

THE UNIVERSITY OF ALBERTA
THE SØFTESTAD IRON ORE DEPOSIT,
SOUTHERN NORWAY

by

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Abstract

The Sjøftestad iron deposit of southern Norway is found associated with hornblende gneisses that form part of the Nissedal supracrustal series. The deposit is for the most part a magnetite-titanhematite ore rich in apatite. Sulfide-rich patches, formed later than the oxides, are found associated with the ore.

Works done on samples collected from the area include: (i) thin and polished section study of ore and country rocks; (ii) microprobe analysis for Fe, Mg, and Ti of coexisting Ca-amphiboles and biotites to determine the distribution of these elements between the two mineral phases; (iii) analysis for Fe, Ti, Mg, Mn, Al and V of coexisting iron oxide phases; (iv) sulfur isotope measurements on pyrite-chalcopyrite pairs from sulfide-rich sections of the ore body.

Fe and Mg show an equilibrium trend in their distribution between the Ca-amphiboles and biotites analysed, while the distribution of titanium is found to be irregular. The magnetites of the ore, hornblende gneiss, amphibolites, and skarns are found to be poor in Ti, V and Mg, while the coexisting titanhematites exhibit a wide variation in their Ti content. The metagabbroic magnetites coexist with hemoilmenites, and show a distinct chemical difference compared to the above magnetites by having higher V and Al, and lower Mg contents. The reason and implications of all these findings are

discussed with the use of relevant data from the literature.

The sulfur isotope values for pyrites and chalcopyrites indicate that the sulfur falls within the magmatic-hydrothermal range. Values from two pyrite-chalcopyrite pairs indicate temperatures of recrystallisation of about 550°C and greater than 700°C.

Comparisons are made between the Sjøftestad deposit and the Gallivare iron ore body of northern Sweden, and evidences for a possible metasomatic replacement origin is discussed.

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CHAPTER I

INTRODUCTION

Location

The Sjøftestad iron ore deposit is located in the Precambrian of southern Norway, associated with rocks of the Nissedal supracrustal series. The mine itself is situated on a ridge east of Nisser Lake, about $\frac{1}{2}$ Km. south of Nissedal village at approximately $59^{\circ} 8'N$ latitude and $8^{\circ} 33'E$ longitude. The ore outcrops about 200m. above the lake shore, and generally strikes N-S, dipping about 40° eastwards (Kvien, 1961). Figure 1 shows its position within the supracrustals.

Previous Work

Published works on the actual deposit are either lacking or not easily available, but according to Mitchell (1967) the earliest description of the ore was given by Vogt in 1895, but he did not elaborate on the geology of the area. Aamo (1950) gave a brief outline of the history and production of the mine up to that time, and described a few features of the deposit. Kvien (1961) made a survey of the deposit and briefly described some of the surrounding amphibolites and gneisses. Mitchell (1967) conducted a study of the Nissedal supracrustal series, and gave a

brief description of the Sjøftestad ore.

Mine History, Size and Production

According to Aamo (1950), the Sjøftestad deposit was discovered in the 1860s, but there was no mining activity until 1914. The deposit, first owned by the Naes Jernverk Company, was sold to Sjøftesand Gruber in 1916.

Between 1914 to 1921 some ore was produced (data lacking on the exact tonnage); all mining activity ceased at the end of this period, and production didn't start again until April, 1939. A magnetic survey conducted in the summer of this year indicated the presence of another ore lens south of the three earlier known ones. Ore reserve was estimated to be about 100,000 tons, but by the end of 1949, 195,000 tons had been delivered. The yearly production in 1950 was 26,000 tons, which increased to 34,000 tons by 1957 (Vokes, 1960). Production continued until 1964, when the mine was again closed down.

Table 1 shows a few analyses of the average shipped ore between 1957 and 1964, and one analysis made by Vogt in 1889. The total Fe content of the shipped ore ranges from about 52% to 56%. The high P and Ca content, the low Ti content, and the presence of Zn in some of the analyses are noteworthy, and will be discussed at a later stage.

Saebø (1966) made the following estimate of the amounts of minerals of economic interest contained in the ore based on previous productions:

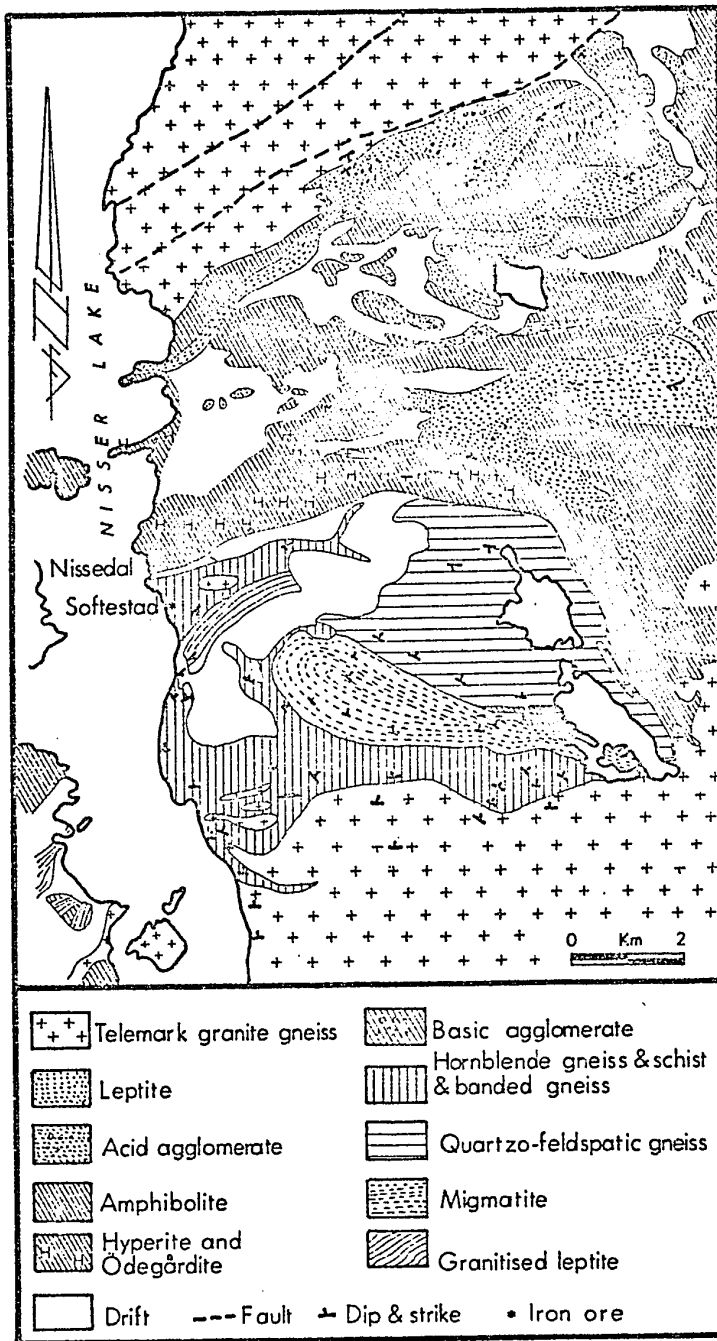


Fig. 1. Geological map of the Nissedal area. (From Mitchell, 1967)

TABLE 1: Analyses of Sjøftestad Ores*

	1	2	3	4	5
Fe ₂ O ₃	-	-	54.69	57.26	55.14
FeO	-	-	21.55	20.74	20.84
Total Fe	58.8	51.98	55.0	56.17	54.77
MnO	00.08	-	0.15	0.14	0.14
CaO	8.60	6.17	4.64	5.13	5.88
MgO	0.40	1.32	1.03	1.21	1.00
Al ₂ O ₃	Trace	3.51	2.70	1.8	1.9
SiO ₂	2.40	11.37	9.60	8.13	8.28
TiO ₂	-	-	-	0.46	0.42
P ₂ O ₅	?	?	2.75	2.81	3.44
P	2.50	1.29	1.20	1.23	1.50
As ₂ O ₅	-	-	0.08	0.00	0.00
As	-	-	0.05	0.00	0.00
ZnO	-	-	0.11	0.01	0.01
CO ₂	-	-	-	0.10	0.09
S	0.025	0.05	0.28	0.29	0.25

1. Vogt (1889)
2. Average shipped ore (1958)
3. Average shipped ore (1959)
4. Average shipped ore (May, 1964)
5. Average shipped ore (June, 1964)

*From R. Mitchell, personal notes.

Minerals of economic interest: Magnetite, Hematite, Apatite, Fluorite, Lombaardite, Beryl.

100,000 tons of ore would give:

50,000 tons/yr. 68% Fe

5,000 tons/yr. 90% Apatite

1,000 tons/yr. 90% Lombaardite

300 tons/yr. 95% Fluorite

? tons/yr. 95% Beryl

40,000 tons/yr. waste (mainly hornblende, with minor feldspar, sphene, mica, epidote, chlorite, zeolites, etc.). The total rare earths, determined gravimetrically by Dr. Alstad at the Kjernekjemisk Institute, University of Oslo, is: $RE_2O_3 = 0.17\%$ (but Ce negligible).

Geologic Setting and Age of Rocks

A. Introduction: The Precambrian of Norway is part of the great 'Baltic Shield' that makes up most of Sweden, the whole of Finland, and adjacent parts of Russia. Barth and Dons (1960) have given an extensive report on the general geology, structure, metallogenic provinces and relative ages of the Precambrian within Norway, and they describe the two main areas of occurrence: the southern part, shown in Fig. 2, and the northern part, of which no mention will be made.

B. The Southern Precambrian: This is broadly divisible into the Telemark area, the Bamble area and the Kongsberg

area as shown in Fig. 2. Absolute age determinations of various minerals and rocks give a common range of age of between 800 to 1100 m.yrs. (Barth and Dons, 1960).

The Telemark region makes up the largest part of the Precambrian of southern Norway, forming a large granitised area separated from the Kongsberg-Bamble area by major fault zones that show conspicuous brecciation and mylonitisation. A large part of the Telemark is occupied by non-granitised or slightly granitised supracrustal rocks surrounded by large areas of granites and granitic gneisses. The Nissedal supracrustals are one such series of rocks with which the Sjøftestad ore deposit is associated.

Barth and Dons (1960) state that no stratigraphic division has as yet been possible within the highly granitised sequence surrounding the supracrustals, but they recognize three groups in the supracrustals, with a total thickness exceeding 4,000 meters:

Bandak group (acid and basic lavas, quartz-rich sediments).

Unconformity

Seljord group (quartzites, conglomerates, schists).

Unconformity

Rjukan group (Vemork formation (basic lavas, sediments)

(Tuddal formation (acid lavas, tuffs).

(oldest)

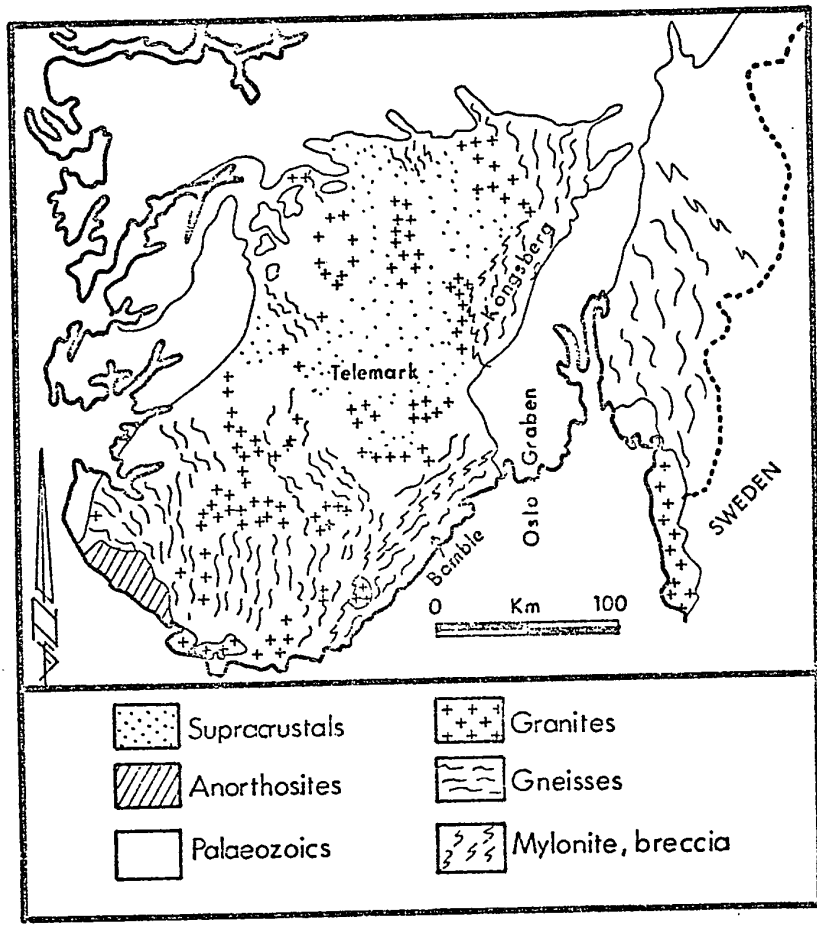


Fig. 2. The Southern Precambrian of Norway.
(From Barth and Dons, 1960)

C. The Nissedal Series: The stratigraphic position of the Nissedal supracrustals is uncertain (Mitchell, 1967). Barth and Reitan (1963) state that the Seljord quartzites are terminated to the south (i.e., some 20 km. north of the Nissedal area) by gneiss and granite, but that "metasediments reoccur in synclines within the gneiss granite, e.g. the Nissedal area on both sides of Lake Nisser." Mitchell (1967) noted the apparent complete lack of metasediments in the Nissedal series, stating that the only rocks that could be sedimentary in origin are quartzite blocks in the agglomerates, which could be derived from an equivalent of the Seljord quartzites. As will be discussed later, skarn mineral assemblages have been observed in several rock samples studied for this work from the Sjøfestad area, indicating that metasediments do form at least a small portion of the Nissedal rocks.

Stretching from the west shore of Nisser north-eastwards towards the south shore of Flavann, the Nissedal supracrustals cover an area of approximately 200 sq. km. The most important rock types are: amphibolites, hyperites and gabbroic rocks, leptites, ödegardites (hornblende-scapolite rocks), agglomerates (basic and acidic), quartzofeldspathic gneisses and schists, migmatites, hornblende schists and gneisses and banded gneiss (Mitchell, 1967). Their distribution is shown in Fig. 1 (Mitchell, 1967). The amphibolites are the most abundant rock types within

the region.

Three periods of deformation are recognized in the Precambrian of southern Norway (Wegmann, 1960): the first two periods involved folding, metamorphism and emplacement of intrusions at varying levels in the crust; the superposition of these two deformation periods produced complex structures. The third period produced large shear zones that dissected the whole recrystallised basement into great rhombohedral slices, with some later deformations and adjustments continuing into the Caledonian and Permian ages. These three periods of deformation are recognizable in the Nissedal rocks and within the ore body, (Mitchell, 1967), indicating an identical tectonic history for both the ore-deposit and the country rocks.

CHAPTER II

THE ORE DEPOSIT AND ASSOCIATED COUNTRY ROCKS

Introduction

The Sjøftestad ore lies within the so called "Sjøftestad gneiss" (Kvien, 1961), which is a quartz-biotite-hornblende gneiss having varying amounts of sphene, apatite and iron oxides as accessories. Mitchell (1967) concludes that the hornblende gneisses and schists of Sjøftestad represent amphibolites which have been slightly granitised, and indeed sections of the associated amphibolites studied resemble the hornblende gneisses in general texture and mineralogy; the hornblende gneisses show a more pronounced gneissosity and a slightly higher content of quartz and biotite.

With more intensive granitisation, the hornblende gneiss grades into granite gneiss to the south (Fig. 1), with much of the hornblende being converted to biotite, and an increase in the amount of quartz taking place. To the north, hyperites, metagabbros and ødegaardites occur in close association surrounded by amphibolites, and Mitchell (1967) states that an intrusive contact could not be observed between these two groups due to intense metamorphism and recrystallisation.

Ore Deposit and Configuration with
Respect to Wall Rocks

The field aspects of the Sjøftestad deposit have been described by Kvien (1961) and briefly by Mitchell (1967). The ore lies on the western limb of an easterly trending syncline, striking generally N-S and dipping 40° eastwards on the surface. Thin and tabular in shape, the ore has a sinuous outcrop, due to the effects of the two periods of folding.

The ore body is totally conformable with the enclosing gneiss, and for the most part contacts are sharp, but Kvien (1961) notes that in places a considerable evidence of replacement of the hanging wall is seen.

In the immediate vicinity of the mine, the ore could be considered to be made up of three "lenses," one in the northern sector, one in the central region and the third in the southern part. Considerable compositional variations are observed throughout the mine, but generally the northern lens is mainly composed of banded, fine grained magnetite and apatite, with very little hematite; the central lens is composed of a "brecciated," medium grained magnetite-hematite ore (60:40), while the southern lens is mainly composed of a striped hematite-magnetite ore which exhibits a gradual increase of hematite with depth. Kvien (1961) concludes that the ore of the northern lens is a "relict" ore which has been altered to the medium grained, brecciated ore of the type seen in the central lens.

An interesting aspect of the ore-zone lithology is the frequent occurrence of skarn rocks especially in the central banded magnetite-hematite sector of the mine. Kvien (1961) noted their presence, but gave no explanation regarding their origin. Several samples studied for this work, both from within the ore zone and country rocks, were found to be essentially skarn mineral assemblages. As will be discussed more fully in a later section, it is possible that the Sjøftestad deposit is a skarn type deposit formed by the metasomatic introduction of iron by fluids mobilised prior to or during the process of metamorphism and granitisation. This raises the possibility that at least a small portion of the original rocks that make up the supracrustals in the Sjøftestad region could have been calcareous sediments whose primary features have been destroyed by intense metasomatic activity and metamorphism. Such rocks provide suitable sites for the deposition of iron, and explain the presence of skarn rocks associated with the ore.

Younger lithologies of the ore zone are represented by pegmatite and lamprophyre dykes that cut the ore. Pegmatites are especially common in the southern part of the mine, the largest measuring 17 m. in thickness and striking N 15°E with a dip of 40° eastwards (Kvien, 1961). The lamprophyre dykes are the youngest rock type, cutting both ore and pegmatites. Kvien (1961) describes these as camptonites, but from the absence of olivine, the presence of

clinoamphibole (cummingtonite) and diopside as phenocrysts in thin sections of lamprophyre samples studied, this rock type is determined to be spessartite.

Finally, chlorite and calcite veins that parallel the ore zone in the footwall cut the pegmatites and the lamprophyre dykes, and appear to occupy minor faults (Kvien, 1961).

The Country Rocks

Thin sections of over 60 rock samples from the ore and adjacent country rocks, as well as from more distant locations within the supracrustals were studied for this work. These rock types include: biotite-hornblende gneisses, amphibolites, skarns, metagabbroic rocks, ödegardites, quartzo-feldspathic gneisses and lamprophyres. The general distribution of the main rock types within the Nissedal series is shown in Fig. 1.

A. Biotite-hornblende Gneiss and Amphibolites

As mentioned earlier, these two rock types have a very similar mineralogy, and differ only in texture and normative contents of the constituent minerals. The hornblende gneisses show a more pronounced gneissosity and slightly higher contents of quartz and biotite, grading into amphibolites with less granitisation (Table 2).

Texture: Hornblende gneiss, with varying amounts of biotite, forms the country rock immediately adjacent to the

ore deposit. Dark grey in colour, and generally fine grained, the rock exhibits a moderate to fairly well defined foliation due to the alignment and banding of biotite and hornblende porphyroblasts. The texture is generally crystalloblastic (Plate II(c)) with both plagioclase and oriented calcic amphibole grains having comparable sizes. For the amphibolites, the texture becomes more porphyroblastic, with coarser, sub-oriented and sub-idioblastic hornblende grains occurring in a finer grained plagioclase.

Mineralogy: The essential constituents of both rock types are plagioclase and calcic amphibole (mainly hornblende), with minor biotite, quartz and potash feldspar (in the hornblende gneiss) occurring in varying amounts. Sphene, epidote, apatite and iron ore form the important accessories. In ore impregnated samples, allanite forms rims around iron oxide grains.

Plagioclase forms the dominant mineral of the hornblende gneiss and amphibolite. In the hornblende gneiss, it occurs as interlocking, generally xenoblastic to sub-idioblastic grains; untwinned grains are encountered much more frequently than twinned ones. In the amphibolite samples, the plagioclase tends to be more idioblastic, and twinning is more frequent. In both rock types, twinning usually follows the simple Albite law, with Pericline twinning subordinate and Carlsbad-albite twinning rare. From

the maximum extinction angles of albite twins, the composition of the hornblende gneiss plagioclase ranges from albite to oligoclase ($An_5 - An_{16}$); most of the amphibolite plagioclases show pretty much the same range, but for two samples (nos. 268 and 270) compositions of An_{65} to An_{80} (calcic labradorite to bytownite) were recorded.

Saussuritisation is a common feature of the plagioclases in these rock types, and frequently blebs of epidote after plagioclase are seen in the altered sericitic mass, probably due to the effects of retrograde metamorphism. Undulatory extinction and bending of the twin planes due to the effects of strain are frequently seen features in the hornblende gneiss plagioclases.

Potash feldspar (microcline) occurs in minor amounts in most hornblende gneiss samples (rare in the amphibolites), and forms xenoblastic grains interlocked with the plagioclase grains. It also occurs as coarser prophyroblasts. Saussuritisation is less common than in the plagioclases, and in many cases the potash feldspar remains unaltered next to the saussuritised plagioclase. Evidence of plagioclase being replaced by potash feldspar is seen in some sections.

Hornblende forms the dominant ferromagnesian mineral, occurring as fine to medium-grained porphyroblasts. In the hornblende gneisses, it forms clusters, pods or discontinuous bands of oriented, xenoblastic grains, while in

the amphibolites it tends to be coarser, more idioblastic and less oriented giving the rock a poorly defined foliation. In rarer cases, it occurs as coarse, poikiloblastic grains in both rock types.

The calcic amphibole in some samples is markedly blue, indicating a more hastingsitic composition. In a few oxide-rich hornblende gneiss samples, ferroactinolite coexists with blue-green hornblende; such two calcic amphibole assemblages imply the existence of a solvus, provided the two amphibole phases are in mutual equilibrium (Ernst, 1968). In some sections of ore zone samples, hornblende is seen replacing biotite. Kvien (1969) has noted that away from the ore-body, the gneiss is richer in biotite compared with the wall rocks.

In most ore sections, the oxide phases grow in close association with the ferromagnesian minerals, and in some the iron oxide minerals are found replacing the ferromagnesians (Plate I(g)).

Biotite forms fine, oriented idioblasts disseminated throughout the hornblende gneiss; it is less abundant in the amphibolites, but occurs in the same manner. In both rock types, its normative content varies widely, being absent or rare in some samples and constituting up to 20% of the rock in others. Prehnitisation and chloritisation of the biotite grains are frequently observed (Plate I(a), (e)), indicating retrograde metamorphic effects.

In some hornblende gneiss samples, phlogopite (ZY - pale red brown, X-pale yellowish to colourless) has been observed in minor amounts.

Quartz is very minor to absent in the amphibolites, while it forms up to 8% of the total mineral constituents in the hornblende gneisses. It occurs as fine xenoblastic grains with the plagioclase.

Sphene is an important accessory in both hornblende gneisses and amphibolites, occurring in amounts of up to 5% in some samples. It occurs either as fine euhedral grains or forms rims around titanhematite grains (Plate I(g)), biotites and plagioclases (Plate I(h)). It is among the latest formed minerals in these rock types.

Apatite is another common accessory mineral, especially in the ore-impregnated samples. In the amphibolites, it occurs as individual grains scattered through the rock, while in the ore-rich hornblende gneisses, it forms short bands closely associated with the oxide bands (Plate I(d)).

Epidote is also a common accessory, occurring as fine, disseminated grains in both hornblende gneisses and amphibolites. It also occurs as fine, needle-like grains with sericite as an alteration product of plagioclase, and in some cases forms rims around iron oxide grains (Plate III(g)). In ore-rich hornblende gneiss samples, allanite frequently forms rims around oxide grains; it is possible that this mineral is lombardite, a rare-earth rich variety of allanite as indicated by Saebo (1966) (see Ch. I).

TABLE 2. Modal Analysis Figures for Hornblende-Gneisses and Amphibolites

Rock Type	Hornblende-gneiss							Amphibolite					
	22	213	241	255	287	168	171	180	268	270			
Sample No.	22	213	241	255	287	168	171	180	268	270			
Mineral% Plagioclase	25	30	42	35	42	58	55	49	42	45			
K-feldspar	15	5	8	2	-	-	-	<1	-	-			
Ca-amphibole	22	25	17	23	10	18	35	25	50	40			
Biotite	16	-	5	20	18	9	-	8	2	4			
Quartz	4	6	8	5	5	2	1	~2	<1	1			
Epidote group	2	2	1	4	<1	-	2		2	-			
Sphene	2	4	5	3	2	4	5	6		4			
Apatite	<1	4	2	<1	4	1	<<1	1	<1	<1			
Fe-oxide	13	22	13	9	18	8	2	8	3	3			

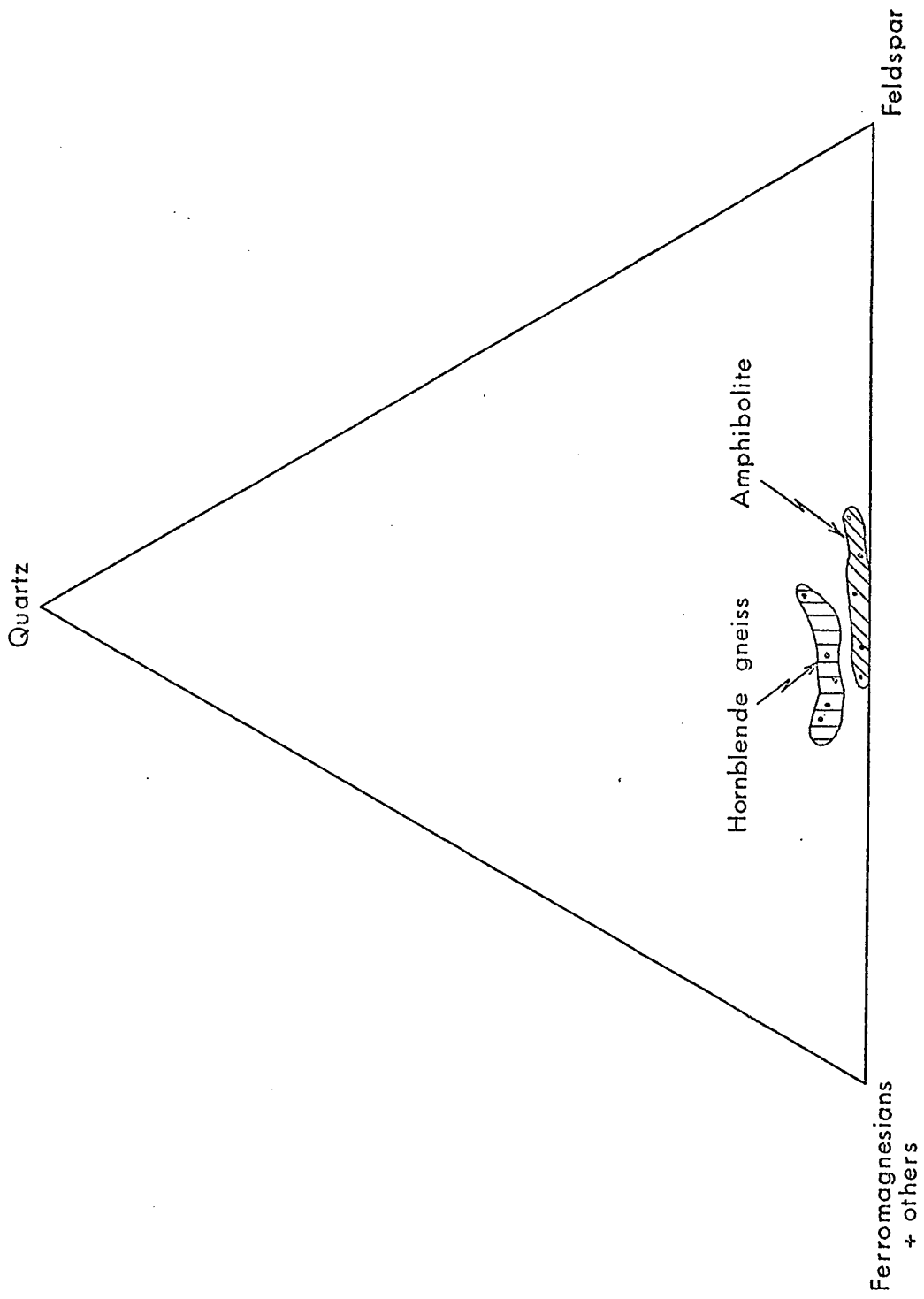


Fig. 3. Modal analysis of hornblende gneiss and amphibolite

Other less important accessories noted include: rare garnet grains in hornblende in one amphibolite sample (Plate I(f)); gypsum and anhydrite in one ore body sample (Plate I(g)), in which the very fine-grained gypsum appears as a secondary mineral filling in the space between the hornblende gneiss mineral grains.

Modal analysis figures and a modal analysis diagram for the amphibolites and hornblende-gneiss are given in Table 2 and Fig. 3.

B. Skarn Rocks

Two types of skarn are recognized in samples from the Sjøftestad area: andradite-bearing pyroxene skarn and pyroxene-amphibole skarn. Both types are well-foliated and rich in ore minerals (mainly titanhematite).

The andradite-pyroxene skarn has a porphyroblastic texture, with a wide variability in the grain sizes of the constituent minerals. The pyroxene is hedenbergite (R.I. > 1.75, positive, biaxial, $2V - 60^{\circ}-80^{\circ}$, $Z - 42^{\circ}$), which is markedly pleochroic (yellow-green). It occurs as fine to medium grained porphroblasts in a xenoblastic, saussuritised finer grained plagioclase. The plagioclase composition ranges from An_{25} to An_{35} (calcic oligoclase to sodic andesine); the albite twinning appears to be on the verge of being destroyed due to the general saussuritisation.

Andradite-rich garnet (determined from x-ray powder photographs) occurs both as fine idioblastic to sub-idioblastic grains and as coarse xenoblastic aggregates that form crude bands between oxide layers (Plate II(e), (h)). Many grains have small plagioclase cores, indicating that the garnet formed by replacing the altered plagioclase.

Epidote occurs both within the cores of garnet grains (Plate II(e)) and as an outer rim on other garnet grains; this either indicates a coincidental growth of the two minerals (both formed by replacing plagioclase), or that some epidote formed later by filling in vugs that are formed when the higher density garnet replaces the plagioclase.

Medium to coarse grained apatite occurs as bands closely associated with the iron oxides, in much the same way as the other ore-body samples. Allanite fills in cracks and inter-granular space of the apatite grains, and also grows along the grain boundaries of the oxide bands by replacing the oxides.

Calcite and quartz are rare minerals of this rock type; calcite in part forms by replacing some garnet. The ore oxides form crude bands that wind along the plagioclase grain boundaries, and almost appear to have formed by filling spaces between grains (Plate II(h)).

The pyroxene-amphibole skarn again exhibits a porphyroblastic texture. The principal mafic mineral is

hedenbergite, occurring as fine grained aggregates. Fine-grained laths of ferroactinolite frequently rim the hedenbergite grains, and have grown by replacing them (Plate III(a)).

Plagioclase is much less abundant in this type of skarn compared to the andradite-pyroxene skarn described above. It has undergone almost complete alteration, and twinning has generally been destroyed. Epidote has formed by replacing the altered plagioclase, and calcite occurs in minor amounts with the epidote. Apatite and allanite occur in much the same manner described above.

The iron oxides in this type of skarn are found closely associated with the ferromagnesian minerals, and at least in part they have formed by replacing them; they form fine to medium grained porphyroblasts, with their longest dimensions oriented parallel to those of the amphiboles.

C. Metagabbroic Rocks

These rock types are found distributed in the area north of the ore body, enclosing Nissedal village and extending north and north-east (Fig. 1). Unlike the hornblende gneisses and amphibolites, whose original features have been destroyed by metamorphism, the metagabbroic rocks exhibit textures indicative of the gabbroic nature of the parent rocks.

One of these is a corona texture, typical of metamorphosed gabbros (Plate III(f)). Clusters of fine,

7

xenoblastic hornblende grains, usually with central cores of ilmenite, are rimmed by radially arranged hornblende laths. The whole circular patch of hornblende grains is enclosed in coarse, sub-idioblastic laths of plagioclase of oligoclase-andesine composition.

In samples that have undergone more severe deformation, a symplectitic texture has developed, with very fine-grained untwinned plagioclase intimately intergrown with fine hornblende.

A high degree of alteration is observed in all the metagabbroic samples examined. Scapolitisation and saussuritisation of the plagioclase, and leaching of magnetite from composite grains of magnetite and ilmenite are commonly seen features. Skeletal grains of ilmenite are enclosed in fine hornblende and biotite, which themselves have five rims of untwinned secondary plagioclase (Plate III(e)). Saussuritised and scapolitised plagioclase forms the groundmass.

Altered pyroxene (hypersthene?) occurs in some amounts, and is being replaced by hornblende.

The approximate normative constituent of the metagabbroic minerals is, on the average: Plagioclase (50%), hornblende (35%), biotite (2%), pyroxene (3%), scapolite (5%) and Fe-ore (5%).

D. Other Rock Types

Although not immediately associated with the Sjøftestad ore, other rock types from the Nissedal area have been examined in order to obtain a better understanding of the varied lithologies of the supracrustals. These include quartz-feldspathic gneisses and ödegardites.

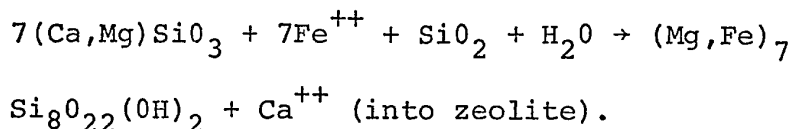
The quartzo-feldspathic gneisses exhibit a crystalloblastic texture (Plate III(d)), with plagioclase, microcline and quartz forming the dominant constituents. Quartz (15%) occurs both as coarser xenoblasts interlocked with the feldspars, and as finer, probably later-formed generation of grains in between the coarser mass.

Plagioclase (60%) is the principal feldspar, and is of oligoclase composition. It is for the most part untwined and occurs as anhedral grains with the quartz. Microcline (15%) occurs in the same manner as the plagioclase. Both feldspars show slight saussuritisation. The rock is given a schistose aspect by the alignment of fine grained biotite (7%) and hornblende (3%) laths.

Two samples of ödogardite (hornblende-scapolite rocks) have also been examined in the course of this work. This is a coarse grained rock composed of lenticular patches of hornblende (~40%) 2 to 4 cm. in length set in a greyish-white scapolite (~50%). Magnetite (10%) occurs as accessory. The scapolite has formed by extensively replacing original feldspar. Mitchell (1967) suggests that the

Nissedal ödogardites were formed by secondary pneumatolytic alteration of pegmatitic rocks.

In addition, samples from the lamprophyre dykes that cut both the ore and pegmatites have been examined. The texture of the lamprophyre is porphyritic (Plate III (g) and (h)), with fine grained anhedral laths of brown cummingtonite (biaxial, positive, $2V \sim 90^\circ$, length fast, extinction angle $\sim 14^\circ$) and anhedral colourless diopside (high $2V$, positive, biaxial, length slow, extinction angle $\sim 45^\circ$) in a fibrous zeolite plus possibly a little plagioclase ground mass. The zeolite is identified as thomsonite (biaxial, positive, length fast, birefringence ~ 0.007 , R.I. ~ 1.55). Some biotite and Fe-oxide occur as accessories. Diopside is being altered to cummingtonite in places, and the magnetite (?) appears as skeletal grains. The probable reaction can be represented as:



From the mineral associations, the lamprophyre is identified as spessartite rather than camptonite (Mitchell, 1967; Kvien, 1961), since olivine is absent and clinoamphibole and pyroxene occur as phenocrysts.

E. Metamorphic Grade of the Nissedal Rocks

The hornblende gneisses and amphibolites form the dominant rock types of the Nissedal supracrustals. The

important mineral associations in these is as follows:

Hornblende - Oligoclase - Biotite - Quartz

Hornblende - Oligoclase - Epidote (- ± Biotite)

Hornblende - Oligoclase - Biotite - Chlorite -
- Prehnite

Hence an amphibolite to epidote amphibolite facies of metamorphism is indicated. Chlorite and prehnite are the result of retrograde metamorphism. Mitchell (1967) states that the superposition of two metamorphic periods accounts for the occurrence of amphibolite grade rocks intermixed with rocks of epidote amphibolite grade. The author attributes the second period of metamorphism as causing the retrograde metamorphism of pre-existing amphibolite grade rocks, since the granitisation that accompanied this second period indicates temperatures of 500-600°C (as opposed to 550°- 750°C for the amphibolite facies).

The Iron Oxide Phase

The general mineralogy and texture of the three ore lenses in the immediate vicinity of the Sjøftestad mine has been briefly described in a previous section. A considerable compositional variation of the ore is observed throughout the mine with respect to the amounts of magnetite vs. titanhematite, and such variations are evident even in different sections of the same sample.

A. Ore Mineralogy and Texture of the
Ore Zone

The Sjøftestad oxide phase has a relatively simple mineralogy: magnetite and titanhematite, occurring in varying proportions, constitute the iron and titanium bearing ore minerals. The ore samples range from a totally magnetite to titanhematite-rich ores. Ilmenite is totally absent in all the examined sections of samples from the mine. Hercynite makes a rare appearance in the magnetites of a few samples, occurring as tiny spot or lens-shaped blebs (Plate VI(c)).

(Note: the term 'ore' is applied here to those samples in which the iron oxides constitute greater than 40% of the rock.)

Texturally, the ore can be classified into the following types:

(a) Layered, well-foliated magnetite-apatite or magnetite-apatite-titanhematite ore (Plate IV(a)), which, from Kvien's (1961) description of the mine, represents the type of ore encountered in the northern lens. The magnetite and titanhematite grains are generally flattened and elongated in the direction of the foliation, and quite frequently form coarse to medium sized bands whose direction is again parallel to the foliation direction in the rock. Magnetite predominates over titanhematite. Medium sized grains of apatite, with their longest dimensions oriented parallel to the banding, occur in close association with

the oxide grains. Other minerals that occur in between the oxide bands are plagioclase, hornblende, ferroactinolite, sphene, allanite and epidote, and rarely gypsum.

Polysynthetic twinning in the coarser titanhematite grains, especially those that make up the coarse bands, is frequently observed in this type of ore (Plate V(a), (b)). Such twinning is absent in the finer grained titanhematites that occur disseminated between the bands. This feature has been noted not only in the ore body samples, but also in ore-rich amphibolites and skarns. Edwards (1960) attributes such secondary twinning in some ore minerals to slight slipping movements of the order of one atom diameter during a deformation period, giving an apparent rotation of adjacent portions of the deformed grain through a 180° angle. In the case of titanhematite, such movements preferentially take place along the (0001) plane in the direction of $(10\bar{1}0)$ (Edwards, 1960); the shearing forces set up by the interference of crystal grains during growth becomes sufficient to induce twinning. The twin planes of many such grains are bent and broken in the ore body samples due to continuing stress after twinning.

The reason why secondary twinning is restricted to the coarse titanhematite grains that form the bands is not altogether clear. It may reflect a post-deformational growth of finer grained titanhematite in between earlier deformed bands. No twinning is observed in the magnetite grains, probably due to the greater hardness and more

brittle nature of magnetite which breaks or fractures during deformation rather than glide.

Replacement relationships between magnetite and titanhematite are hard to find in the ore body samples. Granular bands of magnetite coexist side by side with granular bands of titanhematite with no recognizable replacement features at the grain contacts. In a few samples martitisation of some magnetite grains is observed (Plate IX (g)); the alteration to hematite is most pronounced at the margins of the grains or along cracks, extending irregularly into the magnetite crystal. In the particular example given (Plate IV(g)), the TiO_2 content of the martite was found to be only 0.2 wt %, while other coexisting titanhematite grains contained up to 1 wt % TiO_2 in solid solution.

Edwards (1960) states that when hematite gradually replaces magnetite by oxidation, there comes a stage when no magnetite is left, but its previous presence is indicated by a "Widmanstätten" -like texture in the hematite. Such textures have not been observed in any of the titanhematite grains from Sjøftestad. It is possible that the titanhematites represent a primary oxide phase like the magnetite, though some hematite was formed by direct oxidation of magnetite.

The second type of texture is the granular type of ore texture, with fine to medium grained oxide grains showing little or no foliation (Plate IV(b),(c)). Both

titanhematite-rich and magnetite-rich ores show this type of texture which probably results from a more intense deformation and recrystallisation of the above type of layered texture. Such ores are representative of the central ore lens described by Kvien (1961). The grains are more rounded and finer grained than the above type, and tend to be equidimensional (Plate IV(b)); in some samples, the transition from the more foliated texture to the granular type texture is seen by the coexistence of finer, more rounded grains with coarser, tabular, oriented grains (Plate IV (c)).

It is notable that in this type of ore, lamellar twinning in the titanhematite grains is rare or absent, which is possibly an indication of the greater degree of recrystallisation that destroys such textures.

At the extreme end of the deformation process, magnetite ores exhibit a cataclastic texture (Plate IV(d)), due to the hardness of the magnetite grains which shatter into fine fragments under intense stress. No titanhematite ores with this type of texture have been observed; the reason probably lies in the fact that beyond a certain critical amount of deformation, less brittle minerals (like hematite) anneal and grow by a process of gradual absorption of some grains by others through boundary migration (Edwards, 1960).

B. Ore-Mineralogy and Texture of
the Country Rocks

Magnetite and titanhematite are abundant accessories of the hornblende gneisses, amphibolites and skarns. They normally occur in amounts of up to 10%, though in some ore-impregnated samples, the proportion rises to greater than 25% of the rock.

The iron oxides generally occur as fine to medium grained porphyroblasts, oriented parallel to the foliation in the same way as the ferromagnesian. In oxide-rich samples, the ore minerals form short wavy bands (as in Plate IV(e)). No ilmenite has been found in these rock types as in the ore samples.

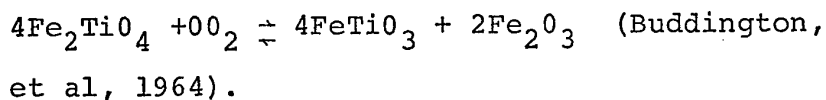
The metagabbroic samples, on the other hand, show a notable difference in ore mineralogy and texture compared to the above. Hemo-ilmenite is present in all samples of these rock types examined, occurring in one of three ways:

(i) As fine "exsolution" lamellae in magnetite (Plate V(c)), in which the hemo-ilmenites, elongated parallel to their (0001) directions, lie in the (111) directions of the magnetite (Friedrich, 1966). The origin of such textures in magnetites is attributed to the oxidation of exsolved ulvöspinel originally present in the magnetite (Buddington, et al, 1964), and will be discussed more in Ch. III.

(ii) As granular aggregates closely associated with magnetite grains that have hemo-ilmenite lamellae. The

lamellae are in optical continuity with the hemo-ilmenite grains in contact with the magnetite, and in many cases are connected to them (Plate VI(a)), indicating that the hemo-ilmenite grains have formed from the magnetite by an outward ionic diffusion to the magnetite grain borders.

(iii) As skeletal, magnetite-free grains (Plates V(d), III(e)), in which the sites occupied by the former magnetite of the above type (ii) now have amphiboles and biotites for the most part, and occasionally fine grains of goethite-rimmed pyrite. This is a common form of occurrence of the metagabbroic ilmenites, and the fact that magnetite has been extensively leached out from these rock types may have a significance with regards to a source for the iron deposit, as is discussed in Ch. IV. Titanhematites or hematite is absent in these rocks, but in one sample (no. 181, Plate V(e)) extremely fine, discontinuous lamellae of probably titanhematite occurs within the ilmenite. The lamellae disappear in the vicinity of the magnetite-ilmenite grain contact, and are the product of direct exsolution from an ilmenite hematite solid solution which formed by the oxidation of the original ulvöspinel component of the magnetite:



The Sulfide Phase

Sulfide-rich patches are reported to occur in certain sections of the ore-body (Kvien, 1961; Aamo, 1950). Pyrite,

chalcopyrite, bornite and chalcocite are the principal sulfides that occur with the ore, and Mitchell (1967) considers those to be primary minerals formed with the ore.

In the samples examined, bornite and chalcopyrite together with pyrite form the dominant sulfide minerals; chalcopyrite occurs as coarse granular patches with irregular boundaries in the bornite host. The two minerals form extensive solid solutions at temperatures above 475°C, at which temperature they are in a state of disorder (Edwards, 1960). Rapid cooling results in unmixing, with the development of crystallographic intergrowths of one mineral in the (111) plane of the other. However with slow cooling, the exsolution bodies diffuse rapidly to the grain boundaries of the host, and segregate as free grains, so that a granular texture results.

Brett (1961) states that in general, the more concentrated the initial solid solution, the less common are lamellae as the final exsolution product, since lamellae can result in less concentrated solid solutions experimentally cooled at a very slow rate. Hence at Sjøftestad, it is reasonable to assume that the initial concentration of the solid solution was high and cooled at a rate slow enough to produce the lenses or mutual boundary textures rather than lamellae.

Narrow, elongated strings of unconnected chalcopyrite grains form broad "rims" around sections of the bornite

host. This is the typical "rim texture" (Edwards, 1960) that results from the unmixing of chalcopyrite-bornite solid solutions. Locally, bornite is altered to covellite.

Pyrite forms well-shaped grains in close association with chalcopyrite (Plates V(g), (f)), and appears to have formed later than the chalcopyrite, it frequently occurs at the edge of chalcopyrite grains, and has grown inwards. Plate V(g) illustrates this well, in which the shape of the original outline of the chalcopyrite grain remains essentially unchanged as the pyrite grows inwards into the chalcopyrite. Often, there is a rim of goethite around pyrite grains.

In rare cases, chalcopyrite is seen to replace earlier formed pyrite. This appears to indicate more than one period of chalcopyrite formation.

Sulfide-oxide Relationship, and Paragenetic Sequence of Crystallisation

Magnetite occurs as well-shaped grains with the sulfides, but hematite is absent. From polished section studies, it appears that the sulfide phase was introduced after the crystallisation of the magnetite; this is well exemplified by Plate V(h), which shows bornite in contact with magnetite (to the bottom left, bornite-chalcopyrite unmixing is seen). Veins of bornite and covellite extend into magnetite from the bornite-magnetite contact, indicating the

bornite to be the later formed phase, filling in cracks and irregularities in the magnetite.

Chalcopyrite forms partial rims around magnetite grains, and like the bornite, extends into the cracks and broken edges. In the metagabbroic ilmenites, pyrite and hornblende fill in the space formerly occupied by magnetite.

Hence, the paragenetic sequence of ore mineral crystallisation at Sjøftestad appears to be:

- 1) Magnetite (+ titanhematite)
- 2) Ilmenite (in the metagabbroic oxide phase)
- 3) Chalcopyrite - bornite
- 4) Pyrite, goethite

CHAPTER III

ANALYTICAL RESULTS AND DISCUSSION

Introduction

After initial examination of the samples collected from the Sjøftestad area, about fifty were selected to represent the main mineral assemblages with which the Fe-Ti oxide phases were associated. Polished sections of these samples were prepared and grouped according to rock type, and major and trace element analyses were performed for coexisting oxide phases of about thirty samples, utilizing an electron probe microanalyser. Analysis for Fe, Ti and Mg was also performed on coexisting calcic amphiboles and biotites of some Sjøftestad rocks.

The results of the amphibole-biotite analysis are given in Table 3. Three of the samples (nos. 22, 40, 287) are ore-rich hornblende gneisses, one sample (255) is an amphibolite, sample 211 is a metagabbroic rock and sample 177 is an acid gneiss. Tables 4-7 show the results of the oxide analyses; thirteen of the samples listed in Table 4 come from various sections of the ore body, the country rock being hornblende gneiss. The amphibolite oxides are grouped in Table 5, while Table 6 shows the results of the skarn oxide analysis. The metagabbroic oxides are listed

in Table 7.

Sulfur isotope measurements were conducted on coexisting pyrite and chalcopyrite pairs, and the results are given in Table 8.

Distribution of Fe, Mg and Ti in
Coexisting Calcic-amphiboles and
Biotites of some Sjøftestad Rocks

Partial probe analyses for Fe, Mg and Ti was performed for coexisting biotites and hornblendes from three hornblende gneiss samples (nos. 22, 40, 287), one amphibolite (no. 255), one granitic gneiss (no. 177) and one hyperite (no. 211). The result of the analyses is shown in Table 3.

The geologic applications of elemental distribution ratios between coexisting minerals have been reviewed by McIntyre (1963) and Moxham (1965), and numerous works on major, minor and trace element distribution between coexisting silicate phases are found throughout the geologic literature of the past two decades. Works by Kretz (1959, 1960), Mueller (1960, 1961) and Ramberg (1952) have yielded much information about the equilibrium relationships of metamorphic minerals as expressed by their compositions and the chemical potentials of their constituent elements. These researchers have shown that:

(i) For a range of concentrations, if an element forms an ideal mixture in two coexisting mineral phases, the concentration of the element in one mineral plotted against

the concentration of the same element in the coexisting mineral defines a straight line of unit slope.

(ii) At dilute concentrations (i.e. for minor elements), the curve becomes a straight line passing through the origin, whether or not the element forms an ideal solution in one or both phases; the slope defines the distribution coefficient K_d .

(iii) Temperature is a major factor that controls the slope of the distribution curve, with pressure having a negligible effect. An element distribution is also dependent upon the ratios of other elements, especially the major elements, in the coexisting mineral phases.

(iv) The chemical potential of an element decreases with decreasing concentration. Small concentrations of an element should be expected to strive towards an equilibrium distribution. (It is assumed that at some very low concentration, material transfer of components becomes kinetically inhibited, and equilibrium will no longer be achieved [Moxham, 1965].)

Figures 4, 5 and 6 show the distribution of Fe, Mg and Ti respectively between coexisting calcic amphiboles and biotites of some Sjøftestad rocks. The three elements, together with Al, Mn, Cr, Li, etc., occupy the octahedral positions in both biotites and amphiboles, and are 6-coordinated.

TABLE 3. Microprobe Analytical Result of Coexisting
Ca-Amphiboles and Biotites

Wt. % Sample NO.	Fe		Mg		Ti		K		Fe/Fi+Mg+Ti		Mg/Mg+Fe+Ti		Ti/Ti+Mg+Fe	
	Ca- amph	Biot	Ca- amph	Biot	Ca- amph	Biot	Ca- amph	Biot	Ca- amph	Biot	Ca- amph	Biot	Ca- amph	Biot
22	9.95	8.96	9.61	11.46	0.11	0.55	0.58	8.21	0.506	0.427	0.489	0.547	0.006	0.026
40	7.57	7.15	10.76	12.67	0.06	0.60	0.38	7.90	0.412	0.350	0.585	0.621	0.003	0.029
255	11.22	9.44	8.09	11.10	0.18	0.58	1.24	8.14	0.575	0.447	0.415	0.525	0.009	0.028
287	10.61	9.36	7.93	10.21	0.17	0.60	0.05	8.32	0.565	0.464	0.423	0.506	0.009	0.030
211	14.32	9.16	11.86	11.46	0.05	0.67	0.07	6.53	0.546	0.430	0.452	0.538	0.002	0.032
177	15.55	16.20	5.77	6.95	0.44	1.19	0.89	7.63	0.715	0.666	0.265	0.284	0.020	0.049

A. Distribution of Iron

To show the distribution of iron between the two mineral phases, the ratio of the concentrations $Fe/Fe+Mg+Ti$ in calcic amphibole is plotted against the same ratio in biotite as seen in Fig. 4. Curve A defines the best fit line between all the plotted points, while curve B represents the trend of the iron distribution for five of the six analysed samples (without sample 177).

Sample 177 is from an acid gneiss, and stands out from all the other samples by having a substantially higher Fe concentration (as well as a relatively higher Ti and lower Mg contents) in both the calcic amphibole and biotite; it is the only sample in which secondary ilmenite occurs as tiny grains in close association with the ferromagnesians, but none with the Ti-poor magnetites as for sample 211. The ilmenite grains appear to have formed from the ferromagnesian phases rather than by the oxidation of the original ulvospinel component of the magnetite. As will also be shown later from the Ti distribution, the high Fe recorded for this sample appears to be due to contamination from sub-microscopic ilmenite phases within the ferromagnesians.

Kertz (1960) has shown that the plot of the iron distribution points for coexisting biotites and hornblendes from skarns is affected by the amounts of aluminum, sodium and potassium in the amphiboles. Those amphiboles having a relatively high Al, Na and K content show a curvilinear

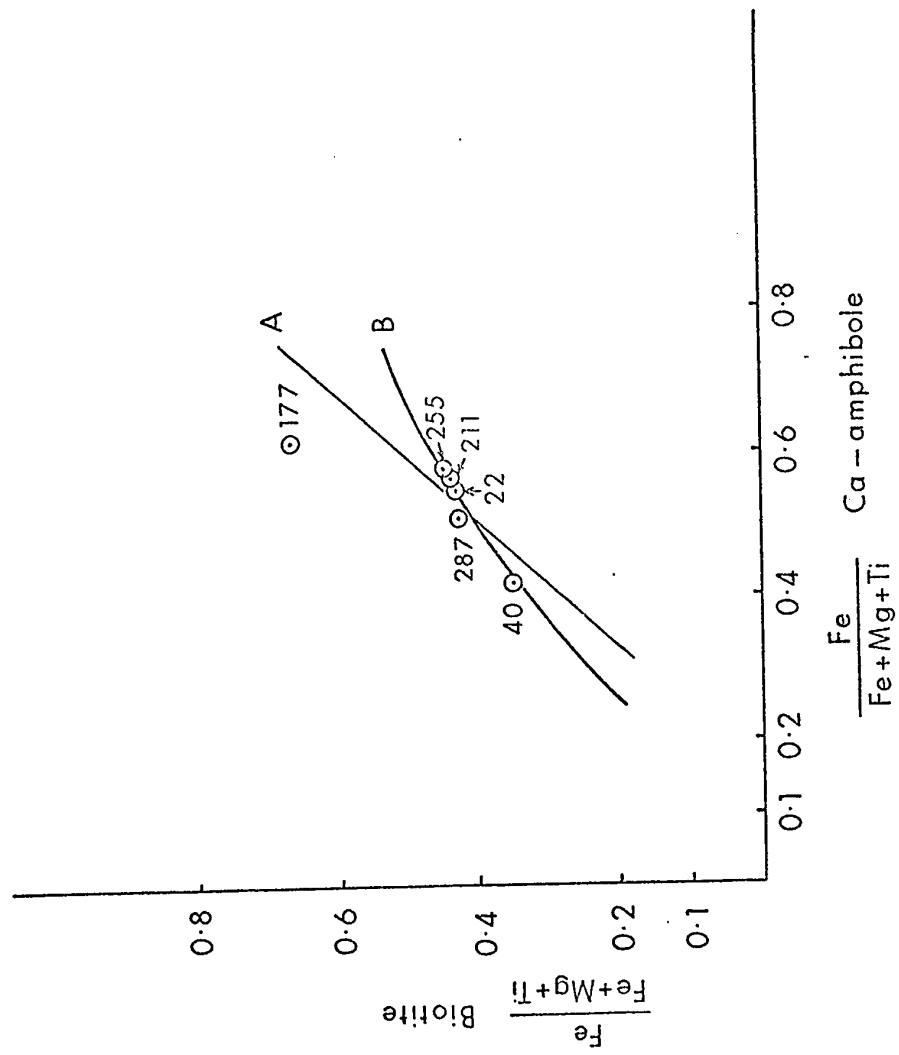


Fig. 4. Distribution of iron between coexisting Ca-amphiboles and biotites

relationship, with the ratio $Fe/Fe + Mg + Ti$ increasing at a faster rate than the same ratio in the coexisting biotite phase; the distribution curve is mainly dependent on the amount of aluminium rather than sodium and potassium, since the variation in the Al content of amphiboles is greater than that of Na or K (Kretz, 1960).

The curve B in Fig. 4 appears to indicate that such a non-linear relationship exists in Fe distribution between the calcic amphiboles and biotites of the Sjøftestad rocks. From Kretz's (1960) findings, one could expect the calcic amphiboles of the hornblende gneisses, amphibolites and hyperites to be relatively enriched in aluminium, so that they acquire a higher concentration of iron relative to the coexisting biotites. (The markedly blue colour exhibited by the calcic amphiboles in many hornblende gneisses and amphibolites indicate a more Al-rich, hastingsitic composition.) On the whole, an approach towards equilibrium is observed with respect to the distribution of iron between the calcic amphiboles and coexisting biotites at Sjøftestad.

B. Distribution of Magnesium

Table 3 shows that magnesium is generally slightly more enriched in the biotites than in the coexisting amphiboles. Together with Fe, it forms the dominant cation occupying the Y-site in both mineral phases.

For the Sjøftestad calcic amphiboles and biotites, magnesium values range from lows of 5.8 wt. % in the

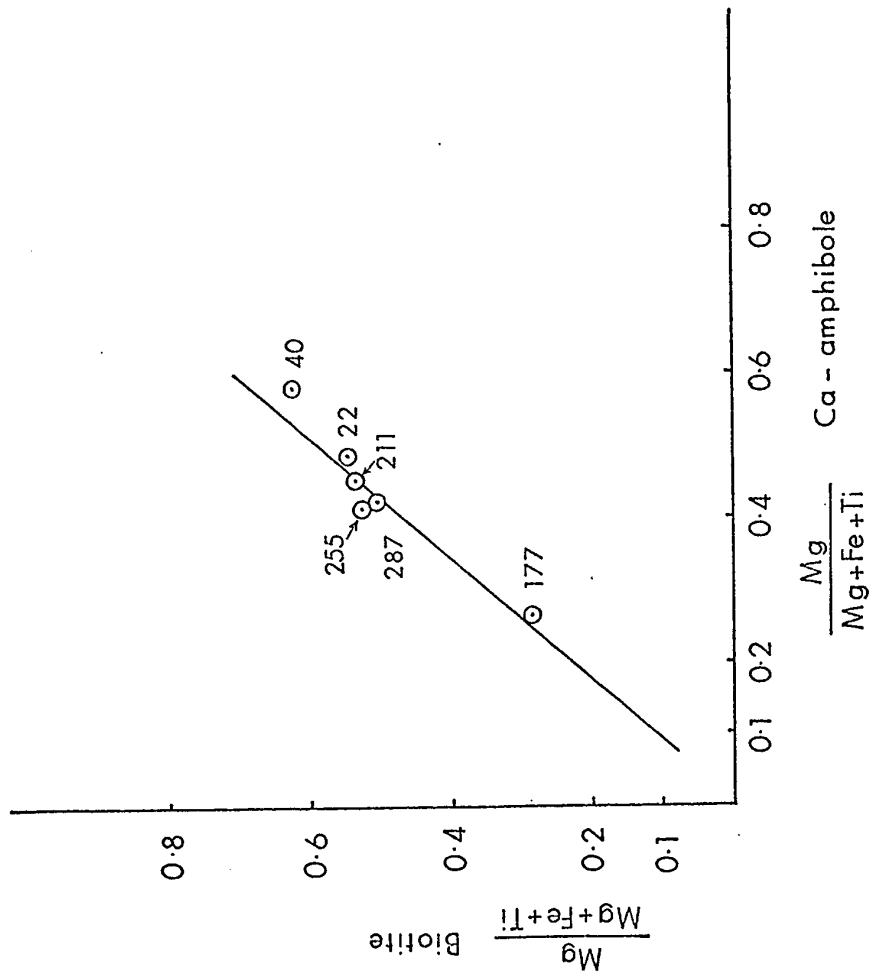


Fig. 5. Distribution of Magnesium between co-existing Ca-amphiboles and biotites

amphiboles and 7 wt. % in the biotites (sample 177), to high values of 11.8 wt. % in the amphiboles (sample 211) and 12.8 wt. % in the biotites (sample 40). Magnesium distributions are shown in Fig. 5, which shows the concentration ratio $Mg/Mg + Fe + Ti$ in Ca-amphibole plotted against the same ratio in biotite.

The distribution points lie close to a line having a slope of almost one and passing very close through the origin, indicating that equilibrium has been approached with respect to Mg distinction between the two coexisting mineral phases. It is notable that sample 177, whose Fe and Ti distribution points fall out of the general trend in Figs. 4 and 6 plots very near the equilibrium distribution curve in Fig. 6. The reason appears to be that whatever magnesium present in the ilmenite phase within the ferromagnesians (described above) is negligible compared to the magnesium present in the calcic amphiboles and biotites, and hence does not affect the distribution ratio too much.

C. Distribution of Titanium

From the analytical results of Table 3, it is seen that the titanium content of the Sjøftestad biotites is far greater than that of the coexisting calcic amphiboles. In the biotites of the hornblende gneisses and amphibolites (samples 22, 40, 255, 287), skarn (252) and hyperite (211), the range in the Ti-content is fairly narrow (0.54 to 0.68 wt. %); the calcic amphiboles show values ranging from less

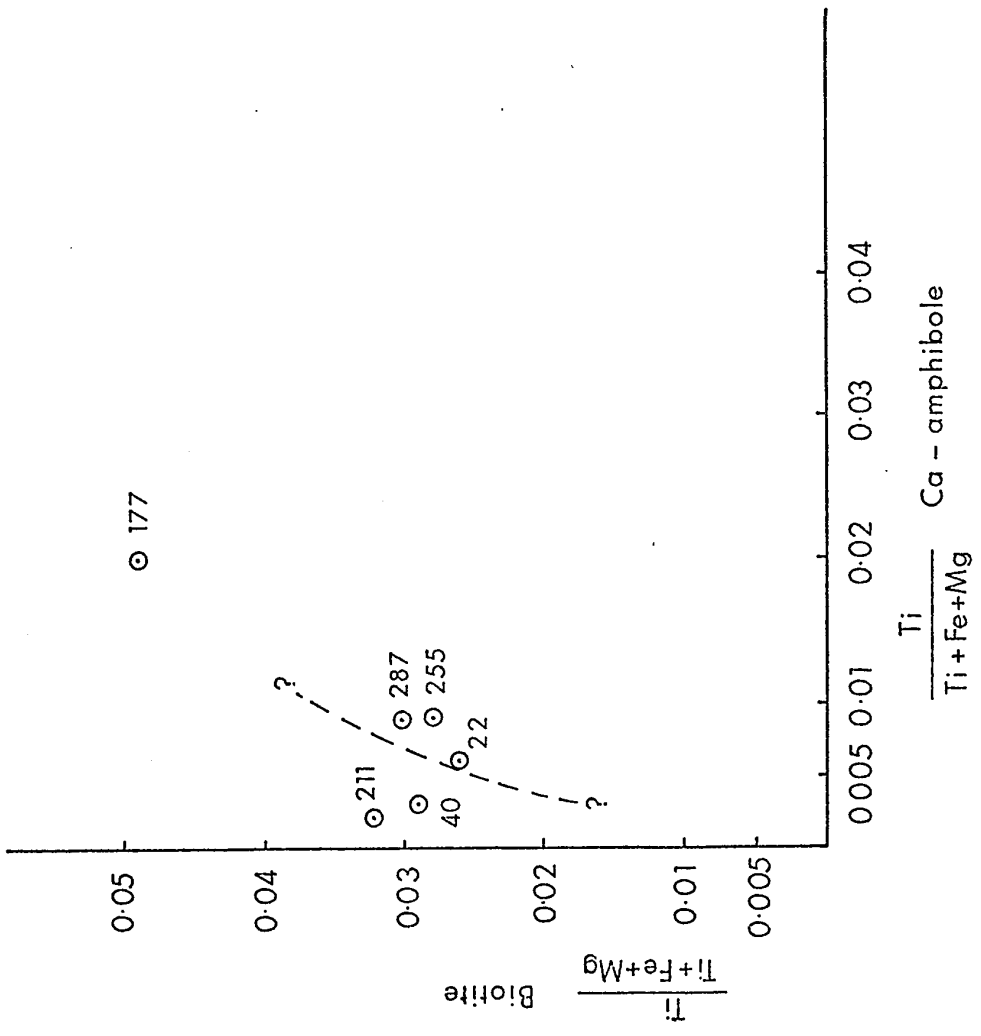


Fig. 6. Distribution of titanium between coexisting Ca-amphiboles and biotites

than 0.1 to about 0.2 wt. %.

The position occupied by titanium in the biotite structure has been subject to some debate. It appears likely that titanium is present both as Ti^{3+} (0.68\AA) replacing octahedrally bound Al^{3+} (0.51\AA), and Fe^{3+} (0.76\AA), and as Ti^{4+} replacing some Si^{4+} in the tetrahedral position in spite of the large ionic difference as suggested by Saxena (1966) (in Kwak, 1968), though no confirmation of this has been made. The content of titanium in biotite has been suggested to be directly dependent on the prevailing parameters (especially temperature) during crystallisation; Kwak (1968) found that the Ti content of biotites increases with increasing grade of metamorphism.

Titanium distribution between coexisting calcic amphiboles and biotites from Sjøftestad is shown in Fig. 6. A considerable scatter is observed, and there appears to be no pattern in the distribution plots. Similar situations have been encountered in titanium distributions between hornblendes and biotites by Kretz (1960), Annersten (1968) and to a lesser extent by Moxham (1965). Kretz (1960) partially attributes such scatter to be due to the aluminium, sodium and potassium contents of the calcic amphiboles, - i.e. an increase in the contents of these elements increases the affinity of amphiboles for both Fe and Ti. This does not appear to be a satisfactory explanation in this case, since the calcic amphiboles show a higher enrichment in

iron and a much lower enrichment in titanium compared to the coexisting biotites. A more likely explanation for the apparent non-equilibrium distribution with respect to titanium may be obtained from petrographic evidence: reaction rims of late-formed sphene around biotites are common features in many hornblende gneiss and amphibolite samples (e.g. Plate I(h)), and small secondary ilmenite grains are found closely associated with the ferromagnesian phases in some samples (e.g. sample 177). This could indicate that if chemical equilibrium was established with respect to titanium between the two mineral phases during crystallisation, it has not been retained due to subsequent reactions; in this case, the abundance of sphene in most of the samples examined in close association with the ferromagnesian and oxide phases (titanhematites) shows the oxidation of the titanium phase in these minerals to sphene. Moxham (1965) also suggests the existence of Ti as microlites of ilmenite or rutile within the ferromagnesian minerals as the most likely explanation for the scatter in the titanium distribution points between hornblendes and biotites. Sample 177 is a convenient example, which, as for the Fe distribution plot of Fig. 4, the titanium distribution falls out of the general plot area in Fig. 6 due to the presence of ilmenite microlites.

Oxide Analytical Results

Tables 4-7 show the results of the analyses on the iron-titanium oxide phase at Sjøftestad. Before a detailed discussion of the results of the analysis, a brief review of the Fe-Ti oxide phases in the literature is given, since the significance of the results are better understood by considering previous works on these mineral phases.

Initial surveys of the elements Ni, Co, Cu and Zn indicated that these elements are present in concentrations well below the detection limit of the electron probe microanalyser. The conditions of the analysis for the rest of the elements, together with an evaluation of the accuracy of the results are given in Appendix A.

Since microprobe analyses indicate only the total Fe concentration in a given mineral and cannot differentiate between Fe^{2+} and Fe^{3+} contents, it was necessary to recalculate the total Fe results in order to obtain the true chemical composition of the oxides. The procedure adopted here is that proposed by Rumble (1971), and the complete scheme is given in Appendix B.

A. The Fe-Ti Oxide Phases in the Literature

As noted by Abdullah (1964), knowledge of the $\text{FeO} - \text{Fe}_2\text{O}_3 - \text{TiO}_2$ system and the nature and conditions of formation of the Fe-Ti oxide phases has been derived solely from work on the high temperature igneous rocks, and little

attention has been paid to the role of these oxides in metamorphic rocks.

The above phases fall into two crystallographic types: the Rhombohedral Series (ilmenite-hematite) and the inverse-spinel type Magnetite Series (magnetite-ulvöspinel). Complete solid solution exists between magnetite and ulvöspinel on the one hand, and ilmenite and hematite on the other, while there is limited solubility at high temperatures between magnetite and hematite (Abdullah, 1964).

The principal phases in the $\text{FeO} - \text{Fe}_2\text{O}_3 - \text{TiO}_2$ system are: Fe_2O_3 (hematite), $\text{FeO}, \text{Fe}_2\text{O}_3$ (magnetite), $2\text{FeO} - \text{TiO}_2$ (ulvöspinel) and $\text{FeO} \cdot \text{TiO}_2$ (ilmenite).

The ilmenite-magnetite system was considered to be the least known and most controversial of all, since all attempts at homogenising the two phases below the melting point were unsuccessful (Lindsley, 1961). Hence most authors have concluded that the frequently encountered ilmenite lamellae in magnetite are the result of oxidation of original ulvöspinel in magnetite, rather than exsolution of ilmenite from an ilmenite-magnetite solid solution (Lindsley, 1961, 1962; Buddington, et al, 1963, 1964).

The fact that oxidation-reduction processes play a dominant role in the development of the various oxide phases have been stressed by several investigators (Lindsley, 1962; Buddington, et al, 1963, 1964; Vaasjoki, et al, 1963; Annersten, 1968). Hence the Fe-Ti oxide phases provide a useful oxygen barometer for the rocks in which they occur,

and in favourable cases, the compositions of magnetite - ulvöspinel ss¹ and those of the coexisting hematite - ilmenite ss are found to be unique functions of fO_2 and temperature (Lindsley, 1962, Buddington, et al, 1964).

In the literature, precise data is lacking on the equilibration of the Fe-Ti oxides under metamorphic conditions. Buddington, et al (1964) propose that under dry conditions of metamorphism, if the Fe-Ti oxides are the only significant iron-bearing minerals, the effective P_{O_2} is that of the oxide minerals and an exchange reaction takes place: the magnetite gains Fe_2TiO_4 and the ilmenite gains Fe_2O_3 . Under wet conditions when a water-rich fluid is in equilibrium with the oxides, the fluid becomes more reducing relative to the oxides upon heating, and again the magnetite becomes enriched in Fe_2TiO_4 . Upon cooling, ilmenite is "exolved" in the same manner as for primary igneous rocks. If the maximum temperature of metamorphism is lower than that reflected in the primary minerals and there is no great change in the redox conditions, the magnetite becomes depleted in TiO_2 reflecting the new conditions. The authors note that the oxide minerals reflect the conditions of metamorphism only if they have undergone complete recrystallisation.

¹ss = solid solution

TABLE 4. Analytical Result of Ore-Body Oxides
in Hornblende Gneiss

Sample NO.	2		11		22		38		40		47		54	
	Mt	Hem	Mt	Hem	Mt	Hem	Mt	Hem	Mt	Hem	Mt	Hem	Mt	Hem
TiO ₂	Tr	1.04	0.01	2.65	0.01	8.43	Tr	0.80	0.04	9.66	0.02	0.52	0.01	0.19
FeO*	28.85	1.62	28.67	2.39	28.81	7.16	30.07	1.55	30.24	6.78	28.97	1.14	30.02	0.84
Fe ₂ O ₃ *	70.61	96.67	70.53	94.36	70.67	83.98	69.39	97.22	69.79	81.33	70.05	97.89	69.27	98.56
MnO	0.23	-	0.39	0.03	0.19	0.04	0.19	-	0.15	1.30	0.42	-	0.25	0
MgO	0.03	0.02	0.02	0.02	0.03	0.04	0.03	0.02	0.02	0.04	0.06	0.03	0.08	0.01
Al ₂ O ₃	0.08	0.34	0.13	0.21	0.12	0.10	0.17	0.20	0.31	0.15	0.33	0.21	0.23	0.16
V ₂ O ₃	0.19	0.31	0.25	0.35	0.16	0.25	0.15	0.21	0.24	0.24	0.15	0.22	0.15	0.24

* Total Fe recalculated to Fe²⁺ and Fe³⁺ (See Appendix B).

Mt = Magnetite

Hem = "Hematite" (including ilmeno-hematites)

TABLE 4 (continued)

Sample No.	213		235		241		215*		248*		287
	Mt		Mt	Hem	Mt	Hem	Mt		Mt		Hem
TiO ₂	0.07		-	0.24	0.01	4.03	0.01		0.02		7.48
FeO*	30.09		30.04	0.63	30.17	3.76	30.04		30.60		6.34
Fe ₂ O ₃ *	69.43		69.46	98.67	69.30	91.63	69.33		68.57		85.46
MnO	0.13		0.15	-	0.12	-	0.22		0.16		0.13
MgO	0.03		0.02	0.02	0.02	0.02	0.06		0.03		0.06
Al ₂ O ₃	0.20		0.15	0.18	0.12	0.13	0.19		0.19		0.24
V ₂ O ₃	0.15		0.19	0.27	0.26	0.44	0.16		0.22		0.28

* Total Fe recalculated to Fe²⁺ and Fe³⁺ (See Appendix B).

TABLE 5. Analytical Result of Oxides from Amphibolites

Sample No.	265		296		264	360	255	267
	Mt	Hem	Mt	Hem				
TiO ₂	0.01	0.41	-	0.07	7.12	2.65	6.10	0.09
FeO*	29.04	0.86	29.85	0.43	6.20	2.49	5.47	30.18
Fe ₂ O ₃ *	70.23	98.14	69.51	99.16	86.17	94.44	87.82	69.28
MnO	0.29	-	0.25	-	-	-	0.02	0.14
MgO	0.02	0.02	0.07	0.01	0.04	0.07	0.05	0.04
Al ₂ O ₃	0.15	0.24	0.22	0.17	0.21	0.17	0.29	0.11
V ₂ O ₃	0.24	0.33	0.10	0.16	0.27	0.18	0.26	0.16

* Total Fe recalculated to Fe²⁺ and Fe³⁺ (See Appendix B).

TABLE 6. Analytical Result of Skarn Oxides

Sample No.	138		141	251	252
	Mt	Hem			
Wt. %					
TiO ₂	-	0.62	0.63	-	6.28
FeO*	29.01	1.30	1.22	30.17	5.39
Fe ₂ O ₃ *	70.17	97.43	97.66	69.49	87.74
MnO	0.39	0.01	0.02	0.09	0.06
MgO	0.02	0.01	0.02	0.01	0.06
Al ₂ O ₃	0.41	0.34	0.34	0.09	0.22
V ₂ O ₃	0.23	0.31	0.23	0.15	0.27

* Total Fe recalculated to Fe²⁺ and Fe³⁺ (See Appendix B).

TABLE 7. Analytical Result of Metagabbroic Oxides

Sample No.	209		211		181		145 ¹
	Mt	Ilm	Mt	Ilm	Mt	Ilm	Ilm
TiO ₂	0.07	49.17	0.03	49.46	0.11	44.62	49.36
FeO*	29.64	38.11	30.02	36.77	29.83	34.65	37.02
Fe ₂ O ₃ *	68.98	10.18	67.83	9.91	68.85	18.35	10.31
MnO	0.04	1.61	0.21	1.55	0.05	1.56	2.63
MgO	0.07	0.16	0.08	0.69	0.02	0.09	0.10
Al ₂ O ₃	0.69	0.09	0.74	0.07	0.26	0.06	0.02
V ₂ O ₃	0.52	0.68	1.54	1.09	0.48	0.67	0.55

* Total Fe recalculated to Fe²⁺ and Fe³⁺ (See Appendix B).

¹ Only ilmenite present

B. Discussion of Oxide Analytical Results

As noted in a past section, there is considerable variation in the normative amounts of magnetite vs. coexisting titanhematite, even in different sections of the same sample, for most of the ore-samples studied. The samples vary from totally magnetite to almost wholly titanhematite ores, and no chemical correlation could be established in the compositions of the rhombohedral phases coexisting with magnetite or occurring without magnetite.

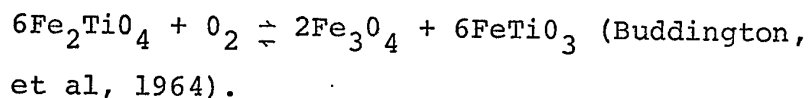
Tables 4 to 7 show the microprobe analytical results of four groups of selected samples. A discussion on the elemental variations in oxide phases within each group and with respect to other groups follow.

(i) Titanium

In magnetite, Ti occurs either as Fe_2TiO_4 (ulvöspinel) in solid solution with the magnetite, or as ilmenite lamellae in the magnetite host, while in the hematites Ti is present as ilmenite in solid solution.

As the magnetite analyses in all the tables show, the Ti content of the Sjøftestad ore magnetites as well as those of the associated country rocks is very low, being less than 0.1% in almost every case. This indicates that there is very little ulvöspinel left in solid solution with the magnetites. The magnetites of Table 7 (from metagabbros and hyperites) coexist with hemo-ilmenite, which occur

either as finely exolved lamellae in magnetite or as granular and skeletal aggregates, while those of Tables 4 to 6 do not have any associated hemo-ilmenites. The hemo-ilmenites of Table 7 are the product of oxidation of an original magnetite-ulvöspinel ss, whereby the ulvöspinel component is oxidised to ilmenite in the following manner.

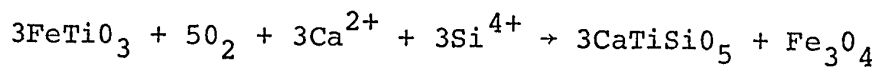


A high oxygen pressure is indicated for these samples, since the magnetites are almost devoid of ulvöspinel component.

In the other groups of samples, however (Tables 4 to 6), which have again very little ulvöspinel in solid solution with the magnetite, hemo-ilmenite or ferrian-ilmenite is absent either as lamellae in magnetite or as separate grains. It appears that either:

1) Any original titanium present has diffused out of the magnetite grains in response to complete recrystallisation (Buddington, et al, 1964), and incorporated into other coexisting phases (e.g. titanhematites, ferromagnesian), or

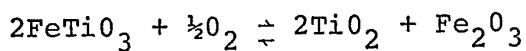
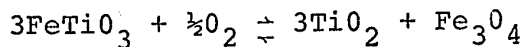
2) A high degree of oxidation has occurred, whereby any ilmenite that may have formed by the oxidation of original was in turn oxidised to magnetite + sphene according to the following relationship:



(Magnetite grains with sphene rims have been observed.) A third possibility:

3) The magnetites originally crystallised as Ti-poor magnetites. However, it is unlikely that the Ti-content of the original magnetites would be as low as the analyses show, since sphene does form reaction rims around some magnetite grains and titanium values this low are not recorded for natural magnetites unless they have undergone a high degree of oxidation.

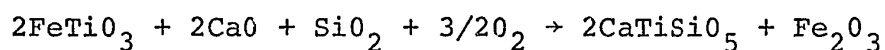
The absence of rutile associated with the Fe-Ti oxide phases in any of the samples examined shows that at Sjøftestad, ilmenite did not oxidise to rutite + Fe_3O_4 or rutite + Fe_2O_3 according to the relationships given by Annerston (1968) for the Gällivare region of northern Sweden.



It should be noted that the Ti content of magnetites is not an absolute measure of PO_2 unless ilmenite is present (Abdullah, 1964; Annersten, 1968).

The titanium content of the titanhematites shows a considerable variation, ranging from about 0.2 wt. % TiO_2 (sample 54) to almost 10 wt. % TiO_2 (sample 40) within the ore-body (Table 4). Similar variations are observed in the skarn associated oxides of Table 6 (0.6 to 6.3 wt. % TiO_2), and the amphibolite titanhematites of Table 5 (<0.1 to 7.1 wt. % TiO_2). In all the titanhematites, titanium is present

as FeTiO_3 (ilmenite) in solid solution with Fe_2O_3 , and no exolved ilmenite is observed. Sphene is a characteristic mineral in most of the ore-samples, forming rims around titanhematite grains (Plate I(g) and (h)), indicating that some of the ilmenite in solid solution with the hematite has been oxidised, combining with calcium and silica (either from solutions or from plagioclase and ferromagnesian):



The above reaction implies that the Ti content of the titanhematites with sphene rims was higher than that shown by the present analyses.

The reason for such wide variations in the TiO_2 content of the titanhematites, not only within the country rock samples but also within different sections of the ore-body, is almost certainly due to local inhomogeneities of oxygen fugacity. This may be caused by several factors, including different periods of formation and subsolidus late stage oxidation or reduction due to influx of water from surrounding rocks (Kretschmar, et al, 1971). Most titanhematites show no visible replacement relationships with the coexisting magnetites, but some did form by replacing magnetites (Plate IV(g)). As was observed for one particular sample in the course of the probe analysis (sample 215), it is possible that the hematites having low titanium contents are those that formed by subsolidus oxidation

of magnetite; but this could not be confirmed in other samples since evidences for replacement could not be established.

(ii) Manganese

The ionic radius of Mn^{2+} is 0.80 Å, close to that of Fe^{2+} (0.74 Å), and hence the two ions are mutually replaceable. With a similar electronegativity to Fe^{2+} , manganese occurs in solid solution in minerals containing ferrous ion (Mason, 1966). In the rhombohedral phase, it can occur as pyrophanite ($MnTiO_3$) which is isomorphous with ilmenite ($FeTiO_3$) and geikielite ($MgTiO_3$) (Friedrich, 1966).

As the analytical results in the tables show, manganese is enriched in the magnetites rather than the coexisting titanhematites as would be expected. Magnetite samples from the ore-body show a range of 0.1 to 0.4 wt. % MnO , and similar ranges are observed in the skarn samples and amphibolites. The titanhematites show only trace amounts of manganese, with the notable exception of sample no. 40 (Table 4), which contains almost 10 wt. % TiO_2 and 1.3 wt. % MnO dissolved in the hematite. Manganese in this case is likely present as pyrophanite ($MnTiO_3$) in the same way as ilmenite in solid solution. However other samples having similar high TiO_2 contents (samples 22, 287 from Table 4, 252 from Table 6, 264, 255 from Table 5) don't show similar high manganese contents.

In the metagabbros and hyperites (Table 7), the hemo-ilmenites show higher MnO contents, ranging from 1.5 to 2.6 wt. %. In these too, manganese is present as $MnTiO_3$ in solid solution with ilmenite.

Temperature has been recognized as a major factor in controlling the distribution of MnO between coexisting magnetite and ilmenite (Buddington, et al, 1964). The metagabbros and hyperites of Sjøftestad show the following ranges in the ratio of MnO in hemo-ilmenites vs. coexisting magnetites:

Sample #	Wt.% MnO "ilm" ¹	Wt.% MnO Mt ²	Ratio MnO "Ilm": Mt
209	1.61	0.04	>40
211	0.21	1.55	~7.4
181	0.05	1.56	>30

Buddington, et al (1964) found that the ratios of MnO in ilmenite: magnetite are low (1-5 - 5) for oxides associated with diabase gabbros and anorthosites, intermediate (5 - 5.5) for quartz-bearing syenites and granites, and high (5 - 15 or greater) for high grade metamorphic rocks and some granite pegmatites. Estimated temperature ranges for these three groups are found to be 700 - 890°C for the gabbros and associated rocks, 630 - 675°C for the granitic rocks, and 550 - 635°C for the metamorphic rocks.

¹"ilm" refers to hemo-ilmenite.

²Mt = magnetite

In accordance with this data, the Sjøftestad metagabbros and hyperites, which all show the high ratios of MnO in the hemo-ilmenites to the magnetites, probably experienced temperatures similar to those indicated for the metamorphic ores cited by Buddington, et al (1964).

(iii) Magnesium

On the basis of its ionic radius ($0.66\overset{\circ}{\text{Å}}$), magnesium could be expected to proxy for Fe^{3+} , but its electronegativity is lower (1.2 compared to 1.8 for Fe^{3+}) which greatly limits its exchange capacity with Fe^{3+} (Mason, 1966). It replaces Fe^{2+} instead, which has a lower electronegativity than Fe^{3+} and occurs as magnesioferrite ($\text{MgFe}_2^{3+}\text{O}_4$) with the rhombohedral phases (isomorphous with ilmenite and pyrophanite).

Almost without exception, magnesium is present in only trace amounts in both the magnetites and titanhematites of all groups (<0.1 wt. %). It shows slightly higher concentrations in the hemo-ilmenites of Table 7, which show an average concentration of 0.1 wt. % except sample 211, which contains about 0.7 wt. % MgO. In this latter case, the hemo-ilmenite can be considered a mixed oxide phase composed mainly of FeTiO_3 and Fe_2O_3 , with appreciable amounts of the other two members of the isomorphous series: MnTiO_3 and MgTiO_3 (pyrophanite and geikielite respectively).

(iv) Aluminium

In the Fe-Ti oxide phases, aluminium is present as the Al-spinel hercynite (FeAl_2O_4), substituting for Fe^{3+} . Within the ore-body magnetites, the concentration of Al_2O_3 ranges from <0.1 to >0.3 wt. %, averaging about 0.2 wt. %. Comparable concentrations are obtained for the coexisting titanhematites, but there is a general slight tendency for aluminium to be enriched in the titanhematites than the coexisting magnetites. The amphibolites and skarns show similar variations of aluminium in their oxides. It is the magnetites of Table 7 (metagabbros and hyperites) that show a distinct difference, with Al_2O_3 concentrations of up to 0.74 wt. % in the magnetites.

In order to determine the variations (if any) of aluminium with respect to hemo-ilmenite lamellae in the magnetite host, a qualitative electron beam scan was performed simultaneously for Al, Fe and Ti on a magnetite grain similar to Plate V(c). The result of the scan is shown in Fig. 7. In this particular section of the grain, aluminium is present in relatively trace quantities as can be seen from the relative intensity scale; but when the $\text{AlK}\alpha$ radiation achieves a recordable intensity in the course of the scan, its position coincides with the $\text{TiK}\alpha$ peaks that represent the position of the hemo-ilmenite lamellae. (This is indicated by areas labelled C on the chart.) It is noted that not all of the hemo-ilmenite lamellae have exolved hercynite

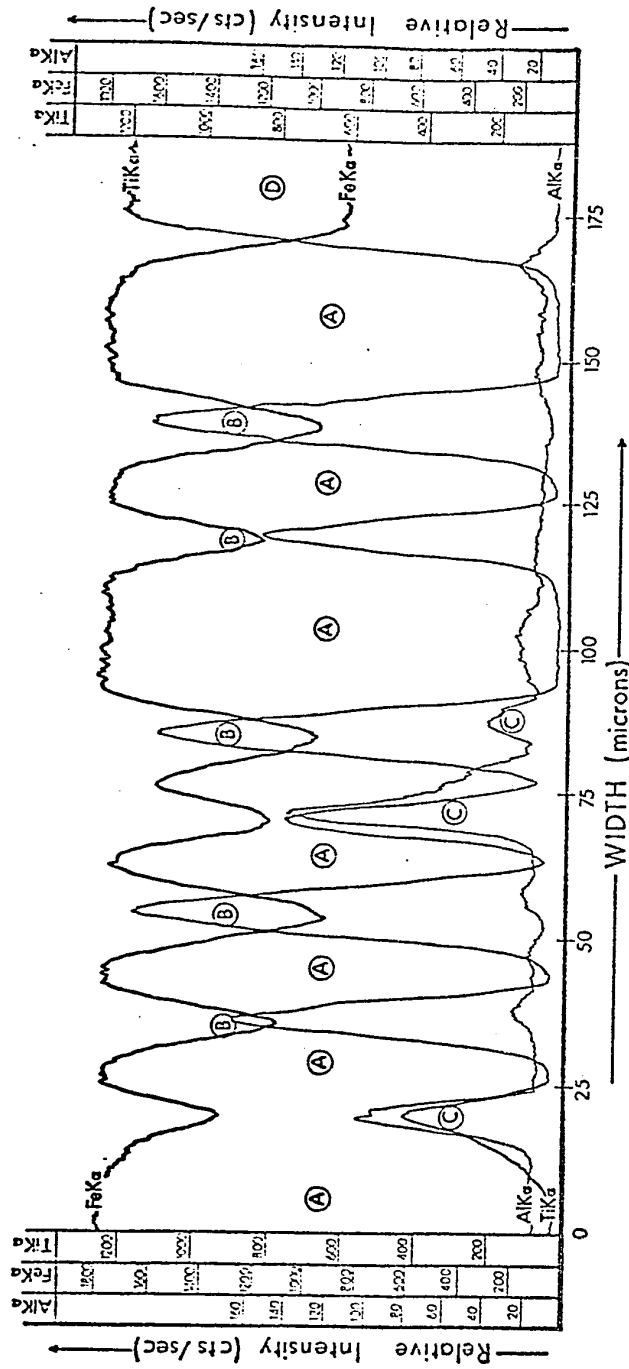


Fig. 7 Qualitative electron-beam scan of magnetite with ilmenite lamellae from a hyperite (Sample 211).
 (A) = Magnetite (B) = Evolved ilmenite (C) = Evolved ilmenite + Al-spinel (D) = Adjacent ilmenite grain.
 (Scanning speed: 16 μ / minute.)

associated with them, - in fact the granular aggregates of hemo-ilmenites are free of aluminium. Hence exsolution of hercynite probably took place with the original ulvöspinel, which has since been oxidised to hemo-ilmenite.

Another mode of occurrence of hercynite in some grains of magnetite within the ore-body is illustrated by Plate VI(c), which is an AlK α image of part of a magnetite grain (sample 47). Here hercynite is exsolved as tiny discs or blebs less than 5 microns in diameter; these blebs are microscopically invisible, and their presence was only indicated in the course of the aluminium microanalysis. Only three samples show this phenomenon (40, 47 and 54), and the slightly higher Al₂O₃ indicated for the magnetite analyses of these samples is due to contamination from such discs of hercynite, which could not be avoided.

(v) Vanadium

Vanadium in the iron oxides is probably present as V³⁺ with an ionic radius of 0.74Å; its ionic radius is greater than Fe³⁺ (0.64Å), but its electro-negativity is much less, this latter factor being responsible for its enrichment in early-formed magnetites (Mason, 1966).

The Sjøftestad magnetites have been reported to be rich in vanadium compared to magnetite associated with gabbroic rocks (Mitchell, 1967). This was found not to be the case, as the vanadium analyses in the tables show.

The magnetites from the ore body, amphibolites and skarns (Tables 4 to 6) have vanadium contents ranging from 0.1 to 0.3 wt. % V_2O_3 , while the coexisting titanhematites, having a greater proportion of Fe^{3+} ions, have vanadium contents of $1\frac{1}{2}$ to twice that of the coexisting magnetites. It is also interesting to note that the concentration of V_2O_3 in the titanhematites generally varies in proportion to the titanium concentration; the ones with the highest TiO_2 values have lower V_2O_3 contents because of the proportional lowering of the Fe_2O_3 content of the titanhematites.

The metaigneous samples of Table 7 again show the distinct difference of their oxides from those of the ore-body and country rocks, this time with respect to vanadium. The vanadium contents of the magnetites range from about 0.5 to 1.1 wt. % V_2O_3 , while the coexisting hemo-ilmenites surprisingly show slightly higher V_2O_3 contents than the magnetites. Fe_2O_3 in the ilmenites occurs as hematite, either exolved as tiny discontinuous lamellae (Plate V(E)) or in solid solution. Most of the metagabbro and hyperite samples examined show that extensive leaching of the magnetite has occurred, leaving skeletal grains of ilmenite (Plates III(e), V(d)). It is possible that during the leaching process, some of the vanadium was accommodated into the hematite phase of the hemo-ilmenites, thus enriching them in vanadium. This could not be checked by direct analysis of the hematite lamellae in the hemo-ilmenite due

to their extremely small size. (The high Fe_2O_3 content recorded in the analysis for sample 181 is due to this difficulty.)

Figure 8 shows the distribution of vanadium between coexisting magnetites and rhombohedral phases (titanhematites and hemo-ilmenites). The enrichment in vanadium of magnetites from the metagabbros and hyperites relative to those from the ore-body and country rocks is evident.

Vaasjoki, et al (1963) have discussed the possibility of using the relative amounts of vanadium and iron in magnetites as a possible genetic indicator of the ore formation concerned. Taking the vanadium module, which is defined as the percentage of vanadium contained by the total iron in a magnetite, the authors established distinct genetic groupings of iron-ore bodies from different parts of the world. The magnetites that show high vanadium modules (>1.1) were all found to be those associated with basic eruptive rock series and dykes. Metamorphosed ores associated with basic intrusives gave somewhat lower varying values (~ 0.5 to 0.6), while the lowest values (~ 0.17 to 0.20) were obtained for ores that originated by a process of metasomatic replacement.

At Sjøftestad, the ore-body magnetites have vanadium modules ranging from 0.11 to 0.17; the skarn magnetites show values of 0.15 to 0.16, while magnetites from the amphibolites have V-modules of 0.10 to 0.16. The magnetites

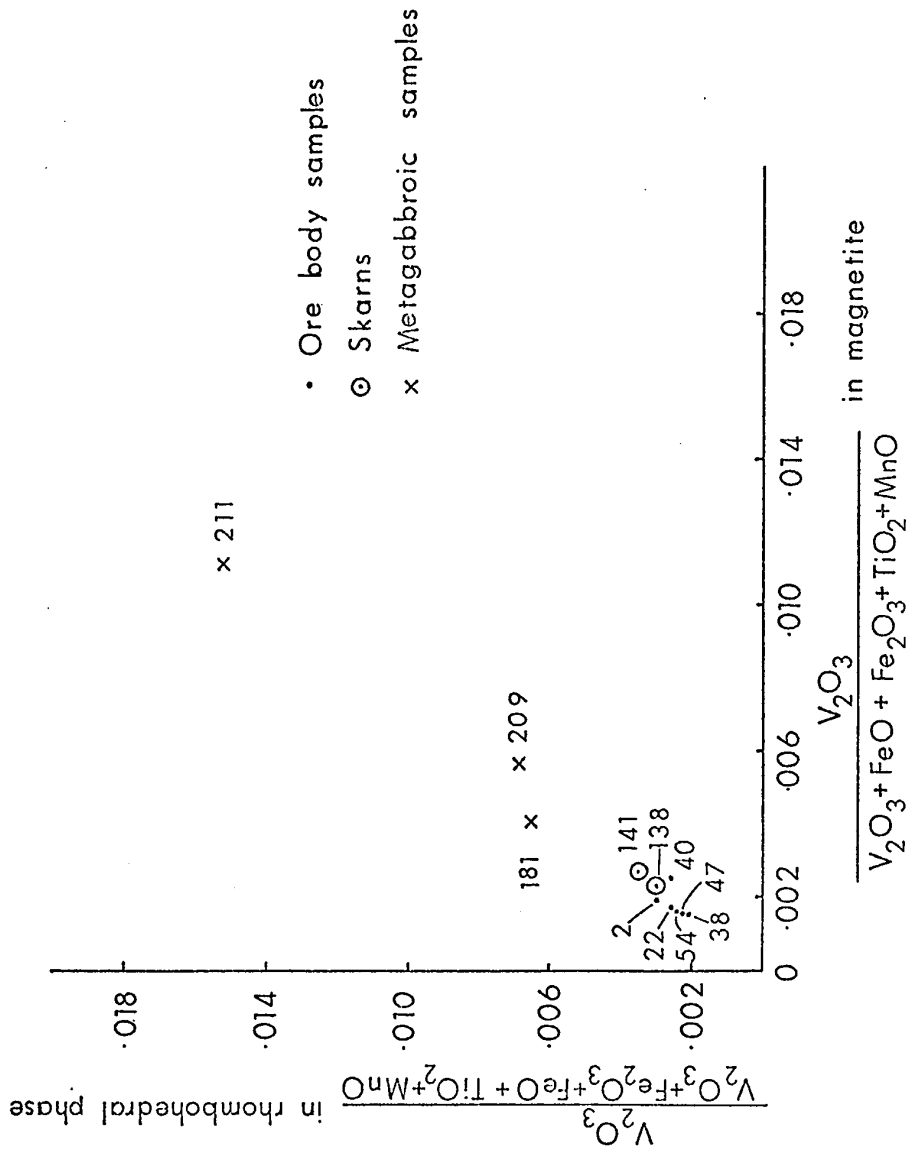


Fig. 8. Distribution of Vanadium between coexisting magnetite and rhombohedral phases

associated with the metabasics give higher values (~ 0.4 to 0.75), similar to the metamorphosed ores associated with basic intrusives cited by Vaastoki, et al (1963). The low vanadium module values obtained for the magnetites from the ore-body, amphibolites and skarns are similar to those iron-ore bodies formed by a metasomatic replacement process (Vaasjoki, et al, 1963).

In summary, the Sjøftestad magnetites are almost pure magnetites; the ore magnetites and those associated with skarns and amphibolites are chemically and texturally very similar with very little Ti and V dissolved in them, while the coexisting titanhematites show a wide variation in the amounts of ilmenite in solid solution with the hematite phase. This is considered to be due to local variations in oxygen fugacity.

The magnetites associated with the metagabbros and hyperites show distinct chemical and textural difference compared to those above. They are again Ti-poor, but coexist with hemo-ilmenites, which resulted from the oxidation of the original ulvöspinel dissolved in the magnetites in response to the high oxygen pressure conditions that prevailed during metamorphism. Compared to the ore-body magnetites and those from skarns and amphibolites, these magnetites show higher Al and V, and lower Mn contents.

These differences probably indicate that the ore-body magnetites at Sjøftestad do not owe their origin to

direct magmatic processes as those of the metagabbros and hyperites, but rather to metasomatic replacement processes; the skarn oxides that formed metasomatically show similar chemistry and texture to the main ore.

Sulfur Isotope Measurements and Discussion

Mass spectrometer measurements of sulfur isotope values were performed on coexisting pairs of pyrite and chalcopyrite from a sulfide-rich section of the ore-body under the supervision of Prof. Krouse of the Dept. of Physics, University of Calgary.

The results are given in Table 8 below. Due to difficulties in measurements, values are not given for Py 116(b) and Cpy 123(b).

The δS^{34} values range from -3.8 to -2.6, within the magmatic hydrothermal type deposits (Barnes, 1968), and it is likely that the mineralising solutions leached the sulfur from the surrounding magmatic rocks.

TABLE 8. Sulfur Isotope Values for Coexisting Pyrite-Chalcopyrite Pairs from the Sjøftestad Ore

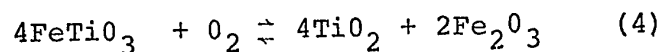
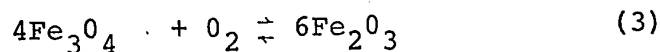
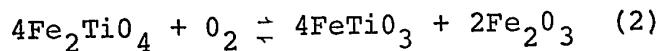
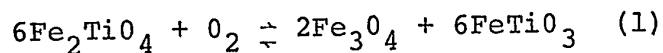
Sample No.	δS^{34}		$\Delta \delta S^{34}_{py-cpy} \text{ ‰}$
	Cpy	Py	
116(a)	-3.7	-3.1	0.6
116(b)	-3.8		
123(a)	-3.2	-3.1	0.1
123(b)		-2.6	

A good agreement is obtained for the chalcopyrites of 116(a) and 116(b), and for the pyrites of 116(a) and 123(a). But the δS^{34} pyrite value of 123(b) shows a difference, though still not too far out of the standard deviation. The reason for such a conspicuous difference is not clear.

Using Kajwara, et al's (1971) temperature scale for coexisting pyrite-chalcopyrite pairs, a temperature value of about 550°C is obtained for the sample 116 pair, while the sample 123 pair give a value greater than 700°C. The former temperature is a realistic value that reflects the epidote-amphibolite to amphibolite grade of metamorphism the ore-body has experienced. From the latter value, it appears that equilibrium relationships have not been met between pyrite and chalcopyrite in certain cases with respect to their S^{32}/S^{34} ratio.

Application of Experimental and Theoretical
Data to Ore and Country Rocks

A. Progressive Oxidation of Oxides: Buddington, et al (1964) gave the following relationships with progressive oxidation:

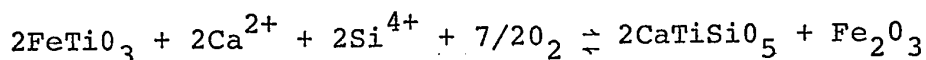


In the Sjøftestad area, reaction (1) is displaced towards the right hand side. Samples from the meta-basics just north of the ore-body contain magnetite with ilmenite lamellae (Plate V(C)). The titanium content of these magnetites is very low, indicating that almost no ulvöspinel remains unoxidised in the magnetites in accordance with the high oxygen pressures that prevailed during metamorphism.

Reaction (2) is interesting for one particular sample (no. 181, Plate V(e)) having ilmenite with tiny discontinuous lamellae of hematite. The ilmenite in this sample occurs both as lamellae in the magnetite and as granular and skeletal aggregates in contact with the magnetite that bears the lamellae of ilmenite. As noted in the section on ore description, the hematite lamellae disappear in the vicinity of the ilmenite-magnetite contact. Kretchmar, et al (1971) have noted this same feature in magnetite-hemo-ilmenite grain contacts in samples from the Whitehorse anorthosite. The hemo-ilmenite grains are inhomogeneous with regard to the amounts of hematite in solid solution, being poorer near the magnetite grain contacts and more saturated away from the contacts. Kretchmar, et al (1971) suggest that this 10 to 15 micron ferrian-ilmenite rim is due to the migration of ilmenite during the oxidation of ulvöspinel into surrounding hemo-ilmenite grains, where it dissolves and forms a homogeneous ferrian ilmenite rim.

Reaction (3) has been observed in a few specimens (Plate IV(g)), indicating that locally the oxygen pressure was high enough to oxidise some of the magnetite to titanium-poor hematite.

Neither the rutile-hematite association of reaction (4), nor the rutile-magnetite association indicated by Annersten (1968): $3\text{FeTiO}_3 + \frac{1}{2}\text{O}_2 \rightleftharpoons 3\text{TiO}_2 + \text{Fe}_3\text{O}_4$ has been observed in the Sjøftestad suite. As noted previously, the reaction rims of sphene around titanhematite grains in the hornblende gneisses and amphibolites indicates the oxidation of the ilmenite fraction of titanhematites in a calcium-silica-rich environment.



The above reaction is consistent with the high oxygen pressures indicated by the other assemblages.

Buddington, et al (1964) give a graph (Fig. 5) that shows that specified compositions of magnetite ss and ilmenite ss are unique functions of temperature and oxygen fugacity. Attempts to utilise their curves for determining the temperature and $f\text{O}_2$ that prevailed during the metamorphism of the Sjøftestad ore proved unsatisfactory, largely because the oxide mineral compositions of Sjøftestad do not fall in the range of the available experimental data. Extrapolation of the curves by a method suggested by Smith (1972) was attempted, but the range of values obtained combined with the uncertainties in extrapolation (both the

magnetite ss and hematite ss curves approach parallelism with a decrease of ulvöspinel and an increase in Fe_2O_3 in hematite ss) proved to be without meaning.

Buddington, et al (1964) encountered the same difficulty in the oxide compositions of some metamorphic and partly metasomatised paragneisses. The authors explain the high values of excess TiO_2 in many of the rhombohedral phases by the presence of meta-ilmenite; the Fe_2O_3 content of the original phases was lower and the FeTiO_3 content higher than the analyses presently indicate. This would fit with the suggestion presented here that whatever ilmenite was formed from the original titanhematites was oxidised in a calcium-silica rich environment to form Fe_2O_3 + sphene.

B. Estimation of the Range of $f\text{O}_2$

Assuming a realistic temperature of metamorphism of around 550°C for the Sjøftestad ore and country rocks (as indicated by sulfur-isotope data, metamorphic mineral assemblages of upper epidote-amphibolite to amphibolite facies [Winkler, 1967], and MnO distribution between the coexisting oxide phases [page 61]), a general limit can be placed on the oxygen pressure that prevailed during metamorphism. Annersten (1968) gives the theoretical stability fields for the system Magnetite-Hematite-Ilmenite-Rutile as a function of oxygen pressure between $400 - 600^\circ\text{C}$. The coexistence of magnetite + hematite + ilmenite at Sjøftestad indicates a

P_{O_2} range of about 10^{-18} to 10^{-16} atm. for a temperature range of 530 to 570°C. That higher oxygen pressures prevailed is indicated by the oxidation of ilmenite to sphene (instead of hematite + rutile). An approximate upper limit of 10^{-10} atm. P_{O_2} could be placed for the Sjøftestad rocks since ilmenite-free hematite does not coexist with either sphene or rutile.

As noted by the author, the theoretical stability field given for this system is strictly valid for pure phases; the extent to which titanium, entering as a third component in solid solution to stabilise hematite at lower oxygen pressures (Lindsley, 1962) is not known.

CHAPTER IV

COMPARISON OF SØFTESTAD ORE WITH METAMORPHOSED
KIRUNA TYPE ORES, AND POSSIBLE SKARN ORIGIN

Types of Fe Deposits: From the numerous works on descriptions and modes of origin of iron ore bodies in the literature, it is evident that most of the important deposits fall into one of the following general categories:

(i) Magmatic deposits, which could be of two main types:

(a) The "Kiruna type" deposits, which are phosphatic (apatite-rich) deposits formed by a magmatic differentiation process. Magnetite is the principal ore mineral, with hematite subordinate and secondary.

(b) The "Hauki type" ores (Frietsch, 1970), which are genetically connected to the Kiruna type ores, but formed at later stages and lower temperatures. These types of deposits are believed to be hydrothermal-metasomatic in origin (Frietsch, 1970), representing later stages in the magmatic differentiation process that gave rise to the main mass of deposits of the Kiruna type, and are mainly hematite ores.

(ii) The "Cornwall type" contact or skarn deposits, formed by pyrometasomatic processes; such deposits owe their origin to magmatic activity nearby, which supplies the necessary ore materials, volatiles and heat that alter the

nature of the country rocks the magma invades. Both magnetite and hematite can occur in those type of deposits.

(iii) The "Lake Superior type" deposits, which are stratified, usually quartz-rich ores of sedimentary origin. The iron in these is present as hematite, magnetite, pyrite, siderite or one of the iron silicates (Park, et al, 196).

Of the above categories of iron-ore bodies, the Sjøftestad deposit could belong to either the Kiruna type or the skarn type deposit, since it exhibits features that resemble these two types. Due to repeated metamorphism and deformation, all original features of the ore and country rocks have been obliterated, and hence its mode of origin cannot be stated with certainty, but some comparisons could be made with similar ore bodies elsewhere.

Kvien (1961) has concluded that the Sjøftestad deposit resembles the Gällivare iron-ore body of Sweden, and most probably belongs to the Kiruna type ore as originally suggested by Vogt (1895; in Kvien, 1961). The main features of the Gällivare ore deposits have been described by Geijer (1910; 1967) and Annersten (1970).

The Gallivare Deposit

Situated in northern Sweden at approximately 67°N latitude, 45 miles north of the Arctic circle, the Gällivare iron formation occurs in gneisses and leptites. Both the ore and country rocks are of Precambrian age. The leptites are described as metamorphosed acid and intermediate

volcanics, and contain plagioclase (An_{10}), calcic amphibole, diopside, sphene, apatite, scapolite, quartz and iron oxides (Annersten, 1968). Bodies of granite, which appear to be the youngest rocks in the field, are enclosed within the leptites and gneisses, and even in the ores. The contact between the ore and leptites is not sharp everywhere, and metamorphism occurred in connection with strong deformation and folding. Annersten (1968) gives a temperature of metamorphism of approximately 550°C.

The ore is mainly composed of magnetite, but in the western part, some hematite occurs with the magnetite; locally, some martitisation of the magnetite is seen. The so-called "Main Ore" of the Gällivare field comprises a chain of sill-like ore bodies in which the banding of apatite is very common. The apatite grains lie in parallel rows, with many short gaps, and bands of pure apatite rock, reaching several centimeters in width are not rare (Geijer, 1967). In addition, Frietsch (1967) notes the presence of sillimanite, andalusite and corundum associated with the ore; he suggests that these minerals formed as a result of wall-rock alteration by intruding pegmatites and granites rather than as a result of earlier magmatic activity.

Annersten (1968) states that the pre-metamorphic history of the Gällivare deposit still remains an open question, since skarn rocks rich in hornblende and pyroxene occur at the border between the ore and the leptites,

suggesting a genetic connection between the two. But other workers (Magnusson, 1953 [in Annersten, 1968]; Frietsch, 1967, 1970; Geijer, 1967) have concluded that the deposit is a Kiruna type deposit which formed by a magmatic differentiation process, and was later metamorphosed.

Comparison of Sjøftestad with Gällivare

The similarities between the Sjøftestad and Gällivare iron-ore deposits can be summarised as follows:

- (i) Both deposits are of Precambrian age and have experienced similar temperatures of metamorphism.
- (ii) They are both apatite-rich deposits.
- (iii) Both deposits show low sulfur contents.
- (iv) The magnetites of both ores are impoverished in TiO_2 , indicating comparatively high oxygen fugacities during metamorphism.
- (v) The Al_2O_3 content of the Gällivare magnetites average around 0.2 wt. % (Frietsch, 1970), which is about the same as for the Sjøftestad magnetites.
- (vi) Pegmatite dykes cut the ores of both deposits.

The differences observed between the two deposits include:

- (i) Hematite is rare at Gällivare, while it forms an important constituent of the ore at various sections of the Sjøftestad deposit.
- (ii) The magnetites of Gällivare show a slight chemical difference to those of Sjøftestad in that they contain a higher proportion of trace elements. Ni and Co are present

in amounts of up to 0.1 wt. % at Gällivare (Frietsch, 1970) while these elements could not be detected in measurable amounts on the microprobe in the Sjøftestad magnetites. In addition, the Gällivare magnetites contain slightly higher V_2O_3 contents (up to 0.4%) and lower MnO contents (<0.05 to 0.1%) compared to the Sjøftestad magnetites (0.1 to 0.2% V_2O_3 and 0.2 to 0.4% MnO).

(iii) The host rocks are leptites at Gällivare and hornblende gneisses at Sjøftestad.

(iv) Ilmenite is found associated with the ore magnetites in sections of the Gällivare deposit (Annersten, 1968), while it is completely absent in the Sjøftestad ore-body samples.

(v) Sillimanite, andalusite and corundum are found associated with the ore at Gällivare (see above); these minerals have neither been reported or observed in the Sjøftestad ore.

(vi) Sulfide-rich patches composed mainly of pyrite, chalcopyrite, bornite and chalcocite are found associated with parts of the ore at Sjøftestad (Kvien, 1961). Gypsum and anhydrite have also been found associated with the ore (Plate I(b); also Mitchell, 1967). These are features that have not been reported in the Gällivare deposit. The above similarities and differences between the two deposits do not in themselves indicate that the Sjøftestad ore either is or is not a Kiruna type ore as is believed for Gällivare. On the basis of its high phosphorous and low sulfur contents,

and the apparent lack of metasediments in the Nissedal supracrustals (Mitchell, 1967), the Sjøftestad deposit could be classified as a metamorphosed Kiruna type deposit. On the other hand, there are certain aspects of the ore that could probably be better explained by considering a skarn origin hypothesis (see below). As noted by Frietsch (1970), skarn deposits could contain relatively high phosphorous contents (up to 2%); the andradite-bearing pyroxene and amphibole skarns from Sjøftestad have a significant amount of layered apatite grains associated with the iron oxides in much the same way as other parts of the ore. It should also be noted that sulfides and sulfates are by no means rare in the Sjøftestad ore, even though in localised patches (Kvien, 1961).

Evidences for Skarn (or Metasomatic) Origin

The occurrence of skarn rocks, especially in the central banded magnetite - hematite sector of the mine (Kvien, 1961) is a significant evidence for a possible skarn origin for the deposit. A few samples from outside the ore zone have also been found to be skarn mineral assemblages. The mineralogy of the skarns have been noted in the previous section. They are essentially andradite-bearing amphibole and pyroxene skarns containing in addition some epidote, apatite, allanite, plagioclase and some calcite. Both magnetite and hematite occur in these rocks

either as bands associated with apatite grains or disseminated through the rock replacing the amphiboles. Phlogopite has also been noted in some ore-impregnated hornblende gneiss samples.

It is significant that the magnetites from the ore, hornblende gneiss and skarns show a striking similarity in chemistry compared to those of the metabasics to the north of the ore-body. Since it can be assumed that all the rock types in the area have experienced the same degree of metamorphism and recrystallisation, such chemical differences are due to different modes of origin. Especially significant are the high V and Al, lower Mn, and the association of ilmenite with the magnetites in the metabasics (Table 7); this suggests that the ore body oxides did not form from a direct magmatic source as those of the metabasics. Furthermore, the extensive leaching of the magnetite from the metabasics leaving skeletal grains of ilmenite (Plate V(d)) indicates that these rocks at least in part provided a source from which iron was removed by fluids mobilised prior to or accompanying metamorphism and granitisation. The iron could then be deposited by a metasomatic replacement process at suitable sites in the surrounding country rocks whose original features have been obliterated by metamorphism.

Works by Davidson, et al (1968) on the formation of iron oxide deposits near intrusive masses in Pennsylvania

give added support to such a suggestion. The authors found that during the stage of iron-enrichment in the igneous differentiation sequence, an aqueous fluid capable of dissolving iron from the magnetite-ilmenite intergrowths developed within the crystallising magma, leaving skeletal grains of ilmenite; they suggest that such iron-bearing solutions could escape to higher levels where they would be a potential source of magnetite deposits of the Cornwall (or skarn) type.

From hypersolidus and subsolidus hydrothermal quenching experiments, Martin, et al (1969) demonstrated that iron is among the constituents that tend to fractionate into an aqueous gas phase at geologically reasonable temperatures (450 - 580°C); The iron-bearing solutions migrate readily towards lower temperature zones. There appears to be an apparent lack of dependence on rock type, the presence of a melt phase or of chloride, fluoride, sulfide or carbonate anions, which strongly suggests the applicability of the experimental results to many contact metasomatic ore deposits (Martin, et al, 1969). The authors' findings further indicate that hematite can be deposited directly from the iron-bearing solutions, which could be applied to the Sjøftestad deposit since replacement relationships between magnetite and hematite are hard to find. The magnetite-rich zones of the ore could represent depositions at lower oxygen fugacities while the more hematite-rich sectors represent depositions at higher f_{O_2} , reflecting spatial and

time differences during the ore-formation process.

The concordant nature of the Sjøftestad deposit to the hornblende gneiss host rocks, and the fact that considerable replacement of the hanging wall is seen at various sectors of the mine (Kvien, 1961) further suggest a genetic connection between the ore and wall rocks. However, Mitchell (1967) suggests that this may be due to alterations accompanying the metamorphic process, and may not be connected with the actual ore genesis.

Finally, the plot of the compositions of coexisting magnetite ss and hematite ss on the temperature - fO_2 graph given by Buddington, et al (1964) falls outside the range of available experimental data, in much the same way as those oxides of metasomatic origin (Buddington, et al, 1964).

CHAPTER V

SUMMARY AND CONCLUSIONS

Ore and Country Rocks

The Sjøftestad iron ore deposit is found associated with hornblende gneisses that form part of the Nissedal supracrustals of the Telemark region. Other rock types that form the supracrustals are amphibolites, metagabbroic rocks, leptites, ödogardites, quartzo-feldspathic gneisses, migmatites and meta-agglomerates.

Both the ore and country rocks are of Precambrian age, and have undergone metamorphism of the epidote - amphibolite grade. The intense metamorphism and granitisation has obliterated all the primary textures of the host hornblende gneisses and amphibolites. At least two periods of metamorphism and intrusive emplacement are recognised in the rocks, with a third period of deformation that produced large shear zones in the recrystallised basement (Wegmann, 1960). Retrograde effects, probably due to the superposition of the two periods of metamorphism, are evidenced by the chloritisation and prehnitisation of the ferromagnesian minerals in the hornblende gneisses and amphibolites.

The iron-ore body is totally conformable with the enclosing gneiss, and is made up of lenses that grade into

one another. In places, there is a considerable replacement relationship between the ore and the gneiss. On a textural basis, the ore is classified into 3 types: (i) A layered, foliated magnetite-apatite-titanhematite ore. (ii) An unfoliated (or poorly foliated), granular magnetite and/or titanhematite ore. (iii) A fine grained, intensely sheared magnetite ore.

Such textural differences are interpreted as a reflection of the degree of deformation and recrystallisation that various sections of the ore have experienced. The layered type of ore recrystallised into the more granular coarser type, which upon shearing formed the cataclastic, fine grained type of magnetite ore.

Sulfide-rich patches occur in certain sections of the ore body; the main sulfide minerals are pyrite, bornite, chalcopyrite and chalcocite. They are considered to have formed later than the oxide phase since they fill in cracks and irregularities in coexisting magnetite grains.

Analytical Works

Microprobe analysis for Fe, Mg and Ti has been performed on coexisting calcic amphiboles and biotites of some ore-associated country rocks to determine the distribution of these elements between the two mineral phases. Analyses have also been made for Fe, Ti, Mn, Mg, Al, and V on the iron oxide phases from the ore and hornblende gneiss, amphibolites, skarns and metagabbroic rocks. The amounts of Ni,

Co and Cu in the oxides were found to be negligible.

A. Fe, Mg and Ti in Coexisting Ca-Amphiboles and Biotites

The distribution of Fe between the two coexisting phases (Fig. 4) is found to be a non-linear relationship, with the calcic amphiboles having a tendency for a higher enrichment in Fe relative to the biotites. Kretz (1960) has shown that the Al, K and Na contents of the two minerals (especially Al) has a significant effect on the distribution of Fe between them: the higher the Al (K and Na) content of the amphiboles, the greater the tendency for iron enrichment. This would probably explain the iron distribution observed for the Sjøftestad minerals, though Al analysis was not performed. On the whole, there is a close approach towards equilibrium with respect to the distribution of iron between the two phases.

The Mg distribution curve (Fig. 5) shows a fairly linear distribution of the element between the two coexisting minerals, and again a close approach towards equilibrium with respect to magnesium is indicated.

In the Ti distribution however (Fig. 6), there is a considerable scatter of the distribution points, indicating that chemical equilibrium with respect to titanium was either not established or was not retained between the amphiboles and biotites. The reason is probably due to the presence of titanium microlites in the form of sphene or

ilmenite in the ferromagnesian; this is evidenced by reaction rims of sphene around biotites, and the close association of sphene and secondary ilmenite with both ferromagnesian phases.

B. Oxide Analytical Results

The results of the oxide analyses (Tables 4-7) show that the oxides from the ore body, amphibolites and skarns have a similar chemistry and differ from the metagabbroic oxides.

The magnetites of all the rock types and ore are universally poor in titanium, indicating a relatively high oxygen pressure during metamorphism. Titanhematites coexisting with magnetite show a wide variation in their TiO_2 contents (from 0.3 to almost 10 wt. %), most probably due to local inhomogeneities of oxygen fugacity.

Manganese is enriched in the magnetites relative to the coexisting titanhematites. The magnetites of the metagabbros show a generally lower MnO content than those of the ore, amphibolites and skarns. The metagabbro hemo-ilmenites, however, show MnO contents of 1.6 to 2.6 wt. %, which is assumed to be present as pyrophanite ($MnTiO_3$) in solid solution with ilmenite.

Magnesium is present in only trace amounts in both magnetites and titanhematites of ore and country rocks (<0.1 wt. %); in the hemo-ilmenites of the metagabbroic rocks, the MgO content goes up to 0.7 wt. %, with magnesium

present as geikielite (MgTiO_3) in solid solution with ilmenite and pyrophanite.

Aluminium is present in minor amounts in both magnetites and titanhematites of the ore, amphibolites and skarns (0.2 wt. % Al_2O_3), while the metagabbro magnetites have up to 0.74 wt. % Al_2O_3 . In a few ore samples, aluminium is exolved as small discs of hercynite.

Vanadium is present in amounts ranging from 0.1 to 0.3 wt. % V_2O_3 in the magnetites of the ore, while it is slightly enriched in the coexisting titanhematites; this is considered to be due to the greater proportion of Fe^{3+} ions in titanhematites which vanadium replaces. Similar concentrations are observed for the oxides of the amphibolites and skarns. The metagabbroic oxides, on the other hand, show V_2O_3 contents of 0.5 to 1.5 wt. % in the magnetites and 0.5 - 1.1 wt. % in the ilmenites. The possibility of using vanadium in magnetites as a genetic indicator as suggested by Vaasjoki, et al, is investigated; the low V-module of the ore magnetites is found to be similar to ores formed by a metasomatic replacement process, while the higher module of the metagabbroic magnetites is formed to be within the range of metamorphosed ores associated with basic intrusives.

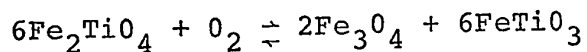
C. Sulfur Isotope Results

Sulfur-isotope measurements for pyrite-chalcopyrite pairs from the ore-body indicate δS^{34} values that range

from -3.8 to -2.6, which fall in the general range of magmatic sulfur. The mineralising fluids probably leached the sulfur from the surrounding magmatic rocks. From Kajwara, et al's (1971) temperature scale for coexisting pyrite-chalcopyrite pairs, temperatures of 550°C and >700°C are obtained for two pairs of the sulfide minerals; it appears likely that equilibrium relationships were not met in the latter pair.

Application of Some Literature Data
to Ore and Country Rocks

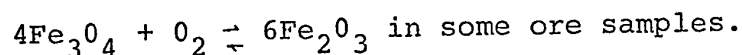
From the relationships between the iron and titanium oxide phases with progressive oxidation given by Buddington, et al (1964), a high rate of oxidation is indicated during the metamorphism of the Sjøftestad ore and country rocks. Very little ulvöspinel remains unoxidised in the magnetites, in accordance with the relationship:



Since no ilmenite occurs with the ore, it is assumed that any ilmenite formed by the above process was either incorporated in the hematite phase to form hematite-ilmenite solid solution, or was oxidised to sphene.

With higher oxidation, the relationship:

$4\text{Fe}_2\text{TiO}_4 + \text{O}_2 \rightleftharpoons 4\text{FeTiO}_3 + 2\text{Fe}_2\text{O}_3$ is observed in the metagabbroic samples. The highest oxidation of the oxides at Sjøftestad is represented by:



The absence of rutile with either magnetite or hematite in any of the samples places a limit on the high oxygen pressure reached during metamorphism. With these considerations, a P_{O_2} range of 10^{-18} to 10^{-10} atm. is estimated for the Sjøftestad area from the theoretical stability fields for the system Magnetite-Hematite-Ilmenite-Rutile given by Annersten (1968).

Attempts to utilise the temperature- f_{O_2} -Oxide-solid solution graph of Buddington, et al (1964) was unsuccessful, largely because the compositions of the coexisting oxides do not plot within the range of the available experimental data.

Conclusions on Origin of Deposit

Comparisons between the Sjøftestad deposit and metamorphosed deposits of the Kiruna type (the Gällivare iron ore field of N. Sweden) show similarities and differences that do not conclusively indicate whether or not the Sjøftestad ore should be considered a metamorphosed Kiruna type ore.

The significant similarities between the two deposits are: both have experienced similar grades of metamorphism, show low sulfur contents, have magnetites that are impoverished in titanium, and are apatite-rich.

The notable differences between the two deposits are: the rarity of hematite at Gällivare compared to Sjøftestad; the slight enrichment in Ni, Co and V in the

Gällivare magnetites relative to those at Sjøftestad; the association of ilmenite, sillimanite and andalusite with the Gällivare ore, while these minerals are absent at Sjøftestad; the presence of sulfide-rich patches associated with the Sjøftestad ore.

Since the Sjøftestad ore exhibit some skarn features, the possibility of a skarn origin is investigated; skarn deposits could in some cases be apatite-rich (Frietsch, 1970), and sections of the Sjøftestad ore are sulfur-rich. Evidences for a metasomatic replacement (skarn) origin are:

- (i) The association of skarn rocks with the ore.
- (ii) The low vanadium module of the ore magnetites, similar to ores formed by a metasomatic process.
- (iii) The chemical differences observed between the ore magnetites and those of the metagabbroic rocks (especially with respect to Al, V and Mn), indicating that the ore magnetites do not owe their origin to direct magmatic processes like those of the metagabbros.
- (iv) The leaching of the magnetite fraction of the magnetite-ilmenite grains of the metagabbros, indicating these rock types as possible source rocks for the ore.
- (v) The concordant nature of the ore to the host hornblende gneiss, and the extensive replacement of the hanging wall seen in various sections of the mine, suggesting a genetic connection between ore and hornblende gneiss.

(vi) The plot of the Sjøftestad oxide compositions outside the range of most coexisting oxide phases in the data given by Buddington, et al (1964), who also found that oxides of metasomatic origin have this same tendency.

While a Kiruna type origin for the Sjøftestad deposit is possible, the points noted above provide sufficient evidence to suggest a metasomatic replacement origin.

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Appendix A

Microprobe analysis

An ARL/Electron Probe X-ray Analyser was utilised for all the analysis performed for this work. The conditions of the analysis were:

Operating voltage: 15 Kv

Beam current: 0.1 Micro-amps

Beam width: 1 micron

Analysing crystals: RAP for Mg and Al

LiF for Mn, Fe, Ti and V

A computer program (Smith, et al, 1970¹) was utilised to process and correct the microprobe data. For every sample, three grains of whatever mineral phase to be analysed were selected, and every grain was probed three to five times for all the analysed elements. The average counts were then used for processing towards the final results.

Accuracy of the results

In general, higher concentrations of an element in a mineral phase or a large number of counts increases the accuracy of the microprobe results. From an evaluation of the percentage error guide tables provided for the electron-probe microanalyser, the following estimates are made for the accuracy of the results (at 99% confidence level):

For Fe: generally 3% by wt. of the oxides

For Ti, Mn, Mg, Al and V: generally 5% by wt. of the oxides.

¹Smith, D.G.W. and Tomlinson, M.C., 1970. An APL language computer program for use in Electron Microprobe Analysis. Computer Contribution 45, State Geological Survey, The University of Kansas, Lawrence.

APPENDIX B

Recalculation of Microprobe Analysis

Microprobe analysis gives the total concentration by weight of an analysed element irrespective of its valence state or states. It therefore becomes necessary to recalculate the values for iron obtained for the oxides to determine the ferrous-ferric oxide proportions.

Procedure: The procedure adopted is that described by Rumble (1970). The number of cations in the oxides are normalised to their ideal values of 2 for the rhombohedral phases and 3 for the spinels. The cations present are then assigned to various components:

(i) In magnetite, Mg is assigned to MgFe_2O_4 , Mn to MnFe_2O_4 , Al to FeAl_2O_4 , Ti to Fe_2TiO_4 , V to FeV_2O_4 and the remaining Fe to Fe_3O_4 .

(ii) In the rhombohedral phases, Mg is assigned to MgTiO_3 , Mn to MnTiO_3 , Al to FeAlO_3 , V to FeVO_3 , the remaining Ti to FeTiO_3 and the remaining Fe to Fe_2O_3 .

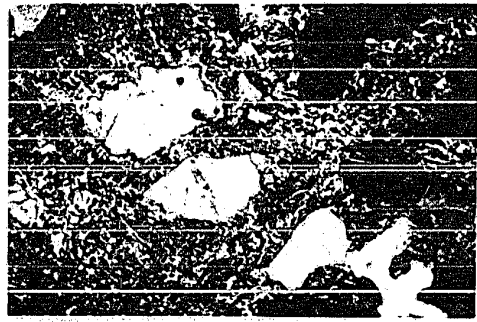
The $\text{Fe}^{2+} : \text{Fe}^{3+}$ ratio is then determined by adding up the number of these cations in each component; the result is used to recalculate the microprobe iron analysis in terms of FeO and Fe_2O_3 .

PLATE I

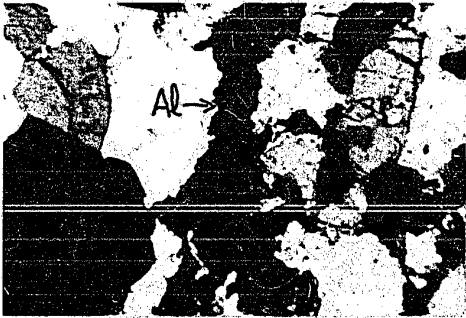
- (a) Reaction rim around apatite (central, dark grey grains) in prehnitised biotite. (Sample 40)
- (b) Anhydrite (centre left grains) and plagioclase (twinned grain, centre right) altering to gypsum (ground mass). (Sample 40)
- (c) Allanite (Al) forming rims around apatite (grey) and iron oxide (lower centre, within allanite). (Sample 40)
- (d) Apatite bands (high relief, bottom) growing in close association with oxides (black, some sphene-rimmed) in hornblende gneiss. The white grains (low relief) are plagioclase, the elongated grey ones are hornblende. (Sample 213)
- (e) Texture of biotite-amphibolite (white is plagioclase). Note that some of the iron oxide grains have grown by replacing the ferromagnesians - e.g. elongated oxide lath in contact with biotite (right of tope centre) maintains original shape of biotite. (Sample 125).
- (f) Garnet (central, small euhedral grain) in hornblende in amphibolite. (Sample 268)
- (g) Iron oxides (black) replacing hornblende (grey) in acid gneiss. Sphene forms rims around titanhematite (centre). (Sample 177)
- (h) Sphene (Sp) growing along plagioclase (light) grain boundaries, rimming oxides (black) and forming drop shaped grains in hornblende gneiss. (Sample 241)



(a) 200 μ



(b) 400 μ



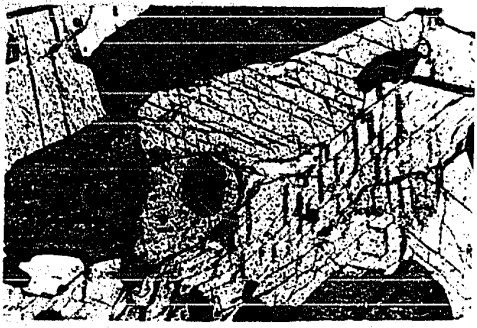
(c) 150 μ



(d) 400 μ



(e) 200 μ



(f) 150 μ



(g) 200 μ



(h) 200 μ

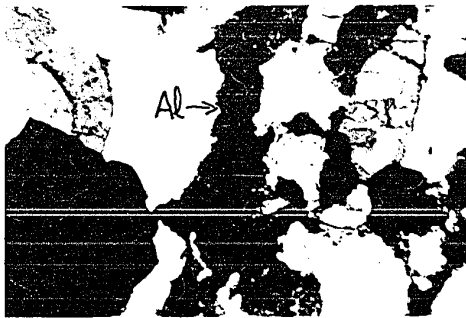
PLATE I.



(a) 200 μ



(b) 400 μ



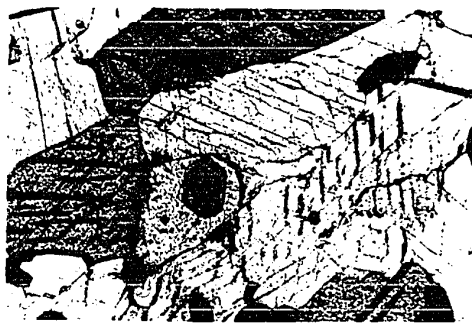
(c) 150 μ



(d) 400 μ



(e) 200 μ



(f) 150 μ



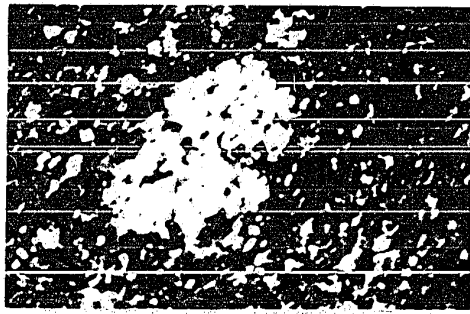
(g) 200 μ



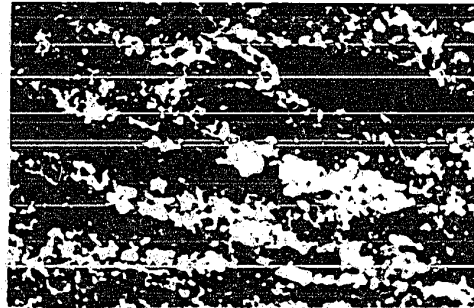
(h) 200 μ

PLATE II

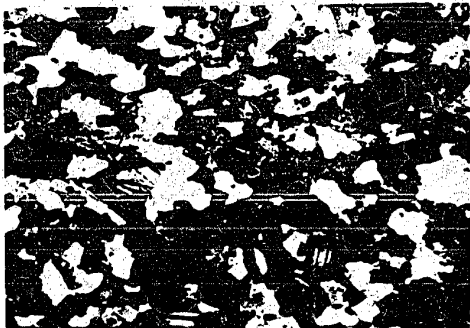
- (a) Scapolite (central light patch) in fine grained amphibolite. (Sample 210)
- (b) Texture of a fine grained, sphene-hornblende gneiss. Dark bands sphene and hornblende, light patches plagioclase. Coarse Porphyroblast at top right is sphene. (Sample 90)
- (c) Tendency for grain size uniformity of plagioclase (light and grey) and hornblende (darker grey) in a typical hornblende gneiss. (Sample 272)
- (d) Relict diopside (Di) in hornblende in ore-rich hornblende gneiss. (Sample 213)
- (e) Andradite-rich garnet (Andr) occurring as euhedral grains and forming epidote (Ep) cores in skarn. Both minerals occur in altered plagioclase (plag). (Sample 138)
- (f) Crude bands of andradite garnet (grey) in between bands of iron ore (black) in skarn. (Sample 141)
- (g) Allanite (centre, dark) with epidote rim in skarn. (Sample 141)
- (h) Oxide bands winding along altered plagioclase grain boundaries in skarn. (Sample 138)



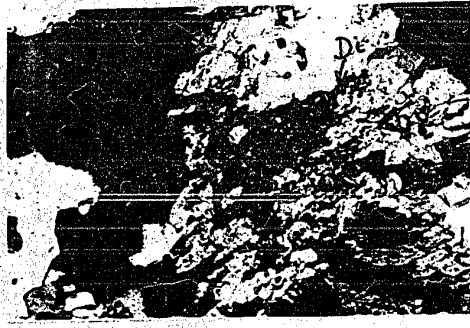
(a) $\text{---} 500\mu$



(b) $\text{---} 500\mu$



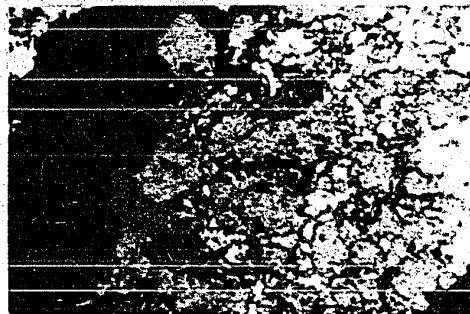
(c) $\text{---} 600\mu$



(d) $\text{---} 200\mu$



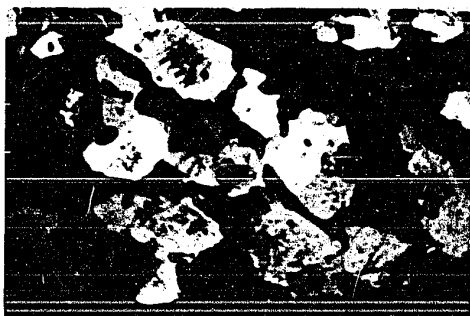
(e) $\text{---} 100\mu$



(f) $\text{---} 500\mu$

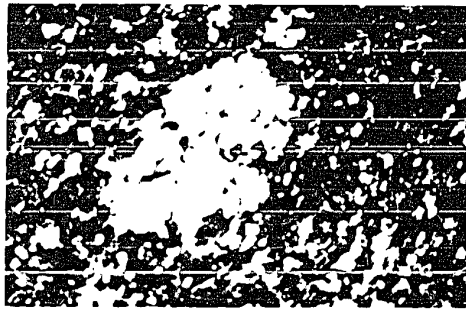


(g) $\text{---} 200\mu$

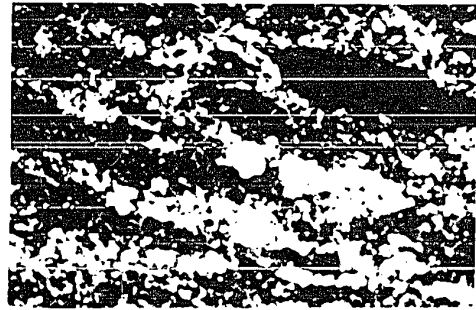


(h) $\text{---} 600\mu$

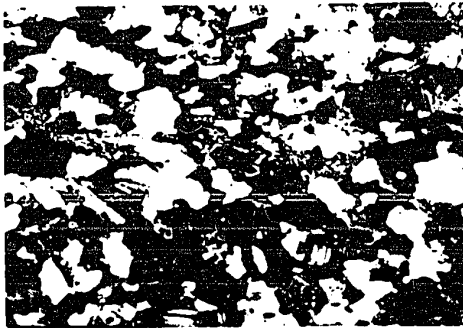
PLATE II.



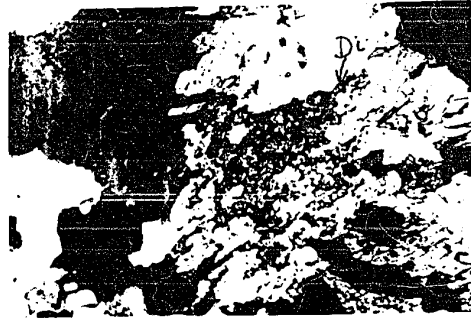
(a) 500 μ



(b) 500 μ



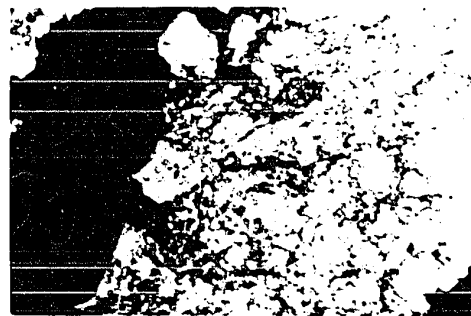
(c) 600 μ



(d) 200 μ



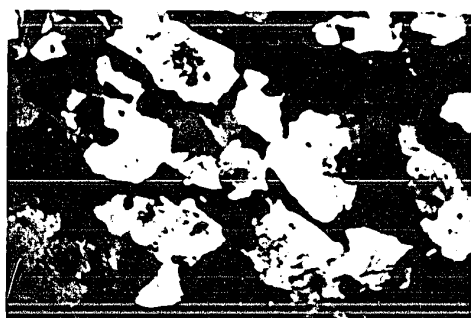
(e) 100 μ



(f) 500 μ



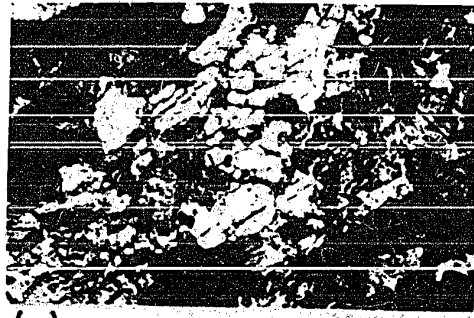
(g) 200 μ



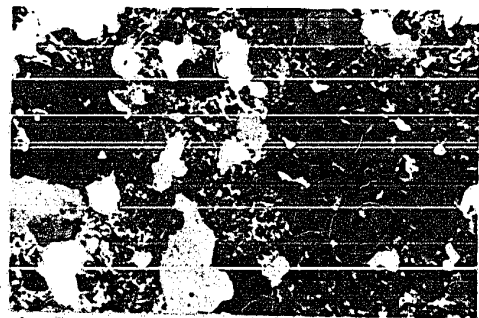
(h) 600 μ

PLATE III

- (a) Hedenbergite (central light, cleaved grains) being in part replaced by ferroactinolite (grey) in pyroxene-amphibole skarn. Black ore iron oxides. (Sample 252)
- (b) Iron oxides (black) altering to epidote in highly sericitised quartzo-feldspathic gneiss. (Sample 174)
- (c) Granoblastic texture in granitised leptite. (Sample 335)
- (d) Granular-porphyroblastic texture of quartzo-feldspathic gneiss. Finer grained quartz grow in between grain boundaries of coarser grained, earlier quartz and plagioclase (Sample 393)
- (e) Skeletal ilmenite (black) altered to hornblende (ho). White rim at left centre is secondary plagioclase growing along the patch of hornblende. From a metagabbro.
- (f) Coronite metagabbro, with radial laths of hornblende (surrounding black ilmenites at centre) enclosed in coarse porphyroblasts of twinned plagioclase. (Sample 113)
- (g) Texture of lamprophyre (spessartite). Euhedral grains of cummingtonite (white, fine grained laths) and diopside (bottom left) in a zeolite ground mass. (Sample 200)
- (h) Fibrous zeolite ground mass (centre) in spessartite. Black are iron oxides. (Note: scale at the bottom should read 100 μ .) (Sample 185)



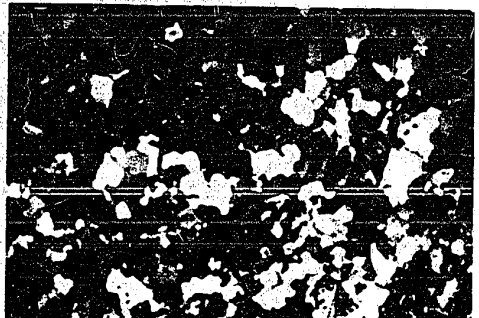
(a) 150 μ



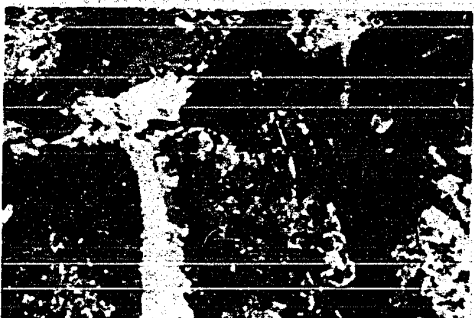
(b) 200 μ



(c) 500 μ



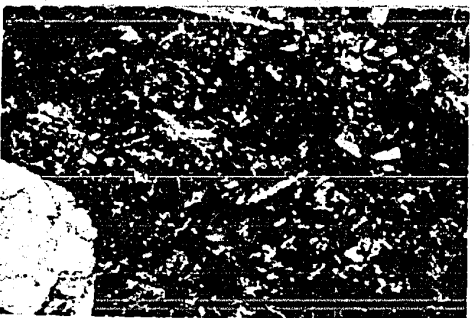
(d) 500 μ



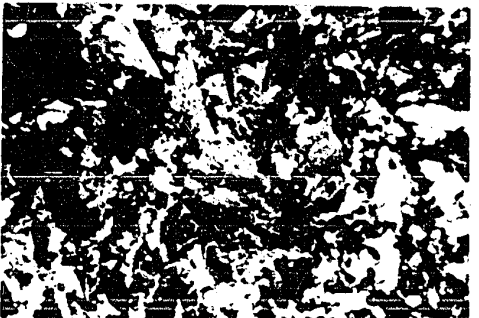
(e) 400 μ



(f) 500 μ

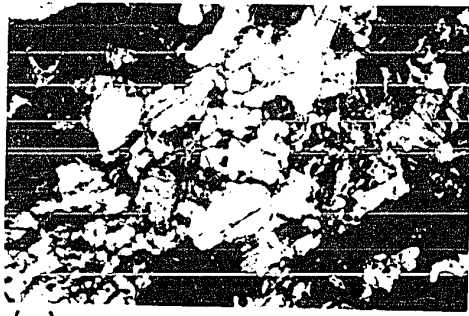


(g) 400 μ



(h) 1000 μ

PLATE III.



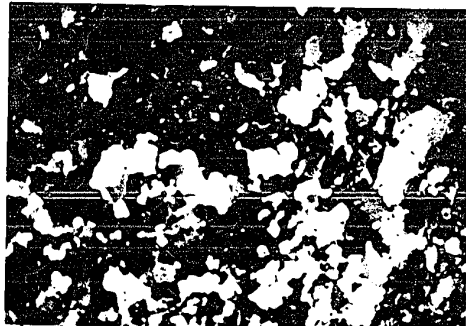
(a) $\text{---} 150\mu$



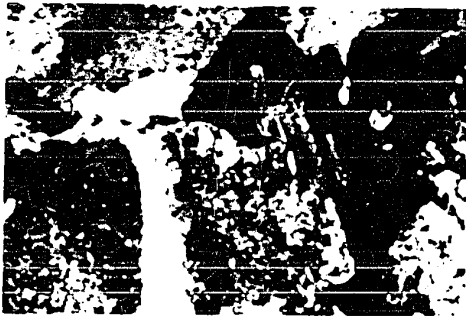
(b) $\text{---} 200\mu$



(c) $\text{---} 500\mu$



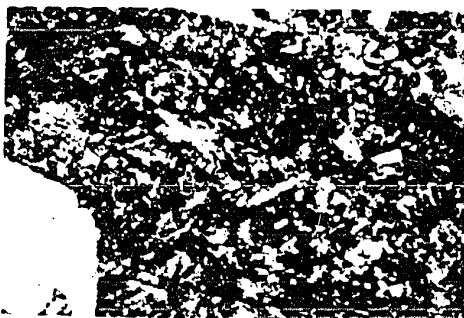
(d) $\text{---} 500\mu$



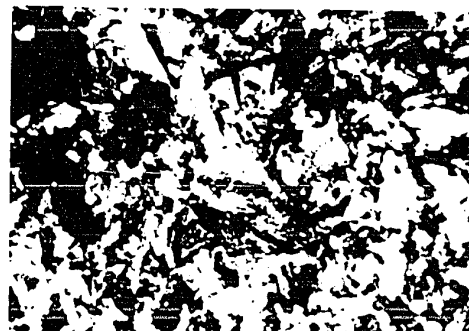
(e) $\text{---} 400\mu$



(f) $\text{---} 500\mu$



(g) $\text{---} 400\mu$

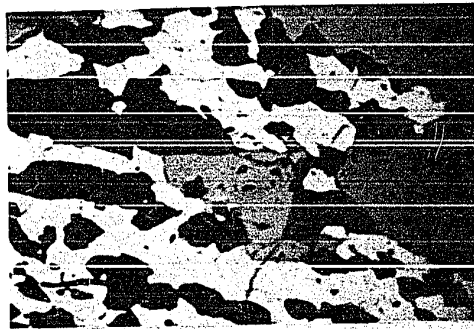


(h) $\text{---} 1000\mu$

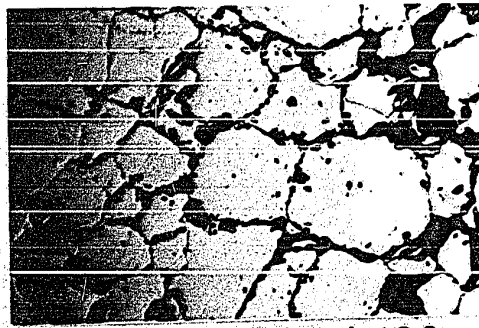
PLATE III.

PLATE IV

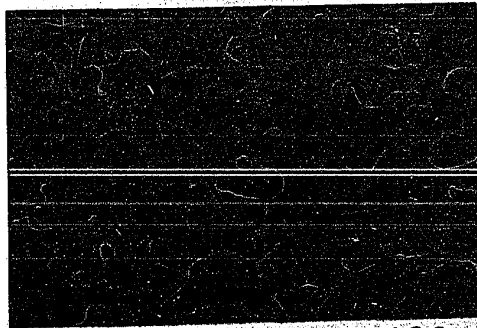
- (a) Banded texture of magnetite (light grey) - titanhematite (white) - apatite (dark grey) ore. (Sample 54)
- (b) Unfoliated, granular magnetite ore. (Sample 237)
- (c) Granular titanhematite ore showing slight foliation. (Sample 2)
- (d) Cataclastic texture in magnetite ore. (Sample 267)
- (e) Sphene-rimmed titanhematite (left) in ore-rich hornblende gneiss. Magnetite grain (right) is free of such rims. (Sample 272)
- (f) Elongated pod-shaped grains of magnetite (grey) in titanhematite (light grey). (Sample 265)
- (g) Martitisation of magnetite (light-grey). White is martite. (Sample 298)
- (h) Bands of sphene-rimmed titanhematite in ore-rich amphibolite. Bands are parallel to the foliation in the rock. (Sample 252)



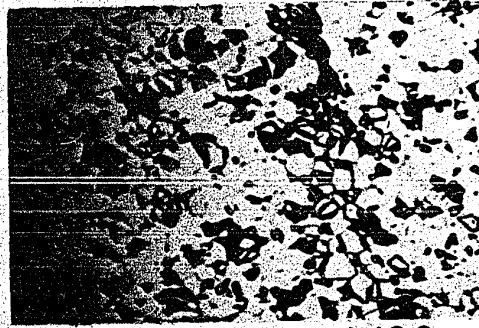
(a) 200 μ



(b) 400 μ



(c) 400 μ



(d) 400 μ



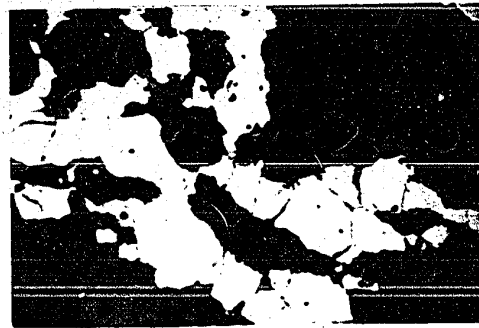
(e) 100 μ



(f) 400 μ



(g) 150 μ



(h) 500 μ

PLATE IV.

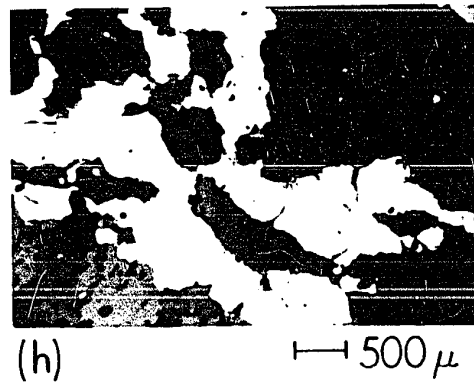
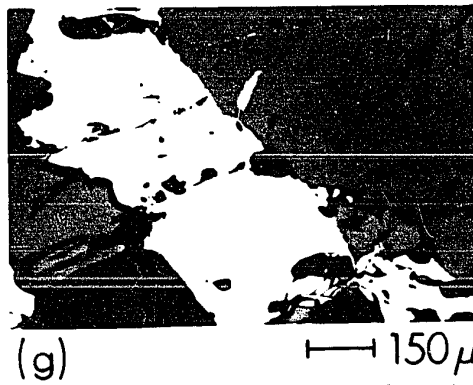
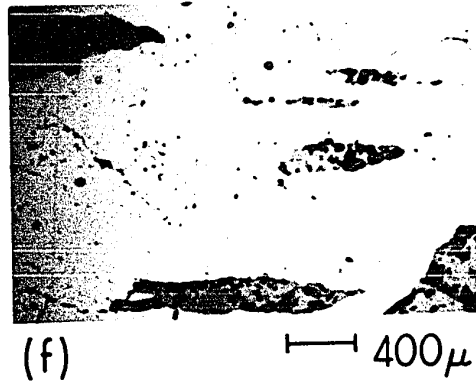
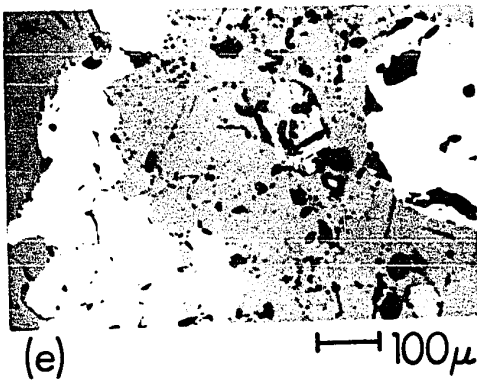
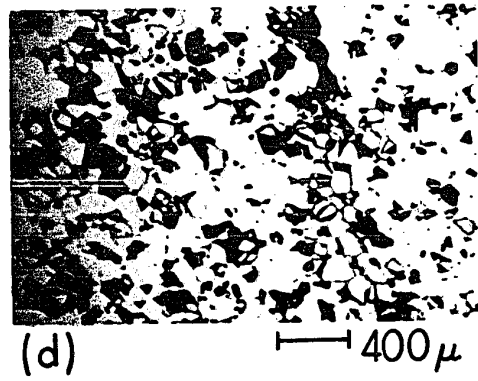
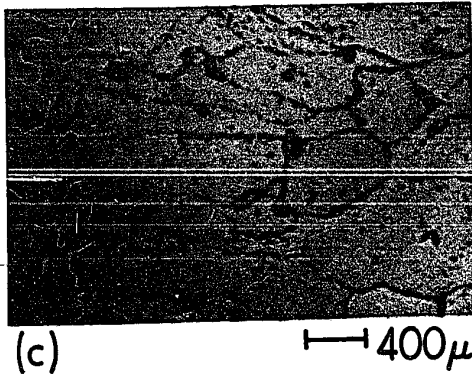
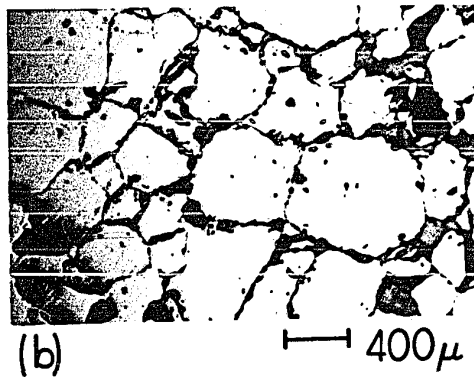
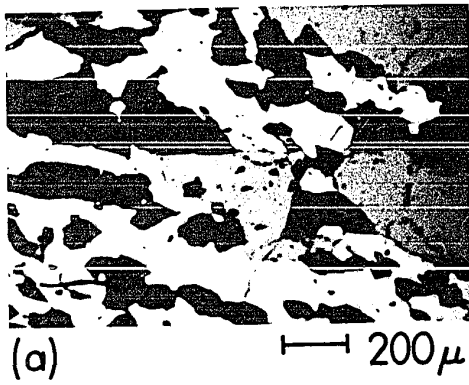
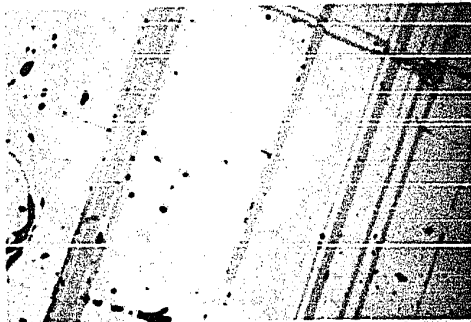


PLATE IV.

PLATE V

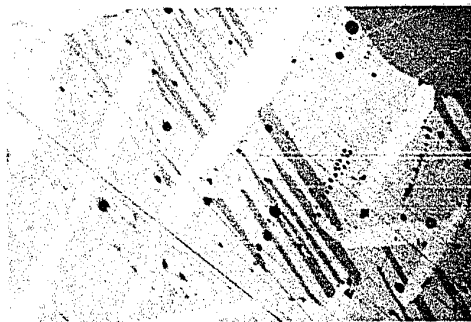
- (a) & (b) Polysynthetic twinning in coarse titanhematite grains. (Samples 40 and 141)
- (c) Hemo-ilmenite lamellae in magnetite from a hyperite. Partially crossed polars. (Sample 211)
- (d) Skeletal hemo-ilmenite of a metagabbro. The bulk of the magnetite host has been leached out, leaving hemo-ilmenite lamellae that outline the shape of the original magnetite. (Sample 145)
- (e) Extremely fine, discontinuous lamellae of hematite (light dots) in ilmenite. Note that the lamellae disappear in the vicinity of the magnetite (Mt) - ilmenite contact. (Sample 181)
- (f) Well-developed crystal of pyrite (greyish white) in chalcopyrite (grey). (Sample 287)
- (g) Pyrite (Py) replacing chalcopyrite (ch); the shape of the original chalcopyrite grain remains unchanged. Grey rim around the pyrite is goethite. (Section provided by R. Mitchell, from sulfide-rich section of ore body).
- (h) Vein of bornite (bn) extending into magnetite (Mt). To bottom left, bornite-chalcopyrite intergrowths. (Same section as in (g) above).



(a) 200 μ



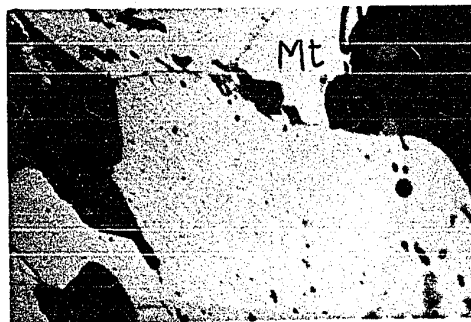
(b) 400 μ



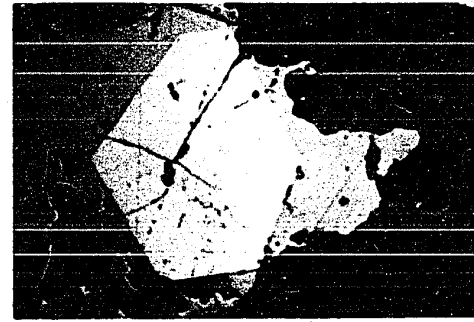
(c) 50 μ



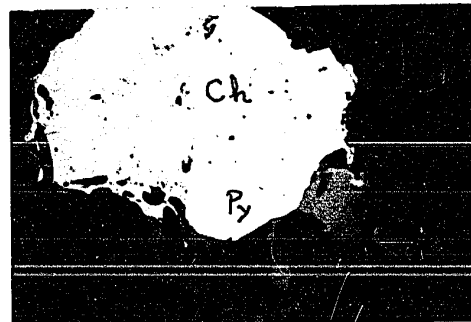
(d) 300 μ



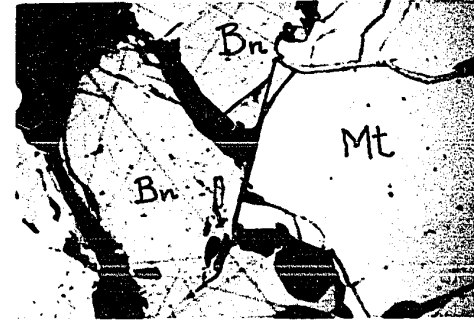
(e) 50 μ



(f) 100 μ



(g) 150 μ



(h) 300 μ

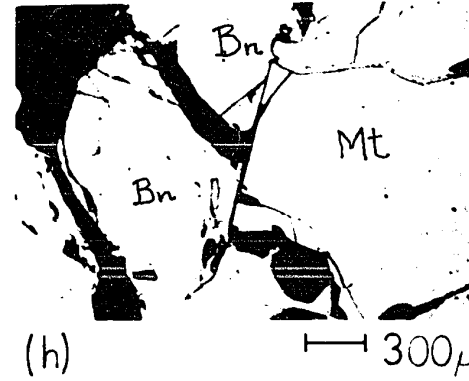
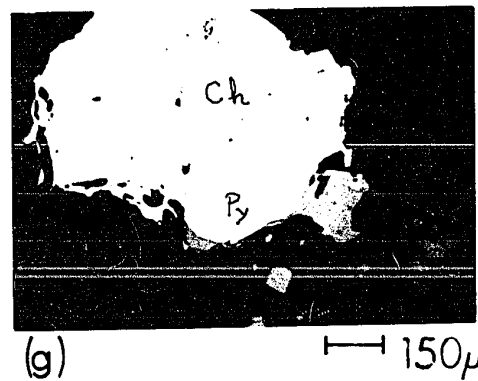
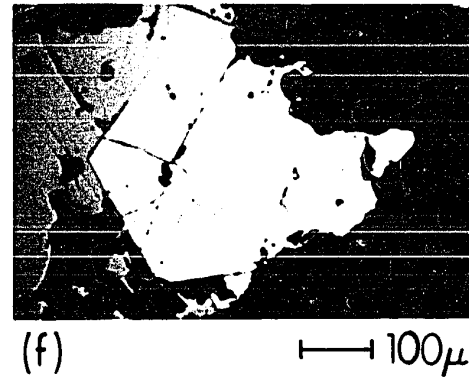
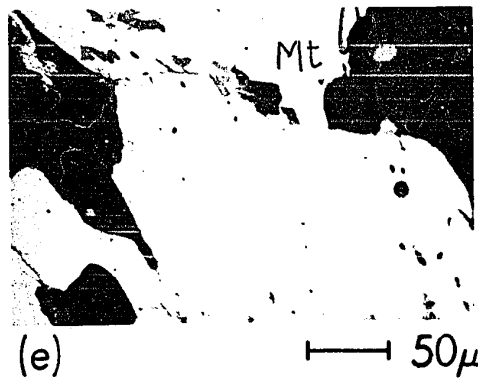
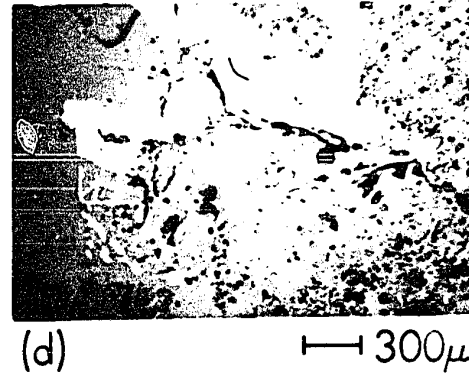
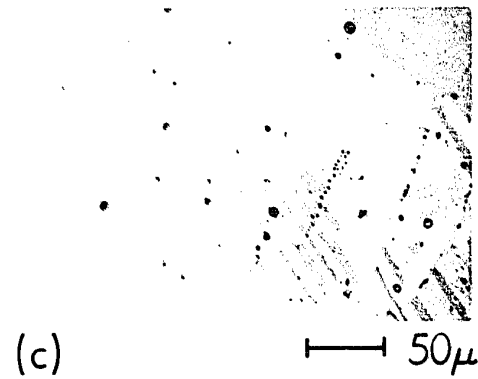
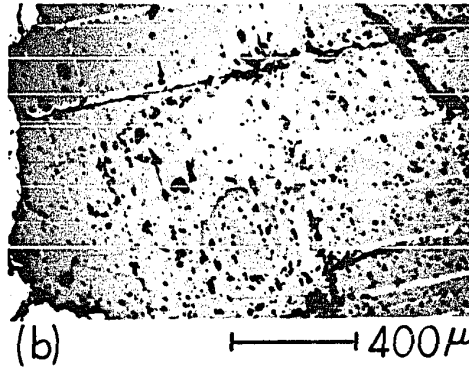
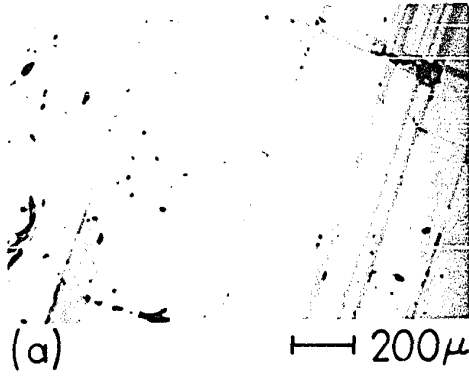
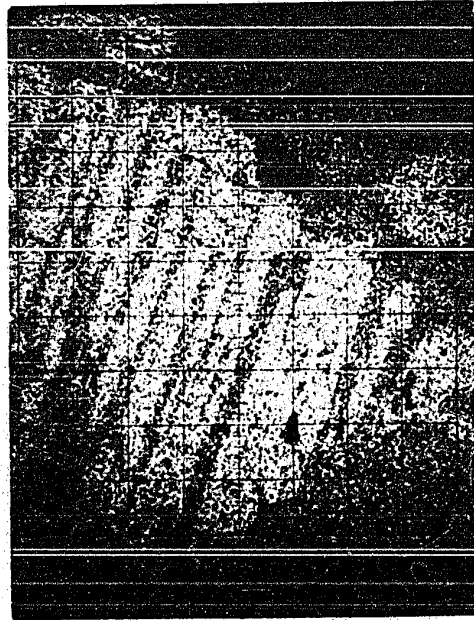


PLATE VI

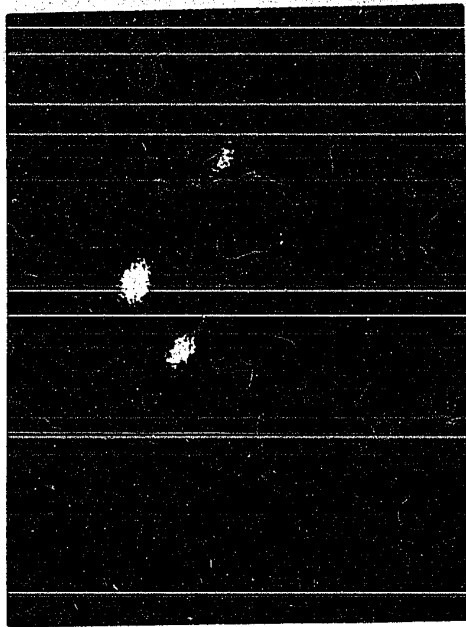
- (a) & (b) Ti $K\alpha$ and Fe $K\alpha$ images (respectively) of part of a magnetite grain with hemo-ilmenite lamellae in contact with hemo-ilmenite grains. Note that the lamellae are connected to the outer hemo-ilmenite grains in (a), where both are indicated by the white areas. (Sample 181)
- (c) An Al $K\alpha$ image of part of a magnetite grain showing blebs of hercynite (white patches) in magnetite (black). (Sample 54)
- (d) A Ti $K\alpha$ image of a sphene-rimmed titanhematite grain. Note that some small, Ti-rich patches occur within the titanhematite. (Sample 252)



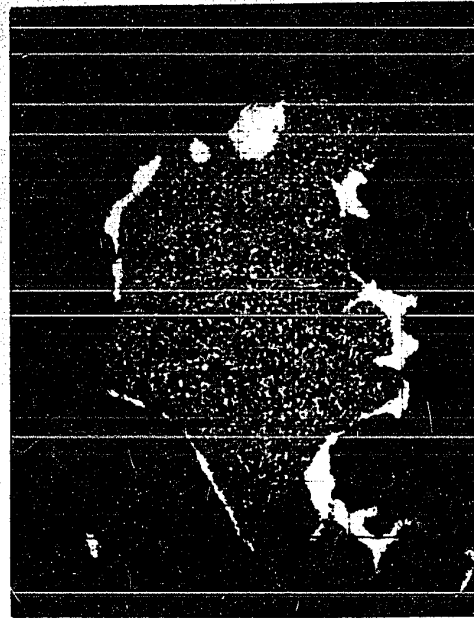
(a) $\text{---} 50\mu$



(b) $\text{---} 50\mu$



(c) $\text{---} 10\mu$

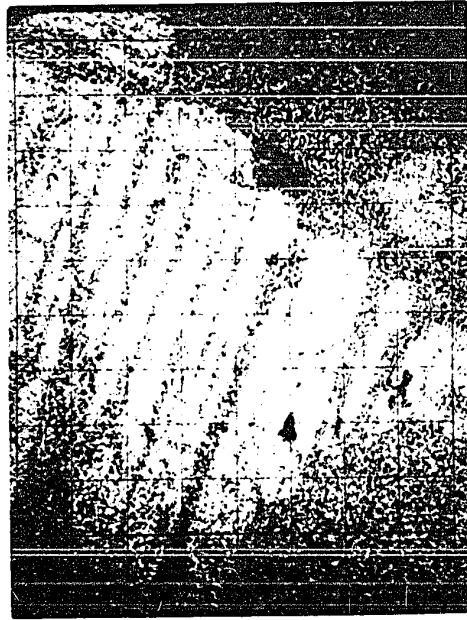


(d) $\text{---} 30\mu$

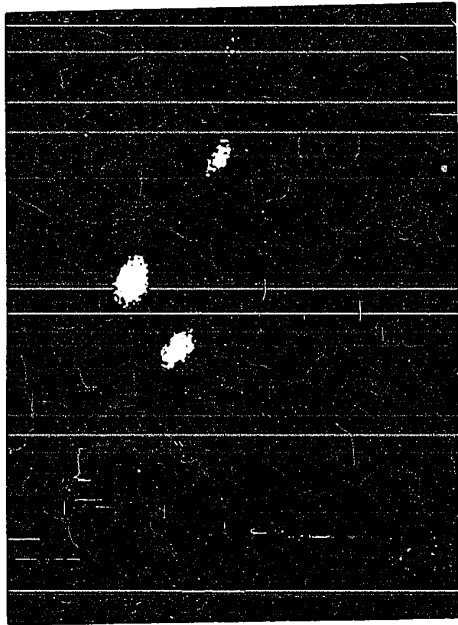
PLATE VI.



(a) $\text{H} \text{---} 50\mu$



(b) $\text{H} \text{---} 50\mu$



(c) $\text{H} \text{---} 10\mu$



(d) $\text{H} \text{---} 30\mu$

PLATE VI.