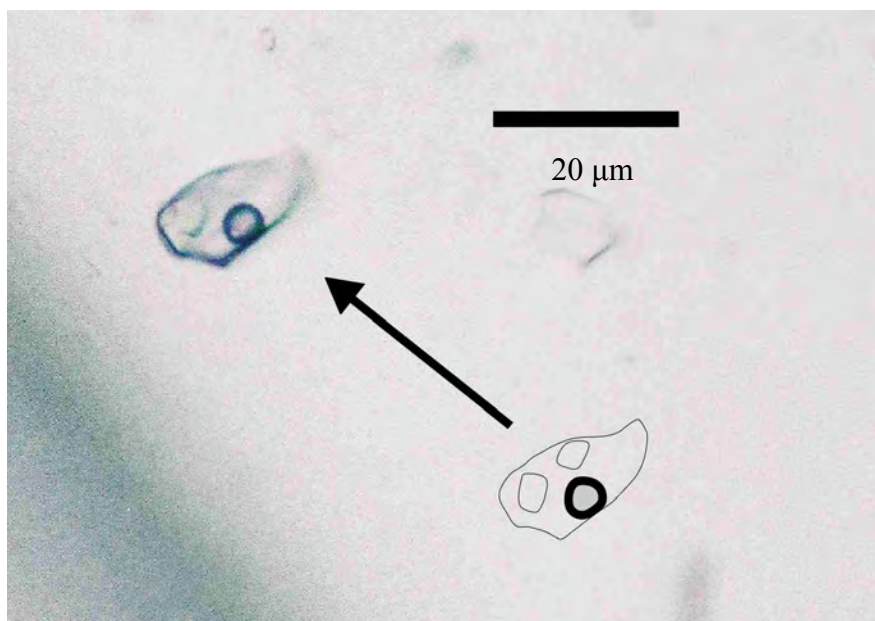


Figure 3.50: Type V: Liquid + vapour + solid fluid inclusion, related to a trail of secondary inclusions. There appear to be two solid phases. (Sample R7-292).

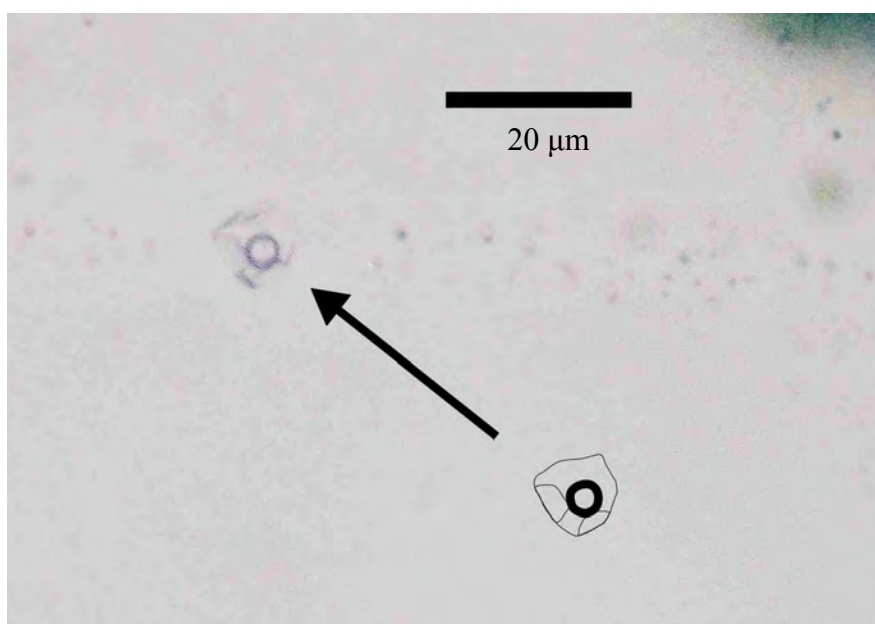


Figure 3.51: Type V: Liquid + vapour + solid fluid inclusion, related to a trail of secondary inclusions. There appear to be two solid phases. (Sample R7-292).

As with the more primary looking inclusions described earlier, there is typically no visible CO_2 present in the secondary inclusions, indicating low concentrations of CO_2 . However, in a few Type III two phase, $\text{V}>\text{L}$ inclusions, bulging of the meniscus may indicate the presence of minor CO_2 (Figure 3.52). A number of these inclusions would need to be observed and analysed in order to confirm this. Analyses of the larger Type II two-phase, $\text{L}>\text{V}$ inclusions (Table 3.30), show that they are low salinity inclusions with 2.4 wt% equivalent NaCl, and a considerable range in measured homogenisation temperatures ($T_{\text{h-liq}} = 179 - 230\text{ }^{\circ}\text{C}$). The low salinities observed in these essentially aqueous inclusions

contrasts strongly with the Type V inclusions containing a solid salt phase. The secondary inclusion trails are therefore likely to represent a phase separation, and that the overall initial salinities of the inclusion trails are likely to have been considerably higher than the limited analyses suggest. Due to the low number of late-phase, secondary inclusions that were analysed, it is not possible to categorically state that the inclusions in the secondary planes have been necked, however, the wide range of inclusion types, compositions and morphologies does add credence to this possibility. Consequently, it is possible that with necking some of the contents (including CO₂) have been lost.

Sample	Dimensions	Shape	Type	$T_{m \text{ ice}}$	wt% NaCl	$T_{h \text{-liq}}$
R7-292 (area 1)	15 x 10 (μm)	negative crystal	Type II: L > V	-1.4 °C	2.4	179 °C
R7-292 (area 2)	10 x 5 (μm)	negative crystal	Type II: L > V	-1.4 °C	2.4	230 °C

Table 3.30: Description of fluid inclusions from secondary inclusion trails through quartz in quartz vein. The equivalent wt% NaCl has been determined from freezing point depression tables calculated by Bodnar (1992).

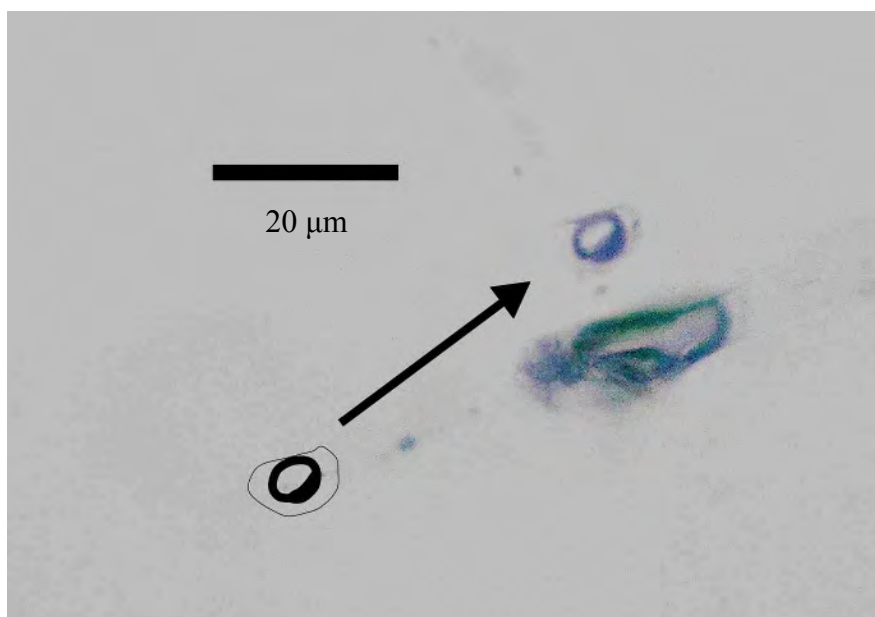


Figure 3.52: Type III: two-phase, $V > L$ fluid inclusion, related to a trail of secondary inclusions. It is possible that this inclusion does contain some minor CO₂. (Sample R7-292).

4 P-T-X CONDITIONS

In order to generate an accurate model of the geological evolution of the Haveri deposit, it is necessary to determine and understand the various *P-T-X* conditions that the rocks have been subjected to during the various stages of their formation. This requires the accurate identification and interpretation of the various mineral assemblages associated with the main geological events (Chapter 2). The *P-T-X* conditions of the mineral assemblages can then be determined either by indirect methods, such as using petrogenetic grids or mineral chemistries (geothermobarometry), or where circumstances permit, by more direct methods where fluids present during the event have been trapped and preserved as fluid inclusions (Section 3.4). Most commonly a combination of these techniques will be used to construct a model for the evolution of the *P-T-X* conditions over time.

Petrogenetic grids depict the stable mineral assemblages for all bulk compositions over P-T space (Spear, 1993) and, by accurately describing the equilibrium mineral assemblages present in the rock, it is possible to constrain a range of P-T conditions that describes the observed mineral assemblage. Furthermore, textural interpretations of the mineral reactions that have taken place permit inferences to be made about the evolving path (within P-T space) that the rock has been subjected.

Geothermobarometry is the calculation of pressures and temperatures of equilibration, using the temperature and pressure dependence of the equilibrium constant for the coexisting minerals of interest (Spear, 1993). The pressures and temperatures are calculated by the combination of thermodynamic variables and the mineral analyses, and then applied to a calibrated formula which has been determined either theoretically or experimentally. With these data, a line of constant equilibrium can be drawn on a P-T diagram, and it is inferred that the sample must have equilibrated somewhere along this line (Spear, 1993). Different types of geothermobarometers are available, depending on the assemblages and reactions that are present in the sample. These include exchange reactions and net transfer reactions. Exchange reactions are commonly temperature dependent, such as the exchange of Fe and Mg between garnet and biotite, and thus make good geothermometers, whereas net transfer reactions, that cause the production and consumption of phases, commonly result in large volume changes, and hence are more sensitive to pressure and make good geobarometers. By using a number of different equilibria (and calibrations) it may be possible to generate a number of lines of equilibrium on a P-T diagram that have an approximate point of intersection.

Other geothermobarometers include solvus thermometers, which are based on the compositions of phases coexisting across a solvus. The two exchanges used in solvus thermometry are Ca-Mg exchange in the pyroxenes and carbonates and K-Na exchange in the micas and feldspars (Spear, 1993). Other more specific geothermobarometers applicable to the Haveri region include the cation site occupancy in chlorite as a geothermometer (Cathelineau, 1988), and the FeS content in sphalerite as a geobarometer (Lusk and Ford, 1978; Scott, 1976).

There are numerous geothermobarometers available within the literature, often with a considerable number of separate calibrations. Some of these thermobarometers are based on simple thermodynamic models, and some are based on more complex models; some of these have been theoretically and some experimentally determined. The overall result is a multitude of different calculated P-T conditions for any supposed event. To be able to provide some insight into which results are “correct” and which are “incorrect”, it is necessary to appreciate the reasons for the variability in the results. It is therefore important to have an understanding of the assumptions that have been made in calibrating the particular thermobarometers, as well as their limits and constraints. Likewise, it is also necessary to understand likely causes of error and deviation in the results produced.

The goal in geothermobarometry is to infer the conditions at which a sample equilibrated by measuring the value of the equilibrium constant and using that value combined with the other thermodynamic variables (free energy, enthalpy, entropy, volume and heat capacity of reaction), to calculate temperature, pressure or both (Spear, 1993). The equilibrium constant (K) for a reaction can be divided into an “ideal” term (equal to the distribution coefficient K_D), and a “non-ideal” term which involves the activity coefficients ($K\gamma$). In studies that assume ideal mixing in equilibrium phases, the activity coefficients ($K\gamma$) equal 1, and thus the equilibrium constant (K) equals the distribution coefficient (K_D). This assumption can be justified if the deviations from ideality in the phases are small, or the deviations from ideality tend to cancel out (Chipera and Perkins, 1988). However, naturally occurring minerals are usually more complex than the simple mineral chemistries, with impurities and non-ideal mixing on cation sites. Consequently, activity models must be derived for these more complex phases to determine the activity coefficients. Unfortunately, such relations are poorly understood for most minerals, and thus the determination of the activity coefficients to account for all the additional components in a mineral is one of the principal sources of deviation between different calibrations of geothermobarometers.

Geothermobarometers are usually calibrated on systems within a restricted range of mineral compositions, and are commonly valid only to a stated range of pressure or temperature. The extrapolation of calibrated thermobarometers to include other compositions, or to assemblages with temperatures or pressures beyond that intended is another source of possible error in geothermobarometry (Essene, 1982).

One of the greatest potential sources of error in geothermobarometry is in the assumption that the minerals in question are in equilibrium and represent the metamorphic conditions being investigated. Accurate petrography is therefore critical in determining the relationship of the phases, and in assessing whether mineral chemistries are likely to have been retrograded or reset by chemical diffusion. In many cases, minerals such as garnets may display a chemical zonation, and a decision needs to be made as to whether the core or the rim represents the metamorphic condition of interest (Essene, 1982).

A final (but perhaps understated) source of error is in the determination of the mineral compositions themselves. Most mineral analyses are presently measured by electron microprobe which is unable to measure valence states, and most significantly $\text{Fe}^{3+}/\text{Fe}^{2+}$. Although there are several methods for estimating the Fe^{3+} content of ferromagnesian minerals, these are estimates which depend on the accuracy of the initial microprobe analysis. Many calibrated geothermobarometers that use Fe^{2+} (e.g. garnet – biotite) do not take into account the Fe^{3+} content of the mineral, and fail to specify if the calibration refers to total Fe or Fe^{2+} . Consequently, this may be considered a potentially large source of error in estimating P-T conditions.

4.1 METAMORPHISM

The establishment of the metamorphic conditions that the rock suite has been subjected is an important part of reconstructing the geological history. Unless the rock suite has been heavily retrograded, the P-T conditions that are determined refer to the peak metamorphic conditions, and/or cooling conditions, that the rocks have experienced. In the Haveri area, the principal rock types useful for determining the metamorphic conditions are the pelitic metasedimentary rocks of the Osara Formation and the amphibolitic metavolcanic rocks of the Haveri Formation.

4.1.1 Pelites

The mineral assemblages in pelitic metasedimentary rocks are particularly sensitive to changes in temperature and pressure, and are thus good monitors of metamorphic histories (Spear, 1993). The numerous phases and reactions that can be present in pelitic rocks, mean that it is usually possible to accurately constrain the P-T conditions, by a number of different techniques.

4.1.1.1 Petrogenetic grids

Over 90 % of the mass of pelitic schists are represented by the chemical system $\text{SiO}_2\text{-Al}_2\text{O}_3\text{-MgO-FeO-K}_2\text{O-H}_2\text{O}$, which will be referred to as the KFMASH system, which can itself be simplified for graphical purposes as comprising two sub-systems $\text{SiO}_2\text{-Al}_2\text{O}_3\text{-FeO-K}_2\text{O-H}_2\text{O}$ (KFASH) and $\text{SiO}_2\text{-Al}_2\text{O}_3\text{-MgO-K}_2\text{O-H}_2\text{O}$ (KMASH) (Spear, 1993). Like many published versions of petrogenetic grids for pelites, the grid published by Spear and Cheney (1989) takes account of Fe-Mg solid solution, and is based on experimental reaction determinations in the KFASH and KMASH subsystems, which are used to constrain the limits of reactions in the full KFMASH system.

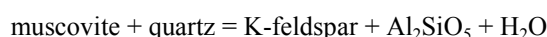
The pelitic mineral assemblage at Haveri is quite varied (Figure 2.18), and it is important to determine which minerals are related to the peak metamorphic event. Furthermore, it is equally important to determine which of these minerals are in equilibrium conditions. The most common minerals observed in sections from Haveri pelites consist of biotite, garnet and cordierite, with lesser amounts of quartz,

feldspar, apatite, spinel, andalusite, magnetite and ilmenite. However, although all of these minerals are present, they are not all ubiquitous to any particular sample. Of the minerals in the KFMASH system, biotite is the only mineral observed in all of the pelitic sections. In some lithologies, biotite is associated with almandine garnet (Fe rich), and in other lithologies biotite is associated with cordierite (Mg rich). It is only rarely that the three minerals are observed together (e.g sample P9-27). Likewise, the presence of andalusite is also sparse and, consequently, its presence in any particular sample may have to be inferred on the basis of nearby lithologies, which may be lacking in either garnet or cordierite.

This prograde pelitic assemblage observed at Haveri can be constrained on a P-T grid in the KFMASH subsystem (Figure 4.1) by the presence of biotite (Ann), andalusite (And), and porphyroblasts of almandine garnet (Alm) as well as the absence of staurolite (FeSt). In the KMASH subsystem (

Figure 4.2) the assemblage can be constrained on a P-T grid by the presence of phlogopite (Phl), andalusite (And) and porphyroblasts of cordierite (MgCrđ), and the absence of chlorite (MgChl).

It is possible that the assemblages may be further constrained by the absence of muscovite (Ms) controlled by the dehydration reaction:



However, the paucity of K-feldspar and Al_2SiO_5 occurring together in the pelitic assemblages, and in particular in the samples with almandine garnet, suggests that there is not enough evidence to use this reaction as a constraint.

In the full KFMASH system (Figure 4.3) the combined assemblage of biotite, garnet, and cordierite (with or without observable andalusite) is a further constraint on the P-T conditions, with the assemblage being controlled by the reaction:



Although this reaction normally only has three of the four minerals present, it is possible for all four to be present. This is due to extra components (such as Mn in the garnet) which makes the assemblage divariant rather than univariant (Spear, 1993). The P-T conditions as determined by the petrogenetic grids indicates that the regional metamorphism at Haveri is characterised by pressures below 2 kbar, and a broad temperature range of 550 °C to 750 °C

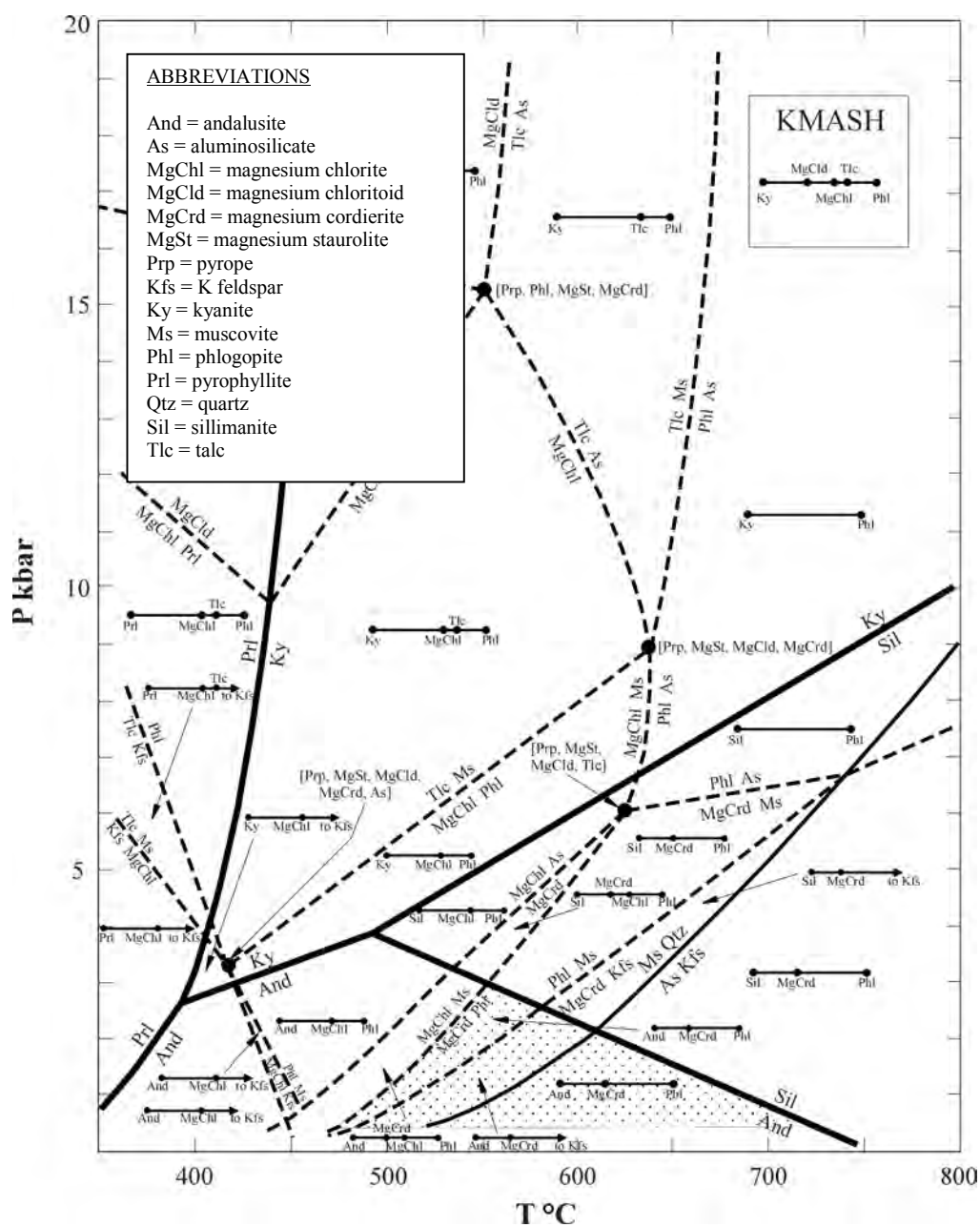


Figure 4.2: P-T grid for pelites in the KMASH subsystem. The shaded area represents the peak metamorphic assemblage observed at Haveri (after Spear and Cheney, 1989)

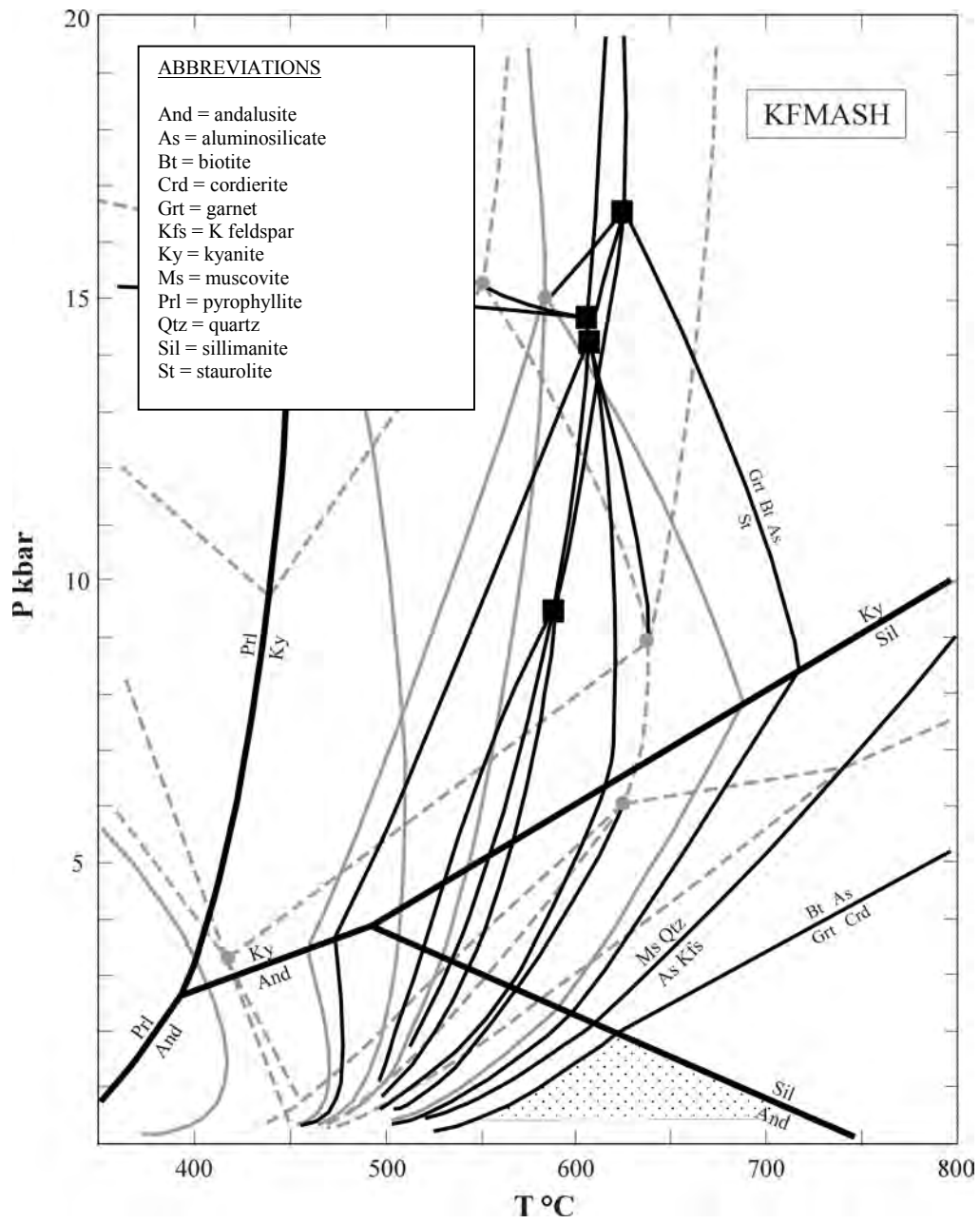


Figure 4.3: *P-T* grid for pelites in the KFMASH system. The shaded area represents the peak metamorphic assemblage observed at Haveri. The solid and the dashed grey lines are the reaction curves for the KFMASH system and the KFMASH system respectively (after Spear and Cheney, 1989).

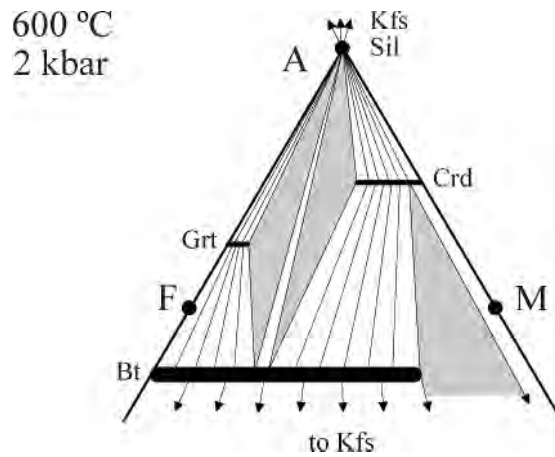


Figure 4.4: AFM diagram for pelites at 2 kbar and 600 °C (after Spear and Cheney, 1989).

The mineral assemblage observed in the pelites is controlled by a combination of bulk composition, pressure and temperature. Figure 4.4 shows an AFM diagram for pelites at 600 °C and 2 kbar pressure. It is apparent from this diagram, that at these P-T conditions, the assemblages are likely to be either biotite + garnet +/- Al_2SiO_5 , or biotite + cordierite +/- Al_2SiO_5 . At these conditions, an assemblage biotite + cordierite + garnet cannot exist in equilibrium. Compositions of pelites from Haveri (Table 4.1) show that a sample with both garnet and biotite (P5-16) is more Fe-rich than a sample with biotite and cordierite (P6-43), and is thus in agreement with the AFM diagram in Figure 4.4. In addition, it is apparent from Figure 4.4 that the presence of Al_2SiO_5 is entirely dependent on the bulk composition of the rock. The Mg^* 's of coexisting cordierite and biotite are too high in sample P6-43 to permit the three-phase assemblage of biotite + cordierite + Al_2SiO_5 , confirming the observed lack of andalusite, and the presence of a two-phase assemblage of biotite + cordierite (the garnet is discussed below). Sample P9-27, however, has more Fe-rich cordierite, plotting within (but close to the edge of) the three-phase field of coexisting biotite + cordierite + Al_2SiO_5 , thus accounting for the presence of all three minerals. The relative paucity of observed andalusite in sample P9-27 most likely reflects a low-Al bulk composition within this field.

Sample	P5-16	P6-43	P9-27
Assemblage	Gt + Bt	Crd + Bt (+ Gt)	Crd + Bt + And
Mg^*_{Gt}	12.3 – 17.5	16.0 – 24.3	
Mg^*_{Bt}	49.2 – 53.3	59.6 – 60.7	
Mg^*_{Crd}		74.9 – 75.5	59.8 – 61.0

Table 4.1: The range of values for magnesium numbers ($\text{Mg}^* = 100 \times (\text{Mg}/\text{Mg} + \text{Fe}^{2+})$) of various pelitic minerals from different pelitic assemblages at Haveri. Analyses by electron microprobe.

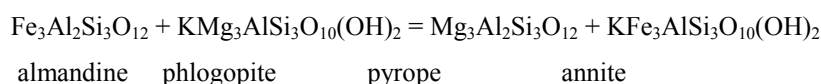
The AFM diagram (Figure 4.4) explains why garnet and cordierite (with biotite) have rarely been observed together at Haveri. In the one sample where they have been observed together (P6-43), there

is only one garnet crystal in a biotite + cordierite assemblage. In this situation, where the garnet appears to be texturally in equilibrium with the biotite and cordierite, it is necessary to consider alternative explanations, such as extra components in the garnet, or a heterogeneous sample, with the garnet occurring in a more Fe-rich portion of the sample.

The AFM diagram in Figure 4.4 is a reasonable representation of metamorphic conditions at Haveri, and accurately explains the observed assemblages and the general lack of observed Al_2SiO_5 within the pelitic lithologies. However, one discrepancy that is apparent, is that the composition of the biotites coexisting with the garnets (Table 4.1) is more Mg-rich than expected from Figure 4.4. Possible reasons for this are: (1) the P-T conditions at Haveri are not 600 °C and 2 kbar, and that the actual P-T conditions (higher P or lower T) cause a shift in the fields to more Mg-rich compositions. The presence of andalusite rather than sillimanite would also suggest this. (2) the garnet and the biotite may not represent an equilibrium assemblage. The more Mg-rich nature of the biotite compositions may reflect a re-equilibration at lower temperatures. In this case, andalusite may represent a retrograde replacement of sillimanite.

4.1.1.2 Garnet – biotite exchange thermometer

A geothermometer commonly applied to pelites is the biotite-garnet exchange thermometer which is based on the temperature dependent exchange of Mg and Fe between coexisting biotite and garnet. With increasing temperature, biotite becomes more Fe-rich and garnet more Mg-rich as the following continuous exchange reaction occurs:



Numerous calibrations of this geothermometer have been proposed: Thompson (1976), Goldman and Albee (1977), Ferry and Spear (1978), Hodges and Spear (1982), Perchuk and Lavrent'eva (1983), Ganguly and Saxena (1984, 1985), Indares and Martignole (1985), Hackler and Wood (1989), Williams and Grambling (1990), and Bhattacharya et al. (1992). These calibrations vary in the assumptions of how Fe and Mg substitute into the garnet and biotite, as well as how they have been calibrated to certain P-T conditions. Table 4.2 shows the results of a number of different garnet – biotite calibrations for several pelitic samples from Haveri calculated at 1.5 kbar. Garnet and biotite analyses were averaged from each sample to minimise errors associated with the analyses themselves. In addition, as garnet has been shown to be compositionally zoned (Section 1.1.1.1), the geothermometric calculations for sample P6-43 were made separately using an average of all of the garnet analyses as well as averages of garnet analyses from the core and from the rim.

<i>Analytical data</i>		<i>P5-16</i>	<i>P6-43</i>		
		All Garnets (n=14)	All Garnets (n=25)	Garnet cores (n=15)	Garnet rims (n=8)
Average Fe _{Gt} ± 1 Std.Dev.		4.40 ± 0.09	4.24 ± 0.16	4.15 ± 0.12	4.41 ± 0.10
Average Mg _{Gt} ± 1 Std.Dev.		0.87 ± 0.07	1.12 ± 0.14	1.20 ± 0.07	0.96 ± 0.09
		All Biotite (n=12)	All Biotite (n=5)		
Average Fe _{Bi} ± 1 Std.Dev.		2.42 ± 0.11	2.02 ± 0.03		
Average Mg _{Bi} ± 1 Std.Dev		2.61 ± 0.03	3.05 ± 0.03		
<i>Calibration</i>	<i>Error</i>	<i>P5-16</i>	<i>P6-43</i>		
Ferry and Spear (1978)	± 50 °C	578 °C	561 °C	591 °C	500 °C
Indares and Martignole (1985) – Model B	± 50 °C	571 °C	546 °C	576 °C	486 °C
Bhattacharya et al. (1992) – G & S Model	± 50 °C	602 °C	603 °C	618 °C	572 °C
Bhattacharya et al. (1992) – H & W Model	± 50 °C	596 °C	591 °C	607 °C	557 °C

Table 4.2: Peak metamorphic temperatures for pelitic samples from Haveri, calculated at 1.5 kbar pressure using various calibrations of the garnet-biotite geothermometer.

The Ferry and Spear (1978) experimentally calibrated the partitioning of Fe and Mg between garnet and biotite and assumed ideal mixing in garnet – biotite solid solutions. Most subsequent workers have demonstrated that Fe and Mg mixing in garnet and biotite is non-ideal, and that Fe-Mg partitioning may be affected by elevated levels of Mn and Ca in garnet and Ti in biotite. Indares and Martignole (1985) combined the experimental data of Ferry and Spear (1978) with data from natural garnet-biotite occurrences to derive an empirical thermometer that corrected for dilutants in both biotite and garnet. Bhattacharya et al. (1992) used the activity models for non-ideal mixing in garnet proposed by Ganguly and Saxena (1984, 1985) and Hackler and Wood (1989), combined with thermochemical data to produce two garnet-biotite calibrations, within the temperature range 575 – 950 °C, whilst still recognising the limitations provided by more complex substitutions in both garnet and biotite.

It is apparent from Table 4.2 that the zonation in garnet compositions does have an affect on the calculated temperatures. The Fe²⁺ enrichment (and Mg²⁺ depletion) in the rims of the garnets relative to the core produces a calculated temperature which, depending on the calibration used, is 46 – 91 °C lower than the temperatures calculated from the cores alone. Not surprisingly, the average of all the analyses produces an intermediate calculated temperature biased towards the higher proportion of core analyses. The implications of this are that the cores most likely reflect the peak metamorphic conditions, whereas the lower temperatures from the rims indicate there has been some re-equilibration of the Fe and Mg along the contact of the grains as the rock cooled. It is not possible to see if the same pattern exists in the biotite, due to difficulties in obtaining acceptable microprobe analyses from biotites.

The range of P-T conditions calculated by the various garnet-biotite thermometers is consistent with the P-T conditions determined using the petrogenetic grid (Figure 4.5) and suggests that the peak metamorphic temperatures (as calculated from pelitic assemblages) are in the range of 570 to 620 °C at pressures of less than 2 kbar. The implications of the low-P high-T metamorphic conditions will be discussed further in Section 4.5

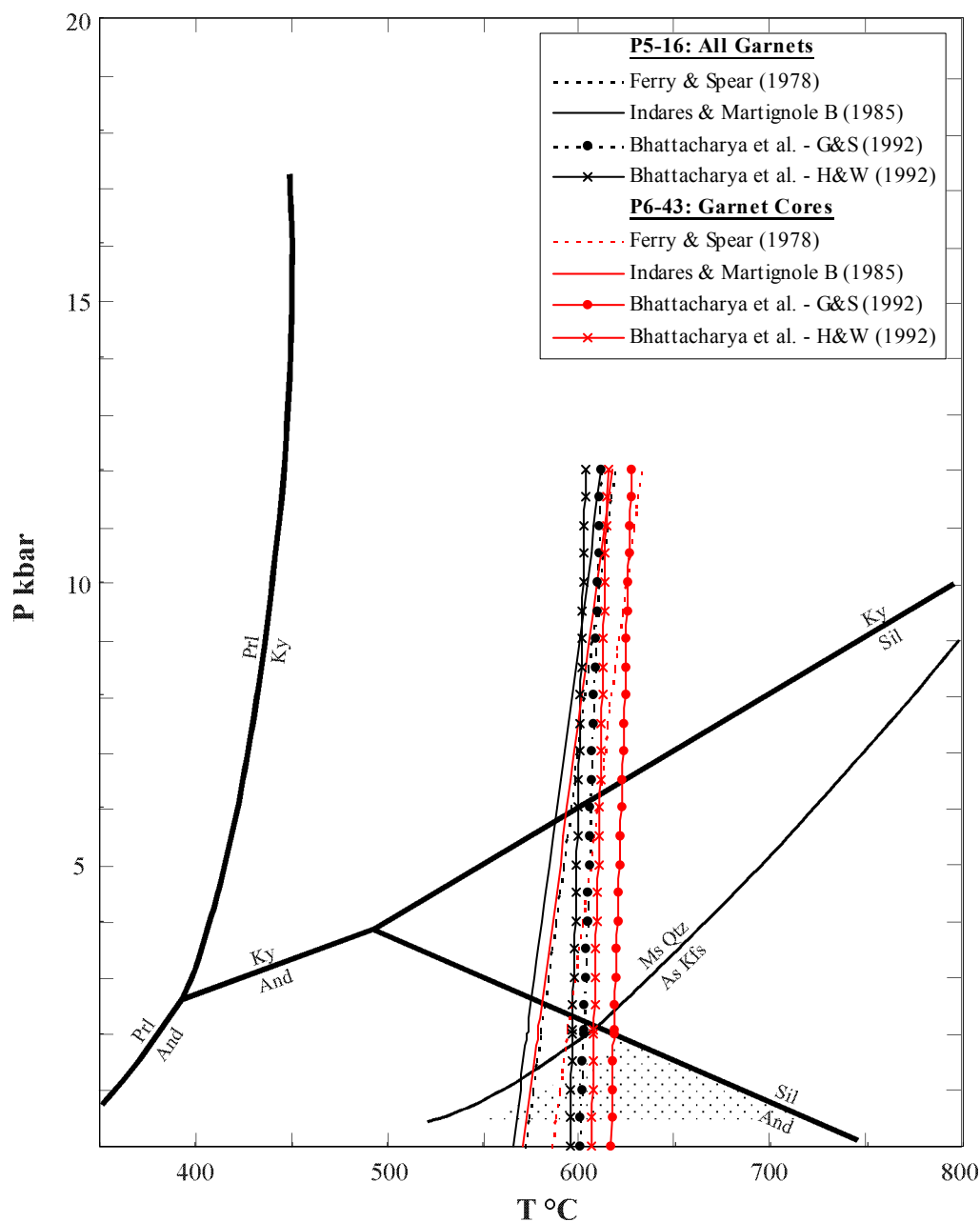


Figure 4.5: P-T diagram displaying the garnet-biotite P-T equilibrium curves calculated from the various calibrations for samples P5-16 (all garnets) and P6-43 (garnet cores). Also shown is the calculated P-T region from the petrogenetic grids.

4.1.1.3 fO_2 in pelites during peak metamorphism

In addition to indicating P-T conditions of metamorphism, mineral equilibria can be useful indicators of the nature and composition of the metamorphic fluids. Minerals that contain Fe will have a stability field that is a function of the fO_2 of the system. Minerals that contain mostly ferrous iron (Fe^{2+}) will have a larger thermal stability at low values of fO_2 , whereas minerals that contain mostly ferric iron (Fe^{3+}) will have a larger thermal stability at high values of fO_2 (Spear, 1993). Figure 4.6 shows $\log fO_2$ versus T diagrams for almandine and annite. Almandine ($Fe_3Al_2Si_3O_{12}$) contains little or no ferric iron, and it can be seen that the thermal stability is limited to fO_2 conditions near the quartz-fayalite-magnetite (QFM) buffer. Annite ($KFe_3AlSi_3O_{10}(OH)_2$) also nominally contains all ferrous iron, but invariably contains ferric iron. The upper thermal stability of annite therefore ranges to a higher value of fO_2 and it can be stable in the hematite field, albeit at low temperatures (Spear, 1993).

The coexistence of almandine and annite in the pelitic rocks at Haveri enables a reduced field of $\log fO_2$ versus T to be plotted. This can be further constrained by the presence or absence of magnetite, being buffered by the QFM buffer (Figure 4.6c). The majority of pelites are lacking in magnetite, and are thus below the QFM buffer. The assemblage almandine-biotite in these pelites thus defines a stability field with a range of $\log fO_2$ values of -17.0 to -28.0 , and temperatures between 560°C and 820°C . Using an estimated temperature of 600°C (based on the range of temperatures in Figure 4.5), the range of $\log fO_2$ values is approximately between -21.0 and -26.0 . The metasedimentary lithologies that have been observed with coexisting magnetite are however lacking in garnet and, thus, fall above the QFM buffer (Figure 4.6a). Therefore, these magnetite-bearing metasedimentary rocks likely represent conditions of higher oxygen fugacity, or perhaps, regions of lower temperatures of formation.

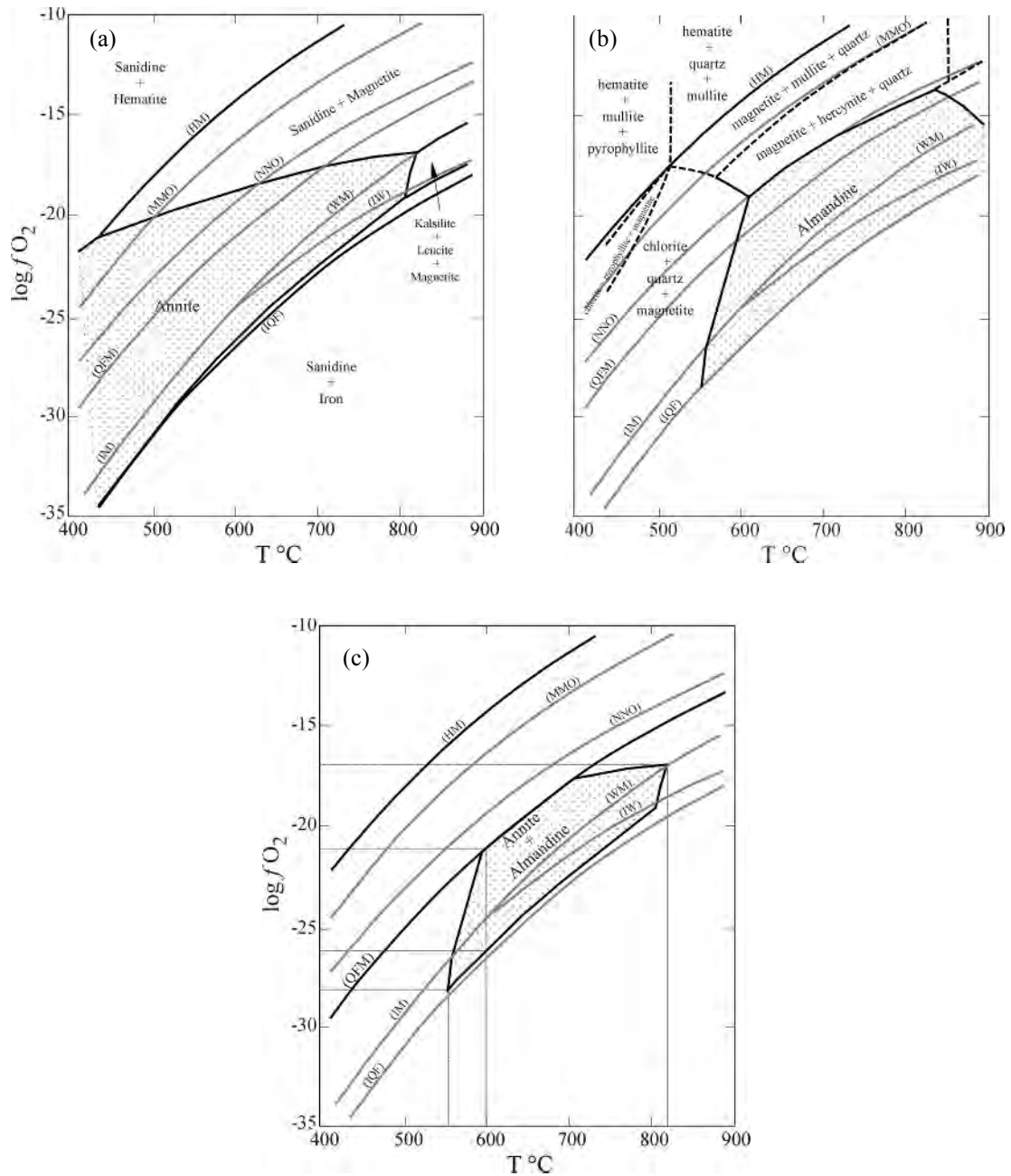


Figure 4.6: Plots of $\log (fO_2)$ versus temperature showing the effect of oxygen fugacity on the stability of a) annite; b) almandine (after Spear, 1993); and c) the assemblage annite-almandine-magnetite.

4.1.1.4 **Discussion**

Pelitic assemblages typically offer numerous reactions which enable reasonably accurate calculations of metamorphic P-T conditions. The Haveri samples are principally composed of the silicates biotite, garnet and/or cordierite, and andalusite. This assemblage allows close constraints to be put on the peak metamorphic P-T conditions using petrogenetic grids. The use of geothermobarometric calculations is less accurate due to the reliance on only the garnet-biotite thermometer. A number of other calibrated reactions are available for use in pelitic rocks, which typically put closer constraints on the calculated P-T conditions. The garnet-cordierite and cordierite-biotite exchange reactions (Hensen and Green, 1973; Holdaway and Lee, 1977) are potentially useful thermobarometers relying on Fe-Mg exchange. However, they are inapplicable to the Haveri assemblage due to the higher P-T limits and assemblages for which they have been calibrated. The net transfer reaction garnet-plagioclase-quartz- Al_2SiO_5 (GASP) is a commonly used thermobarometer, with particular sensitivity to pressure. Many calibrations have been calculated for this reaction using different activity models for garnet and plagioclase (Ghent, 1976; Ghent et al., 1979; Newton and Haselton, 1981; Hodges and Spear, 1982; Ganguly and Saxena, 1984; Hodges and Royden, 1984; Koizol and Newton, 1988; Powell and Holland, 1988; Koizol, 1989; Spear, 1993). Unfortunately, pelitic samples from Haveri containing garnet are devoid of coexisting plagioclase, and the range of plagioclase compositions from other samples (Figure 3.36) is too great to justify the inference and assumption that they represent the plagioclase compositions in the garnet-bearing pelites.

4.1.2 **Amphibolites**

The metamorphism of mafic rocks is not characterised by a large number of diagnostic mineral assemblages as in pelites, but by a limited number of mineral assemblages with the minerals (especially amphibole) displaying a wide range of composition (Spear, 1993). Consequently, P-T diagrams are much simpler, which led to the development of the concept of metamorphic facies (Figure 4.7), with the facies being separated by gradational boundaries signifying the major reactions.

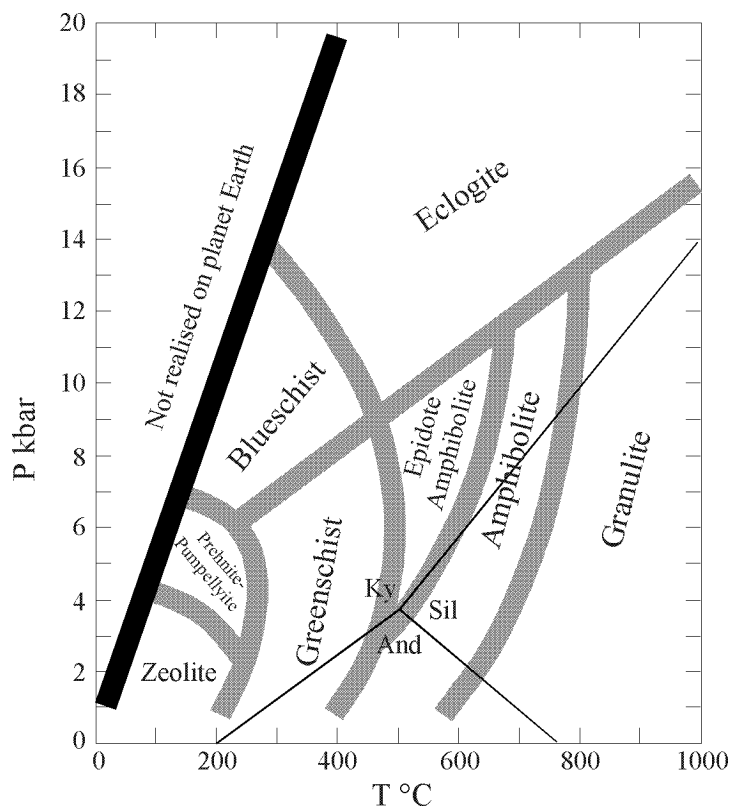


Figure 4.7: *P-T diagram showing the principal metamorphic facies and the Al_2SiO_5 triple point (after Holdaway, 1971; Spear, 1993)*

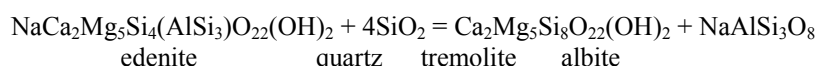
The mineral assemblage of the Haveri amphibolites consists of amphibole, plagioclase, quartz, sphene, epidote, magnetite and ilmenite. This assemblage, as defined by the absence of chlorite and the presence of epidote, indicates that the Haveri amphibolites are in the epidote amphibolite facies. This is clearly at odds with the P-T conditions determined from the pelites. Although the absence of epidote typically marks the amphibolite facies, epidote is still stable in the amphibolite facies in rocks with Ca-rich bulk compositions (e.g. calc-silicate rocks) (Spear, 1993). However, the Haveri amphibolites cannot be considered as calc-silicate rocks. Also, epidote is not a common constituent (<2 %) of the mafic assemblage, and is recognised as a distal alteration phase in both the amphibolitisation and the calcic-skarn type alteration. It must therefore be considered that epidote is not part of the metamorphic equilibrium assemblage.

Studies on the compositions of plagioclase with increasing temperature in basaltic rocks show that at 7 kbar, plagioclase is $An_{3.5}$ in the greenschist facies up through the epidote amphibolite facies and increases to nearly An_{30} in the amphibolite facies as epidote reacts out of the assemblage (Apted and Liou, 1983). At the lower pressure of 2 kbar, there is an even greater increase in the An content of plagioclase, from An_5 in the greenschist facies (475 °C) to An_{45} in the amphibolite facies (550 °C) (Liou et al., 1974). The rapid increase in An content of plagioclase at low pressure compared with intermediate pressure is a general feature in the metamorphism of mafic rocks and reflects the relative stabilities of anorthite versus epidote. That is, epidote reacts out quickly at low pressure leaving An-

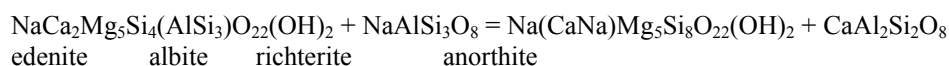
rich plagioclase, whereas at intermediate pressure epidote reacts out slowly and plagioclase remains An-poor to much higher temperature (Spear, 1993). The measured An contents of plagioclase from Haveri amphibolites are approximately An₂₀₋₅₅, suggesting that epidote has reacted out of the mafic rocks, and that therefore the epidote observed in thin section is not in equilibrium with the rest of the metamorphic assemblage that has been metamorphosed to the amphibolite facies.

4.1.2.1 Hornblende – plagioclase thermometer

An amphibole-plagioclase thermometer was developed by Blundy and Holland (1990), based on the reaction:



After much comment on the simplicity and limitations of this thermometer (Blundy and Holland, 1992a,b; Hammarstrom and Zen, 1992; Poli and Schmidt, 1992; Rutherford and Johnson, 1992), Holland and Blundy (1994) formulated two new thermometers based on more complex (non-ideal) mixing models for amphibole and plagioclase. These thermometers were based on the original reaction for thermometer A (for assemblages with quartz), as well as a new reaction for thermometer B (for assemblages with or without quartz):



The application of these thermometers requires the estimation of Fe³⁺ in amphibole. Although the procedure is similar to that proposed by Schumacher (in IMA, 1997), there are some minor discrepancies, and ferric iron has been calculated according to the method outlined in Holland and Blundy (1994, corrections in Dale et al., 2000). Both the amphibole and the plagioclase analyses were averaged from each section.

The results of the amphibole-plagioclase thermometers for the Haveri amphibolites are shown in Table 4.3 (calculated at 1.5 kbar pressure), and displayed on a P-T diagram in Figure 4.8. It is apparent from the data that results from thermometer A (edenite-tremolite) are consistently higher than those from thermometer B (edenite-richterite). This discrepancy was recognised by Holland and Blundy (1994), who ascribed it to the samples being silica-undersaturated, in which case the activity of silica is less than the assumed value of 1 ($a_{\text{SiO}_2} < 1$) in thermometer A, thus producing anomalously high temperatures. The results for sample R7-296 appear to be anomalously high, and are likely to be caused either by minerals that are not in equilibrium with each other, or by the minerals having been affected by hydrothermal alteration, and/or retrograde effects.

<i>Sample</i>	<i>Hornblende analyses</i>	<i>Plagioclase analyses</i>	<i>Error</i>	<i>Therm. (A)</i>	<i>Therm. (B)</i>
R8-219 All	2	2	± 40 °C	666 °C	627 °C
R8-216b	Pair 1	3	± 40 °C	704 °C	608 °C
	Pair 2	3	± 40 °C	684 °C	644 °C
	All	12	± 40 °C	708 °C	647 °C
R6-284	Pair 1	1	± 40 °C	678 °C	613 °C
	Pair 2	2	± 40 °C	650 °C	595 °C
	Pair 3	2	± 40 °C	644 °C	572 °C
	All	6	± 40 °C	664 °C	593 °C
R7-296	Pair 1	3	± 40 °C	739 °C	673 °C
	Pair 2	2	± 40 °C	732 °C	669 °C
	Pair 3	3	± 40 °C	790 °C	714 °C
	All	18	± 40 °C	729 °C	683 °C

Table 4.3: Peak metamorphic temperatures for amphibolites from the Haveri Formation, calculated at 1.5 kbar pressure using the edenite-tremolite geothermometer (Therm. A) and the edenite-richterite geothermometer (Therm. B) of Holland and Blundy (1994). Calculations have been made using metamorphic hornblende-plagioclase pairs, with mineral grains in contact. Also supplied are calculations using averages of all analysed plagioclases and metamorphic amphiboles from each sample.

The amphibole-plagioclase geothermometry indicates peak metamorphic conditions similar to those determined for the pelites, and corresponds to an approximate temperature of 620 °C, and pressures of 1 – 1.5 kbar. Although these P-T conditions fall within the granulite facies field of Figure 4.8, the field boundaries represent gradational zones, and are imprecisely determined. The absence of pyroxene in the mineral assemblage means that the rock does not fall within the granulite facies, (defined where amphibole breaks down to pyroxene), and can be classified to upper amphibolite facies (Spear, 1993). It is therefore likely that the amphibole-plagioclase temperatures are slight overestimates (although they are within error) of the peak metamorphic conditions.

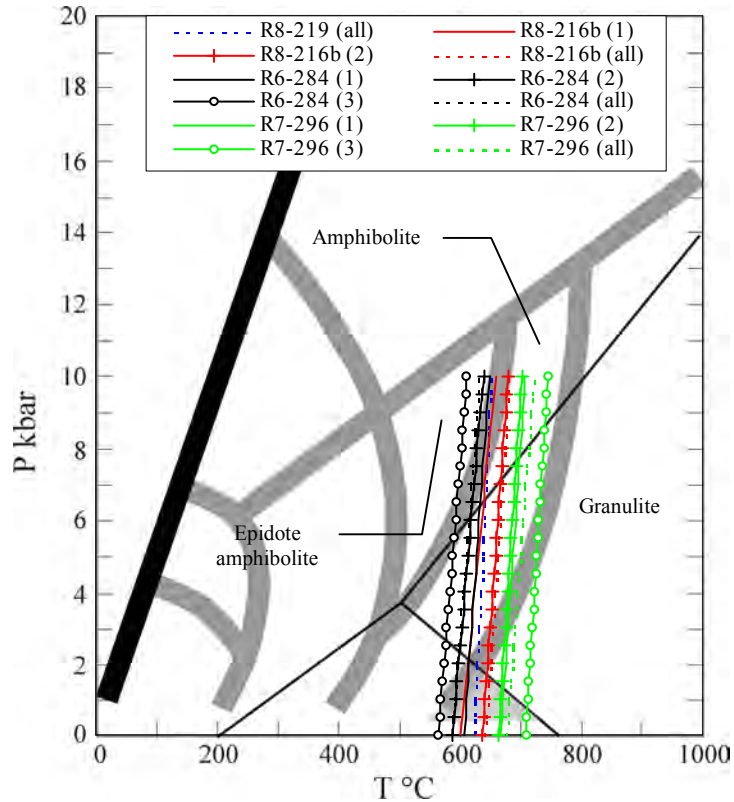


Figure 4.8: *P-T* diagram displaying the hornblende-plagioclase *P-T* equilibrium curves calculated using the edenite-richterite thermometer (thermometer B) of Holland and Blundy (1994). Also shown is the calculated *P-T* region from the petrogenetic grids for the pelites, and the ranges of the metamorphic facies (after Holdaway, 1971, in Spear, 1993).

4.1.2.2 fO_2 in amphibolites during peak metamorphism

As with the pelites it is possible to constrain the conditions of the metamorphic fluids in the amphibolites using mineral equilibria and $\log fO_2$ versus temperature diagrams. The equilibrium mineral assemblage in the amphibolites consists of amphibole-plagioclase-quartz-magnetite-ilmenite-sphene. This assemblage is buffered by the hematite-magnetite (HM) buffer and can be constrained on the $\log fO_2$ versus temperature diagram of Spear (1981, 1993) at 1 kbar pressure by the presence of sphene (Figure 4.9) to a thermal stability field with a range of $\log fO_2$ values of approximately -12.4 at 670°C to less than -24.0 at temperatures below 500°C . At the calculated metamorphic temperature of 620°C , the range of possible values of $\log fO_2$ is reduced to between approximately -13.8 and -16.2 . At the metamorphic temperatures calculated for the pelites the range of possible values of $\log fO_2$ is between approximately -14.5 and -17.5 .

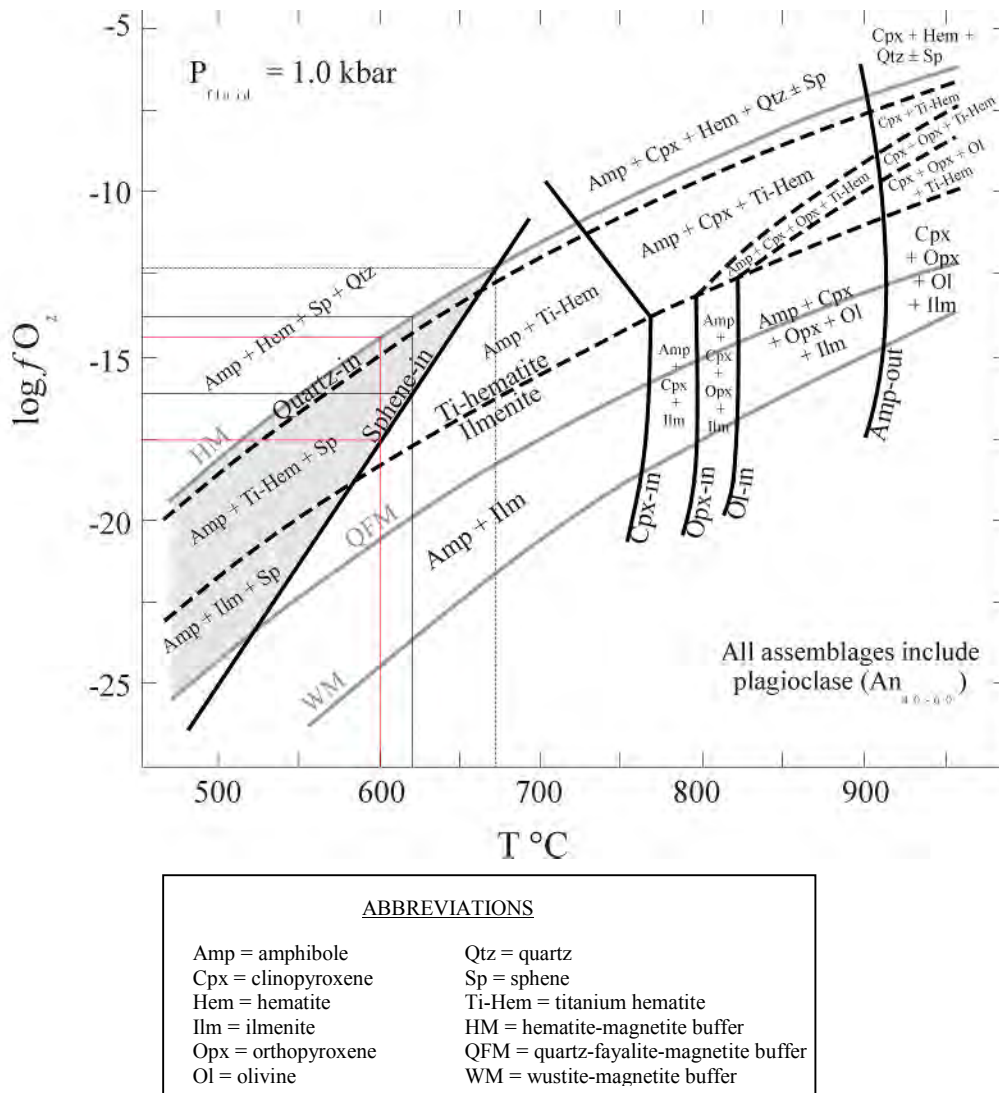


Figure 4.9: *Log fO₂ versus temperature plot showing the effect of oxygen fugacity on the amphibolite to granulite facies transition in a typical basalt (after Spear, 1981).*

4.1.3 Discussion

The P - T - X conditions for peak metamorphism have been fairly well constrained in both the amphibolites of the Haveri Formation and in the pelitic rocks of the upper Haveri Formation and the Osara Formation. A number of different methods have been used to estimate the P - T conditions, and although there is a certain amount of deviation between the methods used, they are in broad agreement. The range of metamorphic temperatures in the amphibolites calculated at 1.5 kbar, and using thermometer B of Holland and Blundy's (1994) amphibole-plagioclase thermometer, are consistently higher than the temperatures calculated using the garnet-biotite thermometers in the pelites. However, the results do overlap within calculated error (Figure 4.10), and both fall within the P - T ranges estimated using the petrogenetic grids.

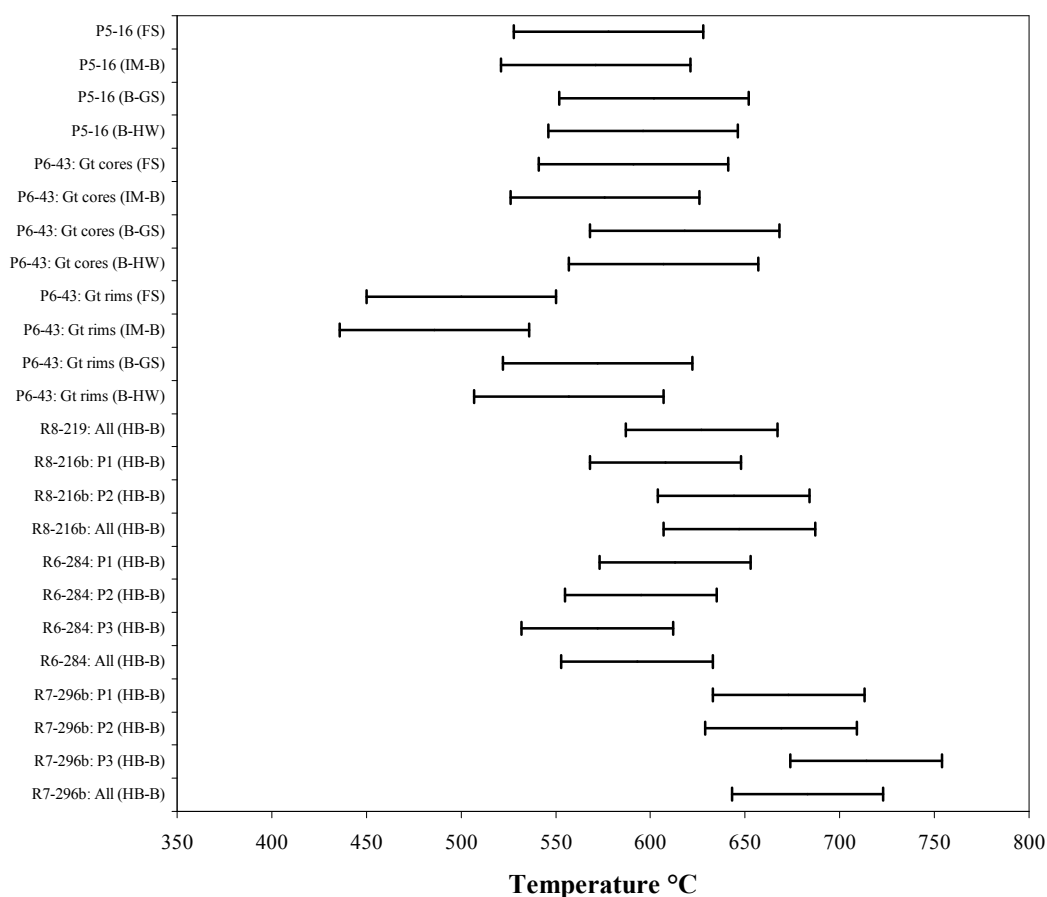


Figure 4.10: The range of calculated temperatures of both amphibolites and pelites at 1.5 kbar for the various geothermometers and calibrations. (FS) Ferry and Spear (1978); (IM-B) Indares and Martignole (1985); (B-GS) Bhattacharya et al. (1992, using Ganguly and Saxena, 1984); (B-HW) Bhattacharya et al. (1992, using Hackler and Wood, 1989); (HB-B) Holland and Blundy (1994) thermometer B.

On the basis of geothermobarometry, mineral equilibria and petrogenetic grids, it is possible to estimate peak metamorphic conditions to a P-T field of 550 – 650 °C, at pressures of less than 2 kbar. The amphibolite-granulite facies transition (Spear, 1993) constrains this to a maximum of approximately 600 °C and 1.5 kbar pressure. The implications of these metamorphic conditions will be discussed in Section 4.5. The oxygen fugacity of the metamorphic fluids can be constrained at 600 °C to a log fO_2 value of approximately –21.0 to –26.0 for the pelitic assemblages and a log fO_2 value of approximately –14.5 to –17.5 for the mafic metavolcanic rocks. The contrast between the more oxidized mafic metavolcanic rocks and the more reduced metasedimentary rocks has important implications for mineralisation, and will be discussed in detail in Section **Error! Reference source not found.**

4.2 AMPHIBOLITISATION

Amphibolitisation is the dominant form of alteration observed at Haveri, and is characterised by the silicate rock assemblage becoming dominated by calcic amphiboles, and displaying increasing Fe-enrichment with increasing degrees of alteration. Also associated with amphibolitisation (although more distal) is deamphibolitisation and albitisation of the volcanic host-rock. Amphibolitisation is closely associated with both sulphide and oxide mineralisation, and amphibolitised rocks are also the most important host for gold mineralisation. Pyrrhotite is the dominant sulphide mineral with lesser chalcopyrite and minor pyrite, and occurs as massive zones, veins and disseminations. Magnetite is also an important hydrothermal phase, forming intergrowths with pyrrhotite and chalcopyrite as well as forming patches of massive magnetite. Fine amphibole veinlets commonly pass laterally into and out of quartz $\pm$ plagioclase veinlets, suggesting that at least some of the quartz veinlets are genetically related to the alteration. This is further supported by the presence of amphibole in the quartz veins which have similar compositions to the amphibolitisation.

Although plagioclase is part of the alteration assemblage (i.e. albitisation), it is most commonly discernible as an outer halo around amphibole veinlets, and thus reflects a different portion of the thermo-chemical gradient associated with the alteration. Due to the predominance of the amphibole in the more pervasive (and proximal) expressions of the alteration, coarse-grained plagioclase suitable for analysis by electron microprobe is rarely observed. Therefore in terms of P-T calculations, the hornblende-plagioclase thermometer is not applicable to the alteration phase. The predominance of amphibole, and the absence of other suitable minerals in equilibrium with amphibole (e.g. ferromagnesian minerals) means that there are no mineral-based geothermobarometers applicable to the alteration assemblage.

The use of stable isotopes and fluid inclusions are among the other methods available for geothermometry. Oxygen isotope thermometry has been applied to a sample of intense amphibolitisation (sample P7-82), containing patches of massive magnetite (Section 3.3.3.3). Using the mineral-pairs hornblende and magnetite, the 5 acceptable calibrations produce reasonably consistent temperatures with an average of 530 °C (standard deviation = 17 °C). This is the only direct measurement of temperatures from the amphibolitisation itself. However, two other temperature calculations were made on quartz vein samples associated with amphibolitisation (Section 3.3.3.4). One of these (sample R8-168a) was based on the ^{18}O compositions of quartz, magnetite and feldspar contained within the vein. The 15 calibrations produced a calculated mean temperature of 610 °C (standard deviation = 70 °C). The other calculation was based on the ^{18}O compositions of quartz in the vein and hornblende in the selvage of the vein. The two calibrations produced temperatures of 518 °C and 553 °C (average = 536 °C).

Despite checking a number of sections, no fluid inclusions were identified in the amphiboles themselves, and thus fluid inclusion studies were limited to the quartz veins associated with both amphibolitisation and gold mineralisation (Section 3.4). The fluid inclusion studies identified only a very few fluid inclusions that could potentially be of a primary nature, the majority being later-stage secondary inclusion trails. The primary-looking inclusions have homogenisation temperatures of approximately 136 – 172 °C; these are likely to reflect reequilibration to lower P-T conditions. The compositions of these inclusions suggest that they are < 10 wt% equivalent NaCl-H₂O inclusions with a low concentration of CO₂. The late-stage secondary trails contain a greater variety of inclusions, including primarily aqueous two-phase inclusions as well as some with solid phases indicating the presence of hypersaline fluids. As with the primary-looking inclusions, no CO₂-rich inclusions were observed, suggesting low concentrations of CO₂.

Reequilibration textures indicative of internal underpressure along a pathway of isobaric cooling were also observed, and support a model of contact metamorphism, with the intrusion of hot granitoids at shallow levels (low P) within the crust.

The whole-rock geochemistry and mineral chemistry of the amphibolitisation suggests that this alteration phase is characterised by an enrichment in calcium, sodium, iron (Fe²⁺ and Fe³⁺) and potassium. The K-enrichment appears to be a distal expression of the amphibolitisation (see Section 3.1.3 for discussion on the chemistry).

4.3 *CALCIC-SKARN*

The calcic-skarn alteration typically displays a marked mineral zonation away from a calcite-quartz vein. This is most commonly represented by the proximal-to-distal assemblage of garnet-pyroxene-epidote-amphibole. However, calcic skarns typically have a high variance of mineral assemblages which reflects the varying physio-chemical conditions encountered with both distance from the vein, and the temporal evolution of the skarn system. Consequently, as local conditions vary, other minerals such as scapolite are occasionally present in the proximal assemblage.

4.3.1 Phase equilibria

As skarns are chemically open (metasomatic) systems, there are potentially many physio-chemical variables in addition to pressure and temperature that can be important in defining the mineralogy and mineral compositions. These variables include $X(\text{CO}_2)$, $f(\text{O}_2)$, $f(\text{S}_2)$, pH and activities of non-volatile aqueous species. Consequently the thermodynamic variance of skarn mineral assemblages is normally high, which commonly restricts definition of physio-chemical variables to upper or lower limits. Indeed, it is often not possible to determine specifically which variables are responsible for observed

zoning (Bowman, 1998a). Although in some mineralogically zoned skarns it is possible to place closer limits on the P - T conditions of skarn formation by phase equilibria alone, more usually the evaluation of P - T conditions should involve the application of phase equilibria to both the observed skarn assemblage as well as the contact metamorphosed rocks in the thermal aureole, as well as applications of fluid-inclusion and isotope geothermometry to the skarns (Bowman, 1998a).

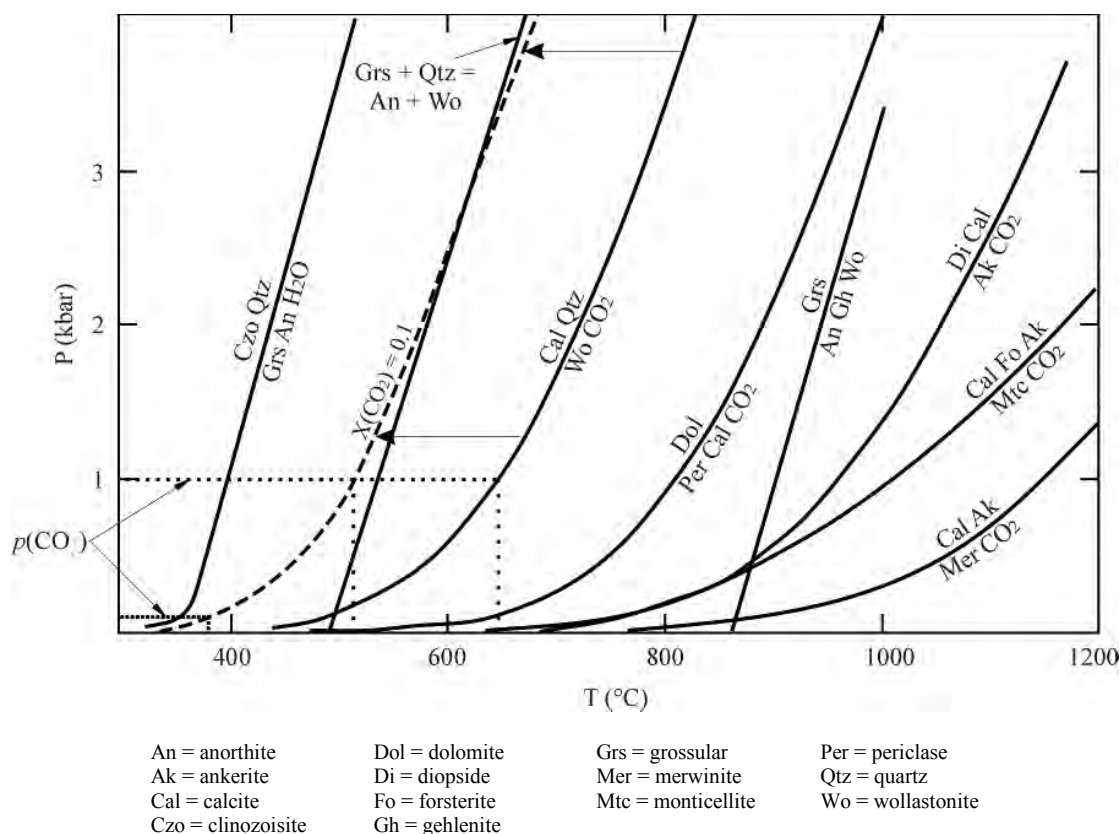
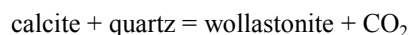


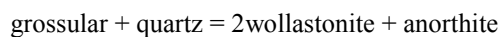
Figure 4.11: Pressure-temperature diagram with a variety of phase equilibria used for placing limits on the P - T conditions of skarn formation. Decarbonation equilibria are at the condition $P_T = p(\text{CO}_2)$ (solid lines) and at $X(\text{CO}_2) = 0.1$ (dashed line). As the pore fluids are diluted by H_2O , the $p(\text{CO}_2)$ decreases and the reaction is shifted to lower temperatures (after Bowman, 1998a).

The presence of wollastonite in a skarn mineral assemblage has been used for placing minimum T limits of skarn formation according to the equilibrium reaction:



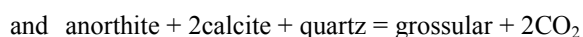
However, decreasing the $X(\text{CO}_2)$ of the skarn fluids (by greater dilution by H_2O) shifts the decarbonation equilibria to lower temperatures (Figure 4.11). This indicates that the above reaction provides well defined minimum T limits on wollastonite only where the slope of the reaction curve is steep (at pressures greater than 1 kbar) and where $X(\text{CO}_2)$ is well known. This demonstrates that in addition to P and T , the $X(\text{CO}_2)$ of a skarn fluid is an important variable in controlling skarn mineralogy and zoning.

It is perhaps possible to put an upper T limit for the conditions of skarn formation at Haveri by making a few assumptions. Wollastonite is absent from the calcic-skarn assemblages, but grossular garnet and quartz are present as proximal mineral phases. Therefore the absence of wollastonite can provide a maximum temperature limit according to the reaction



Furthermore, from the metamorphic studies, it is possible to place an upper limit for the pressure at 1.5 kbar. If we assume the fluid has $X(\text{CO}_2) = 1$ (i.e. no H_2O in the fluid), and $P_{\text{total}} = P_{\text{fluid}} = p(\text{CO}_2) = 1.5$ kbar then we can assume a maximum temperature of approximately 560 °C for the proximal calcic skarn fluids, although this is clearly an overestimate of the actual temperature of formation. At an $X(\text{CO}_2) = 0.1$, the maximum temperature would be approximately 515 °C. The presence of calcite and scapolite in the proximal assemblage of the calcic-skarn is indicative of a CO_2 -rich fluid, which would have an $X(\text{CO}_2)$ somewhere between the two examples used above. This corresponds to a formation temperature of between 515 and 560 °C. A better estimate of the $X(\text{CO}_2)$ of the fluids would more accurately constrain the P - T conditions of skarn formation.

The locations of skarn phase equilibria in P - T - X space are, however, also dependent on other variables such as mineral solid solutions and on the concentrations of dissolved salts in the H_2O - CO_2 fluid. The degree of solid solution in calc-silicate minerals, such as the dilution of the grossular component in garnet with andradite or the dilution of the clinozoisite component in epidote with pistacite or the hedenbergite component in pyroxenes with diopside, all serve to shift the equilibria limiting the stability of these minerals to higher T and $X(\text{CO}_2)$ values. Figure 4.12 shows the effect on the T - $X(\text{CO}_2)$ equilibria for the reactions :



at varying degrees of andradite-grossular substitution in garnet. As grossular becomes increasingly diluted by andradite, the field of garnet stability increases. Consequently, the presence of grossular garnet only provides clear evidence of low- $X(\text{CO}_2)$ fluids when the grossular component in garnet is greater than 90 % ($X_{\text{Grs}} > 0.90$) (Bowman, 1998a).

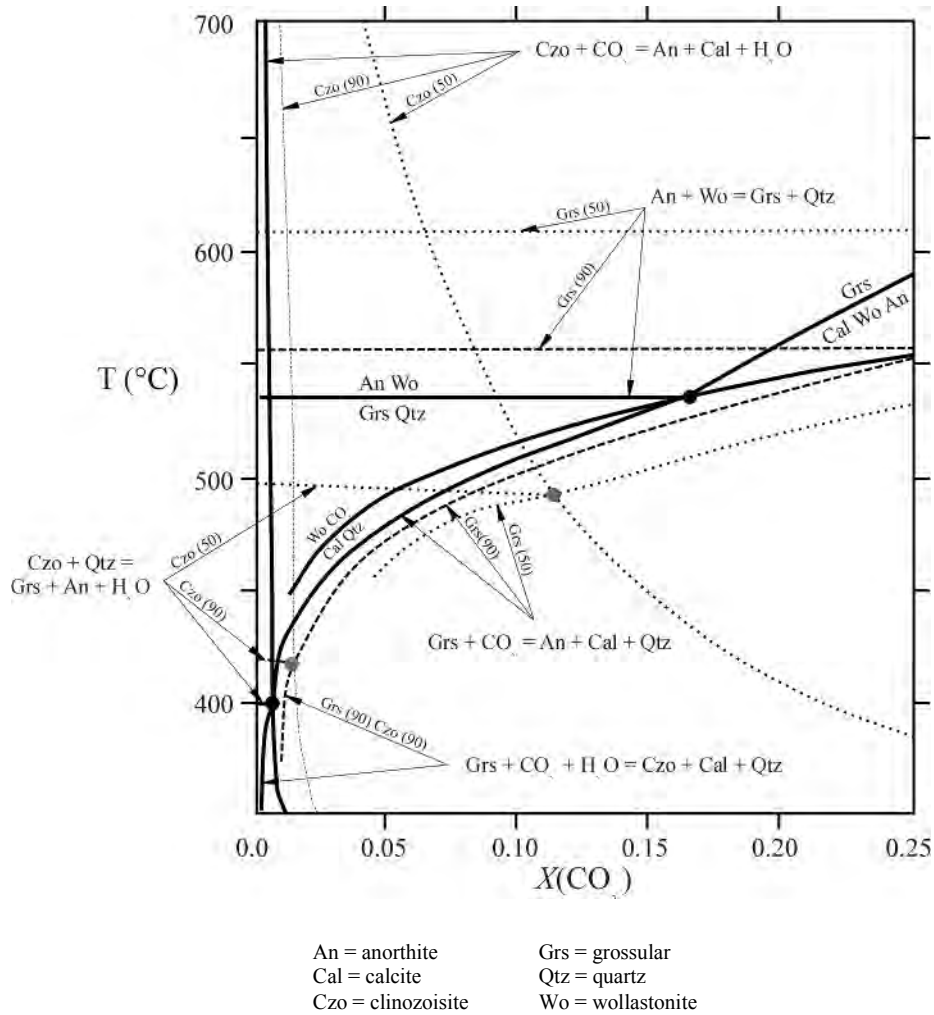


Figure 4.12: Water-rich [0.0 to 0.25 $X(\text{CO}_2)$] portion of an isobaric ($P_T = P_f = 1$ kbar) T - $X(\text{CO}_2)$ diagram illustrating the effects of solid solution (dilution of Grs with Adr, and Czo with Ps) on the phase equilibria in the system $\text{CaO-MgO-Al}_2\text{O}_3\text{-SiO}_2\text{-H}_2\text{O-CO}_2$ relevant to skarn mineralogy (after Bowman, 1998a)

The presence of dissolved salts, such as NaCl , and perhaps MgCl_2 and CaCl_2 , have both thermodynamic and physical effects on skarn-forming fluids. As the presence of dissolved salts increases the $a(\text{CO}_2)$ in $\text{H}_2\text{O-CO}_2$ fluids increases, and therefore the equilibrium boundaries of mixed-volatile phase equilibria shift to progressively higher T and/or lower $X(\text{CO}_2)$ (Bowman, 1998a).

Another important effect of increasing $a(\text{CO}_2)$ is to expand the immiscibility gap in the $\text{H}_2\text{O-CO}_2$ system (CO_2 -brine) to progressively higher P - T conditions, overlapping the P - T conditions of metamorphism and skarn formation. This expansion has several potential impacts for skarn forming processes (Bowman, 1998a). First, the development of phase immiscibility in skarn-forming fluids (by increasing salinity and/or $X(\text{CO}_2)$ or decreasing T) may have significant effects on the solubilities of both silicate gangue and ore minerals. Second, immiscibility will produce a low-salinity, low-density, CO_2 -rich vapour phase. The positive volume change accompanying separation of this CO_2 -rich vapour phase may cause hydrofracturing (carbofracturing) and brecciation.

The importance of $f(\text{CO}_2)$ in skarn mineralogy has been demonstrated above. However, the observed skarn mineralogy is also a function of oxygen and sulphur fugacities. Bowman (1998a), using reaction topologies on isobaric T - $f(\text{O}_2)$ diagrams and isothermal and isobaric $\log f(\text{O}_2)$ - $\log f(\text{S}_2)$ diagrams, again shows the importance of all the physio-chemical variables P , T , $X(\text{CO}_2)$, $f(\text{O}_2)$, $f(\text{S}_2)$, pH, mineral solid solutions and activities of non-volatile aqueous species on skarn mineralogy. It is therefore apparent that to constrain the P - T - X conditions of skarn formation, it is necessary to utilise other geological and geochemical techniques, such as the phase equilibria of the metamorphic rocks, fluid inclusion studies, stable isotope studies and mineral chemistry to better define and limit some of the variables, and so thereby define the positions of the phase equilibria in P - T - X space. Even with broad limits placed on the values of many of these variables, the actual determination of the locations of the equilibria in P - T - X space requires complex thermodynamic calculations beyond the scope of this study.

4.3.2 Sphalerite geothermobarometry

The FeS content of sphalerite was recognised as a potentially useful geobarometer by Barton and Toulmin (1966) and Scott and Barnes (1971), and calibrated by (among others) Scott (1973) and Lusk and Ford (1978). The FeS content of sphalerite is a function of temperature (T), pressure (P) and FeS activity (a_{FeS}). The a_{FeS} is not necessarily dependent on P and T , and for geobarometry must be buffered by divariant assemblages in the Fe-S system such as pyrite + pyrrhotite (Scott, 1976). The necessity of buffering the sphalerite by direct mutual contact with both pyrite and pyrrhotite was demonstrated by Scott (1976)

The calcic-skarn assemblage at Haveri includes sphalerite, pyrrhotite and pyrite. However, sphalerite is an uncommon mineral, and has not been observed in mutual contact with pyrrhotite *and* pyrite. Indeed, all of the analysed sphalerites are in contact with only pyrrhotite, and do not even have pyrite in the same sections. The FeS content of the sphalerites (Figure 3.38) range between 10.5 and 14.8 mol% FeS, which corresponds to pressures of 4.8 and 9.6 kbar using the thermometer of Lusk and Ford (1978). These pressures are much greater than the expected pressures defined by the peak metamorphism (and amphibolitisation). This overestimate of the pressure was explained by Scott (1976) who showed that sphalerite coexisting with pyrite only contains much less FeS (and thus gives underestimates of the pressure) and that sphalerite coexisting with pyrrhotite only contains much more FeS, and thus gives overestimates of the pressure. Consequently, unless sphalerite can be observed in direct mutual contact with both pyrite and pyrrhotite, sphalerite geobarometry will remain inapplicable to Haveri.

The FeS content of sphalerite in equilibrium with either pyrrhotite or pyrite can, however, provide an indication of both the $a(\text{S}_2)$ and the temperature of formation of the sphalerite, provided that there is an independent estimate of the pressure. Figure 4.13 shows a plot of the mole percent FeS in sphalerite at equilibrium with pyrite and/or pyrrhotite at 1 bar pressure (Vaughan and Craig, 1997). Sphalerites with an FeS content in the range of 10 – 15 mol% FeS, and in equilibrium with pyrrhotite, can only have

formed at temperatures in excess of 600 °C, and $\log a(\text{S}_2)$ values greater than -2 , at a pressure of 1 bar. However, with an increase in pressure, and at a constant temperature, the FeS content of sphalerite in equilibrium with iron sulphide assemblages decreases (Barton and Toulmin, 1966). The extent of this decrease is approximately 3 mol% FeS for a pressure increase from 0 kbar to 2.5 kbar, at 600 °C (Scott, 1976). Consequently, the lower border of the shaded field in Figure 4.13 will shift to lower temperatures and $a(\text{S}_2)$, resulting in a minimum temperature of formation less than 600 °C. This shift is partially offset by the actual FeS contents of the sphalerites being in the 10 – 15 mol% FeS range rather than at 20 mol% FeS. Unfortunately, without a more accurate and applicable calibration, it is not possible to state a minimum temperature for the formation of the sphalerite in the calcic skarn. However, it is apparent that the sphalerite in the calcic skarn, and thus the calcic skarn itself, are likely to have formed at high temperatures, with minimum formation temperatures likely to be in the region of 500 – 600 °C.

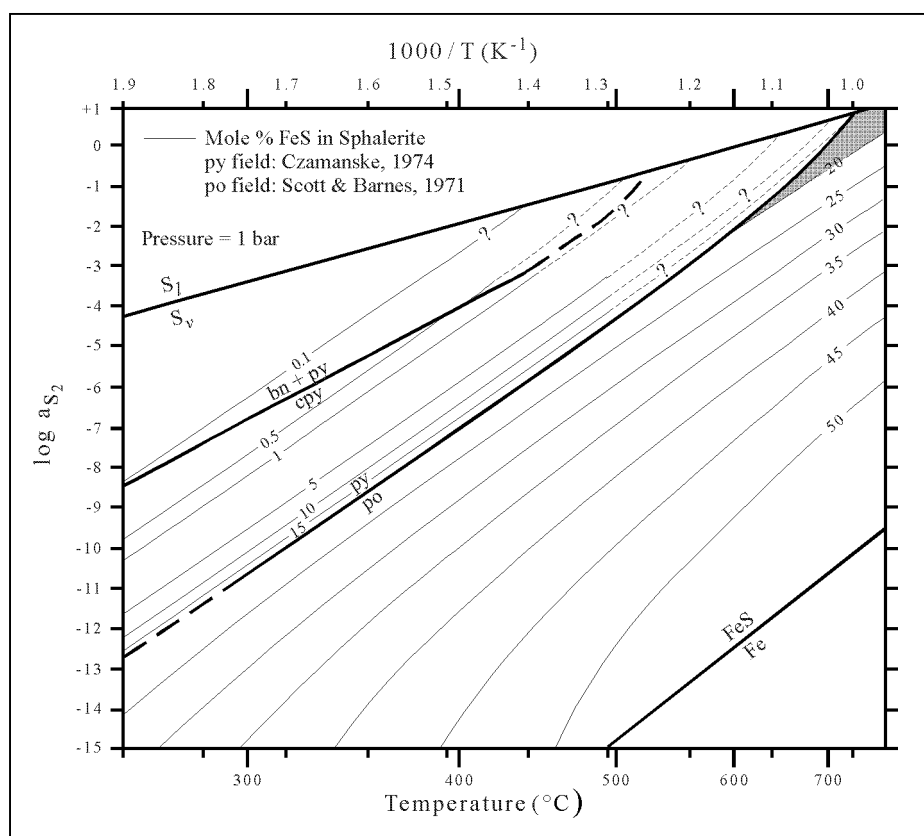
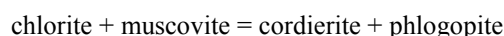


Figure 4.13: The mole percent FeS in sphalerite at equilibrium with pyrite and/or pyrrhotite versus T and $a(\text{S}_2)$. Abbreviations are: bn = bornite; py = pyrite; cpy = chalcopyrite; po = pyrrhotite; S_l = sulphur liquid; S_v = sulphur vapour (after Vaughan and Craig, 1997).

4.4 RETROGRADE CHLORITE

Chlorite is a common retrograde mineral within the mineral paragenesis at Haveri, occurring in all of the observed rock types. As well as occurring in late chlorite veinlets representing channelised fluid flow, it also occurs as a more pervasive retrograde replacement after earlier higher temperature minerals. Chlorite is commonly seen replacing biotite and amphibole, and in the metasedimentary rocks is seen as a common break-down product of cordierite. However, the wide stability range of chlorite on a petrogenetic grid, combined with difficulties in determining equilibrium minerals usually means that only very broad P-T conditions can be defined by the presence of chlorite. In the pelitic rocks of the Osara Formation, and assuming a decreasing P-T path since peak metamorphism (as suggested by the isobaric-cooling fluid inclusion decrepitation textures (Section 3.4.2)), a maximum temperature of approximately 500 °C at 1 kbar pressure can be placed on the stability of chlorite (Figure 4.2) (Spear, 1993). This is defined by the breakdown of cordierite in the KMAH system according to the reaction:



In rocks of basaltic composition, chlorite is stable throughout the greenschist facies (Figure 4.7), correlating to a maximum temperature of approximately 400 – 450 °C at 1 – 2 kbar pressure (Spear, 1993).

A number of chemical variables in chlorite have, however, been shown to be strongly temperature dependent (McDowell and Elders, 1980; Cathelineau and Nieva, 1985; Cathelineau, 1988), and therefore have the potential of being effective geothermometers. As temperature increases, there is an increase in tetrahedral aluminium (Al^{IV}), Fe, Fe/Mg and the molar fraction of the end-member chamosite molecule, and a corresponding decrease in the octahedral vacancy, octahedral aluminium (Al^{VI}) and the molar fraction of the end-member pyrophyllite-2 gibbsite molecule.

Cathelineau (1988) calibrated the chlorite geothermometer using chlorites from a geothermal system and a combination of direct temperature measurements from a thermal probe and temperature estimates from fluid inclusion data. The best correlation was with Al^{IV} , which showed a positive linear correlation between the aluminium in the tetrahedral site (Al^{IV}) and temperature of formation within the temperature range of 130 – 310 °C. The coefficient of the regression line is 0.97, and the relation between T and Al^{IV} is the following:

$$T\ (^{\circ}\text{C}) = -61.92 + 321.98(\text{Al}^{\text{IV}})$$

Furthermore, Cathelineau (1988) states that numerous data from the literature support the extrapolation of the regression line to temperatures higher than 350 °C, and that the chlorite geothermometer could probably be applied to chlorites of diagenetic, hydrothermal or metamorphic origin, as the Al in the

tetrahedral site seems to be independent of the rock lithology or the fluid composition. However, authigenic chlorites formed from biotite alteration may inherit chemical features of the parent mineral if alteration has not been complete.

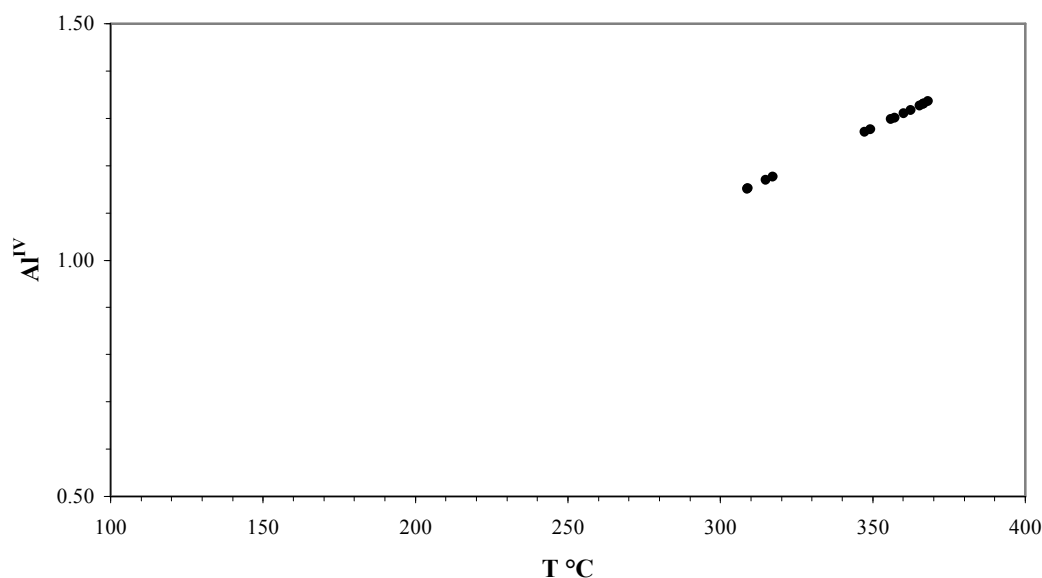


Figure 4.14: Plot of Al^{IV} vs. calculated temperature for chlorites from Haveri (after Cathelineau, 1988).

Figure 4.14 is a plot of Al^{IV} contents of chlorite versus calculated temperature. It is apparent that there are two populations of chlorites; a higher temperature range (347 – 368 °C) and a lower temperature range (309 – 317 °C). The lower temperature cluster (n=4) represents analyses of chlorites that have replaced biotite, and may thus be showing an inherited chemistry effect. The higher temperature cluster (n=10) is from a chlorite vein (and hence non-authigenic). However, the two clusters also define the two samples from which they were analysed. Therefore it is not possible to say whether the genetic difference of the chlorites is responsible for the observed temperature difference, or whether this observed temperature difference is a result of spatial differences at Haveri. Regardless, the temperature range of approximately 309 – 368 °C calculated using chlorite thermometry is consistent with the stability fields for chlorite on the petrogenetic grids, and thus can be taken as a lower temperature limit for hydrothermal alteration (which it cross-cuts), and hence also ore formation at Haveri.

4.5 DISCUSSION

The P-T conditions of metamorphism at Haveri have been reasonably well constrained to a maximum limit of 600 °C at 1.5 kbar, which corresponds to an approximate depth of 5 km. These P-T conditions are in broad agreement with other estimates of P-T conditions for the area.

Campbell (1978) estimated conditions between 500 °C, 1.5 kbar and 600 °C, 3 kbar for the peak of metamorphism in the Takamaa – Siivikkala area (to the east of the Hämeenkyrö Batholith, west of Lake Näsijärvi). On the basis of mineral parageneses in the Haveri Formation, Mäkelä (1980) has estimated metamorphic conditions to be approximately 550 °C and 2.5 kbar pressure. In addition, using the “aluminium in hornblende” geobarometer of Johnson and Rutherford (1989), Clark (1997) calculated a confining pressure of ca. 2.2 ± 0.5 kbar for both monzogranitic and granodioritic facies of the Hämeenkyrö Batholith. Furthermore, Clark (1997) “carried out electron microprobe analysis on coexisting (i.e. strictly in contact) sphalerite and hexagonal pyrrhotite from four areas of the inetermediate depth zones of orebody IV” for Stage 2 mineralisation at Paroinen, and calculated pressures of approximately 1250 ± 250 bars, using the equation of Bryndzia et al. (1990).

All of these separate studies, supported by this study, indicate that metamorphic (and post-metamorphic) environment in the broad Haveri region of the Tampere Schist Belt, was one of low pressure and high temperature conditions. Such metamorphic environments, with a high geothermal gradient, are typical of tectonic settings which include island arcs, ocean ridges and contact aureoles (Figure 4.15). Of these, on the basis of the observed geology, with the presence of the nearby granitoids, and the hornfelsic textures observed in thin section, the most likely metamorphic environment for the Haveri rocks is within a contact aureole.

The post-metamorphic P-T path is also consistent with the interpretation of a contact aureole and a cooling history that can be related to the emplacement and cooling of a plutonic mass (Pirajno 1992). With the intrusion of a plutonic mass, there is isochemical thermal metamorphism, as exemplified by the hornfelsic textures and mineralogy. This is followed by metasomatism and the generation of hydrothermal systems (represented by amphibolitisation and the calcic-skarn), and associated mineralisation. As the pluton finally cools, and the hydrothermal system contracts, there is retrograde hydrothermal alteration, as seen by chlorite and epidote replacement. The temperatures calculated by chlorite thermometry (309 – 368 °C), place a lower limit for the mineralisation and main alteration phases encountered at Haveri. The hydrothermal systems responsible for the mineralisation (and associated alteration) were thus high-temperature systems, with temperatures in excess of 309 – 368 °C, and likely to be in the region of 530 – 610 °C (as calculated by oxygen isotope thermometry). This wide range in temperatures may reflect a gradation from higher temperature, proximal zones to somewhat cooler distal zones, although insufficient temperature calculations and spatial information on the distribution of the proximal/ore zones is available to confirm this. The genetic models for the

alteration and mineralisation, and their relationship to the regional geology, will be examined in greater detail in Section **Error! Reference source not found.**

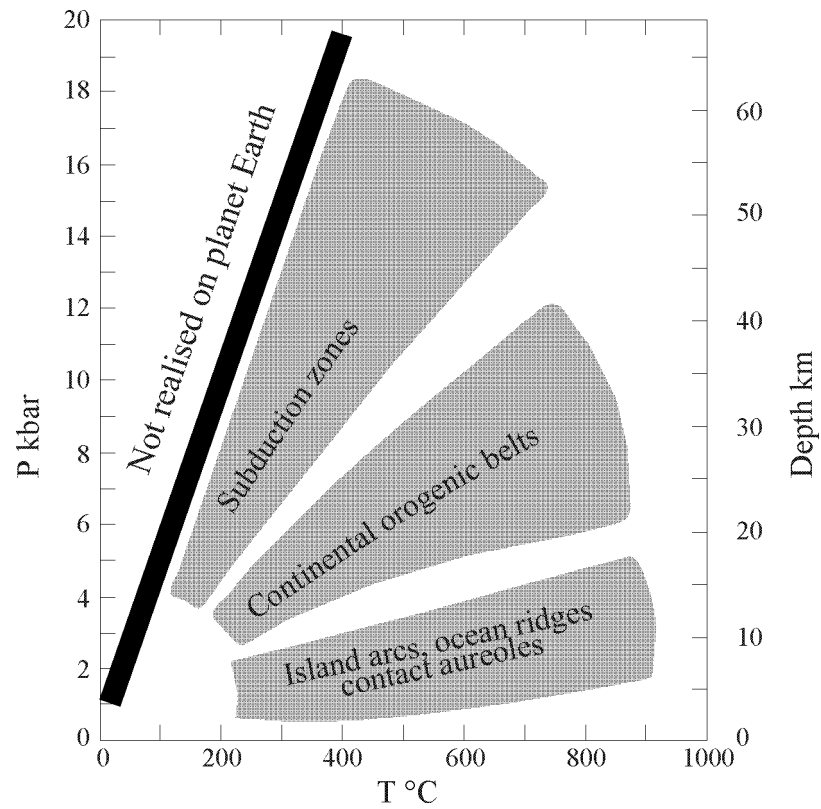


Figure 4.15: A pressure-temperature diagram showing the general regions for metamorphism in the crust (after Spear, 1993)

5 GOLD MINERALISATION

5.1 ORE DEPOSIT MODELS

Haveri has been the site of periodic mining since the 18th century. During its history, Haveri has been exploited for magnetite iron ore, gold and copper, and to a much lesser degree, cobalt. The first attempt to describe the geology and construct a genetic model for the Haveri mineralisation was by Stigzelius (1944). Since then, ore deposit geology has continued to evolve, and geologic models for the origin of ore deposits have waxed and waned in popularity. Likewise, the Haveri deposit has over the years been attributed to different genetic models. Stigzelius (1944) regarded the mineralisation as epigenetic, although at a later stage he proposed a syngenetic model for the mineralisation (Stigzelius, 1976, in Mäkelä, 1980).

A syngenetic approach was also supported by Mäkelä (1980), who compared Haveri to the Cyprus-style volcanogenic massive sulphide (VMS) deposits. The syngenetic model adopted by Mäkelä (1980), was formulated at a time when geological research was focussed on sea-floor mineralisation and the discovery of “black smokers”, resulting in an upsurge in the popularity of syngenetic models for ore deposits. Mäkelä (1980) based his syngenetic interpretation on a number of observations, including: the geochemistry and geotectonic environment of the pillowed mafic metavolcanic rocks (i.e. initial stage of island-arc development in a sea-floor spreading environment); the presence of chemical sediments (i.e. limestones); the interpretation of the massive sulphides and cataclastites as syngenetic flow breccias; and the sulphur isotope data as being indicative of seawater sulphur. Furthermore, he explained that the “magnetite disseminations in the metabasalts appear to be primary magmatic segregations (resulting from the high concentration of Fe in these tholeiitic rocks), whereas those in the metasedimentary rocks correspond to the oxide facies of chemical sedimentation”. However, Mäkelä (1980) failed to demonstrate a syngenetic relationship between the ore (and alteration) and the mafic metavolcanic rocks, and this study has shown the sulphide bands (cataclastites and massive sulphide) to be at least syn-D₂ in origin. In addition, this study has shown on the basis of field relations and carbon isotopes, that the “limestone” bands are in fact late stage (post-D₂) carbonate-barite veins. The interpretation of the sulphur isotope data is not conclusive of a seawater origin. A magmatic or magmatic-hydrothermal origin for the sulphur is equally, if not more, plausible than a seawater origin. Finally, Mäkelä (1980) failed to adequately account for the abundant magnetite (which he ascribed to primary magmatic segregations, resulting from the high concentration of Fe in the tholeiitic rocks) and the widespread amphibolitisation.

Although this study has clearly dispelled the notion that the gold mineralisation at Haveri is syngenetic, it must be stated that it is possible that syngenetic sulphides were indeed deposited during the formation of the pillowed volcanics in a back-arc basin environment, and that they have subsequently been remobilised into the sulphide veins and cataclastites observed today. However, no evidence for this has been observed, and an epigenetic origin for the gold and the alteration is incontrovertible.

A number of ore-deposit classifications exist for epigenetic Au deposits, of which three broad deposit types can be considered as having certain characteristics in common with Haveri, in terms of their lithological, alteration or mineralisation associations. These deposit types include orogenic lode-Au deposits, skarn deposits and Fe-oxide copper-gold deposits. These will be reviewed briefly below.

5.1.1 Orogenic Lode-Gold Deposits

5.1.1.1 Introduction

Orogenic lode-gold deposits (otherwise known as turbidite-hosted, greenstone-hosted, synorogenic or “mesothermal” gold deposits) are found in regionally metamorphosed terranes, whether of dominantly volcanic or sedimentary rocks, of all ages from the mid-Archaean to the Tertiary/Quaternary. The deposits are typically found within high strain zones in brittle (lower greenschist facies), brittle-ductile (mid-upper greenschist facies), or ductile (amphibolite / lower granulite facies) structures (Eilu et al., 1997; Pirajno 1992). Mineralisation in lode-gold deposits is typically hosted by quartz-dominated vein systems. The gold deposits were formed over a crustal continuum, with depositional conditions interpreted to represent depths of 15-20 km to the near surface (Groves et al., 1998). The actual origin of the hydrothermal ore fluids is still a somewhat contentious issue, and numerous genetic models have been proposed that include magmatic, metamorphic, and meteoric sources for the fluids, as well as combinations of the three.

5.1.1.2 Tectonic setting and metamorphism

At the broadest scale, the ore deposits were formed during compressional to transpressional deformation processes at convergent plate margins in accretionary and collisional orogens, which are typified by emplacement of abundant granitoid plutons and batholiths (Groves et al., 1998). The gold deposits are typically produced in the late-collisional stages of the development of orogenic belts (Partington and Williams, 2000). Pre-metamorphic protoliths for the auriferous Archaean greenstone belts are predominantly volcano-plutonic terrains of oceanic back-arc basalt and felsic to mafic arc rocks. Metamorphism associated with orogenic lode-gold deposits is most commonly greenschist-facies, although significant deposits also occur in higher-grade rocks extending into the granulite-facies (Eilu et al., 1999). At deeper crustal levels, the ore fluids were in thermal equilibrium with their host rocks, and thus the deposits were broadly synchronous with the peak of metamorphism, whereas at upper crustal levels the ore fluids were about 50-100 °C below the peak-metamorphic temperature of the host-rocks and thus slightly post-peak metamorphic in timing (Eilu et al., 1997; Groves et al., 1998). Many Proterozoic lode-gold deposits are also spatially associated with granite intrusion (especially in Australia, with reduced, fractionated, I-type granites), some of which are also mineralised, suggesting a complex temporal relationship between the end of orogenesis, granite intrusion, contact metamorphism, and gold mineralisation (Partington and Williams, 2000).

Mueller and Groves (1991) recognised that the lode-gold deposits of the Archaean Yilgarn Block of Western Australia can be divided into two groups: high-temperature deposits (400 – 700 °C) formed at pressures of 3 – 5 kbar in the lower parts of the greenstone belts (10 – 15 km depth), and medium-temperature or “mesothermal” deposits (250 – 400 °C) formed at pressures of 1 – 2 kbar in the upper parts of these belts (3 – 7 km depth). The high-temperature deposits are usually located in amphibolite facies terranes, whereas the mesothermal deposits are usually located in greenschist facies terranes.

5.1.1.3 Mineralisation and alteration associated with orogenic lode-gold deposits

Mineralisation in orogenic lode-gold deposits is primarily structurally controlled, with the mineralising fluids being channelled along the structures. Most deposits are localised in secondary or subsidiary faults or shear zones, or third-order structures, most commonly near regional-scale (typically transcrustal) deformation zones. The deposits can occur in virtually all rock types within the hosting volcano-sedimentary rock sequences, and in any metamorphic-grade environment, although greenschist-facies domains are the most common (Eilu et al., 1997). The styles of mineralisation include disseminations in shear zones, stratabound/stratiform replacements and massive to laminated quartz veins (Pirajno 1992). The deposits are typified by quartz-dominated vein systems, which develop according to whether the structural environment is brittle, brittle-ductile, or ductile in nature. The types of structures hosting lode-gold mineralisation include (1) brittle faults to ductile shear zones with variable kinematics, (2) fracture arrays, stockwork networks or breccia zones, (3) foliated zones with pressure solution cleavage, or (4) fold-hinge zones (Eilu et al., 1997; Witt, 1993). Veins systems typically contain 3-5 percent sulphides (normally dominated by Fe-sulphides) and 5-15 percent carbonate minerals under PT conditions where these are stable. Arsenopyrite is the most common sulphide mineral in metasedimentary country rocks whereas pyrite or pyrrhotite are more typical in metamorphosed igneous rocks. Gold-bearing veins exhibit variable enrichments in As, B, Bi, Hg, Sb, Te and W; Cu, Pb and Zn are only slightly elevated above regional backgrounds (Groves et al., 1998).

Hydrothermal alteration associated with orogenic lode-gold deposits typically forms an envelope that extends laterally up to hundreds of metres from ore and, commonly, also beyond the structure (shear zone) hosting gold mineralisation or, at least, beyond the sulphidised part of the system. In general, the shape of the alteration envelope is similar to the form of the gold mineralisation: i.e. it is roughly tabular and follows the strike and dip of a quartz vein or structure hosting the mineralisation. Alteration is typically developed perpendicular to the major fluid-flow direction, producing lateral zonations in the alteration assemblages which reflect decreasing fluid/rock ratios and related chemical gradients. Lateral zonations in the alteration assemblages can be complicated where two or more alteration haloes overlap. At the deposit scale, zonations parallel to the fluid-flow direction (vertical zonations) are virtually never seen, except where the metamorphic grade varies, and a mineral isograd cross-cuts the deposit. Vertical variations in the alteration assemblages may also be produced when the fluid conduit transects host lithologies with differing bulk compositions (Mueller and Groves, 1991; Witt, 1993; Eilu et al., 1997; Groves et al., 1998).

Hydrothermal alteration associated with orogenic lode-gold deposits generally includes carbonatisation, silicification, sulphidisation and alkali metasomatism (Pirajno, 1992). However, the actual mineral assemblage produced varies according to the metamorphic grade, and the composition of the host rocks (e.g. Eilu et al., 1997, 1999). In terms of relevance to this study, it is the high-temperature group of deposits (400 – 700 °C) described by Mueller and Groves (1991), that requires particular attention. This group of deposits formed in mid-amphibolite to lower-granulite facies host rocks, (as opposed to the more “normal” greenschist facies) and as such has significant differences in terms of their alteration mineralogy, mineralisation and hosting structures.

The “norm” for orogenic lode-gold deposits is in greenschist-facies domains, where associated alteration assemblages include calcite, dolomite, ankerite, chlorite, sericite or fuchsite. As metamorphic grade increases (up to lower-amphibolite), biotite replaces sericite as the dominant potassic mineral. Ca-bearing amphiboles and calcic plagioclase also become more prevalent in the alteration assemblages, and carbonate species other than calcite disappear. In addition as metamorphic grade increases, the extent of the alteration halo diminishes. At even higher metamorphic grades (mid-amphibolite to lower-granulite) the calc-silicate alteration assemblages are very skarn-like (Mueller and Groves, 1991), with diopside and Ca-amphibole as the most characteristic alteration minerals (Figure 5.1). A distal alteration zone of biotite may also be discernible. At such high metamorphic grades, the alteration haloes rarely extend beyond the mineralised shear zones, and more than a few tens of metres from the gold mineralisation, unless there are several sub-parallel mineralised structures (Eilu et al., 1997).

In the greenschist-facies host rocks, pyrite is the dominant sulphide. However, as metamorphic grade increases (up to lower-amphibolite) pyrrhotite rather than pyrite becomes the dominant Fe-sulphide mineral. At even higher metamorphic grades (mid-amphibolite to lower-granulite) pyrrhotite is commonly accompanied by arsenopyrite and/or loellingite. Magnetite is also increasingly important at higher temperatures (Clark, 1997). At the higher temperatures and pressures associated with the amphibolite – granulite facies metamorphism, the structures hosting mineralisation are more typical of brittle-ductile and ductile conditions, with shear zones containing fragments of wall rock and/or quartz veins (Eilu et al., 1997).

On the basis of the skarn-like alteration mineralogy of the high-temperature orogenic lode-gold deposits, Mueller and Groves (1991) suggested that a genetic link may exist between these two deposit types, and that the more typical “mesothermal” gold deposits at higher levels are part of continuous, large-scale hydrothermal systems, that at deeper levels become gold skarns.

Alteration zone Mineral	Distal	Intermediate (where present)	Proximal
SILICATES			
Talc ¹			
Tremolite ^{1,2}			
Hornblende			
Anthophyllite ¹			
Cummingtonite ¹			
Grunerite ³			
Clinopyroxene ²			
Orthopyroxene ⁴			
Olivine ^{1,3}			
Garnet ⁵			
Titanite ⁶			
Biotite/Phlogopite ²			
Quartz ⁷			
Plagioclase ⁷			
K-feldspar ⁸			
OXIDES			
Magnetite			
Ilmenite			
CARBONATES			
Calcite ⁶			
SULPHIDES			
Pyrrhotite ²			
Arsenopyrite ⁶			
Loellingite ⁶			
GOLD			

- 1) Only in ultramafic rocks, if present at all. Olivine in the distal zone is a regional metamorphic mineral.
- 2) May be part of thin, mono- and bi-mineralic subzones within the domain of proximal alteration. Note that the variation shown here in their volume in the proximal zone is only schematic, not exact.
- 3) Only in BIF; olivine (fayalite, if present) only in the proximal zone.
- 4) Only in the highest-temperature rocks.
- 5) In distal and intermediate zones, only regional metamorphic garnet is present.
- 6) Common, but not present in all cases.
- 7) Generally not present in ultramafic rocks.
- 8) Present throughout the sequence if in unaltered rock; otherwise only in the proximal zone.

Figure 5.1: Schematic summary of the paragenetic alteration sequence around orogenic lode-gold deposits in mid-amphibolite to lower-granulite facies environments. Proximity to mineralisation increases to the right. Black represents the common case, and grey indicates a less common occurrence. Zone widths shown bear no resemblance to actual widths in the field. After Eilu et al. (1999), and references therein.

5.1.1.4 Physico-chemical conditions of formation

Fluid inclusion studies of orogenic lode-gold deposits indicate that these deposits are characterised by $\text{H}_2\text{O}-\text{CO}_2\pm\text{CH}_4$ ore fluids with low to moderate salinities (typically between 2 and 10 wt.% NaCl equivalent), and the almost ubiquitous presence of CO_2 -rich inclusions. Fluids consistently have CO_2 concentrations in the order of 5-25 mole percent (Eilu et al., 1997). Homogenisation temperatures (Figure 5.2) span a wide range, but are mostly between 200 and 450 °C (e.g. McCuaig and Kerrich, 1998).

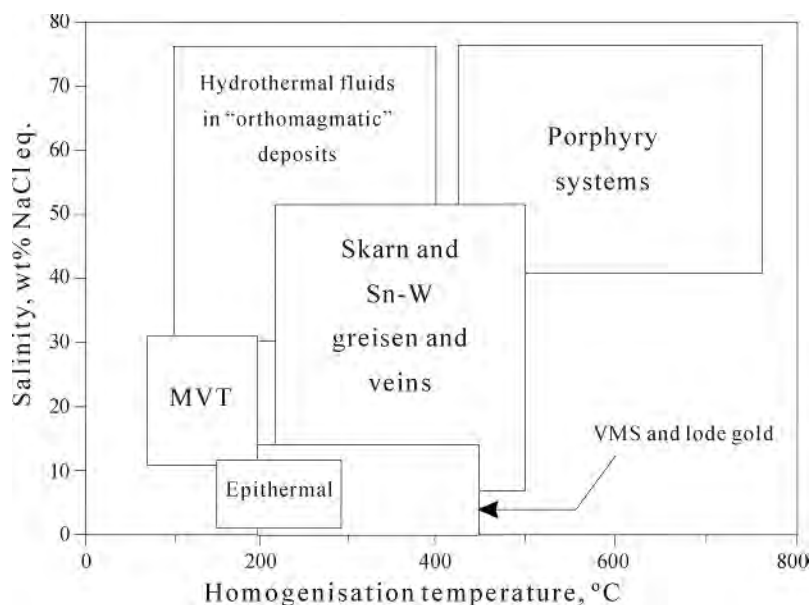


Figure 5.2: Schematic representation of salinity-temperature relationships of fluid inclusions typically associated with some deposit types. The boundaries for each field are somewhat arbitrary, and serve only as a broad approximation. Modified after Lattanzi (1994).

The CO_2 -rich, low-salinity nature of fluid inclusions in lode-gold deposits, has often been used as evidence for a nonmagmatic origin for the ore-forming fluids (e.g. Groves et al., 1998; Mueller and Groves, 1991; Phillips and Powell, 1993). Baker (2002) has, however, suggested that the fluids may in fact be magmatic in origin, and are the result of a complex interplay between exsolution of different volatiles (CO_2 , H_2O and Cl) from felsic magmas emplaced at different crustal levels.

Stable isotope studies of a number of Precambrian lode-gold deposits (Figure 5.3) show that the ^{18}O isotopic composition of vein quartz is fairly narrow, with $\delta^{18}\text{O}_{\text{quartz}}$ values of between approximately +9 to +15 ‰ (Figure 5.3). The values for the Mesozoic deposits are generally some 1-8 ‰ higher, and can be attributed mainly to higher reported temperatures of vein formation in the Archaean deposits, although differences in the $\delta^{18}\text{O}$ of the ore fluid, caused by variations in wall-rock lithology and water/rock ratios, may also be partially responsible (Taylor, 1987).

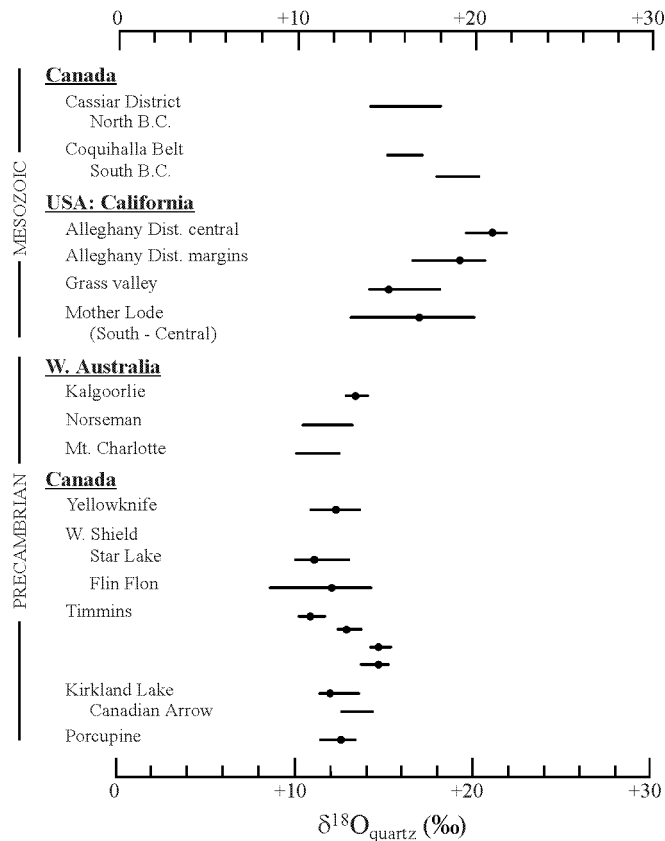


Figure 5.3: Plot of $\delta^{18}\text{O}$ for vein quartz from lode-gold vein deposits. Closed circle indicates average value. After Taylor (1987) and references therein.

The range of calculated $\delta^{18}\text{O}_{\text{fluid}}$ values are typically 5-8 ‰ in volcanic belts and 7-10 ‰ in sedimentary belts (Eilu et al., 1997). Many studies have shown that there are often several generations of fluids, and that the isotopic composition of the fluids may have changed with time. In many cases the compositions of the late gold-depositing fluids are lower by approximately 1-2.5 ‰ than the earlier fluids (Taylor, 1987). This has been taken as evidence that, as the hydrothermal system evolves, there is a greater input of (isotopically lighter) meteoric water. This interpretation is also supported by the hydrogen isotopes which also show a depletion in δD in the later fluids (e.g. Kyser, 1986; Taylor, 1987). Taylor (1987) uses the oxygen and hydrogen isotope data to support the conclusion that fluids of mixed and complex origin (a deeper-seated, isotopically heavy, evolved or “metamorphic” fluid, and an isotopically light meteoric fluid) were responsible for hydrothermal alteration, gold transport and deposition.

The mean values of $\delta^{34}\text{S}_{\text{pyrite}}$ are broadly similar for a number of orogenic lode-gold deposits (Figure 5.4), and fall in one of two categories: (1) -0.5 to +3.5 ‰, for most, and (2) < -3 ‰, for a few Archaean deposits (Taylor, 1987). Deposits with $\delta^{34}\text{S}_{\text{pyrite}} < -3$ ‰, are further distinguished by the presence of sulphate.

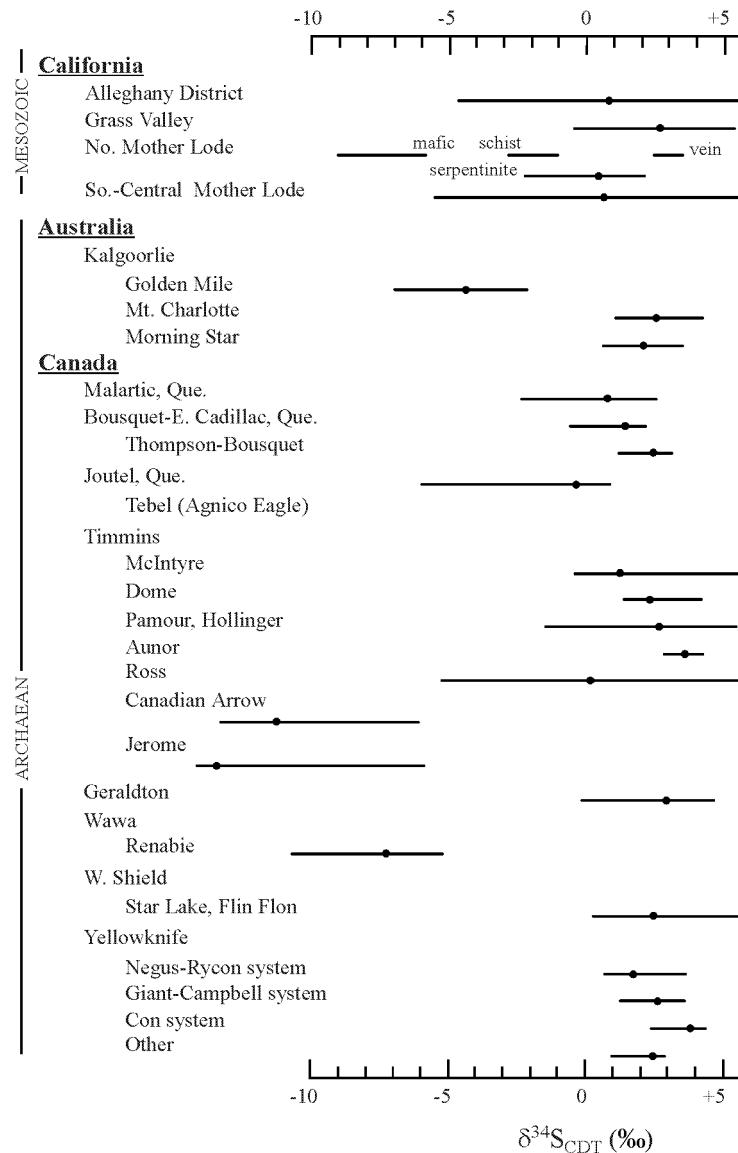


Figure 5.4: Plot of $\delta^{34}S_{py}$ for lode-gold vein deposits. Bar and filled circle indicate range and mean values of $\delta^{34}S$ respectively. After Taylor (1987) and references therein.

Wallrock alteration assemblages and fluid inclusion studies indicate that the ore fluid responsible for deposition in orogenic lode-gold deposits, was typically a low-salinity, near-neutral, $H_2O-CO_2 \pm CH_4$ fluid which transported gold as a reduced sulphur complex (Eilu et al., 1997; Groves et al., 1998).

There is a very strong structural control on the localisation and deposition of gold, with cyclic fault-valve pressure release being an important depositional mechanism (Cox et al., 1990). In such a mechanism, upward fluid flow is impeded by decreasing the permeability of conduits due to progressive mineral deposition. When supralithostatic fluid pressures are attained, hydrofracture networks form in the wall-rocks below the fault valves. At this stage, fluid pressure differences may cause fluid infiltration of wall-rocks adjacent to faults, leading to fluid-rock interaction. Fault zone failure and valve breaching causes an abrupt drop in fluid pressure within the fault zone, especially in dilatant jogs. This causes reacted fluids in the wall-rocks to be drawn back into the transiently lower

pressure fault zone where they mix with primary hydrothermal fluids. This process can be very important in promoting efficient gold deposition. Furthermore, such a pressure release model explains why these deposits are found in second- and third-order structures which are easier to seal, rather than the larger first-order regional structures which remain “open” (Cox et al., 1990). Phillips (1990) has suggested that interaction with, and sulphidation of, the mafic wall-rocks are extremely potent mechanisms for decreasing the sulphur activity of the fluids, and destabilising the reduced sulphur complex, leading to gold precipitation.

5.1.1.5 Genetic models for orogenic lode-gold deposits

The most contentious issue in formulating a genetic model for orogenic lode-gold deposits is the source of the ore fluids. The mechanisms for the actual transport and deposition of the gold, are themselves more broadly accepted (except the syngenetic models), although some of the finer points are perhaps not yet fully understood, such as the behaviour of sulphide complexes above 400 °C (Mueller and Groves, 1991; Groves et al., 1998). Therefore most of the models vary only to the extent of different sources for the ore fluids. However, due to the complex geological setting of orogenic lode-gold deposits, a large number of genetic models (Kerrick, 1990; Mueller and Groves, 1991) have been invoked that include: (a) magmatic fluids, such as from lamprophyres, tonalite-trondhjemite-granodiorite magma suites or oxidised felsic magmas; (b) metamorphic fluids, derived from the devolatilisation of mafic / ultramafic rocks during medium-grade metamorphism or devolatilisation of the lower crust during high-grade metamorphism; and (c) meteoric fluids that have been undergone deep circulation.

An in-depth review of the genesis of orogenic lode-gold deposits is beyond the scope of this study, and thus will not be examined further, as the issue of genesis will no doubt continue for some time to come. However, it is sufficient to say that the orogenic lode-gold deposits form in a complex geologic environment, where magmatism, deformation and metamorphism are occurring contemporaneously, and all are changing character in space and time (Kerrick, 1990). In such a scenario, it is not unlikely that a combination of processes are at work.

5.1.2 Skarn Deposits

The term *skarn* originated in central Sweden, where it was used by miners to refer to the coarse-grained calc-silicate gangue associated with iron ore (Geijer and Magnusson, 1952). Since that time the term has been expanded to include a large variety of calc-silicate rocks that are rich in calcium, iron, magnesium, aluminium, and manganese regardless of their association with economically interesting minerals, and that formed by replacement of originally carbonate-rich rocks (Einaudi et al., 1981).

Skarns occur on all continents and in rocks of almost all ages. Although the majority of skarns is found in lithologies containing at least some limestone, skarns can form in almost any rock type, including shale, sandstone, granite, iron-formation, basalt, and komatiite (Einaudi et al., 1981). Skarns can form during regional or contact metamorphism and from a variety of metasomatic processes involving fluids of magmatic, metamorphic, meteoric, or marine origin. Skarns are found adjacent to plutons, along faults and major shear zones, in shallow geothermal systems, on the bottom of the seafloor, and at lower crustal depths in deeply buried metamorphic terranes (Meinert, 1992). In addition to the geological variability, skarns have been mined for a variety of metals, including Fe, W, Cu, Pb, Zn, Mo, Ag, Au, U, REE, F, B, and Sn (Meinert, 1998). What links these diverse environments, and what defines a rock as skarn, is the mineralogy, which includes a wide variety of calc-silicate and associated minerals, but is usually dominated by garnet and pyroxene. Thus the presence of skarn does not necessarily indicate a particular geologic setting or particular protolith composition. Rather, its development indicates that the combination of temperature, pressure, fluid and host rock composition was within the stability range of the identified skarn minerals (Meinert, 1992).

Skarns can be subdivided according to several criteria. Exoskarn and endoskarn are common terms used to indicate a country-rock and intrusion protolith, respectively. Magnesian and calcic skarn can be used to describe the dominant composition of the protolith and resulting skarn minerals. In addition, there is a continuum between purely metamorphic (thermometamorphic) and purely metasomatic processes that generate skarns, creating calc-silicate hornfels, reaction skarns, and skarnoid as the metasomatic component increases. The economically most important skarn deposits result from large-scale metasomatic transfer, where fluid composition controls the resulting skarn and ore mineralogy (Einaudi et al., 1981; Meinert, 1992).

Skarn ore deposits are best classified on the basis of the dominant economic metal (Einaudi et al., 1981), and then further subdivided on the basis of compositional, tectonic or genetic variations (Meinert, 1992). There are thus seven major skarn types (Fe, Au, W, Cu, Zn, Mo and Sn) and several others of local importance (including F, C, Ba, Pt, U and REE) (Meinert, 1992).

The formation of skarn deposits is a dynamic process, and for skarns related to plutons there is a parallel relationship between the sequence of emplacement, crystallisation, alteration and the cooling of

the pluton and the corresponding metamorphism, metasomatism and retrograde alteration in the surrounding rocks. In addition, in most large skarn deposits, there is a transition from early/distal metamorphism resulting in hornfels, reaction skarn and skarnoid, to later/proximal metasomatism resulting in relatively coarse-grained, ore-bearing skarn (Meinert, 1992). Towards the final phases of cooling, a retrograde stage occurs, during which the early skarn assemblages are destroyed and replaced by hydrous minerals (Pirajno, 1992).

The degree to which a particular stage is developed in a specific skarn will depend on the local geologic environment of formation. For example, metamorphism will likely be more extensive and of higher grade around a skarn formed at relatively great crustal depth than one formed under shallow conditions. Conversely, retrograde alteration during cooling, and possible interaction with meteoric water, will be more intense in a skarn formed at relatively shallow depths in the earth's crust compared with one formed at greater depths (Meinert, 1992). The depth of skarn formation also affects the mechanical properties of the host rocks. In a deep skarn environment, rocks tend to deform in a ductile manner, rather than fracture. This results in sub-parallel intrusive contacts with the bedding of host rocks. In occurrences such as these, skarn is usually confined to a narrow, but vertically extensive, zone. In contrast, host rocks at shallow depths will tend to deform by fracturing and faulting, rather than folding, resulting in intrusive contacts sharply discordant to bedding. In such occurrences, skarn cuts across bedding and massively replaces favourable beds. In addition, the hydrofracturing typically increases the permeability of the host rocks, not only for igneous-related metasomatic fluids but also for later, possibly cooler, meteoric fluids. The influx of meteoric water and the consequent destruction of skarn minerals during retrograde alteration is one of the distinctive features of skarn formation in a shallow environment (Meinert, 1992).

5.1.2.1 Gold Skarns

Gold is a common by-product of many skarns. The term “gold skarn” is commonly used in the economic sense suggested by Einaudi *et al.* (1981) to refer to ore deposits that are mined solely or predominantly for gold, and which exhibit calc-silicate alteration, usually dominated by garnet and pyroxene, that is related to mineralisation (Meinert, 1992, 1998; Theodore *et al.*, 1991).

Although skarns can form in virtually any tectonic environment, there is a certain tectonic control on the metallogenic associations of skarns (Figure 5.5). Many economic gold skarns appear to have formed in back-arc basins associated with oceanic island arcs. In addition, calcic Fe-Cu skarns are the principal skarn types which have significant gold as a by-product (they are also typically enriched in Co, Ni and Cr). This is virtually the only skarn type found in oceanic island-arc terranes. Some of the key features that characterise these skarns are their association with gabbroic and dioritic plutons, abundant endoskarn, widespread sodium metasomatism and the absence of Sn and Pb. These features reflect the primitive oceanic nature of the crust, wall rock and plutons (Meinert, 1992).

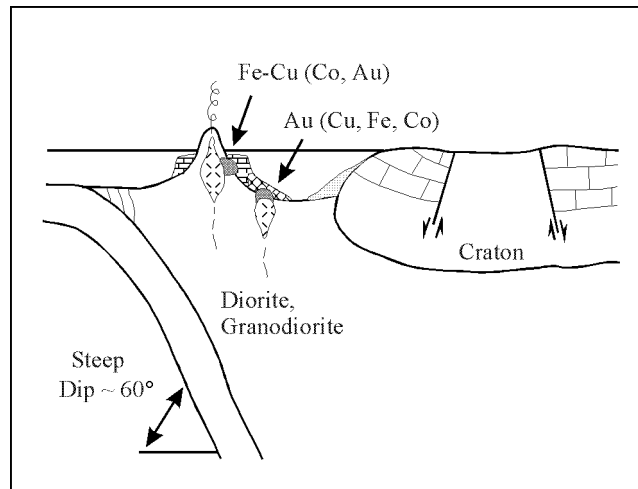


Figure 5.5: *Idealised tectonic model for the formation of gold skarns in oceanic subduction and back arc basin environments. After Meinert (1992).*

One of the main characteristics of all gold skarns is the surrounding biotite $\pm$ K-feldspar (potassic) alteration as a distal alteration zone. In most cases, the biotite $\pm$ K-feldspar alteration is formed in relatively fine-grained, clastic host rocks resulting in a hornfelsic texture. The K in biotite $\pm$ K-feldspar alteration seems to be, at least in part, internally derived from the host stratigraphy (Meinert, 1998). In most cases, the original clastic sedimentary rocks contain abundant K-rich argillaceous and arkosic material which has been replaced by amphibole and/or pyroxene and then garnet in inner or more intense zones of alteration. As garnet, pyroxene or amphibole do not contain K, the K originally in the host rock is liberated during skarn formation and may be incorporated into the distal biotite $\pm$ K-feldspar alteration. Therefore, although the distal biotite $\pm$ K-feldspar alteration can be considered to be metasomatic on the outcrop scale, on the larger deposit scale it is more the case of *in situ* redistribution of elements during mineral reactions (Meinert, 1998).

If the biotite $\pm$ K-feldspar alteration represents the distal alteration zone of Au skarns, then the proximal zone may be represented by relatively coarse-grained garnet and pyroxene. This proximal skarn is zoned from an internal garnet-dominated zone, closest to the pluton or fluid pathway, to a more distal zone dominated by pyroxene (Meinert, 1998). The relative proportions of garnet and pyroxene are a complex function of protolith composition, activity of components in the hydrothermal fluid, and overall oxidation state as influenced by magmatic sources, wallrock composition, and mineral reactions (Einaudi et al., 1981; Meinert, 1992, 1998).

5.1.2.1.1 *Classification of gold skarns*

The majority of gold skarns share a number of common features, such as biotite hornfels, garnet-pyroxene alteration, clastic- or volcanoclastic-rich protoliths, and a Au-As-Bi-Te geochemical signature. However, they also exhibit significant differences, which form the basis for the four major subdivisions described by Meinert (1998): reduced Au skarns, oxidised Au skarns, magnesian Au skarns and metamorphic Au skarns.

Most Au skarns are calcic skarns. However, a few magnesian Au skarn deposits do exist, such as at Cable Mountain and Butte Highlands in Montana. Most magnesian skarns form from dolomitic protoliths and exhibit a diagnostic mineralogy that includes forsterite, spinel (usually dominated by magnetite) and serpentine. Early magnesian skarn minerals are often overprinted by later calcic skarn minerals (Meinert, 1998). As is commonly the case with most gold skarns, there is a mineralogical association of gold with retrograde alteration.

Although most skarns are associated with relatively shallow plutons (typically of Phanerozoic age) that have intruded previously unmetamorphosed sedimentary rocks (Einaudi et al., 1981), Meinert (1998a) describes a disparate group of gold skarns, classified as metamorphic gold skarns, that have a common link to regional metamorphism. This group includes skarns from Precambrian orogenic belts where the skarn is associated with both plutonism and high-T-P metamorphism. In addition to these plutonic/metamorphic occurrences there are several “mesothermal” lode-gold deposits with skarn alteration in Precambrian terranes without associated intrusive rocks (Meinert, 1998). Although little is known about the geological relations of the skarn alteration or the connection between gold mineralisation and skarn formation, these deposits have several characteristics uniting them. They have a mineralogy dominated by very Fe-rich (pyrrhotite dominated) and reduced assemblages, including almandine-spessartine garnet, hedenbergitic pyroxene, and Fe-rich amphibole (Figure 5.6). They are generally formed from Fe-rich rocks, such as iron-formation, komatiite or metabasite, (although there are exceptions such as Navachab which is formed in carbonate rock) at considerable depth (2 – 6 kbar pressure) in regional metamorphic terranes.

Many of these metamorphic Au skarns, are classified as such purely on the basis of the calc-silicate (skarn) mineralogy. However such mineral assemblages are also commonly formed in orogenic gold deposits under amphibolite-facies PT conditions. Consequently many deposits that have been classified by Meinert (1998) as metamorphic Au skarns, such as Lupin and Nevoria (Figure 5.6), actually have many features in common with amphibolite-facies orogenic gold deposits (such as those in the Southern Cross greenstone belt, in Norseman, and certain deposits in Pilbara) and the genetic classification therefore depend on the temporal and spatial relationships between mineralisation and alteration (Bloem et al., 1994; Bullis et al., 1994; Eilu et al. 2000; Hagemann et al., 1998; McQuaig et al., 1993; Mueller, 1997; Neumayr et al., 1993; Robert and Poulsen, 1997; Witt et al., 2001; Witt and VanderHor, 1998).

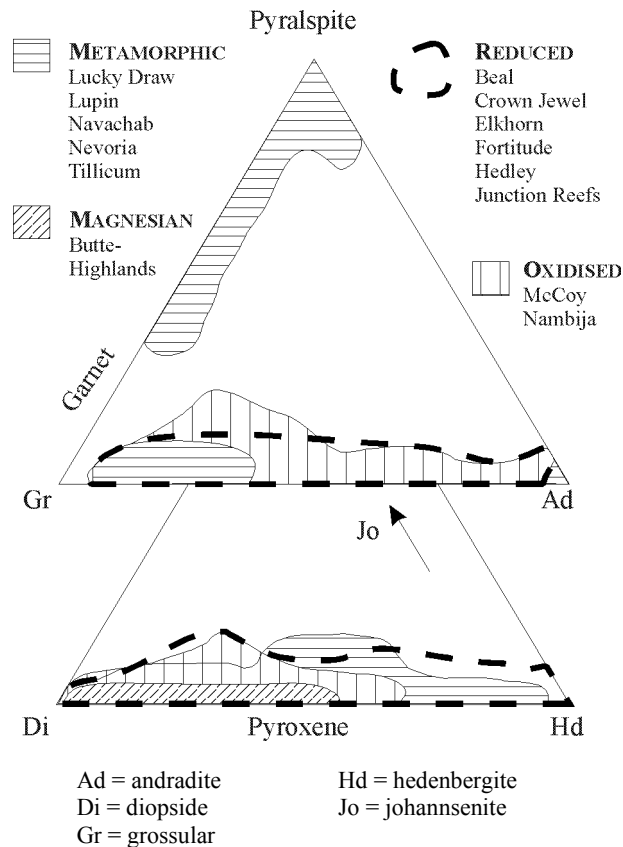


Figure 5.6: Comparison of garnet and pyroxene compositions from reduced, oxidised, magnesian and metamorphic Au skarns. Also listed are examples of each category of gold skarn. After Meinert (1998) and data therein.

The majority of gold skarns are related to shallow plutons (≤ 5 km depth of formation) and are represented by reduced and oxidised Au skarns. Most plutons associated with reduced Au skarns are ilmenite-bearing mafic diorites and granodiorites, whereas most plutons associated with oxidised Au skarn are more silicic and magnetite-bearing. As with metamorphic gold skarns, the compositions of the garnet, pyroxene and amphibole in the skarn mineral assemblage are characteristic of the type of skarn (Figure 5.6). Although most sulphide mineralisation occurs at lower temperatures after the main stage of garnet and pyroxene formation, the silicate, oxide, and sulphide mineral assemblage tends to reflect the overall oxidation and sulphidation state of the hydrothermal system (Figure 5.7). Reduced gold skarns contain hedenbergitic pyroxenes, intermediate grandite garnet, actinolite – ferro-actinolite, pyrrhotite $\pm$ magnetite. Oxidised gold skarns, on the other hand, are characterised by major amounts of andraditic garnet, diopsidic pyroxene, tremolite – actinolite, hematite and pyrite (Meinert, 1998).

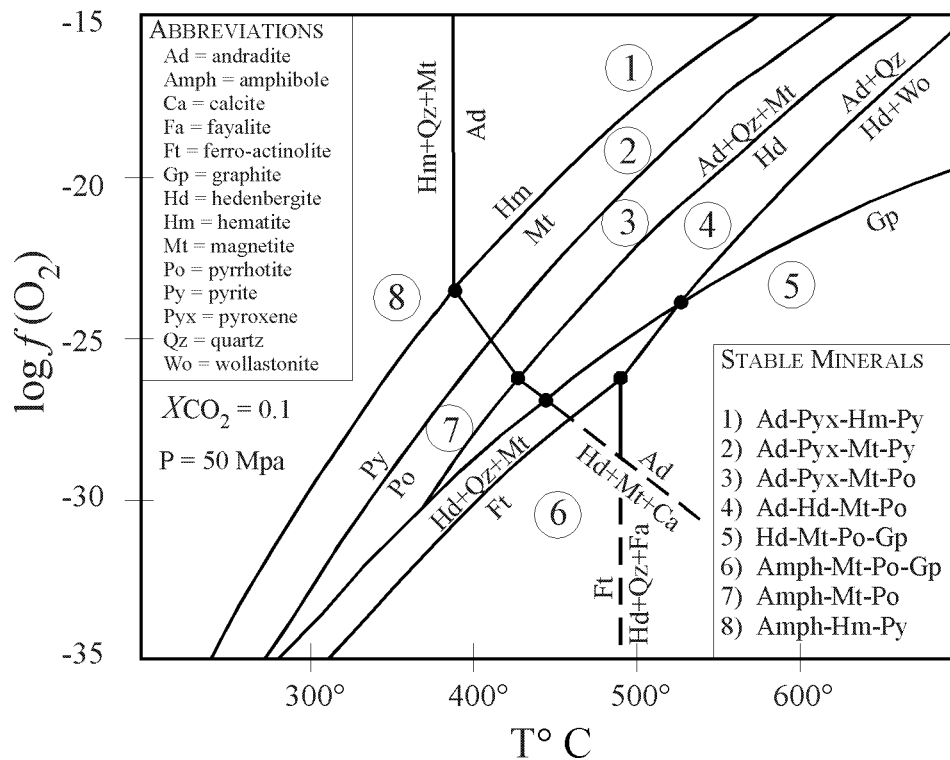


Figure 5.7: T - $f(O_2)$ diagram showing the stability fields of major skarn silicate, oxide, and sulphide minerals. Oxidised Au skarns typically contain associations 1, 2, and 8. Reduced Au skarns typically contain associations 3, 4, and 7. Metamorphic Au skarns, typically contain associations 4, 5, 6, and 7. Associations 5 and 6 are not stable in oxidised Au skarns because of the presence of graphite. Associations 1 and 8 are not stable in metamorphic Au skarns due to the presence of hematite. After Meinert (1998)

5.1.2.1.2 Conditions of formation of gold skarns

Fluid inclusion studies of skarn deposits in general (summarised by Kwak, 1986; Meinert, 1992) consistently indicate that prograde skarn assemblages form at high temperatures (with homogenisation temperatures up to and in excess of 700 °C) from high salinity (up to and in excess of 50 wt.% $NaCl_{equiv.}$) fluids, commonly with salt daughter minerals present. As a skarn evolves from a prograde to retrograde system (either spatially or temporally), different mineral assemblages become stable. Typically, fluid inclusions from distal/retrograde stages of skarn formation have lower homogenisation temperatures and salinities than from proximal/prograde stages (Meinert, 1992).

Meinert (1992) summarises the fluid inclusion studies of a number of gold skarn deposits. The fluid inclusions in prograde garnet and pyroxene have homogenisation temperatures up to 730 °C and 695 °C respectively, with salinities up to 33 wt.% $NaCl_{equiv.}$. In contrast, scapolite, epidote and actinolite formed during the retrograde skarn events have homogenisation temperatures of 320-400 °C, 255-320 °C and 320-350 °C respectively.

Stable isotope studies of skarns are consistent with fluid inclusion and mineral equilibria studies which demonstrate that most large skarn deposits form from diverse fluids, including early, high temperature highly saline brines directly related to crystallizing magma systems (Meinert, 1992). Skarn silicate minerals are markedly depleted in ^{18}O compared to metamorphic calc-silicates, being buffered by fluids in oxygen isotope equilibrium with adjacent intrusive rocks ($\delta^{18}\text{O}_{\text{water}} = +8.0$ to $+11.0$) (Taylor, 1987). More specifically, garnet, pyroxene and associated quartz all have $\delta^{18}\text{O}$ values in the $+4$ to $+9$ range consistent with derivation from magmatic waters (Meinert, 1992). Hydrous minerals from the retrograde/distal stages of skarn formation display a progressive decrease in the δD value over a limited range of $\delta^{18}\text{O}$ values. The latter correspond to values in oxygen isotope exchange equilibrium with the adjacent intrusion, at the temperature of that stage of skarn formation. This trend of declining δD at relatively constant $\delta^{18}\text{O}$ value can result either from mixing of ^{18}O -enriched meteoric water with magmatic water or by isotopic evolution of one water source (most likely meteoric water) at low but variable fluid/rock ratios (Bowman, 1998b; Meinert, 1992; Taylor, 1987).

Sulphur isotope studies of skarns indicate that during the early stages of skarn formation, skarn S is dominated by magmatic S. However, variations in the $\delta^{34}\text{S}$ values of this magmatic S from one skarn to another probably indicate the importance of assimilation, source mixing, or magma mixing prior to magma emplacement into the upper crust and the onset of skarn formation. Variations in the $\delta^{34}\text{S}$ values of sulphide minerals spatially and temporally within skarn indicate either progressive contributions of wallrock S (source mixing) or changes in the physico-chemical conditions [T , $f(\text{O}_2)$, $f(\text{S}_2)$, pH] of the skarn fluids. Both of these changes can result from progressive interaction of magmatic fluids with the carbonate wallrocks which the skarns replace (Bowman, 1998b).

Meinert (1989, 1998) suggested that gold in skarns is transported as a chloride complex at high temperatures, and as a sulphide complex at lower temperatures. Due to the high temperature (350 – 650 °C), hypersaline fluids that are associated with gold skarns, high concentrations of gold are able to be transported as chloride complexes (Gammons and Williams-Jones, 1995, 1997). However, as temperatures decrease (both temporally and spatially) below 400 °C, the solubility of gold chloride complexes is greatly diminished. At the lower temperatures (< 300 °C) associated with the retrograde stage of a skarn system, a thiosulphide complex $[\text{Au}(\text{HS})_2]^-$ is more likely (Hayashi and Ohmoto, 1991). Precipitation of the gold can be caused by cooling, or by the oxidation of a reduced hydrothermal fluid (Meinert, 1998).

5.1.3 Fe-Oxide Copper-Gold Deposits

The Fe-oxide Cu-Au deposits have only emerged as a distinctive grouping during the last two decades, although individual mines have been known of longer. Within the class, the individual deposits are extremely variable in many of their characteristics, and as such, fundamental uncertainties remain regarding the metallogensis of these deposits (Williams, 1999b). The principal characteristics of this diverse class of deposits are 1) their localisation by tectonic (including volcanic structures in high level systems) rather than plutonic processes, 2) abundance of iron minerals other than sulphides, especially magnetite and/or hematite, 3) vein-breccia-replacement style ores, 4) the ore elements Cu with Au and/or Ag, commonly (but not invariably) accompanied by a specialised suite of minor elements including Co, U, LREE, Ba and F, and 5) the presence of regionally extensive Na-Ca-Cl metasomatism (Hitzman et al., 1992; Williams, 1999b).

Much of the literature regarding the Fe-oxide Cu-Au deposits as a metallogenic class, focuses on the deposits in the Australian Proterozoic terrains. These are highly variable in character, ranging from very large deposits (e.g. Olympic Dam), to small, high grade deposits such as those in the Tennant Creek Inlier and the Cloncurry district of the Mount Isa Inlier. The Fe-oxide association of these deposits range from hematite-dominated (e.g. Olympic Dam) to magnetite-dominated (e.g. Ernest Henry). Fe-sulphide may be present as either pyrite, pyrrhotite or both. In addition, most of the deposits contain chalcopyrite as the only significant Cu-phase, although some contain hypogene bornite and chalcocite. Furthermore, the Cu-Au ratios vary substantially between deposits, and there is no single consistent minor element association, although there is a distinctive association with F, Ba, LREE and U, and commonly also Co, Mo and Bi (Hitzman et al., 1992; Williams, 1999b). Despite these variations, it is still possible to characterise many features common to deposits within this class.

5.1.3.1 Age and setting of Fe-oxide Cu-Au deposits

The Fe-oxide Cu-Au deposits fall into two main temporal groupings of Proterozoic and Phanerozoic age. In both cases, there seems to be a link to the magnetite-apatite deposits of the “Kiruna-type”, which in some cases occur in the same districts. Most of the Proterozoic examples occur in extensional tectonic regimes either within intracratonic settings or at continental margins. These deposits commonly display spatial associations variously with anorogenic magmatism or magmatism associated with extensional deformation syn- to post-regional metamorphism (Hitzman et al., 1992; Pollard, 1999). There is a strong concentration of these deposits in late Palaeoproterozoic to early Mesoproterozoic rocks (1.8 – 1.4 Ga) (e.g. Hitzman et al., 1992; Williams, 1999b). Proterozoic Fe-oxide Cu-Au deposits have been recognised in many Proterozoic terrains including some in Australia (Cloncurry district of the Eastern Fold Belt of the Mount Isa Inlier; the Tennant Creek inlier; and Proterozoic components of the Gawler Craton), North America (Yukon Territory; North West

Territory; Midwest USA), Scandinavia (Kiruna district; Bergslagen district), Brazil (Carajás district) and South Africa (Vergenoeg) (Williams, 1999b). Phanerozoic examples are found in extensional settings at destructive plate margins and range in age from early Cambrian to late Tertiary. The most significant Phanerozoic examples include deposits in the coastal belt of Chile (e.g. La Candelaria) and Peru (Williams, 1999b).

5.1.3.2 Nature of associated intrusive rocks and metamorphism

The intrusive rocks which are associated with Fe-oxide Cu-Au mineralisation range from diorite to granite in composition. These commonly occur as extensive granitoid batholiths composed of multiple intrusions, commonly in association with basaltic and/or ultramafic rocks (Pollard, 1999). Magmatism in the Cloncurry district has been shown to have been long lived (ca. 40 Ma) and to have involved multiple magma sources (Pollard et al., 1998).

In intracratonic settings the granites and associated Fe-oxide Cu-Au mineralisation were commonly emplaced during or slightly after the peak of metamorphism, and metamorphic assemblages are generally indicative of high-temperature, low-pressure metamorphism. Such a style of metamorphism requires a significant contribution of heat from the mantle and suggests that the major thermal events, granitoid formation and mineralisation are ultimately driven by intrusion of mantle-derived magmas into the crust (Pollard, 1999).

5.1.3.3 Alteration styles associated with Fe-oxide Cu-Au deposits

Fe-oxide Cu-Au deposits are a highly variable class of deposit that formed under a range of conditions in diverse geological environments and in many different host rocks. In addition, there is plenty of evidence for variable fluid chemistry from deposit to deposit. This variability consequently produces many different alteration assemblages, that are often further complicated by a regional-scale metasomatism (Williams, 1999a).

A fundamental characteristic of the class as a whole, is the (relatively) sulphur-poor nature of the deposits, with excess iron present as the oxides hematite and magnetite. Hitzman et al. (1992) recognised that there is a variation from relatively deep-seated deposits in which magnetite predominates (e.g. Cloncurry district) to deposits formed at shallower levels in which hematite is more abundant (e.g. Olympic Dam). This is paralleled by a variation from higher- to lower-temperature assemblages in the associated silicate alteration (Figure 5.8). The alteration styles tend to be sodic at depth, potassic at intermediate to shallow levels, and sericitic and silicic at very shallow levels (Hitzman et al., 1992).

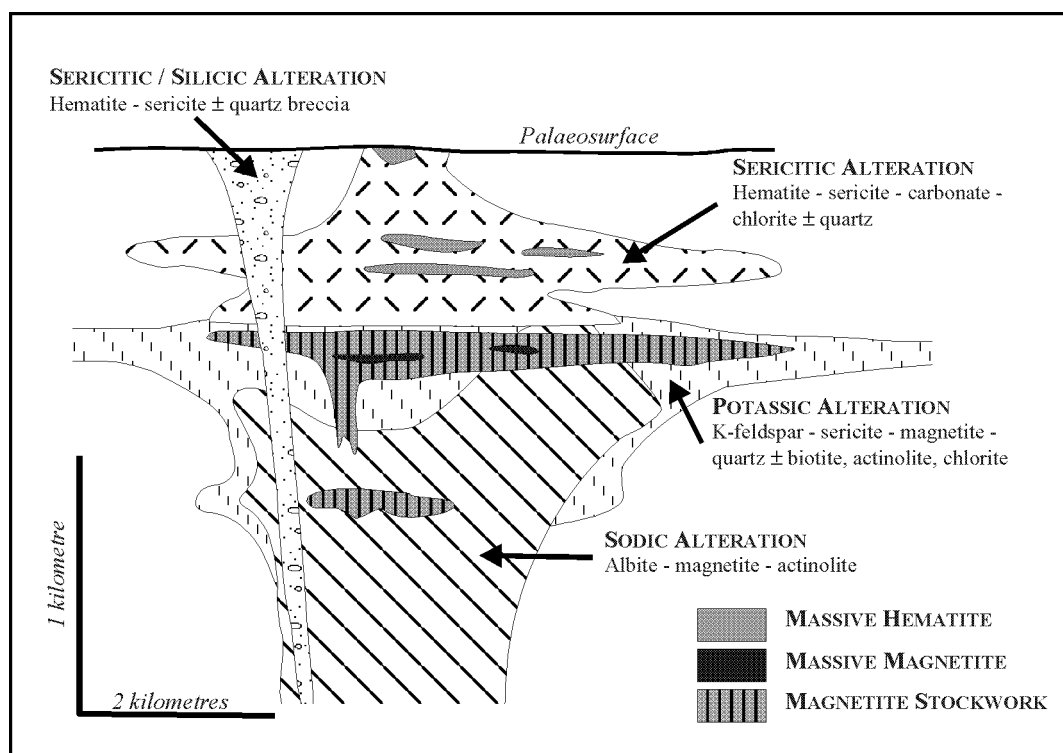


Figure 5.8: Schematic cross section illustrating alteration zoning in iron oxide-Cu-Au deposits. This section extends from near surface to several kilometres below surface (after Hitzman et al., 1992).

The major alteration style associated with Fe-oxide Cu-Au deposits is Na-(Ca-Fe-Cl) alteration. The characteristic mineral of this type of alteration is sodic plagioclase (albite or oligoclase). Other minerals which are commonly present include, quartz, scapolite, actinolite, diopsidic clinopyroxene, paragonite, sphene, magnetite, hematite, apatite and carbonates. K-feldspar is typically destroyed and plagioclases are commonly transformed to more albitic compositions. This transformation involves the consumption of silica, and some altered rocks may become silica-undersaturated leading to the formation of corundum or Al-silicates such as andalusite (Williams, 1999a).

Sodic (-calcic) alteration occurs in association with a wide variety of brittle and ductile structures, and in some cases there is evidence for multiple, overprinting, stages of sodic (-calcic) alteration. The most common types of structural controls on alteration include: brittle-ductile shear zones, ductile fabrics (cleavages) linked to regional deformation, breccia complexes and brittle fracture systems (De Jong and Williams, 1995; Ettner et al., 1993; Frietsch et al., 1997; Hitzman et al., 1992; Mark, 1998; Oliver et al., 1993; Pollard, 1999; Williams, 1999a). In addition, pervasive sodic (-calcic) alteration in the form of clinopyroxene-albite±sphene alteration has commonly been observed in Fe-oxide Cu-Au deposits, such as at the Lightning Creek prospect in the Mount Isa Eastern Fold Belt (Perring et al., 1999), and the Candelaria deposit in Chile (Marschik and Fontboté, 2001). In northern Fennoscandia, pervasive sodic (-calcic) alteration is also accompanied by the formation of scapolite (Frietsch et al., 1997).

Another common alteration style observed in Fe-oxide Cu-Au deposits is high temperature (>400 °C) K-Fe alteration. This is similar to assemblages observed in the potassic alteration zone around porphyry copper deposits and is characterised by biotite, K-feldspar and magnetite, which are typically associated with plagioclase destruction. Other minerals that may be present include amphiboles, scapolite and almandine-spessartine garnet. Biotite is the most indicative mineral of high temperature potassic alteration as K-feldspar can also form at low temperatures. The high temperature K-Fe alteration has a key relationship with the magnetite-rich deposits, and is thus most extensively developed in the deeper deposits such as Ernest Henry and Candelaria (Williams, 1999a).

Variations in fluid chemistry are capable of producing Ca- and/or Mg-rich variants of K-Fe alteration, with mineralogical characteristics that are transitional to skarn. Alteration assemblages including tremolite-anthophyllite-biotite are present at Osborne and hornblende-biotite-magnetite±epidote at Eloise and Aitik (Baker, 1998; Williams, 1999a,c).

Skarn assemblages characterised by diopside-series clinopyroxenes and grandite garnets occur in several Fe oxide Cu-Au deposits. The primary skarn assemblages are typically of high temperature (400-600 °C), and the occurrences are associated with the deeper-seated magnetite-rich Cu-Au systems. Associated minerals include amphiboles, scapolite, magnetite, carbonates and sulphides, all of which may be retrograde products of skarn. In some cases (e.g. Candelaria), the skarn assemblages seem to relate directly to the presence of carbonate rocks, whereas in others (e.g. Mount Elliott) it is possible that Ca±Mg has been added (along with Fe) from the fluid phase (Williams, 1999a,c).

In addition to enveloping (and in some cases hosting) Fe-oxide Cu-Au deposits, sodic (-calcic) alteration is also recognised as a regional alteration feature affecting large areas such as Northern Fennoscandia, the Cloncurry district and the Chilean coastal belt. At the deposit-scale, however, it is commonly the K-Fe alteration (and its variants that are Ca- and/or Mg-rich) that tend to host the ore. This applies particularly to the larger examples such as Aitik, Candelaria, Ernest Henry and Olympic Dam, which, despite displaying a range of different alteration assemblages, are all characterised by orebodies hosted by even larger volumes of rock in which plagioclase (primary and/or product of earlier Na metasomatism) has been comprehensively destroyed (Williams, 1999a).

The origin of the extensive sodic (-calcic) alteration is still the subject of much debate, and three main models have been proposed. Barton and Johnson (1996) proposed that the high-salinity (Na- and Cl-rich) fluids responsible for the sodic (-calcic) alteration were amagmatic in origin, resulting instead from igneous-driven circulation of fluids interacting with evaporites to produce the observed fluid salinities. Hitzman et al. (1992) and De Jong and Williams (1995) envisage a model whereby magmatic and/or non-magmatic fluids are drawn into intrusive-related hydrothermal systems to cause sodic (-calcic) alteration on a downward/inward heating path and potassic alteration on an upward/outward cooling path. Pollard (1999) describes an alternative mechanism caused by the unmixing of a

magmatic-hydrothermal H₂O-CO₂-salt fluid as a function of temperature and pressure. Fluid unmixing results in partitioning of CO₂ to the vapour phase and salts to the liquid phase (Bowers and Helgeson, 1983), with an increase in pH in the aqueous phase. Following unmixing, the CO₂-poor aqueous phase will have Na/(Na + K) greater than the equilibrium value for a Cl-bearing fluid in equilibrium with two alkali feldspars (i.e. in granites). In order to approach equilibrium, the fluid must contribute sodium to, and gain potassium from the wall rocks (albitisation). Due to the temperature dependence of equilibrium Na/(Na + K), isobaric cooling will result in the sodic (-calcic) alteration being succeeded by potassic alteration as temperatures fall. This can explain why early albitisation in Fe-oxide Cu-Au deposits is commonly overprinted by K-rich metasomatism.

5.1.3.4 Conditions of formation of Fe-oxide Cu-Au deposits

Fluid inclusion studies on Fe-oxide Cu-Au deposits within the Cloncurry district consistently identify the presence of both (hyper-)saline inclusions and CO₂-rich inclusions. In total, 6 types of fluid inclusions (Figure 5.9) can be recognised in the Cloncurry ore systems (Pollard, 1999; Pollard et al., 1997). The fluid system of the saline inclusions can be inferred as H₂O-NaCl-KCl-CaCl₂-FeCl₂, based on the presence of various daughter minerals, and has been sub-divided into three types of inclusion, according to homogenisation temperatures and composition. Type 1 inclusions are multiphase and have high Th (263 – 550 °C; average 450 °C) and high salinities (30 – 60 wt.% NaCl equiv.). The Type 1 inclusions have been interpreted as pre- to syn-mineralisation inclusions. Type 2 inclusions are liquid-rich two-phase (L+V) or multiphase inclusions with moderate Th (308 – 380 °C) and salinity (24 – 37 wt.% NaCl equiv.). Type 2 inclusions have only been observed as clusters within quartz grains from only a few samples from two deposits in the Cloncurry district. They have been interpreted as syn-mineralisation inclusions. Type 3 inclusions are liquid-rich two-phase (L+V) or multiphase inclusions with low Th (97 – 311 °C, mostly below 200 °C) and similar salinities to Type 2 inclusions (28 – 39 wt.% NaCl equiv.). Type 3 inclusions are the dominant inclusion type in most of the Cloncurry deposits. They commonly occur as secondary inclusions along multiple arrays of healed microfractures in quartz associated with pre-mineralisation assemblages. Type 3 inclusions are also reported as primary inclusions in fluorite and quartz which precipitated during syn- and post-mineralisation stages. Pollard et al., (1997) interprets Type 3 inclusions as being trapped during both the late-mineralisation and post-mineralisation stages. The CO₂-rich fluid inclusions have been classified as Type 4, and are commonly found in nearly all the Cloncurry ore systems. Type 4 inclusions are generally monophase, liquid CO₂, although a vapour CO₂ bubble is occasionally present. Type 4 inclusions occur either as primary inclusions spatially associated with Type 1 multiphase inclusions, or as pseudo-secondary and secondary inclusions spatially associated with Type 3 inclusions. Type 4 CO₂-rich inclusions were trapped during all stages of the hydrothermal system, from early alteration through mineralisation, to post-mineralisation stages. In a few deposits in the Cloncurry district, CH₄-rich inclusions with or without a small proportion of aqueous liquid, have been observed. These CH₄-rich inclusions have been classified as Type 5. Finally, Type 6 inclusions are H₂O-rich, containing a vapour bubble (10 – 99

vol.%) and a liquid phase. Type 6 inclusions have low salinities (<5 wt.% NaCl equiv.) and contain neither CO₂ nor CH₄.

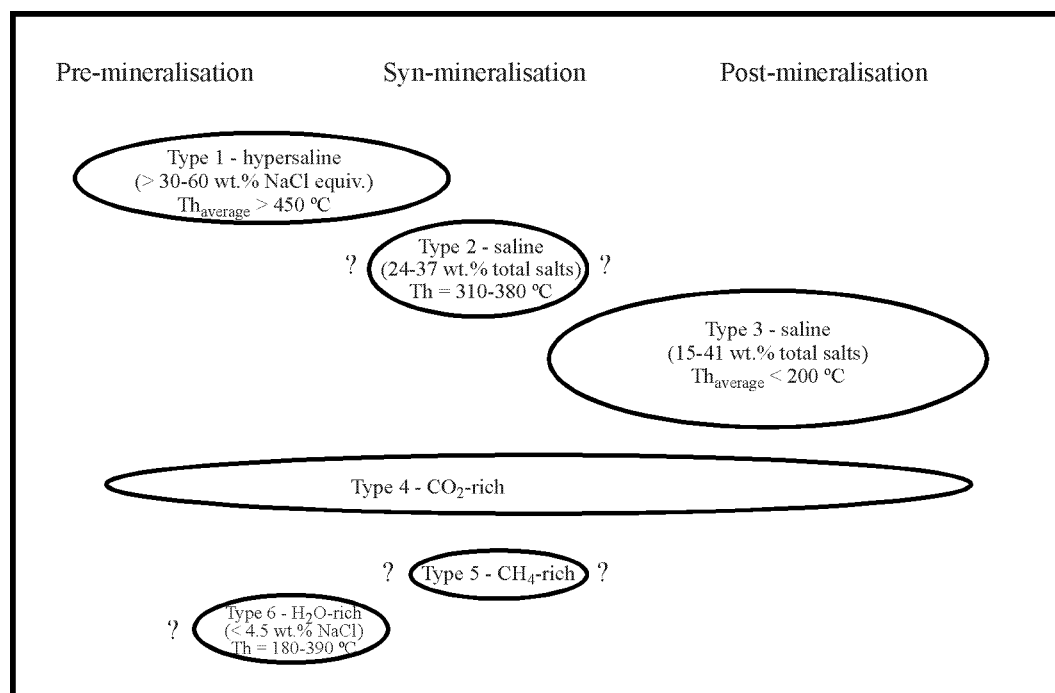


Figure 5.9: Schematic diagram to illustrate the timing of fluid inclusion types in Cloncurry mineral deposits relative to a generalised paragenetic sequence. After Pollard et al. (1997).

In summary, the fluid inclusion data from the Cloncurry district support the theory of a CO₂-rich vapour unmixing from a high temperature (> 450 °C) H₂O-CO₂-salt fluid as a result of changes in temperature and/or pressure, producing the CO₂-rich inclusions (Type 4). The hypersaline fluid evolved to progressively lower temperatures and salinities during and after mineralisation producing saline inclusions (Type 1 – 3). However, the fluid inclusions are unable to indicate whether the source of the fluids is magmatic or magmatic-hydrothermal in origin, and it is necessary to look at the stable isotope data in order to determine this.

Studies of oxygen and hydrogen isotopes in deposits of the Cloncurry district have shown that there is a remarkable similarity in calculated $\delta^{18}\text{O}_{\text{fluid}}$ and $\delta\text{D}_{\text{fluid}}$ compositions, not only from different deposits, but also from different stages of alteration and mineralisation. This similarity in fluid compositions suggests that although the timing of individual alteration and mineralisation events may vary, they ultimately share similar origins (Pollard, 1999).

The calculated $\delta^{18}\text{O}_{\text{fluid}}$ compositions of both pre- and syn-ore stage fluids within the Cloncurry district are generally between 7.4 and 11‰, and thus overlap the fields for magmatic and metamorphic water as described by Taylor (1974). Data from hydrous minerals show that there is a greater variability in the calculated $\delta\text{D}_{\text{fluid}}$ values (Figure 5.10). Hydrous minerals from the pre-ore alteration stages commonly display an extreme depletion in the calculated $\delta\text{D}_{\text{fluid}}$ values. Depletion in $\delta\text{D}_{\text{fluid}}$ values is commonly attributed to interaction with meteoric water. However, in the Cloncurry district, late-stage epidote interpreted as representing a meteoric water input suggests that local meteoric water has been enriched in δD relative to magmatic fluids (Pollard, 1999). Dong et al. (1997) suggested that the low δD at Cannington may reflect unmixing of a CH_4 -rich vapour from the H_2O - CH_4 -salt fluid. Low values from other deposits may reflect the composition of degassed magmatic fluids as suggested by Mark et al. (1999) for the Roxmere Diorite (Pollard, 1999).

In several deposits in the Cloncurry district, oxygen isotope studies have shown that the Na-Ca alteration, magnetite-quartz-feldspar veining, and sulphide mineralisation have been formed by fluids with $\delta^{18}\text{O}_{\text{water}}$ compositions similar to those of spatially related granites (Perring et al., 1997a,b; Pollard et al., 1997; Pollard, 1999). In some cases, the calculated $\delta^{18}\text{O}_{\text{water}}$ compositions of pre- and syn-mineralisation stages are slightly elevated compared to magmatic fluids. Taylor (1974) suggests an upper limit of 9 ‰ for magmatic fluids. Pollard et al. (1997) attribute this increase in $\delta^{18}\text{O}_{\text{water}}$ composition to partial isotopic exchange between a dominantly magmatic-derived water and isotopically heavier metamorphosed host rocks (Figure 5.11). With increasing fluid/rock ratios, the fluid phase becomes volumetrically more dominant, and thus the effect on the initial fluid composition is lessened. The amphiboles from the regional Na-Ca alteration typically display heavier $\delta^{18}\text{O}_{\text{water}}$ isotopic compositions (Figure 5.10), suggesting a greater degree of interaction with the host rocks than with the amphiboles from the ore deposits themselves (Pollard, 1999). When the oxygen isotope data are plotted against temperature (Figure 5.11), it becomes apparent that there is a progressive decrease in calculated $\delta^{18}\text{O}_{\text{water}}$ compositions with decrease in temperature, indicating mixing with an isotopically lighter fluid. Below about 350-300 °C, the data suggest that there is an increasing meteoric component in fluid, which typically circulated post-Cu-Au ore formation. The fluid-inclusion data also support this interpretation, with all deposits showing an early high-temperature hypersaline fluid becoming progressively less saline with decreasing temperature. This is attributed to both precipitation of abundant Cl-rich minerals and mixing with more dilute fluids (Pollard, 1999).

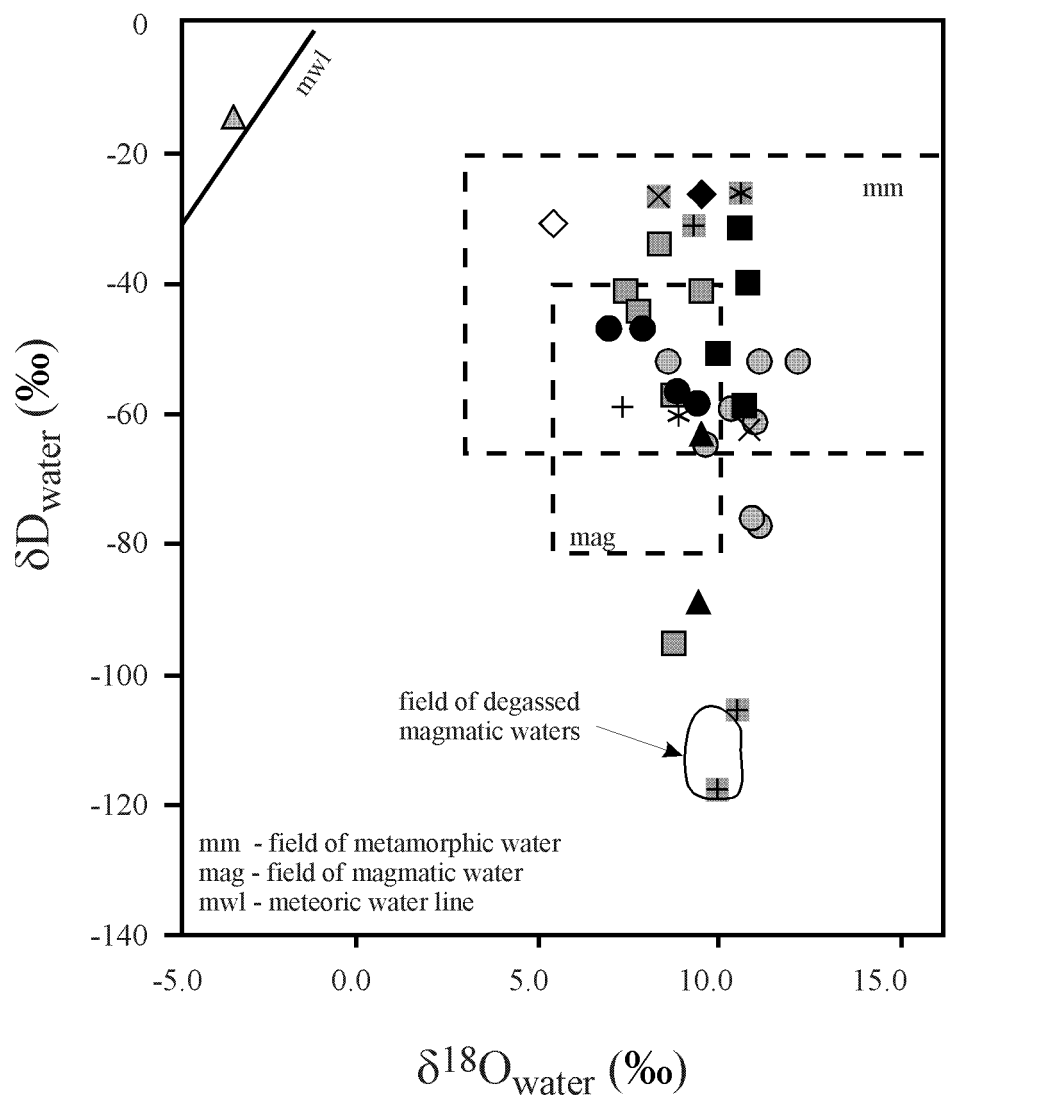


Figure 5.10: Oxygen versus hydrogen isotope data for hydrous minerals from Cloncurry mineral deposits. Also shown is the field of degassed magmatic waters from the Roxmere Diorite (Mark et al., 1999). Modified after Pollard et al. (1997).

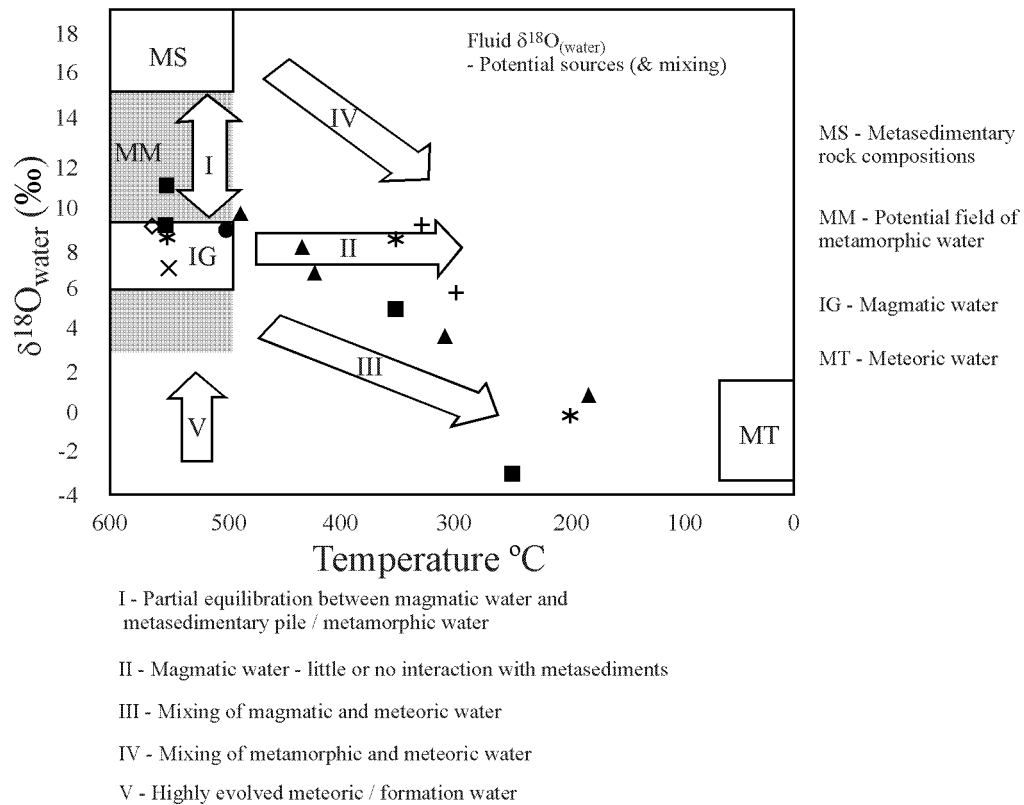
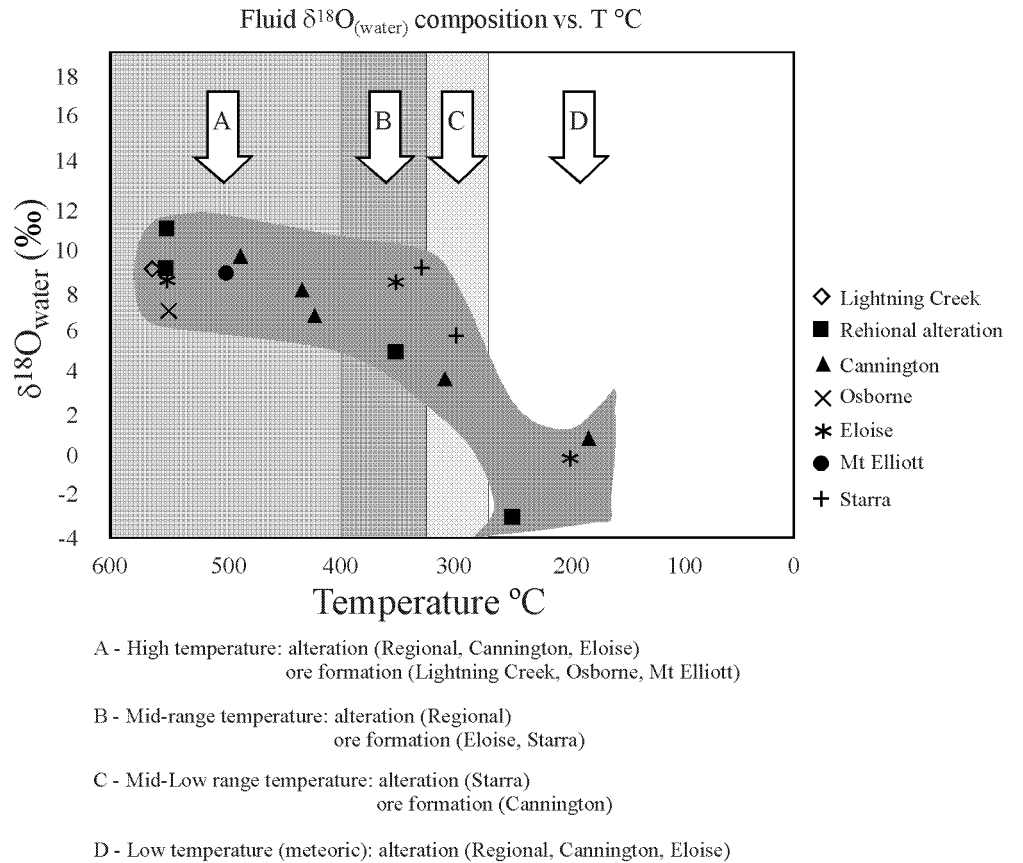


Figure 5.11: Temperature vs. oxygen isotope composition for fluids from Cloncurry mineral deposits and a range of possible sources and evolutionary pathways. After Pollard et al. (1997).

Sulphur isotope studies have been used to determine the source of the sulphur. However, as discussed in Section 3.3.8, the $\delta^{34}\text{S}$ of sulphides precipitated from an aqueous fluid is a function not only of the original isotopic composition of the sulphur-bearing fluid, but also of the relative proportions of the ionic species present, as well as pH, temperature, oxygen fugacity and the total sulphur content. Figure 5.12 shows the combined sulphur isotope data for the Cloncurry mineral deposits. With the exception of the sulphide mineralisation at Greenmount, and the Pb-Zn-Ag deposits, the sulphur isotope data for the Cu-Au deposits fall within a relatively narrow range of $\delta^{34}\text{S}$ values, with average values close to the magmatic value of 0‰ (Pollard et al., 1997; Pollard, 1999). Those deposits with negative $\delta^{34}\text{S}$ values (compared to magmatic values), are interpreted to represent sulphides precipitated at high oxygen fugacities and/or high temperatures of formation, or to reflect a mixed sulphur source, with magmatic fluids reacting with and leaching isotopically light sulphur from surrounding sediments (Pollard et al., 1997; Pollard, 1999). Conversely, elevated $\delta^{34}\text{S}$ values are likely to represent sulphide precipitation at lower temperatures and/or oxygen fugacities, or by mixing magmatic source sulphur with isotopically heavier sulphur from surrounding sediments. The sulphide mineralisation at Greenmount, (as well as the Pb-Zn-Ag deposits) are all enriched in ^{34}S relative to magmatic values. At Greenmount, the local presence of the Marimo Shale and Stavely Formation, containing isotopically heavy diagenetic pyrite ($\delta^{34}\text{S} = 15\text{-}25\text{‰}$), supports the interpretation of mixing between igneous and metasedimentary derived sulphur (Pollard, 1999). The isotopic values of the Pb-Zn-Ag deposits can be similarly explained as mixing between magmatic and sedimentary derived sulphur, or alternatively may be as a result of formation at lower temperatures, or a combination of the two.

In summary in Cloncurry, the Fe-oxide Cu-Au deposits have average $\delta^{34}\text{S}$ values close to 0‰, while those from the Pb-Zn-Ag deposits are approximately 5‰ heavier. When considered in conjunction with fluid inclusion and oxygen isotope data, the sulphur isotope data indicate a dominantly magmatic (or leached igneous) sulphur source for the deposits, with variation in individual deposits a result of differing temperatures, pH, oxygen fugacity and, most importantly for the Pb-Zn-Ag members of this deposit class, mixing with metasediment derived sulphur (Pollard, 1999).

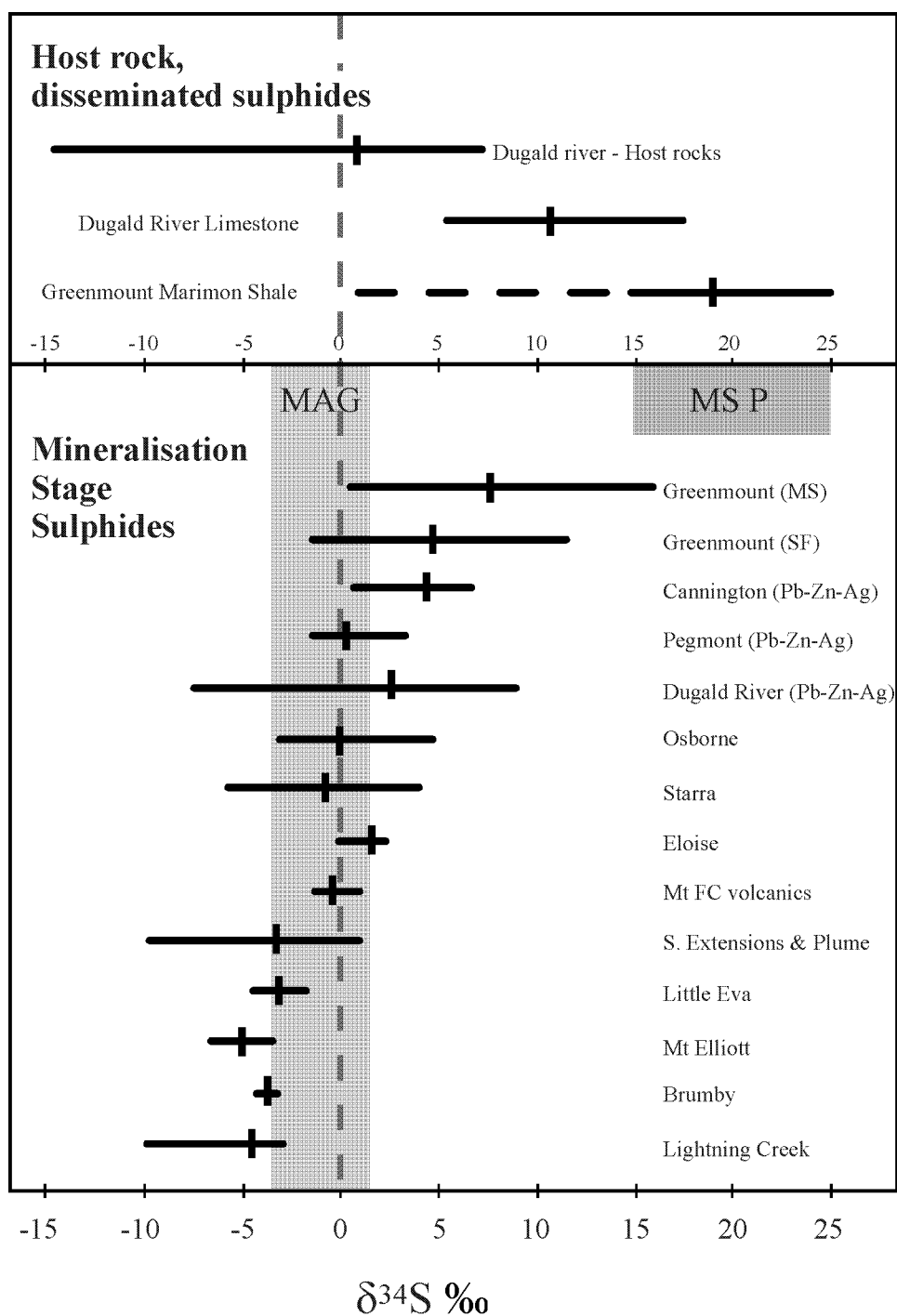


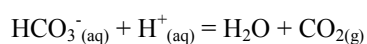
Figure 5.12: Sulphur isotope data for sulphides from a number of Cloncurry mineral deposits. After Pollard et al. (1997). MAG = range of $\delta^{34}\text{S}$ for magmatic sulphur; MS P = range of $\delta^{34}\text{S}$ from diagenetic pyrite in the local Marimo Slate (MS). SF = Staveley Fmn., FC = Fort Constantine.

5.1.3.5 Structural and physiochemical controls on mineralisation

Mineralisation in Fe-oxide Cu-Au deposits is commonly controlled by regional-scale structures which were active during mineralisation. In deeper-level environments, these structures commonly had both brittle and ductile components, and in some cases represent reactivation of older ductile shear zones. Mineralisation typically occurs in second-order structures near the main regional structures. At the local scale, mineralisation occurs in a wide variety of structural settings that include features such as fault jogs, fault intersections, intersections of faults and lithological contacts or brittle lithologies, and subsidiary structures linking strands of major fault systems. Hydrothermal intrusive breccias, break-up breccias, crackle zones and veins are the most common local fluid channelways (Pollard, 1999).

The physicochemical characteristics of the hydrothermal fluids (high temperature and high salinity) responsible for Cu-Au mineralisation in the Cloncurry district (and indeed in most Cu-Au deposits) favour transport of Cu and Au dominantly as chloride complexes (Pollard, 1999). Therefore, the most likely controls on the deposition of Cu and Au in Fe-oxide Cu-Au deposits relate to processes that contribute to decreasing temperature and pressure, changes in oxidation state and/or changes in pH. Among the variety of factors that could cause changes in these parameters, the most important appear to be fluid migration, fluid unmixing, wall rock interaction and fluid mixing. The key underlying factors are the requirements for a supply of Cu and Au in a physicochemical regime where their solubilities decreased in tandem with the availability of reduced sulphur for sulphide precipitation (Pollard, 1999).

Fluid inclusion and stable isotope data from the Cloncurry district indicate unmixing of magmatic-hydrothermal H₂O-CO₂-salt fluids in all Cloncurry Cu-Au deposits. Fluid unmixing (or boiling) results in the formation of a CO₂-rich fluid (or vapour) and a highly saline H₂O fluid. This reaction involves an increase in pH of the aqueous fluid due to the reaction:



An increase in pH decreases the solubility of metal chloride complexes (Figure 5.13), and thus contributes to the precipitation of Cu and Au. This decrease in solubility of Cu and Au chloride complexes is of a similar order of magnitude over a wide range of oxygen fugacities, which may explain similarities in Cu:Au ratios (2:1 as wt.% g/t) observed in many Cloncurry ore deposits (Pollard, 1999). Fluid unmixing in Fe-oxide Cu-Au deposits most likely occurs as a result of a combination of decompression and cooling, with decompression dominating in deposits with extensive sodic-calcic dominated assemblages, and cooling dominating in deposits with extensive potassic assemblages (Pollard, 1999).

Another method of inducing Cu and Au precipitation, is by decreasing the oxygen fugacity of the fluid. The most likely method in which this can be done is by the interaction of hot, relatively oxidised fluids with more reduced host rocks, including graphitic rocks, amphibolites and magnetite ironstones (or magnetite-rich alteration) (Pollard, 1999). A number of the Cu-Au deposits in the Cloncurry district have characteristics that suggest that mineralisation has been caused by such an interaction of oxidised fluids with more reduced host rocks. These include Starra, Tennant Creek, Ernest Henry and Osborne (interaction with magnetite-rich rocks), Cannington (interaction with graphite-rich sediments), and Greenmount and Mount Dore (interaction with carbonaceous wall rocks). However, the lack of widespread evidence for the oxidation of the wallrocks during mineralisation in a number of these deposits suggests that fluid reduction via interaction with reduced rocks was probably not the main mechanism for metal precipitation (Pollard, 1999). Pollard (1999) also suggests that precipitation of metals may be caused by fluid interaction with amphibolites, due to an increase in pH with the alteration of amphibole to biotite.

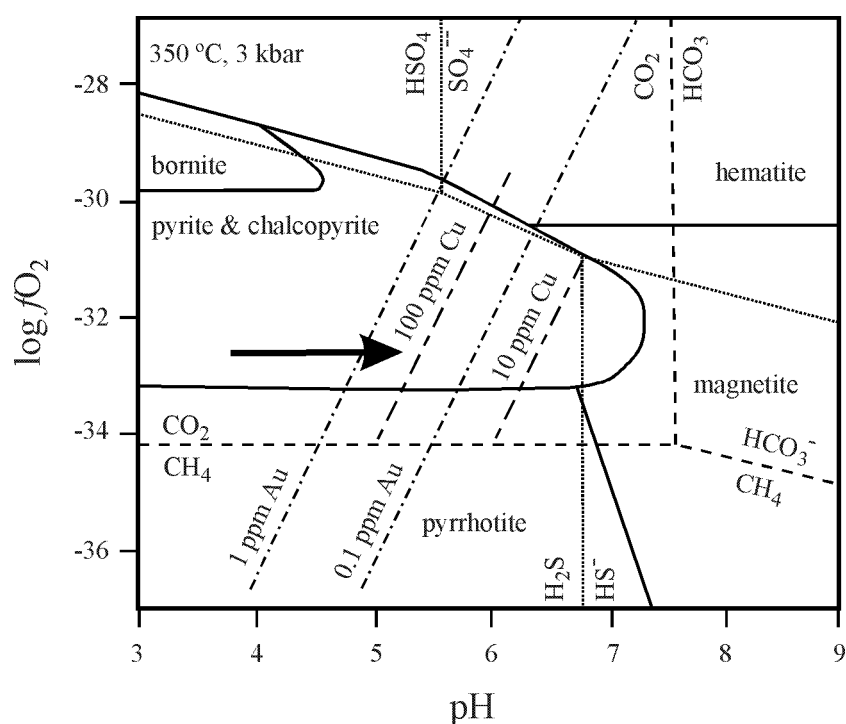


Figure 5.13: *Log fO_2 versus pH diagram for the Fe-S-O system for 350 °C and 3 kbar for 25 wt.% NaCl solution with $\Sigma S = 0.01m$ and showing Cu- and Au-chloride solubilities (diagonal dashed lines). Arrow shows possible fluid path caused by unmixing of CO_2 -rich vapour phase from an H_2O - CO_2 -salt fluid. After Pollard (1999) and Ettner et al. (1993).*

The physico-chemical conditions that are associated with the formation of Fe-oxide Cu-Au deposits are similar to those associated with a number of other deposit types including “Kiruna-type” magnetite-apatite deposits, skarns, porphyries and to certain extents, orogenic lode-gold deposits. The similarities in many of the conditions, and in particular of the sources and nature of the fluids, result in similarities in alteration and mineralisation assemblages, as well as in isotopic and fluid inclusion data. There is therefore a certain degree of overlap in the classification of these deposits, and they may be genetically

related (Pollard, 1999; Williams, 1999b). Williams (1999a) suggests that most Fe-oxide Cu-Au deposits are fault/shear-zone-controlled distal magmatic hydrothermal products of systems that may have been characterised by significant amounts of nonmagmatic fluids (Figure 5.14).

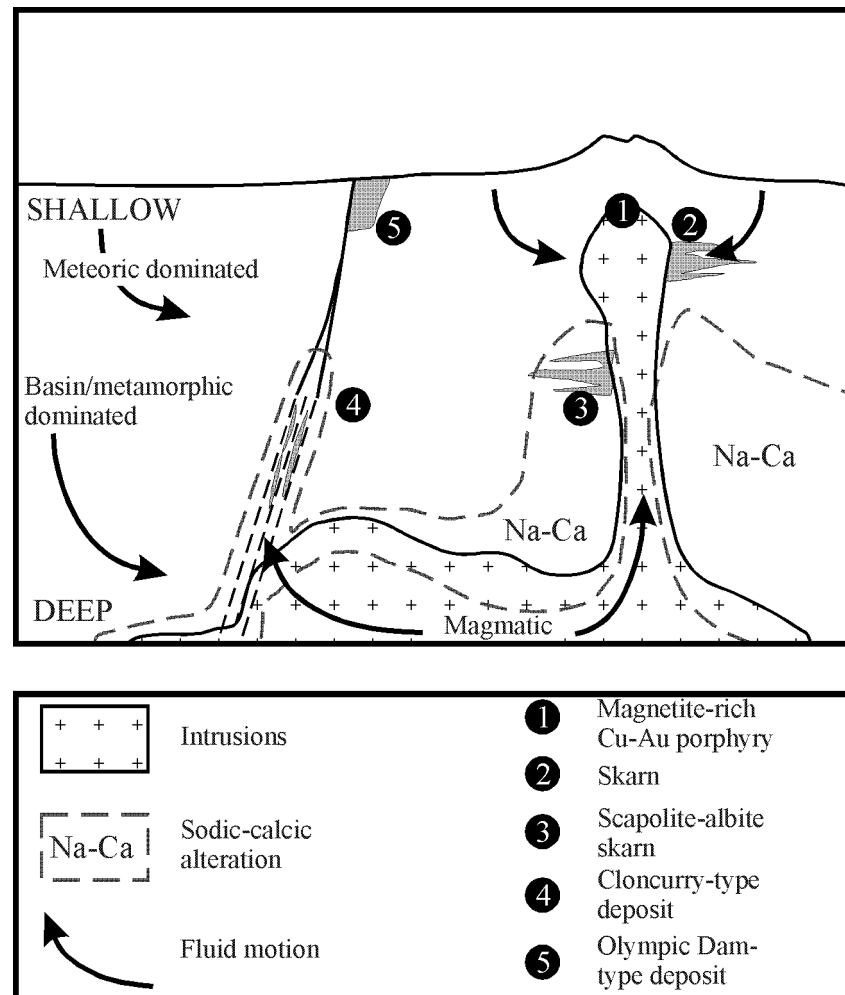


Figure 5.14: Schematic comparison of possible settings of Fe-oxide-Cu-Au deposits, “scapolite-albite iron skarns, typical iron skarns and magnetite rich porphyries (after Williams, 1999b).

5.2 DISCUSSION: A CRITICAL APPRAISAL OF THE DATA FROM HAVERI

The geology and gold mineralisation at Haveri have a number of characteristics that are common to one or more of the genetic models described above. This is perhaps not surprising, considering that a number of different processes can produce similarities in mineral assemblages or geochemistry. In order therefore, to draw valid conclusions from the available data, it is necessary to examine the data in its entirety, and to balance the results from all of the different methods and observations, before inferring conclusions on the genesis of the Haveri deposit. However, as is the nature with geology, there are always likely to be gaps in the data, and thus to arrive at a satisfactory conclusion (albeit, with reservations), a certain subjectivity in the interpretation of the data may also be required to cover the areas of uncertainty.

5.2.1 Pre-mineralisation geology, and tectonic setting

The supracrustal lithologies of the Haveri and Osara Formations have mineral assemblages and compositions that are consistent with their formation in a back-arc basin tectonic setting, with the initial formation of pillow basalts, and the subsequent deposition of volcanoclastics grading upwards into the epiclastic turbidites, deposited in the later stages of basin development as back-arc spreading declined. This basic sequence of events (Figure 5.15) fits in with a model of the Svecofennian Domain consisting of a number of allochthonous arcs being accreted onto margins of the Archaean craton (Figure 1.6). A back-arc tectonic setting for the Haveri Formation, which has been dated at 1990 ± 25 Ma (Vaasjoki and Huhma, 1987), requires the contemporaneous existence of a volcanic arc. The Tampere Schist Belt is therefore widely regarded (e.g. Kähkönen, 1989; Lahtinen, 1996; Lahtinen and Huhma, 1997) as representing a mature island arc, or a volcanic arc at an active continental margin. Evidence for this is provided by the calc-alkaline geochemistry of the volcanics in the main part of the TSB. Although older rocks have not been identified *in situ* in the Tampere Schist Belt, evidence for a pre-existing arc has been inferred from the zircon dating in the metasedimentary rocks ($\frac{1}{3}$ Archaean) (Huhma et al., 1991). The whole-rock geochemistry and, in particular, the REE and the trace element geochemistry of the Haveri Formation generally support a back-arc basin interpretation for the tectonic setting, as does its geographical position on the northern margin of TSB. Back-arc basin settings are consistent with the tectonic setting of skarn deposits enriched in Au, Cu, Fe and Co (Figure 5.5), an assemblage which is observed at Haveri. Mäkelä (1980) proposed that the Haveri Au-Cu mineralisation was syngenetic, and similar to the Cyprus-type massive sulphide deposits. During the course of this study, it has become apparent that the gold mineralisation (at the least) is epigenetic in nature, and is both spatially and temporally related to the amphibolitisation and D₂ structures. It is possible, however, that a certain proportion of the sulphides observed at Haveri, were deposited syngenetically in a similar setting to the Manus back-arc basin (Moss et al., 2001), and have been subsequently remobilised.

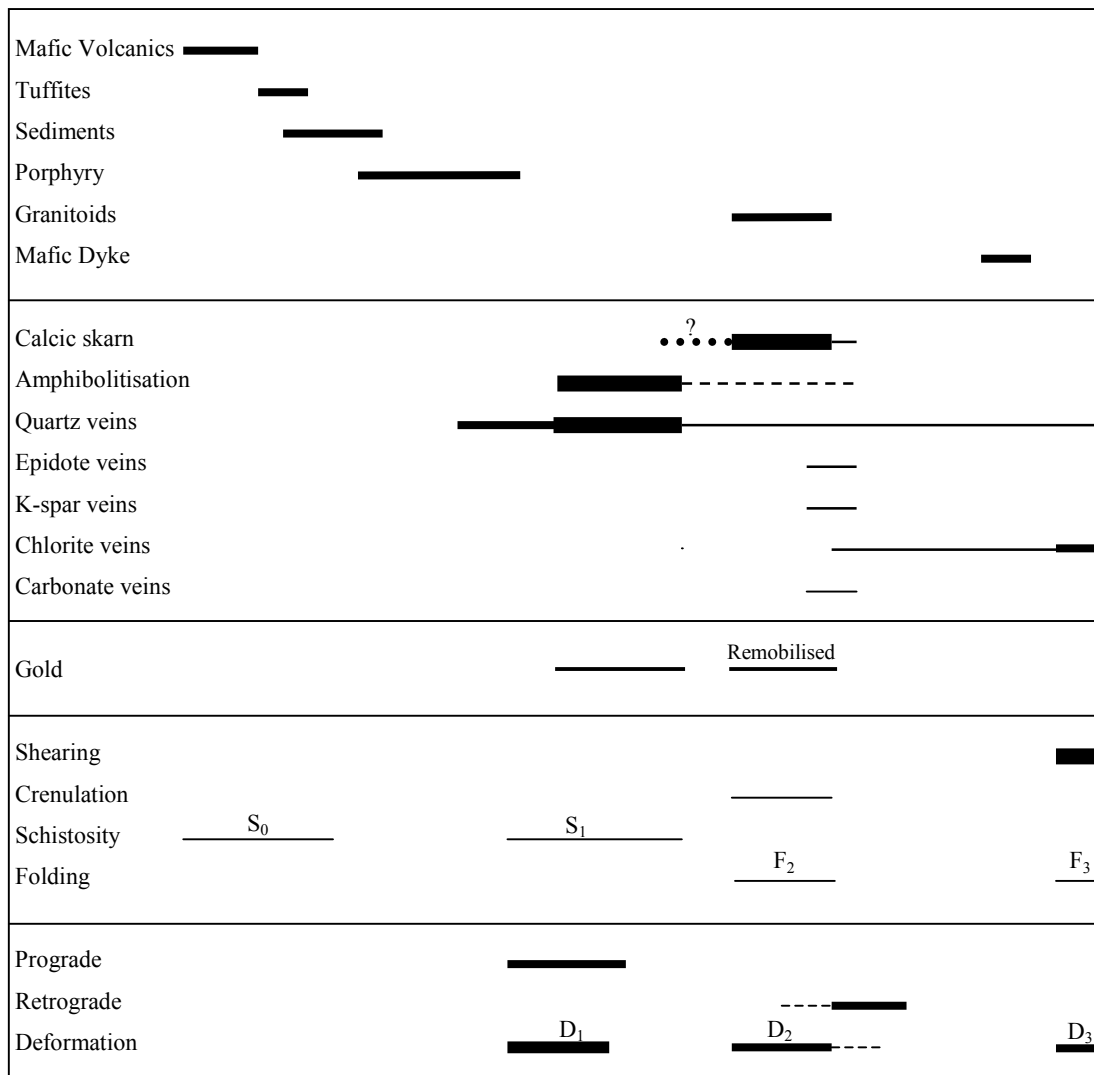


Figure 5.15: Chronological sequence of events at Haveri

This supracrustal sequence was subsequently intruded by the sub-volcanic quartz-feldspar porphyries and the voluminous granitoids of the CFGC. The quartz-feldspar porphyries have been intruded concordantly along the contact zone between the Haveri Formation and the Osara Formation. The porphyries have been folded along with the supracrustal sequence and hence were intruded before the termination of D₁. The spatial proximity between the porphyries and the Au-Cu mineralisation does pose the question as to whether they are genetically related, and whether the porphyries are the source of the mineralising fluids. This will be discussed later.

The main granitoids observed in the vicinity of Haveri are the granodiorite and the microcline granite to the east and north of the Haveri mine (Figure 2.2). Both of these granitoids are I-type, magnetite-series granitoids that form part of the CFGC. The CFGC granitoids are primarily synkinematic granitoids intruded 1.9 – 1.87 Ga ago. No radiometric dating has been attempted during the course of this study, and no record of radiometric dates for these two intrusive bodies have been found in the literature. However, radiometric ages have been determined for other intrusives within the CFGC and

for the Hämeenkyrö batholith, south of Haveri. Nironen (1989) demonstrated the synkinematic nature of the Hämeenkyrö batholith which was emplaced at 1882 ± 6 Ma (Patchett and Kouvo, 1986) during regional metamorphism and the major D₁ deformational event in this segment of the Tampere Schist Belt. This places an age constraint for prograde metamorphism and D₁ deformation at approximately 1.88 – 1.89 Ga. Nironen (1989) has also demonstrated that the batholith was affected by D₂ foliation that Clark (1997) attributes to *ca.* 1860 Ma, during the waning stages of the thermal evolution of the Svecofennide volcano-plutonic arc. The generation of the granitoids is believed to have been initiated 1.89 Ga ago by delamination of the base of the lithosphere, and replacement by a hot asthenosphere (Nironen, 1997). Decompression melting of the asthenosphere produced a mantle magma that rose to the crust-mantle boundary, remelted the lower crust and generated a variety of magmas by mixing and mingling of the crustal and mantle magmas. The high heat flow associated with crustal underplating and intraplating, and resultant magmatism, generated high-temperature low-pressure metamorphism that culminated coevally with the emplacement of the magmas and the formation of the CFGC. The emplacement of the granitoid intrusions is likely to have created steep geothermal gradients that would have been responsible for the hornfelsing observed in the country rocks, as well as the generation of hydrothermal alteration resulting from thermally driven magmatic-hydrothermal fluid convection cells. In addition, the magnetite-rich, I-type character of the intrusions is considered favourable in terms of potential magmatic-hydrothermal Cu-Au mineralisation.

The peak metamorphic conditions for the Haveri area have been calculated using a combination of geothermobarometry, mineral equilibria and petrogenetic grids, to a P-T field of 550 – 650 °C, at pressures of less than 2 kbar. This can be further constrained by the amphibolite-granulite facies transition to a maximum of approximately 600 °C and 1.5 kbar pressure, which is equivalent to an approximate depth of 5 km. These P-T conditions are consistent with those calculated by Campbell (1978), Mäkelä (1980) and Clark (1997). The oxygen fugacity of the metamorphic fluids can be constrained at 600 °C to a log f_{O_2} value of approximately –21.0 to –26.0 for the pelitic assemblages and a log f_{O_2} value of approximately –14.5 to –17.5 for the mafic metavolcanic rocks. This contrast between the more oxidized mafic metavolcanic rocks and the more reduced metasedimentary rocks, is an important chemical control for the precipitation of Au and Cu from ore-bearing fluids. The controls on ore transport and deposition will be discussed further in the next section.

5.2.2 Alteration styles and relationships

The predominant form of hydrothermal alteration observed at Haveri is the widespread amphibolitisation of the metavolcanic rocks. This is characterised by a mineral assemblage dominated by calcic amphibole, with lesser amounts of sphene, quartz, and plagioclase, as well as variable amounts of diopside and biotite. Magnetite is ubiquitous among the opaque phases commonly forming massive accumulations in intensely and pervasively altered rock. Pyrrhotite and chalcopyrite are the dominant sulphide phases associated with amphibolitisation. Utilising a combination of petrography,

whole-rock geochemistry and mineral chemistry, it is possible to characterise the alteration by significant enrichments in Fe, Ca, and Mn and a partial enrichment in K. The only significant element showing consistent depletion is Na. This is similar to the alteration described for Stage I at Paronen (Clark, 1997).

As discussed earlier (Section 3.1.3), a potential problem is that the more altered rocks display a depletion in the Na and Sr content of the rock suggesting feldspar removal, whereas the positive REE europium anomaly of the altered rocks indicates feldspar addition. Feldspar addition is also supported by the plagioclase electron microprobe analyses, that indicate sodium enrichment and albitisation in the more altered samples. This apparent contradiction can be explained in a number of possible ways. It is possible that albitisation (Na enrichment) is taking place with amphibolitisation (Fe-Ca enrichment), but that due to the predominance of calcic amphibole (containing no Na), the bulk composition of the altered rock actually reflects a Na depletion. However, as plagioclase is the only mineral in the assemblage able to accommodate Na, with increasing amphibolitisation (at the expense of plagioclase), the remaining plagioclase becomes more sodic in composition.

Another possibility is that albitisation of the mafic metavolcanic rocks occurred earlier than the Fe-Ca metasomatism, perhaps even on a regional scale, producing sodic plagioclases (i.e. similar to sub-seafloor spilitisation). In such a scenario, the later Fe-Ca metasomatism (amphibolitisation) would generate proximal plagioclase compositions more calcic than the distal sodic (albitic) compositions, thus giving the appearance of Na-depletion with increasing alteration. This is supported by the plagioclase microprobe analyses, which show the rims of the metamorphic plagioclases to be more calcic than the cores. The areas of deamphibolitisation and bleaching of the metavolcanic rock (which are spatially removed from areas of amphibolitisation) may then be characterised by preferential leaching of Fe-Mg-Ca, and thus further changing the feldspars to more sodic compositions (albitisation).

Pre-ore albitisation is a ubiquitous and global feature of Fe-oxide Cu-Au deposits, (being associated with the early regional Na-Ca alteration phase) as well as many orogenic lode-gold deposits (e.g. Davidson, 1994; Groves et al., 1998; Eilu, 1997; Hitzman, 2001; Hitzman et al., 1992; Rotherham, 1997; Williams, 1999a,c, 2001). It is therefore considered that pre-ore albitisation is the most likely explanation for the sodic plagioclases observed in the amphibolitisation, and that the plagioclase feldspars with low An contents represent remnants from this earlier alteration phase not yet fully equilibrated with the later ore-related Fe-Ca alteration. This is supported by the calcic rims which are interpreted as representing partial re-equilibration.

The whole-rock geochemistry also showed that there was a partial enrichment in K in the altered amphibolites, affecting only the less amphibolitised samples. This has been interpreted as K-metasomatism expressed as biotitisation. It is possible that the lack of K-enrichment in the more intensely altered samples simply reflects the volumetric affects of massive Fe-Ca metasomatism.

Alternatively, the biotitisation (K-metasomatism) may be either a more distal alteration phase, or may represent a later, less widespread, fluid phase. Potassic (K-Fe) alteration is a common feature associated with a number of magmato-hydrothermal alteration systems, which in deeper, hotter systems are manifested by the presence of biotite (rather than K-feldspar). At Paroinen, Clark (1997) also noted the temporal relationships between the Fe-Ca metasomatism and the K metasomatism observed in the first stage of alteration. As at Haveri, there is local evidence of biotite replacing amphibole, even though they normally appear to be in equilibrium. Clark (1997) suggested that at Paroinen, the different mineral enrichments either reflected heterogeneities in the host-rocks or varying degrees of Fe, Ca, and K metasomatism. Clark (1997) favoured a model in which early Fe-Ca metasomatism was followed by strong K metasomatism, based on alteration sequences observed in several porphyry copper deposits. This sequence has also been widely reported in Fe-oxide Cu-Au deposits, such as in the Cloncurry district of Australia (Williams, 1999a,c). The K-enrichment observed at Haveri may therefore represent a weaker expression of the K-alteration phase observed at Paroinen.

The widespread amphibolitisation (Fe-Ca metasomatism) observed at Haveri, is also closely associated with quartz veining, with amphibole veinlets laterally transforming into quartz veinlets. In addition, the quartz veinlets commonly contain amphibole mineral inclusions of similar chemistry to that observed in the amphibolitisation, as well as displaying a halo of amphibole alteration surrounding the vein. In terms of gold mineralisation, gold has commonly been observed in both the quartz veinlets and in the more intensely amphibolitised host-rocks, further supporting a genetic link between the amphibolitisation and the quartz veins. The variable presence (albeit minor) of K-feldspar, plagioclase and calcite in some of the quartz veinlets suggests that the fluids responsible for the amphibolitisation (Ca dominated) also carried minor amounts of Na and K. Therefore, it is possible that both the albitisation and the biotitisation represent slight temporal and/or spatial variations of the main Ca-Fe metasomatism.

The amphibolitisation, sulphide, magnetite and gold mineralisation, as well as the bulk of the quartz veining clearly occurred post-D₁, and pre-D₃ (Figure 5.15), and have been interpreted to have occurred in the early stages of D₂ (as the sulphide bands do show minor folding by F₂) (Section 2.5). This is similar to the timing of Stage I alteration and mineralisation that Clark (1997) interpreted for the Paroinen deposit.

The relationship between the main phase of alteration / mineralisation and the calcic-skarn alteration is not totally clear. No clear-cut field or microscope relationships have permitted a relative timing to be identified between the two alteration phases. No amphibole veins cut the calcic-skarn/carbonate zones, indicating at least that the amphibolitisation is not later than the calcic-skarn. However, no observations have been made of the calcic-skarn unambiguously cutting amphibole veinlets. There are therefore several possibilities remaining. Either the calcic-skarn is a later, separate, unrelated event to the amphibolitisation, or, alternatively, the calcic-skarn is genetically related to the amphibolitisation, and that the calcic-skarn represents: (a) a proximal part of the amphibolitisation; (b) a temporal evolution of

the fluids responsible for the amphibolitisation; (c) a product of fluid unmixing, in which a hydrous phase and a CO₂-rich phase have separated, resulting in the fluids responsible for the amphibolitisation and the calcic-skarn respectively.

Two distinct and genetically unrelated alteration events are considered unlikely. This is because the markedly different mineral assemblages of the two alteration styles, combined with the lack of any unambiguous cross-cutting relationships, would suggest that if there were two genetically distinct fluids, they must be contemporaneous. On the basis of the different *P-T-X* conditions determined for the two styles of alteration, this is considered unlikely, and thus there is likely to be some sort of genetic relationship between the two mineralogically separate events.

One of the main differences is that the amphibole alteration is always devoid of epidote, whereas epidote is ubiquitous to the calcic-skarn. Another difference is that the amphibolitisation has an alteration halo of deamphibolitisation, with a consequent enrichment in plagioclase content, whereas the calcic-skarn has an alteration halo with saussuritisation, and the presence of hydrous lower-temperature minerals such as zoisites. Finally, the amphibole alteration is commonly associated with gold, whereas the calcic-skarn is invariably devoid of gold mineralisation. These differences would argue against a simple proximal / distal (i.e. spatial) relationship. However, these mineralogical differences would also be consistent with a temporal evolution of a fluid within one broad alteration event, possibly related to the emplacement and cooling of the granitoids, with the calcic-skarn alteration being caused by a later, more evolved, lower temperature phase of the amphibolitisation. The mineralogical differences would also be consistent with a model of fluid unmixing, as suggested for the Cloncurry Cu-Au deposits (Pollard, 1999). In such a scenario, fluid unmixing (or boiling) would result in the formation of a CO₂-rich fluid (or vapour) causing the calc-silicate alteration and a highly saline H₂O fluid causing amphibolitisation.

The *P-T-X* conditions of the amphibolitisation have been constrained using a combination of stable isotopes and fluid inclusions. Oxygen isotope thermometry on mineral separates from samples of intense amphibole alteration and associated quartz veining has indicated formation temperatures in the range of 530 – 610 °C. Limited fluid inclusion studies on quartz veins associated with both amphibolitisation and gold mineralisation identified only a few “primary” looking inclusions. Thermometric studies indicate that these are low to moderate salinity (< 10 wt.% equiv. NaCl) hydrous inclusions with no detectable CO₂ phase. The homogenisation temperatures of approximately 136 – 172 °C are interpreted to reflect re-equilibration to lower P-T conditions. This is supported by decrepitation textures indicative of internal underpressure along a pathway of isobaric cooling, consistent with a tectonic model of contact metamorphism at shallow levels (low P) within the crust.

The numerous late-stage secondary trails contain a greater variety of inclusions, including primarily aqueous two-phase inclusions as well as some with solid phases indicating the presence of hypersaline fluids. The secondary inclusion trails are therefore likely to represent a phase separation, and that the

overall initial salinities of the inclusion trails are likely to have been considerably higher than the 2.4 wt.% equiv. NaCl suggested by the limited analyses. These secondary trails are consistent with the evolution of the fluid to more saline conditions. As with the primary-looking inclusions, no CO₂-rich inclusions were observed, suggesting low concentrations of CO₂ in the fluid. The trails themselves are commonly randomly orientated (although parallel planes are also not uncommon) and cut across both grain and sub-grain boundaries in the quartz, indicating that they are relatively late in terms of the deformation, and thus are likely to be at least late-D₂ deformation. These secondary trails are similar in morphology and composition to the Type 3 fluid inclusions reported for the Fe-oxide Cu-Au deposits of the Cloncurry district (Pollard et al., 1997).

The *P-T-X* controls on the calcic-skarn alteration assemblage are less well constrained, although phase equilibria put a maximum temperature at 560 °C (based on $X(\text{CO}_2) = 1$), which is clearly an overestimate. At an $X(\text{CO}_2) = 0.1$, the maximum temperature for the assemblage is approximately 515 °C. The presence of calcite and scapolite in the proximal assemblage of the calcic-skarn is indicative of a CO₂-rich fluid, which would have an $X(\text{CO}_2)$ somewhere between the two examples used above. This corresponds to a formation temperature of between 515 and 560 °C. This compares favourably with temperatures calculated from sphalerite compositions in the range of 500 – 600 °C. Although it has not been possible to determine the values of all of the physico-chemical variables (*P*, *T*, $X(\text{CO}_2)$, $f(\text{O}_2)$, $f(\text{S}_2)$, pH, mineral solid solutions and activities of non-volatile aqueous species) applicable to the calcic-skarn assemblage, the maximum temperatures calculated do indicate high-temperature formation, albeit slightly lower than those calculated for amphibolitisation (530 – 610 °C). In addition the presence of scapolite in the proximal skarn assemblage does indicate that they were formed from CO₂-rich saline fluids.

These limited results are not at odds with a theory whereby the calcic-skarn represents a later, more evolved, lower-temperature phase of the amphibolitisation. However, as before, the lack of an unambiguous cross-cutting relationship between the two alteration events, although not discounting this theory, would tend to argue against the calcic-skarn event being later than the amphibolitisation. It is therefore considered more likely that the CO₂-rich fluids responsible for the calcic-skarn are the product of fluid unmixing as suggested by Pollard (1999) from the fluids responsible from the amphibolitisation. It is also possible that the calcic-skarn alteration represents the late-D₂ phase of evolved fluids in the secondary inclusion trails described above. However, it should be noted that there is no evidence at this stage to support (or refute) such a theory, and further studies need to be completed in order to verify this.

5.2.3 Gold Mineralisation at Haveri

The gold mineralisation at Haveri is related primarily to the amphibolitisation stage of alteration, with gold being hosted by the intensely and pervasively amphibolitised volcanics and volcanoclastics, as well as in the associated quartz veins. These features are considered to be syn- to post-D₁. Gold mineralisation is also heavily concentrated in the sulphide bands and cataclastites, with lesser concentrations along the contact with the porphyries, as well as in other low strain zones such as breccias. These features are all interpreted to be post-D₁ and pre- to syn-D₂, although the shearing of the cataclastites (and hence potential remobilisation of the gold) may have continued post-D₂.

The location and distribution of the gold in the various ore types, and absence of visible gold in the sulphide ore types has important implications on the genesis of the gold. The consistent inability to microscopically identify visible gold in the sulphides with the lower gold grades (< 1.5 g/t Au), supports the notion that there is “invisible gold” in the sulphides, at least at the lower grades. At higher gold grades (> 6 g/t Au), the gold is visibly associated with amphibolitisation and quartz veins. At intermediate gold grades (2 – 6 g/t Au), where there is massive sulphide associated with amphibolitisation and quartz veins, and the location of the gold can still not be identified, there are several possibilities. Either the gold is located within the quartz veins and amphibolitised host rock, and due to the nuggety nature of the gold (as observed elsewhere), is just not being observed in certain sections, or is distributed too finely to be recognised. The second possibility is that the gold is hosted within the sulphide, and is either too fine, or is in solid solution with the sulphide.

If the gold is in the quartz veins and amphibolitised host rock, then there are two possible sources for the gold. The gold may have been introduced into the system with the hydrothermal fluids responsible for the amphibolitisation and quartz veins (which may or may not have a genetic relationship with the massive sulphide). Alternatively, the fluids responsible for the amphibolitisation and quartz veins may have remobilised “invisible gold” from pre-metamorphic sulphides. This latter scenario excludes the cataclastites and main massive sulphide bands as a possible source of the gold, as they have been shown to post-date the emplacement of the porphyries, as well as D₁ deformation (Section 2.5). However, the unaltered host rocks (both the metavolcanic rocks and the metasedimentary rocks) typically have Au concentrations below 0.01 ppm. This therefore suggests that the gold was introduced to the system during the amphibolitisation event, and that the likely source of the gold (and possibly the other metals) is external in origin, such as the granitoids, rather than the supracrustal rocks. However, it cannot be discounted that there was a pre-concentration of Au within the supracrustal sequence, as has been shown to be the case in the Manus back-arc basin (Moss et al., 2001), which has subsequently been leached, transported and redeposited by the hydrothermal solutions. The ultimate “source” of the gold at Haveri is unlikely to be definitively solved, however, there is little doubt that it is during the amphibolitisation (syn- to post-D₁) and associated quartz veining that the majority of the gold was transported and deposited, with possible remobilisation of the gold during the cataclastite formation (pre- to syn-D₂) (Figure 5.15).

The fluids responsible for the amphibolitisation and the quartz veins (as discussed above) are therefore likely to represent the primary ore-bearing fluids. On the basis of fluid inclusions and oxygen isotope thermometry, these can be characterised as high-temperature (530 – 610 °C), low- to moderate-salinity, low CO₂, fluids which may possibly display an evolution of increasing salinity (from the secondary inclusion trails in the quartz veins), and into higher CO₂ contents (calcic-skarn).

The presence of H₂O-CO₂ inclusions, which contain both liquid and vapour CO₂, and low salinities are a ubiquitous feature of orogenic lode-gold deposits, and are generally taken to infer generation in a deep (at least several kilometres) environment (Groves et al., 1998). Conversely, halite-bearing fluid inclusions, coexisting with vapour-rich inclusions containing little or no CO₂ would indicate a relatively shallow environment, probably associated with emplacement of silicic plutons (Roedder and Bodnar, 1997). Despite the limited nature of the fluid inclusion studies at Haveri, the data support formation from fluids with low CO₂ contents, and salinities which are at least low to moderate, but which may in fact be considerably higher depending on the timing of the secondary inclusion trails. Such inclusions are therefore likely to represent an ore genetic model with a shallow, magmatic affinity, as opposed to a deep, crustal source. Such a fluid source is also more in line with the regional geotectonic environment, of an evolving arc on a continental margin, with voluminous high-level granitoid emplacement. A magmatic-dominated fluid source is also supported by the oxygen-hydrogen isotope studies of the amphiboles. These data suggest a predominantly magmatic source for the fluids which has had a minor interaction with meteoric fluids.

There are two possible magmatic sources for the ore-bearing fluids: the porphyries and the granitoids. The amphibolitisation (and mineralisation) at Haveri has been shown to affect the quartz-feldspar porphyries (generating the uraltite porphyries), and thus there are two possibilities. Either, the amphibolitisation is genetically related to the porphyries, and the alteration in the porphyry represents an endoskarn, or alternatively the alteration and the porphyries are genetically unrelated, and the amphibolitisation is a later, separate event, related to the emplacement of the granitoids. The fact that only the narrow porphyries display significant alteration, and that the larger porphyry bodies remain unaltered, does suggest that the alteration is an external event that post-dates the porphyries. This is also supported by considerable circumstantial evidence. The Au mineralisation has been clearly observed in the amphibolitisation, as well as in the sulphides and quartz veins, all of which affect the porphyries. However, these features are not restricted to the immediate vicinity of the porphyries, and, in many cases where there is significant mineralisation, no porphyry body has been observed. In addition, the porphyries appear to be volumetrically minor in relation to the extent of the widespread amphibolitisation. Although at this stage it is not possible to quantify the full spatial extents of the alteration (due to a fragmented geological picture, and unknown structure), it is apparent that the observed distribution of the alteration would require a volume of fluid flow unlikely to be generated by the porphyries.

Isotopic similarities between the quartz veins and the porphyries, if supported by other evidence would add credence to the notion of genetic relationship. But in isolation, is not considered sufficient evidence for this, rather it is considered that the similarity may be due to isotopic exchange or buffering or is merely coincidental.

The spatial association of the mineralisation (and alteration) with the porphyries is therefore considered to be genetically unrelated, and is more likely to be caused by the ore-bearing fluids exploiting the relatively low stress areas provided by the competency contrasts between the more competent porphyries and the more plastic metavolcanic rocks, metatuffs and metasedimentary rocks. Such structural controls on mineralisation are especially common in orogenic lode-gold and Fe-oxide Cu-Au systems. The nearby, voluminous, granitoids have a geochemistry (I-type, magnetite-series) which is considered favourable for the generation of hydrothermal fluids capable of forming ore deposits enriched in Cu and Au, and are thus considered the most likely source for the ore-bearing fluids. Furthermore, the emplacement of such large plutonic bodies is far more likely to generate sufficient fluids and heat to drive a hydrothermal system large enough to account for the extent of the alteration observed at Haveri, and to have occurred over enough time to generate an economic concentration of metals.

The ore-carrying capacity of a hydrothermal fluid is largely determined by the physico-chemical conditions of the fluids, and the activity of the main ligands that form metal complexes, rather than the abundance of the metals themselves (e.g. Pirajno, 1992). Of the numerous ligands which are able to form stable gold complexes, there are (on the basis of their abundance in hydrothermal fluids) two main classes of complexes important for the transport of ore metals in hydrothermal solutions, namely: bisulphide complexes and chloride complexes. Both these complexes are capable of transporting large quantities of metals in solution, although under different *P-T-X* conditions. Other complexes involved in the transport of gold include tellurium and thioarsenide complexes (Romberger, 1990), which are not considered applicable for gold at Haveri. This is because no significant Te or As was encountered during the recent gold exploration at Haveri (Strauss, 1998).

Numerous studies on the solubility of gold in hydrothermal solutions under different *P-T-X* conditions all broadly agree that bisulphide complexes are dominant at lower temperatures, less than approximately 350 – 400 °C. Above these temperatures chloride complexes become the dominant gold-bearing species (Figure 5.16) (Henley, 1973; Seward, 1973; Barton, 1980; Shenberger and Barnes, 1989; Zotov and Baranova, 1989; Hayashi and Ohmoto, 1991; Gammons and Williams-Jones, 1995, 1997; Benning and Seward, 1996; Seward and Barnes, 1997; Gibert et al., 1998). The actual boundaries where one species becomes dominant over another vary, depending on the particular study and the *P-T-X* conditions assumed. In a temperature range of 250 – 400 °C, gold hydrosulphide complexes (AuHS^0) become the dominant gold-bearing species in most ore-forming conditions and at low pH conditions, although the gold thiosulphide species $[\text{Au}(\text{HS})_2^-]$ becomes more important at near neutral to weakly alkaline pH (above 5.5) and low salinity conditions (Seward and Barnes, 1997;

Gibert et al., 1998). At higher temperatures ($> 400\text{ }^{\circ}\text{C}$), gold thiosulphide $[\text{Au}(\text{HS})_2]^-$ only plays a part in gold transport at higher pH conditions (Seward and Barnes, 1997). The solubility of gold in AuHS^0 species is independent of pH and salinity (Gammons and Williams-Jones, 1997). However, the solubility of gold in all bisulphide complexes is very sensitive to the activity of reduced sulphur species (H_2S), and the concentration of reduced sulphur atoms in solution must be far greater than that of the metals, if the complexes are to remain stable.

At temperatures in excess of about $350\text{ }^{\circ}\text{C}$, gold chloride species (AuCl_2^-) become important only in H_2S -poor brines, but at temperatures in excess of $400\text{ }^{\circ}\text{C}$, and in acidic solutions, gold chloride species become the dominant mechanism for Au transport (Gibert et al., 1998). Gold solubilities in chloride complexes increase with increasing temperature, and highest gold solubilities ($> 100\text{ ppm}$) are achieved by fluids that are hot ($\geq 500\text{ }^{\circ}\text{C}$), oxidised, highly saline, and acidic, and that have a high KCl/NaCl ratio (Gammons and Williams-Jones, 1997).

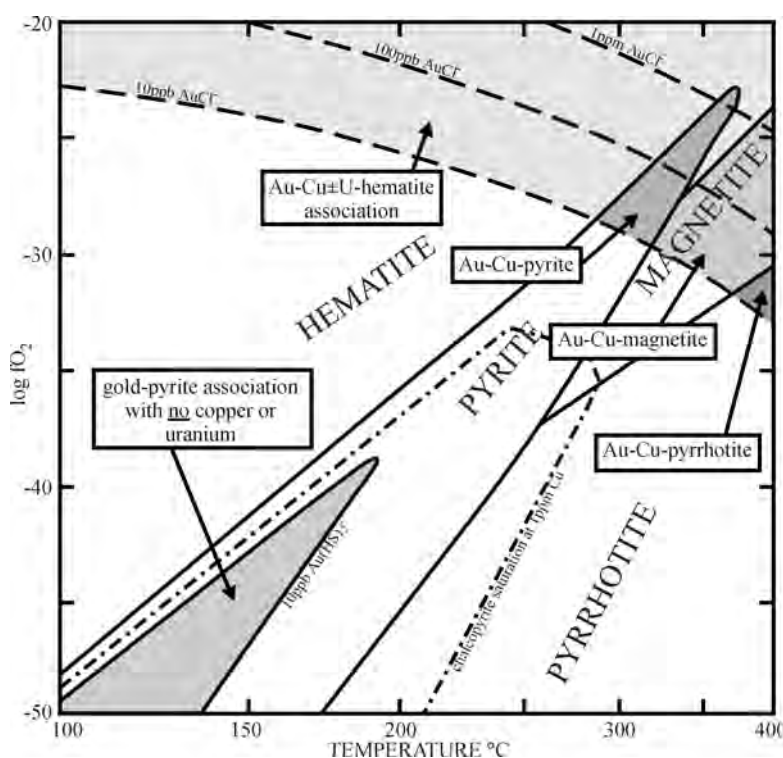


Figure 5.16: Fields of gold transport and deposition with Cu and Fe, in $f\text{O}_2$ -temperature space. The gold-copper association is controlled by the stability of AuCl_2^- at high temperature and $f\text{O}_2$. The gold-copper-pyrrhotite and gold-copper-magnetite associations are absent where gold transport is controlled by $\text{Au}(\text{HS})_2^-$. Conditions are $a_{\text{S}} = 10^{-3}$, $a_{\text{Cl}} = 10^{0.5}$, $\text{pH} = 4$. After Davidson and Large (1994). The Haveri data plot to the right of this diagram at higher temperatures, and into the expanded Au-Cu-magnetite and Au-Cu-pyrrhotite fields, depending on the local $f\text{O}_2$ conditions: the higher $f\text{O}_2$ conditions of the metavolcanic rocks leading to the Au-Cu-magnetite association; the lower $f\text{O}_2$ conditions of the pelites leading to the Au-Cu-pyrrhotite association.

Gold can be deposited from sulphide complexes by a number of geological mechanisms that cause a reduction in the solubility of the bisulphide gold complexes. A reduction in the activity of H_2S (and hence gold solubility) in the hydrothermal fluid can be caused by direct precipitation of sulphides or sulphate, dilution and mixing with an H_2S -poor brine (seawater), boiling, sulphidisation of the wall rocks or a change in $f(\text{O}_2)$. Boiling or phase separation produced by sudden pressure release (perhaps assisted by increase in temperature and/or volatile build up), results in water, H_2S and acidic volatiles partitioning into the vapour phase, thus decreasing the H_2S contents and increasing the pH of the metal-bearing fluid, causing metal deposition (Seward and Barnes, 1997). During boiling, $f(\text{O}_2)$ also increases, and the oxidation ($\text{HS} \rightarrow \text{H}_2\text{S} \rightarrow \text{H}_2\text{SO}_4$) and destruction of thiocomplexes will take place, due to the rapid disappearance of HS (Pirajno 1992). The principal mechanisms for triggering gold deposition from chloride complexes are by reduction in temperature (the most important), decreased chloride concentration (e.g. by fluid mixing), increasing H_2S concentration, and by acid neutralisation by increase in the pH (e.g. caused by boiling or phase separation) (Gammons and Williams-Jones, 1997; Seward and Barnes, 1997).

It has long been suggested that both chloride and sulphide complexes are active within a hydrothermal system, albeit over separate P - T - X - t conditions. If a high temperature, gold-chloride enriched, fluid (e.g. magmatic) was cooled, gold solubility would steadily decrease, until such a point where the dominant complex switches to a bisulfide, at which point gold solubility may actually increase with cooling. In addition, if fluid immiscibility occurs at high temperature ($> 400^\circ\text{C}$), the gold will follow chloride into the brine phase, and water, H_2S and any acidic volatile phases will partition into the vapour phase. As this vapour rises and cools it may recondense, forming a sulphur-rich, chloride-poor solution capable of dissolving gold and transporting it to shallower crustal levels (Gammons and Williams-Jones, 1997). Thus in a downward collapsing magmatic-hydrothermal system, retrograde events that produce sulphur-rich, chloride-poor solutions may remobilise gold into the epithermal environment.

From the above discussion it is apparent that different complexes are responsible for transporting the gold (and other metals) under varying P - T - X conditions, and that it is therefore possible to predict which gold-complexes (and hence depositional controls) may be more or less applicable to different ore deposit types. Consequently, the high-temperature, high-salinity fluids, dominated by chloride complexes, are most valid in hotter and deeper regions of magmatic and hydrothermal systems, and in hydrothermal fluids originating during metamorphic dewatering. Sulphide complexes become dominant in the lower temperature ($< 350^\circ\text{C}$) environments (e.g. epithermal) as well as those with low salinities (e.g. orogenic lode Au deposits) (Pirajno, 1992; Seward and Barnes, 1997).

At Haveri, considerable evidence supports the transport of gold in hydrothermal solutions as a gold chloride complex, especially where the gold is associated with the amphibolitisation and the quartz veining. Such evidence includes the high temperatures ($> 500^\circ\text{C}$) of formation, saline fluids, high fluid $f\text{O}_2$ and low activities of sulphur, as indicated by the abundant magnetite and the high Au:Ag

compositions of the gold (electrum) grains. However, it is also possible that the sulphide cataclastites represent a lower temperature remobilisation of gold in fluids with a higher sulphur content, and in which gold is transported as a bisulphide complex. Unfortunately, the actual siting of gold within the cataclastites is still undetermined.

As mentioned previously, the gold is located in zones of intense amphibolitisation, in quartz veinlets, in cataclastites / massive sulphide bands, in brittle breccias, along the boundaries of porphyries, and in a much broader sense along the lithological contact between the metavolcanic rocks and the overlying metasedimentary rocks (i.e. in the tuffs). Therefore, it appears that there are potentially a number of depositional mechanisms applicable for the Au-Cu mineralisation at Haveri. As discussed above, the gold in the amphibolitisation (and associated quartz veins) is most likely to have been transported as chloride complexes. A decrease in temperature is one of the principal mechanisms by which a gold chloride complex becomes less soluble in the hydrothermal fluid. Haveri is situated less than a kilometre from the granitoids of the CFGC, and is thus likely to be along a steep temperature gradient away from the intrusive heat source. Consequently, one may expect that cooling as a mechanism for deposition would be significant. Although a decrease in temperature is one of the principal mechanisms for depositing gold from a chloride complex, cooling alone, without any mechanism for trapping the fluids, is unlikely to produce significant metal concentrations. Fluid traps may be structurally and/or lithologically controlled. Gold enrichment along the porphyry contact is perhaps an example of this style of deposition. The contact between the porphyry and the supracrustal rocks, would have acted as a fluid conduit. Trapping of the fluids would have taken place in structural traps resulting in the decreased permeability of the surrounding rocks (such as in antiforms), and/or in zones of low strain provided by the competency contrast of the porphyry and the supracrustals. Such traps may also be one of the causes for pooling of the fluids leading to pervasive amphibolitisation.

Another important mechanism for the deposition of the gold is through fluid unmixing as suggested for the Cloncurry ores (Pollard, 1999). Fluid unmixing is a function of both temperature and pressure, where fluid saturation (due to cooling in low strain zones) of brecciation and pressure release (due to fluid overpressure and structural fracturing) cause the fluid to separate, with the formation of a CO₂-rich aqueous vapour (or fluid) phase and a highly saline aqueous fluid phase. This fluid unmixing results in an increase in the pH of the saline fluid phase, decreasing the solubility of chloride complexes, and contributing to the precipitation of gold and copper.

Another potentially important mechanism for the deposition of the Haveri ores is provided by the REDOX contrasts between the relatively oxidised amphibolites ($f(\text{O}_2) = -14.5$ to -17.5 at 600 °C) and the relatively reduced metasedimentary rocks ($f(\text{O}_2) = -21.0$ to -26.0 at 600 °C). The interaction of a hot, relatively oxidised ore-bearing fluid with reduced host rocks, or with fluids equilibrated with reduced host rocks (fluid mixing), will decrease the solubility of chloride complexes, and further contribute to the precipitation of gold and copper. The location of the Haveri mineralisation along the

broad contact between the metavolcanic and the metasedimentary rocks, as defined by the interlayered tuffites, is therefore a likely contributory factor for the gold-copper mineralisation.

Finally, if the cataclastites / massive sulphide-hosted gold does in fact represent a remobilisation of the gold by bisulphide complexes, then gold deposition may be induced by direct precipitation of sulphides or sulphate, dilution and mixing with an H₂S-poor brine (seawater), boiling, sulphidisation of the wall rocks or oxidation of the fluid. The brecciation and sulphides associated with this style of mineralisation would suggest that boiling processes and the direct precipitation of sulphides would be the dominant depositional mechanisms for the Haveri cataclastite / massive sulphide bands.

There thus appear to be several possible causes for the mineralisation at Haveri, with a combination of lithological (REDOX) control, structural control (providing permeability traps and decompression related precipitation), and temperature decrease causing the mineralisation. The necessity for the structural traps for cooling fluids, reducing permeability, and causing fluid unmixing (by pressure release and low strain zones) suggests that, at the deposit scale at least, structural controls are the primary causes for ore deposition, and are critical for providing effective traps for the ore. However, the REDOX conditions are clearly an important contributory factor in ore-deposition.

The association of Au with Cu ($\pm$ Mo), as observed at Haveri, is common in I-type porphyry environments formed in island-arc settings. The sources of the metals in such scenarios has also been a subject of much debate, with two main schools of thought (Pirajno, 1992): one envisages that the metals originated in the source region of the magmas, with subsequent assimilation of crustal material and scavenging of metals. In such a scenario, high Mo/Cu ratios would indicate a thick sialic crust, whereas, high Au and high Cu/Mo ratios is indicative of a thin and more mafic crust. In the latter case, the source of the metals would be provided by the more mafic components of the crust such as oceanic crust and the predominantly dioritic plutons. The second school of thought suggests that a subcrustal source might be involved and is the prime supplier of the metals. In such a scenario, the metal sources are believed to include the subducted oceanic crust, the upper mantle wedge and possibly the zone of underplating by the rising mafic melts beneath the crust.

Although this discussion on the source of metals in porphyry environments is still unresolved, it is perhaps likely that a combination of the two processes are at work. Likewise at Haveri (with similarities in metal associations, and to a certain extent, tectonic setting), it is possible that a combination of processes are at work, with the metals being supplied by both leaching of the country rocks combined with magmatic sources. However, a definitive answer on the source of the metals is likely to remain unresolved.

The alteration and mineralisation observed at Haveri is better interpreted as a progressive evolution of a single event, rather than as individual events. To summarise then, the alteration commenced with a pre-ore albitisation phase (Na-Ca alteration) that was closely followed by a strong Ca-Fe alteration which

was associated with Au-Cu-magnetite mineralisation with the ore-minerals being transported as chloride complexes at high temperatures ($> 400\text{ }^{\circ}\text{C}$). Metal deposition was induced by a variety of mechanisms including mainly structural controls, resulting in local variations in temperature, pressure and pH of the fluids, and consequent metal precipitation. A broad chemical control also contributed to metal deposition caused by a reduction of oxidised fluids at the contact with the relatively more reduced metasedimentary rocks. Post-ore phase alteration includes an evolution to more K-rich alteration (biotitisation) and to the development of the lower temperature calcic-skarn alteration. This evolution of the alteration stages and the mineralisation is interpreted to be broadly synchronous with early, pre- to syn- D_2 deformation (with the calcic-skarn phase perhaps outlasting D_2) and the emplacement, crystallisation and cooling of the granitoids (Figure 5.15).

At a later stage, post- D_2 and pre- D_3 , the carbonate-barite veins were formed represents an evolution to more oxidised conditions and lower temperatures, as evident from the presence of sulphate and pyrite (rather than pyrrhotite). These are interpreted as very late emanations from the cooling granitoids (supported by C-O isotopes and field relationships), perhaps interacting with meteoric waters. This stage was followed by the emplacement of the narrow dolerite dykes into the cooled stratigraphy (chilled margins) and then a weak D_3 deformation, resulting in shearing and open folding of the dykes, and structural reactivation along the carbonate-barite bands. The D_3 was accompanied by pyrite deposition.

5.3 A GENETIC MODEL FOR GOLD MINERALISATION AT HAVERI

The data obtained from Haveri indicate that there are a number of similarities with ore deposit models that include orogenic lode-gold, Au-skarn and Fe-oxide Cu-Au deposits. The structural controls on mineralisation observed at Haveri are similar in many respects to those structures in the orogenic lode-gold deposit class (Mueller and Groves, 1991; Eilu et al., 1997, 1999; Partington and Williams, 2000). In addition, the isotopic data from Haveri are not dissimilar to those for many orogenic lode-gold deposits. Furthermore, the alteration assemblages and metamorphic temperatures are broadly compatible with the deeper, hypozonal orogenic lode-gold deposits formed at and/or subjected to mid-amphibolite facies metamorphism. However, the calculated maximum pressure for Haveri is 1.5 kbar, which is equivalent to a depth of only 5 km. This is too shallow a depth for hypozonal orogenic lode-gold deposits to be formed, that produce the mineral assemblages that are observed at Haveri. In the case of amphibole-pyroxene-dominated assemblages, a formation depth of at least 10 km is required (Mueller and Groves, 1991). Furthermore, the fluid inclusion studies at Haveri have failed to reveal the presence of CO₂-rich fluid inclusions, which are ubiquitous to orogenic lode-gold deposits, and have indicated the presence of hypersaline fluid inclusions with daughter minerals (albeit in secondary trails). These features are strongly indicative of a magmatic source for the hydrothermal fluids. A magmatic source for the hydrothermal fluids, combined with a relatively shallow formation depth, although not ruling out formation as an orogenic gold deposit, do strongly support a magmatic-hydrothermal ore genetic model, such as suggested for many skarns and Fe-oxide Cu-Au deposits.

Another important difference between Fe-oxide Cu-Au deposits and orogenic lode-gold deposits is the tectonic setting. Whereas orogenic lode-gold deposits are typically formed in compressional or transpressional tectonic regimes (Groves et al., 1998), Fe-oxide Cu Au are more typically formed in extensional tectonic regimes near destructive plate margins or in intracratonic settings (Hitzman et al., 1992; Pollard, 1999). Nironen (1997) suggested that from about 1.89 Ga, an extensional tectonic regime existed across the Central Svecofennian, resulting from the delamination of the lithosphere and asthenospheric upwelling, resulting in the numerous granitoids of the CFGC. Furthermore, the Tampere arc itself was the site of extensional tectonism 1.87 Ga ago (Figure 1.6d). It is therefore apparent that on the basis of tectonic setting, Haveri has more in common with Fe-oxide Cu-Au deposits than with orogenic lode-gold deposits.

The Haveri deposit shows a number of similarities with skarn, including the calc-silicate nature of the proximal alteration assemblage (being dominated by calcic-amphibole after clinopyroxene), and the distal potassic alteration. The mineral chemistry of the garnets and pyroxene are indicative of the more reduced versions of gold skarn deposits. In addition, fluid-inclusion studies indicate that skarn deposits are typically formed at high temperatures, and from highly saline fluids, indicative of a magmatic source. The isotopic data from skarns are also indicative of a dominantly magmatic source, with a meteoric input in the cooling / retrograde stages. These are not dissimilar to the data obtained from

Haveri. However, skarn deposits typically display a strong chemical control on their distribution, being created by metasomatic replacement of carbonates and thus, despite a few exceptions, the majority of skarns are found in lithologies containing at least some limestone. The absence of limestone in the pre-alteration and pre-mineralisation lithologies at Haveri favours an alternative ore genetic model, but does not, as such exclude skarn. Furthermore, a reduced gold-skarn model is unable to effectively account for the extensive magnetite observed at Haveri.

Of the described deposit types, the Fe-oxide Cu-Au deposits appear to have the greatest affinity with the Haveri deposit, in terms of tectonic setting, physico-chemical conditions of formation (fluid inclusions and stable isotopes), mineral assemblages, fluid evolution, and distribution and controls on mineralisation. On the basis of the available data, Haveri fits most easily into the Fe-oxide Cu-Au deposit class, and for the most part, is similar to descriptions of the magnetite-dominated Fe-oxide Cu-Au deposits of the Cloncurry region. However, Fe-oxide Cu-Au deposits are highly variable in many of their characteristics, and many fundamental uncertainties exist regarding their metallogenesis. However, despite this, it is possible to formulate a probable genetic model for Haveri with reference to the regional geological picture, and discuss how this model relates to other deposits and deposit types.

The principal difference between Haveri and “typical” Fe-oxide Cu-Au deposits (Hitzman et al., 1992; Pollard, 1999; Williams, 1999a) is that the main alteration phase is more calcic (rather than sodic or potassic) at Haveri, producing the widespread amphibolitisation. There are few examples of deposits where Ca-rich alteration assemblages are dominant. However, at the Eloise Cu-Au deposit in the Cloncurry district, the alteration assemblage is dominated by amphibole $\pm$ clinopyroxene (Baker, 1998), and at Osborne, the Eastern High Grade Lode has a Ca-rich alteration assemblage consisting of anthophyllite–hornblende–biotite (Adshead, 1995). Although calcic-rich systems as a whole are uncommon, many deposits have mineralogical characteristics that are transitional to skarn assemblages, and at Mount Elliott (Cloncurry district), a true skarn assemblage (diopside and grandite garnet) hosts the ore (Partington and Williams, 2000). The majority of Fe-oxide Cu-Au deposits do, however, have minor Ca-rich phases, and there are numerous examples where these calcic-rich alteration phases are significant to the extent that they are directly associated with the magnetite and Au-Cu mineralisation, such as throughout the Stuart Shelf, South Australia (Davidson and Large, 1994; Gow et al., 1994), where magnetite-rich, skarn-like assemblages are formed in the hornblende-hornfels facies. Such alteration assemblages have been linked to low pressure, high temperature (400 – 600 °C) metasomatic processes related to the main phase of felsic intrusive activity (Gow et al., 1994).

The cause of Ca enrichment accompanied with a relatively minor sodic enrichment has not been discussed in the literature. However, one of the main issues of contention in the formation of Fe-oxide Cu-Au deposits, and the widespread sodic (-calcic) alteration in particular, is the source of the fluids. There are two principal theories for fluid source, which more than likely represent end-members of a continuum between magmatic and amagmatic theories. The high salinities of the ore-bearing fluids have been variously ascribed to the circulation of non-magmatic evaporitic fluids through volcano-

plutonic complexes (e.g. Barton and Johnson, 1996, 1997; Johnson and Barton, 1997) or to magmatic sourced fluids (e.g. Gow et al., 1994; Baker, 1996, 1998; Rotherham et al., 1996; Rotherham, 1997; Pollard, 1999, 2001; Perring et al., 2000). Intermediate models, with variable degrees of mixing of fluid sources have also been suggested (Davidson and Large, 1994; Barton and Johnson, 1998). In the case of Haveri (Figure 5.17) the calcic-rich nature with only an early phase of sodic alteration is perhaps more indicative of either a magmatically-dominated source, or a metamorphic source in non-evaporitic rocks (e.g. clastic sediments), rather than an evaporitic source. This is also supported by the sulphur isotope data and the lack of evaporite sequences either close to Haveri or in the TSB in general. The high-temperatures associated with the alteration (530 – 610 °C) tend to favour a magmatic over a metamorphic source.

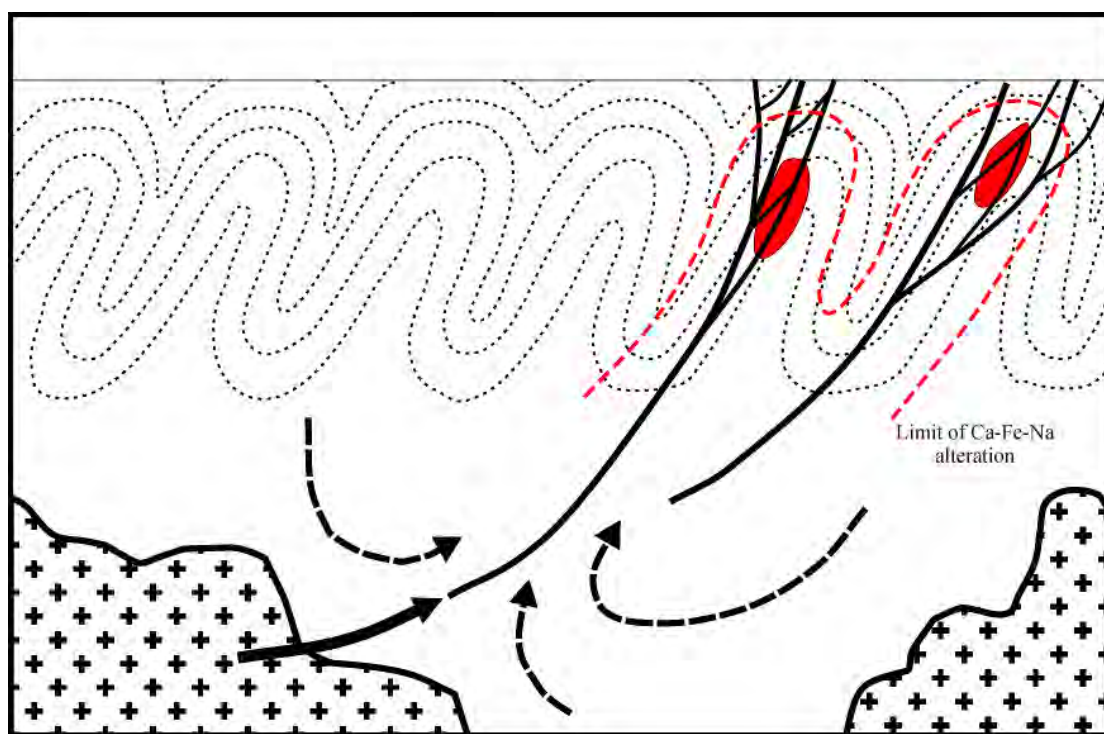


Figure 5.17: A schematic interpretation of the source of the ore-bearing fluids at Haveri, and the importance of structural control for the transport and localisation of the alteration (red dashed line) and mineralisation (areas in red).

The interpretation that Haveri is a member of the Fe-oxide Cu-Au class of deposit, related to the emplacement, crystallisation and cooling of the CFGC in an extensional tectonic setting, firmly places it within a broader group of magmatic-hydrothermal deposits that show many similarities, and are thus likely to indicate a genetic relationship. The most proximal members of the magmatic-hydrothermal system are represented by the porphyry deposits themselves. Further from the intrusion, when suitable country-rock lithologies are present, skarn deposits predominate, where ore deposition from magmatic fluids is primarily chemically controlled. More distal still, are the Fe-oxide Cu-Au deposits, dominated by magnetite at deeper levels and by hematite at higher levels (Hitzman et al., 1992; Williams, 1999a). The ore fluids potentially have a greater input of non-magmatic fluids (especially in the shallower, hematite-dominated deposits), and ore deposition is largely structurally controlled, with similar

structural styles to orogenic lode-gold deposits. It is perhaps not inconceivable, that the orogenic lode-gold deposits are the most distal expression of the magmatic-hydrothermal systems, allowing the fluids to develop their more complex characteristics, depending on the variable lithological and fluid interactions encountered along the fluid pathway. As these fluids are channelled upwards, so hypozonal, mesothermal and finally epizonal deposits are developed. Such a view of the genesis of magmatic-hydrothermal ore-deposits is supported by the “blurring” of deposit-styles that is observed with increasing temperature, either caused by depth or proximity to other heat sources. Mueller and Groves (1991) noted that many of the deep, high-temperature orogenic lode-gold deposits have alteration mineral assemblages that qualify as skarn deposits in the non-genetic descriptive sense of Einaudi and Burt (1982). Likewise, a number of Fe-oxide Cu-Au deposits display skarn assemblages, whereas others display features characteristic of orogenic lode-gold deposits.

Epigenetic Au ± Cu mineralisation is a common feature of the Fennoscandian Shield, and deposit types from the broad spectrum of epigenetic, magmatic-hydrothermal deposits discussed above are common. Of the many deposit types present in the Fennoscandian Shield, Haveri can be most closely compared to the numerous deposits of the Nordkalott Cu-Au district, including Bidjovagge, Pahtohavare, Nautanen and Saattopora (Table 5.1). Similarities also exist with several deposits within the Bergslagen district.

The Bidjovagge gold-copper deposit in northern Norway has been the subject of many detailed studies (e.g. Bjørlykke et al., 1987; Ettner et al., 1994; Nie et al., 1999), and despite a few clear differences, can be seen to have many similarities to Haveri. The deposit was mined until 1992 and contained approximately 3 million tonnes of ore with an average grade of 0.5 g/t Au and 1.8% Cu (Bjørlykke et al., 1987). The deposit is hosted by the Early Proterozoic Kautokeino greenstone belt, with mineralisation mainly occurring within graphitic schist, albite felsite, meta-diorite sills and amphibolite. Metamorphic grade varies from upper greenschist facies (to the east) to lower amphibolite facies (to the west). Mafic-to-granitoid stocks and dykes occur frequently at Bidjovagge. Syenodioritic dykes were emplaced during or between gold-rich and copper-rich mineralisation (Ettner et al., 1994). The major structure at Bidjovagge is a tight, gently plunging anticlinorium, with a discontinuous shear zone (consisting of slip and reverse faults, folds, tectono-clastic rocks, tectonic lenses, breccia and micro-breccia) running parallel to the antiform limbs. All these tectonic events and relevant metamorphism occurred during the Svecokarelian orogeny at approximately 1900-1850 Ma (Nie et al., 1999). The ores at Bidjovagge can be roughly divided into gold-rich (5 – 20 g/t Au and 0.1 – 0.5% Cu) and copper-rich (2 – 5% Cu and less than 1 – 2 g/t Au) ores, although these are commonly mixed together (Bjørlykke et al., 1987). The most common ore minerals are pyrite, chalcopyrite, pyrrhotite, magnetite and hematite, with accessory native gold and tellurides. The main gangue minerals are albite, actinolite, calcite and quartz (Nie et al., 1999). Alteration around the ore bodies consists mainly of albitisation, carbonatisation and scapolitisation, as well as less intense mica alteration and hematitisation. The albitisation is the most intensive, and is closely associated with gold-copper mineralisation (Nie et al., 1999). Although the gold-copper mineralisation as a whole was formed

during or slightly after the peak of the Svecokarelian orogeny, the copper-rich assemblage cross-cuts the gold-rich assemblage, indicating a later phase of mineralisation (Ettner et al., 1994; Nie et al., 1999). Fluid inclusion studies suggest that the ores were formed from high salinity (30 – 45 wt% NaCl equiv.) fluids at temperature of approximately 300 – 375 °C, between 2 and 4 kbar (Ettner et al., 1994). The ore fluids were emplaced along the shear zone, and REDOX reactions between the fluids and the graphitic schists resulted in gold precipitation from chloride complexes (Bjørlykke et al., 1987; Ettner et al., 1994; Nie et al., 1999). The source of the fluids was interpreted by Nie et al. (1999) to be derived from igneous sources related to syenodiorite dykes and meta-diorite sills, on the basis of Nd isotope data and REE geochemistry.

The Pahtohavare Cu-Au deposit near Kiruna, in Sweden, is another deposit of the Nordkalott district displaying similarities to Haveri. Pahtohavare is estimated to have contained some 1.15 million tonnes of ore with a grade of 0.9 g/t Au and 2% Cu. It is hosted by a tuffaceous unit with mafic units composed of basaltic lavas above and below. The tuffaceous unit itself has intercalations of black schist, carbonate rocks and magnetite-sulphide ores. The Pahtohavare area has been subject to folding in response to shearing along a sequence of WNW-trending faults or shear zones (Lindblom et al., 1996). The tectonic influence produced favourable zones of increased permeability along which the ore fluids were channelled and deposited. The ore minerals consist primarily of pyrite and chalcopyrite, with pyrrhotite being locally significant. Magnetite is also reported as a common constituent. The main gangue minerals consist of ferro-dolomite or ankerite, calcite, quartz and some scapolite with minor biotite and albite. Alteration includes albitisation, scapolitisation and minor biotitisation (Lindblom et al., 1996; Frietsch et al., 1997). Fluid inclusion studies at Pahtohavare indicate that there was early formation of quartz and pyrite at temperatures near 500 °C and pressures of 2 – 2.4 kbar, from a supersaturated solution of magmatic origin. The main stage of chalcopyrite and gold deposition is characterised by aqueous fluids of variable salinity (up to 30 eq. wt% NaCl including CaCl₂) at temperatures below 350 °C and pressures between 1 and 2 kbar (Lindblom et al., 1996). Gold was transported as a chloride complex, and precipitated due to an increase in pH, caused by CO₂ unmixing from the ore fluid, as well as cooling and dilution of the solution (Lindblom et al., 1996). Frietsch et al. (1997) also suggested that REDOX reactions between fluids and graphite in the black schists were an important cause of metal precipitation.

Although there are obvious differences between the many mines and deposits within the Nordkalott district, it is possible to make a generalised model for a “typical” Nordkalott Cu-Au deposit, that can be seen to have a number of similarities with Haveri: these similarities include the presence of abundant magnetite, widespread sodic (-calcic) alteration and albitisation, and moderate- to high-temperature, saline ore-bearing fluids carrying gold as chloride complexes. However, the “typical” Nordkalott deposit differs significantly from Haveri in that there is abundant scapolitisation and tourmalinisation (perhaps evidence of fluid interaction with evaporites) (Frietsch et al., 1997). Furthermore, the Nordkalott deposits are not as calcic (being more sodic) than Haveri, and were generally formed at lower temperatures (250 – 350 °C) and in many cases slightly higher pressures (Frietsch et al., 1997).

Despite this, a similar ore genetic model is evident, with deep-seated crustal magmatic processes initiated during low-P metamorphism, and the subsequent channelling of the ore-bearing fluids into fault zones, with shearing, fracturing and brecciation (Frietsch et al., 1997). Mechanisms for gold precipitation are also similar to those observed at Haveri, including REDOX reactions between the ore fluids and country rocks, and fluid unmixing, as well as a variety of structurally controlled depositional mechanisms.

Within the TSB itself, there are two other significant deposits: Paroinen and Kutemajärvi. Haveri displays many features comparable to the early, Stage I alteration and mineralisation at Paroinen as described by Clark (1997). This includes early Fe-Ca metasomatism in the vicinity of evolving D₂ shear zones, creating skarn-like mineral assemblages associated with gold enrichment in a sulphide-poor (i.e. magnetite-enriched) assemblage. Ore fluids for Stage I were estimated to have been high-temperature (at least 600 °C) involving saline fluids. At Paroinen, the early Fe-Ca metasomatism was followed by strong K-metasomatism which was only weakly developed at Haveri. Paroinen then continued into the Stage II alteration and mineralisation characterised by the main Cu-ore forming stage at lower temperatures (*ca.* 455 °C) and by moderately saline ore fluids. This phase is apparently absent at Haveri. Hydrothermal activity progressed overall from high (>600 °C) to very low (<150 °C) temperatures, accompanied by major decompression from *ca.* 1.5 kbar to *ca.* 400 bars. Despite failing to unequivocally identify the magmatic source, Clark (1997) clearly favours a proximal magmatic-hydrothermal model for the generation of the Paroinen deposit, and gold precipitation being controlled by the simple cooling of ore fluids. However, the presence of shearing, and the lack of any detailed description of the host lithologies at Paroinen (beyond being hosted by metabasaltic volcanic rocks), do not rule out other mechanisms of ore precipitation, such as REDOX reactions, pressure release and fluid unmixing that have been observed elsewhere in the development of Cu-Au deposits of the Fennoscandian.

The Kutemajärvi Au deposit further east in the TSB, has many features which are significantly different to Haveri (and also Paroinen). The Kutemajärvi deposit is situated in a part of the TSB that was subjected to peak metamorphic conditions of 470-570 °C and 3-4 kbar, corresponding to a low-pressure domain transitional from greenschist to amphibolite facies (Kilpeläinen et al., 1994). The host rocks of the deposit are strongly hydrothermally altered, intermediate metavolcanic rocks, in a sequence that is considerably more felsic than the rock sequence hosting the Haveri deposit. Hydrothermal alteration is characterised by silicification and sericitisation with minor chloritisation. Gold mineralisation is hosted in discrete vertical pipes associated with intense alteration (especially silicification) and deformation. Gold is present as native gold, as well as gold tellurides, the latter representing a later stage than the main gold and sulphide event. This is a feature absent at Haveri but present at Paroinen. Eilu et al., (2003) has interpreted Kutemajärvi as a metamorphosed high-sulphidation epithermal-style deposit. Epithermal-style deposits by their nature, typically occur at shallow depths in the crust, and thus have a poor preservation potential in metamorphosed, deformed and eroded terranes, unless accompanied by rapid subsidence (Groves et. al., 1998; Eilu et al., 2003).

Consequently, Eilu et al., (2003) has suggested that the Kutemajärvi deposit was formed in an extensional (post-orogenic) setting such as during arc extension.

However, despite these obvious differences there are still some significant similarities with both Haveri and Parainen. The close spatial relationship between the deposit and the abundant plutonic rocks is a common feature to all the deposits, as are the shear zones and deformational structures that are likely to have acted as fluid conduits. At Kutemajärvi, there was also a progression of increasing sulphidation from early oxide-dominated phases with magnetite present in the assemblage to a sulphide-dominated assemblage containing chalcopyrite and pyrrhotite. Poutiainen and Grönholm (1996) interpreted the ore deposition mechanism as occurring due to a combination of phase separation (fluid unmixing) and reaction with wall rocks. These are features recognised at Haveri. Finally, and perhaps most significantly, these deposits all formed in an extensional tectonic setting, with similar structural controls on the fluid pathways and sites for precipitation of mineralisation, with the main differences being the ultimate source of the fluids and the crustal level of formation. Consequently, despite the differences exhibited by Kutemajärvi compared to Haveri (and Parainen), there are enough similarities to suggest that a broad genetic relationship may indeed exist between these deposits.

In summary, the Haveri deposit is closely related to the many magmatic-hydrothermal Au±Cu deposits that are common throughout the Early Proterozoic of Fennoscandia (Table 5.1), and indeed of other Early Proterozoic terranes globally. These deposits, on a district scale (as opposed to individual mine scale) are commonly spatially associated with synkinematic, I-type granitoids that are typically formed in extensional (or transtensional) tectonic settings. Although these granitoids may not always be the source of the ore fluids themselves, their proximity to the ore deposits is not coincidental and, as such, are highly likely to be a major contributory factor in the genesis of the deposits. The deposits themselves (Figure 5.18) tend to have strong structural controls, with shear zones, along which ore fluids can be channelled. The ore zones themselves are controlled by physico-chemical changes leading to metal precipitation. Such changes include REDOX boundaries, structural sites such as folds (which cause fluid traps, where cooling and fluid unmixing can occur), and shear zones and fractures (where seal-and-rupture mechanisms lead to pressure release).

	HAVERI	PAROINEN	KUTEMAJÄRVI	PAHTOHAVARE	BIDJOVAGGE
Host rocks	Intercalated tuffs at contact of mafic metavolcanic rocks and clastic sediments	Mafic metavolcanic rocks and metabasaltic tuffs	Intermediate volcanics and tuffs	In tuffaceous unit within a sequence of mafic metavolcanic rocks. Graphite horizons also present.	Graphitic schist, albite felsite, meta-diabase sills and amphibolite (mafic metavolcanic rocks)
Metamorphic facies	Amphibolite	Amphibolite	Greenschist – amphibolite	Upper greenschist (locally reaching lower amphibolite)	Upper greenschist to lower amphibolite
Metamorphic conditions	600 °C, 1.5 kbar	500 – 600 °C, 1.5 – 3 kbar	470 – 570 °C, 3 – 4 kbar		
Intrusives	Quartz-feldspar porphyries and Synkinematic, I-type granitoids of the CFGC	Synkinematic, I-type granitoids of the CFGC and post-D ₁ porphyry dykes	Synkinematic, I-type granitoids of the CFGC	Iron-ore bearing porphyries of Kiruna	Mafic-to-granitoid stocks and syenodioritic dykes
Dating	1890 – 1880 Ma (D ₁ and granitoids emplacement)	1885±20 Ma (D ₁ & Hämeenkyrö Batholith). 1864±11 Ma (stage II hydrothermal uraninite)	1.88±0.01 Ga (thermal metamorphism)	1.91 – 1.89 Ga (age of porphyries)	1885±18 Ma (mineralisation, from davidite associated with Au)
Principal ore assemblage	Po+Mt+Cp+Py+Au	Stage I: Mt+Au Stage II: Cp+Sch+Tour+Asp+Ag Stage III: Troilite + Po	Au + Au-Te tellurides Py+Po+Cp+Sp	1 st phase: Py+Mt 2 nd phase: Cp+Au	Py+Cp+Po+Mt+Hm
Alteration	Pre-ore (Na-Ca) albitisation (Fe-Ca) Amphibolitisation Calcic-skarn (Fe-Ca) Biotitisation (K) Late chloritisation	I: Early Fe-Ca metasomatism Strong K-metasomatism II: Tourmalinisation Silicification III: Carbonate veins	Silicification Sericitisation Chloritisation	Albitisation Scapolitisation Epidotisation	Albitisation Carbonatisation Scapolitisation
Fluids	Low CO ₂ aqueous fluids <10 eq. wt% NaCl (early) becoming more saline	I: 20 – 27 eq. wt% NaCl II: 12 – 18 eq. wt% NaCl	<7 eq. wt% NaCl H ₂ O-CO ₂ -CH ₄ (Metamorphic)	~ 30 eq. wt% NaCl (+ CaCl ₂) Minor CO ₂ phase	30 – 45 eq. wt% NaCl H ₂ O-CO ₂ -CH ₄
Mineralisation conditions	530 – 610 °C Post-ore chlorite: 309 – 368 °C	I: >600 °C II: 489 – 145 °C, 1.25 kbar III: <145 °C	320 – 380 °C, 2.0 – 2.8 kbar (Metamorphic)	1 st phase: ~ 500 °C, 2.4 kbar 2 nd phase: <350 °C (1-2 kbar)	300 – 375 °C, 2 – 4 kbar
Ore hosting structures	D ₂ folding and shear zone	D ₂ shear zones and breccias	Discrete vertical pipes associated with intense shearing and silicification.	Folding, faults and shear zones	Shear zones and breccias
Precipitation mechanisms	REDOX contrasts, pressure release leading to fluid unmixing	Cooling of ore fluids	Phase separation due to pressure release, as well as fluid mixing	CO ₂ unmixing and fluid dilution	REDOX contrasts

Table 5.1: Summary table of the characteristics of a number of Fennoscandian Cu-Au deposits comparable to Haveri. Data taken from Bjørlykke et al. (1987), Ettner et al. (1994), Lindblom et al. (1996), Poutiainen and Grönholm (1996), Clark (1997), Frietsch et al. (1997), Nie et al. (1999).

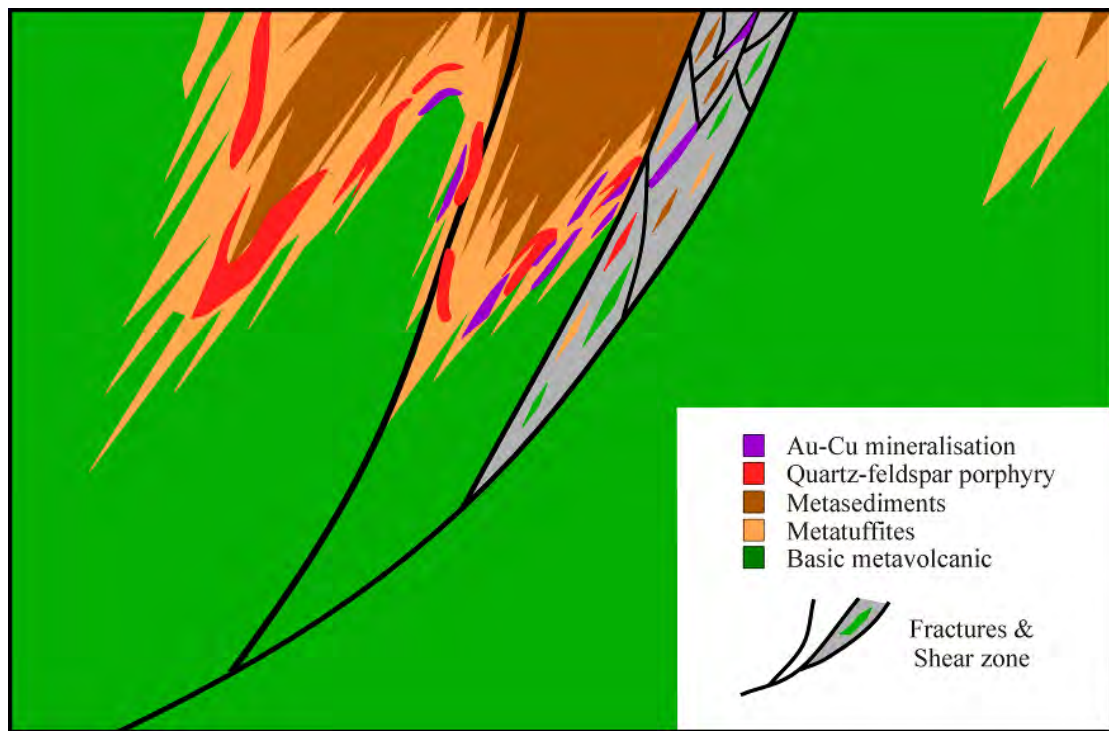


Figure 5.18: Schematic representation of Au-Cu mineralisation at Haveri, being controlled by the proximity to shear zones and fractures, coupled with the strong REDOX contrasts offered by the metatuffite unit. Other styles of mineralisation, including pressure-release in shear zones, as well as structural traps offered by folds and the competency contrasts of the porphyries are also depicted.

Consequently, the Svecofennian arc terranes are considered highly prospective regions for exploration for Fe-oxide Cu-Au deposits similar to Haveri, due to the high heat flow created by the emplacement of the voluminous synkinematic I-type granitoids. Exploration criteria within these terranes should focus on regions with widespread Fe-Ca-Na alteration, expressed by albitisation, amphibolitisation and/or the presence of abundant iron oxides. In addition, within the broader zones of alteration, favourable sites for ore deposits would include structural features such as shear zones and folds, and REDOX boundaries provided by certain lithological contrasts.

6 CONCLUSIONS

During the early stages of the formation of the Tampere volcanic arc approximately 1.99 Ga ago, the tholeiitic basalts and tuffites of the Haveri Formation were formed. There is a continuous gradation upwards from the predominantly volcanoclastic Haveri Formation into the overlying epiclastic meta-greywackes of the Osara Formation. The Haveri deposit is hosted in this contact zone. Whole rock geochemical studies (including REE and trace elements) indicate that the metabasaltic rocks of the Haveri Formation have a geochemistry similar to E-type MORB, and thus they are interpreted as forming in a back-arc basin setting between a mature island arc or continental margin volcanic arc to the south (to later become the Tampere Schist Belt), and the Archean craton and Karelian formations to the north. After initial sea-floor volcanism, active magmatism waned, and the depositional environment became dominated by turbidites of the Osara Formation. This supracrustal sequence was subsequently intruded concordantly by the sub-volcanic, calc-alkaline, dacitic quartz-feldspar porphyries, that are interpreted as having an arc-type origin. This is consistent with the evolution from an initial back-arc basin, through a period of passive margin and/or fore-arc deposition represented by the Osara Formation greywackes and the basal stratigraphy of the TSB, prior to the onset of arc-related volcanic activity characteristic of the TSB and the Svecofennian proper.

Approximately 1.89 Ga ago, delamination of the base of the lithosphere and replacement by a hot asthenosphere are believed to have initiated granitoid formation. Decompression melting of the asthenosphere produced a mantle magma that rose to the crust-mantle boundary, remelted the lower crust and generated a wide variety of magmas by mixing and mingling of the crustal and mantle magmas. The high heat flow associated with crustal underplating and intraplating, and resultant magmatism generated high-temperature low-pressure metamorphism that culminated coevally with the emplacement of the magmas and formation of the CFGC synkinematic, I-type, magnetite-series granitoids 1.90 – 1.87 Ga ago. In the Haveri area, with the emplacement of the Hämeenkyrö batholith at 1882 ± 6 Ma, prograde metamorphism and D_1 deformation can be constrained to 1.89 – 1.88 Ga ago, coeval with the main phase of synkinematic granitoid emplacement.

By a combination of petrogenetic grids, cation exchange thermometry and isotope thermometry, peak metamorphism in the Haveri area can be constrained to a maximum of approximately 600 °C and 1.5 kbar pressure, which is equivalent to an approximate depth of 5 km. The metamorphic mineral assemblages for the various rock types can constrain the oxygen fugacity of the metamorphic fluids (at 600 °C) to a $\log f(\text{O}_2)$ value of approximately –21.0 to –26.0 for the pelitic assemblages and a $\log f(\text{O}_2)$ value of approximately –14.5 to –17.5 for the mafic metavolcanic rocks. This indicates that there is a significant REDOX boundary along the gradational contact of these rock types.

At some time after D_1 deformation and prograde metamorphism, during crystallisation and cooling of the granitoids, a magmatic-dominated hydrothermal system became prevalent, and was responsible for considerable hydrothermal alteration and Cu-Au mineralisation. This alteration is, for the main part,

interpreted to have commenced and evolved through late- and post-D₁ deformation, perhaps continuing through to the early stages of D₂ deformation. The alteration commenced with a pre-ore Na-Ca alteration resulting in albitisation of the host rock. This was closely followed by a strong Ca-Fe alteration phase, responsible for the voluminous and pervasive amphibolitisation and quartz veining. The amphibolitisation and quartz veining was associated with abundant sulphide (mainly pyrrhotite and chalcopyrite) and magnetite deposition as well as gold mineralisation. Oxygen isotope studies, combined with very limited fluid inclusion studies suggest that the Ca-Fe alteration formed at high temperatures (530 – 610 °C) from low CO₂, low- to moderately-saline, magmatic dominated fluids. Fluid inclusion decrepitation textures in the quartz veins suggest isobaric decompression. This is compatible with formation in high-T / low-P environments such as contact aureoles and island arcs.

The calcic-skarn assemblage, combined with phase equilibria and sphalerite geothermobarometry, are indicative of formation at high temperatures (500 – 600 °C) from fluids with higher CO₂ contents and more saline compositions than those responsible for the Fe-Ca alteration. Limited fluid inclusion studies have identified hypersaline inclusions in secondary inclusion trails within quartz. The presence of calcite and scapolite also support formation from CO₂-rich saline fluids. It is suggested that the calcic-skarn alteration and the amphibolitisation evolved from the same fluids, and that *P-T* changes led to fluid unmixing resulting in two fluid types responsible for the observed alteration variations.

Both sulphides and oxides are common at Haveri, with ore types varying from massive sulphide and/or magnetite, to networks of veinlets and disseminations of oxides and/or sulphides. Cataclastites, consisting of deformed, brecciated bands of sulphide, with rounded and angular clasts of quartz vein material and altered host-rock are an economically important ore type. Ore minerals are principally pyrrhotite, magnetite and chalcopyrite with lesser amounts of pyrite, molybdenite and sphalerite. There is a general progression from early magnetite, through pyrrhotite to pyrite indicating increasing sulphidation with time. Gold is typically found as free gold within quartz veins and within intense zones of amphibolitisation. Considerable gold is also found in the cataclastite ore type either as invisible gold within the sulphides and/or as free gold within the breccia fragments.

The gold mineralisation at Haveri is related primarily to the amphibolitisation stage of alteration. At such *P-T* conditions, and in moderately saline fluids, gold and copper are soluble as chloride complexes. Metal deposition was induced by a variety of mechanisms, including mainly structural controls, resulting in local variations in temperature, pressure and pH of the fluids, and consequent metal precipitation. Decrease in pressure and temperature also led to fluid unmixing and the generation of the CO₂-rich fluids causing calcic-skarn alteration. The fluid unmixing also caused an increase in pH, decreasing the solubility of metal chloride complexes, and contributing to the precipitation of Cu and Au. A broad chemical control also contributed to metal deposition, caused by changes in the oxygen fugacities of fluids at the REDOX boundary represented by the contact between the relatively oxidised metavolcanic rocks and the relatively reduced metasedimentary rocks. Post-ore alteration includes an evolution to more K-rich alteration (biotitisation). During the latter stages of D₂

deformation, with progressive cooling to temperatures below 400 °C, hydrothermal fluids may have remobilised gold and copper as bisulphide complexes and deposited them in the late-D₂ shear-controlled cataclastites and massive sulphides. As regional temperatures cooled below 400 °C, retrograde, lower temperature minerals such as chlorite also began to replace the earlier higher-temperature minerals such as amphibole and biotite. Chlorite geothermometry on retrograde chlorite indicates temperatures of 309 – 368 °C. As chlorite represents the latest hydrothermal alteration event, this can be taken as a lower limit for hydrothermal alteration and mineralisation at Haveri.

At a later stage, post-D₂ and pre-D₃, the carbonate-barite veins were formed. This represents evolution to more oxidised conditions and lower temperatures, as evident from the presence of sulphate and pyrite (rather than pyrrhotite). These, along with the epidote veinlets seen in the field, are interpreted as very late emanations from the cooling granitoids (supported by C-O isotopes and field relationships), perhaps interacting with meteoric waters. This was followed by the emplacement of dolerite dykes into the cooled sequence (chilled margins) and then the weak D₃ deformation followed, resulting in shearing and open folding of the dykes, and structural reactivation along the carbonate-barite bands. The D₃ shearing-phase was accompanied by pyrite deposition.

The Haveri deposit has many similarities with ore deposit models that include orogenic lode-gold deposits, certain Au-skarn deposits and Fe-oxide Cu-Au deposits. Although these deposit types form distinct genetic classes, they also share many characteristic features, thus blurring the boundaries between them. However, if the evidence accumulated from Haveri is reviewed in its entirety, it is apparent that the Haveri deposit has numerous features in common with many Fe-oxide Cu-Au deposits. These features include: formation in an extensional tectonic regime; proximity to I-type granitoids; alteration types including pre-ore albitisation and the main Na-Ca-Fe alteration associated with magnetite, iron-sulphides and gold in a relatively sulphur deficient system; the *P-T-X* conditions of mineralising fluids; and finally, similar mechanisms for gold precipitation from Au chloride complexes. All of these features compare favourably with other Early Proterozoic deposits within the TSB and Fennoscandia, as well as many of the deposits in the Cloncurry district of Australia. Consequently, the Haveri deposit can be seen to represent a high-T, Ca-rich member of the recently recognised Fe-oxide Cu-Au group of deposits.

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APPENDIX A: SAMPLING

SAMPLE LOCATIONS AND NAMING

The samples selected for this study were largely collected during and resulting from the exploration conducted by Glenmore Highlands Inc. Consequently, the sample names adopted for this thesis have (where possible) been chosen to be consistent with the naming of samples used for the exploration programme.

Sample names in this thesis are of two types: those that refer to borehole samples (e.g. H5-280 or P22-72a) and those that were selected specifically for whole-rock analyses and / or individual “grab” samples (e.g. TS-27). These include borehole samples, underground grab samples and surface outcrop samples.

The majority of samples were taken from boreholes drilled during the exploration programme. The initial letter(s) and number before the hyphen refer to the borehole number, with the letter indicating the drilling programme. The various drilling programmes are detailed in Table A-1. The number after the hyphen refers to the downhole depth of the sample rounded to the nearest metre. If two samples have been taken from the same metre interval then they are labelled with a letter suffix (e.g. P22-72, P22-72a, P22-72b etc.).

Sample	No of Holes	Metres drilled	Description
Rx-xxx	15	3853.30	Regional drilling phase
Hx-xxx	6	1374.05	Drilling concentrated to the S and W of Haveri
Hmx-xxx	2	205.80	Drilling concentrated at the Haveri Mine
Px-xxx	22	2519.62	Drilling concentrated at Peltosaari
TS-xx			Grab samples and borehole samples for analysis

Table A-1: Description of sample numbering employed during this study.

The details of each of the boreholes are provided in Table A-2.

<i>Hole</i>	<i>North</i>	<i>East</i>	<i>Elevation</i>	<i>Depth</i>	<i>Azimuth</i>	<i>Dip</i>
R1	6845194	2461048	92.5	229.85	201	45.3
R2	6845249	2461036	92.5	251.70	207	44.5
R3	6845255	2461008	92.5	195.70	214	45.1
R4	6845261	2460901	86	206.20	196	47.1
R5	6845310	2460788	86	205.10	205	44.8
R6	6845180	2460427	101	362.00	193	45
R7	6845198	2460196	95	342.50	152	45.3
R8	6845149	2460471	101	300.50	335	44.2
R9	6844908	2460140	86	214.80	293	44
R10	6844917	2460125	86	241.00	121	45.3
R11	6844912	2460132	86	258.00	212	45
R12	6844810	2460732	90	245.05	207	44.8
R13	6844659	2461018	87.5	218.20	242	44.5
R14	6843853	2460687	100	301.20	230	45.3
R15	6843690	2460383	100	281.50	239	45.5
H1	6844870	2459740	80	181.85	240	45
H2	6844641	2459745	80	238.20	60	49
H3	6844808	2459878	80	202.00	240	45.6
H4	6844676	2459912	80	200.00	240	44.6
H5	6844679	2459917	80	298.80	65	44.5
H6	6843387	2460473	90	253.20	212	45.4
P1	6845085	2461000	85	99.30	34	45
P2	6845096	2460984	85	100.70	34	45.3
P3	6845071	2460967	85	103.40	34	46
P4	6845046	2460951	85	101.80	34	47.9
P5	6845118	2460951	85	107.85	34	45.7
P6	6845093	2460934	85	109.85	34	46.8
P7	6845093	2460934	85	110.85	34	61.2
P8	6845068	2460917	85	130.00	34	62.1
P9	6845115	2460901	85	111.05	34	47.6
P10	6845099	2460890	85	137.40	34	61.3
P11	6845074	2460873	85	160.70	34	62.7
P12	6845162	2460884	85	124.30	34	59.7
P13	6845138	2460868	85	129.95	34	60.7
P14	6845096	2460840	85	174.60	34	61
P15	6845243	2460818	85	99.40	34	47.5
P16	6845218	2460801	85	130.00	34	60.5
P17	6845202	2460718	86	85.45	34	60.4
P18	6845168	2460696	86	115.00	34	60
P19	6845094	2460959	85	87.97	34	45
P20	6845094	2460959	85	85.45	34	60
P21	6845096	2460912	85	101.80	34	45
P22	6845096	2460912	85	112.80	34	60
HM1	6845054	2460259	92	71.90	154.6	45
HM1A	6845054	2460259	92	133.90	154.6	65

Table A-2: Locations of all diamond drill holes drilled by Glenmore Highlands since acquiring the Haveri property. These total 7952.77 metres of drilling.

APPENDIX B: WHOLE-ROCK SAMPLES (TS-1 TO TS-32)

Sample	Location	Northing	Easting	Rock Type	Description
TS1	NW of Haveri	6845479	2460000	Amphibolite	Close to granodiorite contact. "vein" of GRD cuts outcrop.
TS2	NW of Haveri	6845578	2459984	Granodiorite	Close to amphibolite contact (80 metres from TS1).
TS3	Ansomäki	6843964	2460383	Amphibolite	Altered amphibolite 1 metre from porphyry.
TS4	Ansomäki	6843876	2460387	Porphyry	Narrow porphyry cutting amphibolites, dark matrix.
TS5	Ansomäki	6844380	2460430	Amphibolite	North slope of Ansomakki, in "pillowed" amphibolites
TS6	Peltosaari	6845140	2460980	Mafic Dyke	Sample of the dyke in the Peltosaari trench
TS7	Peltosaari (E)	6845170	2461210	Amphibolite	"Pillowed" volcanic from east end of Peltosaari
TS8	Haveri mine	6844969	2460258	Amphibolite	Plug drill of intense amphibolite on west of pit
TS9	Haveri mine	6844993	2460273	Porphyry	Fresh QFP from west of pit
TS10	Haveri mine	6845012	2460259	Mafic Dyke	Plug drill of main dyke through pit outcrop
TS11	Haveri mine	6845045	2460262	Blotchy C-S	Plug of blotchy calc-silicate wedst of north trench
TS12	Haveri mine	6845045	2460277	Carbonate	Main band of carbonate from north trench
TS13	Haveri mine	6844973	2460271	Carbonate	Main band of carbonate from south trench
TS14	R7 trench	6845200	2460220	Porphyry	Porphyry taken from 3m from e end of main trench
TS15	H6-131			Porphyry	Fresh QFP.
TS16	P8-26			Porphyry	Fresh QFP. Close to sample in SA.
TS17	P22-20			Uralite Porphyry	UPOR (to 20.74) in contact with QFP. Analysed 20.5 to 20.63
TS18	R13-215			Granodiorite	Fresh granodiorite.
TS19	R1-225			Granodiorite	Fresh granodiorite.
TS20	P16-36			Porphyry	QFP with pink potassic alteration.
TS21	P5-107			Amphibolite	Vesiculated amphibolite
TS22	P5-96			Amphibolite	Light network skarn in amphibolite.
TS23	P21-78			Amphibolite	Dense network skarn in amphibolite.
TS24	P21-72			Amphibolite	Intense amphibole skarn.
TS25	P21-53			Sediments	Fresh biotite hornfels (sediments)
TS26	P22-31			Sediments	Skarned biotite hornfels
TS27	P19-66			Amphibolite	Amphibole - pyroxene skarn in amphibolite
TS28	P19-67			Amphibolite	Amphibole - pyroxene skarn in amphibolite
TS29	Haveri mine	6844942	2460282	Sediments	West of pit near steel laers.
TS30	Haveri mine	6844969	2460255	C-S / sediments	Grey banded unit by intense amphibole (west)
TS31	Haveri mine	6844975	2460246	C-S / sediments	Grey banded unit by intense amphibole (east)
TS32	Haveri mine	6844986	2460253	Sediments	Rusty yellow weathering of sediments
TS33	Haveri mine	6845001	2460262	Sediments	"Coarse" sediments with pale yellow weathering
TS34	Haveri mine	6845021	2460268	C-S	Banded unit
TS35	Haveri mine	6845032	2460286	Sediments	Rusty unit between the bands of banded calc-silicate
TS36	Haveri mine	6845032	2460344	C-S	Rusty calc-silicate east of sheared porphyry north of pit.
TS37	Haveri mine	6845024	2460376	Sediments	Folded sediments north east corner of open pit
TS38	Haveri mine	6844946	2460368	Amphibolite	Spotted amphibolite on east of open pit.

Table B-1: Location and descriptions of samples for whole-rock analyses.

SAMPLE	TS1	TS2	TS3	TS4	TS5	TS6	TS7	TS8
SiO ₂	50.040	59.400	51.460	63.000	48.360	54.350	51.200	44.730
Al ₂ O ₃	13.910	16.870	13.330	16.800	12.770	14.420	13.900	8.560
FeO	8.940	6.000	12.160	3.620	13.030	7.990	8.850	11.560
Fe ₂ O ₃	3.805	0.222	1.176	0.147	1.969	1.270	1.235	2.673
MnO	0.160	0.120	0.220	0.150	0.260	0.170	0.140	0.160
MgO	6.020	2.440	5.640	3.390	5.200	5.060	5.930	14.350
CaO	9.880	5.570	10.050	4.480	10.780	9.770	11.310	9.200
Na ₂ O	2.910	3.570	1.490	1.180	2.600	2.660	3.200	0.630
K ₂ O	0.650	1.980	0.570	4.860	0.410	0.590	0.460	2.440
TiO ₂	1.410	0.750	1.590	0.450	1.940	0.750	1.530	0.660
P ₂ O ₅	0.160	0.220	0.150	0.110	0.150	0.110	0.130	0.130
LOI	0.730	0.960	0.550	1.150	0.260	0.580	0.530	1.790
TOTAL	99.610	98.780	99.740	99.337	99.180	98.600	99.400	98.170
Sc	42.000	15.000	41.000	11.000	44.000	34.000	41.000	13.000
Be	2.000	2.000	2.000	2.000	2.000	1.000	2.000	3.000
V	338.816	127.812	346.424	89.706	369.275	221.084	320.584	729.930
Cr	91.035	21.595	68.961	60.483	27.043	133.222	138.257	141.301
Co	30.524	13.045	40.429	9.687	40.907	24.680	19.298	30.700
Ni	62.071	30.921	73.227	29.492	86.706	38.738	75.571	150.995
Cu	85.163	23.735	101.582	26.599	114.526	23.313	16.548	99.751
Zn	28.604	62.066	91.505	66.189	97.641	51.231	27.550	130.699
Ga	15.302	18.292	19.382	18.186	16.881	15.864	14.408	16.891
Ge	1.493	1.193	1.538	1.815	1.646	1.423	0.816	3.633
As	<5	<5	9.371	12.892	9.223	6.742	<5	11.730
Rb	11.887	48.242	18.441	157.577	5.175	6.656	9.763	121.921
Sr	261.764	577.686	190.200	121.732	165.673	404.385	220.018	11.908
Y	27.463	18.712	28.402	14.060	33.464	17.853	28.419	37.239
Zr	98.130	105.895	108.635	128.804	135.818	73.504	97.699	93.888
Nb	6.682	7.780	7.337	11.327	8.859	5.717	5.782	7.390
Mo	3.395	1.503	1.416	1.699	0.694	0.850	0.275	78.813
Ag	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5
In	<0.1	<0.1	<0.1	<0.1	<0.1	0.141	<0.1	0.171
Sn	0.782	1.313	0.984	1.946	1.469	<0.5	1.245	4.226
Sb	0.465	0.240	2.062	1.185	2.878	1.194	0.551	1.422
Cs	0.402	1.452	0.958	2.398	0.172	0.465	0.461	4.018
Ba	135.705	726.514	97.723	1,170.000	148.509	103.720	71.466	213.234
La	14.464	24.886	9.965	27.456	13.751	11.425	9.891	14.459
Ce	30.859	47.382	23.849	49.947	34.053	25.901	23.226	23.888
Pr	3.777	5.312	3.428	5.258	4.301	3.205	2.931	3.373
Nd	16.956	21.727	16.568	19.766	18.751	12.563	13.782	14.327
Sm	4.594	4.985	4.652	4.019	5.499	3.073	4.231	4.126
Eu	1.409	1.228	1.262	0.848	1.667	1.016	1.336	0.867
Gd	4.400	4.026	4.625	3.261	5.386	3.201	4.142	4.112
Tb	0.801	0.608	0.891	0.445	0.964	0.493	0.816	0.684
Dy	5.077	3.540	5.365	2.532	6.166	2.884	5.278	4.537
Ho	1.022	0.670	1.047	0.481	1.256	0.638	1.046	1.030
Er	2.812	1.868	3.108	1.406	3.387	1.875	2.799	2.926
Tm	0.408	0.259	0.482	0.217	0.507	0.276	0.420	0.425
Yb	2.877	1.724	2.859	1.440	3.180	1.731	2.857	3.000
Lu	0.443	0.280	0.444	0.235	0.507	0.289	0.445	0.572
Hf	2.760	2.918	3.164	3.216	3.762	1.860	2.747	2.421
Ta	0.432	0.543	0.478	0.647	0.589	0.301	0.414	0.516
W	3.928	<0.2	0.593	1.589	2.245	<0.2	0.318	0.334
Tl	<0.05	0.267	0.158	1.145	0.063	<0.05	<0.05	1.134
Pb	<5	7.909	<5	24.531	6.254	5.349	<5	<5
Bi	0.247	0.208	0.215	0.337	0.259	0.510	0.108	0.476
Th	0.680	4.023	1.102	4.005	1.588	1.566	1.061	4.594
U	0.265	1.513	0.300	1.842	0.364	0.806	0.301	30.962
S (%)	0.070	0.030	0.020	0.470	0.020	0.040	0.010	0.190

Table B-2: Whole-rock major and trace element data. Oxide concentrations are reported as percentages. Trace element data are reported as parts per million.

SAMPLE	TS9	TS10	TS11	TS12	TS13	TS14	TS15	TS16
SiO ₂	67.050	52.660	52.110	5.040	7.790	59.270	68.010	68.630
Al ₂ O ₃	16.150	14.750	13.020	0.140	2.740	15.340	16.410	16.360
FeO	3.300	8.050	5.660	2.250	3.070	5.330	3.340	1.830
Fe ₂ O ₃	0.983	2.494	3.660	4.129	2.558	0.327	0.068	1.636
MnO	0.050	0.220	0.060	0.440	0.200	0.090	0.080	0.040
MgO	1.370	5.680	8.870	13.640	6.610	3.460	1.600	1.170
CaO	3.400	8.660	8.340	30.540	17.610	7.570	3.560	3.970
Na ₂ O	6.010	1.500	3.700	0.050	0.050	5.400	4.470	4.260
K ₂ O	0.510	1.540	0.530	0.030	0.030	0.550	1.690	1.500
TiO ₂	0.250	1.090	0.660	<0.01	0.130	0.520	0.270	0.240
P ₂ O ₅	0.050	0.170	0.080	0.020	0.050	0.160	0.060	0.060
LOI	0.760	1.050	2.090	34.480	15.410	0.660	0.940	0.940
TOTAL	100.240	98.750	99.400	91.020	56.570	99.280	100.880	100.820
Sc	8.000	40.000	20.000	<1	4.000	20.000	9.000	7.000
Be	1.000	2.000	5.000	<1	1.000	2.000	1.000	1.000
V	59.180	282.260	1,290.000	10.033	260.272	198.939	72.093	53.845
Cr	37.426	181.210	150.106	<10	33.778	55.623	31.945	27.101
Co	13.867	28.989	43.154	<0.5	19.567	12.423	6.703	12.875
Ni	13.815	52.600	127.252	21.331	42.249	31.811	21.324	12.223
Cu	202.591	13.763	353.520	50.583	208.654	28.833	29.366	54.516
Zn	55.259	102.214	37.644	49.638	<10	48.522	67.952	24.583
Ga	17.392	19.894	17.730	<1	3.854	19.307	16.779	18.282
Ge	0.976	1.476	1.904	<0.5	<0.5	3.044	1.115	0.674
As	11.797	6.522	7.221	<5	12.135	9.067	9.929	5.228
Rb	15.339	57.216	18.652	0.715	1.303	12.260	44.246	66.477
Sr	538.546	439.059	224.819	390.536	1,510.000	426.643	689.687	861.666
Y	9.587	23.124	48.391	4.623	13.143	17.188	8.933	8.410
Zr	95.915	117.930	116.751	3.873	14.056	116.823	101.781	97.837
Nb	8.971	7.455	11.010	<0.5	1.835	9.068	8.004	9.125
Mo	4.050	1.586	86.438	7.490	28.258	1.932	0.749	1.639
Ag	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5
In	<0.1	<0.1	<0.1	<0.1	<0.1	<0.1	<0.1	<0.1
Sn	0.658	1.033	0.882	<0.5	<0.5	1.785	0.739	<0.5
Sb	1.174	0.762	2.563	0.193	1.051	1.488	0.837	0.474
Cs	0.626	2.324	0.360	<0.1	<0.1	0.404	3.258	1.477
Ba	325.745	355.838	105.186	36,800.000	129,000.000	999.513	1,450.000	1,080.000
La	12.413	29.694	57.326	5.223	4.110	30.570	11.177	10.783
Ce	21.497	56.426	85.751	6.057	5.714	58.130	20.427	19.554
Pr	2.433	6.656	10.260	0.666	1.008	7.141	2.238	2.348
Nd	9.267	28.797	36.268	2.712	5.437	25.955	9.232	8.618
Sm	1.867	6.536	6.984	0.659	2.244	4.886	2.002	1.705
Eu	0.463	1.725	1.162	<0.005	<0.005	1.005	0.514	0.449
Gd	1.819	5.354	7.730	0.477	2.181	4.631	1.644	1.546
Tb	0.275	0.805	1.088	<0.01	0.259	0.579	0.271	0.241
Dy	1.547	4.508	6.412	0.477	1.588	2.900	1.548	1.339
Ho	0.307	0.807	1.492	0.094	0.328	0.588	0.291	0.263
Er	1.034	2.296	4.543	0.274	1.054	1.898	0.867	0.861
Tm	0.155	0.316	0.699	0.032	0.146	0.281	0.149	0.143
Yb	1.063	2.135	4.220	0.198	0.946	1.645	1.084	0.955
Lu	0.176	0.328	0.794	0.028	0.156	0.281	0.166	0.151
Hf	2.407	3.144	3.006	1.371	6.479	2.872	2.678	2.480
Ta	0.511	0.452	0.774	0.661	2.899	0.553	0.500	0.535
W	0.815	<0.2	2.432	3.677	12.281	22.921	0.801	0.402
Tl	0.121	0.664	0.135	<0.05	<0.05	<0.05	0.484	0.239
Pb	10.574	5.243	<5	<5	6.125	5.907	14.277	<5
Bi	0.486	0.239	0.362	0.242	0.584	0.560	0.394	0.063
Th	2.287	4.032	11.147	0.138	2.512	5.419	2.345	2.312
U	1.981	2.506	21.380	0.159	5.342	1.581	1.366	1.343
S (%)	0.340	0.060	2.600	1.450	6.500	0.100	0.450	0.310

Table B-2 (cont.): Whole-rock major and trace element data. Oxide concentrations are reported as percentages. Trace element data are reported as parts per million.

SAMPLE	TS17	TS18	TS19	TS20	TS21	TS22	TS23	TS24
SiO ₂	55.940	65.350	64.950	64.350	49.910	50.980	41.930	41.180
Al ₂ O ₃	14.100	15.210	15.000	15.740	13.170	12.770	15.450	11.880
FeO	9.330	5.050	5.030	2.100	10.160	9.940	12.990	15.890
Fe ₂ O ₃	0.611	0.067	0.090	2.966	1.509	2.553	3.094	4.441
MnO	0.110	0.090	0.090	0.060	0.180	0.150	0.180	0.250
MgO	6.490	1.820	1.890	1.810	5.190	5.720	5.680	5.850
CaO	4.980	3.960	3.830	2.710	11.550	9.470	13.010	11.890
Na ₂ O	1.390	3.610	3.510	5.530	2.730	3.050	1.610	1.470
K ₂ O	4.380	3.680	3.680	2.110	0.480	0.900	1.130	1.390
TiO ₂	0.510	0.630	0.630	0.260	1.780	1.770	1.950	1.480
P ₂ O ₅	0.170	0.120	0.260	0.060	0.120	0.120	0.150	0.110
LOI	1.160	0.530	0.760	1.790	2.120	0.660	0.350	0.230
TOTAL	100.210	100.117	100.280	99.720	100.020	99.220	98.970	97.830
Sc	25.000	15.000	15.000	8.000	44.000	43.000	51.000	36.000
Be	2.000	2.000	2.000	1.000	2.000	2.000	2.000	2.000
V	158.851	88.146	88.359	67.238	376.951	323.131	383.529	281.619
Cr	288.482	35.748	48.156	32.027	77.794	53.317	117.385	94.739
Co	27.926	10.363	12.143	11.866	27.303	19.525	18.915	23.067
Ni	43.568	25.084	40.641	13.300	69.412	65.730	83.554	50.728
Cu	11.164	23.811	27.432	342.985	17.308	33.444	14.036	16.918
Zn	52.971	67.507	86.797	43.961	60.651	57.510	71.856	46.327
Ga	15.640	18.187	20.554	16.110	17.770	17.758	19.012	14.816
Ge	1.694	1.373	1.515	0.847	1.584	1.601	2.021	1.847
As	16.273	<5	<5	5.935	11.921	20.084	<5	<5
Rb	166.534	104.580	129.788	88.331	5.914	26.282	13.079	13.981
Sr	465.174	419.521	474.150	336.254	189.561	194.097	148.847	63.148
Y	14.844	23.248	24.553	9.298	32.238	33.882	35.659	24.656
Zr	131.716	183.289	184.144	91.370	135.814	149.593	129.059	101.895
Nb	6.902	9.871	12.281	8.309	8.149	7.520	8.271	6.035
Mo	2.565	2.642	2.130	6.833	1.492	5.044	1.343	0.929
Ag	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5
In	<0.1	<0.1	<0.1	<0.1	0.111	0.128	0.192	0.349
Sn	1.063	1.742	1.994	0.838	1.282	1.781	1.762	2.556
Sb	1.288	0.370	0.385	0.610	2.414	2.064	1.124	0.619
Cs	4.332	2.901	3.305	1.298	0.243	0.503	0.695	0.652
Ba	2,400.000	974.979	1,100.000	988.426	120.789	1,060.000	80.739	72.201
La	26.777	37.306	36.467	12.618	10.320	10.728	14.374	12.192
Ce	52.718	70.728	67.208	20.243	24.635	25.134	33.331	29.016
Pr	6.035	7.353	7.964	2.462	3.477	3.527	4.338	3.609
Nd	24.874	25.829	29.738	9.401	14.618	16.589	18.581	14.163
Sm	5.536	5.637	5.761	1.966	4.354	5.333	5.428	3.657
Eu	1.067	1.157	1.084	0.523	1.434	1.774	2.863	2.136
Gd	3.989	4.383	4.861	1.760	4.640	4.721	4.832	3.794
Tb	0.583	0.665	0.767	0.277	0.864	0.996	0.970	0.691
Dy	2.922	4.111	4.307	1.529	5.364	6.607	6.520	4.347
Ho	0.506	0.783	0.803	0.282	1.151	1.246	1.316	0.953
Er	1.407	2.148	2.482	0.932	3.234	3.313	3.537	2.522
Tm	0.219	0.334	0.402	0.151	0.451	0.524	0.579	0.389
Yb	1.330	2.345	2.464	1.017	2.897	3.505	3.868	2.450
Lu	0.191	0.366	0.359	0.162	0.468	0.497	0.571	0.389
Hf	3.391	4.631	4.799	2.283	3.417	4.101	3.699	2.802
Ta	0.442	0.742	0.875	0.488	0.468	0.519	0.527	0.385
W	0.404	0.302	0.535	0.827	0.658	1.317	1.553	0.991
Tl	0.812	0.559	0.707	0.413	<0.05	0.074	0.053	0.051
Pb	5.753	13.961	19.701	5.234	<5	5.522	<5	<5
Bi	0.378	0.121	0.157	0.364	0.120	0.604	0.357	0.093
Th	5.833	8.589	8.698	2.513	1.430	1.447	2.368	1.518
U	1.956	2.555	3.139	1.299	0.718	1.430	3.410	3.347
S (%)	0.010	0.030	0.040	0.230	0.020	0.050	0.010	0.020

Table B-2 (cont.): Whole-rock major and trace element data. Oxide concentrations are reported as percentages. Trace element data are reported as parts per million.

SAMPLE	TS25	TS26	TS27	TS28	TS29	TS30	TS31	TS32
SiO ₂	56.030	36.320	43.790	41.910	46.130	52.280	54.850	58.260
Al ₂ O ₃	17.720	8.490	10.020	9.940	11.430	18.610	18.480	18.090
FeO	7.770	11.900	17.340	15.960	11.820	8.370	5.380	2.860
Fe ₂ O ₃	0.975	7.865	3.699	6.783	6.404	1.268	2.041	4.442
MnO	0.090	0.180	0.220	0.240	0.070	0.060	0.050	0.020
MgO	3.770	14.920	5.380	5.480	7.490	3.970	3.250	3.100
CaO	4.730	6.500	13.630	14.050	5.610	3.060	4.390	1.060
Na ₂ O	1.330	0.330	1.460	1.430	2.100	4.500	4.850	1.670
K ₂ O	3.100	2.280	0.660	0.620	2.750	3.780	2.900	5.180
TiO ₂	1.460	0.710	0.890	1.360	0.460	1.730	1.750	1.440
P ₂ O ₅	0.110	0.050	0.090	0.090	0.120	0.050	0.120	0.060
LOI	1.130	6.040	0.500	0.060	3.550	1.090	1.090	2.980
TOTAL	98.215	96.920	99.610	99.690	99.240	99.690	99.730	99.480
Sc	43.000	23.000	24.000	33.000	14.000	53.000	32.000	50.000
Be	2.000	1.000	1.000	2.000	4.000	2.000	4.000	2.000
V	238.314	205.973	277.771	277.914	868.218	308.277	626.920	319.521
Cr	179.347	171.762	34.380	39.894	96.222	144.170	227.505	148.462
Co	41.379	90.107	56.950	42.371	83.609	27.769	17.695	27.013
Ni	80.203	70.303	42.603	54.024	41.081	116.654	100.069	55.338
Cu	11.002	1,150.000	132.306	71.881	467.894	340.280	106.945	264.198
Zn	46.591	144.149	70.118	85.915	58.371	91.433	51.687	38.712
Ga	23.100	15.039	17.521	18.827	15.490	24.285	25.387	21.923
Ge	1.677	1.779	4.635	3.360	0.935	2.415	1.884	2.058
As	103.425	34.373	8.441	5.178	<5	6.997	6.434	9.525
Rb	106.447	97.522	14.434	7.555	122.539	146.822	126.220	106.570
Sr	168.160	13.818	110.572	102.287	159.340	303.262	367.672	93.675
Y	23.690	52.058	17.206	24.778	49.985	42.658	46.392	27.117
Zr	141.115	60.851	62.669	97.482	104.972	148.525	182.901	136.459
Nb	10.256	13.905	3.570	5.796	10.692	12.723	16.362	10.469
Mo	1.186	4.338	2.441	1.960	59.789	1.604	104.024	5.397
Ag	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5
In	<0.1	0.280	0.318	0.337	<0.1	<0.1	<0.1	<0.1
Sn	0.656	1.152	1.401	1.896	<0.5	3.124	1.780	<0.5
Sb	0.641	2.089	0.822	1.635	0.916	0.722	1.477	2.941
Cs	3.231	2.550	0.989	0.613	2.244	4.153	3.270	2.136
Ba	700.106	5,030.000	116.321	96.462	899.976	568.387	555.458	547.894
La	9.453	5.195	10.104	21.238	36.636	10.352	29.375	10.243
Ce	20.964	16.195	17.565	37.233	57.616	21.864	53.271	23.890
Pr	2.829	2.965	2.046	4.503	7.968	3.119	7.725	3.237
Nd	13.176	16.751	8.691	19.717	29.365	12.802	32.775	12.999
Sm	3.987	6.897	2.373	5.249	6.732	4.071	7.516	3.475
Eu	0.918	1.170	0.871	1.695	1.026	1.998	2.528	1.036
Gd	3.538	7.041	2.275	4.441	7.376	4.942	7.156	3.667
Tb	0.748	1.610	0.461	0.800	1.296	0.987	1.331	0.672
Dy	4.921	10.057	2.987	4.886	7.659	6.184	7.973	4.089
Ho	0.892	1.793	0.591	0.902	1.634	1.468	1.578	0.924
Er	2.426	4.292	1.616	2.436	5.046	4.279	4.628	2.599
Tm	0.396	0.646	0.250	0.375	0.792	0.569	0.689	0.348
Yb	2.588	4.019	1.700	2.544	4.959	3.708	4.320	2.280
Lu	0.363	0.525	0.268	0.377	0.867	0.611	0.646	0.389
Hf	3.706	1.840	1.751	2.681	2.980	3.964	4.943	3.575
Ta	0.700	0.519	0.217	0.405	0.904	0.649	1.035	0.587
W	1.576	1.837	0.776	1.006	1.450	1.258	1.283	1.514
Tl	0.701	0.916	0.070	<0.05	0.590	1.648	0.973	1.159
Pb	7.731	<5	<5	<5	<5	6.592	7.498	<5
Bi	0.144	1.061	0.310	0.210	0.264	0.278	0.290	0.200
Th	1.896	0.320	0.663	1.236	12.010	1.982	4.568	1.964
U	1.316	0.533	0.637	1.277	15.748	11.381	51.690	6.175
S (%)	0.030	0.030	1.150	0.590	6.000	0.940	0.530	2.550

Table B-2 (cont.): Whole-rock major and trace element data. Oxide concentrations are reported as percentages. Trace element data are reported as parts per million.

SAMPLE	TS33	TS34	TS35	TS36	TS37	TS38
SiO ₂	62.880	54.850	63.120	55.310	48.030	54.420
Al ₂ O ₃	15.800	8.690	14.570	13.140	12.150	13.030
FeO	1.240	7.780	3.200	7.280	9.710	10.500
Fe ₂ O ₃	5.342	3.694	3.394	3.239	3.109	1.521
MnO	0.020	0.220	0.030	0.080	0.120	0.150
MgO	2.740	9.900	4.430	6.090	10.080	4.490
CaO	0.810	10.150	4.510	4.300	6.940	8.420
Na ₂ O	0.360	1.730	2.880	2.130	1.890	3.750
K ₂ O	4.030	0.340	1.450	3.870	2.460	0.690
TiO ₂	1.400	0.310	0.450	0.500	0.520	1.720
P ₂ O ₅	<0.01	0.070	0.060	0.070	0.120	0.120
LOI	3.780	1.470	1.540	2.070	1.660	0.240
TOTAL	98.510	100.070	99.990	98.870	97.860	100.230
Sc	49.000	9.000	13.000	14.000	16.000	42.000
Be	2.000	3.000	6.000	4.000	4.000	2.000
V	288.189	582.486	1,120.000	998.212	733.315	348.008
Cr	134.151	55.765	86.095	107.475	112.120	69.292
Co	26.566	39.477	14.951	51.010	33.434	28.519
Ni	68.114	78.603	121.015	126.933	179.273	78.322
Cu	386.413	272.507	182.524	473.909	111.645	50.782
Zn	40.491	326.461	42.704	73.485	64.681	83.810
Ga	23.568	13.125	18.973	15.224	16.621	14.185
Ge	1.737	3.104	1.242	1.419	1.465	1.710
As	13.647	15.221	14.393	5.192	7.278	8.718
Rb	96.594	6.720	62.002	86.570	124.032	16.497
Sr	43.071	135.731	194.116	131.260	132.709	149.241
Y	26.015	42.082	74.981	51.127	31.009	33.765
Zr	131.441	109.803	257.936	158.775	99.352	119.629
Nb	11.651	10.336	17.390	12.440	8.178	8.006
Mo	21.508	58.652	119.132	96.085	56.740	3.508
Ag	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5
In	0.153	0.158	<0.1	<0.1	<0.1	<0.1
Sn	3.756	2.007	<0.5	0.747	0.801	1.567
Sb	1.957	1.880	5.365	1.319	1.577	2.231
Cs	0.802	0.138	1.625	2.417	3.738	0.194
Ba	1,020.000	298.416	689.701	2,720.000	387.931	101.302
La	16.241	12.870	78.602	48.184	33.422	8.736
Ce	31.366	24.780	117.972	80.838	53.041	23.551
Pr	4.390	3.778	15.478	10.612	7.180	3.165
Nd	19.815	17.152	59.612	36.414	26.516	14.018
Sm	5.075	5.311	12.400	7.321	6.121	4.377
Eu	1.043	0.867	1.111	1.123	1.191	1.639
Gd	4.549	4.770	11.064	8.064	5.529	4.802
Tb	0.821	0.891	1.864	1.161	0.759	0.897
Dy	4.850	5.908	11.830	6.745	4.340	5.689
Ho	0.906	1.243	2.306	1.515	0.887	1.191
Er	2.586	3.698	6.761	4.602	2.548	3.263
Tm	0.378	0.574	1.057	0.644	0.407	0.440
Yb	2.308	4.148	6.905	4.062	2.981	2.886
Lu	0.341	0.713	1.044	0.792	0.648	0.479
Hf	3.823	2.838	7.823	4.513	2.741	3.278
Ta	0.727	0.616	1.499	0.976	0.533	0.452
W	5.706	1.058	3.168	1.034	32.238	1.441
Tl	0.436	0.083	0.904	0.558	0.706	0.164
Pb	<5	6.738	11.719	13.885	<5	5.956
Bi	0.496	0.859	0.310	0.314	0.166	0.361
Th	2.077	10.850	16.411	14.315	10.414	1.464
U	11.879	8.763	18.916	16.977	12.420	1.981
S (%)	3.500	1.350	1.600	2.100	0.780	0.130

Table B-2 (cont.): Whole-rock major and trace element data. Oxide concentrations are reported as percentages. Trace element data are reported as parts per million.

APPENDIX C: MINERAL CHEMISTRY

Sample	Amp	Pyx	Gt	Plag	Crd	Bio	Ep	Chl	Scp	Sph	Au
H5-180		11	13								
HM1-69		3	1	2			2				
P5-16			14			12					
P6-43			25		2	5					
P6-91	14	9									
P8-46	3										
P8-68		6									
P9-27					3						
P10-103	4	2									
P13-39				4		5		4			
P13-67a	3					1					
P22-70	8										7
P22-70a	10										
P22-72											6
R6-284	24			12		2					
R7-104b				1		1	3	10			
R7-296	29			15							
R7-320	9										
R8-169a		4	9	2			3		6		
R8-193		3		4							
R8-216b	18			17							
R8-219	6			1							3
R8-240	6										
R8-253	1	1									4
R12-19	19						3				
R12-27	6						5				
R12-51	8									18	
R12-119	7	8	4				4				
R12-219	7	3									

Table C-1: Summary table of electron-microprobe analyses, detailing the number of analyses performed on different minerals from each section.

Sample	Description
H5-180	Garnet-pyroxene-sphene (calcic skarn) in amphibolite.
HM1-69	Zoned garnet-pyroxene-epidote calcic skarn vein in porphyry.
P5-16	Biotite hornfels (metasediment) with garnet and biotite.
P6-43	Biotite hornfels (metasediment) with garnet, cordierite and biotite.
P6-91	Amphibolitisation (amphibole associated with pyroxene) of metavolcanic.
P8-46	Intense amphibolitisation with associated pyroxene and sulphides. In amphibolite.
P8-68	Pyroxene-amphibole alteration in metavolcanic portion of tuffite.
P9-27	Biotite hornfels (metasediment) with cordierite and biotite.
P10-103	Intense amphibolitisation with associated pyroxene, magnetite and sulphides. In amphibolite.
P13-39	Porphyry with skarn vein and sulphides. Retrograde chlorite replacing biotite.
P13-67a	Amphibole and biotite clasts in a sulphide cataclastites.
P22-70	Intense amphibolitisation with magnetite and gold. Retrograde amphibole in cores.
P22-70a	Intense amphibolitisation with magnetite and gold. Retrograde amphibole in cores.
P22-72	Gold in fractures in intense amphibolitisation.
R6-284	Amphibolite with amphibole veinlets (including core and rim type amphiboles).
R7-104b	Retrograde biotite hornfels. Epidote and chlorite replacing cordierite.
R7-296	Amphibolite with amphibole and quartz veinlets.
R7-320	Retrograde amphibole in cores. Amphibole vein associated with sulphides.
R8-169a	Calcic skarn vein (garnet-pyroxene-epidote-scapolite). In amphibolite.
R8-193	Spotted amphibolite, with pyroxene in some of the spots.
R8-216b	Metamorphic amphibole. Amphibole veinlets. Amphibole in quartz vein.
R8-219	Amphibolite with amphibolitisation and quartz veins. Gold in alteration.
R8-240	Amphibolite with amphibole veinlets (including core and rim type amphiboles).
R8-253	Amphibolite veins with sulphides. Gold in pyrrhotite in amphibole vein.
R12-19	Amphibole alteration associated with carbonate in amphibolite.
R12-27	Amphibole alteration associated with carbonate, epidote and sulphide veins. In amphibolite.
R12-51	Coarse amphibole with sulphides (+ sphalerite) in carbonate skarn.
R12-119	Calcic skarn vein (garnet-pyroxene-amphibole-epidote).
R12-219	Veinlet of amphibolitisation (amphibole + pyroxene) in porphyry.

Table C-2: Description of samples used for electron-microprobe studies

METAMORPHIC AMPHIBOLES										
Section	R7-296	R7-296	R7-296	R7-296	R7-296	R7-296	R7-296	R7-296	R7-296	R7-296
Analysis	(1-1) 1	(1-1) 2	(1-1) 1	(1-1) 2	(1-1) 3	(1-2) 1	(1-2) 2	(1-2) 3	(1-2) 4	(1-2) 5
SiO ₂	45.662	45.695	45.363	45.813	45.398	46.839	45.180	45.122	45.423	46.090
TiO ₂	0.806	0.855	0.973	0.916	0.883	0.839	0.857	1.051	0.903	0.970
Al ₂ O ₃	8.737	8.743	8.401	8.136	8.357	7.431	8.734	8.131	8.111	8.234
Cr ₂ O ₃	0.113	0.000	0.100	0.000	0.000	0.000	0.000	0.000	0.000	0.076
FeO	19.403	19.672	19.242	19.416	19.670	19.289	19.708	19.672	19.484	19.673
MnO	0.218	0.284	0.220	0.224	0.270	0.256	0.206	0.274	0.303	0.192
MgO	8.973	9.018	9.559	9.692	9.553	10.255	9.373	9.703	9.685	9.700
CaO	11.361	11.732	12.084	11.778	11.022	10.566	10.610	11.301	10.903	10.998
Na ₂ O	0.881	1.010	0.978	1.027	1.018	0.843	1.060	0.986	0.947	1.041
K ₂ O	0.627	0.666	0.533	0.543	0.532	0.330	0.388	0.397	0.434	0.283
Cl			0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
F			0.113	0.162	0.000	0.191	0.000	0.000	0.000	0.000
Total	96.781	97.675	97.453	97.545	96.703	96.648	96.116	96.637	96.193	97.257
-O=Cl,F	0.000	0.000	0.048	0.068	0.000	0.080	0.000	0.000	0.000	0.000
Total	96.781	97.675	97.405	97.477	96.703	96.568	96.116	96.637	96.193	97.257
Number of ions on the basis of 23O using the calculated average Fe ³⁺ content										
Si	6.90	6.87	6.82	6.87	6.85	7.02	6.84	6.81	6.87	6.89
Al ^{iv}	1.10	1.13	1.18	1.13	1.15	0.98	1.16	1.19	1.13	1.11
sum T	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
Al ^{vi}	0.46	0.42	0.31	0.31	0.34	0.33	0.40	0.26	0.32	0.35
Ti	0.09	0.10	0.11	0.10	0.10	0.09	0.10	0.12	0.10	0.11
Fe ³⁺	0.24	0.21	0.30	0.32	0.44	0.42	0.48	0.51	0.49	0.43
Cr	0.01	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.01
Mg	2.02	2.02	2.14	2.17	2.15	2.29	2.11	2.18	2.18	2.16
Fe ²⁺	2.16	2.24	2.11	2.09	1.95	1.86	1.90	1.91	1.89	1.93
Mn	0.01	0.02	0.01	0.01	0.02	0.00	0.01	0.02	0.02	0.01
sum C	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00
Mg	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Fe ²⁺	0.05	0.03	0.01	0.04	0.08	0.14	0.12	0.06	0.09	0.10
Mn	0.01	0.02	0.01	0.01	0.02	0.03	0.01	0.02	0.02	0.01
Ca	1.84	1.89	1.95	1.89	1.78	1.70	1.72	1.83	1.77	1.76
Na	0.09	0.06	0.03	0.06	0.12	0.14	0.15	0.09	0.12	0.13
sum B	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00
Na	0.17	0.24	0.26	0.24	0.18	0.11	0.16	0.20	0.15	0.18
K	0.12	0.13	0.10	0.10	0.10	0.06	0.07	0.08	0.08	0.05
sum A	0.29	0.36	0.36	0.35	0.28	0.17	0.24	0.27	0.24	0.23
Total	15.29	15.36	15.36	15.35	15.28	15.17	15.24	15.27	15.24	15.23

Table C-3: Amphibole electron-microprobe analyses, with stoichiometry based on 23O, and using the average calculated Fe³⁺ content according to Schumacher (IMA, 1997).

METAMORPHIC AMPHIBOLES										
Section	R6-284	R8-216b	R8-216b	R8-216b	R8-216b	R8-216b	R8-216b	R8-216b	R8-216b	R8-216b
Analysis	(4-1) 2	(1-1) 1	(1-1) 2	(2-1) 1	1	2	3	4	5	6
SiO ₂	45.969	47.353	47.687	47.473	47.913	46.428	47.720	45.682	45.969	46.690
TiO ₂	0.987	0.704	0.820	0.712	0.518	0.982	0.823	0.684	0.784	0.474
Al ₂ O ₃	7.611	6.209	6.850	7.031	7.019	6.913	6.564	9.081	8.369	8.509
Cr ₂ O ₃	0.000	0.225	0.000	0.076	0.081	0.116	0.161	0.000	0.000	0.000
FeO	18.915	15.917	16.363	16.803	15.194	15.424	15.123	17.011	16.077	15.965
MnO	0.213	0.208	0.200	0.227	0.253	0.253	0.277	0.154	0.284	0.217
MgO	10.262	12.637	12.528	12.278	12.339	11.980	12.728	10.657	11.342	11.419
CaO	11.889	11.931	12.199	12.157	11.934	11.787	12.129	12.007	12.157	12.008
Na ₂ O	0.962	0.735	0.705	0.825	0.828	0.912	0.934	1.079	1.024	0.929
K ₂ O	0.440	0.240	0.373	0.345	0.287	0.406	0.368	0.530	0.518	0.444
Cl	0.000	0.348	0.000	0.425	0.000	0.000	0.000	0.000	0.000	0.000
F	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Total	97.248	96.159	97.725	97.927	96.366	95.201	96.827	96.885	96.524	96.655
-O=Cl,F	0.000	0.079	0.000	0.096	0.000	0.000	0.000	0.000	0.000	0.000
Total	97.248	96.080	97.725	97.831	96.366	95.201	96.827	96.885	96.524	96.655
Number of ions on the basis of 23O using the calculated average Fe ³⁺ content										
Si	6.89	7.04	6.99	6.96	7.11	7.01	7.06	6.83	6.88	6.95
Al ^{iv}	1.11	0.96	1.01	1.04	0.89	0.99	0.94	1.17	1.12	1.05
sum T	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
Al ^{vi}	0.24	0.13	0.18	0.18	0.34	0.24	0.21	0.43	0.36	0.44
Ti	0.11	0.08	0.09	0.08	0.06	0.11	0.09	0.08	0.09	0.05
Fe ³⁺	0.38	0.49	0.47	0.49	0.25	0.26	0.27	0.24	0.24	0.25
Cr	0.00	0.03	0.00	0.01	0.01	0.01	0.02	0.00	0.00	0.00
Mg	2.29	2.80	2.74	2.68	2.73	2.70	2.81	2.38	2.53	2.53
Fe ²⁺	1.96	1.46	1.51	1.55	1.60	1.66	1.59	1.86	1.76	1.72
Mn	0.01	0.01	0.01	0.01	0.02	0.02	0.02	0.01	0.02	0.01
sum C	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00
Mg	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Fe ²⁺	0.03	0.03	0.03	0.03	0.03	0.03	0.02	0.03	0.01	0.03
Mn	0.01	0.01	0.01	0.01	0.02	0.02	0.02	0.01	0.02	0.01
Ca	1.91	1.90	1.92	1.91	1.90	1.91	1.92	1.92	1.95	1.91
Na	0.05	0.05	0.04	0.05	0.06	0.05	0.04	0.04	0.03	0.05
sum B	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00
Na	0.23	0.16	0.16	0.19	0.18	0.22	0.23	0.27	0.27	0.22
K	0.08	0.05	0.07	0.06	0.05	0.08	0.07	0.10	0.10	0.08
sum A	0.32	0.20	0.23	0.25	0.24	0.30	0.30	0.37	0.37	0.31
Total	15.32	15.20	15.23	15.25	15.24	15.30	15.30	15.37	15.37	15.31

Table C-3 (cont.): Amphibole electron-microprobe analyses, with stoichiometry based on 23O, and using the average calculated Fe³⁺ content according to Schumacher (IMA, 1997).

METAMORPHIC AMPHIBOLES										
Section	R8-216b	R8-216b	R8-216b	R7-296	R7-296	R7-296	R7-296	R7-296	R7-296	R7-296
Analysis	9	11	12	1.1	1.2	1.3	2.1	2.2	4.1	4.3
SiO ₂	47.025	48.092	48.497	45.196	45.016	45.252	46.247	45.942	44.754	44.912
TiO ₂	0.587	0.691	0.398	0.924	1.021	0.707	0.729	0.685	0.788	0.732
Al ₂ O ₃	7.089	6.825	6.655	8.333	8.133	8.222	7.640	7.484	8.278	8.492
Cr ₂ O ₃	0.058	0.170	0.000	0.071	0.106	0.000	0.000	0.049	0.000	0.000
FeO	15.613	14.838	14.994	19.239	19.471	19.454	18.884	19.258	19.831	18.860
MnO	0.312	0.273	0.277	0.291	0.287	0.385	0.326	0.212	0.145	0.232
MgO	12.350	12.513	12.675	9.516	9.660	9.656	10.042	10.151	9.761	9.667
CaO	11.872	11.814	12.081	11.564	11.383	11.779	11.511	11.858	11.591	11.866
Na ₂ O	0.879	0.863	0.741	1.181	1.158	1.078	1.006	1.016	1.120	1.143
K ₂ O	0.392	0.381	0.271	0.437	0.483	0.437	0.406	0.394	0.522	0.483
Cl	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
F	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Total	96.177	96.460	96.589	96.752	96.718	96.970	96.791	97.049	96.790	96.387
-O=Cl,F	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Total	96.177	96.460	96.589	96.752	96.718	96.970	96.791	97.049	96.790	96.387
Number of ions on the basis of 23O using the calculated average Fe ³⁺ content										
Si	7.01	7.13	7.16	6.84	6.81	6.83	6.96	6.91	6.77	6.82
Al ^{iv}	0.99	0.87	0.84	1.16	1.19	1.17	1.04	1.09	1.23	1.18
sum T	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
Al ^{vi}	0.25	0.32	0.32	0.33	0.26	0.29	0.31	0.23	0.24	0.34
Ti	0.07	0.08	0.04	0.11	0.12	0.08	0.08	0.08	0.09	0.08
Fe ³⁺	0.39	0.19	0.26	0.32	0.41	0.43	0.34	0.42	0.51	0.32
Cr	0.01	0.02	0.00	0.01	0.01	0.00	0.00	0.01	0.00	0.00
Mg	2.74	2.76	2.79	2.15	2.18	2.17	2.25	2.28	2.20	2.19
Fe ²⁺	1.53	1.61	1.57	2.08	2.00	2.01	1.99	1.97	1.95	2.05
Mn	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.01	0.01	0.01
sum C	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00
Mg	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Fe ²⁺	0.03	0.04	0.02	0.04	0.05	0.02	0.05	0.03	0.05	0.02
Mn	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.01	0.01	0.01
Ca	1.89	1.88	1.91	1.88	1.85	1.90	1.86	1.91	1.88	1.93
Na	0.06	0.07	0.05	0.07	0.08	0.05	0.08	0.05	0.07	0.04
sum B	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00
Na	0.20	0.18	0.16	0.28	0.26	0.26	0.22	0.25	0.26	0.30
K	0.07	0.07	0.05	0.08	0.09	0.08	0.08	0.08	0.10	0.09
sum A	0.27	0.25	0.22	0.36	0.35	0.35	0.29	0.32	0.36	0.39
Total	15.27	15.25	15.22	15.36	15.35	15.35	15.29	15.32	15.36	15.39

Table C-3 (cont.): Amphibole electron-microprobe analyses, with stoichiometry based on 23O, and using the average calculated Fe³⁺ content according to Schumacher (IMA, 1997).

METAMORPHIC AMPHIBOLES							CORE OF AMPHIBOLE VEINS			
Section	R7-296	R6-284	R6-284	R6-284	R6-284	R6-284	R6-284	R6-284	R6-284	R6-284
Analysis	4.4	1.2	2.6	2.7	3.1	3.2	(1-1) 1	(1-1) 2	(1-1) 3	(1-1) 4
SiO ₂	45.229	46.539	46.480	47.844	48.261	47.831	50.530	50.896	51.056	50.664
TiO ₂	0.835	0.821	0.866	0.874	0.754	0.827	0.227	0.223	0.084	0.214
Al ₂ O ₃	7.865	7.254	7.531	6.932	6.634	6.714	4.520	4.548	4.599	4.413
Cr ₂ O ₃	0.022	0.035	0.040	0.031	0.000	0.027	0.000	0.000	0.000	0.000
FeO	19.676	18.815	18.627	18.538	17.719	18.237	16.674	16.613	17.543	16.647
MnO	0.369	0.208	0.228	0.236	0.189	0.181	0.225	0.220	0.253	0.200
MgO	9.955	10.143	9.841	10.218	10.503	10.537	11.985	12.052	11.763	12.265
CaO	11.616	11.345	11.201	11.477	11.802	11.842	11.559	11.888	11.609	12.131
Na ₂ O	1.141	1.040	1.115	1.021	0.937	1.075	0.454	0.460	0.725	0.478
K ₂ O	0.430	0.311	0.301	0.312	0.233	0.303	0.000	0.000	0.000	0.109
Cl	0.000	0.000	0.000	0.000	0.000	0.000	0.343	0.000	0.000	0.476
F	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Total	97.138	96.511	96.230	97.483	97.032	97.574	96.174	96.900	97.632	97.121
-O=Cl,F	0.000	0.000	0.000	0.000	0.000	0.000	0.077	0.000	0.000	0.107
Total	97.138	96.511	96.230	97.483	97.032	97.574	96.097	96.900	97.632	97.014
Number of ions on the basis of 23O using the calculated average Fe ³⁺ content										
Si	6.81	7.02	7.03	7.13	7.20	7.13	7.50	7.51	7.50	7.47
Al ^{iv}	1.19	0.98	0.97	0.87	0.80	0.87	0.50	0.49	0.50	0.53
sum T	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
Al ^{vi}	0.20	0.30	0.37	0.35	0.37	0.31	0.29	0.30	0.29	0.23
Ti	0.09	0.09	0.10	0.10	0.08	0.09	0.03	0.02	0.01	0.02
Fe ³⁺	0.51	0.31	0.25	0.20	0.10	0.14	0.21	0.17	0.21	0.21
Cr	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Mg	2.23	2.28	2.22	2.27	2.34	2.34	2.65	2.65	2.57	2.69
Fe ²⁺	1.93	2.00	2.05	2.07	2.09	2.10	1.82	1.85	1.90	1.82
Mn	0.02	0.01	0.01	0.01	0.01	0.01	0.00	0.01	0.02	0.01
sum C	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00
Mg	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Fe ²⁺	0.04	0.07	0.06	0.04	0.02	0.03	0.04	0.03	0.05	0.01
Mn	0.02	0.01	0.01	0.01	0.01	0.01	0.03	0.01	0.02	0.01
Ca	1.87	1.83	1.81	1.83	1.89	1.89	1.84	1.88	1.83	1.92
Na	0.07	0.09	0.11	0.11	0.08	0.07	0.09	0.07	0.11	0.06
sum B	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00
Na	0.27	0.21	0.21	0.18	0.19	0.24	0.04	0.06	0.09	0.08
K	0.08	0.06	0.06	0.06	0.04	0.06	0.00	0.00	0.00	0.02
sum A	0.35	0.27	0.27	0.24	0.23	0.30	0.04	0.06	0.09	0.10
Total	15.35	15.27	15.27	15.24	15.23	15.30	15.04	15.06	15.09	15.10

Table C-3 (cont.): Amphibole electron-microprobe analyses, with stoichiometry based on 23O, and using the average calculated Fe³⁺ content according to Schumacher (IMA, 1997).

CORE OF AMPHIBOLE VEINS										
Section	R6-284	R6-284	R6-284	R6-284	R6-284	R6-284	R6-284	R8-240	R8-240	R8-240
Analysis	(1-1) 5	(1-2) 1	(1-2) 2	(3-1) 4	(3-1) 5	(3-1) 6	(3-1) 7	(1-1) 4	(1-1) 5	(1-1) 6
SiO ₂	50.437	50.343	50.409	52.104	49.553	51.324	52.284	44.377	44.146	46.948
TiO ₂	0.213	0.264	0.222	0.204	0.479	0.359	0.291	0.675	0.736	0.602
Al ₂ O ₃	4.721	4.759	4.818	2.970	5.264	3.959	3.396	7.455	7.302	5.759
Cr ₂ O ₃	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
FeO	17.022	17.478	17.448	15.816	17.134	16.570	16.430	21.109	21.202	19.591
MnO	0.216	0.198	0.202	0.200	0.217	0.259	0.216	0.312	0.226	0.274
MgO	12.341	12.300	11.973	13.412	11.633	12.571	13.166	9.789	9.703	11.320
CaO	11.656	11.491	10.720	10.919	11.346	11.600	11.901	11.144	11.296	11.355
Na ₂ O	0.621	0.659	0.631	0.457	0.687	0.453	0.448	1.119	1.236	1.065
K ₂ O	0.186	0.145	0.000	0.000	0.000	0.000	0.000	0.394	0.375	0.222
Cl	0.448	0.000	0.895	0.361	0.267	0.366	0.000	0.000	0.385	0.350
F	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Total	97.413	97.637	96.423	96.082	96.313	97.095	98.132	96.374	96.222	97.136
-O=Cl,F	0.101	0.000	0.202	0.081	0.060	0.083	0.000	0.000	0.087	0.079
Total	97.312	97.637	96.221	96.001	96.253	97.012	98.132	96.374	96.135	97.057
Number of ions on the basis of 23O using the calculated average Fe ³⁺ content										
Si	7.40	7.39	7.46	7.69	7.36	7.54	7.59	6.74	6.73	7.02
Al ^{iv}	0.60	0.61	0.54	0.31	0.64	0.46	0.41	1.26	1.27	0.98
sum T	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
Al ^{vi}	0.22	0.21	0.30	0.21	0.28	0.22	0.18	0.07	0.05	0.04
Ti	0.02	0.03	0.02	0.02	0.05	0.04	0.03	0.08	0.08	0.07
Fe ³⁺	0.33	0.31	0.31	0.15	0.29	0.21	0.18	0.83	0.78	0.59
Cr	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Mg	2.70	2.69	2.64	2.95	2.58	2.75	2.85	2.22	2.21	2.52
Fe ²⁺	1.72	1.75	1.72	1.67	1.78	1.77	1.77	1.79	1.87	1.78
Mn	0.01	0.00	0.00	0.00	0.01	0.00	0.00	0.02	0.01	0.00
sum C	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00
Mg	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Fe ²⁺	0.05	0.08	0.12	0.14	0.05	0.05	0.05	0.07	0.06	0.08
Mn	0.01	0.02	0.03	0.03	0.02	0.03	0.03	0.02	0.01	0.03
Ca	1.83	1.81	1.70	1.73	1.81	1.83	1.85	1.81	1.85	1.82
Na	0.11	0.09	0.15	0.11	0.12	0.09	0.07	0.10	0.08	0.07
sum B	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00
Na	0.07	0.09	0.03	0.02	0.07	0.04	0.06	0.23	0.28	0.24
K	0.03	0.03	0.00	0.00	0.00	0.00	0.00	0.08	0.07	0.04
sum A	0.11	0.12	0.03	0.02	0.07	0.04	0.06	0.31	0.36	0.28
Total	15.11	15.12	15.03	15.02	15.07	15.04	15.06	15.31	15.36	15.28

Table C-3 (cont.): Amphibole electron-microprobe analyses, with stoichiometry based on 23O, and using the average calculated Fe³⁺ content according to Schumacher (IMA, 1997).

RIM OF AMPHIBOLE VEINS										
Section	R6-284	R6-284	R6-284	R6-284	R6-284	R6-284	R6-284	R8-240	R8-240	R8-240
Analysis	(2-1) 1	(2-1) 2	(2-1) 3	(2-1) 4	(3-1) 1	(3-1) 2	(3-1) 3	(1-1) 1	(1-1) 2	(1-1) 3
SiO ₂	48.419	48.726	46.848	48.041	47.314	47.237	47.906	42.992	46.447	46.966
TiO ₂	0.813	0.677	0.910	0.817	0.858	0.921	0.964	0.540	0.484	0.589
Al ₂ O ₃	6.584	6.473	6.939	6.882	7.035	7.245	6.984	7.688	6.152	5.683
Cr ₂ O ₃	0.000	0.000	0.000	0.000	0.063	0.000	0.000	0.000	0.000	0.000
FeO	18.909	18.612	18.105	18.674	18.308	18.222	18.446	22.043	20.440	19.172
MnO	0.233	0.213	0.221	0.262	0.132	0.262	0.116	0.229	0.238	0.252
MgO	10.411	10.660	10.497	10.586	10.453	10.476	10.398	9.329	10.973	11.455
CaO	11.251	11.241	11.545	11.822	11.771	11.675	11.288	11.492	11.654	11.454
Na ₂ O	0.822	0.873	0.942	0.761	0.902	0.843	0.926	1.126	0.908	0.975
K ₂ O	0.189	0.213	0.324	0.285	0.382	0.376	0.316	0.385	0.163	0.109
Cl	0.000	0.000	0.000	0.000	0.385	0.512	0.328	0.349	0.000	0.361
F	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Total	97.631	97.688	96.331	98.130	97.218	97.257	97.344	95.824	97.459	96.655
-O=Cl,F	0.000	0.000	0.000	0.000	0.087	0.116	0.074	0.079	0.000	0.081
Total	97.631	97.688	96.331	98.130	97.131	97.141	97.270	95.745	97.459	96.574
Number of ions on the basis of 23O using the calculated average Fe ³⁺ content										
Si	7.17	7.20	7.06	7.10	7.07	7.04	7.12	6.62	6.94	7.03
Al ^{iv}	0.83	0.80	0.94	0.90	0.93	0.96	0.88	1.38	1.06	0.97
sum T	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
Al ^{vi}	0.32	0.33	0.29	0.30	0.31	0.31	0.34	0.01	0.02	0.04
Ti	0.09	0.08	0.10	0.09	0.10	0.10	0.11	0.06	0.05	0.07
Fe ³⁺	0.30	0.29	0.25	0.29	0.24	0.30	0.29	0.88	0.71	0.63
Cr	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.00
Mg	2.30	2.35	2.36	2.33	2.33	2.33	2.30	2.14	2.44	2.56
Fe ²⁺	1.97	1.94	1.98	1.98	2.01	1.94	1.95	1.91	1.77	1.71
Mn	0.01	0.01	0.01	0.02	0.01	0.02	0.01	0.00	0.00	0.00
sum C	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00
Mg	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Fe ²⁺	0.07	0.07	0.05	0.04	0.03	0.03	0.05	0.05	0.07	0.06
Mn	0.01	0.01	0.01	0.02	0.01	0.02	0.01	0.03	0.03	0.03
Ca	1.79	1.78	1.86	1.87	1.88	1.86	1.80	1.90	1.86	1.84
Na	0.13	0.14	0.07	0.07	0.07	0.09	0.14	0.02	0.04	0.07
sum B	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00
Na	0.11	0.11	0.20	0.15	0.19	0.16	0.12	0.31	0.23	0.21
K	0.04	0.04	0.06	0.05	0.07	0.07	0.06	0.08	0.03	0.02
sum A	0.14	0.16	0.27	0.20	0.26	0.23	0.18	0.39	0.26	0.24
Total	15.14	15.16	15.27	15.20	15.26	15.23	15.18	15.39	15.26	15.24

Table C-3 (cont.): Amphibole electron-microprobe analyses, with stoichiometry based on 23O, and using the average calculated Fe³⁺ content according to Schumacher (IMA, 1997).

RETROGRADE AMPHIBOLE IN CORES										
Section	P22-70a	P22-70a	P22-70	P22-70	P22-70	P22-70	P22-70	P22-70	R7-320	R7-320
Analysis	(1-1) 3	(2-1) 1	(2-1) 2	(2-1) 3	(2-1) 4	(2-2) 1	(2-2) 3	(2-2) 4	(1-1) 1	(1-2) 1
SiO ₂	53.068	52.916	52.006	47.399	52.678	43.908	46.801	52.451	55.022	54.800
TiO ₂	0.000	0.045	0.048	0.187	0.000	0.221	0.285	0.000	0.000	0.000
Al ₂ O ₃	1.422	1.732	1.880	5.375	1.073	8.826	6.105	1.902	0.649	0.937
Cr ₂ O ₃	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
FeO	15.383	14.918	19.255	21.604	18.158	24.037	21.372	18.728	12.063	12.997
MnO	0.280	0.357	0.330	0.318	0.304	0.317	0.318	0.308	0.267	0.263
MgO	13.908	14.218	12.580	10.157	13.070	7.883	10.204	12.780	17.151	16.330
CaO	12.429	12.367	11.779	11.589	11.643	10.799	11.358	11.667	12.298	11.442
Na ₂ O	0.142	0.128	0.164	0.485	0.078	0.726	0.550	0.170	0.054	0.085
K ₂ O	0.000	0.000	0.000	0.342	0.000	0.466	0.170	0.000	0.000	0.000
Cl	0.000	0.000	0.000	0.357	0.000	0.713	0.297	0.000	0.000	0.000
F	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Total	96.632	96.681	98.042	97.456	97.004	97.183	97.163	98.006	97.504	96.854
-O=Cl,F	0.000	0.000	0.000	0.081	0.000	0.161	0.067	0.000	0.000	0.000
Total	96.632	96.681	98.042	97.375	97.004	97.022	97.096	98.006	97.504	96.854
Number of ions on the basis of 23O using the calculated average Fe ³⁺ content										
Si	7.80	7.76	7.66	7.09	7.80	6.66	7.00	7.69	7.88	7.92
Al ^{iv}	0.20	0.24	0.33	0.91	0.19	1.34	1.00	0.31	0.11	0.08
sum T	8.00	8.00	7.99	8.00	7.99	8.00	8.00	8.00	7.99	8.00
Al ^{vi}	0.05	0.06	0.00	0.04	0.00	0.24	0.08	0.02	0.00	0.08
Ti	0.00	0.00	0.01	0.02	0.00	0.03	0.03	0.00	0.00	0.00
Fe ³⁺	0.15	0.18	0.25	0.76	0.15	0.96	0.82	0.29	0.08	0.01
Cr	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Mg	3.05	3.11	2.76	2.27	2.89	1.78	2.28	2.79	3.66	3.52
Fe ²⁺	1.74	1.64	1.99	1.90	1.96	2.00	1.79	1.90	1.26	1.39
Mn	0.02	0.01	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00
sum C	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00
Mg	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Fe ²⁺	0.00	0.01	0.14	0.05	0.14	0.10	0.06	0.11	0.11	0.17
Mn	0.02	0.03	0.04	0.03	0.04	0.04	0.04	0.04	0.03	0.03
Ca	1.96	1.94	1.86	1.86	1.85	1.76	1.82	1.83	1.89	1.77
Na	0.02	0.02	0.00	0.07	0.00	0.11	0.08	0.02	0.00	0.02
sum B	2.00	2.00	2.04	2.00	2.02	2.00	2.00	2.00	2.03	2.00
Na	0.02	0.02	0.05	0.07	0.02	0.11	0.08	0.02	0.01	0.00
K	0.00	0.00	0.00	0.07	0.00	0.09	0.03	0.00	0.00	0.00
sum A	0.02	0.02	0.05	0.14	0.02	0.20	0.11	0.02	0.01	0.00
Total	15.02	15.02	15.07	15.14	15.04	15.20	15.11	15.02	15.03	15.00

Table C-3 (cont.): Amphibole electron-microprobe analyses, with stoichiometry based on 23O, and using the average calculated Fe³⁺ content according to Schumacher (IMA, 1997).

RETROG-RADE		AMPHIBOLE VEINLETS								QUARTZ VEIN
Section	R7-320	R7-296	R7-296	R7-296	R7-296	R7-296	R8-216b	R8-216b	R8-216b	R7-296
Analysis	(1-3) 1	(1-2) 1	(1-2) 2	(1-2) 3	(1-2) 4	(5-1) 2	(3-1) 1	(3-1) 2	(3-2) 1	(4-1) 2
SiO ₂	55.080	45.379	45.164	45.964	46.945	45.722	48.503	48.411	47.948	43.343
TiO ₂	0.037	1.006	0.729	0.953	0.726	0.906	0.720	0.468	0.688	0.890
Al ₂ O ₃	0.662	8.493	8.470	7.791	7.820	8.377	6.067	5.957	6.714	9.627
Cr ₂ O ₃	0.000	0.000	0.000	0.000	0.142	0.000	0.000	0.000	0.000	0.000
FeO	13.437	19.272	19.099	18.506	18.485	19.829	15.721	15.586	16.300	21.268
MnO	0.253	0.272	0.327	0.182	0.332	0.246	0.174	0.238	0.188	0.212
MgO	15.995	9.319	9.585	9.739	9.798	9.779	13.004	13.292	12.662	8.103
CaO	11.818	11.763	11.575	11.671	11.585	10.594	12.176	11.919	11.738	13.314
Na ₂ O	0.074	0.990	0.861	0.924	0.955	1.030	0.646	0.630	0.679	1.126
K ₂ O	0.000	0.666	0.575	0.669	0.605	0.468	0.220	0.224	0.338	0.576
Cl	0.000					0.000	0.000	0.403	0.000	0.000
F	0.000					0.000	0.000	0.000	0.000	0.067
Total	97.356	97.160	96.385	96.399	97.393	96.951	97.231	96.725	97.255	98.459
-O=Cl,F	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.091	0.000	0.028
Total	97.356	97.160	96.385	96.399	97.393	96.951	97.231	96.634	97.255	98.431
Number of ions on the basis of 23O using the calculated average Fe ³⁺ content										
Si	7.95	6.85	6.84	6.97	7.03	6.86	7.12	7.12	7.04	6.56
Al ^{iv}	0.05	1.15	1.16	1.03	0.97	1.14	0.88	0.88	0.96	1.44
sum T	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
Al ^{vi}	0.06	0.36	0.35	0.36	0.41	0.34	0.17	0.15	0.20	0.28
Ti	0.00	0.11	0.08	0.11	0.08	0.10	0.08	0.05	0.08	0.10
Fe ³⁺	0.00	0.24	0.41	0.16	0.17	0.50	0.43	0.54	0.52	0.35
Cr	0.00	0.00	0.00	0.00	0.02	0.00	0.00	0.00	0.00	0.00
Mg	3.44	2.10	2.16	2.20	2.19	2.19	2.84	2.91	2.77	1.83
Fe ²⁺	1.49	2.17	1.97	2.15	2.11	1.86	1.47	1.33	1.43	2.35
Mn	0.00	0.02	0.02	0.01	0.02	0.01	0.01	0.01	0.01	0.03
sum C	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	4.93
Mg	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Fe ²⁺	0.12	0.03	0.04	0.03	0.03	0.12	0.03	0.04	0.06	0.00
Mn	0.03	0.02	0.02	0.01	0.02	0.02	0.01	0.01	0.01	0.00
Ca	1.83	1.90	1.88	1.90	1.86	1.70	1.91	1.88	1.85	2.16
Na	0.02	0.05	0.07	0.06	0.09	0.15	0.05	0.07	0.08	0.00
sum B	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.16
Na	0.00	0.24	0.19	0.21	0.19	0.15	0.14	0.11	0.11	0.33
K	0.00	0.13	0.11	0.13	0.12	0.09	0.04	0.04	0.06	0.11
sum A	0.00	0.37	0.30	0.34	0.30	0.24	0.18	0.16	0.17	0.44
Total	15.00	15.37	15.30	15.34	15.30	15.24	15.18	15.16	15.17	15.53

Table C-3 (cont.): Amphibole electron-microprobe analyses, with stoichiometry based on 23O, and using the average calculated Fe³⁺ content according to Schumacher (IMA, 1997).

IN QUARTZ VEIN						ON RIM OF QUARTZ VEIN				
Section	R8-216b	R8-216b	R8-216b	R8-219	R8-219	R7-296	R7-296	R7-296	R7-296	R7-296
Analysis	(4-1) 2	(4-2) 1	(4-3) 1	(1-2) 1	(1-2) 2	(2-1) 1	(2-1) 2	(2-1) 3	(2-1) 4	(2-1) 5
SiO ₂	48.635	47.167	49.114	51.918	52.430	44.703	45.184	45.488	45.670	44.937
TiO ₂	0.186	0.592	0.491	0.103	0.140	0.643	0.784	0.904	0.802	0.864
Al ₂ O ₃	5.649	7.009	5.438	3.086	3.389	8.557	7.840	8.192	8.297	8.643
Cr ₂ O ₃	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
FeO	15.486	16.981	15.646	15.243	16.070	19.737	19.133	18.679	18.930	19.520
MnO	0.174	0.217	0.228	0.240	0.174	0.313	0.305	0.350	0.273	0.241
MgO	13.745	12.223	13.454	14.675	14.212	9.129	9.674	9.333	9.369	9.145
CaO	11.714	11.726	12.152	11.468	11.355	12.015	11.945	11.949	12.099	12.350
Na ₂ O	0.634	0.688	0.620	0.412	0.398	0.923	0.835	0.945	0.974	1.039
K ₂ O	0.193	0.275	0.162	0.000	0.000	0.711	0.621	0.667	0.592	0.606
Cl	0.467	0.443	0.000	0.000	0.000					
F	0.000	0.000	0.000	0.000	0.000					
Total	96.416	96.878	97.305	97.145	98.168	96.731	96.321	96.507	97.006	97.345
-O=Cl,F	0.105	0.100	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Total	96.311	96.778	97.305	97.145	98.168	96.731	96.321	96.507	97.006	97.345
Number of ions on the basis of 23O using the calculated average Fe ³⁺ content										
Si	7.14	6.97	7.18	7.55	7.57	6.80	6.87	6.92	6.91	6.80
Al ^{iv}	0.86	1.03	0.82	0.45	0.43	1.20	1.13	1.08	1.09	1.20
sum T	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
Al ^{vi}	0.12	0.19	0.12	0.08	0.15	0.33	0.27	0.39	0.39	0.34
Ti	0.02	0.07	0.05	0.01	0.02	0.07	0.09	0.10	0.09	0.10
Fe ³⁺	0.65	0.62	0.49	0.35	0.25	0.37	0.37	0.14	0.17	0.24
Cr	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Mg	3.01	2.69	2.93	3.18	3.06	2.07	2.19	2.12	2.11	2.06
Fe ²⁺	1.19	1.42	1.39	1.38	1.53	2.14	2.05	2.23	2.23	2.23
Mn	0.01	0.01	0.01	0.00	0.00	0.02	0.02	0.02	0.02	0.03
sum C	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00
Mg	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Fe ²⁺	0.06	0.05	0.03	0.13	0.17	0.00	0.01	0.00	0.00	0.00
Mn	0.01	0.01	0.01	0.03	0.02	0.02	0.02	0.02	0.02	0.00
Ca	1.84	1.86	1.90	1.79	1.76	1.96	1.95	1.95	1.96	2.00
Na	0.08	0.08	0.05	0.06	0.06	0.02	0.03	0.03	0.02	0.00
sum B	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00
Na	0.10	0.12	0.12	0.06	0.06	0.25	0.22	0.25	0.26	0.30
K	0.04	0.05	0.03	0.00	0.00	0.14	0.12	0.13	0.11	0.12
sum A	0.13	0.17	0.15	0.06	0.06	0.39	0.34	0.38	0.38	0.42
Total	15.13	15.17	15.15	15.06	15.06	15.39	15.34	15.38	15.38	15.42

Table C-3 (cont.): Amphibole electron-microprobe analyses, with stoichiometry based on 23O, and using the average calculated Fe³⁺ content according to Schumacher (IMA, 1997).

Section Analysis	RIM OF QUARTZ VEIN		NEAR QUARTZ VEIN		ASSOCIATED WITH PYROXENE					
	R8-219 (2-1) 1	R8-219 (2-1) 2	R8-219 (3-1) 1	R8-219 (3-1) 2	P6-91 (1-1) 1	P6-91 (1-2) 1	P6-91 (1-1) 2	P6-91 (1-2) 2	P6-91 (1-2) 3	P6-91 (1-3) 1
SiO ₂	48.659	48.382	47.980	49.594	40.095	39.434	42.865	40.909	43.662	40.504
TiO ₂	0.731	0.721	0.696	0.715	0.765	0.777	0.586	0.649	0.520	0.631
Al ₂ O ₃	6.119	6.332	6.712	6.101	12.580	12.941	11.278	12.068	9.408	12.292
Cr ₂ O ₃	0.000	0.000	0.000	0.099	0.000	0.000	0.000	0.151	0.000	0.000
FeO	17.024	17.307	17.320	16.928	20.604	20.735	17.915	19.800	18.292	19.341
MnO	0.212	0.273	0.162	0.180	0.450	0.255	0.337	0.272	0.223	0.290
MgO	12.595	12.374	12.092	12.655	9.083	8.875	9.749	8.325	10.207	8.444
CaO	11.622	11.771	11.785	11.547	12.447	12.483	12.086	11.675	12.211	11.982
Na ₂ O	0.786	0.783	0.860	0.680	1.323	1.229	1.088	1.077	1.015	1.247
K ₂ O	0.000	0.000	0.086	0.000	1.332	1.318	1.197	1.469	0.966	1.456
Cl	0.000	0.000	0.000	0.299						
F	0.000	0.000	0.000	0.000						
Total	97.748	97.943	97.693	98.499	98.679	98.047	97.101	96.395	96.504	96.187
-O=Cl,F	0.000	0.000	0.000	0.067	0.000	0.000	0.000	0.000	0.000	0.000
Total	97.748	97.943	97.693	98.432	98.679	98.047	97.101	96.395	96.504	96.187
Number of ions on the basis of 23O using the calculated average Fe³⁺ content										
Si	7.11	7.06	7.04	7.18	6.02	5.95	6.46	6.28	6.62	6.24
Al ^{iv}	0.89	0.94	0.96	0.82	1.98	2.05	1.54	1.72	1.38	1.76
<i>sum T</i>	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
Al ^{vi}	0.16	0.15	0.20	0.22	0.24	0.26	0.46	0.46	0.30	0.47
Ti	0.08	0.08	0.08	0.08	0.09	0.09	0.07	0.07	0.06	0.07
Fe ³⁺	0.55	0.57	0.50	0.43	0.93	0.98	0.45	0.57	0.50	0.51
Cr	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.02	0.00	0.00
Mg	2.74	2.69	2.64	2.73	2.03	2.00	2.19	1.90	2.31	1.94
Fe ²⁺	1.46	1.49	1.57	1.53	1.65	1.64	1.81	1.95	1.82	1.98
Mn	0.01	0.02	0.01	0.00	0.06	0.03	0.02	0.02	0.02	0.03
<i>sum C</i>	5.00	5.00	5.00	5.00	5.00	4.99	5.00	5.00	5.00	5.00
Mg	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Fe ²⁺	0.07	0.06	0.06	0.09	0.00	0.00	0.00	0.02	0.00	0.00
Mn	0.01	0.02	0.01	0.02	0.00	0.00	0.02	0.02	0.01	0.01
Ca	1.82	1.84	1.85	1.79	2.00	2.02	1.95	1.92	1.98	1.98
Na	0.10	0.08	0.08	0.09	0.00	0.00	0.03	0.04	0.01	0.01
<i>sum B</i>	2.00	2.00	2.00	2.00	2.00	2.02	2.00	2.00	2.00	2.00
Na	0.13	0.14	0.17	0.10	0.38	0.36	0.29	0.28	0.29	0.36
K	0.00	0.00	0.02	0.00	0.25	0.25	0.23	0.29	0.19	0.29
<i>sum A</i>	0.13	0.14	0.18	0.10	0.64	0.61	0.52	0.57	0.48	0.65
Total	15.13	15.14	15.18	15.10	15.64	15.62	15.52	15.57	15.48	15.65

Table C-3 (cont.): Amphibole electron-microprobe analyses, with stoichiometry based on 23O, and using the average calculated Fe³⁺ content according to Schumacher (IMA, 1997).

ASSOCIATED WITH PYROXENE										
Section	P6-91	P6-91	P6-91	P6-91	P6-91	P6-91	P6-91	P6-91	P10-103	R12-219
Analysis	(1-3) 2	(1-3) 3	(1-3) 4	(1-4) 1	(1-4) 2	(1-5) 1	(1-6) 1	(1-6) 2	(2-1) 3	(1-1) 1
SiO ₂	40.917	48.108	51.354	41.256	43.072	50.993	44.754	42.661	48.964	54.593
TiO ₂	0.711	0.258	0.077	0.672	0.651	0.139	0.572	0.417	0.245	0.089
Al ₂ O ₃	12.204	6.696	4.279	10.961	10.578	4.192	9.354	9.878	4.750	1.830
Cr ₂ O ₃	0.000	0.000	0.000	0.000	0.000	0.000	0.097	0.000	0.070	0.000
FeO	19.295	16.151	14.061	19.920	19.630	13.750	19.098	18.913	19.611	9.824
MnO	0.354	0.306	0.348	0.377	0.428	0.307	0.241	0.323	0.313	1.145
MgO	8.694	12.452	13.968	8.810	9.495	14.302	10.192	9.597	11.016	17.725
CaO	11.949	12.541	12.798	11.882	12.358	13.018	12.348	12.348	12.342	12.655
Na ₂ O	1.174	0.739	0.469	1.083	1.114	0.422	0.947	1.001	0.544	0.098
K ₂ O	1.385	0.407	0.276	1.311	1.174	0.226	1.006	1.050	0.125	0.000
Cl									0.966	0.000
F									0.000	0.000
Total	96.683	97.658	97.630	96.272	98.500	97.349	98.609	96.188	97.980	97.959
-O=Cl,F	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.218	0.000
Total	96.683	97.658	97.630	96.272	98.500	97.349	98.609	96.188	97.762	97.959
Number of ions on the basis of 23O using the calculated average Fe ³⁺ content										
Si	6.25	7.07	7.47	6.34	6.44	7.43	6.65	6.52	7.25	7.72
Al ^{iv}	1.75	0.93	0.53	1.66	1.56	0.57	1.35	1.48	0.75	0.28
sum T	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
Al ^{vi}	0.45	0.23	0.21	0.32	0.31	0.15	0.28	0.30	0.08	0.02
Ti	0.08	0.03	0.01	0.08	0.07	0.02	0.06	0.05	0.03	0.01
Fe ³⁺	0.57	0.38	0.12	0.66	0.58	0.19	0.51	0.56	0.47	0.24
Cr	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.01	0.00
Mg	1.98	2.73	3.03	2.02	2.12	3.11	2.26	2.19	2.43	3.74
Fe ²⁺	1.90	1.61	1.59	1.90	1.88	1.48	1.86	1.85	1.96	0.93
Mn	0.03	0.03	0.04	0.03	0.05	0.04	0.02	0.04	0.02	0.07
sum C	5.00	5.00	5.00	5.00	5.00	4.98	5.00	4.99	5.00	5.00
Mg	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Fe ²⁺	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Mn	0.02	0.01	0.00	0.02	0.01	0.00	0.02	0.00	0.02	0.07
Ca	1.96	1.98	2.00	1.96	1.98	2.03	1.96	2.02	1.96	1.92
Na	0.02	0.01	0.00	0.02	0.01	0.00	0.02	0.00	0.02	0.01
sum B	2.00	2.00	2.00	2.00	2.00	2.03	2.00	2.02	2.00	2.00
Na	0.32	0.20	0.13	0.30	0.31	0.12	0.25	0.30	0.13	0.01
K	0.27	0.08	0.05	0.26	0.22	0.04	0.19	0.20	0.02	0.00
sum A	0.59	0.27	0.18	0.56	0.54	0.16	0.44	0.50	0.16	0.01
Total	15.59	15.27	15.18	15.56	15.54	15.18	15.44	15.51	15.16	15.01

Table C-3 (cont.): Amphibole electron-microprobe analyses, with stoichiometry based on 23O, and using the average calculated Fe³⁺ content according to Schumacher (IMA, 1997).

ASSOCIATED WITH PYROXENE							ZONED WITH GARNET AND PYROXENE			
Section	R12-219	R12-219	R12-219	R12-219	R12-219	R12-219	R12-119	R12-119	R12-119	R12-119
Analysis	(1-1) 2	(3-1) 1	(3-2) 1	(3-3) 1	(3-4) 1	(4-1) 2	(1-1) 1	(1-1) 2	(1-1) 3	(1-1) 4
SiO ₂	54.945	54.554	54.796	54.811	55.543	53.891	45.904	45.411	47.165	45.163
TiO ₂	0.088	0.057	0.122	0.095	0.095	0.130	0.799	0.751	0.280	0.483
Al ₂ O ₃	1.963	1.267	1.733	1.580	1.413	1.795	9.274	9.638	7.994	10.069
Cr ₂ O ₃	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
FeO	9.549	9.660	10.067	9.672	9.734	10.204	18.435	18.724	21.052	18.804
MnO	1.203	1.289	1.316	1.251	1.116	1.507	0.367	0.307	0.294	0.396
MgO	17.299	17.589	17.278	17.634	18.092	17.353	9.529	9.234	8.744	9.194
CaO	12.348	12.836	12.642	12.451	11.792	12.033	11.605	11.813	11.867	11.801
Na ₂ O	0.086	0.061	0.079	0.081	0.077	0.069	0.817	0.844	0.664	0.968
K ₂ O	0.000	0.000	0.000	0.000	0.000	0.000	0.430	0.432	0.309	0.383
Cl	0.000	0.000	0.000	0.331	0.000	0.333				
F	0.000	0.000	0.000	0.000	0.000	0.000				
Total	97.481	97.313	98.033	97.575	97.862	96.982	97.160	97.154	98.369	97.261
-O=Cl,F	0.000	0.000	0.000	0.075	0.000	0.075	0.000	0.000	0.000	0.000
Total	97.481	97.313	98.033	97.500	97.862	96.907	97.160	97.154	98.369	97.261
Number of ions on the basis of 23O using the calculated average Fe ³⁺ content										
Si	7.82	7.78	7.77	7.79	7.86	7.72	6.86	6.81	7.02	6.76
Al ^{iv}	0.18	0.21	0.23	0.21	0.14	0.28	1.14	1.19	0.98	1.24
sum T	8.00	7.99	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
Al ^{vi}	0.14	0.00	0.06	0.05	0.09	0.02	0.50	0.51	0.42	0.54
Ti	0.01	0.01	0.01	0.01	0.01	0.01	0.09	0.08	0.03	0.05
Fe ³⁺	0.03	0.17	0.15	0.14	0.03	0.23	0.29	0.29	0.36	0.35
Cr	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Mg	3.67	3.74	3.65	3.73	3.81	3.70	2.12	2.06	1.94	2.05
Fe ²⁺	1.11	0.98	1.04	1.01	1.05	0.99	1.98	2.03	2.23	1.98
Mn	0.04	0.10	0.09	0.06	0.00	0.04	0.02	0.02	0.02	0.03
sum C	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00
Mg	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Fe ²⁺	0.00	0.00	0.00	0.00	0.07	0.00	0.04	0.03	0.03	0.02
Mn	0.10	0.05	0.07	0.09	0.13	0.14	0.02	0.02	0.02	0.03
Ca	1.88	1.96	1.92	1.90	1.79	1.85	1.86	1.90	1.89	1.89
Na	0.02	0.00	0.01	0.01	0.01	0.01	0.07	0.05	0.06	0.06
sum B	2.00	2.02	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00
Na	0.01	0.02	0.01	0.01	0.01	0.01	0.16	0.19	0.13	0.22
K	0.00	0.00	0.00	0.00	0.00	0.00	0.08	0.08	0.06	0.07
sum A	0.01	0.02	0.01	0.01	0.01	0.01	0.24	0.27	0.19	0.30
Total	15.01	15.03	15.01	15.01	15.01	15.01	15.24	15.27	15.19	15.30

Table C-3 (cont.): Amphibole electron-microprobe analyses, with stoichiometry based on 23O, and using the average calculated Fe³⁺ content according to Schumacher (IMA, 1997).

ZONED WITH GARNET AND PYROXENE				ASSOCIATED WITH CARBONATE						
Section	R12-119	R12-119	R12-119	R12-19	R12-19	R12-19	R12-19	R12-19	R12-19	R12-19
Analysis	(1-1) 5	(1-2) 1	(1-2) 2	(1-1) 2	(1-1) 4	(1-1) 5	(1-1) 6	(1-1) 7	(1-1) 8	(1-1) 10
SiO ₂	45.724	50.839	52.513	52.767	55.237	56.485	56.791	54.162	55.450	56.939
TiO ₂	0.335	0.115	0.000	0.168	0.071	0.000	0.000	0.108	0.069	0.042
Al ₂ O ₃	11.259	3.158	3.078	6.045	3.123	0.186	1.864	2.998	3.035	1.583
Cr ₂ O ₃	0.000	0.000	0.134	0.000	0.000	0.000	0.000	0.000	0.000	0.000
FeO	18.930	18.678	19.716	4.810	4.315	6.271	2.831	7.799	3.281	3.171
MnO	0.338	0.569	0.397	0.141	0.193	0.191	0.105	0.216	0.065	0.170
MgO	8.888	10.644	10.502	19.852	20.953	20.094	21.917	17.966	21.430	22.036
CaO	12.560	12.369	12.257	13.495	13.042	13.250	12.917	12.737	12.856	12.704
Na ₂ O	0.961	0.204	0.223	0.819	0.451	0.000	0.303	0.278	0.456	0.254
K ₂ O	0.487	0.102	0.000	0.258	0.100	0.000	0.000	0.000	0.188	0.000
Cl										
F										
Total	99.482	96.678	98.820	98.355	97.485	96.477	96.728	96.264	96.830	96.899
-O=Cl,F	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Total	99.482	96.678	98.820	98.355	97.485	96.477	96.728	96.264	96.830	96.899
Number of ions on the basis of 23O using the calculated average Fe ³⁺ content										
Si	6.71	7.63	7.71	7.32	7.66	7.98	7.86	7.72	7.70	7.87
Al ^{iv}	1.29	0.37	0.29	0.68	0.34	0.02	0.14	0.28	0.30	0.13
sum T	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
Al ^{vi}	0.66	0.19	0.24	0.30	0.17	0.01	0.16	0.23	0.19	0.13
Ti	0.04	0.01	0.00	0.02	0.01	0.00	0.00	0.01	0.01	0.00
Fe ³⁺	0.21	0.08	0.06	0.07	0.09	0.01	0.04	0.04	0.08	0.03
Cr	0.00	0.00	0.02	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Mg	1.95	2.38	2.30	4.10	4.33	4.23	4.52	3.82	4.43	4.54
Fe ²⁺	2.11	2.26	2.36	0.48	0.39	0.73	0.28	0.89	0.29	0.29
Mn	0.03	0.07	0.02	0.02	0.01	0.02	0.00	0.02	0.00	0.00
sum C	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00
Mg	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Fe ²⁺	0.00	0.00	0.00	0.00	0.02	0.00	0.01	0.00	0.02	0.04
Mn	0.01	0.00	0.03	0.00	0.01	0.00	0.01	0.01	0.00	0.02
Ca	1.98	1.99	1.93	2.00	1.94	2.00	1.91	1.95	1.91	1.88
Na	0.01	0.01	0.04	0.00	0.04	0.00	0.07	0.04	0.07	0.06
sum B	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00
Na	0.26	0.05	0.02	0.22	0.09	0.00	0.01	0.03	0.06	0.01
K	0.09	0.02	0.00	0.05	0.02	0.00	0.00	0.00	0.03	0.00
sum A	0.35	0.07	0.02	0.27	0.10	0.00	0.01	0.03	0.09	0.01
Total	15.35	15.07	15.02	15.27	15.10	15.00	15.01	15.03	15.09	15.01

Table C-3 (cont.): Amphibole electron-microprobe analyses, with stoichiometry based on 23O, and using the average calculated Fe³⁺ content according to Schumacher (IMA, 1997).

ASSOCIATED WITH CARBONATE										
Section	R12-19	R12-19	R12-19	R12-19	R12-19	R12-19	R12-19	R12-19	R12-19	R12-19
Analysis	(1-1) 11	(1-2) 4	(1-2) 5	(1-3) 2	(1-3) 5	(1-3) 9	(2-1) 1	(2-1) 2	(2-1) 3	(2-1) 4
SiO ₂	53.931	54.080	53.333	46.226	48.098	50.116	52.824	56.148	55.221	48.269
TiO ₂	0.041	0.000	0.086	0.247	0.330	0.133	0.130	0.051	0.000	0.307
Al ₂ O ₃	3.943	1.961	2.736	9.253	7.585	5.433	3.361	1.583	1.179	7.644
Cr ₂ O ₃	0.000	0.000	0.073	0.098	0.000	0.000	0.064	0.000	0.065	0.000
FeO	6.317	11.515	11.856	15.482	14.173	13.186	11.151	11.346	12.893	14.590
MnO	0.121	0.311	0.366	0.243	0.334	0.244	0.259	0.294	0.350	0.326
MgO	18.761	15.799	15.427	11.421	12.574	13.827	15.897	16.881	16.075	12.935
CaO	12.710	12.650	12.819	12.557	12.559	12.633	12.636	12.983	12.610	12.343
Na ₂ O	0.370	0.151	0.179	0.825	0.722	0.422	0.293	0.084	0.105	0.624
K ₂ O	0.165	0.000	0.000	0.338	0.188	0.176	0.000	0.000	0.000	0.205
Cl										
F										
Total	96.359	96.467	96.875	96.690	96.563	96.170	96.615	99.370	98.498	97.243
-O=Cl,F	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Total	96.359	96.467	96.875	96.690	96.563	96.170	96.615	99.370	98.498	97.243
Number of ions on the basis of 23O using the calculated average Fe ³⁺ content										
Si	7.63	7.83	7.72	6.86	7.10	7.37	7.63	7.87	7.87	7.05
Al ^{iv}	0.37	0.17	0.28	1.14	0.90	0.63	0.37	0.13	0.13	0.95
sum T	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
Al ^{vi}	0.29	0.16	0.18	0.48	0.41	0.31	0.20	0.13	0.06	0.36
Ti	0.00	0.00	0.01	0.03	0.04	0.01	0.01	0.01	0.00	0.03
Fe ³⁺	0.05	0.02	0.04	0.29	0.19	0.15	0.10	0.01	0.06	0.38
Cr	0.00	0.00	0.01	0.01	0.00	0.00	0.01	0.00	0.01	0.00
Mg	3.96	3.41	3.33	2.53	2.77	3.03	3.42	3.53	3.41	2.81
Fe ²⁺	0.69	1.37	1.40	1.63	1.56	1.47	1.24	1.32	1.45	1.39
Mn	0.01	0.03	0.04	0.03	0.03	0.03	0.02	0.01	0.00	0.02
sum C	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00
Mg	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Fe ²⁺	0.01	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.02	0.01
Mn	0.01	0.01	0.01	0.00	0.01	0.00	0.02	0.03	0.04	0.02
Ca	1.93	1.96	1.99	2.00	1.99	1.99	1.95	1.95	1.92	1.93
Na	0.06	0.03	0.01	0.00	0.01	0.01	0.02	0.02	0.01	0.04
sum B	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00
Na	0.04	0.01	0.04	0.24	0.20	0.11	0.06	0.00	0.01	0.14
K	0.03	0.00	0.00	0.06	0.04	0.03	0.00	0.00	0.00	0.04
sum A	0.07	0.01	0.04	0.30	0.23	0.15	0.06	0.00	0.01	0.18
Total	15.07	15.01	15.04	15.30	15.23	15.15	15.06	15.00	15.01	15.18

Table C-3 (cont.): Amphibole electron-microprobe analyses, with stoichiometry based on 23O, and using the average calculated Fe³⁺ content according to Schumacher (IMA, 1997).

ASSOCIATED WITH CARBONATE										
Section	R12-19	R12-19	R12-51	R12-51	R12-51	R12-51	R12-51	R12-51	R12-51	R12-51
Analysis	(2-1) 5	(2-1) 7	(2-1) 1	(2-1) 3	(2-2) 2	(2-2) 3	(2-2) 4	(1-1) 1	(1-2) 1	(1-3) 1
SiO ₂	52.862	54.815	43.376	55.696	48.462	46.983	55.757	42.054	42.450	41.659
TiO ₂	0.146	0.000	0.905	0.149	0.687	0.779	0.047	0.842	0.783	1.030
Al ₂ O ₃	2.737	1.963	11.933	1.191	7.560	8.844	0.933	13.232	13.938	13.657
Cr ₂ O ₃	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
FeO	13.174	12.708	13.105	10.772	11.531	10.766	10.651	13.834	11.639	12.661
MnO	0.297	0.426	0.418	0.453	0.497	0.428	0.465	0.374	0.369	0.405
MgO	15.002	15.887	12.853	17.528	15.509	15.511	17.541	12.163	13.455	12.791
CaO	12.442	12.332	13.039	13.185	12.596	12.995	13.061	12.348	12.289	12.696
Na ₂ O	0.222	0.153	1.207	0.062	0.855	1.025	0.076	1.235	1.356	1.357
K ₂ O	0.000	0.000	0.499	0.000	0.275	0.413	0.000	0.829	0.681	0.765
Cl			0.337	0.000	0.327	0.000	0.000	0.000	0.000	0.370
F			0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Total	96.882	98.284	97.335	99.036	97.972	97.744	98.531	96.911	96.960	97.021
-O=Cl,F	0.000	0.000	0.076	0.000	0.074	0.000	0.000	0.000	0.000	0.083
Total	96.882	98.284	97.259	99.036	97.898	97.744	98.531	96.911	96.960	96.938
Number of ions on the basis of 23O using the calculated average Fe ³⁺ content										
Si	7.67	7.82	6.35	7.81	6.93	6.75	7.86	6.20	6.18	6.12
Al ^{iv}	0.33	0.18	1.65	0.19	1.07	1.25	0.14	1.80	1.82	1.88
sum T	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
Al ^{vi}	0.14	0.15	0.41	0.01	0.21	0.25	0.02	0.50	0.57	0.48
Ti	0.02	0.00	0.10	0.02	0.07	0.08	0.00	0.09	0.09	0.11
Fe ³⁺	0.16	0.05	0.55	0.14	0.49	0.47	0.11	0.65	0.66	0.65
Cr	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Mg	3.24	3.38	2.81	3.67	3.31	3.32	3.69	2.67	2.92	2.80
Fe ²⁺	1.43	1.43	1.06	1.12	0.88	0.82	1.14	1.05	0.74	0.91
Mn	0.01	0.00	0.05	0.04	0.03	0.05	0.04	0.02	0.02	0.05
sum C	5.00	5.00	4.98	5.00	5.00	5.00	5.00	5.00	5.00	5.00
Mg	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Fe ²⁺	0.01	0.04	0.00	0.00	0.00	0.00	0.00	0.00	0.02	0.00
Mn	0.02	0.05	0.00	0.01	0.03	0.00	0.02	0.02	0.02	0.00
Ca	1.93	1.88	2.05	1.98	1.93	2.00	1.97	1.95	1.92	2.00
Na	0.03	0.03	0.00	0.01	0.04	0.00	0.01	0.03	0.04	0.00
sum B	2.00	2.00	2.05	2.00	2.00	2.00	2.00	2.00	2.00	2.00
Na	0.03	0.02	0.34	0.01	0.20	0.29	0.01	0.33	0.34	0.39
K	0.00	0.00	0.09	0.00	0.05	0.08	0.00	0.16	0.13	0.14
sum A	0.03	0.02	0.44	0.01	0.25	0.36	0.01	0.48	0.46	0.53
Total	15.03	15.02	15.46	15.01	15.25	15.36	15.01	15.48	15.46	15.53

Table C-3 (cont.): Amphibole electron-microprobe analyses, with stoichiometry based on 23O, and using the average calculated Fe³⁺ content according to Schumacher (IMA, 1997).

ASSOCIATED WITH CATACLASTITE				ASSOCIATED WITH SULPHIDE						
Section	P13-67a	P13-67a	P13-67a	R12-27	R12-27	R12-27	R12-27	R8-253	R7-320	R7-320
Analysis	(1-1) 1	(2-1) 1	(3-1) 1	(1-1) 1	(1-1) 3	(1-1) 6	(1-1) 7	(1-1) 1	(2-1) 1	(2-1) 2
SiO ₂	44.157	41.460	43.014	42.938	43.830	43.659	43.215	48.581	47.613	46.353
TiO ₂	0.752	0.793	0.940	0.671	0.456	0.853	0.652	0.138	0.558	0.470
Al ₂ O ₃	8.621	10.142	8.995	10.713	9.848	10.421	10.552	4.895	6.705	8.014
Cr ₂ O ₃	0.000	0.000	0.000	0.097	0.090	0.070	0.075	0.000	0.000	0.000
FeO	19.915	21.363	20.275	19.198	18.285	18.831	19.010	19.685	17.428	18.826
MnO	0.307	0.306	0.224	1.229	1.120	1.412	1.384	0.288	0.308	0.331
MgO	10.298	8.840	9.464	9.446	9.596	9.976	9.221	11.581	11.965	10.696
CaO	11.954	11.625	11.767	11.595	11.769	11.388	11.650	11.725	11.132	12.057
Na ₂ O	0.846	0.888	0.882	1.245	0.990	1.133	1.091	0.602	0.682	0.703
K ₂ O	0.854	1.160	1.059	0.324	0.316	0.274	0.320	0.170	0.000	0.000
Cl	0.965	0.825	1.091	0.355	0.000	0.000	0.000	0.000	0.000	0.000
F	0.000	0.000	0.062	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Total	97.704	96.577	96.620	97.456	96.300	98.017	97.170	97.665	96.391	97.450
-O=Cl,F	0.218	0.186	0.272	0.080	0.000	0.000	0.000	0.000	0.000	0.000
Total	97.486	96.391	96.348	97.376	96.300	98.017	97.170	97.665	96.391	97.450
Number of ions on the basis of 23O using the calculated average Fe ³⁺ content										
Si	6.61	6.34	6.55	6.42	6.62	6.46	6.49	7.18	7.06	6.86
Al ^{iv}	1.39	1.66	1.45	1.58	1.38	1.54	1.51	0.82	0.94	1.14
sum T	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
Al ^{vi}	0.13	0.17	0.17	0.31	0.37	0.28	0.35	0.04	0.23	0.26
Ti	0.08	0.09	0.11	0.08	0.05	0.09	0.07	0.02	0.06	0.05
Fe ³⁺	0.78	0.92	0.68	0.83	0.65	0.89	0.76	0.68	0.59	0.67
Cr	0.00	0.00	0.00	0.01	0.01	0.01	0.01	0.00	0.00	0.00
Mg	2.30	2.01	2.15	2.11	2.16	2.20	2.06	2.55	2.64	2.36
Fe ²⁺	1.69	1.78	1.88	1.58	1.66	1.44	1.62	1.70	1.48	1.64
Mn	0.02	0.02	0.01	0.09	0.10	0.09	0.12	0.01	0.00	0.02
sum C	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00
Mg	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Fe ²⁺	0.02	0.02	0.02	0.00	0.00	0.00	0.00	0.05	0.10	0.02
Mn	0.02	0.02	0.01	0.07	0.04	0.09	0.06	0.03	0.04	0.02
Ca	1.92	1.90	1.92	1.86	1.90	1.81	1.87	1.86	1.77	1.91
Na	0.04	0.05	0.04	0.08	0.05	0.10	0.07	0.07	0.10	0.05
sum B	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00
Na	0.20	0.21	0.22	0.29	0.24	0.22	0.25	0.10	0.10	0.15
K	0.16	0.23	0.21	0.06	0.06	0.05	0.06	0.03	0.00	0.00
sum A	0.36	0.44	0.42	0.35	0.30	0.27	0.31	0.14	0.10	0.15
Total	15.36	15.44	15.42	15.35	15.30	15.27	15.31	15.14	15.10	15.15

Table C-3 (cont.): Amphibole electron-microprobe analyses, with stoichiometry based on 23O, and using the average calculated Fe³⁺ content according to Schumacher (IMA, 1997).

ASSOCIATED WITH SULPHIDE					ASSOCIATED WITH GOLD					
Section	R7-320	R7-320	R7-320	R7-320	P22-70	P22-70	P22-70a	P22-70a	P22-70a	P22-70a
Analysis	(2-1) 3	(2-2) 1	(2-2) 2	(2-2) 3	(1-1) 2	(1-1) 3	(3-1) 3	(4-1a) 1	(4-1a) 2	(4-1a) 3
SiO ₂	48.010	47.693	47.005	47.403	41.153	40.792	41.022	41.160	41.466	40.880
TiO ₂	0.526	0.515	0.486	0.652	0.576	0.560	0.897	1.038	1.122	1.246
Al ₂ O ₃	7.634	6.210	6.946	6.893	10.849	10.683	11.164	11.001	10.878	11.259
Cr ₂ O ₃	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
FeO	18.510	19.003	18.685	18.968	24.282	24.973	23.290	23.201	23.301	22.337
MnO	0.233	0.357	0.266	0.406	0.250	0.297	0.284	0.256	0.311	0.288
MgO	11.077	11.117	11.165	10.906	6.841	6.740	6.519	6.824	6.673	6.956
CaO	12.309	12.324	12.179	11.589	10.878	10.937	11.281	11.808	11.766	11.040
Na ₂ O	0.763	0.608	0.738	0.737	0.935	0.940	0.929	0.931	1.033	1.094
K ₂ O	0.000	0.000	0.000	0.000	0.699	0.804	1.133	1.377	1.678	1.522
Cl	0.000	0.000	0.000	0.000	0.858	0.775	0.817	0.715	0.983	0.777
F	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.108
Total	99.062	97.827	97.470	97.554	96.463	96.726	96.519	97.596	98.228	96.622
-O=Cl,F	0.000	0.000	0.000	0.000	0.194	0.175	0.184	0.161	0.222	0.221
Total	99.062	97.827	97.470	97.554	96.269	96.551	96.335	97.435	98.006	96.401
Number of ions on the basis of 23O using the calculated average Fe ³⁺ content										
Si	6.99	7.04	6.96	7.01	6.33	6.28	6.36	6.33	6.37	6.34
Al ^{iv}	1.01	0.96	1.04	0.99	1.67	1.72	1.64	1.67	1.63	1.66
sum T	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
Al ^{vi}	0.30	0.13	0.17	0.21	0.30	0.22	0.39	0.32	0.34	0.39
Ti	0.06	0.06	0.05	0.07	0.07	0.06	0.10	0.12	0.13	0.15
Fe ³⁺	0.47	0.59	0.63	0.61	1.03	1.13	0.67	0.62	0.46	0.53
Cr	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Mg	2.40	2.45	2.46	2.40	1.57	1.55	1.51	1.56	1.53	1.61
Fe ²⁺	1.76	1.75	1.67	1.69	2.01	2.01	2.30	2.36	2.52	2.31
Mn	0.01	0.02	0.02	0.03	0.02	0.02	0.02	0.02	0.02	0.02
sum C	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00
Mg	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Fe ²⁺	0.02	0.00	0.02	0.05	0.08	0.07	0.04	0.01	0.01	0.06
Mn	0.01	0.02	0.02	0.03	0.02	0.02	0.02	0.02	0.02	0.02
Ca	1.92	1.95	1.93	1.84	1.79	1.81	1.87	1.95	1.94	1.83
Na	0.04	0.03	0.04	0.09	0.11	0.10	0.07	0.03	0.03	0.09
sum B	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00
Na	0.17	0.15	0.18	0.12	0.17	0.18	0.21	0.25	0.27	0.24
K	0.00	0.00	0.00	0.00	0.14	0.16	0.22	0.27	0.33	0.30
sum A	0.17	0.15	0.18	0.12	0.31	0.33	0.44	0.52	0.60	0.54
Total	15.17	15.15	15.18	15.12	15.31	15.33	15.44	15.52	15.60	15.54

Table C-3 (cont.): Amphibole electron-microprobe analyses, with stoichiometry based on 23O, and using the average calculated Fe³⁺ content according to Schumacher (IMA, 1997).

ASSOCIATED WITH GOLD					ASSOCIATED WITH EPIDOTE	
Section	P22-70a	P22-70a	P22-70a	P22-70a	R12-27	R12-27
Analysis	(4-1b) 1	(4-1b) 2	(4-1d) 1	(4-1d) 2	(2-1) 1	(2-1) 2
SiO ₂	40.427	40.686	40.201	40.682	51.368	50.416
TiO ₂	0.793	0.818	0.423	0.517	0.167	0.076
Al ₂ O ₃	11.100	11.301	11.656	11.743	4.348	4.208
Cr ₂ O ₃	0.000	0.069	0.000	0.000	0.000	0.114
FeO	22.959	23.437	21.889	22.411	15.992	15.087
MnO	0.235	0.291	0.298	0.262	1.499	1.139
MgO	6.594	6.648	6.898	6.951	13.428	13.740
CaO	11.685	11.285	11.860	11.343	12.153	12.640
Na ₂ O	0.946	0.905	0.892	1.014	0.400	0.321
K ₂ O	1.665	1.497	1.647	1.681	0.000	0.086
Cl	0.943	0.524	0.642	0.997	0.000	0.000
F	0.158	0.091	0.000	0.000	0.000	0.000
Total	96.404	96.937	95.764	96.604	99.355	97.827
-O=Cl,F	0.279	0.157	0.145	0.225	0.000	0.000
Total	96.125	96.780	95.619	96.379	99.355	97.827
Number of ions on the basis of 23O using the calculated average Fe³⁺ content						
Si	6.32	6.29	6.29	6.30	7.36	7.32
Al ^{iv}	1.68	1.71	1.71	1.70	0.64	0.68
<i>sum T</i>	<i>8.00</i>	<i>8.00</i>	<i>8.00</i>	<i>8.00</i>	<i>8.00</i>	<i>8.00</i>
Al ^{vi}	0.36	0.35	0.44	0.45	0.09	0.05
Ti	0.09	0.10	0.05	0.06	0.02	0.01
Fe ³⁺	0.57	0.73	0.58	0.61	0.51	0.53
Cr	0.00	0.01	0.00	0.00	0.00	0.01
Mg	1.54	1.53	1.61	1.61	2.87	2.98
Fe ²⁺	2.43	2.27	2.29	2.25	1.41	1.30
Mn	0.02	0.02	0.03	0.02	0.10	0.13
<i>sum C</i>	<i>5.00</i>	<i>5.00</i>	<i>5.00</i>	<i>5.00</i>	<i>5.00</i>	<i>5.00</i>
Mg	0.00	0.00	0.00	0.00	0.00	0.00
Fe ²⁺	0.00	0.04	0.00	0.04	0.00	0.00
Mn	0.02	0.02	0.01	0.02	0.08	0.02
Ca	1.96	1.87	1.99	1.88	1.87	1.97
Na	0.02	0.07	0.01	0.06	0.06	0.02
<i>sum B</i>	<i>2.00</i>	<i>2.00</i>	<i>2.00</i>	<i>2.00</i>	<i>2.00</i>	<i>2.00</i>
Na	0.26	0.20	0.26	0.24	0.06	0.07
K	0.33	0.30	0.33	0.33	0.00	0.02
<i>sum A</i>	<i>0.59</i>	<i>0.50</i>	<i>0.59</i>	<i>0.57</i>	<i>0.06</i>	<i>0.09</i>
Total	15.59	15.50	15.59	15.57	15.06	15.09

Table C-3 (cont.): Amphibole electron-microprobe analyses, with stoichiometry based on 23O, and using the average calculated Fe³⁺ content according to Schumacher (IMA, 1997).

INTENSE AMPHIBOLE						
Section	P10-103	P10-103	P10-103	P8-46	P8-46	P8-46
Analysis	(1-1) 1	(1-1) 2	(1-1) 3	(1-1) 1	(1-1) 2	(1-2) 3
SiO ₂	40.844	40.559	40.370	41.102	40.568	41.890
TiO ₂	0.810	0.771	0.852	0.922	1.028	0.583
Al ₂ O ₃	10.415	10.250	10.382	10.364	10.345	9.304
Cr ₂ O ₃	0.000	0.000	0.066	0.000	0.000	0.000
FeO	23.967	23.942	24.120	23.186	23.017	23.365
MnO	0.235	0.266	0.250	0.505	0.561	0.558
MgO	6.635	6.648	6.971	7.271	7.249	7.752
CaO	11.973	11.746	11.459	11.662	11.331	11.835
Na ₂ O	1.177	1.193	1.058	0.914	0.842	0.944
K ₂ O	1.104	1.061	1.084	1.170	1.161	1.014
Cl	0.716	0.925	0.697	0.936	0.454	0.586
F	0.000	0.000	0.000	0.000	0.000	0.000
Total	97.160	96.436	96.612	97.096	96.102	97.245
-O=Cl,F	0.162	0.209	0.157	0.211	0.102	0.132
Total	96.998	96.227	96.455	96.885	96.000	97.113
Number of ions on the basis of 23O using the calculated average Fe ³⁺ content						
Si	6.33	6.33	6.26	6.33	6.30	6.43
Al ^{iv}	1.67	1.67	1.74	1.67	1.70	1.57
sum T	8.00	8.00	8.00	8.00	8.00	8.00
Al ^{vi}	0.24	0.22	0.16	0.21	0.19	0.11
Ti	0.09	0.09	0.10	0.11	0.12	0.07
Fe ³⁺	0.68	0.73	0.94	0.83	0.91	0.90
Cr	0.00	0.00	0.01	0.00	0.00	0.00
Mg	1.53	1.55	1.61	1.67	1.68	1.77
Fe ²⁺	2.43	2.39	2.17	2.15	2.07	2.10
Mn	0.03	0.02	0.02	0.03	0.04	0.05
sum C	5.00	5.00	5.00	5.00	5.00	5.00
Mg	0.00	0.00	0.00	0.00	0.00	0.00
Fe ²⁺	0.00	0.00	0.03	0.00	0.02	0.00
Mn	0.00	0.02	0.02	0.03	0.04	0.03
Ca	1.99	1.96	1.90	1.92	1.89	1.95
Na	0.01	0.02	0.05	0.04	0.06	0.03
sum B	2.00	2.00	2.00	2.00	2.00	2.00
Na	0.35	0.34	0.27	0.23	0.19	0.25
K	0.22	0.21	0.21	0.23	0.23	0.20
sum A	0.57	0.55	0.48	0.46	0.42	0.45
Total	15.57	15.55	15.48	15.46	15.42	15.45

Table C-3 (cont.): Amphibole electron-microprobe analyses, with stoichiometry based on 23O, and using the average calculated Fe³⁺ content according to Schumacher (IMA, 1997).

Section	P6-91	P6-91	P6-91	P6-91	P6-91	P6-91	P6-91	P6-91	P6-91	R12-119
Analysis	(1-6) 2	(1-6) 3	(2-1) 1	(2-1) 2	(1-5) 1	(1-5) 2	(1-5) 3	(1-6) 4	(1-6) 5	(1-2) 1
SiO ₂	51.819	50.950	52.016	52.504	52.594	52.296	52.799	52.092	52.527	50.300
TiO ₂	0.062	0.126	0.000	0.000	0.068	0.060	0.052	0.057	0.055	0.149
Al ₂ O ₃	0.898	1.754	0.360	0.300	1.087	0.790	1.058	0.684	0.931	1.132
Cr ₂ O ₃	0.000	0.000	0.000	0.000	0.000	0.069	0.089	0.079	0.000	0.000
Fe ₂ O ₃	2.017	2.603	3.079	2.847	1.327	1.742	1.474	1.561	0.262	0.000
FeO	7.202	6.349	3.552	1.713	7.903	7.165	8.010	7.067	8.961	21.368
MnO	0.409	0.337	0.410	0.337	0.272	0.332	0.287	0.351	0.346	0.607
MgO	12.650	12.683	14.732	16.209	12.804	12.564	12.265	12.700	12.188	4.458
CaO	24.033	23.911	24.599	24.558	24.187	24.646	24.884	24.608	24.153	22.466
Na ₂ O	0.230	0.239	0.099	0.081	0.208	0.241	0.246	0.163	0.190	0.292
K ₂ O	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Total	99.320	98.952	98.847	98.549	100.450	99.905	101.164	99.362	99.613	100.772
Number of ions on the basis of 6O										
Si	1.967	1.941	1.966	1.970	1.969	1.972	1.969	1.973	1.982	1.983
Al ^{IV}	0.033	0.059	0.016	0.013	0.031	0.028	0.031	0.027	0.018	0.017
Al	0.008	0.020	0.000	0.000	0.017	0.007	0.016	0.004	0.024	0.035
Fe ³⁺	0.057	0.074	0.087	0.080	0.037	0.049	0.041	0.044	0.007	0.000
Cr	0.000	0.000	0.000	0.000	0.000	0.002	0.003	0.002	0.000	0.000
Ti	0.002	0.004	0.000	0.000	0.002	0.002	0.001	0.002	0.002	0.004
Fe ²⁺	0.228	0.201	0.111	0.053	0.247	0.225	0.249	0.223	0.283	0.705
Mn	0.013	0.011	0.013	0.011	0.009	0.011	0.009	0.011	0.011	0.020
Mg	0.716	0.720	0.830	0.906	0.715	0.706	0.682	0.717	0.686	0.262
Ca	0.978	0.976	0.996	0.987	0.970	0.996	0.994	0.999	0.977	0.949
Na	0.017	0.018	0.007	0.006	0.015	0.018	0.018	0.012	0.014	0.022
K	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
TOTAL	4.018	4.023	4.028	4.026	4.012	4.015	4.013	4.014	4.002	3.998
En	35.6	36.0	40.6	44.4	35.9	35.2	34.2	35.7	34.7	13.4
Fs	14.8	14.3	10.3	7.0	14.7	14.2	15.0	13.9	15.2	37.0
Wo	48.7	48.8	48.7	48.3	48.7	49.7	49.9	49.8	49.4	48.5
Ac	0.8	0.9	0.4	0.3	0.8	0.9	0.9	0.6	0.7	1.1
	100	100	100	100	100	100	100	100	100	100

Table C-4: Pyroxene electron-microprobe analyses, with stoichiometry based on 6O, and using the calculated Fe³⁺ content according to Droop (1987).

Section	R12-119	R12-119	R12-119	R12-119	R12-119	R12-119	R12-119	R8-193	R8-193	R8-193
Analysis	(1-2) 2	(1-2) 3	(1-2) 4	(1-3) 1	(1-3) 2	(1-3) 3	(1-3) 4	(1-6) 1	(1-6) 2	(1-6) 3
SiO ₂	49.600	50.858	52.174	49.594	50.543	50.886	50.311	52.491	52.263	51.924
TiO ₂	0.092	0.000	0.061	0.106	0.079	0.086	0.043	0.039	0.037	0.000
Al ₂ O ₃	1.112	0.967	1.100	1.072	1.076	0.899	0.956	0.473	0.483	0.543
Cr ₂ O ₃	0.071	0.090	0.000	0.000	0.071	0.000	0.088	0.000	0.000	0.000
Fe ₂ O ₃	0.680	0.000	0.000	2.121	0.101	0.000	0.000	3.760	2.892	3.211
FeO	18.823	18.821	18.458	20.357	20.914	22.355	22.064	7.208	6.962	6.387
MnO	0.734	0.456	0.620	0.657	0.628	0.527	0.567	0.259	0.177	0.302
MgO	5.465	6.996	6.375	3.582	4.193	3.617	3.669	12.037	12.492	12.859
CaO	22.495	21.451	22.830	24.178	23.786	22.978	22.794	24.847	24.423	23.895
Na ₂ O	0.273	0.219	0.235	0.219	0.218	0.219	0.244	0.441	0.395	0.402
K ₂ O	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Total	99.345	99.858	101.853	101.885	101.609	101.567	100.736	101.555	100.124	99.523
Number of ions on the basis of 6O										
Si	1.974	1.991	2.000	1.959	1.979	1.997	1.992	1.969	1.975	1.972
Al ^{IV}	0.026	0.009	0.000	0.041	0.021	0.003	0.008	0.021	0.022	0.024
Al	0.026	0.035	0.050	0.009	0.029	0.039	0.037	0.000	0.000	0.000
Fe ³⁺	0.020	0.000	0.000	0.063	0.003	0.000	0.000	0.105	0.082	0.091
Cr	0.002	0.003	0.000	0.000	0.002	0.000	0.003	0.000	0.000	0.000
Ti	0.003	0.000	0.002	0.003	0.002	0.003	0.001	0.001	0.001	0.000
Fe ²⁺	0.625	0.617	0.594	0.669	0.685	0.736	0.732	0.224	0.219	0.201
Mn	0.025	0.015	0.020	0.022	0.021	0.018	0.019	0.008	0.006	0.010
Mg	0.324	0.408	0.364	0.211	0.245	0.212	0.217	0.673	0.704	0.728
Ca	0.959	0.900	0.938	1.023	0.998	0.966	0.967	0.999	0.989	0.972
Na	0.021	0.017	0.017	0.017	0.017	0.017	0.019	0.032	0.029	0.030
K	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
TOTAL	4.006	3.995	3.985	4.017	4.001	3.990	3.994	4.032	4.025	4.028
En	16.4	20.9	18.8	10.5	12.4	10.9	11.1	33.0	34.7	35.8
Fs	34.0	32.3	31.8	37.6	36.0	38.7	38.4	16.5	15.1	14.9
Wo	48.6	46.0	48.5	51.0	50.7	49.6	49.5	48.9	48.8	47.9
Ac	1.1	0.8	0.9	0.8	0.8	0.9	1.0	1.6	1.4	1.5
	100	100	100	100	100	100	100	100	100	100

Table C-4 (cont.): Pyroxene electron-microprobe analyses, with stoichiometry based on 6O, and using the calculated Fe³⁺ content according to Droop (1987).

Section	P10-103	P10-103	R12-219	R12-219	R12-219	R8-253	P8-68	P8-68	P8-68	P8-68
Analysis	(3-1) 1	(3-1) 2	(2-1) 1	(2-2) 1	(2-3) 1	(1-2) 1	(2-1) 1	(2-1) 2	(3-1) 1	(3-1) 2
SiO ₂	50.833	51.048	53.852	53.467	52.655	51.825	53.862	54.966	54.338	53.947
TiO ₂	0.043	0.000	0.045	0.042	0.000	0.000	0.000	0.000	0.138	0.227
Al ₂ O ₃	0.889	0.655	0.255	0.494	0.319	0.521	0.314	0.297	1.338	1.800
Cr ₂ O ₃	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Fe ₂ O ₃	3.315	3.027	0.000	0.000	2.369	2.067	2.333	0.000	0.360	1.056
FeO	11.318	11.767	6.712	7.436	5.187	8.931	0.517	2.724	3.534	4.124
MnO	0.286	0.298	1.479	1.735	1.675	0.321	0.315	0.307	0.258	0.257
MgO	9.489	9.859	14.307	13.937	13.645	12.117	17.638	17.892	16.846	15.921
CaO	22.738	23.583	23.664	23.107	24.790	22.439	25.082	23.656	24.295	24.755
Na ₂ O	0.684	0.256	0.054	0.000	0.000	0.511	0.000	0.000	0.033	0.051
K ₂ O	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Total	99.595	100.493	100.368	100.218	100.639	98.732	100.061	99.842	101.140	102.138
Number of ions on the basis of 6O										
Si	1.972	1.965	1.997	1.991	1.971	1.987	1.972	1.999	1.966	1.948
Al ^{IV}	0.028	0.030	0.003	0.009	0.014	0.013	0.014	0.001	0.034	0.052
Al	0.013	0.000	0.008	0.013	0.000	0.011	0.000	0.011	0.023	0.024
Fe ³⁺	0.096	0.087	0.000	0.000	0.066	0.059	0.064	0.000	0.010	0.029
Cr	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Ti	0.001	0.000	0.001	0.001	0.000	0.000	0.000	0.000	0.004	0.006
Fe ²⁺	0.364	0.376	0.208	0.232	0.161	0.285	0.016	0.083	0.107	0.124
Mn	0.009	0.010	0.046	0.055	0.053	0.010	0.010	0.009	0.008	0.008
Mg	0.549	0.566	0.791	0.774	0.761	0.693	0.962	0.970	0.908	0.857
Ca	0.945	0.973	0.940	0.922	0.994	0.922	0.984	0.922	0.942	0.958
Na	0.051	0.019	0.004	0.000	0.000	0.038	0.000	0.000	0.002	0.004
K	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
TOTAL	4.029	4.026	3.998	3.997	4.021	4.018	4.021	3.995	4.003	4.009
En	27.2	27.9	39.7	39.0	37.4	34.5	47.3	48.9	45.9	43.3
Fs	23.3	23.3	12.8	14.5	13.8	17.7	4.4	4.7	6.3	8.1
Wo	46.9	47.9	47.3	46.5	48.8	45.9	48.3	46.5	47.6	48.4
Ac	2.6	0.9	0.2	0.0	0.0	1.9	0.0	0.0	0.1	0.2
	100	100	100	100	100	100	100	100	100	100

Table C-4 (cont.): Pyroxene electron-microprobe analyses, with stoichiometry based on 6O, and using the calculated Fe³⁺ content according to Droop (1987).

Section	P8-68	P8-68	H5-180	H5-180	H5-180	H5-180	H5-180	H5-180	H5-180	H5-180
Analysis	(3-1) 3	(3-1) 4	(1-2) 2	(1-3) 1	(1-3) 4	(1-4) 1	(1-4) 2	(1-4) 4	(1-4) 5	(2-1) 1
SiO ₂	53.244	53.091	49.022	48.664	48.603	48.355	49.343	48.577	48.488	48.956
TiO ₂	0.177	0.081	0.065	0.000	0.044	0.050	0.080	0.089	0.047	0.063
Al ₂ O ₃	1.427	0.783	0.724	0.949	0.738	1.836	0.889	0.914	0.926	0.877
Cr ₂ O ₃	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Fe ₂ O ₃	0.707	1.215	0.000	1.268	4.366	1.013	1.378	2.445	2.312	1.130
FeO	3.289	2.607	24.111	23.390	19.834	22.477	22.382	21.228	21.590	20.616
MnO	0.187	0.147	0.645	0.582	0.740	0.646	0.652	0.603	0.630	0.541
MgO	16.471	16.598	3.600	3.027	3.261	3.897	3.521	3.348	3.204	4.402
CaO	24.190	24.367	20.387	21.511	23.716	20.449	22.151	22.583	22.685	22.214
Na ₂ O	0.000	0.000	0.281	0.271	0.297	0.343	0.297	0.308	0.220	0.243
K ₂ O	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Total	99.692	98.890	98.835	99.662	101.599	99.065	100.693	100.095	100.103	99.042
Number of ions on the basis of 6O										
Si	1.958	1.969	1.993	1.975	1.949	1.958	1.974	1.963	1.962	1.975
Al ^{IV}	0.042	0.031	0.007	0.025	0.035	0.042	0.026	0.037	0.038	0.025
Al	0.020	0.004	0.027	0.020	0.000	0.045	0.016	0.006	0.006	0.017
Fe ³⁺	0.020	0.034	0.000	0.039	0.130	0.031	0.041	0.074	0.070	0.034
Cr	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Ti	0.005	0.002	0.002	0.000	0.001	0.002	0.002	0.003	0.001	0.002
Fe ²⁺	0.101	0.081	0.820	0.791	0.658	0.759	0.746	0.713	0.726	0.694
Mn	0.006	0.005	0.022	0.020	0.025	0.022	0.022	0.021	0.022	0.018
Mg	0.903	0.918	0.218	0.183	0.195	0.235	0.210	0.202	0.193	0.265
Ca	0.953	0.968	0.888	0.935	1.019	0.887	0.950	0.978	0.983	0.960
Na	0.000	0.000	0.022	0.021	0.023	0.027	0.023	0.024	0.017	0.019
K	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
TOTAL	4.006	4.011	3.999	4.010	4.035	4.008	4.011	4.020	4.019	4.009
En	45.5	45.8	11.1	9.2	9.5	12.0	10.5	10.0	9.6	13.3
Fs	6.4	5.9	42.7	42.7	39.7	41.4	40.6	40.2	40.7	37.5
Wo	48.1	48.3	45.1	47.0	49.7	45.2	47.7	48.6	48.9	48.2
Ac	0.0	0.0	1.1	1.1	1.1	1.4	1.2	1.2	0.9	1.0
	100	100	100	100	100	100	100	100	100	100

Table C-4 (cont.): Pyroxene electron-microprobe analyses, with stoichiometry based on 6O, and using the calculated Fe³⁺ content according to Droop (1987).

Section	H5-180	H5-180	H5-180	R8-169a	R8-169a	R8-169a	R8-169a	HM1-69	HM1-69	HM1-69
Analysis	(2-1) 2	(2-1) 4	(2-1) 5	(1-1) 1	(1-1) 2	(1-1) 3	(1-1) 4	(1-1) 1	(1-1) 2	(1-1) 3
SiO ₂	48.842	49.569	49.096	50.786	50.597	51.044	51.406	50.983	51.022	51.001
TiO ₂	0.000	0.052	0.000	0.000	0.000	0.043	0.000	0.000	0.000	0.000
Al ₂ O ₃	0.809	0.836	0.814	0.685	0.641	0.905	0.764	0.895	0.992	0.717
Cr ₂ O ₃	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Fe ₂ O ₃	0.068	2.363	2.618	0.832	2.014	2.499	2.470	1.699	0.890	0.475
FeO	22.294	18.600	18.772	14.974	13.514	12.904	12.131	14.465	15.121	16.199
MnO	0.582	0.424	0.525	0.425	0.441	0.548	0.643	1.097	1.167	1.110
MgO	4.146	4.817	4.866	8.504	8.376	8.522	8.982	7.839	7.690	7.455
CaO	21.153	23.929	23.143	22.497	23.622	24.142	24.276	23.551	23.224	22.858
Na ₂ O	0.222	0.226	0.233	0.290	0.291	0.323	0.340	0.268	0.269	0.235
K ₂ O	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Total	98.117	100.816	100.067	98.992	99.496	100.929	101.012	100.797	100.375	100.050
Number of ions on the basis of 6O										
Si	1.989	1.964	1.964	1.987	1.977	1.966	1.972	1.973	1.979	1.988
Al ^{IV}	0.011	0.036	0.036	0.013	0.023	0.034	0.028	0.027	0.021	0.012
Al	0.027	0.003	0.002	0.019	0.006	0.007	0.006	0.014	0.024	0.021
Fe ³⁺	0.002	0.070	0.078	0.024	0.059	0.072	0.071	0.049	0.026	0.014
Cr	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Ti	0.000	0.002	0.000	0.000	0.000	0.001	0.000	0.000	0.000	0.000
Fe ²⁺	0.759	0.613	0.624	0.489	0.439	0.413	0.387	0.466	0.489	0.527
Mn	0.020	0.014	0.018	0.014	0.015	0.018	0.021	0.036	0.038	0.037
Mg	0.252	0.284	0.290	0.496	0.488	0.489	0.513	0.452	0.445	0.433
Ca	0.923	1.016	0.992	0.943	0.989	0.996	0.998	0.977	0.965	0.955
Na	0.018	0.017	0.018	0.022	0.022	0.024	0.025	0.020	0.020	0.018
K	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
TOTAL	4.001	4.019	4.022	4.007	4.017	4.021	4.021	4.014	4.008	4.004
En	12.8	14.1	14.4	24.9	24.2	24.3	25.5	22.6	22.4	21.8
Fs	39.6	34.6	35.6	26.5	25.5	25.0	23.8	27.6	27.9	29.1
Wo	46.8	50.4	49.1	47.4	49.2	49.5	49.5	48.8	48.7	48.1
Ac	0.9	0.9	0.9	1.1	1.1	1.2	1.3	1.0	1.0	0.9
	100	100	100	100	100	100	100	100	100	100

Table C-4 (cont.): Pyroxene electron-microprobe analyses, with stoichiometry based on 6O, and using the calculated Fe³⁺ content according to Droop (1987).

METASOMATIC GARNETS										
Section	R12-119	R12-119	R12-119	R12-119	H5-180	H5-180	H5-180	H5-180	H5-180	H5-180
Analysis	(1-3) 1	(2-1) 1	(2-1) 2	(3-2) 1	(1-1) 1	(1-1) 2	(1-1) 3	(1-1) 4	(1-2)	(1-2) 2
SiO₂	37.675	38.234	37.908	40.023	36.177	36.236	37.747	36.170	37.910	37.180
TiO₂	1.229	0.228	0.285	0.097	0.633	0.526	0.682	0.703	0.821	0.474
Al₂O₃	8.465	8.120	7.093	16.461	6.811	6.566	6.792	6.841	7.100	6.319
Fe₂O₃	16.113	18.875	20.388	5.435	20.918	19.850	18.583	20.153	18.717	17.910
FeO	3.035	2.271	1.674	9.327	2.444	4.560	4.788	3.564	4.136	5.491
MnO	0.300	0.328	0.271	0.351	0.517	0.665	0.501	0.610	0.551	0.567
MgO	0.040	0.000	0.000	0.036	0.057	0.035	0.064	0.064	0.085	0.036
CaO	33.364	33.813	34.060	29.815	31.814	30.056	31.487	30.899	32.177	30.250
Cr₂O₃	0.000	0.000	0.018	0.018	0.056	0.005	0.000	0.005	0.000	0.038
Total	100.222	101.869	101.697	101.563	99.427	98.499	100.644	99.009	101.497	98.265
Number of ions on the basis of 24O										
Si	6.33	6.38	6.39	6.26	6.30	6.37	6.44	6.32	6.40	6.50
Al^{iv}	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Al^{vi}	1.68	1.60	1.41	3.03	1.40	1.36	1.36	1.41	1.41	1.30
Fe³⁺	1.95	2.26	2.46	0.63	2.59	2.49	2.27	2.51	2.26	2.25
Cr	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.01
Ti	0.16	0.03	0.04	0.01	0.08	0.07	0.09	0.09	0.10	0.06
Fe²⁺	0.41	0.30	0.22	1.20	0.34	0.64	0.65	0.49	0.56	0.77
Mn	0.04	0.05	0.04	0.05	0.08	0.10	0.07	0.09	0.08	0.08
Mg	0.01	0.00	0.00	0.01	0.01	0.01	0.02	0.02	0.02	0.01
Ca	6.00	6.05	6.16	4.99	5.94	5.66	5.75	5.78	5.82	5.67
Total	16.58	16.66	16.72	16.19	16.75	16.70	16.65	16.72	16.65	16.64
pyrope	0.15	0.00	0.00	0.13	0.23	0.14	0.25	0.26	0.33	0.14
almandine	6.33	4.72	3.49	19.25	5.29	9.92	10.02	7.73	8.59	11.73
spessartine	0.66	0.73	0.60	0.74	1.20	1.55	1.11	1.41	1.22	1.29
uvavovite	0.00	0.00	0.06	0.06	0.19	0.02	0.00	0.02	0.00	0.15
grossular	41.24	36.40	32.95	62.66	29.56	24.86	27.61	28.00	29.97	24.58
andradite	51.62	58.15	62.90	17.16	63.53	63.51	61.00	62.58	59.90	62.12
pyralspite	7.14	5.45	4.09	20.13	6.72	11.61	11.39	9.40	10.14	13.16

Table C-5: Garnet electron-microprobe analyses, with stoichiometry based on 24O, and using the calculated Fe³⁺ content according to Droop (1987).

METASOMATIC GARNETS										
Section	H5-180	H5-180	H5-180	H5-180	H5-180	H5-180	H5-180	R8-169a	R8-169a	R8-169a
Analysis	(1-3) 3	(1-3) 5	(1-3) 6	(1-4) 1	(1-4) 2	(1-2) 7	(1-2) 9	(1-1) 1	(1-1) 2	(1-1) 3
	core							core	core	core
SiO ₂	36.203	36.626	36.377	36.226	36.384	36.742	35.703	36.544	35.742	35.777
TiO ₂	0.806	0.619	0.811	0.414	0.488	0.701	0.425	0.568	0.521	0.426
Al ₂ O ₃	7.026	6.839	6.737	6.345	6.562	6.886	5.991	4.406	4.819	4.257
Fe ₂ O ₃	22.283	20.630	20.042	22.466	21.645	21.344	23.332	24.009	22.960	23.539
FeO	0.413	1.918	3.193	1.154	1.195	1.423	1.160	1.875	1.554	2.201
MnO	0.545	0.531	0.558	0.589	0.554	0.559	0.831	0.503	0.595	0.505
MgO	0.065	0.066	0.071	0.036	0.046	0.038	0.043	0.047	0.027	0.045
CaO	33.512	32.610	31.489	32.685	32.866	33.179	31.999	32.580	32.004	31.511
Cr ₂ O ₃	0.003	0.066	0.000	0.000	0.048	0.017	0.046	0.000	0.002	0.000
Total	100.856	99.905	99.278	99.915	99.789	100.890	99.530	100.532	98.225	98.262
Number of ions on the basis of 24O										
Si	6.22	6.33	6.33	6.31	6.32	6.30	6.29	6.41	6.39	6.43
Al ^{iv}	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Al ^{vi}	1.42	1.39	1.38	1.30	1.34	1.39	1.24	0.91	1.02	0.90
Fe ³⁺	2.72	2.54	2.49	2.78	2.67	2.60	2.91	2.97	2.90	2.99
Cr	0.00	0.01	0.00	0.00	0.01	0.00	0.01	0.00	0.00	0.00
Ti	0.10	0.08	0.11	0.05	0.06	0.09	0.06	0.07	0.07	0.06
Fe ²⁺	0.06	0.26	0.44	0.16	0.16	0.19	0.16	0.26	0.22	0.31
Mn	0.08	0.08	0.08	0.09	0.08	0.08	0.12	0.07	0.09	0.08
Mg	0.02	0.02	0.02	0.01	0.01	0.01	0.01	0.01	0.01	0.01
Ca	6.17	6.03	5.87	6.10	6.12	6.09	6.04	6.13	6.13	6.07
Total	16.79	16.74	16.71	16.80	16.78	16.76	16.83	16.84	16.83	16.84
pyrope	0.26	0.27	0.29	0.15	0.19	0.15	0.18	0.19	0.11	0.19
almandine	0.88	4.11	6.87	2.49	2.57	3.03	2.53	3.99	3.39	4.80
spessartine	1.25	1.22	1.28	1.37	1.28	1.27	1.96	1.16	1.40	1.19
uvarovite	0.01	0.22	0.00	0.00	0.16	0.06	0.15	0.00	0.01	0.00
grossular	33.56	31.03	28.98	28.83	30.40	31.79	26.20	19.57	22.32	18.14
andradite	64.02	63.16	62.58	67.16	65.40	63.70	68.98	75.10	72.77	75.68
pyralspite	2.40	5.59	8.44	4.01	4.04	4.45	4.67	5.33	4.90	6.17

Table C-5 (cont.): Garnet electron-microprobe analyses, with stoichiometry based on 24O, and using the calculated Fe³⁺ content according to Droop (1987).

METASOMATIC GARNETS							
Section	R8-169a	R8-169a	R8-169a	R8-169a	R8-169a	R8-169a	HM1-69
Analysis	(1-1) 7	(5-1) 1	(5-1) 2	(5-1) 3	(5-1) 4	(5-1) 5	(1-1) 5
SiO₂	36.204	36.253	35.836	35.862	35.652	35.786	35.970
TiO₂	0.386	0.327	0.433	0.455	0.343	0.360	0.304
Al₂O₃	4.790	4.742	4.685	4.234	4.725	4.525	3.690
Fe₂O₃	22.471	24.153	24.618	26.033	23.782	23.266	24.742
FeO	1.882	2.018	1.056	0.941	2.057	2.745	2.997
MnO	0.609	0.598	0.539	0.655	0.629	0.629	1.714
MgO	0.029	0.025	0.039	0.048	0.031	0.054	0.044
CaO	32.071	31.983	32.446	32.472	31.370	30.938	30.030
Cr₂O₃	0.000	0.014	0.000	0.041	0.046	0.012	0.017
Total	98.442	100.113	99.653	100.740	98.635	98.315	99.508
Number of ions on the basis of 24O							
Si	6.44	6.39	6.36	6.34	6.38	6.42	6.45
Al^{iv}	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Al^{vi}	1.00	0.99	0.98	0.88	1.00	0.96	0.78
Fe³⁺	2.83	3.01	3.08	3.23	3.00	2.95	3.12
Cr	0.00	0.00	0.00	0.01	0.01	0.00	0.00
Ti	0.05	0.04	0.06	0.06	0.05	0.05	0.04
Fe²⁺	0.26	0.28	0.15	0.13	0.29	0.39	0.42
Mn	0.09	0.09	0.08	0.10	0.10	0.10	0.26
Mg	0.01	0.01	0.01	0.01	0.01	0.01	0.01
Ca	6.12	6.04	6.17	6.15	6.02	5.95	5.77
Total	16.81	16.85	16.87	16.91	16.85	16.83	16.87
pyrope	0.12	0.10	0.16	0.20	0.13	0.22	0.18
almandine	4.07	4.35	2.29	2.03	4.50	6.00	6.50
spessartine	1.42	1.39	1.26	1.53	1.49	1.48	4.03
uvarovite	0.00	0.05	0.00	0.14	0.16	0.04	0.06
grossular	21.56	19.66	21.50	18.80	19.62	17.71	10.09
andradite	72.83	74.45	74.78	77.30	74.10	74.54	79.13
pyralspite	5.60	5.84	3.72	3.76	6.12	7.71	10.71

Table C-5 (cont.): Garnet electron-microprobe analyses, with stoichiometry based on 24O, and using the calculated Fe³⁺ content according to Droop (1987).

METAMORPHIC GARNETS										
Section	P5-16	P5-16	P5-16	P5-16	P5-16	P5-16	P5-16	P5-16	P5-16	P5-16
Analysis	(1-1) 1	(1-1) 2	(1-1) 3	(1-1) 4	(1-1) 5	(1-1) 6	(1-1) 7	(1-1) 8	(2-1) 1	(1-1) 9
	rim	rim	rim	rim	core	core	core	core	core	
SiO ₂	37.730	37.464	37.343	37.692	38.003	37.322	37.494	36.955	37.458	36.556
TiO ₂	0.046	0.021	0.010	0.000	0.038	0.010	0.038	0.002	0.000	0.063
Al ₂ O ₃	21.221	20.852	20.391	21.251	20.410	19.713	20.730	19.734	20.505	19.509
Fe ₂ O ₃	0.674	0.612	0.420	0.331	0.922	0.921	1.061	1.117	0.801	1.997
FeO	33.638	33.290	34.108	33.068	33.059	32.362	32.063	31.891	32.667	30.617
MnO	2.164	2.255	2.960	2.214	2.774	2.644	3.215	3.212	2.526	2.855
MgO	3.577	3.550	2.688	3.797	3.798	3.742	3.636	3.515	3.838	3.578
CaO	2.304	2.276	2.158	2.336	2.216	2.285	2.395	2.171	2.127	3.031
Cr ₂ O ₃	0.000	0.000	0.000	0.000	0.024	0.004	0.000	0.000	0.027	0.018
Total	101.355	100.319	100.078	100.689	101.243	99.003	100.631	98.598	99.948	98.224
Number of ions on the basis of 24O										
Si	5.98	6.00	6.04	5.99	6.04	6.07	6.00	6.05	6.02	6.02
Al ^{iv}	0.02	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00
Al ^{vi}	3.95	3.94	3.89	3.98	3.83	3.78	3.91	3.81	3.89	3.78
Fe ³⁺	0.08	0.07	0.05	0.04	0.11	0.11	0.13	0.14	0.10	0.25
Cr	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Ti	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01
Fe ²⁺	4.45	4.45	4.61	4.39	4.39	4.39	4.28	4.35	4.38	4.19
Mn	0.29	0.31	0.41	0.30	0.37	0.36	0.44	0.45	0.34	0.40
Mg	0.85	0.85	0.65	0.90	0.90	0.91	0.87	0.86	0.92	0.88
Ca	0.39	0.39	0.37	0.40	0.38	0.40	0.41	0.38	0.37	0.53
Total	16.02	16.02	16.01	16.01	16.03	16.03	16.03	16.03	16.02	16.06
pyrope	14.14	14.13	10.74	15.02	14.91	14.97	14.47	14.21	15.29	14.62
almandine	74.46	74.25	76.35	73.35	72.65	72.45	71.41	72.11	72.89	69.84
spessartine	4.86	5.10	6.72	4.98	6.19	6.01	7.27	7.38	5.72	6.63
uvarovite	0.00	0.00	0.00	0.00	0.08	0.01	0.00	0.00	0.09	0.06
grossular	4.55	4.68	4.90	5.66	3.39	3.67	3.70	2.83	3.58	2.75
andradite	1.99	1.83	1.30	0.99	2.79	2.89	3.15	3.48	2.43	6.09
pyralspite	93.46	93.49	93.80	93.36	93.75	93.43	93.15	93.69	93.91	91.10

Table C-5 (cont.): Garnet electron-microprobe analyses, with stoichiometry based on 24O, and using the calculated Fe³⁺ content according to Droop (1987).

METAMORPHIC GARNETS										
Section	P5-16	P5-16	P5-16	P5-16	P6-43	P6-43	P6-43	P6-43	P6-43	P6-43
Analysis	(1-1) 10	(1-1) 11	(1-1) 12	(1-1) 13	1	2	3	4	5	6
					rim	traverse	traverse	traverse	traverse	traverse
SiO ₂	36.949	37.701	38.071	37.555	37.950	38.310	38.326	38.466	38.162	38.365
TiO ₂	0.041	0.052	0.000	0.021	0.044	0.000	0.013	0.000	0.010	0.016
Al ₂ O ₃	19.631	20.247	20.495	20.360	21.364	21.277	21.525	21.700	21.453	21.747
Fe ₂ O ₃	0.948	0.120	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
FeO	31.998	33.096	32.830	33.006	31.983	31.749	31.476	30.586	31.198	31.199
MnO	2.689	2.513	2.536	2.604	1.665	1.477	1.447	1.664	1.774	1.760
MgO	3.804	3.813	3.919	3.714	4.203	5.066	5.281	5.319	5.104	5.360
CaO	2.121	2.100	2.087	2.011	1.909	2.028	2.095	2.025	1.989	2.085
Cr ₂ O ₃	0.027	0.051	0.007	0.018	0.041	0.017	0.004	0.046	0.000	0.021
Total	98.208	99.693	99.945	99.289	99.159	99.924	100.167	99.806	99.690	100.553
Number of ions on the basis of 24O										
Si	6.06	6.07	6.09	6.06	6.06	6.06	6.04	6.06	6.04	6.02
Al ^{iv}	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Al ^{vi}	3.79	3.84	3.86	3.87	4.02	3.97	4.00	4.03	4.00	4.02
Fe ³⁺	0.12	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Cr	0.00	0.01	0.00	0.00	0.01	0.00	0.00	0.01	0.00	0.00
Ti	0.01	0.01	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.00
Fe ²⁺	4.38	4.45	4.40	4.46	4.29	4.21	4.15	4.05	4.14	4.10
Mn	0.37	0.34	0.34	0.36	0.23	0.20	0.19	0.22	0.24	0.23
Mg	0.93	0.91	0.93	0.89	1.00	1.19	1.24	1.25	1.20	1.25
Ca	0.37	0.36	0.36	0.35	0.33	0.34	0.35	0.34	0.34	0.35
Total	16.03	16.00	15.98	16.00	15.94	15.97	15.97	15.95	15.97	15.98
pyrope	15.36	15.06	15.49	14.76	17.12	20.08	20.87	21.31	20.34	21.11
almandine	72.31	73.33	72.89	73.61	73.44	70.81	69.93	69.07	69.95	69.06
spessartine	6.17	5.64	5.70	5.88	3.85	3.33	3.25	3.79	4.02	3.94
uvarovite	0.09	0.17	0.02	0.06	0.13	0.05	0.01	0.14	0.00	0.06
grossular	3.09	5.42	5.91	5.69	5.46	5.73	5.94	5.69	5.70	5.84
andradite	2.98	0.37	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
pyralspite	93.84	94.04	94.07	94.26	94.41	94.22	94.05	94.17	94.30	94.10

Table C-5 (cont.): Garnet electron-microprobe analyses, with stoichiometry based on 24O, and using the calculated Fe³⁺ content according to Droop (1987).

METAMORPHIC GARNETS										
Section	P6-43	P6-43	P6-43	P6-43	P6-43	P6-43	P6-43	P6-43	P6-43	P6-43
Analysis	7	8	9	10	11	12	13	14	15	16
	traverse	traverse	traverse	traverse	traverse	traverse	traverse	traverse	rim	rim
SiO ₂	38.061	38.215	38.272	38.365	38.490	38.110	38.556	38.325	38.055	37.785
TiO ₂	0.000	0.086	0.011	0.000	0.000	0.000	0.042	0.000	0.017	0.000
Al ₂ O ₃	21.468	21.420	21.565	21.475	21.508	21.528	21.363	21.552	21.416	21.541
Fe ₂ O ₃	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
FeO	30.965	30.609	30.137	30.784	30.951	30.484	31.263	31.541	33.968	32.633
MnO	1.944	2.176	2.019	2.164	1.949	1.995	1.536	1.555	1.827	1.675
MgO	5.046	5.113	5.268	5.336	5.371	5.488	5.220	5.216	3.637	4.288
CaO	2.033	2.202	2.069	1.905	2.072	1.943	2.031	2.050	1.964	2.012
Cr ₂ O ₃	0.021	0.000	0.000	0.000	0.021	0.042	0.041	0.000	0.000	0.000
Total	99.538	99.821	99.341	100.029	100.362	99.590	100.052	100.239	100.884	99.934
Number of ions on the basis of 24O										
Si	6.04	6.04	6.06	6.05	6.05	6.03	6.07	6.03	6.03	6.01
Al ^{iv}	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Al ^{vi}	4.01	3.99	4.02	3.99	3.98	4.01	3.97	4.00	4.00	4.04
Fe ³⁺	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Cr	0.00	0.00	0.00	0.00	0.00	0.01	0.01	0.00	0.00	0.00
Ti	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Fe ²⁺	4.12	4.06	4.01	4.07	4.08	4.04	4.13	4.16	4.51	4.35
Mn	0.26	0.29	0.27	0.29	0.26	0.27	0.20	0.21	0.25	0.23
Mg	1.19	1.20	1.24	1.25	1.26	1.29	1.23	1.22	0.86	1.02
Ca	0.35	0.37	0.35	0.32	0.35	0.33	0.34	0.35	0.33	0.34
Total	15.97	15.97	15.95	15.97	15.97	15.97	15.95	15.97	15.98	15.98
pyrope	20.16	20.33	21.17	21.13	21.17	21.81	20.75	20.61	14.44	17.13
almandine	69.59	68.47	68.24	68.58	68.60	68.13	69.98	70.08	75.83	73.29
spessartine	4.41	4.92	4.61	4.87	4.36	4.51	3.47	3.49	4.12	3.80
uvarovite	0.07	0.00	0.00	0.00	0.07	0.13	0.13	0.00	0.00	0.00
grossular	5.77	6.29	5.98	5.42	5.80	5.42	5.67	5.82	5.61	5.78
andradite	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
pyralspite	94.16	93.71	94.02	94.58	94.13	94.45	94.20	94.18	94.39	94.22

Table C-5 (cont.): Garnet electron-microprobe analyses, with stoichiometry based on 24O, and using the calculated Fe³⁺ content according to Droop (1987).

METAMORPHIC GARNETS									
Section	P6-43	P6-43	P6-43	P6-43	P6-43	P6-43	P6-43	P6-43	P6-43
Analysis	17	18.1	19.1	19.2	19.3	19.4	18.2	18.3	18.4
	rim		rim			rim	rim	rim	
SiO ₂	38.057	38.175	37.933	38.008	38.055	37.702	37.968	37.803	37.551
TiO ₂	0.000	0.022	0.064	0.043	0.024	0.000	0.044	0.000	0.057
Al ₂ O ₃	21.420	21.627	21.273	21.452	21.482	21.128	21.377	21.140	21.300
Fe ₂ O ₃	0.000	0.000	0.000	0.000	0.000	0.065	0.000	0.000	0.000
FeO	31.906	32.852	33.635	32.639	32.448	33.768	33.202	33.630	32.466
MnO	1.581	1.679	1.762	1.637	1.576	1.968	1.828	2.011	1.591
MgO	4.779	4.452	3.981	4.664	4.934	3.839	3.966	3.603	4.467
CaO	1.917	2.029	1.960	1.994	1.953	1.935	1.910	1.981	1.973
Cr ₂ O ₃	0.000	0.033	0.000	0.029	0.008	0.049	0.029	0.012	0.008
Total	99.660	100.869	100.608	100.466	100.480	100.453	100.324	100.180	99.413
Number of ions on the basis of 24O									
Si	6.04	6.01	6.02	6.01	6.01	6.01	6.03	6.04	6.00
Al ^{iv}	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Al ^{vi}	4.01	4.02	3.98	4.00	4.00	3.97	4.00	3.98	4.01
Fe ³⁺	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00
Cr	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00
Ti	0.00	0.00	0.01	0.01	0.00	0.00	0.01	0.00	0.01
Fe ²⁺	4.25	4.33	4.47	4.32	4.29	4.50	4.42	4.50	4.35
Mn	0.21	0.22	0.24	0.22	0.21	0.27	0.25	0.27	0.22
Mg	1.13	1.05	0.94	1.10	1.16	0.91	0.94	0.86	1.06
Ca	0.33	0.34	0.33	0.34	0.33	0.33	0.33	0.34	0.34
Total	15.97	15.98	15.99	15.99	15.99	16.00	15.97	15.98	15.99
pyrope	19.11	17.58	15.75	18.39	19.39	15.18	15.83	14.37	17.85
almandine	71.79	72.90	74.72	72.29	71.58	74.90	74.54	75.39	72.87
spessartine	3.59	3.77	3.96	3.67	3.52	4.42	4.15	4.56	3.61
uvarovite	0.00	0.10	0.00	0.09	0.02	0.16	0.09	0.04	0.03
grossular	5.51	5.66	5.57	5.56	5.49	5.15	5.39	5.64	5.64
andradite	0.00	0.00	0.00	0.00	0.00	0.20	0.00	0.00	0.00
pyralspite	94.49	94.24	94.43	94.35	94.48	94.50	94.52	94.32	94.33

Table C-5 (cont.): Garnet electron-microprobe analyses, with stoichiometry based on 24O, and using the calculated Fe³⁺ content according to Droop (1987).

PLAGIOCLASE ANALYSES										
Section	R8-216b	R8-216b	R8-216b	R8-216b	R8-216b	R8-216b	R8-216b	R8-216b	R8-216b	R8-216b
Analysis	2	3	4	5	6	7	8	9	10	11
SiO ₂	57.366	56.112	57.191	57.509	58.708	58.765	59.692	57.673	56.979	57.189
TiO ₂	0.000	0.000	0.025	0.000	0.016	0.032	0.000	0.022	0.005	0.000
Al ₂ O ₃	26.209	26.983	26.826	27.158	26.027	26.497	25.879	26.459	27.381	26.800
Fe ₂ O ₃	0.113	0.143	0.167	0.242	0.091	0.186	0.083	0.106	0.041	0.076
MgO	0.000	0.000	0.018	0.009	0.002	0.000	0.014	0.010	0.000	0.000
CaO	7.613	9.081	8.471	8.403	7.319	7.946	7.278	7.838	8.501	8.500
Na ₂ O	7.627	6.372	6.778	6.992	7.663	7.105	7.805	7.183	7.070	7.091
K ₂ O	0.053	0.071	0.049	0.073	0.059	0.082	0.062	0.068	0.052	0.057
Total	98.981	98.762	99.525	100.386	99.885	100.613	100.813	99.359	100.029	99.713
Number of ions on the basis of 32O										
Si	10.38	10.20	10.30	10.27	10.50	10.44	10.57	10.39	10.22	10.29
Al	5.59	5.78	5.69	5.72	5.49	5.55	5.40	5.62	5.79	5.68
Fe ³⁺	0.02	0.02	0.02	0.03	0.01	0.02	0.01	0.01	0.01	0.01
Ti	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Mg	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Ca	1.48	1.77	1.63	1.61	1.40	1.51	1.38	1.51	1.63	1.64
Na	2.68	2.25	2.37	2.42	2.66	2.45	2.68	2.51	2.46	2.47
K	0.01	0.02	0.01	0.02	0.01	0.02	0.01	0.02	0.01	0.01
Total	20.16	20.03	20.03	20.07	20.08	20.00	20.07	20.06	20.12	20.11
Ab	64.3	55.7	59.0	59.8	65.2	61.5	65.8	62.1	59.9	60.0
An	35.4	43.9	40.7	39.7	34.4	38.0	33.9	37.5	39.8	39.7
Or	0.3	0.4	0.3	0.4	0.3	0.5	0.3	0.4	0.3	0.3
Total	100	100	100	100	100	100	100	100	100	100

Table C-6: Feldspar electron-microprobe analyses, with stoichiometry based on 32O.

PLAGIOCLASE ANALYSES										
Section	R8-216b	R8-216b	R8-216b	R7-296	R7-296	R8-216b	R8-216b	R8-216b	R7-296	R7-296
Analysis	13	14	15	4.1	4.2	(1-2) 2	(2-1) 1	12	2.3	4.3
SiO ₂	56.666	56.887	57.439	54.423	53.062	55.802	55.998	53.634	52.876	52.675
TiO ₂	0.046	0.044	0.049	0.012	0.020	0.000	0.064	0.028	0.053	0.038
Al ₂ O ₃	27.076	27.378	25.956	28.560	28.472	27.514	26.326	29.393	29.420	29.176
Fe ₂ O ₃	0.079	0.216	0.216	0.102	0.253	0.279	0.787	0.106	0.192	0.151
MgO	0.006	0.000	0.000	0.008	0.000	0.049	0.278	0.000	0.000	0.000
CaO	8.442	8.945	6.945	9.935	10.150	9.381	8.831	11.132	11.586	10.528
Na ₂ O	7.066	6.557	7.774	6.226	6.040	5.464	5.920	5.402	5.201	5.633
K ₂ O	0.053	0.062	0.092	0.107	0.096	0.000	0.096	0.053	0.074	0.069
Total	99.434	100.089	98.471	99.373	98.093	98.489	98.300	99.748	99.402	98.270
Number of ions on the basis of 32O										
Si	10.23	10.20	10.44	9.88	9.78	10.15	10.23	9.72	9.64	9.69
Al	5.76	5.79	5.56	6.11	6.19	5.90	5.67	6.28	6.32	6.33
Fe ³⁺	0.01	0.03	0.03	0.01	0.04	0.04	0.11	0.01	0.03	0.02
Ti	0.01	0.01	0.01	0.00	0.00	0.00	0.01	0.00	0.01	0.01
Mg	0.00	0.00	0.00	0.00	0.00	0.01	0.08	0.00	0.00	0.00
Ca	1.63	1.72	1.35	1.93	2.00	1.83	1.73	2.16	2.26	2.08
Na	2.47	2.28	2.74	2.19	2.16	1.93	2.10	1.90	1.84	2.01
K	0.01	0.01	0.02	0.02	0.02	0.00	0.02	0.01	0.02	0.02
Total	20.12	20.03	20.14	20.16	20.19	19.85	19.93	20.09	20.11	20.14
Ab	60.1	56.8	66.6	52.8	51.6	51.3	54.5	46.6	44.6	49.0
An	39.6	42.8	32.9	46.6	47.9	48.7	44.9	53.1	54.9	50.6
Or	0.3	0.4	0.5	0.6	0.5	0.0	0.6	0.3	0.4	0.4
Total	100	100	100	100	100	100	100	100	100	100

Table C-6 (cont.): Feldspar electron-microprobe analyses, with stoichiometry based on 32O.

PLAGIOCLASE ANALYSES										
Section	R7-296	R7-296	R7-296	R7-296	R7-296	R7-296	R7-296	R7-296	R7-296	R7-296
Analysis	2.1	2.2	2.4	1.1	1.2	1.3	(1-2) 1	(1-2) 2	(1-2) 3	(1-2) 4
SiO ₂	58.222	56.861	57.843	57.642	59.481	57.888	58.848	58.417	66.136	65.631
TiO ₂	0.020	0.031	0.026	0.024	0.000	0.000	0.000	0.000	0.000	0.000
Al ₂ O ₃	26.299	27.121	26.661	25.858	25.782	26.202	26.524	26.138	22.376	23.195
Fe ₂ O ₃	0.087	0.174	0.162	0.076	0.227	0.022	0.084	0.153	0.271	0.162
MgO	0.000	0.000	0.000	0.000	0.019	0.001	0.000	0.000	0.000	0.000
CaO	7.668	8.809	8.231	7.879	7.326	7.709	7.181	7.341	5.398	5.820
Na ₂ O	7.106	7.087	7.098	7.786	8.027	7.506	6.831	6.890	6.015	6.383
K ₂ O	0.538	0.121	0.078	0.128	0.116	0.084	0.159	0.109	0.109	0.106
Total	99.940	100.204	100.099	99.393	100.978	99.412	99.627	99.048	100.305	101.297
Number of ions on the basis of 32O										
Si	10.44	10.21	10.35	10.41	10.54	10.42	10.51	10.51	11.51	11.36
Al	5.56	5.74	5.62	5.50	5.39	5.56	5.59	5.54	4.59	4.73
Fe ³⁺	0.01	0.02	0.02	0.01	0.03	0.00	0.01	0.02	0.04	0.02
Ti	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Mg	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.00
Ca	1.47	1.69	1.58	1.52	1.39	1.49	1.37	1.42	1.01	1.08
Na	2.47	2.47	2.46	2.73	2.76	2.62	2.37	2.40	2.03	2.14
K	0.12	0.03	0.02	0.03	0.03	0.02	0.04	0.03	0.02	0.02
Total	20.07	20.16	20.06	20.21	20.14	20.11	19.89	19.92	19.20	19.35
Ab	60.7	58.9	60.7	63.7	66.1	63.5	62.6	62.5	66.3	66.0
An	36.2	40.4	38.9	35.6	33.3	36.0	36.4	36.8	32.9	33.3
Or	3.0	0.7	0.4	0.7	0.6	0.5	1.0	0.7	0.8	0.7
Total	100	100	100	100	100	100	100	100	100	100

Table C-6 (cont.): Feldspar electron-microprobe analyses, with stoichiometry based on 32O.

PLAGIOCLASE ANALYSES										
Section	R7-104b	R8-216b	R6-284	R6-284	R6-284	R6-284	R6-284	R6-284	R6-284	R6-284
Analysis	(1-1) 2	1	1.1	1.2	1.3	2.1	2.2	2.3	2.4	2.5
SiO ₂	59.526	60.525	60.865	61.036	60.051	60.513	59.086	60.483	62.800	62.254
TiO ₂	0.000	0.851	0.039	0.000	0.041	0.023	0.061	0.026	0.002	0.011
Al ₂ O ₃	26.944	23.791	24.468	24.250	23.930	24.126	24.401	24.471	23.244	22.821
Fe ₂ O ₃	0.346	0.167	0.250	0.200	0.401	0.189	0.162	0.264	0.253	0.231
MgO	0.000	0.014	0.000	0.000	0.057	0.000	0.000	0.000	0.012	0.000
CaO	7.063	5.779	5.124	5.103	4.941	4.831	5.117	5.203	4.814	5.061
Na ₂ O	6.787	8.648	9.051	8.999	9.161	9.223	9.074	8.780	8.888	8.728
K ₂ O	0.000	0.067	0.107	0.084	0.083	0.068	0.125	0.066	0.073	0.058
Total	100.666	99.842	99.904	99.672	98.665	98.973	98.026	99.293	100.086	99.164
Number of ions on the basis of 32O										
Si	10.51	10.81	10.84	10.89	10.84	10.87	10.75	10.83	11.11	11.13
Al	5.61	5.01	5.14	5.10	5.09	5.11	5.23	5.16	4.85	4.81
Fe ³⁺	0.05	0.02	0.03	0.03	0.05	0.03	0.02	0.04	0.03	0.03
Ti	0.00	0.11	0.01	0.00	0.01	0.00	0.01	0.00	0.00	0.00
Mg	0.00	0.00	0.00	0.00	0.02	0.00	0.00	0.00	0.00	0.00
Ca	1.34	1.11	0.98	0.98	0.96	0.93	1.00	1.00	0.91	0.97
Na	2.32	2.99	3.13	3.11	3.21	3.21	3.20	3.05	3.05	3.02
K	0.00	0.02	0.02	0.02	0.02	0.02	0.03	0.02	0.02	0.01
Total	19.82	20.07	20.14	20.12	20.19	20.17	20.23	20.10	19.98	19.97
Ab	63.5	72.8	75.7	75.8	76.7	77.3	75.7	75.1	76.6	75.5
An	36.5	26.9	23.7	23.7	22.9	22.4	23.6	24.6	22.9	24.2
Or	0.0	0.4	0.6	0.5	0.5	0.4	0.7	0.4	0.4	0.3
Total	100	100	100	100	100	100	100	100	100	100

Table C-6 (cont.): Feldspar electron-microprobe analyses, with stoichiometry based on 32O.

PLAGIOCLASE ANALYSES										
Section	R6-284	R6-284	R8-193	R8-193	R8-193	R8-193	R6-284	R6-284	R7-296	R8-219
Analysis	2.6	3.2	(3-1) 1	(3-1) 2	(3-1) 3	(3-1) 4	(4-1) 1	(4-1) 2	3.1	(3-1) 1
SiO ₂	61.478	62.565	60.532	60.689	62.056	61.934	61.799	61.894	64.738	62.154
TiO ₂	0.055	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.002	0.000
Al ₂ O ₃	22.912	24.077	23.268	23.580	23.551	23.948	24.996	25.081	20.947	23.653
Fe ₂ O ₃	0.639	0.212	0.103	0.126	0.000	0.163	0.365	0.356	0.091	0.422
MgO	0.140	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
CaO	5.456	5.267	5.696	5.481	5.032	5.406	4.973	5.183	1.202	4.824
Na ₂ O	8.558	8.884	8.578	8.562	8.808	8.599	7.847	7.787	11.588	7.903
K ₂ O	0.088	0.039	0.137	0.193	0.130	0.104	0.088	0.000	0.072	0.000
Total	99.326	101.044	98.314	98.631	99.577	100.154	100.068	100.301	98.640	98.956
Number of ions on the basis of 32O										
Si	11.01	10.99	10.95	10.94	11.05	10.97	10.91	10.90	11.56	11.08
Al	4.84	4.98	4.96	5.01	4.94	5.00	5.20	5.21	4.41	4.97
Fe ³⁺	0.09	0.03	0.01	0.02	0.00	0.02	0.05	0.05	0.01	0.06
Ti	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Mg	0.04	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Ca	1.05	0.99	1.10	1.06	0.96	1.03	0.94	0.98	0.23	0.92
Na	2.97	3.02	3.01	2.99	3.04	2.95	2.69	2.66	4.01	2.73
K	0.02	0.01	0.03	0.04	0.03	0.02	0.02	0.00	0.02	0.00
Total	20.02	20.02	20.08	20.06	20.02	20.00	19.81	19.80	20.24	19.77
Ab	73.6	75.2	72.6	73.1	75.4	73.8	73.7	73.1	94.2	74.8
An	25.9	24.6	26.6	25.8	23.8	25.6	25.8	26.9	5.4	25.2
Or	0.5	0.2	0.8	1.1	0.7	0.6	0.5	0.0	0.4	0.0
Total	100	100	100	100	100	100	100	100	100	100

Table C-6 (cont.): Feldspar electron-microprobe analyses, with stoichiometry based on 32O.

PLAGIOCLASE ANALYSES					K-FELDSPAR ANALYSES			
Section	R8-169a	R8-169a	HM1-69	HM1-69	P13-39	P13-39	P13-39	P13-39
Analysis	(6-1) 2	(6-1) 4	(2-1) 1	(2-1) 2	(1-1) 2	(1-1) 3	(2-1) 1	(2-2) 1
SiO ₂	59.850	60.023	58.322	58.132	62.597	62.109	63.163	63.804
TiO ₂	0.000	0.000	0.000	0.000	0.041	0.113	0.078	0.000
Al ₂ O ₃	25.388	25.021	25.593	26.005	18.388	18.309	18.412	18.203
Fe ₂ O ₃	0.171	0.183	0.146	0.096	0.080	0.000	0.120	0.157
MgO	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
CaO	6.326	6.331	7.664	7.263	0.000	0.000	0.000	0.000
Na ₂ O	6.966	7.097	6.600	6.648	0.706	0.792	0.748	0.651
K ₂ O	0.082	0.000	0.000	0.000	16.655	16.844	16.797	16.432
Total	98.783	98.655	98.325	98.144	98.467	98.167	99.318	99.247
Number of ions on the basis of 32O								
Si	10.74	10.78	10.56	10.54	11.84	11.81	11.85	11.93
Al	5.37	5.30	5.46	5.55	4.10	4.10	4.07	4.01
Fe ³⁺	0.02	0.02	0.02	0.01	0.01	0.00	0.02	0.02
Ti	0.00	0.00	0.00	0.00	0.01	0.02	0.01	0.00
Mg	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Ca	1.22	1.22	1.49	1.41	0.00	0.00	0.00	0.00
Na	2.42	2.47	2.32	2.34	0.26	0.29	0.27	0.24
K	0.02	0.00	0.00	0.00	4.02	4.09	4.02	3.92
<i>Total</i>	<i>19.79</i>	<i>19.79</i>	<i>19.85</i>	<i>19.85</i>	<i>20.24</i>	<i>20.31</i>	<i>20.24</i>	<i>20.13</i>
Ab	66.2	67.0	60.9	62.4	6.1	6.7	6.3	5.7
An	33.2	33.0	39.1	37.6	0.0	0.0	0.0	0.0
Or	0.5	0.0	0.0	0.0	93.9	93.3	93.7	94.3
Total	100	100	100	100	100	100	100	100

Table C-6 (cont.): Feldspar electron-microprobe analyses, with stoichiometry based on 32O

Calcic skarn (Amp-Pyx-Gt)							Calcic skarn (+ Scap)		
Section	R12-119	R12-119	R12-119	R12-119	HM1-69	HM1-69	R8-169a	R8-169a	R8-169a
Analysis	(1-2) 1	(1-2) 2	(3-1) 1	(3-1) 2	(1-1) 1	(1-1) 2	(3-1) 1	(3-1) 2	(5-1) 1
SiO ₂	38.312	38.700	38.806	39.388	37.592	38.372	37.621	37.388	56.993
TiO ₂	0.000	0.000	0.124	0.078	0.046	0.047	0.103	0.064	0.060
Al ₂ O ₃	26.379	26.387	23.739	26.361	22.861	24.395	22.530	22.746	14.419
Fe ₂ O ₃	9.708	9.534	13.622	10.038	13.492	10.762	13.219	13.251	9.794
MnO	0.120	0.099	0.064	0.082	0.183	0.233	0.000	0.108	0.198
MgO	0.027	0.056	0.000	0.000	0.000	0.000	0.000	0.000	0.050
CaO	23.671	23.364	23.579	23.878	23.749	24.556	23.836	23.592	15.722
Na ₂ O	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
K ₂ O	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Total	98.217	98.140	99.934	99.825	97.923	98.365	97.309	97.149	97.236
Number of ions on the basis of 12.5O									
Si	3.00	3.02	3.02	3.03	3.00	3.02	3.02	3.01	4.25
Al	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Al	2.43	2.43	2.18	2.39	2.15	2.27	2.13	2.16	1.27
Fe ³⁺	0.57	0.56	0.80	0.58	0.81	0.64	0.80	0.80	0.55
Ti	0.00	0.00	0.01	0.00	0.00	0.00	0.01	0.00	0.00
Mn	0.01	0.01	0.00	0.01	0.01	0.02	0.00	0.01	0.01
Mg	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.01
Ca	1.98	1.96	1.97	1.97	2.03	2.07	2.05	2.03	1.26
Na	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
K	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
TOTAL	8.00	7.98	7.98	7.98	8.01	8.02	8.01	8.01	7.34
Ps	19.0	18.7	26.8	19.6	27.4	22.0	27.3	27.1	30.3

Table C-7: Epidote electron-microprobe analyses, with stoichiometry based on 12.5O.

Section	Associated with Amp + carbonate					Associated with carbonate			Flaser in biotite		
	R12-27 (2-1) 1	R12-27 (2-1) 2	R12-27 (2-2) 1	R12-27 (2-2) 2	R12-27 (2-3) 1	R12-19 (1-3) 1	R12-19 (1-4) 1	R12-19 (1-4) 2	R7-104b (2-1) 2	R7-104b (2-1) 3	R7-104b (2-1) 4
Analysis											
SiO ₂	37.250	39.681	37.127	37.107	36.885	41.319	38.060	38.978	43.411	42.705	42.980
TiO ₂	0.649	0.057	0.183	0.083	0.078	0.077	0.000	0.000	0.489	0.249	0.213
Al ₂ O ₃	21.322	19.010	21.862	21.125	22.329	20.770	27.737	28.362	23.338	22.936	22.746
Fe ₂ O ₃	14.563	14.938	14.206	15.191	13.910	11.539	7.676	7.648	2.176	1.985	2.189
MnO	0.321	0.499	0.371	0.588	0.404	0.153	0.060	0.211	0.000	0.000	0.000
MgO	0.000	2.237	0.000	0.000	0.000	3.165	0.000	0.000	0.290	0.138	0.179
CaO	23.487	21.221	22.559	22.117	23.370	22.076	23.664	23.842	27.689	28.098	28.350
Na ₂ O	0.000	0.045	0.000	0.000	0.043	0.000	0.033	0.000	0.000	0.000	0.000
K ₂ O	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Total	97.592	97.688	96.308	96.211	97.019	99.099	97.230	99.041	97.393	96.111	96.657
Number of ions on the basis of 12.5O											
Si	3.00	3.17	3.02	3.03	2.98	3.21	2.99	3.00	3.36	3.36	3.37
Al	0.00	0.00	0.00	0.00	0.02	0.00	0.01	0.00	0.00	0.00	0.00
Al	2.03	1.79	2.10	2.03	2.11	1.90	2.55	2.57	2.13	2.13	2.10
Fe ³⁺	0.88	0.90	0.87	0.93	0.85	0.67	0.45	0.44	0.13	0.12	0.13
Ti	0.04	0.00	0.01	0.01	0.00	0.00	0.00	0.00	0.03	0.01	0.01
Mn	0.02	0.03	0.03	0.04	0.03	0.01	0.00	0.01	0.00	0.00	0.00
Mg	0.00	0.27	0.00	0.00	0.00	0.37	0.00	0.00	0.03	0.02	0.02
Ca	2.03	1.82	1.97	1.94	2.03	1.84	1.99	1.97	2.30	2.37	2.38
Na	0.00	0.01	0.00	0.00	0.01	0.00	0.01	0.00	0.00	0.00	0.00
K	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
TOTAL	8.00	7.99	7.99	7.98	8.03	8.00	8.01	7.99	7.98	8.00	8.01
Ps	30.4	33.4	29.3	31.5	28.5	26.2	15.0	14.7	5.6	5.2	5.8

Table C-7 (cont.): Epidote electron-microprobe analyses, with stoichiometry based on 12.5O.

Section	R6-284	R6-284	P13-39	P13-39	P13-39	P13-39	P13-39	P13-67a	P5-16	P5-16
Analysis	(4-1) 1	(4-1) 2	(2-1) 1	(2-1) 2	(2-2) 1	(2-2) 2	(2-3) 2	(3-1) 1	(1-1) 9	(1-1) 13
SiO ₂	35.327	35.382	36.05	35.2	33.832	35.771	35.776	34.682	35.617	35.766
TiO ₂	2.422	2.493	1.057	1.051	0.888	0.978	0.749	2.554	1.522	1.621
Al ₂ O ₃	15.952	15.873	16.522	17.085	16.176	16.264	17.233	14.927	17.712	18.331
FeO	22.174	21.779	24.15	25.051	25.027	24.352	24.399	21.347	18.989	18.698
MnO	0.09	0.096	0.284	0.29	0.334	0.333	0.263	0.23	0	0
MgO	9.828	9.607	8.832	8.658	8.82	8.828	8.806	10.975	11.615	11.552
CaO	0.055	0.052	0.083	0	0.064	0	0	0	0	0
Na ₂ O	0.107	0.109	0.122	0.046	0.051	0.104	0.065	0.07	0.307	0.319
K ₂ O	10.278	9.933	7.575	7.282	9.596	9.251	7.924	9.342	9.449	7.936
Total	96.233	95.324	94.675	94.663	94.788	95.881	95.215	94.127	95.211	94.223
Number of ions on the basis of 22O										
Si	5.45	5.49	5.60	5.49	5.39	5.56	5.54	5.45	5.43	5.44
Al	2.55	2.51	2.40	2.51	2.61	2.44	2.46	2.55	2.57	2.56
Total Z	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
Al	0.36	0.40	0.63	0.64	0.42	0.54	0.69	0.21	0.61	0.73
Ti	0.28	0.29	0.12	0.12	0.11	0.11	0.09	0.30	0.17	0.19
Fe ²⁺	2.86	2.83	3.14	3.27	3.33	3.16	3.16	2.80	2.42	2.38
Mn	0.01	0.01	0.04	0.04	0.05	0.04	0.03	0.03	0.00	0.00
Mg	2.26	2.22	2.05	2.01	2.09	2.04	2.03	2.57	2.64	2.62
Total Y	5.78	5.75	5.98	6.08	6.00	5.91	6.00	5.92	5.84	5.91
Ca	0.01	0.01	0.01	0.00	0.01	0.00	0.00	0.00	0.00	0.00
Na	0.03	0.03	0.04	0.01	0.02	0.03	0.02	0.02	0.09	0.09
K	2.02	1.97	1.50	1.45	1.95	1.83	1.57	1.87	1.84	1.54
Total X	2.07	2.01	1.55	1.46	1.98	1.87	1.59	1.89	1.93	1.63
TOTAL	15.84	15.76	15.53	15.54	15.97	15.77	15.59	15.81	15.77	15.55
Mg/(Mg+Fe ²⁺)	0.44	0.44	0.39	0.38	0.39	0.39	0.39	0.48	0.52	0.52
Ti#	0.05	0.05	0.02	0.02	0.02	0.02	0.01	0.05	0.03	0.03

Table C-8: Biotite electron-microprobe analyses, with stoichiometry based on 22O.

Section	P5-16	P5-16	P5-16	P5-16	P5-16	P5-16	P5-16	P5-16	P5-16	P5-16
Analysis	(1-1) 17	(1-1) 18	(1-1) 19	(1-1) 20	(1-1) 26	(1-1) 27	(1-1) 28	(1-1) 29	(1-1) 30	(1-1) 31
SiO₂	35.304	34.675	35.258	36.213	35.877	35.973	35.697	35.787	35.963	36.252
TiO₂	1.458	1.502	1.561	1.59	1.581	1.513	1.476	1.535	1.565	1.588
Al₂O₃	18.159	17.757	17.883	18.357	18.393	18.759	18.402	18.768	18.819	18.532
FeO	19.988	20.686	20.143	18.961	18.512	18.27	18.519	18.732	18.51	19.327
MnO	0	0.115	0.06	0.064	0	0	0	0	0	0
MgO	11.404	11.253	11.109	11.466	11.634	11.688	11.632	11.723	11.812	11.665
CaO	0.057	0.055	0.063	0.036	0	0	0	0	0	0
Na₂O	0.256	0.237	0.226	0.342	0.287	0.279	0.285	0.316	0.361	0.206
K₂O	9.249	8.425	8.505	8.453	7.757	7.873	9.096	9.77	9.659	8.818
Total	95.875	94.705	94.808	95.482	94.041	94.355	95.107	96.631	96.689	96.388
Number of ions on the basis of 22O										
Si	5.36	5.34	5.40	5.46	5.45	5.44	5.41	5.37	5.38	5.42
Al	2.64	2.66	2.60	2.54	2.55	2.56	2.59	2.63	2.62	2.58
Total Z	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
Al	0.61	0.56	0.63	0.71	0.75	0.79	0.70	0.68	0.69	0.69
Ti	0.17	0.17	0.18	0.18	0.18	0.17	0.17	0.17	0.18	0.18
Fe²⁺	2.54	2.66	2.58	2.39	2.35	2.31	2.35	2.35	2.31	2.42
Mn	0.00	0.02	0.01	0.01	0.00	0.00	0.00	0.00	0.00	0.00
Mg	2.58	2.58	2.54	2.57	2.64	2.64	2.63	2.62	2.63	2.60
Total Y	5.90	6.00	5.93	5.87	5.92	5.91	5.85	5.82	5.82	5.89
Ca	0.01	0.01	0.01	0.01	0.00	0.00	0.00	0.00	0.00	0.00
Na	0.08	0.07	0.07	0.10	0.08	0.08	0.08	0.09	0.10	0.06
K	1.79	1.66	1.66	1.62	1.50	1.52	1.76	1.87	1.84	1.68
Total X	1.88	1.74	1.74	1.73	1.59	1.60	1.84	1.96	1.95	1.74
TOTAL	15.78	15.74	15.67	15.60	15.51	15.51	15.69	15.78	15.76	15.63
Mg/(Mg+Fe²⁺)	0.50	0.49	0.50	0.52	0.53	0.53	0.53	0.53	0.53	0.52
Ti#	0.03	0.03	0.03	0.03	0.03	0.03	0.03	0.03	0.03	0.03

Table C-8: Biotite electron-microprobe analyses, with stoichiometry based on 22O.

Section	R7-104b	P6-43	P6-43	P6-43	P6-43	P6-43
Analysis	(2-1) 2	9.1	9.2	9.3	10.2	10.3
SiO ₂	30.271	37.287	37.482	37.011	36.844	37.428
TiO ₂	1.319	1.183	1.279	1.211	1.089	1.215
Al ₂ O ₃	15.996	18.374	18.178	18.211	18.527	18.137
FeO	28.005	16.639	16.102	15.994	16.067	16.332
MnO	0.157	0.047	0.075	0.063	0.004	0.039
MgO	10.055	13.781	13.658	13.529	13.927	13.72
CaO	0.101	0.018	0.018	0.009	0	0.019
Na ₂ O	0.067	0.506	0.518	0.558	0.528	0.506
K ₂ O	4.384	7.168	7.157	7.626	7.086	7.362
Total	90.355	95.101	94.466	94.212	94.07	94.758
Number of ions on the basis of 22O						
Si	5.03	5.53	5.57	5.54	5.50	5.56
Al	2.97	2.47	2.43	2.46	2.50	2.44
Total Z	8.00	8.00	8.00	8.00	8.00	8.00
Al	0.17	0.74	0.76	0.75	0.76	0.74
Ti	0.16	0.13	0.14	0.14	0.12	0.14
Fe ²⁺	3.89	2.06	2.00	2.00	2.01	2.03
Mn	0.02	0.01	0.01	0.01	0.00	0.00
Mg	2.49	3.04	3.03	3.02	3.10	3.04
Total Y	6.74	5.98	5.94	5.91	5.99	5.94
Ca	0.02	0.00	0.00	0.00	0.00	0.00
Na	0.02	0.15	0.15	0.16	0.15	0.15
K	0.93	1.36	1.36	1.46	1.35	1.40
Total X	0.97	1.50	1.51	1.62	1.50	1.54
TOTAL	15.71	15.49	15.45	15.53	15.50	15.49
Mg/(Mg+Fe ²⁺)	0.39	0.60	0.60	0.60	0.61	0.60
Ti#	0.02	0.02	0.02	0.02	0.02	0.02

Table C-8: Biotite electron-microprobe analyses, with stoichiometry based on 22O.

Section	P9-27	P9-27	P9-27	P6-43	P6-43
Analysis	(1-1) 1	(1-1) 2	(1-1) 3	1	3
SiO ₂	48.295	48.705	47.64	48.772	48.908
TiO ₂	0.014	0	0.034	0.009	0.016
Al ₂ O ₃	32.824	33.162	32.977	32.691	33.21
FeO	9.324	9.057	9.397	5.805	5.815
MnO	0.281	0.255	0.219	0.032	0.16
MgO	7.95	7.931	7.826	9.723	10.069
CaO	0.012	0.007	0.021	0	0.028
Na ₂ O	0.081	0.103	0.134	0.251	0.094
K ₂ O	0	0	0	0	0
Total	98.781	99.22	98.248	97.312	98.298
Number of ions on the basis of 18O					
Si	4.98	4.99	4.94	5.02	4.98
Ti	0.00	0.00	0.00	0.00	0.00
Al	3.99	4.00	4.03	3.96	3.99
Fe ²⁺	0.80	0.78	0.82	0.50	0.50
Mn	0.02	0.02	0.02	0.00	0.01
Mg	1.22	1.21	1.21	1.49	1.53
Ca	0.00	0.00	0.00	0.00	0.00
Na	0.02	0.02	0.03	0.05	0.02
K	0.00	0.00	0.00	0.00	0.00
TOTAL	11.03	11.02	11.05	11.03	11.03
Mg*=100 Mg/Mg + Fe	60.3	61.0	59.8	74.9	75.5

Table C-9: Cordierite electron-microprobe analyses, with stoichiometry based on 18O.

RETROGRADE CORDIERITE										
Section	R7-104b	R7-104b	R7-104b	R7-104b	R7-104b	R7-104b	R7-104b	R7-104b	R7-104b	R7-104b
Analysis	(1-1) 1	(1-1) 2	(1-1) 3	(1-1) 4	(1-1) 5	(1-1) 6	(1-1) 7	(1-1) 8	(1-1) 9	(1-1) 10
SiO ₂	25.00	24.48	24.62	24.78	25.38	25.32	25.61	25.12	25.20	25.42
TiO ₂	0.10	0.05	0.12	0.06	0.13	0.14	0.13	0.06	0.10	0.12
Al ₂ O ₃	20.54	20.46	20.24	20.15	20.43	20.06	21.00	21.09	21.14	21.75
Cr ₂ O ₃	0.00	0.00	0.02	0.00	0.03	0.01	0.00	0.00	0.02	0.03
FeO	27.90	27.88	28.08	28.37	27.31	27.46	27.82	28.08	27.71	27.84
MnO	0.25	0.22	0.16	0.22	0.26	0.16	0.20	0.17	0.23	0.22
MgO	13.08	13.21	13.18	13.28	13.65	13.75	14.01	13.64	14.01	13.82
CaO	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Na ₂ O	0.01	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00
K ₂ O	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Total	86.89	86.30	86.42	86.86	87.19	86.88	88.78	88.16	88.41	89.21
Number of ions on the basis of 28O, and assumed 16(OH)										
Si	5.40	5.34	5.36	5.38	5.45	5.46	5.40	5.35	5.34	5.33
Al ^{IV}	2.60	2.66	2.64	2.62	2.55	2.54	2.60	2.65	2.66	2.67
	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
Al ^{VI}	2.64	2.60	2.56	2.54	2.61	2.55	2.61	2.64	2.62	2.70
Cr	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01
Ti	0.02	0.01	0.02	0.01	0.02	0.02	0.02	0.01	0.02	0.02
Fe ²⁺	5.04	5.08	5.12	5.15	4.90	4.95	4.90	5.00	4.91	4.88
Mn	0.05	0.04	0.03	0.04	0.05	0.03	0.04	0.03	0.04	0.04
Mg	4.22	4.29	4.28	4.30	4.36	4.42	4.40	4.33	4.42	4.32
Ca	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Na	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
K	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
TOTAL	11.96	12.02	12.01	12.03	11.95	11.97	11.98	12.00	12.01	11.96
Mg*=(Mg/Mg+Fe)	45.5	45.8	45.6	45.5	47.1	47.2	47.3	46.4	47.4	47.0

Table C-10: Chlorite electron-microprobe analyses, with stoichiometry based on 28O, and assumed 16(OH).

REPLACING BIOTITE				
Section	P13-39	P13-39	P13-39	P13-39
Analysis	(3-1) 1	(3-1) 2	(3-1) 3	(3-1) 4
SiO ₂	24.37	24.58	24.16	24.38
TiO ₂	0.22	0.25	0.22	0.13
Al ₂ O ₃	14.70	15.32	14.85	14.68
Cr ₂ O ₃	0.04	0.04	0.00	0.07
FeO	40.03	39.90	40.38	40.16
MnO	0.16	0.18	0.16	0.20
MgO	7.39	7.56	7.36	7.33
CaO	0.05	0.07	0.03	0.10
Na ₂ O	0.01	0.07	0.00	0.03
K ₂ O	0.00	0.00	0.00	0.00
Total	86.96	87.97	87.17	87.07
Number of ions on the basis of 28O, and assumed 16(OH)				
Si	5.70	5.66	5.65	5.70
Al ^{IV}	2.30	2.34	2.35	2.30
	8.00	8.00	8.00	8.00
Al ^{VI}	1.74	1.82	1.74	1.74
Cr	0.01	0.01	0.00	0.01
Ti	0.04	0.04	0.04	0.02
Fe ²⁺	7.83	7.68	7.89	7.85
Mn	0.03	0.03	0.03	0.04
Mg	2.58	2.60	2.56	2.55
Ca	0.01	0.02	0.01	0.03
Na	0.01	0.03	0.00	0.01
K	0.00	0.00	0.00	0.00
TOTAL	12.24	12.23	12.27	12.26
Mg*=(Mg/Mg+Fe)	24.8	25.3	24.5	24.6

Table C-10: Chlorite electron-microprobe analyses, with stoichiometry based on 28O, and assumed 16(OH).

IN CONTACT WITH CHALCOPYRITE								WITH PYRRHOTITE	
Section	R12-51	R12-51	R12-51	R12-51	R12-51	R12-51	R12-51	R12-51	R12-51
Zn (wt%)	57.095	56.446	56.766	56.494	57.704	58.287	57.44	58.604	60.904
S (wt%)	34.693	34.533	34.699	33.982	34.024	34.025	34.025	34.074	33.754
Fe (wt%)	8.479	8.328	8.118	8.229	6.511	6.897	7.741	7.372	6.119
Total	100.267	99.307	99.583	98.705	98.239	99.209	99.206	100.05	100.777
FeS (mol%)	14.81	14.73	14.34	14.57	11.67	12.17	13.63	12.84	10.52
ZnS (mol%)	85.19	85.27	85.66	85.43	88.33	87.83	86.37	87.16	89.48
	100	100	100	100	100	100	100	100	100
IN CONTACT WITH PYRRHOTITE									
Section	R12-51	R12-51	R12-51	R12-51	R12-51	R12-51	R12-51	R12-51	R12-51
Zn (wt%)	56.643	58.426	57.939	58.868	57.97	58.443	56.482	59.118	57.274
S (wt%)	33.427	34.156	34.138	33.932	34.953	34.537	34.689	35.457	34.298
Fe (wt%)	7.816	7.185	6.951	6.814	7.456	7.724	8.053	7.103	7.824
Total	97.886	99.767	99.028	99.614	100.379	100.704	99.224	101.678	99.396
FeS (mol%)	13.91	12.59	12.32	11.93	13.09	13.40	14.30	12.33	13.79
ZnS (mol%)	86.09	87.41	87.68	88.07	86.91	86.60	85.70	87.67	86.21
	100	100	100	100	100	100	100	100	100

Table C-11: Electron-microprobe analyses of sphalerite grains, associated with pyrrhotite or chalcopyrite.

IN AMPHIBOLE ALTERATION										
Section	P22-70	P22-70	P22-70	P22-70	P22-70	P22-70	P22-70	R8-219	R8-219	R8-219
Analysis	(1-1) 1	(1-1) 2	(1-1) 3	(1-1) 4	(1-1) 5	(1-1) 6	(1-2) 1	(1-1) 1	(1-1) 2	(1-1) 3
Au	97.169	95.501	96.182	95.584	97.246	96.216	95.735	96.896	97.004	95.77
Ag	3.97	4.196	3.806	4.037	3.369	3.635	4.064	2.875	2.813	3.116
Total	101.139	99.697	99.988	99.621	100.615	99.851	99.799	99.771	99.817	98.886
Fineness	961	958	962	959	967	964	959	971	972	968
IN PYRRHOTITE					IN FRACTURES IN AMPHIBOLITE ALTERATION					
Section	R8-253	R8-253	R8-253	R8-253	P22-72	P22-72	P22-72	P22-72	P22-72	P22-72
Analysis	(1-1) 1	(1-1) 2	(1-1) 3	(1-1) 4	(1-1) 1	(1-1) 2	(1-1) 3	(1-1) 4	(1-1) 5	(1-1) 6
Au	72.781	74.201	70.079	69.879	93.529	93.733	94.02	93.896	92.712	93.498
Ag	26.662	26.095	28.834	29.325	6.36	6.57	5.354	6.961	7.341	7.07
Total	99.443	100.296	98.913	99.204	99.889	100.303	99.374	100.857	100.053	100.568
Fineness	732	740	708	704	936	934	946	931	927	930

Table C-12: Electron-microprobe analyses of gold grains and their associations.