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A GEOLOGICAL INVESTIGATION OF THE COBALT - FAHLBANDS
OF THE MODUM AREA, NORWAY.

A thesis submitted for the Degree of
Doctor of Philosophy
in the
University of Durham

by

John Blundell Gammon

Grey College

July 1964



Abstract

The cobalt deposits of Modum, Norway, which were worked from 1772 to 1898, are described. The geology of the area is discussed with particular reference to the Modum Formation which forms the host rock for the greater part of the cobalt mineralisation. The petrography and field relations of this formation suggest that it represents a series of ancient sediments which have been metamorphosed to the upper amphibolite facies of regional metamorphism.

The origin of the word 'fahlband' is discussed and a suggested definition of the term is given. The fahlbands of the original locality of Kongsberg are compared with those present in the Modum area. It is postulated that the fahlband zones represent metamorphosed sulphidic black muds.

Olivine gabbro bodies have intruded the Modum Formation; the mineralogical changes occurring on amphibolitisation of the gabbros are discussed. From a review of the field relations, petrography and chemistry of the amphibolites in the region, it is concluded that some, at least, of these rocks are derived by metamorphosism of basic intrusives.

The cobalt mineralisation occurs as a low grade impregnation in the amphibolites and metasediments. The structural controls of the mineralisation are discussed and a possible zonal arrangement of the cobalt minerals is recognized. Optical properties and textural relations of the cobalt minerals have been determined. Possible genetic hypotheses are reviewed and it is concluded that the deposits probably represent an occurrence originally of the Cobalt, Ontario type which has undergone high grade metamorphism.

The mineralogy of the main ore mineral cobaltite (CoAsS) is considered with particular reference to an order-disorder transformation which occurs in the vicinity of 850°C. This study leads to the conclusion that the crystal structure and type of twinning in cobaltite are useful criteria for the recognition of metamorphism in ore deposits.

The chemical and structural relationships of the minerals in the system FeAsS-CoAsS are also discussed and a method for the determination of cobalt in arsenopyrite is given.

Introduction

The cobalt deposits of Modum are situated in the south-east of Norway in country of considerable natural beauty. The area is at the confluence of several rivers, draining the mountainous region of central Norway, which unite to form the Drammen River before entering the sea by way of the Oslo fjord. The area exhibits a wide variety of geological phenomena, and has been the scene of mining operations for more than three centuries. It has nevertheless received very little attention from geologists. Some reconnaissance studies of the geology have been made, but only the immediate vicinity of the Kongsberg silver mines has been subjected to really detailed investigation. The principal object of this work is to record the results of a detailed study of the geology of the Modum cobalt deposits. In addition, a comparison is made with similar deposits at Kongsberg and in central Sweden, based upon reconnaissances made by the author, and upon the writings of former workers. New aspects of the mineralogy of cobaltite (CoAsS) are also recorded.

The work was carried out during the tenure of a NATO Science Studentship. Part I records the results of field and laboratory investigations into the geology of the Modum cobalt deposits, and concludes with a comparison with similar deposits and possible genetic hypotheses. In Part II the mineralogical studies on cobaltite and related minerals are reported.

One full season was spent in the field, from May 1962 to September 1962. In addition, six weeks in the autumn of 1961, and

two months in the spring of 1963 were spent in Modum. The rest of the 1963 field season was spent on reconnaissance studies at Kongsberg and at Swedish cobalt localities. All the important cobalt workings in Modum were visited and detailed maps, on a scale of 1:1000 were made of their environs. Underground mapping, on a scale of 1:500, was carried out at the Skuterud cobalt mines. Structural data was recorded from all localities and a total of 857 hand specimens were collected. The periods between the field seasons were spent in laboratory examination of material from the area, at the University of Durham and at the University Geological Museum, Oslo.

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PART I
THE GEOLOGY OF THE MODUM COBALT DEPOSITS

INTRODUCTION

Location and accessibility

The Modum district is situated in Buskerud Fylke (County), 45 kilometres west of Oslo. The centre of the area of interest lies, approximately, at $60^{\circ} 2'$ N. Latitude and $0^{\circ} 51'$ Longitude. The mining district can be reached by train from Oslo, via Drammen and Vikersund, to Sysle station on the Krøderen railway, and thence about three kilometres by road. The nearest town is Hønefoss, 20 kilometres by road or train in a north-north easterly direction. There are two good road connections between Modum and Oslo, a northern route along Lake Tyrifjord via Hønefoss and a southern route via Drammen. Each route is about a 100 kilometres. Figure 1 shows the location of the area studied.

The Kommune (administrative district) of Modum comprises the parishes of Heggen, Nykirke and Snarum. It covers an area of about 500 square kilometres of which 300 square kilometres are pine forest. There is a considerable timber and wood pulp industry in the district, and several hydro-electric power stations with a total capacity of more than 30,000 kw. In the area there are two parallel cobalt ore bearing zones, from 50 to 400 metres wide, both extending about 10 kms. in a north - south direction.



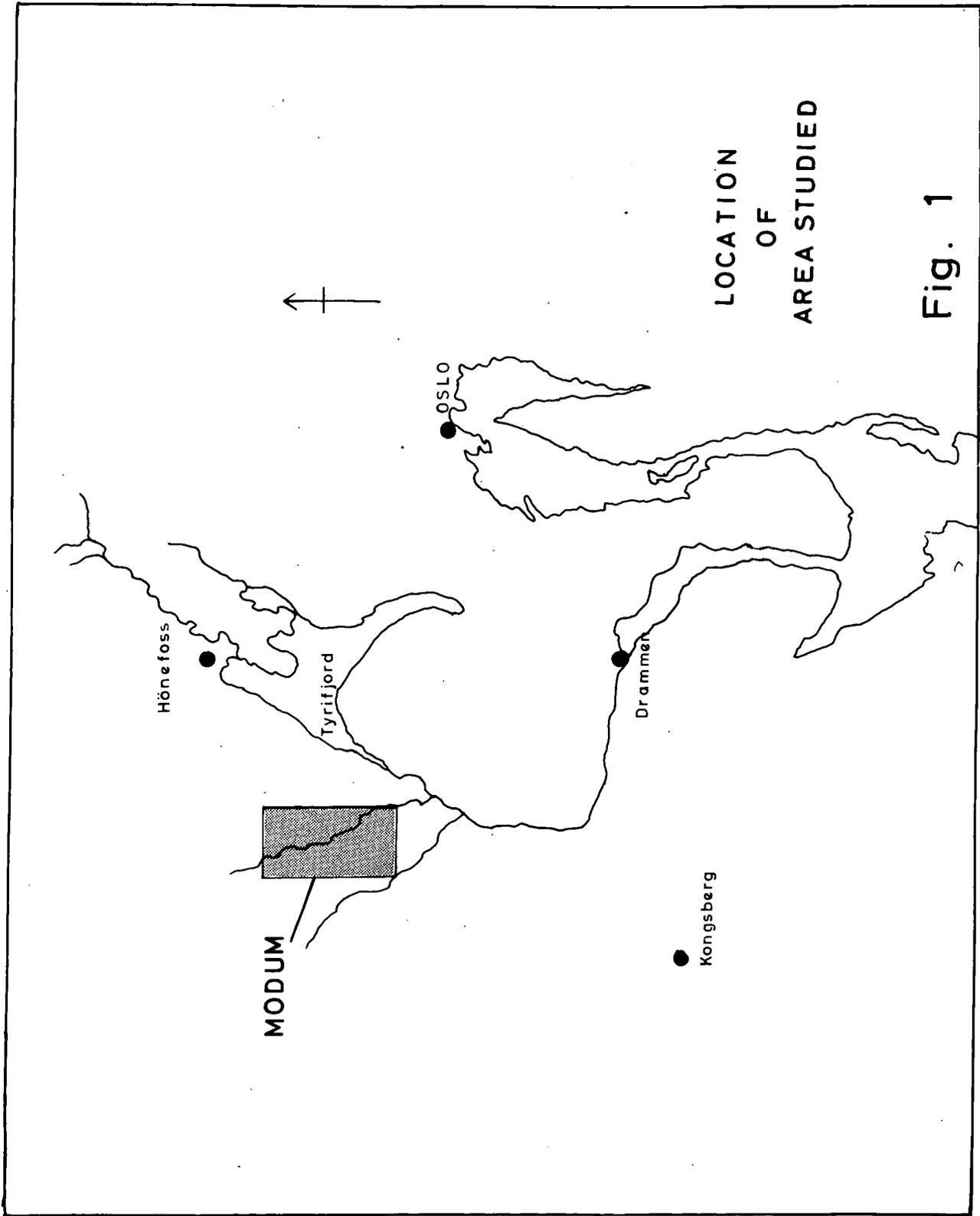


Fig. 1

Topography and exposure

The two mineralized zones are separated by the Snarum River, which here has an altitude of about 100 metres above sea level. Most of the old mines are on the western side of the river, only a few smaller ones having been opened in the eastern mineralised zones. The western mines lie along a forested ridge, which is cut by several cross valleys, dividing the Snarum Valley from the Sigdal Valley to the west. The highest point, on this ridge, is some 460 metres above sea level, so that the maximum relief in the area is about 360 metres. The highest points on this ridge are underlain by basic rocks. To the east of the Snarum River, the area is represented by a rolling surface of low relief, extending eastwards to the Tyrifjord. This surface is the pre-Cambrian peneplane, which is here dissected by subsequent valleys of small relief.

The rock exposure in the area studied varies considerably. Much of the western ridge has been swept clean by the Pleistocene glaciers so that exposures can be fairly good. The Snarum Valley is filled with alluvium and glacial drift and exposure is very poor indeed. Much of the area is thickly forested and the degree of exposure varies considerably, with small drift-filled depressions, separating ice scoured outcrops. The outcrops are generally quite smooth and frequently covered with 'reindeer moss'. The opencast workings, afford extremely useful fresh outcrops in a third dimension, enabling the measurement of the attitudes of minor folds, contacts, etc., and the collection of hand specimens.

Purpose and method of approach

The purpose of this investigation was to determine the age and genesis of the cobalt deposits. Since no single approach provides adequate information, a number of methods, which complement each other, were applied.

The structural setting of the deposits, has been investigated by conventional geological mapping, both at the surface and in the accessible underground workings. A tectonic synthesis is attempted by the statistical analysis of small scale structures, leading to a reconstruction of the geometry of the major structures.

The petrography of the host rocks is described and textural interpretations of the mineral associations present are attempted. These studies lead to suggestions regarding the pre-metamorphic history of these rocks.

The amphibolitic rocks were investigated in the field and laboratory by various techniques to determine whether they represent metasediments or a suite of metamorphosed basic rocks.

The mode of occurrence and texture of the cobalt ore is described, and its relations to the sulphidic-graphite schists and uranium mineralization in the region, is discussed.

Comparisons are made with similar rocks and ores occurring at Kongsberg and in central Sweden, and the validity of the term 'cobalt-fahlband', as applied to this type of deposit, is discussed.

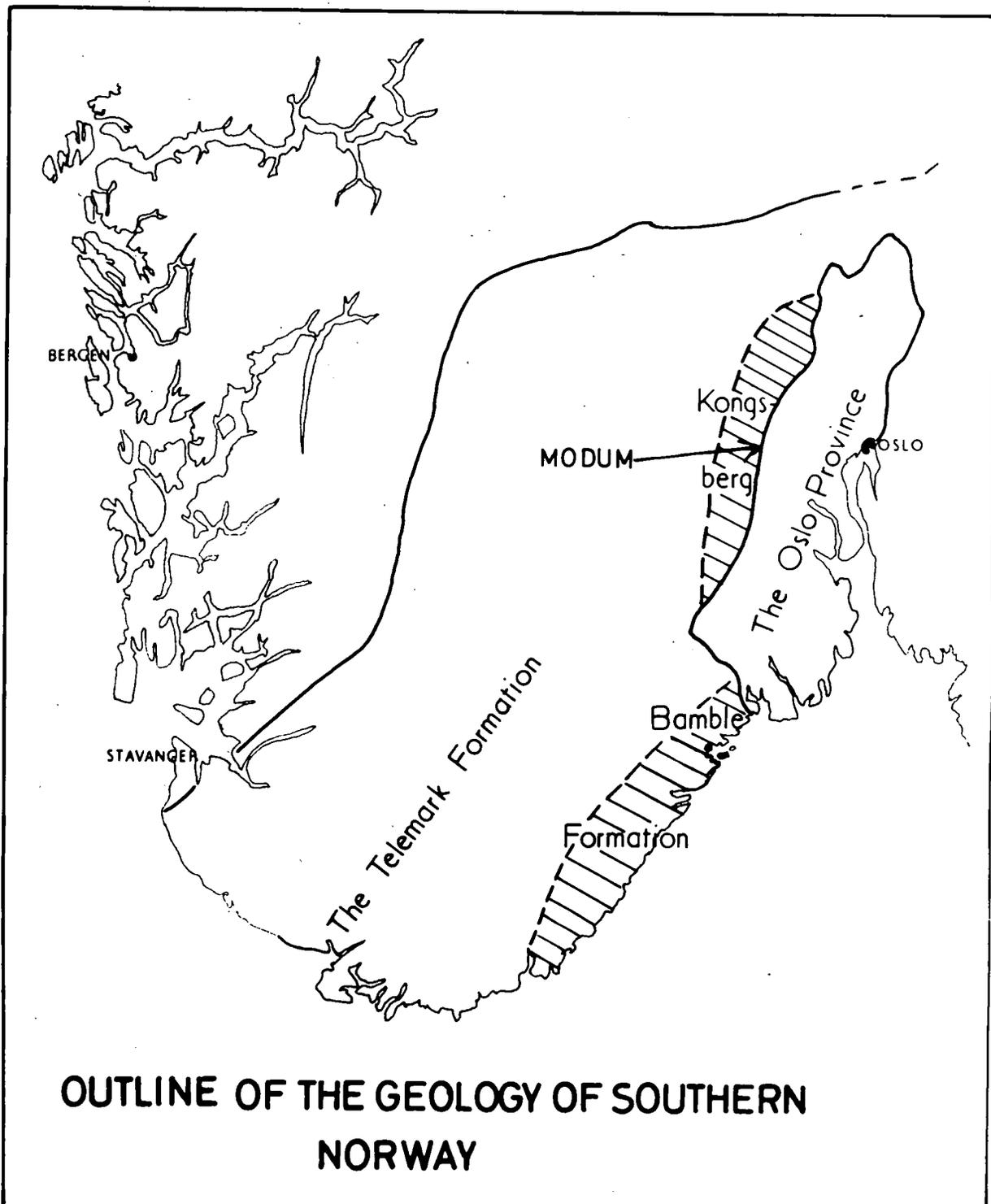
Finally, possible genetic hypotheses are considered in relation to the data obtained from the various investigations.

Regional geological setting

The broad geological relationships can be seen in Figure 2 . The area lies within the southern Norwegian Precambrian shield, which is largely occupied by rocks of the Telemark Formation. The supracrustal and granitic-gneiss area of this Formation is terminated by a prominent mylonite zone to the south and east. This 'Great Friction Breccia' separates the Telemark Formation from rocks of the Kongsberg - Bamble Formation, divided by a wedge of overlying Palaeozoic rocks into a northern and southern area. (A. Bugge 1929, Selmer - Olsen 1950).

The Modum district is situated within the Kongsberg - Bamble Formation, a complex of highly metamorphosed crystalline schists, gneisses and migmatites associated with subordinate amounts of plutonic rocks. A complicated tectonic style with steep fold axes and plastic deformation has developed. Most primary structures have been obliterated by repeated metamorphism in more than one orogenic period. (A. Bugge 1929, J.A.W. Bugge 1943).

Jens Bugge (1960) distinguishes between an old complex of diverse rocks, of which a number are of undoubted supracrustal origin, formed before an older orogenic period, and a younger complex, consisting of migmatites, charnockites, pegmatites and granites, formed during and after a younger orogenic period. The two orogenic periods are separated in time by the intrusion of widely distributed basic rocks, the hyperites. The Telemark Formation may be attributable to the later orogeny.



**OUTLINE OF THE GEOLOGY OF SOUTHERN
NORWAY**

Fig. 2

The Kongsberg - Bamble Formation has been correlated with the 'Leptite Formation' of central Sweden. The younger complex being correlated with the Gothian rocks in Sweden and the older complex with a pre-Gothian (Svionian) cycle. (A. Bugge, 1922).

Radiometric age determinations on rocks from the Kongsberg - Bamble Formation range from 760 - 1040 million years. (Neumann, 1960). The Herefoss granite, an intrusive granite of the younger complex, gave a date of 900 million years (Elders, 1963). Most of these determinations were made on biotites using the K-Ar method, it is possible that the rocks were effectively redated by the younger orogenic episode.

Previous Investigations

Cobalt mining was carried out in Modum between 1776 and 1898, and many travellers visited the mines during this time. The earliest descriptions of the mines contribute little of geological interest. Fabricius (1779) describes the appearance of the ore outcrops at the site of discovery. Strøm (1790) comments on the decline in ore grade at a depth of 30 - 35 metres. Leopold von Buch (1810) and Hausmann (1819) gave the first full geological descriptions of the deposit.

During the first half of the nineteenth century the mines were flourishing and several good geological descriptions from that time are available: Daubrée (1843), Schmidhuber (1847), Durocher (1849) and Müller (1857). The manager of the pigment factory at Haugfoss, Th. Scheerer, who later became Professor of Mineralogy at

Freiberg Bergakademie, published some of the first analyses of the ore minerals skutterudite and cobaltite. (Scheerer, 1838, 1840, 1849; Scheerer and Wohler, 1838; Scheerer and Francis, 1840). The overseer at the mines, Karl Bøbert, published the only extensive contemporary geological descriptions of the deposits (1846), and compared them with the Kongsberg fahlbands (1848) and Swedish cobalt localities (1832).

In the latter half of the nineteenth century the mines declined in importance, and little of significance was added to the older descriptions, most writers offering a compilation from previously published sources. (Stelzner 1866, Helland 1879, Kjerulf 1850, Graff 1883, Beck 1899, Krusch, Beyschlag and Vogt 1913).

Reconnaissance geological studies of the Kongsberg - Bamble Formation (A.Bugge 1936, J.A.W.Bugge 1943) included observations on rocks from the region, and part of the area is described in the Norwegian Geological Survey (N.G.U.) publication covering the Flesberg and Eiker regions (A.Bugge 1937).

The results of the German investigations of 1940-43 are summarised in Rosenqvist (1949), the only published, modern, description of the mines. Unpublished reports were prepared for the E.C.A. Marshall Plan (Adamson 1950), for the Kongsberg silver mines (Heltzen 1952) and for the N.G.U. (Vokes 1957). A radiometric survey of the Modum area is described in a report of the N.G.U. (Van Autenboer 1957).

Jøsang has recently completed a study of the geology of the region for the N.G.U., which is now awaiting publication. The author has had access to the manuscript maps which accompany this publication through the kind co-operation of Mr. Jøsang.

HISTORY OF MINING

Discovery

The original discovery of the ore is described by J.H. Voss, in an early work probably written between 1783 and 1790. He tells of one Ole Withloch, a miner from Kongsberg, who, having been dismissed from the silver mines sought new deposits of silver to make his fortune. In October 1772 he collected samples from Skuterud, which he thought contained large quantities of silver. He had these analysed at Kongsberg, where the ore was found to be rich in cobalt. In return for describing the location of his discovery, he received a handsome monthly allowance and the honorary title of 'Bergmeister'.

Haugfoss pigment factory

Laboratory tests were made at Kongsberg and it was found that a good quality blue pigment could be produced from the ore, so plans went ahead for large scale development. Several further ore bodies were discovered in 1773, and in April 1776 the King of Denmark-Norway gave directions for the compulsory purchase of the mining area, and ordered the building of a factory to produce the blue pigment. Haugfoss was selected as the site of this factory, and work began there in 1777. The crushed and washed concentrate was brought by pack horse, from the mines, to the factory, which lay 10 kms. south of Skuterud. At Haugfoss, the ore was roasted with quartz and potash, to give the cobalt pigment or smalt (potassium cobalt silicate). Most of the blue pigment was sold to

the Royal Danish Porcelain works in Copenhagen, some was also sold in Amsterdam, in England and domestically in Norway.

Early mining operations (1778 - 1814)

Several German and Danish travellers visited the area in its early days and described what they saw in their diaries. (Gram 1776, Fabricius 1779, Jonge 1779, Bradt 1801, Mumssen 1788, Strøm 1790, Sneedorf 1790, von Buch 1810, Hausmann 1819, Otte 1835, Reissegger 1848 and Durocher 1849). These accounts are usually only concerned with the festivities relating to their visit, and a description of the factory at Haugfoss. Some of them made the pilgrimage up to the mine workings however and described a little of what they saw.

The first opencast operations were on the present site of the South Mine at Skuterud, and Fabricius (op.cit.) describes how work had just begun on an adit at a lower level (probably Bennecke Adit). Strøm (op.cit.) describes ore bodies about 5 metres wide at the surface which tail out completely at a depth of about 30 metres. So operations at depth, must have been well under way, at the time of his visit, in 1790.

Norwegian State ownership (1814 - 1822)

On May 17th, 1814 the Norwegian State severed the links of their 500 years subservience to Denmark and became internally self-governing under the Swedish Crown. The export of blue pigment and arsenic from Modum was one of the main sources of revenue for the new state, and in the summer of 1815 the pigment factory and ore reserves were used as security for obtaining credit from abroad. The Norwegian government denationalised the company in 1822 and sold it to the Wegener Brothers of Oslo.

The 'flourishing' period (1822 - 1849)

The twenty years that the operations were directed by the Wegener Brothers were the most flourishing in the history of the mines. At this time more than 1,000 men were employed at the mines and pigment factory. In 1849 the mines were sold by auction. In the description of the deposits prepared for this auction, (Lammers 1849a, 1849b), the development work of the first 50 years of operations are described. Four main groups of mines had been developed, the South, Central and North Mines at Skuterud, and the Middagshville Mines at Saafstad.

The South Mines had been the main producer to that date, reaching a depth of 60 metres, with access by the Bennecke Adit which was provided with a trackway. Underground development had just begun at the Central and North Mines, with the driving of the Klara Adit a length of 5,360 feet by fire setting. This underground development exposed very rich ore, Lammers (op.cit.) considered these mines to be the most promising for future development. The Middagshville Mines were opened in 1843, but the ore was not rich enough to justify further development.

Period of decline (1849 - 1890)

The mining of cobalt and the production of pigment was taken over in 1849, by an English firm, Goodhall & Reeves. They concentrated on the Central and Northern Mines at Skuterud and opened several large stopes below the level of Klara. They sold the mines in 1856, after a rock fall, in one of the opencasts had caused the death of six miners.

From this date, until the end of mining, the operations were carried on by a German firm, "Sächsischer Privat-Blaufarbenwerkverein, who were already engaged in similar operations at Schneeberg in Saxony (Vogt 1899). The cobalt concentrate was sent to Saxony for processing. The discovery of the synthetic blue pigment, ultramarine, in 1848 and the discovery of new deposits overseas, sounded the death knell of the cobalt blue industry in Modum.

In 1878 the mines were still among the largest in Norway, but after this time the ore lenses became more and more difficult to follow. A certain amount of diamond boring was done from the Klara level, but met with no success (Guldbrandsen 1952). After 1894 no new development work was done. The mines were kept open in the hope that new ore would be found. At this time the Ludwig Eugen Level was opened up for transportation of ore broken from the stopes below the Klara Level, and for exploration in depth. No new ore lenses were found, however, and the mines were finally closed in 1898. (Lindemann 1932, Tank 1952, Gottschalk 1914).

The twentieth century

In 1919 the installations at Haugfoss were sold to the Modum Kommune for conversion into a saw-mill. The mines themselves lay derelict for 50 years, until in 1941, the Germans started quite extensive investigations. They were particularly interested in the deposits, as they were cut off from their normal cobalt sources, at this time. The operations were discontinued in 1943, the ore grade being too low to justify a re-opening of the mines. The

same conclusions were reached in 1952 when the deposits were investigated on behalf of the Kongsberg Silver Mines (Heltzen 1952).

Snarum Verk (1822 - 1849)

From 1822 to 1849, a separate company, 'Snarums Blaafarveverk', carried out operations in the area to the north. They mined ore from the Døvikollen and Svartefjell Mines and transported it to a pigment works built near the village of Vestre Snarum. A long legal argument ensued because the Modum company claimed that these mines rightfully belonged to them, although they were not actually working them at the time. (Lindeman 1932). Difficulties with the poor quality of the ore and processing of it, soon led to the failure of the company. It went bankrupt in 1849 causing great hardship to its employees.

PAST PRODUCTION

Production statistics are incomplete, it is therefore difficult to estimate the grade of the produced ore, and the amount of cobalt recovered. Production figures were given largely in terms of tønder (barrels of 0.25 m³ capacity), and assays were made only on the cobalt concentrate.

The mined rock was concentrated by hand sorting, to give a first concentrate. Crushing and washing of this concentrate produced a second concentrate with about 30% cobalt. This concentrate was sent to the pigment factory for the production of cobalt blue. The production statistics for 1840, a year of high production, are shown in table 1.

Table 1

Cobalt production at Modum, 1840

Mined rock		Hand picked concentrate		Washed concentrate		Assayed cobalt
Tønder	Metric tons	Tønder	Metric tons	Tønder	Metric tons	Metric tons
200000	125,000	6,000	4,000	120	80	25

In that year therefore, the crude ore contained only about 0.02% recoverable cobalt, this was increased to 0.6% by hand picking and 30% by crushing and washing. There is no information on the actual grade of the crude ore but it was probably between 0.06% and 0.1% cobalt. Recovery was probably better in later years, Kjerulf (1850) mentions that recovery that year was 0.07% cobalt,

probably due to selective mining of higher grade ores, containing about 0.1-0.2% cobalt.

Adamson (1950) has estimated the total amount of cobalt contained in cobalt products, produced at Modum from 1776 to 1898, as about 1,000 tons. Rosenqvist (in Adamson 1950) estimates the amount of ore and rock material taken out of the Modum mines, on the basis of the size of the old workings, as about 10 - 12 million tons.

On the basis of these figures the overall ore grade would be about 0.001% cobalt. However Adamson (op.cit.) has estimated that the production of ore that underwent concentration by hand picking can hardly have been more than 3-4 million tons, the grade of this would thus be about 0.02 - 0.03% cobalt.

GENERAL GEOLOGY OF THE MODUM DISTRICT

A study of the geology of the district has been recently completed by Jø/sang, the results of which are now awaiting publication. Manuscript copies of Jø/sang's geological map of the Modum district were available to the present investigation. The following summary, is based upon information contained in these maps, and upon the author's own field observations in the area. The author has not had access to the description which will accompany these maps, when they are published, and takes full responsibility for all statements made.

Tectonics and metamorphic grade

The dominant strike in the area is north-south, with the dip varying little from vertical. The various rock units have lenticular outcrops, and cannot, usually, be traced for long distances along the strike. Fold closures are rare, and large scale structures are difficult to decipher. The observed mineral associations suggest that the region has undergone metamorphism of the upper almandine-amphibolite facies.

Rocks of the 'Older Complex'

The area consists mainly of rocks of the 'Older Complex' (Bugge 1943), a series of banded gneisses being exposed from the east of the mineralized zone to the westward margin of the Oslo "graben". The area to the west, including the mineralized zone, consists of rocks, which are still recognisable as metasediments.

The Modum district is characterised by a greater abundance of quartzites than are normally encountered in the Kongsberg - Bamble Formation (Bugge, op.cit.). Very pure quartzites are rare, well-oriented biotite grains usually giving the rocks a foliation.

These quartzites can be traced laterally and vertically into more pelitic rocks containing abundant sillimanite. The sillimanite is frequently present in quartz-sillimanite nodules (faserkiesel), which give the rock a marked lineation. With increasing mica content, the psammitic rocks pass by gradation, into true mica schists, containing muscovite and phlogopitic biotite.

Horizons containing copper and iron sulphides with graphite are rather common in these rocks, rust coloured outcrops are characteristic of these sulphide-graphite schists, which have been called fahlbaands by Norwegian geologists.

The banded gneisses at the east of the area, are composed of alternating light and dark bands, from several centimetres to a few metres in thickness. The individual bands consist of amphibolitic, biotitic, quartzitic, dioritic and granitic gneiss, with frequent pegmatitic bands some of which show a transgressive relationship.

These rocks probably represent a supracrustal succession although no original features are clearly recognisable. The different bands may have been determined by original bedding,

although the chemical composition of the bands may, in many cases, have changed. Dietrich (1960) has proposed a sedimentary origin, for the banded gneisses at Randesund, in the southern area of the Kongsberg - Bamble Formation.

Olivine gabbros

These rocks are probably the equivalent of the olivine norites (hyperites) found elsewhere in the Kongsberg - Bamble Formation. They form concordant bodies whose intrusive nature is suggested by plutonic texture, fine grained border facies and transgressive features in places.

They are emplaced in the metasediments of the 'Older Complex'. A subsequent plastic deformation, when the hyperites behaved comparatively rigidly, causes them to be elongated with their longest axes sub-parallel to the enclosing gneisses. The central parts of the bodies are usually fresh and massive and they pass outwards, by gradual transition, into schistose amphibolites. A metasomatic alteration has resulted in scapolitisation and albitisation of the gabbros in places.

Rocks of 'Younger Complex'

The only representatives of the 'Younger Complex' in the Modum area are the numerous, large, cross-cutting granite-pegmatites. These are massive and undeformed and transgress the olivine-gabbros. They are thus probably late - or post - kinematic.

Rocks of uncertain age relations

There are a number of common rock-types in the region, the origin of which, cannot be assigned to any particular period.

These include cordierite-anthophyllite gneiss, nodular sillimanite 'granite' and magnesite-serpentine lenses.

Cordierite-anthophyllite gneisses are common in the Modum area, especially in the zone of sillimanitic metasedimentary rocks. Elsewhere in the Kongsberg - Bamble Formation they occur in close connection with gabbroid rocks and amphibolites (Bugge op.cit.); this distribution has caused Brøgger (1934) to suggest that they may be derivatives of a basic magma.

Nodular sillimanite 'granites' occur as inter-banded layers in the surrounding gneiss, never showing intrusive contacts. The occurrence of sillimanite in these rocks is restricted to quartz-sillimanite nodules. They possibly represent metasomatic rocks, formed through a granitisation of older aluminous sediments (Bugge 1943).

Magnesite-serpentine deposits are found as small lenses in the metasedimentary sequence, the largest being at Morud where the magnesite has been worked for refractory furnace linings (Jøsang, 1960).

PETROGRAPHY OF THE MODUM FORMATION

Introduction

The rocks of the 'Older Complex' are divisible, in the northern part of the Kongsberg - Bamble Formation, into three distinct types, each type forming a clearly marked zone running north-south, parallel to the regional strike. (A. Bugge, 1943). The western zone comprises rocks of the Kongsberg Formation ('Kongsbergformasjonen' of Bugge, op.cit.), consisting of dioritic and quartz-dioritic orthogneisses with associated chloritic schists. The central zone, occupied by the Modum Formation ('Modumformasjonen' of Bugge, op.cit.), is characterised by a greater abundance of quartzitic rocks with interbanded amphibolitic and granitic layers. This is followed, to the east, by an outer zone of banded gneisses ('amfibolitisk båndgneiss sone' of Bugge op.cit.). The first two of these zones have been recognised in the southern area, where the Modum Formation is correlated with the Bamble Formation. It is possible that these 'Formations' and 'zones' represent different regional metamorphic facies.

The area of interest is situated on the boundary of the Modum Formation with the outer zone of banded gneisses. The cobalt workings are restricted to rocks of the Modum Formation, the only exceptions being unconfirmed traces of 'cobalt mineralisation', from unspecified localities, in the outer zone of banded gneisses (Bøbert 1846). The present investigation has thus been mainly concerned with the Modum Formation.

Rock classification

The rocks were classified in the field and the laboratory into three main groups:-

1. Siliceous
2. Aluminous
3. Calcareous

All gradations exist between the siliceous and aluminous groups. The mineralogy of the calcareous group is always distinctive. The basis of the classification is the contrast between silic and femic minerals and the fabric they impose on the rock. It was found possible to employ this classification in the field and it forms a useful framework for the detailed petrography.

Table 2

Classification of the paragneisses

SILICEOUS GROUP	Quartzite	>80% quartz
	Siliceous granulite	50-80% quartz
ALUMINOUS GROUP	Felspathic granulite	65% quartz + felspar (>50% quartz)
	Semi-pelite	35-65% quartz + felspar
	Pelite	<35% quartz + felspar
CALCAREOUS GROUP	Calc-silicate granulite	

Within the area described siliceous granulites form the major background unit. Felspathic granulite and semi-pelite are frequently interbanded in the siliceous granulites. Quartzites are quite common and form very obvious features in the field. Pelites and calc-silicate granulites are very restricted in their distribution. The mineralogical variation shown by 33 modal analyses is seen in Figure 3.

Quartzites

Field relations

The quartzites are resistant to weathering and form conspicuous light grey ridges. They are quite widely distributed in the mining area, and vary in thickness from 0.5 - 250 metres. The field relations suggest that they are earlier than the amphibolite bodies, one of which is seen to transgress a thick quartzite in the Skuterud district. (Figure 109). Detailed inspection of quartzite-amphibolite boundary relations, suggests that the amphibolites have intruded the quartzites along bedding planes (Figure 40). To the south-east of Morud, and again just west of V.Spone Kapel, quartzite can be traced along the strike into granodioritic gneiss, suggesting that the latter perhaps formed through the metasomatic alteration of quartzite.

At outcrop, the quartzite is seen to be a grey rock with a granular texture and vitreous appearance. Oriented mica flakes give it a poor foliation which is parallel to a mineralogical banding. The mineralogical banding is produced by thin felspar layers (2-3 mm. thick) alternating with quartz-rich layers.

Figure 3: Quartz (Q) - felspar (F) - mafic (M) diagram
to illustrate the mineralogical variation of the paragneisses
of the Modum Formation.

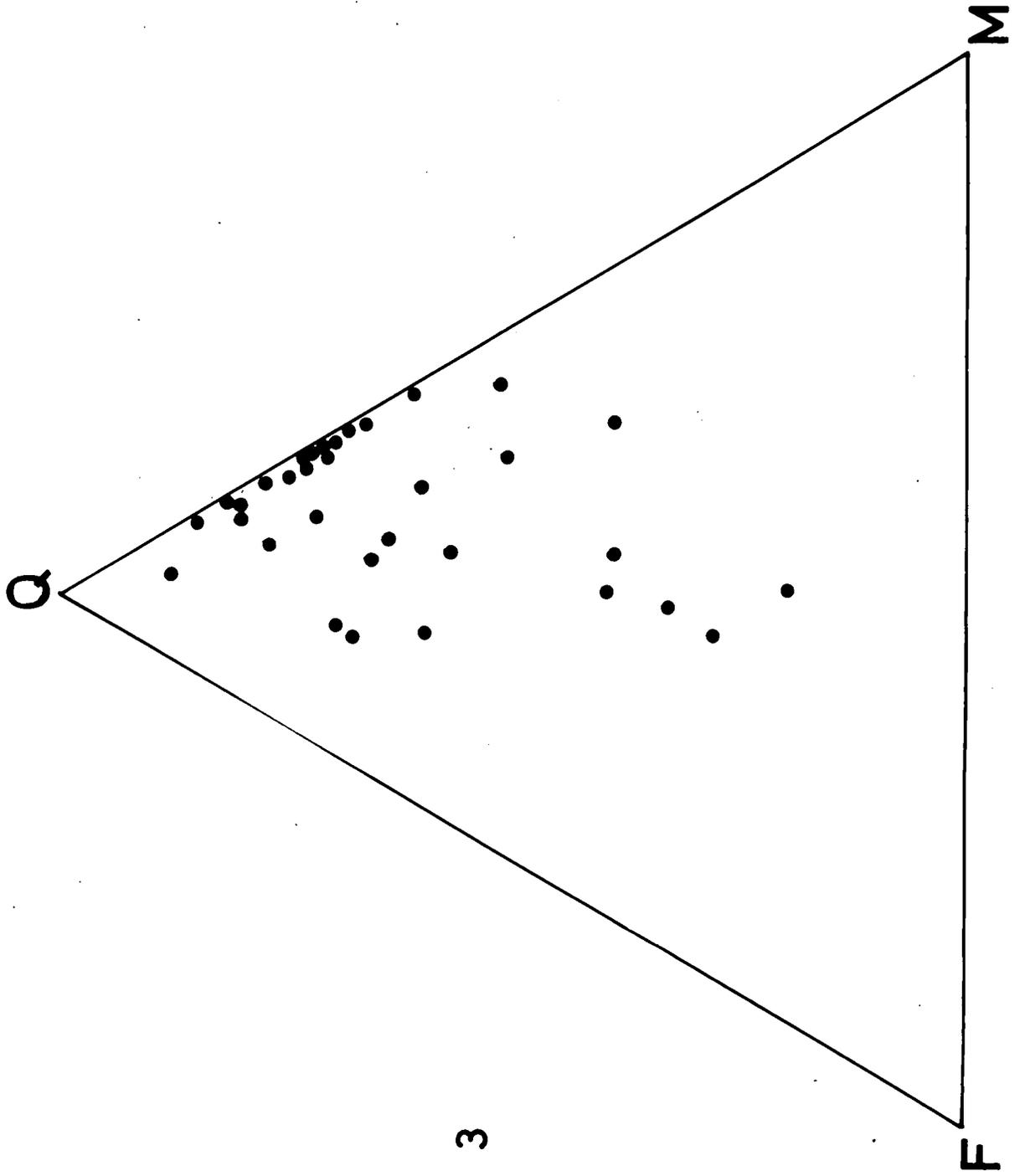


Fig. 3

In addition there is a very marked colour banding, due to slight variations in biotite and tourmaline content. This mineralogical colour banding is considered to represent the original bedding. When examined in detail, it is seen that this banding is not strictly parallel to the local foliation, occasional violent contortions are seen, where the colour bands are thrown into tight minor folds, these are interpreted as being of tectonic origin rather than original sedimentary slump structures.

Dense, monomineralic quartz veins and lenses are restricted to the quartzites, where they are very common. There is little doubt that the quartz has been derived, by segregation, from the surrounding quartzite (Figure 4).

A very characteristic feature of the quartzite outcrops are the numerous, closely spaced, joints. These follow several clearly marked trends (see p.126 for discussion of jointing). The quartz-segregation veins are affected by these joints (Figure 4), a feature which distinguishes them from the later, hydrothermal, quartz veins.

Petrography

All the quartzites are very much dominated by their quartz content, but some variation is seen in the accessory mineral contents. Some typical modes are given in Table 3.

The quartz shows two typical modes of occurrence. The commonest type is found in large grains, ranging in size from 1 to 6 mm., with very sutured margins. This type usually shows a dimensional orientation parallel to the foliation with extremely marked strain extinction (Figure 5). The quartz of this type is

Table 3.

Quartzite Modes.

Specimen number	15	19	461	100	98
Quartz	81.0	80.7	82.3	83.1	86.0
Plagioclase	1.9	2.0	-	-	-
K-felspar	+	-	-	-	-
Biotite	9.7	9.1	-	10.6	8.8
Muscovite	-	-	11.1	-	+
Tourmaline	6.0	2.6	-	1.9	2.0
Amphibole	-	-	4.8	-	-
Sillimanite	0.9	-	-	0.5	0.6
Zircon	+	+	-	-	-
Opaque	0.4	5.6	1.8	3.9	2.6
Plagioclase An content	31	26			
No. of points	854	967	672	1037	798

+ - Present in minor amounts

Localities: 15-1000/240 (Fig.110); 19-1000/205 (Fig.110);
 461- -30/120 (Fig.109); 100-690/260 (Fig.110);
 98-685/265 (Fig.110).

Figure 4: Quartz-segregation lenses in quartzite.

Top: Segregation lenses with many closely spaced fractures, in jointed quartzite. South Mine, Skuterud.

Bottom left: Concordant quartz-vein with replacive apophyses. Forhaabning Adit, Skuterud.

Bottom right: Transgressive quartz-segregation lens, the colour banding in the surrounding quartzite converges on the lens. Forhaabning Adit, Skuterud.

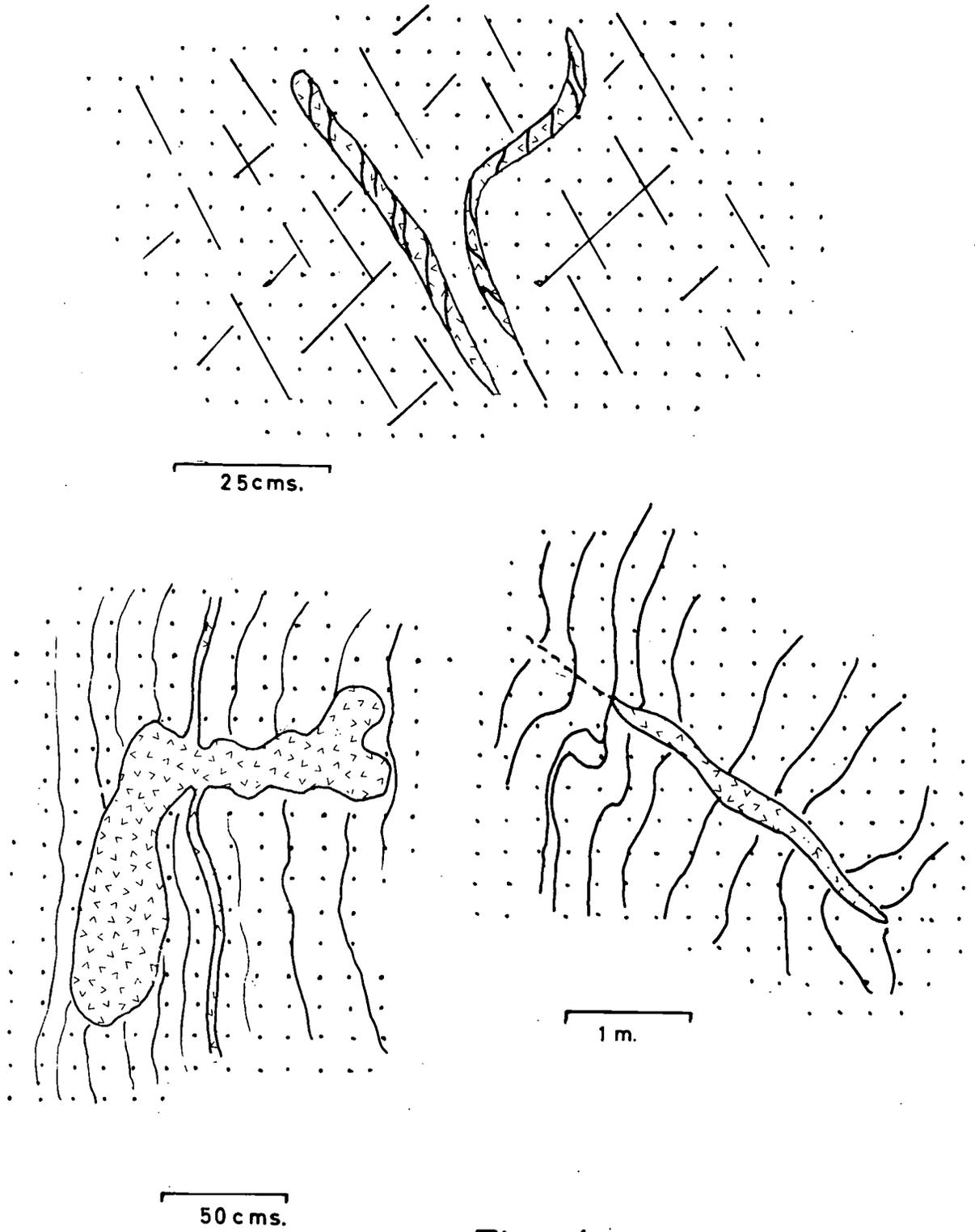


Fig. 4

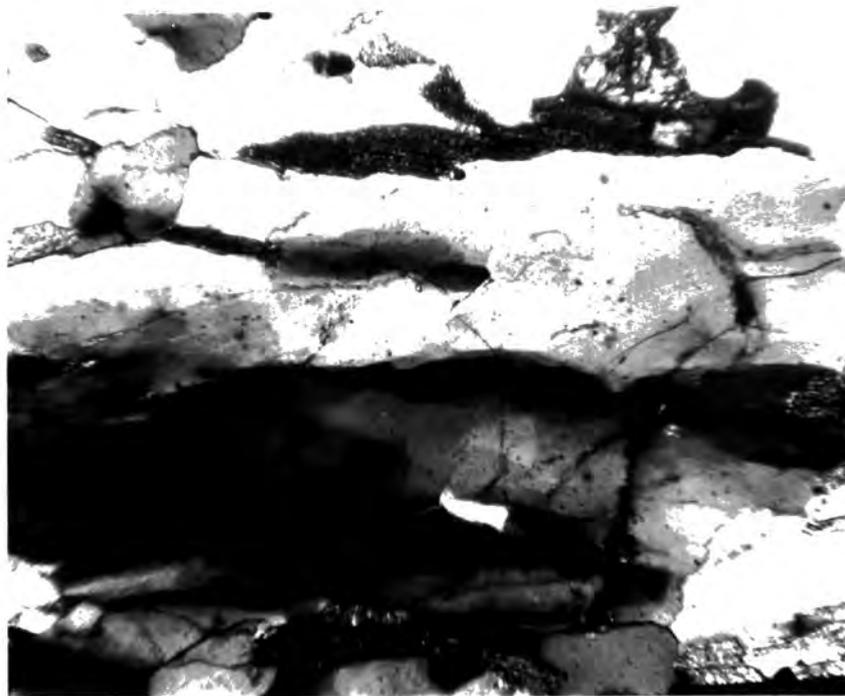


Figure 5: Quartz, showing strained extinction and irregular boundaries.
 Siliceous granulite, Klara Level, Skuterud.
 Crossed nicols. X 40.

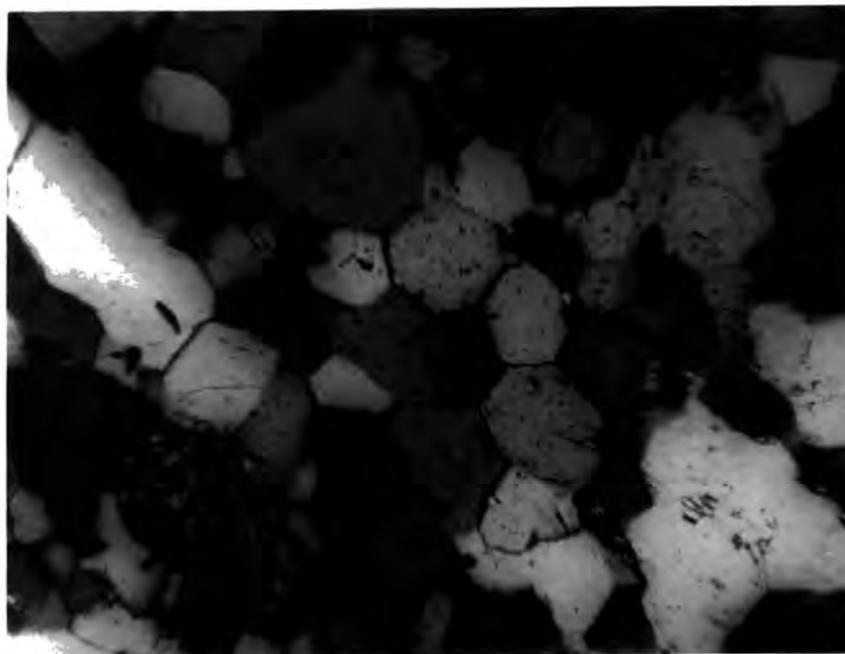


Figure 6: Quartz, with polyhedral crystal boundaries approaching an ideal
 angle of 120° . North Mine, Skuterud.
 Crossed nicols. X 40.

frequently full of fine dusty inclusions. The second type of quartz was found associated with cobalt mineralisation in quartzites from Skuterud. The grain size is much smaller, lying in the range 0.05-2 mm. The rock consists mainly of polygonal quartz grains, often coming together in triple-point junctions approaching an angle of 120° . This quartz is very clear and shows no undulatory extinction (Figure 6).

Plagioclase is a minor constituent and is found in the occasional feldspathic bands. The grains are subhedral, 0.3-0.9 mm. in length, and range in composition from An_{26} to An_{33} . Some grains have bent twin lamellae; sericitisation is very slight.

Muscovite is present as small, ragged, bent plates, 0.1 to .4 mm. in length. A second generation of fresh muscovite reaches 3-4 mm. in size and transects the foliation (Figure 7).

Phlogopitic biotite, pleochroic in pale brown and yellow is commonly present. The flakes, 0.5-3 mm. in length, lie parallel to the foliation. They are frequently altered to a chlorite with faint penninite interference colours.

From inspection of the modes in Table 3 it will be seen that tourmaline is rather common in the quartzites, occurring as dravite concentrated in the darker colour bands. The crystals range in size from 0.05-1 mm, mainly as equant grains, sometimes showing good crystal faces, occasionally very rounded and reminiscent of detrital grains. Textural studies suggest that the tourmaline is fairly late in the crystallisation sequence. It appears to be replacing biotite (Figure 9a,b,d), quartz (Figure 9a,c) and feldspar (Figure 9d) and to be later than the sillimanite and zircon (Figure 9g,h). However, in some sections the tourmaline itself is being replaced by quartz (Figure 9i-9n), by the late muscovite mentioned above (Figure 9o) and by the ore minerals (Figure 9p-q).

The amphibole appearing in the mode of rock 461 is anthophyllite. It occurs as ragged grains, 0.5-3 mm. long, showing the characteristic amphibole cleavage. It shows symmetrical extinction with low second order interference colours, and is biaxial positive with a large 2V.

Small sillimanite needles (fibrolite) are found scattered through the rock, oriented in the plane of the foliation. Well rounded zircon grains, 0.3 - 0.5 mm. in diameter, are present in some sections. (Figure 8).

Siliceous granulites

Field relations

Siliceous granulite is the commonest rock type of the Modum Formation within the area studied. Although very rich in quartz

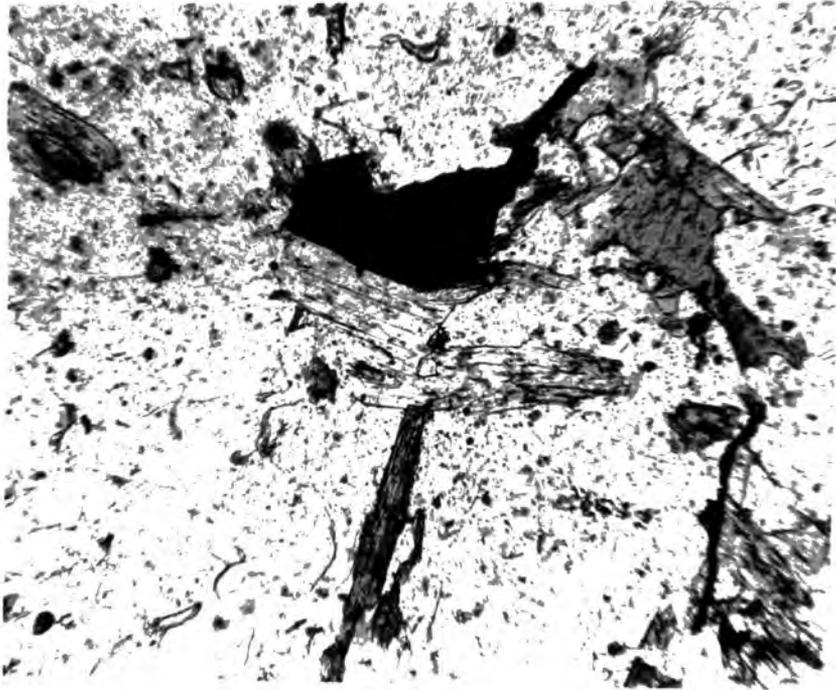


Figure 7: Muscovite transgressing the foliation. Siliceous granulite. Central Mine, Skuterud.

Ordinary light.

X 40.

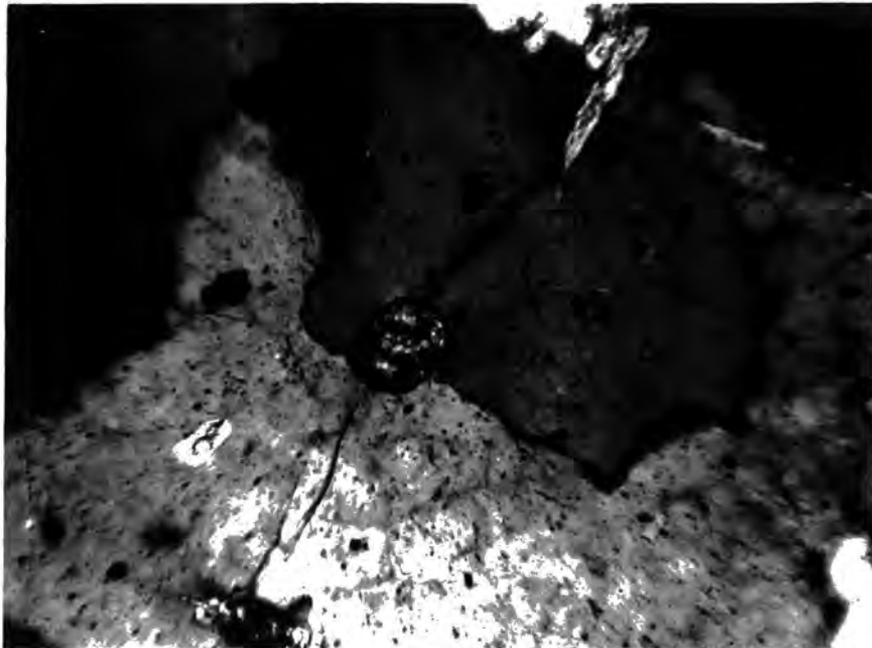


Figure 8: Rounded zircon in siliceous granulite. Central Mine, Skuterud.

Crossed nicols.

X 80.

Figure 9: Textural relations of tourmaline observed in thin section.

a- Large blastic tourmaline growing in quartz-biotite-schist. It cuts across the foliation. b- Tourmaline replacing muscovite. Note the characteristic trigonal outline and poorly defined fractures parallel to the crystal faces. c- Doubly terminated euhedral tourmaline replacing strained quartz. Note poor fractures parallel to the basal plane. d- Euhedral tourmaline growing at the expense of biotite, quartz and plagioclase. e- Small euhedral, equant tourmalines in strained quartz which is also being replaced by tourmaline. f- Blastic tourmaline replacing quartz and muscovite, which are also being attacked by ore minerals. The included quartz grain is in optical continuity with the quartz immediately to the left of the tourmaline. g- Tourmaline surrounding rounded zircon (Z) and being replaced itself by ore minerals. h- Tourmaline growing around earlier formed sillimanite. i- Fragments of one large tourmaline grain in recrystallised quartz, all the tourmaline is in optical continuity. j- Tourmaline being replaced during late recrystallisation of quartz. k- Tourmaline being replaced by quartz and muscovite. l- Prismatic tourmaline crystal replaced by quartz. m- Tourmaline replaced by quartz. n- Lens of recrystallised quartz in biotite-sillimanite-quartz-schist. The quartz contains fragments of tourmaline all in optical continuity. o- Large flake of late muscovite growing in quartzite and containing inclusions of tourmaline and quartz. p, q, and r- tourmaline replaced by ore minerals.

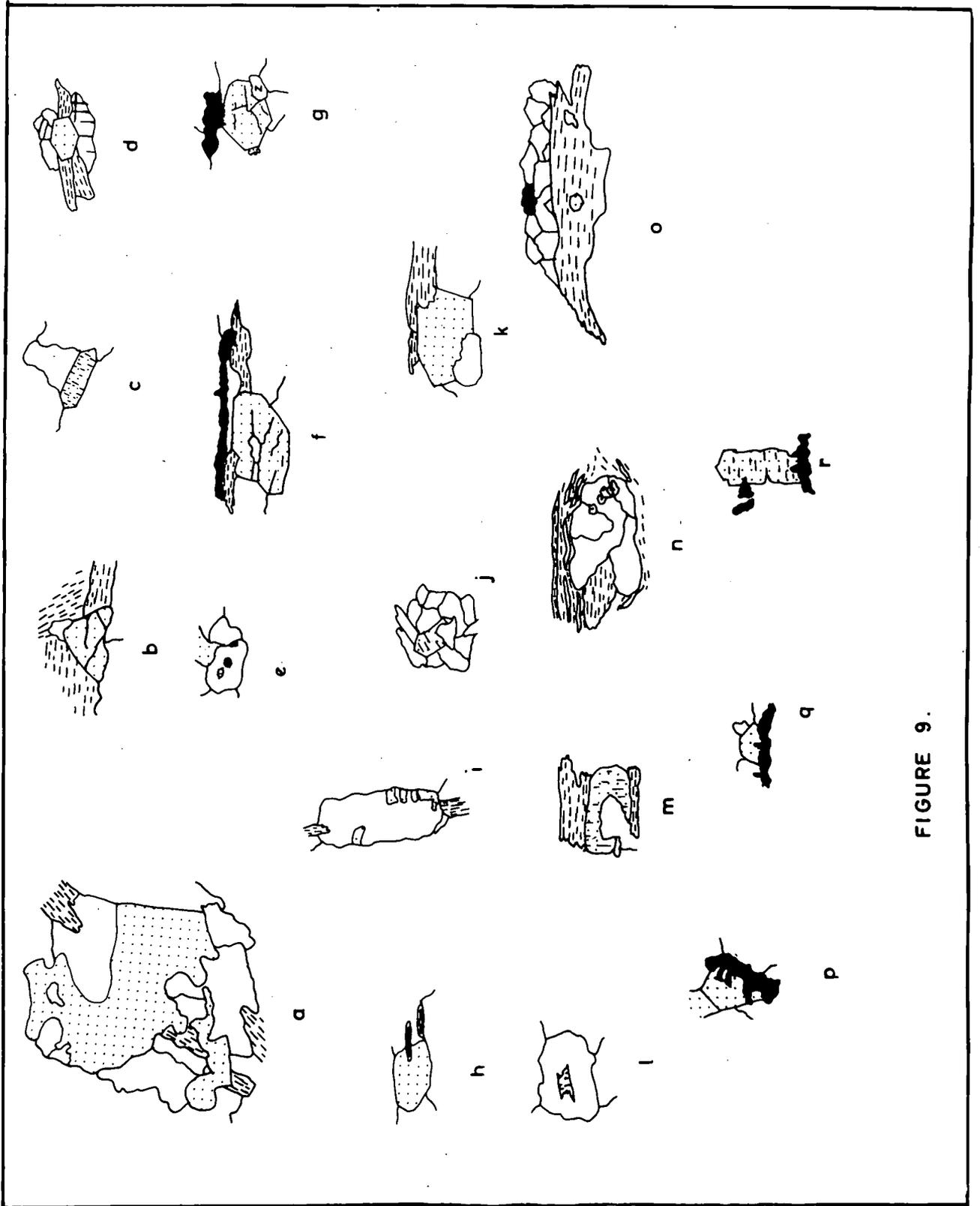


FIGURE 9.

they present a very different field appearance from the true quartzites. The more varied mineralogy expresses itself in outcrop features. The micaceous minerals are often concentrated in bands, to which they impart a marked schistosity. The simple colour banding, observed in the quartzites, is not seen in the siliceous granulites. The most obvious feature in the field, is the greater abundance of sillimanite, which is often concentrated in quartz-sillimanite nodules (faserkiesel). These faserkiesel are flattened in the plane of the schistosity and are ovoid in plan, they are frequently surrounded by a biotite rich envelope (Figure 10). Rapid variations in lithology are common, such that bands of felspathic granulite and semi-pelite are frequently found. No distinctive marker horizons are present, so separate outcrops are not capable of being correlated. Small scale folding is quite commonly seen on suitably oriented outcrops. Jointing, though present, is not so marked as in the true quartzites. Near V. Spone Kapel siliceous granulite can be traced along the strike, into a granitic gneiss, in which quartz-sillimanite nodules stand out as very obvious features.

Petrography

The rocks included within this unit of the classification show rapid variations in lithology on both a microscopic and a macroscopic scale. A common feature is the alternation of graded units, each about 2 cms. wide, which change gradually from a rock consisting essentially of quartz and minor plagioclase feldspar to a

— cms.

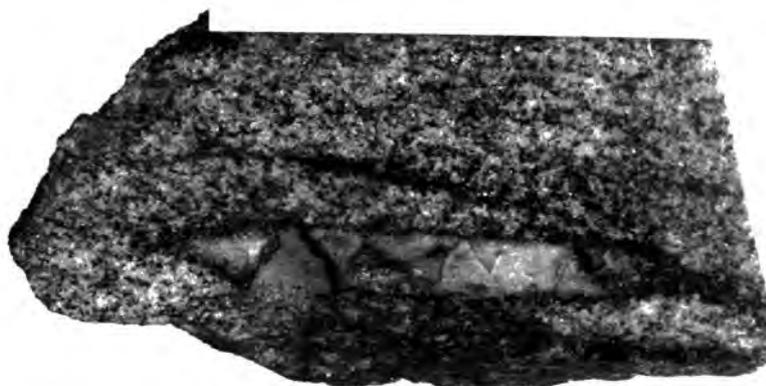


Figure 10: Quartz-sillimanite nodule in siliceous granulite. Note the biotite selvage at the margin of the lens. Nordgruventjen, Skuterud.



1 inch



Scale

Figure 11: Mineralogical banding in siliceous granulite suggestive of sedimentary graded bedding. Central Mine, Skuterud.

schist consisting of essentially phlogopitic biotite (Figure 11). The modes of 18 siliceous granulites are tabulated (Table 4), to illustrate the mineralogical variation.

Quartz is present in grains varying in size from 0.001 to 0.5 mm. It shows an amoeboid development with sutured and interlocking boundaries and marked strain extinction. No rounded quartz grains have been observed, the larger grains are clearly oriented parallel to the foliation. Textural evidence suggests that the quartz has recrystallised last; it is frequently seen replacing biotite.

Plagioclase feldspar is an important member of the mineral assemblage, forming up to 25% of the rock. It is concentrated in feldspar-rich bands paralleling the foliation. Sericitisation is usually advanced, some twin lamellae being attacked preferentially. Determinations on unsericitised grains, using the Michel-levy method, indicates compositions ranging from An₁₅ to An₃₅.

Potassium feldspar is a minor constituent of some of the feldspathic layers. It is present as microcline, showing incipient alteration to clay minerals, with well developed quadrille structure.

Schistose bands of biotite are rather common (see Table 4), the biotite is pleochroic with X - pale yellow or colourless, Y and Z - pale reddish brown. It has moderate birefringence with $2V = 0^\circ$. Pleochroic haloes are observable round small zircon inclusions. The biotite is noticeably paler in colour in proximity to opaque minerals. The grains are clearly oriented parallel to the foliation. The texture suggests that biotite is replaced by tourmaline and quartz (Figure 12). The biotite is frequently observed in parallel intergrowth with chlorite. The chlorite is pleochroic with X - colourless, Y = Z - pale grass green, it has parallel extinction, $2V = 0^\circ$, and anomalous blue interference colours, suggesting penninite.

Muscovite is present in subordinate amounts. It is usually intimately associated with sillimanite, but texturally it is not possible to say which mineral is the earlier. Large poikiloblastic flakes of muscovite are also observable cutting obliquely across the foliation.

Dravitic tourmaline is as abundant in the siliceous granulites as in the quartzites. It is particularly concentrated in biotitic bands and shows the same textural relations to the other minerals as described from the quartzites.

Table 4.

Siliceous Granulite Modes.

Specimen number.	11	12	13	16	17	18	103	106	110	115	116	143	44	484	481	101	47	86
Quartz.	72.1	68.5	59.6	75.2	73.0	67.6	67.0	71.8	63.9	72.5	70.8	60.8	74.3	77.0	65.7	70.0	62.3	72.5
Plagioclase.	-	-	22.1	2.2	1.6	18.5	0.7	7.2	12.6	-	-	9.7	-	7.5	14.3	12.1	-	-
K-felspar.	-	-	2.0	-	-	1.5	-	-	-	-	-	-	-	-	+	-	-	-
Biotite.	21.8	24.0	6.8	15.0	11.8	12.4	6.2	11.8	10.1	23.1	25.0	11.0	13.8	1.0	9.3	-	-	-
Muscovite.	-	-	-	-	5.5	-	-	-	-	-	-	2.0	-	8.0	1.8	2.1	-	4.7
Tourmaline.	1.2	0.5	2.0	0.1	0.7	-	6.6	9.2	-	3.0	2.4	9.2	0.7	-	4.7	0.1	-	15.5
Amphibole.	-	-	-	-	-	-	17.0	-	12.5	0.4	-	-	11.0	6.0	2.1	-	35.6	1.4
Sillimanite.	-	-	0.4	5.6	2.8	-	-	-	-	-	-	7.1	-	-	-	14.2	-	5.0
Zircon.	-	+	-	-	-	+	-	-	-	-	-	-	-	-	-	-	-	-
Opaque.	4.9	7.0	7.1	1.9	4.6	-	2.5	-	0.9	1.0	1.8	0.2	0.3	0.5	2.1	1.5	2.1	0.9
Plagioclase: An content	-	-	28	18	31	24	33	15	35	-	-	22	-	30	32	21	-	-
Number of points	812	935	765	1086	943	1005	782	895	1089	1097	985	992	897	795	943	967	859	1021

+ - Present in minor amounts.

N.B. Specimen localities tabulated below:

Localities: 11-1000/260 (Fig. 110); 12-1000/255 (Fig. 110); 13-1000/252 (Fig. 110); 16-1000/238 (Fig. 110); 17-1000/230 (Fig. 110); 18-1000/226 (Fig. 110); 103-705/260 (Fig. 110); 106-720/260 (Fig. 110) 110-Middagsville Mine; 115-Middagsville Mine; 116-Middagsville Mine; 143-160/130 (Fig. 109); 44-980/240 (Fig. 110); 484-1310/240 (Fig. 110); 101-695/265 (Fig. 110); 47-980/240 (Fig. 110); 86-990/210 (Fig. 110).



Figure 12: Quartz replacing biotite. Siliceous granulite, North Mine, Skuterud. Crossed nicols.

X 40.

Anthophyllite is present in some of the siliceous granulites. The grains have euhedral to subhedral outlines and are colourless. Longitudinal sections have parallel extinction. It has moderate birefringence, is biaxial positive with $2V$ of ca. 75° .

Sillimanite occurs in small, slender prismatic crystals evenly distributed through the rock. It is also found in felted aggregates (fibrolite) associated with muscovite. The individual crystals range in length from 0.2 to 2 mm. They are usually completely surrounded by quartz and are clearly arranged parallel to the foliation. They show parallel extinction, have moderate birefringence and are length slow.

Zircon is the commonest accessory mineral, occurring as well rounded grains in the quartz rich bands and as small inclusions in biotite. Apatite is also present as small euhedral prismatic crystals. It is similar in appearance to the sillimanite from which it is distinguished by its weaker birefringence and length fast character.

Felspathic granulite

Field relations

Rocks of this category were nowhere found in sufficient abundance to form a mappable unit. They always occur as thin, more deeply weathered, bands in the siliceous granulites.

Petrography

The appearance of the felspathic granulites in thin section is very similar to that of the siliceous granulites. The only major difference being the amount of plagioclase feldspar. Two typical modes are shown in Table 5.

Semi-pelite and pelite

Field relations

Micaceous bands are rather common in the siliceous granulites where they vary in width from a few centimetres to about three metres. Within these bands, thin, 0.5-5 cms. thick, layers are found consisting almost entirely of biotite. These semi-pelites

Table 5.

Felspathic Granulite and Semi-Pelite Modes

Specimen Number	Felspathic granulite		Semi-pelite	
	144	495	122	22
Quartz	41.3	27.8	52.0	28.1
Plagioclase	23.5	42.2	5.4	21.0
K-felspar	-	+	-	12.0
Biotite	24.9	12.0	24.7	20.1
Muscovite	-	12.6	-	-
Tourmaline	10.2	-	-	2.0
Amphibole	-	1.4	12.3	-
Opaque	1.0	4.0	5.4	17.8
Plagioclase An. content	26	32	24	33
No. of points	987	1075	1105	998

Localities: 114-150/150 (Fig.109); 495-1310/240 (Fig.110);
122-Middagsvhillie Mine; 22-1000/260 (Fig.110);

and pelites have deeply weathered outcrops, particularly when they are associated with graphite and sulphide minerals (see p.138 , "The Fahlbands"). The very micaceous layers are frequently associated with shear planes.

Also included in this group is a rock with a very striking field appearance encountered at the western limit of the Forhaabning and Klara Crosscuts of the Skuterud Mines. It is a dark biotite schist containing many quartz-sillimanite lenses. (Figure 13). These lenses are smaller in size and of more frequent occurrence than the similar lenses in the siliceous granulites. They are flattened in the schistosity and elongated to impose a lineation on the rock.

Petrography

Quartz - biotite - tourmaline schist is the commonest rock type within the mineralised zone. Inspection of the modal analyses in Tables 4 and 5 illustrates that the distinction between semi-pelite and siliceous granulite is purely arbitrary, depending solely on the relative amounts of quartz and biotite. The texture, grain size, and mineralogy of the rocks are basically the same.

The biotite-schist with sillimanite nodules contains a more deeply coloured biotite than that previously described. It has a deep brown colour with a slight reddish tint and is pleochroic with X: deep brown, Y = Z: pale brown with reddish tint. It extinguishes parallel to the basal cleavage and has a 2V in the range $15-18^{\circ}$, it is associated with muscovite and quartz. The nodules contain no biotite and consist mainly of quartz with sheaths of sillimanite needles radiating out from the centre of the nodules (Figure 14).



Figure 13: Sillimanitic nodular biotite-schist in situ in the Klara Adit.

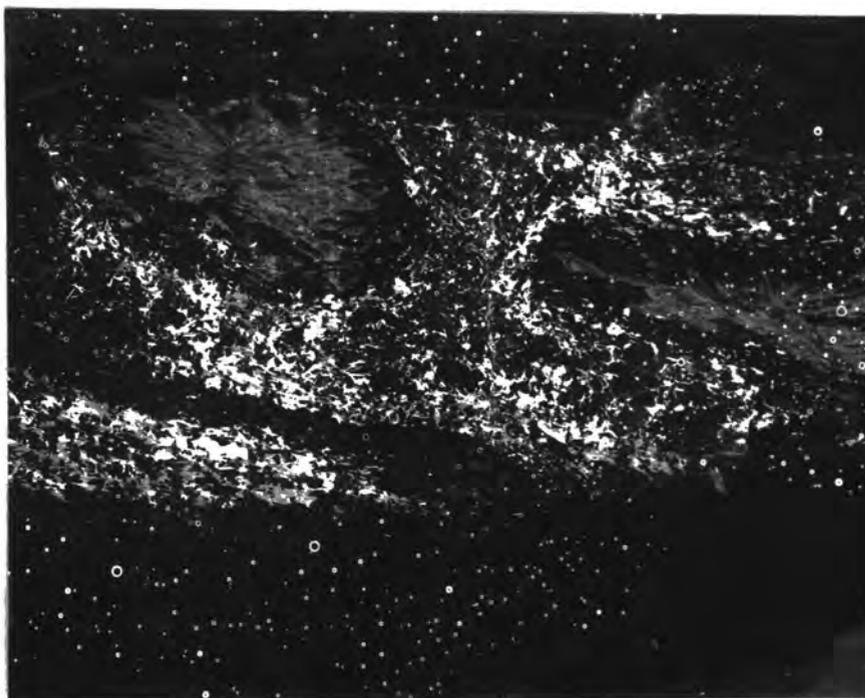


Figure 14: Photograph of whole thin section of the rock shown above. Note the radiating sillimanite in the quartz-sillimanite nodules, and the absence of biotite (white) from these nodules.

Calcareous granulite

Field relations

This rock type is rather uncommon in the Modum area. It consists of alternating bands of felspathic quartzite and calc-silicate minerals. The individual bands vary in width from 0.1 to 4 cms. and are traceable over the complete outcrop length. The best developed calcareous granulite horizon is exposed to the west of the South Mines at Skuterud (Figure 109).

Petrography

In thin section the banded texture is clearly seen, the lighter layers consist of alternations of quartzitic and felspathic horizons. The quartz has very sutured margins and strained extinction. In some bands it is markedly different in having a larger grain size, in others the grains have less sutured margins and appear to be approaching the 'stable' polyhedral intergranular texture. The felspathic bands consist of large interlocking grains of plagioclase. These are badly sericitised, but albite twinning is observable in some crystals indicating a composition of An₆₅ (Michel-Levy method).

Associated with these lighter layers particular concentrations of tourmaline, apatite, zircon and rutile are noticeable. The rutile and tourmaline are particularly associated with the felspathic bands and have a blastic development to the surrounding minerals. The zircon and apatite are in very small, extremely well rounded grains. These are particularly noticeable near the junction of the quartz-felspathic and calcareous bands, they possibly represent original, sedimentary, accumulates of heavy minerals.

The darker bands consist of intergrown quartz and hornblende. The hornblende is in large ragged grains oriented parallel to the foliation, the crystals contain many embayments and rounded inclusions of quartz. It is pale green in colour with X: pale yellow-brown; Y: yellow-green; Z: bluish-green and $\hat{c}Z: 22^\circ$.

Diopside-actinolite skarn

Within the rocks of the Modum Formation small lenses of diopside-actinolite skarn are well developed. These are frequently elongated in the plane of the foliation, the constituent minerals

reach lengths of several centimetres. They are often associated with pyrite, chalcopyrite and cobalt minerals.

The diopside is very pale green in thin section with a positive 2V of 60-70° and $\hat{c}Z$: 40°. It appears to have suffered cataclasis in some areas, chalcopyrite is frequently present as an infilling to the cracks so produced. The quartz present between the diopside crystals is in large unstrained grains reaching 0.8 cms. in length, rounded inclusions of quartz are frequently seen in the diopside. Small rounded zircons are also present in some of the diopside crystals. Calcite is occasionally observed forming an anhedral matrix between the pyroxene crystals.

Between the diopside and the surrounding rock a border zone of actinolitic hornblende is developed. This zone has been observed with a maximum thickness of 6 cms. The hornblende is pleochroic with X: pale yellow-green; Y: pale blue-green; Z: blue-green. It has $\hat{c}Z$: 15° and a large negative 2V.

MINERALOGY OF THE MODUM FORMATION

Mica

In the preliminary inspection of thin section and hand specimen material from the Modum area, it was noticed, that there was considerable variation in the types of mica present. In order to gain some idea of the variation, some typical examples were selected for study. The micas were separated from the rocks, X-ray powder films were prepared, refractive index measurements were made on flakes in immersion oils and the optical properties of the micas were studied in thin section.

Method of separation

The rocks were crushed to -90 mesh by hand, material of the desired size was sieved off, as frequently as possible, to minimise the production of 'fines'. The crushed rock was washed in water and all fine material removed. The sample was then dried and placed in a separating funnel, containing bromoform and the heavy fraction removed. With some samples, it was found that an efficient separation was not obtained, unless the heavy liquid was slightly diluted with acetone. The collected fractions were filtered, washed with alcohol, until free of the heavy organic liquid and dried. The dried heavy fractions were then put through a Franz magnetic separator. It was found that with an inclination of 10° , iron-rich micas were separated at 0.60 amps. and the white micas were preferentially enriched in the fraction left at 1.50 amps. In some cases, a preliminary separation was made using the 'mica separator', this consists of a polished, vibrating aluminium plate on to which the sample was gradually placed. The plate was so inclined, that the rounded minerals rolled straight down, while the micas slowly worked their way across the plate to be collected at the edges.

The separated samples so obtained, proved to be a convenient size for refractive index determinations. For X-ray examination a hand picked sample was obtained using a mounted needle made slightly sticky by plunging it into plasticine, the separated grains were crushed in an agate mortar and mounted for examination in a 9 cm. powder camera. It was found that the rather prolonged grinding

necessary to reduce the mica to a suitable size had no appreciable effect on the intensity of the lines such as described by Yoder and Eugster (1955).

X-ray investigation of the mica polymorphs

Discussion: It has been well known since Hendricks and Jefferson (1939) described several mica polymorphs, that optical examination should be supplemented by X-ray analysis to yield satisfactory information.

All micas can be described by a monoclinic sub-cell, the c-axis of which, is running from one twelve-fold co-ordinated ion through the sheet to the nearest neighbouring twelve-fold co-ordinated ion. Every sheet can be stacked in a number of different ways by rotation around an axis perpendicular to the sheet, yielding two, three and six layer unit cells. Smith and Yoder (1956) reviewed the polymorphic phenomena due to stacking, and concluded that only the following simple stacking polymorphs are possible: $1M$, $2M_1$, $2M_2$, $2O$, $3T$ and $3H$. Except for the $1M$ and $3T$ trioctahedral micas, for which single crystal methods are needed, the polymorphic state of a mica can easily be determined by X-ray powder analysis.

Results: The diagnostically important range, $d = 3.36 - 2.49 \text{ \AA}$, of the powder patterns of some mica polymorphs described in the literature and obtained from the Modum samples are presented in Table 6.

From an inspection of these data, it is apparent, that of the dioctahedral micas, the only polymorph discovered at Modum is muscovite $2M_1$, and for the trioctahedral micas phlogopite $1M/3T$ was the only one present. This result is in accordance with previous

Table 6.

Interplanar spacings (d-values) of some mica polymorphs. Only the diagnostically important 3.36 - 2.49 Å range is included.

1	2	3	4	5	6	7
3.36 vs	3.35 vs	3.33 vs	3.34 vs	3.36 vs	3.33 vs	3.34 vs
3.07 m	3.21 m	3.11 w	3.14 w	3.28 m	3.22 s	3.20 m
2.93 vw	3.00 m	2.88 w	2.92 w	3.16 vw	3.00 m	3.02 w
2.69 vw	2.87 w	2.59 w	2.71 vw	3.04 s	2.85 vw	2.70 vw
2.58 m	2.80 w	2.56 m	2.64 vw	2.93 vw	2.77 vw	2.67 vw
2.56 s	2.59 m		2.62 m	2.82 w	2.57 s	2.58 s
	2.58 m		2.51 m	2.65 w		
	2.56 s			2.62 m		
				2.52 vs		
8	9	10	11	12	13	14
3.48 m	3.48 vs	3.40 vs	3.40 vs	3.40 vs	3.40 vs	3.50 vs
3.42 vs	3.15 vw	3.17 vw	2.82 s	3.18 vw	3.18 w	3.42 m
3.22 m	2.90 vw	2.87 vw	2.48 s	2.90 vw	2.90 w	2.81 vw
3.02 m	2.69 vw	2.82 vw		2.70 vw	2.51 vw	2.59 vw
2.85 vw	2.62 s	2.58 s		2.62 s	2.70 vw	2.55 vw
2.83 vw	2.53 s	2.52 m		2.53 s	2.53 s	2.47 vw

Legend: 1. muscovite 1M (Yoder and Eugster, 1955). 2. muscovite 2M (Yoder and Eugster, 1955). 3. muscovite 3T (Yoder and Eugster, 1955). 4. phlogopite 1M/3T (Smith and Yoder, 1956). 5. phlogopite 2M, (Christie, unpublished data). 6. muscovite 2M1 (spec. 34M1, pegmatite, Stollhammen, Skuterud.). 7. muscovite 2M (spec. 323M4, pegmatite, North Mine, Skuterud.). 8. muscovite 2M, (spec. BSRM1, Forhaabning Adit, Skuterud.). 9. phlogopite 1M/3T, (spec. BSRM2, Forhaabning Adit, Skuterud.). 10. phlogopite 1M/3T, (spec. 705M1, North Mine, Skuterud.). 11. phlogopite 1M/3T, (spec. 323M1, pegmatite, North Mine, Skuterud.). 12. phlogopite 1M/3T, (spec. 322M1, sillimanitic nodular siliceous granulite, North Mine, Skuterud.). 13. phlogopite 1M/3T, (spec. 308M1, felspathic granulite, Svartefjell Mine.). 14. chlorite, (clinchlore), (spec. 323M3, pegmatite, North Mine, Skuterud.).

observations on micas from metamorphic terrains. Hurst (unpublished work quoted in Smith and Yoder, 1956), examined 'a very large collection' of biotites and muscovites, from an area in which 'all grades of metamorphism may be found', and found that all the biotites had the LM or 3T structure and all the muscovites were $2M_1$. Smith and Yoder (1956), found muscovite from the sillimanite zone in Glen Cova, Scotland to have the $2M_1$ structure, phlogopite from a tremolite marble near Chester, Vermont, had both the LM and 3T polymorphs (determined by single crystal methods). Christie (personal communication) examined 69 samples of muscovite from the whole of Norway and found that the majority (63) had the $2M_1$ structure, in the trioctahedral micas that he studied he found that the LM/3T structure was the commonest in a wide variety of Norwegian rocks, pegmatitic, metamorphic and probably magmatic.

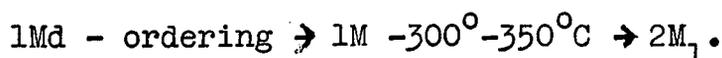
Significance of results: Attempts have been made to correlate the prevalence of muscovite $2M_1$ in metamorphic terrains with physical conditions. Muscovite lMd, an early stage of growth, with disordered stacking structure, is believed, by Yoder and Eugster (1955), to represent a possible low temperature form, with ordering this changes to the LM polymorph. The lMd and LM polymorphs are very common in nature, forming dense pseudomorphous masses or decomposition films on other minerals, indicating formation in a low temperature environment. The LM and lMd polymorphs investigated by Yoder and Eugster (op.cit.) had been variously described in the literature as 'illite', 'sericite', 'hydromuscovite', 'secondary muscovite' or 'pinite'. It is possible, that the fine white mica

formed due to the alteration of feldspars and sillimanite, observed in the Modum rocks, represents a LM or LMd muscovite polymorph, but difficulties of separation prevented confirmation of this.

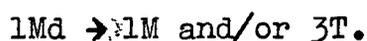
Yoder and Eugster (op.cit.) found that heating synthetic LM muscovite at 350°C and 15,000 bars P_{H_2O} converted it into $2M_1$ muscovite. Christie (unpublished work) heated natural LM muscovite for several weeks at 300°C and found that the X-ray powder pattern was sharper than before, indicating that some slow recrystallisation had taken place. Thus if water pressure is considered to have only a slight influence, the critical temperature for the change $LM \rightarrow 2M_1$ can be considered to lie between 300° and 350°C.

Smith and Yoder (1956) reach the same conclusion regarding the muscovite polymorphs, and consider that in the trioctahedral micas the LMd polymorph changes to either LM or 3T, or a mixture of them. Thus, to summarise current opinion, we can represent the changes thus:

Dioctahedral micas:



Trioctahedral micas:



With increasing temperature (and pressure ?) moving the series to the right in both cases.

Colour and refractive index of the micas

Trioctahedral micas: It was hoped that some idea of the range of chemical composition of the micas in the phlogopite - siderophyllite/lepidomelane series could be obtained by observations of refractive index and colour variations.

Foster's (1960a) work on the trioctahedral micas shows that with pure phlogopite as the prototype, the composition of all the other trioctahedral micas may be derived by replacement of Mg^{2+} in the octahedral position by, most commonly, Fe^{2+} and R^{3+} (Al^{3+} or Fe^{3+}). The mixed crystal series being thus:-

PHLOGOPITE	($>70\%$ Mg^{2+} in octahedral position)			
Mg-BIOTITE	(Mg^{2+} $>$ Fe^{2+})	"	"	"
Fe-BIOTITE	(Fe^{2+} $>$ Mg^{2+})	"	"	"
SIDEROPHYLLITE	(Fe^{2+} & Fe^{3+})	"	"	"
or				
LEPIDOMELANE	(Fe^{2+} & Al^{3+})	"	"	"

In general, there is a tendency for both the refractive index and the intensity of colour of the mica, to increase towards the iron rich end of the mixed crystal series. Winchell, in his studies on the biotite system (1925, 1935), calculated analysed biotites in terms of four theoretical end molecules, and plotted their optical properties on a rectangular diagram. This diagram is reproduced (Figure 15a) with some Modum micas plotted in the appropriate positions.

Hall (1914b) criticised Winchell's diagram, on the grounds that it is an oversimplification of a much more complicated relationship. Hall studied the refractive indices of the system, not in terms of end-member molecules, but rather from the point of view of the effect of the substitution of the various atoms, in the structural formula. Heinrich (1946) also criticises Winchell's diagram, because he considers that the 'end-molecules' have little structural significance, and do not adequately explain the variations in chemical composition. He found that the most

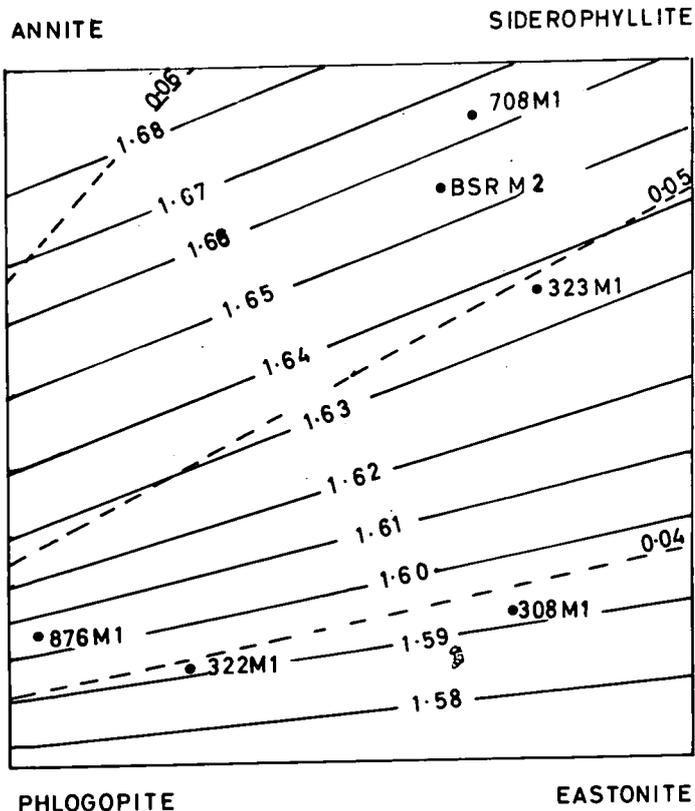
Figure 15: Attempts at correlating biotite optics with chemical composition.

A. After Winchell (1935).

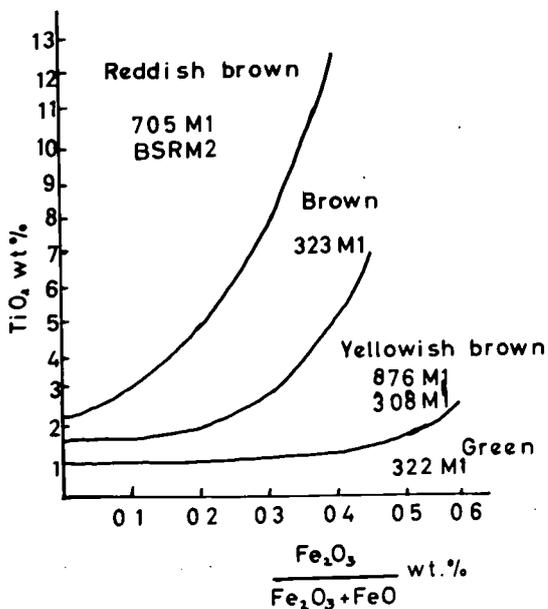
B. After Hayama (1959).

C. After Heinrich (1946).

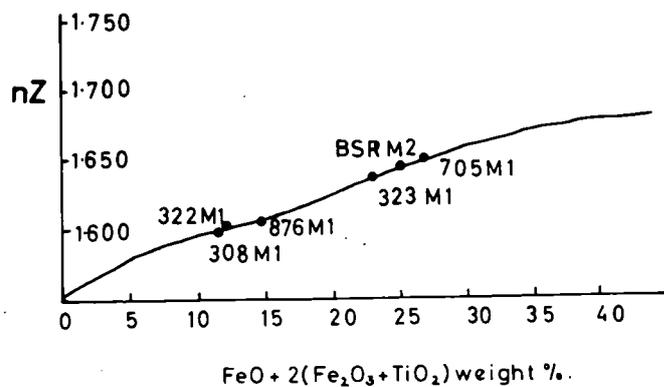
N.B. The specimen numbers on the diagrams refer to micas from Modum Formation rocks mentioned in the text.



A



B



C

Fig. 15

satisfactory relationship of nZ to composition was obtained when it was plotted against weight percent $\text{FeO} + 2(\text{Fe}_2\text{O}_3 + \text{TiO}_2)$. The significant fit obtained, using this formula, shows the relatively greater effect of ferric iron and titanium on nZ . Heinrich's graph is reproduced as Figure 15c with some of the Modum micas plotted.

Attempts have also been made to correlate variations in colour with chemical composition. It has been generally agreed that the reddish tint of biotite is due to high TiO_2 content, the influence of the other constituents has not been properly understood. Grout (1924), stated that Fe_2O_3 gives a greenish tint to biotite, Tilley (1924), considered the brownish tint of biotite, as due to a high FeO content. All these observations were based on very few available analyses. Hall (1941a), discusses the problem and considers the colour to depend on the contents of TiO_2 , total iron, and MgO . He believes that the reddish tints are due to larger contents of TiO_2 . MgO dilutes the reddish tint produced by TiO_2 , unless there is a great deal, in which case, the diluting effect of MgO is negligible. The greenish tints are held to be due to a high content of total iron.

On the basis of more available analyses (41), Hayama (1959), decided that the colour of biotite depended on TiO_2 and the $\text{Fe}_2\text{O}_3 / (\text{Fe}_2\text{O}_3 + \text{FeO})$ ratio. When these are plotted against each other, four fields become distinguishable containing (1) reddish brown, (2) dark brown and brown, (3) yellowish to greenish brown and (4) green to bluish green biotites. Thus he considers the effect of MgO on the colour to be negligible. The fields obtained by Hayama are reproduced as Figure 15b.

Results: Using the data of these authors, it is now possible to gain some idea of the chemical variation of the Modum trioctahedral micas. The relevant data for the Modum material are given in Table 7. The specimens were selected as being typical of the commonly occurring types.

Specimen 705M1 is from a biotite-garnet schist associated with amphibolite at the North Mine, Skuterud. The high values for birefringence and n_z indicate a high ferric iron content (Winchell, Hall), the reddish brown colour indicates the presence of considerable titanium. Its composition probably approximates the formula $\underline{K_2 Fe^{2+}_4 Fe^{3+}_{0.5} Ti_{1.5} Si_6 Al_2 O_{22} (OH) F}$; in Foster's terminology it would be a siderophyllite.

Specimen BSRM2 is from the biotite schist with sillimanite nodules, collected at the western limit of Klara Crosscut. This also has high birefringence and refractive index, the reddish tint is more subdued than in 705M1. Thus BSRM2 is also probably high in total iron with rather less titanium than in 705M1. Its composition probably approximates the formula $\underline{K_2 Fe^{2+}_4 Fe^{3+} Ti Si_6 Al_2 O_{22} (OH) F}$ and is also a siderophyllite in Foster's terminology.

Specimen 323M1 is from a quartz-felspar-muscovite-biotite pegmatite cutting the ore zone at the North Mine, Skuterud. The refractive index and birefringence values suggest that this mica is lower in ferric iron than either 705M1 or BSRM2 (Winchell, Hall). The brown colour suggests that titanium is not as important (Hayama) and the formula probably lies close to $\underline{K_2 Fe^{2+}_3 Mg^{2+} Fe^{3+}_{1.5} Ti_{0.5} Al_2 Si_6 O_{22} (OH) F}$. In Foster's terminology 323M1 is probably an Fe-biotite.

Table 7.

Properties of some trioctahedral micas from Modum.

Specimen Number	322M1.	308M1.	876M1.	323M1.	BSRM2.	705M1.
Polymorph	Phlogopite 1M/3T	Phlogopite 1M/3T	Phlogopite 1M/3T	Phlogopite 1M/3T	Phlogopite 1M/3T	Phlogopite 1M/3T
X	Colourless	Pale Yellow	Pale Yellow	Brown	Pale reddish- brown	Reddish brown
Pleochroism						
Y	Pale brownish green	Pale brownish yellow	Brownish yellow	Deep brown	Reddish brown	Deep reddish brown
Z						
Extinction (Δ°)	0°	0°	0°	0°	0°	up to 4°
2 V _X	0°	5° - 15°	5° - 20°	0°	15° - 18°	ca. 2°
Refractive Indices	nX 1.553 nY 1.592 nZ	1.555	1.553	1.592	1.605	1.607
Birefringence	0.039	0.037	0.049	0.037	0.053	0.055

Localities: 322M1-sillimanitic nodular siliceous granulite, North Mine, Skuterud.
 308M1-sillimanitics siliceous granulite, Central Mine, Skuterud.
 876M1-quartzite, Forhaabing Adit. 323M1-biotite-muscovite-quartz-
 felspar pegmatite, North Mine, Skuterud. BSRM2-sillimanitic nodular
 biotite schist, Klara Adit, Skuterud. 705M1-garnetiferous biotite-
 amphibolite, North Mine, Skuterud.

Specimen 876M1 was separated from a quartzite in Forhaabning Adit, Skuterud. The colour indicates a low Ti content, probably falling in field 3 of Hayama's diagram. The refractive index and birefringence values indicate a low total iron content and probably greater magnesium content (Winchell, Hall). It is considered that 876M1 has a formula approximating:

$$\frac{K_2 Fe^{2+}_3 Mg_2 Fe^{3+}_{0.75} Ti_{0.25} Si_6 Al_2 O_{22} (OH) F.}{}$$
 and lies between Mg- and Fe- biotite in Foster's terminology.

Specimen 308M1 is from a felspathic granulites at Svartefjell Mine. It has similar properties to 876M1 being only slightly paler in colour and lower in relief. This indicates that iron is even less important in this mica, it is probably a Mg-biotite in Foster's terminology with a formula possibly close to

$$\frac{K_2 Fe^{2+}_2 Mg_3 Fe^{3+} Si_6 Al_2 O_{22} (OH) F.}{}$$

Specimen 322M1 is from a sillimanitic siliceous granulite at the North Mine, Skuterud. The greenish tint indicates the virtual absence of Ti, the low relief suggests that iron is even lower than in 308M1 and this mica could probably best be described as a phlogopite with a formula approximating to $\frac{K_2 Fe^{2+}_2 Mg_4 Si_6 Al_2 O_{22} (OH) F.}{}$

Occurrence of the mica types: The Ti rich siderophyllites (e.g. 705M1), are confined to the biotite schists possibly formed through the alteration of amphibolite (see p. 121). The only siderophyllite found in the normal metasedimentary sequence, is that in the sillimanitic modular biotite schist, in Klara Adit (BSRM2).

Fe-biotite (323M1) is not common, being confined to the late, cross-cutting, granite pegmatites.

Fe-Mg-biotite (876M1), Mg-biotite (308M1) and phlogopite (322M1) are the common trioctahedral micas in the quartzites, siliceous and felspathic granulites and semi-pelites of the ore zone rocks.

Alteration of the trioctahedral micas: In about 75% of the specimens examined, many small needles of another mineral were noticed, growing in from the grain boundaries. It was suspected that this was chlorite. This was confirmed during the X-ray work as frequently lines due to chlorite appeared on the films. In one case they were the only lines present, showing that the alteration was quite far advanced in that specimen (323M3, No. 14 in Table 6). The chlorite was identified as a clinocllore on the basis of the powder film.

Dioctahedral micas

The muscovites selected for study have all been X-rayed, where practicable, and only the $2M_1$ polymorph has been found. This is the normal mica for this grade of metamorphism. An unconfirmed possibility, is that the muscovite, formed by the retrograde alteration of feldspar and sillimanite, represents the LM or LMD polymorph.

Muscovites separated from three different rock types agree quite closely in their properties. (See Table 8). When plotted on Winchell's diagram of nZ versus $2V$ (1951, Fig. 254), it is seen that these three micas fall fairly close together, showing approximately 60% of the muscovite and 40% of the phengite molecule (See Fig. 16).

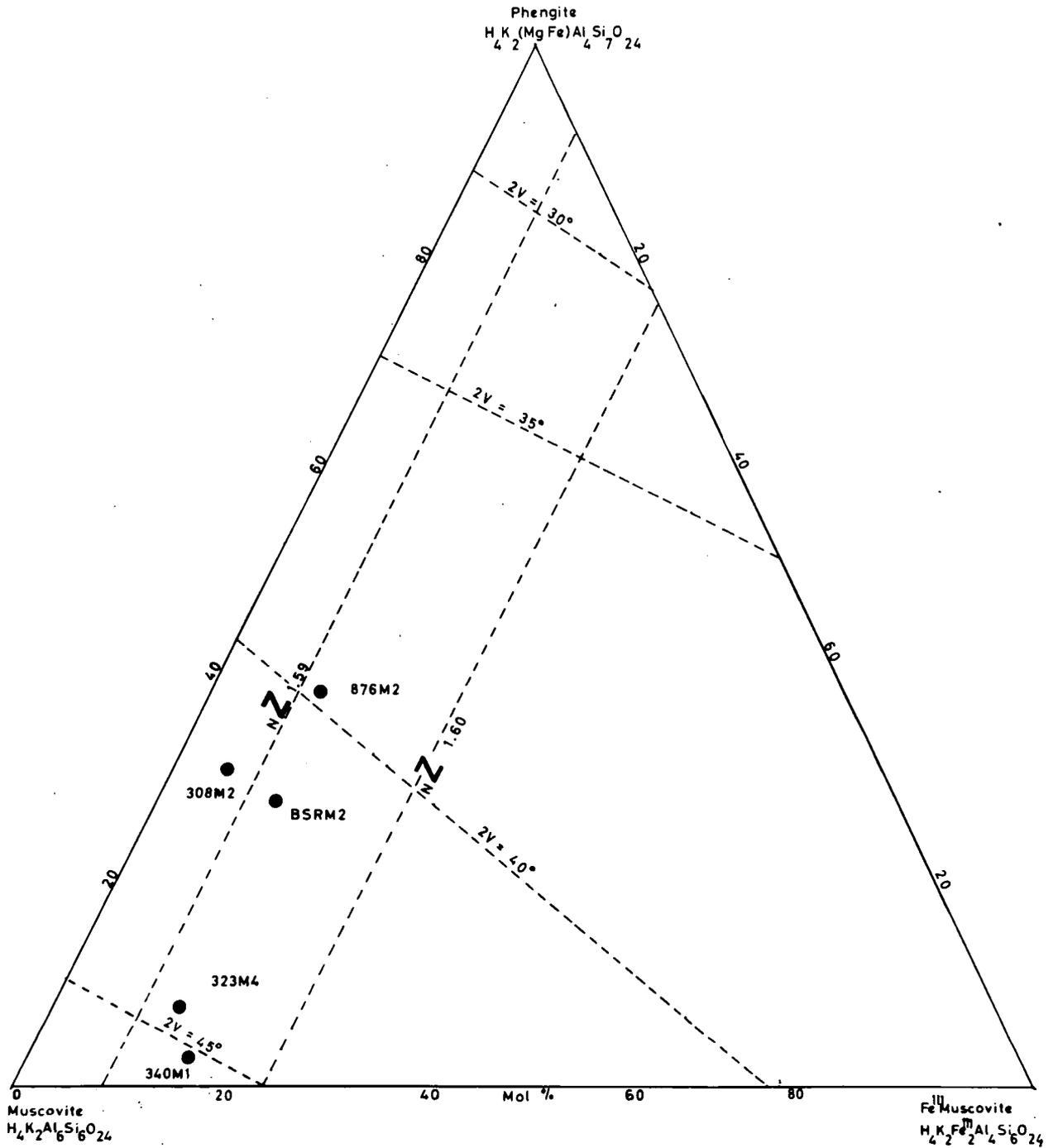
Table 8.

Properties of muscovite.

Sample number.	876M2	308M2	BSRM1	340M1	323M4
Refractive n _X	1.562	1.559	1.563	1.564	1.560
Indices. n _Y =n _Z	1.591	1.589	1.592	1.595	1.593
Birefringence.	0.029	0.030	0.029	0.031	0.033
2 V _X	ca 40°	ca 42°	ca 42°	ca 45°	ca 44°
Optic sign	-ve	-ve	-ve	-ve	-ve
Dispersion	Distinct	Moderate	Distinct	Strong	Strong
	r > v	r > v	r > v	r > v	r > v
Colour	White-pale green	White-pale green	White	White	White
Absorption	X Y = Z	X Y = Z	X = Y = Z	n.o.	n.o.
Relief	Fair	Fair	Fair	n.o.	n.o.
Interference colour	Upper 2nd order	Lower 2nd order	Upper 2nd order	n.o.	n.o.
Extinction	Parrallel	Parrallel	Parrallel	n.o.	n.o.

n.o.: = not observed.

Localities: 876M2-quartzite, Forhaabning Adit, Skuterud. 308M2-sillimanitic nodular siliceous granulite, Skuterud. BSRM1-sillimanitic nodular biotite schist, Klara Adit, Skuterud. 340M1-muscovite-felspar-tourmaline pegmatite, Stolltdammen. 323M4-biotite-muscovite-quartz-felspar pegmatite, North Mine, Skuterud.



Correlation of Optical Properties of Muscovite
 with Chemical Composition.
 (After Winchell 1951)

Specimen numbers refer to muscovites
 from Modum.

Fig. 16

Using these data to convert to Foster's (1960b) structural formula we get: $\underline{\text{Al}_{2.6} \text{Si}_{3.2} \text{Fe}^{2+}_{0.2} \text{O}_{10} (\text{OH})_2 \text{K}}$.

Muscovites separated from two pegmatites in the area are rather similar in their properties (see Table 8). These mica were not observed in thin-section. The tabulated properties would indicate that the mica is fairly close to a pure muscovite from Winchell's diagram.

Muscovite is a common constituent of the normal mineral assemblages of the various schists and gneisses, it is usually subordinate to biotite. Two generations of later muscovite may be observed in these rocks, one is associated with the breakdown of feldspar and sillimanite, the other occurs as large blastic flakes cross-cutting the foliation of the earlier minerals. Muscovite is also common in the pegmatites, several of which were worked for mica during World War II.

Tourmaline

This mineral is nearly ubiquitous in the Modum ore-zone rocks. Sometimes, only a few small crystals of the mineral can be found, in any one thin slice, at other times, tourmaline can form up to 12% of the rock. It is frequently associated with biotite rich layers. Rosenqvist, (1949), mentions the presence of tourmaline at Modum and concludes that it is a dravite with n_o : 1.615 and n_E : 1.637.

Separation

A sample of the ore zone quartzite, carrying cobalt and uranium minerals, was crushed to -90 mesh. The sample was washed

and a heavy fraction collected, using bromoform. This heavy concentrate consisted mainly of tourmaline, ore minerals and zircon. These minerals were separated using a Franz magnetic separator, the tourmaline being obtained with an ammeter reading of 1.50 and a tilt of 10° .

The tourmaline so obtained, was a pale yellow colour rather reminiscent of sphene and consisted of equant grains, looking rather like garnets (Figure 17). The mineral looked so unlike tourmaline, that its identity was checked by picking out some grains for X-ray analysis.

Refractive index measurements

The refractive indices of this sample were determined using a series of immersion oils and a Leitz-Jelley refractometer, with the following results:

$$n_0 = 1.6455 \pm 0.0005$$

$$n_E = 1.622 \pm 0.001$$

$$n_0 - n_E = 0.023$$

Figure 19 is taken from Winchell (1951) and shows the properties of the various tourmaline series. The points marked on this diagram, corresponding to the properties of the Modum tourmaline, indicate that it could either be a dravite containing about 40% of the uvite molecule, or an elbaite containing about 30% of the schorlite molecule. The absence of other lithium compounds in the rock, makes the latter possibility rather unlikely.

Deer, Howie and Zussman (Vol. 1, 1962, p. 312) have plotted the optical properties of several analysed tourmalines against the number of $(Fe^{2+} + Fe^{3+} + Mn)$ ions on the basis of $31(O,OH,F)$. For the iron-poor members of the tourmaline group these properties,

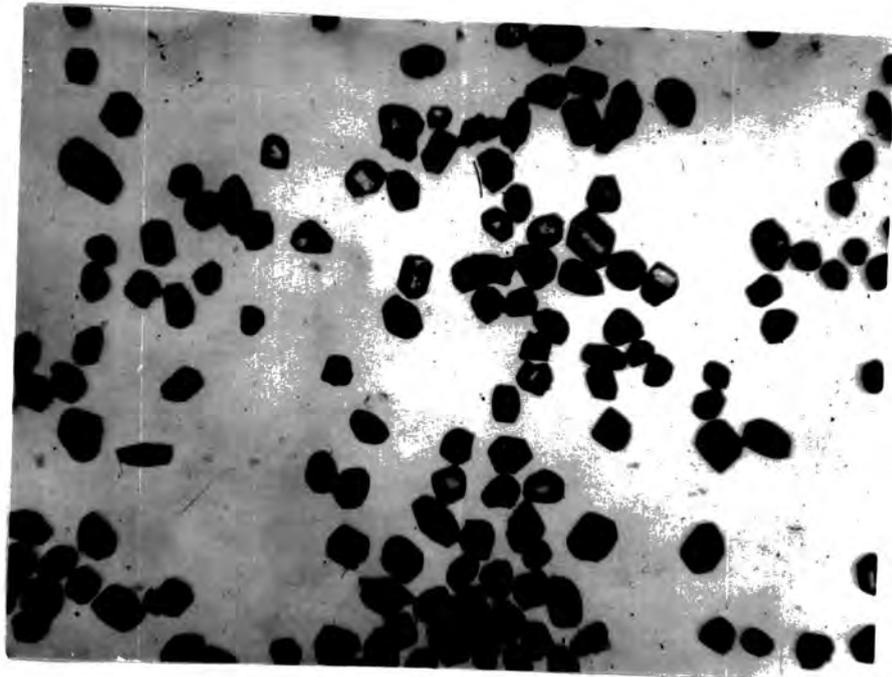


Figure 17: Tourmaline crystals separated from cobaltiferous quartzite, Forhaabning Adit, Skuterud.

Ordinary light.

X 20.

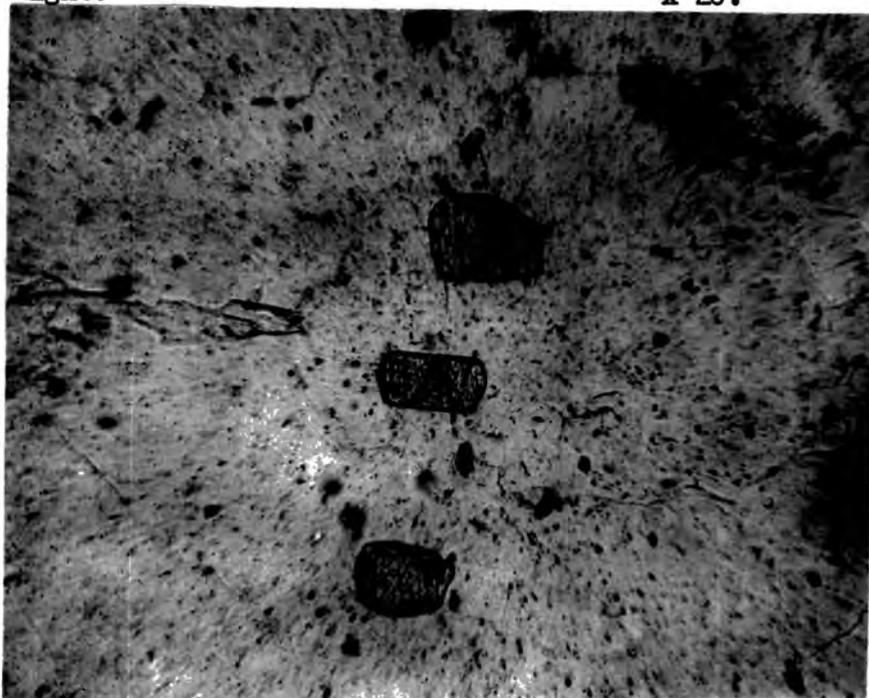


Figure 18: Euhedral tourmaline crystals in siliceous granulite. Central Mine, Skuterud.

Ordinary light.

X 40.

Figure 19: Attempts at correlating the optical properties of tourmaline with chemical composition.

Top: After Deer, Howie and Zussman (Vol. 1, 1962, p. 312).

Bottom: After Winchell (1951).

N.B. The points marked on the diagrams correspond to the optical properties of the Modum tourmaline.

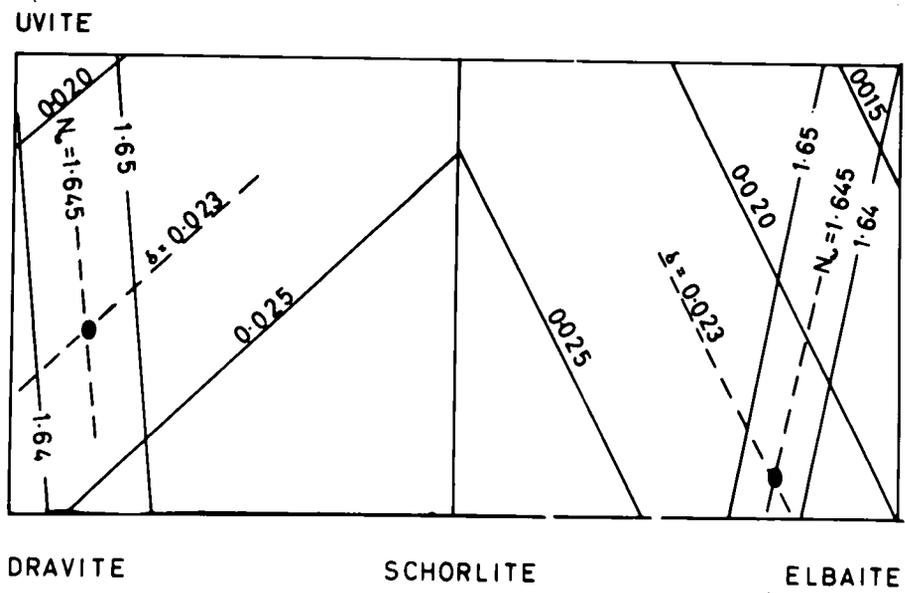
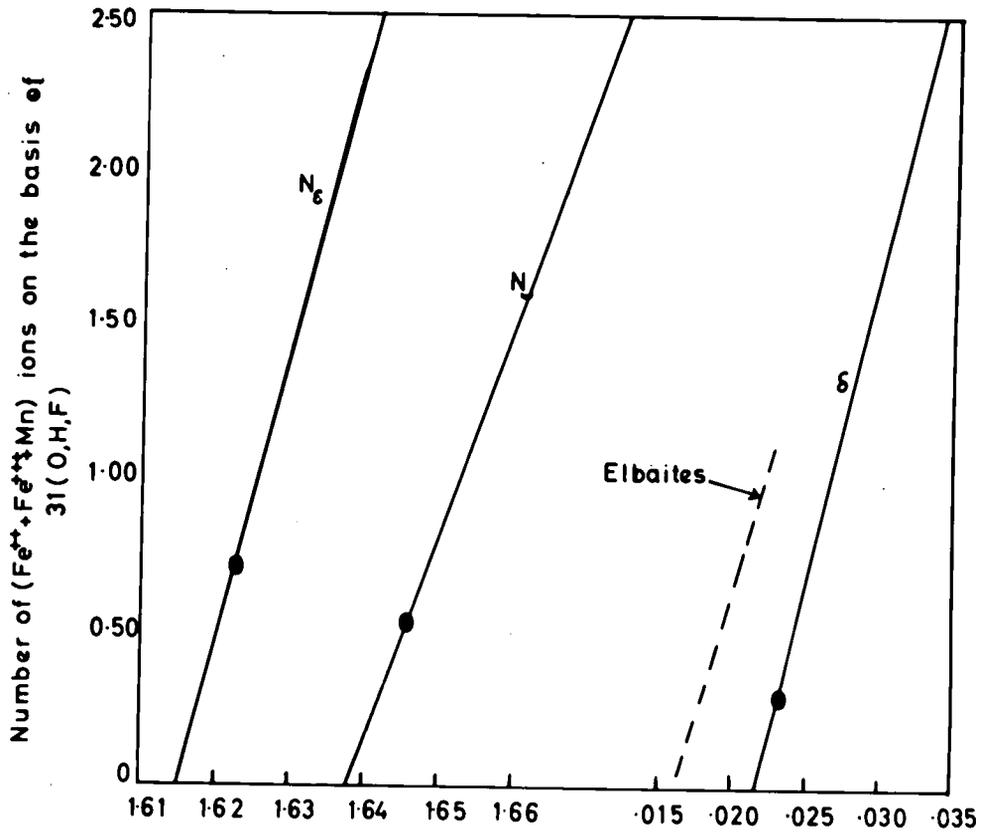


Fig. 19

however, do not distinguish the lithium from the magnesium tourmalines, though the former tend to have lower birefringence than the latter. This diagram is reproduced (Figure 19) with the properties of the Modum tourmaline plotted. It can be seen that the value for the birefringence suggests dravite rather than elbaite.

These data conflict with that of Rosenqvist for the same material. He states (1948 p. 199) "Turmalinen har følgende optiske egenskaper. I makro lys brun, ϵ lys grønnlig nesten fargeløs, ω blågrønn, $\epsilon = 1,637 \pm 0,003$, $\omega = 1,615 \pm 0,003$. Dette er etter Winchell en dravitt."

Presumably there is a mis-print in the above and the last part of the penultimate sentence should read: $\omega = 1.637 \pm 0.003$, $\epsilon = 1.615 \pm 0.003$. However it is certainly not "etter Winchell en dravitt", inspection of Figure 19 shows that Rosenqvist's data fall outside the limits of Winchell's diagram on the dravite side and indicate rather an elbaite. However his data fit Deer, Howie and Zussman's diagram very well (Figure 19) and indicate a pure dravite containing no iron or manganese.

X-ray determination

Epprecht, (1953), carried out precision determinations of the lattice constants on tourmalines of known composition, and demonstrated that two series can be distinguished, between elbaite and schorl and between schorl and dravite. It is thus possible to assign a tourmaline to one or the other of these series, when its lattice constants are known. In view of the ambiguous data derived from the optical determination this method was applied to the Modum tourmaline.

Figure 20: Variation in unit cell parameters of tourmalines in relation to the elbaite-schorlite and schorlite-dravite solid solution series. (After Epprecht, 1953).

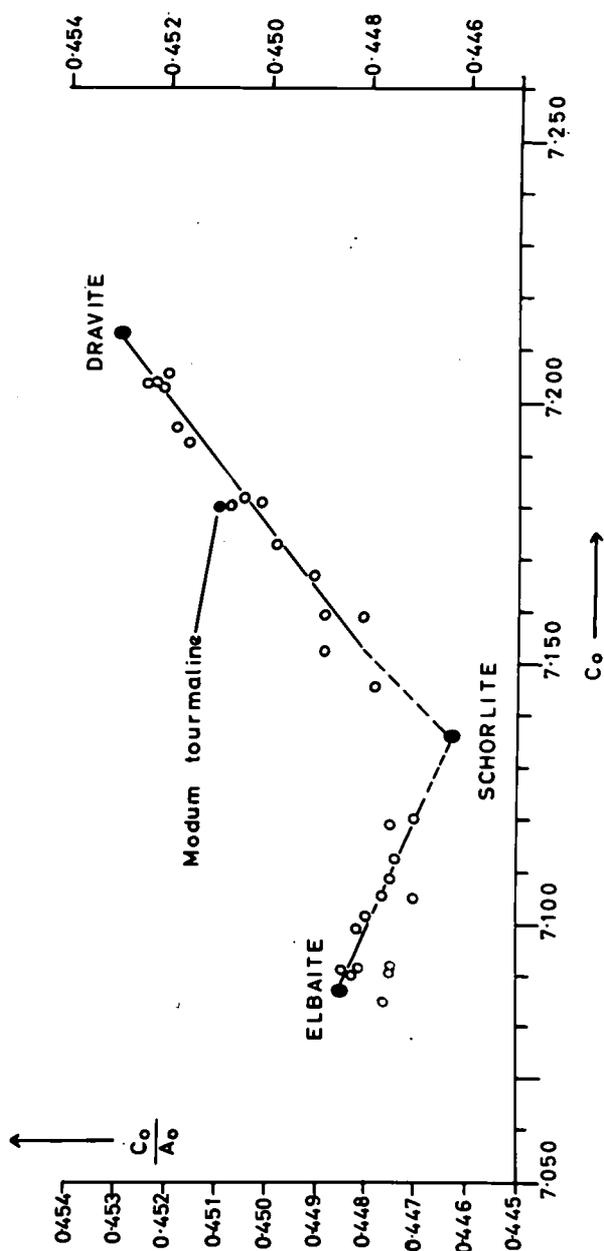
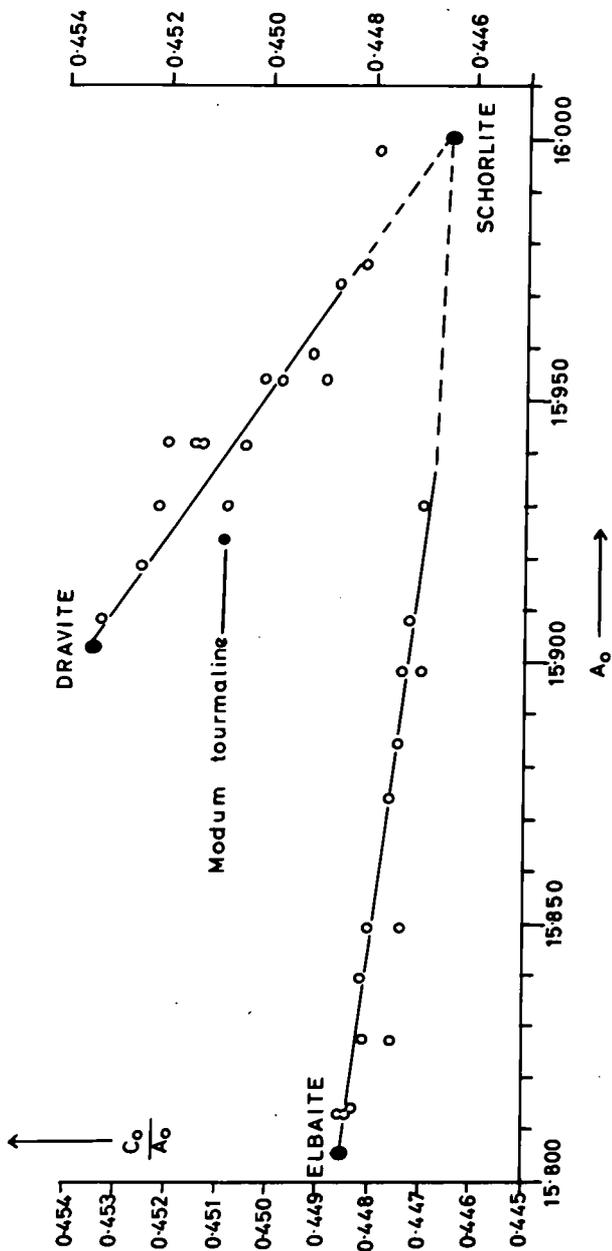


Fig. 20

A powder photograph showing values of up to 43° was taken, using pure $\text{Pb}(\text{NO}_3)_2$ as an internal standard, with a Guinier focusing camera.

The author is indebted to Professor Epprecht for assigning indices to the diffraction lines on this photograph.

By correcting the values of the 'd' spacings with those of lead nitrate, the following cell-parameters were obtained:

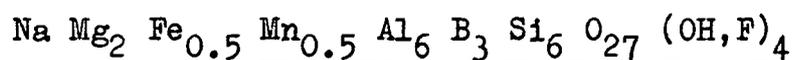
$$a_0 : 15.923 \text{ kx}$$

$$c_0 : 7.180 \text{ kx}$$

$$\frac{a_0}{c_0} : 0.4509$$

Reference to Figure 20 shows that this places the Modum tourmaline as a member of the dravite-schorlite series, with similar lattice parameters to a black tourmaline from Kodarma, India, described by Jacobs (1938). That is, it contains about 65% of the dravite molecule and 35% of the schorlite molecule.

This result is in accordance with the composition derived from the optical data, and indicates a formula probably approximating:



Sillimanite

Sillimanite being such a widespread member of the common mineral assemblages, it was considered desirable to investigate the properties of the mineral.

Optical properties:

$$n_X = n_Y = 1.659$$

$$n_Z = 1.679$$

Birefringence: 0.020

Optic sign: Biaxial positive with $2V \approx 25^\circ$, when measured on cross sections of fairly large crystals. The axial plane was seen to be parallel to the (010) cleavage in two cases.

Dispersion: $r > v$, strong

Cleavage: Good (010). Uneven transverse fractures, inherited by the pseudomorphous 'sericite'.

Colourless and non-pleochroic: Length slow. High relief. Strong (up to 2nd order blue) birefringence except for basal sections which showed low, first order colours.

Extinction: Straight against prismatic crystal edges and (010) cleavage. Symmetrical with crystal outlines in cross-section.

X-ray properties

The diffraction pattern obtained using a 9 cm. powder camera contained the lines characteristic of sillimanite at 'd' values of 1.25, 1.425 and 1.57 Å. This confirms the optical diagnosis and shows that the mineral is not mullite ($3Al_2O_3 \cdot 2SiO_2$). The X-ray diffraction pattern being the only sure way of showing this, as the optical properties are very similar. (Wyckoff et al. 1926; Agrell and Smith 1960).

Table 9
Diffraction data for sillimanite and mullite

I	II	III
3.411 s	3.403 vs	3.401 s
2.88 w	2.89 m	2.89 w
2.691 w	2.686 m	2.683 w
2.541 m	2.543 s	2.543 m
2.42 vw	2.41 w	2.41 vw
2.28 w	2.27 m	2.27 w
2.20 m	2.20 s	2.194 m
2.116 w	2.114 m	2.112 w
	1.97 w	1.97 vw
1.889 vw	1.88 w	1.875 vw
1.833 vw	1.83 w	1.833 vw
1.696 m	1.690 s	1.687 m
1.592 w	1.59 w	1.596 vw
	1.57 w	1.561 vw
1.518 m	1.518 s	1.518 m
1.436 w	1.44 s	1.441 m
	1.425 vw	
1.394 vw	1.39 w	1.393 vw
1.328 vw	1.32 s	1.326 m
1.260 m	1.27 m	1.272 w
	1.25 m	1.250 w

I. Mullite, III. Sillimanite (data from Wyckoff et al. 1926). II. Sillimanite. (Sillimanitic, nodular-biotite-schist, Klara Adit, Skuterud).

N.B: For ease of comparison, the intensity notation used by Wyckoff et al. has been changed as follows:

Faint (f) : Weak (w)
Very Faint (ff) : Very weak (vw)

ESTIMATION OF CHEMICAL COMPOSITION

Recent work (Balasanyan, 1957; Friedman, 1960; Heier, 1961) has shown how the chemical composition of rocks can be estimated from the optically determined mineralogical mode. The accuracy of these determinations depends upon at least two conditions:

- (1) The mineral percentages computed from thin sections must be reliable.
- (2) The optically determined composition of the minerals must be accurate.

Condition 1

The classic procedures described by Chayes (1949) have been followed to determine the relative abundance of each mineral using a 'Swift' point-counter.

Errors in modal analysis are due to two factors, the counting error and the sampling error. Theoretical discussions of these errors are found in Chayes (1956) and Bayly (1960).

As the number of points is increased, the counting error diminishes rapidly. With the medium grain size of the rocks under discussion, it was considered that approximately 1,000 points per section were sufficient to minimise the counting error.

A sampling error occurs when the sample analysed differs from the surrounding rocks. Care was taken to select representative samples and porphyroblastic rocks were not considered amenable to this treatment. Sections were cut normal to mineralogical banding and counting traverses were made at approximately 40° to the banding, as suggested by Chayes (1956).

Condition 2

It has been shown that quartzo-felspathic rocks give better agreement between analysed and calculated composition, than rocks rich in ferro-magnesium minerals (Heier, 1961). This is because of the uncertainties in the optical derivation of chemical composition of ferro-magnesian minerals.

In the rocks of the Modum Formation, the common mineral assemblages are composed of varying amount of quartz, felspar, biotite, muscovite, sillimanite, tourmaline and opaque minerals.

Of these, quartz and sillimanite hardly vary in composition and present no difficulties. The composition of the plagioclase was determined by conventional optical methods. Potassium felspar, which is only a minor constituent of most of the rocks, was assumed to have no soda substitution.

The composition of biotite, muscovite and tourmaline were derived by a variety of methods (see pages 42-64), which enable major chemical variations to be accounted for. This seems to be an improvement on using the 'standard formulae' discussed by Friedman (loc. cit.)

Procedure

The calculation of chemical composition from the modal analysis, involves the conversion of the mode to oxide weight percent by applying the following formula:

$$\text{Wt.\% oxide in a mineral} = \text{Min.vol.\%} \times \text{Min.density} \frac{(\text{Oxide mol.wt.})}{(\text{Min.mol.wt.})}$$

The sum of the derived oxide values from all the minerals are then recalculated to 100%. Uncertainty as to the nature of the opaque minerals has led to their being ignored and included as 'undetermined' in the final analysis.

The mineral formulae and density values employed in these calculations are listed in Table 10.

The parameter (mineral density $\times \frac{\text{Oxide mol.wt.}}{\text{mineral mol.wt.}}$) was calculated for the constituent oxides in each mineral. These values were then set in a matrix of 8 (minerals) by 10 (oxides). Mr. R. Stacey, has developed a computer programme which stores the information in the matrix, and can convert the raw data from a modal analysis, into an estimated chemical analysis in a matter of 20 secs.

Table 10

Mineral formulae and densities used in
estimating chemical composition

MINERAL	FORMULA	DENSITY
Quartz	SiO ₂	2.7
Albite	NaAlSi ₃ O ₈	2.6
Anorthite	CaAl ₂ Si ₂ O ₈	2.6
Potash felspar	KAlSi ₃ O ₈	2.6
Sillimanite	Al ₂ SiO ₅	2.8
Tourmaline(dravite)	NaMg ₂ Fe ^{0.5} Mn ^{0.5} Al ₆ B ₃ Si ₆ O ₂₇ (OH) ₄	3.1
Biotite(phlogopite)	K ₂ Fe ² Mg ₄ Si ₆ Al ₂ O ₂₂ (OH)	2.9
Muscovite	Al _{2.6} Si _{3.2} Fe ^{0.2} O ₁₀ (OH) ₂ K	3.0

Results

The resultant analyses are shown in Table 11 and plotted against weight % silica in Figure 21. The correlation between total iron and magnesium is considered to have little significance as these elements are largely present in the same minerals, the fluctuations merely reflecting the variations in mineral modal proportions.

Table 11.

Estimates of chemical composition.

Specimen number.	22	144	143	13	12	11	116	115	18	19	100	15	98
SiO ₂	56.66	68.52	76.99	77.74	77.99	80.91	81.23	82.50	85.10	86.22	87.90	88.13	90.17
Al ₂ O ₃	10.23	11.77	11.86	7.70	3.07	3.04	3.84	3.96	5.99	2.48	2.28	4.31	2.16
Na ₂ O	1.68	2.31	1.18	2.05	0.02	0.04	0.09	0.12	1.62	0.26	0.07	0.37	0.07
K ₂ O	4.04	2.69	1.46	1.04	2.63	2.38	2.73	2.52	1.61	1.00	1.17	1.06	0.97
CaO	1.33	1.14	0.40	1.29	0.00	0.00	0.00	0.00	0.86	0.10	0.00	0.11	0.00
MgO	4.04	5.58	2.89	1.47	4.60	4.24	4.95	4.68	2.39	1.98	2.20	2.39	1.87
MnO	0.08	0.40	0.36	0.08	0.02	0.05	0.10	0.1	0.00	0.11	0.08	0.24	0.08
FeO	3.57	4.62	3.04	1.24	4.13	3.78	4.37	4.09	2.16	1.68	1.91	1.91	1.60
B ₂ O ₃	0.21	1.03	0.93	0.20	0.05	0.12	0.24	0.35	0.00	0.27	0.20	0.61	0.21
H ₂ O	0.53	0.94	0.70	0.23	0.55	0.52	0.64	0.64	0.28	0.31	0.31	0.45	0.27
Opaque	17.62	0.99	0.20	6.96	6.94	4.90	1.80	1.00	0.00	5.60	3.90	0.40	2.60
Total	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00

N.B. For corresponding modal analyses see Tables 3, 4 and 5.

Localities: 22-semi-pelite, 1000/260 (Fig. 110); 144-felspathic-granulite 150/150 (Fig. 109); 143-siliceous-granulite, 160/130 (Fig. 109) 13-siliceous-granulite, 1000/252 (Fig. 110); 12-siliceous-granulite, 1000/255 (Fig. 110); 11-siliceous-granulite, 1000/260 (Fig. 110); 116-siliceous-granulite, Middagsville Mine; 115-siliceous-granulite, Middagsville Mine; 18-siliceous granulite, 1000/226 (Fig. 110); 19-quartzite, 1000/205 (Fig. 110); 100-quartzite-690/265 (Fig. 109); 15-quartzite-1000/240 (Fig. 110); 98-quartzite-685/265 (Fig. 109)

Figure 21: The estimated chemical compositions of some Modum Formation rocks, as shown in table 11, plotted as oxide weight-percent against SiO_2 weight-percent.

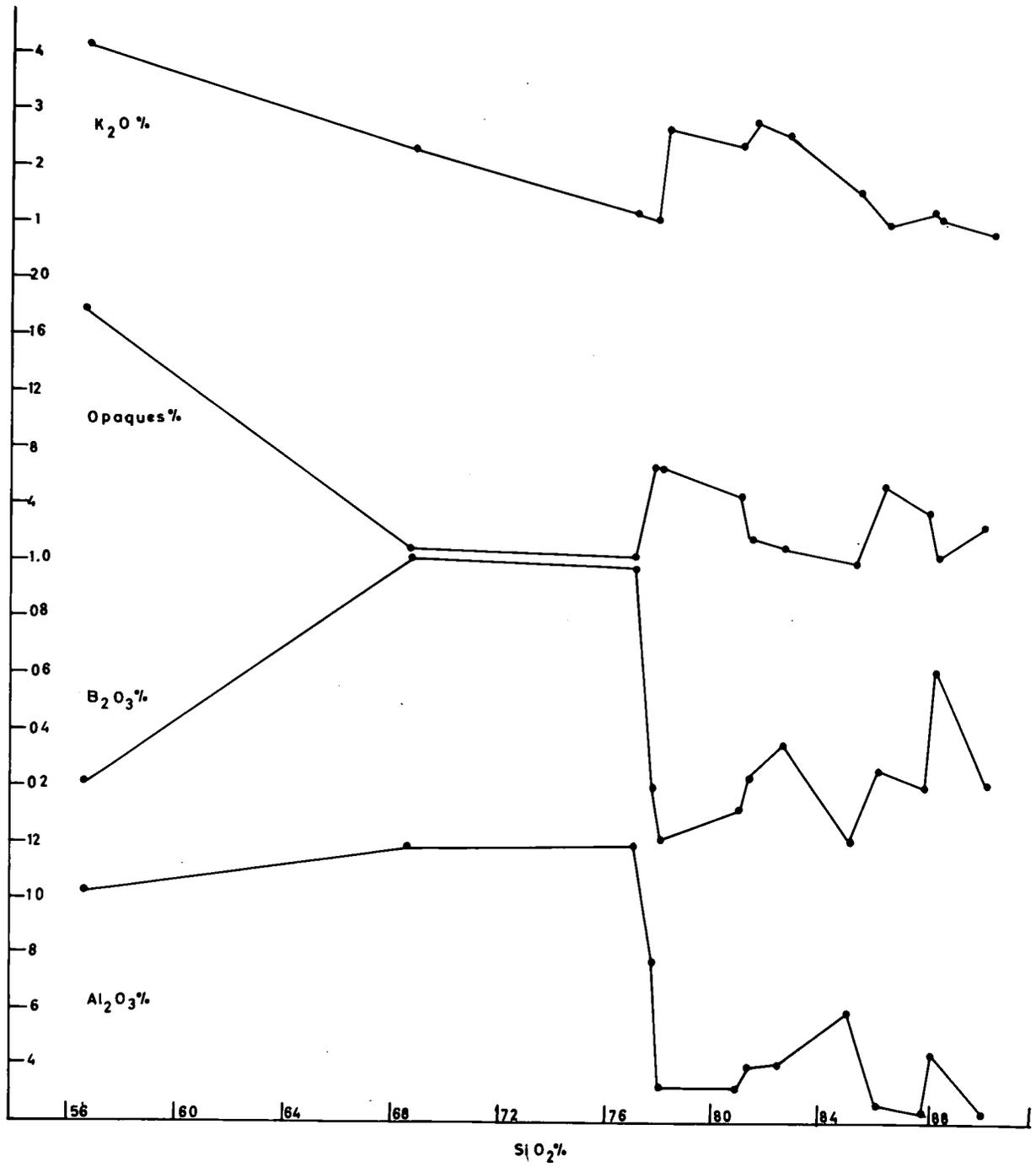


Fig. 21

Of more interest, however, is the correlation between Al_2O_3 and B_2O_3 . The aluminium is present in a large variety of minerals (felspar, sillimanite, muscovite, biotite and tourmaline) while boron is present only in the tourmaline. The method of converting the modes to estimated chemical analysis has, in this instance, been valuable in showing a chemical relationship, which could well have been overlooked, from a simple inspection of the mineral modal proportions.

It must be emphasised that this method is no more accurate than the data that is used. It is no substitute for a conventional chemical analysis and results obtained must be treated as approximations.

PETROGENESIS

The origin of the various rocks is a problem fundamental in any areal geologic study. In deeply eroded areas it becomes largely an evaluation of the metamorphism. Here an attempt has been made to trace the metasedimentary rocks back to their original sediment and delineate their metamorphic history.

Nature of the sediments

The presence of pure quartzites and calcareous horizons, well-developed mineralogical banding, relict graded bedding (p. 31) and rounded zircons (p. 36), leaves little doubt that the Modum Formation represents a metasedimentary sequence.

Quantitatively, the most important rock types are the siliceous granulites together with their felspathic and micaceous variants. Quartz-rich sedimentary rocks are characterised by high K:Na ratios (Pettijohn, 1957). The widespread occurrence of quartz-biotite-oligoclase gneiss with only scarce potash feldspar may thus imply some material transfer.

Rosenqvist (1948) has postulated a process of 'degranitisation' for the Modum rocks similar to Eskola's (1933) differential anatexis. This process implies the removal of alkalis, principally potassium, some silica, alumina and water from the rock complex, leaving it relatively enriched in sodium. Rosenqvist considers the deficiency in water would make sillimanite, muscovite and potash feldspar stable in the same rock. This unusual mineral assemblage is common at Modum (Rosenqvist 1948; p. 200; Bugge 1943 p. 16-18). The potash rich 'granite ichor' produced in this way would then be available for the production of the migmatites,

granites and pegmatites so common in the Kongsberg - Bamble Formation.

Engel and Engel (1953) rejected a similar hypothesis when studying quartz-biotite-plagioclase gneisses of the Grenville Series in the Adirondack region. They considered that the uniform abstraction of material from large areas, and from different metamorphic facies, to produce a gneiss with a uniform chemical composition was unlikely. At Modum the siliceous granulites vary considerably in composition both laterally and across the strike, furthermore the metamorphic grade is believed to be more or less constant, within the area studied. The intrusive relationship into the gneisses of the potash rich rocks in the Adirondacks was considered incompatible with differential anatexis, it is here suggested that if the potash rich 'ichor' was derived from the gneisses to form migmatites the latter should be expected to have apparently intrusive relations. According to Eskola (1933, p. 24) the residue should be higher in the 'excesses' - total silica and aluminium - than the 'exudate', this relationship is not observed in the Adirondacks, but at Modum the 'residual' rocks contain higher percentages of quartz and sillimanite than the 'transformed' rocks. Engel and Engel consider that the volume of felsic material required to explain the observed migmatitic zones is so large that all relict sedimentary structures would have been obliterated during its abstraction from the gneisses. In the Kongsberg - Bamble Formation as a whole the amount of felsic material to be produced is not excessive compared with the widespread tracts of sillimanitic siliceous granulite. That material

can be removed without major textural disturbance is suggested by the common quartz segregation veins in quartzites showing clear metasedimentary colour banding.

Metamorphic rocks with high Na :K ratios could be derived from greywackes of suitable composition. The presence of large thicknesses of greywacke suggests rapid weathering, transport and deposition in an unstable crustal environment (Pettijohn, 1957). It is difficult to reconcile this with the presence of calcareous sediments and well sorted quartzites stratigraphically present in the same succession.

The addition of soda to clastic particles of common clay minerals by diagenetic interaction with sea water trapped in the sediments would also explain the Na:K ratio. However, Grim et al. (1949) have shown that clays observed forming in contemporary shelf seas are relatively impoverished in sodium and especially enriched in potassium.

However, there is a strong possibility that the above arguments are based on a false premise. The high plagioclase: potassium feldspar ratio is usually taken to imply a high Na:K ratio. The estimated analyses of the rocks, discussed on pages 65-70, show that this may well be fallacious as considerable potassium is present in the micas. Thus the Na:K ratio of the quartz rich metamorphic rocks does not necessarily conflict with the ratio observed in quartz rich sediments.

Thus the question of the ultimate origin of the major rock type still remains open, but it is possible that it represents

a normal sequence of shales and argillaceous sandstones, themselves the product of moderate to intensive residual weathering and good sorting. These may have been affected by differential anatexis.

If the mineralogical banding represents a modified relict bedding, then the bulk of the succession consisted of alternating units of shales and sandstone with occasional graded units. Thicker sandstone units are represented by the persistent, uniformly layered clear quartzites. The small diopside-actinolite skarn lenses (p. 41) possibly represent calcareous concretions or thin calcareous horizons in the original sedimentary sequence (Ramberg 1952, p. 122). The calcareous granulites represent more persistent horizons of a similar nature.

The presence of intermittent reducing conditions is indicated by the sulphidic graphite schists common in the sequence (see discussion p. 120).

Salinity

The only clue as to the salinity of the environment is provided by the boron content, now present as tourmaline. This boron could have been introduced from several different sources:

- (1) Pneumatolytically from an igneous source
- (2) Metasomatically from a distant, not necessarily igneous source

or

- (3) As part of the deposited sediments

Evidence in favour of a later introduction for the boron may be found in the textural relations which suggest that tourmaline is one of the latest minerals to crystallise, (p. 28). It has been

suggested (Neumann, personal communication) that the fahlband zones represent metasomatically altered metasediments, into which tourmaline and the sulphide ores have been introduced. The available textural evidence suggests, however, that the ore minerals are later than the tourmaline in every case. Moreover, there is no correlation between tourmaline content and sulphide mineralization (Figure 22).

Tourmaline in the Otago schists of New Zealand is considered by Hutton (1939) to indicate pneumatolytic activity. This conclusion was reached for several reasons which are not valid at Modum. At Modum tourmaline is only found in rocks of obvious sedimentary parentage, while in New Zealand, it is also found in metamorphosed basic intrusives. In the Otago schists the tourmaline is associated with shear zones, while at Modum it is ubiquitous at favourable horizons and clearly follows the mineralogical banding. Metallic sulphides are present at both localities, but, for the reasons advanced above, it is considered that the Modum tourmaline predated the introduction of much of the ore. At Otago there is a possible igneous source close at hand, a condition not fulfilled at Modum. In the Modum Formation tourmaline is found concentrated in micaceous and sillimanitic bands which are comparable to clayey sediments in bulk composition, this is not the case in New Zealand.

If the tourmaline is of pneumatolytic origin, the most likely time for its introduction would be during the formation of the 'Younger Complex', when granites and pegmatites were widely emplaced throughout the region. The granite pegmatites common in Modum, do indeed carry considerable tourmaline, but this is present

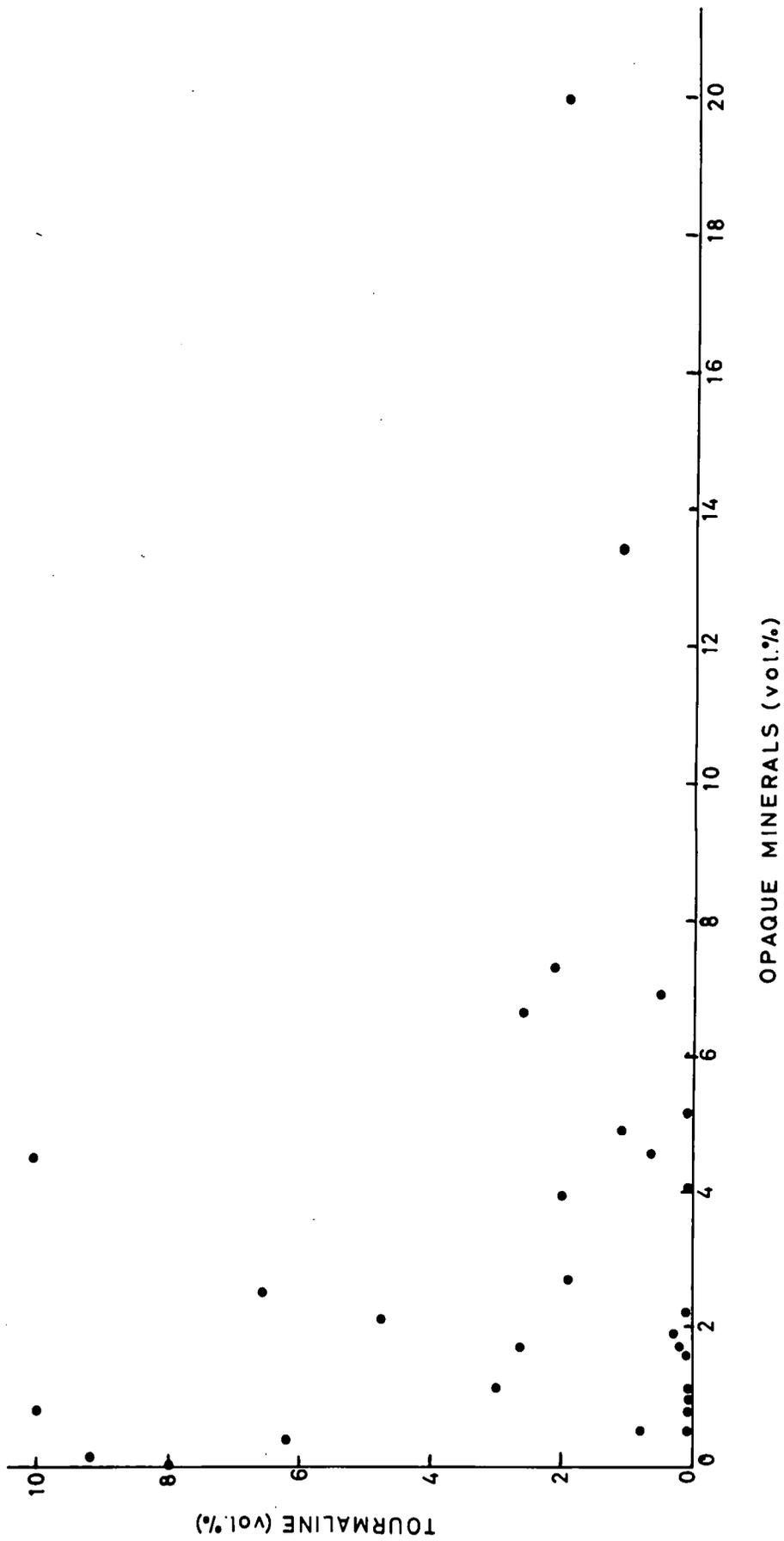


Fig. 22

as black shorlite, totally unlike the pale yellow dravite of the metasediments.

However it is possible that the Mg:Fe ratio of the tourmaline is controlled by the composition of the rock where the tourmaline is developing (Simpson 1931). The availability of Mg rather than Fe at Modum might be indicated by the mineral assemblages - dravite, phlogopite, anthophyllite, cordierite. Thus if boron was introduced from an outside source, either as igneous emanations or as alkali borate fluids derived from the metamorphism of marine clay sediments (FrondeU and Collette (1957), it is possible that a dravitic tourmaline would be the most likely product. Thus while the post-diagenetic introduction of boron is not considered likely, the possibility must remain open.

If the boron was present as part of the deposited sediment it could have originated in detrital material or been absorbed from sea water by the clay minerals.

Serdychenko (1956) has discussed the recrystallisation of original detrital grains of boron containing minerals under metamorphic conditions. The arguments against regarding the Modum dravitic tourmaline as recrystallised detrital material can be summarised thus:-

- (1) The tourmaline rich bands are as frequently associated with 'shale' horizons (micaceous and sillimanitic bands) as with the more 'sandy' (quartzitic) bands, which would be expected to contain the greater amount of 'heavy' detrital grains.

- (2) There is a considerable variation in observed grain size, which is not consistent with a well sorted, sedimentary accumulate.
- (3) The dravite is seen growing, blastically, around detrital zircon grains.

It has long been recognised that small amounts of boron are present in sediments of argillaceous type, the boron being derived from sea water and being held by absorption. (Goldschmidt and Peters, 1932; Landergrén, 1945). Experiments conducted by Goldberg and Arrhenius (1958) indicated that this boron is proxying for silicon in the tetrahedral sheets of the clay minerals. Harder (1959) and Frederickson and Reynolds (1961) independently discovered that most of the boron was present in the clay mineral illite. The boron content of shales is one of the characteristic differences between marine and non-marine sediments (Degens et al., 1958; Krasintseva et al., 1959). It has been shown (Pieruccini, 1950) that there may be as much as 0.15% B_2O_3 in certain marine sediments.

FrondeLL and Collette (1957) conducted experiments on the reaction of mineral grains with $NaCl - H_3BO_3$ solution. They found that tourmaline was readily produced from Al, Fe, Mg and Si bearing minerals, and concluded that:

"....tourmaline can be expected to form in the presence of Na and borate ions during the metamorphism of argillaceous sedimentary rocks, and by the action of these ions on even the high grade equivalents of such rocks."

They considered that a possible source of the necessary boron would be that originally present in the clay minerals of marine sediments. The boron would become available:

".... through desorption attending grain growth, or by recovery from solid solution in unstable phases during their recrystallisation."

Reference to Figure 21 shows that there is a positive correlation between B_2O_3 and Al_2O_3 for rocks of the Modum Formation. Similar relationships for these elements have been noted from analogous metamorphic terrains in Sweden (Hjelmquist, 1938) and Greenland (Ellitsgaard-Rasmussen, 1954). It is here suggested that this relationship is highly significant and indicates that the boron could have been present in the aluminous, clay-rich original sediments.

On recrystallisation and metamorphism this boron would be liberated to react with suitable minerals to form the nuclei for tourmaline crystals. The complete recrystallisation of the other phases present to give sillimanite, mica and quartz would enable the boron solution to be mobile until a late stage in the metamorphism. This could account for the locally very high concentrations of tourmaline, and the textural appearance whereby tourmaline seems to be one of the latest minerals to crystallise.

Conclusions

If this latter hypothesis be correct then we can consider the Modum Formation to represent a sequence of metamorphosed sediments of marine origin. The original nature of these sediments may have been modified by differential anatexis and the liberation of boron containing fluids derived from the reconstitution of illitic clay minerals. The metamorphic reactions resulting in the formation of the present mineral assemblages are considered in the next section.

METAMORPHIC REACTIONS

Quartz

The strained extinction and sutured margins of the quartz grains indicate that the mineral had not reached an ideal, unstrained state at the close of the metamorphic episode. That quartz was one of the most mobile minerals in the rock is indicated by the way that it appears to be replacing all other minerals except the ores.

The texture of some of the very pure quartzites suggests that the quartz in them may have reached a stable state at the close of metamorphism. The quartz is seen in polygonal grains having junctions with other crystals approaching an ideal angle of 120° (Figure 6). This indicates (Voll, 1961; Rast, 1964) that grain boundary migration has taken place towards the centre of curvature of adjacent grains until the ideal case, with stable junctions, is attained.

Biotite

The micaceous bands presumably represent more argillaceous horizons in the original sediment. Considerable interest is attached to the abundance of very pale, phlogopitic, or Mg-biotite as opposed to more iron rich types.

Reference to the literature suggests that Mg-rich biotite is more common at high metamorphic grades (Barth, 1936; Lambert 1959). Engel and Engel (1960) found an antipathetic relationship between Fe and Mg, with Mg predominating at higher metamorphic grades. Tozer (1955) found that the biotite present in the contact aureole of the Donegal granite where it intrudes Dalradian became paler in colour as the amount of sillimanite increased in the rock. It has been suggested (Deer, Howie and Zussman, 1962, Vol. 3, P. 72) that an iron rich biotite is likely to react at lower P-T conditions than iron poor biotite. Thus making iron poor biotite the stable biotite under higher P-T conditions.

However, it should be mentioned that Snelling (1957) found no variation in the Fe:Mg ratio of biotites with increasing grade, while De Vore (1955) and Miyashiro (1956) have demonstrated that in some cases the Fe:Mg ratio increases with metamorphic grade.

At Modum Mg-rich biotites are common only in certain members of the succession, both the sillimanitic, nodular biotite schist (p. 38) and the biotite schists associated with amphibolite (p. 50) have biotite interpreted as being iron-rich. Thus there seems to be no simple correlation of colour with metamorphic grade.

The rocks containing Mg-rich biotites all contain varying amounts of sulphide minerals, and are very similar to the 'pyrrhotite-gneiss' described by Fisher (1962) from the Bethel area, Maine, which also contains very pale biotite compared with that present in other rock types in the area. At Modum it is often noticeable that biotite is bleached in proximity to ore minerals, frequently becoming quite colourless (Figure 23).

Similar observations on the colour of biotite in sulphide rich gneisses have been made by Varynen (1928), Hackman and Wikman (1929), Vahatala (1953), Karl (1952) and Marmo (1957). Varynen (loc. cit.) correlated the colour change with the graphite also present in the sulphidic schists he was studying. Karl (loc. cit.) suggested that the pale biotite was a product of the recrystallisation of normal biotite under conditions where the iron could enter other crystal forms. Marmo (loc. cit.) suggested that the colour was proportional to the sulphide present in the rock, particularly the pyrrhotite content. In this connection it is interesting to note that Deer, Howie and Zussman (1962, Vol. 3, p. 69) suggest that the presence of ferrous iron in biotites facilitates alteration by H_2S to yield pyrite and other sulphides.

Experimental results bearing on the problem have recently been reported by Kullerud and Yoder (1963). Their experiments demonstrate that sulphur reacts readily with many rock forming silicates over wide temperature-pressure ranges. Elements, such as iron, originally present in the silicates react to produce sulphides and oxides. Only moderate amounts of Fe in the original silicates are sufficient to change the mineralogy of the rock profoundly.

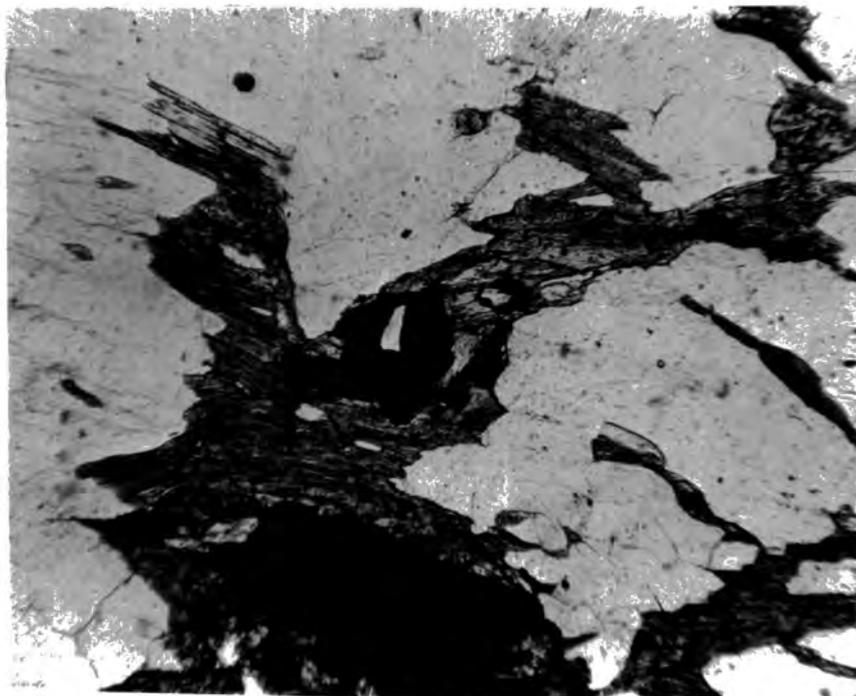


Figure 23: Biotite bleached in proximity to ore minerals.
Sulphidic siliceous granulite, Central Mine, Skuterud.
Ordinary light. X 40.

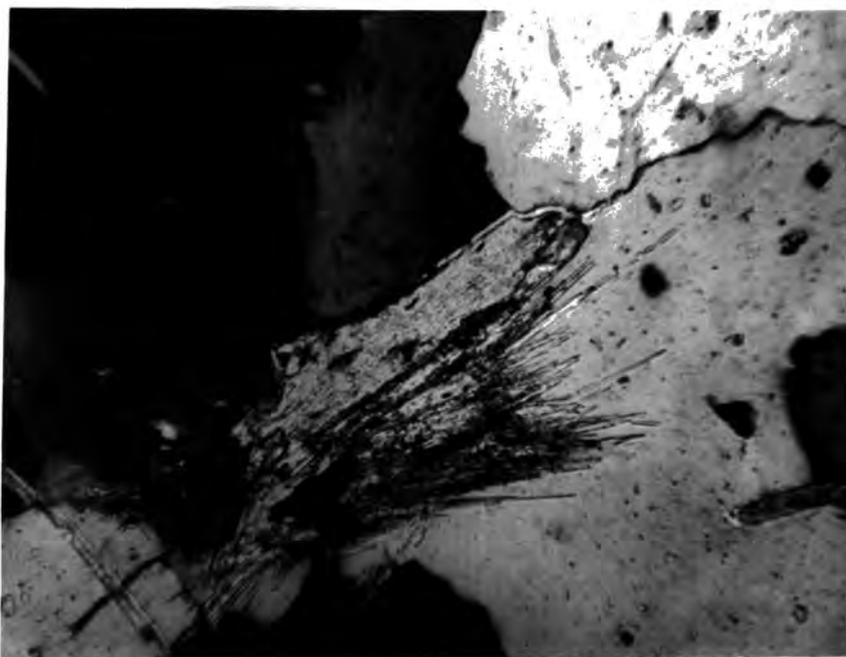


Figure 24: Sillimanite needles growing from a biotite grain. Siliceous
granulite, South Mine, Skuterud.
Crossed nicols. X 40.

It is here suggested that the pale colour of the biotites observed in the sulphide rich assemblages may be due to sulphurization of some of the iron originally present in the biotites. Discussion of the origin of the necessary sulphur will be deferred until a later section (p.202).

This hypothesis may help to explain why the biotite in the sillimanitic, nodular biotite schist is richer in Fe and Ti (see p. 50). Given two pelitic sediments of similar composition, with sulphur present in only one of them, consider what will happen on metamorphism. As metamorphism progresses biotite, in the sulphurous schists, will lose iron by sulphurization; Ti is incompatible with iron-poor biotite and will separate as rutile (Fisher, 1962). In the non-sulphurous sediment both Fe and Ti will be retained in the biotite until high metamorphic grades. At the sillimanite grade the typical assemblages to be expected would be:

(a) Sulphurous:

Mg-biotite + Fe sulphide + rutile ± quartz, feldspar, sillimanite, etc.

(b) Non-sulphurous:

Fe-Ti biotite ± quartz, feldspar, sillimanite, etc. without sulphides.

These assemblages are those observed in the sulphidic siliceous granulites and the nodular sillimanitic biotite-schist respectively.

Sillimanite

Sillimanite has been used as a zonal index mineral in metamorphosed pelitic rocks (Barrow, 1912), the sillimanite zone representing rocks in the highest grade of metamorphism, consisting

typically of micaceous sillimanite schist or coarse quartz-sillimanite gneiss (e.g. Sanders 1954).

The reaction usually considered to take place at the sillimanite isograd is (Turner and Verhoogen 1960):

muscovite + quartz = potash felspar + sillimanite + water.

At Modum, these minerals occur together, but the textural evidence suggests that the muscovite is replacing the sillimanite at a retrograde stage rather than the reverse. Similar associations have been frequently noted in the literature. (Tattam, 1929; Williams, 1934; Billings, 1938; Sanders, 1954, Iyengar et al. 1954 and Tozer 1955).

Sillimanite may also arise from the reaction:

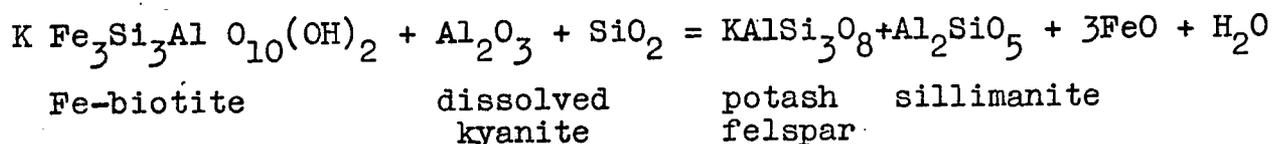
3 staurolite + 2 quartz = almandine + sillimanite + 3 water.

This reaction has been used by Chapman (1952) to define the sillimanite isograd, but is not applicable here since the rocks richest in sillimanite at Modum contain no garnets.

Rather the petrographic evidence suggests that sillimanite arises from the breakdown of biotite. Fibrolitic sillimanite fringes biotite flakes and grows out of their terminations. The proximal ends of the needles are often embedded in the biotite and it is frequently difficult to distinguish the boundaries between them, (see for example, Figure 24).

An examination of the literature dealing with sillimanite in metamorphic terrains shows that a close association has been observed between biotite and sillimanite, (Hall, 1909; Hall and Molengraaf, 1934; Tattam, 1929; Turner, 1933; Williams, 1934; Watson, 1949; and Tozer 1955). The frequency with which this

genetic association occurs in a diversity of environments makes it likely that this is the most common paragenesis of sillimanite. However Ramberg (1952, p. 57) describes the breakdown of biotite at high grades to give potash feldspar and garnet rather than sillimanite. Francis (1956) has suggested that the sillimanitization of biotite is a variant of this reaction, when alumina in solution, derived from kyanite of lower grades, may cause a reaction in which sillimanite appears instead of garnet.



It is considered that this latter reaction is in best agreement with the mineral associations observed at Modum.

It is apparent from the literature that there is often a close coincidence between the position of the sillimanite isograd and the limits of areas of granitic injection, migmatization or granitisation. (Read, 1931; Moore, 1949; Wyckoff, 1952; Smith, 1945; Shackleton, 1946; and Sanders, 1954). Indeed Watson (1948) has advanced the view that the mineral originated at a late stage in the history of migmatization rather than as a high grade metamorphic mineral.

This association of sillimanite with granitic rocks is also true for the Kongsberg - Bamble Formation as a whole. A possible explanation for the genetic association of quartz-sillimanite rocks and potash rich rocks has already been suggested. (p.71-74).

The origin of the quartz-sillimanite nodules so frequently observed presents a problem. In closely adjacent rocks there may be many sillimanite needles scattered throughout the rock as a whole or the virtual restriction of the mineral to quartz-sillimanite nodules. Most probably, these lenses of material rich in silica and alumina were produced during the intense plastic deformation affecting the

rocks, by the disruption of quartz rich layers into a series of disconnected lenses at the same horizon. Alternatively, it is possible that the lenses were an original sedimentary feature representing aluminous inclusions such as clay galls. The abundance of nodular gneisses throughout large areas of the Kongsberg - Bamble Formation makes this latter possibility rather unlikely. It has also been suggested (Starkey, 1960) that similar nodules have been formed through the removal of potash from the rock. The shape of the resulting nodules depending upon the ease with which the migration of material, and the growth of the minerals occurred within the fabric of the host rock.

Tourmaline

The possible origins of this mineral have already been discussed at some length (see p.74-79).

Plagioclase

The anorthite content of plagioclase has commonly been considered to increase with increasing grade of metamorphism, (Ramberg, 1952). In the rocks of the Modum Formation the general range observed is An_{15} to An_{35} , with an average value for the feldspathic siliceous granulites of around An_{30} . It is here suggested that the variations represent differences in the bulk compositions of the rocks, particularly the availability of calcium, rather than a simple response to metamorphic grade, which is believed to be more or less constant throughout the formations investigated.

OLIVINE-GABBRO, AMPHIBOLITE AND BIOTITE SCHIST

Introduction

The field relations suggest that the basic rocks are intrusive into rocks of the Modum Formation and predate the widespread granitic pegmatites. They are therefore correlated with the 'hyperites', which elsewhere in the Kongsberg - Bamble Formation separate rocks of the 'Older' and 'Younger' complexes.

Evidence will be presented to show that there is a genetic connection between the gabbros and the amphibolites, and the amphibolites and the biotite schists. It is thus convenient to consider them together in this section.

Olivine-gabbro

Areas of basic intrusive rocks are common in the Modum district. Similar rocks have been recognised over considerable areas in the Precambrian basement of Southern Norway and Sweden. (Tørnebohm, 1877, Vogt, 1895; A. Bugge, 1922; Brøgger, 1934; C. Bugge, 1918; Hofseth, 1942). These have been termed 'hyperites', a term originally introduced by Swedish geologists to cover rocks of gabbroic, olivine-gabbroic and noritic composition of apparently igneous origin. Hyperite is now defined as: (A.G.I., 1960)

"A plutonic rock composed essentially of calcic plagioclase, augite, hypersthene and olivine. An olivine bearing variety of morite."

However, in the Scandinavian literature the term is very loosely applied and is used more in reference to the field relations than to the mineral composition of the rock concerned. Here it is proposed to ignore the term hyperite and adhere to the more customary classification of the basic rocks. (Turner and Verhoogen, 1960).

Field relations

The gabbros are resistant to weathering, forming sweeping ridges and dome-shaped hills. In shape they are usually elongated parallel to the strike of the surrounding gneisses. The most characteristic feature of these bodies is the zonal arrangement of rock types encountered from the centre to the margin of the intrusive.

In the unfoliated core primary igneous minerals with a typical plutonic texture characteristic of gabbros occur; olivine is clearly visible in hand specimen. A corona becomes increasingly apparent around the olivine crystals in proceeding outwards until a gabbro containing no visible olivine is attained. Then by advancing amphibolitization of the ferromagnesian minerals a marginal, well-foliated amphibolite is produced. In some cases the marginal amphibolite eventually gives place to a biotite schist.

Figure 25 is redrawn from Jø/sang's manuscript and shows two of the intrusive masses studied in the field during the present work. Of particular interest is the fact that the Morud gabbro is emplaced in the nose of a plunging antiform. The unaltered core has been disrupted into three separate areas, and the enveloping amphibolite has been drawn out round the fold closure, suggesting syn-kinematic emplacement. The foliation of the amphibolite and the adjacent gneisses is always concordant.

The Hovdekollen gabbro also has three distinct unaltered regions, which form the present topographic high points. Very good exposures are found in the marginal amphibolite in a road cut south of the bridge across the Snarum River. Mineralogical banding is well developed at this locality with hornblende, feldspar and garnet

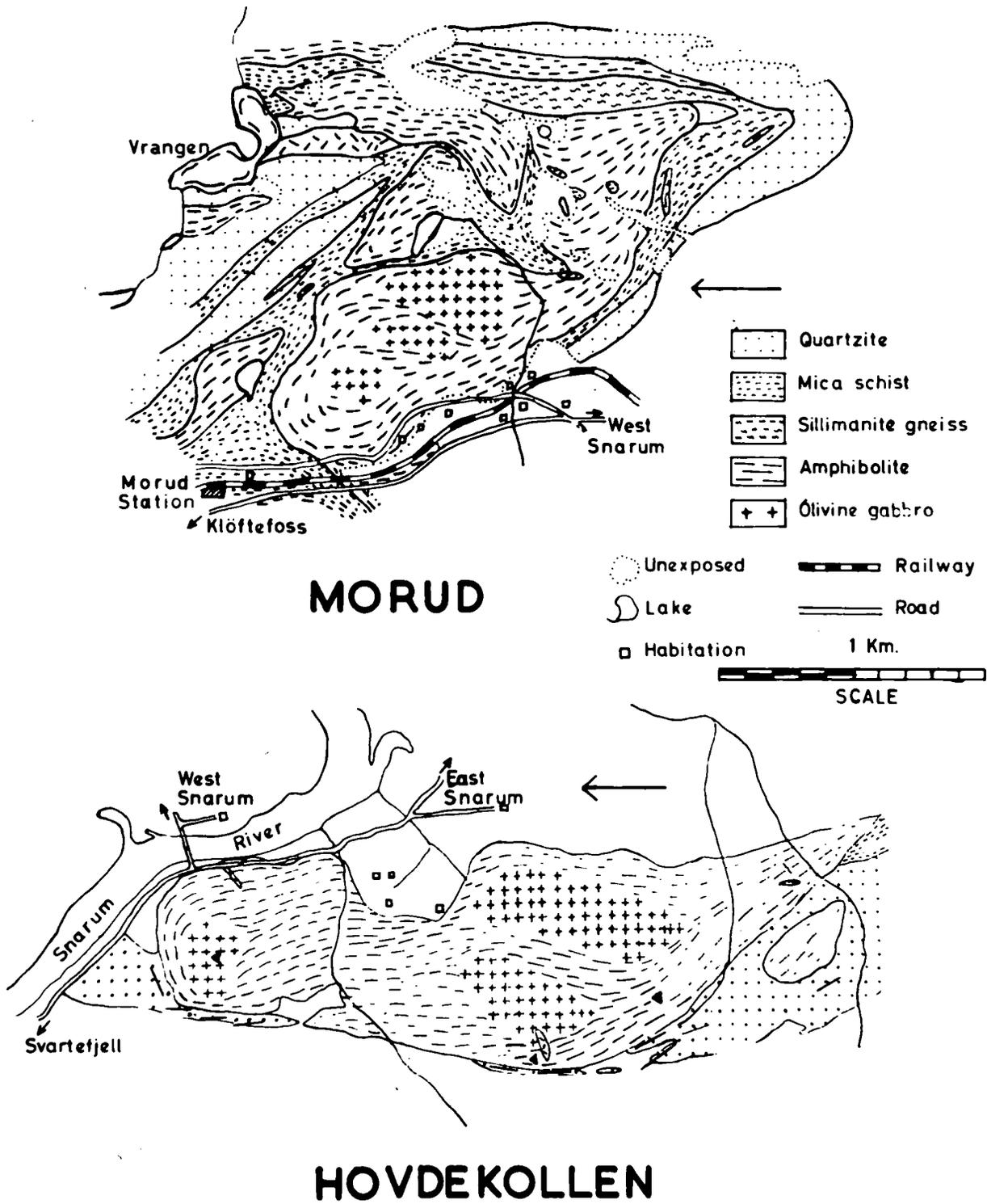


Fig. 25

rich layers alternating. These are of particular interest as they seem to be sub-horizontal with a slight dip in towards the centre of the intrusive. The surrounding gneisses have the same strike direction but are sub-vertical. At this locality the amphibolite is crossed by a granite pegmatite dyke.

Of particular interest to the present study is a small outcrop of olivine-gabbro at the junction of a thick garnet-amphibolite layer with sillimanitic siliceous granulite, due east of the Central Mine at Skuterud. Exposure at this locality is poor but the field relations suggest that the olivine-gabbro passes by gradual transition into the main amphibolite body.

Petrography

The following description is based upon specimens, collected from the core of the Morud olivine-gabbro and the small olivine-gabbro lens at the eastern contact of the Skuterud amphibolite mentioned above. The six thin-sections (M8, M9, 93, 467, 733 and 739) all show similar features.

The igneous texture of the rock is clearly recognizable. Olivine crystals and large plagioclase laths are enclosed by large ophitic crystals of clinopyroxene.

Olivine, shows no sign of zoning and is found in hypidiomorphic grains. It has altered to serpentine and magnetite along fractures. The predominant mineral is plagioclase, forming about 50% of the rock. The large laths (up to 8 x 2 mm.) are elongated parallel to (010). Albite twinning is found in almost every grain and pericline and Carlsbad twins are common. Some laths show a combination of Carlsbad and pericline twins which produce a 'herring bone' pattern (Figure 26). In others pericline twins are superimposed on albite twins in ill-defined patches of the crystal (Figure 27). The crystals are clearly zoned (Figure 28), some zones being particularly rich in a minute dust of brownish opaque material. The cores of the crystals are labradorite (An_{69}) and



Figure 26: Combined Carlsbad and pericline twins producing a 'herring-bone' texture. Olivine gabbro, Morud.

Crossed nicols.

X 40.



Figure 27: Pericline twins superimposed on albite twinning. Olivine gabbro, Morud.

Crossed nicols.

X 40.



Figure 28: Zoned plagioclase, Olivine gabbro, Morud.

Ordinary light.

X 40.

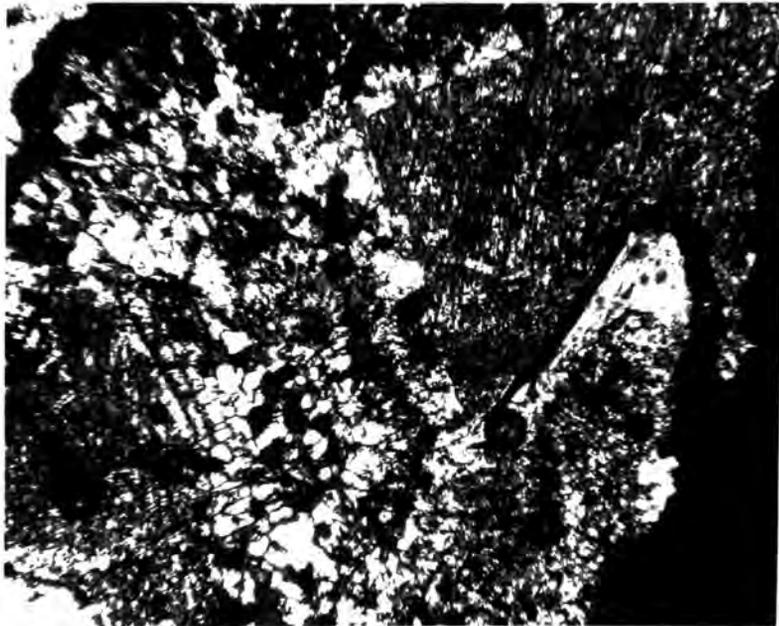


Figure 29: Rims of orthopyroxene developed between olivine and mesostatic clinopyroxene. Olivine gabbro, Morud.

Crossed nicols.

X 40.



Figure 30: Development of conjugate corona between olivine and plagioclase. The inner corona is of orthopyroxene and the outer corona is of fibrous actinolite. Olivine gabbro, Morud.

Crossed nicols.

X 40.

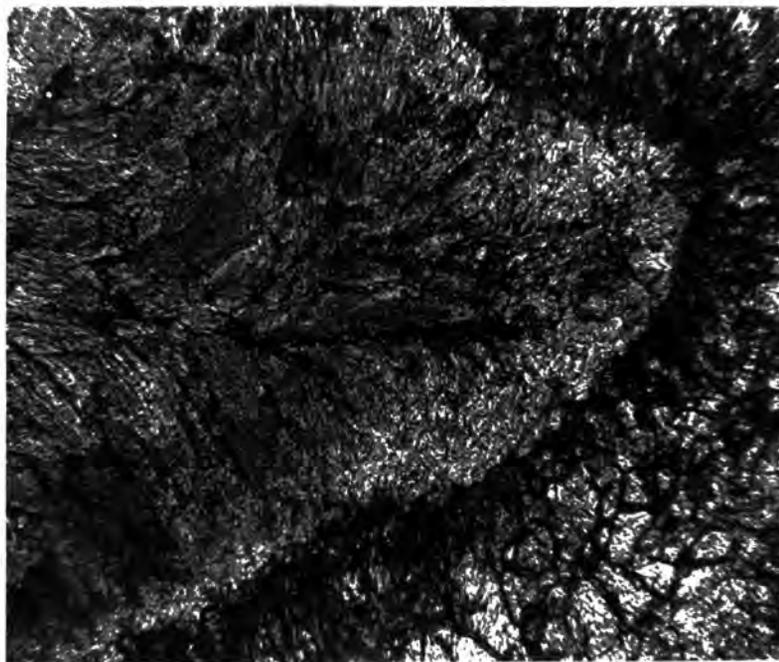


Figure 31: Chains of pleonaste in actinolite corona. Olivine gabbro, Morud.

Ordinary light.

X 80.



Fig. 32: Development of garnet in the actinolite corona. Olivine gabbro, Morud.
Ordinary light. X 40



Fig. 33: Conjugate corona between clinopyroxene and plagioclase. Inner rim of orthopyroxene against clinopyroxene, outer rim of actinolite against plagioclase. Olivine gabbro, Morud.
Ordinary light. X 40



Figure 34: Development of brown hornblende between magnetite and plagioclase. Olivine gabbro, Morud.

Ordinary light.

X 40.

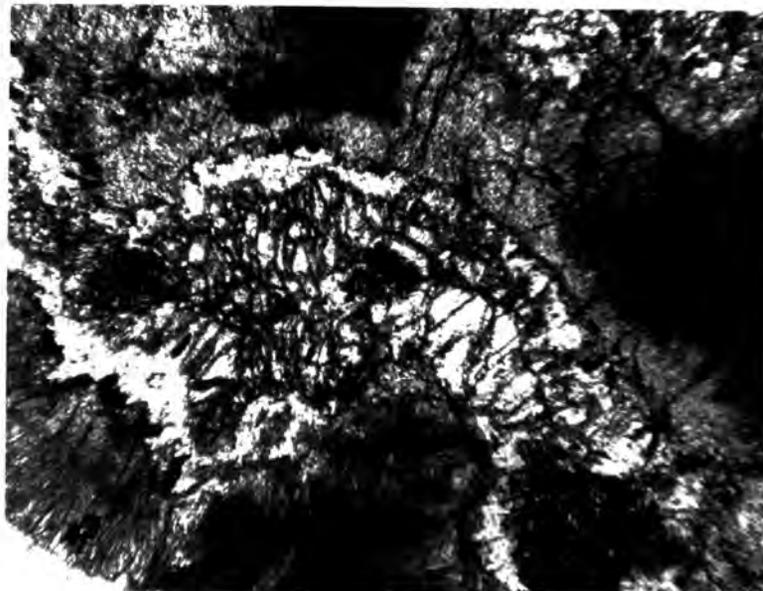


Figure 35: Orthopyroxene pseudomorphing olivine. Olivine gabbro, Morud.

Ordinary light.

X 40.

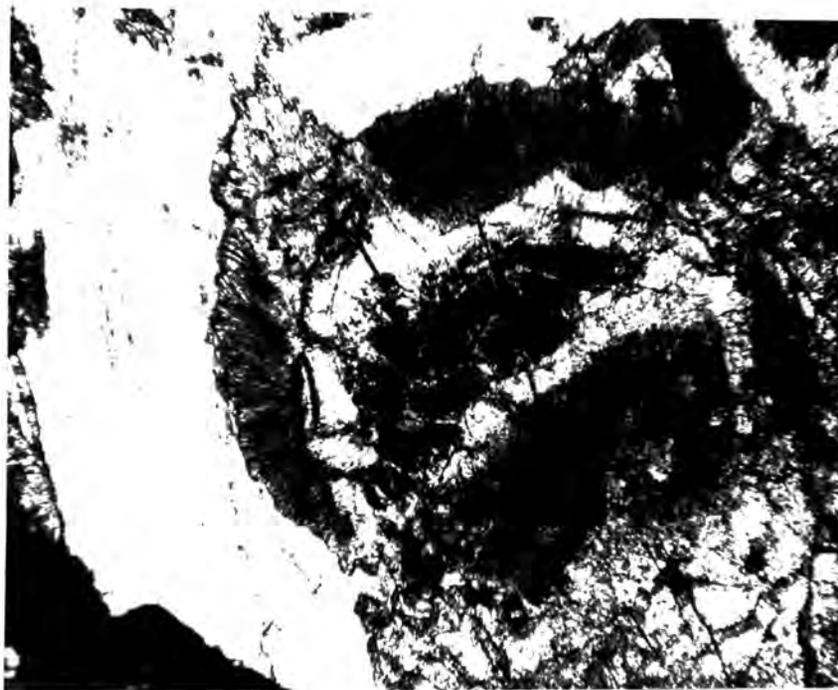


Figure 36: Serpentine and magnetite core to recrystallised orthopyroxene pseudomorphing olivine. Olivine gabbro, Morud.

Ordinary light.

X 40.



Figure 37: Recrystallisation of outer part of actinolite corona to a coarser grained aggregate of a deeper green colour. Olivine gabbro, Morud.

Crossed nicols.

X 40.

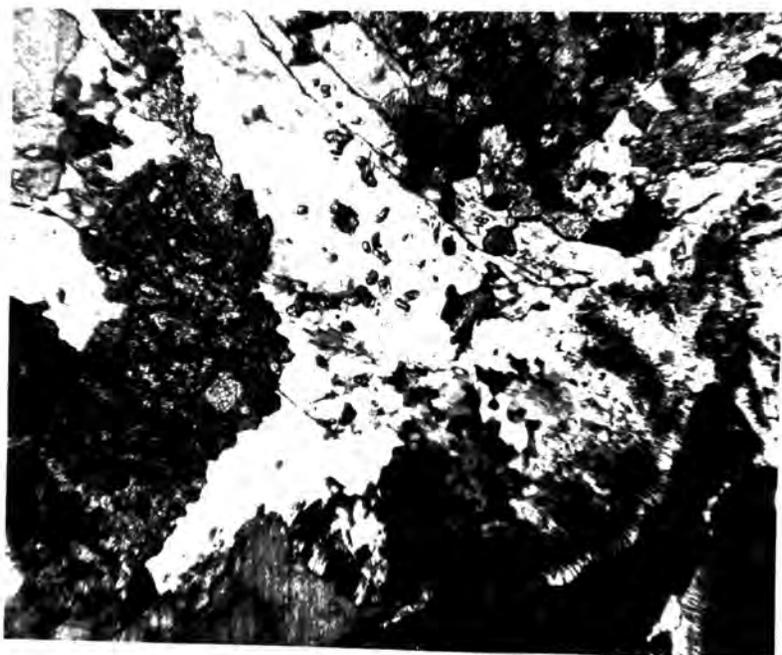


Figure 38: Development of quartz at junction of ferromagnesian and plagioclase rich areas.

Crossed nicols.

X 40.

cms.




Figure 39: Garnet flattened in the plane of the Garnet amphibolite, foliation. South east of Djupedal trail.

grade out into andesine ($An_{34}-An_{48}$) - determined on the universal stage by the maximum extinction method. Some of the laths enclose smaller crystals of early formed plagioclase. The third of the main minerals is the ophitic clinopyroxene. This is a diopsidic-augite with $2V = ca. 60^\circ$; $c\angle Z = 44^\circ$. It is interstitial to both olivine and plagioclase, as large ophitic individuals reaching 1 cm. or more in size. It contains exsolved laths of iron ore parallel to (001). It forms about 20% of the rock.

The three main minerals form about 75-85% of the rock. The other minerals are present as alteration coronas (kelyphitic rims), at the junction of the ferromagnesian minerals and the plagioclase.

Between olivine and mesostatic clinopyroxene small rims of orthopyroxene have formed (Figure 29). The rims are composed of small prismatic crystals oriented parallel to the cleavage (110) directions such that they taper alternately at the distal and proximal ends. Some grains appear to have inclined extinction, but on careful examination it is seen that the cleavage is also inclined to the long axis of the crystal in these grains. A few grains were found to be suitable for the determination of $2V_x$ by the extinction method on the universal stage, this gave a result of 57° . From the diagram in Deer, Howie and Zussman (1962, Vol. 2, Fig. 10) this corresponds to a hypersthene (Fs_{67}). The mineral is not noticeably pleochroic.

Between olivine and plagioclase conjugate corona are produced (Figure 30). The inner one consists of orthopyroxene prisms aligned with their c-axes arranged normal to the olivine margin. The outer corona is much more extensively developed and consists of very fibrous actinolitic hornblende. The fibres are arranged in bunches and fans. It is pleochroic in palish blue-green with extinction $c\angle Z: 16^\circ$. Intergrown with the actinolite fibres are very small crystals of a green mineral with high relief. It is isotropic, and occurs in minute octahedral crystals. This was identified as spinel - probably pleonaste. The pleonaste is particularly abundant along the margin with the inner corona, it is developed in chains parallel to the long axes of the adjacent actinolite (Figure 31). In rare cases larger crystals, colourless, of high relief are associated with the actinolite corona (Figure 32), this is probably garnet.

Where the ophitic clinopyroxene is in contact with olivine the single corona of orthopyroxene is developed, but where it is in contact with plagioclase conjugate corona of the type described above are found (Figure 33). Between iron ore and plagioclase a corona of hornblende occurs which differs from the outer corona of the olivine and clinopyroxene in that the crystals are not as fibrous and are pleochroic in pale brown, $c\angle Z: 13^\circ$ (Figure 34).

A further stage in the alteration of these rocks is the total replacement of the olivine and clinopyroxene by orthopyroxene, which has recrystallised to a granular aggregate (Figure 35). In some cases the innermost parts of these aggregates are very much clouded by fine magnetite dust, so that they appear nearly opaque, in others there is a central core of a serpentine mineral (Figure 36). These aggregates are surrounded by the normal actinolite-pleonaste corona, but the outer parts of this corona are seen to have recrystallised to a coarser grained aggregate of hornblende of a deeper green colour (Figure 37). At this stage quartz becomes noticeable at the edge of ferromagnesian areas (Figure 38).

Thus in these rocks can be seen the breakdown of a typical olivine-gabbro to a mineral assemblage consisting of blue-green hornblende, plagioclase and minor quartz and garnet with remnant orthopyroxene. Brøgger (1934) and Gjelsvik (1952) have described in detail the further changes wrought in these rocks to produce true amphibolites. An intermediate stage where an amphibolite mineralogy is produced but relict igneous textures are preserved is represented by rock 173 described on page 105.

Amphibolite and biotite-schist

Field relations

The amphibolites at Modum have two major modes of occurrence, either as envelopes around the olivine-gabbro masses or as bands intercalated in the metasedimentary sequence. It is sometimes possible to trace a junction between the two types, when for example, the amphibolitized margin of a gabbro is drawn out into a thin layer intercalated in the surrounding gneisses (Figure 25). The two types of amphibolite are indistinguishable in field appearance.

The thicker layers of amphibolite form bold features. A garnetiferous amphibolite several hundred metres wide and more than a kilometre long forms a long ridge to the east of the Skuterud Mines, through which the Klara and Ludwig Eugen Crosscuts have been

driven to intersect the ore zone at depth (Figure 110).

The amphibolites are rather varied in their field appearance as a result of variation in both texture and mineralogy. Massive, unfoliated and unlineated amphibolite is rare, but it is found in the centre of the larger masses and consists of hornblende, plagioclase and garnet with very little biotite. This type commonly contains glomeroblastic plagioclase aggregates, 1-2 cms. long and oriented parallel to the strike of the enclosing rocks. These frequently carry garnet up to 3 cms. in diameter.

Striped amphibolite is the most common type encountered. The individual layers consisting of the alternation of dark, 1 mm. to several cms. thick, hornblende-plagioclase bands, with, thinner, off-white plagioclase rich quartz bands poor in hornblende. Individual bands can be traced for several metres along the strike. The hornblende prisms are dimensionally oriented in the plane of the foliation giving the rocks a lineation. Stereographic projection of the hornblende lineation data suggest that this is a true 'b'-lineation.

Very fine grained amphibolites occur locally containing garnets less than 3 mm. in diameter. These have no relation with the contact against the metasediments. The normal grain sizes observed lie in the ranges: hornblende prisms 5 - 15 mm; quartz and plagioclase grains 1 - 18 mm; garnet porphyroblasts 5 - 40 mm. and biotite flakes 5 - 15 mm. Pegmatitic variants occur conformable to the foliation and vary from a few millimetres to several metres in width. These have been affected by the folding as can be clearly seen on weathered surfaces. The usual mineralogy of the pegmatites

is plagioclase and quartz with clear, red, anhedral garnets, sieved with inclusions. Basic pegmatite layers have been observed on Høgvann, where hornblende crystals measuring 8 x 3 cms. occur in layers several metres long.

Garnetiferous amphibolite is very common. The garnets are usually restricted to individual layers parallel to the foliation, though exceptions to this rule have been noted (Figure 40). Some of the garnets are flattened in the plane of the foliation (Figure 39), but they are usually seen growing across the foliation and blastically enclosing the other minerals (Figure 53b), in such a way that the foliation can be traced without interruption through the garnet crystals. Garnet-amphibolite is patchily developed, some regions of a particular amphibolite carrying large quantities of garnet while others are completely devoid of the mineral.

The different amphibolite varieties are not sharply separable and often show mutual transitions. Transitions are never observed with the surrounding metasediments, the contacts are always knife sharp.

The amphibolites have reacted to deformation, at least part of which is post-dated metamorphism, by folding and shearing. Folding is usually observed in the schistose, biotite-rich bands (Figure 55a). In the more massive types shear zones are commonly developed and are occupied by dilation pegmatites (Figure 53a). Post-tectonic shear planes are occupied by chlorite-schist (p. 110).



Figure 40: Field relations of amphibolites.

Top right: Garnet-plagioclase felspar band forming an angle of 28° with the penetrative foliation imparted by the biotite in the surrounding biotite amphibolite. The garnetiferous band may represent an early formed pegmatite which has been affected by later shearing.

Bottom right: Sketch of hand specimen collected from the amphibolite-quartzite junction exposed at Bennecke Adit portal. The amphibolite appears to transgress the mineralogical colour banding observed in the quartzite.

Left: Amphibolite-quartzite junction sketched in the field from the southern wall of the Forhaabning Crosscut.

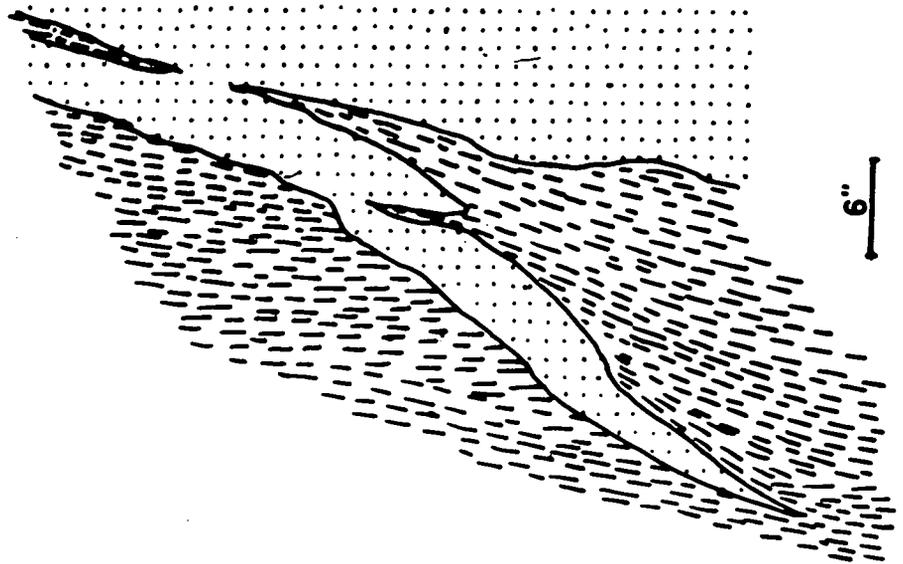
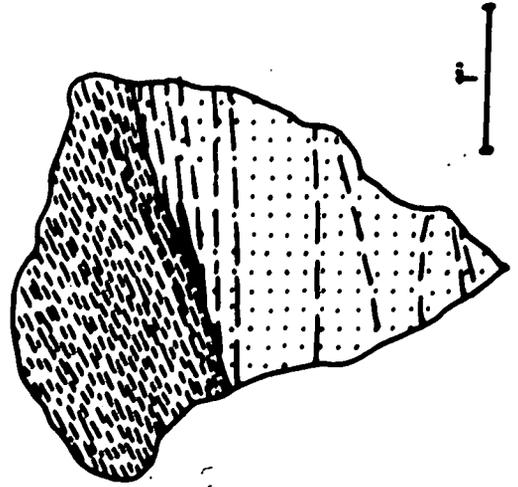
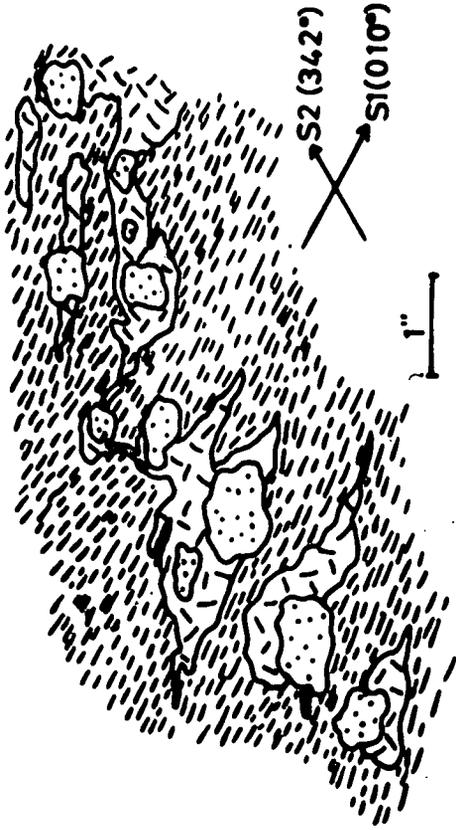


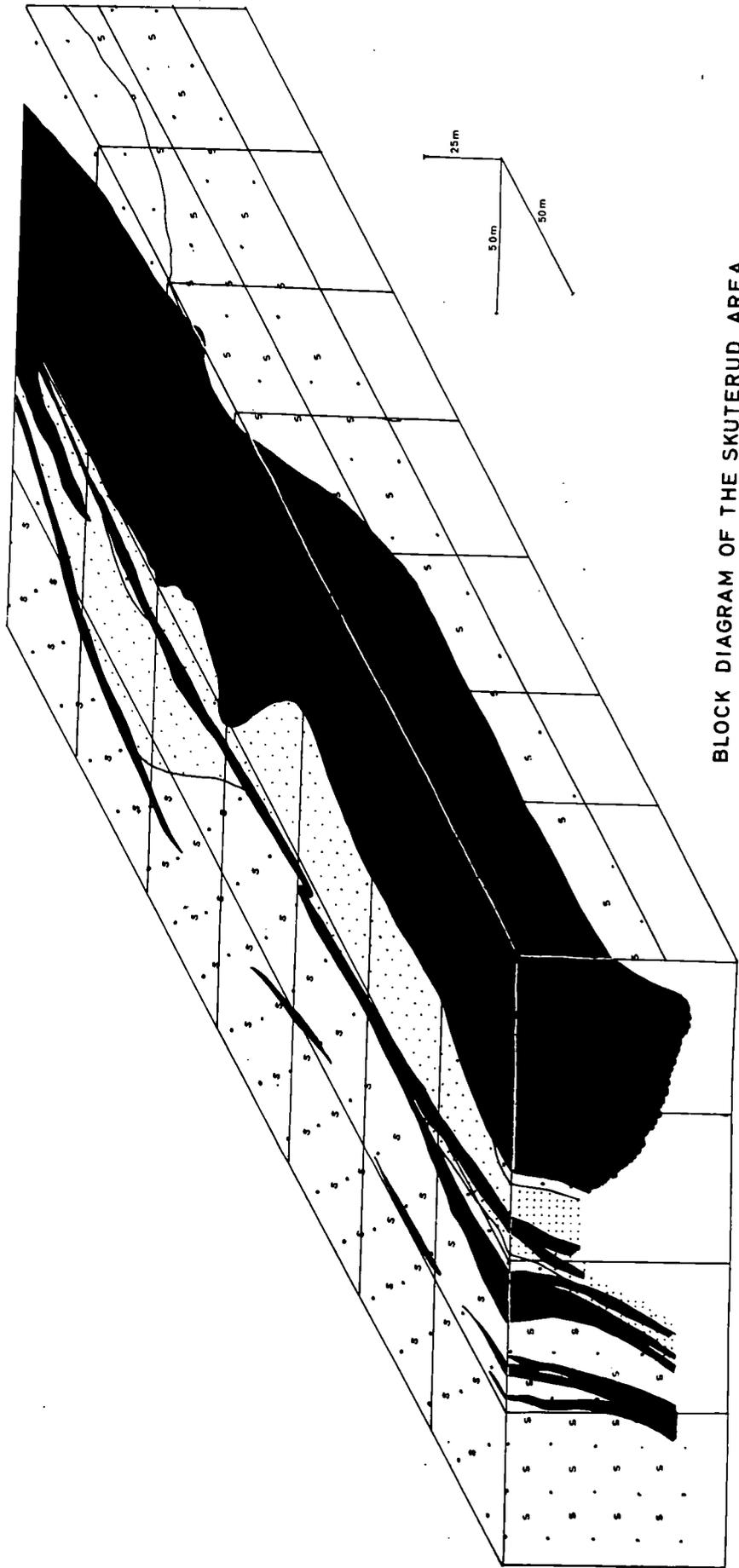
Fig. 40

Biotite is a common mineral in the amphibolites, but in places it becomes dominant and true biotite-schists are developed. These are commonest as marginal facies to the thicker amphibolite layers. In the ore zone at Skuterud there are several small amphibolites which though they rarely reach 30 metres in thickness, are very persistent along the strike. In places these are totally converted to biotite-schist. In view of this close association it is convenient to include discussion of these schists in this section.

The origin of the amphibolites is of fundamental importance to this study, but the distinction between ortho- and para-amphibolites is difficult on textural and mineralogical grounds and considerable weight must be attached to the field relations to other rock types.

The large garnet-amphibolite to the east of the Skuterud Mines is generally parallel to the strike of the enclosing rocks, though detailed mapping suggests that it may be slightly transgressive (Figure 110). The thick quartzite in the vicinity of the South Mines immediately to the west, thins to the north, until amphibolite is in contact with siliceous granulite when the Central Mines are reached. Still further north, however, a similar quartzite appears immediately to the east of the amphibolite and appears to thicken northwards.

The thin amphibolites of the ore zone at Skuterud have been interpreted by Rosenqvist (1948) as unconnected with this garnet amphibolite. The present study indicates rather that they are connected to the main amphibolite just east of the North Mine, and to each other at various places along the strike (Figure 41).



BLOCK DIAGRAM OF THE SKUTERUD AREA.

Fig. 41

The contacts of these thin amphibolites are clearly exposed in some of the workings. At the southern end of the South Mine slight discordances between the foliation of the amphibolite and siliceous granulite and the contacts of the rocks were recorded. At Bennecke Adit portal the foliation in the amphibolite is clearly seen to transect the mineralogical banding of the adjacent quartzite (Figure 40). A few metres from the portal of Forhaabning Adit amphibolite and quartzite are seen to interdigitate (Figure 40). Intrusive relationships might explain the facts.

Petrography

In view of the mineralogical and textural variation exhibited, specimens representative of the various rock types will be described.

Amphibolite marginal to olivine-gabbro massif (173) This specimen was collected from a road cut in the margin of the basic mass at Morud, 40 metres south of Morud railway station (Figure 25). In hand specimen it is a melanocratic rock with some lighter felspathic bands and contains garnet porphyroblasts.

The hornblende is present as an aggregate of polygonal grains, 1-2 mm. in diameter. It is brownish green in colour with pleochroic sheme: X: pale yellow-green, Y: yellowish-green, Z: greenish-brown. Extinction \sphericalangle : 21°.

Plagioclase (An₄₆) is present as interlocking polygonal grains, 1-3 mm. in diameter. The plagioclase and hornblende aggregates form discrete areas in the section (Figure 42). The plagioclase is partly sericitised and is twinned with combinations of albite, pericline and Carlsbad laws.

Some quartz is present in anhedral grains, 1-3 mm. in diameter, it is restricted to the plagioclase rich areas of the section. Minor biotite is present, strongly pleochroic in shades of reddish brown.

Garnet is restricted to the leucocratic areas of the rock. Individual crystals reach 5-6 mm. in diameter and are full of

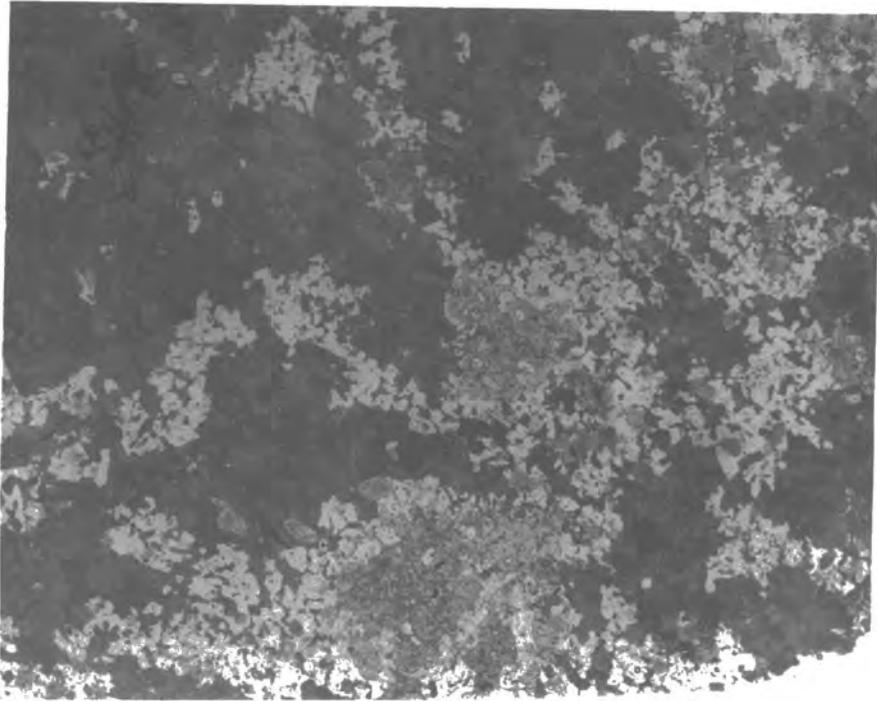


Figure 42: Photograph of whole thin section of rock 173. The plagioclase and hornblende are restricted to discrete areas in the section. These possibly represent a relict igneous texture.

inclusions, some of which are in optical continuity with grains outside the garnet crystals.

Small amounts of opaque mineral are present as skeletal crystals filling the interstices between hornblende.

Striped amphibolite (21) The specimen comes from a thin amphibolite band just east of the main open-cast workings of the Central Mine, Skuterud. The section is cut in one of the leucocratic quartz-plagioclase layers.

Hornblende forms about 30% of the rock. It is present in ragged grains and disconnected fragments in optical continuity. Many of the grains contain small, rounded quartz inclusions. It is pleochroic with X: pale yellow-green, Y: green, Z: deep blue-green. The extinction angle $c \wedge Z$ is 18° .

Plagioclase (An₃₈) is the most abundant mineral forming about 50-60% of the rock. It is in the form of large ragged grains up to 8 mm. in diameter, enclosing small quartz and hornblende grains. Combinations of albite and pericline twins are common. Particularly noticeable are deformation zones in some plagioclase crystals, these are marked by slight strain extinction where deformation is strongest. Individual twin lamellae originate where the crystal is bent and die out away from these zones.

Quartz forms about 15% of the rock and is present as grains 1-3 mm. in diameter. It is clear and unstrained. Small quartz grains are also found in the poikiloblastic plagioclase.

Biotite is a minor constituent, it is pleochroic in reddish brown, individual flakes cut across hornblende crystals and range in size up to 4 mm. Both biotite and hornblende appear to be strongly attacked by the quartz and felspar.

The section also contains minor ore minerals. No garnet is present.

Scapolite-amphibolite (188, 92) Scapolite is found in some of the amphibolites. In rock 92, which was collected from a thin amphibolite band on the eastern slopes of Skuterudåsen, it is present in a plagioclase rich band. In rock 188, which shows the contact of amphibolite and siliceous granulites from the south end of the western open-cast at the South Mine, Skuterud, the scapolite is concentrated near the contact and is associated with both the quartzite and the amphibolite. This rock also contains small, anhedral sphene crystals. The scapolite is clear and colourless.

It occurs in stumpy crystals of high relief, is uniaxial negative and has a good cleavage.

Garnet-amphibolite (109a, 109b, 140) These three specimens were all collected from the thick garnet-amphibolite to the east of the Skuterud Mines. Two come from the portal of the Klara Adit (109 a,b.), and the other from just east of the North Mine (140). They are very similar in thin section.

Hornblende forms between 35% and 45% of the rock. It is pleochroic with X: pale yellow-green, Y: green, Z: brownish-green. Extinction, c Z: 20° . It is present as ragged grains very much corroded by quartz and plagioclase.

Plagioclase (An₄₂) is abundant, usually forming about 50% of the rock. It is in large grains, up to 6-7 mm. in size, containing inclusions of quartz, hornblende and biotite. It is generally fresh, sericitisation being restricted to patches in some of the larger grains. Twinning on albite and pericline laws is well developed. Some crystals appear zoned having combined albite-pericline twins in the darker core, with only the albite twins reaching to the paler, outer zone.

Quartz is present in minor amounts (ca.5%) and is associated with the plagioclase. Biotite is more abundant than in the sections previously described, forming about 10-15% of the rock. It is pleochroic in deep red-brown, frequently altering to chlorite. It is sometimes twinned on the mica law. It is observed both cutting obliquely across hornblende crystals and growing in optical continuity such that the c-axis of hornblende is parallel to the (001) cleavage of the biotite. Small fragments of hornblende are seen enclosed in biotite flakes having the same optic orientation as the adjacent hornblende crystal. Biotite crystals are seen intersecting each other, they reach 4 mm. in length.

Garnet is beautifully developed in large porphyroblasts, frequently more than 1 cm. in diameter (Figure 43). It is very pale pink in colour, with high relief and many irregular fractures. The garnet is full of inclusions of hornblende, quartz, plagioclase and biotite. The crystals have an extremely irregular outline. They appear to have grown by the production of small (0.05 mm.) subhedral crystals at the margin of the main mass; these linked up and surrounded the adjacent minerals which become incorporated into the garnet. The only constituents which appear to have withstood this process are the opaque minerals which are still present as inclusions in the core of the garnets. In rocks containing large porphyroblasts of garnet small euhedral individuals which may be the nuclei of new crystals can be seen. In slide 140 a garnet is seen to abut against a biotite flake with a planar boundary, whilst at the extremities of the biotite flake the garnet has continued its outward growth (Figure 44).

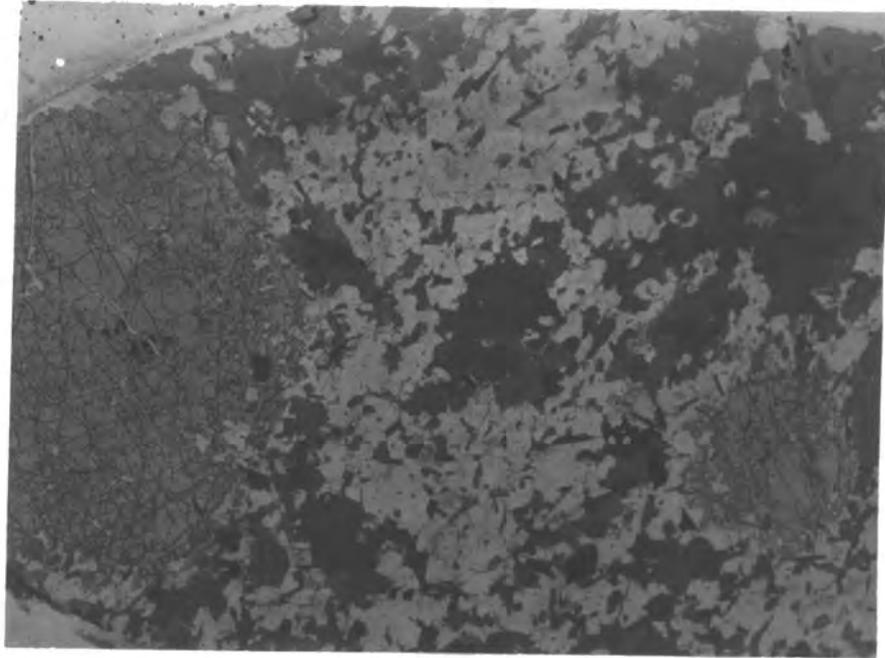


Figure 43: Photograph of whole thin section of rock 109a, to show the large porphyroblastic garnets.

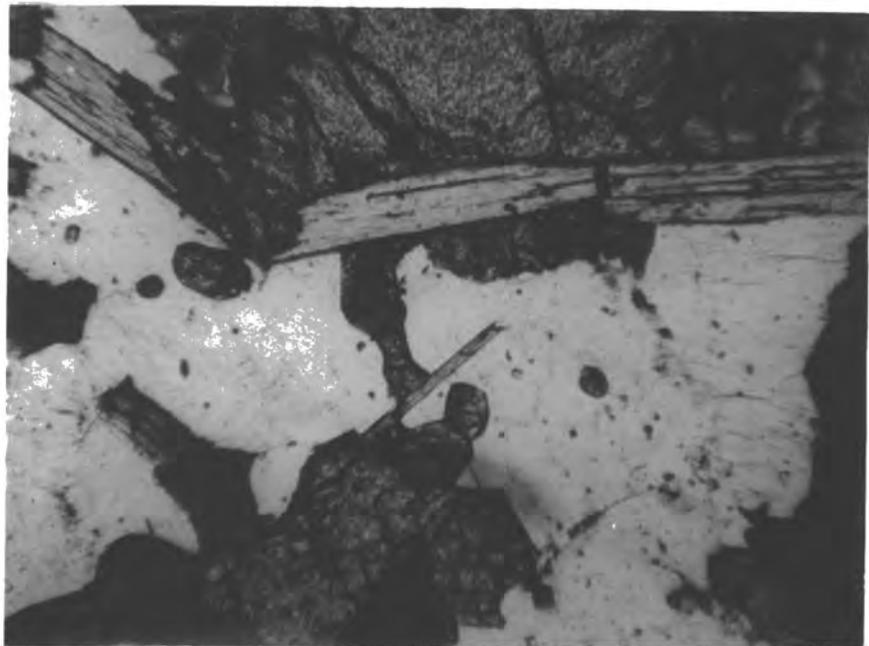


Figure 44: Porphyroblastic garnet in contact with a flake of biotite. It appears that the biotite has impeded the outward growth of the garnet which only continues at the termination of the mica.

Crossed nicols.

X 40.

Biotite-schist (95, 105) These specimens were collected from the thin 'amphibolite' horizons encountered underground in the Klara Level at Skuterud. In hand-specimen they are well-foliated biotite schists.

In thin section the major changes in appearance are due to the development of biotite at the expense of hornblende. The hornblende consists of disconnected fragments and skeletal crystals very strongly corroded by plagioclase and quartz. The biotite is seen developing within the larger hornblende crystals, principally near the margin, with the (001) cleavage plane of the biotite parallel to the c-axis of the hornblende. Biotite forms about 60% of the rock, it is fresh and strongly pleochroic from reddish purple to pale brown. Flakes attain a length of 8-9 mm. In biotite rich areas the flakes are often intergrown such that the (001) planes meet at an angle of 120° . Biotite forms about 60% of the rock.

The plagioclase has a smaller grain size than in the unaltered amphibolites, it now rarely exceeds 4 mm. It is strongly altered to sericite. Where fresh it still exhibits combinations of albite and pericline twinning.

Quartz is more abundant than in the true amphibolites. It is present in small amoeboid grains, up to 0.5 mm. in diameter, associated with plagioclase in lenses and layers parallel to the schistosity imparted by the biotite.

Quartz and plagioclase strongly corrode the biotite and hornblende. Ore minerals are not abundant but appear, texturally, to be the latest mineral.

Chlorite schist (94) This rock was collected from a shear zone near the eastern margin of the thick garnet-amphibolite, 20 metres south-west of Klara Adit portal. The surrounding rock was garnetiferous biotite schist.

Chlorite occupies about 60% of the slide. The individual flakes are twisted and bent and seem to have been affected by micro-folding. It has fair relief and is pleochroic from pale yellow to apple green. It has a perfect basal cleavage and is sometimes twinned on the mica law. Extinction is slightly inclined and anomalous purple interference colours are present. No flakes suitable for obtaining an interference figure were present. On the basis of the inclined extinction the mineral is tentatively assigned to the clinocllore group.

Quartz forms about 15% of the rock, it is present in small patches and lenses. It is full of dusty inclusions and shows strain extinction. The grain boundaries are very sutured, there is quite a variation in grain size, with the majority being about 0.5 mm. but some attaining a size of 4 mm.

The plagioclase is completely altered to fine white mica, no trace of twinning was observed. It is in small anhedral grains forming about 20% of the rock. It appears to be corroding the chlorite.

Skeletal crystals of a very deep brown mineral are associated with the chlorite. It possesses a good cleavage. Birefringence is masked by the strong internal reflections. This is possibly rutile.

Chemistry

Considerable interest attaches to the large garnet-amphibolite situated to the east of the Skuterud Mines (Figure 110). Two good sections through this band are provided by the Klara and Ludwig Eugen Crosscuts. Specimens were collected from these crosscuts for chemical analysis. Location of the samples and a graphical presentation of the data is shown in Figure 45. The results are tabulated below:

Table 12

X-ray fluorescence spectrographic data
for Skuterud garnet-amphibolite

Specimen No.	Weight per Cent						p.p.m.	
	MnO	TiO ₂	CaO	K ₂ O	Fe ₂ O ₃	Cr	Co	Ni
397	0.06	2.66	7.30	1.18	11.10	120	58	50
444	0.20	3.23	8.60	0.83	14.50	200	67	90
443	0.11	2.57	6.56	2.66	14.60	200	60	80
442	0.22	1.23	8.40	1.20	13.60	300	61	115
396	0.20	3.65	8.55	0.90	14.60	90	55	82
395	0.18	2.20	8.86	1.05	13.20	240	48	83

Analyst: T.K. Ball

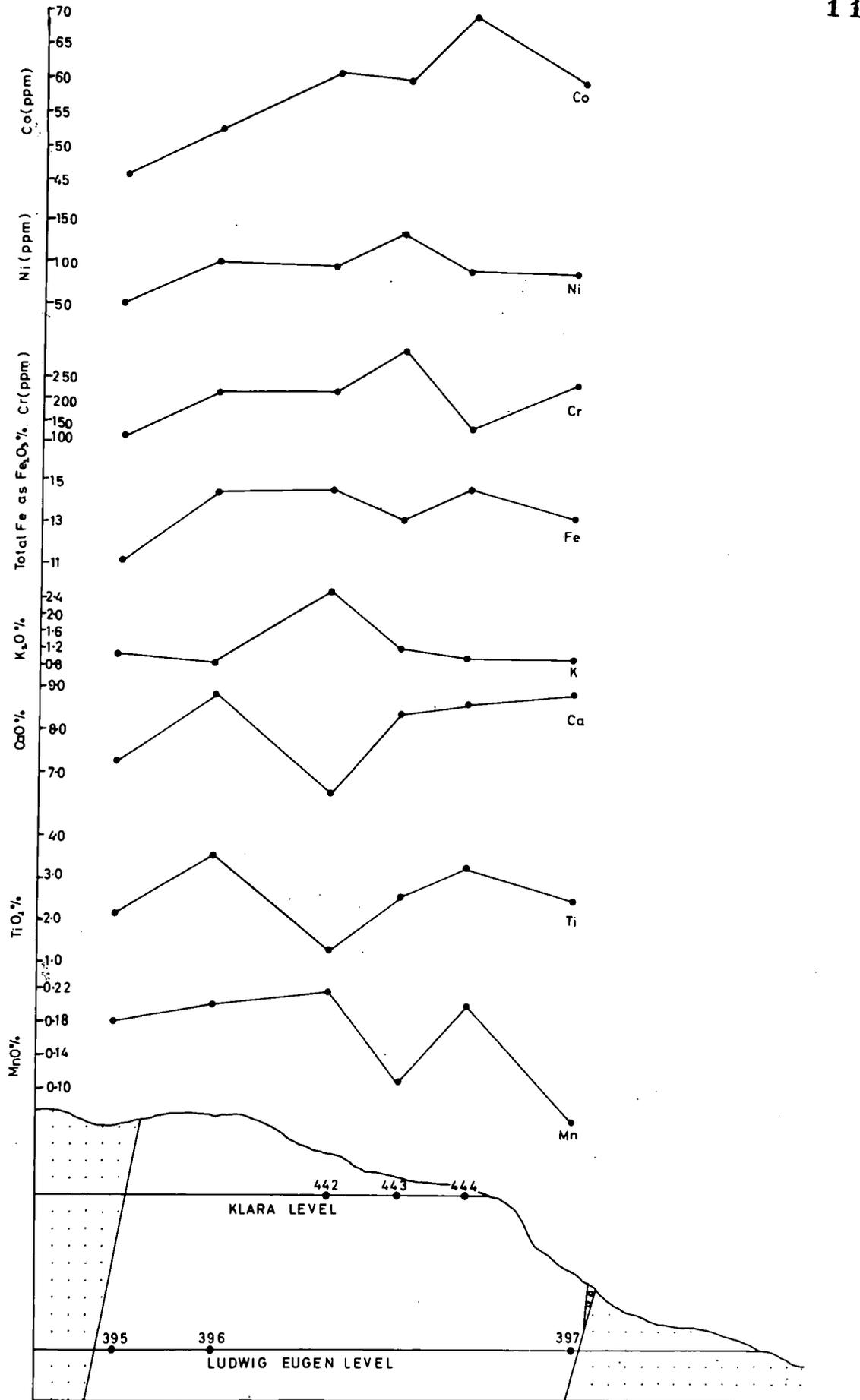


Fig. 45

Opaque mineral constituents of the amphibolites

Several polished specimens of the amphibolites were studied. In all cases the ore minerals are present in minor amounts (up to 4 vol.%) and appear to be corroded by the gangue minerals.

Ilmenite is by far the most common opaque mineral present. It forms skeletal crystals (Figure 46) between the gangue minerals, when associated with hornblende it tends to form large (0.5-1 mm.) discrete crystals, while in plagioclase rich areas it is in smaller (0.05-0.5 mm.) disseminated crystals. It has a sieve texture with many inclusions of gangue mineral (Figure 47), gangue minerals are also seen attacking the ilmenite along the (0001) cleavage direction. In biotite rich amphibolites the ilmenite is altering to rutile (Figure 48), particularly along the (0001) direction. The properties of the ilmenite are as follows:

Colour - In air: pale pinkish-brown, in oil: brown, against gangue minerals in both cases. Reflection pleochroism: in air, very slight; in oil, striking white-brown.

Reflectivity - (Compared with Elba pyrite = 54.5%) 20.1% Bireflectance not measurable.

Anisotropism - (No internal reflections). Deep purple-brown to pale blue-grey. Extinction parallel to twin composition plane (0001). Pale grey in 45° position.

Vickers hardness - (G.K.N. Hardness tester, 100 gm. load.) 546

Isotropic magnetite is present in very subordinate amounts, associated with the ilmenite. Sulphides are not common, they are found associated with grain boundaries and cracks. They are

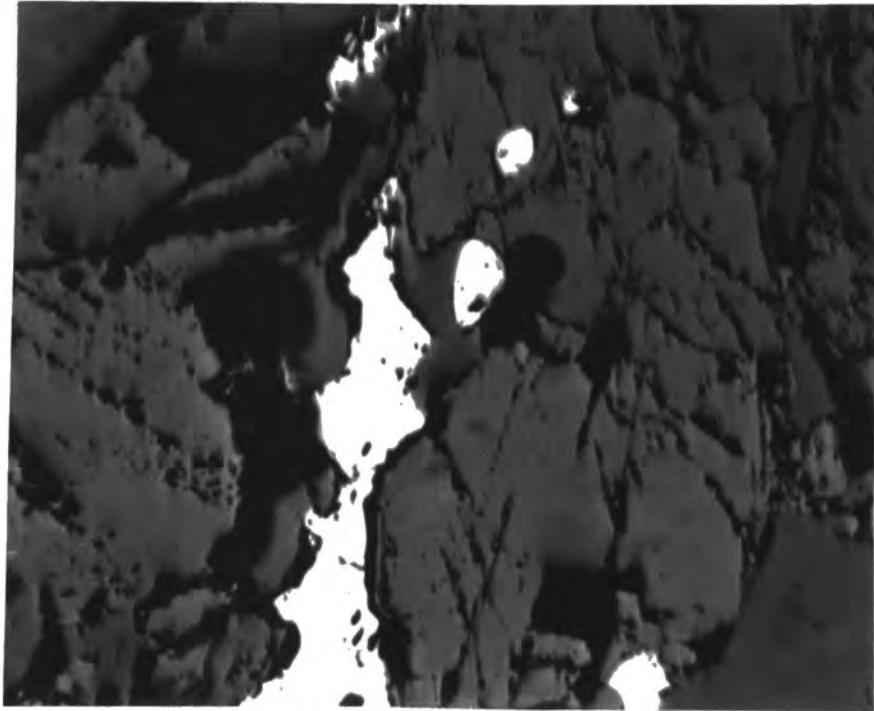


Figure 46: Skeletal ilmenite crystal. Amphibolite, Klara Adit, Skuterud.
Ordinary reflected light. X 50.

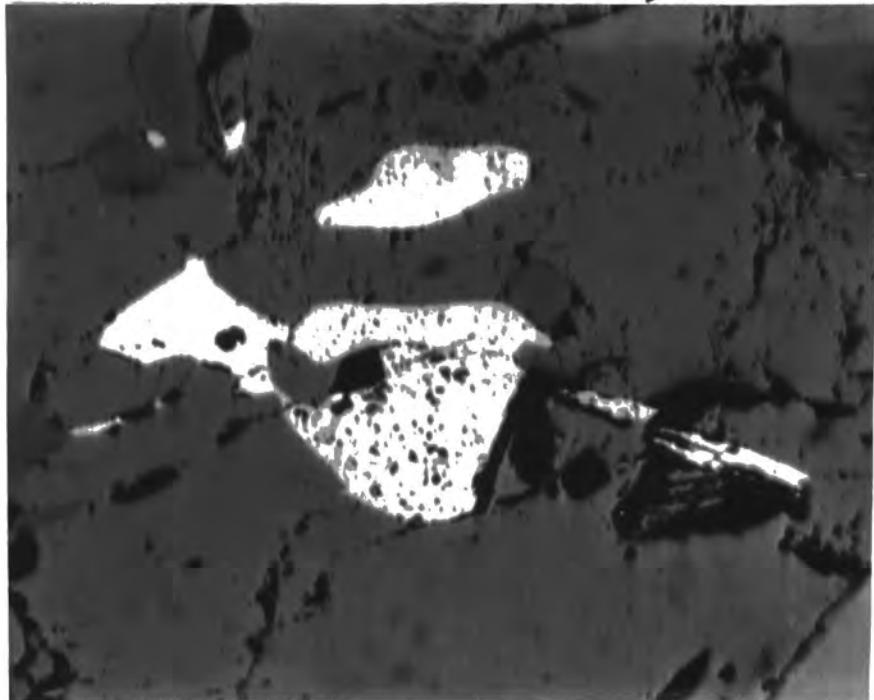


Figure 47: Ilmenite containing inclusions of gangue silicate. Biotite-
amphibolite, Ludwig Eugen Adit, Skuterud.
Ordinary reflected light. X 50.

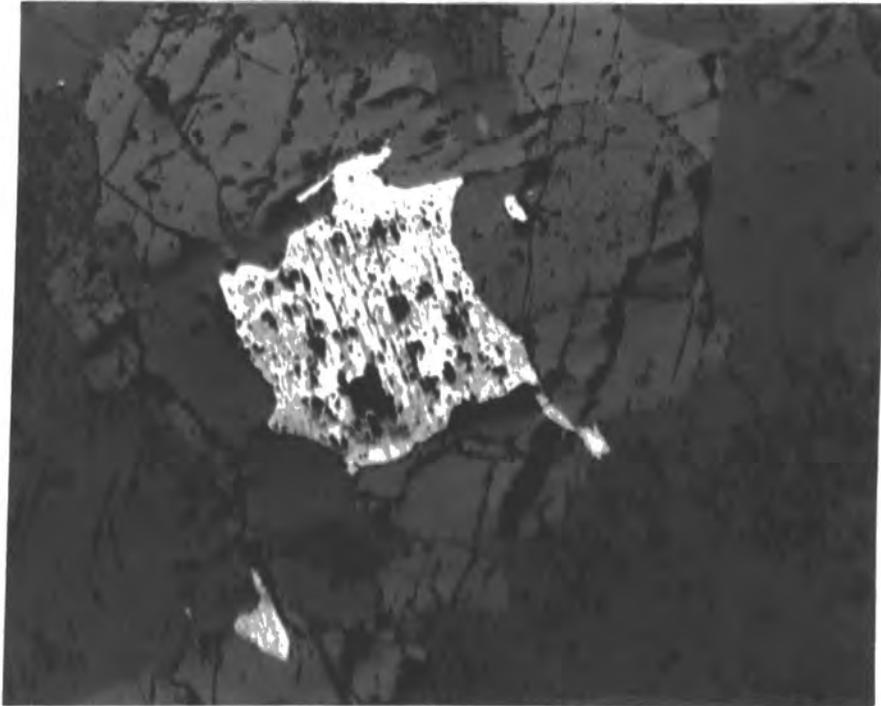


Figure 48: Alteration of ilmenite to rutile along the (0001) cleavage.
Biotite-schist, Forhaabning Adit, Skuterud.
Ordinary reflected light. X 50.

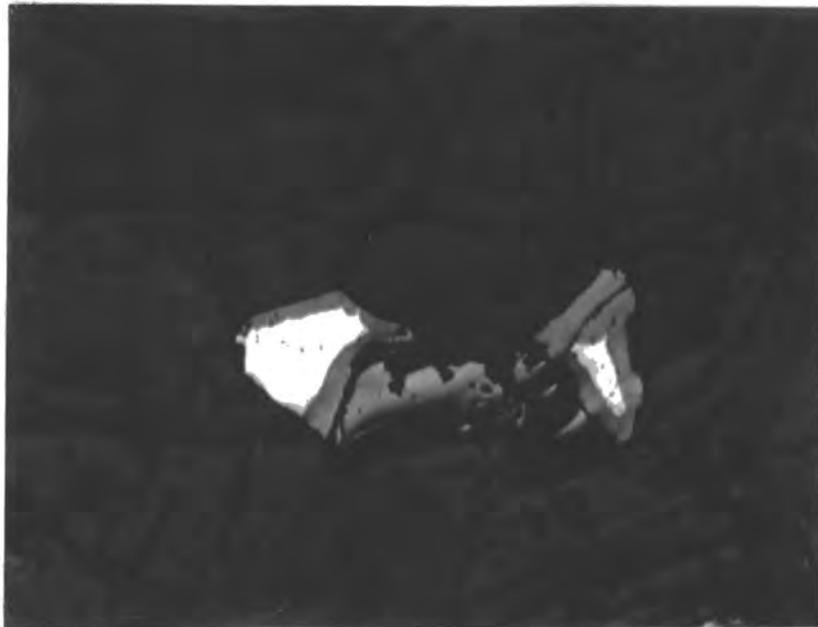


Figure 49: Alteration of pyrite to goethite. Weathered amphibolite,
North Mine, Skuterud.
Ordinary reflected light. X 50.

corroded by the gangue, and have mutual boundaries with the ilmenite. Chalcopyrite is subordinate to pyrite, it is strongly anisotropic and is observed fringing and included in pyrite crystals. The pyrite forms larger grains (up to 1 mm.) showing a weak anomalous anisotropism.

In a rusty weathering amphibolite (92) the pyrite has altered to goethite. The goethite pseudomorphs the original pyrite and is encroaching upon the unaltered core of the crystals, (Figure 49). The goethite has low reflectivity (15.8%) with weak reflection pleochroism, accentuated under oil immersion, from pale grey to yellowish-gray. The goethite rim is formed of feathery crystals intergrown with their long axes perpendicular to the original pyrite crystal boundary. With crossed nicols the mineral is strongly anisotropic: from blue grey to white, the colours being masked by the strong internal reflections.

Petrogenesis

A survey of the relevant literature shows that much attention has recently been directed to the distinction of ortho- and para-amphibolites. (Engel and Engel, 1951; Lapadu-Nagues, 1953; Poldervaart, 1953; Eckelmann and Poldervaart, 1957; Wilcox and Poldervaart, 1958; Evans and Leake, 1960; Walker et al. 1960; Heier, 1962). The majority of these authors have rejected the validity of assumptions based on field, textural or mineralogical evidence, and rely, rather, on a chemical approach.

When all the available evidence concerning the Modum amphibolites is considered, it emerges that there is sufficient to indicate that some, at least, of the amphibolites are derived from basic igneous intrusive rocks.

The presence of amphibolite envelopes around the olivine-gabbro bodies, indicates that these are a metamorphic derivative of the gabbro. Barth (1947) has described a similar body from southern Norway which has become almost totally amphibolitized with only small remnants of the original norite left.

These ortho-amphibolites are indistinguishable from the other amphibolites in the area both in field appearance and in thin-section. Thus the banding of the striped amphibolites does not necessarily imply a sedimentary origin, an assumption made by many workers in the field (e.g. Poldervaart, 1953 p. 268; Wilcox and Poldervaart, 1958, p. 363). Mineral foliation can either be the result of mimetic crystallisation along privileged planes or the result of metamorphic segregation. As plutonic and hypabyssal rocks would not be expected to possess such privileged planes the latter hypothesis is the more likely for the ortho-amphibolites. Minor folding is not well developed in the striped amphibolites, indicating that deformation may have taken place by slip along the schistosity - a condition favourable for metamorphic segregation. The transition from foliated to massive amphibolite may represent a change from suitable to unsuitable conditions for segregation (Evans and Leake, 1960, p. 352). That segregation has indeed taken place is shown by the pegmatitic layers and lenses, quartzo-felspathic pegmatite lenses sometimes show a basic selvage of hornblende.

Having concluded that the textural and mineralogical character of the amphibolites intercalated in the metasediments could be derived from hypabyssal basic rocks as well as from impure calcareous sediments, the field relations of these rocks becomes of paramount importance in interpreting their genesis.

The amphibolites of the Skuterud district have been taken as representative of this type. Detailed mapping of the mining area (Figures 109,110) suggests that the thin individual bands are all interconnected and join the major garnet-amphibolite in the region of the North Mine. This produces a complex outcrop pattern difficult to explain on the basis of tectonics alone. The relation of the various amphibolites is better brought out in Figure 41 which is a block diagram of the area, constructed using data from the underground workings.

The transgressive nature of the major garnet-amphibolite and the contacts of the thin amphibolites with the metasediments have already been commented upon (p.103). These are indicative of an intrusive origin.

Chemical data is abundant in the literature and conflicting conclusions have been drawn from it. However, it is generally agreed that high contents of Ti, Cr, Co and Ni are indicative of ortho-amphibolites (e.g. Engel and Engel (op.cit.)). The Modum garnet-amphibolite is rich in all these elements (Table 12). The variation of cobalt from east to west through the amphibolite is of interest (Figure 45). Ball has studied dolerite sills metamorphosed

to amphibolite facies in Scotland and found a similar variation for cobalt. The shape of the curve indicates that the 'top' of the 'sill' is to the west for the Skuterud amphibolite (Ball, personal communication). This possibility is further strengthened by the presence of the olivine-gabbro lens at the eastern contact of the amphibolite (p.90), i.e. the 'base' of the 'sill'.

Thus there is circumstantial evidence, at least, for the following sequence of events in the Skuterud district:

- (1) Deposition of sedimentary sequence
- (2) Intrusion of thick olivine-dolerite sill, with several connected, thinner sills, emplaced contemporaneously in the overlying rocks. Segregation of olivine rich material to base of sill.
- (3) Isoclinal folding and metamorphism to produce the present mineralogy and attitude of the sill complex.

These conclusions can obviously not be applied to amphibolites elsewhere in the Modum region. Indeed in the beautifully exposed rocks of the Kongsberg - Bamble Formation of the Skaggerak coast several types and generation of amphibolite have been recognised (Wegmann and Schaer, 1962). It is quite possible that similar relations exist at Modum, but limitation of outcrop does not permit their elucidation.

The relations of the various rock types considered in the section can be obtained by attempting a synthesis based on the petrographic data.

The suite of basic rocks was emplaced before or during the major orogenic episode affecting the Modum Formation. It consisted of a series of dolerite and olivine-dolerite dykes and sills with associated gabbro and olivine-gabbro stocks. The coarser grained and olivine rich phases of these intrusions have, to a large extent, resisted the metamorphic transformations. The plagioclase is still present in large, zoned laths. Ophitic clinopyroxene forms the groundmass and olivine crystals are clearly recognizable.

The first stage of the alteration of these rocks is the appearance of coronas around the ferromagnesian minerals. Olivine is gradually replaced by a rim of orthopyroxene, which finally recrystallises to a granular aggregate pseudomorphing the original olivine crystal. At the contact of the pyroxene with plagioclase a corona of hornblende is produced which gradually encroaches upon and replaces much of the plagioclase.

The next stage is probably induced by the effect of tectonic pressures acting on the rock. The plagioclase recrystallise to an equigranular aggregate of polyhedral crystals. At the same time some of the hornblende reacts similarly to the pressure and recrystallises. This process is continued until an amphibolite consisting essentially of amphibole and plagioclase is produced. It is still possible to recognise the igneous texture in the areas of granulated plagioclase which pseudomorph the original laths (Figure 42).

In further response to the deformation a true metamorphic fabric results, with segregation of amphibolitic and felspathic bands. The hornblende grows such that the prisms are elongated in the tectonic 'b' direction. Some garnet grows at this time and is flattened in the plane of the schistosity by further deformation. Some quartz-felspar pegmatites appear by segregation and favour low pressure areas such as fold crests (Figure 55a) and shear zones (Figure 53a) for emplacement.

During the metamorphism and tectonism much of the hornblende is replaced by biotite where conditions are favourable. Such conditions were attained at the contacts of the amphibolites where shearing was most intense and the introduction of water was possible.

The hornblende and biotite remained in a strained state at the close of the deformation and were attacked by the more stable quartz and plagioclase.

The latest event to be clearly recorded in the rocks is the growth of the post-tectonic porphyroblasts of garnet, which replace and surround all the earlier minerals.

Late shearing under much lower pressure-temperature conditions resulted in the breakdown of biotite and residual hornblende to give chlorite. At the same time the plagioclase was greatly sericitised and the quartz became very strained.

LATE PEGMATITES

The late pegmatites are the sole representatives in the Modum region of the 'Younger Complex', which elsewhere postdates the phase of basic intrusions. They are found evenly distributed over the region, transgressing all the rock types so far described. They are distinguished from the segregation pegmatites observed in the amphibolites and quartzites by their field relations.

A single pegmatite can be traced through several different rock types with no change in mineralogy, whereas the mineralogy of the segregation pegmatites is related to that of the host rock. The form of the segregation pegmatites is strictly controlled by the structure of the host rocks whereas the late pegmatites are sharply transgressive. The segregation pegmatites rarely attain a great size while the late pegmatites are found as interconnected lenses and veins over tens of metres.

At the northern extremity of the North Mine at Skuterud (Figure 110) several interconnected lenses of pegmatite can be mapped around interdigitating and enclosed amphibolite and meta-sediment. These pegmatites transgress both the sulphidic-graphite schists of the 'fahlbands' and the cobalt mineralisation. However, a crystal of cobaltite was collected from the centre of one of these pegmatite lenses. Thus the evidence regarding relative age of emplacement of the ore and the pegmatites is here ambiguous.

On the shores of Stolldammen, the artificial lake at the

entrance of the Ludwig Eugen Crosscut, two small quarries have been opened in pegmatite. One quarry is in a felspar-quartz-muscovite-tourmaline pegmatite, worked for the potash felspar, and the other is in a muscovite-quartz-tourmaline-felspar pegmatite, operated by the Germans during World War II for mica. These two pegmatites are unusually rich in microcline and muscovite respectively.

The usual mineralogy of the pegmatites is microcline, plagioclase and quartz, often in graphic intergrowth (Figure 50), with muscovite (p. 56) and biotite (p. 53). Tourmaline is often present in minor amounts. However, some of the pegmatites of Modum, notably in the Snarum region, are famous for the large euhedral crystals of schorlite that they contain. This tourmaline is demonstrably of early formation as large crystals have been deformed and bent by the surrounding media (Figure 51).

The pegmatites transgress all the metamorphic rocks in the region and appear to be completely undeformed. Thus they are later than the last Precambrian orogenic event. They are cut by the quartz-calcite veins (p. 307) and dolerite dykes (p. 274) which are attributed to the Permian. They are also affected by the major shear zones in the region. Similar pegmatites have not been described from the Cambro-Silurian sediments of the Oslo region, nor from the Permian alkaline igneous Oslo province. They are therefore assigned to the later Precambrian. Two radioactive age determinations have been made on minerals from these pegmatites with the following results (Neumann, 1960), 1041 m.y. (phlogopite K-Ar.); 1055 m.y. (Muscovite, K-Ar).

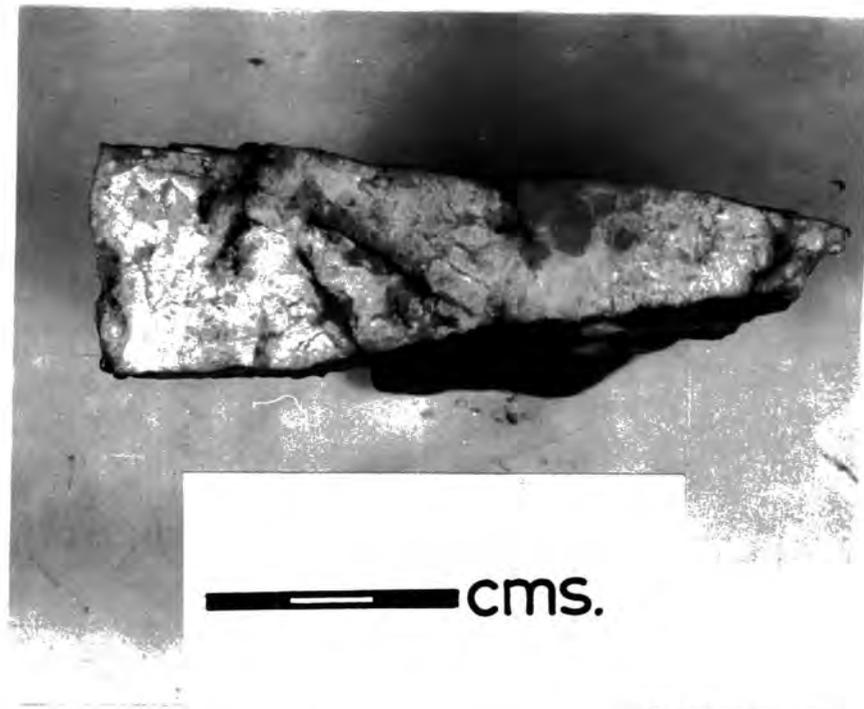


Figure 50: Microcline-plagioclase-quartz-biotite pegmatite, North Mine, Skuterud. The specimen has been stained with sodium cobaltinitrite, to differentiate the potash felspar (dark grey) from the plagioclase (pale grey) and quartz (white).



Figure 51: Tourmaline crystal which has suffered deformation. Plagioclase-tourmaline-muscovite-quartz pegmatite, Stolldammen, Skuterud.

STRUCTURAL DATA

Introduction

As is to be expected in an area of such antiquity the structural relations are very complex. A consideration of the outcrop patterns of the Modum area reveals little in the way of a coherent structural pattern. In the 50 sq.kms. covered by Jø/sangs map only one clear fold closure can be seen, the majority of the area having outcrops of lensoid shape with a dominant north-south trend.

Analysis of small scale structures observable in single outcrops can sometimes lead to a reconstruction of the geometry of the major structures. The methods and theory of this type of analysis were largely developed in the German literature but an increasing amount of work on this subject is now appearing in the English language press. The application of some of these methods has been attempted for the Modum area.

Planar structures

Foliation

The pronounced foliation (S_{11}) is defined by alternating layers of different composition, and a penetrative schistosity in the micaceous rocks. The foliation separates beds of different mineral composition and is parallel to formational contacts, it is thus believed to result from the accentuation of the bedding planes by metamorphic differentiation, emphasising the quartzo-felspathic and calc-ferromagnesian bands. Extreme shearing out of limbs of isoclinal folds and slip of fold cores between foliation surfaces

indicates that the bedding surfaces have been considerably modified during metamorphism. The schistosity is formed by the strong preferred dimensional orientation of the mica flakes. Banding in the quartzites is emphasised by darker micaceous selvages between quartz rich bands.

In some areas primary sedimentational structures may be present. Intense shearing makes any such interpretation dubious, but structures are occasionally observed which suggest graded bedding (Figure 11). The direction of younging inferred from such evidence has very local significance in such a tightly folded terrain.

Axial plane cleavage

Where the first folds (F_1) are tight isoclinal shearing has occurred, preferentially in pelitic horizons, along surfaces parallel to the axial planes of the folds. This gives rise to a poor axial plane cleavage (S'_1) which is usually subparallel to S_1 .

Si-Se fabric relations

The fabric of isolated grains within a porphyroblast, named Si by Sander (1930), may have various relations with respect to the external or Se fabric. The field relations sketched in Figure 53b indicate that in some late garnets Si is parallel to Se of the enclosing amphibolite. This indicates that the garnets grew blastically after the production of S_1 and its subsequent folding (F_1) and that no major deformation has occurred since then.

Joints

Joints are very well developed in the quartzites and

Figure 52: Profile of an F_1 fold developed in siliceous granulite 200 metres south-west of Stolldammen, Skuterud. The quartz rich bands have been thickened in the fold hinges and sheared out on the limbs. The micaceous bands are affected by micro-folding in the fold hinges. Shearing has occurred along the surface parallel to the axial plane of the fold, giving rise to a poor axial plane cleavage (S'_1).

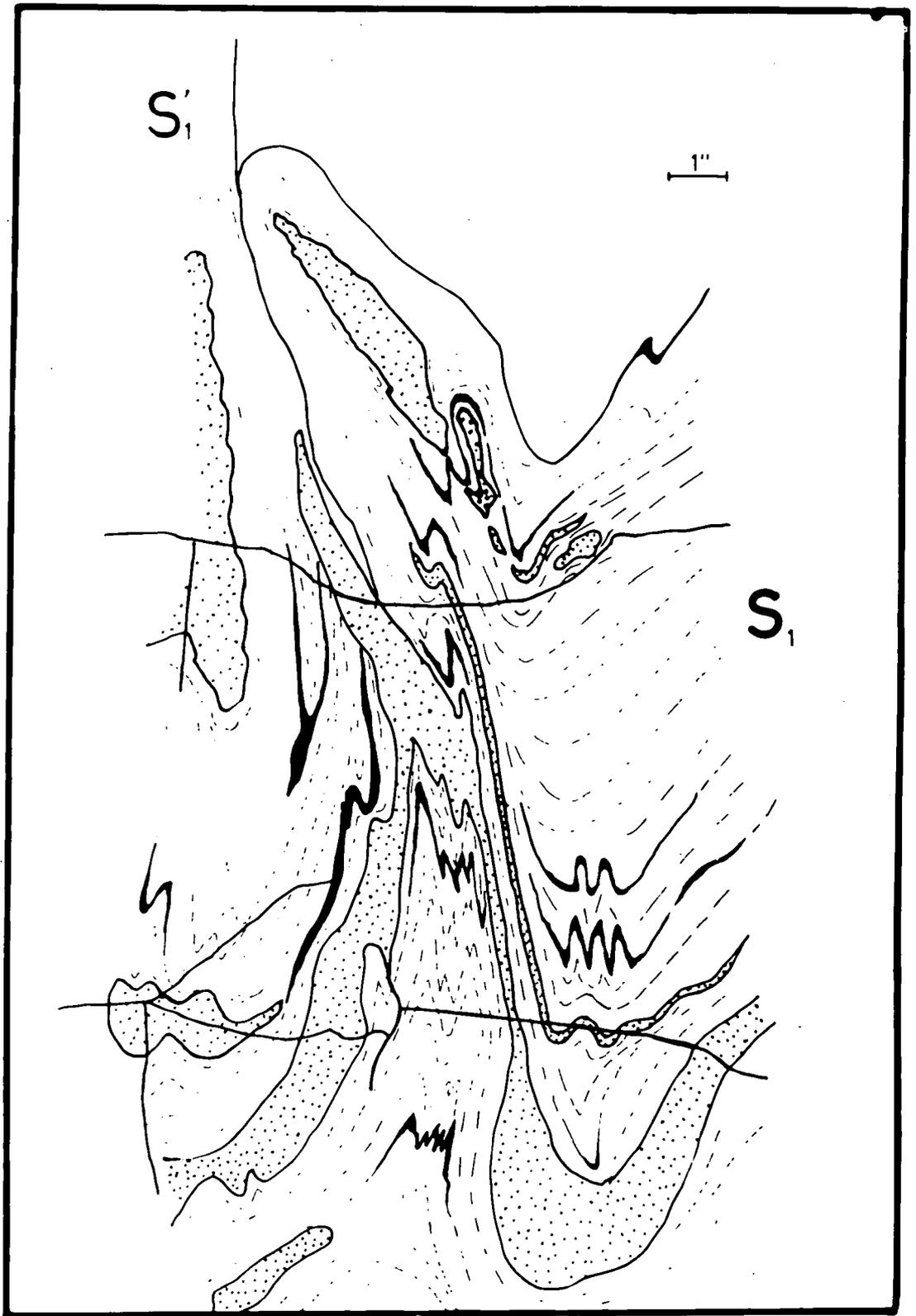


Fig. 52

- Figure 53: a. Shear zone in biotite amphibolite at an oblique angle to the foliation. Dilational, feldspathic pegmatite has preferentially developed in this zone.
- b. Minor folding in garnetiferous biotite amphibolite, BØen Farm. The foliation is traceable through the garnet porphyroblasts.
- c. Shear zone in sulphidic siliceous granulite in a pillar at the Central Mine, Skuterud. A fold, with an axis plunging steeply in a north-western direction, is developed immediately adjacent to the shear zone. The direction and amount of the plunge of this fold axis contrasts strongly with others in the area. It is considered to be the result of drag associated with the shearing.
- d. Part of the Ludwig Eugen Crosscut showing two different generations of jointing.

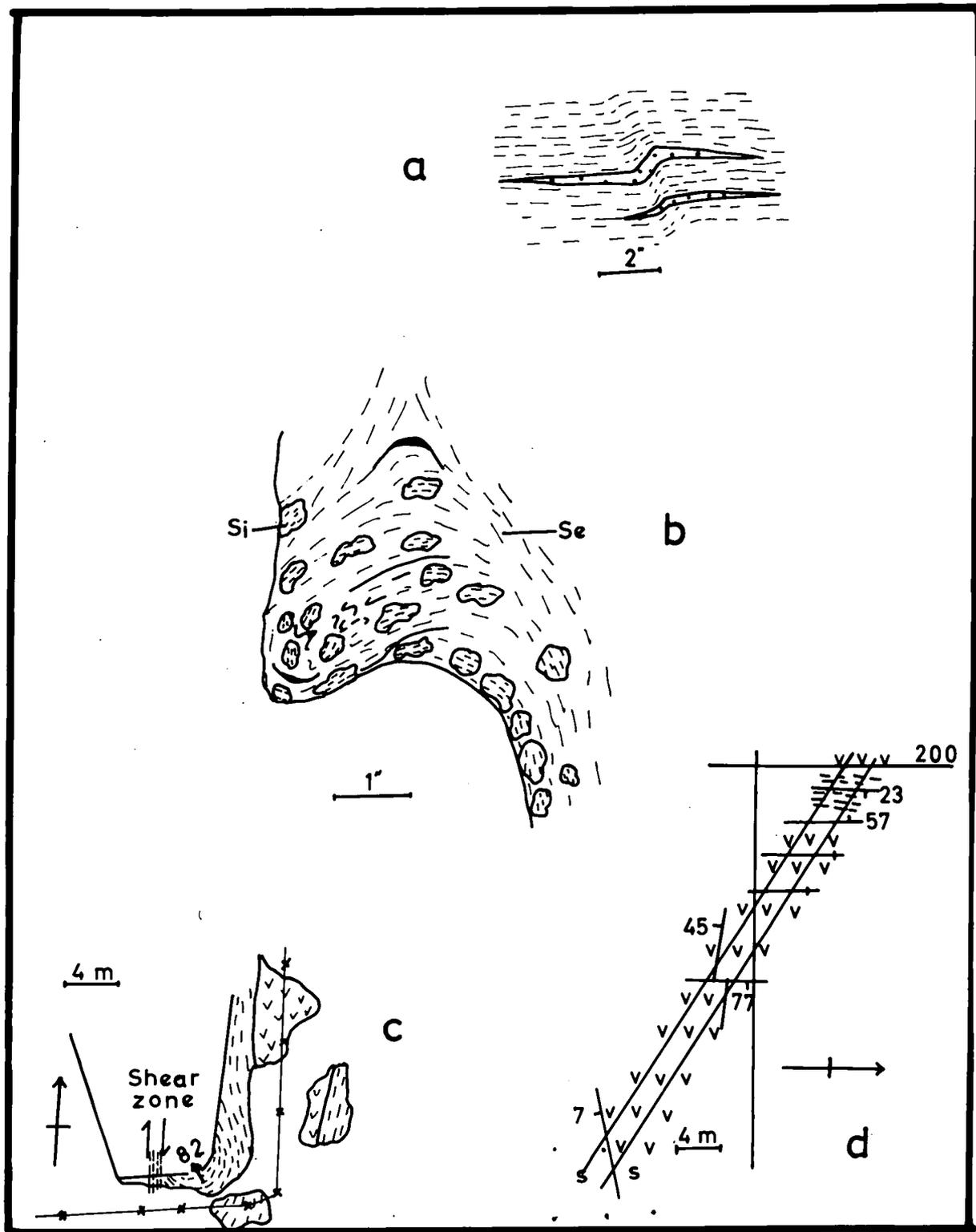


Fig. 53

quartzo-felspathic horizons. Poles to 189 joints planes have been projected using the lower hemisphere of the Lambert equal area net and contoured at $\frac{1}{2}\%$, 1%, 2%, 3% and 5% intervals (Figure 54). From this contoured diagram it is possible to see seven major jointing directions. In order of decreasing frequency these have the following attitudes:

- | | |
|---------------------------------|---------------------------------|
| (1) $042^{\circ}/90^{\circ}$ | (5) $138^{\circ}/70^{\circ}$ NE |
| (2) $076^{\circ}/90^{\circ}$ | (6) $002^{\circ}/75^{\circ}$ E |
| (3) $122^{\circ}/80^{\circ}$ SW | (7) $272^{\circ}/70^{\circ}$ S |
| (4) $152^{\circ}/75^{\circ}$ SW | |

Sets 2 and 7 are normal to the axial planes of the F_1 folds and represent the transverse, 'ac' or 'Q'-joints, common in many such regions, in the quartzites they are frequently infilled with massive white quartz. Sets 1 and 3 form two sets at an angle of 80° to each other. The obtuse angle is bisected by the first set of linear structures and they could be interpreted as shear joints related to the first folding in the area. Set 6 is essentially parallel to the F_1 axial planes, they are well seen in the open cast workings as large slickensided faces. The other joint sets are not easily interpreted in terms of the major fold structures (Figure 54).

Faults and shear zones

No major faults were observed in the area perhaps due to the lack of exposure and good marker horizons. If they played an important role however it would be expected that they would be observable in the underground workings mapped in detail at Skuterud but this was not found to be the case.

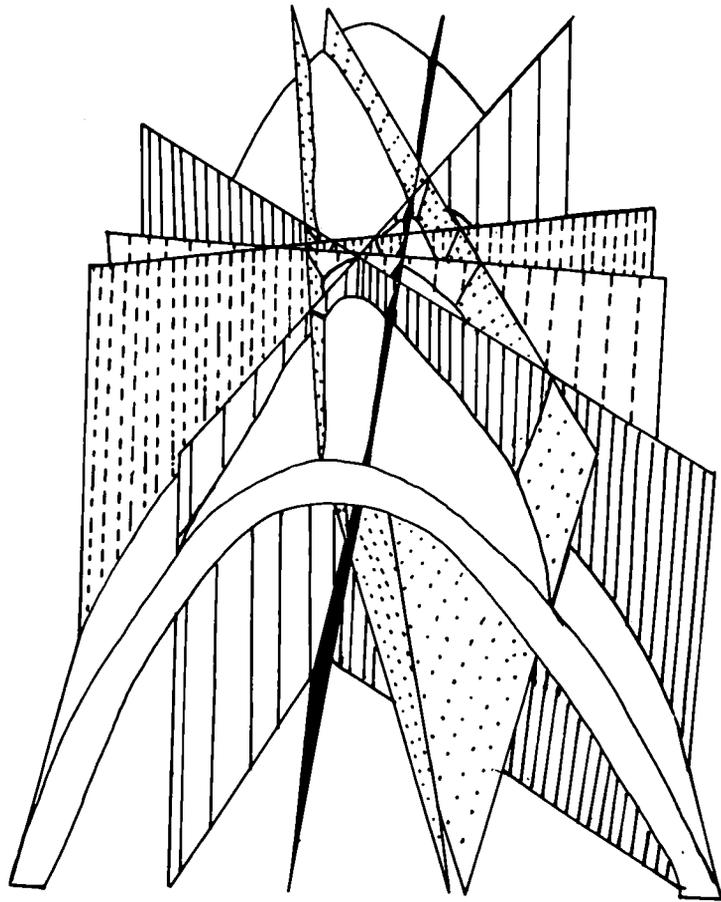
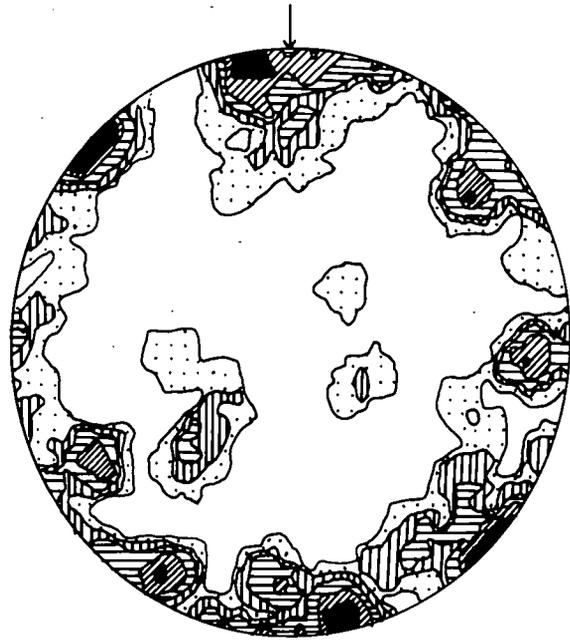


Fig. 54

However the jointing described above as subparallel to the F_1 axial planes is related to major zones of shearing observable in many of the open-cast workings. Shearing has been commonest in the sulphidic graphite schists, but is also observed in the amphibolite (p.116) and other rock types. The pegmatites are affected by the shear zones but are later than the main deformation.

Linear structures

The structural geometry of the Modum area is triclinic due to the presence of more than one set of linear structures. Linear structures formed by elongation of quartz-sillimanite nodules, preferred orientation of hornblende prisms, slickensided movement planes and the axes of minor folds were distinguished. Style of folding proved to be the best criterion for distinguishing folds belonging to different generations.

The first folds F_1

F_1 folds are developed nearly everywhere in the Modum district, where outcrop is suitable. Scale of folding varies from microscopic to macroscopic. The presence of rootless fold hinges indicates that the surfaces between schist and quartzite have become zones of gliding. The general style of the F_1 folds indicates a high degree of plasticity during the deformation. Although fold closures are rarely seen it is clear from the small scale structures that the folding is dominantly isoclinal. Some typical profiles of the first folds are shown in Figure 52, 55a, b, c.

The mineral lineations are found to parallel the F_1 fold axes and, surprisingly, the large slickensides ('glidestriper' Rosenqvist, 1948) also have the same orientation. These three types

Figure 55: a,b,c. Typical F_1 fold profiles.
d. F'_2 fold, with plot of poles to foliation.
Wall of South Mine opencast working, Skuterud.
e. F''_2 fold, with a plot of poles to foliation
and projections of lineations. Muggerud Trial.

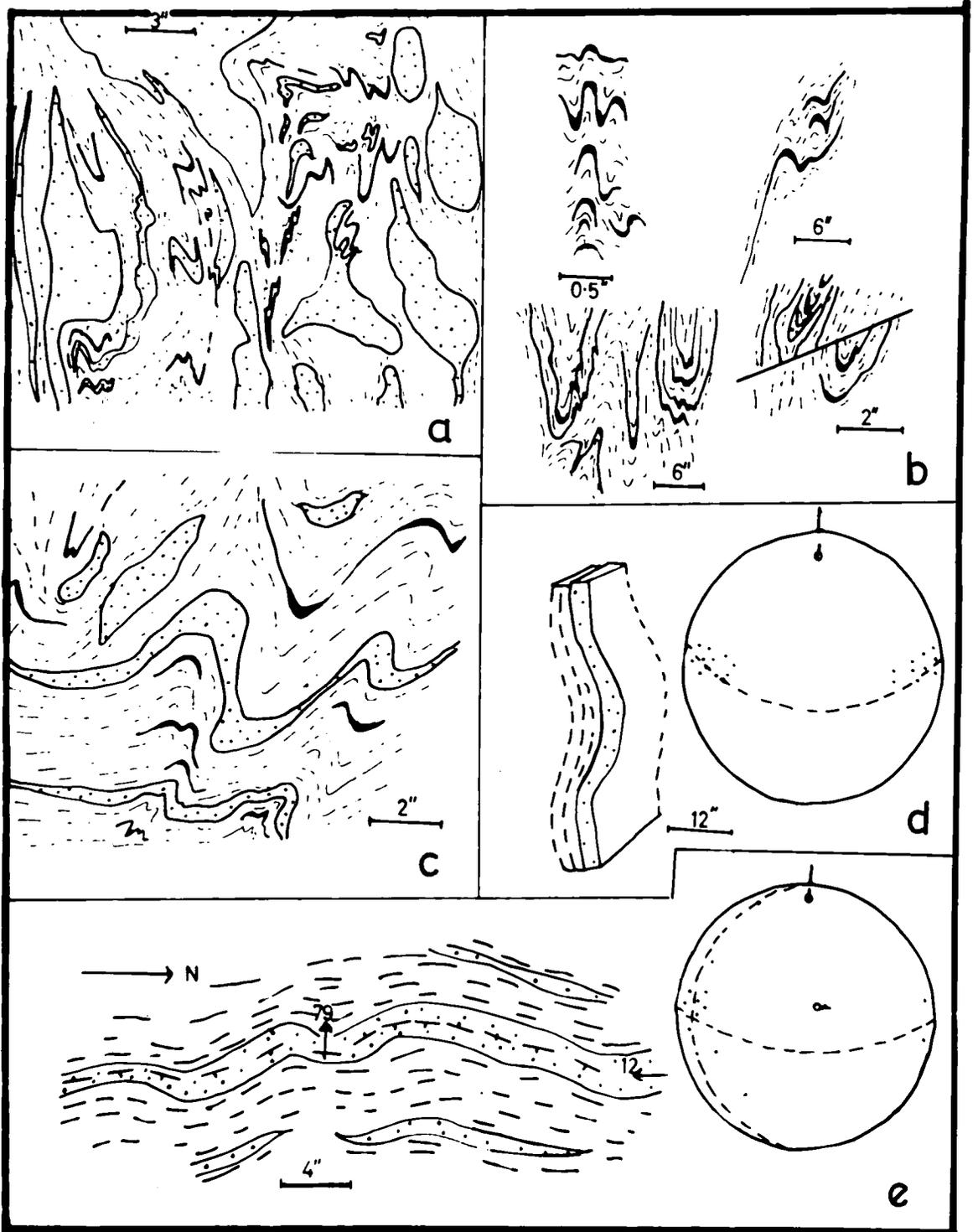


Fig. 55

of linear structure will be collectively referred to as L_1 .

Figure 56a shows all the L_1 data plotted on one synoptic projection, a rather large scatter is obtained. This scatter could be due to data from axially homogeneous sub-areas being plotted on one diagram or could be the expression of several phases of folding. Data for three small sub-areas are plotted on Figures 56b, c, d, it is seen that although the lineations show better approximations to discrete maxima they are still rather widely scattered.

This indicates the existence of more than one phase of deformation and could be explicable by the latest phase being the most intense and obliterating all but a few of the earlier lineations, or the late phase or phases being weak and merely increasing the scatter of the F_1 lineations.

The second folds F_2

The second folds are open, similar-type folds with regular curvature. Some typical profiles of F_2 folds are given in Figure 55d,a. They may be considered as warps of the S_1 planes and three distinct axial trends are observable.

F'_2 folds have a sub-horizontal, north south trending axis with a horizontal axial plane. They are most clearly seen in vertical outcrops as neutral, open folds affecting S_1 which is bent round their closures.

F''_2 folds have a sub-vertical axis with a vertical east-west trending axial plane. They are commonly seen on horizontal outcrops where they warp the S_1 foliation. Figure 55e shows a plot of data from one outcrop showing how the spread of poles to S_1 and

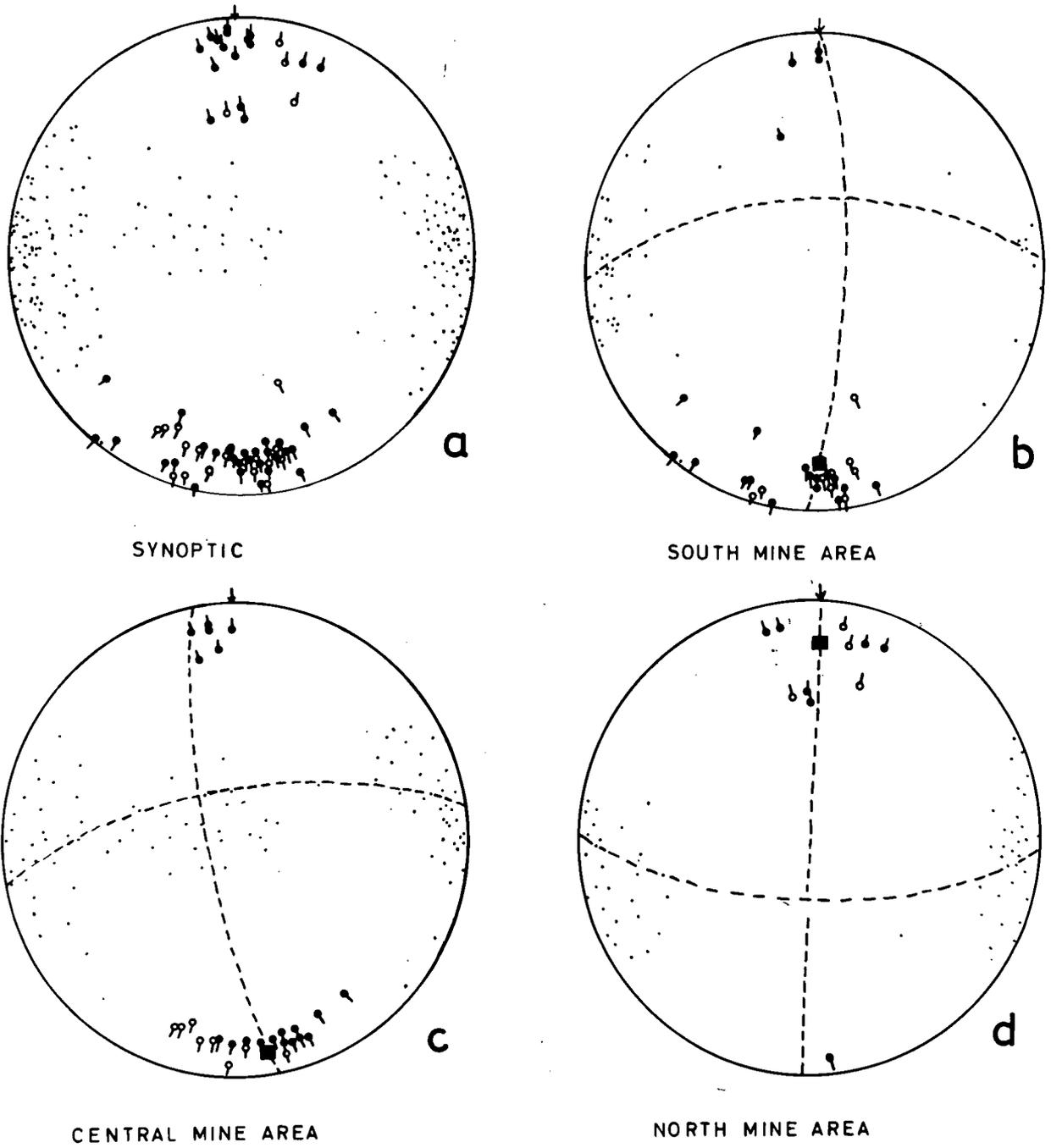


Fig. 56

Figure 57: Effects of F_2 folds on S_1 and L_1 .

F_1 . Idealised plot of poles to foliation (S_1) and 'b' - lineations (L_1) produced by folding about a sub-horizontal, north-trending fold axis (F_1).

F'_2 . Idealised plot of modifications to S_1 distribution produced by further gentle folding about an axis having the same trend and plunge as F_1 but a near horizontal axial-plane (F'_2).

F''_2 . Illustrating the effects on S_1 and L_1 of folding about a vertical axis (F''_2).

F'''_2 . Illustrating the effects on S_1 and L_1 of folding about a horizontal axis with an east-west trend (F'''_2).

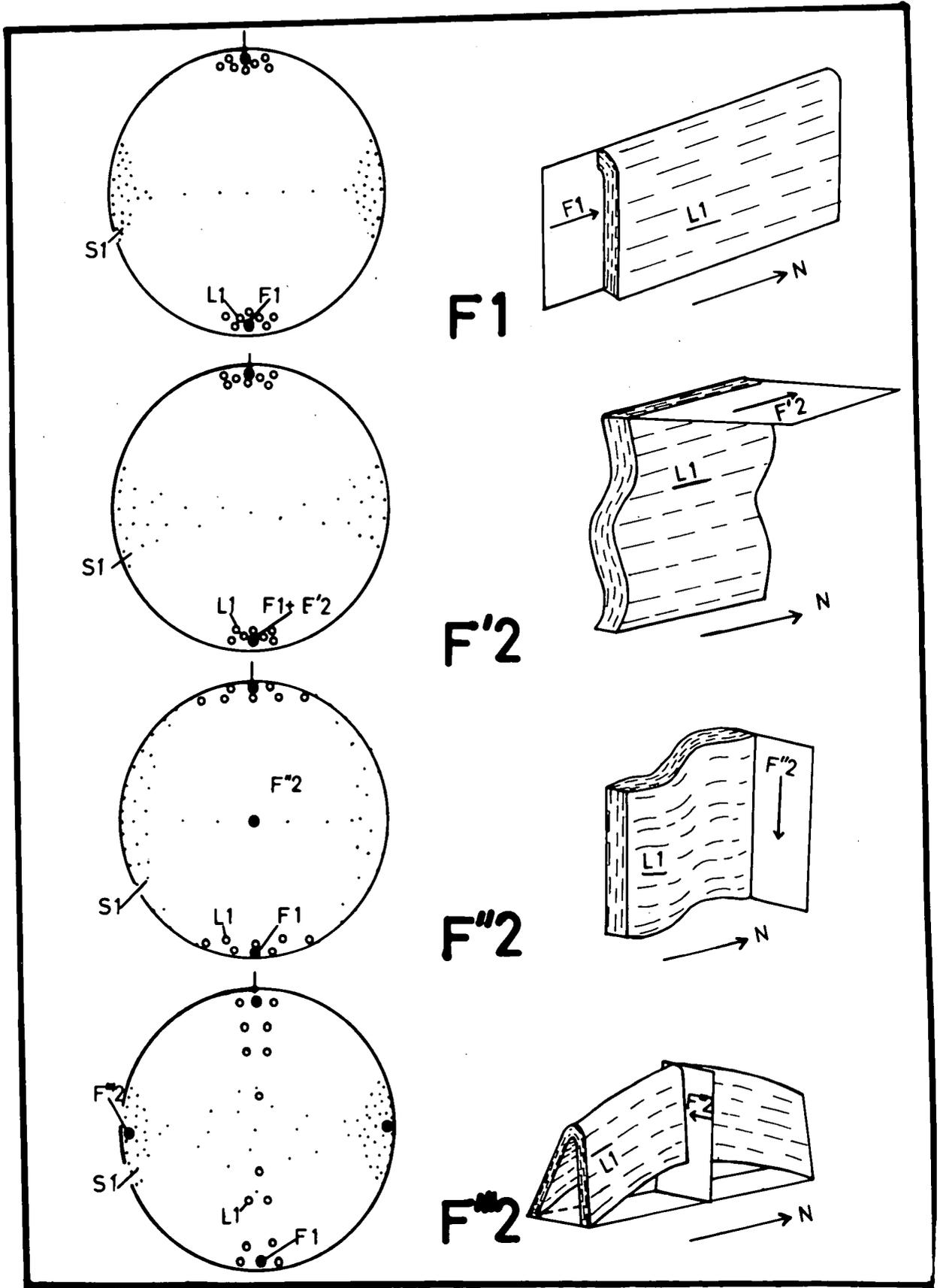


Fig. 57

projections of L_1 are affected by the F'_2 folds.

F''_2 folds have a sub-horizontal east-west trending axis with a vertical axial plane. They are not clearly seen in the field but are considered to be the cause of the variable plunge of L_1 as illustrated in Figure 58.

The effects of F'_2 and F''_2 are clearly seen in the large open workings where their interaction causes the S_1 surfaces to form a series of basins and domes.

It is not possible to say from the field data what the relationships of F'_2 , F''_2 and F'''_2 are in time.

From the evidence of the F_2 folds it is concluded that the spread of data for L_1 is due to a weaker late phase of deformation.

Conclusions

Figures 56b,c,d are synoptic diagrams of structural data for the areas of the South, Central and North Mines respectively at Skuterud. In the light of the foregoing discussion it is possible to say that the distribution of poles to S_1 is largely controlled by the F_1 folding. A spread of the data is caused by the superposed effects of F'_2 and F''_2 .

The axial trend of F_1 and its associated L_1 is dominantly north-south but changes in trend are caused by the effect of F''_2 , and it has a slight plunge either north or south dependant on the effects of F'''_2 .

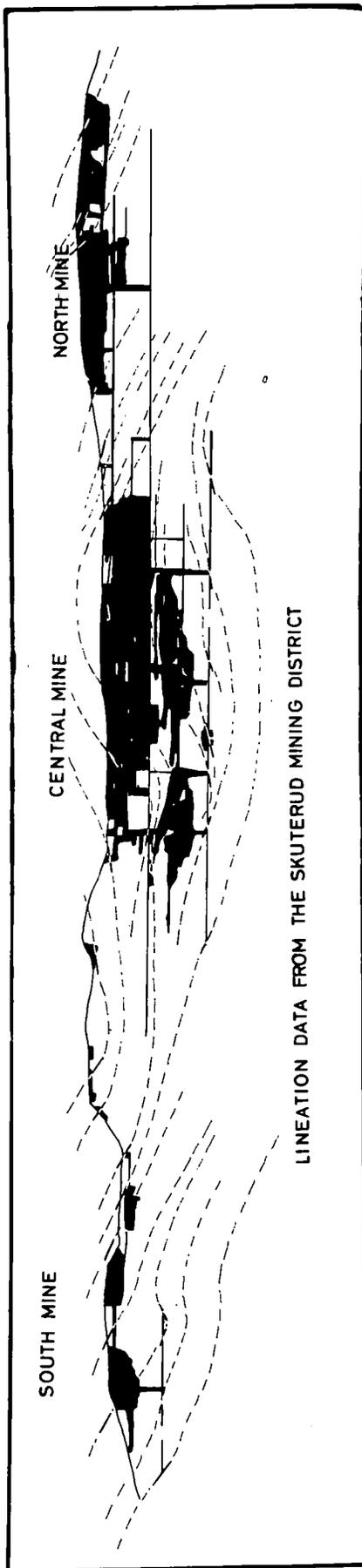


Fig . 58

THE FAHLBANDS

Origin and meaning of the termOriginal meaning of 'fahlband' at Kongsberg

The term was first used at the Kongsberg silver mines, where the common occurrence of stripes of light and dark rock, formed by metamorphic processes were referred to as bånds or baands from the earliest days of mining (C. Bugge, 1918, p. 87).

This nomenclature was extended by the German mine foremen of the 17th and 18th centuries to include baands impregnated by sulphide minerals. These are common in the mining district and have an important effect on the localisation of the argentiferous veins. Crosscutting calcite veins carried workable silver only where they crossed such sulphide rich zones or fahlbaands.

The etymology of the prefix fahl is of some interest as there are several possible derivations of the word. The sulphidic bands were originally called faldbånder, the alternative spelling fahlbånder was first used on Boeck's large manuscript map of the Kongsberg district in the 1790's (C. Bugge, op. cit.). The Royal Commission on the future of the Kongsberg mining district (1833) used faldbånd, and this was continued for some time until Bøbert (1848) decided that fahlbånd was correct and defined the term using that spelling. He was followed by Durocher (1849) and the word, with Bøbert's spelling and definition became popular in the geological literature.

To understand what the word meant originally we must consider the German roots of fahl and fald. There are several possibilities.

The word fall has several meanings in German, two of which are of interest to us. It can mean an event or happening (Muret - Sanders, 1898), thus Norwegianised to fald it could mean a lucky happening. There is an old miners expression common at Kongsberg, "Uden gang og fald, intet sølv." (Literally, "Without vein and luck(?), no silver.") Thus faldbånd could mean the 'lucky' parts of the vein - i.e. where it crossed sulphidic schists.

Fall also means a mineralised vein or horizon (Grimm and Grimm, 1862), but it would seem strange that two words of such similar meanings as fall and baand should be used together.

The German word faul was in common usage to describe decomposed and weathered rocks. This could easily have been Norwegianised to fahl and used in reference to the very weathered appearance of the sulphidic schists.

The German word fahl is used in the sense of faded or ashen coloured, as in fahlerz - 'grey-ore'. (Muret - Sanders, op.cit.) However, it can also mean reddish-brown, rusty or fawn coloured. In this sense it is derived from the Middle High German root val (Davis, 1891), and is closely related to the Dutch word vaal (Cf. Transvaal - 'across the Orange River). Because of this variety of meanings Bøbert (1848) considered it to be the most likely root, hence his decision to standardise the spelling as fahlbånd. He points out that in the sense of grey or bleached it is applicable to some of the weathered sulphide schists, however brownish and rust coloured

outcrops are very common and it adequately describes these too. There is a connection between the depth of weathering and the sulphide content and hence the likelihood of silver. The miners of his day would say "Fahlbaandet blomstrer" ("The fahlband flowers"), referring to the brighter colours observed when the sulphides are abundant, thus suggesting that fahl was used in connection with colour.

Thus Bøbert's term fahlbaand is a meaningful and useful one for describing weathered sulphide-rich horizons, but it was first used as late as 1790 (see above), whereas fallbånd and faldbånd were in use much earlier.

The first reference to faller is by Duwal writing earlier than 1725. (A. Bugge, 1932):

"Ertsfall er et navn som all edel sten hvorudi sølv kan finnes og vokse blir naevnt."

("An ore 'fall' is a name given to all stones where silver is found and grows.") Duwal goes on to say (author's translation):

"These have a different colour from all other stones, and one finds in these above all other stones fine sulphides.....
.... at the surface it is easy to see that they are burnt red by the sun, which calcines and burns the mineralic and martial sulphur, which is in the 'faller'. "

All the faller which contained silver were called edle faller (precious faller) or ertsbaand (ore band). In a map produced in 1771 Madelung collected all the adjacent edle faller together, which, on the German copy, he pluralised as Fallbände and, on the

Norwegian, as fallbänder. This is the first geological map of the mining district. (A. Bugge, 1932, Figure 1).

In 1931 the employees at the silver mines still used fallbånd and fahlbånd in different senses. (Bugge, op. cit.). Thus fahlbånd was used in Bøbert's sense to describe a sulphidic schist with a rusty weathering outcrop, while fallbånd was reserved for such schists which were important in localising the silver deposits.

As fahlbånd; Bøbert's term, gained such popularity and was incorporated in the literature and used to describe other deposits is it suggested that it be retained in preference to fallbånd despite the priority of the latter term.

In the English language literature the word has been occasionally used, with the spelling - fahlband. This spelling will be retained in the present work.

Spreading of term through Scandinavia

Kongsberg was one of the main mining centres of Europe during the 17th, 18th and 19th centuries. Many foreign workers, notably Germans, were employed. A School of Mines, the forerunner of the present University of Oslo, was established, and many distinguished foreigners visited the mines. Probably due to these influences the use of the word fahlband spread to many mining districts of Scandinavia and Germany.

Incorporation into geological literature

Bøbert's (1848) publication on fahlbands was in a German journal with a wide readership, in the following year Durocher gave

the term wide publicity in a leading French journal. The use of the word fahlband to describe 'basement gneiss' containing sulphidic layers, lenses or streaks was thus opportunely presented at a time when many deposits were being geologically described for the first time. The word became 'fashionable' and was applied rather indiscriminately to any deposit bearing a slight resemblance to what Bøbert had described from Kongsberg.

It was used to describe the cupriferous chlorite-schists of Stora Strund, Sweden (Johansson, 1909), the nickeliferous norites of Ringerike, Norway, (C. Bugge, 1914) and Bodenmais in Bavaria (Krusch, Beyschlag and Vogt, 1913). Stelzner and Bergeat (1904) described the Co-Ni deposit of Schladming in Styria and suggested that the Co-Ni veins were enriched where they crossed fahlbands. (This suggestions was later refuted by Schmidt and Verloop (1909). Fahlbands, closely analagous to the Kongsberg type were described from Kandalashka Bay, Russia (Tokarev, 1935) and the Hungerford and Queensboro fahlbands of Ontario were described by Fraleck, (1909).

Thus in textbooks on economic geology published early in the 20th century a section was usually devoted to deposits of "the fahlband type" (Krusch, Beyschlag and Vogt, 1914; Phillips, 1896; Lindgren, 1919). However confusion was beginning to arise on two counts, firstly disseminated sulphides are common in magmatic as well as in metamorphic rocks, and secondly doubt was being cast on Bøbert's syngenetic interpretation of the sulphidic schists. It was suggested that fahlband should be restricted in usage as a purely descriptive petrographic term having no genetic connotations. (Krusch, Beyschlag and Vogt, 1914, p. 46).

Unfortunately this advice was not followed. The trend set by Bøbert himself in describing the skarn iron ores of the Arendal district, Norway and the Cu-Sn deposits of Pitkaranda, Finland, as fahlbands was followed. Ultimately the term was applied to deposits of galena in limestone (Sala, Sweden) and cerargyrite in sandstone (Utah, U.S.A.) (Munster, 1894). The deposits are not restricted to bands in metamorphic gneisses, nor do they have the characteristic 'fahl' weathering, thus it is meaningless to call them fahlbands.

The cobalt ores at Modum are only some 20 kms. from Kongsberg, and they are associated with sulphidic schist. Bøbert (1848) coined the term 'cobalt-fahlband' for these and similar deposits in Sweden. The present author has found no reference to cobalt-fahlbands outside Scandinavia and discussion of these deposits is deferred to a later section.

Decline in usage

The Kongsberg fahlbands were only worked for copper in a few small trial workings, they were of greater interest for their property of causing the silver to precipitate at their intersection with veins. Thus the original fahlbands were sub-economic concentrations of sulphides. In later usage the term has been applied to workable deposits of varying type, this has caused confusion as to the exact meaning of the word. As most of these deposits were worked out, and with a growing knowledge of the different mechanisms of ore genesis, the rather vague term of 'fahlband' was neglected in favour of others for the description of new deposits.

Modern references to the term

The term is very rarely found in the modern literature, being used only in reviews of those deposits to which it was originally applied in older descriptions.

As an illustration of how far removed modern usage is from the original meaning of the Kongsberg faldbånd the following passage from Bateman (1942, p. 139) will serve:

"Many lodes, or fahlbands, formerly thought to have been formed by impregnation of the shear planes are now known to have originated through replacement of the sheared rock."

Suggested definition

It seems a pity that the use of a term of such long standing, having no genetic implications, should be allowed to lapse. Sulphidic schists in layers and lenses are very common in high grade metamorphic terrains in many parts of the world. These have been variously described as sulphide-graphite schists (Marmo, 1960); pyrrhotite-gneiss (Fisher, 1962) etc. It is here suggested that a term already exists to describe such occurrences.

The author suggests the following definition:

A fahlband is a concordant layer or lens of sulphidic impregnation in a metamorphic sequence. The sulphides should be in such a quantity that they are too abundant to be classed as accessory minerals, but too sparse to form a massive ore lens. Fahlbands should have a characteristic rusty-brown appearance on weathering.

THE KONGSBERG FAHLBANDS

Introduction

As Kongsberg is the classic fahlband locality, some time was spent on a field and laboratory study of the district. Kongsberg is situated some 80 kms. west of Oslo in the Fylke of Buskerud. The silver mines are in rocks of the Kongsberg Formation which here consists mainly of banded dioritic and granitic gneiss with amphibolitic ('Vinor diabase') bands, in the form of long dykes, associated with more massive olivine-norite ('hyperite') bodies.

The fahlbands form long, narrow zones where the rocks have been heavily impregnated with sulphides. The later calcite veins are argentiferous only where they cross certain of these fahlbands.

Previous work on the Kongsberg fahlbands

The silver mines have been in operation since 1623 and have been a very important factor in Norway's economy. It is, therefore, not surprising that much has been written on the geology of the district. Munster (1894) reviewed the geology and gave a historical bibliography of works concerning the Kongsberg mines. The most complete, recent account of the geology of the whole district is given by C. Bugge (1917) whose work also includes a full bibliography. The most notable modern work is that of Neumann (1944) who described the mineralogy of the silver veins. The latter is the only English language account of the area.

Helland (1879) considered that the sulphides were of premetamorphic origin as they are found completely enclosed by unfractured quartz and garnet. Vogt (1899b) challenged this hypothesis on the grounds that sulphides are present in the granite-gneiss, which Vogt interprets as intrusive and later than the metasediments. Bugge (1917) considers the sulphidic impregnations to be related to the Vinor amphibolites as zones rich in sulphides are found alongside some of these amphibolites. (Bugge, op.cit., Figures 5 and 6). Neumann (op.cit. p. 15) considers the fahlbands to be in zones of metasomatic alteration where amphibolite has been converted to garnet-amphibolite and dioritic gneiss metasomatised to biotite-garnet-schist, sericite-schist or chlorite-schist.

Field relations

Figure 59 shows the characteristic weathering of a fahlband at Kongsberg. The distribution of such weathering is a useful criterion for deciding the areal extent of the sulphide impregnation.

Weathering of this type is present in some of the olivine-norite bodies, where it derives from the breakdown of nickeliferous pyrrhotite. Sulphidic impregnation of the same type is found also in the surrounding gneisses adjacent to such intrusive bodies. These sulphidic zones differ from the true fahlbands in that they are of very patchy distribution and show no regularity with respect to the strike. At Kongsberg the mineralogy is also different, sulphides associated with the basic intrusives are dominantly pyrrhotite with minor chalcopyrite and very subordinate pyrite, while the true

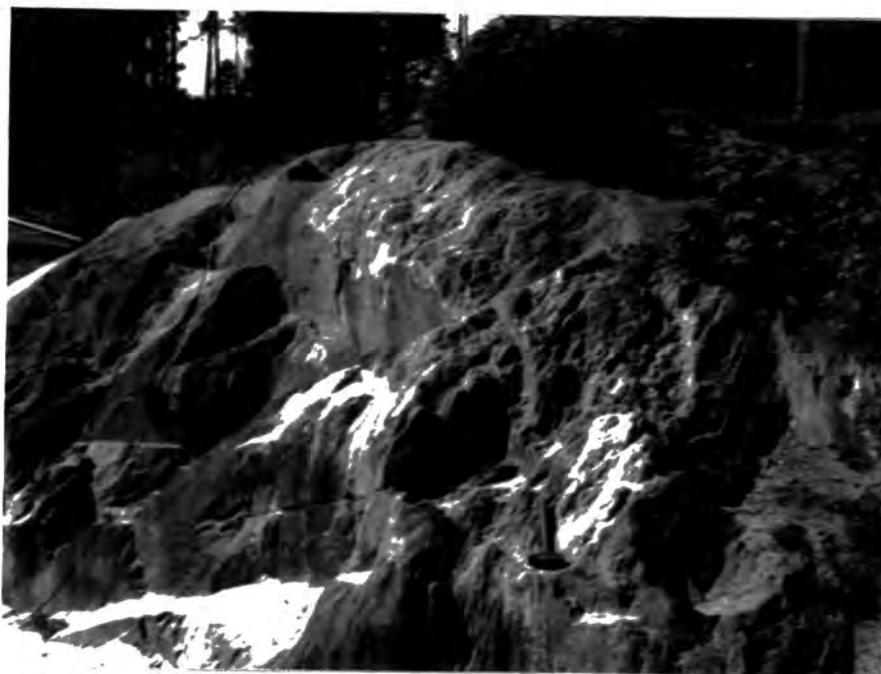


Figure 59: Weathered outcrop of sulphidic schist showing typical 'fahl' weathering. Near Kongens Mine, Kongsberg.

fahlbands are dominantly pyritous with subordinate pyrrhotite and chalcopyrite. As this type of impregnation is not restricted to 'bands' it has been suggested that they be called 'fahler' (C. Bugge, 1917). Some of these fahler have acted in the same way with respect to the late veins as the fahlbands, and silver mines have been developed on them (e.g. Knute Mine and St. Andreas Mine).

The true fahlbands are irregular in distribution on a small scale, but viewed in the perspective of the whole mining district they are seen to be restricted to zones paralleling the north-south regional strike of the metamorphic sequence. There are two main zones of importance to the distribution of silver - the Underberget and Overberget fahlbands. In addition there are about 13 other major fahlbands and innumerable minor ones, of lesser importance as far as the silver mining is concerned (Figure 60).

The Underberget fahlband averages about 50 m. in width with a maximum development of 80 m. in the central part. Sulphides are rather evenly distributed in the whole zone, to the north of Kampenhaugen the zone becomes diffuse and finally dies out at the Svartaas Mine. The sulphides are particularly well developed from Samuels Mine south to the Kobberberg River, with chalcopyrite becoming quite common. In this part of the fahlband, at the Herzog Ulrich Mine, a stope ('Kobberdybet') was unsuccessfully opened for copper (Munster, 1894). To the south of the Kobberberg River the zone splits and becomes diffuse.

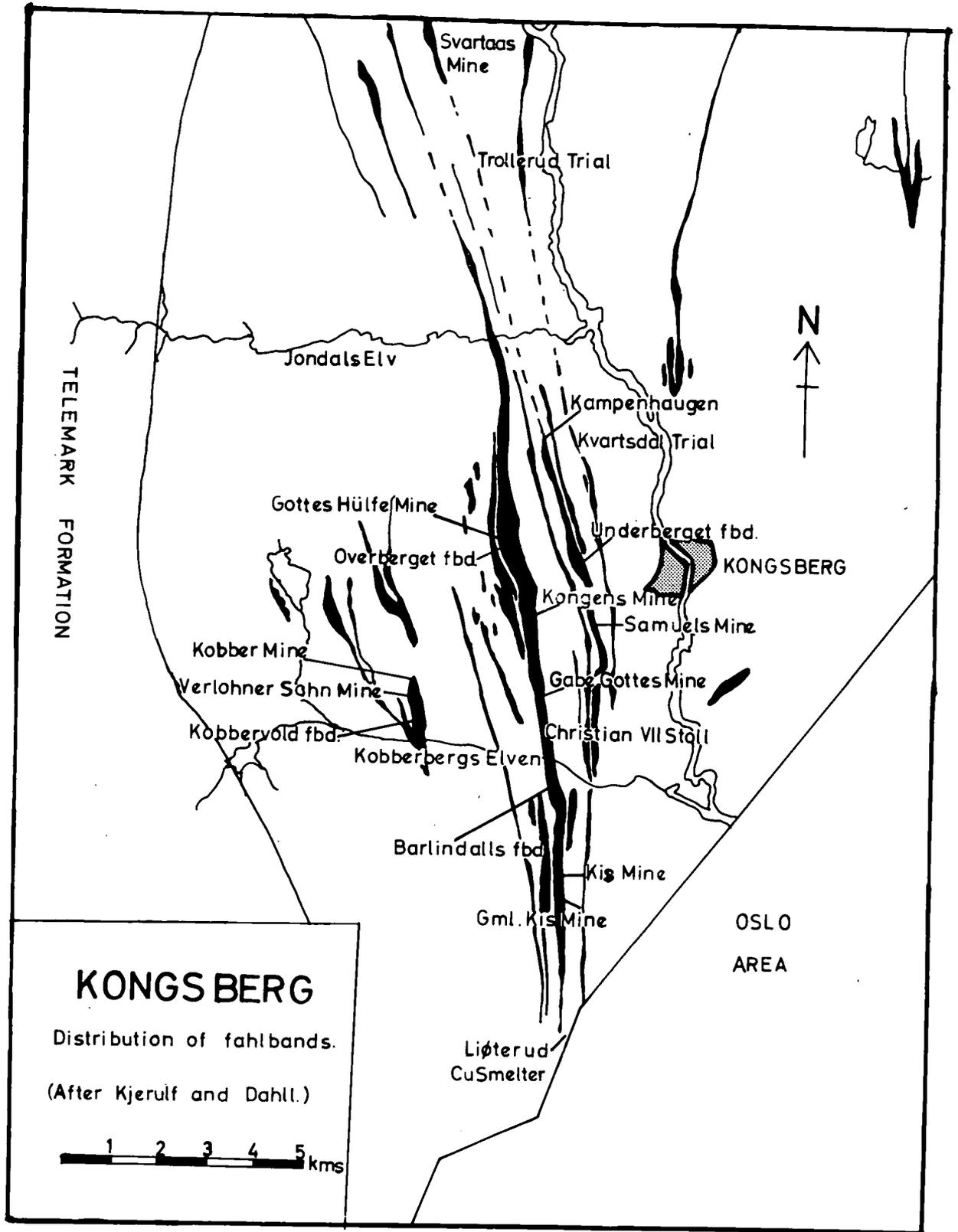


Fig. 60

The Overberget fahlband is the best developed in the region. It is widest at Gottes Hülfe Mine (300 m.), narrows to 120 m at Kongens Mine and is 100 m. wide at Christian VII Adit portal. It is poorly developed north of the Jondals River, but south of there it averages 4-5% sulphide minerals. The stretch from Gabe Gottes Mine to Christian VII Adit portal is particularly rich in sulphide lenses, older descriptions refer to the kisbaandet (sulphide band) which was well developed in Gabe Gottes Mine. The fahlband continues in a regular fashion south of the Kobberberg River where it becomes known as the Barliuddals fahlband.

The Barliuddals fahlband stretches south to the boundary with the Palaeozoic Oslo district rocks near Skrim. Nearly all the silver mines south of the Kobberbergs River are situated on this fahlband. It is particularly sulphide rich at Sandsvaer where a mine (Kisgruben) has been worked for copper from as early as 1490, some 150 years before silver was discovered in the area. The average ore grade was greater than 1.5% Cu (Bugge 1928) and consisted of impregnation ore with no sharp boundaries and occasional lenses of rich ore reaching 1 m. in length. The ore consisted of pyrite with between 3 and 5% chalcopyrite and very minor pyrrhotite. The Liöterud copper smelter was built to treat this ore.

Another fahlband which had workable ore lenses is the Kvartsdal fahlband which is poorly developed south of the Jonsdals River, but becomes very sulphide rich to the north of Trollerud. Trial workings were started at Trollerud and near the Anne Sophie Mine.

To the west of the silver mining area the Kobbervoldets fahlband contains two old copper mines (the Verlohner Sohn and Kobber Mines) near the Kobberbergs River, which derives its name from these deposits. These are stated to have had more than 1.5% copper (Bugge, op.cit.)

Reference to Figure 60 shows that there are many other fahlbands in the district, and in addition one finds many small sulphide lenses sporadically distributed in the area.

The sulphides

The sulphides are not restricted to any one rock type. They have been observed in biotite-schist, biotite-garnet-schist, chlorite-schist, chlorite-garnet-schist, quartz-felspar-biotite-gneiss, felspathic-quartzite and quartzite. In detail, the individual sulphidic bodies are in the form of many lenses lying parallel to the strike, any one fahlband zone being made up of hundreds of these small lenses. The fahlband zones also have the same dip as the enclosing gneisses (Figure 61).

In the central parts of the lenses sulphides form up to ca. 8% of the rock and pass, by gradual transition, into a rock virtually devoid of sulphides before the boundary of an adjacent lens is reached.

The sulphides are usually disseminated as individual crystals forming an integral part of the fabric of the host rock. Small aggregates of several grains are frequently observed, these invariably lie in the plane of the foliation. In thin section mica trains are seen to sweep around such small lenses of sulphide in the same manner as around garnet porphyroblasts (Figure 62).

Figure 61: East-west profile, across the regional strike, at the Kongens Mine, Kongsberg. (From Vogt, 1899b). The plane of the section is along the length of one of the cross-cutting calcite veins of the region. The positions of the workings shown on this profile illustrates how the workable concentrations of silver were restricted to those parts of the vein where the wallrock is fahlband.

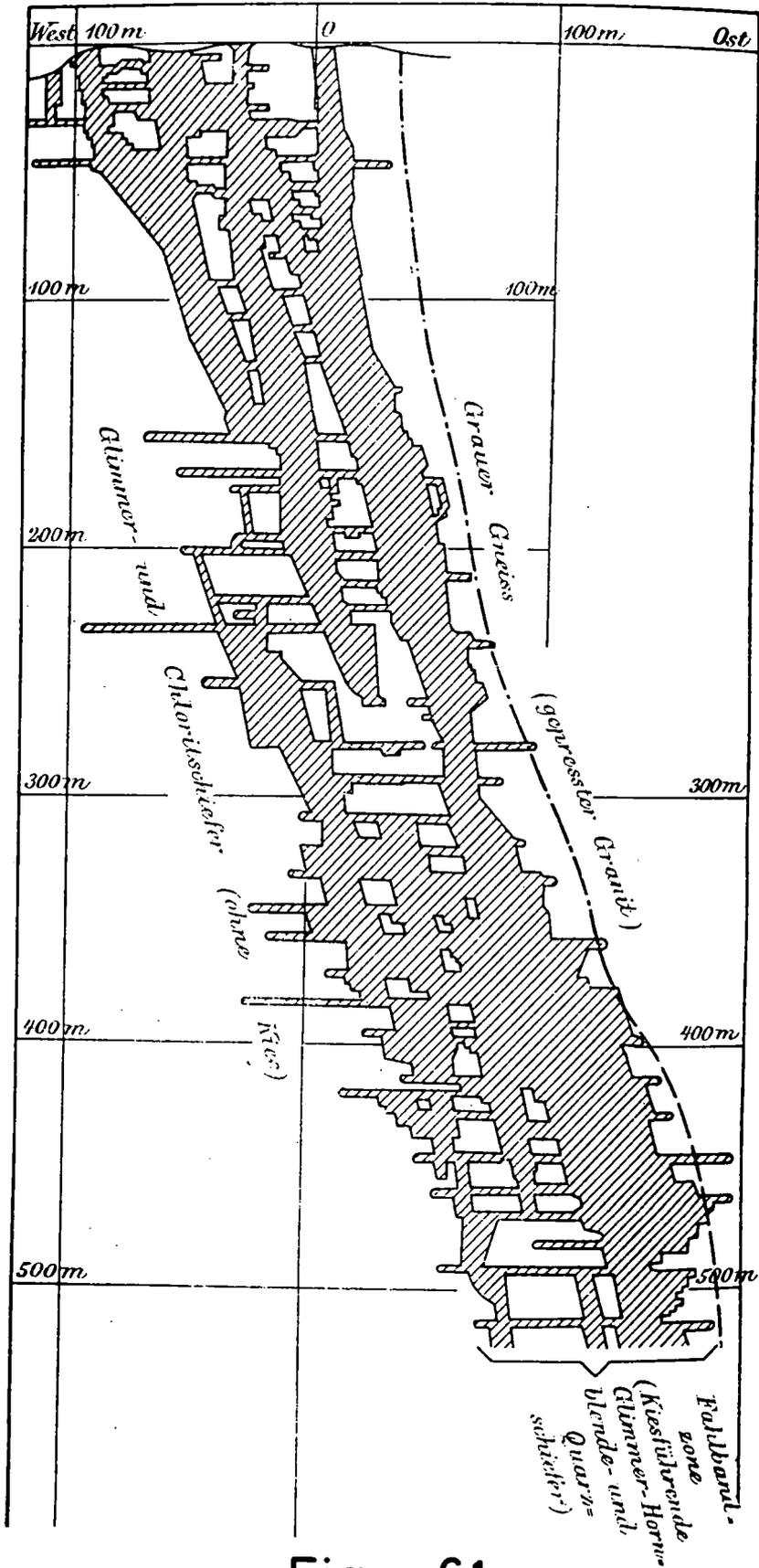


Fig. 61

Garnetiferous bands are frequently coincident with sulphide bands and the two minerals are observed in porphyroblastic intergrowth (Figure 63). The sulphides do not transect the foliation. In some cases thin (1 mm.) bands particularly rich in sulphide lenses were observed. Where these bands are thicker the lenses tend to join together to form a discontinuous layer parallel to the foliation (Figure 64).

In the more quartz rich rocks, small segregation lenses of quartz are developed in the plane of the foliation. These frequently contain sulphides of much larger grain size (up to 1 cm.) and occasionally, small sulphide veinlets are seen cutting across such lenses.

The specimens investigated episcopically have a rather monotonous mineralogy. Pyrite is ubiquitous and, in a majority of samples, is the only sulphide present. When present as disseminated grains it has developed euhedral (100) faces against the surrounding minerals (Figure 65). When it is present as small lenses and layers the crystals have mutually interfering boundaries, but exhibit a subhedral development against the adjacent silicates. The Kongsberg pyrite shows a very noticeable anomalous anisotropism in very pale shades of blue and brown. However, the polarisation figure obtainable showed no separation of the isogyres. The anisotropic effect was useful in textural studies as it was possible to differentiate between adjacent grains on the basis of their different optical orientation. No twinning was observed. Many of the pyrite crystals appear corroded by the gangue minerals (Figure 66). The pyrite intergrown with garnet (see above) is in optical continuity.

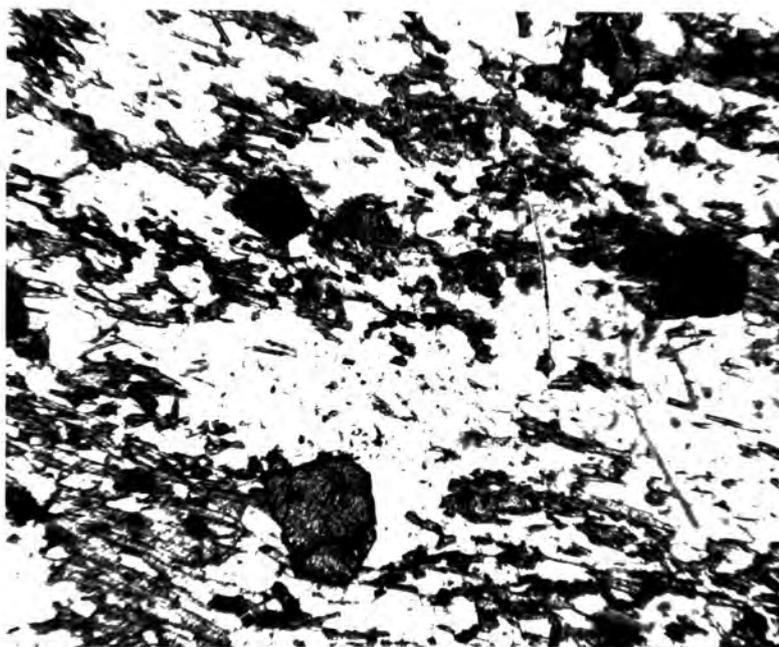


Figure 62: Pyrite porphyroblasts in garnet-mica schist ('fahlband').
Near Kongens Mine, Kongsberg.
Ordinary light. X 40.



Figure 63: Intergrowth of garnet and pyrite. Garnet-mica-schist,
Christian VII Adit portal, Kongsberg.
Ordinary reflected light. X 50.

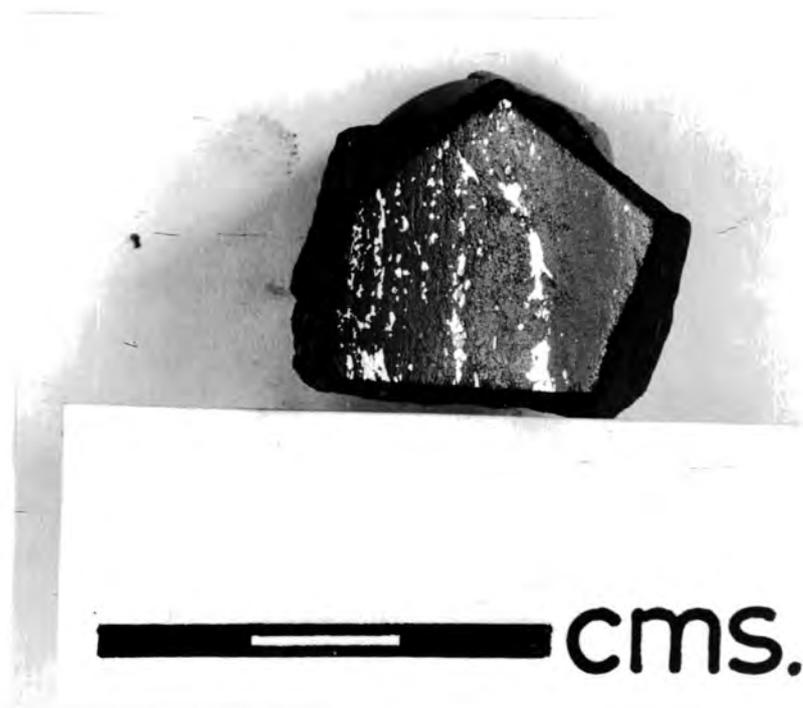


Figure 64: Lenses and discontinuous bands of pyrite developed parallel to the foliation of quartz-mica-schist, Christian VII Adit portal. Kongsberg.



Figure 65: Subhedral pyrite grains disseminated in 'fahlband' quartzite. Kongs Mine, Kongsberg.
Ordinary reflected light. X 50.



Figure 66: Pyrite corroded by gangue, 'fahlband' quartzite,
Kongens Mine, Kongsberg.

Ordinary reflected light.

X 50.

Chalcopyrite was observed in one sample from the entrance to Christian VII Adit. It was present in minor amounts at the grain boundaries of pyrite, and as a thin layer between the pyrite lenses and the surrounding silicates. Pyrrhotite was present in the same specimen as small lenses closely associated with the pyrite. It was badly altered.

Other minerals recorded in the literature but not observed by the author include arsenopyrite, sphalerite, galena, cobaltite, chalcocite, bornite and native copper. These are usually stated to be very rare and most have only been described from localities adjacent to the argentiferous silver veins, from which they may well have been derived. Ilmenite and magnetite are also recorded by many authors, these have only been observed during the present study in association with minor pyrrhotite in the amphibolitic fahler.

Economic importance

The Kongsberg fahlbands have themselves been of minor economic importance. They have been of interest because they provided an environment favouring precipitation of silver in the late veins. The so called 'fahlband rule' has been quoted from the very early days of the silver mining. It can be stated thus:

"The veins ~~are~~ silver-bearing only where they cross a fahlband." Note the spelling - fahlband, this refers only to those fahlbands of importance in localising the deposits (see p.141), thus restating the rule (after Kjerulf, 1865).

"The veins are silver-bearing only where they cross the wall rocks at a point where the vein is silver bearing."

It is, therefore, not surprising that the validity of the rule has not been challenged for three centuries!

What is clear is that all the major mines are situated on the intersections of veins with the Underberget and Overberget fahlbands, the many other fahlbands (s.s.) being relatively unimportant in this respect. Many attempts have been made to explain this association.

The enrichment has been variously ascribed to leaching of silver from the fahlbands themselves (Langberg 1853; 1863), low electrical resistance of the quartz-sulphide lenses in the fahlbands (Munster, 1894), liberation of H_2S from the fahlband pyrite causing precipitation of the silver as argentite (Vogt, 1899), electrolytic deposition of silver by stray earth currents using the fahlbands as electrodes (Munster, 1894), variation in the structure of the wall rocks (Holter, 1931), high thermal conductivity of the fahlbands causing a decrease of temperature in the hydrothermal fluid and consequent precipitation of silver (Neumann, 1944).

The work of Dadson (1937) on the Temiskaming silver ores is of interest in this connection. He arranged the minerals present into a potential series analogous to the potential series of the metals. Silver is quite low in this series and will replace those minerals higher in the series such as pyrite. Furthermore, the growth of the silver into cracks and fissures of the wall-rocks and gangue is explicable by the EMF effect. Silver could be deposited on any mineral acting as a positive pole to the mineral being dissolved by the silver bearing solution. A variant of this hypothesis has already been applied to the Kongsberg occurrence by Tronstad (1932, 1933).

The Modum fahlbands

Introduction

Fahlbands, as defined on page 144, are present in many parts of the Kongsberg - Bamble Formation. They are well developed in the Modum district where fahlband zones of up to 12 kms. strike distance, with a width of 100-200 m., are observed. As at Kongsberg these are of little intrinsic economic importance except where they are associated with workable deposits of other minerals. At Modum two of the fahlbands are closely associated with a cobalt mineralisation, which unlike the Kongsberg silver, is present in the form of impregnations rather than clearly defined veins. These two fahlbands zones have been referred to as cobalt-fahlbands.

Previous descriptions

The first mention of fahlbands at Modum is by Mückler (1799) who was in charge of the cobalt prospecting operations, directed from Kongsberg. Bøbert (1848) coined the term cobalt-fahlband in a paper describing the characteristics of fahlbands in general. He considered the cobalt minerals at Modum to be cogenetic with the sulphides and regarded the occurrence as an interesting variant of the normal fahlband type. In an earlier (1832) paper he had already described similar deposits from Sweden, which he now also classified as cobalt-fahlbands. All succeeding authors (see p. 7) have followed Bøbert in this terminology.

Field relations

The investigations connected with the present study, and the contemporary descriptions of the cobalt ore in the literature, indicate that this mineralisation is not restricted to the fahlbands alone. It is found well developed in pure quartzites containing none of the characteristic fahlband sulphides, it is also often well developed in the ortho-amphibolites and their associated biotite-schists. The distribution of cobalt is not as extensive along the strike as the fahlband, rather it is restricted to the vicinity of large basic intrusives. The majority of 'normal' fahlbands in the area contain no cobalt minerals. The present author therefore feels that it is premature to label Modum as a 'cobalt-fahlband deposit', and prefers to describe the two mineralisations separately.

In general field appearance the Modum and Kongsberg fahlbands are very similar. The extent of the sulphide impregnation is recognized by the 'fahl' weathering, which is very patchily developed within the zone, strongly mineralized rocks grading into barren in the space of a few metres. In general the fahlband zones follow the dip and strike of the Modum Formation rocks, but the lensoid development of the sulphides makes it impossible to trace a single sulphidic horizon for any distance. In the vicinity of Døvikollen the development of lenses, en echelon, in adjacent rock types gives the impression that the sulphidic zone is transgressing the foliation.

There are some major differences between the Kongsberg and Modum fahlbands which can be illustrated by a consideration of one of the main fahlband zones, traceable from south of the Mugerud

cobalt trail to north of the Heggebaek Mines - a distance of over 11 kms. Along this zone many different rock types of the Modum Formation are present and most of them contain more than accessory amounts of sulphide minerals. The zone is discontinuous and cannot be detected in the thickly vegetated cross-valleys cutting across the ridge, the apex of which is coincident with the fahlband zone. Van Autenboer (1957) attempted to detect the fahlband in these hollows by radiometric methods with no success.

Iron sulphide is again the most common mineral observed in the fahlband, but here chalcopyrite is much more abundant than at Kongsberg, it is present in only slightly smaller amounts than the iron sulphide. In the southern part of the zone pyrrhotite is the major iron sulphide, pyrite being very rare or absent, while in the north pyrite is common and the pyrrhotite very subordinate. Thus there appears to be a zonation from south to north.

In contrast to Kongsberg many of the fahlband zone rocks are highly graphitic. Graphite forms up to 5% of some of the rocks and is present in both micaceous and quartzitic types.

Shear zones are very commonly developed in the fahlband. These are clearly seen in some of the open-cast workings (Figure 67). The sheared rock is platy and schistose, the rupture and crushing has taken place in discontinuous patches and the intensity of shearing is very variable. The movement seems to have developed preferentially in the more micaceous and graphitic horizons.



Figure 67: Part of the opencast workings of the Central Mine, Skuterud, showing the development of a shear zone in one of the pillars left in the ore zone.

The intensity of the fahl weathering is largely controlled by a combination of the sulphide content and the ease of access of weathering agencies, controlled by the mica and graphite contents of the rock and the development of shearing. Thus sheared sulphide-graphite-schists are extremely deeply weathered and are friable in the hand, the weathering extends to a considerable depth, making the collection of fresh specimens extremely difficult. By contrast, unsheared sulphidic sillimanite-gneiss has only a very thin coating of weathering products.

Petrography

The sulphide mineralisation is well developed in many varying rock types. It is commonly restricted to the Modum Formation metasediments. Sulphides are found in the amphibolites, (p. 116) but they are very subordinate to the opaque oxides, furthermore they are present in amphibolites in the whole region not just those in the fahlband zones.

Where the fahlband is well developed the only rock which has escaped impregnation is the quartzite, all the others contain sulphides to varying degrees. The petrography of these varying rock types has already been described (pp. 20 - 41). The main features observed in thin-sections of the fahlband rocks are the same as described from these, apart from the development of the sulphides. Texturally the sulphides appear to be the latest phase present (Figure 9), they are frequently associated with mica, replacing it along the basal cleavage. The relationship between the colour of the biotite and the amount of sulphide present has already been commented upon (p. 83).

The most frequently observed rock type in the fahlband zone is a quartz-felspar-phlogopite-sulphide-schist with varying amounts of graphite.

The sulphides

Texture In the siliceous granulites the sulphides are evenly distributed through the fabric of the rock. The disseminated sulphides occur either as isolated grains or as small aggregates of two or three intergrown grains, they are invariably elongated parallel to the foliation.

Small lenses and stringers are sometimes developed with a preferred orientation parallel to the foliation. Such sulphide concentrations are frequently associated with minor quartz-segregation lenses or with the more micaceous bands present. Very rarely small stringers of sulphide are observed cutting across the foliation at a low angle. Sulphides are also observed concentrated in the crests of minor folds.

At the south of the main fahlband zone, where pyrrhotite is the major mineral, rich sulphide lenses are sporadically developed. These may attain a size of 5-6 cms. maximum dimension and frequently contain intergrown chalcopyrite. At the northern end of this fahlband, where pyrite is better developed, definite pyritous horizons are clearly traceable for ca. 100 m. along the strike. The pyrite in these layers can form up to 60% of the rock, the layers are very graphitic and rarely exceed 10 cms. in width.

Mineralogy The following sulphide minerals have been identified in the fahlbands of the Modum district.

Main constituents : Pyrrhotite, pyrite, chalcopyrite

Minor constituents: Marcasite, covellite

Pyrrhotite: The pyrrhotite usually occurs in small, alotriomorphic aggregates consisting of a mosaic of sub-rounded or polygonal grains varying in size from 0.2-5 mm. In the disseminated type of occurrence these aggregates consist of a maximum of 3 or 4 individual grains (Figure 68). In the richer sulphide lenses the pyrrhotite attains its maximum grains size of ca. 5 mm.

There is a very frequent association of pyrrhotite with chalcopyrite in the Modum fahlbands (Figures 69, 70). The textures of these two minerals have proved very inconclusive with regard to age relations. In some specimens the pyrrhotite appears earlier than the chalcopyrite, in others the reverse relationship is observed. In the absence of conclusive evidence the two sulphides are here regarded as essentially contemporaneous. Schwartz (1937) has commented on the "almost universal association" of chalcopyrite with pyrrhotite, this association being mentioned in 79 of the 86 deposits he reviewed.

In chalcopyrite rich specimens the pyrrhotite occurs as more or less rounded grains or aggregates in a smooth, flat field of chalcopyrite. While in the pyrrhotite rich specimens the chalcopyrite appears as a matrix between the pyrrhotite grains. The boundaries are always smooth, even under high magnifications. They are probably good examples of 'mutual boundaries' indicating simultaneous crystallisation.

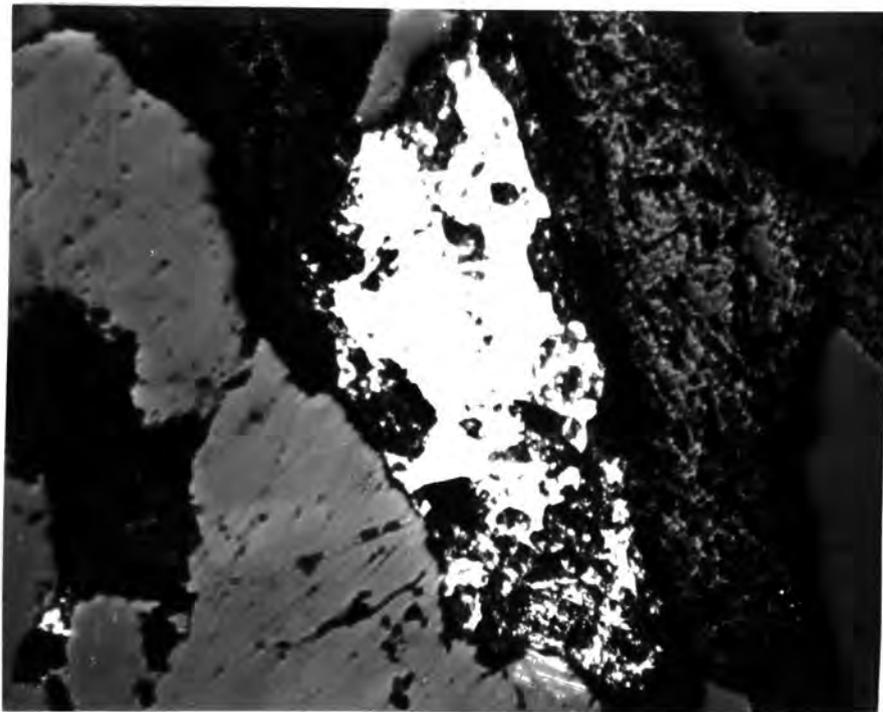


Figure 68: Pyrrhotite aggregates disseminated in sulphidic siliceous granulite. Central Mine, Skuterud.

Ordinary reflected light.

X 50.



Figure 69: Chalcopyrite-pyrrhotite aggregate. Siliceous granulite, Central Mine, Skuterud.

Ordinary reflected light.

X 50.



Figure 70: Chalcopyrite-pyrrhotite aggregate, showing rim of pyrite developed on pyrrhotite as a late alteration.

Ordinary reflected light.

X 50.

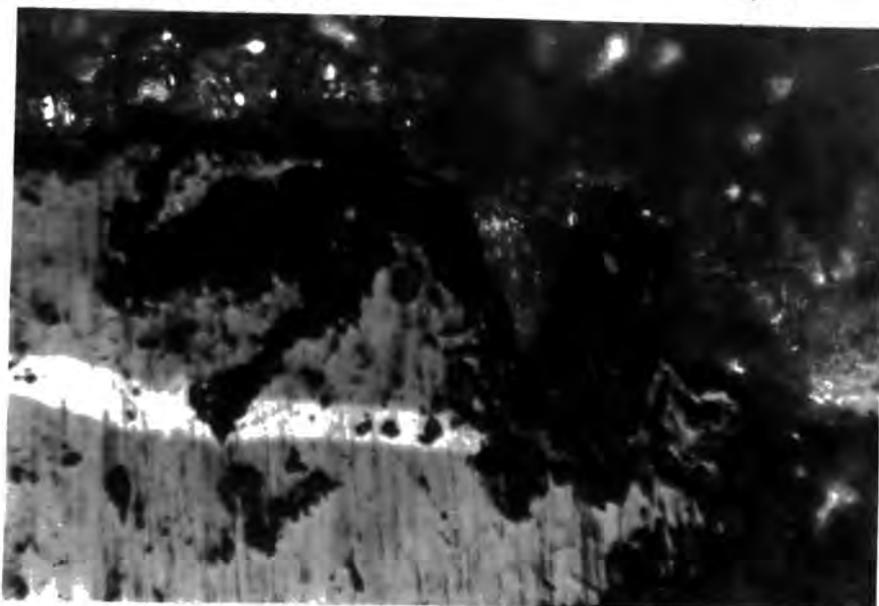


Figure 71: Twinning in pyrrhotite. Sulphidic siliceous granulite, Central Mine, Skuterud.

Reflected light, crossed nicols.

X 50.

The pyrrhotite-chalcopyrite aggregates are elongated parallel to the foliation and appear to be replacing the silicates. Irregular penetration and partial digestion of silicates is observed at grain boundaries. (Figure 70).

The larger aggregates of pyrrhotite have well developed twinning (Figure 71). According to the Schneiderhöhn and Ramdohr (1934) twinning is rare and unimportant and can almost universally be attributed to stress. However the textures of the sulphides at Modum indicate that they probably recrystallised under the influence of the main metamorphism. The twinning possibly resulted from later deformation phases. That the pyrrhotite has been cataclastically deformed subsequent to the twin formation is demonstrated by the gangue mineral along a wedge shaped crack, on either side of this crack the twins have a 12° difference in extinction position. The angular difference between opposite faces of the fracture is also 12° .

In some of the larger grains of pyrrhotite small lamellae of another phase are present (Figure 72). These are present as small sinuous lenses of variable width (0.008-0.015 mm.), always oriented parallel to a common direction. They have never been observed following more than one direction. They are best seen with crossed polars when difference in anisotropism from the host pyrrhotite make them clearly visible. They are barely discernible in ordinary lights, under oil immersion slight differences in pleochroism facilitate the distinction between the lamellae and the host. They are slightly darker in colour than the host pyrrhotite, the Becke line test indicates that they are slightly softer, they are not as anisotropic as the host pyrrhotite, varying in colour between dark and pale brown.

Similar lamellae have been described by many authors. Among those who have speculated on their origin may be mentioned Van der Veen (1925), Scheiderhöhn and Ramdohr (1934), Ramdohr (1950), Scholtz (1936) and Vokes (1957). The properties of the lamellae in the Modum pyrrhotite are closest to those described by Vokes (op.cit) from Birtavarre, Norway. However comparison of polished specimens from the two areas shows the lamellae in the Birtavarre pyrrhotite to be much better developed, they are clearly discernible without the aid of the analyser.

Ramdohr (1950) suggests that the lamellae may be due to segregation of FeS from iron sulphide containing a slight excess of sulphur. Such an excess is often revealed in pyrrhotite analyses. Similar lamellae have been observed during the experimental work of Kullerud and Yoder (1959) and Arnold (1962) on pyrrhotite. The lamellae they observed are brighter, softer and more anisotropic than the Modum and Birtaverre types. They are seen oriented in as many as four directions in the pyrrhotite host. They only develop if the pyrrhotite contains less than 45.6 atomic percent Fe and is rapidly quenched. From the solvus in the pyrrhotite-pyrite binary system (Arnold, op. cit) this corresponds to quenching from a temperature in excess of 666^o C. Annealing at temperatures less than this caused the lamellae to disappear with the consequent unmixing of rounded, scattered grains of pyrite. X-ray analysis (Arnold op.cit.) indicates that the lamellae are sulphur rich. Birchenall (vide Arnold op.cit.) suggests that the lamellar phase is metastable and develops by a martensitic (displacive) mechanism during quenching in response to high supersaturation of sulphur.

The implications of these experimental data are interesting. If the lamellar phase in the Modum pyrrhotite is the same as that described by Arnold then it would imply that the sulphides have cooled from a temperature in excess of 666°C . On geological grounds such a high temperature seems unlikely but is still possible within the estimated limits of the upper amphibolite facies (Turner and Verhoogen, 1960, p. 553).

The lamellae described in the literature have different properties to the synthetic type described above. Vokes (op.cit.) tabulates the results of four different descriptions of the lamellar phase in natural pyrrhotites, these differ from each other and from the synthetic type in colour, hardness and host-daughter relationships. Grönvold and Haraldson (1952) described various inversions in pyrrhotite occurring below 325°C . in which monoclinic pyrrhotites and pyrrhotites with hexagonal super-lattices were encountered. Byström (1945) described natural Swedish pyrrhotites which contained a mixture of hexagonal and monoclinic structures, he did not examine the specimens episcopically, Buerger (1947) investigated pyrrhotite single crystals and found them to possess a mixture of hexagonal and lower symmetry structures. Byström (op.cit.) suggested that the change in structure was due to a polymorphic transformation. Grönvold and Haraldson (op.cit.) conclude that the monoclinic phase has a greater sulphur content than the hexagonal phase and has "either been formed at low temperatures or has cooled down slowly so that the transformation from the B-8 (hexagonal) structure, stable at higher temperatures, to the low symmetric structure has taken place."

Arnold explicitly states that pyrite-pyrrhotite relations were only studied at temperatures in excess of 325°C 'to avoid the complexities involved in the various inversions' encountered below that temperature (op.cit. p. 74). It is here suggested that the lamellae observed in the Modum and Birtavarre pyrrhotites are the result of the polymorphic transformation of sulphur rich hexagonal pyrrhotite to one of the monoclinic phases which develop below 325°C .

A distinctly different lamellar phase is also present in some of the Modum pyrrhotite. This consists of very small lenses of a mineral with high reflectivity, pale yellow colour, hardness greater than pyrrhotite, which is isotropic under crossed polars. The lenses are developed parallel to the (0001) direction of the pyrrhotite host in a very regular fashion (Figure 73). Where the pyrrhotite is weathered the lamellae appear unaffected and are observable in the alteration products (Figure 74). The characteristics of the mineral suggest that it is either pyrite or pentlandite.

A determination of the micro-hardness would have differentiated between these minerals but no area large enough for a suitable test was present. The X-ray fluorescence spectrographic results (p. 264) testify to the extremely low nickel content of the Modum fahlbands. Many analyses were made of the pyrrhotite concentrate during the working period of the mines in the hope that nickel would be obtainable from the otherwise useless pyrrhotite. All the tests were negative. Microchemical tests for nickel (Short, 1948) were performed on the polished specimens with negative results.

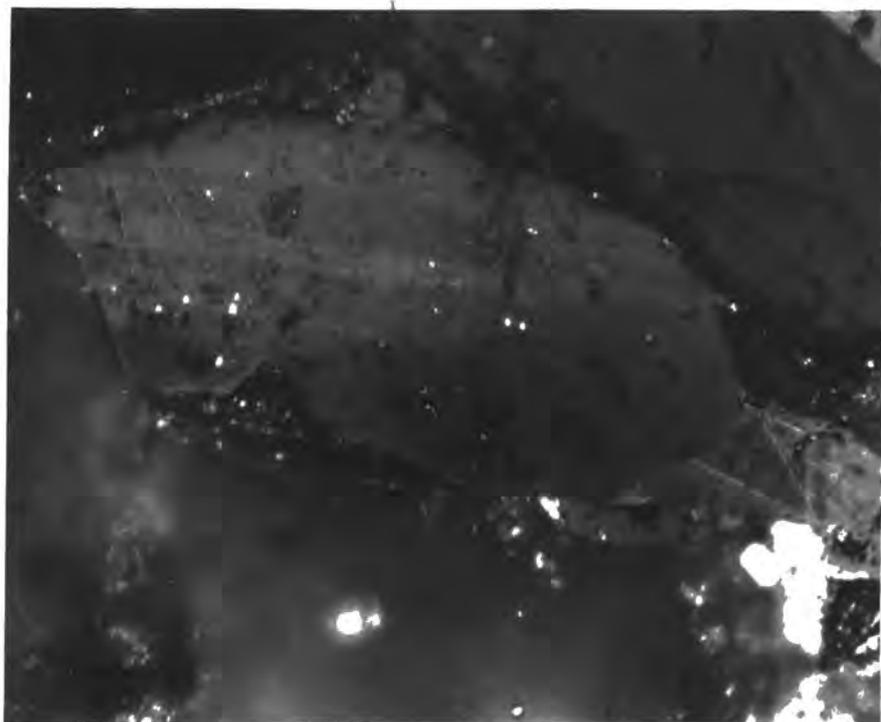


Figure 72: Lamellar pyrrhotite. Central Mine, Skuterud.
Reflected light, crossed nicols. X 50.

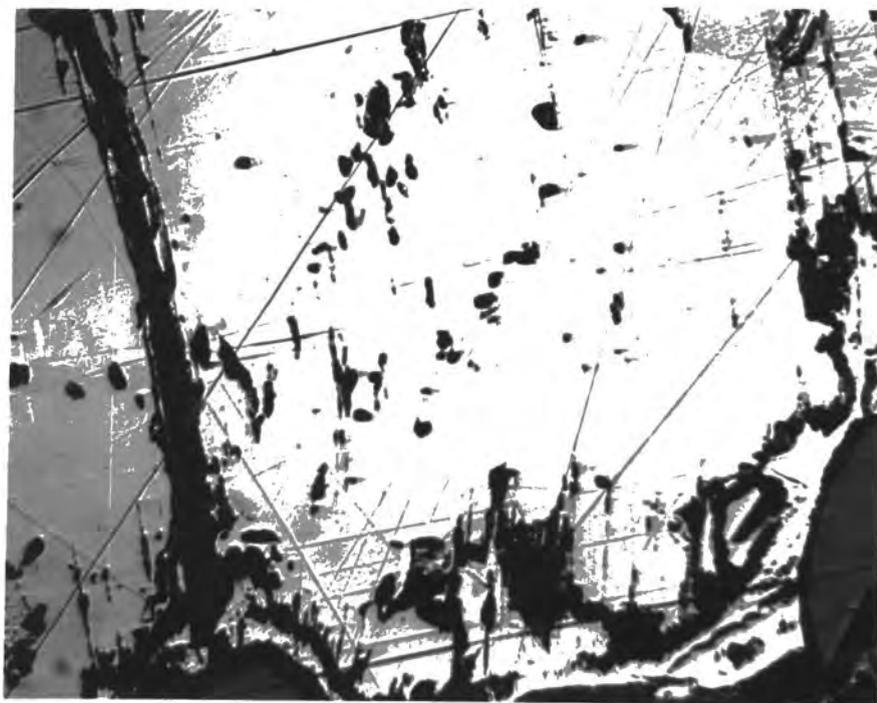


Figure 73: Small lenses of pyrite developed in pyrrhotite parallel to the (0001) direction. Dumps at Bergan Pukkverk, Skuterud.
Ordinary reflected light. X 50.

Thus on the basis of the available evidence it is suggested that these lenses are of pyrite rather than pentlandite. Arnold (1962, Figure 5) illustrates a very similar texture produced by quenching sulphur rich synthetic pyrrhotite in the laboratory. He obtained pyrite plates which were oriented parallel to each other in the pyrrhotite host. He was unable to produce similar textures with natural pyrrhotite of suitable composition.

The Modum pyrrhotite has been examined on the X-ray diffractometer, the pattern produced corresponds to the normal hexagonal type of pyrrhotite structure. Pyrrhotite was separated from rocks in which it was observed in co-existence with pyrite, the lattice spacings were determined for the application of the pyrrhotite geothermometer. (Arnold, 1956, 1957, 1958). By this method the bulk composition of hexagonal pyrrhotite can be determined by measuring the value of the d_{102} reflection in Angstroms. The temperatures of formation indicated by the composition of pyrrhotite in association with pyrite from the solvus for the FeS-FeS₂ system were as follows:

Svartefjell Mine (north end of main fahlband zone)

$$d_{102} = 2.061; \text{ Fe} = 46.9 \text{ at } \%; \text{ T} = 425^{\circ}\text{C}.$$

Central Mine, Skuterud (southern end of main fahlband zone)

$$d_{102} = 2.059; \text{ Fe} = 46.7 \text{ at } \%; \text{ T} = 470^{\circ}\text{C}.$$

The possibility that trace elements are present in excess of the 0.06 wt.% limit imposed by Arnold and Reicher (1962) cannot be excluded. The presence of the possible exsolved pyrite lamellae mentioned above may mean that the pyrrhotite composition has changed since equilibrium was attained. Thus the temperature measured is lower than the true temperature of deposition and can only be

regarded as a minimum value (Arnold, 1962, p. 87).

Estimated temperatures for the upper amphibolite facies of 550-750°C have been obtained (Turner and Verhoogen, 1960, p. 553). If the temperature derived from the pyrrhotite geothermometer is regarded as lower than the true temperature of formation then the results appear consistent with the geological data. The close correspondence of the two temperatures indicates that the pyrrhotite finally equilibrated during a retrograde stage.

The pyrrhotite displays fine examples of a later alteration involving the formation of both marcasite and pyrite. This alteration is most conveniently discussed here. The alteration is proceeding from cracks, en-echelon fractures and grain boundaries to attack the pyrrhotite along the (0001) cleavage planes (Figures 75, 76). The alteration faithfully follows the cleavage directions and is seen to change direction across twin planes (Figure 77). In some zones the replacing sulphide is compact, but in others it is spongy and easily removed by the polishing process.

The spongy, yellow replacing sulphide is harder than the pyrrhotite and is strongly anisotropic from bright yellow to blue grey. Similar alteration products have been described in the literature and usually referred to marcasite. Ramdohr (1950, p. 606) considers the mineral to differ slightly from true marcasite and prefers to call it "marcasite".

Between the marcasite and the pyrrhotite there is frequently a grey zone of minerals referred to as Zwischenprodukte by Ramdohr (op.cit. Figures 373, 376a). Attempts were made to drill out some of this material for X-ray determination. The resulting pattern was confused by pyrrhotite, marcasite and pyrite lines and no definite



Figure 74: Small lenses of pyrite developed in pyrrhotite continuing uninterrupted into the alteration zones.

Ordinary reflected light.

X 50.



Figure 75: Alteration of pyrrhotite proceeding from cracks and grain boundaries. Dumps at Bergan Pukkverk, Skuterud.

Ordinary reflected light.

X 50.

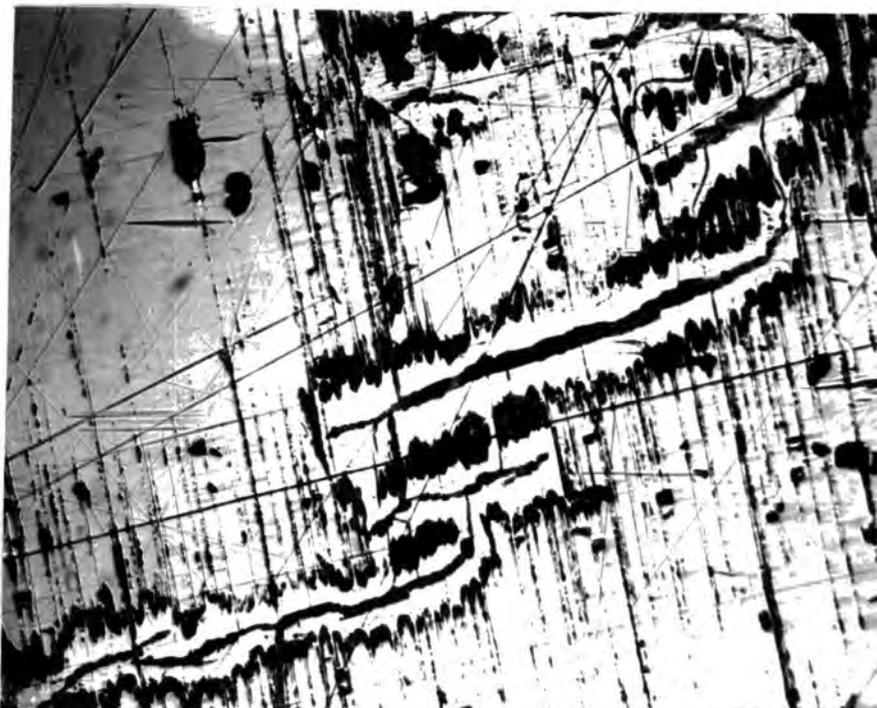


Figure 76: Alteration of pyrrhotite proceeding from en-echelon fractures. Dumps at Bergan Pukkverk, Skuterud. Ordinary reflected light.

X 50.

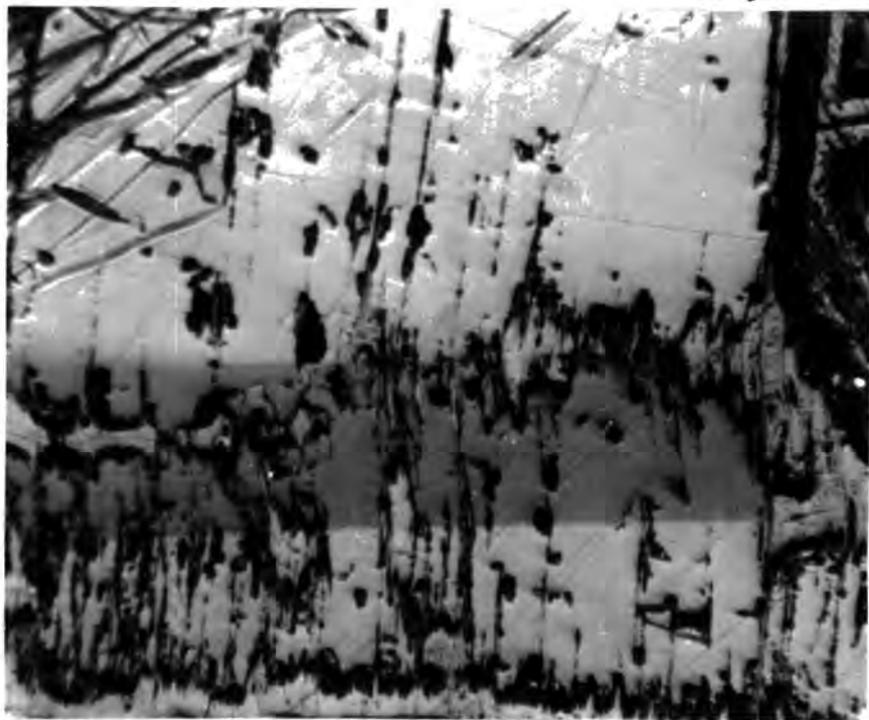


Figure 77: Alteration of pyrrhotite changing direction over twin composition plane. Dumps at Bergan Pukkverk, Skuterud. Reflected light, crossed nicols.

X 50.

conclusions could be reached. It is possible to see in some places that the grey zone is divisible into two parts. That closer to the pyrrhotite being brighter and anisotropic in blue-grey with wine-red internal reflections, while the zone adjacent to the marcasite is darker grey with no noticeable anisotropism and poorer internal reflections. Foslie (1950) and Marmo (1953) have observed similar 'double garlands of marginal zones' and consider the lighter one to be goethite and the darker limonite.

Pyrite has developed in patches in the pyrrhotite, frequently surrounded by a thin layer of chalcopyrite and separated from the fresh pyrrhotite by a zone of marcasite alteration (Figure 78). This texture contrasts strongly with the euhedral pyrite crystals (of an earlier generation?) embedded in a pyrrhotite matrix with no alteration zone separating them (Figure 79).

Fine grained pyrite is also associated with the marcasite in the alteration zones. It is possible to see that the individual crystals of both marcasite and pyrite are elongated normal to the joint surfaces along which the alteration is taking place.

A rib of pyrite develops at both sides of the original crack and grain borders and stands out sharply as thin bright lines on polishing (Figure 70). Aggregates of fine grained pyrite crystals are observed bordering the walls of joints from which alteration has proceeded. These have the appearance of transgressing the earlier marcasite alteration product (Figure 80). Similar veins of chalcopyrite are seen (Figure 81).

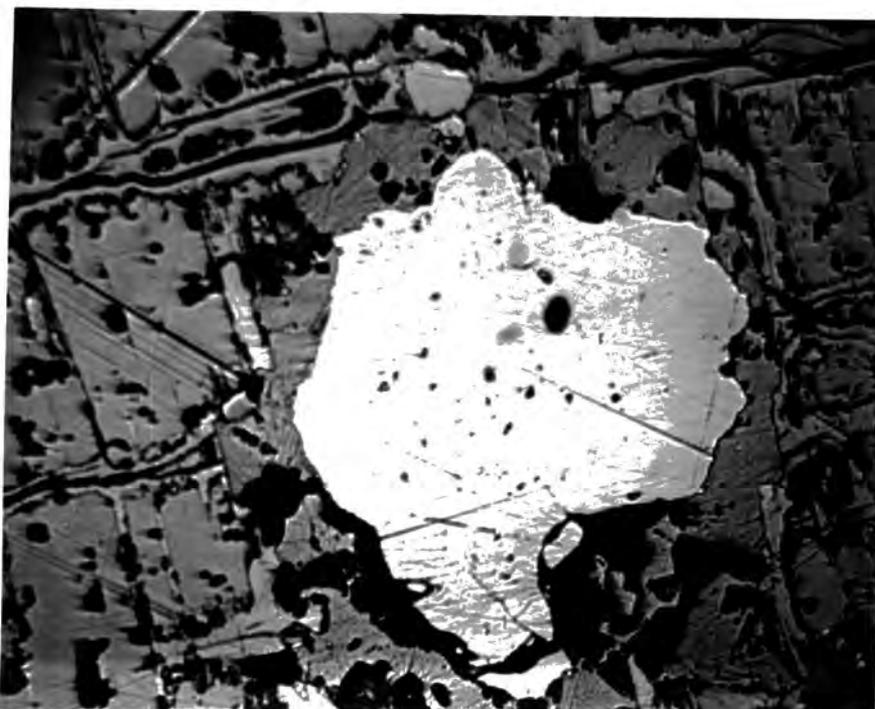


Figure 78: "Birds eye texture" in pyrrhotite. Pyrite surrounded by a rim of chalcopyrite, separated from the fresh pyrrhotite by a zone of alteration. Ordinary reflected light. X 50.

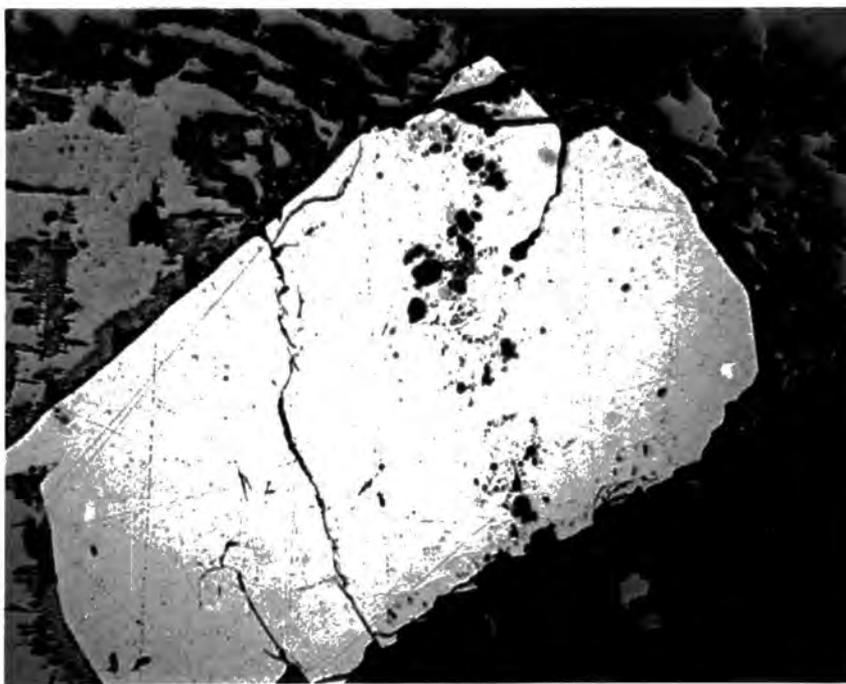


Figure 79: Euhedral pyrite enclosed by pyrrhotite with narrow zone of alteration between the two minerals. Ordinary reflected light. X 50.

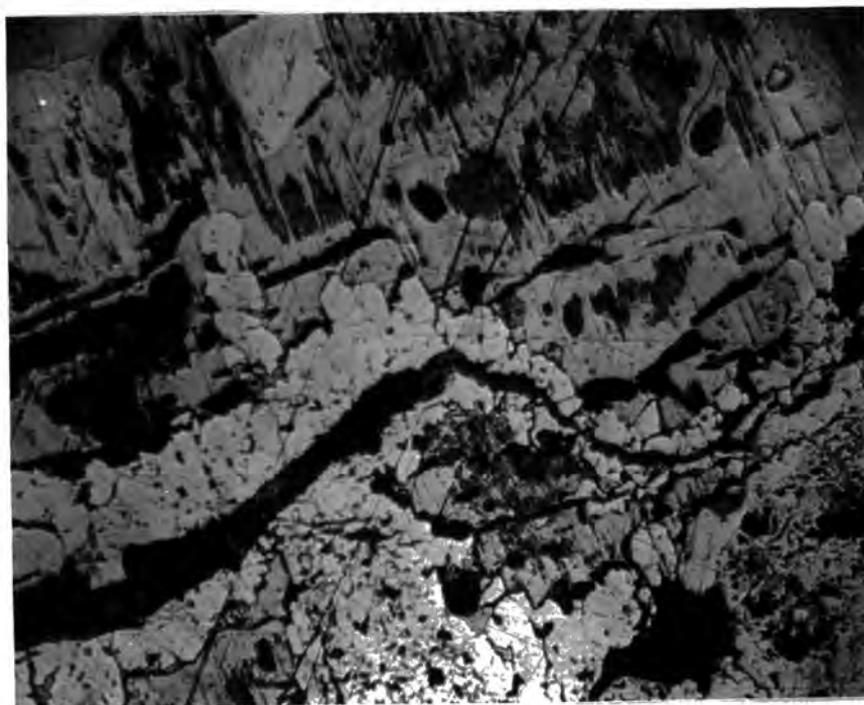


Figure 80: Pyrite developed at the borders of a veinlet transgressing the earlier alteration zone in pyrrhotite.

Ordinary reflected light.

X 50.

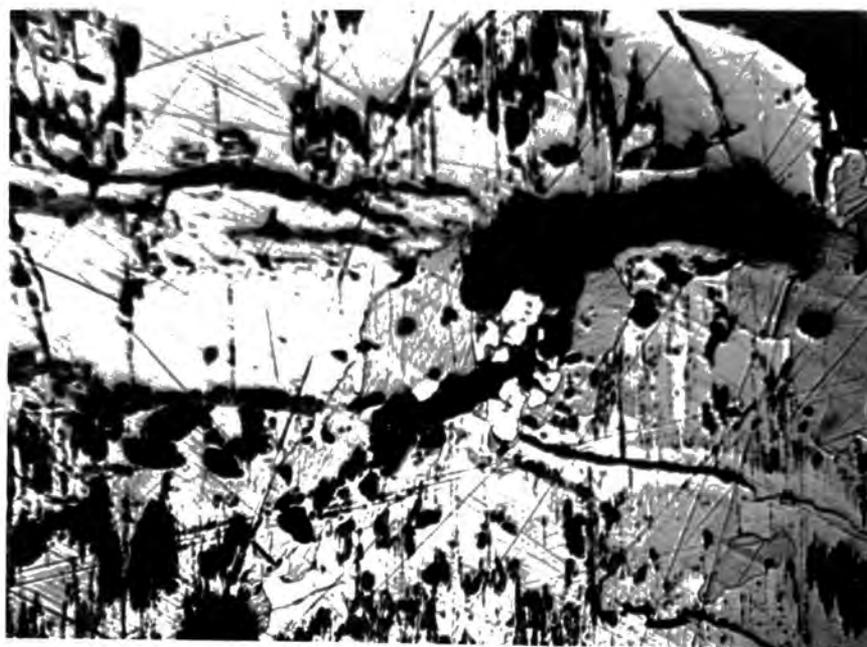


Figure 81: Chalcopyrite developed at the borders of a veinlet transgressing the earlier alteration zone in pyrrhotite.

Ordinary reflected light.

X 50.

There has been some controversy regarding the origin of this alteration in the literature. Ramdohr (1950, p. 412) attributes it to supergene agencies. Edwards (1954) considers that it is due to "changes in the acidity and temperature of the mineralising solutions" and coincides with hypogene carbonate deposition. Wager Vincent and Smales (1957) consider that similar alteration described from the Skaergaard intrusion is not related to weathering, but took place at a late stage in the cooling of the rock.

Marmo (1953) described textures very similar to those observed at Modum from sulphidic-graphite schists in Finland. He postulates that circulating ground waters rich in oxygen and carbon dioxide cause the alteration of pyrrhotite to limonite along cracks and grain boundaries, thus producing the fahl weathering colours. This reaction involves the liberation of H_2S into the groundwater which at lower depths becomes oxygen deficient. This water causes the production of the supergene sulphides by reaction with the earlier formed limonite. Variations in acidity control the relative amounts of pyrite and marcasite produced.

This mechanism is a possibility for the Modum fahlbands. The late pyrite and chalcopyrite veins described above could be attributed to a later alteration effected by groundwater of different pH and composition to that which produced the supergene marcasite.

Pyrite Pyrite is commonest in the fahlbands in the north of the Modum district, where it is very much predominant over pyrrhotite. However it is present in small amounts in the fahlbands of the whole

region, usually as disseminated crystals having a subhedral outline. The individual crystals appear to be heavily corroded by the gangue minerals (Figure 82).

It is also present as concordant aggregates (Figure 83), or as concentrations in discontinuous bands the thickness of which varies from a few millimetres to several centimetres. The pyrite develops euhedral faces at its contact with the surrounding silicates (Figures 83, 84).

It is rarely found in fissure fillings, cutting the plane of the schistosity in varying directions, in association with quartz. The pyrite in these quartz veinlets is euhedral.

Several generations of pyrite are present. The earliest generation has euhedral crystals with the form (100) predominating, these are seen embedded in a matrix of pyrrhotite and chalcopyrite, (Figure 79). Similar crystals very full of inclusions are seen surrounded by a later anhedral aggregate of pyrite free of inclusions in some of the pyritic lenses (Figure 85). Zones of inclusions are seen in some of the grains which are parallel to the crystal boundaries.

Anomalous anisotropism is very marked and individual grains are easily differentiated with crossed nicols. Penetration twins can sometimes be seen (Figure 86).

Anhedral pyrite is seen as a matrix between silicate grains (Figure 87). This is presumably of a later generation than the disseminated euhedral type. The latest pyrite is that associated with the alteration of pyrrhotite (p. 177).



Figure 82: Pyrite corroded by silicate gangue. Svartefjell Mine,
Ordinary reflected light. X 50.



Figure 83: Concordant aggregate of pyrite in sulphide-graphite schist.
Svartefjell Mine.
Ordinary reflected light. X 50.

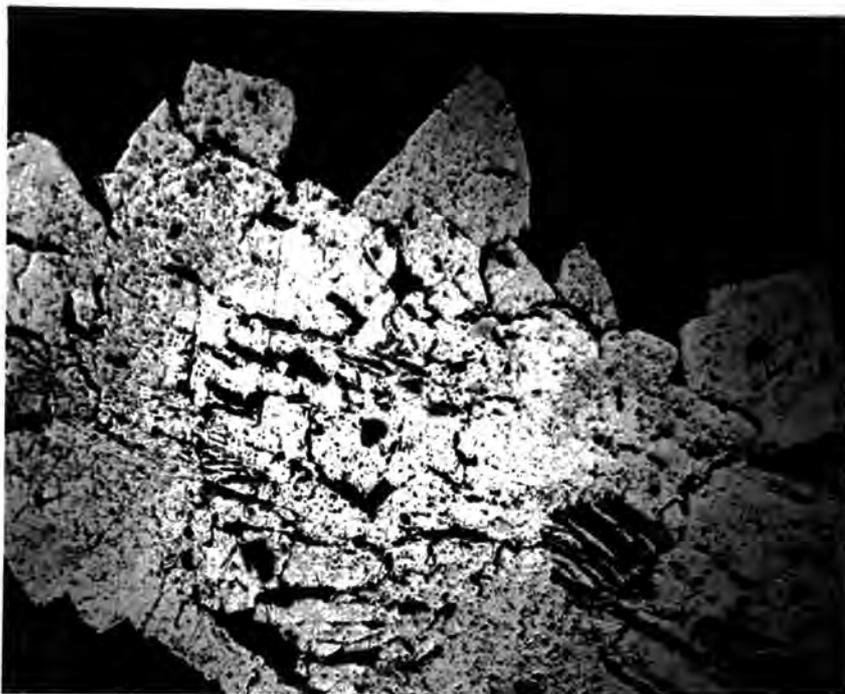


Figure 84: Development of (100) faces in pyrite in contact with silicates.
Svartefjell Mine.

Ordinary reflected light.

X 50.

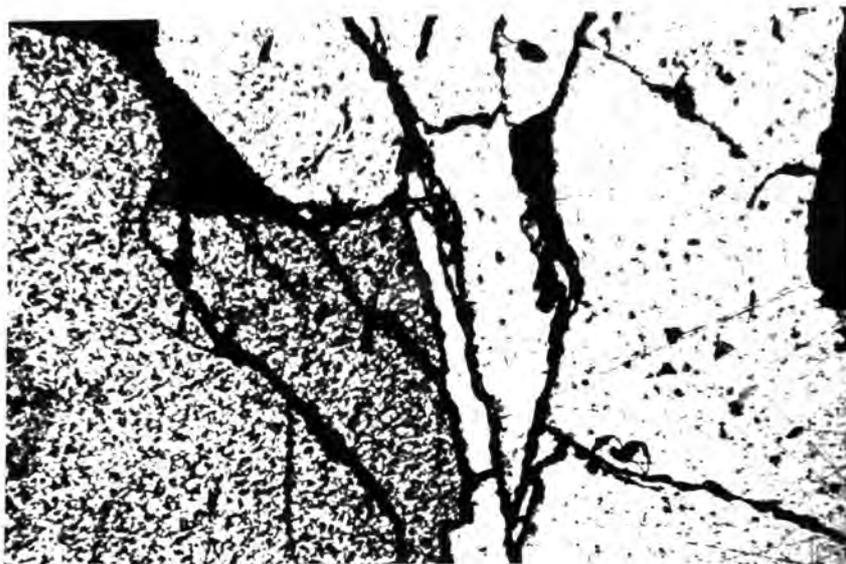


Figure: 85: Pyrite crystal very full of silicate gangue inclusions
surrounded by an anedral aggregate of pyrite free from inclusions.
Svartefjell Mine.

Ordinary reflected light.

X 50.

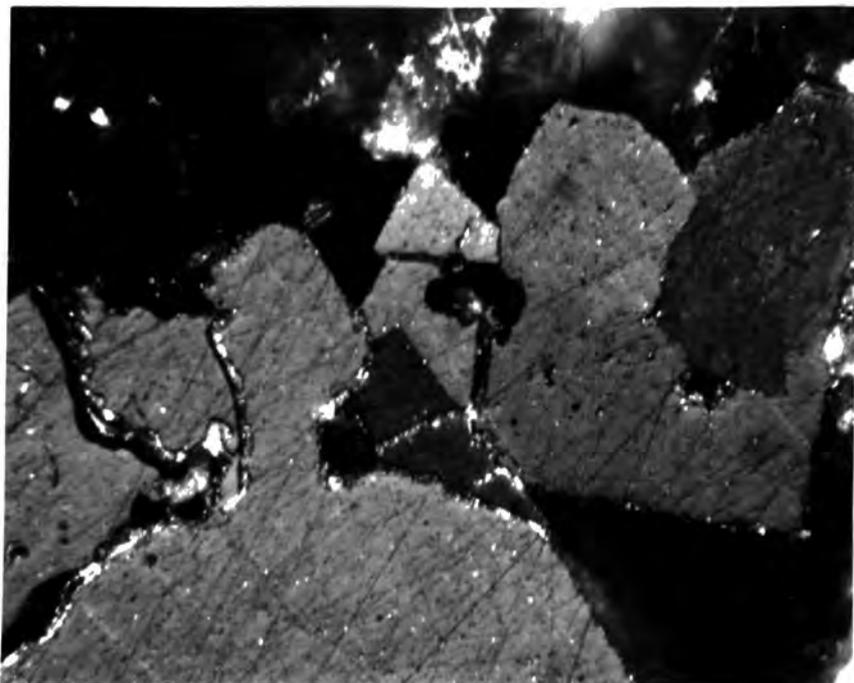


Figure 86: Penetration twins in pyrite. Sulphide-graphite schist, Svartefjell Mine.

Reflected light, crossed nicols.

X 50.

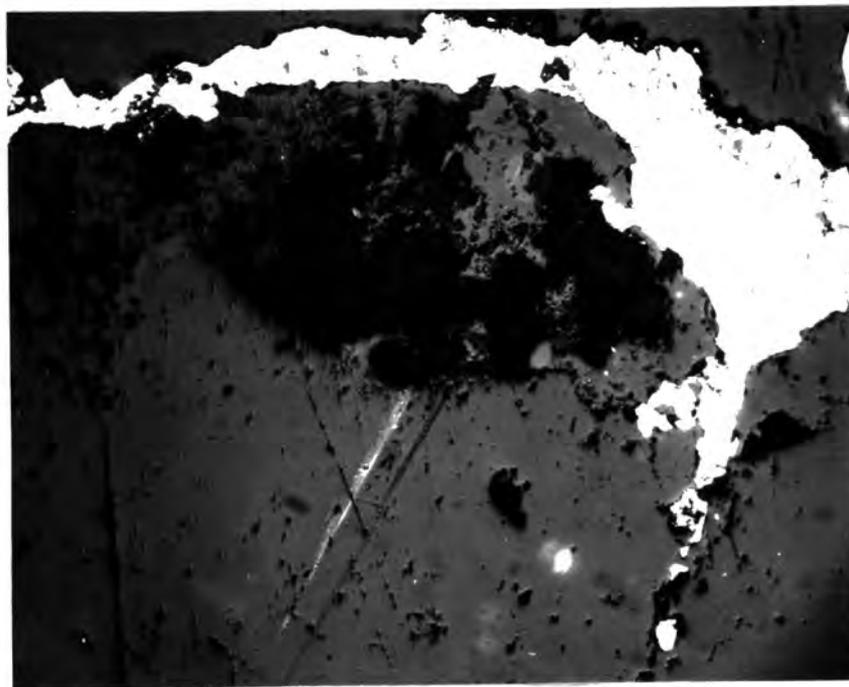


Figure 87: Small veinlets of pyrite between silicate grains. Sulphide-graphite schist, Forhaabning Adit, Skuterud.

Ordinary reflected light.

X 50.

Supergene alteration has also affected the pyrite, but this is less marked than that observed in the pyrrhotite. Crystals are altered along cataclastic cracks to goethite (Figure 88). The pyrite interstitial to silicates has been particularly prone to alteration, minor pyrite residuals being embedded in veins of the goethite (Figure 89).

Chalcopyrite Chalcopyrite is common in all the fahlband rocks of the Modum area, it is usually subordinate to the iron sulphides, but in the vicinity of the Central Mine at Skuterud it is the commonest fahlband sulphide. No euhedral crystals of chalcopyrite have been observed in the fahlbands. It is often present as a matrix between silicate minerals or as plates controlled by the silicate cleavage (Figure 90). It is very frequently intimately associated with pyrrhotite (see p. 165). Chalcopyrite rich areas in polished specimen have a flat even field in which it is difficult to distinguish individual grains.

That chalcopyrite was present during at least some of the deformation stages affecting the rocks is illustrated by Figure 91 where graphite flakes are seen to be moulded around chalcopyrite grains. However it appears to have been mobile at rather a late stage as it is seen penetrating pyrrhotite, pyrite and silicates in small veinlets. Very late-stage chalcopyrite is probably represented by the rims and veinlets associated with the alteration of pyrrhotite.

The chalcopyrite shows a good anisotropism with the occasional development of parallel twin lamellae. No cubanite was observed.



Figure 88: Pyrite crystal altered to goethite along cataclastic cracks.
Weathered sulphidic siliceous granulite, Central Mine, Skuterud.
Ordinary reflected light. X 50.

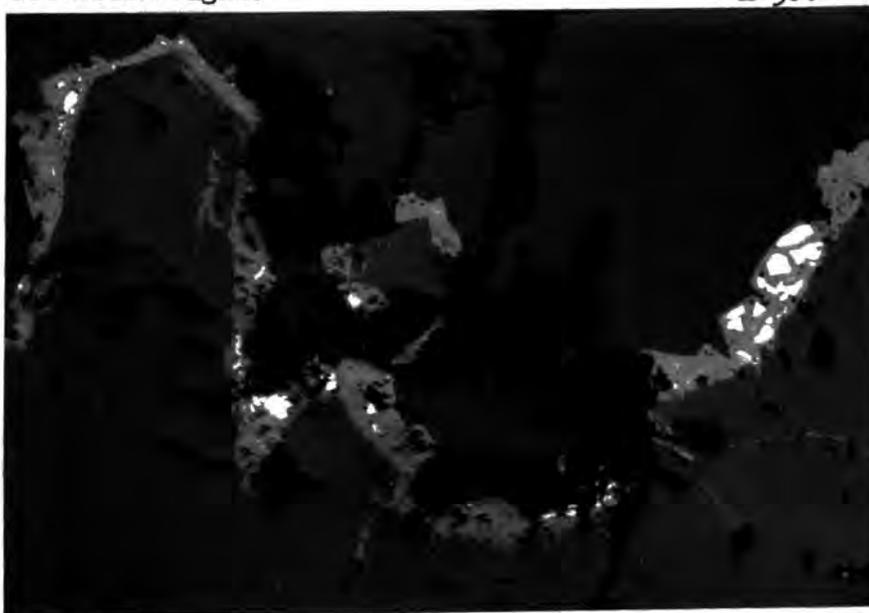


Figure 89: Veinlet of goethite between silicates containing minor pyrite
residuals.
Ordinary reflected light. X 50.

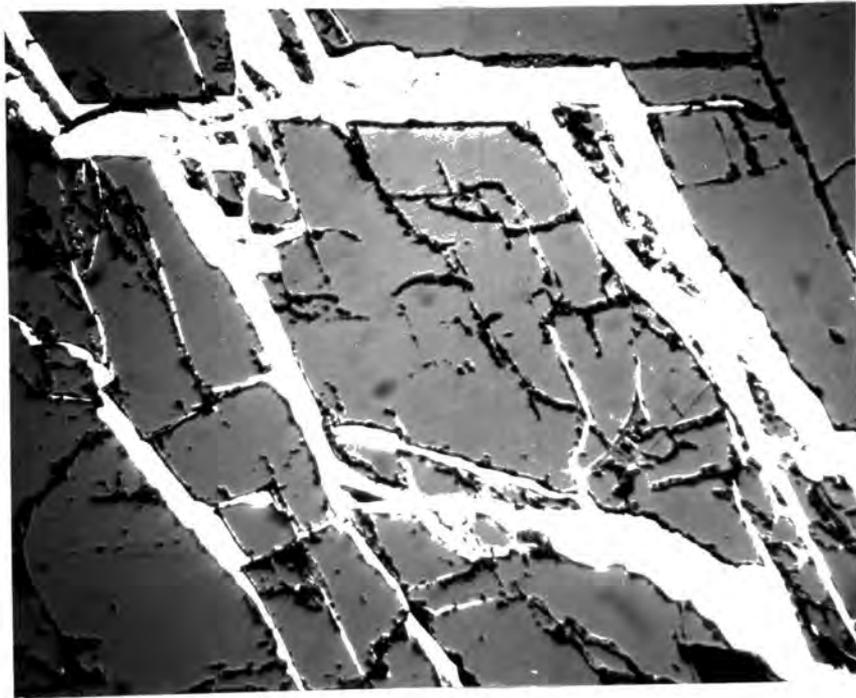


Figure 90: Chalcopyrite forming a matrix for silicate minerals.
Sulphide-graphite schist, Svartefjell Mine.
Ordinary reflected light. X 50.



Figure 91: Flake of graphite moulded around a grain of
chalcopyrite. Sulphide-graphite schist, Svartefjell Mine.
Ordinary reflected light. X 50.

Marcasite The occurrence of marcasite has already been discussed in the section on pyrrhotite alteration. It is quantitatively unimportant in the fahlbands.

Covellite Small irregular areas of covellite were observed as alteration products associated with chalcopyrite. The mineral showed its characteristic pale blue - indigo reflection pleochroism, and deep orange anisotropism. It is quantitatively unimportant.

Other minerals

Apart from the sulphides described above, two other minerals are very commonly seen in the polished specimens from the fahlbands. These are rutile and graphite.

Graphite Graphite can form up to 5% of some of the rocks in the fahlband zones. It is restricted to these sulphidic bands. The individual flakes show the good (0001) cleavage and have a marked reflection pleochroism. The graphite is strongly anisotropic with a pale brownish grey colour. The flakes are all arranged parallel to the foliation but are frequently deformed (Figure 92) and split along the cleavage (Figure 93). The individual crystals have often reacted to the shearing by developing sharp deformation twins (Figure 94).

Rutile Rutile is a common accessory mineral in the Modum Formation rocks and is not restricted to those in the fahlband zones. Modal analyses indicate that it forms up to 0.5% of the rock. It is present as small rounded crystals sometimes elongated parallel to the foliation. It is usually strongly corroded by the silicate minerals and contains frequent inclusions.

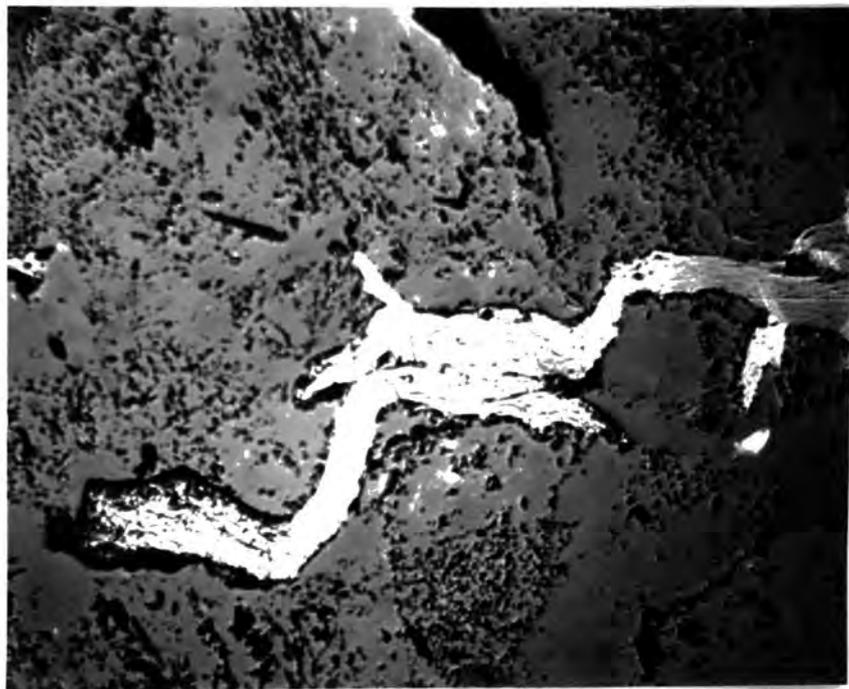


Figure 92: Deformed flakes of graphite in sulphide-graphite schist, Svartefjell Mine.

Ordinary reflected light.

X 50.

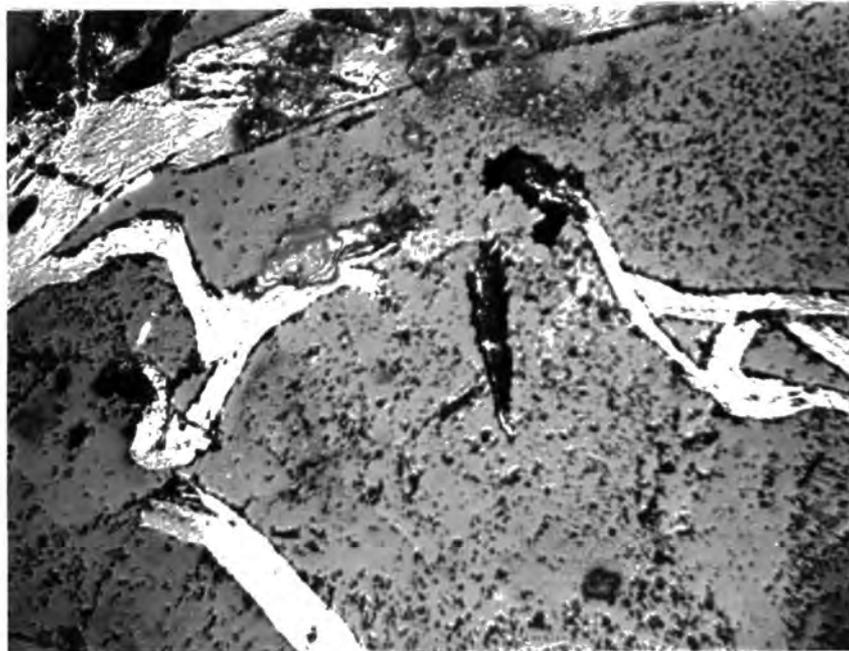


Figure 93: Graphite flakes split along the basal (0001) cleavage. Sulphide-graphite schist, Svartefjell.

Ordinary reflected light.

X 50.

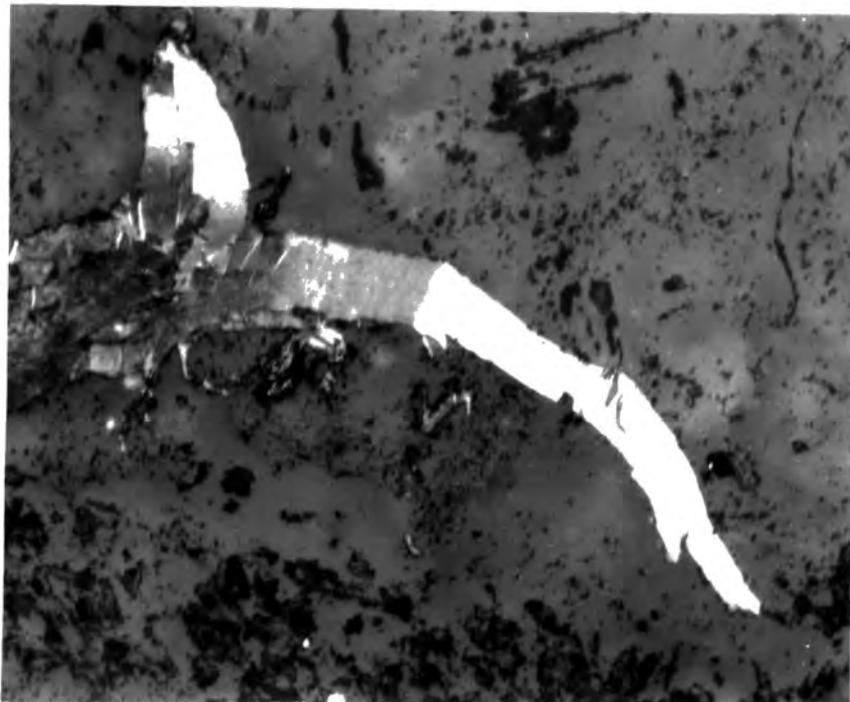


Figure 94: Development of twinning in graphite in response to the deformation.

Reflected light, crossed nicols.

X 50.

In polished specimens it has a low reflectivity and a grey colour slightly tinged by red. It shows weak reflection pleochroism from grey to greyish red. Under crossed nicols it is strongly anisotropic, the anisotropism being frequently masked by the strong red internal reflections.

Other fahlbands

The author has visited several localities in the Precambrian of Scandinavia from which fahlbands have been described. A summary of these occurrences is presented below.

Levang

Hofseth (1942) produced a map of the Levang peninsula, southern Norway on which she indicated the presence of two fahlbands. The first of these was shown as cutting across the strike near the fork on the Levang to Risør and Hellermøyren roads. In the position indicated by Hofseth's map a dolerite dyke is found in association with a quartz-felspar pegmatite containing about 20% magnetite. Both these rocks have a slight rusty weathering which may have been the origin of Hofseth's 'fahlband'.

The second locality is close to the farm Nybu. Here sillimanite-mica quartzites are slightly impregnated with sulphides. Abandoned trial workings are visible in the vicinity of Nybu. Recent mapping by Elder (1964) has shown that this locality is near a major east-west shear zone.

Similar fahlband zones are common over much of the Levang area (Elder, personal communication). Specimens collected on the Portør peninsula, Levang show definite pyrite bands, 0.5 - 2 cms. thick, containing minor chalcopyrite.

Vena

The copper-cobalt deposit of Vena is situated near the village of Ammeberg, Örebro Län, central Sweden. Mining for copper commenced in 1770, in 1805 cobalt was also produced and after 1812 the mines were worked for cobalt alone. Production ceased in 1880. The deposits have been described by Bøbert (1832), Daubree (1849) and Tegengren (1924).

The ore bearing zone contains a great variety of rock types. The main lithology, described by Swedish geologists as 'grey leptite', consists of alternating bands of biotite-gneiss and quartzitic rocks. Thin limestone and dolomite bands are occasionally present, but calcareous layers are more commonly represented by bands of diopside skarn. To the south of this region is a large area of granite and migmatite which covers much of the central part of the Sörmland region. To the north is a region underlain by homogeneous 'red leptite' interpreted by Magnusson (1960) as a metamorphosed volcanic extrusive series.

At Vena a fahlband zone follows the WNW-ESE regional strike and has cobalt and zinc mineralisation associated with it. Chalcopyrite, pyrrhotite and pyrite are all commonly found as impregnations and small concordant lenses in the fahlband rocks. Typical fahl weathering is well developed, particularly in the more micaceous rocks. No graphite was observed. Sphalerite is here a common associate of the other fahlband minerals and several trial workings have been opened in richer concentrations of this mineral. The cobalt mineralisation will be discussed in a later section.

Inglesby

The Inglesby region lies 3.3 kms. north-west of the Vena cobalt mines mentioned above. There are many small pits and trial workings scattered in this tract of forest dating from the 18th and 19th centuries. There has been no geological description of the deposits.

The country rocks are similar to those at Vena. The fahlband consists of an impregnation of sulphides in grey biotite-gneiss and quartzite. The predominating sulphide is chalcopyrite with subordinate pyrite and pyrrhotite. The sulphides are occasionally concentrated in thin bands parallel to the foliation. Sphalerite is as common as the iron sulphides but is subordinate to chalcopyrite. No cobalt minerals were collected.

It is possible that this deposit lies on the northern extension of the Vena fahlband.

Gladhammar

Gladhammar is situated in the south-east of Sweden near the town of Västervik. The large, metasedimentary Västervik quartzite covers most of the region and stretches more than 110kms. along the strike, (Uytenbogaardt, 1960). In the vicinity of the Gladhammar cobalt mines this quartzite is sparsely impregnated with pyrrhotite, pyrite and chalcopyrite, producing the typical fahl weathering products. This fahlband zone has been traced a distance of some 2 kms. along the strike for which distance it appeared to remain conformable with the foliation in the quartzite.

Conclusions

The characteristics of the fahlbands

Having considered brief descriptions of several fahlbands we are now in a position to summarise their characteristic features.

- (1) The sulphides are present in more than accessory amounts, but are only rarely major constituents of the rocks.
- (2) The sulphides are irregular in their distribution, being present in the form of multitudinous lenses in a variety of rock types.
- (3) The zones of sulphidic impregnation are conformable to the dip and strike of the enclosing metamorphic sequence.
- (4) The mineralogy is generally simple consisting of varying amounts of iron and copper sulphides. (Sphalerite at Vena and Inglesby).
- (5) The texture of the sulphides would suggest that they were present during the metamorphism of the host rocks.
- (6) The pyrrhotite geothermometer indicates that pyrrhotite - pyrite equilibration finally took place at a minimum temperature of approximately 450°C. (Modum)
- (7) The sulphides are associated with graphitic schists. (Modum)
- (8) The sulphides are associated with shear zones. (Modum and Levang).
- (9) Fahlband zones are very common in the Precambrian of Scandinavia.

Analogous occurrences

There are many descriptions in the literature of similar deposits in other parts of the world. Some of these are mentioned below.

The 'black-schists' of Fennoscandia These have been well described by Marmo and Mikkola (1951), Marmo (1960) and Peltola (1960), whose works contain a full bibliography of earlier sources. They occur as black, comparatively fine grained schists, usually forming narrow bands of considerable length. These schists characteristically contain graphite and iron sulphides, while chalcopyrite, sphalerite, pentlandite etc. may occur in minor amounts. The sulphides and graphite are present in rocks of very variable composition, they have been described from quartz-biotite schists, diopside skarn, fine grained limestone, amphibolite and quartz-amphibolite.

Loch Fyne, Argyleshire The Dalradian quartzose schists and amphibolites of this region are described as being impregnated by copper and nickel sulphides. (Wilson, 1921). The area was visited by the present writer, exposure is poor, being restricted to adit entrances and mine dumps. The mineralisation appears to be restricted to certain foliation planes in chlorite schist, quartzite and limestone, and consisted of pyrite and chalcopyrite with minor pyrrhotite, chalcocite and bornite. Exposure did not permit the tracing of these horizons along the strike. At McPhuns' Cairn (Wilson, op.cit.p.52) a thick, greater than 3 m., horizon of pyritic schist was poorly exposed under boulders on the shore.

Bethel area, Maine Fisher (1962) discussed sillimanite-grade metasedimentary rocks of the Lower Devonian Littleton Formation from the Bethel quadrangle, Maine. He described a pyrrhotite gneiss exposed for more than 7 miles along the strike, attaining a maximum thickness of 1,500 feet. This is a light- to dark-grey, medium-grained, massive, foliated or bedded micaceous gneiss characterised by the presence of pyrrhotite in the fresh rock. Minor pyrite is also present.

Balmat area, New York This area has been described by Brown and Engel (1956) and is underlain by a Precambrian metamorphic complex possibly equivalent to the Grenville Series of south-eastern Ontario. A pyritic schist horizon outcrops over a distance of approximately 4 miles in a broad arc (ibid. Pl.2). This horizon has an average thickness of about 200 feet with conspicuous rusty brown, yellow and brownish black outcrops. Basically the unit is a fine-grained quartz-mica-felspar schist, slightly graphitic, chloritic, and strongly but erratically pyritic.

The Norwegian 'vasskis' The sulphide ores of the Norwegian caledonides have been divided into gangkis, massive ore lenses possibly of epigenetic origin, and vasskis, thin conformable pyritic bands interpreted as syngenetic. (Carstens 1931, 1944, Foslie, 1938, Bugge 1948). The vasskis bands consist of pyrite, pyrrhotite and subordinate graphite. They are very fine grained and faithfully follow the structural complexities of the host rock, always being parallel to the lithologic variations. The author has seen typical vasskis outcropping in the vicinity of the Skorovass pyrite deposit (Gjelsvik, 1960). At this locality bands of pyrite varying between

0.5 and 3 cms. in thickness could be traced for dekametres along the strike (Figure 95). Individual bands closely follow the sedimentary layering, even around tight isoclinal folds (Figure 96).

The Northern Rhodesian Copperbelt The copper ores of Northern Rhodesia are a classic example of sulphide deposits which are strictly conformable to the slightly metamorphosed shale in which they occur. A comprehensive review of the literature dealing with these deposits has recently been published (Mendelsohn, 1961).

Mount Isa, Queensland The area comprises rocks mainly of Precambrian age including the Mount Isa Shale, which contains rich copper and lead-zinc ore bodies. The Shale has suffered only slight metamorphism and sedimentary structures are still visible (Carter, et.al. 1962). The Shale includes a diversity of lithologies but the most common types are thin-bedded siltstones and shales with pyritic and carbonaceous bands, (ibid.). The beds containing the fine grained pyrite enclose the lead-zinc ore bodies but are much more extensive, and may occur over the whole outcrop length of the Mount Isa Shale (Carter, 1953; Knight 1957). The pyrite of these shales has been shown to occur as infillings of fossil micro-organisms. (Love and Zimmerman, 1961).

Genetic hypotheses

From these brief descriptions it is seen that the fahlbands of the Kongsberg - Bamble Formation have many features in common with a class of deposits which has excited the imaginations of ore geologists for many years.

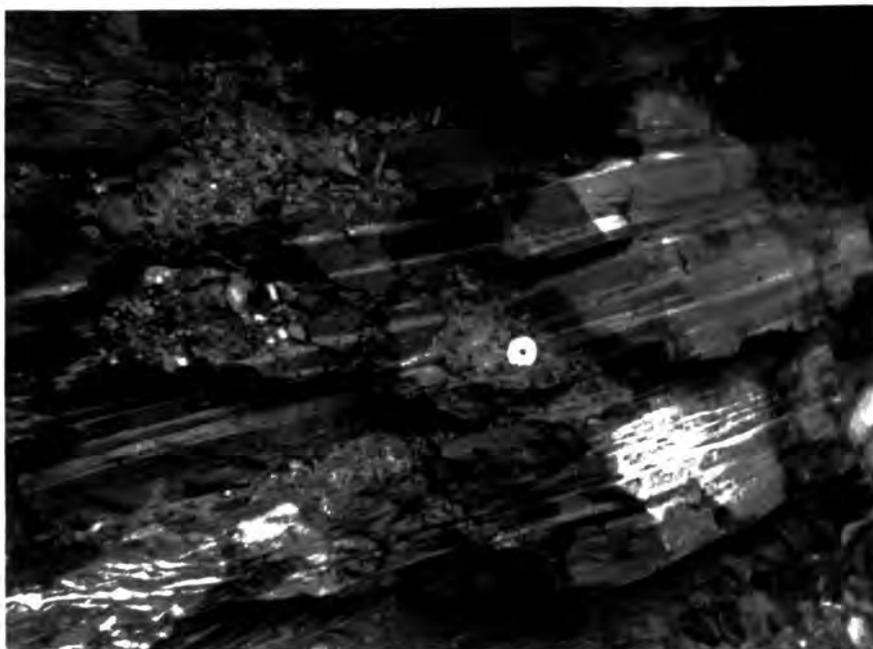


Figure 95: Bands of sedimentary pyrite (vasskis) exposed near the Skorovass pyrite deposit, Grong, Norway.



Figure 96: Detail of hand specimen collected at the above locality showing conformability of the pyrite horizon to the sedimentary layering in the presence of complex folding.

Many hypotheses have been proposed to account for the origin of these deposits. It is proposed to critically review some of these in the light of the field and laboratory data from this investigation.

Hydrothermal origin A post-metamorphic, hydrothermal introduction of the sulphides could be postulated on the grounds that the Modum fahlband zones are associated with shear zones which would provide good channels for the mineralising solutions. Bateman (1959, p.143) considers that disseminated sulphides can originate through "multiple-centre" replacement. Whereby country rock becomes permeated by mineralising solutions, replacement starting simultaneously at innumerable closely spaced centres, with arrested growth resulting in disseminated deposits.

In the author's opinion the main objection to this mechanism having operated at Modum lies in the texture of the ore minerals. They are frequently seen to be corroded by the rock silicates, rather than the converse that would be expected if the sulphides were of late introduction. The texture of the disseminated grains and small lenses is always conformable to the metamorphic fabric of the host rock and suggests that they were present, and recrystallised, under the influence of the metamorphism. With the large scale hydrothermal activity which would be necessary to produce these extensive sulphidic zones one would expect the presence of such gangue minerals as quartz, calcite, etc. to have been introduced contemporaneously. Veins of these minerals are present, but they have an associated pyrite-galena mineralisation of a decidedly different character to that in the fahlbands associated with them.

These veins clearly cross-cut and post-date the fahlbands.

The possibility cannot be discounted that the sulphides were emplaced by hydrothermal agencies some time prior to the main metamorphic episode. This hypothesis would explain many of the textural features observed. If the emplacement of the hydrothermal solutions was controlled by shear zones or faults developed parallel to the strike, with a consequent impregnation of the wall rocks on either side of the resulting lode, on metamorphism the original relations would be obscured by the recrystallisation of the sulphides and silicates and the healing of the original shear zones. This might produce the features observed in the fahlband zones.

However, the parallelism of mineral lineations with major fold axes suggests that the metamorphic recrystallisation was largely contemporaneous with the folding. Thus it follows that a hydrothermal vein developing prior to the metamorphism would be expected also to be affected by the folding. There is no evidence to suggest that this is the case. To account for the multitudinous fahlband zones in the Kongsberg - Bamble Formation it would be necessary to envisage extensive hydrothermal activity over a very large area, with the loci of the resulting deposits all controlled by the regional strike. This possibility appears unlikely. The absence of gangue minerals (quartz, calcite, fluorite, barytes, etc.) and base metal sulphides (sphalerite, galena etc.) accompanying the hydrothermal iron and copper sulphides also makes this hypothesis unattractive.

Syngenetic origin The textural appearance of the sulphides and the field relations of the fahlband zones leads to the speculation that they were deposited, syngenetically, with the sediments. If this were the case it would explain the conformability of the fahlbands with the dip and strike of the metasedimentary sequence, the metamorphic textures and temperatures of formation of the ore minerals and the common occurrence of such zones in the Kongsberg - Bamble Formation.

At first sight this hypothesis does not explain the distribution of the sulphides in widely varying rock types, and their common association with shear zones. A clue to the origin of the sulphides may lie in the three remaining characteristics (p. 194) not mentioned above. These are the presence of sulphides in more than accessory amounts, but only rarely as major rock-forming minerals, the simple mineralogy and the frequent association with graphitic schists. These latter characteristics suggest that the fahlbands may represent metasapropelites.

Sapropelic sediments are known from a wide variety of environments. The necessary conditions for the formation of sapropelites are the restriction, or exclusion, of highly oxygenated water and a source of organic matter. The organic matter must either be accumulating under euxinic conditions or be rapidly covered by succeeding sediments before it can be affected by aerobic bacteria. These bacteria are responsible for eliminating organic carbon as CO_2 or carbonate, hydrogen as water and nitrogen as ammonia.

Environments where sapropelic sediments accumulate have recently been reviewed by Dunham (1961). He lists: poorly drained terrestrial swamps; freshwater lakes with abundant phytoplankton; river estuaries, tidal lagoons and the tops of deltas; arms of the sea with restricted circulation; and basins in the sea bottom.



The lithology of the Modum Formation metasediments suggests that they represent a succession of alternating shales and shaly sandstones with interbedded thick sandstones and calcareous horizons. There is some evidence to suggest that they are of marine origin (p.74). A depositional environment similar to the actualistic example of the Bay of Danzig is envisaged where (Dunham, op.cit.):

"There are substantial areas where the mud is not black; there are banks of lamellibranchs; there are sandy beaches and bars. On the regional scale, a considerable variability of deposits is to be expected. In detail, the black muds may show poor sorting, cross bedding, lamination, slump structures and mud cracks, all reflecting the conditions under which they were formed."

In an environment such as this where rapid sedimentation and oxygen deficient bottom waters are present, anaerobic bacteria take the place of the oxidising type. Large areas of black surface mud have been observed in estuarine-lagoonal environments along the Dutch, Danish and North German coasts. Vann Straaten (1955) states that iron monosulphide (hydrotroilite) is forming in the near surface mud and changing to pyrite in a lower zone. Large areas of black mud contain as much as 5 ml./litre of hydrogen sulphide. The source of this reduced sulphur is of considerable interest.

A group of sulphate reducing bacteria, Desulfovibrio desulfuricans, Beijerinck, is well known. The metabolism of this group is linked to the reduction of the sulphate ion to sulphide. The origin of the sulphide could be from the activity of the true sulphur bacteria (thiobacteria) which decompose the amino-acids of organic material or the sulphate normally present in sea water.

A bacterial production of 2 mgms. of H₂S per litre of water per day has been recorded (Ivanov, 1957) and Rubenchik (1946) notes certain cultures which produce a total of 2,400 mgms. H₂S per litre. Bastin (1926) early demonstrated that H₂S in petroleum reservoirs was derived from sulphate by the action of anaerobic bacteria. Rubenchik (op.cit.) reports that "99.4-99.6% of the total amount of H₂S in the Black Sea was formed from sulphates in the sea water as a result of the activity of sulphate reducing bacteria."

Thus in an anaerobic environment with a source of sulphate ions there seems to be little difficulty in the production of considerable amounts of hydrogen sulphide. However, high concentrations

of molecular H_2S are toxic to sulphate reducing bacteria (Rubenchik, op. cit.). Some means of fixing the H_2S is necessary to allow such a colony to survive. In this context it is of interest to quote Baas Becking et. al. (1960, p. 243):

"The intensity of sulphate reduction depends upon the sulphate content of the water and on the available hydrogen, in both organic and inorganic form. The iron concentration is also important, as iron is the principal acceptor of the H_2S formed. The highly reactive, black iron sulphides may be partly oxidised with the formation of the more stable pyrite and marcasite."

The manner of fixation of the iron sulphide has been recently discussed in the literature.

Pyrite is frequently found with a framboidal habit in black muds (Naumann, 1919; Vallentyne, 1961; Van Straaten, 1955; Love and Murray, 1963). Black shales interpreted as being formed in environments similar to the black muds, are very common in the geological column (Dunham, 1961). These also frequently contain framboidal pyrite (Berg, 1928; Baker, 1960; Ramdohr, 1953; Eagar, 1953; Hemingway, 1934; Love, 1962a). Schneiderhöhn (1923) interpreted these spheres as fossil sulphur bacteria. However Schouten (1946), disputed this interpretation on the grounds that the spheres lacked any resemblance to known sulphur bacteria. Vallentyne, (1963) dissolved the spheres in H_2O_2 followed by HCl and concluded that no significant amount of organic residue was obtainable.

Love (1957) obtained a residue after dissolving the spheres in HNO_3 and HF which was unstable on heating, he concluded that it was of organic composition. From the regularity of the microscopic structure of the remains he concluded that they represented a micro-organism of unknown affinities which he designated Pyritosphaera barbaria (Love, 1962b). Greensmith (1962) has shown that pyrite spheres can also form beneath agitated and highly oxygenated waters in micro-brecciated and oolitic calcareous deposits. Love and Murray (op.cit.) concluded that the spheres grew authigenically within the sediment.

"The pyrite was not present until finely comminuted plant debris and the clay matrix had been deposited as mud rich in iron. As part of the earliest process of diagenesis, authigenic growth of sulphide mineral then occurred.....The formation of iron sulphide continues as long as sulfate was available from the overlying water and iron from the same source or from absorption complexes with organic matter already in the sediment." (Love, 1962a).

Spheres of chalcopyrite have also been described by Schneiderhöhn (1923). If base metals are available then they will presumably react in the same way as iron, indeed D. desulfuricans is capable of precipitating most of the common metals (Baas Becking and Moore, 1961). To explain the exceptional concentration of base metals in the Permian Kupferschiefer Dunham (1964) has postulated the leakage of mineralizing solutions into the Zechstein lagoon during a period when reducing conditions were operative. Gillson (1963) has reviewed the evidence against a syngenetic origin for the Northern Rhodesian Copperbelt, in discussing the source of the necessary copper he concludes:

"The problem, therefore, is not so much the quantity of copper, but concentrating all of it in one very narrow basin.....".

However, White and Northcote (1962) have shown that base metals tend to accumulate preferentially in fine-grained sediments and also show a similarity to the distribution pattern for sulphur. They are particularly concentrated in modern, near-shore, marine environments.

Thus the possibility exists that the Modum Formation was laid down in a near-shore, deltaic environment. Intermittently conditions were such that organic matter was deposited and covered before decomposition was complete. The activity of sulphate reducing bacteria or micro-organisms related to Pyritosphaera caused the liberation of hydrogen sulphide which was authigenically fixed by the iron and any copper present, leading to the production of pyrite and chalcopyrite. Precambrian examples of framboidal pyrite have been described by Baker (1960) and Love and Zimmerman (1961).

On metamorphism the volatile elements of the organic matter (H,O,N) will be driven off to leave a carbonaceous residue (shungite, Marmo, 1953b) which at high metamorphic grades is converted to graphite. The sulphides recrystallise and lose any trace of their original framboidal form. Where conditions are suitable they migrate, together with quartz, to form segregation lenses paralleling the

foliation. The high mobility of the sulphides is reflected in the occasional cross-cutting sulphide veinlets. The relative abundances of pyrite and pyrrhotite could be controlled by variations in the sulphur pressure during metamorphism.

In conclusion, let us review the validity of this hypothesis in relation to the general characteristics of fahlbands as listed on page 194.

- (1) The amount of sulphide present is a function of the reducing environment in the original sediments.
- (2) The irregularity of the sulphide distribution is related to the variable conditions at the time of deposition. Similar distributions are seen in unmetamorphosed black shales.
- (3) The fahlband zones are roughly conformable because suitable conditions (i.e. stagnant bottom waters, etc.) occurred at one time plane, over an area where different sediment types formed the sea bed.
- (4) The mineralogy is simple because iron and copper were the only two elements readily available for H_2S fixation.
- (5) & (6) The sulphides were present in the original sediment and were recrystallised during the metamorphism.
- (7) The graphite represents the metamorphosed, sapropelic, organic detritus.
- (8) Shear zones have developed preferentially in the graphite schists because of their relative incompetence under an applied stress.
- (9) Fahlband zones are common features because they represent the metamorphosed equivalents of the equally common black shales of later geological periods.

THE COBALT MINERALISATION

Distribution of workings

The cobalt mineralisation is concentrated in two narrow zones which stretch for considerable distances parallel to the regional strike of the Modum Formation. Other, uneconomic, zones have been reported to occur between the Snarum River and the Tyrifjord (Bøbert, 1848).

Eastern Zone

The eastern mineralized zone has only been explored by small prospecting pits. The ore in this zone was a poor dissemination with cobaltian-arsenopyrite as the only ore mineral, a combination of circumstances which made the percentage of recoverable cobalt too low to be economic. The workings lie roughly on the same line of strike but are not restricted to the same rock type. The zone is not continuous, there are wide gaps between the various areas where trials were carried out. From north to south the main prospecting areas were the Gubberud, Bakke, Ravbjerg, Olafsby, Skraetterberg and Bøen Trials (Figure 97).

Gubberud Trials These are mentioned in older descriptions as being small trials in the woodland belonging to Gubberud Farm. The present occupants of the farm knew nothing of their whereabouts and a search in the area indicated by Bøbert (1848, p.13) revealed no trace of them.

Bakke, Ravbjerg and Olafsby Trials Bøbert (op.cit.) shows nine separate trial workings in this area. With the aid of the Olafsby family the present writer rediscovered five of these in the woodland to the east of the railway track on Olafsby property (Figure 99).

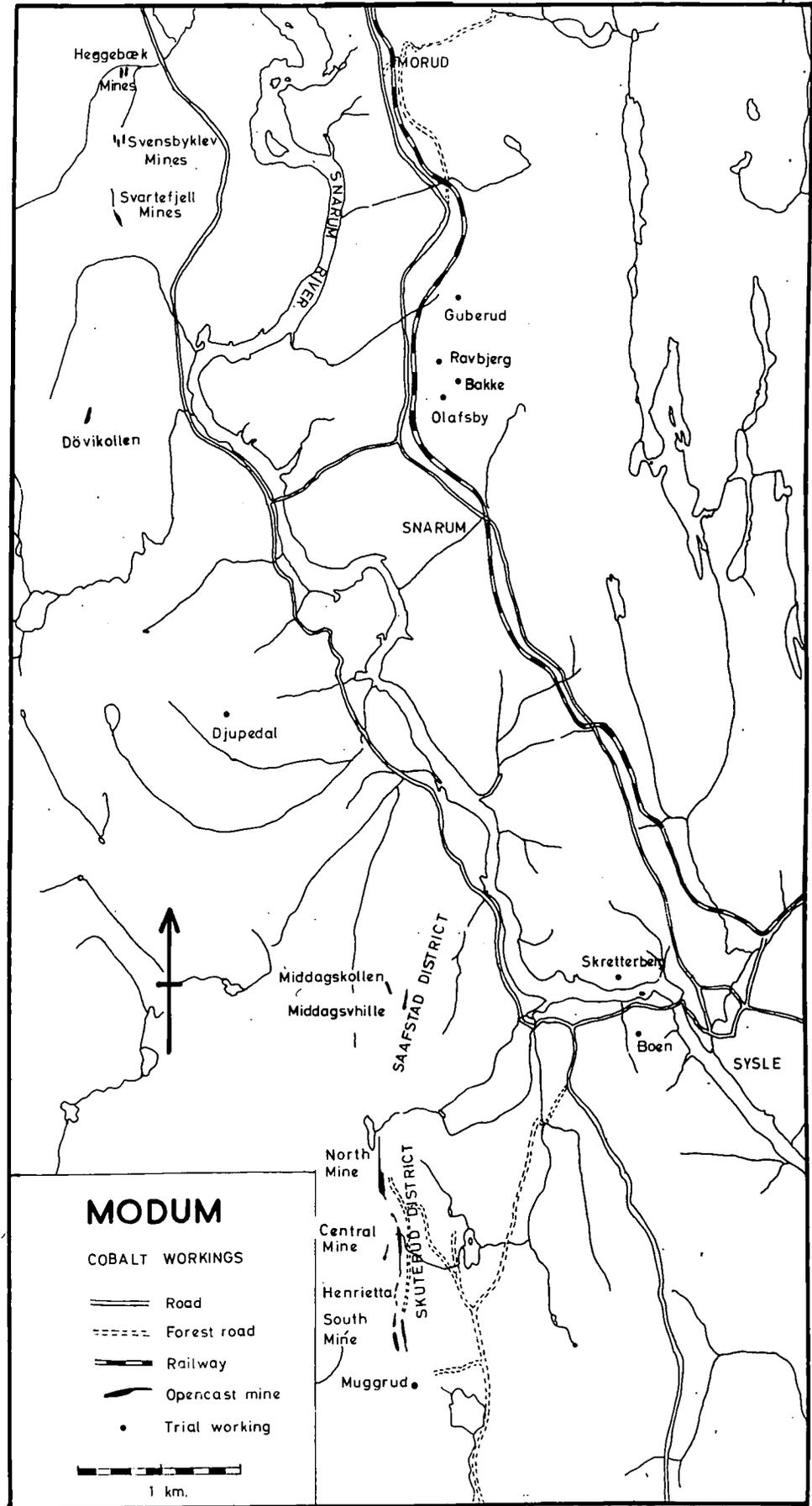
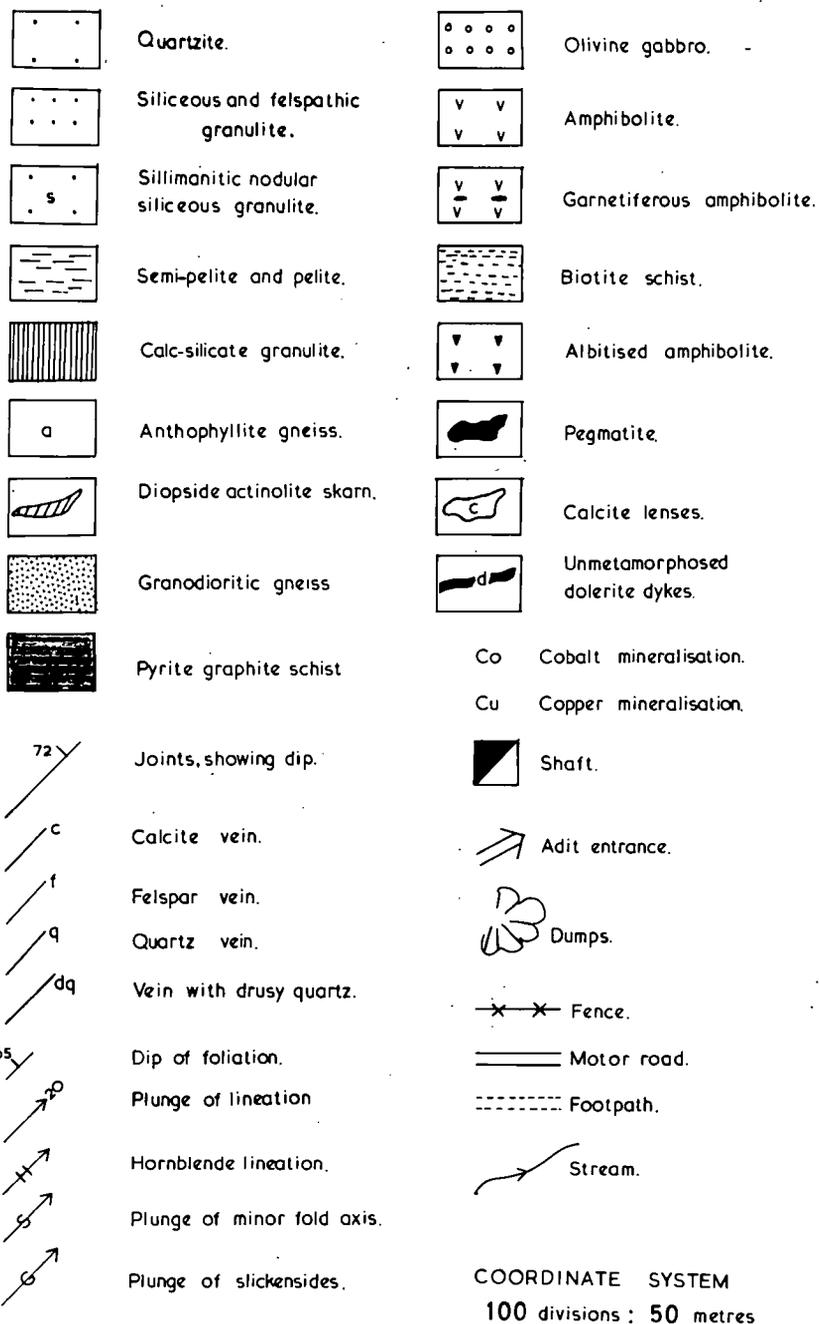


Fig. 97



COORDINATE SYSTEM
100 divisions : 50 metres

KEY TO FIGURES 99 - 115

The workings consist of very small open pits, measuring about three metres square (Bakke and Ravbjerg) and boarded over shafts (Olafsby). They are situated in siliceous granulites near the boundary of these rocks with alternating amphibolites (Figure 99). The dumps at the Ravbjerg Trial provided specimens of quartz-biotite-garnet-felspar gneiss containing many segregation quartz lenses parallel to the foliation. Disseminated in the rock and concentrated in the quartz lenses were pyrrhotite and minor chalcopyrite. The outcrops had the typical fahl weathering. No cobalt bearing minerals were observed.

Skraetterberg Trial This trial is not mentioned in the older descriptions and was probably opened in the 1930's by prospectors from Oslo. (Harald Skuterud - personal communication). It lies on a straight line between the Olafsby and Bøen Trials on a steep cliff on the north bank of the Snarum River (Figure 97). Exposed in the face of the small working is a biotite rich, fahl weathering gneiss. No cobalt minerals and very few sulphides were observed.

Bøen Trial This is a small, water-filled, opening, concealed in woodland to the east of a small footpath 100 m. south of the Sysle-Skuterud road. Remains of lifting gear are visible and the extent of the dumps suggests the shaft was at least 10-15 m. deep.

Biotite rich gneiss of variable lithology is exposed in the surrounding outcrops (Figure 100). All variations are found between biotite-amphibolite and biotite schist to biotite-quartz-garnet-felspar gneiss. Pegmatite is developed as small concordant and larger, crosscutting, lenses. The small quartz-felspar lenses give the rock a lineation. Very slight fahl weathering is developed. The dumps contain the same variety of rock types and have minor pyrrhotite and chalcopyrite, particularly associated with quartz rich lenses.

Western zone

The mines in the Western zone have been the major source of the cobalt produced in Modum. The zone stretches for 12 kms. roughly parallel to the strike of the Modum Formation and is associated with one of the fahlband zones. The main producing mines are restricted to the topographic high points (see profile - Figure 102).

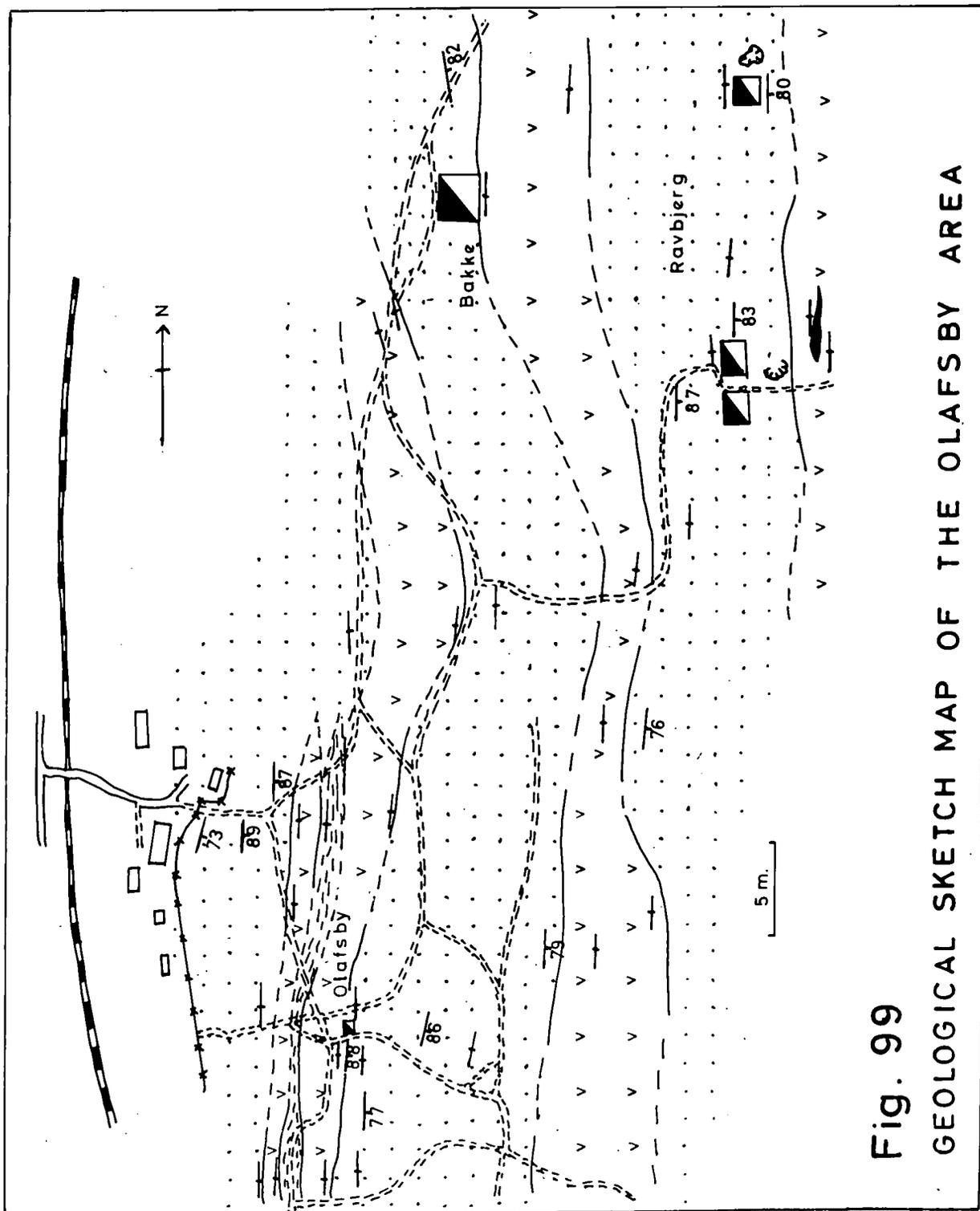
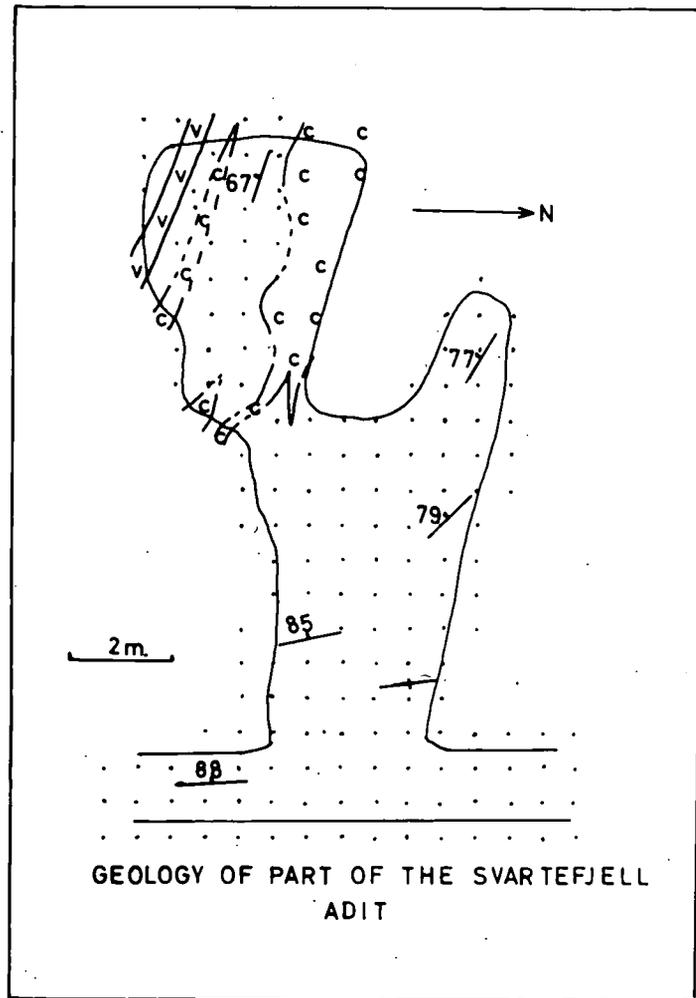
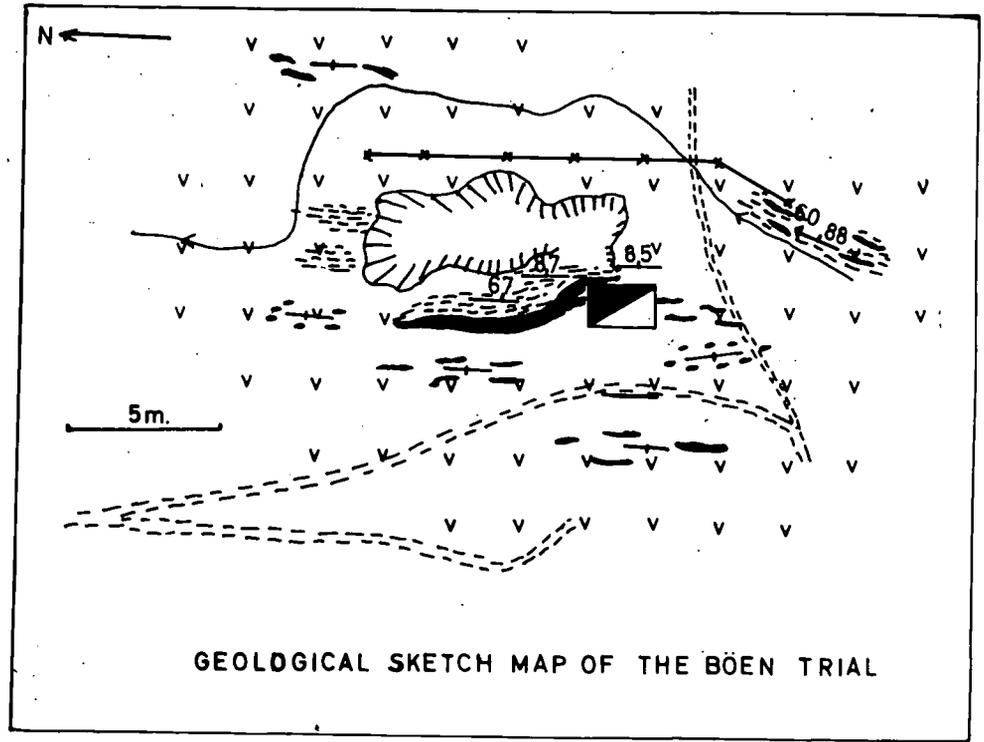


Fig. 99
GEOLOGICAL SKETCH MAP OF THE OLAFSBY AREA



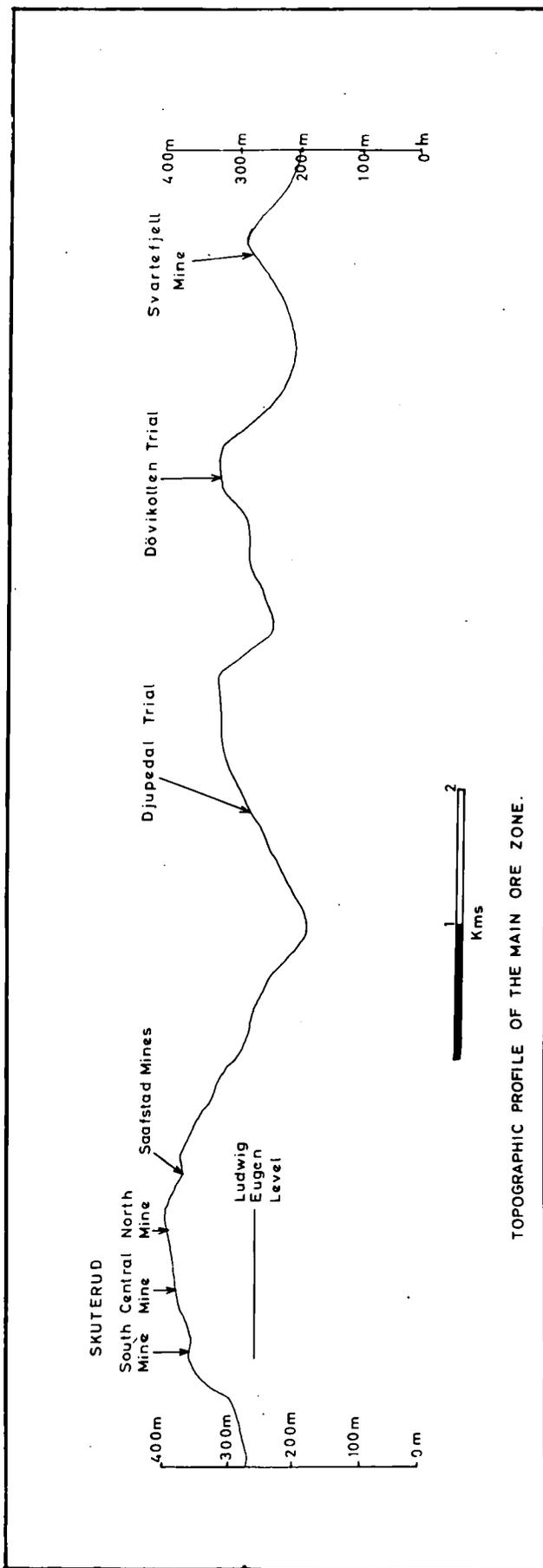


Fig. 102

TOPOGRAPHIC PROFILE OF THE MAIN ORE ZONE.

From north to south the workings are (Figure 97) the Heggebek and Svensby Trials; Svartefjell Mines; Døvikollen and Djupedal Trials; Saafstad Mines; North, Central and South Skuterud Mines and the Muggerud Trial. The southern group of workings, comprising the Skuterud and Saafstad Mines, were the main cobalt producers and were worked by the owners of the Haugfoss Blaufarveverk, while the northern group, notably the Svartefjell Mine, were worked by 'Snarum Verk' between 1822 and 1849 (see p.13).

Heggebek Trials These are the most northerly of the cobalt workings and are situated about 900 m. due north of the Svartefjell Mines. However, mineralisation has been reported even further north by Halvorsen (1954) who states that traces of 'cobalt' were seen on the west bank of the Snarum River during the building of a dam at Ramfoss ($60^{\circ}F'50''N.Lat./0^{\circ}54'15''Long.$). Between this locality and the Heggebek Trials he records that cobalt was seen in the banks of Kvernelven while a log jam was being freed with dynamite. The present writer has been unable to confirm these occurrences.

The Heggebek Trials consist of two parallel opencast workings, about 3 m. wide, driven a distance of about 15 m. into the hillside, in a north-south direction. The eastern of these opencasts is driven along the strike of a strongly sheared, sulphide-graphite schist which is deeply weathered. The east wall of the open cast is in felspathic-granulite containing disseminated pyrite and chalcopyrite. The open-casts are separated by a 10 m. thick striped amphibolite. The western open-cast is mainly in a rather pure quartzite showing good colour banding, highly complex folding and quartz-segregation lenses. The western wall of the working is occupied by another sulphide-graphite schist. The dominant sulphide in both the graphite schists is pyrite. The dumps are very overgrown but show a strongly foliated siliceous-granulite impregnated with pyrite and chalcopyrite.

Svensby Trials These are shown on Bøbert's map (op.cit.) as lying between the Svartefjell and Heggebek workings. The author could find no trace of them in the dense woodland.

Svartefjell Mines These workings were the most extensive of those operated by the Snarum company. Two open-cast mines have been opened along the same line of strike. These are connected by an adit about 15 m. below ground level. The southern mine reaches a depth of 38 m. and is 95 m. long, while the northern mine is about 115 m. long but

only reaches a depth of about 20 m. It is possible that this latter figure is too small as the lower parts of the workings have collapsed. Access to the lower parts of the northern workings was by an adit driven 75 m. through the hillside from the north-west. This adit is no longer passable. A similar adit was driven in to give access to the southern workings at a depth of about 20 m. This adit is still accessible, for part of its length. The geology in part of this adit is shown in Figure 101.

A geological map of the open-cast workings (Figure 103) shows that they were mainly based on a pyrite-graphite schist horizon interbanded with siliceous granulite and amphibolite. Some traces of cobalt mineralisation are visible in the east wall of the southern mine at the junction of siliceous granulites and amphibolites (Figure 104).

It has been estimated (Adamson, 1950) that about 1,000 metric tons of ore have been raised from the Svartefjell Mines.

Døvikollen Trial Ore was taken from here to the Snarum cobalt-blue factory. The workings lie about 80 m. due west of the crest of the hill Døvikollen. Access is by way of the tractor road, followed to its termination beyond Damtjern. From this point the workings lie about ten minutes walk east through woods.

The open-cast has steep walls and the lower workings are flooded and high inaccessible for sampling. The western wall shows typical fahlband weathering developed in sillimanitic-nodular siliceous granulite. This weathering presumably attracted the original prospectors.

There is a marked swing in the direction of the open-cast (Figure 106), this coincides with a swing in the foliation direction of the surrounding rocks. As the other rock units are widest where the foliation swings round it is reasonable to suppose that the siliceous granulite was also at its widest here. Thus from the map it looks as if the workings are mainly in amphibolite whereas they were probably mainly in the siliceous granulite.

From the broken material in the dumps there is very little amphibolite which tends to confirm the above suggestion. There are two main dump areas associated with the workings. One of these, the largest, has large broken pieces of rock and looks like the waste tip, from this were collected anthophyllite rock, brecciated and containing minor arsenopyrite and cobaltite; amphibolite and mica-ceous quartzite with pyrite, pyrrhotite and minor chalcopyrite.

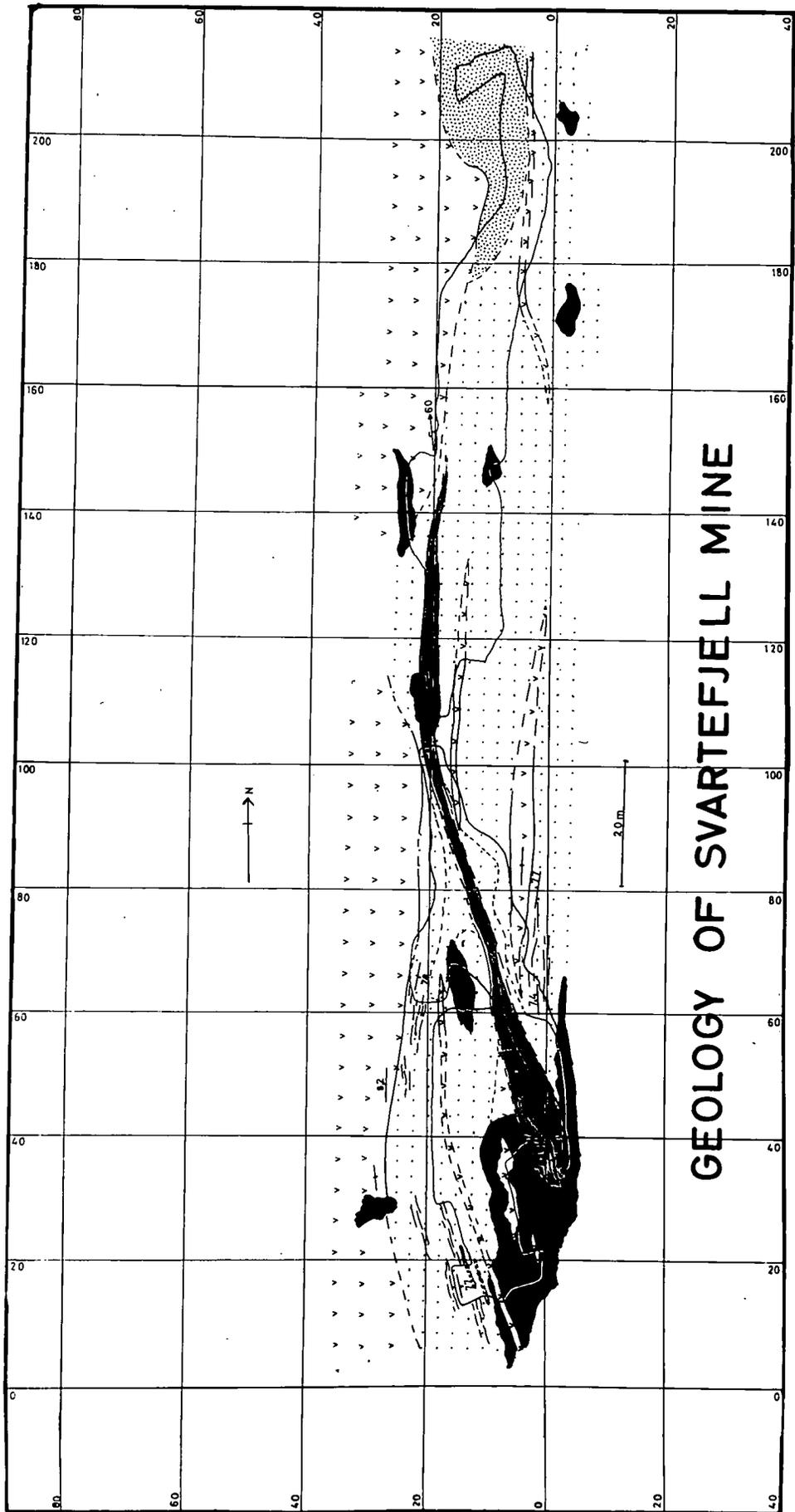


Fig. 103

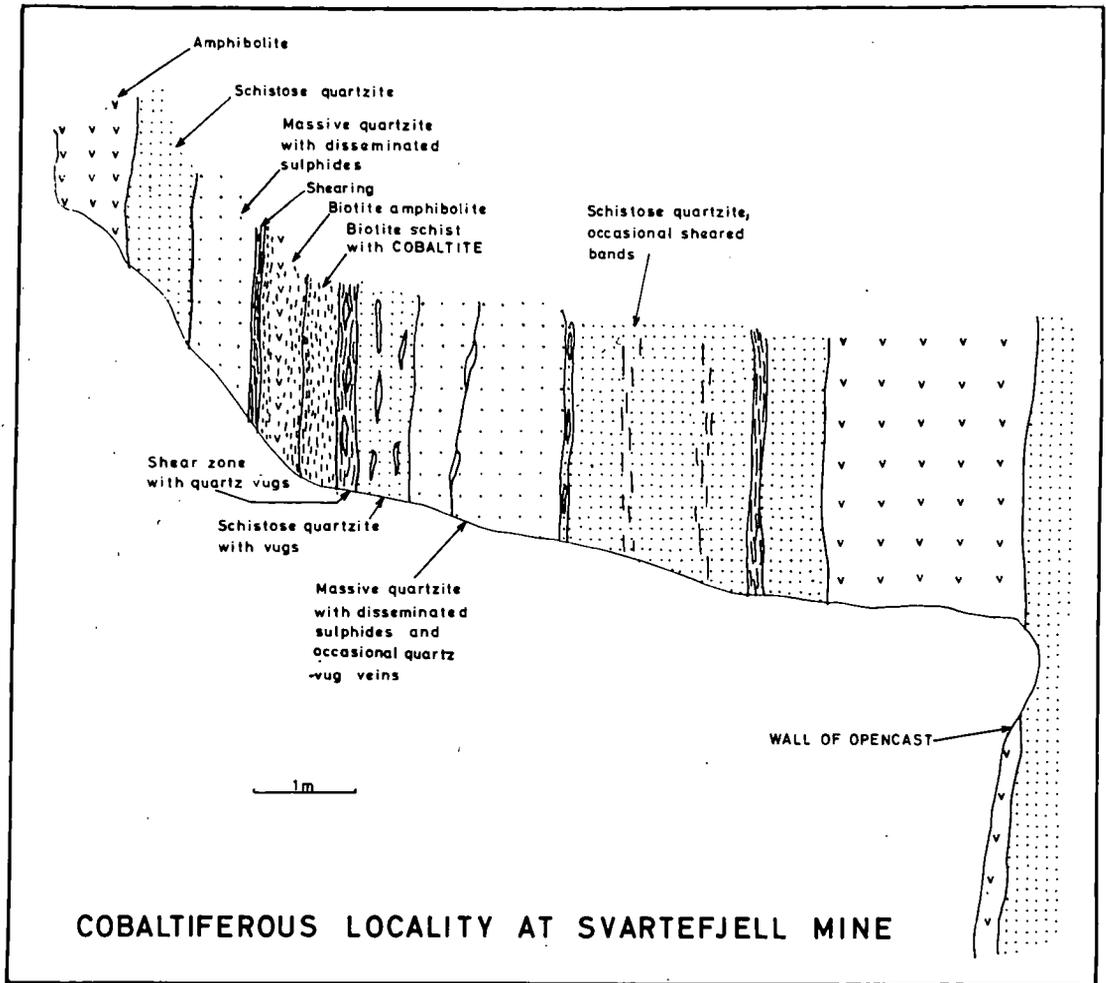


Fig. 104

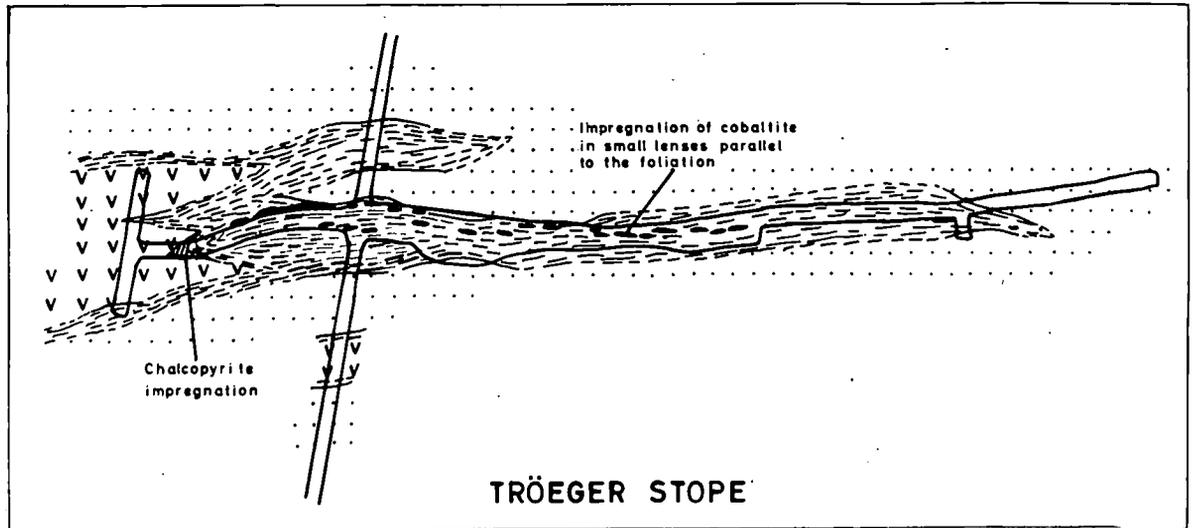


Fig. 105

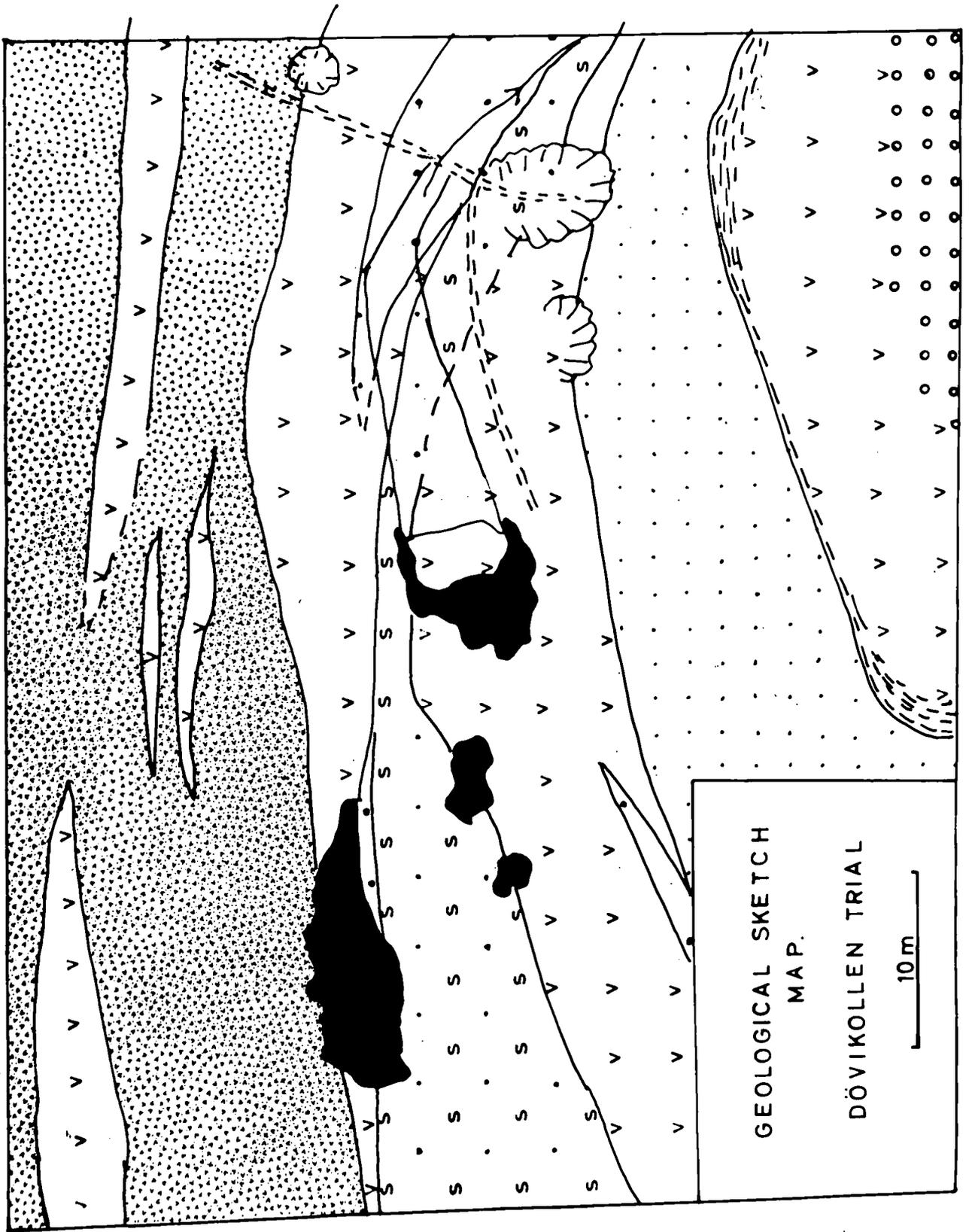


Fig. 106

The other dump contains small pieces, all broken to roughly the same size, and seems to represent one of the final stages of sorting, for nearly all fragments examined from here contain small amounts of ore. This gives some idea of the mineralised rock types, two main lithologies were represented: fine-grained phlogopitic-quartzite containing graphite, pyrite, pyrrhotite, chalcopyrite and arsenopyrite in small stringers, and a sugary white quartzite containing no sulphides but with disseminated arsenopyrite. The cobaltiferous nature of the arsenopyrite is shown by the presence of erythrite on weathered specimens.

Djupedal Trial This is a small open-cast working lying in dense woodland on the eastern slope of the unnamed hill lying immediately south of Hovdekollen. The geological sketch-map (Figure 107) shows the working to be based on a narrow band of sulphidic graphite-schist. The open-pit follows the strike for a distance of about 15 m. with a water filled shaft at the northern end. Also visible in the workings is a fresh dolerite dyke which transgresses the foliation of the graphitic schist and adjacent siliceous - granulite. No cobalt minerals were observed.

The Saafstad Mines These lie about one kilometre north of the North Mine, Skuterud. Bøbert shows about 17 small workings in this area. The largest of these visible today is the Middagshville Mine (Figure 108). This ceased production in 1843 (Lammers, 1849b.), the average ore grade being less than 0.05% cobalt (Adamson, 1950). The southern part of the Middagshville Mine is an open-cast about 120 m. long divided into two halves by a pillar. About 60 m. north, on the same line of strike is another open-cast which stretches for about 80 m. and is flooded in the lower workings. The geological sketch map (Figure 108) shows that this line of strike is occupied by a graphitic mica-schist. The schist has been subjected to heavy shearing and a deep fahl weathering is developed. The siliceous granulites contain disseminated pyrrhotite and chalcopyrite. No cobalt minerals were observed.

The Skuterud Mines The Skuterud mining district is by far the most important in the whole Modum area. This was the site of the original discovery of cobalt, the mines were in continuous production from 1772 until 1898. Apart from the surface workings, underground development has been carried out at three levels. The district is conveniently divided into three regions containing the North, Central and South Mines respectively. The geology of the district is shown on the map (Figures 109, 110).

The North Mine consists of an open-cast which stretches a total distance of 360 m. along the strike. The most northerly section (210 m.) is only a few metres wide while the southern section is very broad, reaching a maximum width of 50 m. Access to the workings can be gained by way of the Ludwig Adit. Cobalt

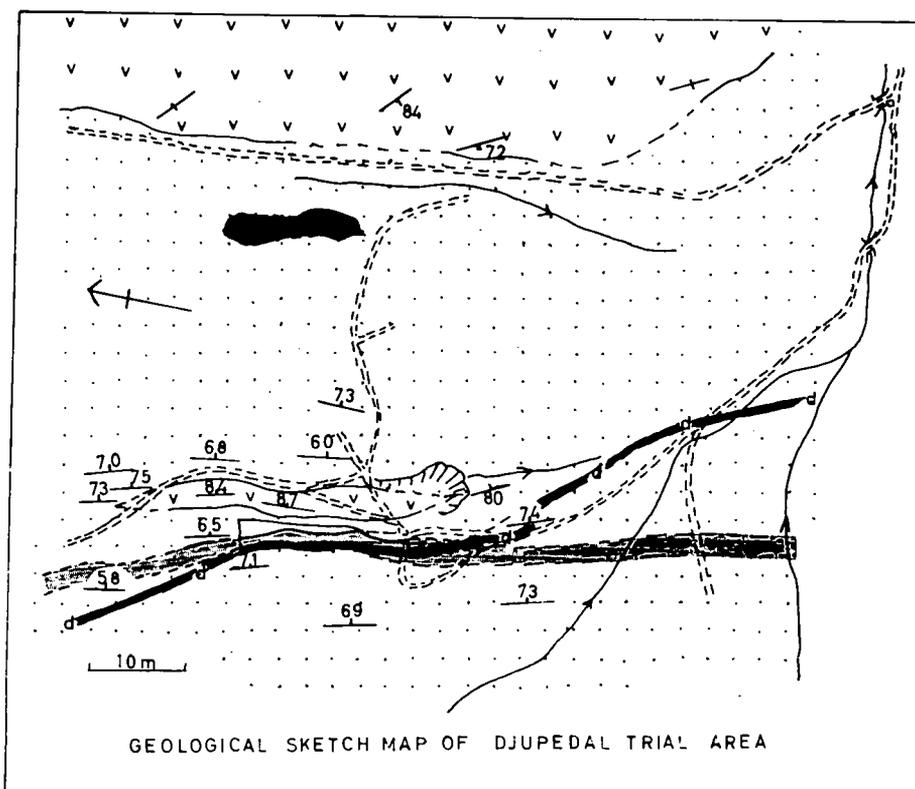


Fig. 107

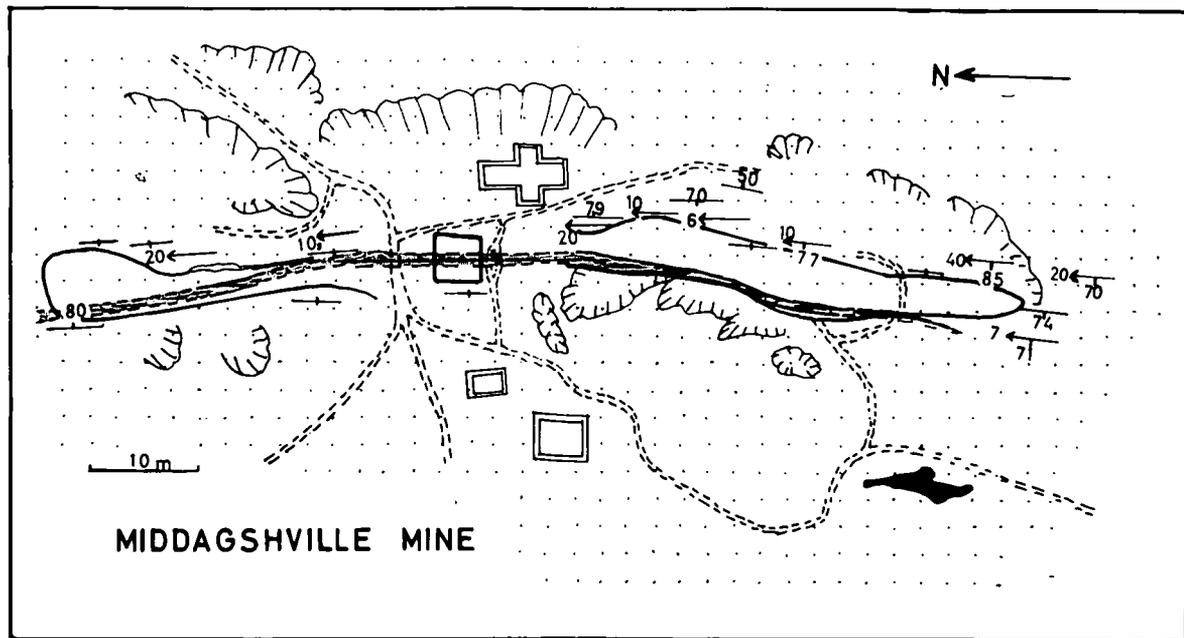
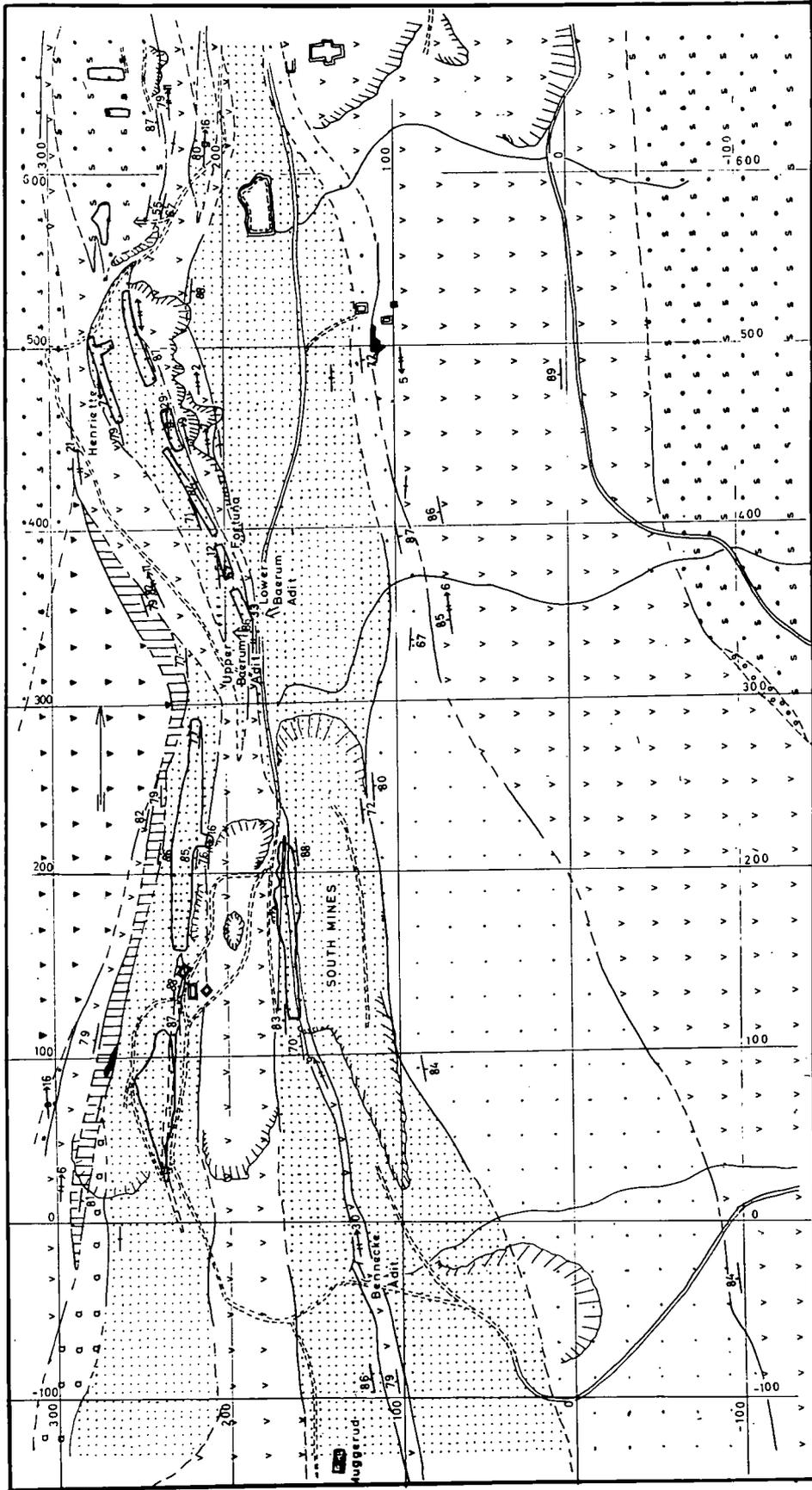
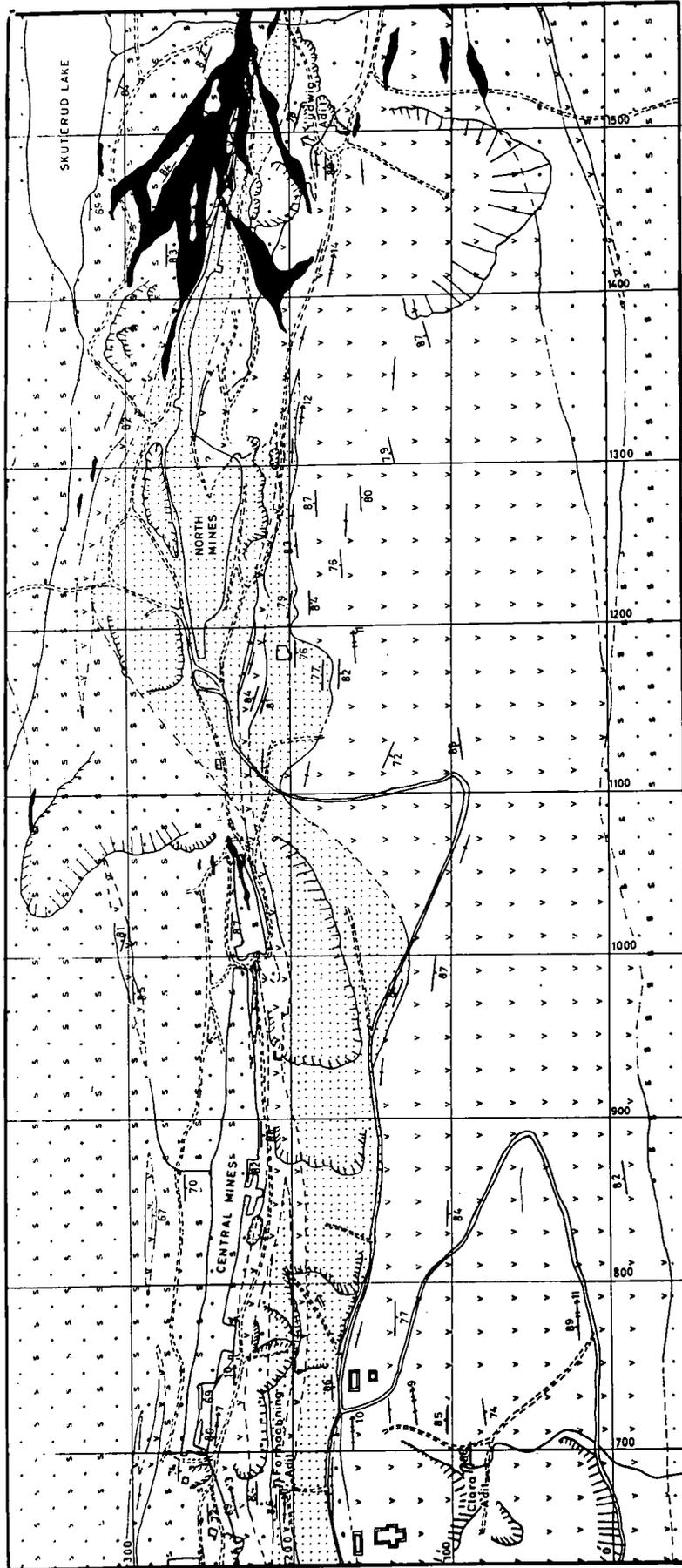


Fig. 108



GEOLOGY OF THE SKUTERUD DISTRICT
SOUTHERN AREA

Fig. 109



GEOLOGY OF THE SKUTERUD DISTRICT
NORTHERN AREA

Fig. 110

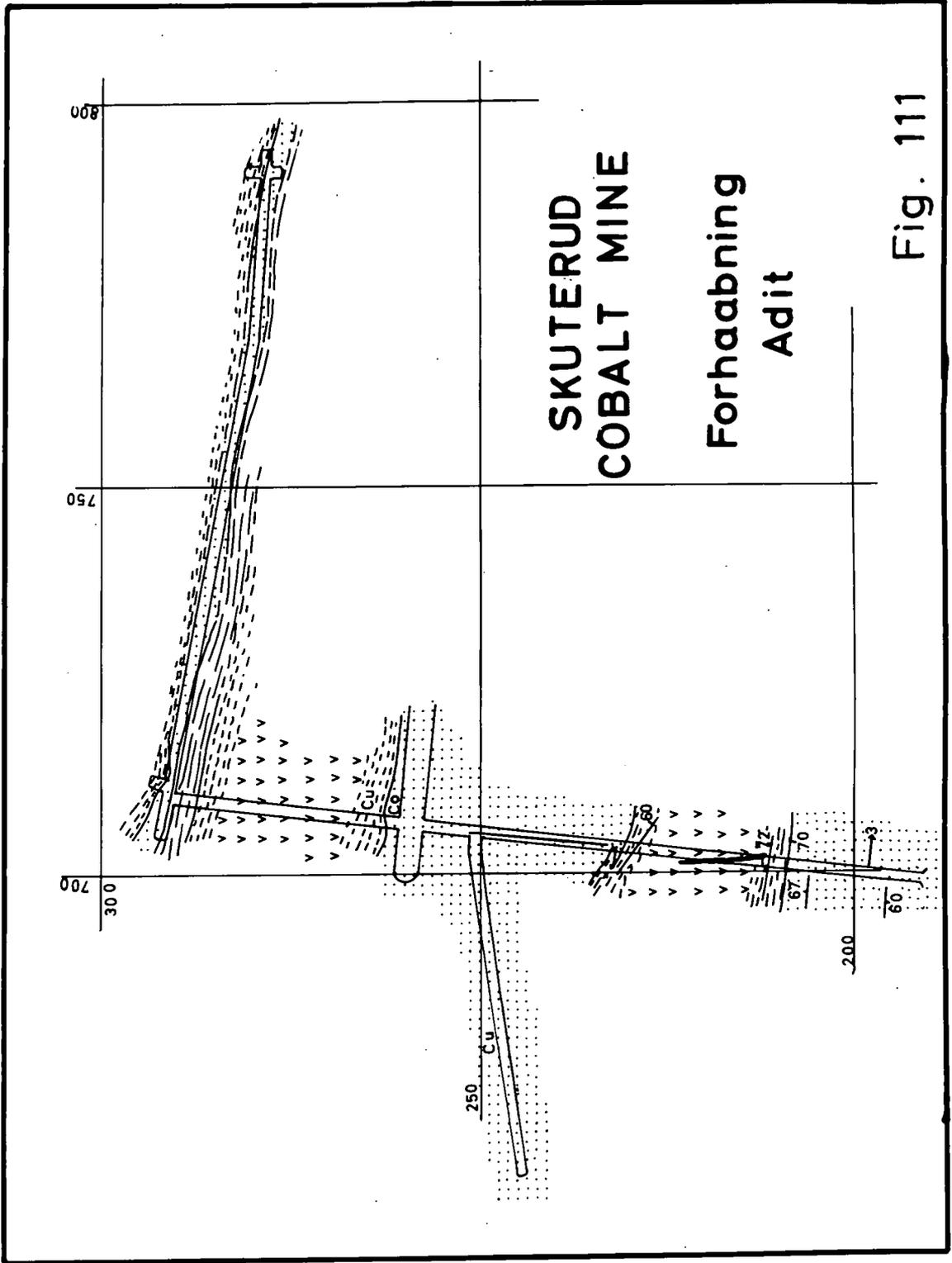
mineralisation is visible in several places in the North Mine, notably in a pillar at the north end of the wider section. A connecting adit is accessible from the south of the North Mine to the northern part of the Central Mine.

The Central Mine has an open-cast pit 370 m. long. The most northerly 70 m. being separated by a pillar. This northern section is completely inaccessible from the surface, however parts of it may be reached from the Klara Level. The main surface workings are accessible from the surface if care is taken and suitable precautions are made.

Between the Central and the South Mines are a series of small open-cast workings along the strike. These have been collectively called the Henrietta Mines.

The South Mines comprise three distinct open-cast workings. Two of these lie on the same line of strike to the west, these are connected by an underground stope which has partially collapsed but is still passable in places. The northern open pit can be reached at two levels by way of the Upper and Lower Baerum Adits. The eastern open cast is about 110 m. long and can be investigated for its complete length. The Bennecke Adit is driven in to give access to the lower workings of this mine, this adit is now flooded.

The Central Mines have been worked at three levels by means of cross-cuts driven in from the east to intersect the ore zone at depth. The Forhaabning Level is 350 m. above sea level. The cross-cut and a drive north parallel to the ore zone are passable. Access to the open-cast can also be gained by this level. The geology of the Forhaabning Level is shown in Figure 111. The Klara Level is driven 330 m. above sea level. The cross-cut is driven a total of 220 m. across the strike. There are three drives south from this cross-cut, the eastern one with traces of cobalt mineralisation, the central one impassable after 25 m. without a rubber dinghy and the western one leading to a ladder raise giving access to the stopes developed between the Klara and Ludwig Eugen levels. The timbering was replaced by the Germans in 1942 and is still in good condition. There are two drives north from the Klara Crosscut, the eastern one is only driven a distance of 25 m. The western drive north is very extensive and stretches to the North Mines, a distance of 800 metres. The first 600 m. of this are still accessible. From this main drive north access to the various open-casts can be gained and the 'Weisscobalt Strossen' can be reached. The geology of the Klara Level is shown in Figure 112. The Ludwig Eugen Level is driven at 270 m. above sea level. The cross-cut is driven 900 metres from the portal at Stolldammen to where it intersects the ore zone beneath the Central Mine. A heading has been driven north in the ore zone a distance of 360 m., a similar drive goes south for 1,180 m. Ore chutes lead into these drives from the overlying stopes. The geology of the Ludwig Eugen Level is shown on Figures(113,114,115).



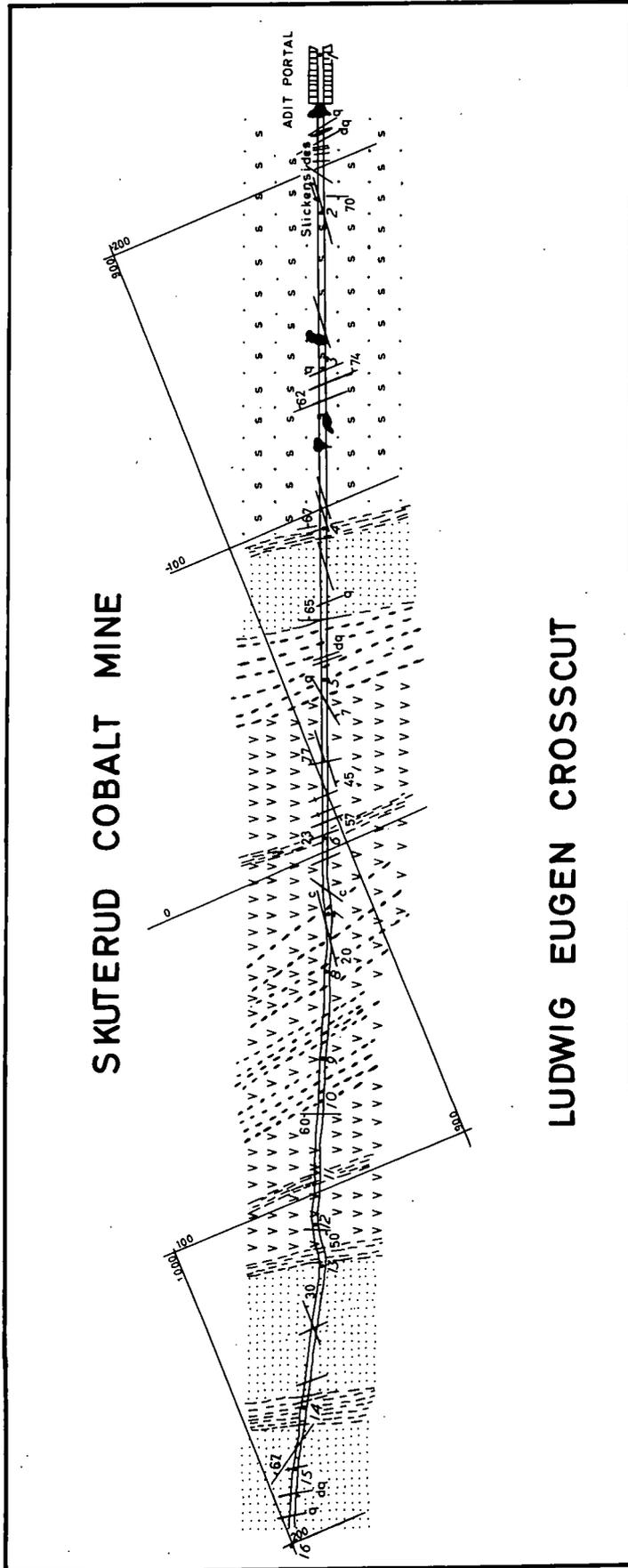


Fig. 113

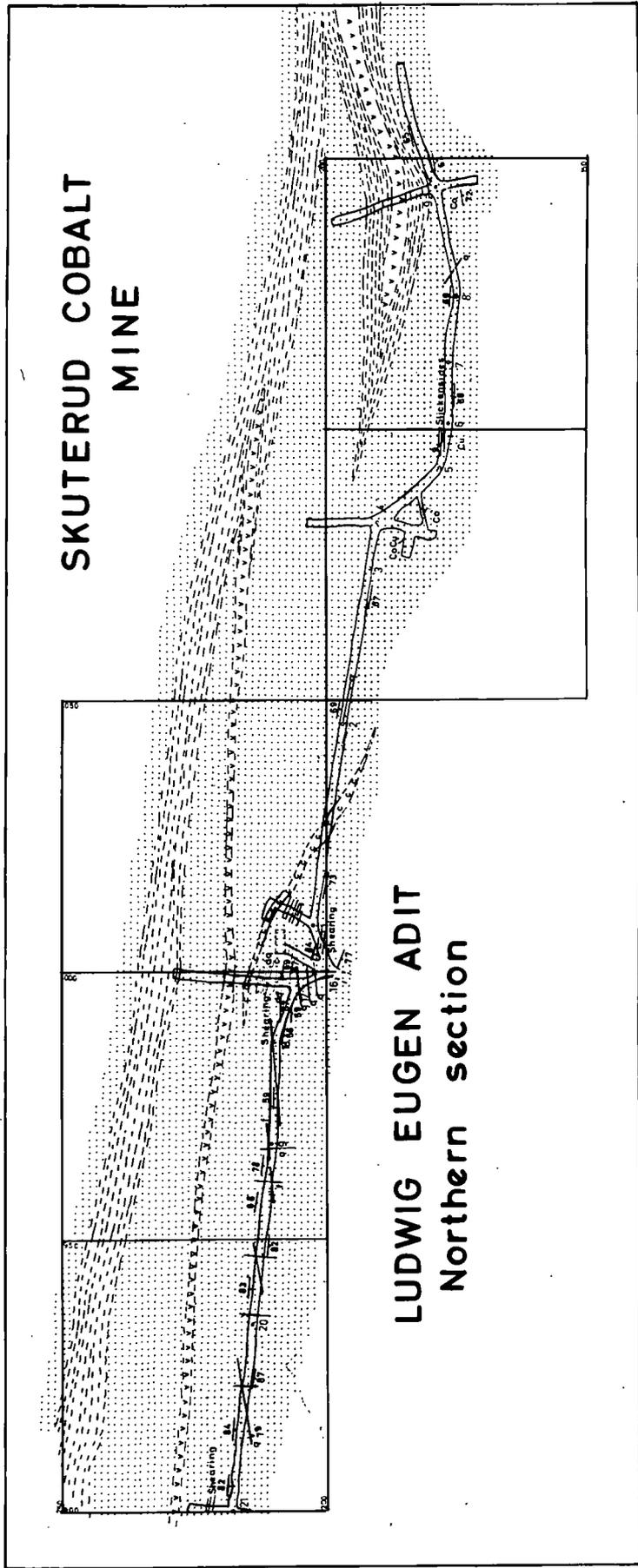
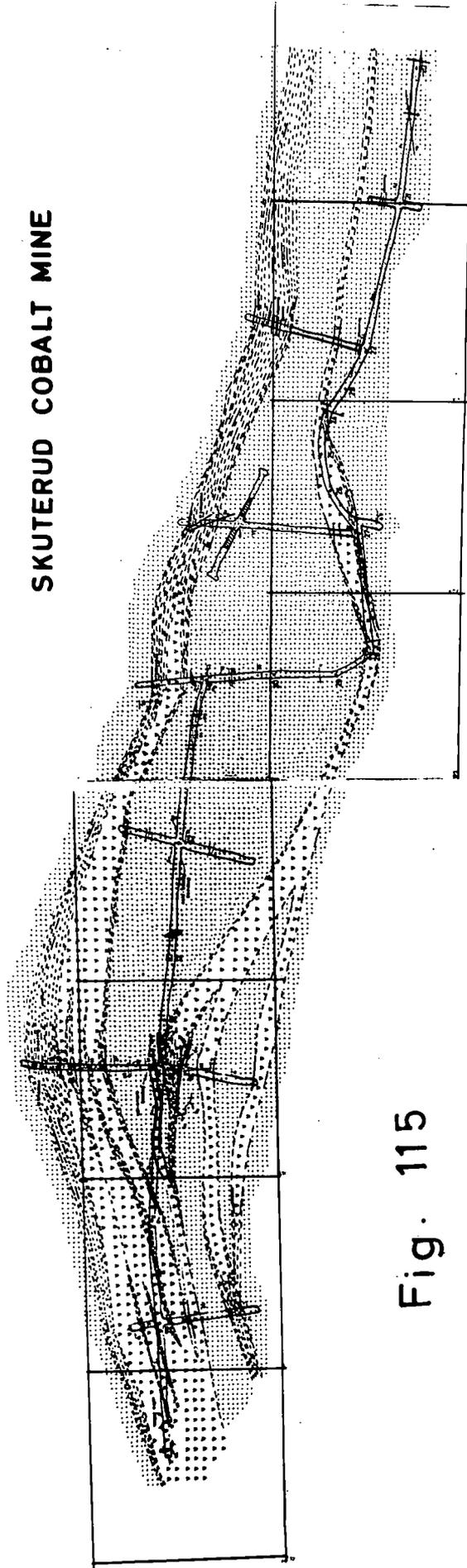


Fig. 114

SKUTERUD COBALT MINE



LUDWIG EUGEN ADIT
Southern section

Fig. 115

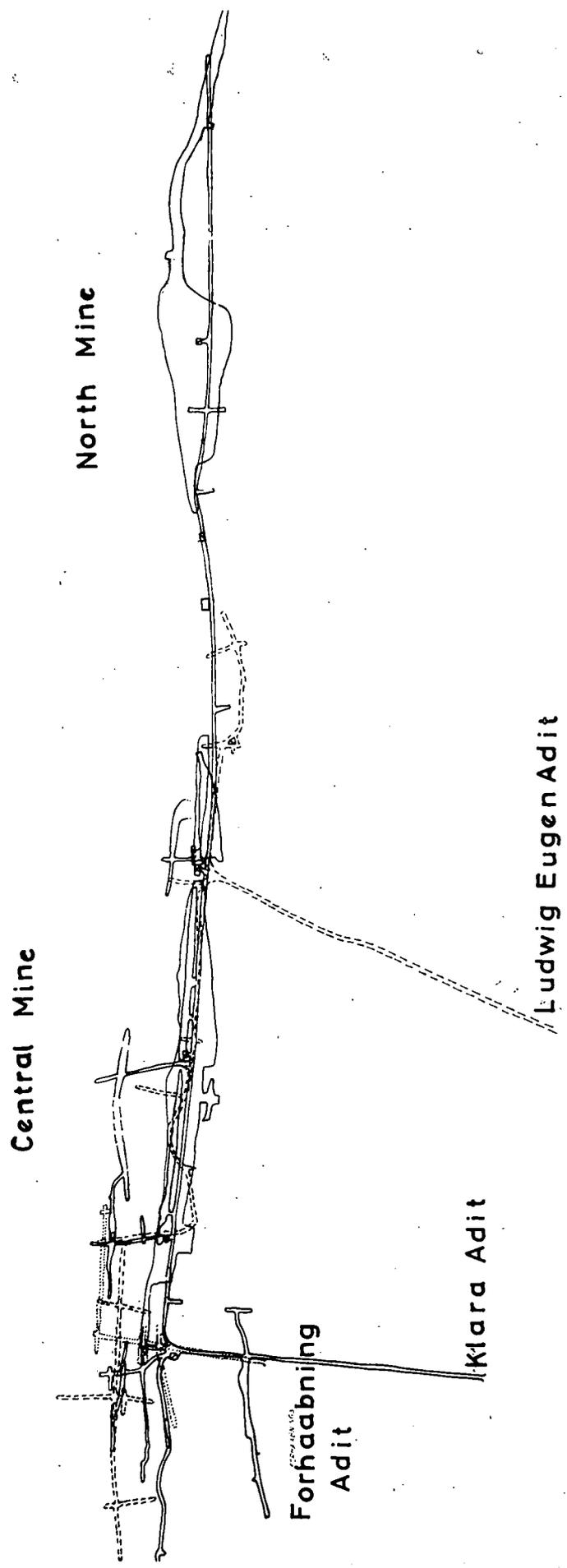


Fig. 116

The ribbon-like nature of the mine plans has necessitated the division of the geological maps of the various levels into several sections. A synoptic plan of the various workings is presented (Figure 11b) showing the mine co-ordinated system to aid in locating any particular map.

The Muggerud Trial This is the most southerly of the workings in the western zone. It is a small open-cast (15 x 3 metres) opened in sulphidic siliceous granulite. No cobalt minerals were observed. Halvorsen (1954) states that the cobaltiferous zone continues south of this point. In a letter in the local paper he states (author's translation);

"The cobalt 'vein' continues from Muggerud to Hugget in Overn Wood. There lived a man called Kristian Madsen who worked at the mines. While digging a drainage ditch he encountered the 'cobalt vein'. He rapidly filled the ditch in because he was afraid that the mine owners would hear of it and start production, in which case he would have to move from the house which he rented from them. In a house a little further north lived a man called Knut. He was a craftsman who made skis and other wooden articles. He said that the 'cobalt vein' was in his cellar. The house is now derelict and grown over so it is impossible to see if there is anything there."

Both of these localities are in amphibolite marginal to an olivine-gabbro body.

Mode of occurrence

The mines were worked by primitive methods, which though expensive were extremely efficient. In consequence there is very little ore left in situ in the workings. It is only visible in the occasional pillars left as supports in the stopes. There are three approaches to deciphering the mode of occurrence of the ore, contemporary descriptions in the literature, study of museum specimens and investigation of the ore remaining in situ.

Contemporary descriptions in the literature In these descriptions the main mineralised zone is referred to as a 'cobalt-fahlband' because of the common association of the cobalt ore with the

sulphidic schist layers. The descriptions in the literature only deal with the Skuterud Mines as these were economically, the most important.

The cobalt was found as streaks and impregnations paralleling the strike of the country rocks. These zones were referred to as ore horizons (ertsbaander) in which the ore occurred as very fine grained impregnations and alternated with barren horizons (felsbaander). The ore horizons varied in thickness from a few centimetres to a maximum of 16 metres, but were usually between 4 and 8 metres wide. They passed over to barren horizons through a decrease in the amount of ore disseminated in the rock. These intervening barren zones also varied in thickness from a few centimetres to between twenty and thirty metres. These wider zones of barren rock were left in situ as strengthening pillars and can be seen in the underground stopes separating the worked out zones. It was necessary to include the thinner of these barren zones during the mining. Voigt (1883) states that of the mined rock only 65% could be considered to be ore grade.

It is stated that these ore horizons maintained their identity over considerable distances, both horizontally and in depth. Only one case is cited of two of the ore horizons joining (Central Mines - Stopes 11 and 13) and where they did the ore was richer than usual. Bøbert (1848) states that the lithology of the cobaltiferous bands did not differ from that of the barren horizons apart from the presence of cobalt minerals. He states that when he arrived the

workings had only reached a depth of 16 metres. From his experience at Vena in Sweden he suspected that the ore would continue at depth. He tested this hypothesis by driving the Ludwig Crosscut 50 m. below ground level at the North Mines and intersected ore. He followed this venture with the even lower Forhaabning Crosscut at the Central Mine and struck a very rich ore lens. The Bennecke Adit was then driven in at the South Mine and also intersected ore. At the time he was writing (1848) ore had been proven to a depth of 120 metres and showed no signs of dying out. Lammers (1849b) commenting on the nature of the ore lenses says that many times it was thought that the deposit had been worked out but further exploration in depth always revealed fresh ore.

Within the ore horizons occasional lenses of compact ore were found. These usually consisted of a granular aggregate of cobaltite and cobaltian-arsenopyrite and were rarely more than 2 cms. wide or 0.5 metres long. These rich ore lenses (rikertsbaander) conformed to the prevailing strike and were often rediscovered at the same horizon. The discovery and exploitation of these rich bands was an important factor in the development of the deposit which would not otherwise have been an economic proposition. One such band of rich ore was exploited in the years 1854-1857 in stopes 7, 8 and 13 of the Central Mine, Skuterud. In places compact cobaltite was found in lenses of "0.5-1 ell" thick and with a length of "2-8 ells" (1 ell = 0.627 metres). Where the massive cobaltite lensed out the rich ore continued as a 0.2-1 m. wide zone richly

impregnated with cobaltite. This changed laterally to a finer grained, normal ore horizon, containing cobaltite, cobaltian arsenopyrite, chalcopyrite, and pyrrhotite. During the mining it was possible to forecast that massive ore would reappear when the rich impregnations of cobaltite were encountered at a likely horizon. Bands of massive ore were particularly common in the Central Mine and much rarer in the South and North Mines.

The mineralisation was described as occurring in all the major rock types of the mining area, including the amphibolitic bands. The most frequent host rock was a fine-grained, foliated quartz-rich phlogopite schist. The ore bands always conformed to the dip of the enclosing rocks, no traces of folding were observed. The late pegmatites cut across the ore zone, Bøbert (op.cit.) states that if an ore band was terminated by such a pegmatite he persevered with the stoping, as the same band was usually found again on the other side of the pegmatite.

Cobaltite was the most important of the ore minerals, it was quite common and contained about 33% cobalt. It usually occurred as impregnations of very small, euhedral crystals the size of which "rarely surpassed that of a pea". Rarely crystals of up to 2cms. diameter were recorded. The cobaltite was particularly common in the rich ore lenses of the Central Mine, an important factor in making this the major cobalt producing working.

Cobaltian arsenopyrite was the most abundant of the ore minerals after cobaltite. As it only contained about 8.5% cobalt much larger quantities were required. Very little was found in the Central Mine, but in the South and North Mines it was quite common while still subordinate to cobaltite, at the Saafstad Mines the

minerals were present in roughly equal proportions. At the northern end of the western zone and in all the eastern zone cobaltian arsenopyrite was the only ore mineral present. The low cobalt percentage of the mineral soon made these workings an uneconomic venture.

Skutterudite, which contains about 20% cobalt was very much rarer than the cobaltite and cobaltian arsenopyrite. It was only found in the Central Mine at Skuterud, being particularly abundant in the 'Weisscobalt Strossen' (White cobalt stope).

Chalcopyrite, a common constituent of the normal fahlband sulphide assemblage, was also present in the cobalt ore lenses. Lenses particularly rich in chalcopyrite were worked to produce both cobalt and copper concentrates.

The ore was graded as to quality when it arrived at the Haugfoss pigment factory according to the following scheme:

Ordinary ore Disseminated cobaltian arsenopyrite with varying amounts of cobaltite but no other ore minerals.

Cupriferous ore The same as 'ordinary ore' but containing enough chalcopyrite to pay for the extraction of a copper concentrate.

'Gulerts' Massive lenses of cobaltite and cobaltian arsenopyrite.

Rich ore Massive lenses consisting essentially of cobaltite alone.

Museum specimens Some excellent specimens of the ore from Modum are housed in the collections of the Mineralogisk-Geologisk Museum, Oslo. The author was able to examine these through the courtesy of Dr. H. Neumann, Chief Curator. Unfortunately they are only labelled as coming from the "Modum cobalt mines", with no further information

on the localities at which they were collected.

The host rock of the majority of these specimens is a grey micaceous quartzite containing phlogopitic bands. Pegmatite and diopside skarn horizons are occasionally present.

The ore minerals, which are mainly cobaltite with very subordinate arsenopyrite, are typically developed as fine grained disseminations throughout the rock. This ore impregnation is noticeably richer in some of the horizons. Small lenses, rarely exceeding a centimeter in length are developed parallel to the foliation, in exceptional cases the ore forms a compact band varying in width between a few millimetres and 2 cms. A photograph of one of the larger Museum specimens shows a very wide band of massive ore (Figure 117), in the same specimen cobaltite is also preferentially developed at the margin of a diopsidic band (Figure 118).

In many of the specimens the cobalt minerals are closely related to small quartz lenses which lie parallel to the foliation, in these lenses the ore minerals attain a size up to 2 cms. (Figure 119). A similar relationship is shown by some specimens where the cobalt minerals are preferentially developed in the gaps between these quartz lenses. This texture is very reminiscent of that commonly seen, on a larger scale, in metamorphic terrains where pegmatitic minerals have developed in the 'low-pressure' areas between boudins (Figure 120).

Many examples are seen where the ore minerals are particularly well developed at one horizon. This is exemplified by the relations shown in Figure 121, where cobaltite is developed in

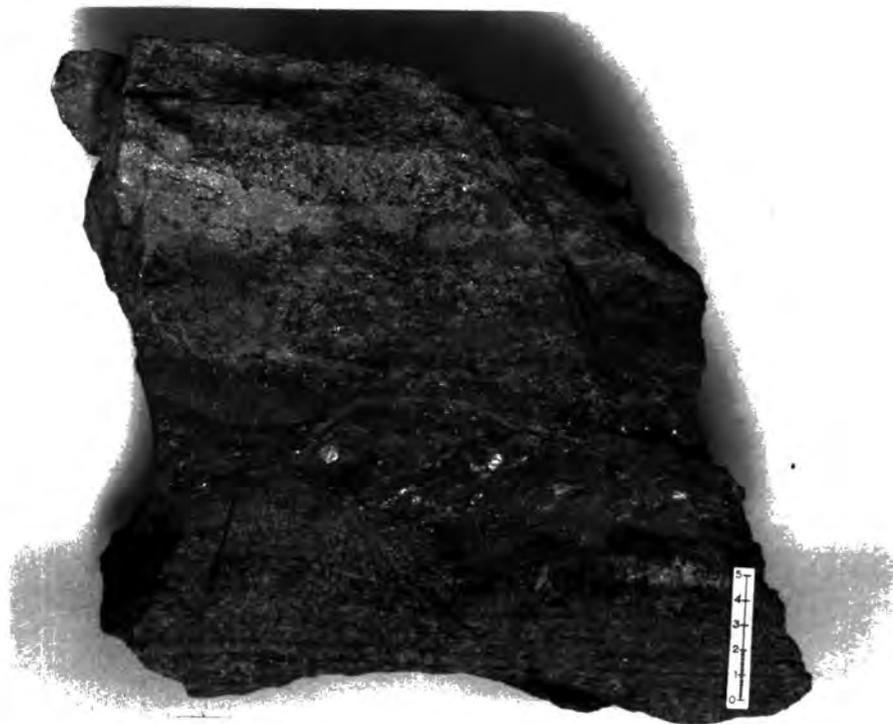


Figure 117: Large specimen of ore from the Skuterud Mines in the collections of the Mineralogisk Museum, Oslo, a band of massive cobaltite ore is visible near the top of the specimen. (Scale in centimetres).

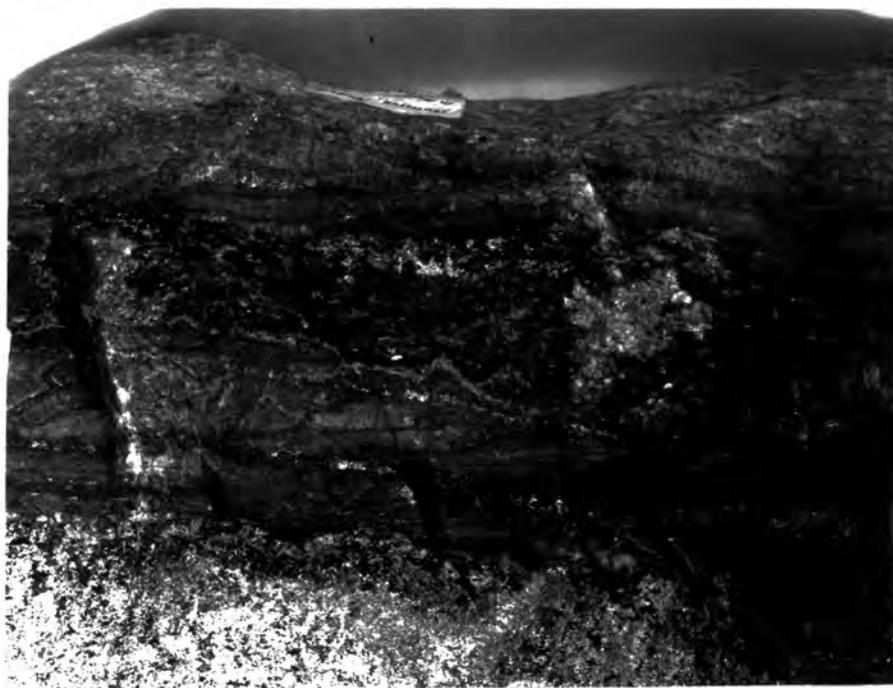
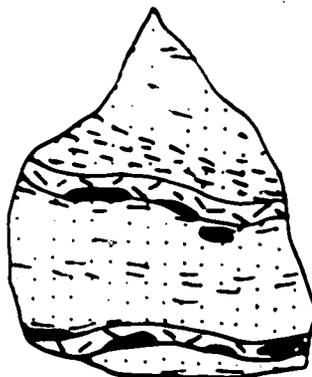
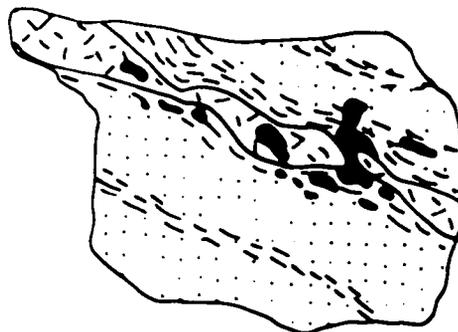


Figure 118: Detail of the same specimen showing the concentration of the ore minerals at the margin of a diopside skarn lens.

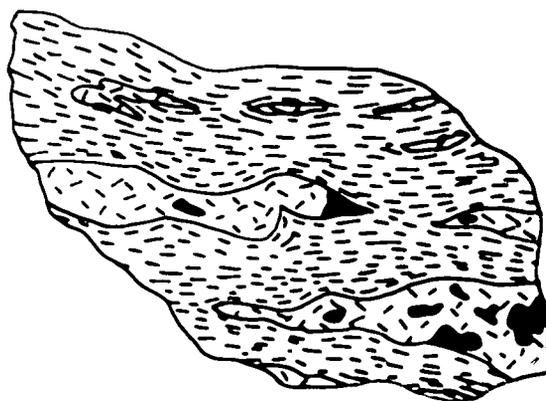
Figure 119: Sketches of ore specimens in the collections of the Mineralogisk-Museum, Oslo. The cobalt minerals (black) are seen to be particularly associated with quartz rich layers and lenses which have developed in the plane of the foliation.



1cm



1cm



1cm

Fig. 119

Figure 120: Sketches of ore specimens from the collections of the Mineralogisk-Museum, Oslo. The cobalt minerals (black) are particularly associated with the gaps between adjacent quartz lenses.

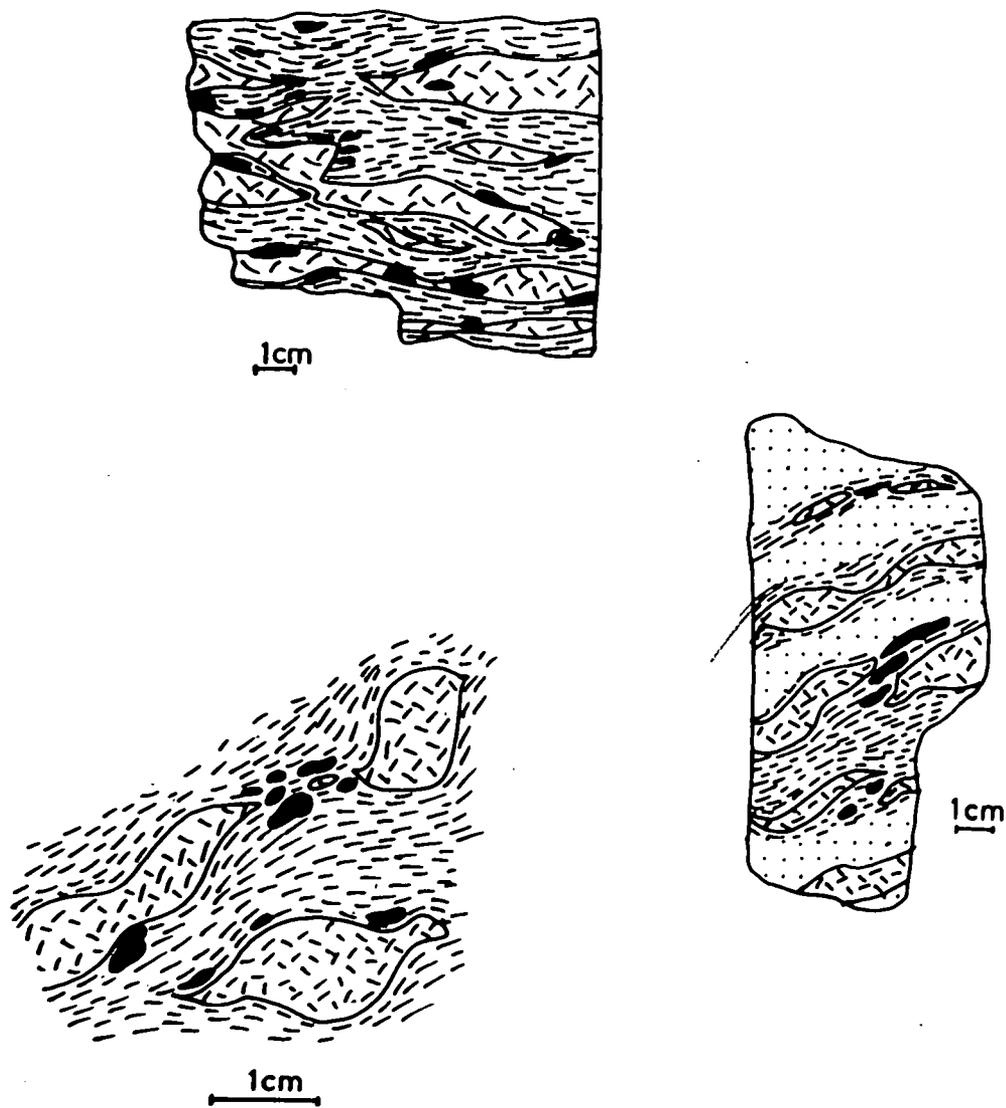
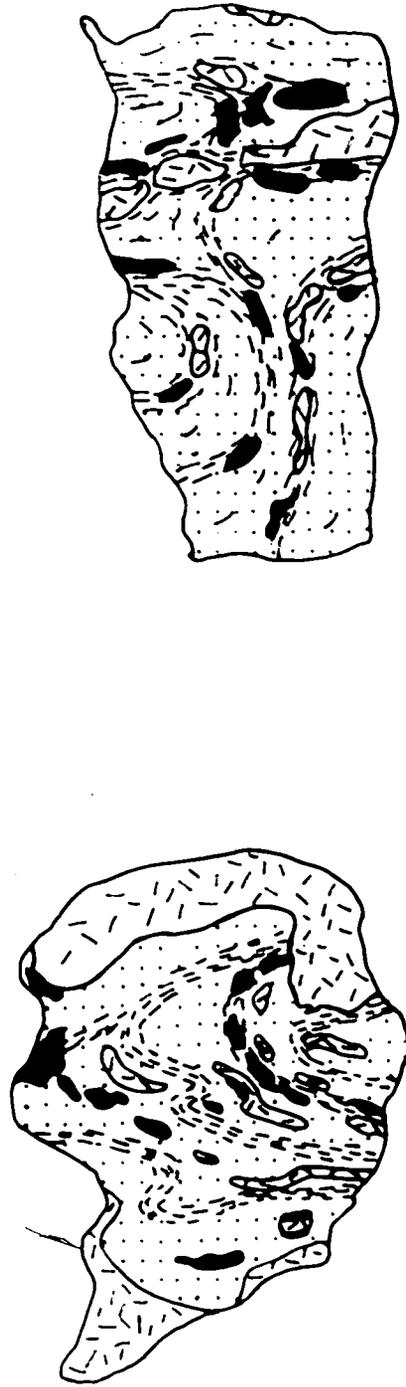


Fig. 120

Figure 121: Sketches of ore specimens from the collections of the Mineralogisk-Museum, Oslo. The cobalt minerals (black) appear to be concentrated in bands which are concordant around minor fold structures.



1 cm

Fig. 121

micaceous bands which can be traced around the closure of a minor fold.

One small lens of massive ore is shown in Figure 122. A zonal arrangement of the minerals is noticeable, large crystals of cobaltite and skutterudite are at the centre of the lens and are surrounded by a fine-grained intergrowth of cobaltite, arsenopyrite and skutterudite. The envelope of the lens is formed of arsenopyrite which here has a prismatic habit, the long axes of the crystals being parallel to the long axis of the lens. The lens lies conformably in the foliation of the host rock.

Ore 'in situ' Cobalt mineralisation is visible in situ in so few places in the mining district that it is worth listing these localities. Nowhere in the eastern zone were cobaltiferous minerals observed, while in the western zone the localities are restricted to those mines which have been most extensively developed, i.e. the Svartefjell and Skuterud Mines.

At Svartefjell mineralisation is visible on the eastern wall of the southern open-cast working (Figure 103 co-ordinates 80/10). The mode of occurrence of the ore is illustrated in Figure 104. Cobaltian arsenopyrite is sparsely disseminated in micaceous quartzite in a major movement zone, erythrite staining is particularly noticeable in a strongly sheared biotite-schist.

In the North Mine at Skuterud a rich cobalt mineralisation is visible in a pillar at the north end of the open-cast working. At this point (Figure 110 co-ordinates 1310/250) the open-cast workings narrow from 50 metres to 5 metres. To the north of this point adits have been driven at the Ludwig Level for short distances in several directions, presumably in a search for further ore. At the surface above this point is a wide outcrop of amphibolite and biotite-schist, while at the Ludwig Level the workings are in quartzite and quartz-mica-schist. The field relations suggest that

Figure 122: Sketch of small lens of cobalt ore from the collections of the Mineralogisk-Museum, Oslo. The cobalt minerals appear to have a zonal arrangement with cobaltite and skutterudite in the centre, followed by a zone of cobaltite, arsenopyrite and minor skutterudite, surrounded in turn by a zone of arsenopyrite. A similar zonation, on a regional scale, is observed in the Skuterud mining district.

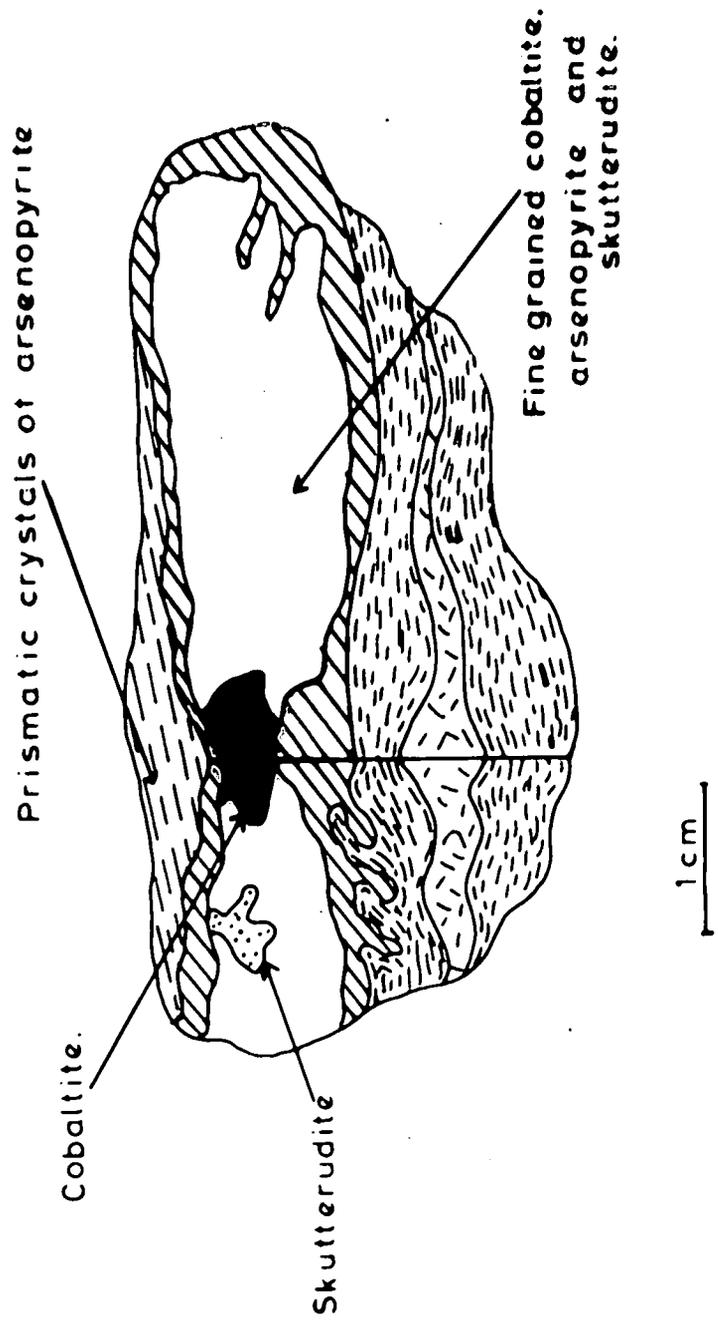


Fig. 122

this area may be at the hinge of an antiform. Mineral lineations suggest that the tectonic 'b' direction is plunging north in this region. Workings continue further north at the lower Klara Level and then also die out. The mineralisation in situ consists of disseminated cobaltite and arsenopyrite in a poorly foliated quartzite (Mode 461: Table 3). The cobaltite crystals are euhedrally developed and attain a size of 3 mm. diameter. Minor chalcopyrite is associated with the cobalt minerals. Erythrite and malachite are extensively developed along joint surfaces.

The Central Mines at Skuterud contained the richest ore lenses, traces of the mineralisation can be seen at six localities in the workings.

At the intersection of the Forhaabning Crosscut with the southern extremity of the open-cast workings (Figure 110 co-ordinates 708/260) cobaltite and cobaltian arsenopyrite are disseminated in a white saccharoidal quartzite. Figure 111 shows the geology of this part of the workings, the cobaltiferous quartzite is separated from an amphibolite to the west by a movement surface dipping east at about 80° . Malachite staining is associated with this movement zone. To the east the pure quartzite gives way to micaceous quartzite containing graphitic and sulphidic zones which have a well developed fahl weathering. Uraninite and chalcopyrite are also disseminated in the quartzite at this locality.

Erythrite staining is visible on the east wall of a stope situated slightly west of the southern extremity of the open-cast, between the Forhaabning and Klara Levels. (Figure 110 co-ordinates 730/265). The erythrite is present on foliation planes and joint surfaces of a very micaceous quartzite. Shear zones are rather common at this locality and follow the trend of the foliation. Some of the shear zones have a thin film of calcite and associated chalcopyrite and minor galena. No primary cobalt ore minerals were visible but broken ore is still present on the floor of the stope, showing arsenopyrite and cobaltite disseminated in white, saccharoidal quartzite. The cobaltite rarely exceeds 3 mm. in diameter but the arsenopyrite crystals attain a maximum size of 1.5 cms. and occasionally occur in small lenses reaching a maximum length of 4 cms. More micaceous quartzite is also present containing a finer grained impregnation of cobaltite, particularly concentrated in the more micaceous bands.

Mineralisation is visible at three places in the Klara adit system. In the roof of the first drive south from the Klara Crosscut (Troeger Stope, Figure 112) about twenty small lenses of cobaltiferous mineralisation are visible. These are developed in coarse grained biotite-schist at the termination of an amphibolite band. The lenses consist of a fine grained impregnation of arsenopyrite with minor cobaltite and chalcopyrite. They are conformable to the foliation and reach a maximum strike length of 0.5 m. and are never more than 5 cms. wide. Erythrite and malachite are well developed. The geology of part of the Troeger Stope is shown in Figure 105.

In the long drive north at the Klara Level erythrite is patchily developed on the walls of the adit. A cross-cut west from this drive (Figure 112 280/770) contains small lenses of cobaltite in a fine grained micaceous quartzite interlayered between two amphibolite bands.

Access to the Weisscobalt Strossen (Figure 110 870/240) is also gained from the Klara Level. The stope is in a white, saccharoidal quartzite containing a rich dissemination of arsenopyrite with minor cobaltite and skutterudite. The quartzite contains many small pegmatitic lenses of quartz and feldspar with cobalt minerals attaining a maximum size of 2 cms. diameter. In the normal quartzite the average size of the crystals is about 5 mm. Graphitic bands are occasionally present in the quartzite. Diopside-actinolite skarn lenses are present at one horizon, these contain large aggregates of chalcopyrite with minor arsenopyrite. The cobalt minerals are occasionally observed in small stringers and veins transgressing the foliation. Calcite-galena veins are present along some of the movement surfaces, these crosscut the cobalt mineralisation.

The Ludwig Eugen Level was used mainly as a transport adit and very little development has been carried out. However, disseminated cobaltite is visible in one of the crosscut into the ore zone (Figure 115 870/270). The host rock is a tourmaline rich quartzite with tightly appressed isoclinally folded banding. Erythrite and malachite are abundantly developed on the walls of the adit at this locality.

Conclusions From these various lines of evidence it is possible to gain some idea of how the cobalt mineralisation occurred. The available textural evidence suggests that the ore was emplaced before the final metamorphic episode which was responsible for producing such features as: the conformability of the disseminated ore bands; the elongation of individual grains and lenses parallel to the foliation; the association of the ore with quartz-segregation lenses, segregation pegmatites and diopside skarn lenses and the conformability with folded bands in the host rock.

The structural control of the mineralisation seems to be mainly the metamorphic foliation of the host rocks. The individual stopes are 'cigar-shaped' with their long axes parallel to the lineation in the adjacent tectonites suggesting that the local 'b'

tectonic direction during deformation was also an important controlling factor. The ore lenses are cut by the post-tectonic pegmatites.

The host rocks of the mineralisation are rather varied. Despite the designation of the deposit as a 'cobalt-fahlband' in the literature it must be emphasised that cobalt mineralisation has nowhere been observed in a typical sulphidic graphite-schist. The host rock of the mineralisation in the Museum specimens is usually a quartz-phlogopite-dravite-schist with varying amounts of muscovite, sillimanite and felspar. The most important cobalt mineral in these specimens is cobaltite. The mineralisation preserved in situ is usually in a saccharoidal, white quartzite containing very minor phlogopite and dravite, with cobaltian arsenopyrite as the main ore mineral.

Figure 123 is a reproduction of one of the old mine plans showing the mineralised areas. It is interesting to note that the major workings at the Central Mine, Skuterud were developed on the 'Glanz koboltzonen' - the cobaltite zone, while parallel to this on the west lay the 'Weiss koboltzonen' - the cobaltian arsenopyrite zone. There is a strong possibility that the museum specimens are representative of the richer cobaltite band while the mineralisation visible in the Forhaabning crosscut, the Central Mine open-cast, the 'Weisscobalt Strossen', the ["]Troeger stope and the North Mine are representative of the cobaltian arsenopyrite band.

If this is the case then it is possible to recognise a zonal distribution of minerals in the deposit, with cobaltite, and possibly skutterudite, forming the central zone surrounded by an

envelope rich in cobaltian arsenopyrite. Thus the main stopes of the Central Mine are in a cobaltite rich area (Glanzkoboltzonen) which is flanked by the arsenopyrite rich areas of the 'Weisskoboltzonen' to the west and the Troeger Stope to the east. The cobaltite gives way to arsenopyrite at the South and North Mines. Arsenopyrite is the dominant mineral in the rest of the western and eastern mineralised zones. This zonation is similar to that exhibited, on a very much smaller scale, by the lens of ore shown in Figure 122.

This hypothesis leads to interesting speculations as to the hypothetical continuation of the cobaltite zone at depth under the North Mine (Figure 124).

The cobalt ore minerals

Cobaltite This is the commonest ore mineral in the Museum specimens and in those collected from the Central Mines, Skuterud. It is typically developed in subhedral and euhedral crystals disseminated in a silicate gangue. The silicates appear to be corroding the boundaries of the cobaltite grains, silicate inclusions are commonly found at the centres of the crystals, (Figure: 125).

In the higher grade ore the cobaltite crystals are in a loose aggregation with silicate minerals separating the individual grains (Figure 126). In the massive ore lenses the grains are tightly pressed together in aggregates with the development of polyhedral grain boundaries (Figure 127).

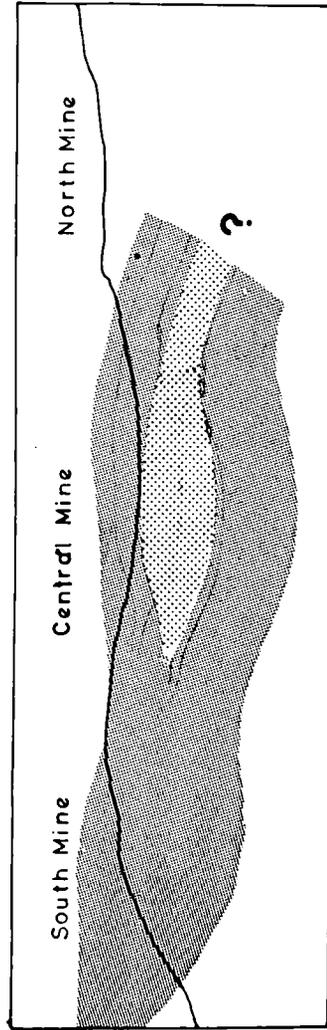
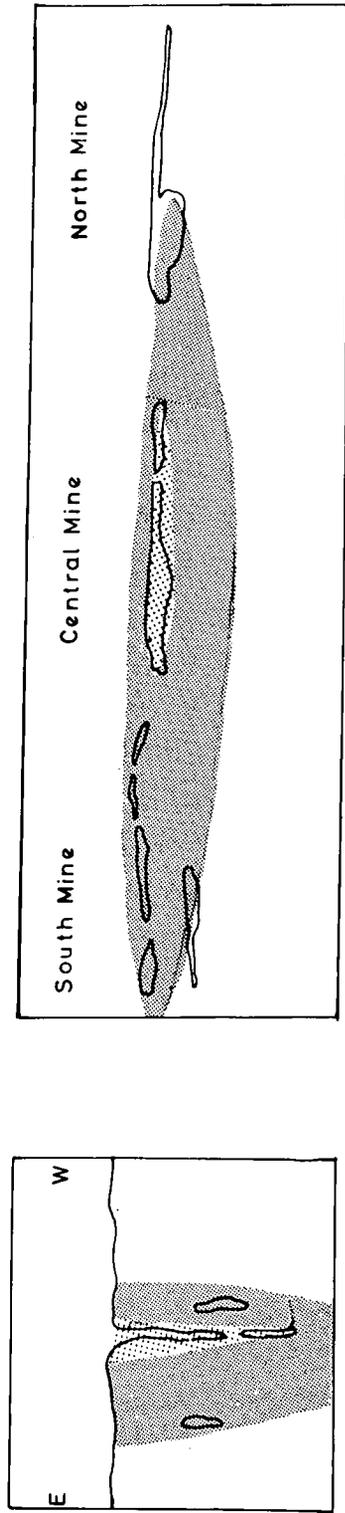


Fig. 124

HYPOTHETICAL MINERAL ZONATION AT SKUTERUD



Figure 125: Cobaltite corroded by gangue and containing silicate inclusions. Specimen from Mineralogisk-Museum collections. Ordinary reflected light. X 50.



Figure 126: Loose aggregation of cobaltite in quartzite. Specimen from Mineralogisk-Museum collections. Ordinary reflected light. X 50.

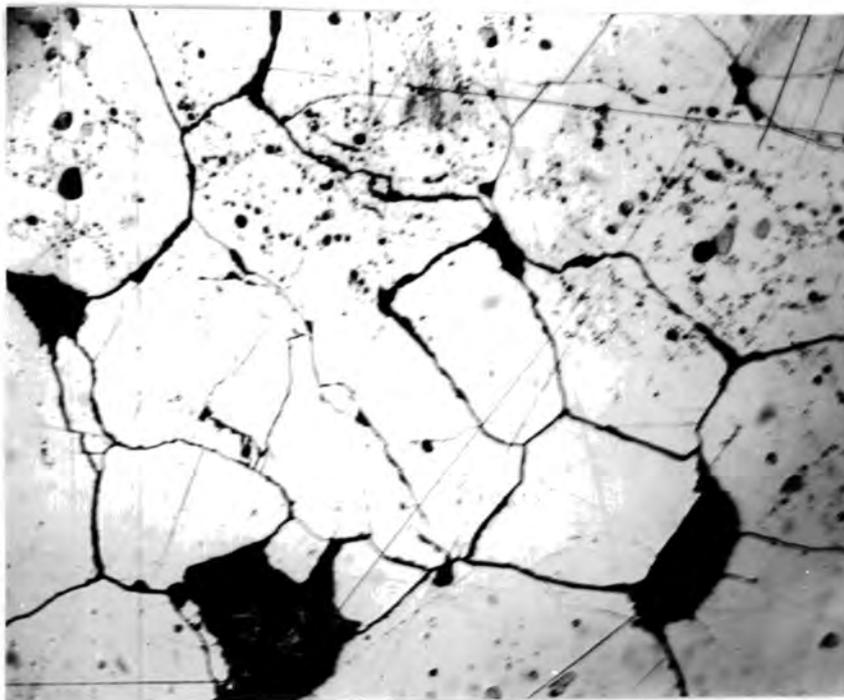


Figure 127: Polyhedral aggregate of cobaltite, massive ore lens in siliceous granulite. Specimen from Mineralogisk-Museum collections. Ordinary reflected light. X 50.

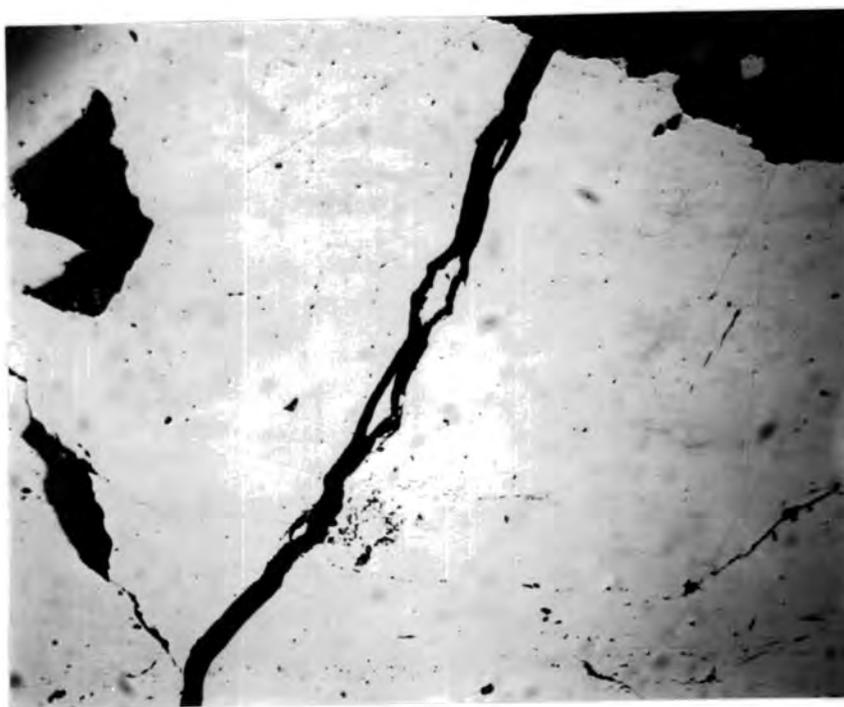


Figure 128: Cataclastic cracks in cobaltite filled by silicate material. Note the corresponding walls of the crack indicating that little or no solution of the cobaltite has occurred. Ordinary reflected light. X 50.

The crystals have suffered cataclasis and are crossed by many anastomosing cracks. In one specimen of ore from a movement zone, which shows a highly polished and slickensided surface, the cobaltite is strongly cataclastically deformed and comminuted into a fine powder adjacent to the movement zone.

Other minerals are found filling the cracks developed in the cobaltite. Most commonly the infilling is of silicate minerals, with no replacement of the cobaltite being visible, the opposite faces of the cracks correspond exactly (Figure 128). In some cases projecting corners of the cobaltite have been broken off and incorporated in the veining silicate (Figure 129). However in some cases indications are seen of the replacement of cobaltite by the silicates (Figure 130). The broken crystals are also invaded by sulphide minerals, pyrite is seen veining and brecciating the cobaltite (Figure 131).

Chalcopyrite is frequently associated with the cobaltite. Where the chalcopyrite is the dominant mineral present the cobaltite is developed as euhedra up to 0.5 cms. diameter in a flat, monotonous field of chalcopyrite (Figure 132). When the minerals are present in roughly equal proportions the boundaries are always smooth, even under high magnifications (Figure 132). Where cobaltite is the predominant mineral the chalcopyrite appears as veins between the subhedral crystals. Small inclusions of chalcopyrite are infrequently seen in the cobaltite.

The cobaltite is distinguishable from arsenopyrite by its slightly pinkish white colour against the blue-white of the arsenopyrite. Reflectivity and hardness are of little diagnostic



Figure 129: Brecciation of cobaltite by veining silicate. Specimen from Mineralogisk-Museum collections.

Ordinary reflected light.

X 50.

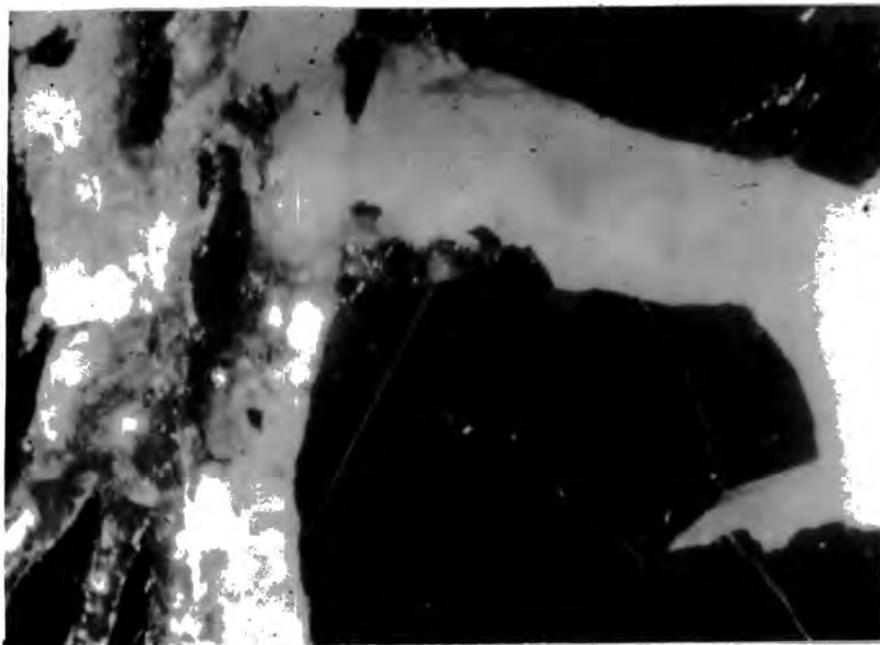


Figure 130: Replacement of cobaltite by silicate gangue. Specimen from Mineralogisk-Museum collections.

Reflected light, crossed nicols.

X 50.

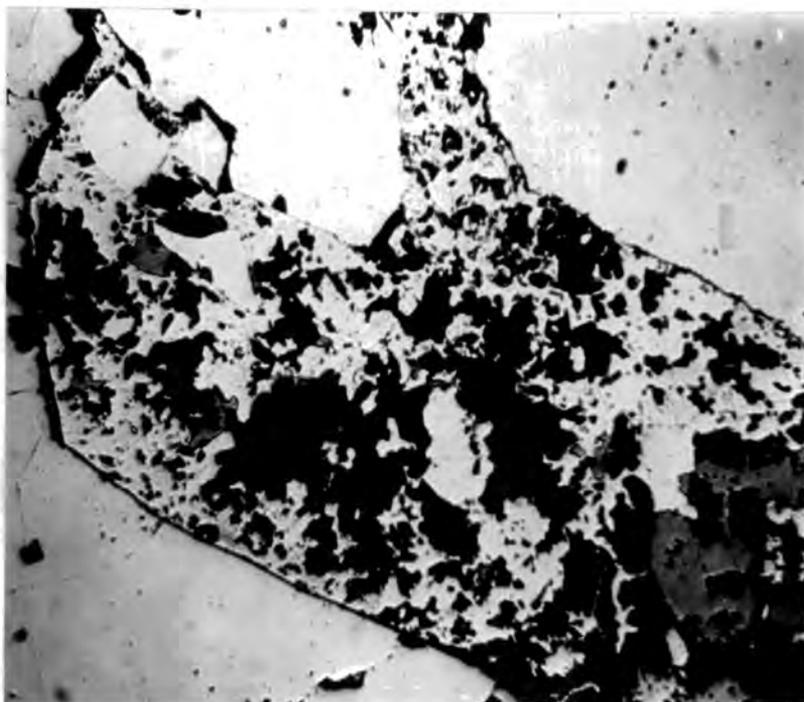


Figure 131: Cobaltite veined and brecciated by pyrite. Specimen from Mineralogisk-Museum collections.

Ordinary reflected light.

X 50.

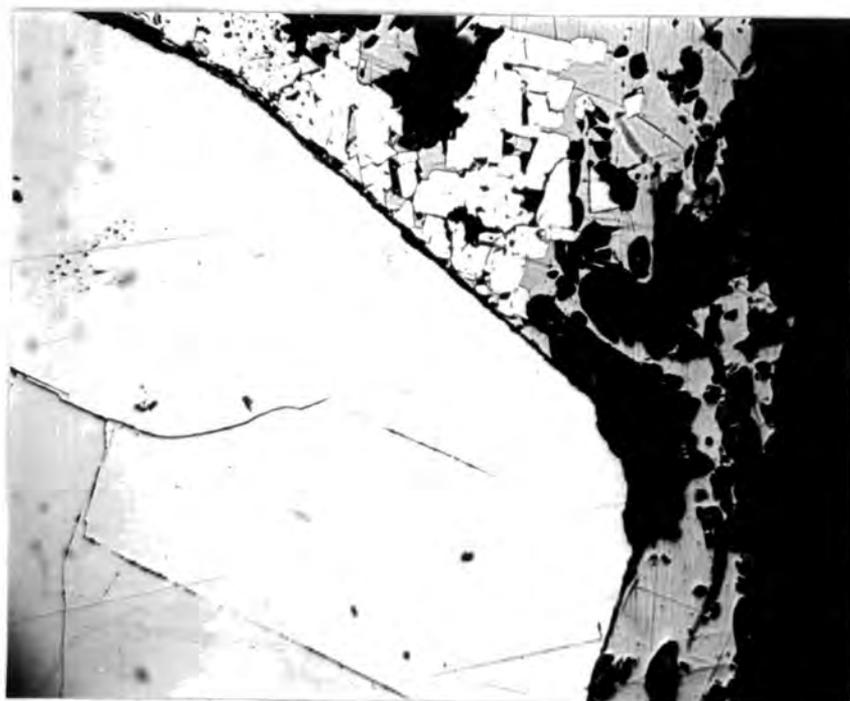


Figure 132: Euhedral crystal of cobaltite in a matrix of chalcopyrite. The boundary between the two minerals is smooth under high magnification. A fringe of pyrite is also developed at this boundary.

Ordinary reflected light.

X 50.

value in routine sample examination. The arsenopyrite has a much stronger anisotropism and is etched by 1:1 nitric acid, which does not affect the cobaltite.

Arsenopyrite is frequently in close association with the cobaltite. They are usually seen as discrete grains in a closely interlocked polyhedral aggregate. However fringes of arsenopyrite have developed on some of the cobaltite crystals (Figure 133). Where these rims become wider they appear as 'veins' between cobaltite crystals (Figure 134). Arsenopyrite is also seen in extremely irregular patches within crystals of cobaltite (Figure 135).

The cobaltite shows a very marked anisotropism. This curious feature in a mineral with such clear cubic morphology is further discussed in Part II of this work. Twinning is well developed in the cobaltite, the twin individuals are feathery lamellae with a preferred orientation parallel to the crystal edges (Figure 136).

Arsenopyrite The Modum arsenopyrite contains varying amounts of cobalt. Cobaltiferous arsenopyrite has been variously designated in the literature as danaite, glaucodot and cobaltian-arsenopyrite. This terminology is further discussed in Part II of this work. As the optical properties of the Modum arsenopyrite give no indication of the cobalt content it is here called simply arsenopyrite.

The arsenopyrite is present in disseminated, subhedral crystals which can attain a maximum length of 2 cms. These appear to be corroded by the surrounding silicate minerals (Figure 137). Where massive lenses of arsenopyrite are observed the grains are closely packed together in a polyhedral aggregate (Figure 138). The occurrence of the arsenopyrite in frequent association with

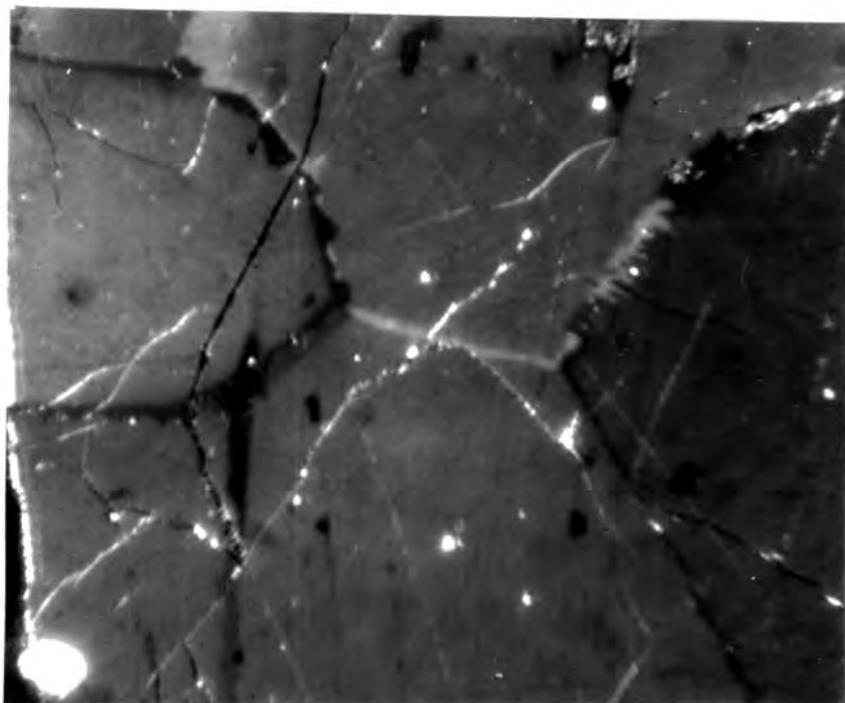


Figure 133: Development of a fringe of arsenopyrite at the borders of cobaltite crystals. Specimen from Mineralogisk-Museum collections. Reflected light, crossed nicols. X 50.

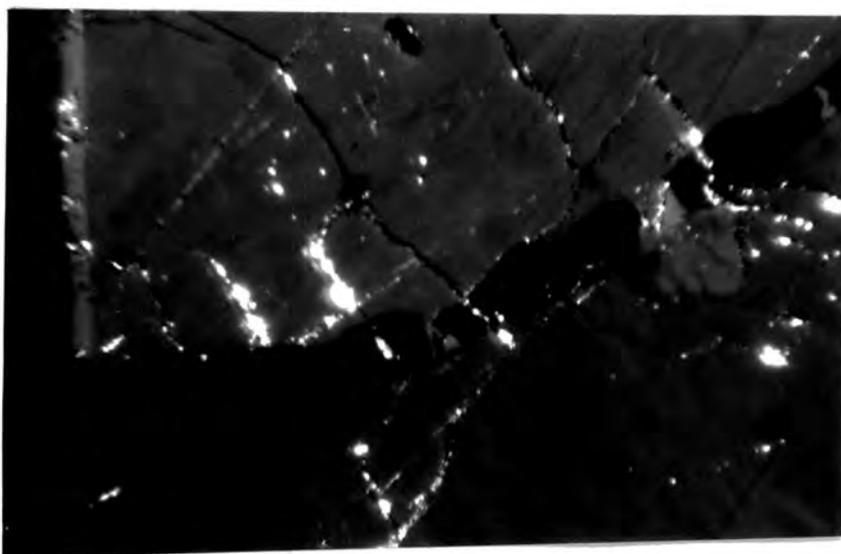


Figure 134: Thicker fringes of arsenopyrite developed at the boundary of cobaltite giving the appearance of arsenopyrite veins. Reflected light, crossed nicols. X 50.

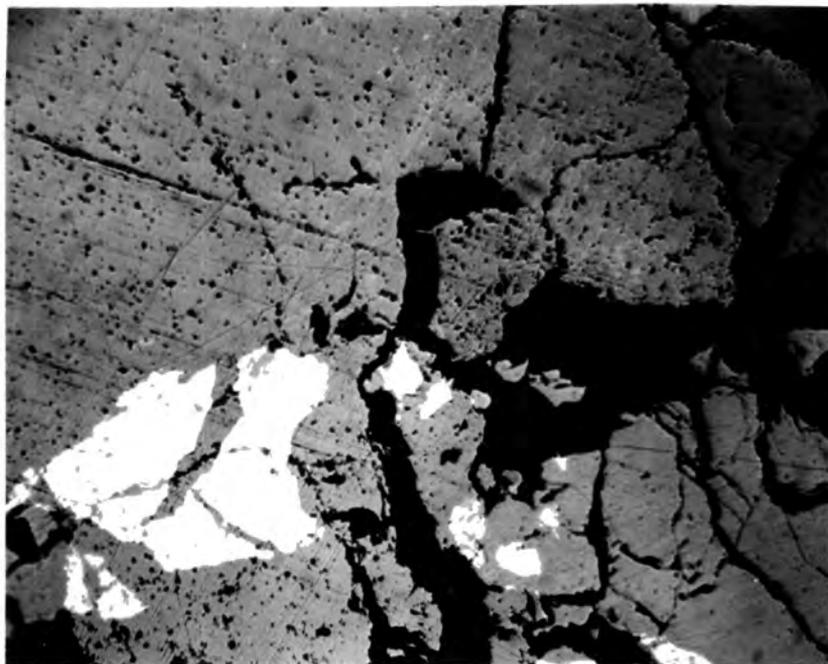


Figure 135: Irregular patch of arsenopyrite in cobaltite. Specimen from Mineralogisk-Museum collections.

Ordinary reflected light.

X 50.

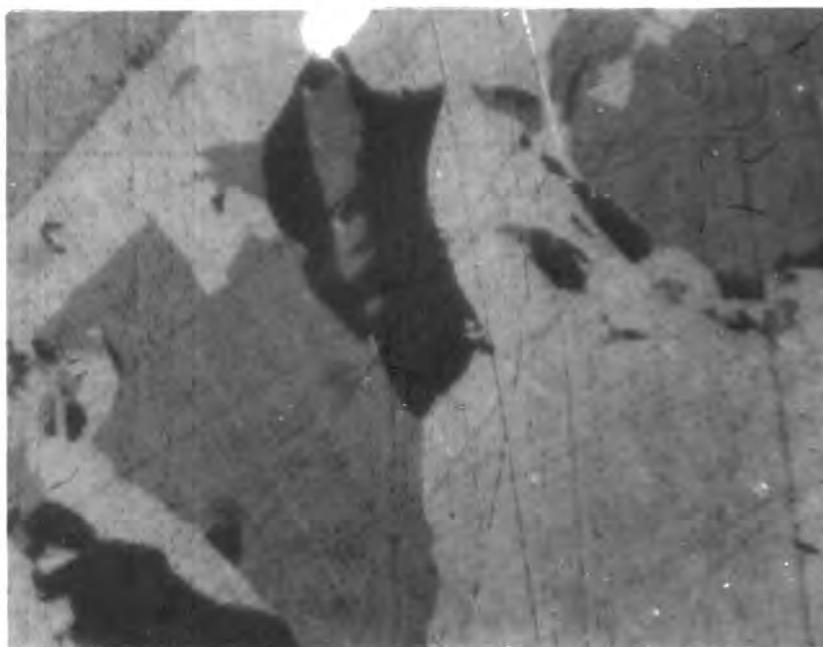


Figure 136: Twin lamellae in cobaltite. Specimen from Mineralogisk-Museum collections.

Reflected light, crossed nicols.

X 50.



Figure 137: Arsenopyrite corroded by gangue silicates, the fragments of arsenopyrite are all in optical continuity. Note the deformed graphite flakes also visible in this photomicrograph. Svartefjell Mine. Ordinary reflected light. X 50.

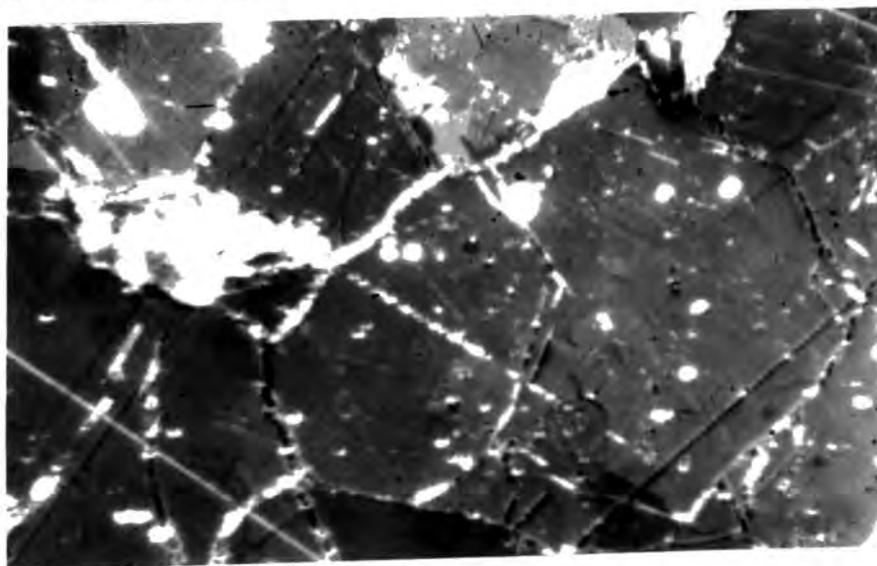


Figure 138: Polyhedral aggregate of arsenopyrite crystals. Weisscobolt Strossen, Central Mine, Skuterud. Reflected light, crossed nicols. X 50.

cobaltite has been commented on above.

In a manner similar to the cobaltite the arsenopyrite has suffered cataclasis. Silicates are seen veining the arsenopyrite along cracks which have exactly corresponding walls (Figure 139), occasionally the arsenopyrite is brecciated by the silicates and small broken off fragments become rotated in the veining silicate mineral (Figure 140). The silicates also appear to be replacing the arsenopyrite in places. Remnants of large original crystals are seen, still in optical continuity, with a matrix of silicate minerals between the fragments, (Figure 137).

The arsenopyrite frequently contains small veins of chalcopyrite (Figure 146), arsenopyrite is present as small, irregularly shaped grains in chalcopyrite rich areas.

The anisotropism is very strong and serves to distinguish the mineral from cobaltite. Twinning is well developed both simple twins (Figure 141) and lamellar twins (Figure 141) are observed.

Some of the larger crystals show an oscillatory zoning at the terminations of prismatic crystal (Figure 142). Some of the zones are distinctly more anisotropic than others, a feature which may be related to differing cobalt contents.

Very small blebs of cobaltite are associated with some of the arsenopyrite. These are clearly seen when the arsenopyrite is etched, leaving the cobaltite unaffected (Figure 143). The blebs are developed in strings parallel to the crystal boundaries of the host arsenopyrite.



Figure 139: Cataclasis in arsenopyrite. Silicate gangue (black) has invaded the arsenopyrite and brecciated it, the fragments of arsenopyrite could be fitted together again to produce an exact fit. Mineralogisk-Museum collections. Ordinary reflected light.

X 50.

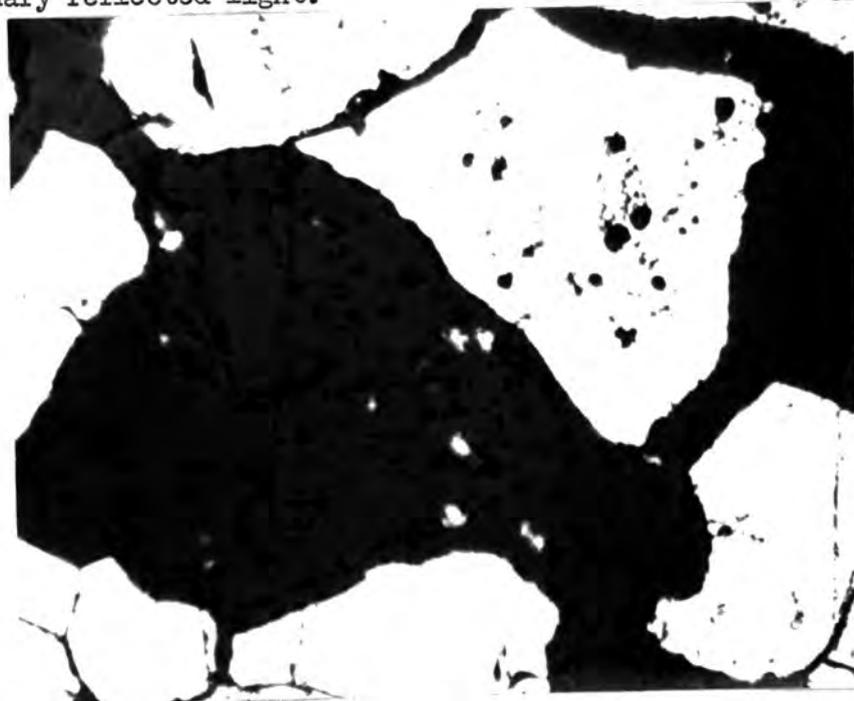


Figure 140: Similar relations to those shown above, with the arsenopyrite fragments rotated in the silicate. Ordinary reflected light.

X 50.

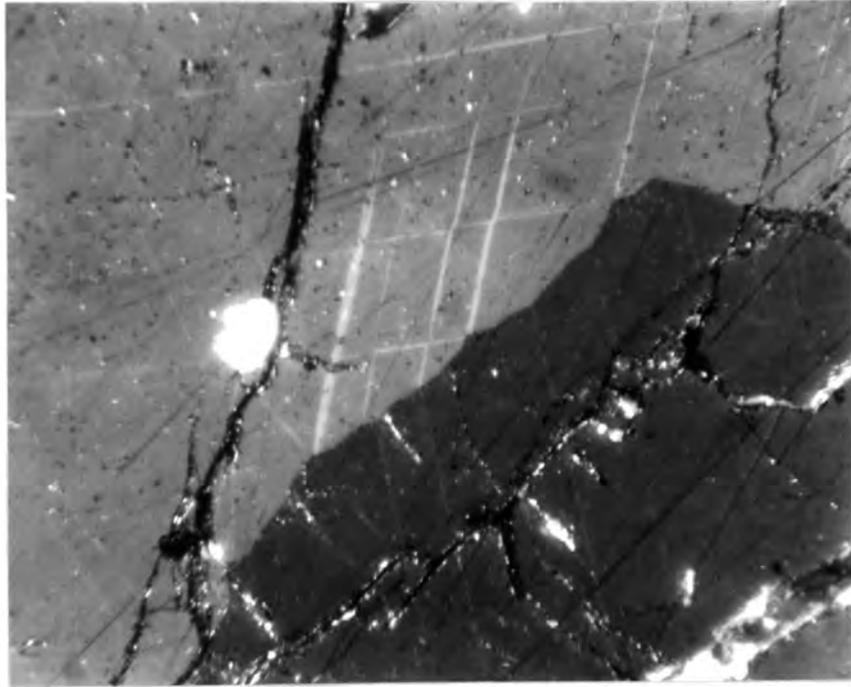


Figure 141: Twinning in arsenopyrite. Both large, simple twins and thin lamellar twins are visible. Forhaabning Adit, Skuterud.
Reflected light, crossed nicols. X 50.

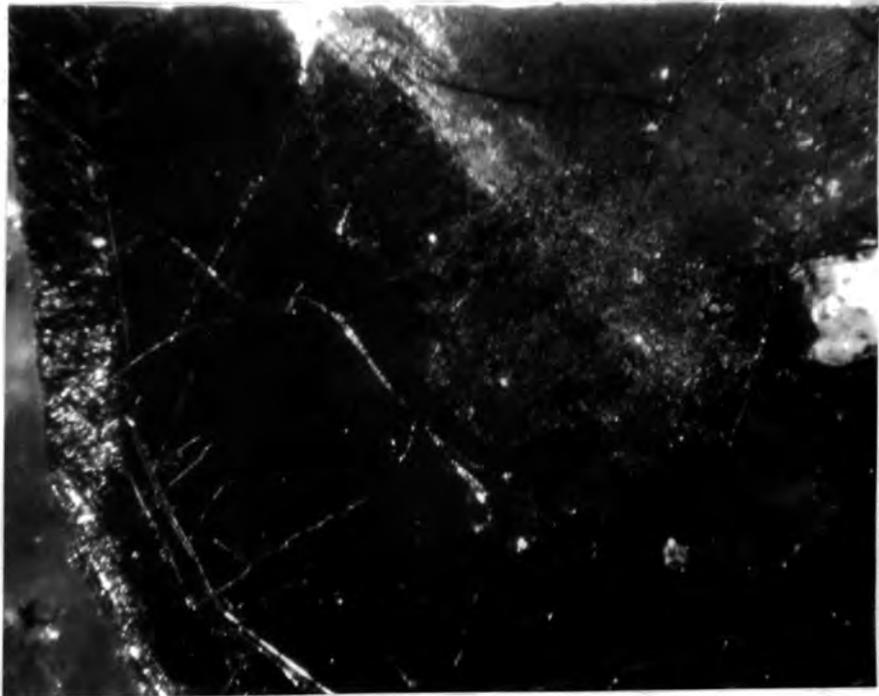


Figure 142: Oscillatory zoning at the termination of a large arsenopyrite crystal. Diopside-skarn, Weisscobolt Strossen, Central Mine, Skuterud.
Reflected light, crossed nicols. X 50.

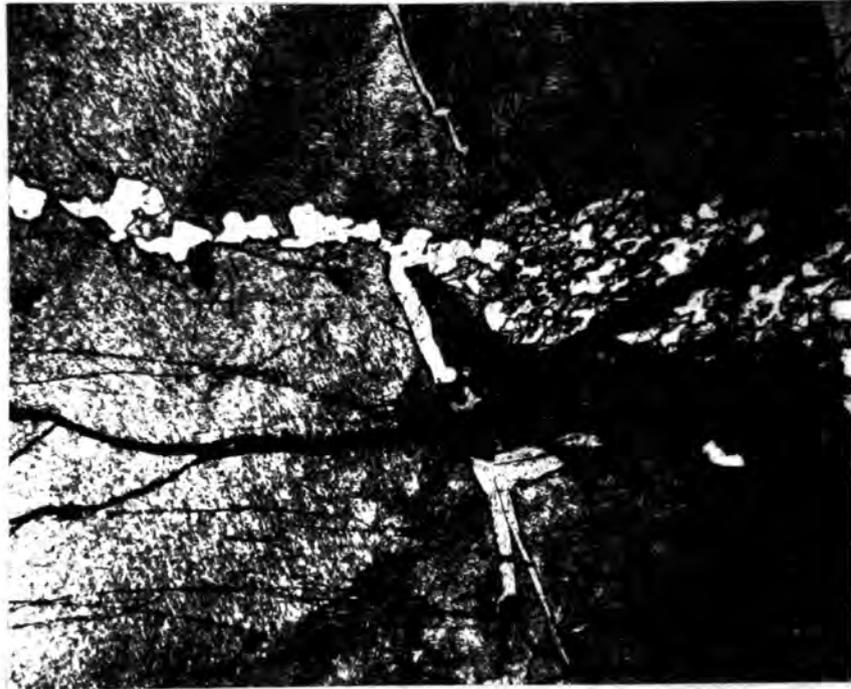


Figure 143: Ore sample etched with 1:1 nitric acid which affects arsenopyrite leaving cobaltite untouched. A string of small cobaltite blebs is visible in the arsenopyrite. Svartefjell Mine.
Ordinary reflected light.

X 50.

In one of the Museum specimens, labelled 'Modum', lamellae of a distinctly different phase are seen in the arsenopyrite. These have been shown to consist of a mineral with a composition midway between cobaltian arsenopyrite and cobaltite (Co: 17%, Fe: 18.5%. As: 46%, S: 18.5%), this association is discussed further in Part II. In the 69 polished specimens of the ore collected from dumps and in situ which have been examined only one further example of this relationship has been found. A single lamella is seen in one crystal of arsenopyrite collected from the dumps in the 'Weisscobalt Strossen' of the Central Skuterud Mine. Skutterudite This mineral is a very rare constituent of the cobalt ore. There are only four authenticated specimens in the collection of the Oslo Mineralogical Museum. The author found one other during the present investigation (Figure 122). The Central Mine at Skuterud is of mineralogical interest as the type locality for skutterudite.

Breithaupt (1823) first described the mineral from material from Skuterud. He suggested two possible names for the mineral, hartkobaltkies because of the hard nature of the mineral, or tesseralkies because of its hemihedral symmetry. The latter name is still in common use in Scandinavia.

The first chemical analysis of the mineral was by Scheerer (1838) who found 79% arsenic and suggested the name arsenikkobaltkies. This name has not gained popularity, possibly because of possible confusion with kobaltarsenkies (cobaltian arsenopyrite).

These three names became superseded in German mineralogical works by skutterudit named after the locality. This in its anglicized form of skutterudite is in common usage today.

Following the works of Holmes (1947) and Roseboom (1962) the name skutterudite embraces all isometric arsenides of cobalt, nickel and iron. These have been variously referred to in the past as smaltite, chloanthite, ferroan, nickelian or cobaltian skutterudite. Following Roseboom (op.cit.) these are all called skutterudite, with the modifier cobalt skutterudite for the only monometallic end member of the series produced synthetically. Analyses show that the material from Skuterud is unusual in approximating to this ideal composition of CoAs_3 , with very little iron substitution and no nickel (Holmes, op. cit.).

In the specimen illustrated in Figure 122 the skutterudite is seen as an irregularly shaped crystal at the centre of a massive ore lens. It is surrounded by a spongy boxwork consisting of an intergrowth of cobaltite and minor arsenopyrite. These minerals have very small areas (ca. 0.02 mm.) of skutterudite associated with them.

Chalcopyrite This is a common associate of the cobalt ore minerals. It usually presents a flat featureless field in which it is difficult to distinguish individual grains. No euhedral crystals have been observed.

In some specimens the chalcopyrite appears to be corroded and replaced by the silicate gangue along fractures and grain boundaries (Figure 144). In the majority of the specimens the reverse relationship is seen. The grain boundaries of the chalcopyrite are

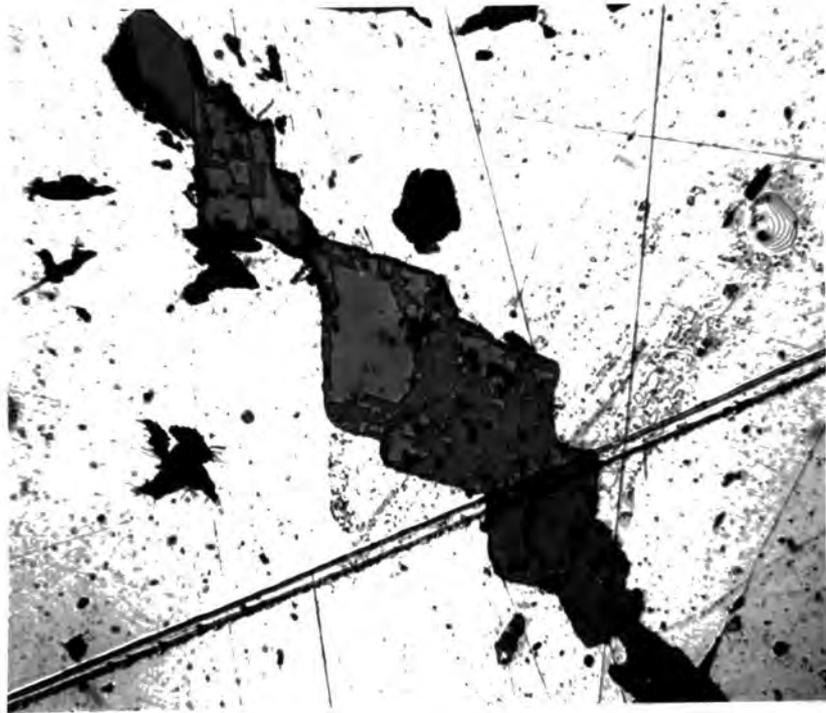


Figure 144: Chalcopyrite with silicate gangue at the grain boundaries.
Svartefjell Mine.
Ordinary reflected light. X 50.

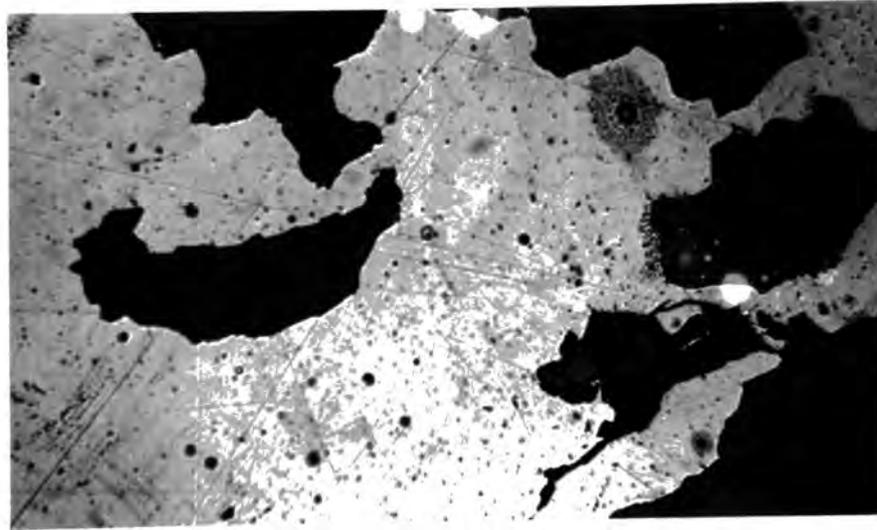


Figure 145: Replacement residual of silicate gangue in chalcopyrite.
Svartefjell Mine.
Ordinary reflected light. X 50.

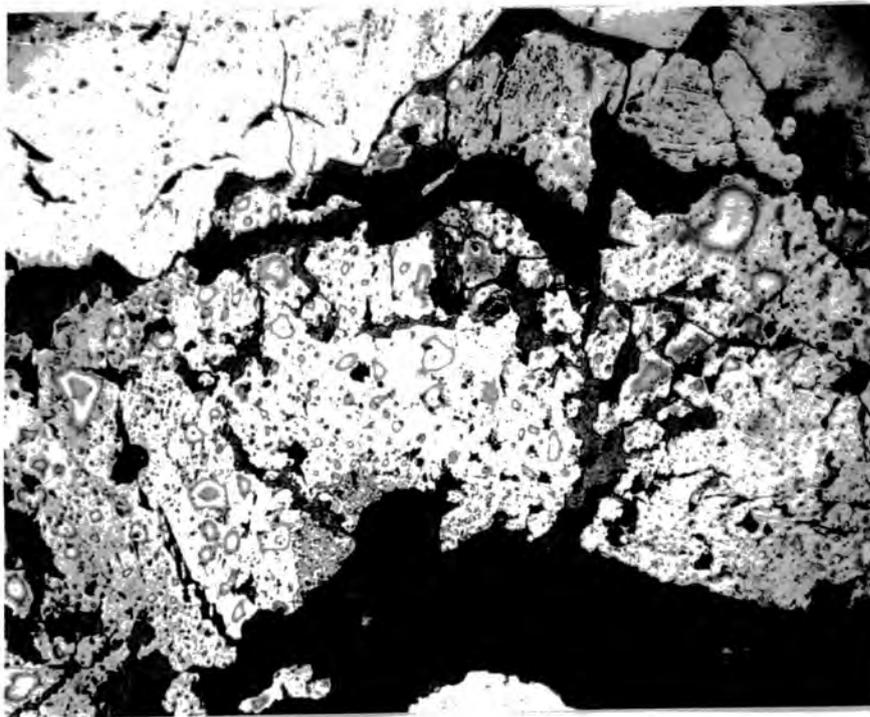


Figure 146: Cataclasis in cobaltite. The cobaltite has been invaded along cataclastic cracks, by sulphides. The sulphide vein has pyrite crystals set in a matrix of chalcopyrite. Mineralogisk-Museum collections.
Ordinary reflected light. X 50.

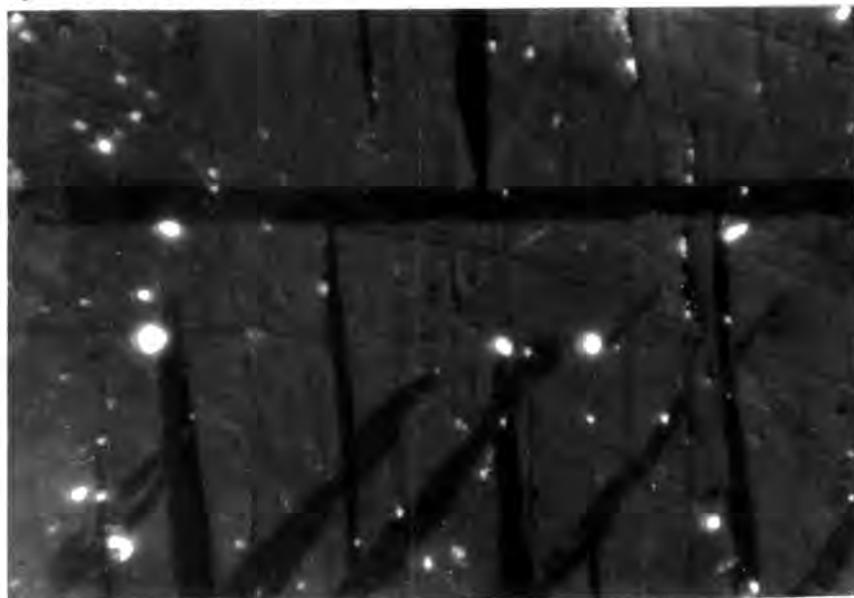


Figure 147: Twinning in chalcopyrite. Svartefjell Mine.
Reflected light, crossed nicols. X 50.

extremely irregular but always appear to be convex outwards to the gangue. Small rounded chalcopyrite grains are observed in the gangue adjacent to the larger grains. Textures are seen where silicate inclusions in the chalcopyrite appear to be suffering replacement by the chalcopyrite (Figures 145).

The chalcopyrite is seen forming a matrix for pyrite crystals in veins cutting across the cobaltite and arsenopyrite (Figure 146). Euhedral crystals of cobaltite are developed in a groundmass of chalcopyrite, frequently with a rim of pyrite separating the two minerals (Figure 132). Extremely irregular crystals of arsenopyrite are seen in the large chalcopyrite individuals.

The chalcopyrite is strongly anisotropic. Lamelliform twinning is well developed, spindle shaped lamellae are observed crossing each other at various angles (Figure 147).

Chemistry

Analyses for cobalt, copper and nickel have been made during the present investigation using standard X-ray fluorescence spectrographic techniques.

The values for cobalt were measured in Oslo on a Norelco spectrograph using cobalt chloride for locating the $\text{CoK}\alpha$ peak. Readings were made on this peak at a value of $52.77^\circ 2\theta$ with a background reading at $54.50^\circ 2\theta$. The tungsten tube was operating at 40Kv and 24mA, using the topaz crystal. The discriminator was employed with amp.28, channel width 16 and attenuation 2. The counts registered in 32 secs. by the scintillation counter were measured and compared with those from a standard sample. A standard was re-measured after every three samples. The standards were prepared by adding known amounts of cobalt metal to one of the Modum quartzites.

The values for copper and nickel were measured in Durham on a Phillips spectrograph using a tungsten tube, topaz crystal, light element collimator, no discrimination, scintillation counter at 900V, rate meter X32 and the tube at 40Kv and 20mA. Counts in

60 secs. were registered with an electronic timer. The samples were measured in rotating, circular holders with mylar windows. The standards used were sandstones containing known amounts of Cu and Ni made up by the addition method. The peak positions measured were: Cu 1 K α low background = $68.37^{\circ}2\theta$, peak = $69.27^{\circ}2\theta$, high background = $70.17^{\circ}2\theta$; Ni 1 K α , background = $74.28^{\circ}2\theta$, peak = $75.43^{\circ}2\theta$. Three readings were made of each value and the average taken. The working curves for Cu and Ni are shown in Figure 148. The standard deviation of the results were: copper - ± 2.1 ppm., nickel 1.1 ppm. The relative deviation of the results were: copper - $\pm 2.2\%$, nickel - $\pm 3.8\%$; measured on a sample containing 94 ppm. Cu and 30 ppm. Ni in both cases.

The results are shown in Table 13. Figure 149 shows the values of Cu:Co:Ni determined for 34 rocks and demonstrates the spread of values between copper-rich and cobalt-rich specimens and the relatively low amounts of nickel. Figure 150b is a plot of logCo: logCu and shows that there is no correlation between these two elements. Figure 150a, by contrast, shows a correlation between cobalt and nickel, these data are replotted as cobalt (p.p.m.) against the Co:Ni ratio in Figure 151. This shows that there is a constant increase in the ratio with cobalt content, suggesting that a fairly constant amount of nickel (20-80 p.p.m.) is available and is being diluted by the addition of cobalt.

Analogous deposits

Five Swedish deposits are frequently included with Modum in discussions on the 'cobalt-fahlbands' in the older literature. These were visited during the present investigation for comparative purposes.

Tunaberg

The Tunaberg cobalt mines are situated near Nykjöping in Södermanland, Sweden. They have been described by Erdmann (1867), Johanssen (1909) and Geiger (1924). The cobalt mineralisation is in a metamorphosed limestone horizon interbanded in a fine-grained,

Table 13

A. Copper analyses.

Sample No.	Cu.%.	Sample No.	Cu.%.	Sample No.	Cu.%.
307	0.01	367	0.00222	398	0.00048
311	0.00927	369	0.0009	406	0.00296
319	0.00192	370	0.00198	409	0.00407
321	0.00342	371	0.00191	475	0.00114
328	0.44	373	0.00082	479	0.00528
329	0.45	375	0.00052	515	0.00085
330	0.44	376	0.0004	517	0.00072
331	0.00708	377	0.00077	518	0.00003
332	0.00418	380	0.00121	520	0.00150
333	0.00323	381	0.00056	521	0.00492
335	0.01121	383	0.00129	522	0.532
336	0.3	384	0.00302	523	0.00332
341	0.16	388	0.545	535	0.0072
343	0.01197	389	0.00338	537	0.00115
350	0.00142	393	0.00172	539	0.00522
346	0.00075			679	0.00214

B. Cobalt analyses.

Sample No.	Co.%.	Sample No.	Co.%.	Sample No.	Co.%.
302	0.00104	367	0.00016	398	0.0004
305	0.00056	369	0.00120	406	0.00152
307	0.00272	370	0.00056	409	0.00104
312	0.00680	371	0.00064	479	0.00296
319	0.00312	372	0.00072	517	0.00072
321	0.00304	373	0.00032	520	0.00112
322	0.22	375	0.00074	521	0.00048
329	0.8	376	0.00152	522	0.00160
330	1.04	377	0.00024	523	0.0008
332	0.36	379	0.00008	524	0.00105
335	2.15	383	0.002	525	0.00112
336	0.14	384	0.00128	535	0.00096
341	0.00552	388	0.00368	537	0.00040
343	0.74	389	0.00064	539	0.00056
364	0.1			578	0.00048

C. Nickel analyses.

Sample No.	Ni (ppm).	Sample No.	Ni (ppm).	Sample No.	Ni (ppm).
307	268	367	28	398	32
311	62	369	30	406	32
319	52	370	40	409	29
321	61	371	45	475	35
328	325	373	39	479	42
329	360	375	32	515	31
330	51	376	45	517	37
331	184	377	41	518	35
332	40	380	45	520	63
333	56	381	31	521	38
335	865	383	42	522	45
336	96	384	64	523	41
341	72	388	75	535	55
343	168	389	45	537	31
350	53	393	30	539	51
364	42			679	214

Localities: 302, 305, 307-graphitic-schist, Svartefjell Mine; 311, 312-quartzite, Svartefjell Mine; 319, 321-siliceous-granulite, Ludwig Eugen Level; 322-siliceous-granulite - North Mine, Skuterud; 328, 329, 330, 331, 332, 333, 335, 336, 341, 343 - ore quartzite, sillar in opencast, North Mine, Skuterud; 350 - siliceous-granulite, Butjern; 364, 367, 369, 370, 371, 372, 373, 375, 376, 377, 379, 380, 381, 383, 384, 388, 389, 393, 398-siliceous-granulite, Ludwig Eugen Level, entral Mine, Skuterud; 406; 409-siliceous granulite-South Mine, Skuterud; 475, 479, 515, 517, 518, 520, 521, 522, 523, 537, 539,-siliceous-granulite, North Mine, Skuterud; 578,-siliceous-granulite, Central Mine opencast, Skuterud; 679-ore-quartzite, Weisscobolt Strassen. Skuterud.

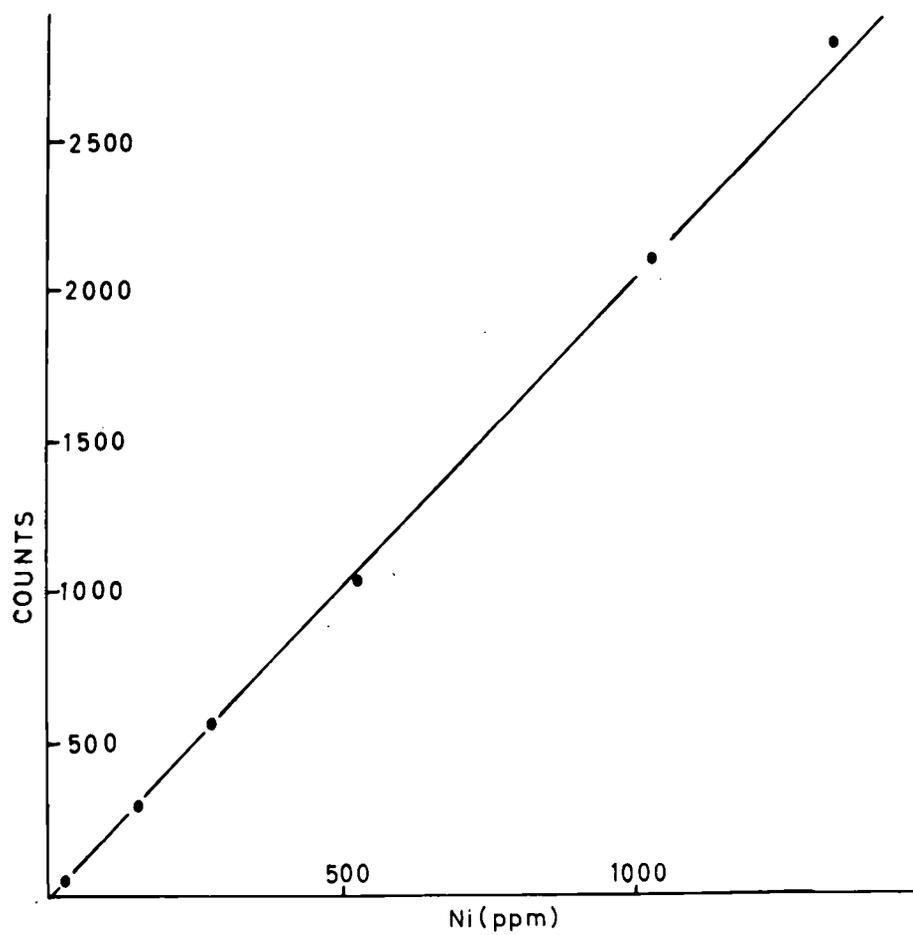
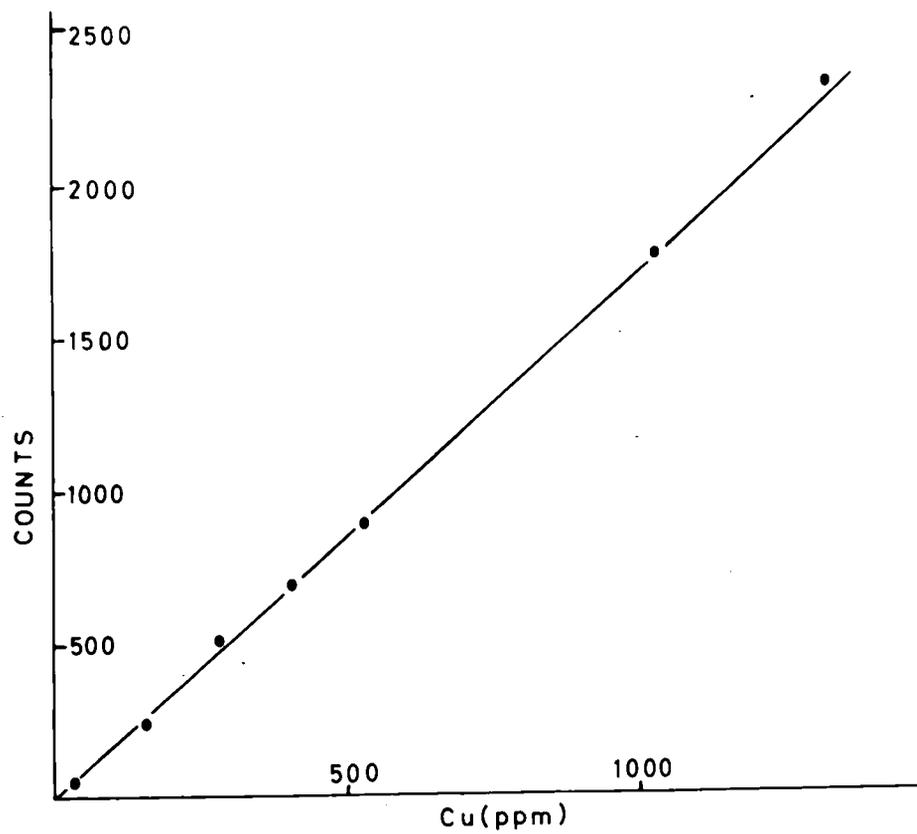
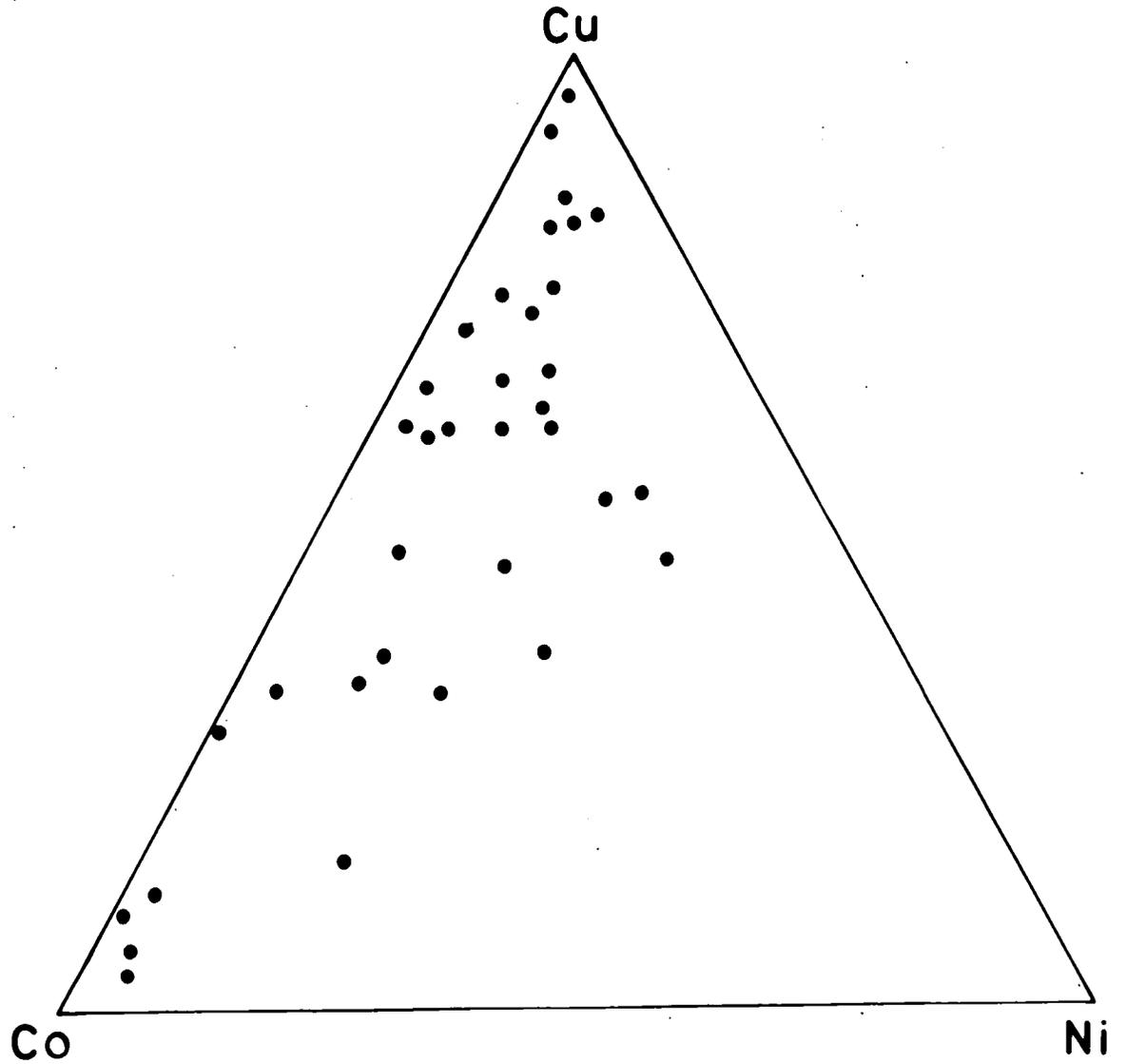


Fig. 148



Co:Cu:Ni for 34 quartzitic rocks from Skuterud

Fig . 149

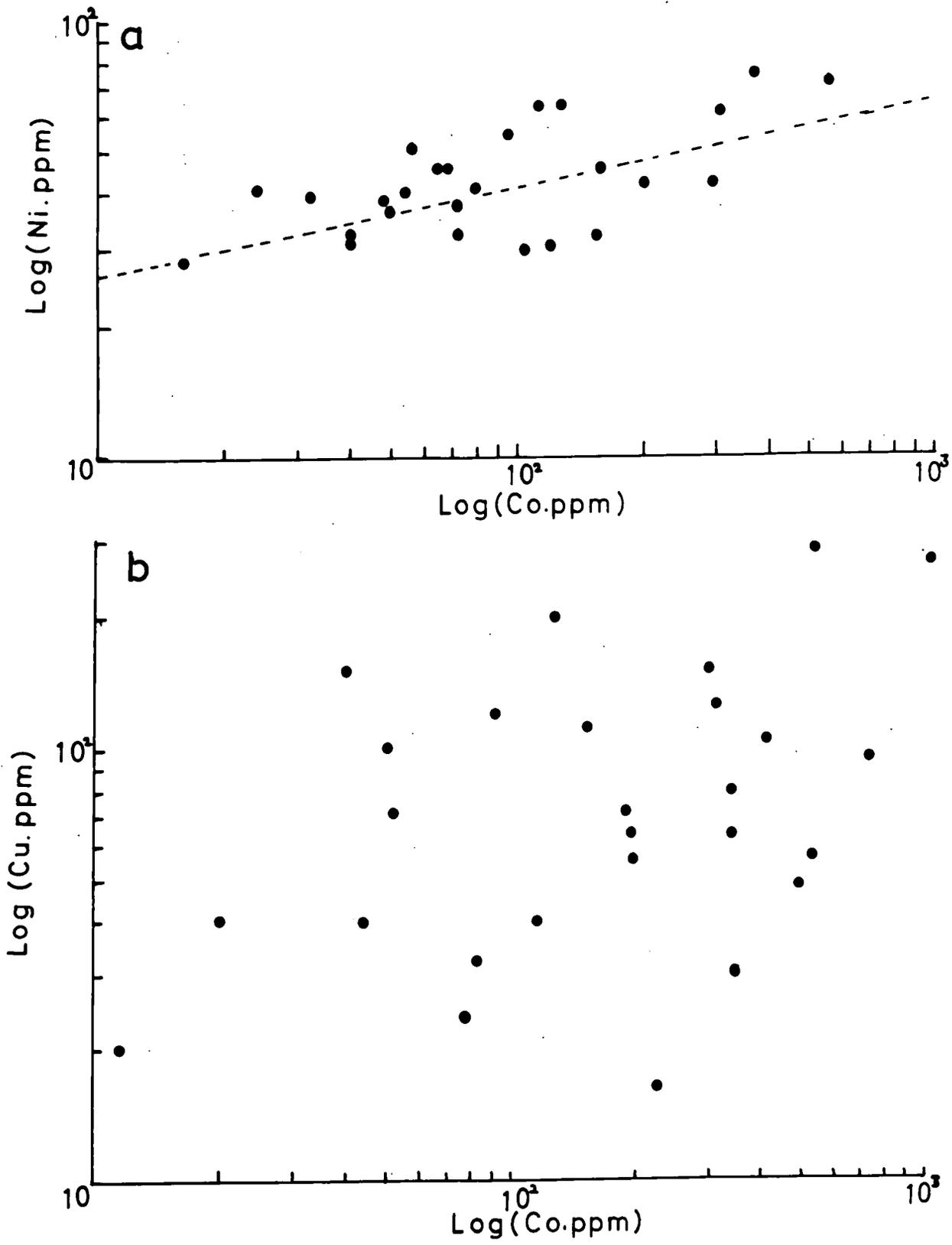


Fig. 150

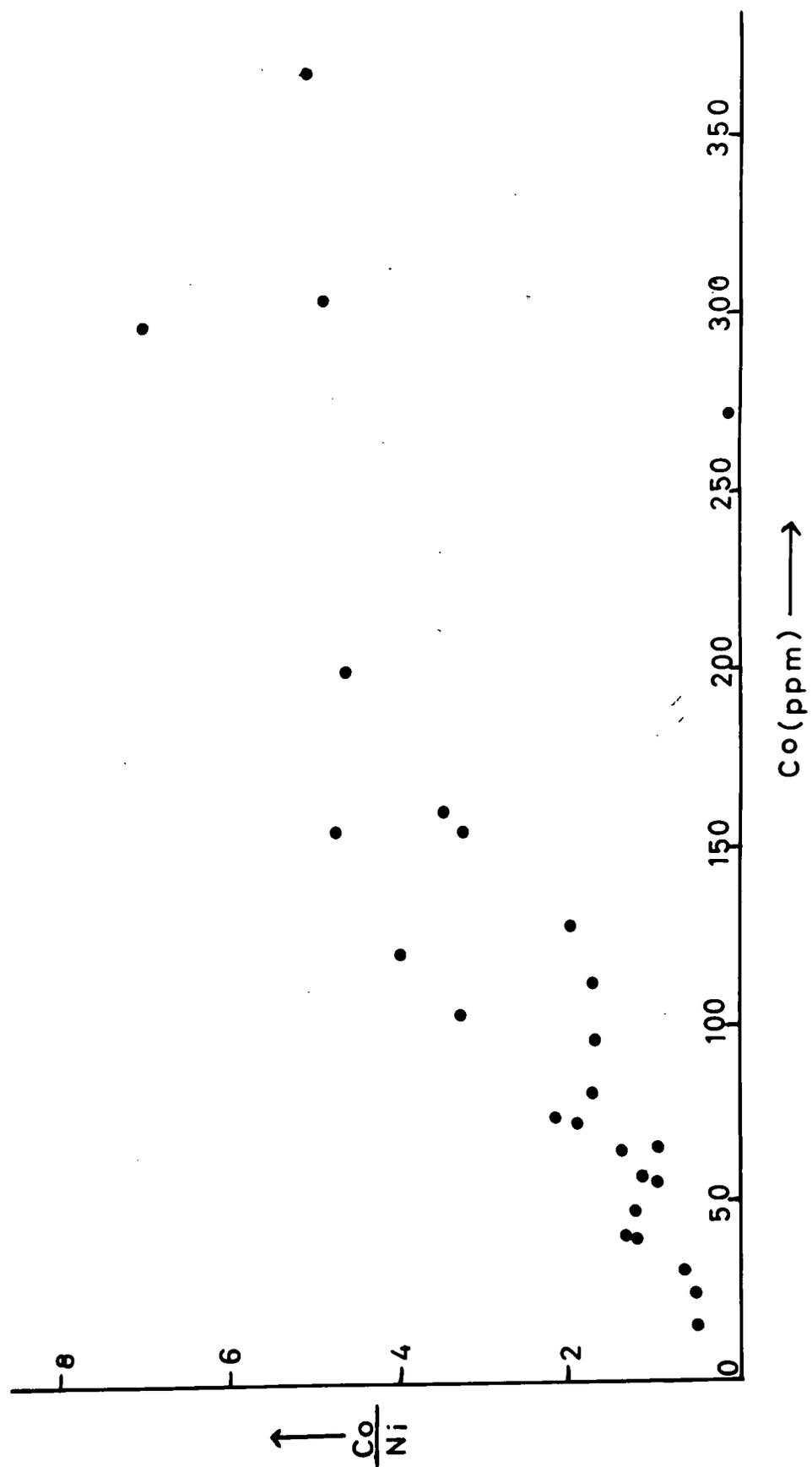


Fig. 151

usually well banded, grey, potash felspar rich gneiss. This gneiss formation also contains garnetiferous bands, pyroxene gneisses and amphibolites. The regional strike is ENE-WSW with a sub-vertical dip, in the ore zone the dip of the foliation flattens to 20° in places. The mineralisation is particularly concentrated in fold closures. It occurs as a dissemination of cobaltite, skutterudite, chalcopyrite, pyrite, galena, bornite, chalcocite and sulphosalt minerals. Where calc-silicate skarn lenses are developed in the limestone the mineralisation is particularly associated with these. The cobaltite is in beautiful euhedral crystals which can attain a maximum size of 6 cms. diameter. Typical fahlbands with pyrrhotite and chalcopyrite are present in the surrounding gneisses but appear to be unconnected with the cobalt mineralisation.

Riddarhyttan

The Riddarhyttan mining district of central Sweden is based on the concordant haematite and magnetite ore bodies extensively developed in an amphibolite grade metamorphic sequence. (Geiger, 1923). Sulphidic impregnations found in some of the iron-ore workings usually consist of chalcopyrite, pyrrhotite and pyrite. Some of these sulphide lenses contain cobaltite, euhedrally developed in large crystals in an anhedral matrix of chalcopyrite. This association was particularly well seen in the stope 'Ohlsons rum' of Pellgruva (Geiger, p. 98) Vena.

Vena

The fahlbands present at this locality have already been mentioned (p. 192). The associated cobalt mineralisation is developed in small lenses surrounded by relatively barren rocks. The workings are in three main zones which are parallel to the strike

and are usually situated in biotite gneiss, however they are also found in quartzite (Långgruvan), mica-schist (Gamla Vena gruvan) and amphibolite (Fredriksgruvan). In the only contemporary map (van Zweigbergk, 1825) the ore is shown to be connected with the more micaceous bands in the biotite gneiss on the 50 m. level of Galtgruvan. Specimens of the ore in the collections of the Riksmuseet, Stockholm show the mineralisation to be closely connected with biotite rich bands even when these bands are tightly folded.

Håkansboda

The Håkansboda copper-cobalt deposit is situated in Örebro "Lan, central Sweden. The ore is found in three en-echelon lenses. Collecting from the dumps suggests that the northern mine (Norgruvan) was based on a massive lens of pyrite, pyrrhotite and chalcopyrite with no visible cobalt minerals. The dumps at the central mine (Kamersgruvan) suggest that the ore here consisted of massive chalcopyrite and pyrrhotite containing euhedral crystals of glaucodot (cobaltiferous arsenopyrite), while the southern mine (Sörgruvan) contained similar crystals of cobaltite set in a matrix of chalcopyrite and pyrrhotite. The country rock surrounding the ore lenses is limestone of a similar type to that at Tunaberg.

Gladhammar

The Gladhammar fahlband has been previously mentioned on page 193. Associated with the iron ore bodies at this locality are sulphidic impregnations containing cobaltite, galena, sphalerite, galenobismutite and linneite in a matrix of pyrite and chalcopyrite. Workings for copper and cobalt have been opened in a zone stretching for 2 kilometres along the strike coincident with the fahlband, these

reach a depth of 180 metres.

Conclusions

It emerges that there are several similarities between these various deposits. They are often associated with limestones (Tunaberg, Håkansboda) and haematite-magnetite ore bodies (Riddarhyttan, Gladhammar), the cobalt minerals occur as 'porphyroblasts' in a matrix of iron and copper sulphides (Riddarhyttan, Håkansboda, Gladhammar, Tunaberg). The only two deposits which can legitimately be called 'cobalt-fahlbands' are Vena and Gladhammar where the cobaltiferous zones are coincident with fahlbands.

These latter two deposits are the only two that bear any resemblance to Modum. Gladhammar lies at the centre of the Västervik metallogenetic province which is particularly rich in the ores of iron, copper, cobalt, molybdenum, uranium, gold and selenium (Uytenbogaardt, 1960). The thick Västervik quartzite is broken up and surrounded by granitic rocks. Uytenbogaardt op.cit. has interpreted the uranium mineralisation as being 'closely related to the granitisation of the quartzite formation'. Thus Gladhammar has many features which differentiate it from the Vena and Modum type of occurrence. It is possible that the cobalt mineralisation has originated in a way similar to that postulated for the uranium.

The characteristics of the Vena deposit are extremely similar to those of Modum, the ores are associated with fahlbands, rich cobaltiferous lenses are surrounded by relatively barren rock but are in zones parallel to the foliation, calc-silicate lenses are developed in the ore zone rocks, the mineral assemblage cobaltite-skutterudite-cobaltian arsenopyrite is present and the deposits have

been worked out at a shallow depth. However at Vena there is no clear field evidence of shear zones coincident with the mineralisation, amphibolite are present and have an intrusive relationship to the metasediments (Henriques, personal communication) but are much less frequent and sphalerite is associated with the impregnated sulphides.

Thus in postulating genetic hypotheses for the Modum cobalt deposits the characteristics of the Vena deposit should also be considered.

Genetic hypotheses

Supergene enrichment

The main cobalt producing mines in Modum are closely related to the present erosion surface (Figure 102). This relationship led A. Bugge to make the suggestion that the deposits may have originated through secondary enrichment (vide Rosenqvist, 1948, p. 210).

Rosenqvist (op.cit.) considers it more likely that a poor Precambrian mineralisation has been enriched in this way, probably under the sub-Cambrian peneplane, which must have been roughly coincident with the present topographic high points. To support this argument he quotes the assay values obtained during World War II which show a slight impoverishment of cobalt with depth.

However the observed mineral association is not compatible with processes of secondary enrichment. Cobaltite, arsenopyrite and skutterudite are all primary minerals found in many clearly hydrothermal deposits. The descriptions of the ore in the literature and the Museum specimens strongly suggest that it was

these primary ore minerals that were being worked. The ores have weathered to erythrite in places but this mineral is not particularly noticeable in the upper levels and, as far as the author is aware, no erythrite ores have been worked in other areas.

It is therefore suggested that this possibility can be firmly rejected.

Hydrothermal introduction during the Permian

Rosenqvist (op.cit. p. 211) states that "the occurrence of a series of cobalt-bearing minerals definitely shows that the mineralisation is of Precambrian age." This unqualified statement has been strongly challenged by Vokes (1957) who suggests that a Permian age is more reasonable for the Modum ores.

Vokes suggests that the mineralisation took place during the Permian when a thick sequence of impermeable shales of Cambro-Silurian age covered the sub-Cambrian peneplane. Hydrothermal solutions coming up planes of weakness would be impeded by this impermeable cover and would spread out to form a series of ore shoots controlled in attitude by the horizontal cover.

This hypothesis is attractive on many counts. It explains the association of the ore bodies with the topography, the declining ore grade with depth and the primary mineral association. The poorly defined 'stratigraphic control' of the cobalt ore zone and its association with the fahlbands are explained by the frequent shear zones associated with the graphitic sulphide schists, these would have formed extremely favourable channelways for the ascending solutions. The tourmaline in the fahlband zone rocks could be a manifestation of a very common type of hydrothermal alteration. The

Permian was a period of widespread igneous activity in the Oslo region, many contact-pneumatolytic deposits were formed at the contacts of the intrusive bodies and the palaeozoic sediments (Goldschmidt, 1911). The nearby Kongsberg silver deposits are attributed to the Permian mineralisation by Neumann (1944), the Kongsberg silver veins contain minor quantities of cobalt minerals.

The mineral zonation from cobaltite to arsenopyrite away from the Central Skuterud Mine suggests that this mine may be located over the main feeder channel. Vokes has suggested that the culmination in the plunge of the lineation (Figure 58) over the Central Mine may have been a factor in localising this feeder channel.

However, this hypothesis fails to explain a number of important factors. The textures of the ore and gangue minerals suggest that mutual replacement has occurred, rather than replacement of the silicates by the ore minerals alone. There is no textural evidence to support the 'ponding-up' of the ore under the Palaeozoic cover . The ore is typically present as a fine grained impregnation particularly concentrated at certain horizons, it is difficult to visualise this texture resulting from selective hydrothermal replacement. The postulated feeder zones, the fahlbands, contain fewer cobalt minerals than the adjacent more massive and unfractured quartzites. Hydrothermal gangue minerals are not noticeably associated with the cobalt ore. The origin of the tourmaline has already been discussed (p.74), it does not seem necessary to invoke a hydrothermal origin to account for its presence. The observed zoning phenomena could equally well be the result of metamorphic processes.

The Permian is indeed the most likely post-Cambrian time for the postulated hydrothermal activity. Quartz and calcite veins with associated galena and chalcopyrite mineralisation are common in the district, these have been tentatively correlated with the very similar Permian veins of the Kongsberg region. These are clearly seen to cross-cut and post-date the cobalt ore. Fresh dolerite dykes of probable Permian age are present in the district, these cut across the fahlband zones at an oblique angle. They have not been observed cutting the cobalt ore, unless it is assumed that the Djupedal Trail contained cobalt minerals, a transgressive dolerite dyke is observed at this locality. These relations suggest that the ore is of pre-Permian and, in all probability, of Precambrian age.

Precambrian mineralisation

If the ore is of Precambrian origin it is possible that it has suffered metamorphism together with the host rocks. The observed textures are strongly suggestive that this is indeed the case. The ore minerals appear to be an integral part of the fabric of the host rock. They display many textural features observed in metamorphic silicate minerals, porphyroblastic habit, segregation into layers and lenses, blastic intergrowths and conformability with minor structures and folds. The mineralogy of the cobaltite and arsenopyrite support the hypothesis that they have undergone metamorphism (Part II).

Perhaps the clearest evidence for the Precambrian and pre-metamorphic age of the mineralisation is the transgressive nature of the post-kinematic late pegmatites. These are described in the

literature and observed in the workings to cross-cut the ore horizons, which continue uninterruptedly on either side.

The main disadvantage of the conclusion that the ore is of pre-metamorphic origin is that its original form becomes more difficult to elucidate. The major mineralogical, textural and structural changes that have been effected in the metasediments suggests that the original nature of the mineralisation will have been strongly obscured by the metamorphic episode.

Syngenetic with the fahlbands The hypothesis that the cobalt minerals are syngenetic with the iron and copper sulphides of the fahlbands was first suggested by Bøbert (1848). The association of both the Modum and Vena cobalt ores with fahlband zones led him to this conclusion. The very low grade of the ores, the preferred association of the ore with certain bands in the metasediments, and the presence of two mineralised zones parallel to the regional foliation are compatible with this hypothesis. It is more difficult to explain the association of the ore shoots with the present erosion surface. It has been suggested that the cobaltiferous horizon has been folded and the ore preserved only in the keel of the major syncline which is coincident with the present erosion surface.

There is no field evidence to support this latter idea, due to the paucity of marker horizons the detailed structure of the deposits has not been elucidated, but at the North Mine, Skuterud there is evidence to suggest that the mineralisation is in the crest of a north plunging antiform (p. 239). The present investigation strongly suggests that the association of the cobalt and fahlband mineralisations is purely coincidental. Fahlbands are

very common features of the Kongsberg - Bamble Formation but it is only at Modum that one of them is associated with cobalt mineralisation. This mineralisation is not confined to the fahlband rocks but is found well developed in quartzites, amphibolites, diopside-skarns and biotite-schists. The fahlband has a much greater lateral extent than the cobalt mineralisation. It is here suggested that too much emphasis has been placed in the past on the association with the fahlbands. Many prospects appear to have been investigated solely on the presence of sulphidic graphite schists. (Heggebek, Djupedal, Muggerrud and others), these trials did not result in the discovery of workable cobalt ore. If the cobalt was linked to copper-iron mineralisation in the fahlbands it would be expected that there should be some correlation between these elements in the more disseminated types of occurrence, Figure 150 shows that there is no correlation between cobalt and copper in the analysed samples. Another strong argument against the sedimentary syngenetic origin of the cobalt minerals is the provision of a suitable source for the cobalt and arsenic. This same problem has been encountered in the Northern Rhodesian Copperbelt where the amount of cobalt present is very much lower. It would seem necessary to invoke some mechanism such as an exhalative-sedimentary origin as the source of the cobalt and arsenic. No actualistic analogue of a sediment with a high cobalt:nickel ratio is known (Davidson, 1962).

Syngenetic with the amphibolites Vogt (in Krusch *et. al.*, 1913)

suggested that both the cobalt and fahlband mineralisation originated through an impregnation of sulphides in the wall rocks adjacent to the intrusive gabbro and amphibolite bodies. It is here suggested

that this mechanism is not necessary to account for the fahlbands but has merit in relation to the cobalt mineralisation.

Two recent publications (Andrews, 1962; Bilbrey, 1962) have reviewed the major world occurrences of cobalt. It is striking how many of these deposits are spatially associated with basic rocks even though they are not, in many cases, considered to have any genetic connection with the deposits concerned. The best documented example of this type of occurrence is the cobalt-silver mineralisation of the Cobalt-South Lorrain-Gowganda area, Ontario.

The Cobalt area is underlain by gently dipping strata of the Cobalt Series laid down on an irregular erosion surface of steeply folded Keewatin lava, tuff and iron-formation. The Nipissing diabase sill, about 1,000 feet thick, intrudes these along a warped plane that passed in part through the Cobalt Series and in part through the Keewatin. The following extracts from Stockwell (1957, p. 68) give a good summary of the mode of occurrence.

"No ores have been found where the diabase sill is absent, and all the ores were found within a few hundred feet of the contact, either above or below, or within the diabase itself.....About 80% of it was found in the Cobalt Series within 100 to 200 feet of the Keewatin contact; the remainder occurred along the upper or lower contact of the sill, mostly within 200 or 300 feet of either margin. Few veins were found near the middle of the sill.....The high-grade veins were mostly narrow, from an inch to a foot in width, though in rare instances, and for relatively short lengths, widths attained several feet.....Some were followed for several thousand feet, but most of them were must shorter. Besides the high-grade veins, in places the wall-rocks were broken by innumerable tiny cracks into which the vein-forming solutions penetrated and deposited silver. Considerable widths of rock were thereby converted into a low-grade ore, locally termed mill-rock."

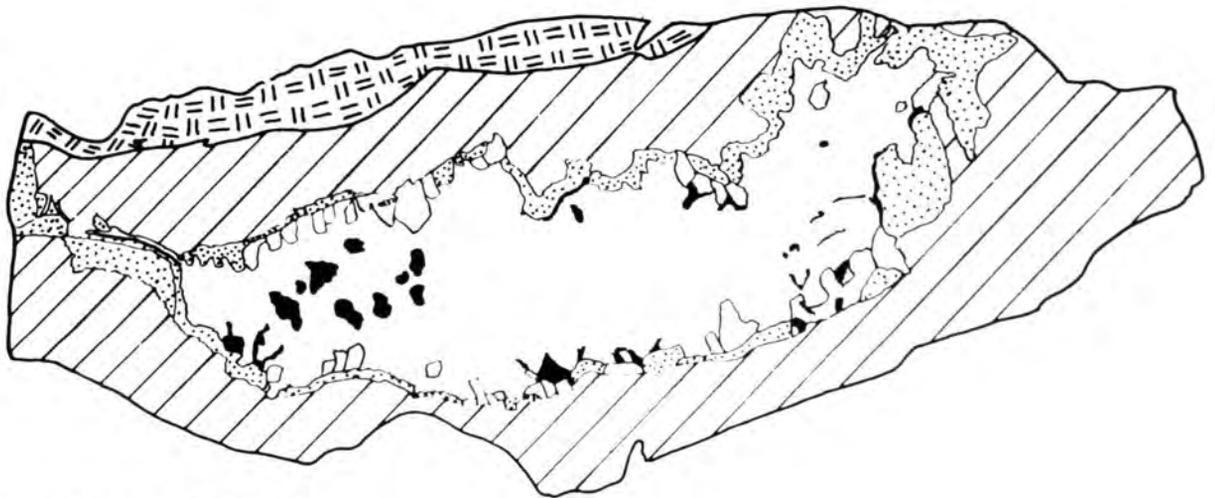
In the South Lorrain area a similar diabase forms an eroded, irregular, elongated dome a few miles long. The productive vein occurred in the sill and in the overlying Keewatin rocks within a small area of about 100 acres. The main vein, which reached a width of 10 feet, occurred in a reverse fault, the mineralisation did not extend from wall to wall but formed a number of stringers that ramified through the shattered country rock.

In the Gowganda area, the diabase sill is probably not more than 500 feet thick; most of the veins lie at the upper contact of the sill in the diabase itself, though one important vein was in Cobalt conglomerate just above the sill.

Sampson and Hriskevich (1957) have described the petrology of the Nipissing diabase. They consider that the original magma was of tholeiitic type and that it has clearly differentiated.

"The upper and lower margins are composed of undifferentiated quartz-diabase. Olivine-bearing hypersthene diabase occurs as saucer-like masses in the lower part of the sheet and these are confined to original structural basins. In the central part of such structural areas the olivine-bearing hypersthene diabase makes up the major portion of the section. In the intervening arched areas it is not present and there the major lower part of the sheet, above the undifferentiated quartz-diabase, consists of quartz-bearing hypersthene diabase. This is overlain by a quartz-bearing diabase, which shows extreme variations in texture from medium to very coarse and is known locally as 'varied texture' diabase. The upper portions of the arched areas are nowhere preserved, but in all probability consisted in large part of granophyric materials that were concentrated there during the differentiation of the sheet."

Aplite dykes are found in the upper parts of the diabase sheet, these are interpreted by Sampson and Hriskevich as being the final differentiates of the diabase magma. Associated with these dykes are metallic minerals containing the elements Co, Fe, As and S. These minerals are found in small veins and disseminations and as vug fillings. Figure 152 is taken from Sampson and Hriskevich's



EXPLANATION

 Diabase

 Albite

 Carbonate

 Aplite

 Quartz

 Metallic minerals

Fig. 152

paper and shows a cross-section of a vug from the central portion of an aplite dyke about one foot wide. These textural relations are taken by them to imply the following sequence of events:

"The material filling the vugs was an integral part of the aplitic magma and as crystallisation proceeded upon the emplacement of the magma, the crystallisation of the major portion of the aplite was followed by segregation of a residual liquid rich in volatile components. As the temperature decreases albite, and slightly later, quartz crystals grew inward from the vug walls. The crystallisation of carbonate and metallics followed with some replacement of the comb quartz and albite until the vug was completely filled."

The consider that these relations indicate that:

"normal processes of differentiation that segregate an aplite magma also bring about a concentration of Co and As as well as some other ore forming elements. Because differentiation within the sheet is capable of producing on a small scale a later residual liquid that deposited minerals so closely similar to those that form the ore deposits, it is logical to think that the main cobalt-silver veins of the district were also derived from a diabase magma - the parent diabase magma from which the diabase sheet at Cobalt is but a small offshoot."

The Cobalt area has been described in some detail because it shows many similarities to the Modum mineralisation. These similarities do not become apparent until the effects of the metamorphism are taken into account.

Consider a sedimentary sequence consisting of alternating sandstones, shales, thin limestones, calcareous sandstones and sapropelic muds. If an olivine-dolerite sill is intruded such that it slightly transgresses a thick sandstone horizon and has connected, smaller sill-like intrusions in the overlying sediments, if this is allowed to differentiate in an analogous manner to the Nipissing diabase and also has an exactly analogous cobalt-mineralisation associated with it, on metamorphism and folding a deposit of the Skuterud type will develop.

The sandstones, shales, thin limestones, calcareous sandstones and sapropelites recrystallise to give, in the upper amphibolite facies, quartzites, mica-schists, diopside-skarn, calcareous granulite and fahlbands respectively. The dolerites will be converted into amphibolites, except for the olivine rich accumulates which tend to resist the metamorphism to some extent. If the whole succession is folded so that the originally horizontal beds are vertical the result would be an area similar to the Skuterud mining district.

The effect of these transformations on the relatively mobile cobalt minerals must be considered. The cobalt minerals in veins near the upper contact of the dolerite could become transformed to give the rich ore impregnations observed at the contacts of the amphibolites and the adjacent metasediments. Veins in the overlying sediments would also lose their identity such that the cobalt minerals appear in rich lenses, remnants of the thin rich veins, and lower grade disseminations, remnants of the ramifying, low grade 'mill-rock', type of ore.

This sequence of events would explain a greater number of the features observed at Skuterud, than any other hypothesis. It has been previously proposed that the main garnet-amphibolite and the associated amphibolite - biotite-schist represent a folded and metamorphosed sill complex (p.119). The cobalt minerals are closely associated with the 'top' contact of this meta-dolerite. The relationship of the mineralisation to the topography is also of interest; the major hills in the mineralised zone are based on either quartzite or gabbro-amphibolite masses, only those features based on

basic intrusives have important cobalt workings associated with them. The mineralised areas are of restricted occurrence, but are always associated with basic rocks.

This hypothesis has some drawbacks. The mineralisation is only associated with the 'top' contact of the 'sill', whereas at Cobalt both contacts are equally important. It may be that the South Lorrain and Gowganda areas are more closely analogous as the mineralisation there is connected with the top contact of the diabase only. The lack of gangue minerals which should be associated with the original cobaltiferous veins is noticeable, it is possible that the nature of these would become obscured during the tectonism and metamorphism. The association of cobalt minerals with certain of the metasedimentary horizons is another anomalous feature, it is possible that this is a purely metamorphic feature, being the result of the remobilisation of the minerals into favourable, micaceous horizons. The brief period spent in the field at Vena would suggest that amphibolites are not as well developed there as at Modum and it may be difficult to postulate the same hypothesis to account for both deposits.

The most telling argument against this theory is probably the geochemical one. The ores of the Cobalt area are a typical example of the Co-Ni-Ag type of mineralisation. This association of elements has been reported from many mining districts all over the world. Modum is unique in the preponderance of cobalt bearing minerals, the skutterudite is the monometallic cobalt end-member containing no nickel or iron, the arsenopyrite is very cobaltiferous, no nickel or silver minerals have been recorded in this study or

reported in the literature.

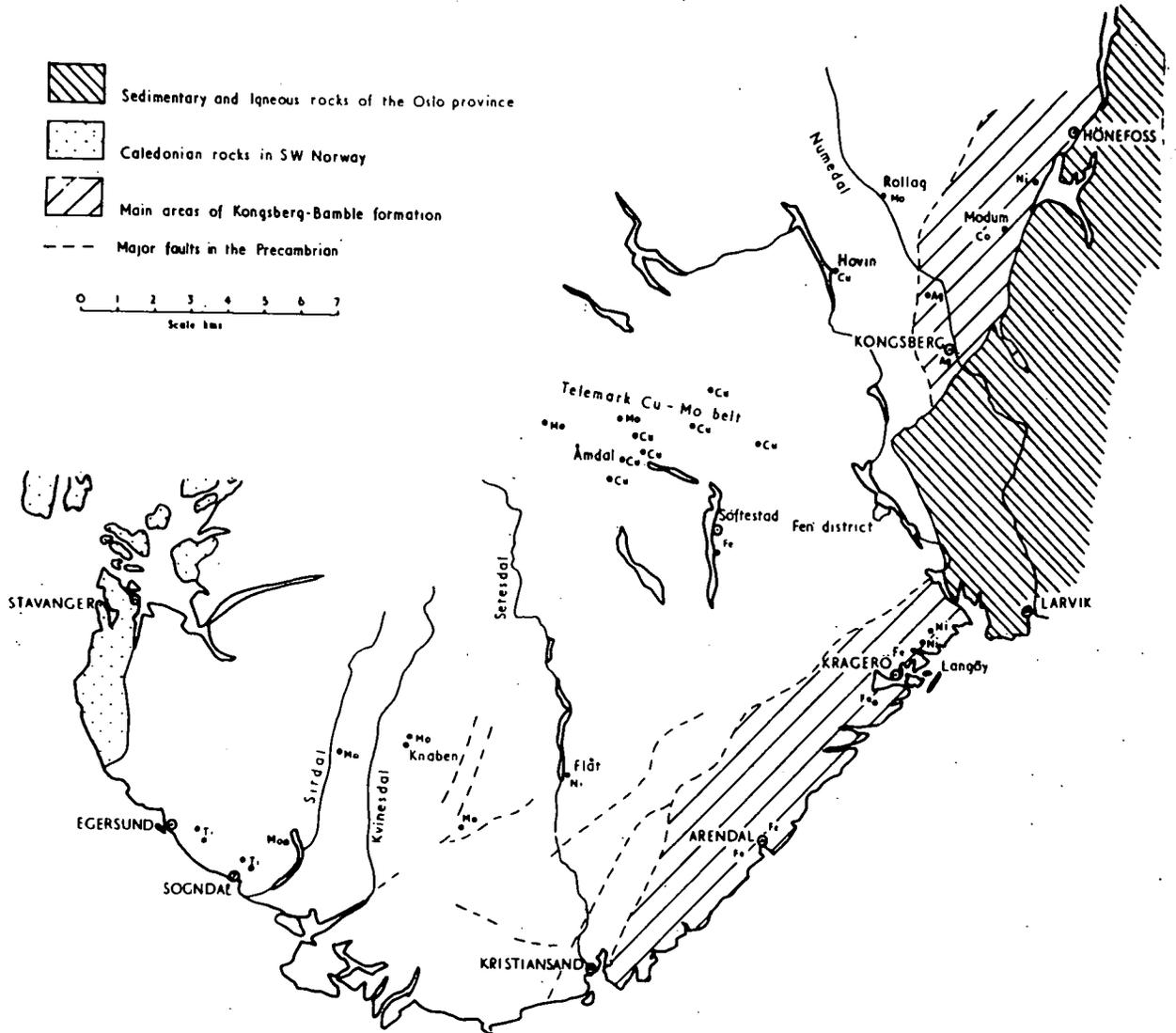
In this connection Vokes' (1961) paper on metallogenetic province in Norway is of interest. The northern segment of the Kongsberg - Bamble Formation contains three mining districts of major importance: the Erteli nickel mines, the Modum cobalt mines and the Kongsberg silver mines. The Erteli deposit is a complex of 'hyperites' containing rich lenses of nickeliferous pyrrhotite-pentlandite ores. This characteristic elemental association in such a small area (Figure 153) is surely significant, but it is difficult to see how deposits of such dissimilar character could be related to each other. The possibility exists that in the large magma chamber, parental to the ubiquitous 'hyperite' bodies of the Kongsberg - Bamble Formation, some form of differentiation occurred such that a nickeliferous sulphide phase separated out, tapping of this differentiate giving rise to the nickeliferous bodies. By this process cobalt could become enriched relative to nickel in the remaining magma, but would probably not be present in sufficient quantities to produce the Modum deposits from a sill of the observed dimensions.

It is possible that a process, similar to that postulated by Sampson and Hriskevich has occurred, where a volatile residuum enriched in cobalt and arsenic is produced as the end product of differentiation in the parental magma chamber. It is not easy to see, either at Cobalt or Modum, how these volatiles come to be deposited in such close association with the basic sills.

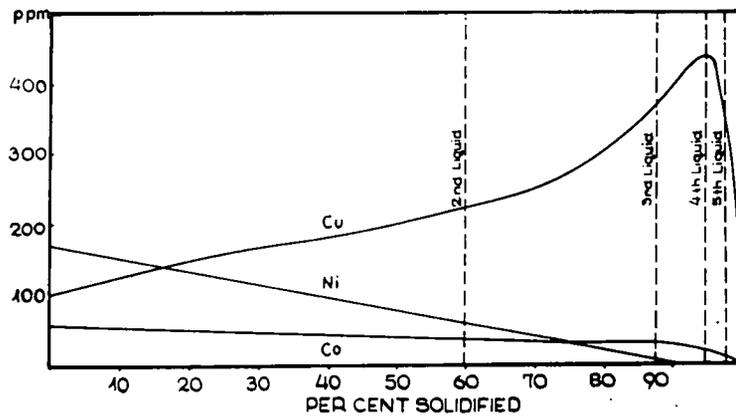
An argument for considering the mineralisation to have its source in either the granophyric differentiate of the gabbroic magma

Figure 153: Top: Map of South-western Norway taken from Vokes (1958), showing the distribution of ore deposits in the region. The northern segment of the Kongsberg-Bamble Formation is characterised as a nickel-cobalt-silver metallogenetic province by Vokes (op.cit.). The mode of occurrence of the Kongsberg (Ag), Modum (Co) and Erteli (Ni) ores is discussed in the text.

Bottom: Illustrating the decrease in nickel and cobalt contents of the residual liquid during igneous differentiation , with a corresponding increase in the cobalt: nickel ratio in the more acid rocks.



Geological sketch-map of part of South-western Norway showing the locations of the more important ore-deposits.



Nickel, copper and cobalt in residual liquids of the Skaergaard intrusion. Data from Wager and Mitchell (1951).

Fig. 153

or a completely unrelated acid igneous source is the high cobalt:nickel ratio of the deposit. The significance of cobalt:nickel ratios has been recently discussed by Davidson (1962), from a review of published analyses of igneous rocks he concludes:

"that the tenor of both cobalt and nickel in igneous rocks increases from acid to ultrabasic types, and that the ratio of cobalt to nickel decreases progressively in the same direction."

Thus ore deposits derived from acid differentiates or granitic rocks should show characteristically high cobalt:nickel ratios. Davidson (op.cit.) Table 3, lists such deposits, rather unfortunately he chooses the copper sulphide deposit of Outokumpu and the Skuterud occurrence as examples, neither of these can be unequivocally shown to be derived from an acid igneous source (Vokes and Gammon, 1962), A better summary of cobalt:nickel ratios in ores in relation to the associated rock type is given by Wilson and Anderson (1959) for basic and ultrabasic rocks.

The clarke of nickel (0.008%) is much greater than the clarke of cobalt (0.0023%). Analyses of sedimentary rocks with very few exceptions, show a similar relationship between the tenors of these elements. Thus Davidson considers that the strata-bound copper - cobalt deposits of Central Africa and the U.S.S.R. have no actualistic analogues in later sediments and are more likely to have been derived from hydrothermal emanations from granitic intrusives, possibly modified by later lateral secretion.

The cobalt:nickel ratio of the Modum deposit varies from 0.5 to 7 with an average value of around 1.5 (Figure 151), thus it would seem, if Davidson's argument is acceptable, an acid hydrothermal source was the most likely.

In the normal differentiation process nickel is preferentially concentrated in the early formed silicate minerals, leading ultimately to a stage where cobalt is present in excess over nickel in the residual liquid (Figure 153).

It is possible that this process is aided by the separation of an immiscible sulphide phase. Hakli (1963) has shown that chemical equilibrium is attained between silicate and sulphide phases in the 25 Finnish Precambrian nickeliferous basic intrusives he studied. Kullerud (1963) has shown that if sulphur is introduced into a basic magma nickel and iron will be extracted from the silicate phases until the stoichiometric requirements of the sulphide phase components are met. The sulphide phase produced is highly nickeliferous with nickel:cobalt ratios varying between 14 and 28 (Hakli, op. cit.). The source of the sulphur in the hyperites may well be from assimilated fahlband rocks.

Thus by a combination of these processes the last differentiation stages of the parent magma can become relatively enriched in cobalt. Escape of volatiles carrying cobalt and arsenic from the magma chamber could occur at this time giving rise to the cobalt-arsenic lodes. That the Modum 'sill' itself may not be the source of the observed cobalt is indicated by the analyses (Figure 45) which shows the cobalt content to be about the same as in other intrusions of this type and age (Ball, personal communication).

Davidson (1962b), in reply to Vokes and Gammon (op. cit.), made the interesting suggestion that the spatial relationships between gabbro and ore at Skuterud are reminiscent of the controls exerted by diabase, amphibolite, and lamprophyre wall-rocks in

localising ore-shoots within the Hercynian vein mineralisation of Central and Western Europe, "the granitic derivation of which is altogether free from doubt", in his opinion.

THE URANIUM MINERALISATION

Introduction

It has been suggested by van Autenboer (1957) and Rosenqvist (1949) that the cobalt ore at Modum is accompanied by a uranium mineralisation.

Van Autenboer carried out a radiometric survey of the Modum area and found that the fahlband zones showed a slightly higher activity than the surrounding rocks. Anomalies were found of quite a high magnitude in the cobaltiferous areas, but the anomalies were of small extent, on the walls of the opencasts or in the adit systems where these crossed one of the mineralised zones. He also found that several of the cobalt ore specimens in the Mineralogical Museum, Oslo, showed a high radioactivity. Maps and sections were prepared by the N.G.U. party showing the anomalous zones delineated by their survey. These have since been mislaid but are said to show that the radioactivity decreases with depth.

Rosenqvist had access to captured German assays and mine plans at the cessation of hostilities in 1945. These are said to show that the cobalt values were decreasing in depth. Many of these reports are also no longer available.

On the basis of the above evidence van Autenboer has concluded that the uranium and cobalt mineralisations were contemporaneous and that the two elements have a similar distribution in the deposits.

An attempt has been made during this investigation to check and enlarge upon these conclusions by the determination of the uranium distribution in the field, the relation of this to the cobalt mineralisation, the paragenetical relationships of the cobalt and uranium minerals and the mutual relationship of the distribution of the two elements.

Field determination of the uranium distribution

The loss of the anomaly maps produced by the N.G.U. meant that it was virtually necessary to start from scratch. The Central Mine, Skuterud, was selected for study as this provides the best section through the ore zone in a vertical sense and several radioactive anomalies were known from the text of van Autenboer's report.

The method followed was to traverse along the crosscut noting the scale reading every few metres and relating this to the geology of the wall rocks. When an area of high activity was encountered more closely spaced readings were made in an attempt to locate the source and collect a specimen for laboratory analysis.

The radiation detectors used were a Harwell type 1280B ratemeter and a Phillips PW 4014 radiation monitor. The Harwell instrument was found to be the most convenient in making traverses but the Phillips instrument, fitted with a probe type PW 18504, was used for measuring the beta radiation and localising the source.

Results

The measurements made using the Harwell ratemeter were of use only in a comparative sense as the scale had not been calibrated, but when using the Phillips beta detector readings could be made up to activities of 30 mr/Hr. The results are shown in Figure 154.

Ludwig Eugen Crosscut The general level of activity here was rather low. Background at the portal was about 2 scale divisions, and readings remained at this level during the traverse of the sillimanitic nodular siliceous granulite forming the first section of the crosscut. In crossing the large body of garnet-amphibolite readings fell to nearly zero on the scale. To the west of this amphibolite the readings, reached a level of 5 scale divisions in the quartzites and dropped again in traversing amphibolite layers. The highest readings obtained in the crosscut were under the stopes worked for cobalt, where up to 8 scale divisions were recorded.

Klara Crosscut Although slightly higher than in the Ludwig Eugen Crosscut the readings here showed the same pattern, i.e. they were low during traverses of amphibolite (1-3 scale divisions) and higher in the quartzites (5-7 scale divisions). In the worked zone at the extreme western end of the crosscut readings reached a maximum of 11 scale divisions.

Measurements were also made in a general way of the large stopes (Vestre no 11 and Østre no 11) reached by ladders from the overlying Klara Level. In these stopes, activity ranged from 8-16 scale divisions. In one place, on the west wall of Vestre no 11, a scale reading of 47 scale divisions was obtained over an area of about one square metre. The smooth nature of the stope wall at this locality made it impossible to sample.

Forhaabning Crosscut This small crosscut, at a higher level than Klara, gave some rather interesting results. The same general pattern was obtained, low values for the amphibolites (1-3 scale divisions), higher for the quartzites (7-9 scale divisions) and the worked zones gave readings of about 12 scale divisions. At the junction with the south end of the main opencast working the readings rapidly increased. Using the beta probe it was possible to delimit at least 5 separate areas showing activities of up to 28 mr/Hr. As can be seen from the graph (Figure) of this traverse the Harwell instrument reached its maximum reading of 50 scale divisions. The areas of high activity were found to be associated with a fairly pure and massive white quartzite containing disseminated cobaltite. A small charge was used to blast away a sample of this quartzite, which was collected for laboratory analysis.

Surface traverse One traverse made in an east-west direction across the ore zone at the southern end of the Central Mine confirmed van Autenboer's results. Readings of about 2 scale divisions were obtained on the garnet-amphibolite, these increased in the vicinity of the mines (mostly underlain by dumped material) to about 5 scale divisions. In the area of the worked zone readings reached a maximum of 10 scale divisions and then tailed off to the west.

Relation to cobalt mineralisation

From Figure 154 it could be concluded that there is a tendency for the radioactivity to decrease with depth in the mine. This effect is not very marked and could well be spurious, particularly when it is considered that in the stopes lying between the Klara and Ludwig Eugen Levels quite high anomalies were obtained.

In this study all high indications of uranium were connected with known areas of cobalt mineralisation, this could be taken to indicate that there is some genetic connection between the two. However elsewhere in the Modum area radioactive anomalies were found which are not connected with cobalt mineralisation, for example to the south-east of the path leading to Skuterudsaeter a pegmatite was found which contained isolated crystals of uraninite giving an overall anomaly of some 20 scale divisions on the Harwell instrument. In the North Mine radioactive anomalies were found connected with the pegmatites which cut the workings at this locality. Van Autenboer in his report gives several other examples of radioactive anomalies in the Modum area which are not connected with the cobalt mineralisation.

If there is some genetic connection it would be reasonable to expect that in those areas of the mine where known cobalt ore remains the radioactivity should be higher. This was not found to be the case, in the Troeger Stope there is abundant evidence of cobalt mineralisation but the level of activity was very low, this was also found to be the case with known ore on the Ludwig Eugen Level.

Thus on the basis of the field survey it has not been possible to show a definite genetic connection between the cobalt and uranium mineralisations.

Mineralogy

Preparation of samples The radioactive samples collected from the Forhaabning Crosscut were checked for radioactivity with a sensitive Geiger-counter in the radiochemistry laboratory of the Mineralogical Museum, Oslo. Faces showing high activity were selected for polishing.

Autoradiographs Autoradiographs were produced of selected hand and polished specimens, the following procedure was found to give the best results. The specimen was laid, face downwards on a photographic plate in the dark room. A weak light was turned on for a few seconds to give an outline of the specimen for orientation purposes. The plate and specimen were then left for 11 days in a light proof box. After this period of time the plate is developed and used as a negative for preparing prints.

In Figure 155 two autoradiographs prepared in this way are shown. Figure 155a shows the radioactive areas connected with a polished specimen from the Forhaabning Adit quartzite, strong activity is seen to be connected with a crack cutting diagonally across the specimen and also with small grains disseminated within the quartzite. Figure 155b shows a hand specimen from Middagskollen Mine, the radioactive minerals are clearly seen to be present in a band which parallels the foliation of the rock.

Separation of the radioactive minerals The mineral responsible for the radioactivity was separated out of the rock by the following technique.

Approximately 3.5 kgs. of the rock collected from the Forhaabning Crosscut were mechanically crushed. The crushed rock was then put through a series of sieves to give several fractions of roughly the same grain size. The various size fractions were then put on a 'Superpanner' and the heavy mineral fraction pipetted off and collected. The residual rock showed negligible radioactivity when the heavy minerals had been removed. The collected concentrates were dried and then put through a Franz magnetic

Figure 155a: Autoradiograph of specimen collected from cobaltite-uraninite-quartzite in Forhaabning Adit, Skuterud. The radioactivity is seen to be connected with a crack running diagonally across the specimen. In addition small , disseminated specks are visible in the surrounding rock, these were found to be due to small uraninite crystals.

Figure 155b: Autoradiograph of siliceous granulite, Middagskollen Mine. The radioactive minerals are seen to be restricted to a band paralleling the foliation in the rock.

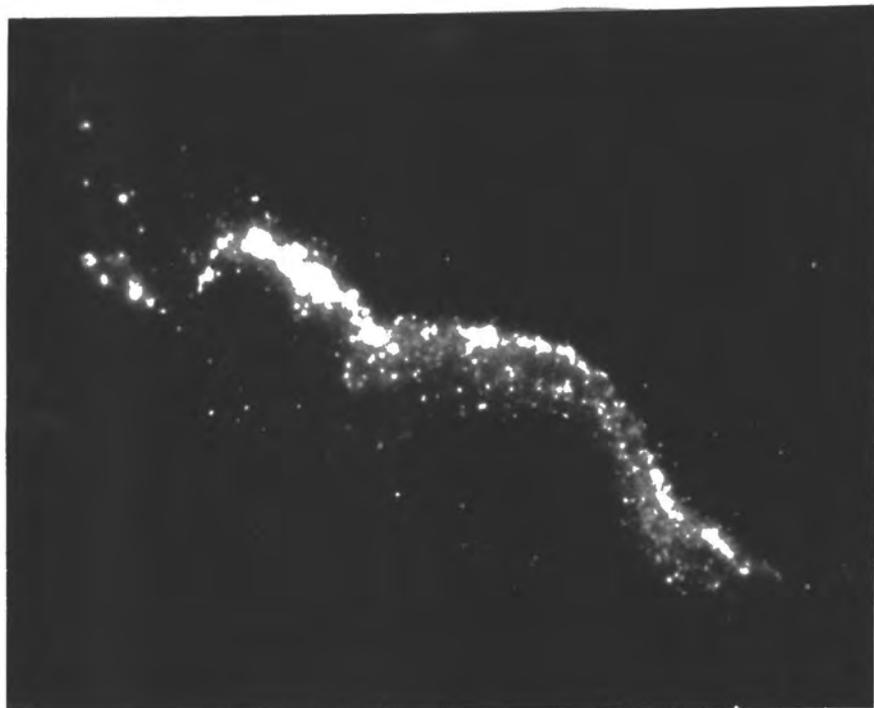


Figure 155a



Figure 155b

separator, fractions of different magnetic susceptibility were collected and examined optically and radiometrically.

The uraninite Under the binocular microscope the radioactive mineral showed a euhedral crystal habit, the cubic form (100) dominating all the crystals. The cubes were all a grey-black colour and showed no signs of alteration.

Several of the crystals were plucked out by hand, crushed and X-rayed. The diffraction pattern confirmed the mineral to be uraninite. The spacing of the d_{151} line was measured on a film taken with a Guinier powder camera, using $\text{Pb}(\text{NO}_3)_2$ as an internal standard, it was found to measure 1.055 \AA corresponding to a cell edge of $a_0: 5.482 \text{ \AA}$.

Berman (1957) has shown that the naturally occurring uranium oxides are found in three main parageneses. These are vein and sedimentary pitchblendes and pegmatitic uraninites. Furthermore he demonstrated that there is constant variation in cell dimensions for the three groups. In Figure 156a Berman's frequency histograms have been converted to frequency curves and the positions of the Modum uraninite indicated. It is seen that the Modum material falls well out of the pitchblende range and nearly coincides with the maximum for unit cell size for uraninites from pegmatites.

In recent years considerable attention has been directed to the reasons for this variation in the unit cell parameter. It has been found that there is linear change in a_0 in the solid solution series between UO_2 ($a_0: 5.42$) and ThO_2 ($a_0: 5.60$). Montgomery (1957) has measured both the thorium content and cell size for a number of uraninites and uranothorianites, if his results are

Figure 156: a. Variation of unit cell parameter with natural occurrence for the uranium oxides. The unit cell dimension of the Modum uraninite roughly coincides with the maximum frequency for uraninites from pegmatites. (Adapted from Berman, 1957).

b. Variation of unit cell parameter with thorium content from data given in Montgomery (1957). The gamma scintillation spectroscopy described on page 298 indicates that the Modum uraninite contains about 30% thorium.

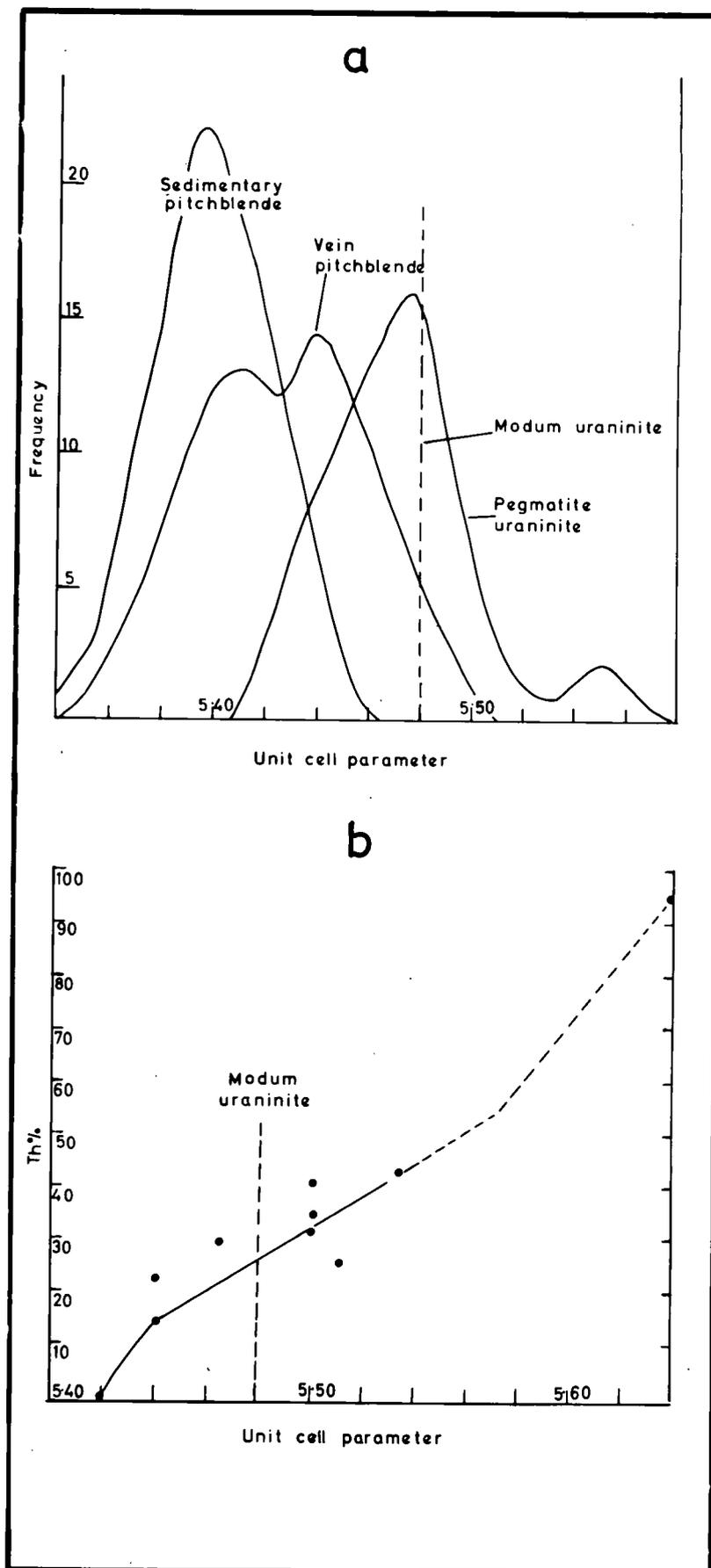


Fig. 156

plotted on a graph (Figure 156b) it is seen that they do not coincide with the results for the synthetic end-members, there is a tendency however for specimens with a higher Th content to have larger values of a_0 . The length of the unit cell edge is thus a useful rough guide to the thorium content of the mineral. Using this relationship it is possible to correlate the unit cell size (5.482) with a thorium content of 20-40% for the Modum material.

This estimate of the thorium content has some confirmation from the results of the gamma-spectrometry described in the next section. Using this technique it was found that specimens rich in uraninite had a Th:U ratio of 0.37:1.

In polished specimen the uraninite is grey in colour with a slight brownish tint against the gangue minerals. It shows no reflection pleochroism. Under oil immersion the colour is a much darker grey. The reflectivity is low (15.5% compared with standard Elba pyrite: 54.5%). The mineral takes a poor polish, the Vickers hardness value (G.K.N. microhardness tester, 100 gm. load) was in the range 780-802. Under crossed nicole the mineral is isotropic with occasional brown internal reflections. Dispersion of the reflection rotation (DR_r) was $v \gg r$, weak.

In none of the specimens examined episcopically were cobaltite and uraninite seen in contact. They were frequently present in the same section, for example the radioactive quartzite from Forhaabning Crosscut contains both cobaltite and uraninite in disseminated euhedral crystals slightly corroded by the quartz

gangue. The observed textures suggest that the uraninite is being attacked and replaced by the silicates.

Gamma scintillation spectroscopy

As a gamma-ray spectrometer was available at the Geological Museum, Oslo, a technique was evolved for determining the amounts of uranium and thorium in some of the Modum rocks. At the time of the investigation cobalt had been determined, by X-ray methods, on a series of amphibolite samples so these same samples were used to compare the distribution of the cobalt and uranium.

Principles of the method Each radioactive isotope has a unique disintegration scheme which proceeds at a characteristic rate by the ejection of one or more particles or rays of discrete or distributed energy. By the use of a detector, whose response is proportional to the energy of the incident particle, discrimination between radiations on an energy basis and therefore between different radioactive isotopes becomes possible.

Gamma-ray discrimination depends upon the fact that the reaction of the gamma-ray with a thallium activated sodium iodide crystal to produce photons is an essentially linear process. The photons are detected by a photomultiplier tube and converted into a pulse of current whose amplitude is in turn proportional to the energy of the original gamma-ray. The resulting voltage is then amplified linearly by the detector pre-amplifier.

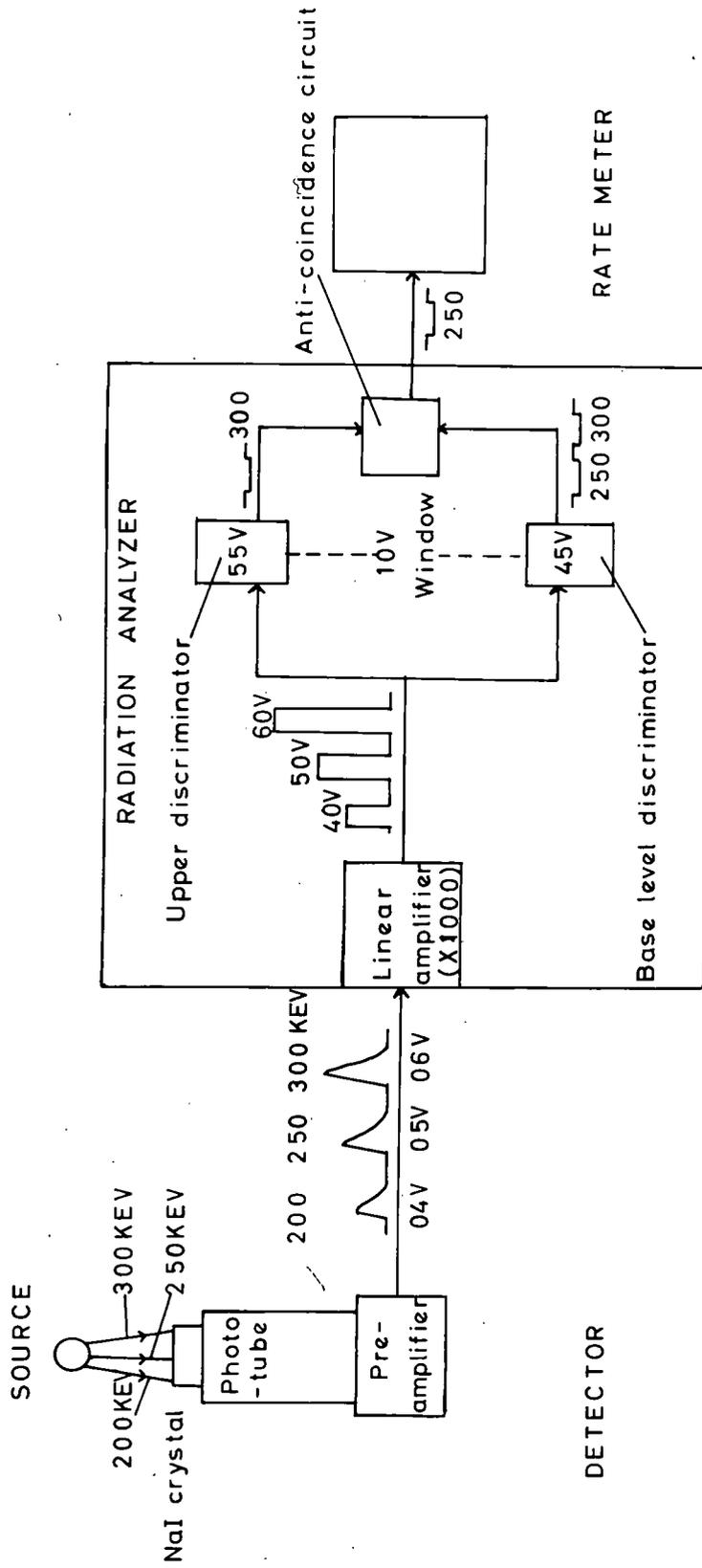
The amplified voltage is passed into a radiation analyser which provides the means of discrimination between gamma-rays of different energies. This discrimination permits the reduction of background and Compton scattering errors in routine gamma

analysis. The voltage pulses are fed into the analyser in which an adjustable base-level threshold eliminates all pulses below a certain amplitude. Riding above this base-line threshold is another adjustable voltage level ('the window') above which all pulses are eliminated. The instrument therefore provides a channel, of adjustable width, which can be set at any voltage and used to scan the spectrum of pulses. Pulses passing through the window are recorded on a rate meter. Channel width and the voltage level at which it operates may be calibrated in terms of gamma-ray energy (Figure 157).

Instrumentation and method The following analytical method was employed. The rock sample was crushed to pass a B.S.S. 100 mesh sieve and homogenised in a rotary blender, 5 gms. of the sample were placed in a sealed polythene tube. The tube was allowed to stand for a week to attain equilibrium with respect to radon. The tube was designed to fit snugly into a well-type detecting crystal. The detector unit consisted of a Nuclear-Chicago Model DS5-5 scintillation detector with a gamma detecting, well-type crystal of $1\frac{7}{8}$ " diameter and $2\frac{1}{4}$ " width. The well was $2\frac{1}{32}$ " in diameter and $1\frac{1}{2}$ " deep, the detector was surrounded with a specially designed lead shield. The photomultiplier was a Dumont type 6292; interfaces of the crystal units, light pipes and photomultiplier tubes were optically coupled with a clear silicone compound. The crystal, which consists of thallium activated sodium iodide, was hermetically sealed in a spun aluminium can with a glass window..

The radiation analyser was a Nuclear-Chicago model 1810 single channel differential pulse height analyser, the base level discriminator could be varied from 1 to 100 volts and the upper discriminator (or window) was variable from 0 to 10 volts.

The pulses which passes through the 'window' of the pulse height analyser were recorded on a Nuclear-Chicago model 186 decade scaler. The number of pulses counted was indicated on three decade strips and a four digit register, allowing a maximum reading of 10 counts. The time elapsed during a counting period was indicated on a six digit timer. All samples were counted for periods exceed-1,500 minutes.



GAMMA SCINTILLATION APPARATUS USED IN THE PRESENT STUDY

Fig. 157

Gamma-spectrum of quartzite from Forhaabning Crosscut All elements found in nature with an atomic number greater than 83 (bismuth) are radioactive. They belong to chains of successive decays, and all species in one such chain constitute a radioactive family or series. The two common series are the U^{238} family, which after 14 transformations, reaches a stable end product Pb^{206} , and the Th^{232} family of which Pb^{208} is the stable end product. Of the remaining elements only potassium with a mass of 40 on the atomic weight scale is commonly found in nature as a radioactive isotope.

Thus the gamma-ray spectra obtained from rock samples are likely to consist of contributions from the U^{238} and Th^{232} families and K^{40} . This is illustrated by the spectrum obtained from a sample of the uraniferous quartzite collected in Forhaabning Crosscut, this spectrum was measured by the following procedure:

The background was measured by placing an empty polythene tube in the detector and measuring the integral gamma count above a base level of 100 KeV (442 counts/min.). Integral counting on the sample gave the following results: base-level 10 KeV- 11,600 c/m; 100 KeV-9,092 c/m; 1,000 KeV- 882 c/m. The instrument was calibrated by finding the correct settings for a Ce^{137} standard. Readings of the gamma-spectrum were then made on the sample from 0.000 to 2.800 MeV using a window width of 5 volts with gain X8, 2.5 volts with gain X4, 1.25 volts with gain X2 and 0.625 volts with gain X1. Counting time was 0.5 mins. for each reading.

The resulting spectrum, plotted on a log.-normal scale is shown in Figure 158. It will be noted that it has been possible to account for most of the peaks as being due to members of the U^{238} or Th^{232} families, with the 1.46 MeV peak being due to K^{40} . A spectrum such as this thus has contributions from three major sources, if two of these were present in constant amounts then the heights of the peaks would be proportional to the concentration of the third. To illustrate this spectra were measured on a series of

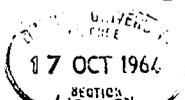
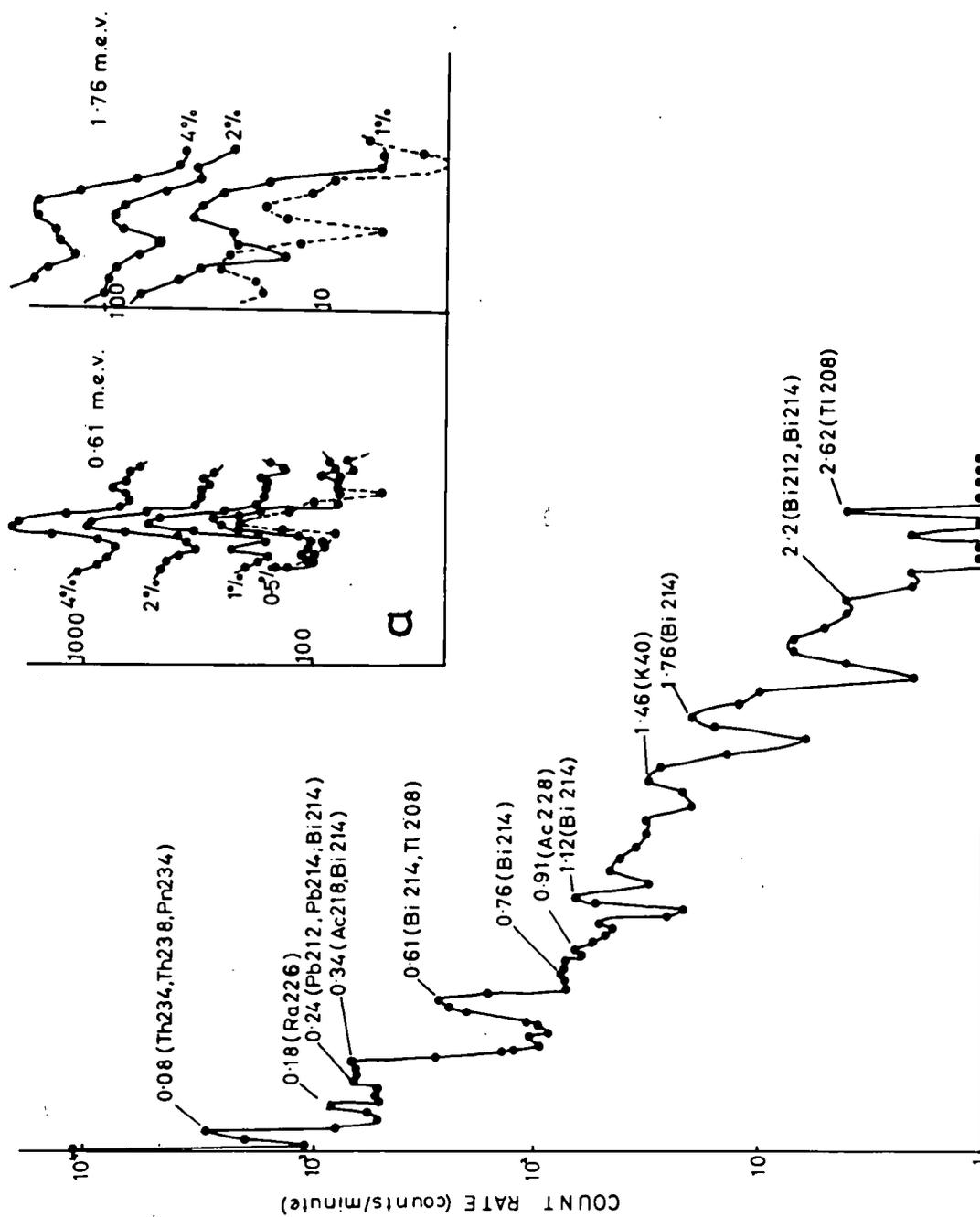


Figure 158: Gamma energy spectrum obtained from the uraninite bearing quartzite collected from Forhaabning Adit.

Figure 158a: (inset) Peaks obtained for the 0.61MeV and 1.76MeV regions for a series of uranium standards. The 0.61MeV peak is due to Bi²¹⁴ in the uranium series and Tl²⁰⁸ in the thorium series, the 1.76MeV peak is also due to Bi²¹⁴.



GAMMA-RAY ENERGY (m.e.v.)

Fig. 158

uranium standards prepared by the National Bureau of Standards, New Brunswick. Figure 158a shows the 0.61 MeV peak, due to Bi^{214} in the uranium series and Tl^{208} in the thorium series, and the 1.76 MeV peak due to Bi^{214} , as obtained from these samples. As these standards contain no appreciable thorium or potassium the peaks are roughly proportional in height to the uranium concentration. These peaks may be compared with those obtained from the uraniferous quartzite (Figure 158), a marked difference in geometry is observed due to the influence of thorium and possibly potassium. Thus simple peak height cannot be used to determine the uranium content, assuming the uranium to be proportional to the amount of the daughter element (Bi^{214}) being measured.

Channel selection and calculations Uranium was determined from the 1.76 MeV peak (Bi^{214} ; Tl^{208}); the following conditions were employed, base-level 38.75 v., window 8.5 v., gain X1, equivalent to a base of 1.55 MeV with window at 0.35 MeV in terms of gamma energy. Thorium was determined from the 2.62 MeV peak (Tl^{208}); with settings, base level 62.5 V (2.50 MeV), window 5v (0.20 MeV.), gain X1. All counting rates were corrected for background. The nearest K^{40} peak occurs at 1.46 MeV but the intensity falls very steeply to zero at about 1.65 MeV and causes no interference with the peaks being measured here.

Let U be the uranium content of the uranium standard in ppm; U_1 the counting rate of this standard in channel 1 (1.76 MeV peak); U_2 the counting rate of the standard in channel 2 (2.62 MeV peak). Let T be the thorium content of the thorium standard in ppm; T_1 the counting rate of this standard in channel 1, corrected for the 0.002% uranium that it contains, T_2 the counting rate of this standard in channel 2, similarly corrected for uranium. Let R_1 and R_2 be the counting rates from the unknown sample in channels 1 and 2 respectively.

In the value found for R_1 , if X and Y are the counts/minute due to uranium and thorium respectively, the following equations will give the required result:

$$X + Y = R_1$$

$$(U_2/U_1)X + (T_2/T_1)Y = R_2$$

It can be shown that:

$$X = \frac{(T_2/T_1)R_1 - R_2}{(T_2/T_1) - (U_2/U_1)}$$

and

$$Y = \frac{(U_2/U_1)R_1 - R_2}{(U_2/U_1) - (T_2/T_1)}$$

Hence:

$$\text{Uranium} = X(U/U_1) \text{ p.p.m.}$$

$$\text{Thorium} = Y(T/T_1) \text{ p.p.m.}$$

As the calibration constants were determined on the basis of a standard sample weight, the unknown must also be of this weight. In the spectrometer used in this investigation the following values were obtained for the various constants:

$$U_2/U_1 : 0.00367. \quad T_2/T_1 : 0.294. \quad U/U_1 : 25.2857$$

$$T/T_1 : 77.042 \quad . \quad U_2/U_1 - T_2/T_1 : -0.29033.$$

Average backgrounds were 7.87 c/m and 0.784 c/m for channels 1 and 2 respectively.

Results The data obtained in this investigation are shown in table 14. The amphibolites contain very little uranium and thorium, the amounts present are close to the detection limit of this method, (0.5 ppm U and 1 ppm Th), little reliance should be placed on numerical values of less than about 4-5 ppm. Th and 2 ppm U. However it is clearly seen that there is no correlation between the uranium and cobalt concentrations. Samples 419 and 280, collected from a cobaltiferous working, contain negligible uranium, while sample 442,

Table 14.

SPECIMEN NUMBER	Co. (ppm.)	U. (ppm.)	Th. (ppm.)
Standard Wl.	51	n.d.	n.d.
4% U. Standard	n.d.	38,912	5
2% U. Standard	n.d.	18,782	3
1% U. Standard	n.d.	9,362	1
0.5% U. Standard	n.d.	4,047	0
0.002% U.) 0.05% Th.) -Standard	n.d.	17	479
0.0004% U) 0.01% Th.) -Standard	n.d.	1.5	89
419	5,700	1.5	0
280	6,500	3	0
365	59	n.d.	n.d.
363	80	2.3	0
397	57	2.8	0
391	55	0	0
411	94	1.4	0
433	65	0	0.1
444	67	4.5	0.5
382	73	0	0
445	48	1	0
396	51	0	3.4
395	47	0.8	0
407	48	0	0
705	58	0	0
352	47	3	6
400	64	0	0
432	153	n.d.	n.d.
318	58	n.d.	n.d.
362	58	6	5
442	62	179	0
381	69	n.d.	n.d.
351	48	n.d.	n.d.
353	50	4	6
443	60	0	0.1
390	42	n.d.	n.d.
Uraninite quartzite (876)	4,030	339	126

Localities: Amphibolites:- 419-Klara Adit; 280-South Mine; 365, 363, 397, 391-Ludwig Eugen Adit; 411, 433, 444-Klara Adit; 396, 395-Ludwig Eugen Adit; 407-South Mine; 705-North Mine; 352-Høgaas; 400-Stolldammen pukverk; 432-Klara Adit; 318-Ludwig Eugen Adit; 362, 442-Klara Adit; 381-Ludwig Eugen Adit; 351-Butjern; 353-Muggerud Trial; 443-Klara Adit; 390-Ludwig Eugen Adit. Uraninite quartzite (876) - Forhaabning Adit, Skuterud.

collected from the non-cobaltiferous centre of the main garnet-amphibolite at Skuterud, contains 0.018% uranium.

Conclusions

From the present investigations it has not been possible to draw any definite conclusions regarding the relationship between the uranium and cobalt mineralisations. Both have a similar mode of occurrence in the Modum Formation rocks, they are both present in the same localities and appear to have similar distributions.

Nowhere has it been possible to determine the relative ages of the two mineralisations where they occur in the same rock. The presence of uraninite in the post-kinematic pegmatites may suggest that the uraninite was introduced after the metamorphism of the cobalt ores. However, several generations of uranium may have been introduced, and van Autenboer (op.cit.) regards much of the uranium as being of Permian age.

A sample of the uraninite, associated with cobaltite in the quartzite from the Forhaabning Crosscut, has been submitted for a radiometric age determination.

THE LATE VEINS

A series of quartz and calcite veins are common in the northern part of the Kongsberg - Bamble Formation. At Kongsberg these veins are argentiferous where they cross some of the fahlband zones and have been studied by many authors. An analogous set of veins is present at Modum.

Field relations

The majority of the veins have a trend of 255° to 275° and follow the transverse 'ac'-joints (p.126). They vary in thickness from a few millimetres to about 20 cms. Many veins are also observed which have a trend of 350° to 010° and follow the axial plane of the major fold structure (joint-set 6, p.127). This latter group of veins frequently contain vugs with well developed quartz, calcite and pyrite crystals. Vugs have been observed with a maximum dimension of some 40 cms.

A subordinate group of anastomosing veins is also present. These follow no clearly defined direction and brecciate the country rock. They are usually very thin and contain few vugs, and are in close association with the two major sets mentioned above. The relationships of these three vein types are particularly clearly seen in the eastern opencast working of the South Mine, Skuterud.

A rather different type of vein is associated with some of the major shear zones in the underground workings. Large, massive lenses of calcite, which appear to have replacive relations to the surrounding rocks are developed. Such lenses are clearly seen in the Ludwig Eugen Level (Figure 114) and in the southern adit of the Svartefjell Mine (Figure 101).

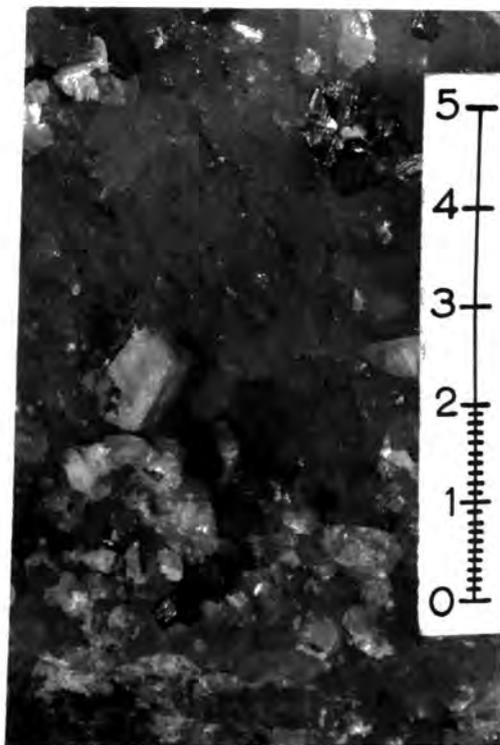


Fig.159: Rhombohedra of calcite growing on early quartz. Specimen from the Mineralogisk-Museum collections, labelled "Skuterud-gruben, Modum".
Scale in centimetres.

Mineralogy

Calcite Calcite is the commonest mineral in the veins. In the large massive calcite lenses it is present in an intergrown, anhedral aggregate with very few associated minerals. In the vugs calcite crystals are developed, commonly as rhombohedra growing on early formed quartz crystals (Figure 159). A variety of calcite, described from Kongsberg as skiferspat ('slate-spar') by Münster (1894) is also present. The 'slate-spar' has extremely well developed (0001) faces, while the hexagonal prism faces are poorly developed, the resulting platy crystals may be arranged parallel to each other or crossing at various angles to form a network which may be infilled by quartz or other minerals.

Quartz Quartz is extremely well developed in some of the vugs, small euhedral crystals, measuring up to 1.5 cms. in length, are observed. These are demonstrably earlier than some of the calcite, terminations of the quartz crystals are seen protruding through a later covering of calcite (Figure 160).

Pyrite Pyrite is the commonest sulphide mineral in the late veins, it is euhedrally developed in pyritohedra which can attain a size of 1 cm. in diameter. They are seen growing, in association with the calcite, on the early formed quartz (Figure 161).

Some pyrite crystals from the South Mine at Skuterud were electrolytically etched, following the method described by Suzuki (1962), very fine growth zones become apparent as a result of this treatment. These zones are developed in two stages, an earlier core containing parallel zones terminating in one clearly marked period of arrested growth, followed by an area containing growth zones paralleling the external crystal faces. Similar zones are also seen

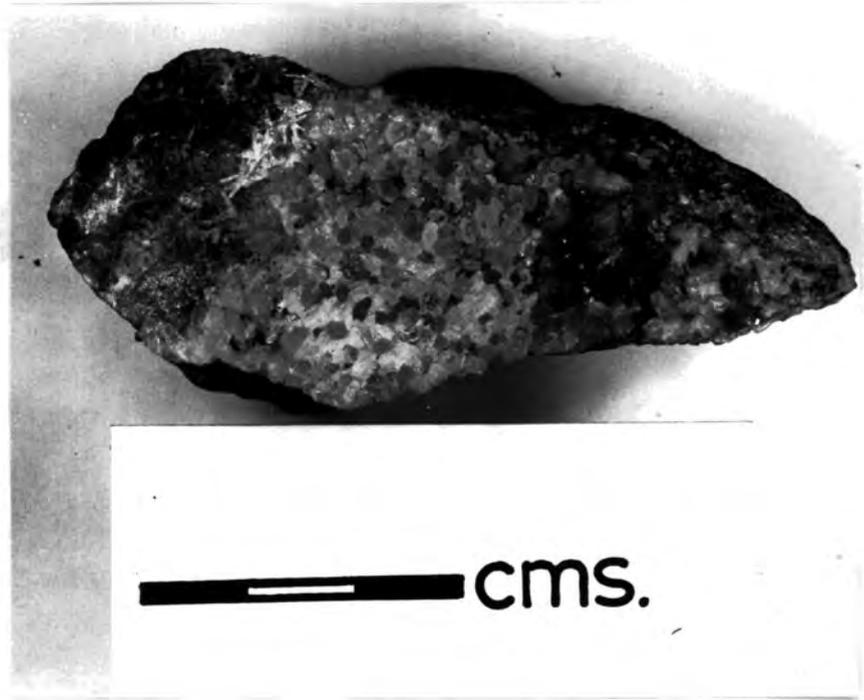


Fig.160: Terminations of quartz crystals protruding through a later covering of calcite. South Mine, Skuterud.

on freshly fractured crystal surfaces (Figure 159).

Galena Galena is the commonest sulphide in the veins after pyrite. It has only been observed in the calcite rich veins associated with joint-set 6. One example of a euhedral crystal was collected, from the veins in the 'Weisscobolt Strossen', which had dominant (100) faces modified by (111) faces. In polished specimen the galena shows a flat featureless field with no associated sulphides.

Chalcopyrite Chalcopyrite is a minor constituent of the late veins. It was most frequently observed in the opencast workings of the Central Mine, Skuterud where it is associated with thin films of calcite and minor galena on joint surfaces.

Chlorite Chlorite is present in the calcite veins of the Svartefjell Mine. It occurs in thin films along cleavage planes and crystal faces of the calcite. It is also present in pockets as a green powder covering all other minerals. In some cases it forms small spherules reaching 0.8 cms. in diameter. The chlorite is very similar to that described in detail from Kongsberg by Neumann (1944).

Native silver No native silver has been observed during this study. Bøbert (1846) states that it was present, as a mineralogical rarity in some of the veins encountered during stoping at the Central Mine.

Temperature of formation

The early formed quartz crystals contain some very good fluid inclusions. A preliminary run, made by Dr. F.J.Sawkins on material from the South Mine, Skuterud, indicates that filling temperatures lie in the range $245^{\circ} - 255^{\circ}\text{C}$.

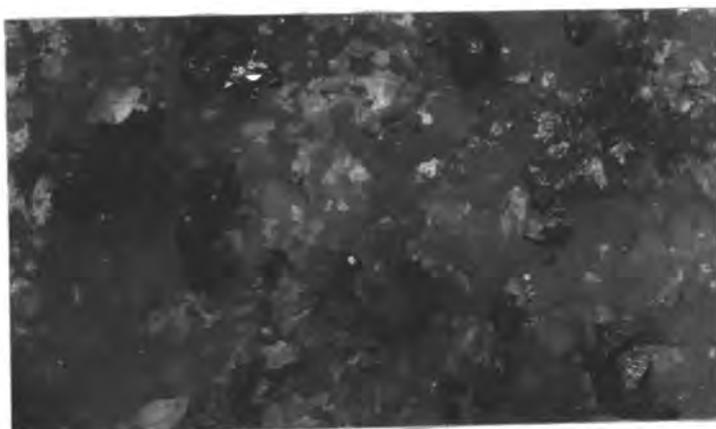


Fig.161: Pyrite crystals euhedrally developed in quartz-calcite-pyrite vug, showing pyritohedral habit. The pyrite appears to be later than the quartz. The unfractured crystal at the top of the photograph measures 5 mm. in diameter.

Time of deposition

The veins at Modum are similar in every respect to those described from elsewhere in the region, notably from Kongsberg. There is general agreement that these veins are of Permian age connected with hydrothermal activity subsequent to the emplacement of the igneous rocks of the Oslo area. (Neumann, 1944).

Moorbath and Vokes (1963) have recently obtained radiometric ages on galenas from these veins which are compatible with a Permian date for emplacement.

PART II

SOME MINERALOGICAL OBSERVATIONS IN THE SYSTEM CoAsS - FeAsS

THE ANOMALOUS PROPERTIES OF COBALTITEConflicting properties of the Modum cobaltite

In the investigation of the cobalt ore deposits of Modum, Norway the anomalous characteristics of the main ore mineral, cobaltite (CoAsS), were noted.

The cobaltite is commonly found in well developed euhedral crystals showing a cubic morphology, combinations of the cube (100) and pyritohedron (210) being particularly common. Episcopic examination showed the mineral to have a marked anisotropy. These observations being quite contradictory two working hypotheses were formulated. These were, either the cobaltite was of a lower symmetry pseudomorphing a cubic habit, or the observed anisotropism was a false one due to minute polishing scratches or to having the nicols slightly uncrossed.

Previous work on crystal structure

Cobaltite, and the related minerals gersdorffite (NiAsS) and ullmannite (NiSbS), show close morphological similarity to pyrite and other isometric diarsenides and disulphides, they were accordingly all placed in the isometric diploidal class ($2/m\bar{3}$), (Dana, 1892). In the more recent edition of Dana (1944) these three minerals have been removed to form a 'Cobaltite Group' with chemical type AX₂Y, proposed by Mechling (1921). This group is stated to differ from the 'Pyrite Group' in that the S-S pairs of pyrite are

replaced by As-S and Sb-S pairs. This substitution in the structure makes each trigonal axis of the space lattice a polar axis with a consequent lowering of the symmetry to the tetartoidal class (23).

Most of the crystalline members of this group show only such forms as the cube, octahedron, rhombic dodecahedron and pentagonal dodecahedron which are common to both the diploidal class ($2/m\bar{3}$) and the tetartoidal class (23). However some crystals of ullmannite show tetrahedral development, and others, described by Miers (1891) are evidently twinned complexes of tetartoidal individuals. These distinct indications of tetartoidal symmetry in ullmannite would seem to verify the validity of the 'Cobaltite Group' erected from crystal-chemical considerations.

The validity of this 'Cobaltite Group' was tested by Peacock and Henry (1948) by a comparison of observed and calculated intensities for X-ray powder patterns. The pattern they obtained conformed to the expected systematic extinctions of the 'Cobaltite Group' space group ($P2_13$) with certain discrepancies. These discrepancies were largely removed when the observed pattern was compared with the calculated intensities for the pyrite space group ($Pa3$). It was therefore concluded that cobaltite had a pyrite type of structure, with equal numbers of As and S atoms statistically distributed over the eight-fold position, a similar conclusion was reached in the case of gersdorffite. Ullmannite was found to have the expected $P2_13$ structure thus being the only valid member of the supposed 'Cobaltite Group'. Bokii and Tsinober (1954) also found that cobaltite has the pyrite type of structure and belongs to space-group $Pa3$.

Thus on the basis of morphology, crystal chemistry and X-ray evidence cobaltite appears to be well established as a cubic mineral. However the anomalous optical properties of the mineral have stimulated renewed interest in the crystal structure. Takeuchi (1957) determined the crystal structure of ullmannite by means of the two dimensional Fourier method and confirmed the space group to be $P2_13$ by single crystal methods. He attempted a similar analysis on cobaltite but "the X-ray diffraction pattern of the former (cobaltite) suggested the presence of some kind of disorder in the structure prohibiting the straightforward analysis."

Onorato (1957) concluded that cobaltite has the pyrite structure with ordered substitution of As-As pairs for half of the S-S pairs in the pyrite structure. Determination of the space-group suggests that cobaltite must be considered as a polysynthetic twin of monoclinic crystals belonging to the space-group $P2_1/c$.

Wintenberger (1962) using Weissenberg techniques concluded that cobaltite is orthorhombic and belongs to the space-group $Pca2_1$. Oftedal and Saebo["] (1963) have reported on work done in the 1920's which suggests that the cubic space-group ($P2_13$) suggested by Mechling (op.cit.) may be substantially correct.

It may be concluded from these studies that two views are held on the crystal structure of cobaltite, either it is cubic with space-group $P2_13$ (Mechling, Oftedal and Saebo["]) or $Pa3$ (Peacock and Henry, Bokii and Tsinober), or it is of lower symmetry with space-group $P2_1/c$ (Onorato) or $Pca2_1$ (Wintenberger).

Previous episcopic observations.

"Schneiderhohn (1922) first reported that cobaltite did not possess cubic symmetry. His episcopic examination revealed a structure made up of intimately twinned orthorhombic individuals. He concluded, by analogy with polymorphic pyrite and marcasite, that the isometric crystals had undergone an inversion at a temperature below that at which they were formed. These observations on the anisotropy of cobaltite were confirmed by Florke (1923) and Ramdohr (1931, 1950). The inversion temperature was provisionally fixed at 850°C by Florke (1926) who heated anisotropic polished specimens to various temperatures, cooled them rapidly in air and repolished the surfaces. He found that the cobaltite started to decompose at 500°C when As_2O_3 was given off but, providing the heating was sufficiently rapid, a core remained which became isotropic when the inversion temperature was exceeded.

Ramsdell (1925) suggested that the surface of the isometric cobaltite crystals may have been affected during polishing producing a surficially twinned orthorhombic structure in the surface layers of atoms. Cobaltites from Cobalt, Ontario are reported to be isotropic under crossed nicols (Thomson, 1930, 1925; Short, 1948).

Outline of present investigation

The Modum cobaltite was studied under carefully controlled optical conditions to determine whether the anisotropy was spurious or not and to determine the optical properties of the mineral. X-ray investigations were made with the powder camera, and diffractometer to determine the crystal structure. Differential thermal analysis and heating experiments were carried out to investigate

the nature of the possible transformation. The properties of the Modum cobaltite were compared with material from Cobalt, Ontario.

LABORATORY INVESTIGATION OF THE MODUM COBALTITEX-ray determinationUse of the powder-camera

A large pyritohedral crystal of cobaltite was selected. On polishing one of the faces the characteristic anisotropy was observed. Some powder was drilled from this specimen, mounted in a powder-camera and exposed to Fe-filtered Co $K\alpha$ radiation for 6 hours. The resulting film was measured, corrected, for shrinkage and the d-spacings calculated. By comparison with the results of earlier authors and the A.S.T.M. index the substance was conclusively identified as cobaltite.

Following the standard procedure for indexing the lines obtained from a cubic substance it was found possible to assign hkl values to all the lines on the film. The unit cell parameter (a_0) of 5.575 ± 0.005 compares well with previously determined values. There are some interesting discrepancies between the values obtained for the d-spacings of the Modum cobaltite and the material from Cobalt, Ontario described by Peacock and Henry (1948). These data are shown in Table 15. Table 16 shows the expected intensities calculated by Peacock and Henry for (a) Mechling's proposed 'Cobaltite Group' ($P2_13$) and (b) the conventional 'Pyrite Group' ($Pa3$).

By inspection of Table 15 it will be obvious that the values obtained are at considerable variance. This may be due to the use of different wavelength radiations, or may be a reflection of different structures in the two samples. The cobaltite investigated by Peacock and Henry is described as "quite homogeneous and isotropic in polished specimen", a description which would not

Table 15.

Powder Diffraction data for cobaltite from Modum, Norway and Cobalt, Ontario.

COBALT MATERIAL (Peacock and Henry, 1948)			MODUM MATERIAL (This study)		
I.	d(meas)	h k l	I	d(meas)	h k l
			m	5.575	100
			wm	3.91	110
$\frac{1}{2}$	3.27	111	wm	3.19	111
6	2.82	002	s	2.875	002
			w	2.815	
10	2.53	012	vvs	2.497	012
			w	2.470	
8	2.29	112	vs	2.275	112
			w	2.263	
2	1.984	022	m	1.960	022
			wm	1.853	300, 221
			vww	1.757	310
9	1.687	113	vs	1.671	113
$\frac{1}{2}$	1.616	222	wm	1.603	222
3	1.556	023	m	1.54	023
4	1.493	123	ms	1.483	123
$\frac{1}{2}$	1.400	004	w	1.386	004
			w	1.3599	410, 322
			vww	1.3271	114, 330
			vww	1.276	133
$\frac{1}{2}$	1.250	024	wm	1.244	024
1	1.220	124	m	1.218	124
$\frac{1}{2}$	1.196	332	wm	1.190	332
$\frac{1}{2}$	1.146	224	wm	1.139	
			vw	1.138	224
			vww	1.121	005, 034
			vw	1.094	015, 134
4	1.078	511, 333	ms	1.0775	511, 333
2	1.040	025, 234	m	1.0365	025, 234
1	1.022	125	m	1.022	125
3	0.989	044	ms	0.987	044
			vvw	0.9715	522, 441
			vvw	0.9550	530, 433
			vvw	0.9415	531
$\frac{1}{2}$	0.933	006, 244	m	0.9315	006, 244
			wm	0.9298	
$\frac{1}{2}$	0.920	061	wm	0.9175	061
2	0.908	116, 235	m	0.906	116, 235
			wm	0.903	
$\frac{1}{2}$	0.853	335			
1.	0.833	063, 245			
1	0.825	136			

Table. 16.

Calculated intensities for (A) the 'Cobaltite Structure' ($P2_13$) and (B) the 'Pyrite Structure' ($Pa3$), from Peacock and Henry 1948.

hkl.	(A)	(B).
001	0	0
011	2.3	0
111	0.8	0.5
002	5.9	6.5
012,021.	12.0.	11.0.
112	7.3	7.7
022	2.8	3.1
122	0.1	0.1
013,031	0.5	0
113	8.0	8.4
222	0.6	0.6
023,032	2.7	2.8
123,213	3.3	3.2
004	0.3	0.3
014,223	0.1	0.1
114	0.2	0.1
033	0	0
133	0.3	0.2
024,042	0.7	0.6
124,214	1.6	1.4
233	0.7	0.8
224	0.9	0.9
005	0	0
034,043	0.1	0.1
015,051	0.1	0
134,314	0.2	0.1
115	2.3	2.7
333	1.2	0.5
025,052	0.4	0
234,324	1.5	1.8
125,215	1.0	1.1
044	2.4	2.6

apply to the Modum material.

In their study Peacock and Henry compared the observed intensity of their lines with uniformly reduced calculated intensities (Table 16) so as to have a direct comparison. They concluded that their data indicated the space-group $Pa\bar{3}$.

It will be noted that the Modum cobaltite cannot be assigned to this space-group. Good reflections for 100, 110, 310, 114 and 033 were obtained, these are forbidden by space-group $Pa\bar{3}$. Similarly the presence of the 100 reflection excludes the possibility of space-group $P2_1\bar{3}$.

Diffractometer

Cavity mounts of cobaltite from Modum and Cobalt were prepared. The diffractometer traces were made using Cu $K\alpha$ radiation. Peaks were obtained for all the lines previously measured on the powder film of the Modum material, the Cobalt material confirmed the results of Peacock and Henry (see above). It is concluded that there is a fundamental structural difference in the cobaltite from the two localities.

Optical properties

The Modum cobaltite has the following optical properties, all data being obtained using the methods described in Cameron (1961).

Colour The colour is silvery-white with slight tints of pink, brown or violet. When in contact with arsenopyrite it is noticeably pinker, against pyrite it is a very bright white. Under oil immersion the pinkish tint against arsenopyrite is accentuated.

Bireflectance Under oil immersion the mineral shows slight

reflection pleochroism from white to pinkish white, this is particularly noticeable at boundaries with arsenopyrite and along twin composition planes. Quantitative determinations of the apparent uniradial reflectivities of randomly oriented grains of cobaltite were made in monochromatic light (559m) following the method described by Cameron (1963). The following values for R_1 and R_2 were obtained.

Table 17

Apparent uniradial reflectivities of cobaltite.

R_1 (%)	48.5	51.6	49.2	52.4	48.6	50.1	51.3	49.6	48.7	52.1	50.8
R_2 (%)	52.8	54.5	53.2	54.3	52.9	53.5	53.3	53.9	52.9	54.1	53.4

These data suggest that neither R_1 nor R_2 has a constant value, indicating biaxial rather than uniaxial symmetry. It will be noted that all values of R_1 are equal to or less than 52.4%, while all values of R_2 are equal to or greater than 52.8%. By analogy with the refractive indices of transparent minerals this intermediate value of the reflectivity is referred to as R_m . Similarly R_p , the minimum value, is equal to or less than 48.5%, and R_g , the maximum value, is equal to or greater than 54.5%. As R_m is closer in value to R_g than to R_p the mineral is considered to be biaxial negative at $\lambda = 549_{\text{m}\mu}$.

Reflectivity The mean uniradial reflectivity was measured in white light in air, using Elba pyrite ($R=54.5\%$) as a standard. The value was 53.1%, with a range from 52.5% to 53.6%. In oil immersion the corresponding values were: mean- 44.5%, range- 44.1-45.3%. The spectral reflectivity was measured in air, on a face polished parallel to the (100) direction, using a constant interference

filter. The resulting data are shown in Table 18.

Table 18.

Spectral reflectivity of cobaltite.

Wavelength ($m\mu$)	400	450	475	500	525	550	575	600	650	700	740
Reflectivity (%)	42.5	47.6	48.1	50.8	52.8	56.2	57.8	58.15	7.6	54.9	52.5

The above values were measured with a bandwidth of $\pm 5 m\mu$, the individual reflectivity values are the mean of four readings.

Anisotropism In observing the anisotropism great care was taken in ensuring that the microscope was correctly adjusted and that the specimen was accurately levelled. The position of the analyser was found by using a Nakamura plate and an anisotropic section of ilmenite, and checked by observing the polarisation figure of galena.

Under these conditions it was confirmed that the mineral is indeed anisotropic. On rotation of the stage pale blues and browns are observed, these are of varying intensity in sections of different orientation. Four positions of even illumination are observed (the 'extinction' positions) alternating with four positions of maximum illumination (the '45°' positions). When the stage is rotated to place the mineral in a 45° position the strongest brown or blue hues are observed, on rotation of the analyser in a clockwise direction the polarisation colour becomes slightly more intense before fading out completely, on rotating the analyser in an anti-clockwise direction the mineral changes colour (i.e. from blue to brown or vice versa) before fading out.

Polarisation figures were obtained for many grains of diverse orientation, on rotation of the stage the isogyres showed only a slight separation. The colour fringes, due to dispersion, were observed. With the mineral in the extinction position the dispersion of the reflection rotation (DRr) was observed by rotating the analyser, the colour fringes were distinct and indicated that DRr is violet > red. With the analyser and polariser exactly crossed the dispersion of the apparent angle of rotation (DAr) was observed by rotating the mineral on the stage, the colour fringes were very indistinct but appeared opposite in sense to DRr. According to Cameron (1961, p. 119) this indicates that both DRr and DAr are v > r.

A qualitative determination of the sign of the phase difference $\Delta_{x,y}$ was made using the gypsum plate (Cameron, *ibid* p.139). This indicated that $\Delta_{x,y}$ was negative.

Hardness The mineral is very hard and difficult to polish. The Becke line test indicates that it is harder than arsenopyrite and skutterudite but softer than pyrite. Measurement of the Vickers hardness was made using a G.K.N. hardness tester with a 100 gm. load, the observed values ranged from 957 to 1,180 with a mean of 1,112. The quality of the indentation was poor.

Form and twinning. The crystals were frequently euhedral the forms $\{100\}$ and $\{210\}$ being particularly common. The individual euhedra consisted of an intimately twinned intergrowth of individual crystals. The twin lamellae were frequently sub-parallel to the adjacent grain boundary. The twins were indistinctly visible as a result of differences in reflection pleochroism but

were clearly seen under crossed nicols (Figure 136).

Electrolytic etching Single crystals of cobaltite were etched in dilute H_2SO_4 solution following the electrolytic method described by Suzuki (1962). The following alterations to his method were made: a carbon rod was used as the cathode instead of a carbon plate as the anode, the reason for the necessary reversal of current was not investigated, it was not found necessary to solder the connecting wire on to a previously prepared Cu electro-plated surface of the specimen as described by Suzuki, adequate contact was maintained by pressing the specimen into plasticene covered by silver foil, the connecting wire was attached to this silver foil.

The etched surfaces produced were useful in studying the internal form of the crystals, for irregularly shaped lamellae were observed. By polishing previously etched specimens it was confirmed that these lamellae represented twinned individuals within the larger crystal.

Discussion

The only conclusion that can be drawn from this study is that, despite the external crystal morphology, the Modum cobaltite is not now in a cubic form.

The X-ray determinations have eliminated the only two likely cubic space-groups ($Pa\bar{3}$ and $P2_1\bar{3}$). The optical properties indicate that the mineral has either orthorhombic, monoclinic or triclinic symmetry. The form of the twins suggests that these may represent transformation twins (see p. 338) which have developed on the inversion of a cubic form to a polymorph of lower symmetry.

Considered in terms of crystal structure the most likely reason for this inversion is the ordering of the arsenic and sulphur atoms in the lattice. Mr. R. Phillips has determined four possible ways in which this ordering could take place if the original structure was of the pyrite type. The structures which result are shown, as projections on 100, 010 and 001, in Figure 162. It will be noticed that only one of these possibilities retains cubic symmetry, this is the structure proposed by Mechling (1921) for the 'Cobaltite Group' and shown by Takeuchi (1957) to be the structure possessed by ullmannite. This arrangement can be eliminated from further consideration as the diffraction data contains reflections forbidden by this space-group (p.322).

Preliminary calculations of expected d-spacings and intensities for these various arrangements by Mr. Phillips indicate that the orthorhombic ($Pca2_1$) structure compares favourably with the observed X-ray pattern. Onorato (1957) concluded from the use of Weissenberg techniques that there were five possible space-groups consistent with his results, these included $Pca2_1$ and $P2_1/c$, of which he considered the latter to be the most likely. Le Damany and Wintenberger (1962) have shown that on Weissenberg diagrams reflections are limited by the following conditions:

$$h \ 0 \ l \ - \ h \ \text{even}$$

$$0 \ k \ l \ - \ l \ \text{even}$$

This observation leads to the consideration of only two possible space-groups - $Pca2_1$ and $Pbcm$. The latter space-group does not permit a structure similar to that of pyrite and they conclude that $Pca2_1$ is the most likely space-group.

	100	010	001	SPACE GROUP	
I				Pca21	ORTHORHOMBIC
II				P2₁/c	MONOCLINIC
III				P2₁3	CUBIC
IV				P₁	TRICLINIC

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Fig. 162

Very slight splitting of high angle reflections has been observed on some of the films and diffractometer traces during the present study. Similar splitting has been observed by Saebo["] (personal communication) and Wintenberger (1962). The latter author estimates the difference in cell parameters as being of the order of 2×10^{-3} .

Thus it is concluded that the Modum cobaltite is in the form of a complexly twinned intergrowth of orthorhombic individuals. Despite the ordering of the As and S atoms the distortion of the lattice is slight and the cell parameters vary insignificantly from the cubic form. The Modum cobaltite probably originated, in its present form, from the inversion of a disordered (cubic) structure.

THE CUBIC-ORTHORHOMBIC TRANSFORMATION OF COBALTITE

Canadian cobaltite

The results of the investigation of the Modum cobaltite suggest that the orthorhombic structure has resulted from the ordering of a cubic form.

The descriptions of the cobaltite from Cobalt, Ontario in the literature are compatible with this material having the disordered, cubic, structure. Some crystals of cobaltite from the Columbus Mine, Cobalt, Ontario were kindly made available by the Royal Ontario Museum, Toronto for a comparison with the Modum material.

Polished specimens of these crystals were prepared. The cobaltite had the same reflectivity (mean uniradial reflectivity: 52.9%) and hardness (V_{100} : 1,090) values as the Modum material. However no reflection pleochroism was observed. With crossed nicols the mineral was evenly illuminated during a 360° rotation of the stage, the polarisation figure was a black cross showing no separation of the isogyres on rotation of the stage.

Diffraction traces, using Cu $K\alpha$ radiation gave a pattern differing only in intensity values from that reported by Peacock and Henry (op.cit.). In particular the 100, 110, 310, 114 and 033 reflections were absent.

Electrolytic etching, under the same conditions as those employed in studying the Modum cobaltite, revealed a zonal structure very similar to that observed in the hydrothermal pyrite from the late veins at Modum.

These observations strongly suggest that the Canadian cobaltite has a disordered structure, with a statistical distribution of As and S atoms in the lattice, producing cubic symmetry with space-group $Pa\bar{3}$.

Heating experiments

If these conclusions regarding the crystal structures of the Modum and Canadian cobaltites are correct, it should prove feasible to convert the ordered to the disordered form by heating.

Preliminary attempts at achieving this transformation were hampered by the strong oxidation of the cobaltite at temperatures above 450°C . Florke (1926) overcame this difficulty by sufficiently rapid heating of the specimen such that an unoxidised core remained which was suitable for polishing. Heating rates were not sufficiently rapid during this investigation and the complete crystal was oxidised.

Differential thermal analysis was attempted to elucidate the nature and temperature of the possible transformation. The powdered sample rapidly oxidised above 400°C . Attempts were made to carry out the analysis in a nitrogen atmosphere but a sufficiently good seal could not be achieved with the apparatus used.

Through the courtesy of Dr. Hayhurst, a sample of the Modum cobaltite was heated in the research laboratory of Messrs. Pickford, Holland and Co. Ltd. The source of the nitrogen was ordinary cylinder nitrogen passed over copper foil in a small furnace to remove the oxygen present. Initially the specimen reacted with the supporting platinum wire, this difficulty was overcome by supporting it on an alumina rod. A successful run was achieved

when the specimen was heated slowly, over a period of 4 hours, to 900°C. After 25 minutes there were indications of fumes coming from the furnace so the sample was removed quickly and quenched in air. The sample showed slight signs of cracking and oxidation.

The sample concerned was one half of a single crystal of cobaltite from the Geologisk-Museum, Oslo, labelled "Kobaltglanz, Skutterudgruben." The crystal measured 1 cm. in width and showed equal development of (100) and 210) faces. Fragments of each half of the crystal, heated and unheated, were used for preparing smear mounts for the diffractometer; the remaining parts were mounted and polished.

The polished specimen of the unheated half showed the typical development of the anisotropism and lamellar twinning described above, while the heated half, within a 0.5 mm. oxidised zone, was completely homogeneous and isotropic.

The diffractometer trace of the unheated portion of the crystal gave the normal pattern for Modum cobaltite. The heated portion gave a pattern similar in all respects to that obtained from the Canadian cobaltite. Figure 163 shows part of the diffractometer traces and illustrates this difference in pattern obtained from one and the same crystal under identical diffraction conditions.

It is concluded that the previous deductions are valid, i.e. the Canadian and Modum cobaltites are polymorphic forms having a cubic (disordered) and orthorhombic (ordered) structure respectively.

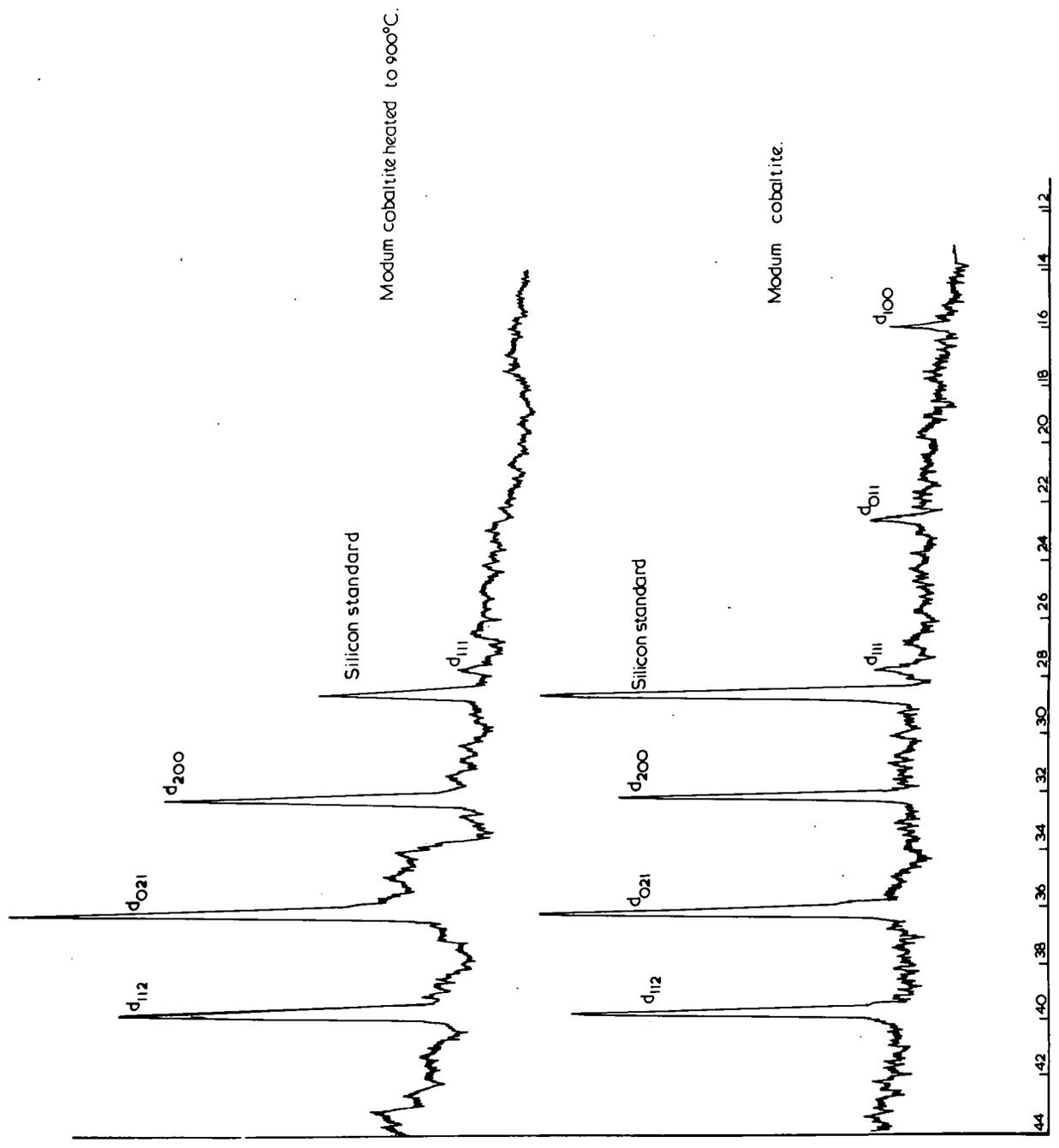
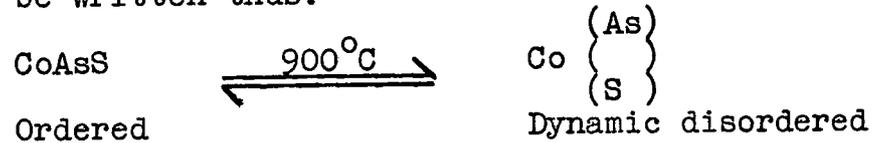


Fig. 163

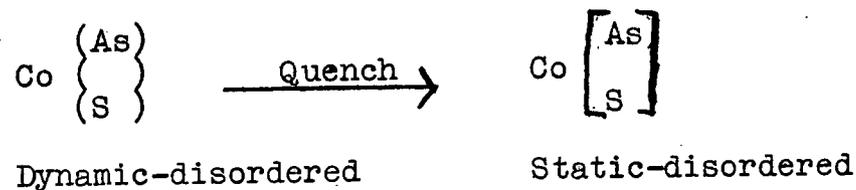
DEGREES $2\theta(\text{Cu K}\alpha)$

Discussion

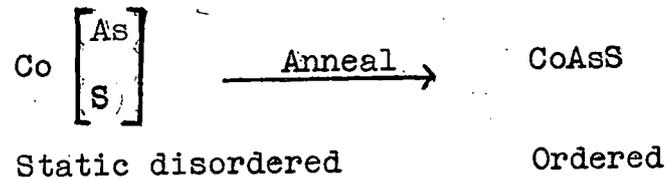
Following Buerger (1945) the heating experiment described above can be written thus:



If this disordered form is cooled rapidly it will retain its structure:



Annealing at elevated temperatures less than the critical point (i.e. as in metamorphism) can cause the following reaction:



The types of polymorphic transformation have been reviewed by Buerger (1951). The cubic-orthorhombic transformation of cobaltite is probably an example of substitutional transformation of disorder.

The transformation can be considered in terms of a flow of As and S atoms among fixed Co atoms. From the point of view of symmetry the random flow of the As and S atoms is equivalent to causing the two species to become statistically equivalent. For this reason the structure statistically assumes the symmetry it would have if all the As and S atoms were of one species. It follows that this could be the equivalent structure to pure CoS_2 (cattierite) which has the pyrite-type (Pa3) structure. Thus the dynamic-disordered $\text{Co} \begin{Bmatrix} \text{As} \\ \text{S} \end{Bmatrix}$ and static-disordered $\text{Co} \begin{Bmatrix} \text{As} \\ \text{S} \end{Bmatrix}$ forms of cobaltite

have statistically assumed the pyrite type symmetry.

In the ordered structure the As and S atoms alternate in a regular way (Figure 162). This reduces the free energy of the structure as it reduces repulsive interactions between identical neighbouring atoms.

The disordered state can be visualised in terms of a distorted network of atoms oscillating between two extreme structures (Figure 164b and d). The effect of this vibration is that the structure statistically assumes the intermediate form or 'open' structure (Figure 164 a). The ordered state can be visualised in terms of one of the extreme 'collapsed' structures in which the constituent atoms of the structure are displaced such that no inter-atomic bonds are broken but the non-contacting atoms are displaced slightly with respect to each other.

A substitutional transformation of this type involves the interchange of As and S atoms in the structure producing an intermediate state in which these atoms will have a lower than normal co-ordination. Therefore the transformation is relatively sluggish, particularly in the down-temperature direction (Buerger, 1951).

Buerger (op.cit.) has shown that, in general, the open form is a form of high residual energy and the collapsed form has a lower residual energy. Thus, in the neighbourhood of absolute zero, the collapsed form is the more stable of the two. On thermal agitation the As and S atoms in the lattice acquire energy, a temperature is reached when enough energy is acquired to overcome the non-nearest neighbour attractions and to swing past the neutral point

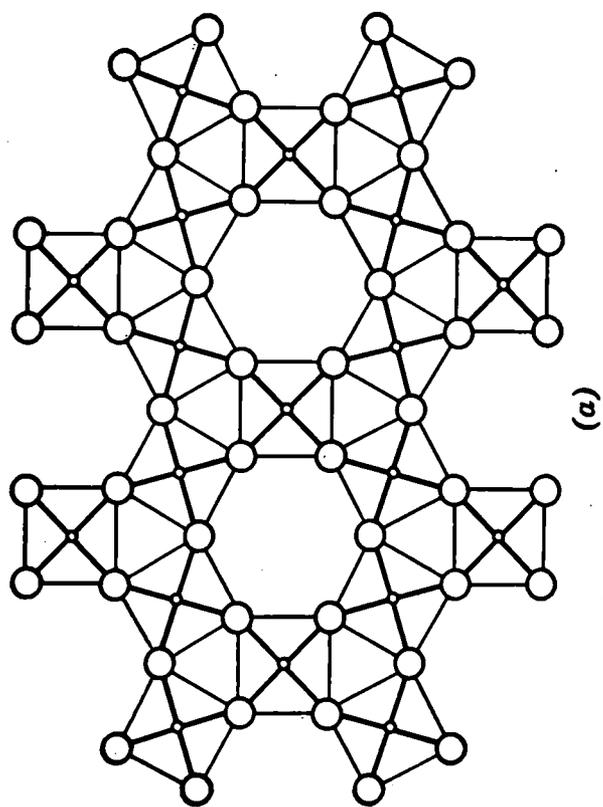
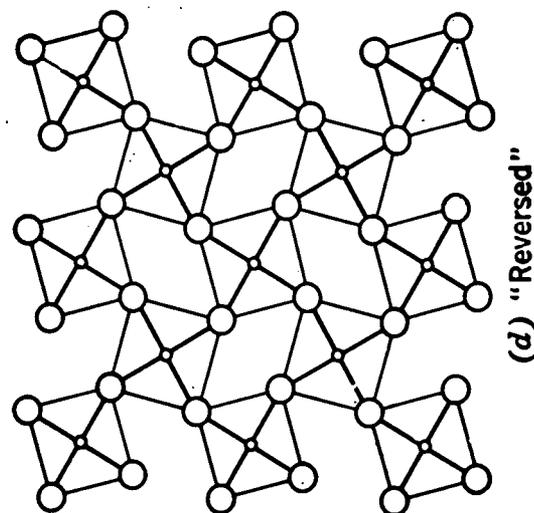
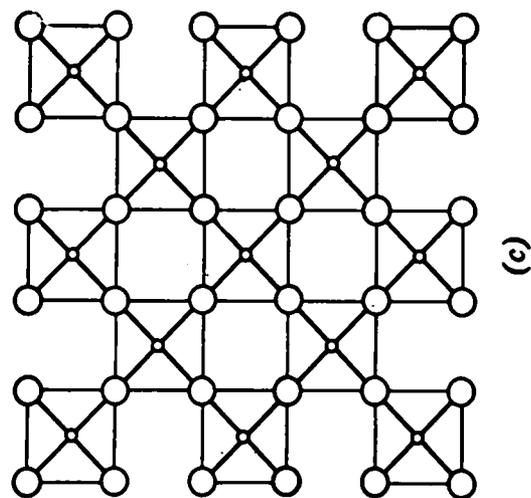
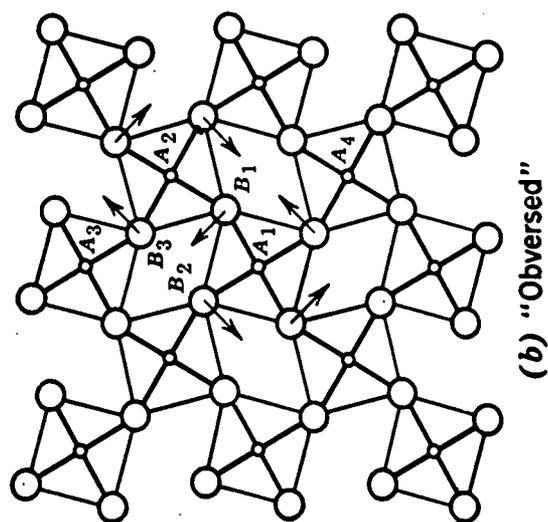


Fig. 164



into the reversed configuration (i.e. from Figure 164b to Figure 164d). When the position is reached that the As and S atoms freely between the two extreme conditions they can be statistically regarded as having assumed the 'open' form - $\text{Co} \begin{Bmatrix} \text{As} \\ \text{S} \end{Bmatrix}$. Thus the open form can be attained by intense thermal agitation of the atoms in the collapsed form and may be regarded as the highly disordered equivalent of the collapsed form. It is also the most probably form, since it can result from innumerable different individual mixtures of obversed and reversed forms, plus connecting regions temporarily in the open form. If this structure is retained by rapid quenching it produces the form $\text{Co} \begin{bmatrix} \text{As} \\ \text{S} \end{bmatrix}$.

Thus in transformations of this type the high temperature form is usually the open form and has the higher symmetry. The reason being that the high temperature form is an average of the obversed and reversed collapsed forms, which is a position of higher symmetry.

The symmetry of the low temperature form is a derivative structure of the basic structure possessed by the open, high temperature, form. The symmetry content per unit volume of the low form is a submultiple of the high form (Buerger, 1947).

Thus the orthorhombic form of cobaltite has a symmetry which is related to the symmetry of the cubic form. In the high-low inversion the general structure remains the same but some of its symmetry operations become suppressed, such that the orthorhombic form has more than one orientation with respect to the cubic form.

With slow cooling, or annealing, the high-low transformation takes place and is propagated as a wave as rapidly as heat can be transmitted (Buerger, 1945). Unless the crystal is extremely small, transformation nuclei appear in various parts of the crystal spontaneously, particularly at the edge, and the transformation spreads from these centres. These nuclei can have different orientations with respect to each other, the number of orientations depending on the difference in symmetry between the high and low forms. These nuclei grow until they make contact with one another, from the mechanism of formation of these crystals their contacts are often irregular. They are in twinned orientation with respect to each other and could be brought into coincidence by the operation of one of the symmetry operations of the cubic form which vanished in the transformation to the orthorhombic form. This is considered to be the fundamental cause of the twins frequently seen in the polished specimens of the Modum cobaltite.

It is here suggested that cobaltites of hydrothermal origin, which show a zonal structure indicative of intermittent deposition, possess a static-disordered ('open') form in which As and S atoms in the lattice become statistically equivalent leading to a structure with cubic ($Pa\bar{3}$) symmetry.

Similarly it is here suggested that cobaltites from high-grade metamorphic terrains have been effectively annealed at temperatures less than the critical point (ca. 850°C) such that the high-low substitutional transformation to the orthorhombic ($Pca2_1$) structure has taken place. In large, originally cubic, crystals the nuclei of the transformation have appeared at different loci

and spread to form the observed transformation twins.

Thus the crystal structure, and type of twinning present, in cobaltite is useful in elucidating the geological history of a deposit. The ordered crystal structure and transformation twinning of the Modum cobaltite strongly suggest that the mineral has been affected by prolonged heating at a temperature less than the critical point. The necessary conditions would be fulfilled by metamorphism at the upper amphibolite facies of regional metamorphism. These relations are considered to be important evidence that the cobalt mineralisation is of pre-metamorphic origin.

ARSENOPYRITE

Crystal structure

The unit cell of arsenopyrite was first investigated by de Jong (1926), using powder and rotating crystal methods he assigned arsenopyrite to the orthorhombic group which includes marcasite. In a later study (de Jong, 1928) on cobaltian arsenopyrite from Sulitjelma, Norway, he assigned the following dimensions to the orthorhombic cell axes: $a=6.45$; $b=9.54$; $c=5.53$ Å. The space-group, determined from indexing 17 lines of a powder photograph, is given as $Pmmm$ or $Pm\bar{m}n$.

Buerger (1936) showed that no orthorhombic space-group can satisfy the intensity data obtained by the Weissenberg method. He showed that the arsenopyrite structure can be considered as a derivative superstructure based on the marcasite type, and concluded that, ideally, it has monoclinic symmetry with space-group $B2_1/d$. He found that the mineral is also commonly triclinic (space-group $B\bar{1}$) and concluded that this was the result of variations from the ideal chemical formula.

Huggins (1937) criticised Buerger's conclusions on the grounds that the proposed structure led to interatomic differences not in harmony with those calculated from electron-pair bond radius sums. Buerger (1937) confirmed the abnormal interatomic distances of marcasite by refining the structure. A similar refinement of the arsenopyrite structure was hampered by the twinned nature of the naturally occurring crystals. On the basis of the crystal structure of gudmundite ($FeSbS$) he suggested qualitative corrections to his published arsenopyrite parameters and confirmed that the interatomic

distances of the arsenopyrite group minerals are quite different to those found in minerals with the pyrite-type structure (Buerger, 1939).

Morimoto and Clark (1961) have shown that arsenopyrite, although appearing to be of orthorhombic symmetry, is composed of twinned monoclinic individuals. The following transformations are employed: de Jong (orthorhombic) cell - Buerger (double monoclinic, metrically orthorhombic) cell - $001/100/010$; Buerger cell - Morimoto and Clark (primitive monoclinic) cell - $\frac{1}{2}0\frac{1}{2}/0\bar{1}0/\frac{1}{2}0\frac{1}{2}$. In all subsequent discussions the latter monoclinic orientation will be retained. The twin planes of the monoclinic individuals (Type I twins) are $(10\bar{1})$ or (101) with twin axis $[10\bar{1}]$ or $[101]$.

Morimoto and Clark (op.cit.) have shown, on the basis of oscillation, precession and Weissenberg methods, that splitting of some reflections occurs which can only be accounted for by a second type of twinning (Type II twins). This requires the assumption that arsenopyrite is triclinic; the postulated twin law is twin axis $[010]$ with twin plane $(010)^*$. The deviation of α^* and γ^* from 90° are less than 30° .

Since $(10\bar{1})$ reflections were obtained for all samples studied by Morimoto and Clark none had truly monoclinic symmetry. They consider that the replacement of S by As causes a gradual change in the properties of arsenopyrite which raises the symmetry of the structure from triclinic to monoclinic. They consider that this change might also be achieved by the incorporation of Co and Sb in the structure rather than calling on further arsenic substitution.

These researches can be summarised thus:

- (1) Arsenopyrite is commonly triclinic but has a pseudomonoclinic symmetry resulting from the Type II twinning;
- (2) Arsenopyrite of suitable composition may be truly monoclinic;
- (3) The monoclinic arsenopyrite has a pseudo-orthorhombic symmetry resulting from the Type I twinning.

It is interesting to speculate on the origin of the Type I twins. These have the appearance in polished specimen of being transformation twins of the type observed in cobaltite. An analogy may exist between the cubic (pyrite-type) structure of disordered cobaltite and the pseudo-cubic (twinned orthorhombic) structure of ordered cobaltite and an orthorhombic (marcasite-type) structure, so far undescribed, and the pseudo-orthorhombic (twinned monoclinic) structure of the ordered arsenopyrite described by Morimoto and Clark.

Morimoto and Clark (op.cit. p. 1,462) investigated this possibility by heating Freiberg arsenopyrite at 600°C for 30 days in a sealed, evacuated, silica-glass tube followed by rapid cooling in cold water. No difference in Weissenberg or precession photographs was observed between the heated and unheated specimens. However these conditions would not have been sufficient to achieve the disordering of orthorhombic cobaltite (see p. 334). The possibility still exists that an orthorhombic-monoclinic polymorphic inversion may be possible in arsenopyrite.

Relationship of the arsenopyrite and cobaltite structural types

It has been shown that the structure of cobaltite (CoAsS) can be derived from the pyrite structure, similarly the arsenopyrite (FeAsS) structure is closely related to the marcasite structure.

For a better understanding of the relationships of these structures it is useful to consider the polymorphic transformation of pyrite to marcasite. The unit cell of pyrite (Pa3, a_0 : 5.417, Z:4) does not bear any obvious relationship to the marcasite unit cell (orthorhombic, Pnmm, a:4.436, b:5.414, c:3.381, Z:2). However intergrowths of the two minerals are common which have marcasite (101) parallel to pyrite (001) suggesting that a similar structural environment of Fe and S is present in these respective planes. Huggins (1922) postulated that this was achieved if the arrangements of Fe atoms about S-S pairs in the pyrite structure were rearranged to give an orthorhombic lattice. Buerger (1931) showed more rigorously that this was in fact the marcasite structure. He showed that the relationship of marcasite to other structures could best be visualised by detaching a sulphur pair, with its immediate iron environment, which consists of a capping at each end by iron triads rotated 180° with respect to each other. These same units, in the undistorted, ideally symmetrical condition, are distinguishable in pyrite. The difference in the structures is essentially one of a different linking of these units. Buerger further compared certain geometrical properties of pyrite with those of the marcasite structure (1931, Table 10) to bring out the essential identity of this common unit. The structure of marcasite projected on (101) is similar to that of pyrite on (001) probably accounting for the observed mineral intergrowths.

These dimorphs can be considered as being related by a reconstructive transformation (Buerger, 1951), whereby the entire structure is unlinked, the relinked again to form a new and different network, yet one having first co-ordination identical with that of

the first structure. Thus the S-S pairs are relinked during the transformation pyrite-marcasite in the manner, diagrammatically shown, in Figure 164a, c.

The density of pyrite is 4.95-5.10, that of marcasite 4.85-4.90 so that pyrite is probably the high pressure polymorph. This hypothesis agrees well with the geological observation that marcasite is, in general, formed only under near surface conditions.

However, as no examples have yet been described of either cobaltite assuming a marcasite type structure or arsenopyrite assuming a pyrite type structure, it follows that there is some fundamental chemical difference which makes both structures stable under identical geological conditions.

The similarities of the two structures may be reviewed. In the arsenopyrite lattice each iron atom has six neighbours at the corner of a somewhat distorted octahedron. One face of the octahedron is a triangle of 3 arsenic atoms while the opposite face is a triangle of 3 sulphur atoms. This same arrangement is found, in an ideal and undistorted form, in the ordered, orthorhombic cobaltite structure (Figure 162). In the arsenopyrite structure each sulphur atom is surrounded by 3 iron and one arsenic atom at the corners of a somewhat distorted tetrahedron. Each arsenic atom is similarly surrounded by 3 iron and one sulphur atom. These arrangements are analogous to those found in the cobaltite structure. A similar relationship between the As-S pairs with their immediate metal atom environment is found between the arsenopyrite and cobaltite structures as that found for the S-S pairs in the marcasite and pyrite structures.

Buerger (1936) postulated that the arsenopyrite structure becomes unstable for large metal atoms due to a close approach near their shared octahedron co-ordination edge. He developed this idea in a later paper (Buerger, 1939) in which he reviewed the limits of the arsenopyrite structural type.

Members of this group have the general formula $AB'B''$. If the relationship of atomic diameters is such that packing of the arsenopyrite type would give rise to a very close approach between A atoms, the arsenopyrite packing type cannot be assumed by the compound and the cobaltite packing type must be assumed instead.

A close approach between A atoms is favoured by:

- (1) Increase in radius of A atoms relative to the radius of B' and B'' atoms;
- (2) Increase in the radius ratio $R_{B'}/R_{B''}$.

Thus it would seem that it is the difference in atomic radii of cobalt and iron which is responsible for stabilising the cobaltite and arsenopyrite structures respectively.

COBALTIFEROUS ARSENOPYRITE

Nomenclature

Analyses of arsenopyrite frequently reveal slight amounts of cobalt. This cobalt is presumably replacing some of the iron in the lattice.

Arsenopyrites containing between 3 and 9 weight percent cobalt are commonly reported. These have been termed either cobaltian-arsenopyrite or danaite. Danaite has been shown to have the same crystal structure as common arsenopyrite (de Jong, 1926).

Arsenopyrite containing 12-17 weight percent cobalt or more have been termed glaucodot (Lewis, 1877). The crystal structure of glaucodot has been investigated by Ferguson (1947). He finds no significant departure from orthorhombic symmetry using rotating crystal and Weissenberg methods. The powder pattern is not indexable using de Jong's (op.cit.) cell dimensions, and Ferguson regards the glaucodot structure as being a superlattice based on the arsenopyrite structure. The cell dimensions in Table 19 are taken from Ferguson's paper and illustrate that the cell of glaucodot has the appearance of a superstructure with sides a , $3b$, c , where a , b , and c are the sides of the rectangular cell of arsenopyrite in the classical orthorhombic setting used by de Jong.

Table 19
Cell dimensions of glaucodot and arsenopyrite

a - 6.63	b - 28.33	c - 5.63 kX	-GLAUCODOT (Ferguson, 1947)
a - 6.67	6b - 28.86	c - 5.73 A	-GLAUCODOT (de Jong, 1926)
a - 6.44	2b - 9.52	c - 5.63 A	-ARSENOPYRITE (de Jong, 1926)
a - 6.42	b - 9.55	c - 5.71 A	-ARSENOPYRITE (Buerger, 1936)
a - 6.42	b - 9.51	c - 5.65 A	

Shishkin (1961) has described 'high-cobaltian glaucodot', in association with ordinary glaucodot, from disseminated ore lenses in Cambrian eruptive rocks from Russia. The high cobalt material is reported to contain 25 weight percent cobalt.

Alloclasite, with 32 weight percent cobalt, has been described from Oraviczabanya, Hungary. Ferguson (op.cit. p. 43) is suspicious of the reported analysis as it deviates distinctly from the ideal cell content of $12 \text{ CoFeAs}_2\text{S}_2$ which is closely approached by material from other localities. Dana and Ford (1958) regard 'alloclasite' as being a mixture of glaucodot with "some other mineral."

Danaite from Modum

The arsenopyrite was worked as a cobalt ore mineral from many of the Modum mines. Specimens of the mineral, from the collections of the Mineralogisk-Museum, Oslo, were selected for a study of the range of composition represented.

Determination of the cobalt content

Preparation of standards Mixtures of iron and cobalt metal powders were prepared in varying proportions, and analysed on a Norelco X-ray fluorescence spectrograph. When absolute peak heights for the two elements (Figure 165a) are plotted for different concentrations, with the same size sample, straight lines are obtained indicating that interference is negligible. Admixture of 20%, 40% and 80% arsenic to the $\text{Fe}_{95}\text{Co}_5$ sample indicated that no interference was obtained from arsenic. Approximately 50% sulphur was added to the standard containing an 80% arsenic admixture, the powder was fused, reground and measured again, the resulting peak height ratio for Co and Fe showed no variation. Thus the ratio of peak heights ($F_{\text{Co}}' : F_{\text{Fe}}$) can be used

Table 20.

Peak heights for standard iron-cobalt mixtures.

Standard	Absolute Peak Height		F _{Co} :F _{Fe} ratio.
	F _{Co} .	F _{Fe} .	
Fe 100 Co 0	0	66	0
Fe 95 Co 5	23.5	64.5	0.364
Fe 90 Co 10	51.0	59.5	0.857
Fe 80 Co 20	98.0	53.0	1.849
Fe 70 Co 30	84.0	26.5	3.920
Fe 50 Co 50	96.5	11.5	8.391

Table 21.

Determination of iron: cobalt ratio in the Modum danaite.

Sample Number.	F _{Co}	F _{Fe}	F _{Co} :F _{Fe} ratio.	Fe: Co ratio.
Arsenopyrite (Bleikvassli)	0	20	0	100:0
1034-1	42	31	1.355	83.6:16.4
1034-2	62	41	1.561	82.0:18.0
1034-3	25	23	1.087	86.7:13.3
1034-4	28	13	2.153	78.1:21.9
1034-5	21	9	2.333	77.1:22.9
1034-6	60	15	4.000	67.4:32.6
1034-7	9	7	1.285	80.5:19.5
1034-8	8.3	5	1.660	81.2:19.8
1034-9	37	30	1.230	86.1:13.9
1034-10	79	17	4.614	66.4:33.6
1034-11	16	14	1.121	86.1:13.9
1034-12	26	16	1.625	81.6:18.4
1034-13	21	16	1.287	84.5:15.5
1034-14	7	8	0.840	87.0:13.0
1034-15	48	43	1.118	86.0:14.0
1034-16	29	26	1.100	86.4:13.6
1034-17	71.6	29	2.467	76.3:23.7
1034-18	71	53	1.353	85.8:14.2
1034-19	24	37	0.654	91.8: 8.2
1034-20	46.5	49	0.949	88.2:11.8
1034-21	56	47	1.200	85.2:14.8
'Glaucodot' (Håkonsboda)	24	38.5	0.628	92.2: 7.8

Table 22.

Modum danaite analyses recalculated in terms of atomic percent cobalt.

| Sample No. at%Co |
|------------------|------------------|------------------|------------------|------------------|
| Bleik.Asphy. 0 | 1034-5 7.8 | 1034-10 11.4 | 1034-15 4.8 | 1034-20 4.0 |
| 1034-1 5.6 | 1034-6 11.1 | 1034-11 4.7 | 1034-16 4.6 | 1034-21 5.0 |
| 1034-2 6.1 | 1034-7 6.6 | 1034-12 6.2 | 1034-17 8.1 | Håkonsboda 2.6 |
| 1034-3 4.5 | 1034-8 6.4 | 1034-13 5.3 | 1034-18 4.8 | |
| 1034-4 7.4 | 1034-9 4.7 | 1034-14 4.4 | 1034-19 2.8 | |

15 4 2

Figure 165: a. Peak heights obtained from diffractometer traces of standard iron-cobalt mixtures.
b. Peak height ratio as a function of cobalt content.
c. Frequency distribution of cobalt values obtained from the Modum danaite.

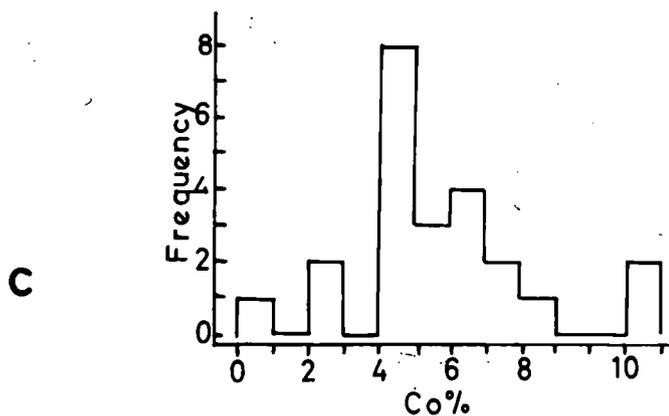
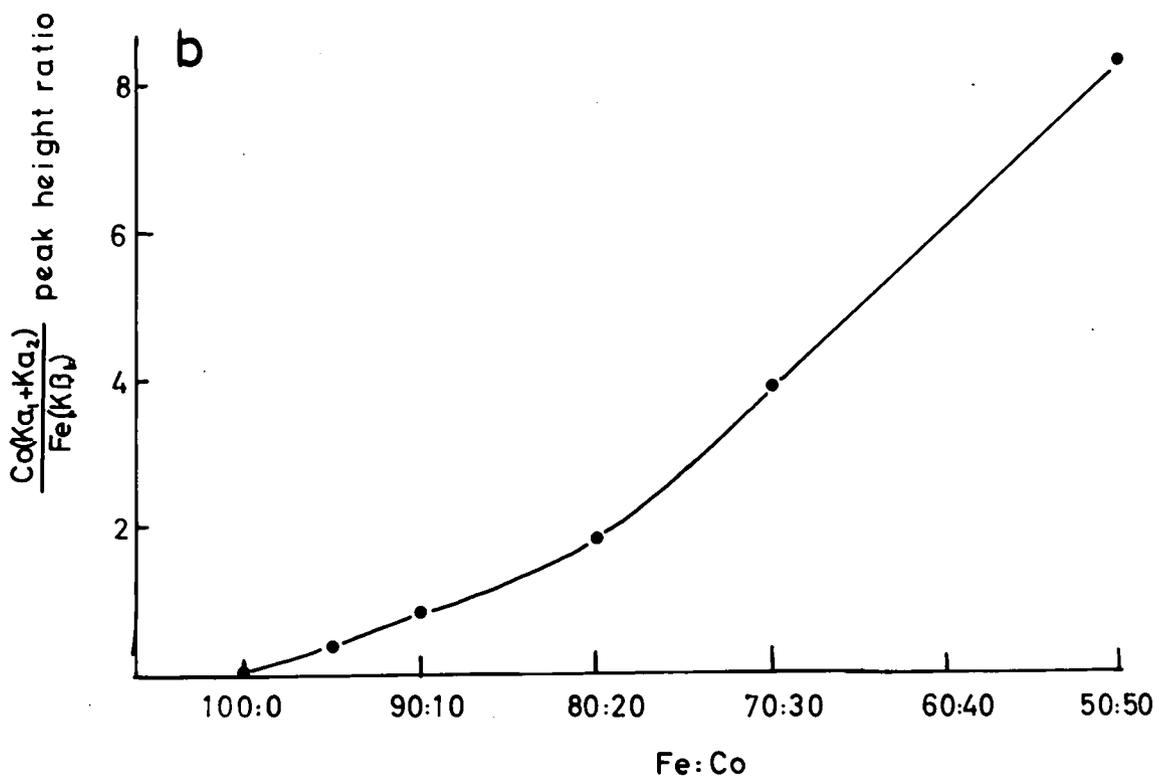
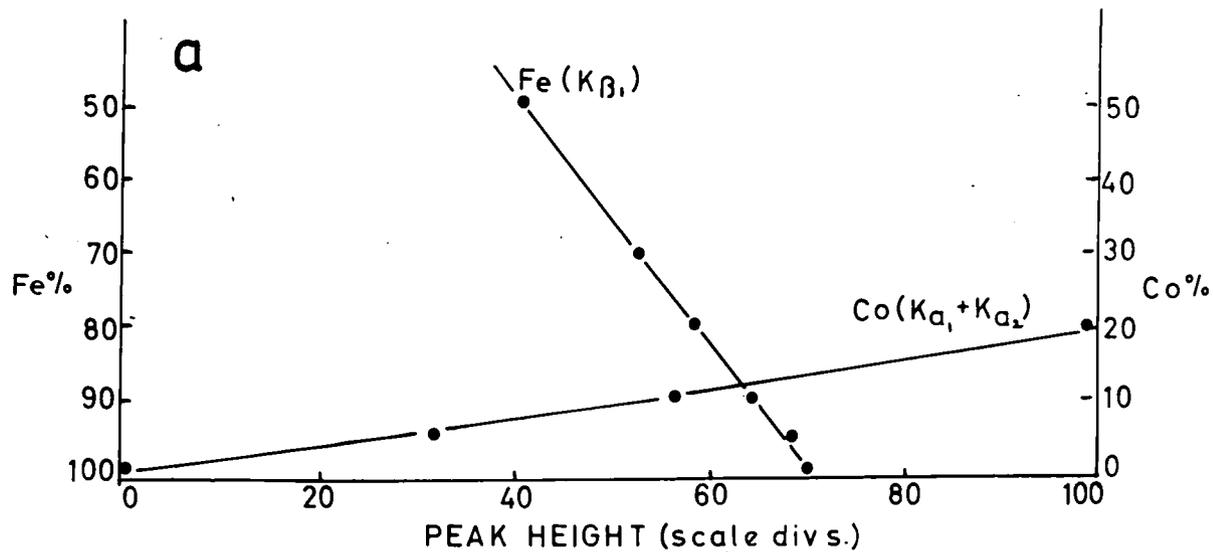


Fig. 165

to determine the Fe:Co ratio in an unknown sample. The determinative curve is shown in Figure 165b, and the data upon which it is based is shown in Table 20.

Results 21 crystals of cobaltian arsenopyrite from the Skuterud Mines were analysed by this method. A similar specimen from the Håkansboda cobalt mines was included and the results obtained were compared with a cobalt free arsenopyrite from the Bleikvassli Mine, North Norway, kindly supplied by Dr. F.M.Vokes. The results are shown in Table 21.

Discussion Taking the formula of arsenopyrite as FeAsS with an ideal cation:anion ratio of 1:2, there should be an iron content of 33.7 atomic percent. Morimoto and Clark (1961) quote 18 reliable analyses of arsenopyrite, essentially cobalt free, and find that the average atomic percentage of iron is 33.3 ± 0.6 . They consider this small variation from an ideal iron content to be due to sampling and analytical errors.

Ferguson (1947) gives four analyses of glaucodots with cobalt varying between 15.00 and 18.65 weight percent, and an average value for (Fe + Co) of 35.6 atomic percent. The analyses are all rather old, three of them being made before 1880 and Ferguson considers them to be close enough to the ideal composition of (Co,Fe)AsS to warrant a structural formula of $12 [\text{CoFeAs}_2\text{S}_2]$ for the ideal cell content. The ideal atomic percentages calculated from this formula are Co-17.93%, Fe-16.99% and Co+Fe-34.92%.

Thus an introduction of as much as 18% cobalt into the arsenopyrite lattice has the effect of only raising the cation percentage from 33.7% to 34.9%, this small increase would seem to support the hypothesis that cobalt comes into the lattice at the expense of

iron, the sum of the cations remaining essentially the same. Thus knowing the iron:cobalt ratio in the mineral, and assuming the sum of iron and cobalt to be 34.0 ± 0.9 atomic percent, the atomic percentages of iron and cobalt can be calculated. The recalculated analyses for the Skuterud danaites are shown in Table 22.

When the percentages of cobalt in danaite are plotted on a frequency histogram (Figure 165c) it is interesting to see how they approximate to a normal distribution which tails off towards a value of 9% cobalt. This coincides with the boundary between danaite and glaucodot as suggested by Ferguson (1947).

It is concluded, since the composition range is within the limits of 0-11% cobalt and the diffraction pattern is not significantly different from common arsenopyrite, that the cobaltiferous arsenopyrite at Modum is danaite rather than glaucodot.

Variation in lattice spacings with cobalt content

Specimens of danaite, with varying amounts of cobalt as determined by the X-ray method described above, were studied with a Guinier focussing camera, using lead nitrate as an internal standard. Some of the reflections showed a rough correlation with the postulated cobalt content. The same samples were later run on a diffractometer, using silicon powder as an internal standard, the lattice spacings also showed a variation with cobalt content. Figure 166 shows the d-spacings in Ångstroms plotted as a function of cobalt content for these samples.

Variation in physical and optical properties with cobalt content

Since the variation in lattice spacings recorded above reflects a change in the crystal structure it was hoped that a

Figure 166: Variation in lattice spacings of danaite as a function of cobalt content. The values for the d_{200} , d_{022} and d_{131} spacings are those obtained using the diffractometer, the d_{212} and d_{102} spacings were obtained from Guinier powder films.

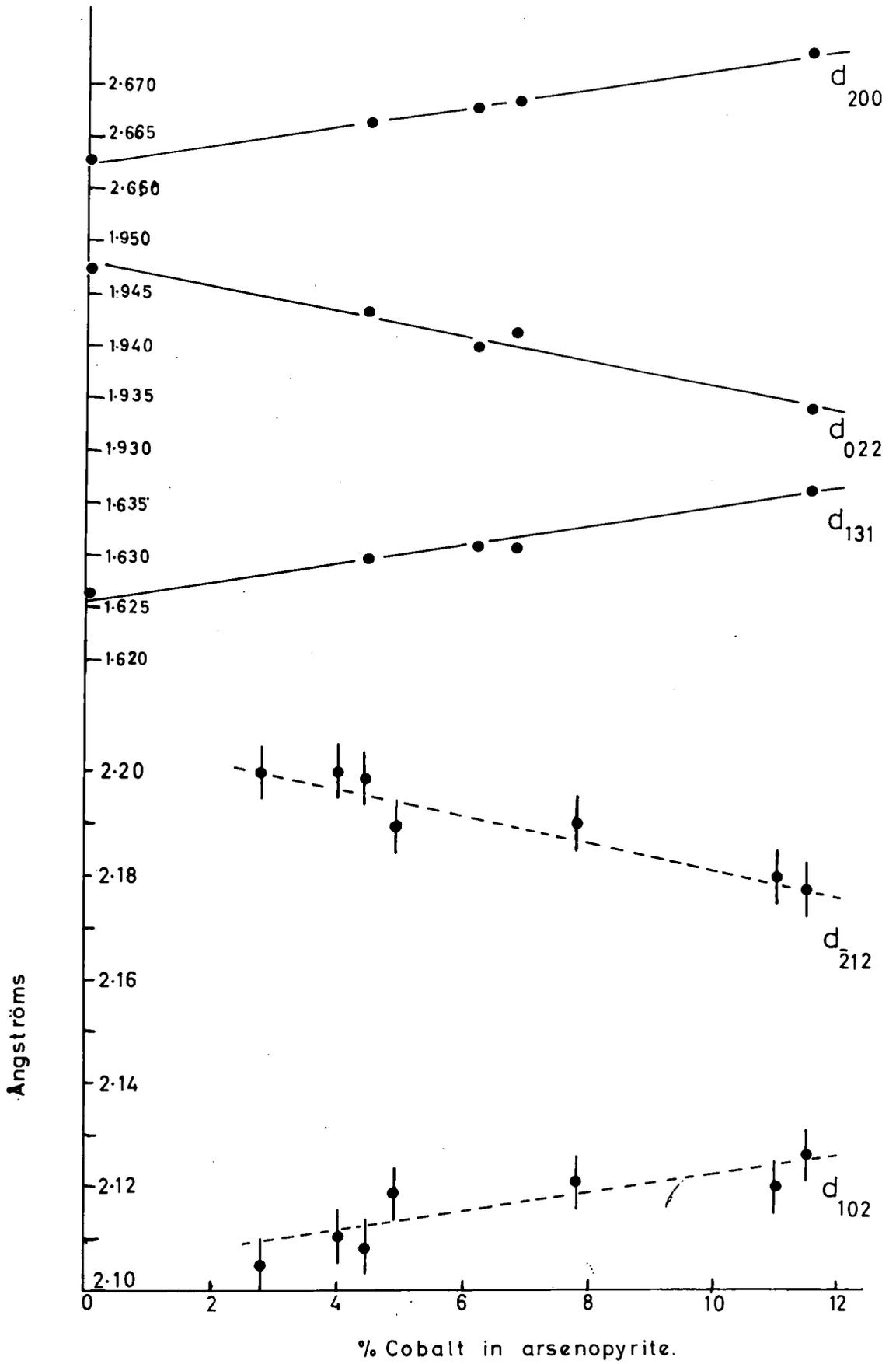


Fig. 166

corresponding change in physical or optical properties would be determinable.

Vickers hardness For this study four euhedral crystals of danaite which showed a variation in cobalt content between 2.5 and 11%, as determined by the fluorescence method, were selected. To compare possible differences in hardness it was considered necessary to ensure that all crystals were measured in the same crystallographic orientation. On these crystals the rhombic prism faces (110) and (012) ((101) and (120) in Buerger's monoclinic setting) were well developed. The (012) faces were lightly ground down using fine carborundum on a glass plate, the crystals were then embedded in plastic and polished parallel to the previously ground surface.

The polished surfaces thus prepared on the four crystals were all in the same crystallographic orientation, parallel to (012), and were bounded by faces of the rhombic prism (110). They also showed the traces of the prismatic (110) cleavage which was clearly picked out by plucking of cleavage fragments during polishing.

The measurement of the Vickers hardness was made with a GKN microhardness-tester. The specimens were levelled in a sample press and reference lines, to show the trace of (110) on the surface, were inscribed on the mount. The specimen was then placed on the stage and oriented by means of the reference lines such that the indentation would be made with one diagonal parallel to the (110) trace. Several indentations were made on each specimen and each diagonal of the impression was measured four times and the average taken.

The following conclusions were drawn from this study:

(1) The average value of all the indentations was 1014, with a range of 715-1354. Compared with the value quoted by Bowie and Taylor (1958) for 'glaucodot' of 1124 with a range of 1071-1166 and 'arsenopyrite' of 1094 with a range of 1048-1127, this is rather low.

(2) The difference in length between the diagonals was sometimes significant. In some series, for example, one diagonal was found to be consistently a little longer than the other. In other series however, the opposite condition prevailed in spite of the fact that the impressions were made on the same surface and with the same orientation. The cause of these divergencies must probably be sought in irregularities in the material and not in the crystal structure as such.

(3) The Vickers hardness is probably independent of the cobalt content. The results obtained are shown in Table 23.

Table 23

Vickers hardness of danaita

<u>Specimen No.</u>	<u>Surface orientation</u>	<u>Co.at.%</u>	<u>Mean V.H. Range (100gm. load)</u>	
1034-19	(012)	2.8	974	715-1354
1034-11	(012)	4.7	1064	974-1168
1034-17	(012)	8.1	1167	1048-1246
1034-10	(012)	11.4	949	762-1267

Other properties The following optical data are those obtained from specimen 1034-11, which contains about 8.1% cobalt. No significant variation in properties were found in specimens with differing cobalt contents.

Colour White, with a slight bluish tint against cobaltite. Reflection pleochroism very weak. In oil immersion, white colour with slightly brownish tint, reflection pleochroism slightly stronger.

Reflectivity The mean uniradial reflectivity in air, using white light and Sulitjelma pyrite as a standard, was 64.8%. The corresponding value in oil was 50.7%. The apparent uniradial reflectivity was measured, with monochromatic light ($\lambda = 549\text{m}\mu$), for a large number of grains. No constant value for R_1 or R_2 was found, indicating a biaxial nature. The values found indicate that R_p is equal to or less than 62.3%, R_g is equal to or greater than 65.7% and R_m lies in the range 63.4%-63.9%. Thus danaite is optically positive in the sense of Cameron (1963).

Anisotropism The danaite is very anisotropic in shades of olive-brown and greenish-blue. The polarisation figure shows a separation of the isogyres on rotation of the stage. The magnitude of this separation would indicate a $2V$ value of ca. 20° with a transmitted light polarisation figure. The dispersion of the reflection rotation (DR_r) and the dispersion of the apparent angle of rotation (DAR) were distinct and opposite in sense indicating that in both cases the dispersion is $v > r$.

Glaucodot from Modum

It has been shown that most of the arsenopyrite from the Modum mines is more properly referred to danaite than to glaucodot on the basis of composition and crystal structure. However the presence of a lamellar phase in the danaite from the Modum mines has been commented upon in a previous section. This phase has the appearance of being exsolved from the host danaite (Figure 167).

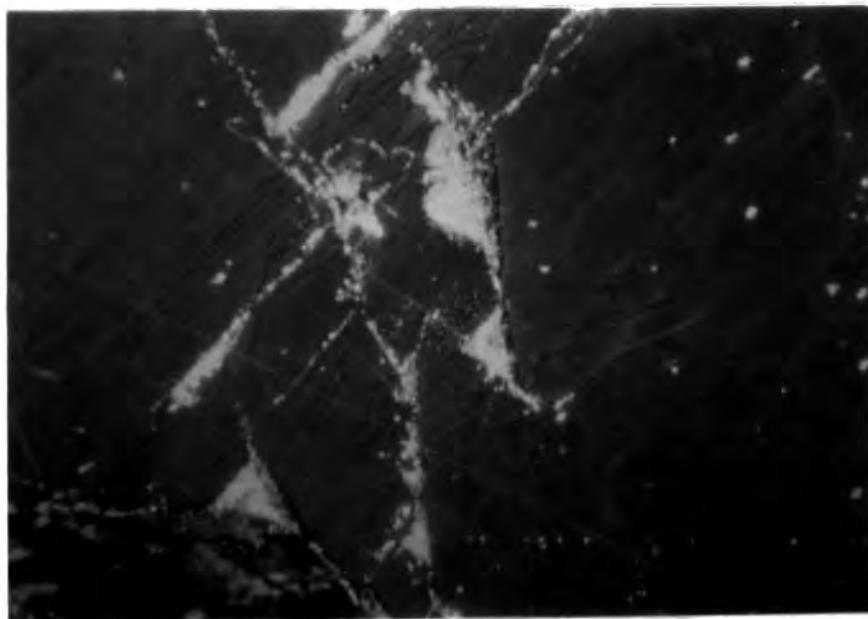


Fig. 167: Exsolution bodies of *glaucodot* (purple) in host *danaite* (blue). Specimen from collections of the Mineralogisk Museum, Oslo.
Reflected light, crossed nicols. X50

The exsolution bodies are too small to be removed for X-ray analysis but it was suspected that these may represent a cobalt-rich phase, possibly glaucodot, which had exsolved from danaite containing too much cobalt for the normal arsenopyrite structure to be stable.

An opportunity to test this hypothesis was presented when the Cambridge Instrument Co. Ltd. arranged a demonstration to show the capabilities of their 'Microscan' X-ray analyser. The instrument demonstrated had no provision for viewing specimens under crossed nicols so it was impossible to find the area of interest, however Miss P.J.Killingwoth of the Microscan Advisory Service kindly undertook the microanalysis at the Company's laboratories in Cambridge (Report No. M7.3.64). The following account is a condensed version of this report.

Preparation A line of conducting carbon was made from the area of interest to the edge of the sample.

Investigation A preliminary qualitative analysis of the area of interest showed that the elements Fe, Co, As and S were the main constituents of both phases. Photographs were taken showing the distribution of the elements S, Fe and Co (Figure 168). A line crossing both phases was then selected, and the intensity variations along it were recorded for each element (Figure 169). These were then compared with those from standards, and element concentrations were calculated. Since no suitable As standard was readily available, it was assumed that this element constituted the remainder of the material. This assumption was consistent with concentration variations indicated by the line scans.

Results Quantitative analysis of phase A was calculated from the larger of the two sections traversed by the line scan. Phase B did not appear to be completely homogeneous, but to show slight concentration gradients for Fe and Co, the composition was therefore calculated for a point mid-way between the two regions of phase A.

	Fe	Co	As	S
Phase A	18.5% \pm 1%	17% \pm 1%	46%	18.5% \pm 1%
Phase B	28% \pm 1%	7.5% \pm 0.5%	47%	17.5% \pm 1%

Results are corrected for inter-element absorption effects on the assumption that the concentration of arsenic may be taken as given.

Figure 168: Photographs of the images produced on the 'Microscan' electron probe, covering a part of the area shown in figure 167.

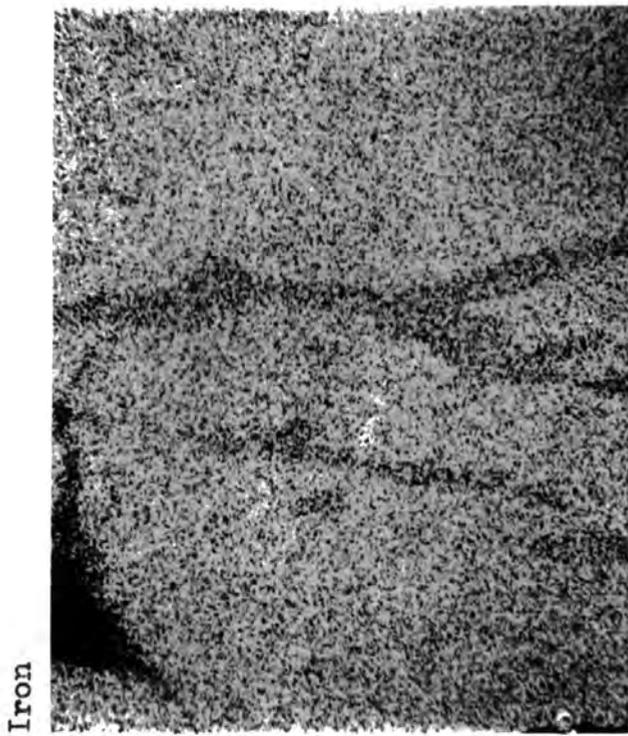
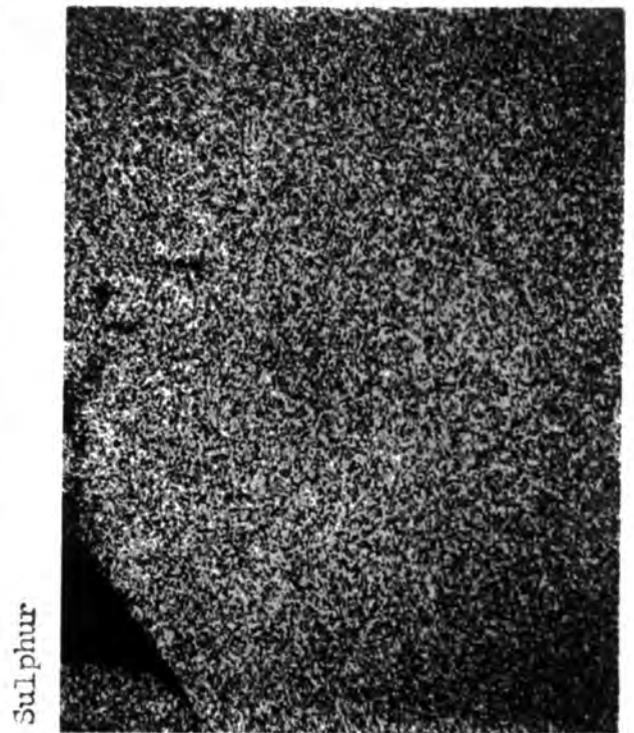
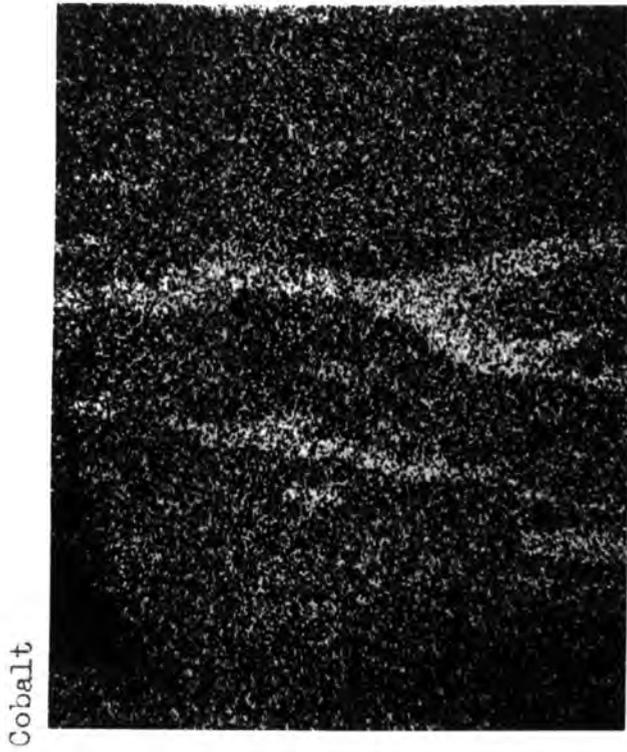


Fig. 168

Figure 169: Line scans produced by the 'Microscan' electron probe from traverses across the exsolution bodies seen in figures 167 and 168.

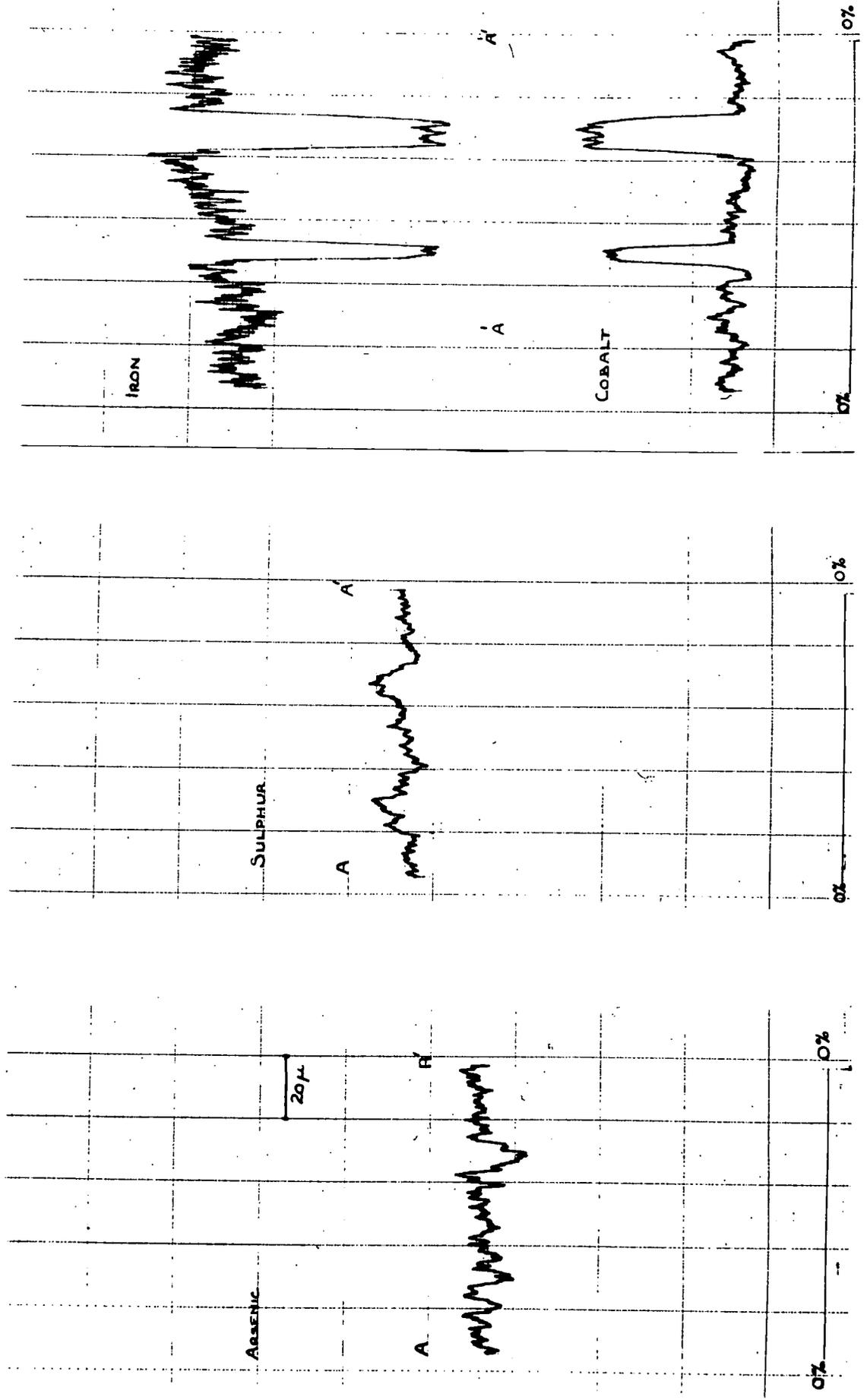


Fig. 169

This analysis is consistent with 'glaucodot' (phase A) being exsolved in host danaite (phase B). The 'glaucodot' is anisotropic in shades of deep purple-red and blue. It has a slight reflection pleochroism and can be indistinctly distinguished from the host danaite in plane-polarised reflected light. It is of slightly lower reflectivity and the Becke line test indicates that it is slightly softer.

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