The iron ore deposits of Ulu Rompin, Malaya

Bean, J. H.

Bean, J. H. (1975) *The iron ore deposits of Ulu Rompin, Malaya*, Durham theses, Durham University. Available at Durham E-Theses Online: http://etheses.dur.ac.uk/8144/

THE IRON ORE DEPOSITS
OF
ULU ROMPIN, MALAYA

by

J.H. BEAN

being a Thesis submitted to the Faculty of Science,
University of Durham for the fulfilment of the
Ph.D. degree.

DURHAM, ENGLAND 25th NOVEMBER, 1975.

External Examiner

ABSTRACT

The Ulu Rompin iron deposits are located in an area of rugged topography in the State of Pahang, Malaya. There are several high grade primary ore bodies and superficial sheets of lateritic ore.

The area is underlain by sheared alkalic rhyolite lavas and tuffs in which there are small lenses of sedimentary rocks, carbonates being the most important. There are several small stocks of hornblende granodiorite which probably merge into one larger parental mass at depth.

The Bt. Ibam ore body, which is the largest, is a tabular mass of magnetite in a highly magnesian gangue, originally chlorite and actinolite but now largely altered to talc. Underneath the ore there is a thin intermittent layer of calcic skarn, while massive chlorite appears along parts of the hangingwall. A short distance below the footwall there is granodiorite. The Bt. Batu Puteh deposit is magnetite associated with calcite marble, and at Bt. Hitam magnetite is found just above granodiorite. At Bt. Pesagi and Bt. Sanlong the primary ore is haematite with very little magnetite. These bodies are located in shear zones in the volcanic rocks, and are not associated with calcic or magnesian minerals.

All the ore bodies and country rocks have been pyritized, and other sulphides and sulphosalts have impregnated the Bt. Ibam body. These have produced zones contaminated by copper and zinc, and to a lesser degree bismuth and lead.

Hypogene and supergene alteration have extensively affected the iron ores and rocks and caused some redistribution of the impurity elements.

The iron ores probably originated in the cooling granodiorite magma and were transported as chlorides or chloride complexes. The control over the loci of mineralization was partly lithological - magnetite bodies in calcic and magnesian rocks, and partly structural - haematite bodies in shear zones.

ACKNOWLEDGEMENTS

The writer is indebted to many people for assistance during different stages of this study. To the directors of Rompin Mining Company for permission to study the deposits and the provision of accomodation at Bt. Ibam, and to the Mine Superintendent, Mr. D. Newick, for help at the mine. The writer has held discussions with many of the company geologists over a period of several years, but is especially indebted to Mr. B.Mellor, Mr. D. Taylor, Mr. M. Muthuppalianappan and Mr. P. Muthuveerappan.

Professor K.C. Dunham kindly accepted the writer as a research student, and his successor, Professor G.M.Brown, continued this support. To Mr. R. Phillips, for help and guidance throughout the laboratory studies, the writer is especially grateful. Many colleagues at the University of Durham helped with various aspects of the work, but of particular assistance were Dr. E.H. Emelaus, Mr. G. Rowbotham, Dr. J.G. Holland, Dr. J. Aucott, Dr. Tarik Tugal and Mr. M. George. The laboratory technical staff had to cope with a large quantity of material which was highly decomposed and very difficult to prepare, and to all of them the writer expresses his gratitude.

TABLE OF CONTENTS

GEOLOGY OF ULU ROMPIN

V INTRUSIVE IGNEOUS ROCKS

VI STRUCTURAL GEOLOGY

IRON ORE DEPOSITS

VII GENERAL DESCRIPTION

VIII BUKIT IBAM ORE BODY

XIV GEOCHEMISTRY OF THE BUKIT IBAM DEPOSIT

Minor Elements in Ores and Rocks

Distribution of Iron and Trace Elements

XV THE DEPOSITS AT BUKIT PESAGI AND BUKIT SANLONG

Ore Deposits

XVI THE BUKIT HITAM DEPOSIT

LIST OF FIGURES

LIST OF TABLES

I INTRODUCTION

Location of Area and Access

The iron ore deposits of Ulu Rompin are located within a belt of country measuring approximately eight miles north to south and four miles east to west, which lies 45 miles northwest of Kuala Rompin, a small fishing village on the east coast of the State of Pahang in Malaya - Fig. 1. The largest individual deposit, Bukit* Ibam, is located towards the centre of the eastern edge of the mineralized zone, and has the co-ordinates 3° 10' N; 102° 59' E - Fig. 24. The name Ulu Rompin, which is Malay for "upper reaches of the Rompin (river)", properly refers to a very large area in central Pahang, but when used as a locality name for the iron ore deposits it is restricted to the small area mentioned above. Due to the paucity of good reference localities in deep jungle country there is frequently considerable confusion over place names, but the names used throughout this thesis are those adopted by the mining company.

Until the area was opened up by the Rompin Mining Co. Ltd. access was extremely difficult, and at the time of writing there is no road linking the mine with the coast. During the many years of exploration and prospecting the only feasible access route was by boat up the Sungei* Rompin as far as Kuala Keratong, and thence overland to the deposits. This route was extremely laborious; the river journey alone was 120 miles due to large meanders in the lower stretches, and many shallows developed in dry weather. Soon after the decision to develop the deposits had been taken the company constructed a railway from Bt. Ibam to the coast, and this is used to supply the mine with most of its requirements and to transport the ore to the coast. An earth landing strip suitable for small aircraft was later constructed at Bt. Ibam, and this greatly facilitated the movement of staff and light supplies. The

---

* In written Malay, Bukit (= hill) is usually abbreviated to Bt. and Sungei (= river) to Sg. These abbreviations are used throughout this thesis.

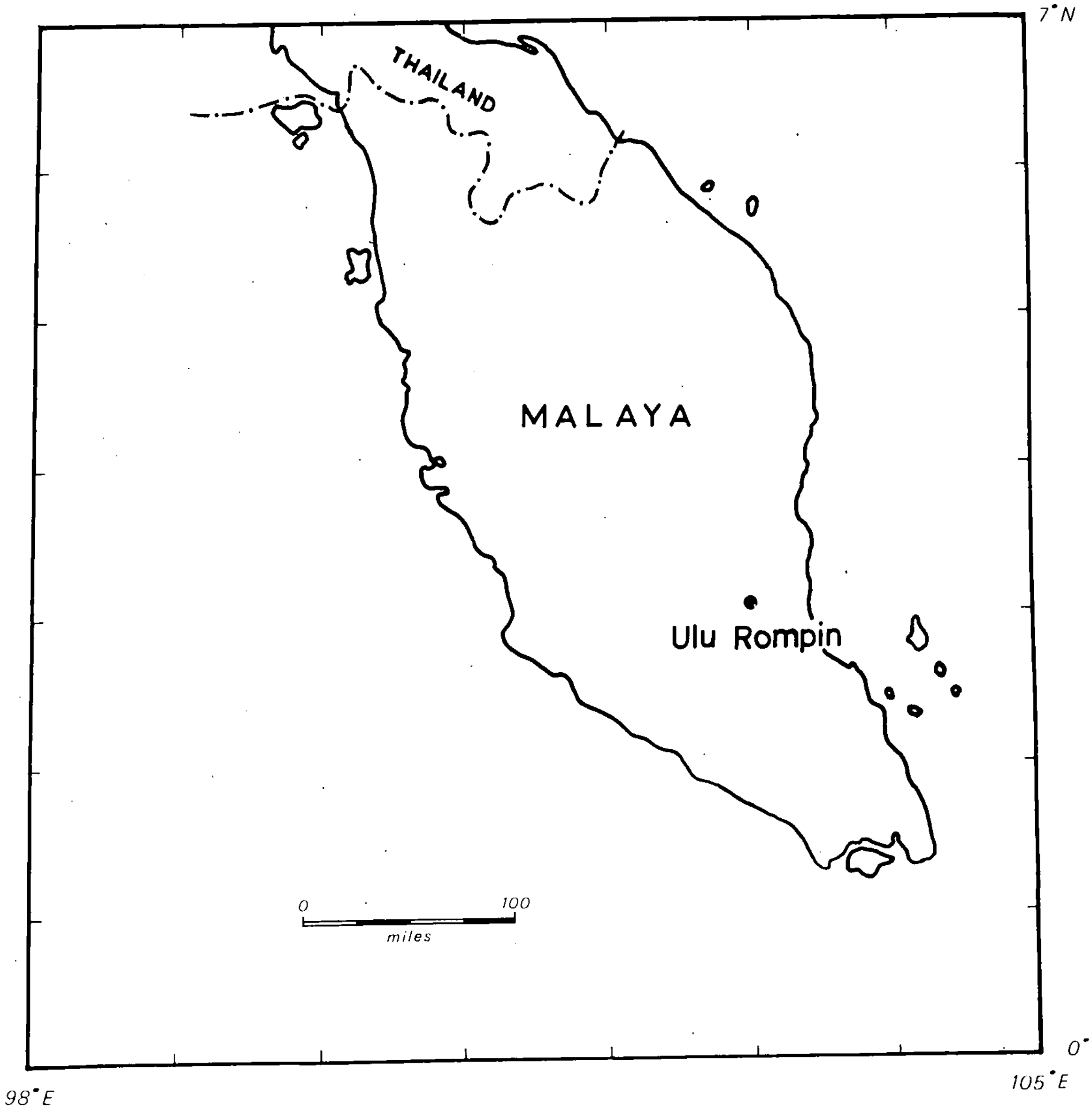


Fig. 1. Sketch map showing the location of Ulu Rompin.

inaccessibility of the iron ore deposits was for many years a handicap to their exploitation, but it is hoped that the mining operations may act as a catalyst in the development of the vast hinterland which is at present uninhabited, except for a few tribes of aborigines.

Physiography

The iron ore deposits are found scattered along a north to south trending ridge which rises to a peak at Bt. Sembilan 1642 ft, one mile west of Bt. Ibam. The ridge extends over a distance of nine miles with an average width of two miles, but becomes somewhat wider towards the north. There is a strongly marked crest line along much of its length, but it divides into several subsidiary ridges at the northern end as the elevation drops - Fig. 2. The ridge forms a marked barrier to east-west movement in the region, and the lowest pass is situated half a mile northwest of Bt. Ibam, at a height of 750 ft. There are a number of minor peaks within the main ridge mass, and many of the iron ore deposits are located on such peaks and take their names from them. The ridge has very steep sides, and is dissected by numerous small, fast-flowing streams, which run almost due east or due west until they meet the main drainage valleys running from north to south.

The streams on the western flanks of the ridge drain into the Sg. Jeram, which follows a meandering course along the western edge of the Ulu Rompin area and then joins the Sg. Keratong, the two together forming the Sg. Rompin. The confluence of these rivers is 120 miles upstream from the mouth of the Rompin river but at an elevation of only 19 ft above sea level. The streams flowing off the eastern flanks of the Bt. Sembilan ridge drain into the Sg. Tepisoh, which runs from north to south in a narrow, flat-bottomed valley, and eventually joins the Sg. Jeram. Flanking the middle reaches of the Tepisoh valley on the east is the mountain mass of Bt. Permandan 1856 ft, but elsewhere the valley is bordered by low, rolling country. The main rivers and streams are shown in Fig. 24.

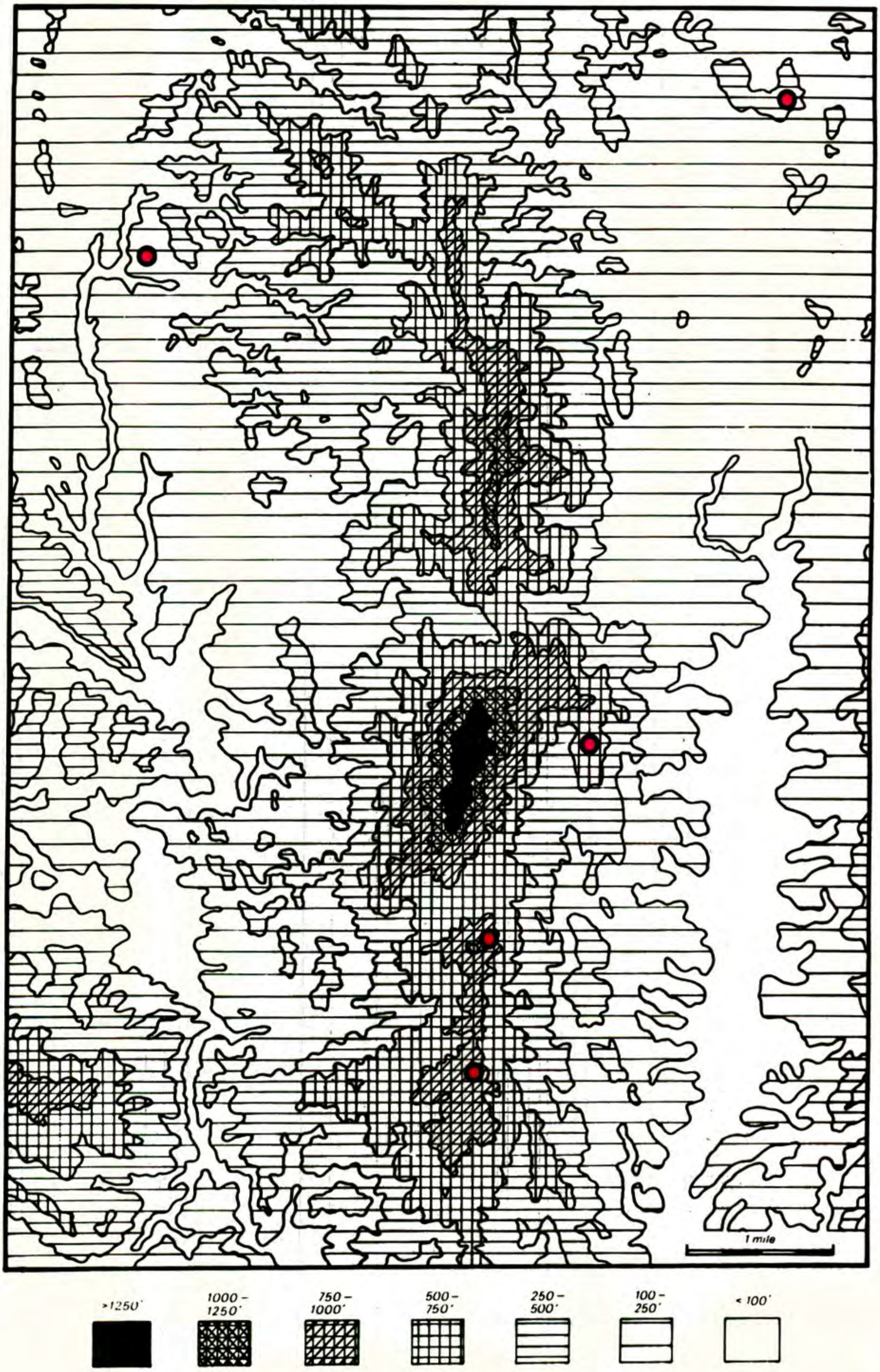


Fig 2 Topography within the Ulu Rompin mineralized zone. The main primary ore localities are shown in red.

As the Ulu Rompin area lies close to the east coast the most marked seasonal influence on the climate is the northeast monsoon, which blows from November to March. During this period torrential rain can severely limit, and on occasions prohibit, mining operations, largely due to the fact that the haulage roads become unusable. For the rest of the year there is fairly regular rainfall in the form of short heavy showers, and hot sunshine. The average annual rainfall is 150 inches, and the relative humidity is extremely high throughout the year. This typical Tropical Rainforest type of climate produces extremely deep and extensive weathering of the surface rocks, and encourages a luxuriant vegetation.

The mineralized area was originally covered by primary jungle or smaller growth, except for a few small areas which had been cleared by aborigines for cultivation. The steep hill slopes are forested by extremely tall trees with little undergrowth, but the lower slopes and valley floors can be covered with a dense cover of shrubs, small trees and grasses. The centres of mining operations have been stripped of their original cover, but much of the area still retains its natural vegetation, and there is some exploitation of valuable timbers.

History of the Ore Deposits

The first indication that there were large deposits of iron ore in the Ulu Rompin area came in 1930 when Che Abdul Rahman, a retired Pahang Forester, showed some samples of iron ore he had received from local aborigines to staff of a Japanese mining company, Ishihara Sangyo Koshi. This firm made a reconnaissance survey in July 1930, and on the strength of the results applied for a prospecting licence, which was granted in January 1932. The only deposits known at that time were at the southern end of the Bt. Sembilan ridge near Bt. Pesagi, and the prospecting was confined to this area. After sinking a large number of pits and driving five adits, the Japanese company concluded that there were about six million tons of medium grade ore, but as this was not sufficient to justify the very large capital costs necessary to develop the deposits the area was abandoned.

The subsequent discovery that there were further large deposits north of Bt. Pesagi led the Japanese to start prospecting once again. A comprehensive exploration extending from 1937 to 1941 located all the major deposits, and from the prospecting results the company concluded that there were 20 million tons of proven ore and 10 million tons described as "hidden reserves". The boulder ore averaged 50% Fe and the primary ore 65% Fe. Ishihara Sangyo Koshi began preparations to open up the deposits, but due to the unsettled conditions then prevailing in the Far East the Malayan Government withheld the mining permit and all operations were suspended.

At the conclusion of the war in 1945 the Malayan Government confiscated all Japanese assets in Malaya, and the Ulu Rompin deposits, together with several other mining properties throughout the country, were put up for sale by tender. Broken Hill Proprietary Ltd., the giant Australian company, showed some interest, but after a brief examination of the property did not pursue the matter further.

In 1951 the property was taken over by Eastern Mining & Metals Co. Ltd., the largest iron ore miners in Malaya. This company began prospecting on a small scale in 1952, and initiated a diamond drilling programme to supplement the data available from Japanese records. All the early work was centred on Bt. Ibam, which obviously contained by far the greater part of the ore reserves. A report prepared from this early work was used as a basis for discussions to raise some of the capital necessary for development, but it became apparent that a thorough survey of the whole of the Ulu Rompin mineralized area was required. Rompin Mining Co. Ltd. was formed in 1958 as a subsidiary of Eastern Mining & Metals Co. Ltd., to continue and extend the prospecting programme.

This programme consisted of a geological examination of the mineralized zone, a check on old Japanese pits and the sinking of new ones, the driving of some new adits and cleaning out of old Japanese adits, interpretation from geophysical data, and an extensive diamond drilling programme. The company's prospecting programme had the effect of reducing the original

Japanese figures for ore reserves; the reasons for this were firstly the discovery at Bt. Ibam of large masses of fine friable ore, which at the time was not a saleable product, and secondly, the delimiting of patches of copper-contaminated ore. Later work showed that zinc and bismuth also formed impurities in some sections of the ore bodies. The results of the prospecting programme have not been released, but taking the Ulu Rompin deposits as a whole, sufficient reserves were available to sustain mining operations. A fortuitous event was the increased use made by the steelmills of pelletized ore, for which fine ore is a necessary raw material.

The construction of a railway from Bt. Ibam to the coast began in 1961, but it encountered enormous difficulties as it had to pass over wide areas of fresh water swamp. The mine installations and staff quarters were built at Bt. Ibam, where the early mining operations were concentrated. Development was later extended to Bt. Pesagi. Production began in 1962 and by 1966 had reached 3 million tons per annum, all of which was exported to Japan.

Purpose of Present Study

Although Malaya has been an important iron ore producing country since the early 1920's no major deposit has been the subject of a detailed examination, and mineralogical studies in particular have been very sparse. The Ulu Rompin deposits are the most recent to be brought into production and offer an interesting subject for study. Within a relatively small mineralized zone there are several high grade deposits which have some noticeable differences in spite of their proximity to one another. The main deposit located at Bt. Ibam deposit is contaminated with Cu, Zn, and to a lesser extent Bi and Pb. The writer was conversant with the area from work carried out while a member of the Malayan Geological Survey, and returned for further studies after leaving government service. The major part of the fieldwork was centred on the area from Bt. Pesagi to Bt. Ibam, the two largest deposits and both in production at the time, but all except one of the subsidiary deposits were also examined. The exception was Batu Puteh, which lies some miles from the other deposits and was not easily accessible. To supplement the surface studies representative drill core from all the deposits was examined and sampled.

The laboratory studies were undertaken in the University of Durham. All the X-ray work (diffraction and fluorescence analysis) and polished and thin section studies are the writer's own, but some use has been made of chemical data obtained in the company's laboratory and due acknowledgement is made in the requisite place.

The aim of the present investigation was to study the geology and geochemistry of the ore bodies and their immediate host rocks, with a special emphasis on the examination of the ores in polished section. Of particular interest was the identity of the minerals responsible for the contamination in the Bt. Ibam ore body as previous information was extremely limited. From the combined results of the field and laboratory studies it was hoped that some conslusions could be drawn regarding the genesis of the iron ores.

II REGIONAL GEOLOGY & STRATIGRAPHY

Regional Geological Environment

The Ulu Rompin mineralized zone lies along the eastern edge of the only large area in southeast Pahang for which geological information is available in some detail. The area is one of the most inaccessible in Malaya but was mapped some years ago by the Geological Survey. For the country east of Ulu Rompin geological information is fragmentary, and limited to field data of a reconnaissance nature plus photogeological interpretation. Fig. 3 shows the outline geology of part of southeast Pahang round Ulu Rompin. No fossils have yet been located within a 30 miles radius of Bt. Ibam and the age of the local rocks is conjectural. The nearest fossils to the north range in age from Permian to Upper Trias-Rhaetic, while to the south, flatlying sandstones of Wealden age lie on a marked erosion surface.

The Ulu Rompin area lies just to the east of a major fault zone which follows a bearing of approximately 350°. East of the fault the major structural features trend north to south but swing slightly west of north towards the northern end of the area, whereas west of the fault the dominant trend is northwest. Fourteen miles west of this fault and parallel to it there is another large fault, the full extent of which is not known. To the west of this second fault the structural trend is again north to south.

Between the two faults there is an elongated basin, the axis of which can be traced over a distance of 25 miles, and which is probably terminated at both ends by the large fault zones. The centre of the basin exposes a monotonous sequence of purplish-red siltstones and shales which outcrop over an area of 50 square miles, and these are underlain by a thick series of sandstones, greywackes, and tuffaceous sandstones. The basal member of the rocks forming the basin is a massive sequence of purplish-red

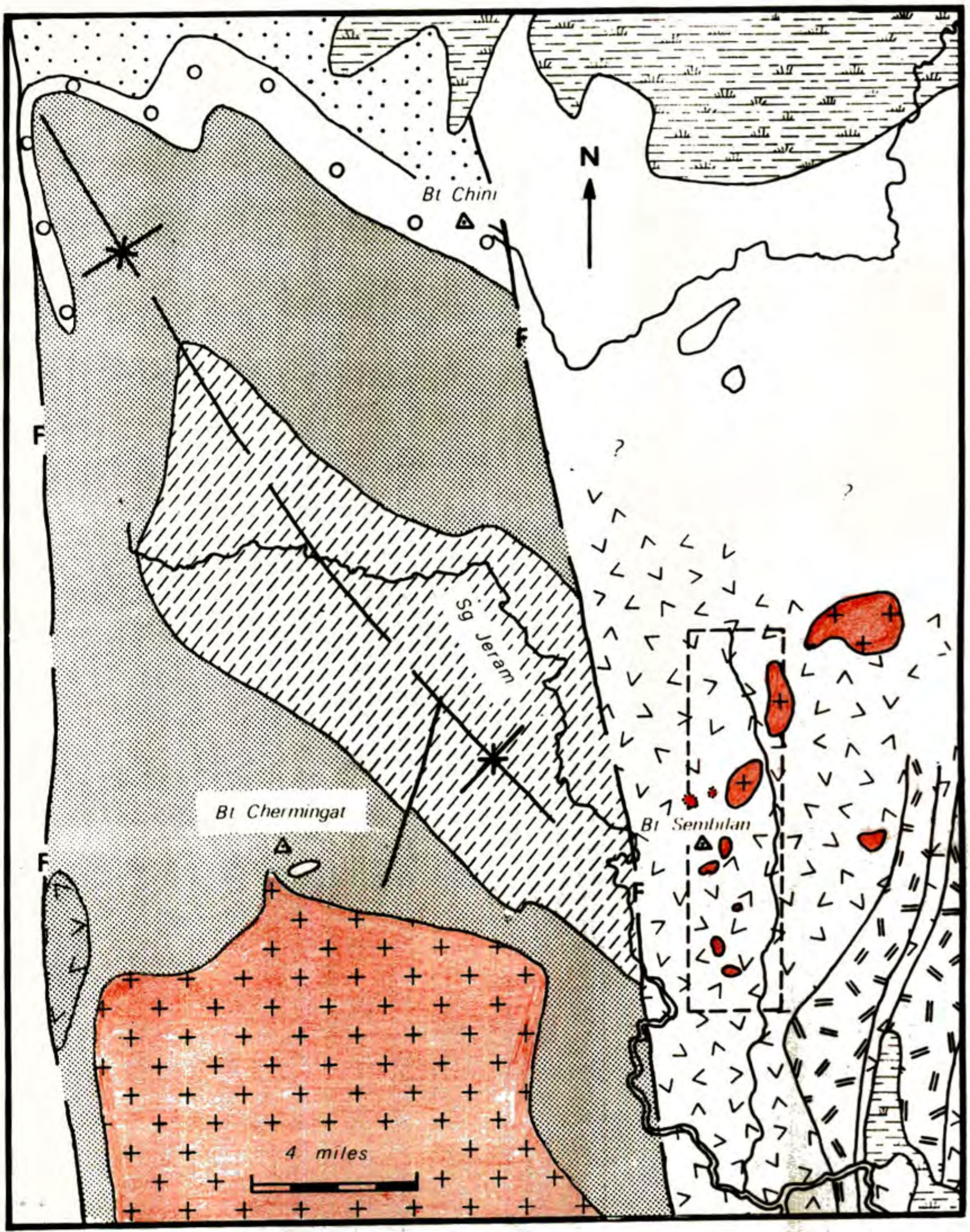
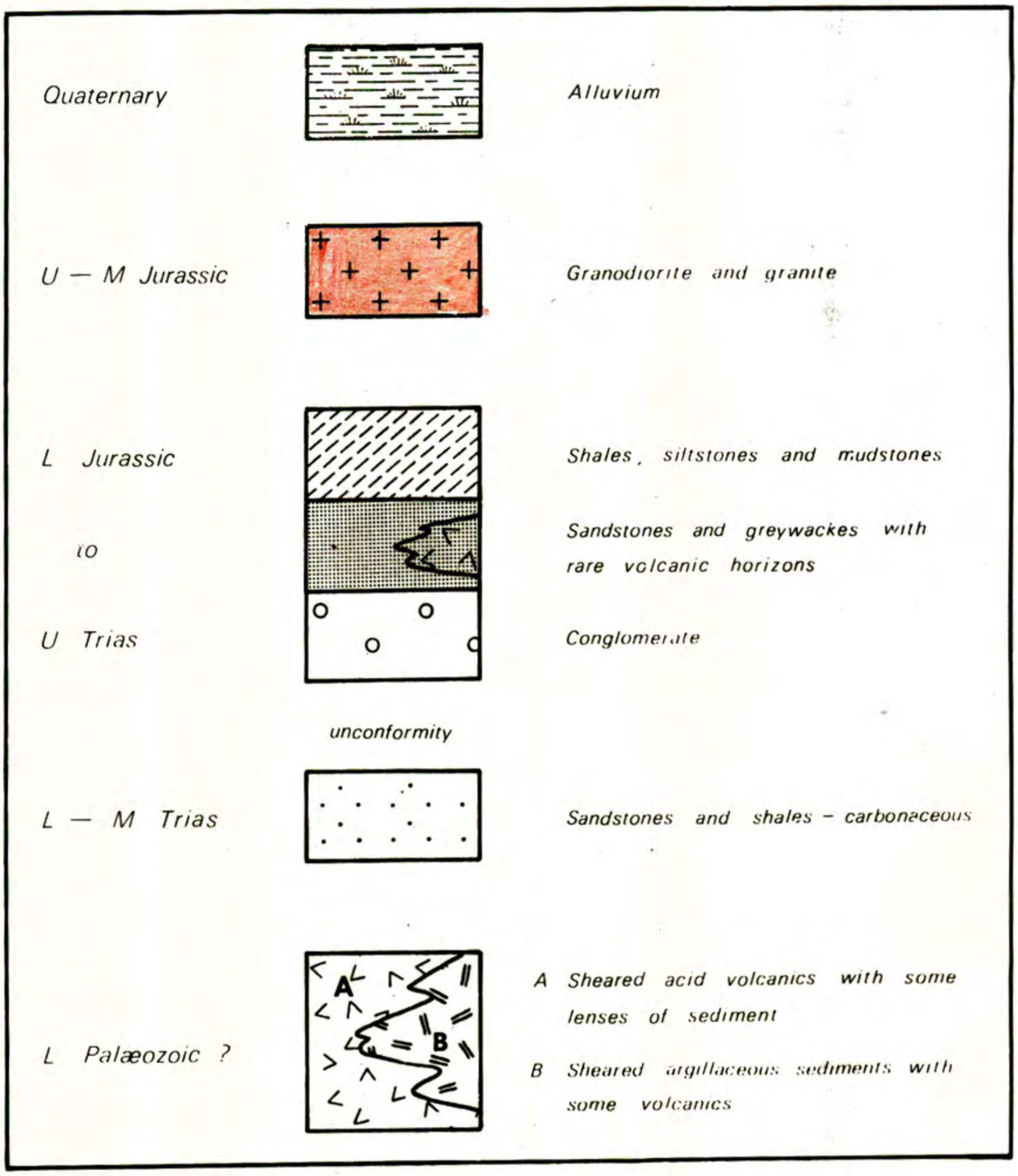


Fig. 3 *Geological sketch map of part of south–central Pahang : the key is overleaf. The figure is based on Geological Survey maps and ROMPINCO plans. The rectangle shown on the right contains most of the primary ore bodies.*



Suggested stratigraphical column for Ulu Rompin: palaeontological data are lacking.

coarse conglomerates and grits, which only outcrop along the northern rim of the basin where they form a very distinct ridge, Chini 2101 ft. The base of the conglomerate is marked by a thin band of rhyolitic crystal tuff. The basin as a whole is a simple open structure, with a few superimposed subsidiary folds and small faults. Metamorphism of the rocks is minimal, being confined to the recrystallization of clay material to chlorite and mica.

The massive basal conglomerate lies with a marked angular unconformity over a series of black shales and sandstones, which are shown on the published geological map as being contiguous with the sandstones and red shales above the unconformity. The writer mapped a large area containing the full rock sequence and does not agree with this correlation, as the unconformity is a major structural break and separates unrelated stratigraphical units. There are no fossils to assist in dating the rocks above the unconformity, but the writer suggested a Lower Jurassic age. A group of rocks in central Pahang which have a strong lithological similarity to the sequence above the unconformity, and appear to occupy the same stratigraphical position, have been described by Koopmans (1968) as "Tembeling Formation". He dates this formation as Upper Triassic to Jurassic, which agrees with the writer's interpretation of the rocks forming the basin to the northwest of Ulu Rompin. The rock sequence found in the basin does not reappear within the Ulu Rompin mineralized zone, which suggests that there is a fault separating the areas. The published geological map does not show a fault extending as far south as the Ulu Rompin area, but evidence from photogeology suggests that the fault is more extensive than the map depicts.

South of the basin there is the northern extremity of a very large intrusion of hornblende granite, quartz monzonite and granodiorite. The granitic mass occupies low swampy country, and contact relationships with the surrounding rocks are unknown. Further south the same granite is unconformably overlain by sediments of Cretaceous age, and it is probable that the granite is of Middle to Upper Jurassic age, which is typical of many of the Malayan granites.

The country lying to the east of the Ulu Rompin area is largely unexplored. There is freshwater swamp extending for an average of 16 miles from the coast, and this is flanked inland by a series of low rounded hills and few ridges which expose sandstones with intercalated volcanic rocks. A belt of undulating swampy country further west is underlain by shales, but towards the Ulu Rompin area acid volcanic rocks become the major rock type, and they form several strong ridges with a general north to south alignment. Some of the volcanic rocks are intensely sheared but others show no evidence of having been subjected to stress.

Previous Geological Studies in Ulu Rompin

The earliest investigations were made by staff of Ishihara Sangyo Koshi at Bt. Pesagi, where they concluded that the area was made up of quartzite and shale which has been intruded by liparite, and they linked the iron ore genetically to the intrusion. No records of Japanese geological work on the other deposits have been seen by the writer. Fitch (1941) examined the area and confirmed the thoroughness of the Japanese prospecting but doubted some of the ore reserve figures. He also gave a brief account of the general geology of the area and concluded that the rock enclosing the iron ore deposits was a sheared quartz porphyry, possibly with tuff in places. He believed that an epidote granite which intruded the quartz porphyry was the source of the mineralizing solutions, which had been injected along shear planes in the quartz porphyry. He rejected a replacement origin for the iron ore as he could find no specimens showing a halfway stage between ore and country rock, but Fitch was restricted in his observations to surface outcrops and adit dumps, the adits themselves having collapsed.

Macandie and Canavan (1948) examined the deposit on behalf of Broken Hill Proprietary Ltd., and concluded that the Bt. Ibam ore had been formed by the injection of magnetite veins along shear zones in quartz porphyry.

MacDonald spent some time in the Ulu Rompin area in 1954 while prospecting operations were under way. He examined some of the surrounding country rock as well as the ore deposits, and prepared a report which was later incorporated into an account of the geology of a large area in Pahang mapped jointly by several members of the Geological Survey of Malaya. (This report was eventually published in 1970) MacDonald was strongly influenced in his interpretation of the geology of Ulu Rompin by the work of Landergren (1948) on some of the central Swedish ores, and suggested that the Rompin ores were originally of sedimentary origin but had been altered when the area was subjected to partial granitization. He described the porphyritic acid rocks as "metasomatic quartz porphyries" which had been formed by granitization of sediments.

Most of the geological work in the Ulu Rompin zone has been carried out by, or on behalf of, the Rompin Mining Co. Ltd. Not all of this information is available to outside personnel, but the writer has been able to consult several reports, drill logs, assay figures etc., and discussed the geology with many company geologists over a period of years. Two early reports, by Hitchen and Moss (1954) and Tillia (1956), suggested that the ore had segregated from a basic magma, and both regarded the serpentinous and chloritic rocks associated with the iron ore as being genetically related to the same magma. Consequent to the accumulation of a mass of new data during prospecting and early mining operations, ideas regarding the genesis of the ore bodies have gradually changed, and the company now believe that the Bt. Ibam ore body formed by replacement of calcareous rocks, and the Bt. Pesagi tabular haematite bodies by oxidation of massive pyrite. No detailed mineralogical studies have been made by company geologists, but a small number of polished and thin sections were examined by Australian Mineral Development Laboratories and the Commonwealth Scientific and Industrial Research Organization, Melbourne, during investigations into impurities in the ore, and the Geological Survey of Malaya examined a very small number of thin sections of drill core samples.

Stratigraphy of the Mineralized Zone

There are no fossils within the mineralized zone so the stratigraphy has to be deduced by comparison and conjecture. There are only two major units involved, a thick sequence of porphyritic acid volcanic rocks with intercalations of sediments, and hornblende granodiorite. The porphyritic rocks are difficult to classify, but the writer believes that they are dominantly waterlain acid pyroclastics with subordinate lavas.

The porphyritic rocks typically have a moderately to steeply dipping foliation, but the sediments immediately to the west in the valley of the Sg. Jeram are only gently flexured and have no vestige of a foliation, which suggests that the Ulu Rompin pyroclastics are older than the Sg. Jeram sediments i.e. pre-Jurassic. There are rhyolitic tuffs interbedded with Middle and Upper Triassic sediments lying about 30 miles northwest of Ulu Rompin, but again these rocks have no foliation. There are two extensive areas of rocks which are virtually indentical with the foliated porphyritic rocks of Ulu Rompin, both in the State of Perak and about 150 to 200 miles northwest of Ulu Rompin. There is obviously a long gap between these areas and Ulu Rompin, but the similarity between the rocks is very marked. It has been shown by Jones (1970) that in the more northerly area the rocks are of Lower Palaeozoic age, but in the other area the age is uncertain. Savage (1937) stated that the rocks were "post-Trias and pre-granite", but he believed that the oldest associated sediments were Permo-Carboniferous whereas recent work has shown that the rocks extend back to the Ordovician, so the porphyritic rocks could be of Lower Palaeozoic age. Up to 1964 it was believed that Lower Palaeozoic rocks did not occur anywhere in Malaya east of the Main Range batholith, but in that year Silurian fossils were located in west Pahang. It is tentatively suggested that the foliated acid porphyritic volcanic rocks in the Ulu Rompin mineralized zone may be of Lower Palaeozoic age, and can be correlated with those in northern Perak.

The thickness of the beds cannot be determined with accuracy, but drilling at Bt. Ibam has proved 950 ft, and this is only a small part of the sequence. The porphyritic rocks have an outcrop width of at least three miles in some places, and if an average dip of 45° is taken, which is probably too low, it gives a thickness of the order of 10,000 ft. This of course ignores any doubling or repetition due to folding, but as discussed later under "Structure" there is no evidence that this has been a significant factor.

The granodiorite is found as small stocks which are probably apophyses of a larger body at depth. It intrudes the porphyritic rocks, and there is no reason to believe that it differs greatly in age from the main Malayan acid intrusives i.e. Middle to Upper Jurassic. In several parts of Malaya acid intrusives have been eroded and over lain by Lower Cretaceous sediments, and radiometric dating confirms the Jurassic age.

III ACID VOLCANIC ROCKS

Introduction

These rocks are by far the most abundant in Ulu Rompin, and make up the greater part of the Bt. Sembilan ridge which runs as a north to south axis through the area. Good natural exposures are only found in stream sections where they frequently form small waterfalls, but the mine workings, some deep road cuttings and drill core provide additional exposures and samples. Intercalated with the volcanic rocks are lenses of sedimentary material, which appear to become more important towards the northwest of the zone. The typical volcanic rock consists of a pale to dark grey or greenish-grey fine grained matrix, in which are set fine to coarse quartz grains, sometimes accompanied by phenocrystic feldspars. Although the normal rock is pale coloured, many are iron stained to various shades of yellow, orange, brown, red or purple - Fig. 4.

Some previous workers have used the presence or absence of visible quartz as the main criterion for naming these rocks in the field, the presence of quartz classifying the rock as a "quartz porphyry". As a field term this is acceptable, but any genetic implication is misleading. In many parts of the Ulu Rompin area the rocks have a pronounced foliation. This is particularly well shown round the ore bodies at Bt. Pesagi and Bt. Sanlong, where lensing of the minerals can occasionally be displayed; some of the coarser rocks could almost be classed as augen schists. Most of the earlier reports suggest that all the acid volcanic rocks in the Ulu Rompin area are foliated, but this is not the case, and particularly round the Bt. Ibam deposit the country rock is largely unsheared. The reason for this is discussed later under "Structure". Some of the finer grained rocks contain no visible quartz grains and look like a hornfels or a schist depending on the degree of foliation; a few of these rocks may have been clastic sediments, but the majority have been ashes and fine equigranular tuffs. The presence of agglomerate has been recorded by mine staff in drill core from Bt. Ibam, but

Fig.4. View looking west across part of the ore body at Bt. Ibam. The rapid change in colour in the acid volcanic rocks is well displayed in the face below the drill.

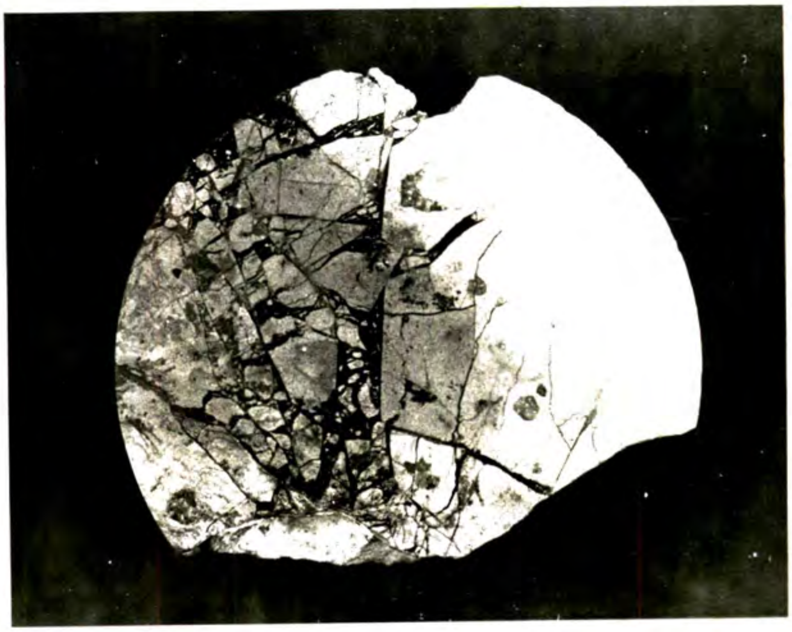


Fig. 5. Drill core sample which demonstrates the development of an agglomeratic appearance in a normal porphyritic lava. The unaltered rock appears at the top right, but the left half of the sample has been stained and veined by goethite, which has in places caused some displacement of rock fragments. x 2.4

no surface exposures have been located, and it is possible that the agglomeratic appearance is deceptive. The writer has examined specimens from Bt. Ibam, and also the iron ore mine at Sri Medan in Johore, which superficially appear to be a breccia or agglomerate, but close inspection shows that they are normal extrusive rocks in a state of partial alteration and replacement. Weathering solutions penetrate along a system of interlocking joints, leaving isolated unaltered and unstained fragments between the joint planes, and the whole rock develops an agglomeratic appearance - Fig. 5.

It is extremely difficult to detect contacts between the various members making up the group of acid volcanic rocks, and it is possible that many of the contacts were originally gradational. Alteration has further tended to mask contact relationships. In some of the deep road cuttings in the mining areas it is possible to observe distinct textural and mineralogical variants within the volcanic rocks, but it is impossible to trace any firm line of junction between them. The reverse position is not uncommon, and what appears from a distance to be a sharp contact proves to be merely an effect of differential weathering and staining. Where the volcanic rocks are not too deeply weathered they often show a profusion of haphazardly orientated joints, which can be so closely spaced that the rock disintegrates into small angular fragments when struck with a hammer. The presence of faults can be established in the country rock adjacent to the main Bt. Ibam and Bt. Pesagi ore bodies, but elsewhere it is difficult to detect fault planes. This is partly due to the fact that the lithological similarity makes displacements difficult to detect, and also that faults tend to encourage accelerated weathering. No unequivocal macroscopic evidence of bedding has been seen by the writer, but it is likely that it coincides with the foliation planes, which strike in a general north to south direction and dip at moderate to high angles in an easterly, and less frequently westerly, direction. Bedding contacts have been observed in a few thin sections, but their direction cannot be ascertained from unorientated core.

Classification

The volcanic rocks are a difficult group to classify with certainty and have been subject to several interpretations. Very similar rocks occur in two areas in northern Perak, and some of the petrographic description in Savage (1937) could be applied verbatim to the Ulu Rompin rocks. The Perak rocks were originally classified as quartz porphyries, but recent work in the Grik area (Jones, 1970) has proved that the rocks are pyroclastic in origin. The main difficulties in classifying the Ulu Rompin rocks are the absence of data on contact relationships, the high degree of recrystallization and alteration the rocks have undergone, and the effects of shearing in some samples. The gross textures are often preserved, but the detail in the fine grained groundmass has been obliterated.

The consistent characteristics of the majority of the rocks are their porphyritic texture, their acidic composition, and, with important exceptions, some degree of foliation. The rocks have been classed as quartz porphyries by some previous workers, and the published geological map describes them as "metasomatic quartz porphyries". The latter classification is susceptible to critical examination as it implies that the coarse grained crystals are porphyroblastic. Examination of thin sections proves that this is not the case, and in fact precisely the opposite process has been in operation and the large crystals have often broken down into finer grained aggregates. The rocks are of normal volcanic origin and the evidence from a large number of thin sections shows that a high proportion of the rocks are pyroclastic, a small number are lavas, and the balance cannot be assigned to either group. Alteration is so advanced that it is impossible to determine what was the precise mineralogical composition of the original rocks beyond the fact that they were acid. Chemical data shows that the rocks were rich in alkalies, particularly potash. There are lenses of sedimentary rocks in the volcanic sequence and some rocks show the intermingling of clastic and pyroclastic debris. The rocks appear to have been deposited in shallow water conditions.

The volcanic rocks of the Ulu Rompin area appear to be identical with leptites, the dominant rock type in that part of Central Sweden where iron ores are common. Mineralogically and chemically the two groups of rocks show many similarities, and the photomicrograph of a leptite shown as Fig. 27 in the comprehensive description of the Central Sweden area by Magnusson (1970) could have been of a rock from Ulu Rompin - c.f. Fig. 9 in this thesis. With regard to the original nature of the Swedish rocks Magnusson says "The leptites .... were, to use international terms, rhyolitic lavas and corresponding volcanic sediments (tuffs and agglomerates).", which is in complete agreement with the writer's interpretation of the Ulu Rompin rocks, with the qualification that here the pyroclastic group appear to be more abundant than the lavas.

Petrography

The rocks are white or coloured in pale shades of grey and green, but can be deeply iron stained. Fig. 4 shows the rapid variation in colour the rocks can display. Quartz, accompanied by feldspar in some specimens, is usually visible to the naked eye. The rocks vary from massive to strongly foliated, and some of the tuffs have a laminated texture due to the parallelism of tabular fragments and bands of different grainsize. The near-surface rocks weather to a clay plus quartz, or form laterite. When fresh the rocks are hard, and have a glassy appearance on broken surfaces. The majority of the rocks have a simple mineralogy, many now consisting of quartz plus clay and sericite, but feldspars can be preserved. The ferromagnesian minerals are largely of secondary origin, and it is unlikely that the original rocks contained many dark minerals.

The bulk of the rocks are porphyritic, the ratio of phenocrysts to matrix varying from 5% to 50%. Some of the sparsely porphyritic types strongly resemble recrystallized sediments, and probably represent the intermingling of clastic and pyroclastic material. The groundmass has completely recrystallized to a granular mosaic and all fine detail has been destroyed. Pyritization has been widespread and calcitization can be strong, especially near the main ore bodies. In a few thin sections it is possible to detect sharp changes in grain size, which mark original depositional horizons. Many of the rocks are crystal tuffs, but lithic fragments up to 2 cm across have been seen.

Quartz.

Quartz is prominent as phenocrysts, and forms a large part of the groundmass. The phenocrysts range in size from 1 mm to 2 mm, and vary from perfect euhedra to almost spherical bodies or rounded fragments. Euhedral crystals often show the bipyramidal form of β quartz - Fig. 6, but short prismatic faces can also be developed. The quartz has often reacted with the groundmass and developed a crenulated edge; the reaction can be very extensive with deep embayments into the quartz. "Veining" by the groundmass has cut some quartz grains into isolated fragments - Fig. 7. Stress has produced an undulatory extinction, a good rhombohedral fracture pattern in some quartz grains - Fig. 8, and a few boehm lamellae. The quartz can occur as single homogeneous grains, but some have partically or completely recrystallized to a fine grained aggregate, with the original crystal outline clearly preserved - Fig. 9. This process often starts with the quartz fracturing into angular fragments controlled by the $\{10\bar{1}1\}$ direction, the fragments later recrystallizing independently in different orientation. Within one thin section it is possible to observe all stages in this recrystallization process. The phenocrystic quartz grains are rarely clear of inclusions, and many are crowded with them. Some inclusions appear as curving linear bands of minute cavities or dark bodies, some as recognizable mica crystals, particularly near the margins, and some as dendritic coatings of iron oxides or hydroxides following fracture planes. Reaction between the quartz and the groundmass can produce a peripheral zone full of minute embryonic crystals, largely micas, with a relatively clear core - Figs. 10 & 11.

The distribution of the phenocrysts can be even or extremely irregular, the latter being typical of the tuffs. The tuffs usually contain a proportion of broken grains, the angular corners having been rounded by reaction - Fig. 12.

In those rocks which have been subjected to strong stress there is some lensing of the quartz grains, but much of the detail is lost in the densely developed sericite. There is sometimes a suggestion that one side of a grain has been more reactive than

Fig. 6. Phenocryst of bi-pyramidal quartz which has been slightly veined and corroded by the groundmass. Crossed nicols x 35.

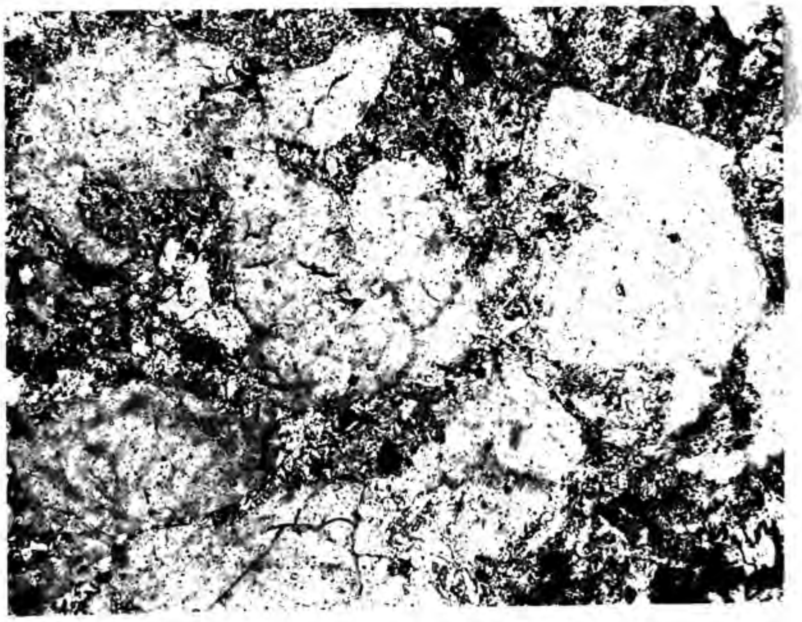


Fig. 7. Phenocryst of quartz which has been extensively replaced by groundmass material. Crossed nicols x 85.

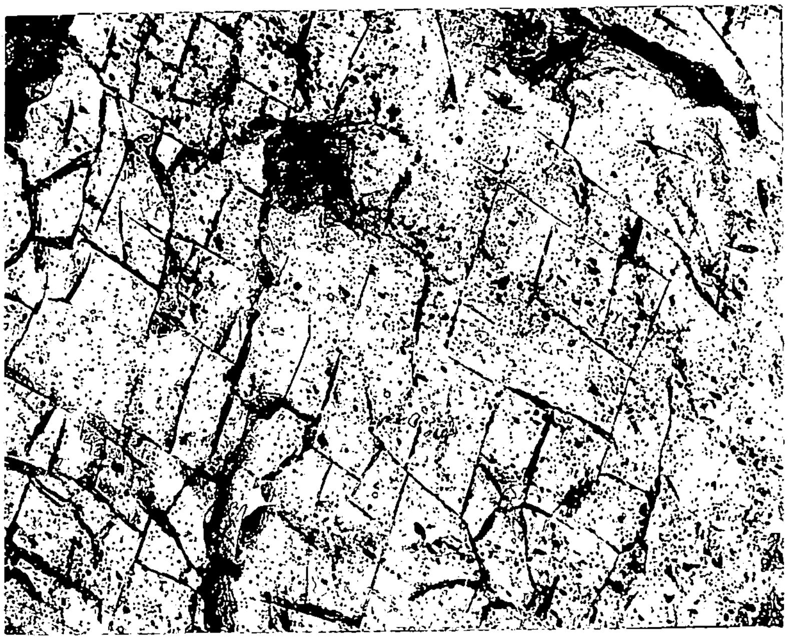


Fig. 8. Rhombohedral fracture pattern in a quartz phenocryst as a result of stress. Plain light x 90.

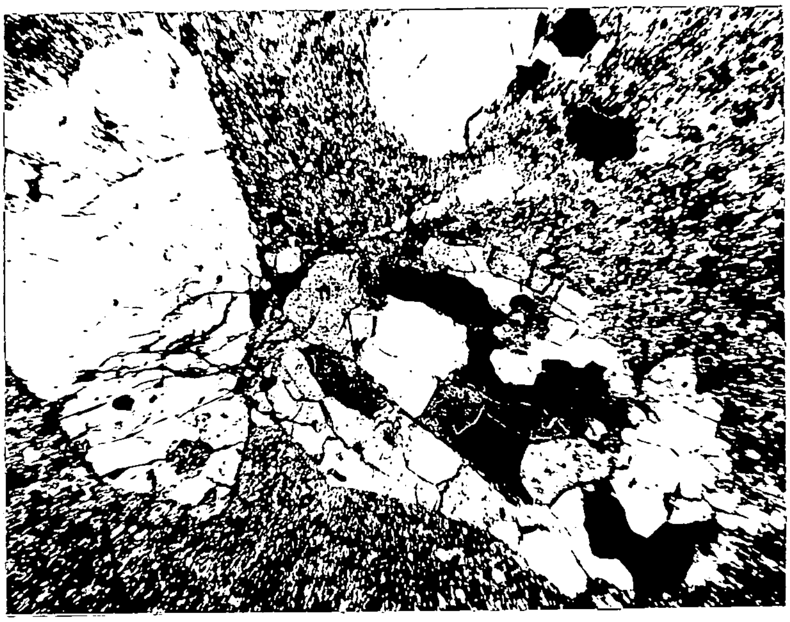


Fig. 9. Two quartz phenocrysts, one of which has recrystallized to a fine grained quartz aggregate. Crossed nicols x 45.

Fig. 10. Narrow reaction zone between a quartz phenocryst and the groundmass in an acid volcanic rock. Crossed nicols x 35.

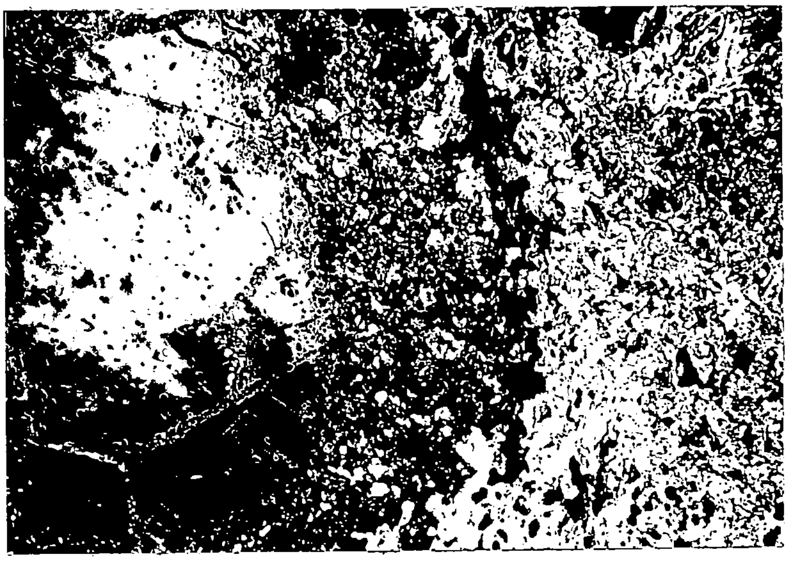


Fig. 11. Enlarged view of part of the reaction zone shown in Fig. 10. Quartz is on the left and the groundmass on the right. Crossed nicols x 180.

the opposite side, presumably due to a strongly directional stress, but the phenomenon is not marked.

Feldspars

Feldspars occur as phenocrysts and in the groundmass, but are usually difficult to identify precisely due to the high degree of alteration they have suffered. In rare specimens the feldspars form the sole phenocrysts, but they are usually accompanied by quartz. The phenocrysts vary in size from 0.5 mm to 4 mm, occurring singly and in small aggregates. They are never fresh, and show all stages of alteration from a cloudiness due to kaolinite to complete pseudomorphing by sericite. The feldspars are rarely corroded to the same extent as the quartz, but may have rounded outlines while still preserving the general crystal shape, broad columnar crystals or more equant grains. Twinning is not well developed and is usually diffuse. Untwinned or carlsbad-twinned orthoclase is the most common feldspar, but plagioclase with albite and pericline twinning does occur. Some diffraction patterns suggest the presence of sanidine, but it has not been identified microscopically. The plagioclase is richly sodic, varying from almost pure albite to oligoclase, Ab_{80}, but one unusual specimen contains andesine.

Several different types of alteration can be seen in the feldspars. Some are almost opaque due to the development of brownish kaolinite, through which the cleavage planes stand out as colourless lines, while others alter to sericite or coarser flakes of muscovite. An uncommon line of alteration shows feldspar recrystallizing to leave a granular quartz-feldspar mosaic, some of the alumina and the potash having been removed; the resulting intergrowth of quartz and unaltered feldspar can display a micrographic texture.

Mica & Chlorite

Sericite is ubiquitous in the volcanic rocks and in some specimens almost obliterates the other minerals. It partially or completely replaces some of the phenocrystic feldspars and is common in the groundmass, and is also found as veinlets. The sericite forms masses of minute flakes which usually have no preferred orientation, but in the more strongly sheared rocks it occurs in parallel bands. There are often a few coarser flakes of muscovite associated with the sericite. Biotite is rare, but forms the dominant mica in one or two specimens. It is found as fine to medium sized flakes which usually show some alteration to chlorite. Muscovite can also alter to chlorite. It is clear that pyrite has been introduced after the micas formed, as it can be seen moulded round some of the larger muscovite crystals.

Chlorite is present in many of the rocks in small amounts, and can be locally abundant. It has formed by alteration of micas and feldspar, and some has been introduced by metasomatism and in late calcite veins. Small flakes are often seen growing round the opaque iron minerals, and chlorite forms veinlets in the volcanic rocks round the Bt. Ibam ore body.

Clay

The groundmass of some of the rocks contains abundant very fine grained dusty material which cannot be identified optically, but diffractometer studies prove that this is kaolinite. There could also be some illite, but its diffraction pattern would be masked by sericite. Kaolinite pseudomorphs some of the phenocrystic feldspars forming dense brownish dusty masses. Kaolinite veinlets are common, and many of the joints in the volcanic rocks round the Bt. Ibam ore body are lined with pure white kaolinite.

Accessory Minerals

Epidote is absent from the majority of the rocks but it can be important, particularly close to the Bt. Ibam ore body where it has formed as a result of calcium metasomatism during the mineralization of the calcareous rocks. Some of the volcanic rocks which lie very close to the ore body are heavily impregnated with epidote, and as a result are pale lime green in colour. The epidote can be disseminated, but in some rocks it is associated with the opaque minerals. Epidote is found as veinlets along joints near the Bt. Ibam ore body.

Apatite has been identified in one or two thin sections as minute prismatic crystals in feldspar. Sphene is present in many of the rocks as small granules formed during the breakdown of biotite, or associated with the opaque minerals. Haematite is common in the rocks close to the primary ore bodies at Bt.

Pesagi, occurring as small plates, and is obviously related to the main haematite mineralization. Pyrite is found in the majority of the volcanic rocks as fine disseminated grains and occasional coarse crystals. It is normally anhedral but can develop its crystal shape. Pyrite forms numerous thin veinlets along joints, and small isolated lenses in the rocks near the Bt. Ibam ore body.

Groundmass

The groundmass has invariably recrystallized to a mosaic with the grain size varying from 0.05 mm to 0.5 mm, but within a single thin section the rock is usually roughly equigranular. All trace of any original texture - flow, glass, glass shards etc. - has been obliterated, but in one thin section the sericite flakes form a swirling pattern round the feldspar phenocrysts, which may be due to the preservation of an original flow texture. The groundmass quartz is accompanied by feldspars, clay and sericite. One sample from close to the Bt. Ibam granodiorite contained embryonic feldspar crystals which were growing as a result of metasomatism.

Lithic Fragments

Some of the rocks contain rock fragments, which have recrystallized to a very fine grained quartz mosaic and their margins may show signs of reaction, but they stand out clearly from the rest of the rock - Fig. 12. These fragments are valuable evidence for the pyroclastic origin of some of the volcanic rocks.

Veins

Quartz, calcite, kaolinite, siderite, zeolite, chlorite, epidote, pyrite, geothite and manganese oxides have all been found in veins in the volcanic rocks. Quartz is the most important and ranges from microscopic veinlets to veins and lenses several feet wide. Where there is a marked foliation the quartz usually follows it, and it also fills some joints. Calcite is found as veins of variable size, but often tends to impregnate irregular masses of the rocks. Siderite and zeolite have only been identified microscopically from one or two samples near the Bt. Ibam ore body.

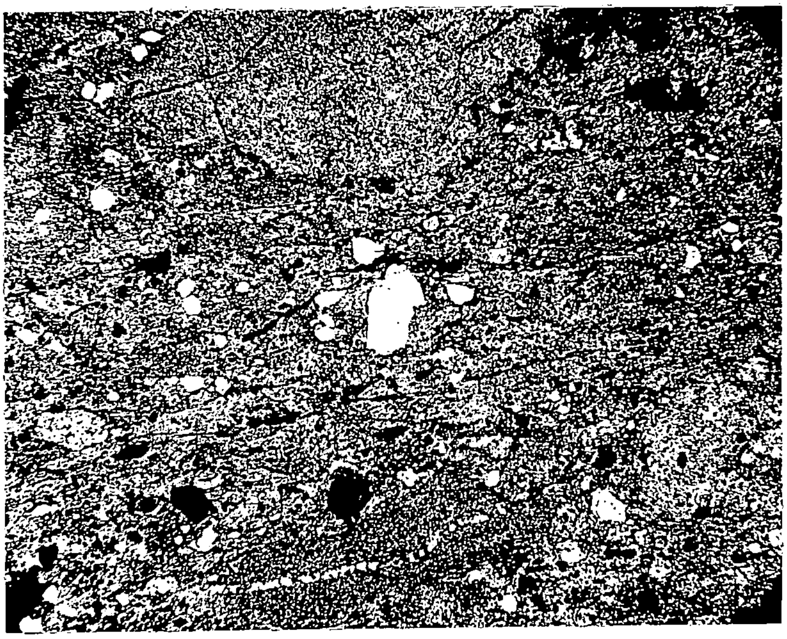


Fig. 12. Crystallo-lithic tuff with broken quartz grains and recrystallized lithic fragments. Crossed nicols x 5.

Diffraction Studies

A number of crushed volcanic rocks were examined by X-ray diffractometer to supplement the thin section studies, and the traces confirm that quartz and sericite are the most abundant minerals and kaolinite is locally important. A number of samples contain only quartz and sericite. Where feldspar patterns appear they are rarely easily identifiable as a particular species, but orthoclase and albite have been recognized, which agrees with the microscopic determinations. One or two traces suggest that sanidine is present, and although this mineral has not been identified optically its presence is compatible with the general petrology. Chlorite appears on a small number of traces.

Chemical Analyses

A number of samples of the volcanic rocks from Bt. Ibam and Bt. Pesagi were analysed by the writer by XRF methods, and the results are shown in Tables 1, 2 & 3 together with the norms.* The Bt. Ibam rocks are split into two groups, those from the footwall and those from the hangingwall, and are as far as possible listed in order of increasing depth from the surface. One or two samples are placed out of order as specimens from a single drill hole have been kept together. The details of the footwall specimens are given below.

No.	Depth (ft)	Remarks
162	127	400 ft from the ore body, just above granodiorite.
165	150	
197	198	45 ft from the ore body.
51	250	Immediately below skarn.
49	270	50 ft from ore body, just below skarn.
136	250	15 ft below skarn.
137	275	40 ft from ore body.
138	285	55 ft from ore body.
142	440	240 ft from ore body.

* Chemical analyses were made by the writer using XRF equipment in the department and methods described by Jugal (PhD thesis, 1969).

Norms are C.I.P.W. calculated as described by Johannsen, 1937, Vol. I. "Descriptive petrography of the igneous rocks".

No.	Depth (ft)	Remarks
278	318	North of Main ore body, below a thin ore lens.
262	450	Between Nos. 162/165 and the ore body.
263	475	
188	530	170 ft from ore body.

Details of the hangingwall samples are:

78	50	225 ft east of the ore body.
190	57	Close to massive chlorite rock.
145	140	235 ft from the ore body.
331	160	350 ft from the ore body.
62	230	25 ft from the hangingwall contact.
220	272	300 ft east of the ore body.
221	325	
224	480	

As can be seen from the above two lists, the samples come from a variety of depths and locations relative to the ore body, and they represent a good selection of the volcanic rocks round the Bt. Ibam ore body. The original rocks vary from those which are very soft due to weathering to others which are hard and ring when struck with a hammer.

Taking the analyses as they stand and without any qualifications the general characteristics of the Bt. Ibam rocks are seen to be the importance of the alkalies compared to lime, a variable iron and magnesia content, and high silica and alumina. The excess of alumina is reflected in the presence of normative corundum, often in appreciable amounts, in all but two of the samples. The high alumina is due to the presence of kaolinite in the rocks, but there is no consistent relationship between the alumina or normative corundum content and the depth of a sample. Much of the ferric iron is present in goethite but a little occurs in chlorite. Ferrous iron is mainly located in the pyrite but a little accompanies the magnesia. Chlorite is responsible for much of the magnesia, and there is evidence that where the

Table 1

Chemical Analyses of Footwall Volcanic Rocks at Bt.Ibam

	162	165	197	51	49	136	137	138	142	278	262	263	188
SiO_2	65.96	73.56	69.67	70.87	61.43	69.52	67.73	70.47	78.50	71.64	66.68	73.11	72.10
Al_2O_3	16.97	14.03	13.58	13.16	13.15	14.38	14.94	11.09	11.77	13.76	16.26	9.57	15.41
Fe_2O_3	1.83	0.57	1.17	0.52	3.21	1.52	2.13	0.29	0.11	0.88	1.72	1.25	1.02
FeO	1.35	1.82	1.59	2.65	4.57	0.97	3.09	4.86	0.65	1.18	1.97	1.63	0.67
MgO	2.50	0.73	4.75	2.02	7.56	0.78	2.06	2.60	1.40	2.70	3.24	3.33	1.25
CaO	0.09	0.29	0.63	0.84	0.50	1.27	0.80	0.83	0.47	1.09	1.30	3.08	1.40
Na_2O	1.76	3.11	1.04	3.84	0.08	2.22	1.35	2.36	0.55	0.47	2.25	2.69	0.04
K_2O	5.77	3.79	2.94	2.25	4.96	6.39	1.93	2.23	4.13	4.53	3.14	1.62	4.88
TiO_2	0.50	0.33	0.35	0.68	0.30	0.50	0.54	0.62	0.07	0.05	0.45	0.31	0.12
MnO	0.09	0.12	0.06	0.06	0.22	0.04	0.17	0.11	0.05	0.05	0.06	0.08	0.04
S	0.66	0.17	0.88	1.07	0.34	0.90	0.91	2.13	0.46	0.88	0.40	1.30	0.51
H_2O	2.52	1.48	1.59	2.04	3.68	1.51	4.35	2.41	1.84	2.80	2.53	2.03	2.56
q	28.4	36.5	42.3	32.6	27.3	27.1	46.5	38.5	54.7	44.0	31.8	38.9	48.0
or	35.5	23.0	18.0	13.5	30.5	38.5	12.0	13.5	25.5	28.0	19.0	9.5	30.0
ab	16.0	28.5	10.0	35.5	-	20.0	13.0	22.0	5.5	4.0	20.5	25.0	-
an	.5	1.5	3.0	4.5	2.5	6.5	4.0	4.0	2.5	6.0	6.5	10.0	7.5
c	8.8	4.9	8.6	3.0	8.0	1.9	10.8	3.8	6.3	6.9	8.2	-	8.8
en	7.0	2.0	13.8	5.6	22.2	2.2	6.0	7.4	4.0	7.8	9.2	4.8	3.6
fs	-	2.6	.6	1.0	4.8	-	2.8	4.8	-	-	1.8	-	-
wo	-	-	-	-	-	-	-	-	-	-	-	2.2	-
il	-	-	-	1.0	.4	-	.2	-	-	-	-	-	-
mt	2.0	.6	1.3	.6	3.4	.9	2.2	.3	.2	.6	2.0	1.2	.3
hm	-	-	-	-	-	.4	-	-	-	.3	-	.2	.5
pr	1.8	.4	2.4	2.7	.9	2.5	2.5	5.7	1.3	2.4	1.0	3.4	1.3

Table 2

Chemical Analyses of Hangingwall Volcanic Rocks at Bt. Ibam

	78	190	145	331	62	220	221	224
SiO_2	69.67	58.00	72.79	73.06	74.03	72.31	73.82	75.35
Al_2O_3	21.37	15.96	15.20	17.85	14.71	12.73	13.89	13.31
Fe_2O_3	0.56	2.18	0.92	0.40	0.75	6.73	1.53	0.80
FeO	0.03	2.95	0.43	0.38	0.63	0.56	0.68	0.78
MgO	1.12	4.62	1.43	1.00	1.11	0.69	1.43	1.55
CaO	n.d.	5.33	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Na_2O	0.07	2.52	0.16	0.15	3.75	0.09	0.15	0.12
K_2O	2.53	2.52	7.03	3.19	2.78	2.81	5.70	4.49
TiO_2	0.09	0.80	0.08	0.14	0.07	0.04	0.05	0.05
MnO	n.d.	0.15	0.02	0.01	0.01	0.02	0.02	0.03
S	0.05	2.25	0.08	0.04	0.51	0.08	0.09	0.29
H_2O	4.51	2.72	1.86	3.78	1.65	3.94	2.64	3.23

q	57.9	15.1	41.7	57.4	37.9	61.3	48.0	54.2
or	16.0	15.0	43.0	20.5	16.5	18.5	36.0	28.5
ab	.5	22.5	1.5	1.5	34.0	1.0	1.5	1.5
an	–	25.5	–	–	–	–	–	–
c	21.7	–	8.5	16.4	6.3	11.4	8.6	9.5
en	3.4	13.0	4.2	3.0	3.2	2.0	4.2	4.6
fs	–	.2	–	.6	–	–	–	–
wo	–	.4	–	–	–	–	–	–
il	–	–	–	–	–	–	–	–
mt	.3	2.4	.6	.6	.3	1.2	1.5	.9
hm	.2	–	.3	–	.3	4.4	–	–
pr	–	5.9	.2	–	1.5	.2	.2	.8

magnesia is significantly high some has been added by metasomatism. A little talc is present in some of the rocks. Lime is usually very low and frequently absent; the rocks which are rich in lime contain epidote, and a little lime is present in plagioclase. Some lime has undoubtedly been added to a small number of the rocks, probably at the time of the main iron mineralization. The alkalies are important constituents with potash usually predominating over soda, but the latter exceeds or equals potash in a few rocks. Potash is present in sericite and orthoclase while the soda is contained in sodic plagioclase. The sulphur content is a reflection of the degree of pyritization the rocks have undergone. The mineralogy of the rocks has undergone a distinct change due to alteration, but it is thought that the quoted analyses give a reasonable indication of the original composition. The excess of alumina shows that some other components have been removed, and petrographic evidence suggests that it is part of the alkalies which has been taken out during the breakdown of the feldspars.

The one rock which is significantly different from the rest is No. 190, with its high lime and low silica. The mineralogy of this specimen is unlike any of the other volcanic rocks examined by the writer in that it contains amphibole; it is rich in epidote and pyrite. An X-ray diffractometer trace showed that the feldspar is andesine, and there is no doubt that this rock is much more basic than the usual type, although some of the lime and magnesia are metasomatic in origin. The rock is a porphyritic lava and probably represents an original andesite.

All the rocks save No. 190 are clearly acidic in composition and are alkali rhyolites, dominantly of the potassic variety. The alkalinity is emphasized in Fig. 13 which shows the normative feldspar components plotted onto a triangular diagram. There is a cluster of samples near the or pole (with one exception these are hangingwall rocks) while the remainder show a reasonable spread between or and up to 70% ab; the an is usually low. The positions of the samples would have been the same if cationic proportions instead of normative minerals had been plotted, except for the two samples with the highest an content, and even these would have moved little.

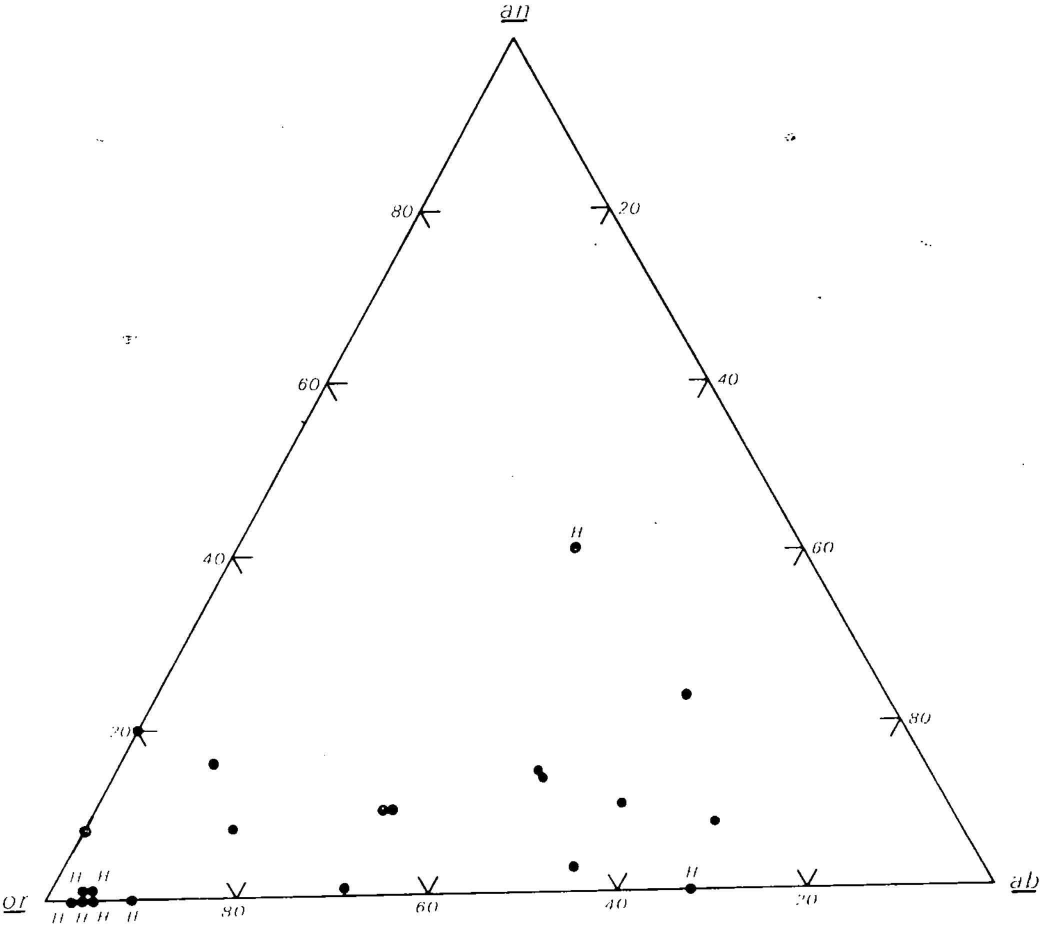


Fig. 13 Triangular diagram showing the normative feldspar in the volcanic rocks at Bt. Ibam.
H = hangingwall rocks.

The analysis of a sheared rhyolitic tuff from the Grik area of northern Malaya quoted in Jones (1970) is similar to those fo the typical Ulu Rompin rocks, with high potash and excess alumina. The Grik sample shows 10% normative corundum. The analyses of the Ulu Rompin rocks bear some strong similarities to those of the leptites from Central Sweden quoted in Magnusson (1970), and as a whole come into the alkali rhyolite group or clan.

Ignoring sample No. 190, which is quite different from the predominant rock type, there are some significant chemical dissimilarities between the rocks from the footwall and hangingwall sides, which suggests thattwo groups of rocks are involved, although their gross features are alike and such a division cannot be made on petrographical evidence. The main point of difference is the lime content; in the footwall it is present in all samples and averages 0.97% whereas it is virtually absent from the hangingwall rocks. The hangingwall rocks contain more silica and alumina but less magnesia, soda and titania, the latter showing a fivefold drop. There is a distinct difference in the degree of pyritization in the two groups of rocks, markedly so if No. 190 is excluded. The averages for a number of oxides in both groups of rocks are shown below, but the figures for iron are not included as they are greatly influenced by the development of pyrite and goethite.

	Footwall	Hangingwall
SiO_2	70.10	73.00
Al_2O_3	13.70	15.58
MgO	2.68	1.19
CaO	0.97	nil
Na_2O	1.67	0.64
K_2O	3.74	4.07
TiO_2	0.38	0.07
S	0.81	0.17

The other analysed volcanic rocks came from Bt. Pesagi and Bt. Sanlong, and the details are given below.

No.	Depth (ft)	Remarks
170	-	Bench at the southern end of Pesagi workings.
172	-	Bench at the northern end of Pesagi 6 workings.
286	87	Near Sanlong B ore body.
288	184	
300	200	Below Pesagi 6 ore body.
304	166	Short distance east of Pesagi 6 ore body.

The analyses are shown in Table 3 from which it can be seen that the chemistry of these rocks is extremely uniform, two samples having reduced silica but enhanced alumina. The rocks bear a closer resemblance to the hangingwall group at Bt. Ibam than to those of the footwall, the most striking feature being the complete absence of lime. The rocks are distinctly potassic and soda is subordinate. The magnesia is low, and below the level of the Bt. Ibam rocks, but the titania is higher. All the rocks show a high level of normative corundum, which is some indication of the alteration they have suffered. The averages for the main components are shown below.

SiO_2	70.40
Al_2O_3	18.19
MgO	0.63
CaO	nil
Na_2O	0.19
K_2O	3.94
TiO_2	0.60
S	0.17

Fig. 14 is an interesting triangular diagram showing all the twenty seven analyses of the volcanic rocks made by the writer. There are three "erratic" samples, but the rest fall into three well defined zones, and although these are close together they show that on chemical evidence one can usually differentiate

Table 3

Chemical Analyses of Volcanic Rocks at Bt. Pesagi & Bt. Sanlong

	170	172	286	288	300	304
SiO_2	71.52	61.56	75.92	73.05	73.55	66.89
Al_2O_3	17.90	22.32	16.23	17.22	14.76	20.62
Fe_2O_3	1.65	3.88	1.15	1.29	1.72	3.64
FeO	0.51	0.75	0.47	0.55	0.81	0.51
MgO	0.68	0.99	0.49	0.43	0.64	0.57
CaO	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Na_2O	0.25	0.30	n.d.	0.12	0.25	0.23
K_2O	3.18	5.97	3.69	3.90	3.85	3.07
TiO_2	0.32	0.96	0.20	0.60	0.67	0.80
MnO	0.02	0.02	0.01	0.02	0.01	0.01
S	0.05	0.02	n.d.	0.17	0.70	0.06
H_2O	3.92	3.23	1.84	2.65	3.04	3.60

Norms

	170	172	286	288	300	304
q	56.4	34.3	59.5	56.2	56.0	53.1
or	21.0	37.5	22.0	24.0	24.0	20.0
ab	2.5	3.0	-	1.0	2.5	2.0
an	-	-	-	-	-	-
c	16.3	17.5	14.0	14.5	11.9	19.5
en	2.0	3.0	1.4	1.2	1.8	1.6
fs	-	-	.4	.8	-	-
il	.4	1.4	.4	1.0	-	.8
mt	.6	1.5	1.3	.9	-	-
hm	.8	1.8	-	-	1.3	2.8
pr	-	-	-	.4	2.0	-
ru	-	-	-	-	.5	.2

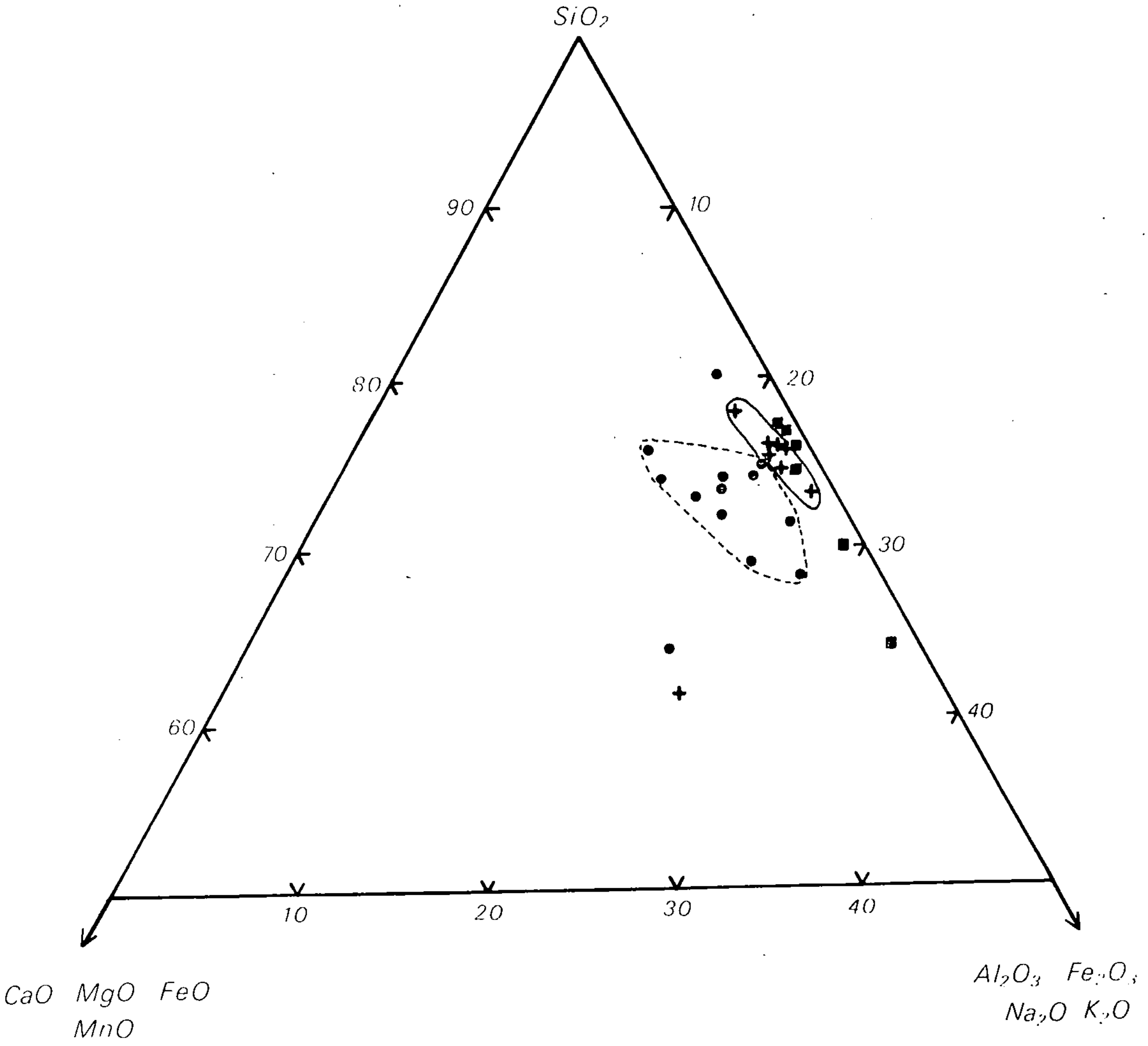


Fig. 14 Triangular diagram for the volcanic rocks at Bt. Ibam & Bt. Pesagi.

Bt. Pesagi
■

	footwall	hangingwall
Bt. Ibam	●	+

hangingwall from footwall rocks at Bt. Ibam. It also demonstrates that the Bt. Pesagi rocks are similar to the hangingwall rocks at Bt. Ibam. The Bt. Ibam footwall group shows a small spread of values, but the hangingwall and Bt. Pesagi samples have an almost linear distribution based on an inverse relationship between silica and alkalies plus alumina plus ferric iron.

Metamorphism Of the Volcanic Rocks

The acid volcanic rocks show little sign of the effects of rising temperature, but in many areas they have been strongly affected by stress. The main mineralogical changes have resulted from the breakdown of some minerals to sericite and kaolinite, and in a few samples from the introduction of lime and magnesia. Reaction between the original components in response to rising temperature seems to have been slight or the evidence has been obliterated, but there has been widespread recrystallization of both the matrix and phenocrysts.

The banded rocks have a schistose appearance but are not true schists as there is no megascopic development of mica or other sheet silicates along the foliation planes. The rocks are cataclasites and can be classed as semischists, using the term as described in Williams, Turner and Gilbert (1954). The foliation is coincident over a wide area and must have developed from uniform stress conditions. The regularity of the foliation in one direction suggests that the causative agent was a simple shearing stress and not a pure stress.

IV SEDIMENTARY ROCKS AND SKARNS

Introduction

There are minor intercalations of sandstone and siltstone, or their metamorphic equivalents, within the acid volcanic rocks in the area from Bt. Pesagi to Bt. Ibam, and they become more important towards the northwest corner of the mineralized zone. The only rocks which show a marked contrast to the general lithology are the calcareous lenses found at Bt. Ibam and Batu Puteh. There may also have been some calcareous rocks at Bt. Pesagi, Bt. Merah, Bt. Sanam and Bt. Hitam, but the evidence is not conclusive.

Arenaceous Rocks

Apart from the skarns, the clearest example of an original sedimentary rock came from a depth of 310 ft in the hangingwall rocks at Bt. Ibam, along section N. 10,400. Although the sample must come from a very small lens, as the same rock has not been encountered in any of the adjacent drill holes, it offers some interesting petrological evidence. The rock is an impure fine grained metaquartzite, in which the interlocking granular texture is clearly seen in plain light as limonite has stained the intergrain boundaries. When the nicols are crossed, a quartz mosaic which bears no relation to the texture seen in plain light becomes visible - Figs. 15 & 16. This thin section provides evidence of two separate phases of recrystallization; a first phase which is now preserved in the skeletal limonite framework, and a second phase which produced the granular mosaic seen under crossed nicols. The surrounding volcanic rocks must have been subjected to the same two metamorphic phases, but the individual phases cannot be differentiated in these rocks.

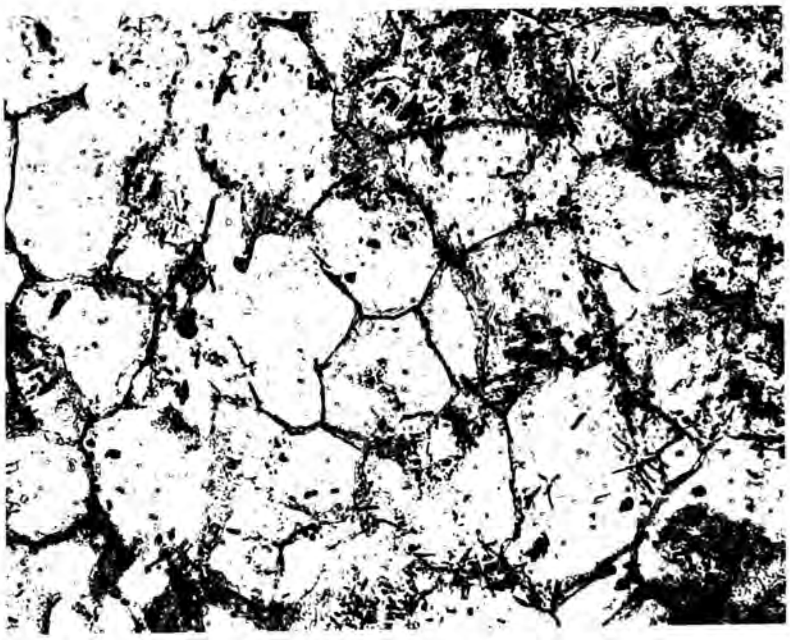


Fig. 15. Fine grained metaquartzite with the intergranular boundaries delineated by goethite. Note a roughly hexagonal grain towards the centre of the photograph. Plain light x 210.

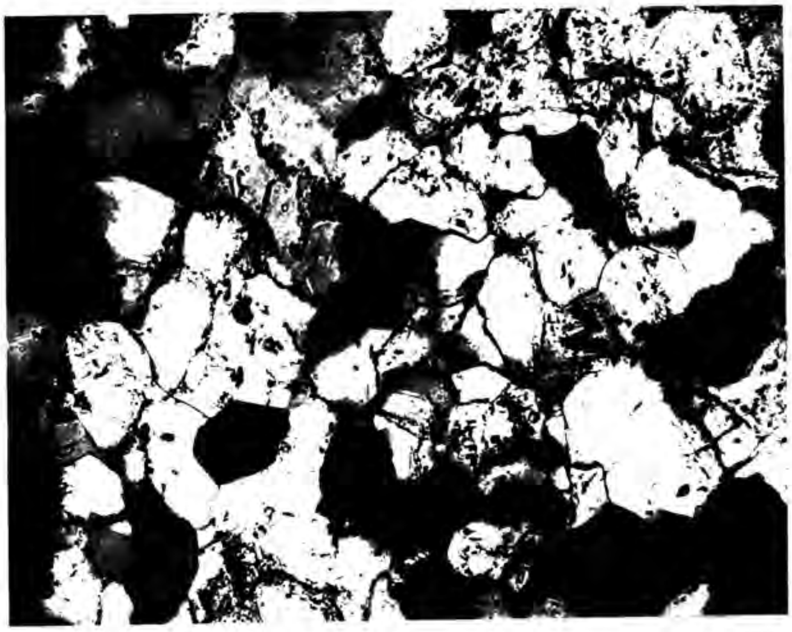


Fig. 16. Same area as Fig. 15 but with the nicols crossed. Note that the texture is different to that seen in plain light. The hexagonal grain towards the centre is clearly made up of smaller grains in different orientations. Crossed nicols.

An analysis of the metaquartzite is shown in Table 4.

Table 4

Chemical Analysis of Impure Metaquartzite
from Bt. Ibam

SiO_2	91.16
Al_2O_3	nd
Fe_2O_3	3.40
FeO	0.25
MgO	2.90
CaO	nd
Na_2O	0.04
K_2O	nd
TiO_2	nd
MnO	1.39
S	0.03
H_2O	0.83

Except for the magnesia content the composition of the metaquartzite is compatible with the evidence as seen in thin section. The iron and manganese are present in the intergranular goethitic material; the magnesia is likely to be present in very fine dusty talc, although none has been identified under the microscope. An X-ray diffractometer trace revealed only a quartz pattern.

Marble

Drilling on the Bt. Batu Puteh deposit has proved the presence of medium to coarse grained crystalline marble, which is white, pink, grey or iron stained brown and reddish-brown. The full dimensions and shape of the marble are unknown, but one drill hole was still in marble after an intersection of 360 ft, while other drill holes indicate a much smaller thickness. No calc-silicate or other metamorphic minerals have been detected in the Bt. Batu Puteh deposit. The country rocks are highly decomposed acid volcanics.

SKARNS

Introduction

These rocks are developed along much of the footwall of Bt. Ibam ore body - Figs. 25 & 31, and similar calc-silicate rock has been seen in core from one drill hole at Bt. Pesagi. Skarn was also located in drill core from a depth of 420 ft in the footwall along section N. 10,400 at Bt. Ibam. The rock lies 220 ft below the main skarn, and it must represent a very small lens as it has not been located in other drill holes. A company geologist has reported skarn from Bt. Hitam, but the writer is dubious about the evidence produced to support this conlusion.

The Bt. Ibam skarns, shown in Figs. 17 & 18, have a maximum thickness of 140 ft and are well bedded, and although slightly folded and faulted are conformable with the ore body; their relationship to the underlying volcanic rocks is less clear. In some parts the contact is faulted, but elsewhere there can be a transition zone, only a few inches wide in some cases, in which the volcanic rocks are impregnated with calcareous minerals, notably epidote, but clearly preserve their original nature. The writer believes that there was a normal sedimentary contact with calcareous rocks on top of tuffs and lavas, and there has been lime metasomatism into the volcanic rocks. The skarns are cut by numerous small faults, most of which are narrow, but examples up to 1 ft wide with a good fault breccia have been observed. The breccias are usually cemented by quartz and/or goethite.

Petrography

The rocks are coloured in various shades of green and brownish-buff, and can be banded in these colours. The green is largely due to epidote with some chlorite, actinolite and mica, while the brownish rocks are rich in garnet. One good specimen showed alternating green and brown bands from $\frac{1}{4}$" to $\frac{1}{2}$" wide, composed of almost solid epidote and garnet respectively. The banding follows the original bedding planes. In this specimen

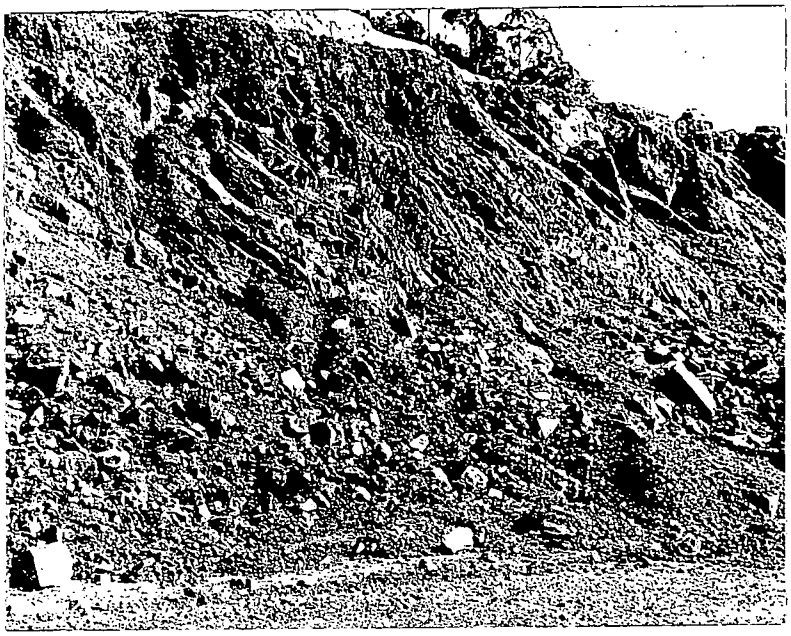


Fig. 17. Bedded skarns dipping east at about 45°. The more massive bed which can be seen at the extreme right of the photograph at the top of the face is the lowest part of the ore body.

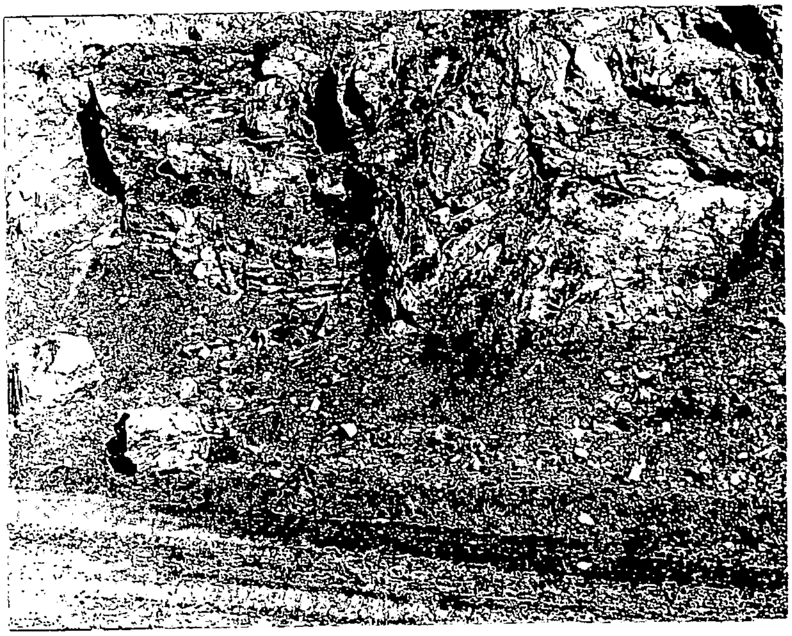


Fig. 18. Bedded skarns which are slightly flexured and dipping east at a gentle angle. A small vertical fault runs through the centre of the outcrop.

there are also rare bands of megascopic garnet crystals set at an angle of about 45° to the main bedding, and which probably represent crystallization in open spaces along small fractures. The rocks which are immediately adjacent to the footwall of the ore body contain a high proportion of chlorite and fibrous amphibole, some bands containing only these minerals, and must approximate closely to the original composition of the gangue in the overlying ore body. The typical skarn is fine to medium grained, with individual crystals often visible to the naked eye although rarely exceeding 1 mm.

The skarns have been affected by oxide and sulphide mineralization, but it is haphazard in its distribution. Magnetite is only found in the magnesian rocks close to the footwall of the ore body as disseminations of small grains, roughly following the bedding directions. The main ore minerals are pyrite and micaceous haematite, which are usually accompanied by quartz and sometimes epidote. The mineralization with pyrite and haematite is found in isolated and irregular pockets up to 5 ft across. It is interesting to note that micaceous haematite is absent in the main ore body just above the skarns.

The most abundant minerals are epidote and garnet, but the rocks also contain chlorite, fibrous amphibole, pyroxene, mica, very rare olivine, quartz and calcite, together with some talc formed by alteration. Thd texture is fine to medium granoblastic, but garnet can form small porphyroblasts.

Epidote

Epidote is the most abundant mineral and usually occurs as anhedral granular aggregates, but some thin sections contain very good euhedral crystals - Fig. 19. The epidote varies from deep pistachio green to colourless, the former being strongly pleochroic; the colour can be irregularly or zonally distributed within a single grain. There is a wide variation in the interference colours displayed, the colourless variety often showing the anomalous colours of clinozoisite, and the zonal texture shows that there must have been some fluctuation in composition during crystallization. The typical epidote-rich rock consists of granular anhedral masses, with the grains varying from 0.1 mm to 0.5 mm. Coarser varieties

Fig. 19. A good example of a twinned and zoned euhedral epidote crystal from the footwall skarns. Crossed nicols x 40.

show subhedral and rarer euhedral crystals up to 2 mm. The well formed epidotes show the pseudohexagonal cross section, or occur as columnar crystals with marked $\{110\}$ terminations, and some that appear to be $\{210\}$. The $\{010\}$ face is supressed. Twinning is not common but when observed is on the $\{100\}$ direction. Zoning rarely shows in plane light, but under crossed nicols some striking examples of rhythmically zoned crystals can be seen - Fig. 19. The granular aggregates usually contain a little interstitial quartz and calcite, and patches of garnet, pyroxene and amphibole. One sample contains a little forsterite.

Epidote can occur as a subordinate mineral in the massive garnet rocks, and has probably formed both before and after the garnet. Small internal relics of epidote in garnet appear to be remnants of larger epidotes which have been largely replaced by garnet, but some peripheral patches of epidote have formed by alteration of earlier garnet. Epidote occupies a central position between two ages of amphibole formation, but is roughly contemporaneous with the pyroxene. Much of the epidote has been slightly altered, but a few specimens show only isolated relics of the mineral in a mass of fibrous amphibole, and it also alters to chlorite. Epidote has clearly crystallized before some of the ore minerals, as they can be seen moulded round euhedral crystals.

Garnet

There is a wide variation in the mode of occurrence of garnet, from isolated euhedral porphyroblasts to massive granular aggregates, and in colour from colourless to pinkish-brown, yellow and yellowish-green. The latter variety looks very like epidote in plane light. The garnet varies in size from 0.005 mm to 2 mm, and is largely anhedral or subhedral. The crystals are all dodecahedral $\{110\}$ and can show a very strong zonal texture. Many grains are slightly anisotropic, and rare examples show interference colours up to first order orange-purple. In zoned crystals the paler coloured zones show the higher anomalous birefringence. The wide variation in colour suggests that there is some variation in composition, and this is confirmed by diffraction studies. Three samples, one from isolated porphyroblasts (No. 45), one from massive granular garnet (No. 52), and one of yellowish-brown garnet (No. 134), were powder photographed and their cell sizes calculated. The cell sizes and the seven strongest lines are shown in Table 5.

Table 5

Cell Size and Seven Strongest Lines
Bt. Ibam Garnets

45 dÅ	52 dÅ	134 dÅ	134 Rel. I
3.012	3.004	3.010	6
2.697	2.685	2.690	10
2.460	2.451	2.457	6
1.953	1.948	1.952	3
1.672	1.668	1.669	3
1.609	1.606	1.609	9
0.819	0.819	0.820	3
a	a	a	
12.043	12.042	12.054	

The size of the cell edge shows that the mineral is an andradite, the typical garnet found in skarns associated with iron ores. The cell edge for synthetic andradite is given as 12.048Å in Deer, Howie & Züssman vol. 1 (1962). The cell edge for No. 134 is slightly larger than this figure but coincides with the 12.054 ± 0.003Å for topazolite, a honey-yellow andradite, quoted by Donnay and Nowacki (1954).

Some of the garnet is clear, but examples full of inclusions are not uncommon. Particles of quartz and epidote are present in the garnet, and one sample contained thin plates of a blood red mineral, probably haematite. Much of the garnet is fresh, but it can show alteration to chlorite, epidote and fibrous amphibole, as well as replacement by carbonates. Alteration to amphibole begins with the peripheral development of dense, translucent greyish-white patches with a fine fibrous texture, which later crystallize into coarser grains of tremolite-actinolite. This alteration often results in the release of fine iron ore. A number of samples rich in garnet show evidence of fine fracturing and impregnation by calcite and, rarely, siderite. Replacement by calcite releases masses of dusty iron ore.

Amphibole

The amphibole is tremolite-actinolite, and it is most abundant immediately adjacent to the footwall of the ore body, where the rocks can have a visible fibrous texture. There are some cross-cutting veinlets of asbestiform amphibole with fibres up to 5 mm long. Some of the amphibole has formed during the early stages of metamorphism and some has formed by alteration of epidote and garnet.

The tremolite-actinolite is typically seen as fibres, but rather stubby prismatic crystals with a fibrous texture also occur. It is colourless to medium green, the colour often being patchily distributed. Iron ore is released when the amphibole replaces garnet, and iron ore plus kaolinite when it replaces epidote.

Tremolite-actinolite alters to chlorite with a little serpentine, and some thin sections show the development of talc. The latter is usually formed when there has been impregnation by calcite. There are a few patches of what may be anthophyllite, but its identity has not been confirmed.

Mica

Micas are not common, but phlogopite has been identified in a few specimens. One unusual sample showed coarse porphyroblasts of phlogopite, up to 1 cm across, set in a fine grained matrix of fibrous amphibole. The porphyroblasts appear as lustrous green crystals, the colour being due to the partial alteration to chlorite. It is not clear why this phlogopite forms such large crystals, as the micas do not normally develop as porphyroblasts. Phlogopite has also been seen as ragged-ended columnar crystals or aggregates associated with massive granular garnet, and as masses with a decussate texture in which fine disseminated magnetite is found. The phlogopite varies in colour from very pale green to pale brown with many almost colourless. The coloured variety is weakly pleochroic. It is clear that the magnetite is later than the mica as it grows round some of the columnar crystals, while the larger magnetites are poikiloblastic towards the mica.

Pyroxene

There are occasional grains of pyroxene associated with the granular epidote masses, but it is a very rare mineral in the Bt. Ibam skarn. One isolated sample from Bt. Pesagi consisted of a granular pyroxene aggregate. The pyroxene is colourless with a distinctly high relief, and is hedenbergite or an iron-rich diopside.

Quartz & Carbonate

Quartz is found as recrystallized interstitial patches or inclusions in the garnet and epidote-rich rocks, and is present as veins and irregular patches. The latter, which reach 5 ft in size, are often associated with pyrite and micaceous haematite. Calcite occurs as veinlets and impregnations, and represents a very late stage in the hydrothermal activity. It has sometimes followed along quartz veins, replacing the original quartz, and it has also impregnated patches within the skarns, replacing all the silicate minerals. Siderite is found as small veinlets and impregnations, but is uncommon.

Chlorite

Much of the chlorite has formed by alteration of other minerals, but it has also crystallized during the early stages of metamorphism, only to be replaced by other silicates as the temperature increased. Chlorite is usually a subordinate mineral, but it is abundant in some patches close to the footwall of the ore body.

Chemical Composition & Origin

Two samples from Bt. Ibam and one from Bt. Pesagi were analysed, the results being shown in Table 6.

The Bt. Ibam rocks largely consist of calcium-iron silicates, but in the Bt. Pesagi sample magnesia is more important, replacing ferric iron and lime. There are no unaltered relics of the rocks which originally filled the space now occupied by the Bt. Ibam skarns and the ore body, but they were obviously carbonates. They were highly reactive compared with the enclosing volcanic rocks, and as lime (in the skarns) and magnesia (in the ore body) are both abundant, the rock must have been a dolomite or dolomitic

Table 6

Chemical Analyses of Skarns

	Bt. Ibam		Bt. Pesagi
	52	141	308
SiO_2	46.01	52.50	46.83
Al_2O_3	5.21	3.69	8.33
Fe_2O_3	15.83	12.30	5.74
FeO	3.26	0.93	5.52
MgO	5.18	1.71	14.02
CaO	20.42	23.24	12.04
Na_2O	n.d.	0.04	0.17
K_2O	n.d.	n.d.	0.02
TiO_2	0.13	0.05	0.13
MnO	1.17	0.97	1.46
CO_2	1.30	3.20	2.38
S	0.11	0.45	1.02
H_2O	1.38	0.92	2.34

52 is from a bench in the main Bt. Ibam pit.
141 is drill core from a depth of 493 ft.
308 is drill core from a depth of 156 ft.

limestone. The skarns have formed from a calcareous horizon low in magnesia, or the magnesia could have been expelled during the metamorphism of a dolomite. There is no evidence of the latter, and it is more probable that there was a limestone horizon in an otherwise dolomitic body. The only major metasomatic additions to the skarns have been silica and iron.

The Bt. Pesagi skarn is found over a short interval in one drill hole only, which passes downwards through a magnesia-rich hybrid into quartz diorite. There is no lateral extension to the skarn and it may represent a calcareous inclusion which was caught up in the quartz diorite magma, although it is possible that there was an extremely small calcareous body which has been largely digested by the intrusive rock. What is certain is that there was no large body of calcareous rock in the Bt. Pesagi area.

V INTRUSIVE IGNEOUS ROCKS

Introduction

There are several small stocks of granodiorite in the Ulu Rompin area, and it is probable that in depth they merge into one larger parental mass. Fig. 3 shows the location of these small bodies. The plutonic rocks are very poorly exposed, and some which do not reach the surface are known only from drill core or adits. As the information regarding the boundaries of the bodies is often poor their precise shapes and dimensions are unknown, but the overall increase in size from south to north, which is clearly seen on Fig. 3, suggests that any underlying parental mass becomes deeper towards the south. In addition to those stocks known in outcrop or from subsurface data, there are several localities throughout the Bt. Sembilan ridge where accumulations of boulders prove that plutonic rocks must occur, although they have no in situ surface expression. From south to north the important plutonic bodies are:

evidence?

Bt. Pesagi.	Two small bodies have been located in drill holes; one is immediately south of the main primary ore body, and the other along the eastern margin of the mining area. In both bodies the maximum dimension is under half a mile.
Bt. Sanlong.	Just to the north of the primary ore body there are accumulations of boulders and rare outcrops in the headwaters of the Sg. Sanlong.
Bt. Ibam & Bt. Mungus.	Adits and drill holes have proved the presence of plutonic rocks below the ore body in the northern half of the mine, while in the Sg. Mungus just to the west of the ore body there were rare outcrops of granitic rocks in an area which is now flooded. There are also a few granitic dykes in the sheared country rock. The full dimensions of the Bt. Ibam intrusive body are unknown.

Bt. Sanam.	Accumulations of boulders, and evidence from drill core, prove the presence of a stock, which appears to trend in a northeasterly direction into the low ground in the Sg. Tepisoh, where one drill hole encountered plutonic rock.
Bt. Hitam.	Drilling has proved the existence of a body which is almost certainly continuous subsurface with that at Bt. Sanam, and also with the larger body of Bt. Chepemak 664 ft, which lies just outside the northeast corner of the mineralized zone proper. The Bt. Chepemak body covers an area of approximately three square miles, and is the largest of the stocks.

The granodiorite is intrusive into the acid volcanic and associated rocks, but the nature of the contact is known only in the Bt. Ibam - Bt. Mungus area. In the stretch of the Sg. Mungus which is now flooded there were exposures which showed granodiorite intrusive into sheared volcanic rocks. The contact was extremely sharp and followed the general direction of the foliation in the volcanic rock, but as the contact was exposed over a distance of a few feet only it is impossible to say whether this was a general feature or merely of local significance. The small intrusive bodies found along the Bt. Sembilan ridge could be interpreted as following two parallel lines aligned roughly with the foliation, but the larger masses between Bt. Sanam and Bt. Chepemak cut across the regional trend. The regional foliation was developed before the emplacement of the intrusive rocks and could have exerted some control, but there is no evidence that the bodies are in any way tabular in form. One drill hole into the footwall at Bt. Ibam shows granodiorite in contact with porphyritic volcanic rock. There is a sharp change from one rock type to the other, and what appears to be the whole of the contact zone is contained within one thin section. There is evidence of intermingling of material, and there has been some felspathization of the volcanic rock. Where the granodiorite has encountered calcareous rocks there has been mobilization of calcite, which now impregnates the surrounding rocks, and calc-silicate minerals have formed. The granodiorite encountered in Adit 12 is noteworthy for the presence of patches of solid epidote, probably derived from calcareous fragments caught up in the magma.

The smaller intrusive bodies tend to form slightly lower ground than the surrounding country rock, which suggests that there has been some hardening of the rocks adjacent to the intrusions by baking or silicification or both, which has increased their resistance to erosion.

There is a strong spatial relationship between the presence of intrusive stocks and the primary ore deposits, which the writer believes is also a genetic association.

Most of the acid plutonic masses in Malaya appear to be fairly high level intrusions, and in this respect the Ulu Rompin granodiorites are typical. Buddington (1959) defined three levels of granite emplacement (Epizone, Mesozone and Catazone) and gave depth estimates for them. He also listed a number of factors which he believes are indicative of a particular level of emplacement, and the Ulu Rompin rocks fit into the Epizone with a few Mesozone features. There are no features of Catazone bodies. Buddington states that the Epizone extends from surface to about six miles, but the intrusions are typically emplaced at about four miles.

Petrography

In colour the rocks vary from greyish-white to dark greyish-green, sometimes with a tinge of pink, and the grain size varies from fine-medium to medium-coarse. The average grain size is about 1 to 2 mm, but some rare grains reach 1 cm. The texture varies from equigranular to inequigranular, but is never porphyritic. In the coarser grained rocks the ferromagnesian minerals stand out clearly against the pale background to produce a mottled appearance, but the finer grained varieties have a more homogeneous colour. One drill hole in the Bt. Ibam area revealed a highly leucocratic variety.

Most of the rocks are hornblende granodiorites, but some of the Bt. Pesagi samples contain less quartz and verge towards quartz diorites, while one sample from Bt. Ibam is an adamellite. The rocks have suffered extensive alteration, and the feldspars in particular are never fresh. Thin veinlets cut some of the rocks, quartz and calcite being predominant, but siderite with barytes, and prehnite have been detected in the Bt. Ibam granodiorite. Texturally the rocks show considerable variation, from hypidiomorphic equigranular to some specimens which are extremely inequigranular, the latter showing a continuous range of grain size. The essential minerals are sodic plagioclase, orthoclase which is slightly perthitic, biotite, sericite, hornblende, chlorite which has formed from other ferromagnesian minerals, and quartz. Also present are epidote, apatite, sphene, magnetite, pyrite and tourmaline, the magnetite occasionally reaching the proportions of an essential mineral.

Plagioclase

Plagioclase is the most abundant constituent and occurs as euhedral and subhedral crystals, which vary from rectangular columnar grains to equidimensional grains with a square or slightly rhombic outline. In some of the coarser grained rocks there is a tendency towards anhedral crystals, but it is not common. When developed as columnar crystals there is little or no parallelism of the individual grains. The plagioclase is invariably twinned and frequently zoned; albite twinning is ubiquitous, but combined carlsbad-albite twins occur rarely, while pericline twinning is slightly more common. Determinations on sections normal to $\{010\}$ show a wide range in composition, from Ab_{95} to Ab_{60}, but the majority appear to lie close to the oligoclase-andesine boundary at Ab_{70}. Zoning, which can be very strongly developed, is of the normal progressive type, and the zonal texture can be accentuated by selective alteration. There is often a wide variation in grain size, from 4 mm down to a fraction of a millimetre, but some specimens e.g. from Bt. Hitam, tend to be equigranular.

Many of the plagioclase crystals have an overgrowth of potassic feldspar; the junction between the two is often indistinct, but it can sometimes be determined by the termination of the plagioclase twin lamellae.

Alteration is never absent but is variable in intensity. The most common alteration product is sericite, which typically occurs as scattered small flakes, with or without orientation. There are also small parallel veinlets of sericite which cut across the albite twin lamellae, probably following the (001) cleavage. The zonal texture has played an important role in controlling the loci of alteration, and beautiful examples of selective zonal replacement or replaced cores with unaltered rims can be seen. The zones of alteration usually show diffuse outlines, but some have very sharp edges with perfect crystal outline. When sericitization is advanced, there can be complete pseudomorphing of the plagioclase. Epidote is common as an alteration product, usually occurring as scattered anhedral grains, but rectilinear networks controlled by the cleavages and twin planes have been seen. Some epidote has been introduced into the rocks by veining, but it has also formed by alteration in situ. Chlorite is found as an alteration product, and on rare occasions forms complete pseudomorphs. Where calcitization has been an important process the plagioclase can be completely replaced by calcite. There is little doubt that calcitization can play an important role in the overall processes of alteration, as there appears to be a sympathetic relationship between the presence of abundant calcite and the intensity of alteration. The writer has noticed the same correlation in an adamellite on Pulau Tioman, an island off the east coast of Malaya (Bean, 1966).

Plagioclase was one of the earliest minerals, probably the first, to form. It is clearly much earlier than quartz and orthoclase, and in some specimens it is earlier than the ferromagnesian minerals, which are moulded onto it.

Potash Feldspar

In the Ulu Rompin rocks potash feldspar is a late-stage mineral which has crystallized with quartz during the last phases of solidification. This is clearly shown in samples from the Bt. Pesagi area, in which quartz and perthitic orthoclase in micrographic intergrowth fill the interstices between the earlier minerals - Fig. 20. The potash feldspar is orthoclase, with some development of fine vein and patch perthite. It is never twinned, and is most frequently seen as anhedral grains or forming rims on plagioclase. It varies in size from 3 mm grains which can be poikilitic towards

Fig. 20. Quartz and orthoclase in micrographic intergrowth filling the spaces between plagioclase crystals. Crossed nicols x 90.

small plagioclase crystals, to a fraction of a millimetre. It is extensively altered, and can be distinguished from plagioclase by its alteration products. In plane light the orthoclase has a dusty reddish-brown colour (it can be almost opaque) through which the cleavages show as clear lines. The alteration product is largely kaolinite in an extremely finely divided state, but there can be sericite in addition. Some of the orthoclase exhibits a slightly wavy extinction, as seen in strained quartz. One sample from Bt. Ibam which appears to be derived from a small offshoot of the main granodiorite body is enriched in orthoclase, which is probably the result of potassium enrichment during the final stages of solidification.

Quartz

Quartz forms from 10% to 40% of the rock, and is a late mineral. In the Sg. Tepisoh granodiorite the quartz occurs as large grains, up to 3 mm, surrounding the plagioclase crystals, and in some sections the texture is almost ophitic. The quartz is always anhedral and can be seen enclosing all the other minerals poikilitically, and it also forms micrographic intergrowths with orthoclase. It is typically clear, but in the Bt. Ibam granodiorite can contain numerous small dark inclusions. In all the bodies the quartz shows a slightly undulatory extinction, and in the Bt. Hitam mass many grains display a rectilinear fracture pattern, probably following the $\{10\bar{1}1\}$ direction.

Biotite

Fresh biotite is rare, but sufficient remains to determine its original properties. It occurs as ragged or cleanly terminated laths up to 3 mm long, irregular flakes or aggregates of flakes, and rare euhedral basal sections. It must have had at least two phases of crystallization as it is both pre- and post-quartz. Biotite is later than the plagioclase and hornblende, as it can be seen moulded round the former and growing peripherally on hornblende. The biotite frequently grows round, or adjacent to, iron ore. When it develops on hornblende the (001) cleavage of the mica and the (110) cleavage of the amphibole coincide. Most of the biotite is earlier than quartz, as spindle-shaped bodies of quartz are frequently observed along (001) cleavage planes. These small bodies have distorted the cleavage planes adjacent to them, and show every sign of having grown in situ, forcibly pushing aside the cleavage planes. On the other hand there are very rare examples of small biotite crystals poikilitically enclosing quartz, and here the biotite is later.

The darkest biotite is a medium brown colour with no tinge of red, but in the majority of the grains the colours have been bleached by alteration. There are rare dark haloes round minute crystals, presumably zircons. Alteration varies from a few chlorite lamellae to complete pseudomorphs. When partially altered, the mica is made up of alternating biotite and chlorite lamellae or spindles parallel to (001), which are sometimes accompanied by muscovite lamellae. During the alteration to chlorite there is frequently a release of sphene and epidote, which occur as trains of small granules along the cleavage traces. The alteration to chlorite must involve some volume change as the cleavage traces are gently distorted, but the sharper kink bends which are seen on rare occasions are due to external stresses and not alteration. Biotite, or its alteration products, is seen in all the plutonic rocks in Ulu Rompin, but it is more abundant in the north, being replaced by hornblende in the rocks from the south.

Chlorite

Most of the chlorite has been derived from the biotite as described above. It also forms by alteration of hornblende, and more rarely plagioclase. The chlorite is a penninite, with straight extinction, and it usually displays anomalous purplish-blue interference colours. In addition to the chlorite formed by alteration there are small interstitial patches which look primary; they show anomalous interference colours, but have a radiating texture. In the quartz diorite at Bt. Pesagi some of the chlorite pseudomorphs of hornblende have themselves been replaced by calcite, with the release of iron ore granules which are scattered through the calcite pseudomorphs or seggregated round the periphery.

Hornblende

Forming up to 10% of the Bt. Pesagi rocks but rarer in the bodies further north, the hornblende is a pale green or brown variety, pleochroic from green or brown to almost colourless. It occurs as irregular grains up to 1 mm, with a few euhedral basal sections. It is untwinned. It is usually later than plagioclase and earlier than biotite, but can enclose quartz and orthoclase poikilitically, which suggests either that a little hornblende is extremely late, or some quartz and orthoclase crystallize before their main period, which is obviously late stage.

In the Bt. Pesagi rocks hornblende usually grows round iron ore granules. It alters to chlorite and epidote, and can be replaced by calcite. It is difficult to tell whether some chlorite grains have been derived from mica or amphibole, and hornblende may have been slightly more abundant than thin section evidence suggests.

Accessory Minerals

Sphene is present as discrete grains, and has also been produced during the chloritization of biotite. It frequently grows round iron ore granules as shapeless masses up to 2 mm across, and more rarely as subhedral crystals. It is pinkish-brown in colour and shows a slight pleochroism. Apatite is present in all the rocks, in very small quantites, as stubby prisms or hexagonal basal sections. It is often seen in close association with the iron ore grains. Epidote is found as an alteration product of feldspar and biotite, and also as small veinlets. In the Bt. Ibam granodiorite there are small solid patches of epidote, thought to have formed from small calcareous inclusions. Rare grains of brown tourmaline have been seen in the Bt. Pesagi and Bt. Ibam rocks.

Magnetite occurs as minute granules released when chlorite is replaced by calcite, and as larger (up to 0.25 mm) euhedral to anhedral grains. There has been some oxidation to haematite at the edges. Magnetite is usually an accessory mineral, but mine reports state that granodiorite samples with as much as 10% magnetite have been located.

Minor Intrusion

The only minor intrusions are found in the valley and tributaries of the Sg. Mungus, which drains along the western edge of the Bt. Ibam-Bt. Mungus ridge. There are a numer of small microgranodiorite dykes intruded into the volcanic rocks; some of the dykes follow the foliation but others are cross cutting.

Chemical Composition

Seven samples taken from four of the Ulu Rompin stocks were chemically analysed by the writer using XRF methods, and the results are shown in Table 7. Samples which were as fresh as possible were chosen, and although all the rocks had suffered some degree of alteration, it is thought that the analyses closely approximate to the original compositions.

With the exception of the Bt. Pesagi sample the rocks are similar in composition, and agree closely with some published analyses of typical granodiorites, e.g. in Hatch, Wells & Wells (1961). The similarity in composition supports the suggestion that the stocks are derived from one parental magma. Sample No. 198 from Bt. Ibam is richer in potash relative to the others, and is an adamellite. This rock, which is distinctly pink in hand specimen, has only been encountered in one drill hole, and may represent a late stage apophysis, as it is located a short distance from the main granodiorite mass. Sample No. 164 is slightly low in lime, but this appears to be of local significance as sample No. 167 from the same drill hole has a higher lime content.

The Bt. Pesagi sample represents a more basic variety and the analysis points to a diorite, but thin sections reveal the presence of free quartz.

Stocks and Aeromagnetic Maps

The aeromagnetic map of the Ulu Rompin mineralized zone is highly complex with numerous anomalies, the contributory elements being the magnetic iron ores and the intrusive rocks. The acid volcanic rocks have virtually no magnetic effect. The magnetic iron ore bodies, except for Batu Puteh, are located close to the intrusive rocks, so it is difficult to separate the contribution from each source. The writer has only examined the published aeromagnetic map on the scale of 4 miles to 1 inch,

Table 7

Chemical Analyses of Rocks From Granodiorite Stocks

	117	164	167	198	279	318	311
SiO_2	66.44	68.49	68.52	64.38	64.03	65.85	56.83
Al_2O_3	15.56	16.35	14.63	15.60	15.23	16.06	16.14
Fe_2O_3	1.84	2.31	0.39	1.89	2.41	2.12	3.60
FeO	2.10	1.25	1.12	2.06	3.14	1.97	4.11
MgO	2.00	1.61	2.39	2.40	2.65	1.51	3.60
CaO	2.21	0.71	4.41	3.40	3.49	2.94	6.93
Na_2O	3.88	4.31	4.95	1.94	3.40	4.07	2.98
K_2O	2.99	1.75	1.12	4.71	2.61	1.99	1.64
TiO_2	0.63	0.61	0.62	0.52	0.71	0.59	0.94
MnO	0.06	0.15	0.07	0.09	0.11	0.06	0.16
S	0.13	0.30	0.21	0.67	0.13	0.61	0.60
H_2O	2.16	2.16	1.57	2.34	2.09	2.24	2.27
q	22.4	31.1	22.3	23.5	20.4	24.8	13.5
or	18.5	10.5	6.5	28.5	16.0	12.0	9.5
ab	36.0	40.0	45.0	17.5	31.5	37.0	27.5
an	11.0	3.5	14.2	17.5	18.0	15.0	27.0
c	2.1	6.7	-	1.5	.5	2.2	-
en	5.8	4.6	6.8	6.8	7.6	4.4	10.4
fs	1.2	-	.4	.2	2.2	-	2.0
wo	-	-	3.2	-	-	-	3.4
il	.8	.8	.8	.6	1.0	.8	1.2
mt	1.9	1.5	.4	2.1	2.5	2.2	3.9
hm	-	.6	-	-	-	-	-
pr	.3	.7	.4	1.8	.3	1.6	1.6

117)	Bt. Ibam	135 ft from surface
164)		180
167)		230
198)		370
279	Sg. Tepisoh	(between Bt. Sanam and Bt. Hitam)
318	Bt. Hitam	352 ft from surface
311	Bt. Pesagi	333 ft

but from a study of a more detailed aeromagnetic map prepared privately for Rompin Mining Co. Ltd., the company geologists concluded that the significant anomalies were caused by the intrusive rocks, the contribution from the iron ores being masked. It would be difficult to deduce this solely from the published map, although it conforms with such an interpretation. The suggestion that the anomalies are related to the stocks and not the iron ores is supported by the evidence from Bt. Batu Puteh. The published map shows no anomaly over this body, which contains magnetite but is not close to any known intrusive stock. Extensive fieldwork by the Geological Survey of Malaya elsewhere in the country has shown that most of the significant magnetic anomalies are associated with igneous rocks containing magnetite, and not with iron ore concentrations.

The stocks within the Ulu Rompin zone have small outcrops, so the fact that much of the zone is magnetically disturbed supports the suggestion that there is a larger intrusive mass at depth.

VI STRUCTURAL GEOLOGY

Introduction

In areas which have suffered deep tropical weathering, and in addition have a thick vegetation cover, it is difficult to obtain sufficient detailed primary data on which to base a structural synthesis. Except for foliation, which can be preserved in weathered rocks, all the structural data have been collected in, or close to, the main workings at Bt. Ibam and Bt. Pesagi.

Primary Bedding

Any original bedding structures in the pyroclastic rocks have been lost following dynamic metamorphism and alteration. Even at Bt. Ibam, where the rocks are well exposed, the writer has seen no clear evidence of sedimentary structures, although a few thin sections from drill core contain evidence of bedding planes. The only place where bedding planes can be observed is in the skarns below the ore body at Bt. Ibam, When they are well bedded the rocks strike between 345° and 40° and dip in an easterly direction at angles ranging from 40° to 60° - Fig. 17. There are local variations in the dip due to folding and faulting, but the general attitude is clear, and information from drill core which intersects the beds at depth confirms the easterly dip. The rocks have completely recrystallized and lost any fine sedimentary structures which they may have contained, but occasional banding of the metamorphic minerals represents original compositional variations. Individual beds typically range in thickness from two to six inches, but more massive members up to two feet are present. The maximum thickness of the rocks is 140 ft, but part of the original calcareous lens is now occupied by the ore body, and the original thickness must have been in the order of 350 ft. The strike length is approximately 4,000 ft, but the full extension down dip is unknown.

The only other locality where the presence of sedimentary rocks has been proved is Bt. Batu Puteh, but the drilling information is not sufficient to determine any structural details.

Foliation

Many of the volcanic rocks forming the Bt. Sembilan ridge display a distinct banded texture, which in some cases is a true foliation i.e. the rocks split along a preferred direction. ? The intensity of the development of foliation is very variable; round the Bt. Pesagi and the Bt. Sanlong ore bodies the country rocks are strongly foliated, whereas round the Bt. Ibam body the rocks show little evidence of having been subjected to shearing stress. The development of the foliation took place before the intrusion of the plutonic stocks and the mineralization, as the intrusive rocks and the ore bodies show no evidence of having been sheared.

The texture in the volcanic rocks varies from extremely delicate lamellae to coarse bands with large quartz augen, although the latter are rare. Differential iron staining, quartz lenses and veinlets, and thin greyish-white stringers of clay and mica developed from the breakdown of feldspars, can all help to accentuate the banded appearance. There is a slight, but marked, swing in the strike of the foliation from south to north along the Bt. Sembilan ridge. Round Bt. Pesagi and Bt. Sanlong the foliation usually trends north to south, and has either very steep dips to the east and west or is vertical. Immediately adjacent to the Bt. Pesagi ore body the foliation is vertical, but there is some foliation trending just west of north and dipping west-southwest. Fig. 21 shows a contoured diagram of the poles of the foliation planes in the Bt. Pesagi area. Further north, round Bt. Sembilan and west of Bt. Ibam, the foliation strikes north to south and has a consistent steep easterly dip. At the northern end of the ridge the foliation swings to the west and trends between 330° and 360°, with steep dips in an easterly, and less frequently westerly, direction.

Rupture along the foliation planes has usually been insignificant, but in the Bt. Pesagi and Bt. Sanlong areas it may have been the most important factor in localizing the mineralization. There is a marked increase in the intensity

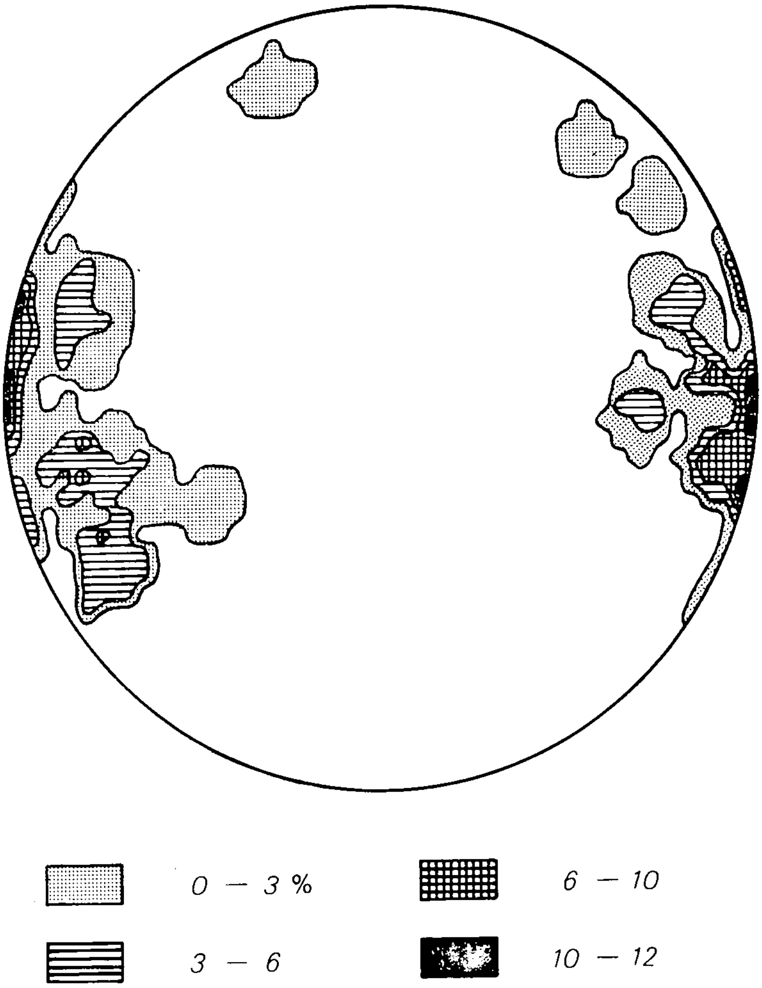


Fig. 21 Contoured diagram of the poles of the foliation planes at Bt. Pesagi.

of foliation towards the ore bodies, and as the iron ore is later than the foliation it is possible than the mineralization followed zones of weakness formed as a result of shearing. The iron ore bodies are parallel with the strike of the foliation, but there may be a slight discordance in dip. Supporting evidence for the existence of a zone of weakness is the fact that two phases of mineralization have followed the same channel, iron oxides first followed by pyrite.

In complete contrast to the rocks surrounding the Bt. Pesagi and Bt. Sanlong ore bodies, the country rock at Bt. Ibam shows little evidence of having been sheared. There is obviously some factor which is unique to the Bt. Ibam area, because outside the immediate environs of the ore body the effects of shearing again become apparent. The writer believes that the large lens of calcareous rocks was the important factor in the structural development of the Bt. Ibam area. It is suggested that the regional stresses were transmitted by the hard volcanic rocks to the calcareous lens, which absorbed the stresses by intergranular and intragranular movement, and possibly some plastic flow. Any sign of deformation in the calcareous rocks has been destroyed during the formation of the skarns. It may be significant that some of the road cuttings between Bt. Pesagi and Bt. Ibam expose volcanic rocks which are largely unsheared, and it has been suggested by some company geologists that calcareous rocks may occur in the Bt. Merah area, which adjoins the road to the west. It is interesting that in describing a contact metasomatic magnetite deposit at Iron Hill on Vancouver Island, Canada, Black (1952) suggests that during deformation a greenstone became closely fractured but the associated limestone merely flowed and recrystallized.

Folding

It is impossible to come to any firm conclusions regarding the regional fold pattern in the Ulu Rompin area owing to the lack of data and marker horizons. There is no evidence that the area has been tightly folded, and at Bt. Ibam it is clear that the folding is a subordinate feature superimposed on the regional easterly dip. The skarn below the ore body can be traced downdip through several hundred feet with no major fold structures affecting it, although there are some flexures. There are a few tight folds, but they are localized and do not materially affect the distribution pattern of the rocks. The dimensions of the folds rarely exceed 10 ft. There has been a gentle warping of the regional dip, as in plan view the ore body has a shallow inverted "S" shape.

Jointing

Natural surface exposures do not usually show good jointing, but some of the fresher rocks in the main pit at Bt. Ibam are heavily jointed. There appears to be an increase in the development of jointing adjacent to the ore body. Fig. 22 is a contoured diagram of the poles of the major joints in the Bt. Ibam pit. This shows that there are four dominant trends;

north	to	south	dipping east
east	to	west	dipping south
northwest	to	southeast	dipping northeast
northeast	to	southwest	dipping northwest

but the first two are the most important. Some of the rocks are cut by myriads of haphazardly orientated joints. Where this type of jointing is profuse there has invariably been a development of kaolinite along the joints, and in many exposures it is clear that there has been late-stage movement, as the kaolinite is full of slickensided surfaces. The joints are often so closely spaced that it is difficult to obtain specimens larger than a half inch diameter, as the rock breaks down to small, angular, kaolinite-coated fragments when struck with a hammer. It is probable that this profuse, small-scale jointing is associated with the process of mineralization and not regional stresses. The volume changes which probably accompanied the mineralization of the calcareous rocks must have produced fluctuating stresses in the surrounding volcanic rocks, which were released by the development of closely spaced and randomly orientated joints. Following the cessation of all hydrothermal activity there was intermittant slumping within the ore body, producing stresses in the country rock and in particular the hanging wall, which were released by slip along the kaolinite-lined joints, as is shown by the slickensided surfaces.

The joints have proved to be easy access channels for mineralizing and weathering solutions, and kaolin, pyrite, quartz, calcite, epidote, chlorite, goethite and manganese oxides, or mixtures of them, have been seen as joint fillings. A convincing proof of the mobility of quartz in weathering solutions is seen in the presence of small, perfect stalactites of opaline quartz along a few joints. These stalactites are hanging vertically and are obviously of very recent origin.

Joints are rare in the skarns underlying the ore body, possibly due to their incompetent nature, and the ore body itself displays little jointing. The softer gangue minerals would absorb any stresses before they affected the harder ore, and any joints which did form would be ideal channels for goethite formation.

Faulting

The Ulu Rompin mineralized zone is probably separated from the structural basin of the Sg. Jeram valley to the west by a normal fault trending approximately 350°, and which has a length of over 25 miles. The age of this fault is uncertain, but if the sediments forming the basin are accepted as being of Upper Trias to Lower Jurassic, then the fault can be dated as late Lower Jurassic, as it does not appear to affect a large granite mass which is found towards its southern end. There are certainly no faults of this size or importance cutting through the Ulu Rompin mineralized zone.

There are very clearly exposed faults in the main primary haematite ore body at Bt. Pesagi. The body, which is an elongated ellipse in plan view, has been cut by a number of small normal cross faults, which have given the body a distinctive stepped appearance - Fig. 113 . There are at least nine of these faults, which strike at 300° and have a vertical dip, with downthrow to the north. The throw is variable, the maximum recorded being 40 ft. Some of these faults can be traced into the country rock for distances up to 150 ft, but others are not discernible in the highly weathered material. Immediately south of the ore body the volcanic rocks are relatively hard due to silicification, and a number of faults following the same general trend are exposed. Some of these show a narrow central band, up to an inch wide, of mylonite along the fault zone, and some have been filled with quartz-pyrite veins, and quartz masses occur along some of the cross faults in the ore body. The faults are clearly later than the primary haematite mineralization, but earlier than the quartz-pyrite veins. The Bt. Sanlong ore body, which lies just under a mile north of Bt. Pesagi and is of the same type, has not yet been mined, but exploration has proved that it too is cut by a number of cross faults similar to those at Bt. Pesagi.

There is abundant evidence of post-mineralization faulting in the main pit at Bt. Ibam, especially in the northern half where exposures are good. Slumping movements within the ore body have produced zones in which small, hard blocks are separated by slickensided surfaces in random orientation, but these zones are not persistent. Some parts of the hanging wall rocks have been broken by collapse due to slumping movements in the underlying ore body.

The most important fault in the Bt. Ibam area cuts through the ore body towards its northern end, and is the only one to cause any marked offset to the ore. This fault can be traced over a distance of 1,500 ft, but it probably has northerly and southerly extensions, and it has been proved from the original ground surface at nearly 700 ft to sea level. The fault follows a bearing of 5° in the south, but swings north in the northern half - Fig. 31. The fault varies from a single plane of movement to a zone of intensely crumpled rocks up to 10 ft wide. The dip is vertical to very steep in an easterly direction. At the southern end the fault forms a single zone, but further north it splits into two parallel faults. In the hangingwall volcanic rocks the fault can be located in two benches, where it forms a 10 ft wide zone of shattered rocks, impregnated by manganiferous limonite and quartz. Quartz is also found along the fault in other localities.

The nature of the crushed material in the fault zone is unusual where the fault cuts through massive chlorite rock. Parallel with the fault the chlorite has been compressed into polished and slickensided layers in which are set polished rounded pebbles of chlorite, giving the rock a distinctive texture. The rock is made up of bulbous, spindle-shaped units with sharply pointed ends, which interlock together as shown in Fig. 23.

Where the fault cuts iron ore it grinds it to a granular mass, but in the lower grade ore the gangue absorbs the movement, and small polished balls of talc have been seen in the fault zone. Chalcopyrite surrounds some of the shattered magnetite so it is obviously later than the faulting.

There is a small fault in hard volcanic rock just to the east of the ore body, and the material in the fault is a strongly banded mylonite, which shows some similarity to the country rock developed at Bt. Pesagi. The rock has been crushed, with fine mica and chlorite accentuating the banding which is parallel with the fault. It is interesting that a smooth surface cut across the banding shows fine joints which are inclined to the direction of the banding. These are probably tension cracks formed in a rotational strain, the direction of the banding being a shear plane (Hills, 1963).

The major fault has experienced more than one phase of movement, as a single phase would not account for the relative disposition of the ore body on either side of the fault. Slickensides and ridges on some of the fault surfaces show that the last movement was a dextral wrench, with a dip slip inclined at 15° to 20° to the horizontal. The overall translation of the ore body is sinistral, so there must have been an earlier phase of sinistral wrenching, probably with a fairly large dip slip component.

Immediately east of the junction between the hangingwall and the main fault at 310 ft elevation, there are a number of small en echelon faults, and where the direction of movement can be determined they seem to be strike slips. In one fault the direction is clearly dextral, but in the others indeterminate. These faults are of limited vertical extent, and have an arcuate shape when viewed along the strike. The small faults make an angle of approximately 15° with the main fault direction, and appear to have formed by drag movements accompanying the dextral phase of wrenching. The fact that there is only one major direction of wrench faulting suggests that the causative stress contained a rotational element, and was not purely compressive.

Small normal faults are not uncommon in the footwall rocks, being easily seen in the skarns, but more difficult to detect in the volcanic rocks. The movements are very small and the faults are of short lateral extent. Some of the faults contain a narrow band of breccia cemented by limonite and/or quartz. The faults do not appear to follow any definite trend, and are probably due to locally developed stresses.

The footwall contact is an important plane of movement, although possibly not a fault in the true sense. The contact shows very good examples of polishing, slickensides and gouges, and without exception the movement has been downdip. There is no doubt that this has been due to slumping within the ore body, causing it to slowly slip down the footwall contact.

Although the evidence is indirect and far from conclusive, it is possible that there is a fault cutting through the ore body near to section N 10,400. This locality marks the topographical low on the ridge between Bt. Mungus and Bt. Ibam, and the ore body here reaches its thinnest dimension. The geology is unique in this area in that siderite is fairly common in the footwall and hangingwall country rocks, and the only known megascopic sphalerite in Ulu Rompin was found here, deep below the ore body. While the writer was on the mine there were no exposures as the area was covered by bulldozed material, and none of the reports written by mine staff during the period of intensive prospecting or early mining refer to a fault, nor do any plans examined by the writer show one. Two early reports however, compiled some time before the precise configuration of the ore body was know, refer to faulting in the area. Tillia (1956) says that there are no positive signs of faulting within the ore deposit, but certain evidence "indicates the presence of shearing or a small displacement fault in the neighbourhood of Section N 10,400", while MacDonald (1970) says "There is evidence of faulting in the ore body, in the section between Bukit Mungus and Bukit Ibam", and mentions slickensides in a pit. There is no offset of the ore body and any fault is probably a pre-mineralization feature. Although the evidence is slight, a fault could help to explain some of the unusual aspects of the geology around N 10,400.

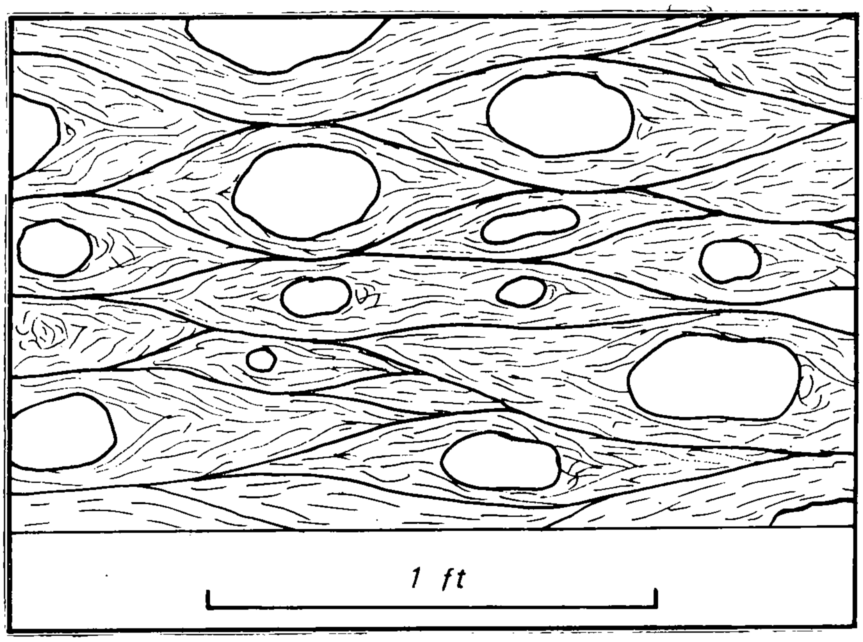


Fig. 23. Structure developed where massive chlorite rock is cut by a fault.

IRON ORE DEPOSITS

VII GENERAL DESCRIPTION

Distribution of the Deposits

Scattered throughout the Ulu Rompin area there are several deposits of iron ore, which can be grouped under nine localities. It is possible that ore will eventually be extracted from all the deposits, but the development of the peripheral ones will necessitate the construction of long haulage roads, as all treatment will have to be undertaken in the plant at Bt. Ibam. The nine deposits are listed below, and their locations are shown on Fig. 24.

1) Bt. Ibam.	The largest deposit and the mainstay of the whole mining operation. This primary ore body is located on two hills, Bt. Ibam and Bt. Mungus, but it is always referred to by the one name. The primary ore is magnetite.
2) Bt. Pesagi.	Iron ore occurs over a wide area round Bt. Pesagi, largely as a thin superfical layer of detrital ore and ore formed by secondary encrichment, but there are some primary magnetite-haematite bodies.
3) Bt. Sanlong.	A small primary magnetite-haematite body with an associated scree of detrital ore, and some ore formed by secondary enrichment.
4) Bt. Hitam.	A small primary magnetite body near a granodiorite stock, with some detrital ore.
5) Bt. Sanam.	A small primary magnetite body near a granodiorite stock. The primary ore has been extensively altered, and there is associated detrital ore.

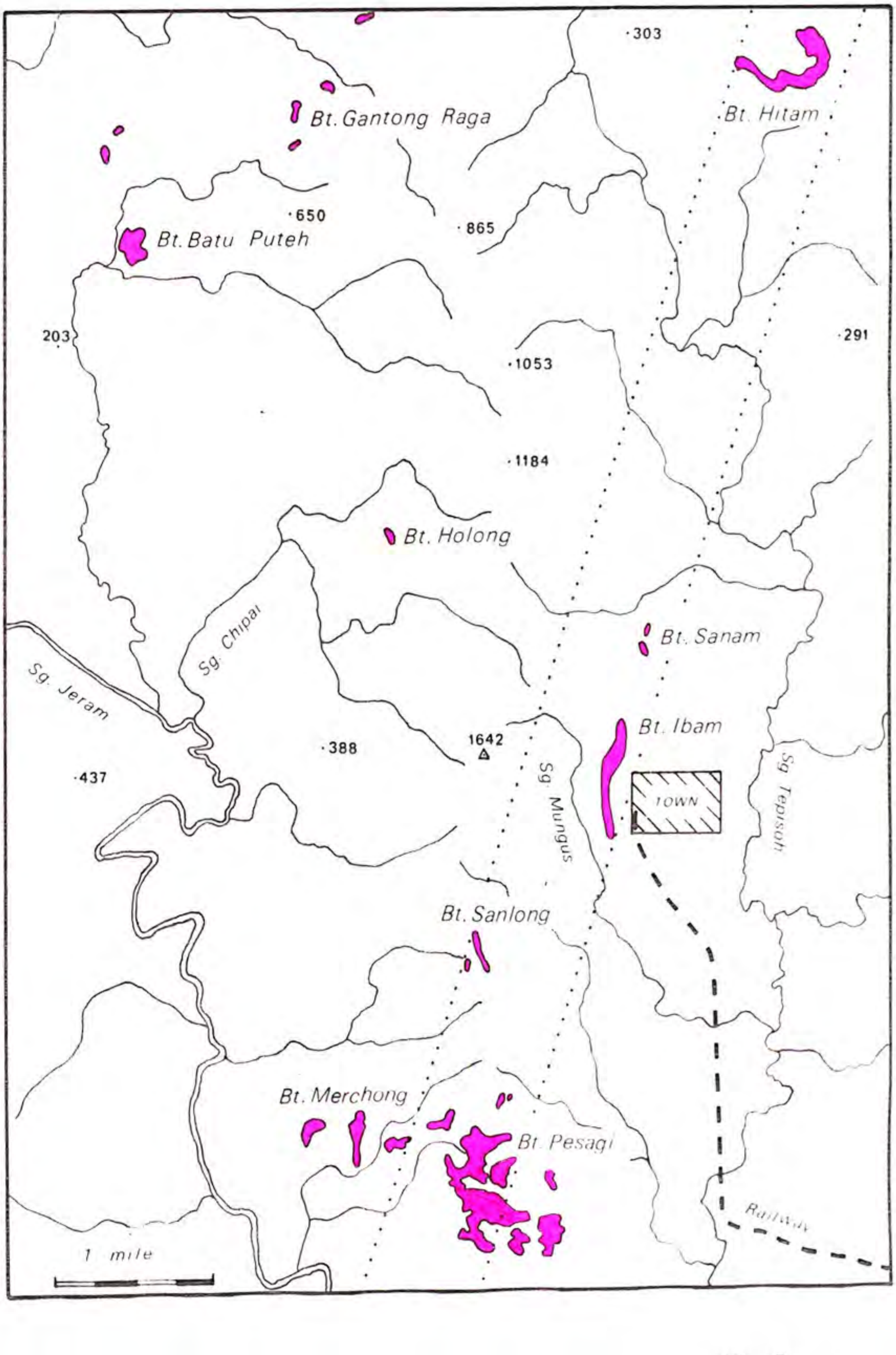


Fig. 24 Map showing the location of the iron ore deposits in Ulu Rompin.

6) Bt. Batu Puteh.	A small primary magnetite body associated with marble.
7) Bt. Merchong.	Small deposits of surface ore formed by secondary enrichment.
8) Bt. Holong.	(Also known as Ulu Chepai or Bt. Pajaid) A small deposit of surface ore formed by secondary enrichment.
9) Bt. Gantong Raga.	A small deposit of surface ore formed by secondary enrichment.

The granodiorite stock associated with the Bt. Hitam deposit extends eastwards out of the Ulu Rompin zone and forms the small hill mass of Bt. Chepemak 664 ft, on which there is some iron mineralization.

In addition to the nine deposits described above there are other small occurrences of iron ore, many of which contain only superficial lateritic material, but there are some patches of primary ore. The latter are found as quartz-rich veins with haematite or magnetite or both, which have been injected parallel to the foliation planes in sheared acid volcanic rocks. The veins are normally of small dimensions, up to 5 ft wide, and are not of commercial value. There are also a few scattered accumulations of boulders of a highly quartzose iron ore, which have obviously been derived from the complete disintegration of primary veins.

Although the main primary deposits lie relatively close together, there are some noticeable differences between them, with the exception of those at Bt. Pesagi and Bt. Sanlong which are identical. Apart from Bt. Batu Puteh which lies isolated three and a half miles to the west, the primary ore bodies are located in a narrow zone just over a half mile wide, which follows a bearing of 15° from Bt. Pesagi in the south to Bt. Hitam in the north, a distance of seven miles - Fig. 24. This zone is not concordant with the regional strike of the foliation, and it does not appear to be a recognizable structural unit, but it does contain many of the intrusive stocks which occur throughout the Ulu Rompin area. There are no known intrusive rocks near the Bt. Batu Puteh deposit.

The primary ore is located in a regional environment of highly altered acid volcanic rocks, but at Bt. Ibam and Bt. Batu Puteh there is evidence that calcareous rocks have been of paramount importance in localizing the mineralization. In the Bt. Pesagi and Bt. Sanlong deposits the control was structural, but at Bt. Sanam and Bt. Hitam the evidence is poor and inconclusive.

Regional Pattern of Mineralization

Although the Ulu Rompin deposits form a fairly well defined group, they are only the southernmost group in a line of deposits developed on a regional scale. Within a narrow belt of country running just west of north from Ulu Rompin for a distance of 40 miles, there are several indications of iron ore, some of which are commercially important, although they are all smaller than the Ulu Rompin deposits. There is a gap in the mineralization immediately south of Ulu Rompin, but 22 miles away, and on a continuation of the line mentioned above, there is a small deposit of iron ore, although it does not appear to be of commercial value. It is possible that the Ulu Rompin deposits and those to the north are related, but it is not known if there is any connection with the deposit further south. All these iron ore deposits lie within a much larger zone which passes roughly north to south through much of central Malaya, in which tin mineralization is virtually absent, but iron, gold, and scattered occurrences of base metals are found.

Immediately west of Ulu Rompin there is a large tract of country which appears to be devoid of significant mineral deposits, but eastwards towards the coast there are scattered deposits of cassiterite, a little tungsten and some poor iron mineralization. These deposits fall within the eastern tin belt of Malaya. A comprehensive explanation for the mineral distribution throughout Malaya is not yet available, but it is quite clear that the presence of hornblende in an intrusive mass virtually precludes the presence of tin mineralization, but such rocks can be associated with iron and gold deposits. All the plutonic rocks in Ulu Rompin are hornblende bearing. Stanniferous granites, which in Malaya usually means a richly potassic variety, can be associated with iron ore deposits, and several which are located in the eastern tin belt are contaminated with cassiterite e.g. Bt. Besi, Bt. Bangkong and Pelepah Kanan.

Note.

When Rompin Mining Co. granted the writer permission to study the Ulu Rompin deposits their one stipulation was that no figures for ore reserves should be quoted; this has been respected. Published figures (MacDonald, 1970) vary from 30 million tons estimated by the original Japanese prospectors to 8 million tons by Fitch (1941).

VIII BUKIT IBAM ORE BODY

Surface Features

This important ore body originally formed an impressive outcrop following the crest of the ridge which runs from Bt. Mungus in the north to Bt. Ibam in the south, with an associated deposit of boulder ore scattered over the eastern and western flanks of the ridge. The ore formed an almost continuous outcrop along its strike length of just over 3,000 ft, and in places formed vertical cliffs which projected as much as 40 ft above ground level. The width of the outcrop varied from approximately 50 ft on Bt. Ibam to just over 350 ft and Bt. Mungus, although the maximum true thickness is about 200 ft. The general strike of the ore body was north to south, but the outcrop was curved into a very shallow inverted "S" shape, convex towards the east in the northern half and the opposite in the south. The original elevation varied from 460 ft on Bt. Ibam to 710 ft on Bt. Mungus, with the lowest point on the col between them at 410 ft. The boulders shed from the main outcrop were more abundant on the eastern hill slopes, where they extended nearly 600 ft downhill. Some of the boulders were of enormous dimensions and easily mistaken for outcrops.

The in situ ore consisted of a hard mass of finely crystalline to earthy, somewhat porous and vughy mixture of haematite and goethite, with some patches of finely crystalline magnetite. Most of the haematite-goethite ore had a clearly detectable remnant magnetism, and had obviously been derived from primary magnetite by alteration. The ore was largely free from visible impurities, but there were patches and veinlets of a highly iron stained clay and quartz, and some parts showed blue and green staining due to copper minerals. The boulder ore on the ridge slopes was of similar composition, but there was also secondary limonitic material formed by lateritization.

Fig. 25 is taken from a company report and shows the surface geology. The map has not been compiled solely from what could be observed on the surface, and some details have been extrapolated

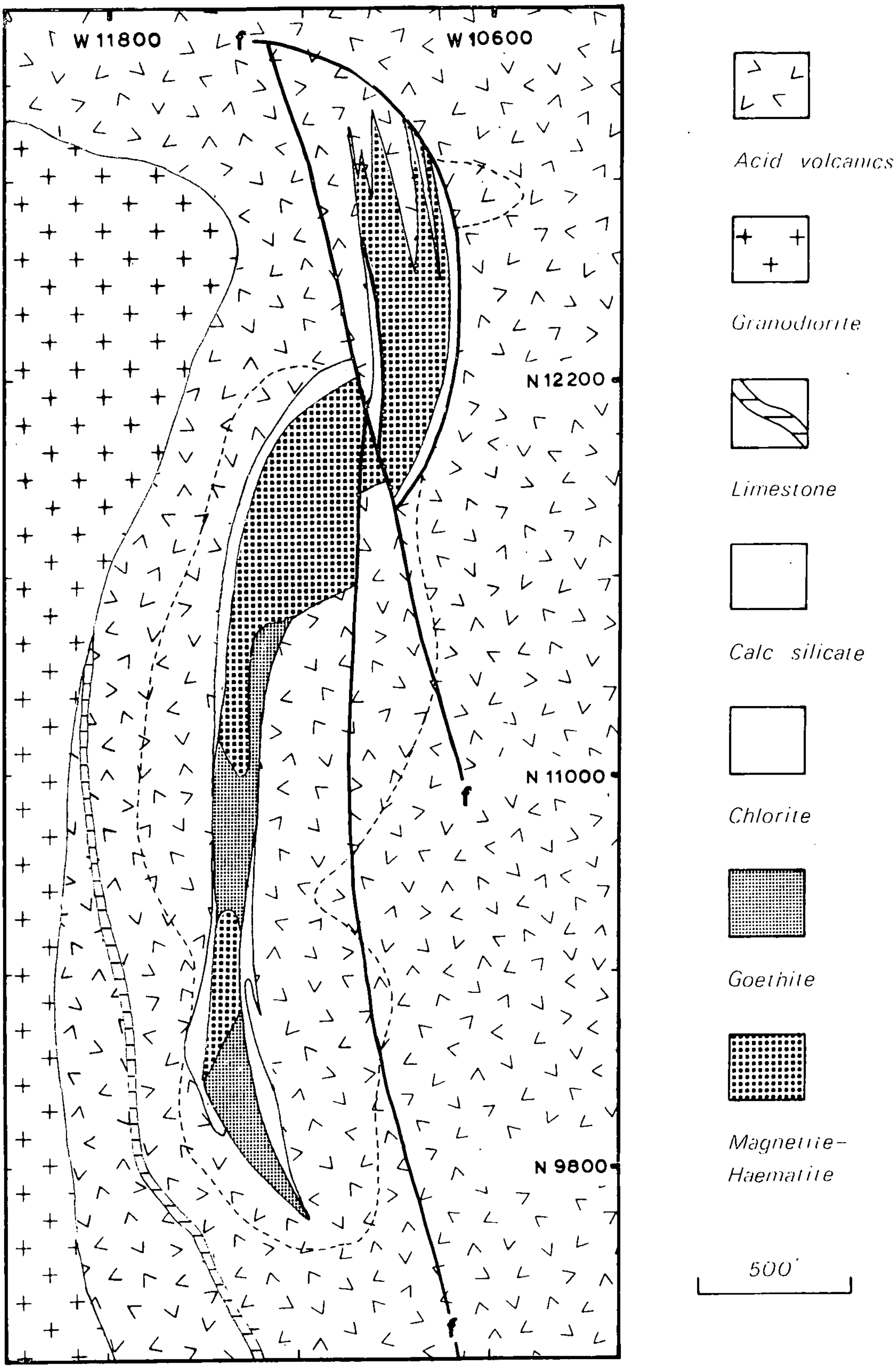


Fig. 25 Geological map of the Bt. Ibam ore body. Reduced from a ROMPINCO map.

from subsurface data. For example, the thin limestone band shown on the map was certainly not discernible in outcrop, and the large faults were not readily apparent. The map must have been compiled when the nature of the ore body and its surroundings was known in some detail, and it gives a clear idea of the overall geological picture prior to mining.

Prospecting

The Bt. Ibam deposit has been subjected to several phases of prospecting, starting in the late 1930's with the Japanese, who relied on pitting to test the boulder ore and aditting to explore the main ore body. They sank 80 pits, and drove 10 adits with a total length of 3150 ft. The present owners checked and expanded this work, but have relied largely on extensive diamond drilling to the explore the area. Up to the end of 1966 the company had sunk over 200 drill holes, the deepest being just under 675 ft. The close drilling pattern was necessary to accurately delineate the chemical characteristics of the ore body, which were complex due to the presence of several impurity elements. The shape of the ore body is fairly simple and was known at an early date, but the chemical complexity only became apparent as prospecting advanced. As the full extent of the contamination became apparent it was imperative to accurately locate the zones of poor ore, plus some areas of high grade but friable ore, which at one time was not acceptable to the buyers. The latter problem was solved when fine ore was required for pelletizing. Various geophysical methods of exploration have been tried, but do not appear to have contributed much to the detailed prospecting.

In order to facilitate their prospecting activities the Rompin Mining Co. laid down their own survey grid over the area, and this is always used to define locations. From the starting point the grid is measured northwards and westwards in feet. This system of reference is used where necessary in this thesis, so, for example, Section N. 10,800 refers to a particular east to west section through the ore body which is 200 ft north of Section N. 10,600.

Shape of the Ore Body

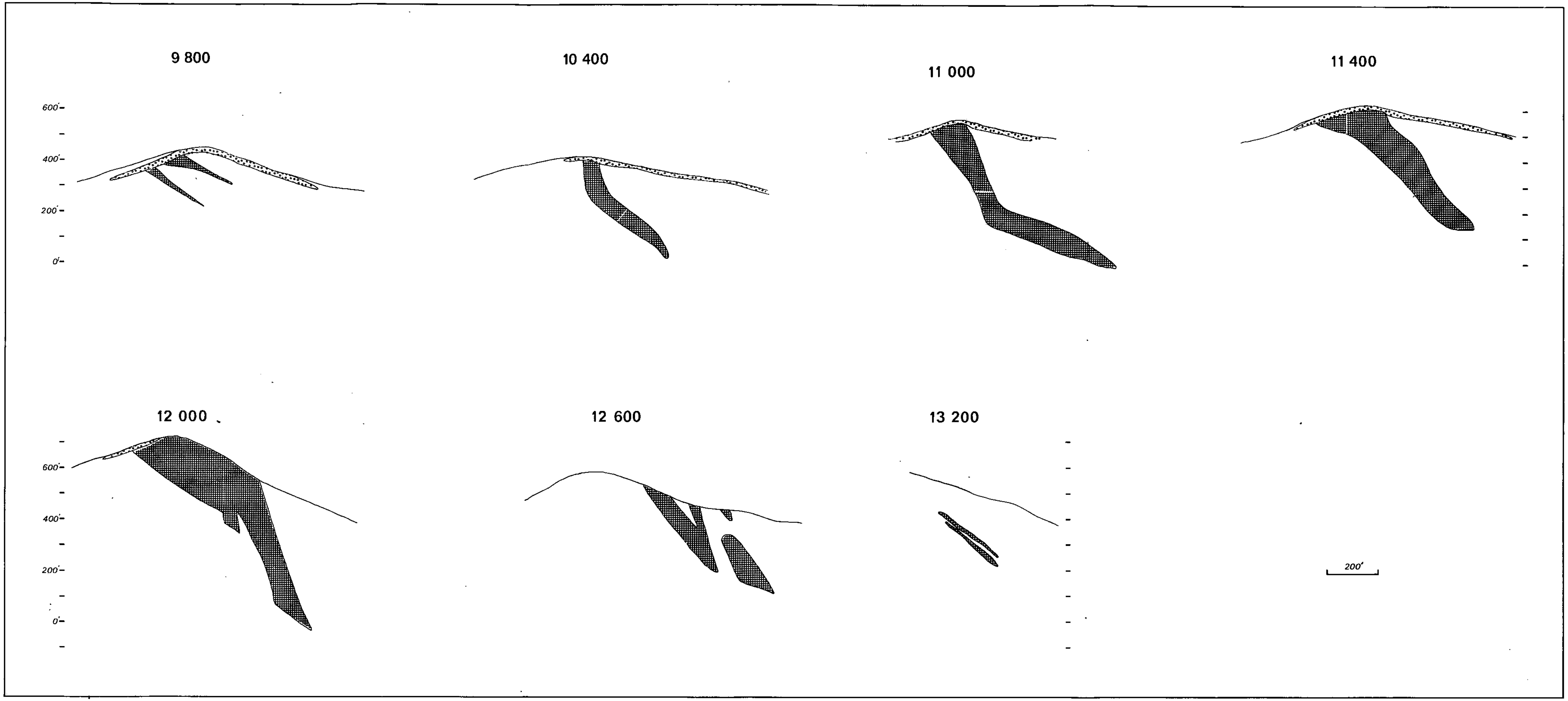
The iron ore forms a tabular body which tapers in depth, strikes north to south, and has an average dip to the east of 45°, although individual dips vary from 25° to 90°. The dip is usually fairly constant throughout the depth of the ore body, but it can show an abrupt change, either steepening as at N 12,000, or becoming shallower as from N 10,800 to N 11,200. At N 10,600 there is a short section in which the hangingwall is overturned, and dips very steeply westwards. Fig. 26 shows a series of cross sections which reveal the variation in shape and attitude of the ore body.

In horizontal section the ore body is in the form of a very shallow inverted "S", as can be seen in Fig. 25. In the higher levels of the body there is a thickening in the Bt. Mungus area, but lower down the thickness is more or less constant throughout the strike length on a particular level.

In longitudinal projection the upper surface coincides with the original ground level, while the lower limits of mineralization are known fairly accurately from diamond drilling. From a point just south of Bt. Ibam the base of the ore body dips northwards to reach 30 ft below O.D. at N 10,800, and then rises again to about 100 ft above O.D. at N 11,600. Around N 11,800 there is a sharp rise, probably caused by the main fault through the ore body, but at N 12,000 the base of the ore body is down to 50 ft above O.D. It rises to 100 ft at N 12,200, and at this line the surface outcrop dies out. From N 12,200 the ore body gradually diminishes into small lenses, and by N 13,200 the mineralization has virtually disappeared. Fig. 27 shows a projection, on a north to south plane, of the outline of the ore body. The maximum extension of mineralization in a vertical plane is found below Bt. Mungus, where it reaches 670 ft.

Dimensions of the Mineralized Body

The essential dimensions of the limits of the mineralization, not all of which can be commercially exploited, are:



modified from ROMPINCO plans

Fig. 26 Representative cross sections through the Bt Ibam ore body from south (9800) to north (13 200), showing the shape of the primary ore and the superficial boulder ore.

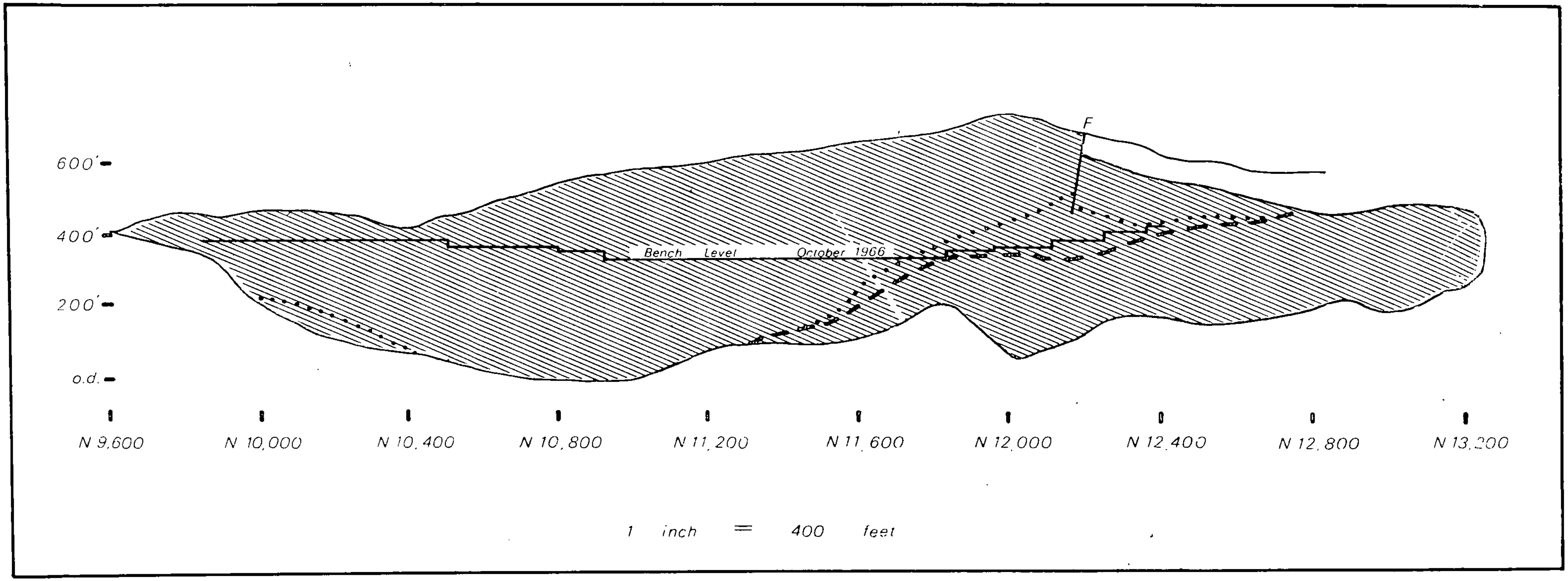


Fig. 27 Longitudinal projection of the Bt. Ibam ore body. The dashed line shows the lower limit of the direct shipping ore and the dotted line the upper limit of low grade ore which requires upgrading.

Maximum length	3,600 ft
Maximum true thickness	200 ft
Maximum thickness in horizontal section	360 ft
Maximum continuous vertical extension	670 ft
Highest elevation	710 ft above O.D.
Lowest elevation	30 ft below O.D.

Types of Iron Ore and their Distribution

The iron ore is found in a country rock of altered acid volcanics, with skarn immediately below the footwall from N. 10,400 northwards. Granodiorite has been proved 200 ft below the mineralization along N. 12,600, and 250 ft below along N. 12,000. When the ore body is viewed as a whole the contacts are very sharp, but they can be complicated in detail. Figs. 28 and 29 show general views of the ore body in September, 1966, one looking north and the other looking south. Fig. 30 is a distant view taken from the town, and shows the mining area in relation to the main Bt. Sembilan ridge. The iron ore is found in one compact body, which facilitates mining operations, but it contains patches of low grade ore and gangue.

The writer's map of the ore body north of section N. 10,600 as exposed in September 1966 is shown in Fig. 31. South of N. 10,600 there had only been surface stripping, and no exposures could be seen due to a layer of dirt and bulldozed material. The iron ore is much more variable than a distant view indicates, and it is possible to recognize three main types of ore - massive, friable and disseminated. The massive ore and the disseminated ore can be further subdivided to give a total of six groups. The change from one type to another can be gradual or abrupt, and the map of the ore body shows the overall distribution of the iron ore types. Table 8 shows the characteristics of the ore types recognized by the writer.

The sixth type is a mixture, partly of type 3 and partly of type 4, which has been silicified, but it is distinctive in the ore body because of its extreme hardness, and the fact that the gangue has retained its original dark green colour; most of the gangue has been altered and bleached to pale green or white.

Fig. 28. A general view of the Bt. Ibam mine looking north. The point where the highest bench passes into the jungle is Bt. Mungus.

Fig. 29. A general view of the Bt. Ibam mine looking south. The curvature of the ore body is revealed in the lowest bench which exposes ore. The rising ground on the extreme left is Bt. Ibam.

Fig. 30. A general view of the Bt. Ibam mine taken from the east. The highest point on the ridge behind the mine is Bt. Sembilan.

Table 8

Iron Ore Types in the Bt. Ibam Ore Body

Massive	1. Hard, blocky magnetite-martite. 2. Hard, massive ore which has been cemented or replaced by goethite.
Friable	3. Dense, finely granular, highly friable magnetite.
Disseminated	4. Fine grained magnetite, which typically shows a relatively even distribution throughout the gangue. 5. Low to medium grade disseminated ore, which contains medium to large crystals of magnetite unevenly distributed throughout the gangue, in addition to fine grained magnetite. This type of ore often contains subhedral or euhedral pyrite crystals. 6. Dense, fine grained magnetite, which has been cemented into very hard masses by silicification.

The map of the ore body at Fig. 31 shows that in spite of some measure of irregularity the distribution of the various types of ore is linear and parallel with the elongation of the ore body. The linear distribution suggests the possibility that some form of lithological control was exerted by banding in the calcareous rocks which filled the space now occupied by the ore body, but the writer can find no contributory evidence that such a control did operate. The gangue is very similar throughout the ore body, so there cannot have been much variation in the chemical composition of the original rocks. In the rare ores which are banded on a fine scale there was no control from the sedimentary rocks - page 100.

The blocky magnetite-martite ore is well developed north of N 11,600 in a band towards the centre of the ore body. The ore is massive, and divided into large blocks by small joints and fractures. It usually has a reddish external surface, but contains a high percentage of relics of magnetite and is usually of high grade. South of N 11,600 the massive ore tends to be concentrated along the footwall and hangingwall contacts. It still contains patches of high grade magnetite-martite, but owes its massiveness more to the cementing effect of goethite. The grade of this ore is very variable, as all types of material have been cemented together, including goethitized gangue. The massive ore along the footwall forms a more or less solid sheet parallel with the contact, and contains stretches of high grade ore. Along the hangingwall the grade is much more variable. Between N 11,600 and N 10,600 the ore is in contact with the acid volcanic rocks, and kaolinite is much more common in this section of the ore body. Some of the heavily iron stained gangue patches appear to be replaced volcanic rocks. At the southern end of the ore body goethite forms most of the ore, and high grade material is sparse.

The highly friable ore is best seen at the northern end of the ore body in a band close to the hanging wall. The ore is distinctive as it forms screes in the mine benches, and is intensely black in colour, in spite of the fact that it has suffered some martitization. There is a small development of friable ore along the footwall north of N 12,100 where it adjoins the main fault, but in the southern half of the ore body the friable ore

Fig 31

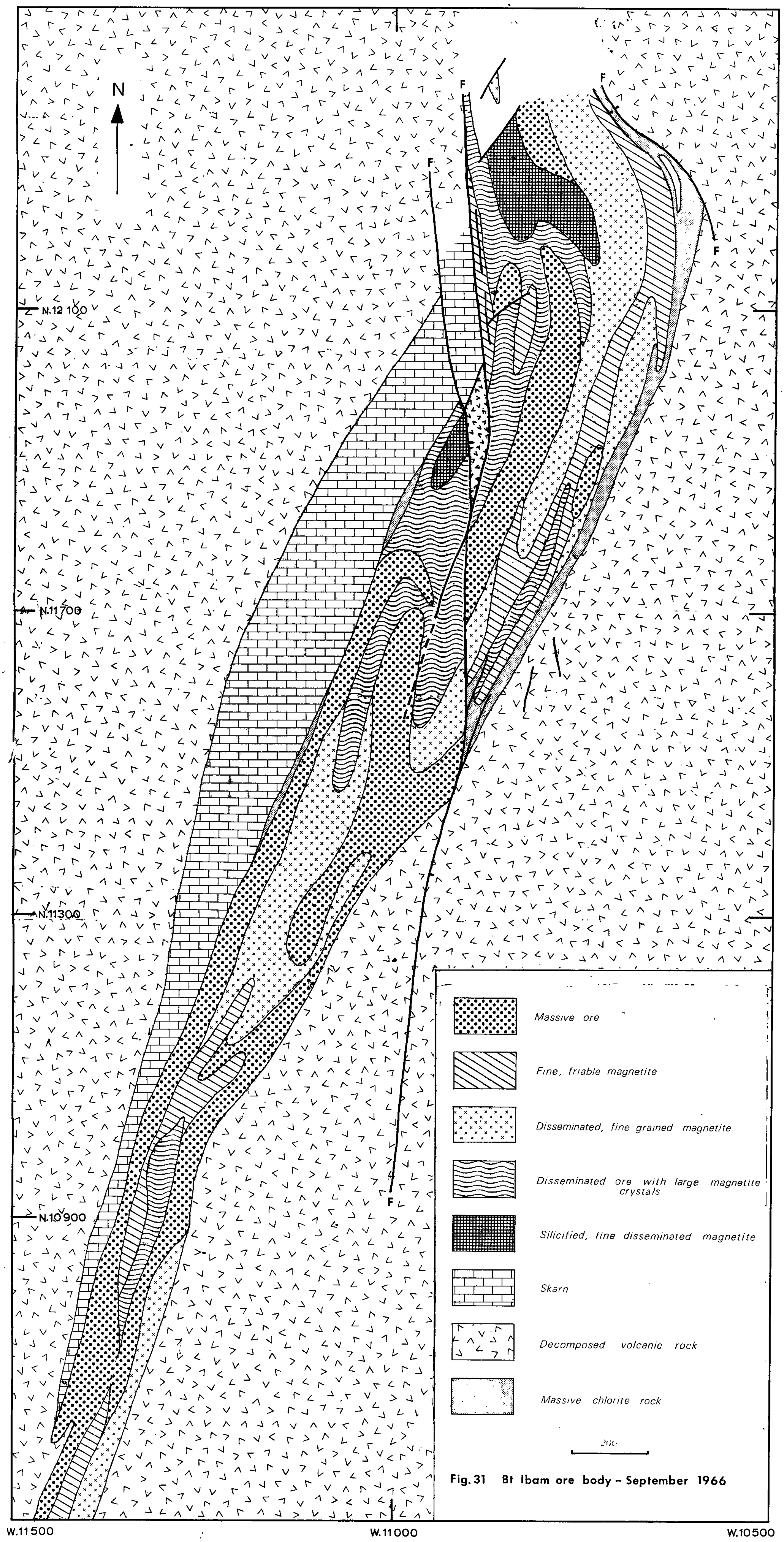


Fig. 31 Bt Ibam ore body – September 1966

is less important. The friable magnetite is usually of very high grade.

The disseminated ore - type 4 - is well developed at the northern end of the mine, but it extends as far as N 11,200 as a band towards the centre of the ore body. From N 11,000 to N 10,600 it is found as a narrow band along the hangingwall. This ore consists of fine grained magnetite set in a soft gangue. The grade varies from high to low as the proportion of gangue increases. The ore has little cohesion due to the soft gangue matrix, but it does not "flow" like the friable magnetite. Type 5 is distinctive in the field as it contains good crystals of magnetite, and is frequently rich in large pyrite crystals. The grade varies from very low to medium. Most of this ore was being stockpiled at the time of the writer's fieldwork, pending the completion of a magnetic separation plant, as the existing plant could not deal with richly pyritic ores. This type of ore is well developed towards the footwall side of the ore body between N 12,400 and N 11,600, although small patches are found elsewhere. The silicified ore - type 6 - was only located in a mass at the northern end of the mine, with a much smaller patch just south of the main fault. This ore is extremely hard, and has a glassy appearance due to the silica impregnation. It is dark green to black in colour, the quartz having preserved much of the gangue and magnetite from alteration.

The distribution of the various types of iron ore below the working levels is not so clear, as the information has to be derived from drill core. The iron ore is not easy to drill effectively, and core recovery was poor in much of the early drilling. The use of more modern techniques, particularly wireline and drilling muds, has increased the core recovery several hundred percent in the recent drilling.

There is a rough vertical zonation from surface to depth of goethite → martite → magnetite, but only the goethite zone is reasonably clearly demarcated. Within approximately 100 ft of the surface the ore was largely massive goethite with relics of martite and magnetite. Some of the goethite was in colloform masses which had obviously been deposited from colloidal solutions flowing within a near-surface supergene environment. Goethite does persist in depth, it has been identified in samples 450 ft from the surface, but colloform structures become less important, and most of the goethite has replaced martite or magnetite in situ.

There is no clear demarcation between heavily martitized magnetite and fresh, or relatively fresh, magnetite, and a series of samples through the ore body can show a haphazard pattern of martitization. The writer has encountered only one vertical sequence of samples, taken from section N 11,000, which shows a progressive change with depth. The top 80 ft consists of massive goethite with relics of martite. Below this comes 120 ft of martitized magnetite with some maghemite and goethite, although there is no systematic relation between depth and the degree of martitization. Below the martite the magnetite is fresh, but has been veined and partially replaced by primary haematite. However, in the same section of the ore body but further east, martite and goethite occur 450 ft below surface. Some cross sections through the ore body constructed by the mine staff show a line called the "base of oxidation", above which the ore is reported to be mainly martite-goethite, and below it magnetite. This line was constructed largely on chemical data, particularly sulphur values, and visual examination of core samples, but the evidence from numerous polished sections is not in good agreement with the position of this boundary. Perfectly fresh magnetite can be found well within the zone of oxidation e.g. fresh magnetite associated with copper and cuprite near the footwall on section N 11,900. The writer can see no consistent relationship between the degree of oxidation and depth, except that magnetite is preserved in only small quantities and colloform goethite is abundant in the near-surface ore. Local controls appear to have been important and have probably extensively modified any evidence of an overall depth control.

IX MINERALOGY OF BUKIT IBAM. (1) IRON MINERALS

Introduction

The Bt. Ibam ore body is a mixture of several oxide and hydroxide minerals, and within a single species it is possible to recognize distinct varieties which have resulted from different methods of formation. The minerals and their characteristic features are shown in Table 9. It is important to clearly distinguish between the two varieties of haematite which have formed from magnetite. One of them, No. 2 in Table 9, has formed by a process of metasomatic replacement under the action of solutions capable of depositing crystalline haematite, but the other, No. 5 in Table 9, has resulted from simple oxidation. The observed reactions between magnetite and the other iron minerals in the Bt. Ibam ore body are shown diagrammatically in Fig. 32.

Magnetite

Macroscopic Features

Magnetite was the only primary ore mineral of economic importance in the Bt. Ibam deposit, but much of it has been altered to martite and goethite. In one locality there is evidence that magnetite has formed from haematite by reduction, but although it is an interesting example of an unusual mineralogical reaction the quantity formed is minute.

The magnetite varies from extremely hard, fine grained, dense metallic masses, to isolated crystals of considerable size. The fresh ore has a metallic steel-blue colour on broken surfaces, but becomes duller and takes on a reddish tinge with oxidation to martite. Really high grade massive magnetite is not abundant, and it is more typical to find patches of gangue minerals throughout the ore. Within the ore body as a whole it is possible to find every gradation from solid ore to isolated crystals of magnetite in a mass of gangue. In hand specimens of the massive ore it is

Table 9

Iron Ore Minerals found in Bt. Ibam Ore Body

PRIMARY	1.	Magnetite	Coarsely to finely crystalline. It is usually dodecahedral but does form octahedra. An abundant mineral.
	2.	Haematite	Finely granular; it veins and replaces magnetite. Rare.
	3.	Haematite	Micaceous type which is only found in the footwall skarns. Uncommon.
SECONDARY	4.	Magnetite	Formed by reduction of micaceous haematite. Found only in the footwall skarns in minute quantities.
	5.	Haematite (Martite)	Formed by oxidation of magnetite and clearly shows a lattice texture. Very common.
	6.	Haematite	Deposited from colloidal solutions. Rare.
	7.	Maghemite	Formed by alteration of magnetite. Widespread, but in small amounts.
	8.	Goethite	Colloform masses deposited from colloidal solutions. Abundant.
	9.	Goethite	Granular and crystalline, formed by alteration of magnetite and haematite. Common.
	10.	'Limonite'	Mixtures of various iron oxides and hydroxides formed from colloidal solutions. Common.

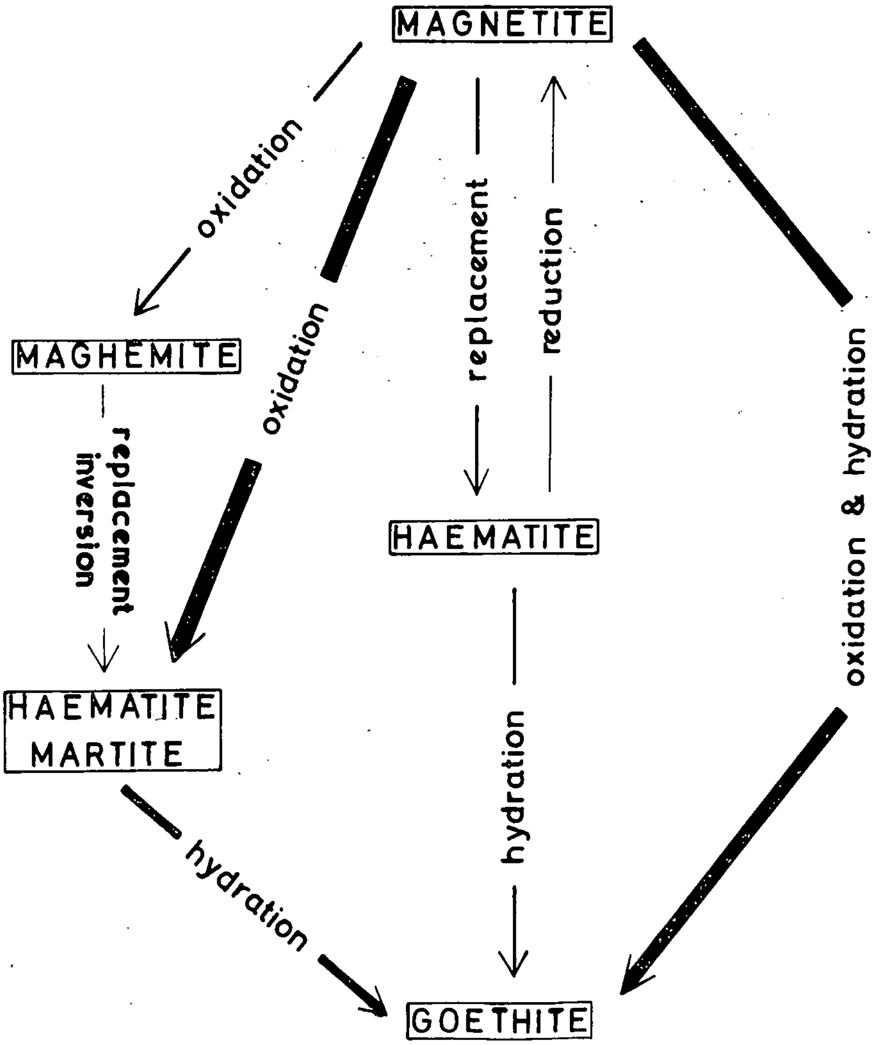


Fig 32 Diagram showing the relationship between the iron ores at Bt. Ibam. Line thickness roughly approximates the relative importance of each reaction.

impossible to detect individual grains, but its crystalline nature is well seen in polished section. As the individual grains become more widely dispersed in the gangue their crystal shape tends to improve, but really well shaped crystals are rare. The best specimens are found in low grade ore, particularly in the Bt. Mungus area, along N 12,300 close to the footwall, where the writer found individual crystals up to three inches across, but most of the euhedral crystals lie in the quarter to three-quarter inch size range. The majority of the crystals are dodecahedral 110 , rare examples being virtually perfect, and octahedra 111 are uncommon. There are also patches of coarse, well shaped crystals in mutually interfering aggregates. Many of the crystals have dull faces due to pitting and/or striations parallel to the long axis on the rhombic faces, and tend to be brittle, some of them shattering with the slightest pressure. The brittleness appears to be due partially to the presence of minute, clay-lined micro-fractures.

One distinctive type of magnetite ore is finely granular and extremely friable. The cohesion of this material is so low that a cut face immediately collapses, the magnetite flowing like dry sand until it reaches its angle of rest. This friable ore occurs in several localities, but is best developed north of N 12,000 towards the hangingwall contact.

Most of the disseminated magnetite is found either regularly distributed throughout the gangue or in pockets, but there are examples of beautifully banded ore, with numerous repetitions of thin layers of ore and gangue. The best examples were seen near the hangingwall along N 11,100, where the banding revealed some highly complicated patterns. In the zones with less contorted structures it is possible to imagine that the banding has been controlled by sedimentary microstructures, but the complex zones precluded this explanation. Some of the structures are too complicated to have been formed by folding, and occasionally completely closed structures are seen. A very good example of the latter is shown in Fig. 33. The texture displayed in this specimen cannot have been controlled by microbedding, except in the highly unlikely event that the magnetite has replaced a banded nodule. Although it is of no economic importance, there is a little magnetite in

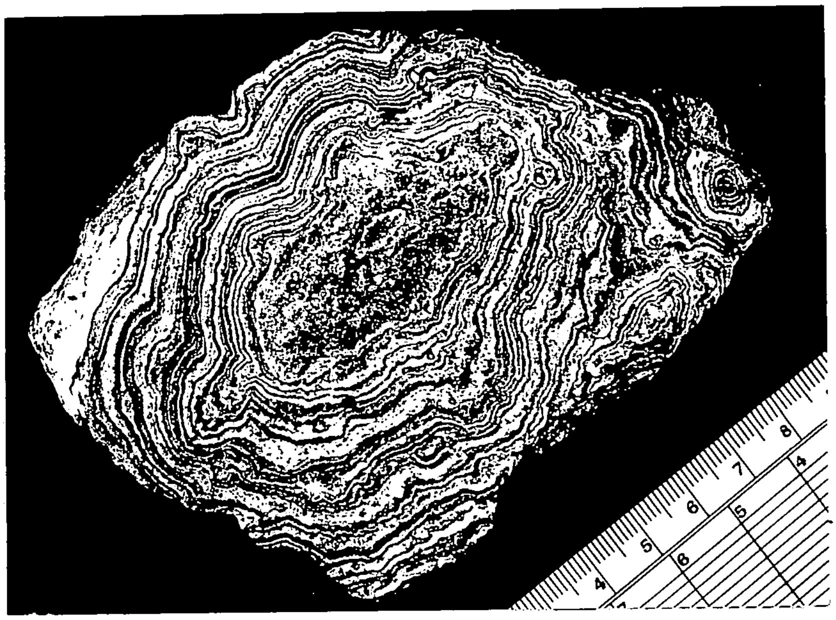


Fig. 33. Photograph of extremely delicate banding in magnetite-talc ore. Completely closed structures as shown here are extremely rare. The scale is in centimetres.

the skarns, and here the bedding has exerted some control over the mineralization. The fine, disseminated magnetite is preferentially concentrated into certain horizons, but the banding is more diffuse, and on a much coarser scale than that described above.

An unusual type of ore, which combines coarse magnetite crystals with a tabular variety, was seen in section N 12,200. The ore occurs in small masses which have a gently curved, almost botryoidal, external surface, and a definite structure. In the centre there is an aggregate of mutually interfering dodecahedral crystals, which is surrounded by a layer of greenish-white gangue (talc) up to two inches wide, and at the outside a zone about a half inch wide with masses of thin, elongated crystals of magnetite. The latter are small, up to a quarter inch long, but clearly visible to the naked eye, and tend to occur in sheaves of radiating crystals. The microscopic features are described later.

Microscopic Features

1. Polish

There seems to be a rough correlation between grain size and the perfection of the polish; the smaller grains usually polish well, but the larger crystals are often full of pits. Some of the pits are due to plucking during polishing, but others are original microcavities, and some show good crystal outlines i.e. they are negative crystals. The irregular pits are largely due to the former presence of unreplaced fragments of gangue. Fig. 34 shows the typical pitted surface on a large single crystal; it can be seen that many of the pits are irregular, but others show the development of straight edges parallel to the parting direction $\{111\}$. In extremely rare specimens small triangular pits are plucked out where the $\{111\}$ planes intersect.

2. Parting

The parting is rarely seen in fresh magnetite, and Fig. 34 shows the clearest example encountered in the Bt. Ibam ore. The parting always shows the typical octahedral pattern, and in oxidized specimens it becomes readily apparent due to selective alteration.

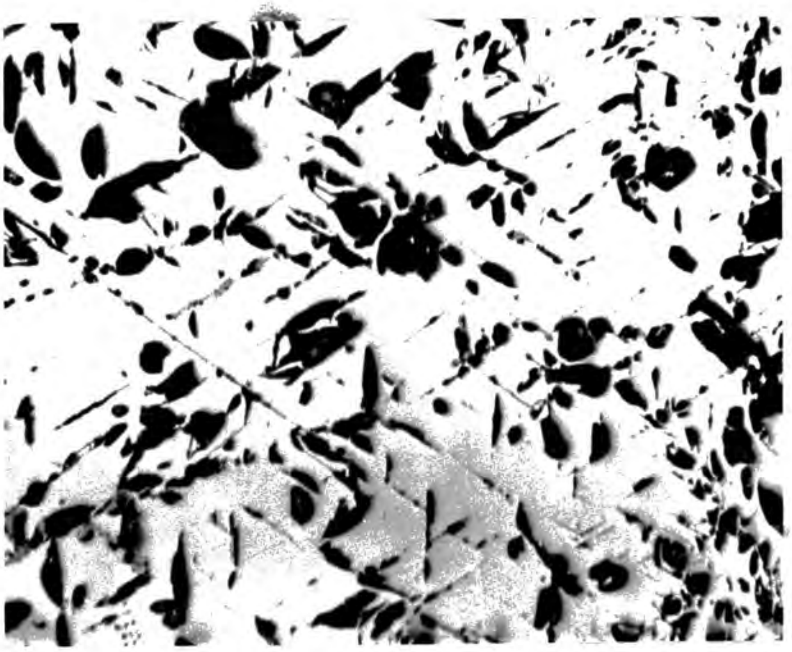


Fig. 34. Typical pitted polished surface in a large magnetite crystal, with delineation of the {111} planes. Some of the cavities are irregular, but others are partially controlled by the {111} planes. x 125.

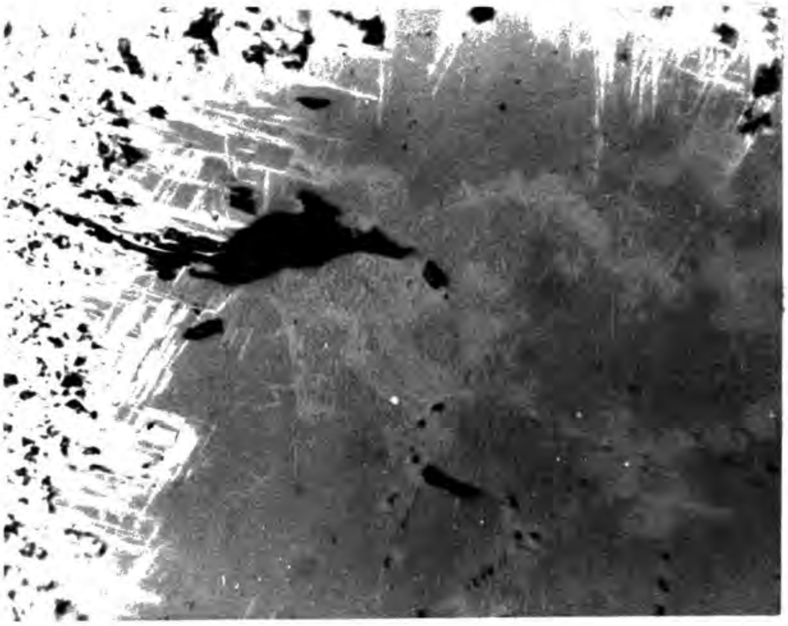


Fig. 35. Development of irregular bluish patches in magnetite. Martite lamellae and the pitted surface in the more solid martite are clearly shown on the left. x 750.

3. Colour

The colour of fresh magnetite is grey with a marked brownish tint, while occasional specimens display a faint violet tinge. The colour appears to be affected to some extent by the polishing procedure, as it was noticeable that some repolished specimens took on a perceptibly different tint. The observed change was from a colour with a violet tint to the more normal brownish-grey, which suggests that the violet colour is not an inherent property.

In the majority of the specimens of fresh magnetite the colour is uniform throughout a single grain, and only two exceptions were noted. In a few samples there is a patchiness of colour, which is difficult to photograph but is clearly seen under the microscope. It is thought that this is due to incipient oxidation, as it strongly resembles bluish patches of maghemite - Fig. 35, although the contrast in colour from the magnetite is much less marked. The other exception is found in specimens in which extremely delicate differences in colour reveal a well developed zonal texture.

4. Crystal Shape

There is every variation in the development of crystal shape, as would be expected in a large ore body, but the majority of the magnetite crystals are anhedral. The preferred crystal shape is dodecahedral, and octahedra are much rarer. In the fresh massive ore it is sometimes difficult to detect the crystal outlines, but peripheral alteration usually outlines the texture. The typical fine grained ore shows an interlocking mass of roughly equant grains which are largely anhedral, but some grains show two or three faces, and rare ones are euhedral - Fig. 36. The latter show the hexagonal shape of dodecahedra. Where there has been a development of goethite along intergrain borders the crystal faces can become rounded, although the granular nature is still apparent. In the highly friable ore approximately 90% of the grains are anhedral, while the rest show rhombs, square sections and hexagons of octahedra and dodecahedra. In the disseminated ore the magnetite is irregular in shape and size, even within a single polished section, but some good euhedral crystals are found - Fig. 37.

The most unusual magnetite was found in one locality only,

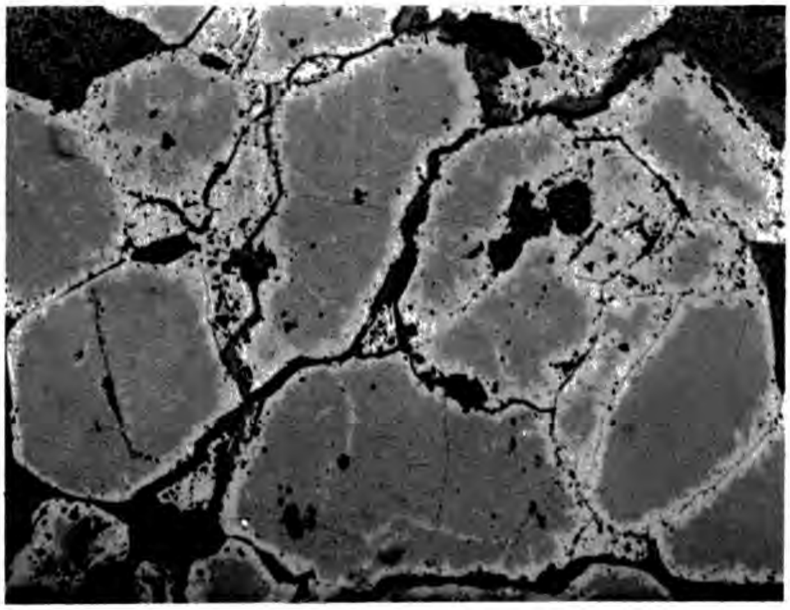


Fig. 36. Granular magnetite with martite (pale grey) developed peripherally and along cracks. The whole mass has been cemented by goethite (mottled dark grey). x 220.

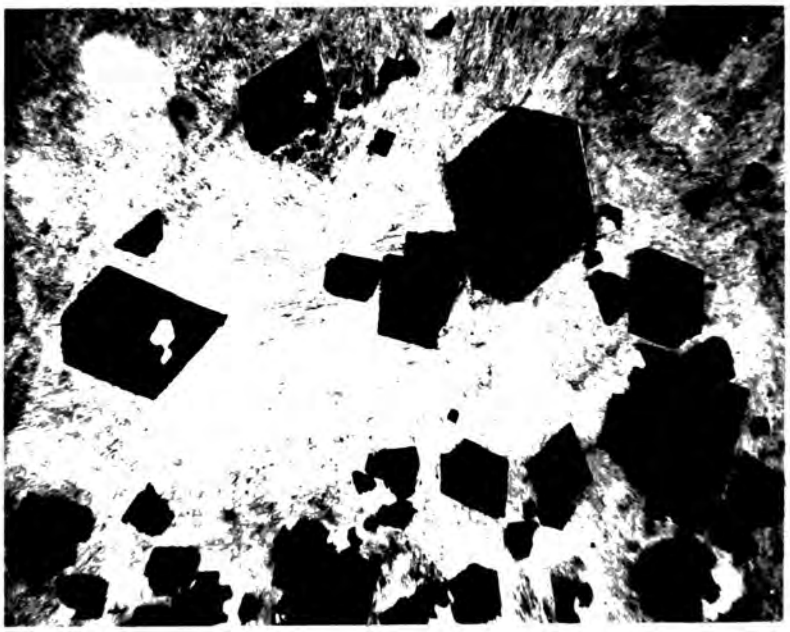


Fig. 37. Disseminated magnetite crystals which show a good euhedral shape. There are a few fragments of unreplaced gangue in the magnetite grains. x 45.

N 12,200: W 10,800, where it occurs as distinctly bladed crystals up to a quarter inch long - Fig. 38. The ratio of the length to width varies from two to approximately twenty. Much of the magnetite is fresh, but it shows some alteration to goethite and rare peripheral specks of haematite. It is most unlikely that magnetite could develop in this form under its own force of crystallization, and some external control must have been operative. Platy magnetite has been described from the contact metamorphic deposit at Cornwall, Pennsylvania, by Cooke (1936) and Davidson & Wyllie (1965), and from Concepion del Oro, Mexico, by Buseck (1966). The magnetite from Cornwall has a similar form to that seen at Bt. Ibam -"Radially arranged aggregates of platy magnetite" according to Cooke- and all the authors attribute the shape of the magnetite to the fact that it has replaced primary haematite. Davidson & Wyllie also suggest that there has been some overgrowth of primary magnetite which has retained the platy habit.

The Bt. Ibam magnetite has a uniform colour throughout, and contains no relics of unreplaced haematite, in which it differs from the Cornwall magnetite. Many of the crystals are not perfectly formed, but contain patches of gangue, some of which cut the crystals into detached fragments - Fig. 38. The writer has found no sign of specularite or micaceous haematite anywhere within the main ore body, and it is unlikely that the bladed magnetite has pseudomorphed haematite. Furthermore, except for the evidence from one thin section of a sample from the footwall skarns, where special conditions were probably operative - page 112, the haematite in the Bt. Ibam deposit is always later than magnetite. It is suggested that the magnetite has not pseudomorphed haematite, but has replaced the original amphibole gangue, assuming its elongated form. The gangue is now a mass of flaky talc, but actinolite was the dominant silicate mineral. Palache, Berman & Frondel vol. 1 (1944) mention that magnetite pseudomorphs of asbestiform minerals have been found, It is not known why the magnetite in this one locality has assumed an elongated habit, as elsewhere in the deposit magnetite has replaced amphibole and still developed its own crystal shape.

Fig. 38. Tabular crystals of magnetite in gangue. x 45.

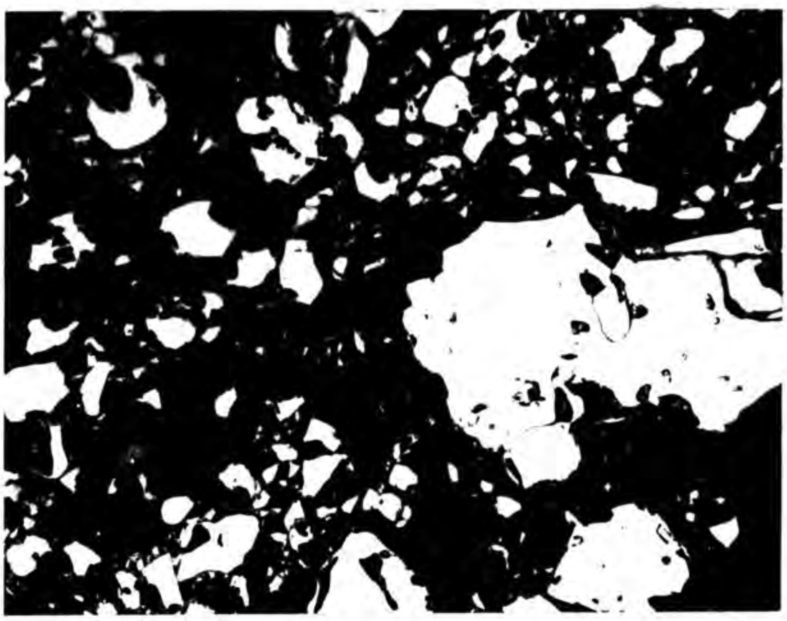


Fig. 39. Magnetite ore which has been cut by a fault. Note the irregularity and angularity of many of the grains. A small mass of chalcopyrite has partly enclosed and replaced two grains of magnetite (centre right). x 220.

Where high grade magnetite ore has been cut by a fault it is broken into irregular and angular fragments, which show a very wide variation in size. The appearance of this broken ore is shown in Fig. 39. In some samples the magnetite has been ground to extremely fine particles which are strung out in linear bands.

5. Zoning

Zoning is extremely rare in the Bt. Ibam magnetite, and it usually requires etching with HBr to develop it, but in two samples from one drill core the texture is visible in the unetched magnetite. The texture is apparent in bands with just perceptible different tints of brownish-grey, and some with a pinkish tinge. In some magnetite samples a crude zonal pattern is revealed by the oxidation products, but in those specimens which were tested with HBr no texture was developed. The only strongly zoned magnetite was found in two specimens from section N 11,900 at depths of 400 ft and 430 ft. The zoning was present in a relatively fine grained, compact ore, and none of the larger euhedral crystals found elsewhere in the ore body which were tested with HBr revealed any texture. In the ores which contain strongly zoned magnetite there are invariably some grains which are completely structureless, even after prolonged etching. Although it is not possible to be certain, it is probable that the ore containing very strongly zoned magnetite comes not from the ore body proper, but from the magnesian skarn immediately below the footwall.

A study of the zoning provides some interesting information on the growth pattern of magnetite. It is noticeable that the zones are rarely of equal width from core to rim, and that there has often been a marked change in the shape of a single crystal at different stages of its crystallization - Fig. 40. There is a marked variation in the growth rate of different faces at different times, producing variable widths for contemporaneous zones throughout a crystal - Fig. 40. Some faces have grown at a fairly constant rate, but others have developed in pulses. It is noticeable that even in crystals which show well developed crystal faces, there have been periods when the growing crystal lost its external form. This is shown in Fig. 41, where an intermediate zone shows a rounded pattern, which is quite different from the core and the rim. In some crystals this sequence is repeated several times. Fig. 41

Fig. 40. Strongly zoned magnetite etched with HBr. Note the change in shape from core to periphery in the main magnetite crystal, and the presence of magnetite which shows no vestige of zoning. x 220.

Fig. 41. Zoned magnetite etched with HBr. Note that some of the intermediate zones do not have well developed faces, and a face present near the centre of the grain (left and vertical in the photograph) becomes progressively supressed towards the edge. x 220.

also shows that growth ceased on a face which was present during the early stages of crystallization, with the result that the face is supressed. Fig. 40 shows that the core of the zoned magnetite is octahedral, but its ultimate form is dodecahedral. If this is a general trend, it could be at least a partial explanation for the preponderance of dodecahedral crystals.

In those samples in which a zonal texture is developed as a result of etching, it can be clearly seen that many of the individual anhedral grains are in fact part of a much larger crystal. Before etching, the specimen shown in Fig. 40 appeared to contain numerous discrete grains of magnetite, and it is only the zonal texture which revealed their true relationship.

The development of the zoned ores proves that there have been rhythmic pulses in the mineralization, and also indicates that there have been at least two phases of magnetite mineralization. Among the zoned magnetites there are many which are completely unzoned, the latter looking like inclusions in the zoned material, as if they belong to an earlier phase of mineralization - Fig. 40.

In a few specimens a zonal texture is revealed by the alteration products of magnetite as they preferentially attack certain bands. The texture always shows a hexagonal pattern. Fig. 42 shows goethite replacing magnetite along $\{110\}$ planes. It is interesting that the magnetites which showed this type of alteration did not develop any zonal texture when etched with HBr.

A third type of zonal texture can be seen in some of the magnetite grains which contain unreplaced fragments of gangue, where the latter are distributed in a clear, but crude, zonal arrangement - Fig. 43.

Texture

The typical texture of the massive magnetite ores is an interlocking mosaic of roughly equant grains, which are usually anhedral of subhedral and more rarely euhedral - Fig. 36. The

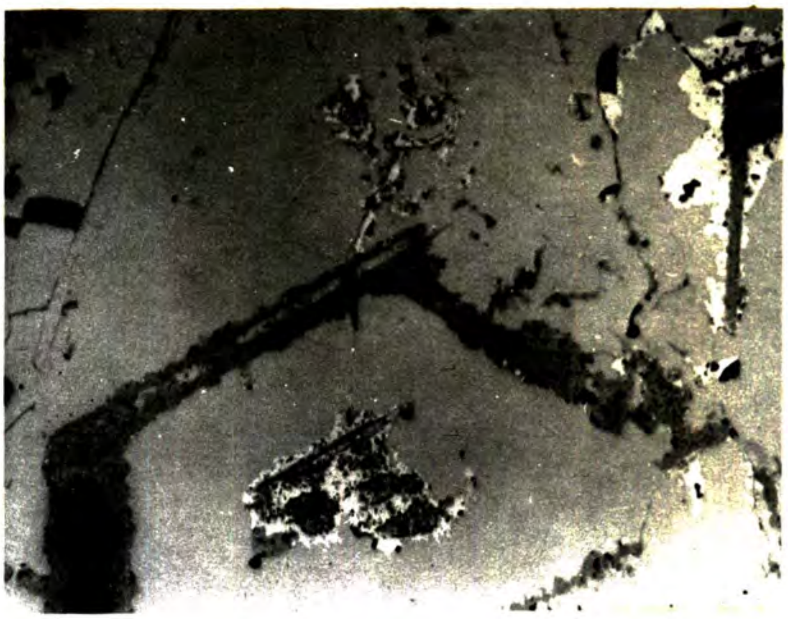


Fig. 42. Goethite (dark grey) replacing magnetite along {110} planes. Martite is pale grey. x 220.

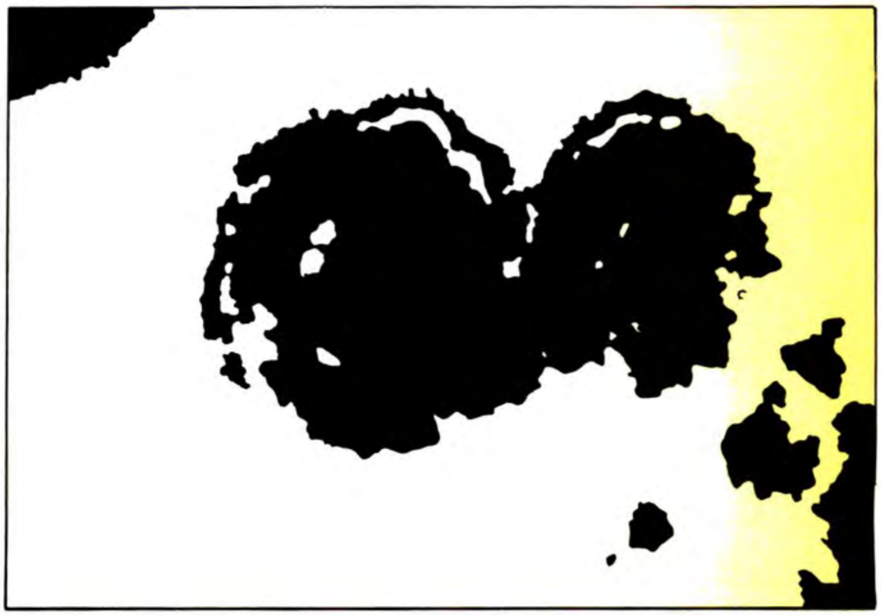


Fig. 43. Drawing taken from a photomicrograph which shows unreplaced gangue fragments in a roughly zonal arrangement in grains of magnetite. x 45.

friable magnetite had a similar texture before the grains lost their cohesion as a result of peripheral martitization - page 129. The disseminated magnetite ores usually show a random distribution of the ore grains in gangue, and the only ones with a distinctive texture are those which are banded. The banding is on a delicate scale, with alternating magnetite and gangue bands from less than 1 mm to 1 cm wide. These strongly banded ores are uncommon.

The crystallographic form of the gangue minerals has usually exerted little control over the growing magnetite, but rare examples of such a control have been observed. In one sample the magnetite had probably pseudomorphed actinolite producing bladed crystals of magnetite - page 106. The magnetite has normally grown under its own force of crystallization and developed its typical equant shape irrespective of the gangue, but in a few polished sections the ore can be seen to have grown round the periphery of some of the gangue minerals. This is well shown in Fig. 44.

Age

Magnetite was the first of all the ore minerals to crystallize. It has been veined and replaced by haematite, and replaced by martite and goethite. Pyrite, chalcopyrite, sphalerite, bismuthinite and tetradymite have all been observed veining and/or replacing magnetite, and those sulphides which have not been seen in contact with magnetite are demonstrably later than pyrite. There is no evidence that magnetite crystallized contemporaneously with any other mineral.

Magnetite after Haematite

One sample of micaceous haematite-pyrite ore which occurs in the footwall skarns shows the development of magnetite after primary haematite, which is unique in the Ulu Rompin deposits. The haematite is seen in polished section as parallel and radiating sheaves of long thin blades, and Fig. 45 shows elongated blades of haematite parallel to a $\{100\}$ face of a euhedral pyrite crystal. When the haematite-pyrite contact area is examined under high magnification it can be seen that the haematite blade immediately

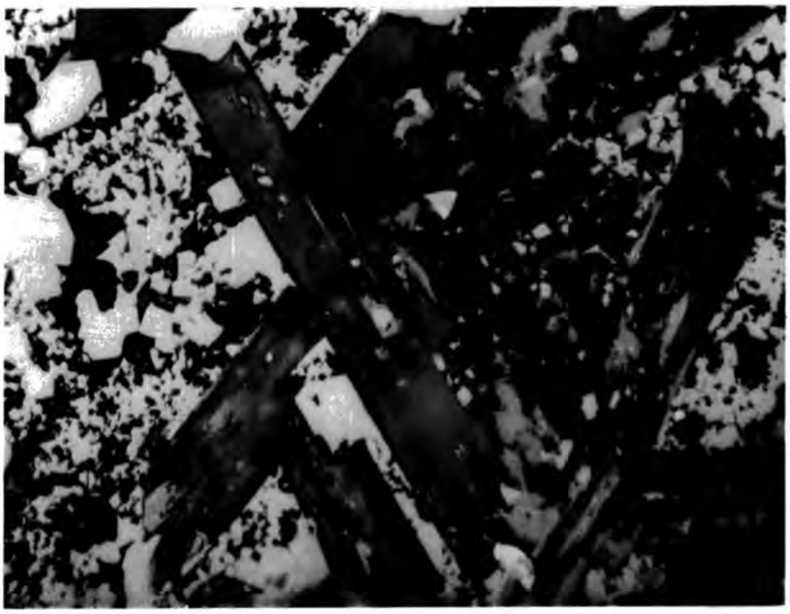


Fig. 44. Magnetite grains growing round the edge of amphibole crystals. x 560.

Fig. 45. Thin blades of micaceous haematite parallel to a {100} face of a pyrite crystal. x 45.

Fig. 46. Magnetite developed in a blade of micaceous haematite which is adjacent to a pyrite crystal. Note that the magnetite is restricted to one blade of haematite. Gangue minerals are black. x 540.

adjacent to the pyrite has been partially altered to magnetite, as shown in Fig. 46. There is no magnetite further away from the pyrite, and there must be a genetic relationship between the pyrite and the development of the magnetite. It is suggested that along the edge of the pyrite crystal a microenvironment developed, in which the haematite was reduced to magnetite. The conditions under which the reaction took place were obviously most unusual, as the phenomenon has been observed in only one polished section, and it is the sole example in all the Ulu Rompin ore deposits of magnetite forming after haematite.

Diffraction Pattern of Magnetite

Table 10 shows the diffraction patterns obtained by the writer for two magnetite samples. The cell size of No. 35 was calculated to be 8.397Å and that of No. 34 to be 8.395Å; these figures agree closely with that given for pure magnetite in Deer, Howie & Zussman vol. 5 (1962), which is 8.396Å, and the value of 8.3963Å given by Basta (1957) and used on ASTM card 11-614. The latter figure has a high degree of precision. The fact that the values for the Bt. Ibam samples are so close to those for pure material strongly indicates that there are few elements in substitution in the lattice, because it is known that such substitutions have a considerable effect on the size of the lattice (Gross, 1965).

Oxidation and Hydration of Magnetite

In spite of the apparently simple chemical reactions which are involved, the oxidation and hydration of magnetite is a complex process, which appears to be sensitive to slight changes in physico-chemical conditions. The oxidation reactions in particular have been studied in detail by numerous workers, e.g. Gruner (1926), Gheith (1952) and Lepp (1957), but the precise nature of the process and the operative controls are still uncertain. Within the relatively small dimensions of the Bt. Ibam ore body the pattern of alteration has varied both in space and time, and the rates of reaction have shown a similar variation. The important alteration products are martite, maghemite and goethite, plus undifferentiated mixtures of secondary oxides and hydroxides classified as 'limonite', which

Table 10

Diffraction Patterns For Magnetite

Co Radiation

35		34		ASTM 11-614	
dÅ	Rel. I.	dÅ	Rel. I.	dÅ	Rel. I.
4.84	5	4.82	4	4.85	40
2.967	8	2.96	7	2.966	70
2.530	10	2.528	10	2.530	100
2.421	3	2.403	1	2.419	10
2.099	7	2.094	7	2.096	70
1.713	6	1.708	3	1.712	60
1.615	7	1.613	6	1.614	85
1.483	9	1.481	7	1.483	85
1.418	1				
1.327	3			1.327	20
1.280	4	1.279	1	1.279	30
1.266	2			1.264	10
1.212	3			1.2112	20
1.122	3	1.118	1	1.1214	30
1.093	6	1.092	2	1.0922	60
1.049	3	1.047	1	1.0489	40
0.9897	2			0.9890	10
0.9693	5	0.9695	2	0.9692	40
0.9387	3	0.9380	2	0.9386	30
a = 8.397Å		a = 8.395Å		a = 8.3963Å	

35 is a dodecahedral crystal.

34 is an octahedral crystal.

collectively show a variety of textural relationships with the primary magnetite. Lepidocrocite, which could be expected to occur in this environment, has not been positively identified in any samples, but it seems unlikely that it is completely absent.

The basic reactions involved in the alteration of magnetite are;

1) $2Fe_3O_4 + O = 3Fe_2O_3$

2) $2Fe_3O_4 + O + 3H_2O = 3Fe_2O_3.H_2O$

There are probably numerous minor variations of these reactions, particularly with regard to the formation of intermediate compounds.

With reference to the second equation Cooke (1936) says "on the face of it, such a reaction appears highly improbable", but several examples of the Bt. Ibam ore demonstrate the simultaneous oxidation and hydration of magnetite. The goethite which is produced by this reaction is quantitatively much more important than that produced by hydration of martite, which is relatively resistent. This agrees with the statement of Gilbert (1925) that "Magnetite is fairly resistent to limonite alteration.... Haematite is much more resistent even than magnetite". There is no evidence in the Bt. Ibam ores that martite is an essential intermediate phase in the alteration of magnetite to goethite. The above remarks apply to the magnetite which has been pseudomorphously altered in situ; where there has been widespread solution of the original iron in magnetite and subsequent redeposition, the sequence of change cannot be followed.

The main controls over the loci and processes of alteration, apart from the pertaining physico-chemical conditions, have been intergrain boundaries, microfractures, crystallographic planes, zoning and cavities. Several of these controls have usually been operative simultaneously, but their individual influence varies from sample to sample. The most important single control has been intergrain boundaries, with the result that peripheral alteration is widespread, and it is extremely rare to see alteration products in magnetite grains which have a fresh periphery. An example of this rare type of alteration is shown in Fig. 47. The initial

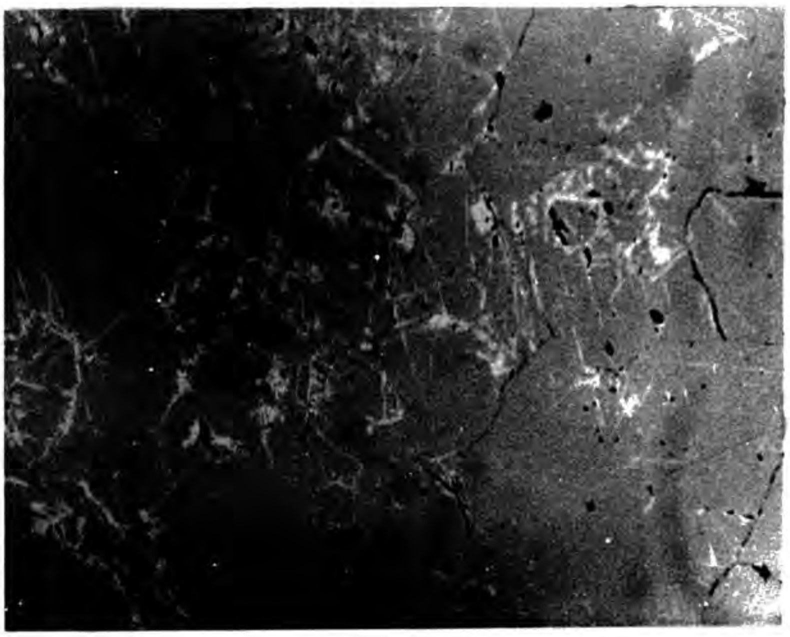


Fig. 47. An unusual type of alteration which shows scattered blades of martite (pale grey) in magnetite grains with a fresh periphery. The grain edges are seen as dark lines. x 125.

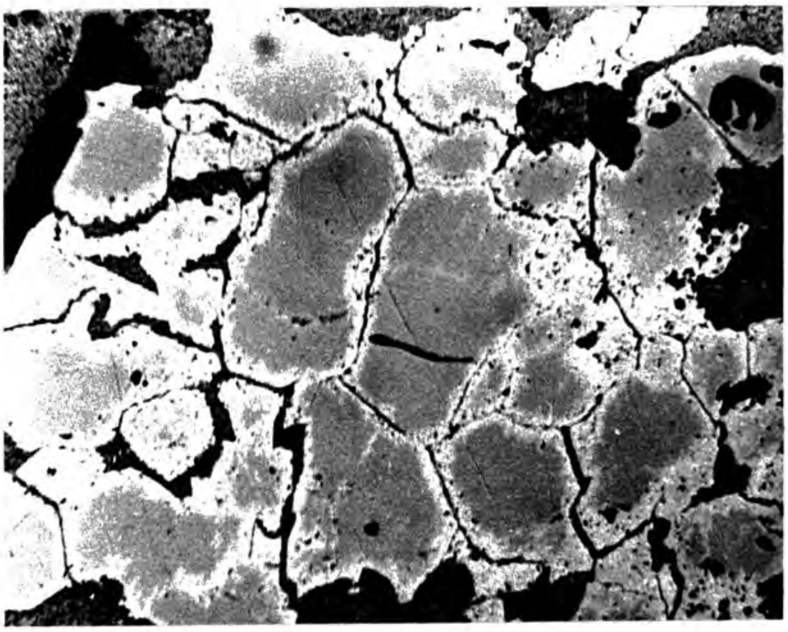


Fig. 48. Grains of magnetite outlined by martite and cemented by goethite. x 125.

alteration along intergrain boundaries produces martite, which varies from thin coronas to complete replacements as alteration procedes towards the centre of a grain. The development of martite rims clearly delineates the shape of the individual magnetite grains. Fig. 48 shows a very typical granular ore in which the magnetite is outlined by martite. Peripheral martite is also shown in Figs. 52, 53, 61 and 63. In some of the magnetite ores which have undergone advanced alteration to goethite the martite rims are clearly preserved, and still reveal the original magnetite outlines. These relics sometimes constitute the only proof that a particular goethite ore has been derived from granular magnetite. Fig. 49 shows the outline of a euhedral magnetite grain preserved by martite in a mass of goethite.

In some samples the development of peripheral martite is relatively uniform, but in others it is highly irregular, even between adjacent grains. Those ores which are composed of grains of approximately equal size and shape tend to show a uniform martite development, while those which contain irregular grains do not. Martite development also tends to be irregular in those ores which contain fractured magnetite grains, as the fractures provide additional loci for alteration. Fig. 50 shows the highly irregular development of martite, while Fig. 48 shows an ore in which all the magnetite grains are in roughly the same stage of alteration.

Microfractures and, to a lesser extent, cavities act in the same way as grain edges in that they provide preferred centres of alteration, although martite round cavities is not a common feature, probably due to the fact that they are often effectively sealed from the exterior of the grain. Figs. 48 and 63 show martite developing along fractures in magnetite.

The peripheral alteration to martite is almost always strongly controlled by the $\{111\}$ parting in the magnetite. This control is clearly seen along the inner margin of the martite rims, where the lamellae show the typical lattice textures - triangular, Fig. 62, square, Figs. 61 and 63, rhombic - depending on the orientation of the magnetite. Where there has been a complete replacement of magnetite by martite the lattice texture is usually clearly

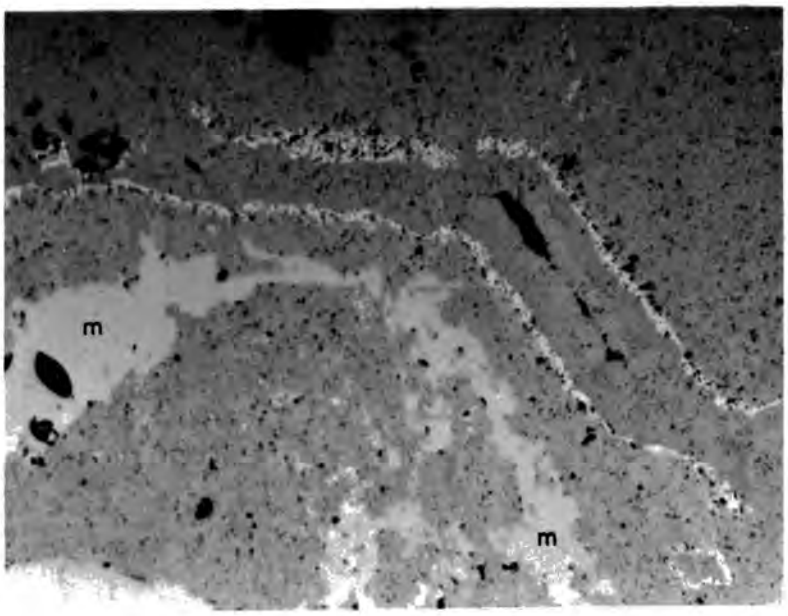


Fig. 49. Outlines of euhedral magnetite crystals preserved by a thin layer of martite (pale grey) in goethite. Relics of magnetite are preserved (lower left). x 220.

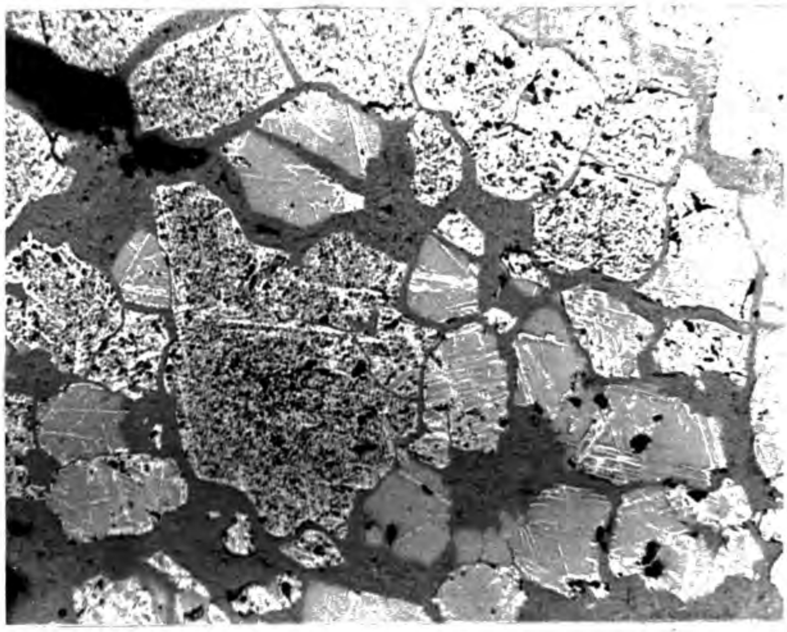


Fig. 50. Highly irregular development of martite (whitish-grey) in adjacent grains of magnetite. Note the pitted surface in the areas of solid martite. The grains have been cemented by goethite. x 125.

distinguishable in polished section, particularly under partially uncrossed nicols. Martite normally develops evenly along all the $\{111\}$ planes, but in rare samples one plane is altered preferentially, the resulting texture being dominantly linear, with a weak lattice texture superimposed on it. This texture is rare, and has been observed only in two or three grains, of which Fig. 51 is a good example.

Goethite is an important constituent of many of the magnetite ores, but this paragraph is only concerned with the goethite formed by alteration of magnetite in situ, and not the more abundant variety deposited from colloidal solutions. Numerous polished sections demonstrate that magnetite readily alters to goethite, but martite is much more resistant. In none of the polished sections is there any visible evidence of the formation of intermediate compounds, and it is concluded that magnetite alters directly to goethite by simultaneous oxidation and hydration. The typical ore shows the original magnetite grains outlined by martite, with their centres partially or completely altered to granular - never colloform - goethite. The final texture is variable, depending on which parts of the magnetite have been replaced. Where the whole periphery of the relict magnetite core has been attached simultaneously the final texture is zonal, showing magnetite → goethite → martite from core to periphery, or there are isolated fragments of magnetite in goethite. In some samples only part of the magnetite has been replaced, leaving a core composed half of magnetite and half of goethite. Figs. 52 and 53 show examples of the replacement of magnetite by goethite. The replacement is irregular, and not controlled by the crystallographic planes in the magnetite.

In a large number of the polished sections which were examined, the peripheral martite rims are complete, and although only a two dimensional view is seen, it is likely that many of the rims are intact over an individual grain. The agents which produced the alteration to goethite must have penetrated through the martite rim to preferentially attack the magnetite core. The facility with which magnetite alters to goethite, and the stability of martite under the same conditions, is convincingly demonstrated in those polished sections which show lattice frameworks of martite set in solid goethite. In Fig. 54 it can be seen

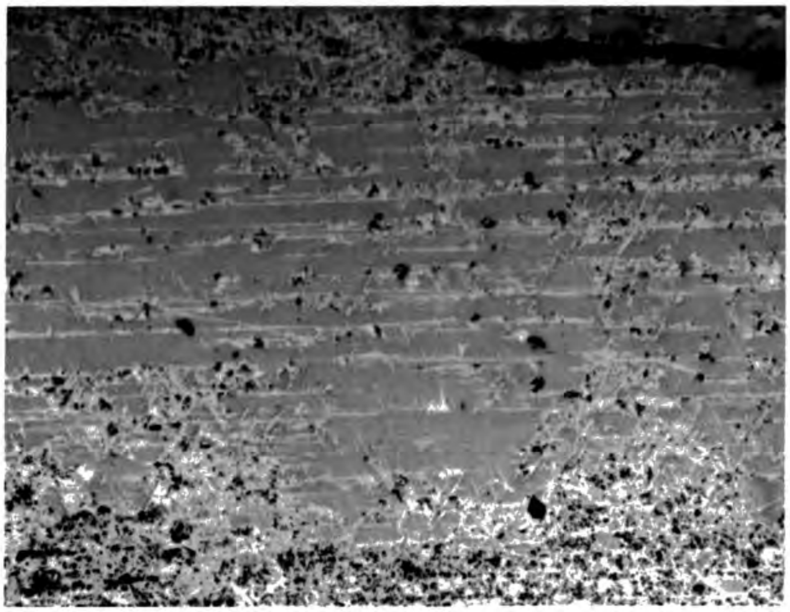


Fig. 51. An unusual linear development of martite (pale grey and horizontal on the photograph) on which a weak lattice texture is superimposed. x 220.

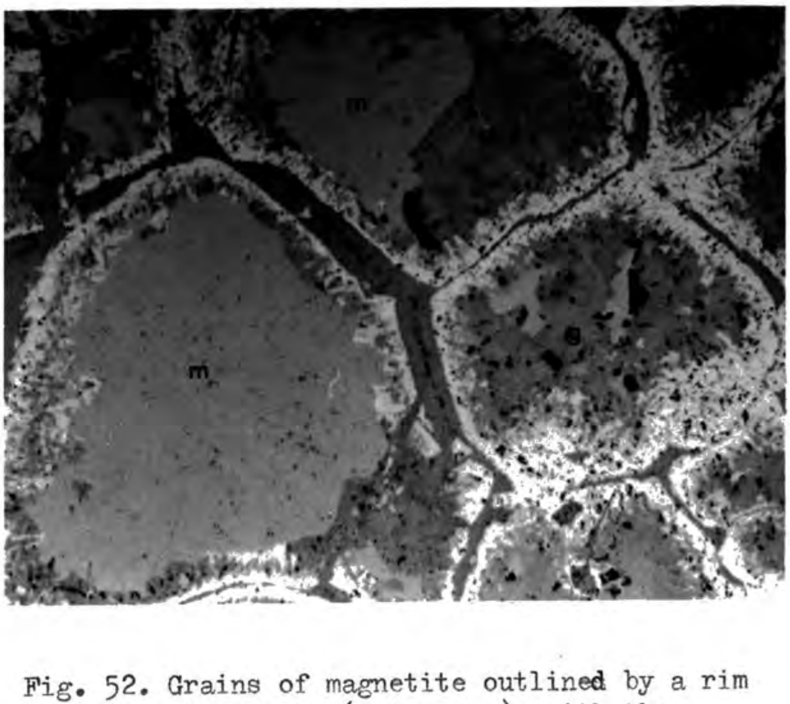


Fig. 52. Grains of magnetite outlined by a rim of martite (pale grey), with the magnetite cores partially replaced by goethite. Note the variation in the development of goethite between adjacent grains. x 220.

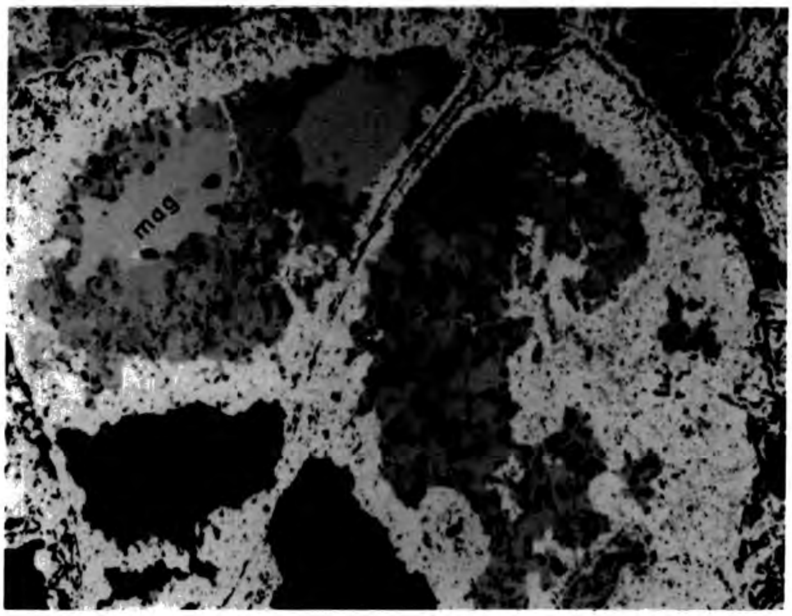


Fig. 53. Magnetite grain with a rim and internal patches of martite (pale grey), and goethite in the centre; the goethite is clearly granular. x 220.

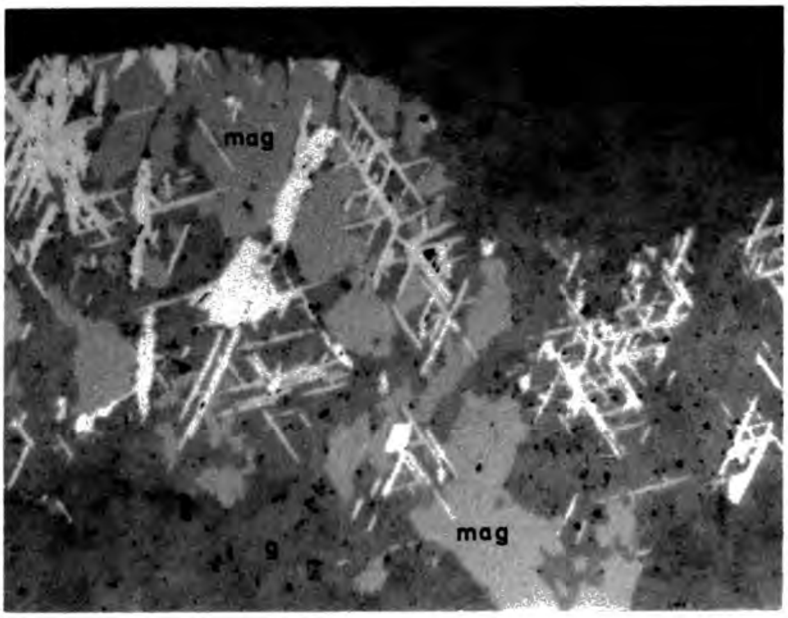


Fig. 54. A photomicrograph which illustrates martite's resistence to replacement by goethite. All the martite lamellae (pale grey) follow one set of {111} planes and have clearly been derived from a single magnetite grain, relics of which still remain. x 550.

that all the martite lamellae are following the $\{111\}$ planes of one magnetite crystal, and that much of the magnetite has been replaced by goethite, leaving the skeletal martite lattices in their original orientation in solid goethite. Such examples prove that magnetite can alter to goethite, and that an intermediate martite phase is not necessary.

A number of magnetite samples show an interesting development of maghemite, although the quantity formed is small. The maghemite usually develops as minute lamellae in the centre of magnetite grains, and although their orientation is clearly controlled by crystallographic planes in the magnetite, they are not the $\{111\}$ planes which control the martite formation. Details of the maghemite occurrences are discussed further under Maghemite - page 132.

Haematite

There are four recognizable varieties of haematite in the Bt. Ibam mine, three in the ore body and one in the footwall skarns. These types are:-

1. Haematite which veins and replaces magnetite (ore body).
2. Secondary haematite (martite) formed by oxidation of magnetite (ore body).
3. Haematite associated with colloform goethite (ore body).
4. Micaceous haematite (footwall skarns).

There appears to be a lack of unanimity in the use of the term "martite". Some authors, e.g. Palache, Berman & Frondel vol. 1 (1944), use the term to describe any haematite which pseudomorphs magnetite or pyrite, but others, e.g. Cooke (1936), restrict it to haematite formed from magnetite by oxidation. Throughout this thesis the writer uses the term martite to describe haematite formed as a result of the oxidation of magnetite.

The haematites in groups 1 and 3 above are rare and have only been observed on a microscopic scale, but martite can form extensive masses. Micaceous haematite is associated with pyrite, quartz, and occasionally epidote, in pods up to 5 ft across in the skarns. Martite takes the physical form of the magnetite it has replaced, i.e. it varies from massive to finely granular. It is easily recognized in the field by its red colour or red streak, and remanent magnetism.

1. Haematite Veinlets and Replacements

This type of haematite is rare, and has only been observed in a few polished sections, most of which originate from the northern half of the ore body and at depths of 200 ft and over. One drill core showed a change from lamellar martite in the higher levels to haematite in veinlets and granular replacements at depth, but this clear succession has not been observed elsewhere.

The haematite has followed incipient small fractures, and penetrated into the mass of the ore along the intergrain boundaries. The veinlets always have irregular edges, due to replacement of the adjoining magnetite - Fig. 55. The intergrain penetrations have started with crystallization in open spaces, but have developed by replacement of the magnetite - Fig. 56. The haematite has advanced progressively into a magnetite grain from the original grain edge or along small cracks, and it is rare to find isolated patches of haematite within a magnetite grain. The replacement of an individual grain is usually only partial, but complete pseudomorphs with small relics of magnetite can be observed. Figs. 56, 57 and 58 show the sequence from initial penetration to almost complete replacement. In some of the polished sections there is no sign of any feeder channel, and the haematite occurs in quite isolated patches.

It can be seen from Figs. 57 and 58 that crystallographic control over the replacement is virtually absent, so the haematite patches tend to be formless, although the original magnetite grain edges are preserved. The junction between the haematite and unreplaced magnetite issharp and smooth in outline. The replacement of magnetite by haematite frequently procedes from one spot on the grain edge, and simultaneous attack of the whole periphery is rare. All these points clearly differentiate this type of

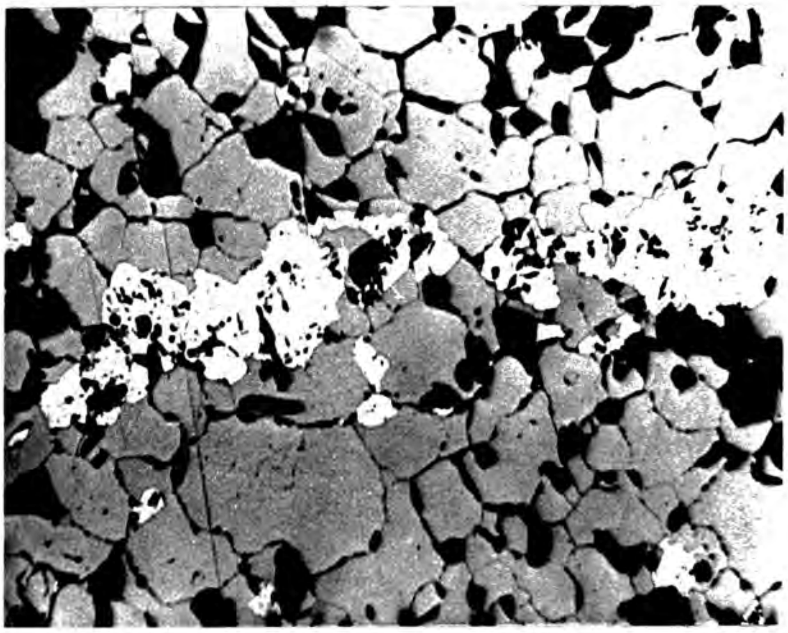


Fig. 55. An irregular haematite veinlet cutting through fresh, granular magnetite. Note the complete absence of martite lamellae x 125.

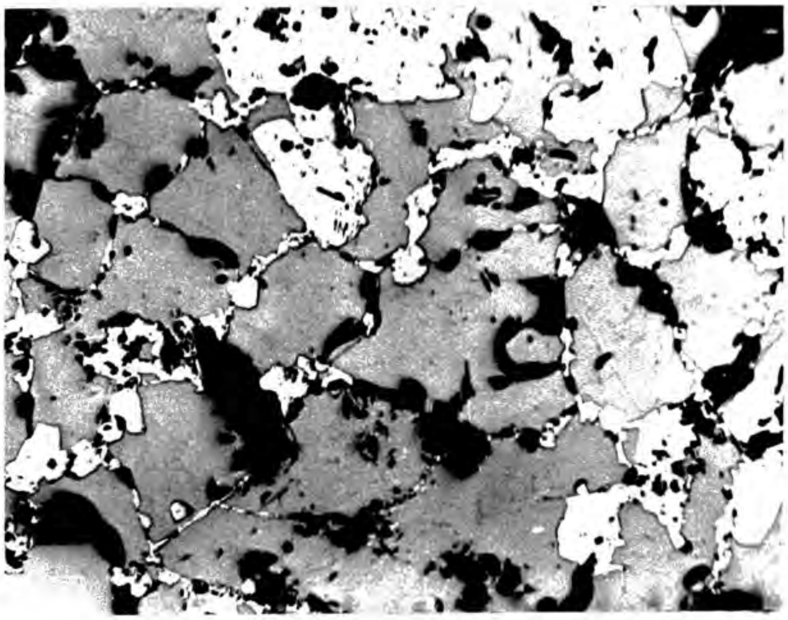


Fig. 56. Haematite penetrating along intergrain boundaries and replacing granular magnetite. It can be seen that some of the haematite has crystallized in open spaces. x 125.

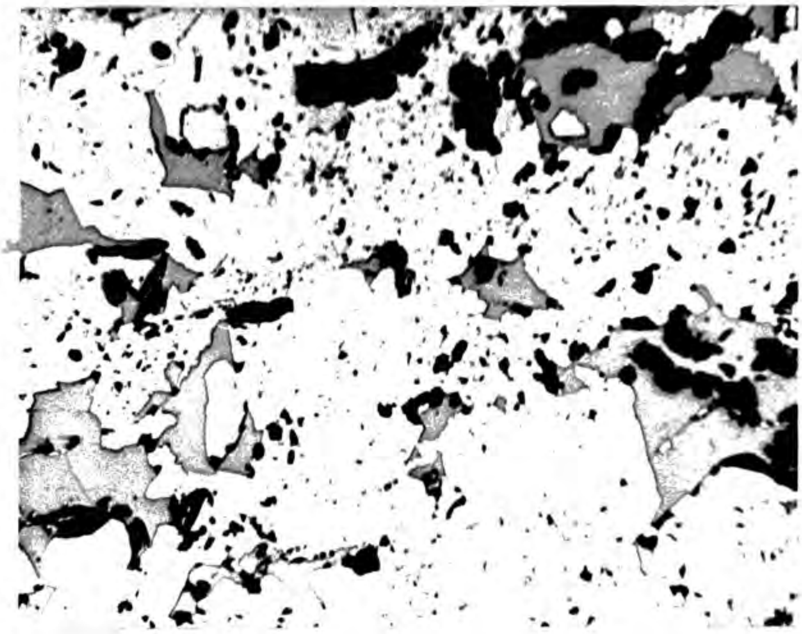


Fig. 57. Granular magnetite which has been largely replaced by haematite. Note the smoothness of the magnetite-haematite interfaces, and the lack of $\{111\}$ control in the magnetite. x 125.

Fig. 58. Small relics of magnetite in haematite. The haematite has preserved the shape of the original magnetite grains. x 540.

haematite from martite.

In polished section it is clear that the replacement has taken place "point by point", and the haematite which has replaced one magnetite grain is itself a homogeneous crystal, with consistent optical properties throughout. On occasions the haematite is twinned, a property never seen in martite, and it can show a "perthitic" texture and, rarely, small exsolved plates of corundum (?). These properties are much better displayed in the massive haematite ore at Bt. Pesagi, and are more fully described later - page 272. There is no doubt that this type of haematite is primary in origin, as some of it has crystallized in open spaces directly from solution, and represents a distinct phase of iron mineralization later than the magnetite. The writer has not seen this type of haematite in association with sulphide minerals, so their relative ages are unknown, but from a study of the paragenetic sequence throughout Ulu Rompin it is almost certain that the sulphides are later.

2. Martite

Martite is a very common mineral and is present in the majority of the magnetite ores examined in polished section, varying from a few scattered lamellae to complete pseudomorphs of magnetite - Fig. 50. It is easily recognized in polished section as it displays lattice textures controlled by the $\{111\}$ planes in the magnetite. Where martite has completely pseudomorphed magnetite the lattice texture is not always visible in plain light, but it becomes apparent with partially uncrossed nicols.

In polished section the martite is white to bluish-white, bireflecting and distinctly anisotropic. The large patches of martite usually have a very pitted, porous surface, which lowers the reflectivity, and this type of ore always shows very strong, deep red internal reflections, which are strongly marked under oil immersion. The pitted surface of martite can be seen in Figs. 35 and 50. Where the lamellae are in an early stage of development there is much less pitting in the martite. Fig. 35 shows pitted martite behind a "front" of lamellae which are advancing into magnetite with only a few, relatively large pits. The pitting is probably due to the fact that the change from magnetite to martite involves a volume increase of just over 5% (Gruner, 1926), which must cause some shattering of the material, the fragments dropping out during polishing. The martite masses are obviously permeable, as in some grains solutions have passed through the martite to attack the magnetite core.

The ends of the martite lamellae are often ragged and irregular, which contrasts with the maghemite lamellae which are finely tapered; the difference can be clearly seen in Fig. 62. Some of the martitized magnetite ores contain small discs of a white material which is transparent or translucent, and reflects light split into spectral colours; the total effect looks like Newton Rings. These are thought to be caused by minute plates of martite on the polished magnetite surface and parallel to it. One of the photographs shown as Fig. 8 by Cooke (1936) is of an artificially oxidized magnetite, and this contains discs which obviously have refracted light into colour bands round their edges. Even in the magnetite grains which look to have been completely martitized, it is usually possible to find small relics of unaltered magnetite. These cause the magnetism in the martite ores.

The writer is convinced that martitization was the main agent which produced the highly friable granular ores. If a polished section of this type of ore is examined, it is observed that the grains have a rim of martite, and some have been completely replaced. The development of martite would break the adhesion between the magnetite grains, because, as stated earlier, the reaction involves a volume increase, which must cause internal stresses. What differentiates the friable from the typical granular ores is the lack of goethite as a cementing medium. In the hard granular ores the partially martitized grains have been welded into solid masses by intergranular goethite - Fig. 48. All the fine grained magnetite ores must have been potentially friable following martitization, but the entry of goethite has cemented most of them into solid masses.

3. Haematite Associated With Colloform Geothite

This type of haematite is only found in the secondary ores which contain well-developed colloform structures. The haematite forms bands and spherical patches of higher reflectivity in the goethite, and there can be alternating bands of haematite and goethite. The colloform texture is present in both minerals - Fig. 59 - and they must have a common mode of origin, i.e. from colloidal solutions. Haematite can form from goethite by dehydration, but such an origin is rejected for the Bt. Ibam secondary ore for two reasons. It is most unlikely that dehydration could preferentially follow certain bands to preserve perfectly the colloform texture, especially as some of the haematite bands are very thin, and there is clear evidence that haematite has replaced earlier deposited goethite, and vice versa. This is well shown in Fig. 59, where the youngest goethite has obviously penetrated into what were originally continuous bands of haematite. The alternating bands of haematite and goethite must represent fluctuations in the environment of deposition. Both haematite and goethite can crystallize from supergene solutions, and among the factors which govern which particular mineral forms are pH, Eh, concentration, other ions in solution, ageing of gels and temperature (Valeton, 1972).

4. Micaceous Haematite

This type of haematite is only found in parts of the footwall skarns, where it is associated with quartz, pyrite, and sometimes epidote, in isolated pods. It is not found in the beds which immediately underlie the ore body, but which contain some fine disseminated magnetite. The mineralized pods are irregular in size and distribution. The haematite forms soft masses made up of very thin scales, which are deep purplish-red in colour.

In polished section the haematite occurs as parallel and radiating sheaves of very thin, elongated crystals - Fig. 45. Very few of the individual crystals are perfect, but contain unreplaced fragments of gangue, and often have irregular edges. Among the long blades there can be many much smaller, irregular grains. Most of the haematite is untwinned, but some shows very good twinning on $\{10\bar{1}1\}$ - Fig. 60. It is interesting that in some cases the twin lamellae can be traced across several individual haematite blades without deviation, even though the blades themselves are not all parallel. This suggests that the elongation of the blades

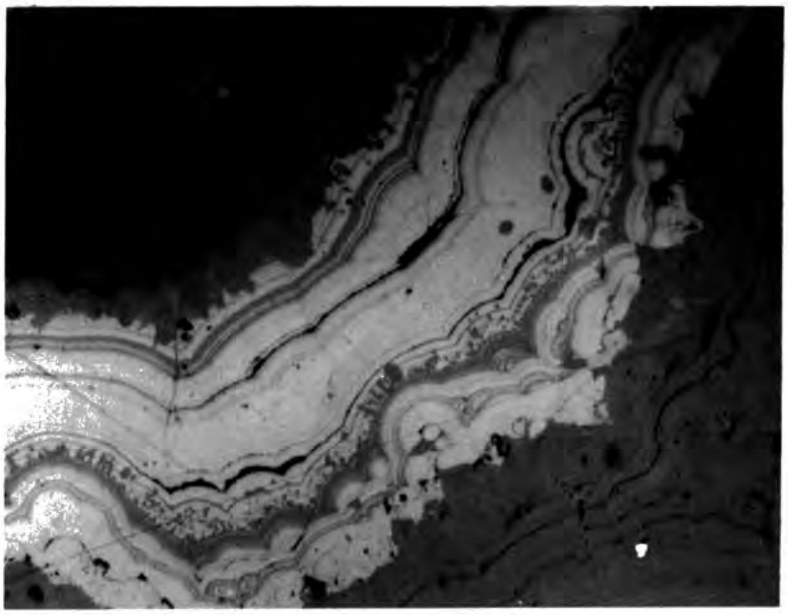


Fig. 59. Colloform banding in haematite (pale grey) and goethite (dark grey). The black patch top left is a cavity. x 560.

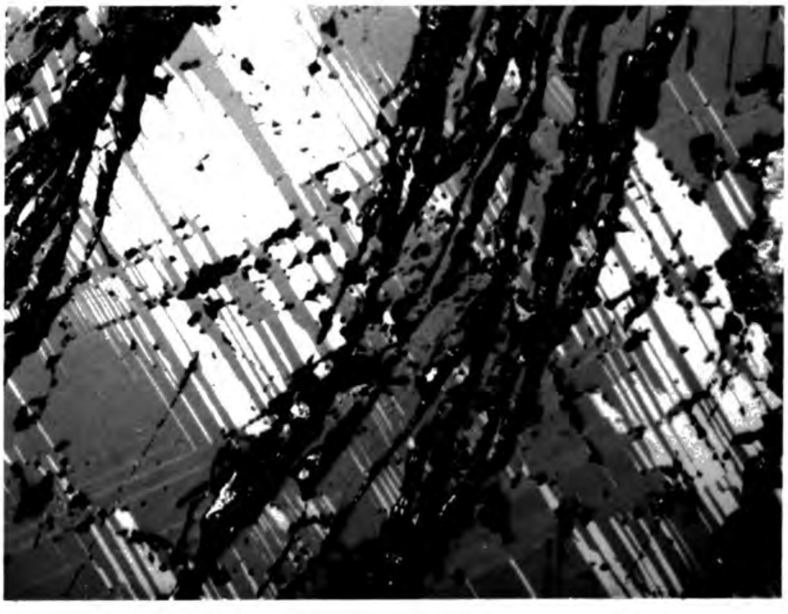


Fig. 60. Twinning on $\{10\bar{1}1\}$ in micaceous haematite. x 530.

is not always perfectly parallel to (0001). Rare examples of the haematite show a "perthitic" texture, which is described more fully under the Bt. Pesagi deposits - page 272. A haematite blade which had been partially altered to magnetite was seen in one polished section and has been described previously - page 112.

The micaceous haematite is associated with pyrite, but their relative age is not always clear. The contact shown in Fig. 45 could indicate that the long blades of haematite have grown parallel to a {100} face of a pre-existing pyrite crystal, but the balance of evidence suggests that pyrite is later. Certainly, what unequivocal evidence there is, shows pyrite later than haematite. Pyrite has been observed penetrating between the haematite blades, and rare pseudomorphs of pyrite after haematite have been seen. Apart from the unique example mentioned above, micaceous haematite has not been seen in contact with magnetite, but indirect evidence suggests that magnetite is older than haematite. In the Bt. Pesagi deposits it can be proved that specularite and micaceous haematite crystallized after the magnetite.

Diffraction Patterns for Haematite

Table 11 shows the diffraction patterns for two haematite samples, one of which was martite formed from friable, granular magnetite, No. 39, and the other micaceous haematite, No. 50, with an ASTM pattern for comparison.

Maghemite

Although maghemite only occurs in minute amounts and is always of microscopic dimensions, it shows some interesting textural features. The mineral has been identified in 18 magnetite samples, but in many of these it is sparsely distributed. It has been located at depths up to 380 ft below surface, but its distribution is irregular. The best examples were located in drill core from N 11,000, between 90 ft and 200 ft from surface.

Table 11

Diffraction Pattern for Haematite

Co Radiation

39		50		ASTM 13-534	
dÅ	Rel.I	dÅ	Rel.I	dÅ	Rel.I
3.68	3	3.68	6	3.66	25
2.70	10	2.70	10	2.69	100
2.52	8	2.52	7	2.51	50
2.281	*	2.288	½	2.285	2
2.205	4	2.206	6	2.201	30
		2.075	1	2.070	2
1.840	3	1.840	6	1.838	40
1.695	6	1.693	7	1.690	60
1.638	½	1.633	2	1.634	4
1.600	*	1.598	3	1.596	16
1.487	4	1.485	5	1.484	35
1.454	4	1.452	6	1.452	35
1.350	½	1.349	1	1.349	4
1.310	1	1.310	3	1.310	20
1.256	1	1.259	2	1.258	8
1.228	½	1.227	½	1.226	2
		1.213	½	1.213	4
1.181	½	1.190	1	1.189	8
1.163	½	1.163	1	1.162	10
1.148	½	1.141	1	1.141	12
1.109	1	1.103	2	1.102	14
1.081	½	1.077	½	1.076	2
1.056	1	1.056	2	1.055	18
		1.043	½	1.042	2
1.035	½	1.040	½	1.038	2
+4 lines		+7 lines		+8 lines	
a = 5.05Å		a = 5.04Å		a = 5.034Å	
c =13.74		c =13.75		c =13.752	
a_r= 5.43		a_r= 5.43			
= 55°24'		= 55°10'			

* = enhanced by other line.

The presence of maghemite has been confirmed by X-ray diffraction, but it is usually easy to identify by its optical properties in polished section. It is always found in a magnetite host and is distinctly bluish-grey in colour, especially under oil immersion, with a reflectivity between that of magnetite and haematite - Fig. 66. The blue colour is much more marked than in haematite. Much of the maghemite is too small to determine all its optical properties accurately, but some displays a weak anitsotropism, which suggests that a little haematite is present, as pure maghemite should be isotropic. No bireflectance or internal reflections have been observed. The most interesting feature of the maghemite is its textural relationship to the host magnetite, and to the martite which usually accompanies it.

It is surprising to find that the maghemite and martite within a single magnetite grain are rarely in contact; martite typically occupies a peripheral position, while maghemite is found in the centre of the magnetite grain - Fig. 61. There are isolated examples of maghemite in contact with martite, but in these the maghemite is in small formless masses, itself an unusual feature - Fig. 66. The typical maghemite is found in the unaltered centre of magnetite crystals as thin lamellae 1μ to 20μ long, and frequently an elongated lens shape, with their orientation clearly controlled by crystallographic planes in the magnetite. It is interesting that these planes are not the $\{111\}$ planes which control the martite, and that although the maghemite usually follows three crystallographic directions the resulting texture is quite distinct from that of the martite. The latter forms geometrical lattice frameworks which are triangular, square or rhomb-shaped, depending on the orientation of the magnetite within the polished section. Maghemite on the other hand does not normally form regular latticeworks, but occurs as isolated or parallel groups of lamellae, and when lamellae do coalesce they tend to join at one point, forming open, symmetrical "Y" shaped structures. The difference between the martite and maghemite textures is clearly illustrated in Fig. 62. This figure, and Figs. 61 and 64, also reveal a texture which martite never shows, in which the maghemite occurs in en echelon, parallel, spindle-shaped lamellae. There is no doubt that this texture is much more typical of exsolution than oxidation. However,

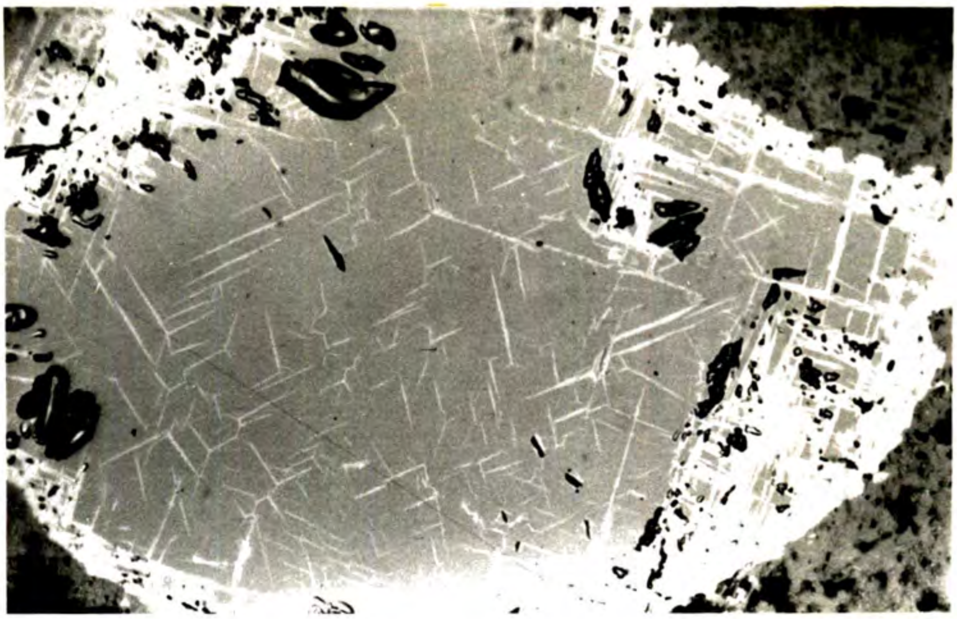


Fig. 61. A magnetite grain with peripheral martite which shows an almost square lattice texture, and fine maghemite lamellae in the centre. Note the difference in orientation of the maghemite and martite lamellae, and the presence of en echelon spindles of maghemite. x 550.

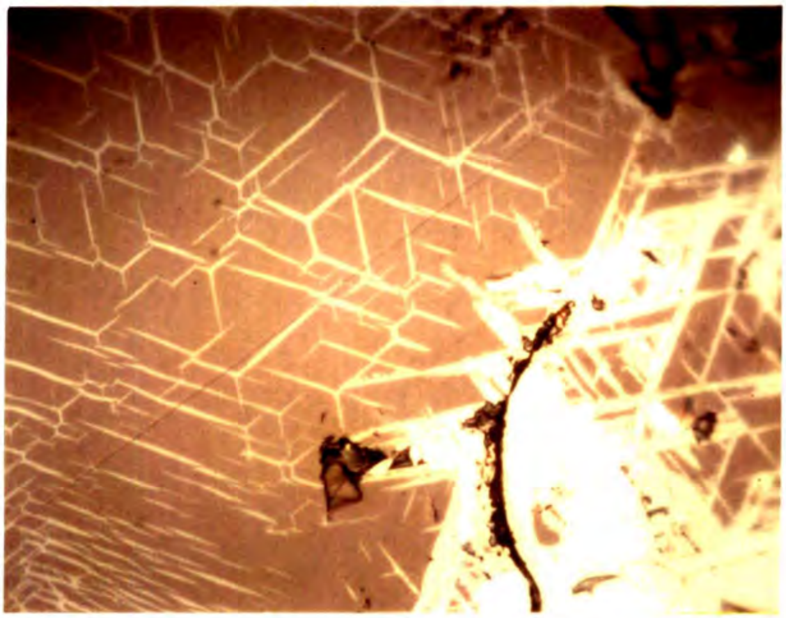
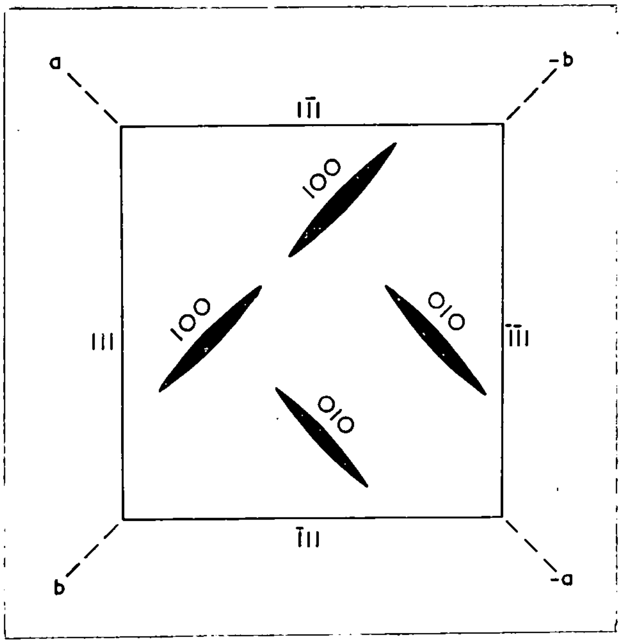


Fig. 62. A photomicrograph which displays the difference between martite and maghemite in magnetite. The martite lamellae form a triangular lattice texture, while the maghemite tends to form "Y"-shaped bodies. The ragged ends of the martite lamellae and smooth outline of maghemite spindles are well shown. x 1350.

if the point where two or more maghemite lamellae coalesce is examined carefully, it is usually observed that there is no marked thinning of the lamellae, a feature which is common in true exsolution lamellae (Schwartz, 1942). It has not been proved conclusively which crystallographic planes the maghemite lamellae follow, but the writer is convinced it is $\{100\}$. In Fig. 63 it can be seen that the peripheral martite alteration produces a square-shaped pattern, which indicates that the magnetite surface is $\{100\}$ and the observer is looking directly down a crystallographic axis. It can be further observed that there are only two directions, almost at right angles, which the maghemite lamellae follow, and these will be 100 planes if the crystallographic axis is vertical to the plane of the photograph. Similar reasoning can be applied to Fig. 61. A diagrammatic representation of the writer's interpretation of the sample in Fig. 63 is shown below. It is worthy of note that when ulvospinel (a spinel with composition Fe_2TiO_4) exsolves from magnetite it follows $\{100\}$ planes (Ramdohr, 1953).



In contrast to the writer's evidence for widespread crystallographic control over the development of maghemite lamellae, van Rensburg (1966) states that in a titanium-free magnetite deposit

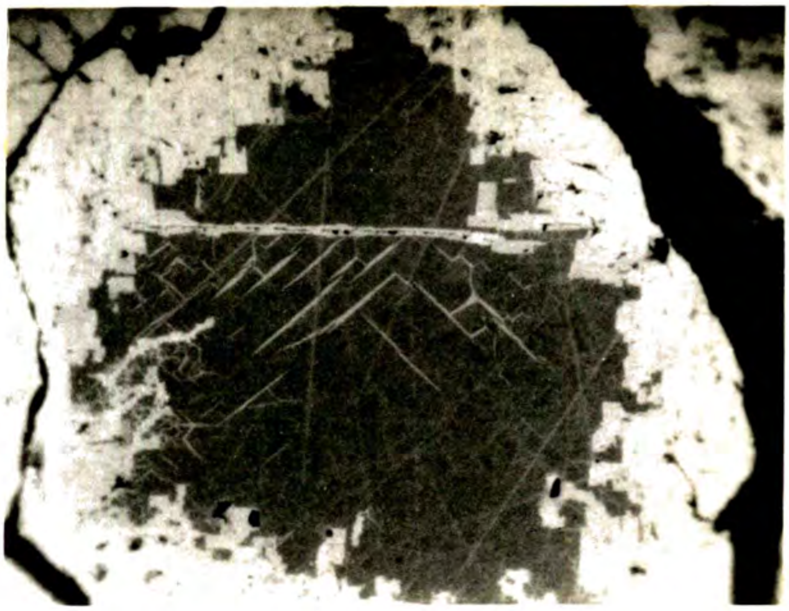


Fig. 63. Magnetite grain with peripheral martite and a few maghemite lamellae. The martite forms a square pattern which indicates that the polished section is a {100} surface. The maghemite follows two directions which are probably {100}. Martite has formed along a small crack (horizontal in the photograph). x 1300.

Fig. 64. Maghemite lamellae in a peripherally martitized magnetite grain. Note the en echelon spindles, top centre. x 1360.

in the Transvaal, Republic of South Africa, "The alteration of magnetite to maghemite takes place irregularly in contraśt with the very regular alteration to martite". Similar comments are made by McLeod (1970) on some Canadian occurrences of maghemite (see below), and Basta (1959) says "The Oxidation of magnetite to maghemite is generally independent of the crystallographic directions of magnetite in contrast to the oxidation of magnetite to haematite - the so-called martitization process".

The development of maghemite varies from a few isolated lamellae - Fig. 63, to zones which are crowded with small lamellae - Fig. 65. The lamellae can be fairly regularly distributed - Fig. 61, or confined to certain areas in the magnetite. On rare occasions the maghemite does not occur as lamellae, but as masses with no detectable texture, although there can be typical lamellae elsewhere in the same grain. Fig. 66 shows a good example of solid maghemite lying between some typical lamellae and lattice martite. It is not easy to determine from the evidence in this polished section if maghemite is replacing the haematite (martite) or vice versa, but on theoretical grounds it is most unlikely that maghemite can replace haematite as an unstable phase would then be replacing a stable phase. However, an examination of the extreme lower portion of Fig. 66 shows martite lamellae engulfed in a mass of maghemite, and it is not unreasonable to interpret this evidence as showing martite in the process of replacement by maghemite. McLeod (1970) studied maghemite in a number of Canadian iron ore deposits and shows a photomicrograph (Fig. 3B) which is very similar to Fig. 66. It is interesting that in the Canadian occurrences it is rare for the maghemite to follow crystallographic planes in the magnetite, and McLeod comments on the specimen mentioned above "this section is unusual in that some magnetite development also follows the magnetite parting planes".

A further variety of maghemite is found in a few magnetite samples which contain irregular patches with a faint bluish tinge, which are thought to be due to the incipient formation of maghemite. The blue colour is often so delicate that it is difficult to to photograph, but Fig. 35 shows a good example. This type of ore is almost certainly the "blue magnetite" described by some early workers in the field of reflected light studies of iron

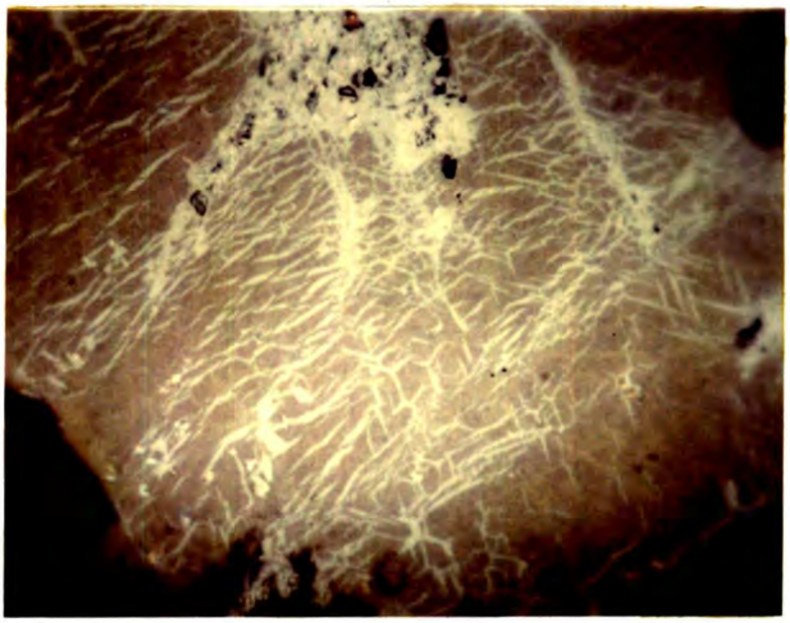


Fig. 65. Photomicrograph which illustrates the dense development of maghemite lamellae in magnetite. The pitted mass, top centre, is martite. x 1350.

Fig. 66. Solid patch of maghemite as well as lamellae in magnetite. The triangular lattice texture of martite can be seen on the left and, faintly, along the bottom. The irregular dark patches are pits in which there are two minute grains of sulphide. The difference in reflectivity of the three minerals is clearly shown. x 1500.

ores, e.g. Callahan & Newhouse (1929) for the Cornwall, Pennsylvania deposit. The presence of irregular bluish patches in magnetite has been reported in some Canadian iron ores (McLeod 1970), and the conclusion reached is the same as the writer's, viz. "The blue magnetite phase presumably represents an intermediate stage in the oxidation of magnetite to maghemite".

Figs. 61 to 64 show that there is no invariable connection between maghemite formation and cracks or pits in the magnetite, and that much of the maghemite is found where the magnetite surface is perfect. In a polished section of a badly pitted single magnetite crystal however, it was observed that there was a development of very delicate maghemite networks between some adjacent pits.

Goethite

There are two recognizable varieties of goethite:-

1. is distinctly granular and has pseudomorphed magnetite and haematite, and sometimes pyrite.
2. consists of finely granular to cryptocrystalline colloform ore, deposited from colloidal solutions.

Goethite is a common mineral which reached its most massive development in the upper levels of the ore body. It is also common just above the footwall contact, where it can form sheets up to 2 ft thick, which mark the base of the ore body, along sections of the hangingwall between N 11,000 and N 11,600, and it is abundant at the southern end of the ore body. The goethite near the footwall and hangingwall contacts has cemented the ore and gangue into solid masses. Goethite is widespread as a staining and impregnating mineral in the gangue.

The ore in the larger masses of goethite varies from a hard, dense, metallic variety, to highly porous and earthy types. The porous ore shows many botryoidal surfaces with cellular, ropy and stalactitic masses. The external surfaces are usually a dull yellowish-brown to rusty brown, but can be shiny, and some are irridescent. In the upper levels of the ore body goethite has been found as pseudomorphs of euhedral magnetite and pyrite crystals.

Goethite has played an important role in the development of the distribution pattern of the impurity elements in the ore body, as due to the mobility of the solutions which deposited the iron hydroxides it has acted as a dispersing agent, particularly for Cu and Zn.

In polished section goethite varies from cryptocrystalline to grains 0.3 mm across, and shows a variety of textures. It is dark grey to medium bluish-grey in colour, and exhibits a wide range of reflectivities, which are affected by the crystallinity, the degree of surface pitting, and the presence of dusty inclusions. The internal reflections are yellowish-brown to brown, and some have a faint reddish tinge. Internal reflections are not always visible, the dense cryptocrystalline ores in particular often lacking them. The goethite is bireflecting and anisotropic, both properties being more easily distinguished in the granular ores, and under oil immersion. Where goethite has replaced magnetite in situ, it forms a mosaic of anhedral grains up to 0.01 mm across - Figs. 52 and 53. This type of goethite is bluish-grey in colour and has a very pitted surface. There seems to be a relationship between the grain size of the goethite and the pits; the coarser grained goethite contains cavities which are of the roughly the same size as the grains themselves and irregularly distributed - Fig. 53, whereas the finer grained ore contains numerous small pits which are regularly distributed throughout the mass - Fig. 49. The goethite does not appear to be controlled by crystallographic planes in the host magnetite. The goethite found in veinlets, which are usually controlled by fractures and intergrain boundaries, can be finely granular, but it also develops with a fibrous texture which is perpendicular to the axis of the vein. The sides of the veins can be sharp, or irregular where replacement has taken place. The goethite has often followed fractures which contain some martite, the end product being a compound vein of martite with a goethite core. The intergranular goethite can be a simple filling of open space, but more frequently there has been a dilational force, which has separated, and sometimes shattered, the originally interlocking grains - Fig. 52. Many of the magnetite ores have been firmly cemented by intergranular goethite - Fig. 36. It is not uncommon to find an intermittent cavity along the centre of goethite veins and intergranular masses; copper can be found in these cavities in the oxidized cupriferous zones - Fig. 82.

Goethite readily replaces magnetite, but martite is much more resistant, and relics of martite lamellae are frequently preserved in goethite when all, or most of, the magnetite has been replaced - Figs. 49 and 54.

A distinctive texture is found in a number of samples from section N 10,400, where granular siderite has been replaced by goethite. In the larger grains the goethite forms a corona, but the smaller grains have been completely replaced - Fig.95 . This material has only been found in drill core, and in hand specimen it looks like lumps of nodular goethite (as it is so described in the company's drill logs), and its true nature is not apparent until it is seen in polished section.

The goethite ores from the upper levels of the ore body are typically non-granular, and are largely composed of colloform masses, in which it is possible to detect relics of martite and magnetite. The ores vary from dense and massive to highly porous, cellular and vughy. The banding in the colloform masses is delineated by zones of slightly different colour or reflectivity, lines of minute pits and/or inclusions, and intercalated haematite bands. The banding can be extremely delicate with bands down to 1μ, or coarse with bands up to 1 cm. In the coarsely banded ores the goethite has a radial fibrous texture, and shrinkage cracks are not uncommon. The colloform masses can contain perfectly concordant, parallel bands, but in some samples it is evident that there has been attack and replacement of the earlier deposited material which has broken the continuity of the banding - Fig. 59. Some of the banded colloform goethite is cut by veins of later goethite, while other masses have been brecciated and the fragments cemented by goethite. The fragments produced by brecciation vary from particles of microscopic size, up to angular blocks over 1 ft across. The brecciation is a late-stage feature, and it is thought to be connected with slumping movements in the ore body.

Some of the porous goethite ores contain thin veinlets and cavity fillings of an unidentified manganese oxide. There are minute specks of sulphide, seen to be pyrite when it is identifiable, in cavities and along shrinkage cracks, which means that this sulphide is of relatively recent origin.

Goethite can be intimately associated with the secondary copper minerals and native copper, and in some samples there is a genetic connection. The secondary copper sulphides and cuprite can be earlier, contemporaneous with, or later than the goethite, but native copper always seems to be later than any associated goethite.

Goethite, and its associated 'limonite', is responsible for the iron staining and impregnation in the gangue and wall rocks. It is found as veins along joints and faults, often cementing the fragments in the latter. In some parts of the gangue there must have been an influx of iron carried in solution, as large masses are stained and impregnated, but elsewhere the goethite is found only round primary iron minerals, from which the iron has been leached.

Table 12 shows the diffractometer pattern for three goethite samples, one of which was granular while the other two consisted largely of colloform ore. The three patterns were virtually identical and have been averaged for the table. In spite of a slight broadening, the peaks on an X-ray diffractometer trace are clear and pronounced, indicating that the goethite is distinctly crystalline, even in the colloform ore.

The d.t.a. curve for goethite shows a sharp endothermic reaction with a peak at 364°C, and an inflection at about 330°C is possibly due to a small percentage of more amorphous "limonite". There is a break in the curve starting at 920°C, which probably marks the commencement of the breakdown of the structure.

Table 12

Diffraction Pattern for Goethite

Co Radiation

Bt. Ibam		B. & T.	
dÅ	Rel.I	dÅ	Rel.I
4.97	17	5.02	2
4.17	100	4.21	10
3.38	11	3.37	2
2.694	35	2.69	8
2.578	12	2.57	2
2.517	9		
2.488	10	2.48	2
2.449	45	2.44	7
2.254	12	2.25	2
2.191	25	2.18	4
2.095	5	2.09	½
		2.00	1
1.921	5	1.920	1
1.802	5	1.803	2
		1.774	½
1.718	20	1.719	5
1.691	10	1.689	2
a = 4.68Å		a = 4.596Å	
b = 9.96		b = 9.957	
c = 2.93		c = 3.021	

B. & T. is Berry and Thompson (1962)

X MINERALOGY OF BUKIT IBAM. (2) PYRITE & CUPRIFEROUS MINERALS

In addition to the iron oxides and hydroxides a wide variety of minerals have been detected in the ore body and its immediate host rocks, but only pyrite is abundant, although copper-bearing minerals are locally important and contaminate parts of the iron ore. The only primary copper mineral is chalcopyrite, but much of the copper is now contained in secondary products.

Pyrite

Pyrite is widely distributed throughout the Bt. Ibam area, being found in the ore body and the wall rocks. In size it varies from grains of a few microns to individual crystals up to two inches, and aggregates which can reach three feet across. Some of the larger crystals are extremely well formed, a few being almost perfect cubes, while smaller modified pyritohedra are not uncommon. The euhedral crystals are found in gangue or low grade ore, in some of the chloritic rocks along the hangingwall contact, and in quartz veins. Some of the crystals have a dull black surface due to a thin coating of secondary copper sulphides, and it is interesting that these crystals can be found in situ adjacent to perfectly normal, untarnished pyrite. Individual crystals are either completely black or completely untarnished, and the fact that the two types are intermingled is prima facie evidence for a mineralization sequence,

pyrite → attack by secondary copper sulphides → pyrite.

This sequence would readily explain the occurrence of the two types side by side, but there is no supporting evidence for a second phase of pyrite mineralization in any of the polished sections studied, and it is consluded that the black coated pyrites have resulted from selective attack by cupriferous solutions.

Microscopic Properties

The optical properties of the pyrite in polished section are completely normal, and no variations from standard textbook descriptions have been observed. A small number of grains exhibit a weak anisotropism, but the majority are isotropic. Some of the pyrite grains have a perfect polished surface, but the majority have some flaws due to cavities and/or fractures. Cavities are not uncommon, and in the larger examples it can be seen that they are lined with small, interlocking fragmentary pyrite faces. Most of the cavities are irregular or rounded in outline, but a few have been controlled by crystallographic planes and are negative crystals. The cavities have sometimes been filled by later minerals. A distinctive type of cavity has been observed in a few samples of pyrite in massive magnetite. The cavities are minute, elongate voids (some have been filled with later mineral), which form swirling "trains" through the pyrite - Fig. 67. The gently curved pattern of the cavities suggests that the pyrite may have been rotating during growth.

Few pyrites are without some sign of fracturing, which varies from hairline cracks to completely shattered crystals. Most of the fractures are irregular, but in some grains there has been a strong crystallographic control - identified as $\{111\}$ in some examples - producing rectilinear networks - Fig. 74. Occasionally two pyrite grains have impinged upon one another producing intense comminution at the point of impact, with a series of cracks radiating from this zone. Fractures have proved to be favourable channels of access, and almost all the post-pyrite minerals, and some gangue material, have been seen as fracture fillings. The causative stresses usually appear to have been of local origin as completely shattered rocks are rare, the brittleness of pyrite probably making it susceptible to relatively slight stresses.

* * * * * * * * *

Pyrite is found in a variety of host materials, and it is convenient to describe it under headings which correspond to the four main environments.

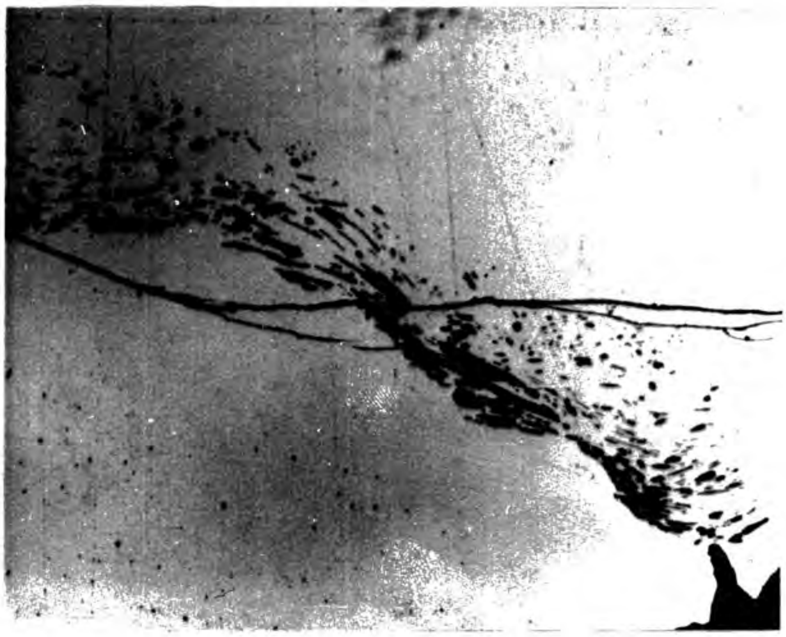


Fig. 67. Small cavities which have a gently curved pattern in a pyrite host. x 220.

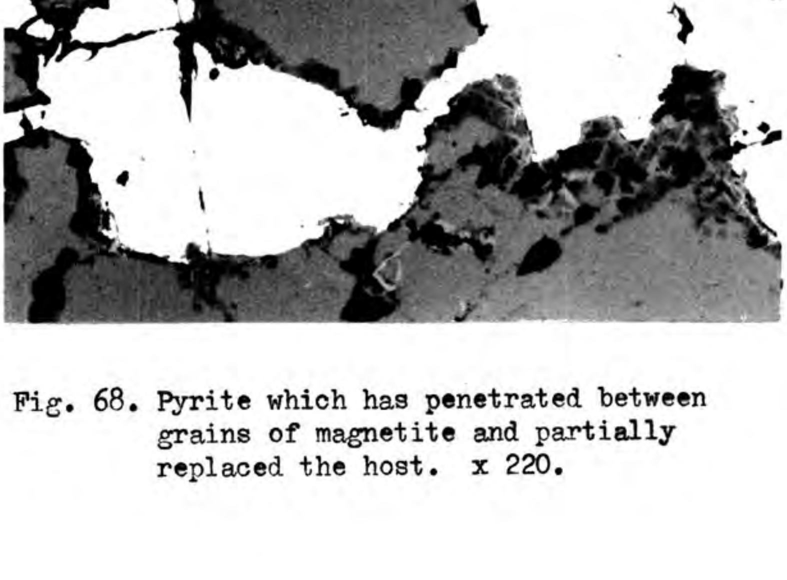


Fig. 68. Pyrite which has penetrated between grains of magnetite and partially replaced the host. x 220.

1. Pyrite in Massive Iron Ore.

Pyrite is found as individual crystals up to half inch across, as veinlets up to quarter inch wide, and as crystalline aggregates up to three feet long. Polished sections show that the pyrite is later than magnetite. The veinlets can be simple fillings of fractures or intergranular spaces, but the larger veinlets have developed by replacement of magnetite. The replacement appears to start from an intergrain penetration or fracture filling and develops laterally - Fig. 68. There are occasional relics of magnetite, but usually the pyrite quickly clears itself of inclusions. The replacement proceeds at a relatively constant rate along the full length of both margins of the vein and perpendicular to its long axis, which preserves the veinlet form. The replacement tends to proceed grain by grain, so that the edge of the pyrite veinlet can preserve the outlines of the original magnetite grains. The replacement is never controlled by the 111 planes in the magnetite. Isolated grains of pyrite have crystallized in cavities or in small gangue patches in magnetite - Fig. 69.

The solid masses of finely crystalline pyrite are found in fine grained friable magnetite, and have developed by crystallization in intergranular spaces and replacement of the adjacent grains. These pyrite masses are formless in outline.

Pyrite is found in some of the massive colloform goethite ores as relics, and as small grains in cavities. The latter are puzzling, as their presence in the cavities and cracks in goethite implies that they have crystallized very late, and certainly after the main phase of pyrite mineralization, and within the main oxidation zone although local conditions may have been anomalous.

2. Pyrite in Gangue and Low Grade Ore.

The pyrite occurs as isolated grains and irregular patches, and veinlets appear to be absent. Most of the good euhedral crystals have grown in the silicate gangue, but much of the pyrite is anhedral and some is highly poikiloblastic. It is possible to find very thin disconnected bands of pyrite which may have crystallized under some form of lithological or structural control, the evidence of which

Fig. 69. Pyrite and chalcopyrite which have crystallized in cavities in magnetite, and partially replaced it. x 535.

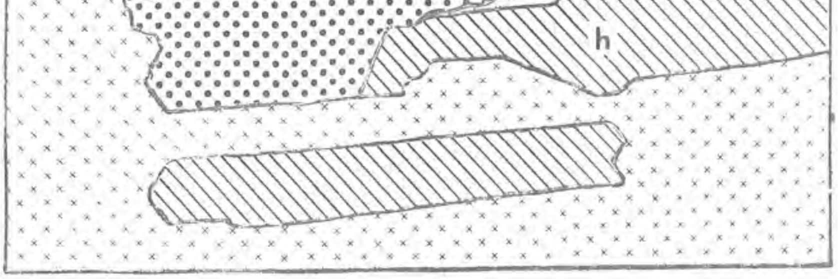


Fig. 70. Drawing taken from a polished section which shows pyrite which has replaced part of a blade of micaceous haematite. x 500 (approx).

is now completely destroyed. The highly irregular shape shown by some of the microscopic sized pyrite grains is partly an original feature, but it has also been accentuated in places by attack from the gangue material. Fractures filled with gangue and the jagged edges of some pyrite demonstrate the activity of the gangue; this probably occurred when the original silicates were altering to talc and associated minerals. The poikiloblastic appearance of some of the pyrite is accentuated by the presence of replacement bodies of sphalerite, which tend to be much larger than those in pyrite grains in high grade iron ore.

In the low grade iron ore the pyrite can grow round the magnetite or develop in cavities and gangue inclusions, but there has also been some replacement. The pyrite does not normally pseudomorphously replace the magnetite but tends to develop its own crystal shape, and this can cause some ambiguity with regard to age relationships as it is possible to see magnetite alongside a well developed {100} pyrite face. At first glance, and taken in isolation, it appears to show magnetite which has crystallized round an earlier pyrite, but the evidence as a whole quite clearly shows that pyrite is later than magnetite.

In some sections of the hangingwall contact there are very good euhedral crystals of pyrite - cubes and pyritohedra - in the massive chloritic rocks and also in quartz veins.

3. Pyrite in Skarn.

Although there is some disseminated pyrite scattered through the skarn, it typically occurs intimately associated with micaceous haematite and quartz in pods of mineralization. The pyrite can show some semblance of crystal shape, but it is usually in anhedral masses. A study of polished sections shows that the evidence for the relative age of the pyrite is not always unequivocal, and may indicate that there have been repeated and overlapping phases of mineralization. One specimen shows blades of micaceous haematite parallel to a {100} face in pyrite - Fig. 45 - which at first sight appears to indicate that the haematite has crystallized later than the pyrite. Other specimens however can show relics of haematite in pyrite grains, and pyrite penetrating between haematite blades.

Fig. 70 shows haematite crystal faces preserved in pyrite, and here there is no doubt that the pyrite has replaced the haematite. The writer believes that all the pyrite is later than the micaceous haematite.

The age of pyrite relative to the silicate minerals, and garnet in particular, is ambiguous. Examples of pyrite euhedral towards garnet and vica versa have been observed in polished section, but on a macroscopic scale pyrite appears to have replaced the silicates. Many of the pyrite grains seen in polished section have replaced the gangue, a poikiloblastic texture being common. It is possible that a little garnet formed after the main phase of metallization.

4. Pyrite In Wall Rock.

Pyritization is widespread in the volcanic rocks throughout the Ulu Rompin area, and oxidation and hydration of the pyrite is responsible for much of the reddish and purplish colouration these rocks display. The pyrite is typically disseminated in small grains, up to an eighth inch, and in clusters, but thin stringers are found along some joint planes and fractures. In polished section the pyrite is usually anhedral or subhedral, and often poikiloblastic. In some specimens the pyrite is clearly moulded round earlier silicates, particularly noticeable with muscovite.

An interesting occurrence of pyrite is found towards the southern end of the ore body, and particularly along section N 10,400. Above and below the ore body there is siderite in which there are relics of pyrite which has been partially replaced by sphalerite - Fig. 96. The siderite is a product of hydrothermal activity and is not wall rock proper, but it lies outside the ore body. Further details are given below under "sphalerite" and "siderite".

Alteration of Pyrite.

Pyrite has played a very important role as a host for many important mineralogical reactions, and most of the other sulphides in the Bt. Ibam area are closely associated with it. Many of the post-pyrite sulphides occur as fracture and cavity fillings and/or replacement bodies in a pyrite host. Cuprite occurs as a replacement of pyrite, and goethite is a common alteration product. Details of the minerals which replace pyrite are given later under their respective headings.

In addition to acting as a host for replacement reactions, pyrite has provided sulphate-bearing solutions which have played an important part in the formation of secondary products, as well as some sulphate crystallization.

Pyrite has been attacked by goethite, and it is not uncommon in polished sections to see veinlets, irregular patches, and coronas of goethite. Some of the gangue has also attacked the pyrite, and one can find fractures which have been filled with silicate minerals.

Diffraction Pattern of Pyrite.

A number of pyrite crystals were powder photographed, and their cell sizes calculated. The investigation was initiated to check if pyrites from different environments and with different crystal forms showed any variation in their cell sizes. If any variation had been detected it was hoped to correlate the cell size with other geological factors e.g. phases of mineralization, but the results show that all the cell sizes are virtually the same.

The diffraction patterns for the Bt. Ibam pyrite specimens are not without some interest, as they contain several additional lines, all extremely weak, to those listed in the standard patterns in the ASTM index and Berry & Thompson (1962). The lines are not due to contamination, and calculation of their N values shows that they are valid reflections from a primitive cubic cell. The interpretation of these lines is complicated by the fact that they are not present on photographs taken with Cu radiation, but are always present when Co radiation is employed. The lines are not β reflections, and sample variation cannot account for the difference between the films taken with Cu and Co radiations. as the same

Table 13

Diffraction Patterns and Cell Sizes for Five Pyrite Samples

Co & Cu Radiation

	342		343	79	60	9	B. & T.	
	dÅ	Rel.I	dÅ	dÅ	dÅ	dÅ	dÅ	Rel.I
	3.126	3	3.127	3.129	3.129	3.125	3.12	2
	2.709	8	2.708	2.710	2.709	2.707	2.70	7
	2.423	7	2.422	2.422	2.423	2.425	2.42	6
	2.211	6	2.212	2.210	2.212	2.211	2.21	5
	1.915	5	1.915	1.915	1.915	1.915	1.917	4
+	1.805	1	1.802	1.803	1.803	1.805	1.803	½
	1.633	10	1.633	1.634	1.633	1.632	1.632	10
	1.564	2	1.564	1.563	1.565	1.564	1.563	1
	1.503	3	1.502	1.502	1.503	1.503	1.504	2
	1.448	4	1.448	1.448	1.448	1.448	1.448	3
+	1.354	1	1.353	1.355	det.	1.355		
+	1.313	1	1.313	1.313	det.	1.314		
+	1.276	½	1.276	1.276		det.		
	1.242	2	1.243	1.243	1.243	1.243	1.243	1
	1.211	2	1.212	1.211	1.211	1.211	1.212	2
	1.183	3	1.183	1.182	1.182	1.182	1.184	2
	1.154	2	1.155	1.155	1.155	1.155	1.156	1
	1.106	3	1.107	1.106	1.107	1.106	1.107	3
+	1.083	½	1.084	det.				
+	1.062	1	1.063	det.		det.	1.065	½
	1.043	8	1.043	1.043	1.043	1.043	1.044	8
	1.006	6	1.006	1.006	1.006	1.006	1.007	6
	0.989	5	0.989	0.989	0.989	0.989	0.990	5
	0.958	7	0.958	0.958	0.958	0.958		
+	0.943	½	0.943	det.				
+	0.929	½	det.					
	0.916	1	0.916	0.916	det.	det.		
	0.903	6	0.903	0.903	0.903	0.903		
*			0.8791	0.8788				
*			0.8579	0.8567				

Table 13 contd.

*		0.8264	0.8260			
*		0.8169	0.8167			
*		0.8078	0.8077			
*		0.7989	0.7988			
a =	5.418	5.418	5.417	5.418	5.419	5.419

\+ = not recorded with Cu radiation.

* = recorded with Cu radiation.

det = line present but it cannot be measured accurately.

342 - cube of pyrite. 343 - black coated cube of pyrite.

79 - pyritohedron. 60 - pyrite separated from skarn.

9 - pyrite from massive magnetite ore.

B. & T. = Berry & Thompson (1962).

samples were used. Although Berry & Thompson (1962) do not list all the weak lines obtained by the writer they record two of them, at 1.803Å and 1.065Å. These authors used Fe radiation, and the writer recorded the lines with Co but not with Cu radiation. As the weak lines are consistently present on films taken with Co radiation, and the calculated N values are nearly whole numbers, it is concluded that they are valid pyrite reflections. The reason for the absence of the lines on films taken with Cu radiation is as yet unexplained.

Table 13 shows the pattern and cell size for five pyrite samples from Bt. Ibam; the patterns and cell sizes are virtually identical. Deer, Howie & Zussman vol. 5 (1962) state that the size of pure synthetic pyrite varies from 5.416Å to 5.419Å, and as can be seen, all the Bt. Ibam samples fall within this range.

Chalcopyrite

Chalcopyrite is the only primary copper mineral in Bt. Ibam, where it is widely distributed but rarely found in rich concentrations. Although secondary copper sulphides slightly complicate the pattern, it is usually found that samples with a significant copper content contain chalcopyrite, the main exception being the cupriferous limonitic ores which rarely contain any recognizable copper minerals. South of section N 10,400 the ore body is low in copper, but to the north it becomes important, and below Bt. Mungus there are some rich pockets. Chalcopyrite can be seen in some hand specimens, usually as small grains, but it can reach nearly a half inch. Specimen No. 17684 in the University of Durham collection is a small specimen of fairly pure chalcopyrite, but such material is extremely rare and the writer found nothing comparable with this specimen during his field studies. Fragments of drill core from 50 ft below the footwall along section N 11,600 consist of pieces of pure chalcopyrite which has a dull black or irridescent surface.

In polished section chalcopyrite shows its normal optical properties, although the anisotropy is very difficult to detect unless two grains with different orientations are adjacent, or else it is twinned. The latter was only detected in the drill core sample, in which the twins vary from fine to coarse parallel lamellae. There can be twinning along two directions at right angles within one body of chalcopyrite. It was noticed that the polishing process could produce anomalous surface effects. In one polished section all the chalcopyrite grains had what appeared to be a core of a pinkish-brown mineral, which did not look entirely natural. Normal hand cleaning methods did not remove this material, but on repolishing it disappeared completely. In the same polished section pyrite was normal, but magnetite also showed anomalous surface effects.

Chalcopyrite occurs as a primary crystallization product and as a replacement mineral. It is frequently found in pyrite as veinlets and replacement bodies, and as small masses growing round pyrite. The veinlets are well developed when the pyrite has been fractured, and chalcopyrite fills the cracks with some replacement along the vein edges. Fig. 71 shows that partial replacement by chalcopyrite along fractures has left rounded relics of pyrite in chalcopyrite. The veinlets are up to 3 mm long and 0.5 mm wide, grading down to a few microns. Where the veining is profuse there is often a small mass of chalcopyrite developed on the edge of the pyrite grain. A few of the chalcopyrite veinlets form complex intergrowths with sphalerite or have sphalerite along the centre, and in rare examples they are accompanied by galena or gold.

Chalcopyrite replacement bodies in pyrite are more abundant than the veinlets, and are very variable in size and shape. The smallest bodies are a few microns across and tend to be concentrated towards the edge of the pyrite. They can be rounded blebs, or have a squarish outline due to crystallographic control exerted by the pyrite. There is evidence of slight crystallographic control in Fig. 72. Some of the larger bodies reach 0.2 mm, but the majority lie in the 20μ to 100μ range. The replacement bodies may have obvious feeder channels or be localized by cavities, but the majority are in solid pyrite. Fig. 72 shows a body with feeder channels. Some pyrite grains contain numerous chalcopyrite bodies, but the latter never form more than a small fraction of the total volume. In a few specimens the chalcopyrite occurs with sphalerite, forming

Fig. 71. Chalcopyrite filling fractures in pyrite. The chalcopyrite has partially replaced the pyrite, and has itself been replaced by neodigenite. x 220.

Fig. 72. Chalcopyrite replacement body in pyrite. The slightly squarish outline of the chalcopyrite suggests some measure of crystallographic control. Bornite and neodigenite have replaced part of the chalcopyrite. x 1310.

irregular patches or a layer between the sphalerite and the pyrite - Fig. 84. The relative age between sphalerite and chalcopyrite is not always clear, but in some samples the chalcopyrite has definitely replaced the sphalerite.

Chalcopyrite develops inthe gangue minerals, and has been observed in a few skarn samples, It forms subhedral and anhedral crystals, and patches up to 1.5 cm across, which are often associated with pyrite. At times the chalcopyrite is moulded round some of the gangue minerals, but can itself be veined by talc. One unusual mode of occurrence shows minute laths of chalcopyrite in parallel growth following cleavage traces in an amphibole gangue. The chalcopyrite in gangue is occasionally veined by sphalerite or galena, and can contain inclusions of sphalerite, galena and tennantite. Chalcopyrite and sphalerite form grains with mutual boundary relationships. Galena forms coronas on a small number of chalcopyrite grains.

Chalcopyrite is found in some of the magnetite ores, having developed in cavities and gangue patches, and at times replaced the magnetite - Fig. 69. One sample of brecciated magnetite contains formless masses of chalcopyrite which enclose the broken magnetite fragments, replacing both the ore and the gangue - Fig. 39. The latter is the only sample seen by the writer in which chalcopyrite is not accompanied by some pyrite. Chalcopyrite forms rare veinlets in magnetite and penetrates along intergrain boundaries.

Completely fresh chalcopyrite is rare and replacement by secondary sulphides is widespread, bornite, covellite and chalcocite-neodigenite being the typical alteration products. Bornite is not abundant but it is always the first to form, and it differs from the others in that its development is usually controlled by crystallographic planes in the chalcopyrite. Covellite and chalcocite-neodigenite normally form on grain edges and along cracks, but an isolated example of centrifugal replacement showed a grain with a core of secondary sulphide and a rim of chalcopyrite. Chalcopyrite has been replaced by galena and gold - Fig. 93, and is commonly found as exsolution bodies in sphalerite - Fig. 84.

The position of chalcopyrite in the paragenetic sequence is fairly clear; it is later than the iron oxides and pyrite, and earlier than the secondary copper sulphides, galena and gold. It is partly contemporaneous with sphalerite, but appears to have continued crystallizing after sphalerite ceased. Its relation to the other sulphides is less certain as contact relationships have not been observed, but it appears to be earlier than bismuthinite.

Bornite

Bornite is a rare mineral in the Bt. Ibam deposit, and is always secondary in origin. It replaces chalcopyrite and is itself replaced by the other secondary copper sulphides. It commonly occurs as broad to narrow wedges and irregular lamellae, which project from the margin towards the centre of chalcopyrite grains. The growth of the wedges is obviously controlled by crystallographic planes in the chalcopyrite - Fig. 73. Bornite also replaces chalcopyrite along irregular cracks.

Chalcocite-Neodigenite & Djurleite

These minerals are described under one heading as they are closely associated in the Bt. Ibam ore body, although djurleite is very much rarer than the other two, and it is often difficult to separate chalcocite from neodigenite by optical methods in polished section. There is no clearly demarcated zone of secondary copper sulphide enrichment, and to a large extent the distribution of the secondary minerals coincides with the distribution of chalcopyrite from which they have formed in situ. In a few samples only is there evidence that there has been a signicant increase in the copper content.

The secondary sulphides are rarely visible in hand specimen; on a cut surface in one altered pyritic magnetite ore they form distinct dark coronas on the small pyrite crystals, and some of the euhedral pyrite crystals have a dull black exterior due to a thin superficial layer of chalcocite-neodigenite plus covellite.

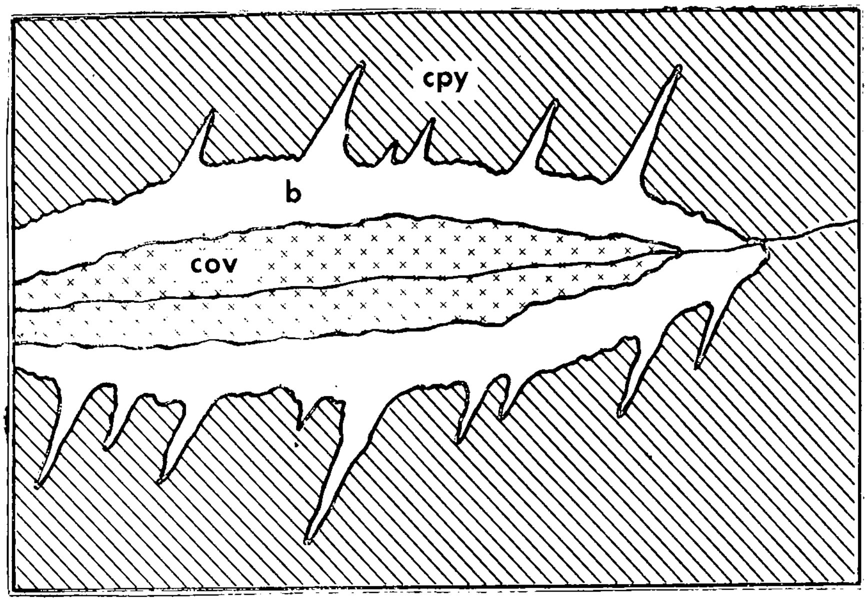


Fig. 73. Drawing taken from a polished section which shows bornite which has followed a fracture in chalcopyrite, and has replaced the chalcopyrite along a selected crystallographic plane. Covellite has later followed the same fracture and replaced the bornite. x 500.

In polished section chalcocite-neodigenite shows some variation in its optical properties, particularly marked in the range of colours displayed. This variation is probably due to slight differences in composition, fine intergrowths, and solid solution mixtures. The colour varies from greyish-white, through greys with a bluish tint to a deep blue. The colour can be even throughout an individual grain or irregularly distributed in patches and streaks. The secondary sulphides which have not formed by replacement of chalcopyrite or pyrite in situ tend to be greyish-white or very pale bluish-grey and have an even colour, replacement of pyrite produces an evenly coloured medium blue chalcocite-neodigenite, while replacement of chalcopyrite gives a medium to dark blue mineral which is often irregularly coloured. This genetic pattern is not without exceptions, but it has a general applicability. Colour is an unreliable guide to identification, except that the greyish-white variety is chalcocite sensu stricto. The anisotropism of chalcocite is only detectable with a strong carbon arc light source, when a purplish-red colour can be seen under nicols which have been fractionally uncrossed, and the anisotropy is weakly displayed. The anisotropy was detected in mineral which was quite clearly blue in colour and would have otherwise been classified as neodigenite. There is no sign in the Bt. Ibam deposit of lamellar chalcocite; in the unevenly coloured varieties the folour distribution is completely irregular. Djurleite was observed in minute amounts in only two or three polished sections, and is always intimately associated with chalcocite-neodigenite. It is greyish-white in colour, with a faint creamy-pink or very pale brownish-violet tint. No anisotropism was detected, although the mineral is orthorhombic.

Chalcocite-neodigenite has usually formed by replacement in situ of chalcopyrite and pyrite and, less frequently, other sulphides, but it also occurs as veinlets and cavity fillings. Rare examples of replacement of magnetite have been seen. It is associated with goethite, covellite, cuprite and malachite. The most extensive masses of chalcocite-neodigenite have formed by replacement of pyrite, some of which have been completely pseudomorphed. Replacement typically starts at the grain edge producing coronas of secondary sulphide, which is sometimes mixed with goethite. In some of the large pyrite crystals alteration has not proceeded beyond a thin surface layer, but it is sufficient to give the crystals a black exterior. Where a pyrite has been shattered the chalcocite-neodigenite penetrates along the cracks, forming networks which can be rectilinear or haphazard or a mixture of the two. The veinlets are partly dilational infillings and partly formed by replacement. The dilational veinlets are extremely small, but the replacement ore can expand into formless masses with relics of pyrite - Fig. 74. In a small number of grains a narrow band at the periphery of the pyrite is left unaltered and replacement follows a crude zonal pattern - Fig. 75. Where there are chalcopyrite bodies in pyrite the secondary copper sulphides tend to preferentially attack the chalcopyrite.

In some ores there is a sympathetic relation between the development of chalcocite-neodigenite and goethite. A series of polished sections taken from a mass of pyritic magnetite along section N 11,500 showed that chalcocite-neodigenite was much more abundant in those ores rich in goethite. It has also been observed that at the intersection of some goethite veinlets and small chalcopyrite bodies in a pyrite host, there tends to be a development of secondary sulphide - Fig. 76. Some of the solutions which deposited goethite must have been cupriferous - e.g. the two small replacement bodies shown in Fig. 76 contain more copper than was originally present in the chalcopyrite - to account for the relation between goethite and secondary sulphide development observed in some ores.

Much of the chalcocite-neodigenite which replaces pyrite is monomineralic, but it can also occur with goethite, cuprite and covellite. The secondary minerals often form intimate mixtures, but the relative ages are revealed where they form veinlets - Fig. 77. In this example the order is clearly

pyrite → cuprite → chalcocite/neodigenite → goethite

and other specimens suggest that cuprite commonly forms before the secondary sulphides. There is no consistent age relationship between chalcocite-neodigenite and goethite, although the bulk of the geothite is later. Covellite is earlier than, and pene-contemporaneous with, chalcocite-neodigenite.

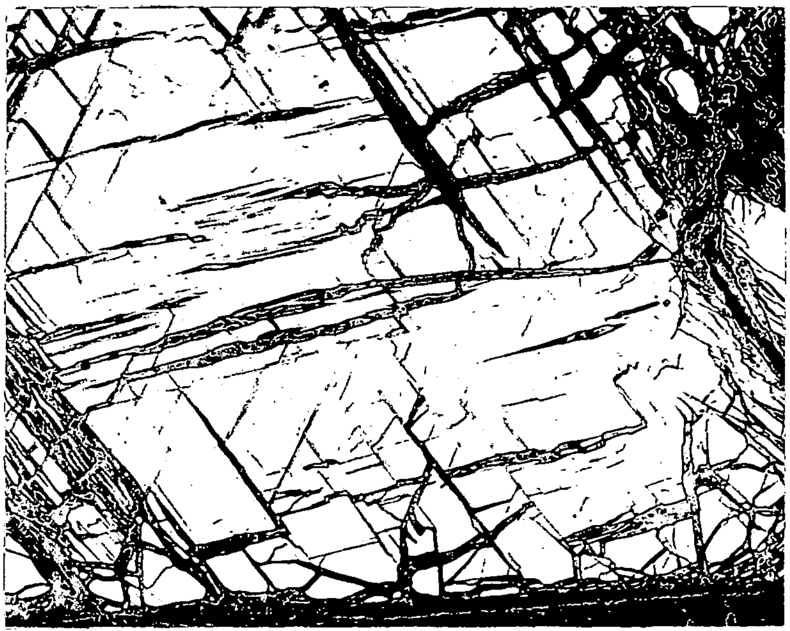


Fig. 74. Pyrite with fractures which show a rectilinear pattern. Chalcocite-neodigenite has filled the fractures and partially replaced the pyrite leaving small, isolated relics (top right). x 125.

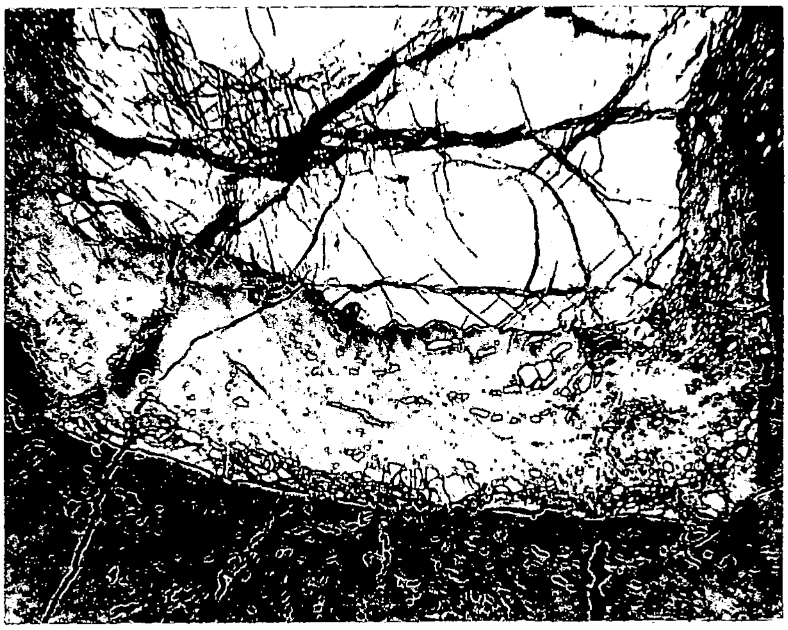


Fig. 75. Chalcocite-neodigenite which has partly replaced a pyrite crystal. Note that the replacement has not always started at the grain edge. The pyrite crystal is in a mass of goethite. x 45.

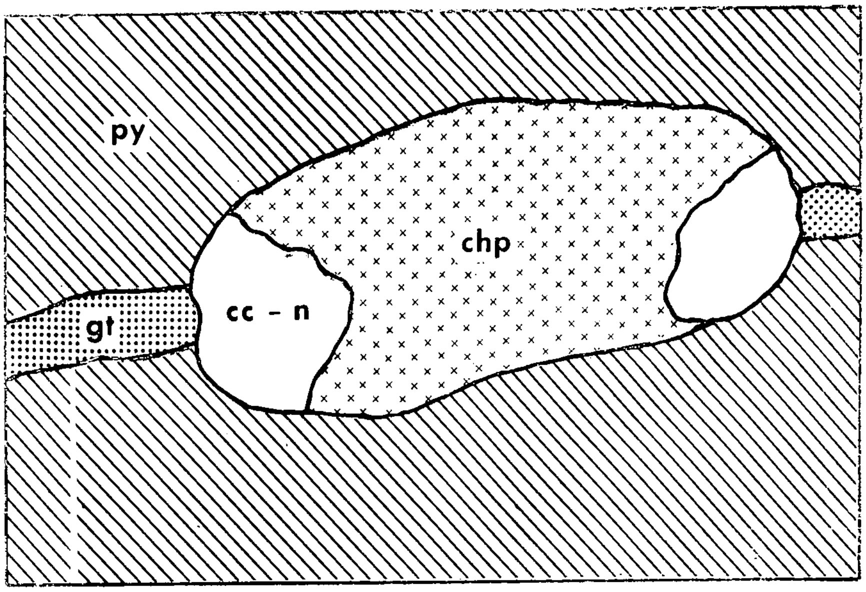


Fig. 76 *Drawing from a polished section which shows chalcocite-neodigenite developed where a goethite veinlet meets a chalcopyrite body in pyrite. x 800 approx.*

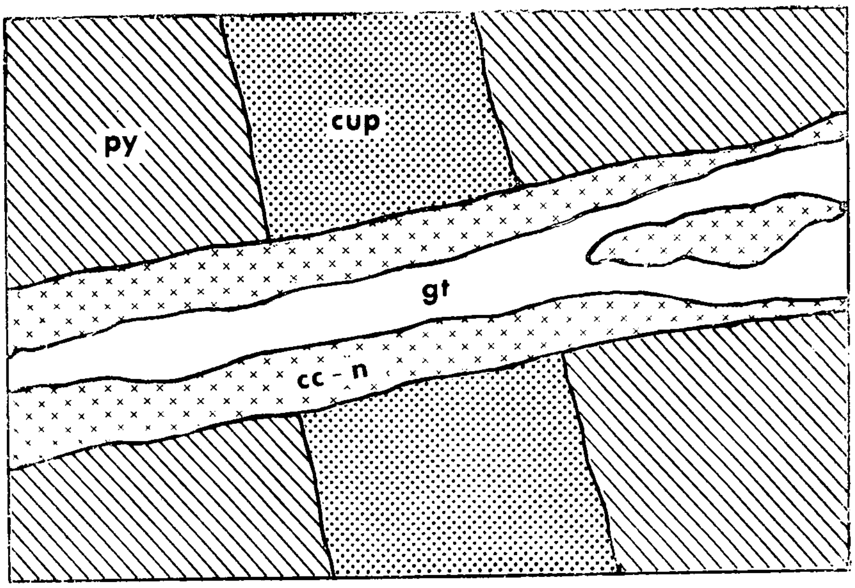


Fig. 77 *Drawing from a polished section which shows intersecting veinlets of goethite, chalcocite-neodigenite and cuprite cutting pyrite. x 500 approx.*

Chalcocite-neodigenite replaces chalcopyrite, usually commencing round the edges or along cracks, the latter being thin and irregular, probably due to the ductility of the chalcopyrite. The secondary copper sulphides form later than bornite - Fig. 72. In some of the chalcocite-neodigenite which has formed directly from chalcopyrite there are minute patches of a purplish-grey mineral which has not been identified, but it appears to be an intermediate phase in the replacement. The only examples of djurleite which have been observed occur as a replacement of chalcocite-neodigenite and enargite, which have themselves replaced chalcopyrite. The djurleite is located round the periphery of the chalcocite-neodigenite and is a later phase - Fig. 78.

Chalcocite-neodigenite replaces sphalerite, some examples showing a core of sphalerite with successive coronas of covellite mixed with neodigenite, and bismuthinite. The secondary copper minerals have also been observed along the contact between galena and tennantite, and also between galena and chalcopyrite.

Occasionally one finds small, formless, poikiloblastic masses of chalcocite-neodigenite which have grown independently of pyrite or chalcopyrite, although the primary sulphides are usually present elsewhere in the same polished section. The writer has seen no evidence that there has been widespread transportation and redeposition of chalcocite-neodigenite to produce markedly enriched zones, and most of the secondary sulphides have formed on, or very close to, pyrite and chalcopyrite. This conclusion agrees with a short report prepared for the mining company by CISRO, Melbourne, in 1961, which states that "The secondary copper sulphides appear to have been formed in situ by replacement of chalcopyrite".

Chalcocite-neodigenite is found as small patches and veinlets in magnetite ore, and has occasionally replaced the magnetite - Fig. 82. Some of the secondary sulphide patches retain straight edges which have been controlled by the $\{111\}$ parting planes in the magnetite, although later attack by goethite has often modified their outline. The chalcocite-neodigenite in veinlets is accompanied by cuprite, covellite and malachite.

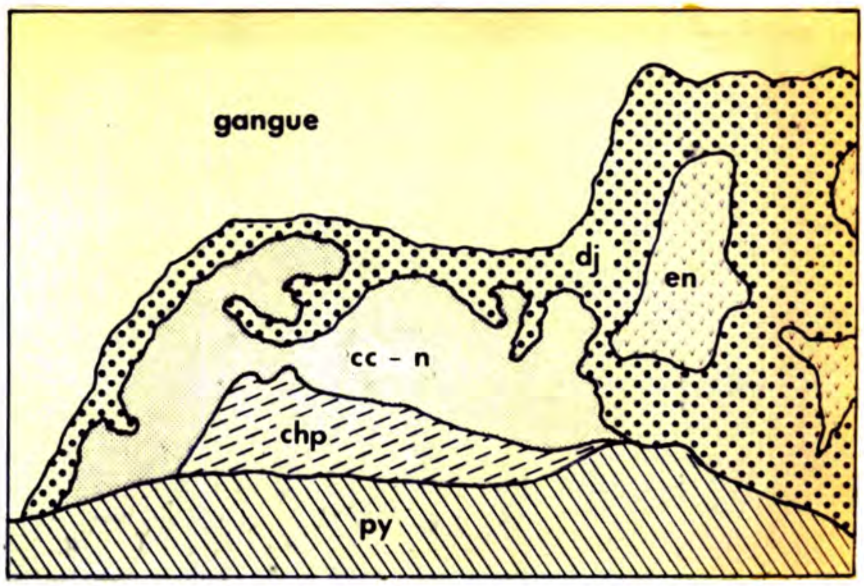


Fig. 78. Drawing taken from a polished section which shows djurleite formed round the periphery of a small patch of chalcocite-neodigenite, which has replaced chalcopyrite. Small relics of enargite can be seen - right. x 800 approx.

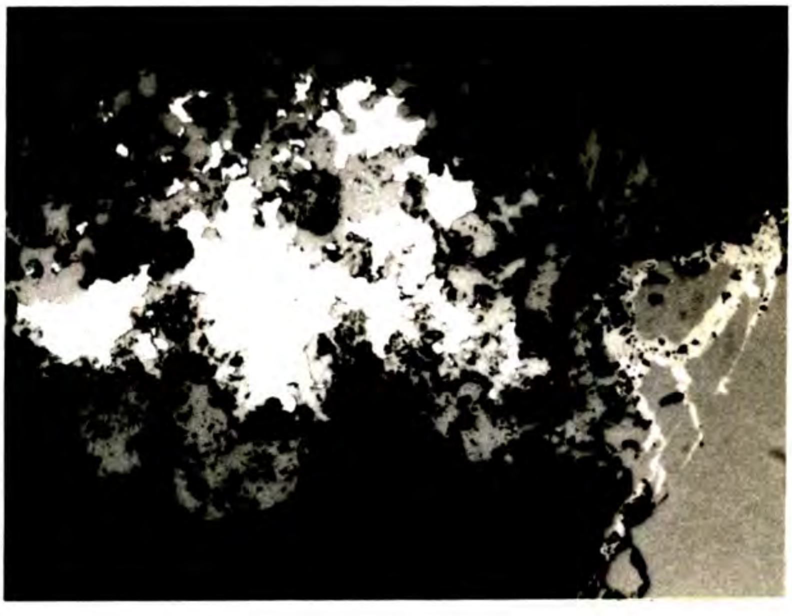


Fig. 79. Formless masses of cuprite formed on native copper. A magnetite grain with a few martite lamellae can be seen lower right. x 560.

Chalcocite-neodigenite is always later than pyrite, chalcopyrite and magnetite and earlier than malachite. It is later than bismuthinite, sphalerite, galena,and some of the cuprite and covellite. It has formed contemporaneously with some goethite, but the latter continued to crystallize after the formation of the secondary sulphide ceased.

Covellite

Although it is much less abundant than chalcocite-neodigenite, with which it is usually associated, covellite forms the predominant secondary copper sulphide in a small number of specimens. Its optical properties agree exactly with textbook descriptions. Unlike chalcocite-neodigenite, covellite is not an important replacement mineral of pyrite, although it does occur in thin coronas and veinlets, but is typically associated with chalcopyrite, or is found in veinlets and cavities in iron ore. Covellite veins and replaces cuprite, and replaces bismuthinite and sphalerite, but is itself replaced by malachite. Covellite is often found in association with chalcocite-neodigenite, but the two minerals do not have a consistent age relationship, and appear to be partly contemporaneous, although a little covellite is certainly earlier.

Covellite forms small veinlets and patches in magnetite, some of the latter showing very straight edges parallel to $\{110\}$ planes, and while possibly formed by replacement they are more likely to be infillings of cavities.

Cuprite

Cuprite is not an abundant mineral in Bt. Ibam and has roughly the same distribution as copper, with which it is often associated. In only one locality, near the footwall along N 11,900, did the writer find cuprite which was visible to the naked eye, where it occurred as masses of minute euhedral crystals along cracks and in cavities, both in ore and gangue. The latter was a rusty brown clay which had been stained purplish-red in patches due to the presence of finely divided cuprite. A sample of the red clay assayed just over 2% Cu. Polished sections show that cuprite occurs in three mineral associations;

1. cuprite-native copper.
2. cuprite-pyrite-secondary copper sulphides.
3. cuprite-malachite in veinlets.

1. Cuprite-Native Copper

In this association the cuprite appears to be younger than the copper, which has influenced cuprite formation in two ways. In some samples the cuprite forms relatively solid masses surrounding irregular relics of copper, and there is no doubt that the cuprite has formed directly from the copper by oxidation – Fig. 79. In other specimens the cuprite forms porous aggregates, the individual crystals tending to be euhedral or subhedral. The associated copper shows little sign of corrosion and can retain a skeletal crystal outline, the cuprite having crystallized round the edges of the copper and in close proximity to it. There is a genetic relationship, but here the copper has acted as a nucleating or precipitating agent and not as a direct source of copper, which must have been introduced in solution. Fig. 81 shows an association of this type. Dana & Ford (1959) state that copper can precipitate cuprite from copper-bearing solutions.

2. Cuprite-Pyrite-Secondary Copper Sulphides.

This association was observed in polished sections from one locality near N 11,500; W 10,900. The host material is a pyritic magnetite, the latter having been extensively altered to martite and goethite. The pyrite has been partially replaced by cuprite, chalcocite-neodigenite, covellite and goethite. The age relationships between these minerals are not always clear, but the normal sequence is

cuprite → secondary copper sulphides → goethite.

Fig. 80 shows a typical occurrence. The cuprite is found as a peripheral alteration of pyrite from which narrow fingers are projected towards the centre, as veinlets cutting pyrite, and as fine grained mixtures with the other minerals in irregular

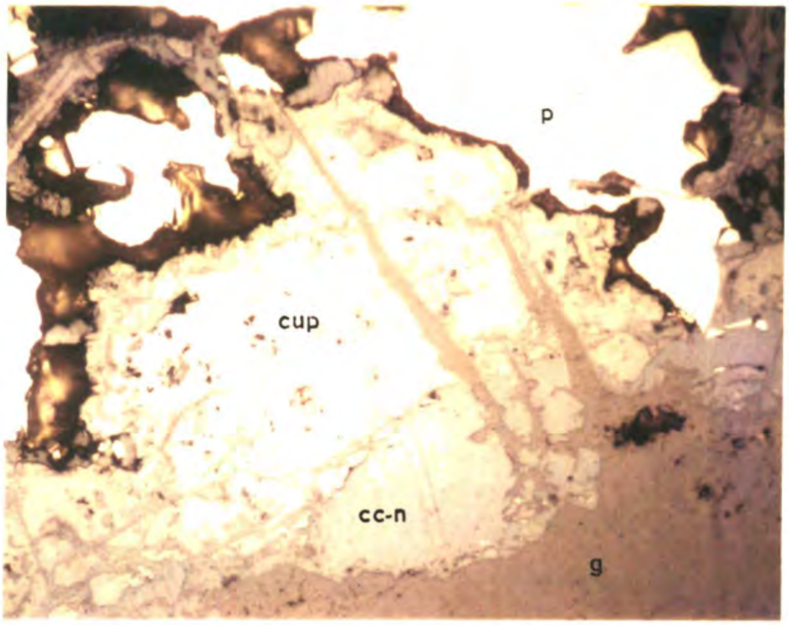


Fig. 80. Photomicrograph which shows the age relationship between cuprite, chalcocite-neodigenite and goethite. The neo-digenite has veined and replaced the cuprite, while goethite has veined them both. x 550.

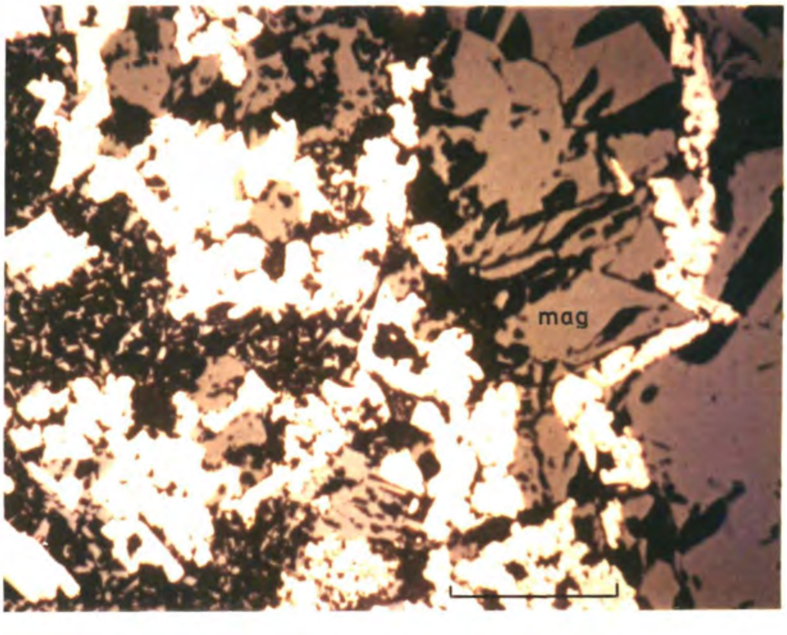


Fig. 81. Native copper which has veined fresh magnetite (right). The bluish-grey mineral towards the left is cuprite, which has grown as a porous mass on copper but not replaced it. x 225.

patches or zones. Where the minerals form veinlets the age relationships are unequivocal - Figs. 77 and 80.

3. Cuprite-Malachite.

A small patch of granular magnetite-martite ore from N 11,800; W 10,800 was heavily impregnated with secondary copper minerals, a bulk sample assaying over 5% Cu. The iron ore has been cut by veinlets and intergranular films of cuprite, which can contain specks of copper. The ore has later been impregnated with malachite, which has replaced much of the cuprite leaving only small isolated relics.

Cuprite must have formed during several phases over a long period of time, as it can be earlier than the secondary copper sulphides and also later than native copper, which appears to be one of the youngest minerals in the Bt. Ibam deposit.

Copper

Native copper is not common, and is usually seen as small specks or thin films and fillings along joints and in cavities, but spongy masses up to 2 cm x 1 cm x 1 cm have been encountered. Copper has also been seen as isolated specks randomly distributed in friable magnetite, and as granules along a contact between porous magnetite and chloritic clay gangue.

In polished section the copper is distinctive with its very high reflectivity and pinkish-red colour. It occurs as porous masses and skeletal crystals, and as intermittent veinlets. In the latter form it can be seen following cracks and intergranular boundaries in magnetite - Fig. 81, and in cracks and cavities in goethite - Fig. 82. The latter type of occurrence shows that copper can be a very late stage product, forming after some goethite has crystallized from colloidal solutions. Fig. 82 shows a magnetite grain which has been fractured and partially replaced by goethite, and there is copper along a cavity in the goethite.

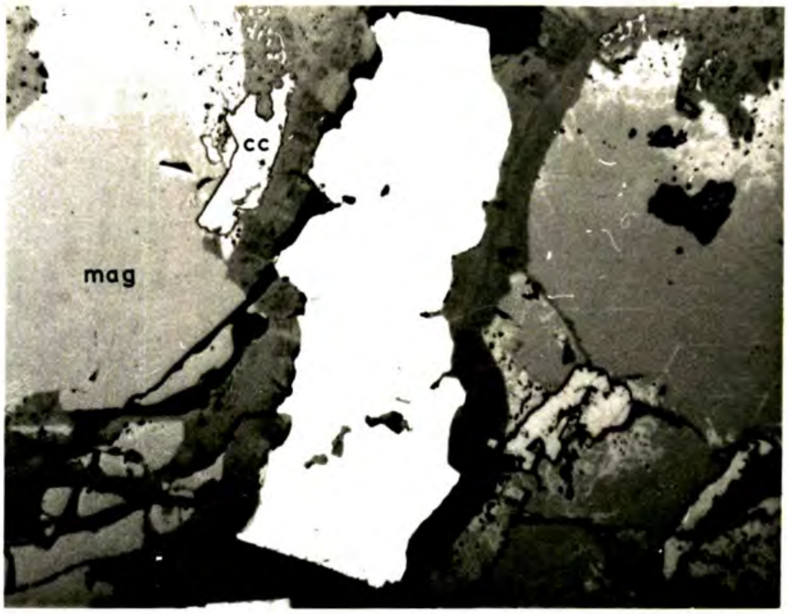


Fig. 82. Native copper crystallized along a cavity in goethite, which has itself partially replaced magnetite. There are small patches of martite and relics of chalcocite. Note some straight edges (probably controlled by 111 planes) in the chalcocite to the left of the copper. x 220.

Fig. 83. Chalcanthite which has crystallized on the surface of magnetite ore.

Copper is usually accompanied by cuprite, some of which has formed by oxidation of the copper, but elsewhere the copper has acted as some kind of catalyst to the cuprite crystallization. It is surprising that some of the copper-cuprite masses are associated with fresh magnetite. The development of cuprite shows that oxidizing conditions have been operative at some stage in the geological history, and yet adjacent magnetite can show little or no alteration to martite, e.g. Fig. 81.

Malachite and Azurite

These two minerals occur together, although azurite is very much rarer. Both carbonates form veinlets in massive martitized ore, while malachite forms small specks in friable magnetite. Polished sections show that they are the youngest of the secondary copper minerals, replacing cuprite, covellite and chalcocite-neodigenite. Malachite is still in the process of formation, as it was observed on the surface of screes of friable magnetite which had collapsed from newly cut benches. The malachite was only present near the surface of the scree material, and absent in the deeper parts. It has crystallized in the spaces between the magnetite grains.

Chalcanthite

In view of its very high solubility this is an unexpected mineral to find on open benches in a region of high rainfall. It was found as feathery surface coatings and along minor cracks in massive magnetite at N 12,000; W 10,800. The mineral gave a distinctive bluish-green colour to the bench face - Fig. 83 - enhanced in some places by the presence of malachite. The chalcanthite must form extremely quickly, as it has been observed within hours of the cessation of torrential rain. The heat of the sun must draw the water held within the rock to the surface where it rapidly evaporates, depositing its dissolved salts. There must be a constant acid attack on the cupriferous minerals to maintain the supply of copper to form the sulphate. The identification of the chalcanthite was confirmed by diffraction, and Table 14 shows the first eight strong lines compared with the ASTM pattern. It can be seen that the two patterns are almost identical.

Table 14

Partial Diffraction Pattern For Chalcanthite

Cu Radiation

Bt. Ibam		ASTM 11-646	
dÅ	Rel.I.	dÅ	Rel.I.
5.72	3	5.73	35
5.49	5	5.48	55
4.73	10	4.73	100
3.99	6	3.99	60
3.71	5	3.71	85
3.31	3	3.30	60
2.829	4	2.824	40
2.750	4	2.749	50

XI MINERALOGY OF BUKIT IBAM. (3) MINERALS OF ZINC, BISMUTH & LEAD

Sphalerite

Sphalerite has been observed in polished sections of samples from several localities within the ore body and its immediate wall rocks between N 10,400 and N 12,300, but the only megascopic occurrence was encountered 130 ft below the footwall along section N 10,400, where veins up to quarter inch wide cut acid volcanic rocks. In the area close to N 10,400 there are some highly zincian wallrock samples which often contain sphalerite, but this mineral does not account for the total zinc content. Some of the goethitic ores are rich in zinc, but sphalerite has not been observed in this type of ore. The zinc content of a sample is not necessarily any indication of its sphalerite content.

In hand specimen sphalerite is medium brown, and in polished section bluish-grey with rare internal reflections coloured white to reddish-brown. Some of the sphalerites show a peculiar optical effect under very high magnification, with myriads of minute bright spots coloured red, blue and green. It is thought that these are due to reflections from cryptocrystalline particles of chalcopyrite, coupled with strong dispersion within the sphalerite.

Except for the vein in acid volcanic rock sphalerite is closely associated with, and subordinate to, other sulphide minerals, and is usually present in small quantities. Sphalerite typically occurs as small bodies and veinlets in pyrite, and less frequently as isolated grains in gangue and iron ore.

Sphalerite has veined pyrite, and there is usually some replacement along the vein edges. The replacement bodies vary in size from a few microns to 0.3 mm. The small bodies tend to have a squarish outline and be concentrated towards the margins of the pyrite grains, while the larger bodies have irregular outlines. The small bodies can be accompanied by chalcopyrite, bismuthinite and galena of similar dimensions. The larger bodies tend to be located internally in the pyrite, but few have any visible feeder channel. One specimen from N 10,400 is unique, as the sphalerite has preferentially replaced the periphery of the pyrite, preserving parts of {100} faces. The larger sphalerite bodies are sometimes accompanied by chalcopyrite, which occurs either as irregular patches or coronas round the edge of the sphalerite - Fig. 84. One patch of sphalerite in a specimen from N 10,400 contained unreplaced relics of bismuthinite. The veinlets in pyrite vary from a few microns to 0.2 mm, and small veinlets have been observed in chalcopyrite. Some of the sphalerite veinlets in pyrite are accompanied by chalcopyrite, either as intergrowths or along the edge of the vein.

One polished section showed orientated lamellae of sphalerite in magnetite, and seemed to show a replacement of the magnetite along crystallographic planes.

Sphalerite is rarely found in the gangue minerals, but can occur as grains up to 0.3 mm across, which vary from anhedra to euhedra. Some has grown with chalcopyrite in mutual boundary relationship, and secondary copper sulphides can develop along the sphalerite-chalcopyrite junction.

Some sphalerites have been replaced by secondary copper minerals, producing concentric structures with coronas of covellite and chalcocite-neodigenite on sphalerite cores. Similar structures show sphalerites with galena or chalcopyrite rims. In some samples from N 10,400 sphalerite has been replaced by siderite.

A feature of many of the sphalerites is the widespread development of orientated lamellae and blebs of chalcopyrite. The particles vary in size from those which are just visible to 0.01 mm, and in shape from rounded to lensoidal. The chalcopyrite particles can be of roughly equal size, or there can be a seriate distribution from fine to coarse within one sphalerite body. Fig. 84 shows the latter type. The coarser chalcopyrite bodies tend to be more sparsely distributed than the fine ones, probably due to the larger bodies "draining" the surrounding sphalerite. The development of orientated chalcopyrite bodies is usually ascribed to unmixing, and this appears to be the case in Bt. Ibam. Some

Fig. 84. Replacement body of sphalerite with exsolved particles of chalcopyrite, in pyrite. Further replacement reaction has produced a rim of chalcopyrite on the sphalerite body. x 750.

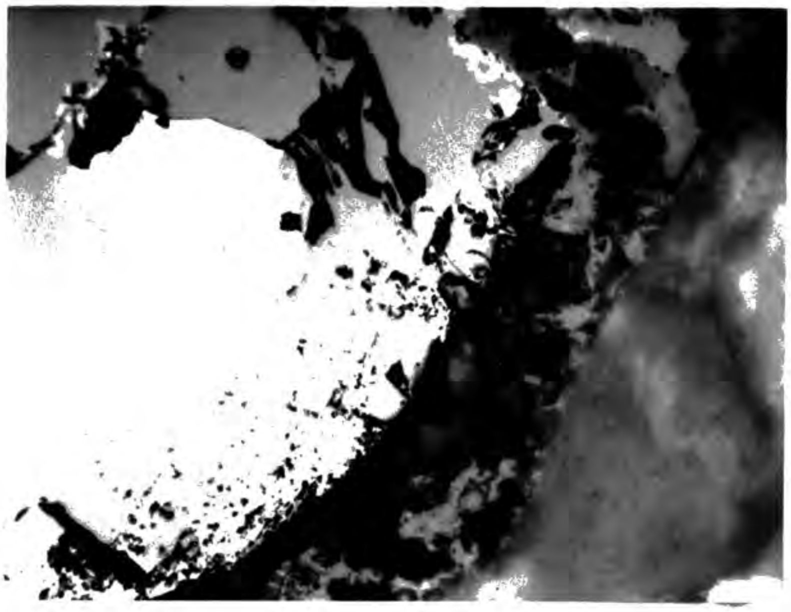


Fig. 85. Bismuthinite which has crystallized in a cavity in magnetite and partially replaced it. The blurred material at the lower right is bismutoferrite. x 530.

of the sphalerite bodies contain nearly 10% exsolved chalcopyrite, which indicates that the original solid solution mixture was virtually saturated with the chalcopyrite component (Donnay & Kullerud, 1958). Another feature which could be connected with an exsolution process is the development of peripheral patches and rims of chalcopyrite due to migration towards the grain edge, although the same thing could result from replacement. The writer believes that the example seen in Fig. 84 is a replacement feature, as there is too much chalcopyrite to have been held in a solid solution mixture, and there is no sign of "draining" of the sphalerite towards the edges.

The position of sphalerite in the paragenetic sequence is reasonably clear. It is later than pyrite, the iron oxides and probably bismuthinite, but is earlier than the secondary copper sulphides and galena. Its relation to chalcopyrite is equivocal. Veinlets of sphalerite in chalcopyrite show that here the sphalerite is later, but chalcopyrite has replaced sphalerite in some samples, while in others the two occur with a mutual boundary relationship. It seems reasonably certain that sphalerite and chalcopyrite were in part contemporaneous, but chalcopyrite continued to crystallize after sphalerite ceased.

A sample from a megascopic vein of sphalerite was powder photographed, and the cell size calculated as 5.416Å. Inserting this figure in the plot of composition against cell edge in Deer, Howie & Zussman vol. 5 (1962), indicates that the sphalerite contains 15 mol% FeS, i.e. the the composition is $Zn_{.85}Fe_{.15}S$. Possible implic tions with regard to temperatures of formation are discussed in a later chapter.

Bismuthinite

Although the presence of bismuth contamination in parts of the Bt. Ibam ore body has been known for some years, it was not until the present study that the causative minerals were positively identified. Apart from traces of tetradymite, the bismuth is present in bismuthinite and its alteration products.

Bismuthinite occurs as rare microscopic veinlets cutting through ore and gangue, as isolated grains and aggregates and, most commonly, as replacement bodies in a pyrite host. The bismuthinite in pyrite occurs as small replacement bodies arranged linearly along cracks, or as bodies in a roughly zonal pattern near the periphery of the pyrite. Isolated grains occur in gangue, where they can show a euhedral or subhedral tabular shape, but are more common in magnetite where they have crystallized in cavities and along cracks, and occasionally replaced the magnetite- Fig. 85. The replacement bodies in pyrite usually have a rounded, but irregular, outline - Fig. 86. The bodies vary in size from barely visible to 0.1 mm, with the majority lying between 30μ and 60μ.

Bismuthinite is later than pyrite and magnetite and earlier than the secondary copper sulphides, and it is probably earlier than sphalerite. Its age relative to chalcopyrite is uncertain. Part of the paragenetic sequence is clearly displayed in the replacement bodies in pyrite, where solutions carrying bismuth have entered fractures, forming small intermittent patches of bismuthinite - Fig. 86. Copper-bearing solutions have later followed the same channels and partially replaced the bismuthinite. The replacement usually commences where the fracture meets the bismuthinite grain, and then spreads out through a series of internal fractures and round the edge of the bismuthinite grain. This alteration pattern is well shown in Fig. 87. There is evidence that some intermediate compounds are formed during the replacement, and although none have been positively identified they are probably Cu-Bi sulphosalts e.g. emplectite, wittichenite. These minerals appear as coronas round the secondary copper sulphides, and also along internal cracks - Fig. 87. The final replacement product is a covellite, chalcocite-neodigenite mixture.

Under oxidizing conditions the bismuthinite can break down to form the very rare mineral bismutoferrite, described later.

Fig. 86. Small irregular replacement bodies of bismuthinite along a crack in pyrite. The small dark spots are cavities. x 560.

Fig. 87. Bismuthinite bodies which have been partially replaced by secondary copper sulphides (almost black). The greyish material which can be seen within the bismuthinite appears to be an intermediate phase in the replacement. x 1360.

Tetradymite

A rare mineral which has been identified from one locality near N 11,900; W 10,900, although the occurrence of bismuth and tellurium in some trace element analyses suggests that it may occur elsewhere. As tellurium minerals have not been recorded previously from Malaya the Bt. Ibam material was checked on a microprobe. Fig. 88 shows the electron image and the X-ray images for Bi and Te for one of the grains.

The tetradymite occurs in a magnetite-amphibole ore which has been strongly brecciated and silicified, and appears to have crystallized before the brecciation. Chalcopyrite, which is the most abundant sulphide, is clearly later than the brecciation, but the tetradymite is earlier than the chalcopyrite. The majority of the tetradymite grains show clear signs of deformation. The mineral occurs as lath shaped grains and more equant masses, and also as minute needles which are in parallel arrangement following the cleavage direction in the amphibole. The tetradymite is intensely white with a high reflectivity, but can show irregular patches of a greyer mineral with a slightly lower reflectivity, which could be due to differences in composition or orientation. The tetradymite displays some highly irregular shapes due to deformation, and penetration of a gangue mineral, probably quartz, along the (0001) cleavage planes causes small fragments to break off. Fig. 89 shows a grain which is clearly bent round a magnetite crystal, and fragmented parallel to the (0001) cleavage. Even more complicated shapes can be produced by deformation, and a patchy appearance in some of them could be a type of pressure twinning.

Bismutoferrite

This mineral has been encountered in only one hand specimen, and although present in very small quantities it is worthy of description in some detail, as it has previously been described as unique to one locality in Saxony. Dana & Ford (1959) do not mention it, but list its antimonian analogue chapmanite. There is a detailed description of bismutoferrite in Milton et al (1958),

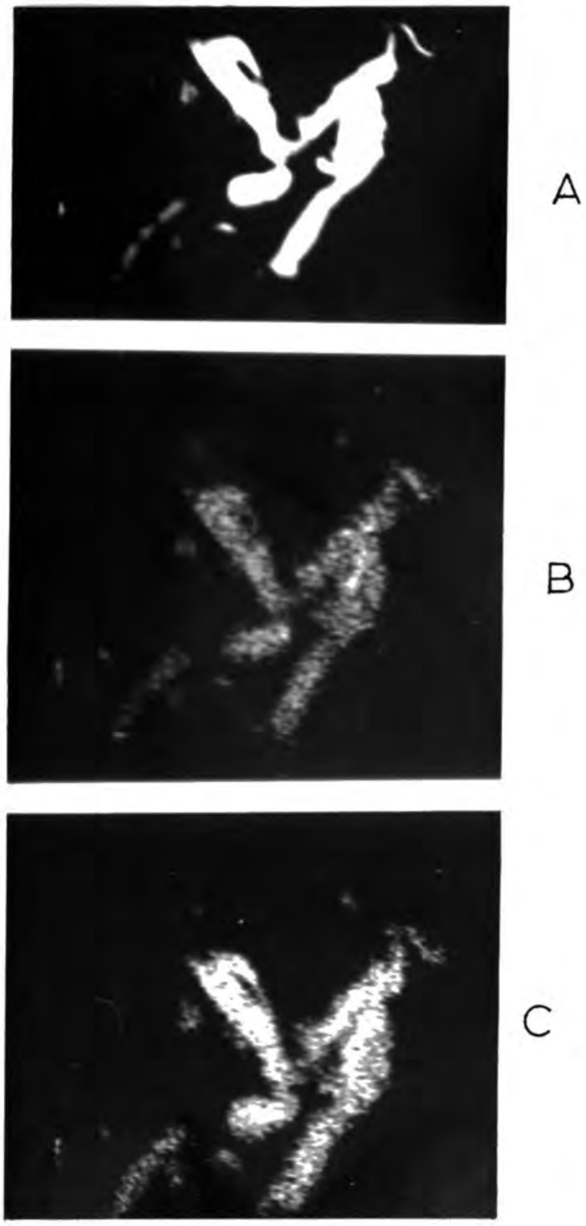


Fig. 88. Photographs taken on a microprobe and showing the electron image and levels of tellurium and bismuth in tetradymite. x 530.

A = Electron Image B = Tellurium C = Bismuth

Fig. 89. Tetradymite bent round a magnetite grain. Note how the tetradymite has been split along its (0001) cleavage. x 550.

their X-ray data being reproduced in ASTM 11-174. Bismutoferrite has the ideal formula $Bi_2O_3.2Fe_2O_3.4SiO_2.H_2O$, while chapmanite has antimony in place of bismuth.

At Bt. Ibam the bismutoferrite is found in disseminated magnetite ore close to the footwall at N 11,500; W 11,100, a composite sample of ore and gangue assaying 1.86% Bi. The magnetite is dense, finely granular and somewhat friable, and is associated with small patches and speckles of white gangue, which is a mixture of talc and kaolinite. No sulphides are visible to the naked eye. The bismutoferrite occurs as a soft, bright yellow, earthy powder filling cracks, and as isolated specks scattered through the ore and, to a lesser extent, the gangue. The colour is striking, but unless the hand specimen is broken along one of the cracks filled with the mineral and a relatively large yellow surface exposed it is easily overlooked.

In polished section the magnetite is finely granular and porous, and has been slightly brecciated. There is a little alteration to martite and goethite. Bismuthinite has formed veinlets intruding the magnetite, but it is now seen as irregular relics, and it also occurs as grains filling cavities in the magnetite - Fig. 85. The bismutoferrite occurs as veinlets and patches, some of which contain the pure mineral but others are mixed with goethite. The bismutoferrite-goethite patches often contain relics of bismuthinite. Bismutoferrite appears as a completely structureless yellowish-white material, with a strong yellow internal reflection. Under oil immersion the internal reflection takes on a greenish tint, the colour being similar to that of a yellowish epidote. Any evidence of bireflectance or anisotropy in completely masked by the internal reflections.

As pure a sample of the bismutoferrite as it was possible to obtain was powder photographed using Co and Cu radiation, and although it was impossible to obtain good back reflections using a 114.6 mm camera, photographs taken with a 57.3 mm camera were suitable for shrinkage corrections. The diffraction pattern is shown in Table 15, together with the ASTM pattern for comparison. A careful study of Table 15 shows that there is a strong coincidence

Table 15

Diffraction Pattern for Bismutoferrite

Cu and Co Radiation

Bt. Ibam		ASTM 11-174	
dÅ	Rel.I	dÅ	Rel.I
7.56	6	7.63	100
-		4.52	5
4.47	1	4.47	7
4.17	7	4.18	20
3.87	8	3.87	100
-		3.79	3
3.56	10	3.58	35
3.17	7	3.18	50
2.89	6	2.90	70
2.65	3	2.66	12
2.58	7	2.59	35
2.52	3	2.53	25
2.44	½	-	
2.374	2	2.383	11
2.240	3	2.251	9
		2.225	3
v.br.	1	2.201	7
2.154	2	2.162	15
2.080	2	2.086	5
2.036	3	2.040	9
2.006	½	2.011	1
br.		1.976	1
1.933	2	1.935	11
1.902	2	1.907	9
		1.891	5
1.861	1	1.863	3
1.834	1	1.835	3
1.783	1	1.785	1
1.738	½	1.744	3
1.715	3	1.719	13
1.691	2	1.694	3
1.683?	2	1.683	5

Table 15 contd.

1.654	½	1.659	1
1.634	3	1.634	12
1.605	½	1.608	9
1.589	3	1.590	12
1.564	1	1.570	1
1.539	1	1.542	1B
1.524	2	1.523	5
1.503	4	1.504	13
1.481	½	-	
1.470	1	1.471	3
1.452	1	(1.456	1
		(1.451	1
1.432	1	1.435	3
1.389	½	1.394	1B
1.366	2	(1.370	3
		(1.360	1
1.325	1	1.329	1
1.296	2	(1.299	1
		(1.291	1
1.280	1	1.282	1
1.253	1	1.257-2	1
1.238	1	-	
1.228)	1	1.224	1
1.221)	1		
+24 lines		+ 24 lines	

between the two patterns, particularly so in view of the large number of lines and the fact that the mineral is an alteration product and probably subject to some variation in composition, and there can be no doubt that the Bt. Ibam material is the same as that described from the type area. Some lines on the Bt. Ibam pattern have been resolved into two closely spaced lines in the ASTM data. There is a discrepency between the two patterns in the spacing of the very first, and one of the two strongest lines, 7.56Å in Bt. Ibam and 7.63Å in ASTM. All the films of the Bt. Ibam sample were carefully checked, and there is no doubt that 7.56Å is the correct value for this sample.

The diffraction patterns for bismutoferrite and chapmanite are vary similar c.f. bismuthinite and stibnite, and it would be extremely difficult to differentiate between the minerals on diffraction data alone. The Bt. Ibam sample is associated with bismuthinite, which indicates the alteration product is bismutoferrite, but a test for the main constituent elements was made with a microprobe. A 2θ scan proved that the only elements present in important amounts were Bi, Si and Fe, that Al was present in trace amounts, and that Sb was absent. Bt. Ibam can, therefore, be classed as the second locality from which bismutoferrite has been recorded. ***See Note below.**

Galena

Galena has been identified in a few polished sections of specimens from the ore body, and one from pyritized volcanic rock below the footwall. The galena is usually found in pyrite, either as veinlets or small replacement bodies, and it rarely reaches more than a few tens of microns in size. The best example of vein galena was seen in pyritized volcanic rock, where small fractures in pyrite and been filled with galena and some tennantite - Fig. 90. The replacement bodies are commoner, and occur as rounded blebs or squarish bodies close to the edge of the pyrite. Some of the galena has replaced chalcopyrite bodies, but most has replaced pyrite directly. Galena was seen outside pyrite in only one specimen, where it occurred as small particles lying between a pyrite vein and the granular magnetite which formed the host. Many of the galena grains contained small relics of sphalerite.

Galena is later than pyrite and chalcopyrite and probably later than sphalerite, but it is penecontemporaneous with, or slightly earlier than, tennantite. It is earlier than the secondary copper sulphides.

*Note.

Since the paragraph on bismutoferrite was written it has been discovered that the mineral was identified in South Terras mine, St. Stephen-in-Brannel, Cornwall (Min. Mag., vol. 33, p. xc, 1964).

XII MINERALOGY OF BUKIT IBAM. (4) OTHER ORE MINERALS

Tennantite

This mineral occurs as grains of microscopic size in polished sections from two localities, and some of the unidentified minute bodies seen in pyrite may be tennantite. It occurs in a magnetite-talc ore with sulphide patches from N 12,300; W 10,700, where it is usually associated with galena, but also occurs as blebs in chalcopyrite. It is younger than the galena, which it appears to have partially replaced. There is often a thin zone of chalcocite-neodigenite between the tennantite and the galena. Tennantite is found with galena filling fractures in pyrite in a pyritized volcanic rock below the footwall along section N 11,900, as shown in Fig. 90. The mineral has a very distinct greenish colour and probably lies close to the antimonian end of the tennantite-tetrahedrite series.

Molybdenite

This mineral has been identified in only one polished section from sample No. 17684 of the University of Durham collection, which comes from "the northern end of the mine". The sample is almost pure sulphide, with abundant chalcopyrite and pyrite. The molybdenite occurs as rare, minute tabular crystals, and as long flakes which have been bent slightly due to pressure. One euhedral crystal completely enclosed in pyrite suggests that the molybdenite crystallized before the pyrite, and as such could be the earliest of all the sulphides in the Bt. Ibam body. This molybdenite grain shows displacement along the (0001) cleavage, possibly caused by the growing pyrite. It is improbable that the displacement took place after the pyrite crystallized as there is no sign of fracturing in the pyrite adjacent to the molybdenite crystal.

Fig. 90. Galena and tennantite filling fractures in pyrite. x 750.

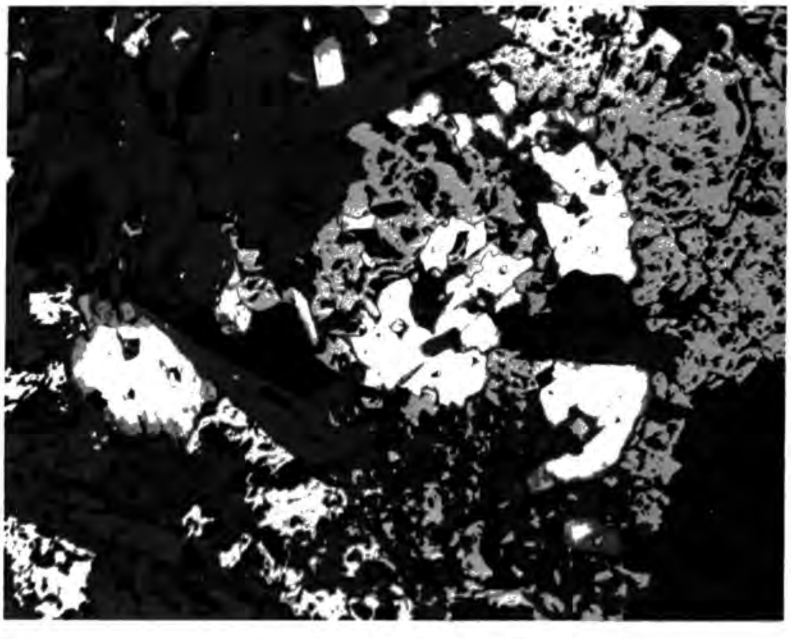


Fig. 91. Arsenopyrite (white) with magnetite (medium grey). The greyish material on the edge of some arsenopyrite grains is an oxidation product. x 125.

Arsenopyrite

This mineral has been observed in one polished section of a sample from N 12,300; W 10,800, where it occurs as anhedral grains associated with fine grained magnetite - Fig. 91. The arsenopyrite is later than the magnetite and contains blebs of a greyish mineral, probably sphalerite. The periphery of some of the arsenopyrite grains has been altered to a greyish mineral with yellow internal relections, probably an oxide of arsenic.

Pyrrhotite

Rare, minute replacement grains of pyrrhotite have been identified in some pyrite crystals.

Gold-Electrum

Gold is a rare mineral in Bt. Ibam, but has been identified in three samples, all from the northern end of the mine. The gold has been introduced at a medium to late stage, as it clearly post-dates pyrite and chalcopyrite, but its age relative to the other sulphides is unknown. It occurs as blebs up to 5μ across and veinlets up to 25μ long in pyrite, and replaces chalcopyrite veinlets in pyrite. In polished section the gold is pale coloured, and shows no reduction of reflectivity in green light, so it is probably electrum rather than gold. Fig. 92 shows a gold bleb in pyrite, and Fig. 93 shows gold replacing chalcopyrite veinlets in pyrite.

Mixed Manganese Oxides

Although they are not abundant, small masses of mixed manganese oxides are found, particularly in the southern half of the mine. The best examples seen by the writer were obtained from drill core from section N 10,400, where they are often associated with high zinc values. The nature of the minerals shows they were

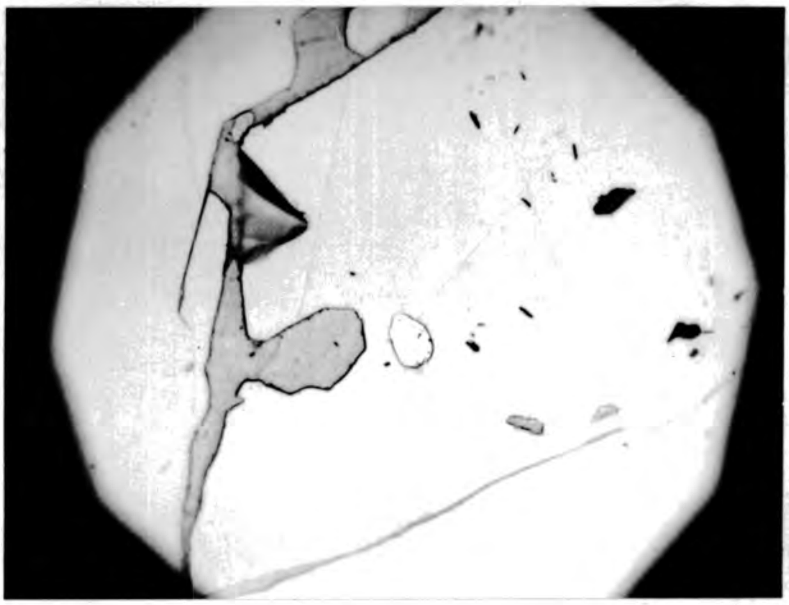


Fig. 92. A chalcopyrite vein and a bleb of gold in pyrite. x 1310.

Fig. 93. Gold which has partially replaced chalcopyrite veinlets in pyrite. x 750.

formed within the zone of oxidation, but they have been located at depths up to 300 ft. In hand specimen the ore consists of black, irregular cellular and porous masses, which can show botryoidal surfaces. They are associated with the normal type of goethitic ore.

It is not easy to identify the component minerals, either in polished sections or by X-ray diffraction. In polished section the ores consist largely of colloform masses of cryptocrystalline material or fine grained crystalline aggregates, often mixed with gangue material. The colloform ore shows a wide range in reflectivity and colour, the latter varying from pale cream to very pale blue. The more crystalline ore varies from stubby laths to acicular crystals, the latter sometimes forming small veins.

It has not been possible to identify the minerals from their optical properties, and although a powder photograph showed a good quartz pattern, the other lines were weak and diffuse, indicating poor crystallinity. The only recognizeable pattern agrees reasonably well with that of cryptomelane, except that a line at 2.14Å is missing. This line is also the strongest one in the psilomelane pattern, so this mineral is either absent or very poorly crystallized.

Manganese Oxides in Veins

In part of the hangingwall near N 11,200; W 10,800 the local bedrock of highly decomposed volcanic rock is cut by quartz veins which are in part associated with earthy manganese oxides. The manganiferous material is later than the quartz, and appears to have been deposited from solutions circulating along the quartz vein system. An analysis for minor elements in the earthy material showed 2.6% Ba, plus small quantities of Bi, Pb, Zn, Cd and Cu. The presence of significant barium indicates that psilomelane is a constituent, but a diffraction trace does not confirm this. The trace is a poor one, but shows a braod peak around 3.11Å and a smaller one at 2.39Å, with other weak peaks. The pattern, which indicates poor crystallinity, does not fit any one manganese mineral, but points to the presence of pyrolusite and possibly some cryptomelane.

Chalcophanite

It has not been possible to confirm the identity of this mineral by X-ray studies, but all the available evidence strongly points to its presence. The fact that some of the secondary manganiferous ores are high in zinc - up to 4.1% - suggests that it may be present, and although the same argument would apply to other zincian manganese minerals, no supporting evidence of their presence has been noted.

A drill core sample from section N 10,400 contains rare acicular crystals, which probably crystallized in a cavity, in a mass of colloform ore. The crystals are greyish-white in colour, with a very strong bireflectance and anisotropy, and a rotation angle of about 15°. The latter appears to be significant as few minerals have a rotation angle of this magnitude, and it points to chalcophanite, which has a rotation angle of about 18° (Cameron, 1961). Apart from a lack of internal reflections the optical properties agree closely with chalcophanite. A microprobe scan for Cu and Zn made across the part of the sample which contained the acicular crystals showed very strong zinc peaks over the crystals, while the copper remained fairly constant - Fig. 94. The proof that the acicular crystals contain significant zinc, plus the evidence of their optical properties indicates that they are chalcophanite.

Strontiobarytes

This mineral has been identified in veinlets which cut the granodiorite below the northern end of the ore body. The sulphate is found in cavities along the centre of siderite veinlets, and is later than the carbonate. The strontiobarytes occurs as very thin transparent discs, up to 0.5 cm across, which have a hexagonal outline. There is a suggestion that the crystals are

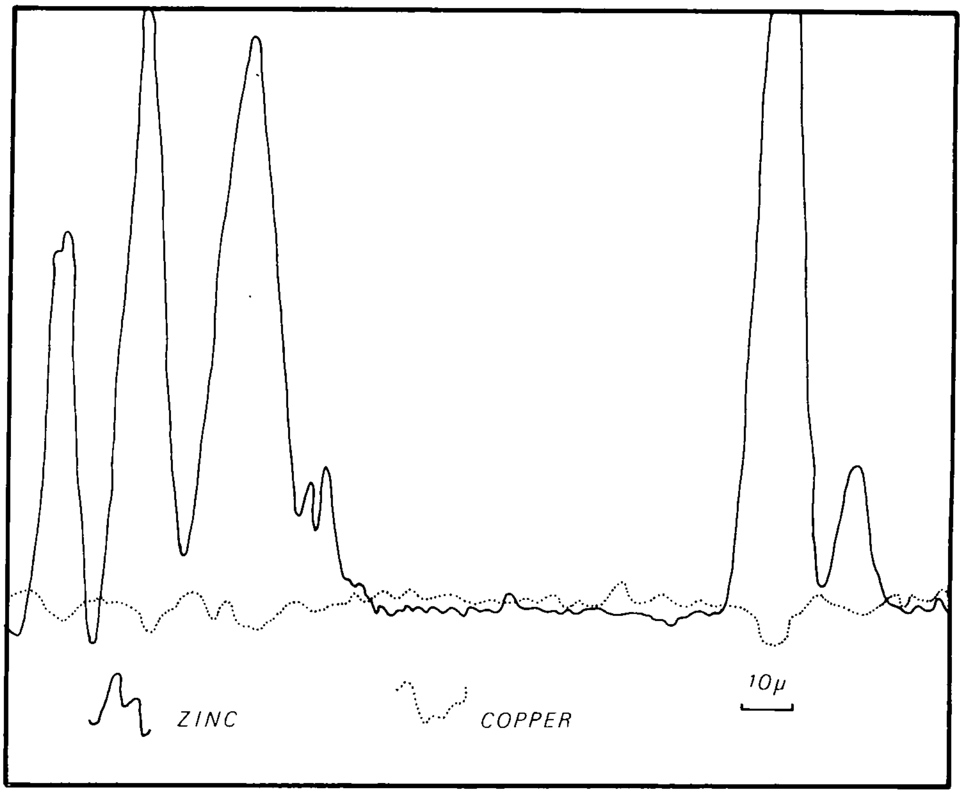


Fig. 94 Microprobe traverse across chalcophanite (?) crystals in colloform ore .

twinned. The X-ray diffraction pattern shows that the composition of the crystals lies between barytes and celestite, and assuming that there is a progressive change in the diffraction pattern from barytes to celestite, the pattern for the Bt. Ibam sample indicates that it contains roughly 25% of the celestite molecule, i.e. it has the composition $Ba_{.75}Sr_{.25}So_4$.

Siderite

This mineral has been identified in veinlets which cut the Bt. Ibam granodiorite, and as somewhat more massive material towards the southern end of the mine, the best specimens coming from drill core along section N 10,400. Siderite has also been identified along section N 10,000, and it is suspected that it may extend to the southern end of the ore body and north to about section N 11,000. The drill core samples are extremely difficult if not impossible to identify in hand specimen due to their variability in appearance, and the fact that many have a superficial coating of goethite. The company drill logs refer to "fragments of nodular limonite", which is an accurate description from external features, and it is only in polished or thin section that the true nature of the samples becomes apparent. The mine authorities recognized that some of the "nodular Limonite" was high in zinc, and assumed that it was contained in the goethite, as they were not aware of the presence of siderite. The writer's investigations have shown that these samples often contain a little sphalerite, but most of the zinc is in solid solution in the siderite itself. The siderite appears to have played an important role in the development of the distribution pattern of zinc towards the southern end of the Bt. Ibam area. The chemical aspects of siderite are considered later.

Along section N 10,400 siderite is found both above and below the ore body, but has only been observed in one specimen which contained magnetite. Along section N 10,000 siderite has been identified in one specimen from below the iron mineralization. Although siderite has been obtained from a considerable length of some drill core it is unlikely that it forms really solid masses. Core recovery in the holes along N 10,400 is often poor, particularly in the upper levels, and compared with the typical decomposed volcanic rock the hard nodules of goethitized siderite would be more likely to be preserved in the drill core. It is probable that the siderite forms veins, small masses and impregnations, but core recovery is not good enough to determine just how extensive it is. The fact that the siderite is restricted to one part of the Bt. Ibam ore zone and cannot be traced through adjacent drill holes to any extent again suggests that the actual siderite occurrences are relatively small. In hand specimen the siderite is variable in appearance; some is light brown, porous and looks like lumps of earthy goethite, while other samples are dense, fine grained and coloured dark brown to black. The latter can look like a mixture of manganese oxides. In some samples siderite is mixed with a flaky brown chloritic material, in which it can also form very thin veinlets made up of small, shiny brown crystals. In polished section the typical siderite sample consists of a fine grained granular aggregate, which is usually somewhat porous. Most of the grains are anhedral or rounded, but euhedral crystals do occur. The siderite is strongly bireflecting and anisotropic, and often has internal reflections which are white to reddish-brown. The smaller grains have been replaced by a mixture of goethite and secondary manganese oxides, but in the larger grains the replacement minerals usually form a corona. A typical siderite sample is shown in Fig. 95. The alteration explains why many of the hand specimens look like lumps of earthy goethite.

Some of the siderite samples from the southern end of the ore body contain sulphides, and a series of progressive mineralogical changes can be traced. Pyrite was the first mineral to form, but was later partially replaced by sphalerite. These samples are the only ones seen by the writer in which sphalerite has extensively replaced pyrite, and it has also retained some of the {100} pyrite faces - Fig. 96. Although the siderite is itself finely granular it sometimes shows gross crystal outlines which must have been inherited from replaced gangue minerals. It is difficult to determine with certainty what the original rocks were, but from their general geological situation they must have been volcanic rocks, although there may also have been some very small carbonate lenses.

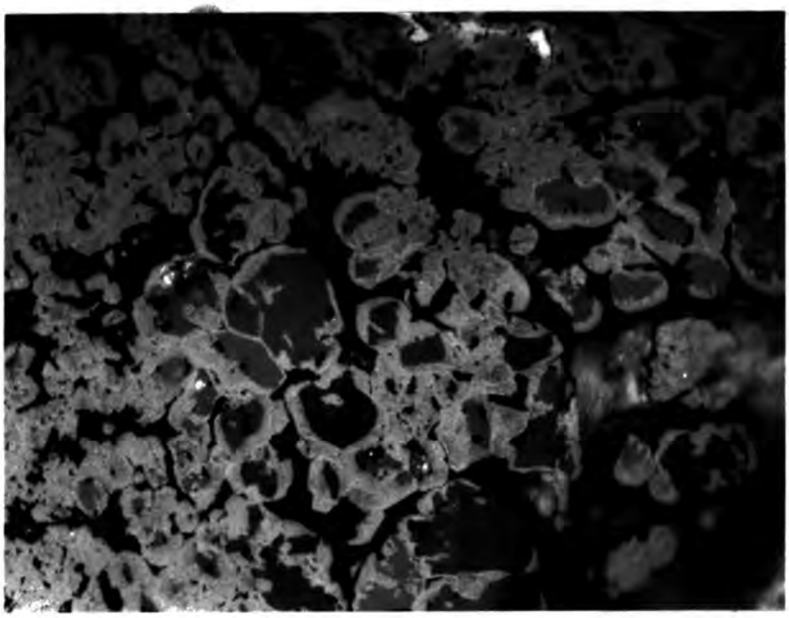


Fig. 95. Granular siderite (dark material in the centre of grains) which has been partial replaced by a mixture of goethite and secondary manganese oxides. x 550.

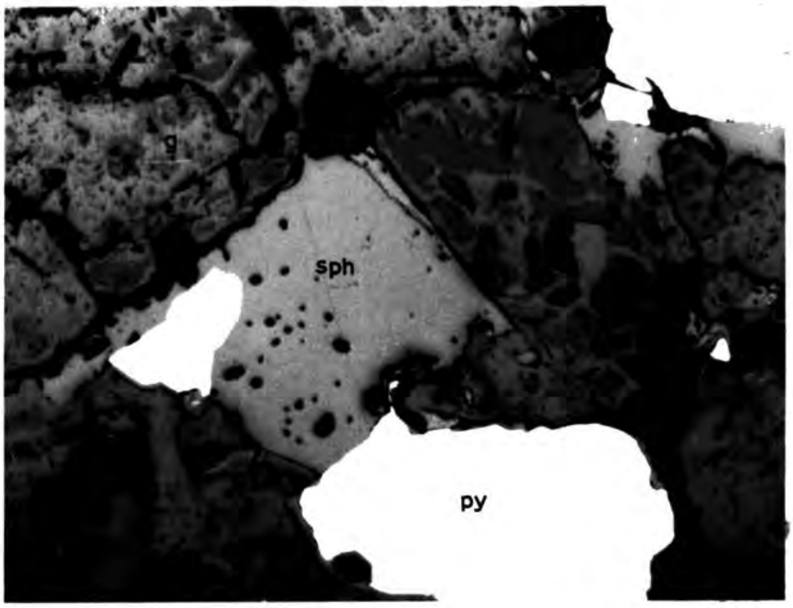


Fig. 96. Pyrite which has been partially replaced by sphalerite. Note the preservation of some {100} faces. Both sulphides have been attacked by siderite, which has been partially altered to goethite. x 210.

The writer believes that the siderite which contains appreciable zinc has obtained at least part from replaced sphalerite, and has acted as a dispersing agent for the zinc. It is significant that section N 10,400, which contains some highly zincian siderites, also contains the only megascopic sphalerite mineralization seen anywhere in the Ulu Rompin area. As siderite is so readily replaced by goethite it would easily give rise to goethites with an elevated zinc content.

Siderite is later than magnetite, pyrite and sphalerite, and must be relatively late in the paragenetic sequence. Its age relative to the cupriferous minerals is unknown.

Chemical Composition

The composition of the zinc-rich siderite is most interesting, and comparable analyses do not appear to have been recorded previously. The writer made a 2θ scan on a Geoscan microprobe to check which elements were present, and this revealed that iron, zinc and manganese were the only important metallic constituents. Dr. K. Ashworth subsequently made complete analyses of four points in the same sample, and his results are shown in Table 16. The lower part of the table shows the elements recast into carbonate molecules.

Table 16

Microprobe Analyses Of Zincian Manganiferous Siderite

	1	2	3	4
Mn	5.81	5.93	7.83	6.42
Fe	26.02	25.72	26.55	24.79
Ca	0.04	0.04	0.72	0.02
Mg	0.06	0.03	0.91	0.03
Zn	17.89	16.25	14.35	19.01
$MnCO_3$	12.13	12.46	16.37	13.41
$FeCO_3$	53.86	53.25	54.96	51.32
$CaCO_3$	0.09	0.09	1.80	0.05
$MgCO_3$	0.21	0.11	3.16	0.11
$ZnCO_3$	34.35	31.19	27.54	36.49
Total	100.64	97.10	103.83	101.38

Dr. Ashworth commented "analyses 1 and 4 seem quite good, and the other two confirm that the composition is uniform throughout". In polished section the siderite looks homogenous, even under the highest magnification. No mineral with the above composition appears to have been recorded, the nearest being an analysis of a zincian siderite quoted in Palache, Berman & Frondel vol. 2 (1951), the analysis being found in the smithsonite section, but the manganese is much lower. Capillitite is a carbonate which contains essential iron, zinc and manganese, but the manganese is the most important constituent and the mineral is described as "ferroan zincian rhodocrosite" by Deer, Howie & Zussman vol. 5 (1962). The Bt. Ibam mineral is a siderite containing approximately 50% siderite, 33% smithsonite and 17% rhodocrosite, and as it is homogenous it must be a solid solution mixture of the three molecules. The solid solution relationships between the main carbonate molecules are not fully understood and data in the literature can be confusing; e.g. Palache, Berman & Frondel vol. 2 (1951) state on page 168 that zinc substitutes for ferrous iron in siderite "in amounts up to about 2 weight per cent ZnO", yet on page 178 they quote an analysis of a siderite with 26.3% ZnO. There is a complete solid solution series from siderite to rhodocrosite, but there is thought to be a gap between siderite and smithsonite. As the Fe" and Zn" ions are exactly the same size, 0.83Å, one would expect the two elements to be completely interchangeable. Deer, Howie & Zussman say that sinc can replace ferrous iron in siderite but do not quote any limits.

Diffraction

The diffraction pattern obtained by the writer for a sample of zincian manganiferous siderite is shown in Table 17A. The sample originated from a depth of 100 ft below surface and 60 ft away from the hangingwall contact of the ore body along section N 10,400, and in hand specimen looked like a piece of nodular goethite. The hand sample contained about 10% zinc. A study of the table shows that the pattern is very similar indeed to the siderite pattern recorded on ASTM 12-531, although the latter has resolved the 1.728Å and 1.082Å lines in the Bt. Ibam zincian siderite pattern into two closely spaced lines; the 1.728Å line

Table 17A

Diffraction Pattern For Zincian Manganiferous Siderite

Co Radiation

122 Bt. Ibam		ASTM 12-531	
dÅ	Rel. I.	dÅ	Rel. I.
3.59	6	3.59	25
2.786	10	2.789	100
2.346	4	2.341	20
2.130	4	2.131	25
1.964	4	1.962	30
1.792	3	1.794	16
1.728	7	1.735	35
		1.730	45
1.527	3	1.527	20
1.507	3	1.505	20
1.425	3	1.425	16
		1.395	8
1.353	1	1.353	20
1.198	1	1.199	18
1.082	2	1.086	18
		1.081	25
1.065	1	1.066	18
0.9812	2	0.9815*	
0.9710	1	0.9717*	

* Taken from ASTM 8-133

is quite broad on the powder photograph and could easily contain two lines. The patterns are remarkably alike when the amounts of zinc and manganese in the Bt. Ibam sample are taken into account, and the lattice appears to have been virtually unaffected by the substitutions. The word "appears" in the last sentence is used purposely, as both zinc and manganese substitution must affect the size of the siderite cell to some extent because the size of the smithsonite and rhodocrosite cells differ from that of siderite. The rhodocrosite cell is larger and that of smithsonite smaller, and it is probable that the tendency of manganese to increase the size of the siderite cell has been balanced by a reduction due to zinc. That this is a possible explanation is supported by a theoretical calculation which can be made. If one assumes a mixture of rhodocrosite and smithsonite in the same proportion that they bear to each other in the Bt. Ibam sample the probable cell size can be calculated, and it is found that this is very close to that of siderite, as shown below:

	$FeCO_3$	$(Zn_{0.66}Mn_{0.34})CO_3$
a_o	4.69Å	4.69Å
c_o	15.30	15.24

Subsequent to the writer's investigation of the diffraction pattern, Dr. R.G. Hardy of the University of Durham made a further study of the sample which was used for the chemical analyses quoted in Table 16. This new data is given in Table 17B, together with the cell dimensions calculated from the diffraction data. It can be seen that the diffraction pattern is very similar indeed to that obtained earlier by the writer, and the cell dimensions are close to those of pure siderite.

Table 17B

Diffraction And Cell Data For Zincian Manganiferous Siderite

Co Radiation

dÅ	Rel. I.	hkl
3.59	30	012
2.788	100	104
2.345	20	110
2.134	20	113
1.961	20	220
1.794	10	024
1.729	70	018, 116
1.505	15	122
1.424	15	214
1.390	5	208
1.353	15	030
1.197	10	128
1.081	15	134
1.066	10	226
0.982	15	404, 2.1.12
0.971	15	318
0.928	15	3.0.12

a_o 4.694Å ± 0.002

c_o 15.321Å ± 0.007

a_{rh} 5.774Å

$_{rh}$ 47°44'

XIII MINERALOGY OF BUKIT IBAM. (5) GANGUE MINERALS

Introduction

The mineralogy of the gangue is relatively simple and uniform, even when the alteration products are taken into account, and must reflect a general homogeneity of the original sedimentary material. The dominant gangue minerals, primary and secondary, are rich in magnesia. The only important primary minerals have been actinolite and chlorite, both of which have suffered extensive alteration to talc with a little antigorite and kaolinite. Quartz is locally abundant, but has been introduced later than most of the metallic mineralization. Nontronite occurs in small patches, and epidote can occur sporadically close to the wall rocks. In general the gangue north of N 11,600 is largely made up of magnesia-rich minerals, while south of this line kaolinite becomes more important in some sections, although magnesian minerals still occur.

Contacts

The contacts are relatively sharp and there is no difficulty in delimiting the ore body. The footwall in particular is well defined; north of N 12,100 the ore body is bounded by the main fault - Fig. 97, while south of this line the contact is a plane of movement parallel with the ore body. The iron ore rests on a surface of soft chloritic material, which shows good examples of gouges, slickensides and polishing - Fig 98, due to slumping movements within the ore body. In some places the actual plane of movement is covered with a bright blue mineral - page 214. The sharp demarcation between ore and wall rock can be clearly seen in Fig 99. The hangingwall contact is not quite as sharp, but is still clear. Except for the zone between N 11,600 and N 10,800 there is massive chlorite between the ore and country rock. The chlorite shows many small planes of movement parallel with the contact, but has not acted as a sliding plane like the footwall contact. There has been some chloritization and epidotization of the acid volcanic rocks close to the contact.

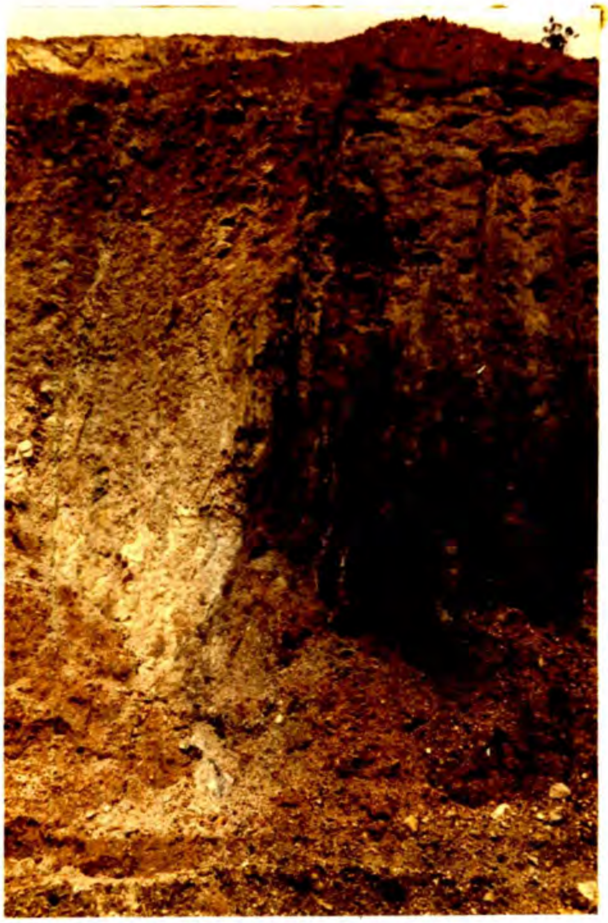


Fig. 97. Iron ore faulted against volcanic rocks.

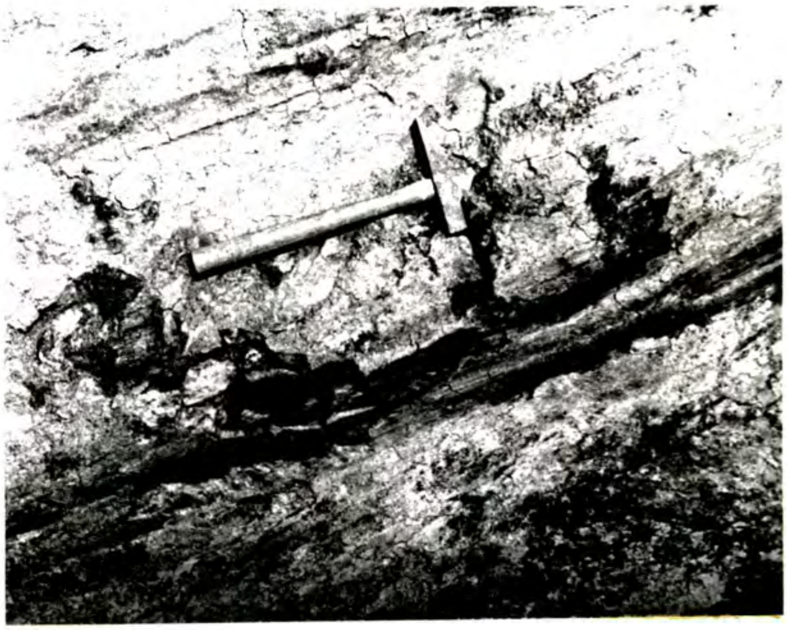


Fig. 98. Gouges developed along the footwall contact.

Fig. 99. View of the footwall contact showing the clean break between ore and the wall rocks.

Between N 11,600 and N 10,800 the iron ore comes directly against acid volcanic rocks, which are highly iron stained, and there are small patches of decomposed country rock within the ore body proper.

Actinolite

Actinolite has been the most abundant mineral formed during the metamorphism and metsomatism of the original carbonate lens, and although much has been subsequently altered relics of the primary material have been preserved. In hand specimen the actinolite is medium to dark green and fibrous, some being asbestiform with fibres up to 1 cm long. The orientation of the fibres can be random or in radiating clusters, the texture often being perfectly preserved throughout the alteration to talc, As alteration progresses the actinolite is bleached through pale green to white, the end product being talc with some chlorite and antigorite. Those mine reports and drill logs which mention tremolite undoubtedly refer to talc-antigorite pseudomorphs of fibrous actinolite. Perfect quartz pseudomorphs of actinolite have also been observed. The best examples of actinolite are found at the northern end of the mine where patches have been preserved by silicification.

The amphibole has exerted some control over the crystallization of the sulphides as pyrite, chalcopyrite and tetradymite have all been observed as parallel growths of minute laths along the amphibole cleavage planes, and goethite can develop in a reticulate pattern under the same control. Large crystals of chalcopyrite moulded round amphibole have also been observed. The actinolite has rarely influenced the crystallization of magnetite, but one occurrence of bladed magnetite could indicate pseudomorphic replacement of actinolite, and magnetite has occasionally grown round the periphery of amphibole grains - Fig. 44. The metallic minerals are usually later than the amphibole, but a little actinolite may have formed contemporaneously with magnetite.

The diffraction patterns for two samples taken from solid actinolite masses, one of which was fibrous and contained disseminated magnetite while the other was asbestiform but lacked ore minerals, were closely comparable with ASTM 7-336. Some lines obtained by the writer and which are not recorded on the ASTM pattern have also been detected during diffraction studies of amphiboles carried out in the University of Durham by G. Rowbotham and P. Broomfield. The cell dimensions for the two samples from Bt. Ibam are shown below, with data from Zussman (1959) for comparism.

	Bt. Ibam		Zussman
a_o	9.861Å	9.845Å	9.861Å
b_o	18.053Å	18.085Å	18.110Å
c_o	5.246Å	5.246Å	5.336Å
β	105°40'	105°18'	104°59'

Chlorite

Chlorite has formed an important gangue mineral throughout the ore body, but it reaches its maximum development along parts of the hangingwall contact. North of N 11,600 a continuous zone of massive chlorite rock, up to 30 ft wide, is found between the ore body and the volcanic rocks. Between N 11,600 and N 10,800 the chlorite is absent from the hangingwall, but reappears to the south, although not as strongly developed. Chlorite is more sporadic along the footwall where it occurs as thin intermittent lenses. Relics of chlorite are found within the ore body, but much has been altered to secondary magnesian silicates. The chlorite along the hangingwall is very distinctive in the mine pit because of its dark green colour, which contrasts strongly with the paler volcanic rocks. The chlorite masses are usually full of polished, slickensided surfaces, which have a general direction parallel to the contact. The hangingwall chlorite proved to be a hindrance to tunnelling operations during prospecting, as it continually caved and large volumes of water flowed through the zone. At the extreme northern end of the ore body where the mineralization has been reduced to very small lenses, drill core samples show the presence of chlorite-calcite rock.

The chlorite forms bands and patches in calcite, and this type of material possibly represents an early stage in the alteration of the original carbonate rocks. The magnesian part of the dolomite has reacted to produce chlorite, while the calcite component has proved less reactive. There is little or no actinolite in this rock, which suggests that chlorite formed before the actinolite.

In thin section the massive chlorite is made up of fine feathery masses of small flakes which are pale green to almost colourless. The interference colours are extremely low, some grains being almost isotropic, and anomalous blues and brown colours are common. The rocks are virtually monomineralic, but a little epidote can occur. The chlorite along the hangingwall north of N 11,600 is noteworthy for the presence of some good euhedral pyrite crystals. Magnetite is typically absent from these rocks, but appears with proximity to the ore body proper, and fine, disseminated magnetite is found in the footwall chlorites. In the higher levels of the ore body the chlorite weathers to a tenacious limonitic clay, but it is often fresh in the lower levels.

Diffractometer traces for several samples rich in chlorite were obtained, and two relatively pure samples were powder photographed. As most chlorites have a similar diffraction pattern an individual pattern is not usually diagnostic, although it can indicate certain chemical characteristics. The diffractometer traces for the four chlorites were essentially similar, although there was some variation in the intensities of the basal reflections. Powder photographs of two samples of nearly pure chlorite, one (No. 189) coming from the hangingwall contact and the other (No. 61) being a hard chlorite "ball" from a fault zone, were obtained, and the diffraction data for No. 189 used to calculate it cell dimensions, which are shown below:

	No. 189	Leuchtenbergite
a_o	5.316Å	5.32Å
b_o	9.207Å	9.21Å
c_o	14.339Å	14.29Å
β	97°00'	97°08'

The figures for No. 189 are extremely close to those for leuchtenbergite (ASTM 12-242), suggesting that the Bt. Ibam chlorite is rich in magnesia.

Samples Nos. 189 and 61 were chemically analysed, the results being shown in Table 18. In the unit formula there should be 12 ions in the octahedral sites, and as can be seen from Table 18 the Bt. Ibam chlorites are in good agreement with this, No. 189 particularly so. The chemical analyses can be used to classify the chlorite species. No. 61 is an oxidized chlorite (> 4% Fe_2O_3) while No. 189 is unoxidized. Putting the unit formula data into the classification diagram proposed by Hey (1954) shows that No. 61 is delessite and No. 189 is clinochlore (syn. leuchtenbergite). The chemical classification of No. 189 agrees precisely with that based on cell dimensions. The chemical and cell size data are also in good agreement with the diagram shown as Fig. 34 in Deer, Howie & Zussman vol. 3 (1962), which relates the b dimension to Fe + Mn content.

General geological considerations suggest that No. 189 is typical of much of the chlorite found at Bt. Ibam, which can therefore be classed as clinochlore. The oxidized nature of No. 61 may be connected with the fact that it has been subjected to stresses in a fault zone.

The d.t.a. curves for the two chlorite samples show a close similarity to curve B in Fig. 75 in Grim (1951), which is for clinochlore. The curves for the two Bt. Ibam samples are similar, but the reaction temperatures are approximately 20° lower for sample 61. This may be connected with the extra iron content, as it is known that in some mineral groups the presence of iron lowers the reaction temperatures e.g. in the montmorillonite-nontronite series (Kerr et al, 1949). There is a small exothermic reaction at 950° in No. 61 which is absent in No. 189.

Talc

Talc is the most abundant gangue mineral and can form large masses of pure mineral, but more typically it contains

Table 18

Chemical Analyses Of Two Chlorite Samples

	61	189
SiO_2	35.09	33.00
Al_2O_3	17.01	18.27
Fe_2O_3	4.88	1.15
FeO	4.84	1.98
MgO	28.16	35.24
CaO	0.04	0.16
Na_2O	0.16	n.d.
K_2O	0.30	0.06
TiO_2	0.46	0.52
MnO	0.39	0.09
S	0.09	0.20
H_2O	8.58	9.33

The ions per unit formula were calculated by assuming 28 O equivalents and ignoring H_2O.

	No. 61		No. 189	
Si	6.41	8.00	5.90	8.00
Al	1.59		2.10	
Al	2.05	11.41	1.96	11.95
Fe^{3+}	0.68		0.14	
Fe^{2+}	0.73		0.30	
Mg	7.71		9.46	
Ca	-		0.03	
Na	0.06		-	
K	0.06		-	
Ti	0.06		0.06	
Mn	0.06		-	
OH		16.00		16.00

a proportion of one or more of antigorite, chlorite and actinolite. The masses of pure talc are off-white, but the associated minerals produce a greenish colour. In mahy places the talc has been iron stained from a pale yellowish-brown to deep rusty or olive brown. Some masses have been completely stained, but in others the staining is restricted to zones surrounding ferriferous minerals.

The talc masses can be structureless, but many have a fibrous or flakey texture preserved from the replaced actinolite and chlorite. Where talc has been caught in a fault or zones of slumping it exhibits many highly polished slickensided surfaces, and sometimes contains rounded, polished "balls" of talc. Talc can form masses up to 25 ft across and varies down to small specks in iron ore.

In thin section talc usually occurs as myriads of randomly orientated minute flakes, but can occasionally be seen in larger flakes due to pseudomorphic replacement of actinolite or chlorite. There has been some mobilization of talc during the process of alteration, as it forms veinlets in some ore minerals, and in rare cases appears to be replacing them. There must have been a very effective leaching action during the formation of talc, as the white masses contain virtually no iron, indicating that this element has been removed during the breakdown of actinolite and chlorite.

The diffraction patterns obtained by the writer for several talc samples agree closely with ASTM 13-558, although the relative intensities are not always consistent, probably due to the effects of preferred orientation in the sample mount. Some of the non-basal reflections are suppressed, particularly when talc is only part of a mixture of minerals, but there is never any doubt as to its identification.

Antigorite

Antigorite has been identified in some diffractometer traces of talcose gangue, but has not been observed visually or microscopically.

Nontronite

Small pockets of nontronite have been observed in the gangue in two or three localities in the northern half of the ore body; one patch was up to 12 inches across and associated with strongly limonitized talc. The nontronite is found as an irregularly distributed, yellowish-green greasy clay (almost epidote green), some patches containing magnetite. Thin yellowish-green coatings seen on magnetite in other parts of the ore body probably contained nontronite.

A diffractometer trace showed a very strong peak around 14.6Å, with weak peaks at 4.56Å and 4.19Å. Treatment with glycerol shifted the basal reflection to 17.9Å; a water-saturated sample showed a strong peak at 19.0Å, which returned to 14.4Å when the sample was dry.

The d.t.a. curve shows peaks at 155° and 465°, representing the loss of absorbed water and lattice (OH) respectively. Grim (1951) states that nontronites lose their lattice water around 550-600°, but the temperature from the Bt. Ibam material is lower, possibly due to weaker Fe-OH bonding (Kerr et al, 1949). A break in the curve at 850° marks the start of the breakdown of the structure, and material heated to 1000° is completely amorphous. A strong exothermic peak at 1153° represents the formation of new compounds, and a sample heated to 1200° was shown by diffractometer to consist of a mixture of haematite and cristobalite.

Kaolinite

Although kaolinite is common in the acid volcanic wall rocks it cannot be called a typical gangue mineral, and is most frequently encountered south of N 11,600. Between N 11,600 and N 10,800 the hangingwall of the ore body is directly in contact with volcanic rocks, and south of N 10,400 the footwall skarn dies out and volcanic rocks come against the ore body. There are small patches of decomposed volcanic rock within the ore body, and a mottled clay forms much of the gangue close to the hangingwall.

This clay can contain small rectangular white bodies which appear to be altered feldspars, and the main constituent of this clay was shown by diffraction studies to be kaolinite. The alteration of chlorite to talc must release some alumina, which probably remains as kaolinite.

In the field it is difficult to identify with certainty any particular patch of clayey gangue as being rich in kaolinite, but two features point to its presence; the kaolinite never has an inherited fibrous or flaky texture, and it tends to be much more variegated in colour, with shades of red and purple being common. In contrast, when the talcose gangue is stained it is usually in shades of brown.

Quartz

Quartz is found throughout the ore body, but is abundant in only rare patches. Where evidence of its age is available it is clearly later than the metallic mineralization, and the deposition of opaline quartz from circulating ground water solutions is still actively in progress. The quartz is found as veins and lenses up to two inches wide, in cavities, in impregnations, and along joints.

The veins cut through ore and gangue with no preferred orientation, and are of short extent. The quartz is normally broken, and stained or cemented by iron minerals. The best examples of quartz impregnation are found around N 11,900; W 10,900 where a brecciated disseminated magnetite ore has been thoroughly impregnated by quartz, and the original actinolite and chlorite gangue has been effectively preserved from alteration. In the same general locality there are impregnations and patches of opaline quartz in a clayey chloritic gangue, which contains magnetite and copper with cuprite. This quartz is of recent origin and has been deposited from circulating ground waters. The quartz is coloured in shades of white, grey and brown, and can show a banded texture.

Near the hangingwall contact at N 11,500; W 10,900 there are small patches and lenses of quartz in friable magnetite, and also present are small cavities within the ore which are lined with water clear crystals of quartz. Some nearby masses of solid magnetite contained small patches of a white fibrous mineral, which proved to be quartz pseudomorphs of actinolite.

Quartz veins are found in the wallrocks. Some of the quartz veins in the skarns have been brecciated and cemented by goethite, while in the hangingwall volcanic rocks quartz has often been deposited along small planes of movement.

Calcite

Calcite is uncommon within the ore body itself and the writer has seen only a few thin veins. In thin section and polished section calcite veinlets and impregnations are occasionally observed, and it is clear that the calcite is a late-stage mineral as it invariably replaces or transects quartz veins where the two occur together. This refers to the hydrothermal quartz, not the opaline quartz of very recent origin.

Calcite is more abundant at the extreme northern end of the ore body, where it occurs with chlorite near the small lenses of mineralization. It is thought that this calcite is the residue from the partial decomposition of the original dolomite.

Blue Mineral along the Footwall

A noteworthy feature of the exposed footwall contact between N 10,600 and N 12,100 is the presence of patches of a strikingly blue mineral. It is usually seen as a thin covering on slabby polished surfaces, and is most distinctive due to its bright and unusual colour - Fig. 100. Some company drill logs refer to "blue chlorite" from the footwall, so it probably occurs sporadically throughout this zone. The blue mineral has also been observed along small faults within the ore body, but again close to the footwall. There seems to be a genetic connection between the

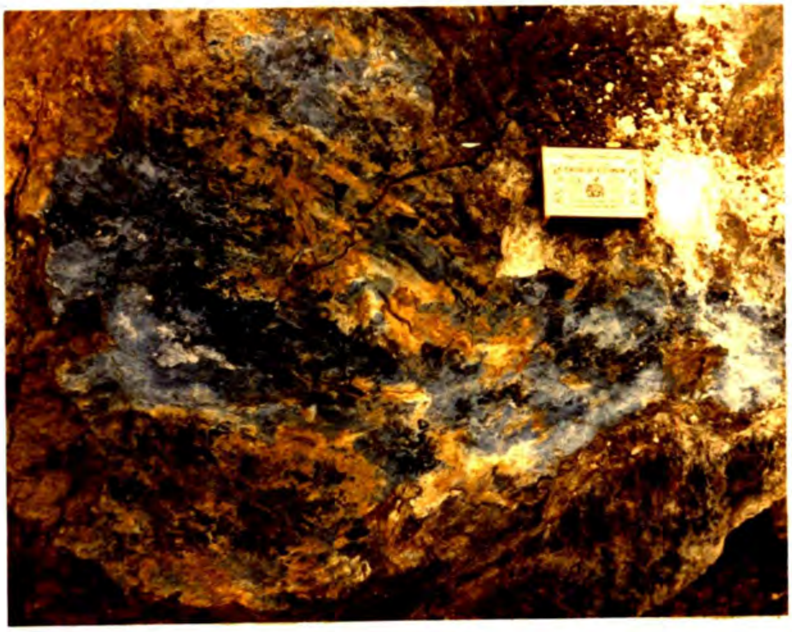


Fig. 100. Bright blue mineral developed on a polished surface along the footwall contact.

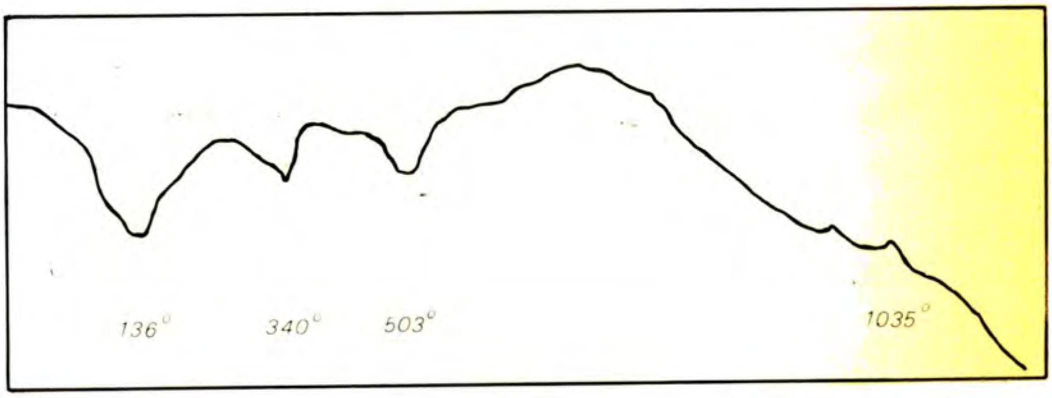


Fig. 101. The d.t.a. curve for the blue mineral found along the footwall contact. The curve suggests a mixture of goethite and a smectite.

development of this blue mineral and planes of movement on and near to the footwall.

It has not proved possible to identify this mineral, even though a number of its properties are known. As pure a sample of solid blue mineral as could be obtained was chemically analysed with the results shown in Table 19.

Table 19

Analysis Of Blue Mineral From The Footwall

SiO_2	57.82
Al_2O_3	1.05
Fe_2O_3	30.31
FeO	0.65
MgO	0.66
CaO	0.25
Na_2O	0.06
K_2O	0.04
TiO_2	n.d.
MnO	0.12
S	0.54
H_2O	8.50

The analysis shows that the mineral has a very simple composition, and is a hydrated iron silicate with the approximate molecular composition $7SiO_2.Fe_2O_3.3H_2O$.

The mineral was also studied by X-ray diffraction, and although the results help to indicate some general structural features the pattern cannot be matched with any of those recorded in the ASTM index. The pattern is shown in Table 20.

Table 20

Diffraction Pattern For Blue Mineral From The Footwall

Co radiation

dÅ	Rel. I.
11.9	100
7.5	10
4.5	10
4.2*	10
3.29	15
1.61	

* this coincides with a goethite peak, but there appears to be a secondary peak superimposed on the goethite. The other goethite lines have been omited.

The diffraction pattern is not well defined and the peaks tend to be broad, but the basal reflection at 11.9Å is very strongly and consistently developed. The pattern suggests that the material is poorly crystalline, and has a well developed layered structure with d_{001} about 12Å. The spacing of the basal reflection does not alter after treatment with glycerol. The mineral must be a clay in which aluminium has been almost completely replaced by iron, and the basal spacing suggests that it is a smectite, as the spacing is too large for kandite, chlorites and illites. The fact that the basal spacing is slightly smaller than is usually encountered in smectites may be due to partial collapse of the structure, and the same reason may explain the non-swelling property. A goethite pattern was clearly recognizable on the diffractometer trace, but the peaks were broad indicating that the mineral was poorly crystalline. This material will account for some of the ferric oxide in the chemical analysis shown in Table 19.

Fig. 101 shows the d.t.a. curve for the blue mineral. The curve shows three endothermic peaks, and appears to be consistent with a mixture of a smectite and goethite. The peaks at 140° and 500° correspond with the loss of absorbed water and lattice (OH) respectively; the temperatures are comparable with those obtained for nontronite - page 212. The endotherm at 340° represents the loss of water from "limonite". Pure goethite has an endothermic peak at 380°-400°, but in limonite the peak occurs at 300°-350°. The d.t.a. data therefore support the diffraction data. The breakdown products after heating to 1,200° were cristobalite and haematite.

It is concluded that the blue mineral is an iron-rich smectite associated with "limonite". The intense blue colour appears to be an inherent feature, and is not due to staining or impurities. Vivianite, which is very similar in colour, is not present, and the concentration of copper is only 34 ppm so copper salts are not responsible.

XIV GEOCHEMISTRY OF THE BUKIT IBAM DEPOSIT

Introduction

The geochemistry of a workable ore deposit can be considered from two different but overlapping points of view. Of primary importance, and crucial to the viability of a mine, is the composition of the mass of ore which can be profitably extracted, processed and sold, but also of interest is the overall distribution and concentration of the major and minor components in the ore, gangue, and surrounding country rocks. The geochemistry of the Bt. Ibam body is comparable with many other iron ore bodies formed in a similar geological environment, but is given additional interest by the number of sulphide and sulphosalt minerals which are associated with the iron oxides. From the results of tens of thousands of chemical analyses performed by the mine operators it is known that the two main metallic impurity elements are copper and zinc, while lead and bismuth occasionally cause trouble but are much less abundant; the writer concentrated on a study of these elements.

Analyses of individual small iron ore samples are of very limited usefulness in themselves unless they illustrate some unusual chemical feature, but they form the vital basic data from which averages are calculated and distribution patterns determined. Averages are essential as a datum against which individual samples can be compared. The figures used in this chapter for average values of the composition of the iron ore have been taken from company sources, and are based on an enormous mass of chemical data: the relatively few analyses which could be undertaken by an individual worker could have no significant effect on such figures. The writer concentrated on determinations of the main impurity elements, in country rock as well as ore, and any unusual chemical features. The whole rock chemistry of the volcanic, plutonic and skarn rocks has been described in earlier chapters.

Bulk Composition of the Iron Ore

The figures which probably most accurately represent the overall chemical composition of a massive, high grade iron ore body are the weighted analyses of the total production. Such figures do not of course give the complete picture, as during treatment the impurities are reduced to a minimum, but the final product represents the bulk of the original ore body. The weighted analyses for the years 1962 to 1965 are shown graphically in Fig. 102. The figures are taken from analyses made by the purchasing steel mills, and are those used for calculating the price of the ore: they are not disputed by the mine authorities, who conduct their own analyses. The figures refer to dry material, and separate curves are drawn for lump and fine ore as they are subject to different contracts. Some ore from Bt. Pesagi is included in the weighted averages, but by far the greater part originated in Bt. Ibam, and the analyses must closely approximate to the overall composition of this body as mined up to the end of 1965.

The curves show some interesting features and it is immediately apparent that during the first three years of operations iron ore of very high grade was produced. This tends to disprove the suggestion made by some mine geologists that the upper levels of the ore body were largely composed of goethite and "limonite", as even if one assumes that the secondary iron ore was pure goethite the bulk composition could not contain more than 60.5% Fe, after allowing for the 4% $Al_2O_3 + SiO_2$. The upper levels obviously contributed most of the early production and must have contained a significant quantity of magnetite and/or martite, a fact which is substantiated by the polished section studies described earlier. The antipathetic relationship between the levels of Fe and $Al_2O_3 + SiO_2$ is to be expected, as apart from trace constituents shipments consist of iron ore and clay. It is clear that considerably more clay survived the treatment procedures in 1965 than during the previous years, possibly due in part to the fact that some alumina and silica is inextricably caught up in goethite.

The curves for lump and fine ore usually show a sympathetic relationship, the main exceptions being copper and zinc in 1965,

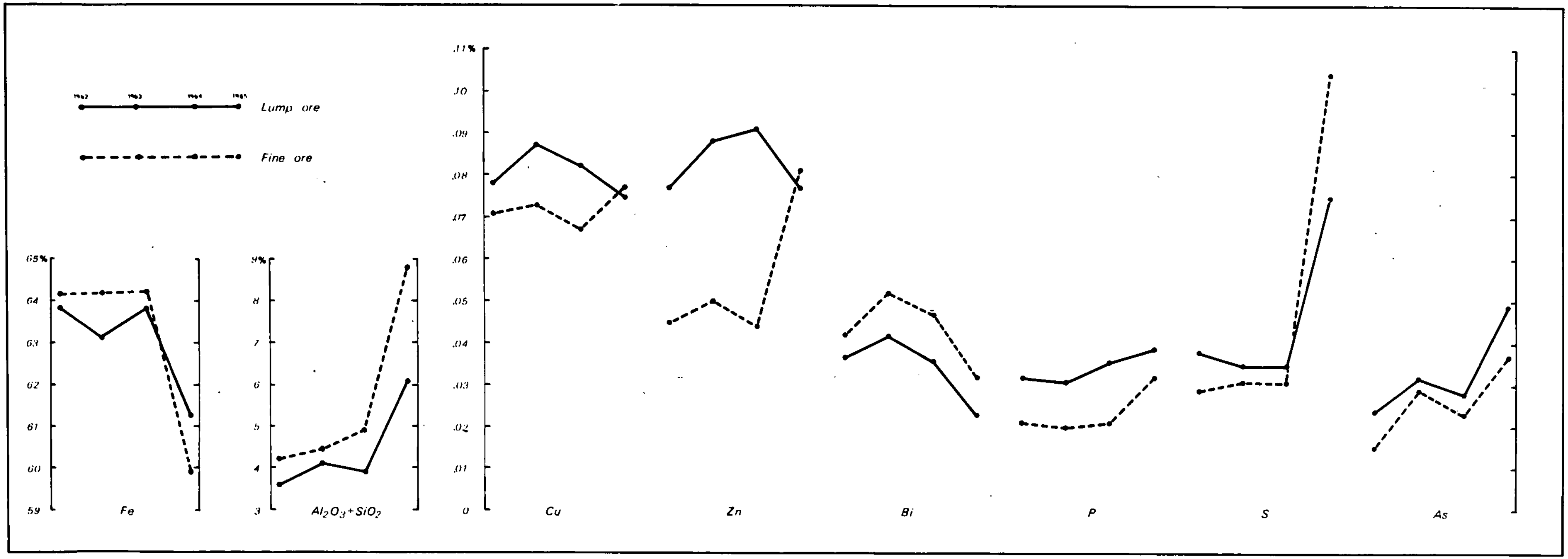


Fig. 102 Graphs showing the weighted averages for the chemical composition of the iron ore exported from 1962 to 1965.

when the concentration in the lump ore fell but rose in the fine ore, markedly so in the case of zinc. The concentrations of the various elements in the lump and fine ore are roughly comparable, usually within 0.01% for the minor impurities, zinc being an exception until 1965.

The curves show that phosphorus and arsenic are low, and sulphur was low until 1965 when it rose in both lump and fine ore. This rise was due to the deeper mining levels reaching the zone where pyrite, the only important source of sulphur, had been less affected by supergene alteration. The only change which appears to be really anomalous is the rise of over 0.03% Zn in the fine ore in 1965. This rise may be due partly to the fact that mining operations at Bt. Ibam gradually progressed southwards towards more zincian ore, but this cannot be the full explanation as much of the zinc is contained in massive limonitic ore, and in fact the zinc in the lump ore shipped from the mine actually fell slightly. During 1965 the company sold some of their highly zincian ore under a special contract in which it was classified as Non-Standard Lump, and the composition of this ore is interesting. The iron content (59.7% Fe) was similar to that in the ore which was sold under the main contract, but the lower $Al_2O_3 + SiO_2$ figure (5.35%) indicates that goethite formed a larger part of the ore. The phosphorus rose slightly while the sulphur fell, arsenic and bismuth stayed roughly the same, copper rose and zinc rose to well over three times its normal level. A comparison of the copper and zinc values in the Standard and Non-Standard Lump for 1965 clearly shows that these two elements, the most important impurities in the Bt. Ibam ore, are not related but vary independently, a fact which has previously been deduced from evidence seen in polished sections.

	Cu	Zn
Standard Lump	0.075%	0.077%
Non-Standard Lump	0.095	0.258
Rise	27%	235%

The analyses for the shipped ore prove conclusively that the metallic impurities are present in substantial quantities in a form other than sulphides. To form chalcopyrite and sphalerite from the copper and zinc in the Non-Standard Lump would require about 0.222% S whereas there is only 0.053%. A similar situation prevails in all the other shipped ore.

The only two important constituents which may occur in the Bt. Ibam iron ore and are not reported in the analyses by the steel mills are lime and magnesia, but these components are not considered deleterious as they help the fluxing process during smelting (United Nations, 1970), which probably explains why they are not recorded. Data in a company file states that the average level of these two components in the shipped ore is MgO 0.3% and CaO 0.1%.

The figures shown below are taken from company files and reflect the overall composition of the ore; they give the calculated composition of all ore remaining as at mid 1965. The word "protore" is used by the mine owners to denote fine, disseminated, largely magnetite ore, which would be susceptible to upgrading by magnetic separation after some grinding. The compositions are given as:

	Fe	Cu	Zn	Bi
High Grade	62.0%	0.09	0.10	0.05
Protore	45.0	0.09	0.03	0.03

As can be seen from these figures, the composition of the remaining high grade ore is similar to that of the lump ore shipped during 1965. The protore will produce a high grade product after magnetic treatment. These figures again show that the levels of copper and zinc are unrelated.

Similar figures are presented in a survey of world iron ore resources by the United Nations (1970), and they must be based on information provided by the mining company.

	Fe	SiO_2	S	P	Cu	Zn	Bi	MgO
Haematite-Goethite ore	62%	8.1	0.05	0.05-0.06	0.05-0.09	0.11	0.04	-
Magnetite ore	47	20.0	0.5	0.03	0.09	-	-	12.0

Minor Elements in Ores and Rocks

Magnetite

Magnetite is the only primary iron ore mineral of any importance in the Bt. Ibam body, and its composition is of considerable interest in view of the number of elements which can substitute for iron in the lattice; a high level of substitutions would profoundly affect the chemistry of the ore body, and any such impurities would not be removable by physical methods. Crushed samples of two single magnetite crystals, one dodecahedral and one octahedral, were scanned (2θ scan) using X-ray fluorescence to check which elements were present in important quantities, and this showed that apart from iron only manganese, and to a much lesser degree copper and zinc, produced recognizable peaks. Apart from the removal of some clay gangue these samples had not been treated in any way.

The writer attempted to extract pure magnetite from a number of samples to compare the trace element content of the pure mineral with that of the original sample. This procedure proved far more difficult than had been anticipated, and even after repeated magnetic separations on one specimen a really pure sample of magnetite was not obtained. This is almost certainly due to the difficulty in separating the magnetite from its alteration products. The original samples were very finely ground (probably 100% was minus 300 mesh) and the powder agitated in water. A hand magnet was suspended in the water, and in theory the magnet should have attracted only the magnetite particles, but experience showed that the initial magnetic fraction needed repeated treatments to remove non-magnetic fragments: conversely, some magnetite could be extracted from the original non-magnetic portion, even after repeated separations.

Two samples, one containing magnetite relics in goethite and the other magnetite disseminated in a clayey gangue, were cleaned repeatedly, and analyses showed that the four main metallic impurity elements were markedly reduced in the magnetic fraction compared with the total sample, while manganese increased slightly. The figures are shown below (in ppm):

		Cu	Zn	Pb	Bi	Mn
9A.	Magnetite in goethite and martite	2375	247	43	32	825
9B.	Magnetite	341	180	18	n.d.	925
	Reduction	86%	27%	58%	100%	
32A.	Magnetite in clay	3.6%	230	34	222	925
32B.	Magnetite	1580	115	n.d.	42	2000
	Reduction	96%	50%	100%	81%	

The results show that most of the manganese is in the magnetite, and zinc is also found in the magnetite lattice in significant quantities. The copper, lead and bismuth on the other hand are largely to be found in the non magnetic fraction, and this is to be expected as none of these elements commonly substitutes for iron in the magnetite lattice.

Tests carried out by consultants to the mining company in connection with ore dressing research confirm that magnetite is depleted in copper and zinc compared with the original samples from which it was obtained. During an investigation into copper-contaminated ore, the British Iron & Steel Research Association found that the distribution of copper in a bulk test sample containing 61.6% Fe and 0.45% Cu was as follows:

	wt%	Cu %	Cu distribution
+100 mesh	24.1	0.63	30.4
-100 +120 "	7.2	.38	5.4
-120 +150 "	14.8	.45	13.2
-150 +200 "	7.8	.33	5.4
-200 "	46.1	.50	45.6
	100.0		100.0

Magnetic separation of the minus 200 mesh fraction gave the following results on analysis:

Non-magnetic	4.40% Cu
Magnetic	0.10

During a similar investigation carried out by Australian Mineral Development Laboratories into a zinc-rich ore the following figures were obtained:

	-80 +100 mesh	-100 +200 mesh	-200 mesh
Head sample	0.242% Zn	0.209% Zn	0.294% Zn
Non-magnetic	0.353	0.342	0.375
Magnetic	0.091	0.098	0.114

It is unlikely that these laboratories made repeated magnetic separations in order to obtain the purest possible samples, as their investigations were directed towards practical processes suitable for a treatment plant. The writer feels that their magnetic fractions probably still contained a significant proportion of martite and goethite caught up with the magnetite.

Four single magnetite crystals, two dodecahedral and two octahedral, were analysed by the writer, and the results confirm the very low levels of lead and bismuth, and the relatively low levels of copper and zinc. Although single crystals were used, they contained some gangue patches and martite and goethite. All had specks of sulphide and No. 17 contained veinlets of pyrite partially replaced by secondary copper sulphides, which accounts for the enhanced copper. In polished section No. 35 was seen to be very fresh and must closely approach to magnetite free of extraneous material. Values below are in ppm:

Method?

Sample No.		Cu	Zn	Pb	Bi	Mn
17	Dodecahedral	2000	179	n.d.	n.d.	4600
34	Octahedral	136	121	n.d.	139	
35	Dodecahedral	283	196	n.d.	17	
67	Octahedral	328	82	n.d.	n.d.	

A number of samples of massive ore which polished sections showed were largely made up of granular magnetite were also analysed, with the following result (values in ppm):

Sample No.	Cu	Zn	Pb	Bi	Mn
10	263	10	n.d.	n.d.	130
13	178	87	n.d.	n.d.	2650
26	77	119	19	23	
179	36	186	n.d.	n.d.	
183	31	84	n.d.	n.d.	
206	46	120	n.d.	n.d.	
228	188	125	175	n.d.	
Average	117	118			

Further information on the composition of the magnetite can be obtained from company data in drill logs. Polished sections were not available for these samples but the iron contents show that the ores have a very high level of purity (pure haematite contains 70% Fe and magnetite 72.4% Fe). The descriptions are taken from the drill logs, and the figures are in percentages (0.1% = 1000 ppm).

	Fe	Cu	Zn	Bi	P	S	As	Mn
Hard magnetite	70.9	0.009	0.007	tr	0.009	-	0.005	0.21
Porous ore	70.5	0.022	0.024	0.011	0.005	0.019	0.005	0.05
Friable magnetite	70.4	0.034	0.024	<0.005	0.009	0.008	-	-
Friable magnetite	70.1	<0.005	0.027	0.011	0.018	0.005	0.005	0.03

The figures for copper, zinc and bismuth are comparable with those obtained by the writer. From a study of all the above figures it can be concluded that the primary magnetite typically contains the four main metallic impurity elements within the following ranges (in ppm):

Cu	50 - 350
Zn	50 - 250
Bi	<50
Pb	<50

The low level of impurities in the magnetite is in agreement with the size of the cell edge reported earlier - Chapter IX - which is close to that of pure Fe_3O_4. It is known that a wide variety of replacement elements affect the size of the unit cell (Gross 1965).

Frietsch (1970) made a detailed study of trace elements in magnetite and haematite in iron ores from northern Sweden, and also reviewed the literature on the subject. The following information which bears on the Bt. Ibam ores has been taken from this work.

Copper

In the upper lithosphere copper averages 70 ppm (Shaw, 1970) but few determinations on magnetites have been made. Hegemann & Albrecht (1954) report that in contact pneumatolytic deposits copper in magnetite can reach 150 ppm, while Frietsch says that in the skarn iron ores of northern Sweden copper occasionally reaches 400 ppm, although the mean value is below 70 ppm.

Zinc

Except in iron ores of contact pneumatolytic or magmatic origin the zinc in magnetite is usually below the average in the upper lithosphere (40 ppm). In contact deposits values up to 9000 ppm have been reported. In the Swedish skarn ores Frietsch records up to 250 ppm in magnetite, but most lie in the 50 to 100 ppm range.

Lead

Most magnetites are low in lead and below the upper lithosphere average of 15 ppm, but values up to 500 ppm in contact pneumatolytic deposits have been recorded.

Bismuth

There are no data given for bismuth in magnetites.

Goethite (including "limonite")

Goethite is found throughout the ore body and far into the country rocks, but reaches its main development in the upper levels of the ore body, especially towards its southern end, and along parts of the hangingwall and footwall contacts. There is also a significant development in the country rock above and below the ore body located along and close to section N 10,400. Much of the goethite occurs in cryptocrystalline colloform masses, and the mobility of the original colloidal solutions has played an important role in the development of the final distribution pattern of the impurity elements. Even without the evidence of chemical analyses it can be deduced that the solutions from which goethite crystallized carried some metallic elements in addition to iron - see e.g. page 162. The obvious chemical feature of the goethite is that it is normally considerably higher in impurity elements than the primary iron minerals, particularly so with regard to copper and zinc, and in some samples lead. There is a very wide range in the metal values carried by the goethite, and some analyses made by the writer are shown below (values in ppm):

Sample No.	Cu	Zn	Pb	Bi	Ba	Mn
6	2800	3500	160	30	10	550
21	1560	4980	535	15	735	-
119	1960	5200	1550	18	n.d.	-
241	1070	1.06%	2350	71	n.d.	-
242	757	1.33%	744	10	14	-
243	755	1690	1222	32	17	-
255	2000	1910	600	45	n.d.	470
256	1070	4180	198	395	n.d.	-
258	4300	3120	1200	40	n.d.	175

It is obvious that any large quantities of goethite carrying the higher values of copper, zinc or lead shown above would be deleterious if included with ore for shipment, and hence the prime importance of strict quality control at the mine.

The following analyses were made in the company's laboratories on samples picked at random by the writer from what is referred to on the mine as the High Zinc Stockpile. The stockpile contained ore excavated towards the southern end of the ore body during the early stages of mining, and was heavily contaminated. Polished sections show that the ore consists largely of goethite with a few relics of magnetite.

	Cu	Zn	Pb	Mn	SiO_2	Al_2O_3	MgO
1.	0.203%	.375%	.04%	.01%	2.2%	.04%	.2%
2.	.255	.268	.18	.06	2.8	.02	.3
3.	.130	.370	.05	.15	1.2	tr	1.0
4.	.190	.340	.02	tr	2.0	.06	.2
5.	.368	.328	.16	.02	1.2	.02	.2
6.	.050	.033	.05	.10	1.1	.04	.6

The above values are roughly comparable with those quoted earlier, although they lack the occasional really high zinc. The two sets of figures show that zinc is high and usually more abundant than copper, the lead values are quite variable and bismuth is relatively low. The manganese figures shown immediately above do not cover the full range as some of the nodular lumps of goethite are manganiferous. This type has usually developed from siderite, and is very high in zinc. There are invariably small unreplaced fragments of zincian siderite and the zinc content of the goethite itself is not known, but it is certainly high.

Except for those goethite samples which contain unreplaced fragments of zincian siderite it is rare for polished sections to display any discrete minerals which could account for the elevated metal values. The conclusion deduced from this evidence is that the contaminating elements are incorporated in the lattice of the goethite. The writer studied a number of goethite samples with the aid of a Geoscan microprobe to determine the distribution of copper and zinc. An analysis of the two sets of figures quoted above for copper and zinc in goethite reveals no consistent relationship between them, and all that can be said is that zinc usually exceeds copper. The lack of a consistent relationship between the concentrations of the two elements has been conclusively proved to extend down to the micro scale.

The writer made microprobe traverses across several goethite samples monitoring the levels of copper and zinc, and these results show that while in some specimens the concentrations of copper and zinc are related in many other specimens they are quite haphazard. Where there is clearly some relationship between the levels of copper and zinc it is not a consistent one, but varies from sympathetic to antipathetic. Fig. 103 shows a virtually perfect antipathetic relationship between copper and zinc, whereas Fig. 104 shows what is essentially a sympathetic pattern, although it is not entirely regular throughout. Fig. 105 shows an interesting traverse across a spheroidal structure in colloform goethite and illustrates the symmetrical distribution of copper and zinc about the centre point. The two elements varied sympathetically in the early phases of development, but later the copper fluctuated while the zinc remained relatively constant. It is clear that the levels of the metallic impurities vary rapidly even over extremely small intervals, and this must be due to fluctuations in the composition of the feeder solutions. ?

Fig. 106 shows a microprobe traverse across a magnetite relic in goethite, and this confirms the low levels of copper and zinc in the magnetite. The zinc curve strikingly demonstrates the marked rise on passing from magnetite to goethite; the difference in the copper levels across the contact is less marked, but it can be seen nevertheless that the magnetite contains a constant low level of copper. The figure incidentally illustrates a completely haphazard relationship between the copper and zinc levels.

Pyrite

Pyrite is the only sulphide which is quantitatively important

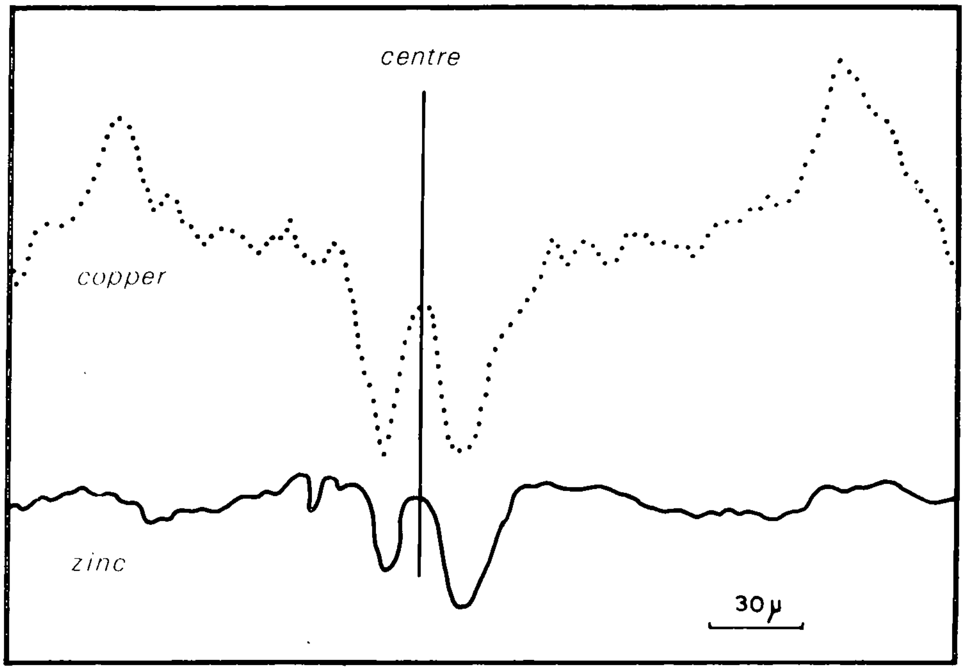


Fig. 105 Microprobe traverse across a spheroidal body in goethite.

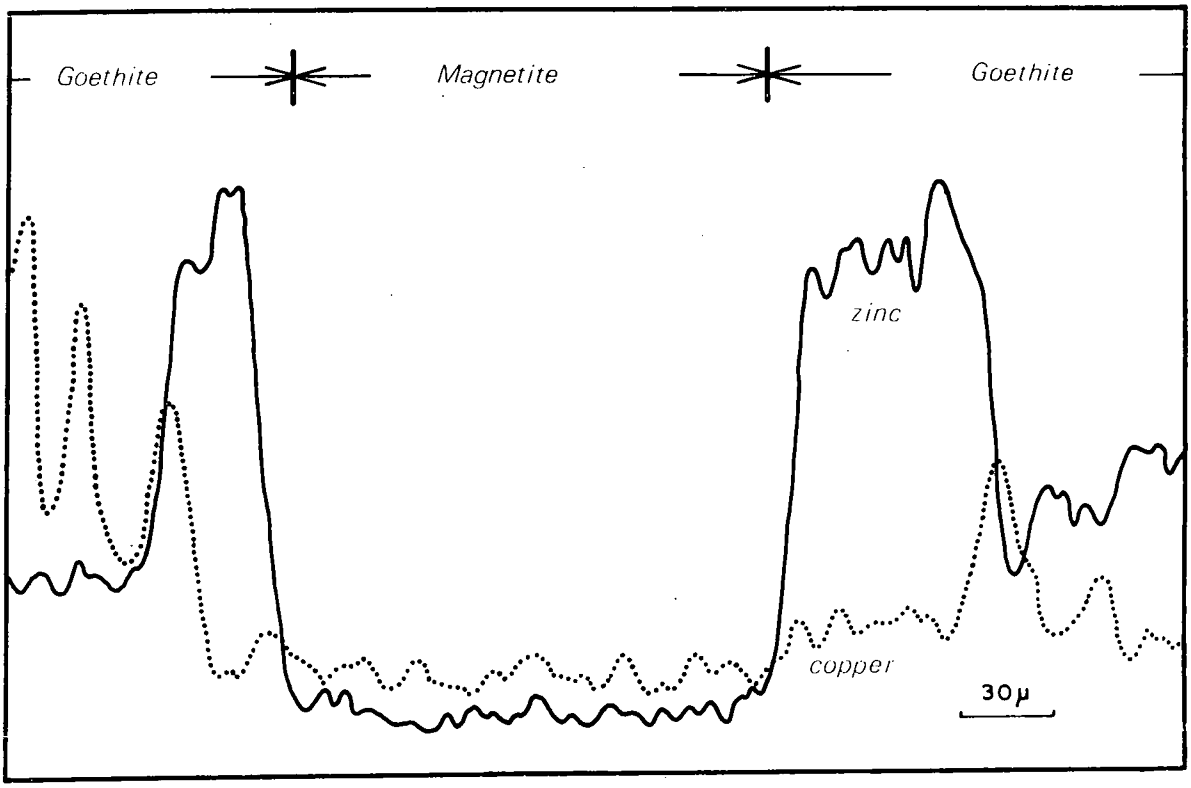


Fig. 106 Microprobe traverse across a magnetite relic in goethite.

and has been observed in polished sections forming a host for microscopic quantities of other sulphides. As has been mentioned previously - chapter X - the well formed crystals of pyrite which form a distinctive feature in some parts of the ore body exist in two differently coloured varieties, one of which is the typical brassy yellow while the other has a black exterior; the black surface can be shiny, dull or irridescent. The two types often occur side by side. Polished sections of single crystals show that the only difference between the two types is the presence of a thin skin of cupriferous material on the black pyrite, and analyses show that the copper content is the only significant distinguishing feature. Readings obtained on a Geoscan microprobe proved that internally the two varieties of pyrite contained precisely the same quantities of copper and zinc, and the excess of copper in the black type is entirely contained in the surface layer. Analyses of one normal coloured and two black pyrite specimens are shown below (values in ppm):

	Cu	Zn	Pb	Bi	Ni
Brassy yellow	203	n.d.	n.d.	n.d.	35
Black cube	4040	18	n.d.	n.d.	80
Black pyritohedron	2260	8	n.d.	52	132

Polished sections showed that these pyrite samples were free of foreign bodies except for the surface skin in the black variety. It is clear that the copper enrichment does not affect the other main impurity elements, but nickel shows a slight increase. 2θ scans on the X-ray fluorescence apparatus confirmed the above results and also proved that no other elements were present in quantity, but there may be a minute amount of arsenic. These low levels of substituted elements are in accord with cell size determinations quoted earlier, which indicate a very high degree of purity in the pyrites.

Data in a company file refers to a richly pyritic iron ore which was briefly studied by the mine staff. The original sample contained 0.172% Cu and 0.042% Zn, and pyrite picked from this sample and cleaned of visible impurities contained 0.066% Cu and

0.022% Zn. These values lie somewhat above those obtained by the writer for normal yellow pyrite, but without a microscopic examination it would be impossible for the mine authorities to see if there were any impurities inside the mineral.

It is apparent that the pyrite contributes little or nothing towards the metallic impurities in the ore body, but it has acted as a host material for several sulphides and sulphosalts, and is the only important source of sulphur.

Acid Volcanic Rocks

The trace elements in a number of acid volcanic country rocks were determined, and the results are shown in Tables 21A and 21B. It has previously been postulated - Chapter III - that there is a slight difference in the whole rock chemistry between rocks from the footwall and hangingwall sides of the ore body, and for convenience this division is retained in Table 21. It is clear however, that with regard to the trace elements determined by the writer the rocks from both divisions are virtually identical. The rocks have been listed in order of increasing distance from the ore body as there is a vague suggestion that the trace element contents (excluding barium) may descrease away from the body, but any such relationship is extremely tenuous, and there is no evidence of any halo or zonation round the ore body. It may be that the inclusion of a number of samples taken from section N 10,400, which has been extensively drilled, has unduly distorted the overall distribution pattern of the original metal values in the volcanic rocks, as it is along this section that zincian siderite is widely found. A small amount of the carbonate, even if now completely replaced by goethite, would elevate the metal values, and markedly so in the case of zinc. For this reason the samples taken from section N 10,400 have been marked in the table with an asterisk. Similar considerations apply to sample No. 203 from the footwall rocks, as it comes from a zone in which zincian siderite has been detected and where the nearby portions of the ore body are contaminated. If these samples are omitted from the calculations the average values for the five elements become:

Table 21A

Trace Elements in the Acid Volcanic Rocks

Footwall

Sample No.	Distance from ore body (ft.)	Cu	Zn	Pb	Bi	Ba	
46B	nil	23	179	135	n.d.	2240	
230	20	82	74	20	n.d.	524	
135	25	170	980	212	<10	1530 *	
136	30	403	292	230	<10	1760 *	
137	40	17	1420	107	n.d.	409 *	
197	45	33	48	<10	n.d.	530	
49	50	235	243	n.d.	n.d.	792	
51	55	246	55	<10	n.d.	412	
138	55	254	149	43	n.d.	464 *	
203	80	630	1140	360	21	1408	
140	130	13	173	<10	n.d.	427	
188	160	10	26	49	<10	359	
333	180	18	43	n.d.	<10	291	
142	240	21	70	83	n.d.	755 *	
160	240	13	385	41	n.d.	701 *	
161	240	24	49	11	n.d.	3500 *	
262	265	12	57	n.d.	n.d.	586	
263	265	22	68	17	n.d.	394	
162	450	20	85	n.d.	n.d.	1050	
165	450	21	30	n.d.	n.d.	2440	
	Average	113	249	67	-	1030	
278		41	29	10	n.d.	675	north of main ore body

* denotes samples from section N 10,400

Table 21B

Trace Elements in the Acid Volcanic Rocks

Hangingwall

Sample No.	Distance from ore body (ft.)	Cu	Zn	Pb	Bi	Ba	
190	nil	16	83	≮10	n.d.	965	
62B	25	17	38	≮10	14	588	
233	25	45	299	n.d.	n.d.	433	
232	90	17	18	n.d.	n.d.	506	
120	90	195	870	118	≮10	720	*
78	225	75	41	n.d.	n.d.	227	
145	235	16	202	≮10	≮10	1375	*
220	300	360	623	67	≮10	536	*
221	300	64	321	≮10	n.d.	1890	*
224	300	13	71	124	n.d.	1495	*
331	300	13	34	n.d.	n.d.	895	
	Average	76	234	30	-	877	
254		13	45	≮10	n.d.	1182	south of ore body

* denotes samples from section N 10,400

	Cu	Zn	Pb	Bi	Ba
Footwall	60	78	24	-	850
Hangingwall	28	79	-	-	687

These figures can be compared with those for the acid volcanic rocks near the Pesagi No. 6 and Sanlong B ore bodies which average:

	Bt. Pesagi	Bt. Sanlong
Cu	26	19
Zn	42	19
Pb	35	-
Bi	-	-
Ba	1363	733

The recalculated averages for the Bt. Ibam rocks show a marked drop compared with those given in Table 21, and are thought by the writer to be closer to the metal values contained in the original volcanic rocks. The higher copper in the footwall rocks is due to enhanced values in two samples, numbers 49 and 51.

It can be stated with some degree of confidence that where the levels of the impurity elements in the Bt. Ibam volcanic rocks are high, there is usually some evidence that they have largely been introduced after the rocks were originally formed. This does not apply to barium which is undoubtedly an original constituent. Although barytes has been noted from Bt. Ibam there is no evidence that it has widely affected the volcanic rocks. If the figures for rocks which have enhanced values due to metasomatism are discounted, then the following ranges cover the probable original values in the volcanic rocks:

Cu	15 - 100 ppm
Zn	30 - 100
Pb	<10 - 50
Bi	<10 - 15

These values can be compared with crustal averages (Mason, 1952) of Cu 70 ppm, Zn 132 ppm, Pb 15 ppm and Bi 0.2 ppm, and with specific averages for acid igneous rocks of Cu 16 ppm, Zn 60 ppm, Pb 19 ppm (Sandell & Goldich, 1943) and Cu 15 ppm (Lundegardh, 1946).

Skarn Rocks

The values below (in ppm) show the concentrations of the five elements in eight samples of calc silicate rocks, although No. 134 was almost monomineralic and composed largely of garnet.

Sample No.	Cu	Zn	Pb	Bi	Ba
45	112	210	13	n.d.	103
50	298	18	76	19	24
52	47	97	n.d.	n.d.	<10
53	41	79	<10	n.d.	<10
134	39	500	<10	n.d.	32
141	35	106	n.d.	n.d.	19
181	18	226	<10	n.d.	12
332	20	244	n.d.	n.d.	113
Average	76	185			

As some parts of the skarn rocks have been extensively mineralized during the main phase of ore genesis by one or more of magnetite, micaceous haematite and pyrite, the above values are considered to be of a relatively low level. A study of polished sections has shown that some of the skarn rocks carry discrete copper and zinc minerals in small quantities, and it is difficult to be sure of the metal content of the original carbonates but the above figures prove it was quite low. Samples No. 52 and No. 141 show no sign of any ore minerals and the figures given for these samples probably closely represent the original concentrations. It is not possible to be certain whether or not the relatively high zinc shown in sample No. 134, which consists largely of garnet, represents an original feature. The fact that the garnet has probably crystallized before the main period of mineralization, although as stated in chapter IV the evidence is sometimes ambiguous, suggests that the zinc may have been incorporated from the original calcareous sediment, on the other hand the sample comes from section N 10,400 where many zinc values are enhanced over a wide zone.

A sample of quartzite shows relatively high zinc, but this is definitely due to the goethite which has impregnated the rock (no. 152). The values are Cu 95 ppm, Zn 860 ppm, Pb nil, Bi nil and Ba nil.

Plutonic Rocks

The values (in ppm) for the trace elements in five samples of intrusive rocks in the Bt. Ibam area are shown below. Four of the samples came from the immediate area of the main ore body, but No. 279 came from some distance to the northeast.

Sample No.	Cu	Zn	Pb	Bi	Ba
117	11	67	37	n.d.	1055
164	17	69	n.d.	20	153
167	13	58	n.d.	n.d.	222
198	16	37	n.d.	n.d.	905
279	35	78	10	n.d.	*
Average	18	62	9	-	584

* Not determined

The levels of copper and zinc in the samples are very consistent and comparable with those in the Bt. Ibam volcanic rocks, and also with the other extrusive rocks in Ulu Rompin. The values can be compared with the averages of 10 ppm Cu and 40 ppm Zn for granites quoted by Turekian & Wedepohl (1961).

Gangue Minerals

The results obtained on two chlorite specimens are shown below:

Sample No.	Cu	Zn	Pb	Bi	Ba
61	14	2355	10	127	62
189	n.d.	255	n.d.	n.d.	n.d.

Sample No. 61 comes from a fault zone and contains more iron than No. 189, some of which is in the form of limonitic material which has contributed to the high zinc figure. Sample No. 189 is one of a very small number of specimens in which copper has not been detected; the zinc is possibly present in the crystal lattice as there is no evidence of mineralization in this sample. The role of goethite in carrying zinc is shown by the analysis of the bright blue smectite which is found along the footwall contact - chapter XIII. This sample contains sufficient goethite for it to be recognizable on a diffraction pattern and contains the following trace elements:

Cu 34 ppm ; Zn 4700 ppm ; Pb 14 ppm ; Bi nil ; Ba 51 ppm

A sample of talc which contains some pyrite showed the following:

Cu 594 ppm ; Zn 274 ppm ; Pb nil ; Bi nil ; Ba nil

Distribution of Iron and Trace Elements

Iron

There is no consistent pattern discernible in the distribution of iron values throughout the ore body, as can be observed from a study of the cross sections shown in Fig. 107A to 107E. For the purpose of constructing the sections the iron ore was divided into four categories, namely +65% Fe, 65% to 60% Fe, 60% to 50% Fe and 50% to 40% Fe. The ore with over 60% Fe must contain

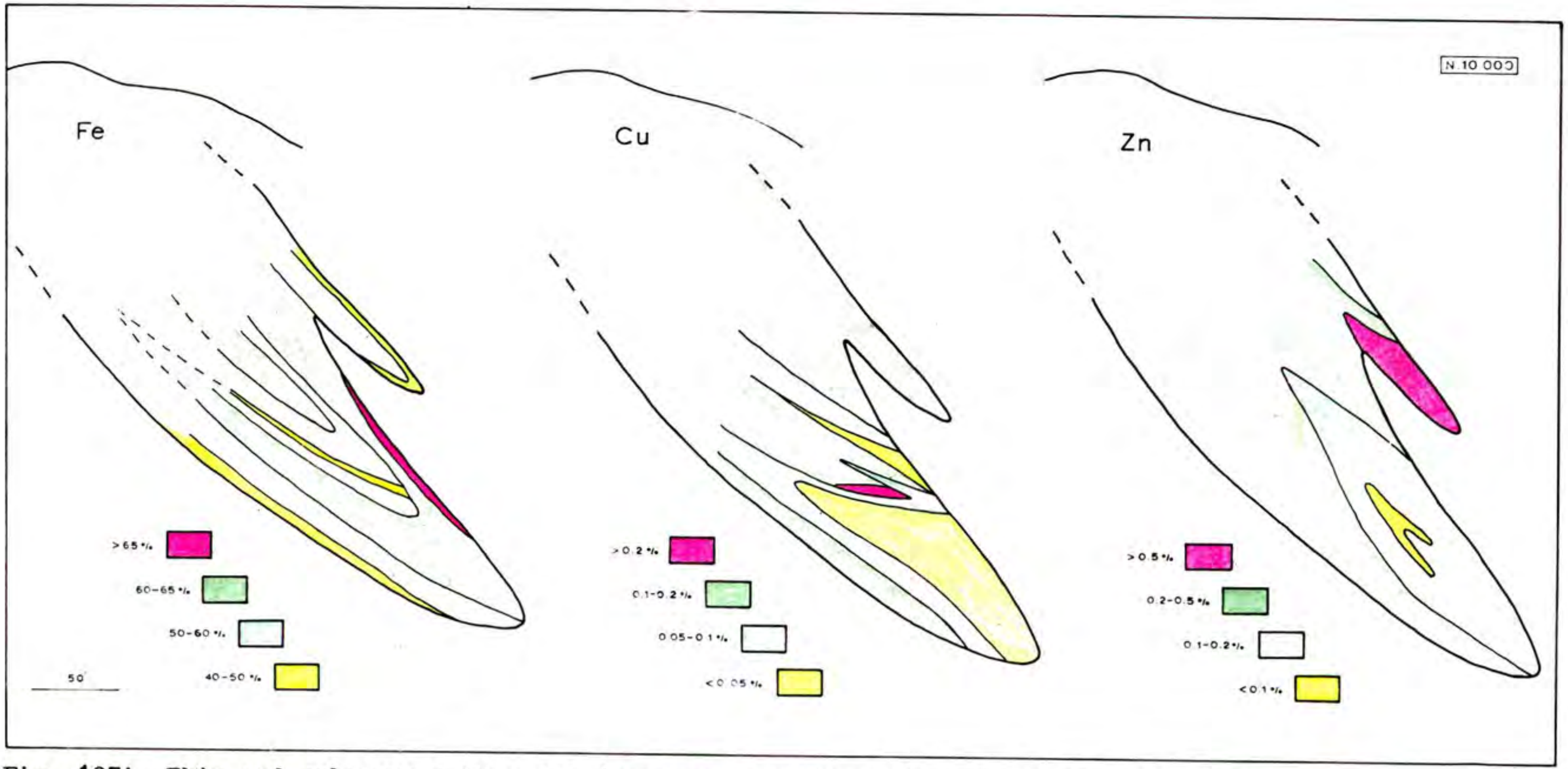


Fig. 107A. This and subsequent Figures show cross sections through the Bt. Ibam body.

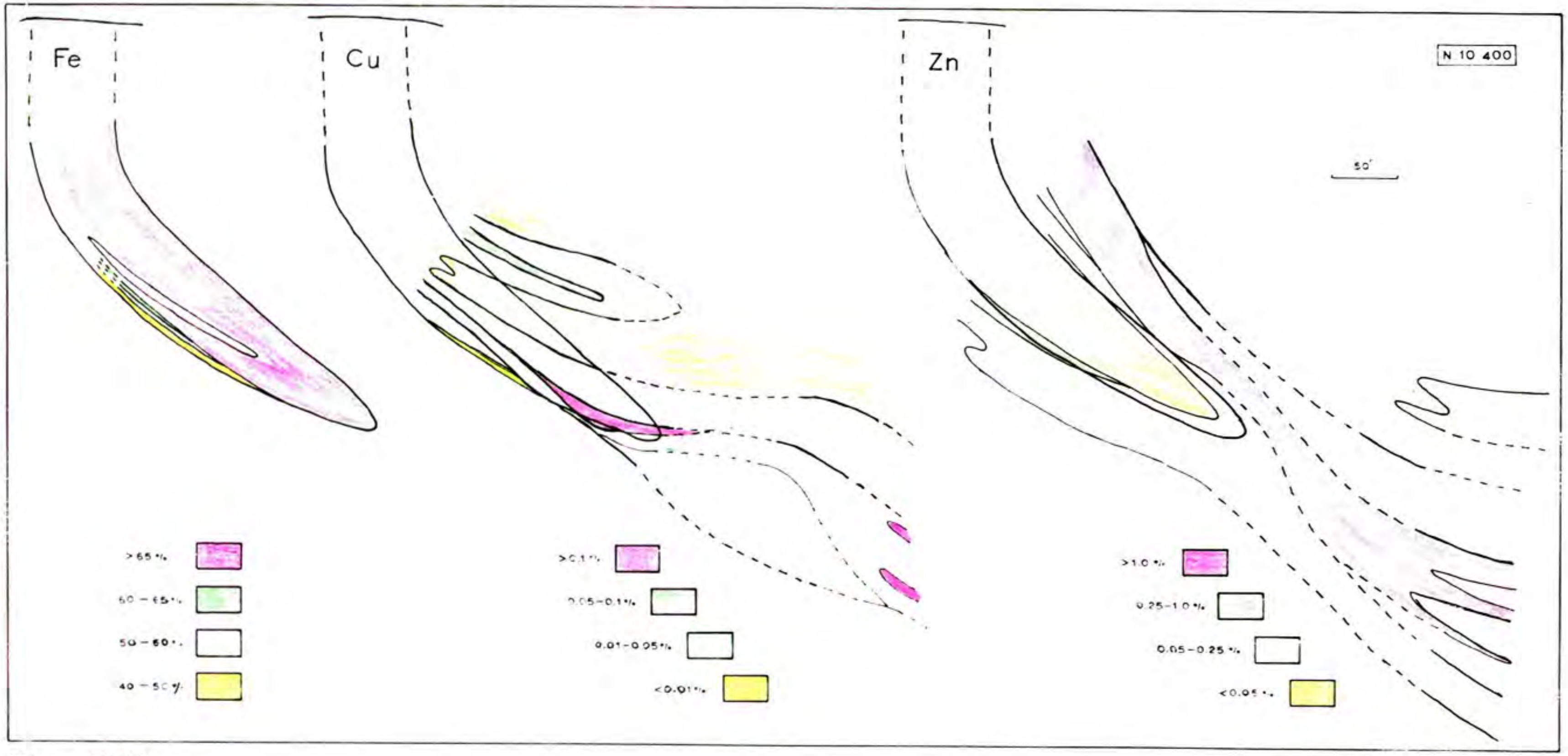

Fe
Cu
Zn
N 10 400
50'
>65 %
60 – 65 %
50 – 60 %
40 – 50 %
>0.1 %
0.05–0.1 %
0.01–0.05 %
<0.01 %
>1.0 %
0.25–1.0 %
0.05–0.25 %
<0.05 %

Fig. 107B.

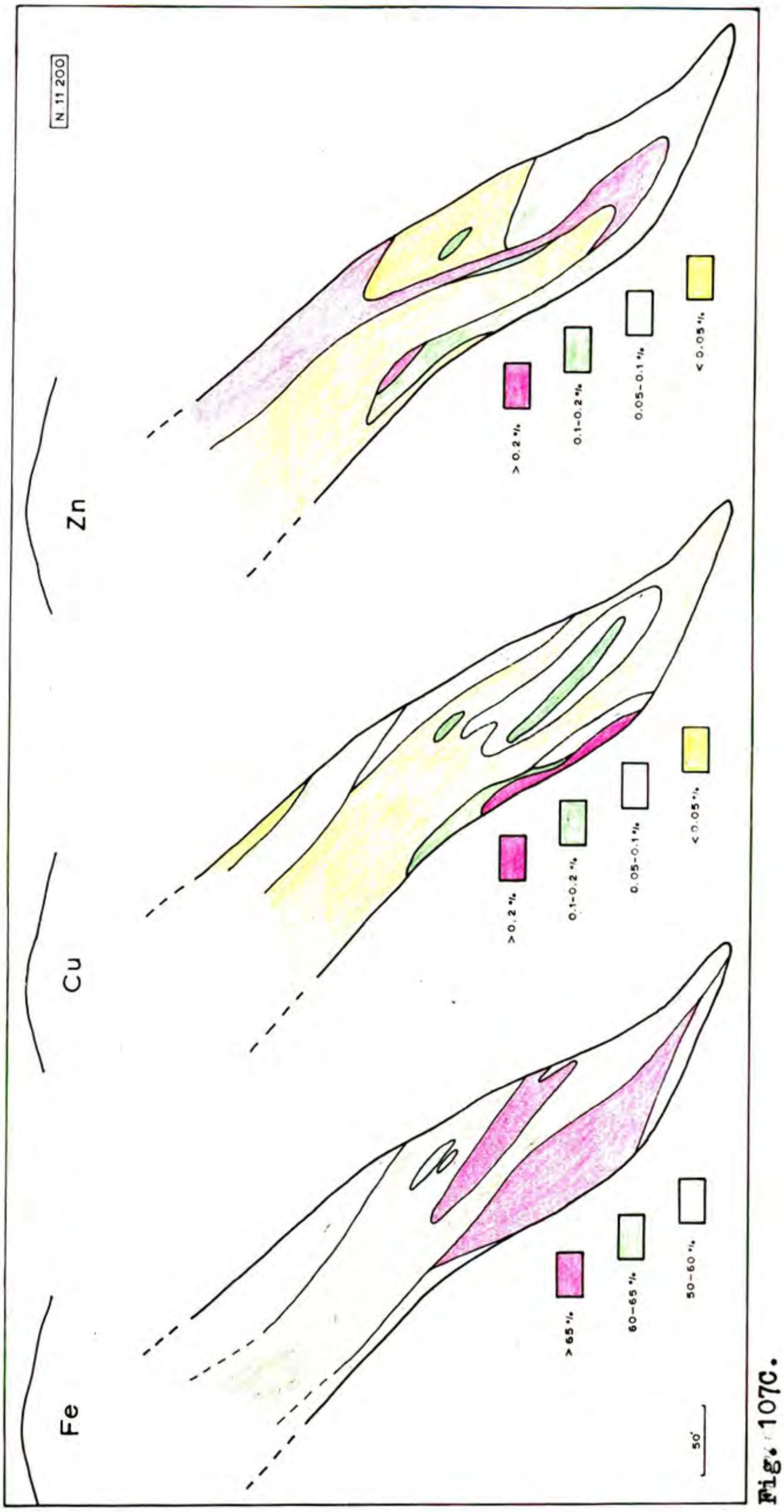

N 11 200
Zn
> 0.2 %
0.1–0.2 %
0.05–0.1 %
< 0.05 %
Cu
> 0.2 %
0.1–0.2 %
0.05–0.1 %
< 0.05 %
Fe
> 65 %
60–65 %
50–60 %
50'

Fig. 107C.

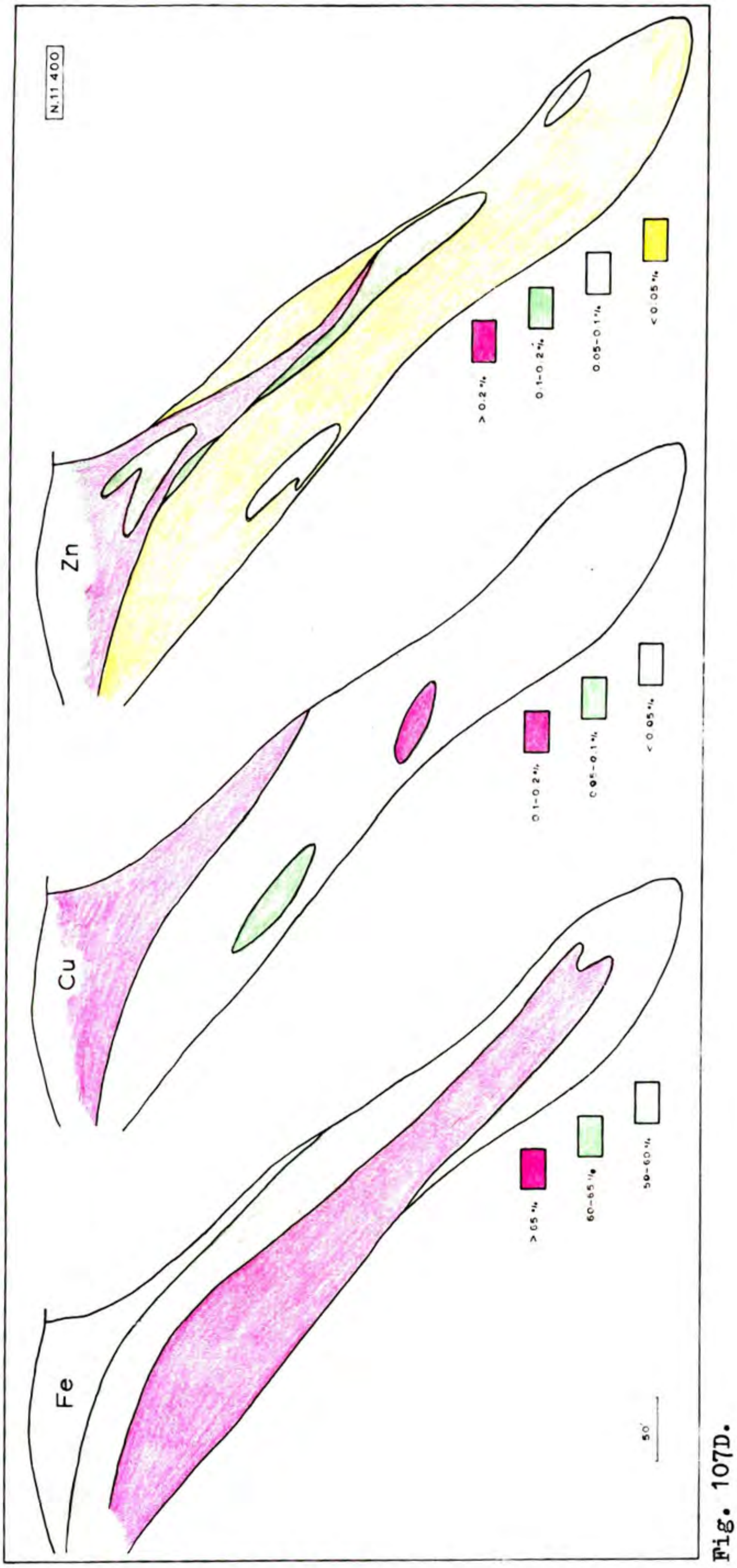

N.11 400
Zn
> 0.2 %
0.1-0.2 %
0.05-0.1 %
< 0.05 %
Cu
0.1-0.2 %
0.05-0.1 %
< 0.05 %
Fe
> 65 %
60-65 %
50-60 %
50'

Fig. 107D.

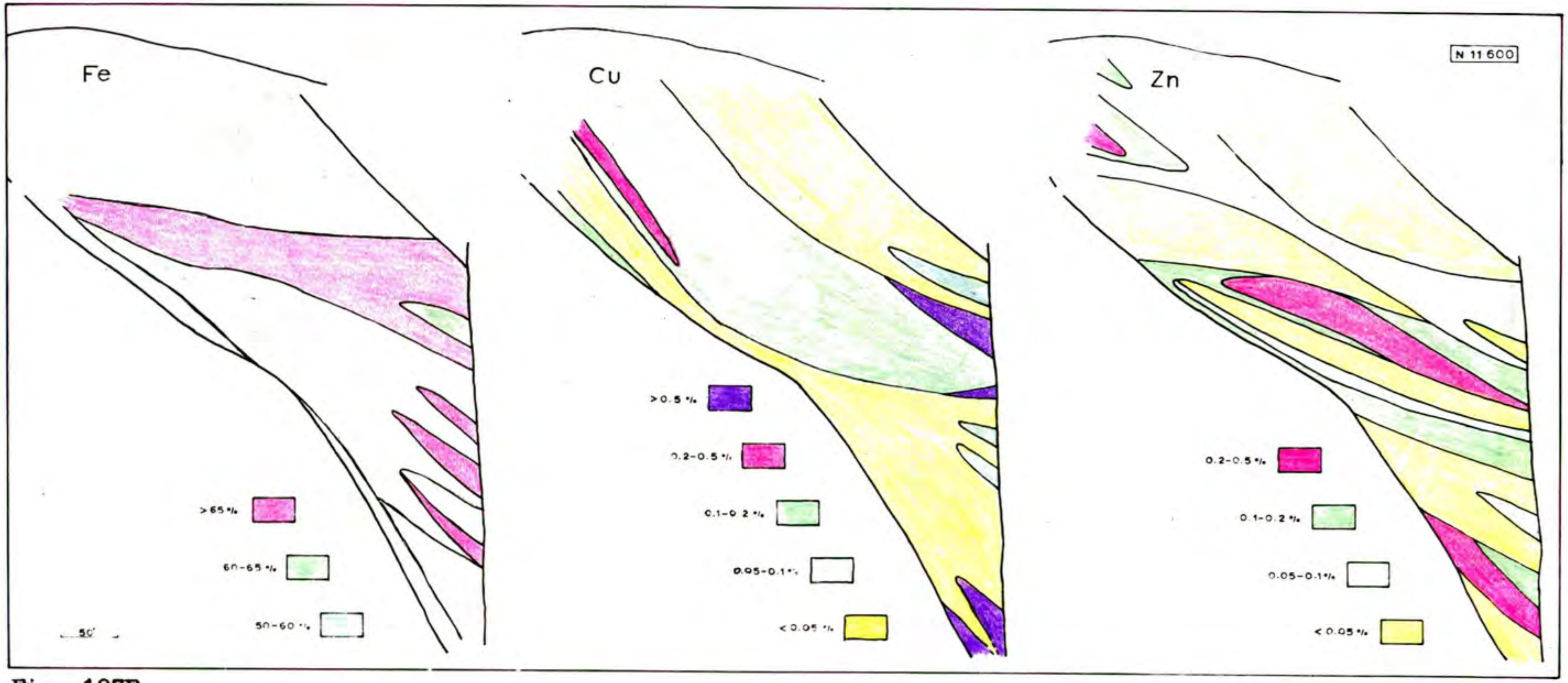

N 11 600
Fe
>65 %
60-65 %
50-60 %
50
Cu
>0.5 %
0.2-0.5 %
0.1-0.2 %
0.05-0.1 %
< 0.05 %
Zn
0.2-0.5 %
0.1-0.2 %
0.05-0.1 %
< 0.05 %

Fig. 107E.

a significant content of magnetite and/or martite, whereas the lower grade ore can consist largely of goethite or of magnetite disseminated in gangue. The cross sections cover the ore body from N 10,000 in the south to about N 12,100 in the north; this zone contains the most solid and important part of the body. To the south iron ore is not abundant and is all goethitic, while to the north the ore body becomes increasingly split by lenses and patches of gangue and country rock, and few large masses of good quality ore remain.

The really high grade +65% ore first appears as a thin band following the lower part of the hangingwall contact at the southern end of the ore body along N 10,000, but to the north it consistently constitutes an important, although variable, part of the ore body. For example, along sections N 10,400 and N 11,800 it forms the bulk of the ore whereas along N 11,600 it makes up a relatively small part. Where the really high grade ore occurs in bands and linear masses these normally follow the general dip of the ore body as a whole, although along N 11,600 there is a broad band with a roughly horizontal attitude. The +65% ore can occur from the highest to the lowest levels of the ore body along an individual section, e.g. N 10,400 and N 12,100, but rarely does so.

The 65% to 60% ore finds its maximum development within the main part of the ore body between approximately N 11,000 and N 11,600, forming a substantial part of the body along the latter section. The 50% and 40% ore is typically found along the hangingwall and footwall contacts, although it does not usually occur throughout the full vertical extent of the ore body. A good example of a near-surface capping of medium grade ore which continues as a band along the hangingwall contact is seen along N 11,400. On N 11,200 bands of low grade ore are found along both contacts, while similar material forms a virtually continuous band on the footwall of N 11,600 and the hangingwall of N 12,000. An unusual situation is found along N 12,100 where a band of low to medium grade ore cuts almost vertically through a mass of +65% ore, but the control here has been a fault, and the zone contains high copper values, as will be described later. Some of the cross sections show relatively complicated patterns of ore distribution, but others, particularly N 11,400, are simple.

If the distribution of the iron values in narrow vertical sections is studied, no overall pattern can be detected. Fig. 108 shows the iron values in five representative vertical drill holes which have long intersections in the ore body. In two holes (3079 and 3152) the iron values remain relatively constant throughout the full depth of the ore intersection, whereas in two others (3135 and 3142) the values are in the 55% to 60% range in the upper levels and change to +60% ore at about 100 ft. This is a fairly common pattern due to goethite near surface giving way to high grade ore lower down. From 100 ft to 350 ft drill hole 3147 shows good ore with bands of poor quality material. Where the drill holes encounter the footwall contact one of three condtions is usually encountered. The ore body can end abruptly against barren country rock (3079 and 3135), there can be a zone of goethite which gives rise to fluctuating values in the 30% to 50% range (3142 and 3152), or there can be irregularly disseminated magnetite which produces thin bands of high grade ore alternating with barren zones (3147).

At the time of the writer's fieldwork the company were exploring deeper levels to the east of some parts of the ore body, and although the individual drill logs were not examined Fig. 109 shows a cross section for N 11,000 compiled by the mine staff incorporating their most recent data. No figures for the different categories of ore are given, but from a comparison with other data it can be assumed that high grade means over 60% Fe, medium grade 50% to 60% Fe and low grade below 50% Fe. The division into massive and friable ore is useful for the planning of mining operations, but it is also a relic from the early days of prospecting when the friable ore was not a saleable product.

Copper and Zinc

These two elements are discussed together, even though they are not necessarily genetically related, as they are by far the commonest metallic impurities, and jointly or severally are

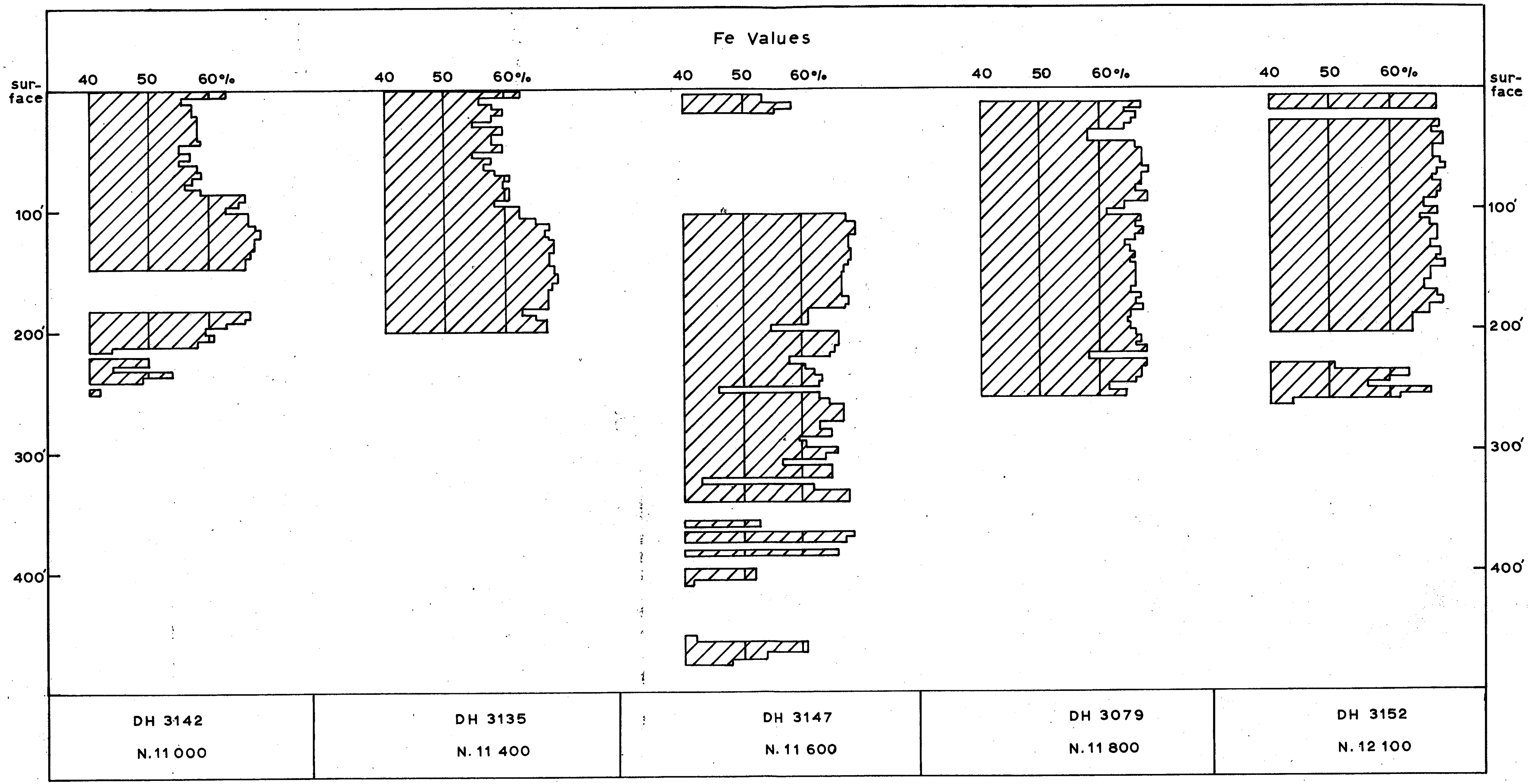


Fig. 108 Iron values in five diamond drill holes.

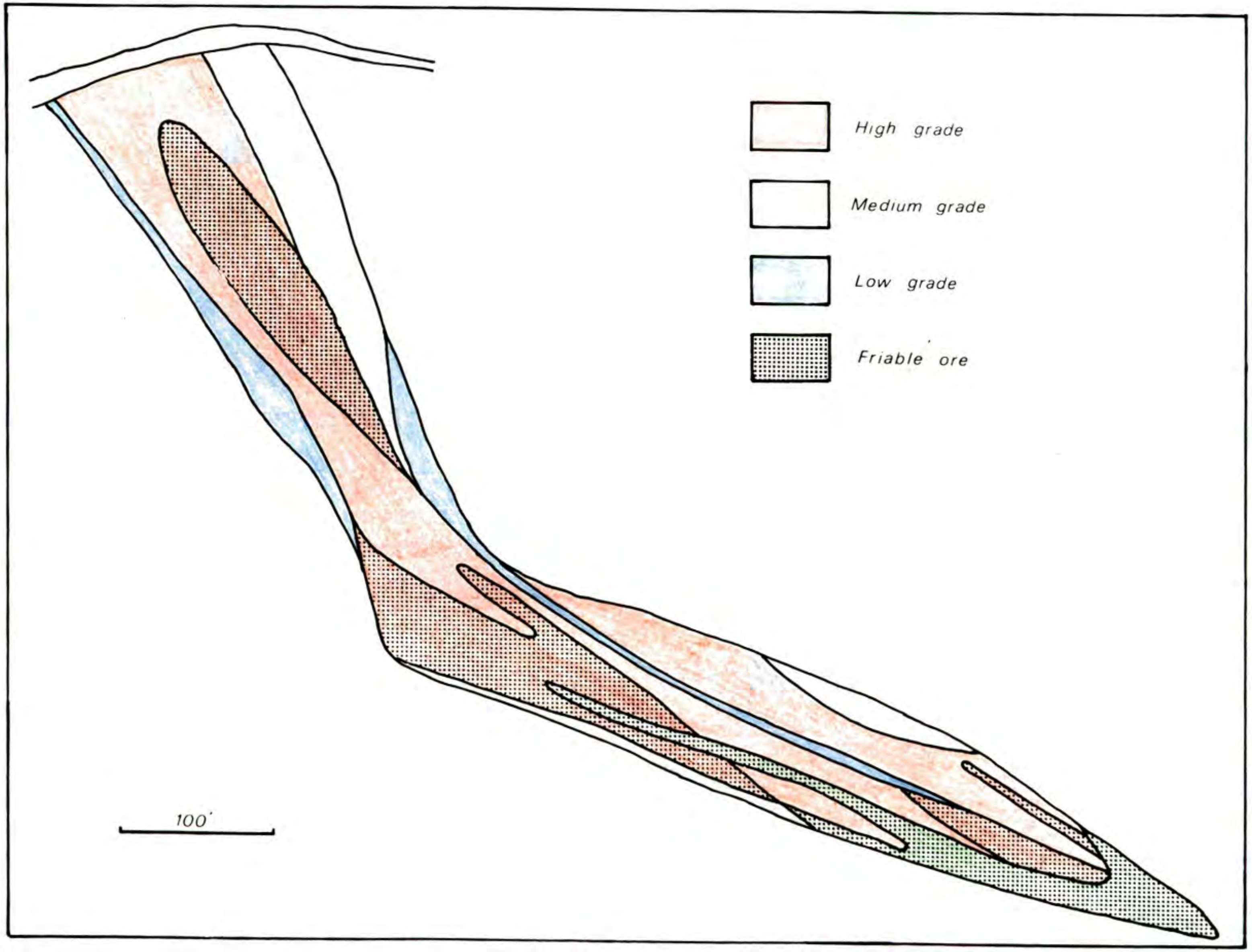


Fig. 109 Cross section through N.11 000 showing the distribution of iron ore.

found throughout much of the ore body. Apart from iron itself, these are the essential elements for which all samples are checked as part of the quality control. Within the ore body it is often possible to detect the presence of copper by eye due to its occurrence in several distinctive minerals, but zinc can only be detected by a chemical analysis. Zinc is almost exclusively found in goethite, and while some copper usually accompanies it, the latter element can occur in a variety of sulphide and oxide minerals, and native copper.

Although some zinc and copper is found throughout the ore body, the deposit can be validly divided into two sections at approximately N 11,500: to the south zinc is important and copper much less so, while to the north the converse is true. This is a direct consequence of the location of the primary sources of zinc and copper, and in spite of considerable modification by secondary processes this fundamental division is still discernible. Zinc has been introduced originally in sphalerite and some related zincian siderite, while copper has been deposited in chalcopyrite. Due to the mobility of the colloidal solutions which deposited goethite and also carried zinc, this element can be found far from the nearest observed primary source (the same applies to the copper which is found in goethite), but the secondary copper minerals are usually found in close proximity to chalcopyrite. Within the limits of the ore body itself it is uncommon to find zinc values exceeding 0.5%, but very high copper values can be found in some localities; this is a result of the difference in mode of occurrence of the two elements. In the surrounding country rocks, which as has already been demonstrated are typically low in impurity elements, the position is reversed, and some extremely high zinc values can be found towards the southern end of the mine whereas copper is usually low.

Figs. 107A to 107E show some cross sections which illustrate the distribution patterns for copper and zinc in selected parts of the ore body. The cross section for N 10,400 in addition shows values in the country rock as more detail is available, and the writer believes that this part of mineralized area has been of prime importance in the development of the zinc pattern.

At the extreme southern end of the ore body the ore is largely goethite which contains both zinc and copper. The former is always more abundant, and along N 10,000 much of the ore contains 0.2% to 0.5% Zn, with a lens of 0.1% to 0.2% Zn ore towards the deepest part. Copper is mostly present in the 0.1% to 0.2% range, but the deepest parts of the ore body are often below 0.1%, and even 0.05% Cu. There is some coincidence between the copper and zinc patterns. In the immediately adjacent country rocks nodular goethite is common, and this contains very much higher zinc than copper; the latter is usually below 0.05% whereas zinc is often over 0.5%, and values up to 3.4% Zn have been found. The very high values are due to the goethite having been developed, largely in situ, from zincian siderite. It is usually possible to detect the goethite which has developed in situ as it tends to be manganiferous, again a feature inherited from the siderite.

Section N 10,400 is most interesting because of the values found in the country rocks rather than those in the ore body itself. The iron ore is nearly all of high grade, and contains less than 0.05% Zn and 0.05% Cu. Zinc values rise to 0.3% along the footwall and in the extreme lower portions of the ore body, and there are a few higher copper values along the deepest parts of the footwall. It is somewhat surprising that zinc is so low in the ore body as there is a zone immediately above and below it which contains values between 0.05% and 0.25% Zn, and just above the hangingwall there is a band in which values exceed 1% Zn. All this high zinc is found in nodular goethite, often manganiferous, which contains relics of zincian siderite. Zinc contents over 10% have been recorded. Fig. 107B shows that this band of high zinc follows the dip of the ore body, and continues well beyond the deepest levels of primary iron oxide mineralization. It has to be emphasized that it is unlikely that every sample in the band contains more than 1% zinc. The explanation is that diamond drilling brought up numerous fragments of goethite although overall core recovery was poor due to the decomposed nature of the country rock, and it is unlikely that the actual volcanic rocks contain exceptionally high zinc. However, the goethite fragments are present in sufficient quantities, and so evenly dispersed throughout the band to justify the belief that this zone has been the most important location of zinc mineralization near the ore body. Fig. 107B shows that below the zincian zone surrounding the ore body the volcanic rocks contain very low zinc values, although megascopic sphalerite veins are found in one locality deep below the iron ore. Sphalerite has been encountered in small amounts as relics in the highly zincian bands - page 196, and the writer believes that the zinc cycle has been:

SPHALERITE
↓
replaced by siderite
↓
ZINCIAN SIDERITE
↓
alteration
↓
ZINCIAN GOETHITE

In addition to the ore which formed in situ, a lot of iron has been transported in colloidal solutions and deposited elsewhere as colloform goethite with a significant zinc content.

The position with regard to copper is entirely different, and as Fig. 107B shows, much of the country rock and nodular goethite contains below 0.01% Cu with further large areas between 0.01% and 0.05% Cu. Although the levels of zinc and copper are so disparate, it can be seen that the zones which contain what can be loosely described as "above background" values are largely coincident.

Along N 10,600 and N 10,800 zinc is the only important impurity, and it affects a considerable part of the ore body. On N 10,800 the zinc shows a marked predeliction for the hangingwall side, and there is a zone which extends at least 350 ft from surface and is up to 100 ft wide which contains over 0.15% Zn. The iron ore is of medium grade and consists largely of goethite. On N 10,600 there are isolated values over 1% Zn in disseminated ore immediately below the ore body proper. N 11,000 continues the pattern seen in N 10,800 with a band of high zinc (+0.15%) along the upper parts of the hangingwall, and small patches of high copper. Recent drilling has proved extensions of the ore body at depth to the east, and parts of this contain up to 0.5% Zn, probably due to goethite derived from zincian siderite.

On N 11,200 the bulk of the iron ore is +60% Fe, but with a thick band of lower grade ore along the upper parts of the hangingwall, and a thinner band along the footwall. This pattern is partially reflected in the distribution of both the copper and zinc values. High values for both copper and zinc are associated with the hangingwall (0.1% Cu and 0.2% Zn) and the footwall (0.1% Cu and Zn) where the iron ore is of medium grade, whereas much of the +60% Fe ore is low (-0.05%) in both elements. One locality on the footwall contact contains over 0.5% Cu. The very lowest parts of the ore body have an elevated zinc level (+0.1%) even though the iron ore is high grade, and copper over 0.05% occupies an elongated patch in the same region.

N 11,400 has one of the simplest distribution patterns for iron ore, and this is reflected in the copper and zinc which show broadly similar outlines. The upper part of the ore body and a band along the higher levels of the hangingwall contact contain +0.2% Zn and +0.1% Cu, and the zinc continues downwards at a slightly lower concentration. The lower boundary of the contaminated ore agrees closely with the junction between 60% and 65% Fe ore. By far the larger part of the ore body is below 0.05% for both copper and zinc.

N 11,600 shows a fairly simple pattern for copper with much of the ore body being below 0.05% Cu, but there is a broad band through the centre with +0.1% Cu which in parts rises to +0.5% Cu. The band of higher values is in no way related to the disposition of the different grades of iron ore. At the very base of the ore body there are small patches of +0.5% Cu, and also in one spot below the footwall where drilling encountered small veins of solid chalcopyrite. The eastern edge of the ore body comes up against a fault, and it is perhaps significant that the highest copper values are found adjacent to the fault. Zinc on N 11,600 shows a more complicated pattern, although much of the ore body contains less than 0.05% Zn with another area containing from 0.05% to 0.1% Zn. A lens containing 0.2% to 0.5% Zn dips gently east of the central part of the ore body, and it lies within a broader band of 0.1% to 0.2% ore. There are small patches of 0.2% Zn ore near the top and bottom of the ore body. The only patch of really high zinc (+0.5%) falls below the footwall and very close to the veins of chalcopyrite.

N 11,800 consists of iron ore which has been little affected by copper or zinc, and the whole ore body contains less than 0.05% Zn. Copper is found in four small patches, two of which contain 0.05% to 0.1% Cu and two 0.1% to 0.2%. A similar situation prevails along N 11,900 where, apart from one very small patch of +0.5% Zn ore, sinc is consistently low. Copper too is only sporadically distributed, but one small patch of +0.2% Cu ore is located along a fault line.

N 12,000 marks the southern limit of a zone which extends northwards for about 300 ft and in which important copper patches are found. On N 12,000 copper over 0.1% is located in lenses roughly parallel with the dip of the ore body and towards its base. The copper extends intermittantly from the hangingwall to the footwall, both contacts being, at least partially, related to faults. Some very high values (+1% Cu) are sporadically distributed in the copper zones. Zinc is typically very low, but a narrow zone along the footwall at the bottom of the ore body contains up to 0.25% Zn. On N 12,100 zinc values are again low apart from one or two isolated samples. Copper values are high (+0.15%) in four lenses, and one lens which is associated with a fault contains patches well above this level. This copper is found in medium grade iron ore, but in the other three zones all grades of ore are affected.

On N 12,200 and N 12,300 the copper zones are still present although diminished in size, but by N 12,400 there is little copper of any significance. Zinc is low throughout. To the north the ore body becomes progressively more divided by lenses and patches of gangue and country rock, and some of the chloritic rocks can contain high copper values due to the presence of chalcopyrite.

Almost all the high zinc values associated with iron ore are contained in goethite and secondary manganiferous material, but copper is present in a variety of minerals. Where high zinc and copper are found together in a sample it usually indicates that goethite forms an important part of the ore. There has been an important redistribution of zinc, but copper has, on the whole, been less mobile. As stated earlier, primary sources of zinc are only important towards the southern end of the ore body, particularly near N 10,400, and even here the zinc has not entered the ore body in appreciable quantities until the colloidal solutions which deposited goethite started to circulate. Copper behaved differently, with most of the chalcopyrite being deposited in the ore body, and the secondary redistribution also taking place within the iron ore. Evidence that zinc has originated separately from copper is found in analyses of the richly zincian samples from the southern end of the mine, which typically contain little copper.

Sample No.	Depth ft.	Zn	Cu
122	130	+10 %	36 ppm
125	155	4.3	48
132	263	2.5	57
155	350	2.2	116
258	186	3.4	64

These figures show that while this type of material could contribute abundant zinc to circulating solutions its affect on copper levels would be insignificant.

The converse of the relationship between the levels of copper and zinc demonstrated above is often shown in samples with high copper contents, as the figures shown below demonstrate.

Sample No.	Cu	Zn
9F	1.5 %	2320 ppm
12A	+5.0	22
32	3.6	230
100	2.1	59

These samples happen to illustrate some of the different associations in which copper is found. No. 9F contains extensively altered magnetite grains among which are set pyrite crystals partially or wholly replaced by chalcocite-neodigenite, with cuprite and native copper in cavities in the goethite. The abundance of goethite is reflected in the relatively high zinc; a much fresher sample from the same zone contained only 247 ppm Zn. No. 12A consists of granular magnetite, with veinlets of cuprite which has been partially replaced by malachite, plus some chalcocite and covellite. This sample contains the highest copper detected by the writer in a sample which completely lacks chalcopyrite. The rapidity with which copper can change is shown by the fact that sample 12B, which originates from within 5 ft of No. 12A, contains only 910 ppm Cu. No. 32 consists of magnetite disseminated in gangue with chalcopyrite, while No. 100 contains chalcopyrite veining and replacing shattered pyrite in magnetite ore. The chalcopyrite in the latter sample has been partially replaced by secondary sulphides.

There is no doubt that the redistribution of copper is still actively in progress otherwise chalcanthite would not be found on bench faces shortly after heavy rain. Also, in some of the cupriferous zones it is possible to follow the development of malachite on freshly cut ore surfaces over a matter of days. During the early days of prospecting it was even suggested that the highly cupriferous iron ore should be stockpiled in the open to allow the copper to be leached by natural processes.

Probably the most striking of all the copper occurrences are those which contain the native metal, although they are not numerous. The copper forms spongy, irregular wirelike masses, and has almost always developed along fractures and joints in the ore body where there has obviously been very easy circulation for the cupriferous solutions. The writer found sizeable masses of copper with cuprite on the footwall along N 12,000, while the mine authorities located native copper in Adit 11 on sections N 12,000 and N 12,100. Native copper has been identified by the writer in polished sections of samples originating from N 11,500 to N 12,000, and even on a microscopic scale the copper is usually located along cracks and in other open spaces. Copper is a late stage product and is still probably in the process of formation: one striking development along the footwall is intimately associated with opaline quartz, which can be proved to be of recent origin.

There is no obvious overall control which has operated on the location of chalcopyrite, but some is found in or in close proximity to zones of movement, and it can be proved that at least part of the chalcopyrite has been deposited after the iron ore has been faulted - see e.g. Fig. 39 . The redistribution of copper has been affected by many factors but the most important have been:

a. original location of the chalcopyrite.
b. presence of passageways for the free circulation of solutions.
c. level of the water table.
d. in certain cases the presence of precipitating agents e.g. carbonates.
e. the presence of pyrite which on alteration can produce active acid solutions, and also act as a host for secondary reactions.

Some maps prepared by the company staff differentiate zones of oxide and secondary sulphide copper, but the writer feels that the oxide copper is so restricted in extent and controlled by local conditions that there is no justification in referring to oxide zones. Polished sections show that oxide and secondary sulphide copper minerals sometimes occur intimately mixed together, and it can be proved that some of the cuprite is older than the sulphide - Fig. 77. The oxide minerals and native copper are regarded by the writer as local modifications of what is largely sulphide copper. Fig. 110 shows a reproduction of a plan prepared by mine geologists which shows copper, zinc, and also bismuth zones in the ore.body. The plan cannot be directly correlated with the writer's cross sections at Fig. 107 as different metal levels have been used, and also the company's plan was prepared before some important drill holes were completed. Nevertheless, the plan

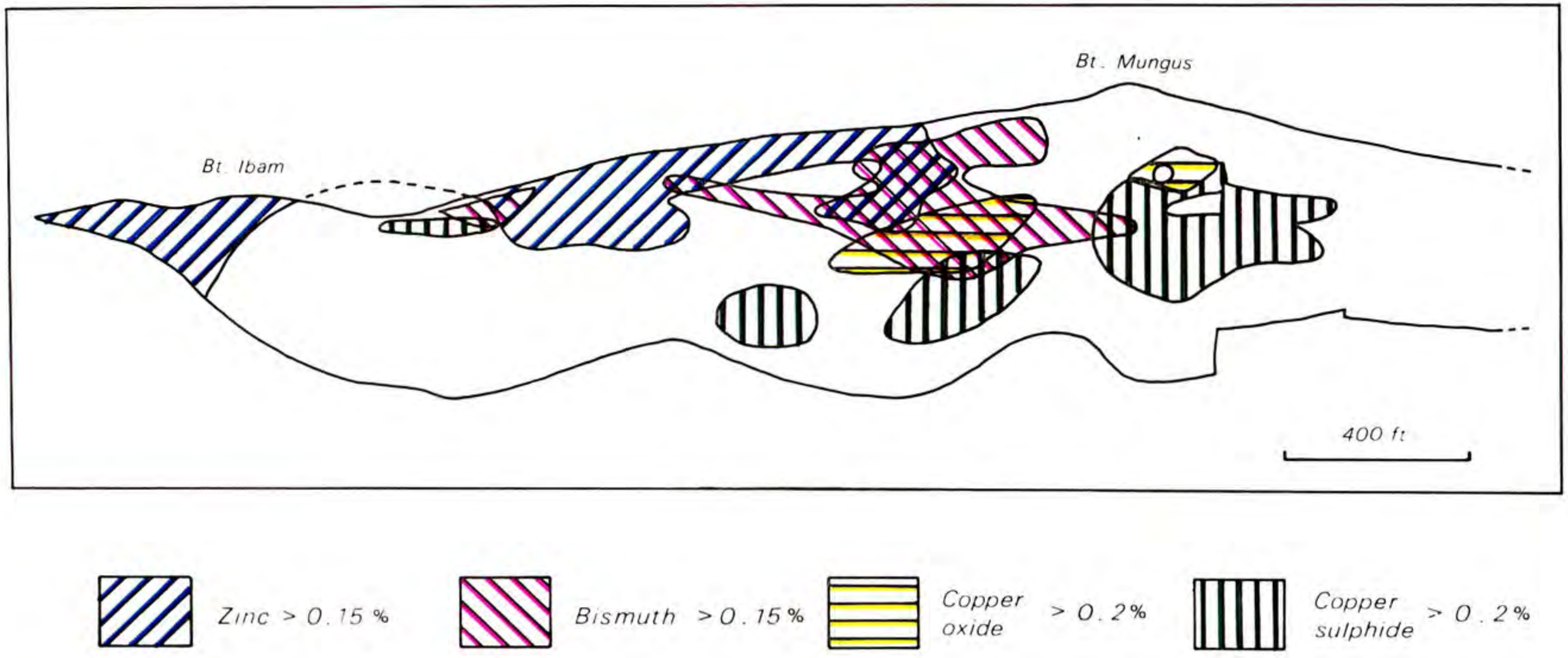


Fig.110 Plan showing contaminated zones at Bt. Ibam. Taken from a ROMPINCO plan.

and the writer's sections are in good agreement as to the location of the main contaminated zones.

The development of the zinc and copper distribution patterns is also considered in chapter XIX.

Bismuth

This element has nothing like the importance of copper and zinc and causes few problems to the mine operators. It has already been demonstrated that there is virtually no bismuth in substitution within the magnetite lattice so it has obviously been introduced separately, which agrees with the evidence from polished sections. Goethite is typically low in bismuth, even when the ore contains high levels of copper and zinc. The country rocks are invariably very low in bismuth.

The values obtained by the writer show a wide range, but most are relatively low and in a number of samples bismuth is not detectable. In those samples with very high bismuth there is always evidence of bismuth minerals to be observed in polished sections, and there is no doubt that apart from an extremely small amount in tetradymite the bismuth has been introduced in bismuthinite. Much of the original bismuthinite remains but some has been altered by secondary reactions.

The bismuth contents in the shipped ore are shown in Fig. 102, and it can be seen that the level of the impurity in both the lump and fine ore dropped markedly in 1965. This is in accordance with the bismuth distribution shown in Fig. 110, as by 1965 much of the high bismuth zone had been mined out. The figures for the exported ore also show that the bismuth in the fine ore is consistently higher than that in the lump ore, and it is lowest of all in the highly zincian Non Standard Lump. A company report dated 1965 states that the average for the ore body is 0.05% Bi and for the protore 0.03% Bi.

Of the samples analysed by the writer six contained more than 1000 ppm Bi, and one had the exceptional content of 1.6%. This sample contained abundant bismuthinite and its alteration product bismutoferrite. Bismuthinite is found along fractures and as replacement bodies in pyrite, and along intergrain boundaries and in cavities in magnetite. All the high values obtained by the writer fall within the bismuth zone shown on Fig. 110. This figure presents a slightly misleading picture of the extent of bismuth contamination because of its method of construction. It is a projection of the bismuth zones onto a vertical plane, but most of the bismuth is in fact contained in narrow lenses which have a much greater vertical than lateral extent. This means that not all of the ore which falls within the area shown as +0.15% Bi in Fig. 110 is so highly contaminated. The figure does however accurately reflect that bismuth is restricted in extent and that nearly all the high values fall within the four hundred feet from N 11,400 to N 11,800, and up to three hundred feet from the surface.

There has been only a limited dispersion of bismuth by secondary processes. A company report states that bismuth tends to accompany the secondary copper sulphides, but this is only a partial truth as many highly cupriferous samples are low in bismuth. In an earlier section it has been shown that bismuthinite can be replaced by covellite and chalcocite with the formation of indeterminate intermediate compounds, and in these examples it is true that the sulphides do contain bismuth. Without exception the samples which have high bismuth values contain bismuthinite, and the writer is convinced that the distribution shown in Fig. 110 is essentially the primary distribution pattern.

Lead

Lead is found in only a few patches in sufficient quantities to constitute a deleterious impurity. Analyses presented earlier in this chapter show that lead is not present in magnetite but can reach significant levels in goethite ore, and within the ore body it is only goethite which is responsible for high lead values.

The sole lead mineral identified by the writer is galena, and that has only been seen in a few samples, where it occurs as veinlets and replacement blebs in pyrite. No evidence of secondary reactions involving galena has been seen, but it is obvious that most of the lead eventually finds its way into goethite from the original galena. The fact that lead is concentrated in secondary products is confirmed by the fact that nodular, secondary manganese oxides in a quartz vein in the hangingwall country rock contain up to 2300 ppm lead. Values up to 2350 ppm Pb have been recorded by the writer, and company drill logs report up to 0.95% Pb, although this is exceptional. A company report gives the average for the ore body as 0.02% Pb.

As most of the lead is found in goethite it is present in the upper levels of the ore body and towards its southern end. The values are sporadic, and there are no real zones of contamination. The highest levels of lead have been found along N 10,000, but it extends northwards to about N 11,500, albeit intermittently.

Arsenic

Arsenic has not been any problem to the mine owners as it is well below the limits set in the specification (0.1%), and no special controls are required. Although all shipments are analysed for arsenic - Fig. 102 - by no means all the drill core samples have been checked. The writer examined company logs which show that arsenic varies from a trace to 0.18%, but most samples fall in the range trace to 0.05% As. A company report also states that there is a divide in the arsenic values at about N 11,600; north of this section arsenic averages 0.02%, but to the south higher values can be found, although the average is not given. The arsenic values for the shipped ore fit in with an approximate range of trace to 0.05% As. The values also show that the lump ore is consistently higher in arsenic than the fines, which suggests that it is probably associated with goethitic ore rather than with magnetite. The only identification of an arsenic-bearing mineral is the writer's discovery of arsenopyrite in one polished section - Fig. 91, and there will be some arsenic in enargite and tennantite which have been seen in minute quantities. A company geologist has suggested that high arsenic is associated with high zinc, but this is not so; the analysis for the Non Standard Lump ore, which is very highly zincian, shows that it contains less arsenic than the normal lump ore. X-ray fluorescence 2θ scans suggest that a little arsenic is found in crystals of pyrite. It is probable that most of the arsenic is found in goethitic ore, although discrete minerals such as scorodite may be present occasionally.

Miscellaneous Metallic Elements

A number of elements which cause no practical problems as impurities can conveniently considered together.

Tungsten was checked during the early prospecting by Malayan Laboratories, a subsidiary of the main mining company, who reported very low concentrations. With a limit of detection of 0.02% only a handful of specimens reached this figure, and the highest reported values were 0.12% W in sludge and 0.10% W in core.

Cobalt was checked by Malayan Laboratories who state that with a detection limit of 0.001% the maximum value obtained was 0.01% and the average for all samples 0.003% Co. Nickel gave precisely the same results as cobalt. The writer analysed for nickel and found very low concentrations in both the ore and country rocks. Most of the results were in the range nil to 20 ppm, and only three samples contained over 100 ppm Ni. One was a pyrite crystal with a black external surface, one was disseminated magnetite in calc silicate, and one manganiferous goethite with relics of zincian siderite.

The writer checked molybdenum, but it was not detected in the majority of the samples. A specimen of mixed manganese oxides contained 126 ppm Mo, and this was by far the highest value recorded. A sample of goethite derived from zincian siderite contained 84 ppm Mo. Cadmium was detected by the writer in only seven samples, and these were rich in zinc and/or manganese. The maximum value was 405 ppm Cd. Tellurium was found in four samples, but at low concentrations and with a maximum of 85 ppm Te. It is probably present in tetradymite.

Chromium was checked in a number of ore samples and nearly all were below 10 ppm Cr. Two crystals of magnetite contained 20 and 17 ppm Cr respectively, and the sample with very high bismuth showed 40 ppm Cr, the highest value recorded by the writer. Manganese is present in the iron ore, and is found in ionic substitution in magnetite in amounts which vary from 0.1% to 0.35% Mn. In those goethites which have not been derived directly from zincian siderite manganese ranges from 100 to 1000 ppm.

Titania is very low and the writer recorded values in iron ore which were below 50 ppm TiO_2. Few determinations are available in company records but one drill log clearly demonstrated an antipathetic relation between iron and titania, showing that the iron ore is low in titania. For +60% Fe ore in one drill core the company records titania values of 0.02% and 0.03% TiO_2.

Fitch (1941) records that samples taken from Japanese prospecting pits were bulked and analysed for tin by the Malayan Geological Survey, with negative results in all samples.

Non Metallic Impurities

There is little to be said about the non metallic impurities as the levels are usually related to the amount of gangue in the ore. Alumina and silica are present in a number of silicate minerals, but predominantly in kaolinitic clay. Sulphur is derived from pyrite with minor contributions from other sulphides and sulphates. Phosphorus is the one element which has not been related to any specific mineral by the writer, and none of the mine records refer to a phosphorus mineral. The writer analysed some of the iron ore for phosphorus and found very low levels, nil to 30 ppm, in magnetite. Goethite contains more phosphorus and an examination of the levels in the shipped ore confirms this - Fig. 102. The lump ore is always higher than the fine ore, and in the Non Standard Lump, which is largely goethite, the phosphorus level rises significantly. From 1962 to 1965 the phosphorus content of the lump ore and fine ore ranged from 0.031% to 0.039% and 0.020% to 0.031% respectively, while in 1965 Non Standard Lump contained 0.071% P. A company report states that a change in the level of phosphorus can be detected at about N 11,600; to the north the level is 0.02%, but to the south there are higher values and phosphorus rises to 0.05% to 0.06% P. The report further states that the higher phosphorus is usually in goethite ore with less than 60% Fe, and that there is a correlation with highly zincian ore. A study of a large number of drill logs does not substantiate this suggestion, and high levels of zinc combined with phosphorus appear to be fortuitous.

* * * * * * * * * * * * * * * *

Fig. 111 shows in diagrammatic form the levels of the main impurity elements in six drill holes which come from different sections of the ore body. The information is taken from company drill logs. The horizontal axis shows increasing depth from left to right, but there is no constant scale; the values have been taken in the order in which they appear in the logs and do not attempt to relate individual values to a specific depth. A vertical line drawn through any graph indicates the values found in the same sample. The diagrams illustrate some of the points made earlier in this chapter, and clearly show that there is no constant relationship between the elements.

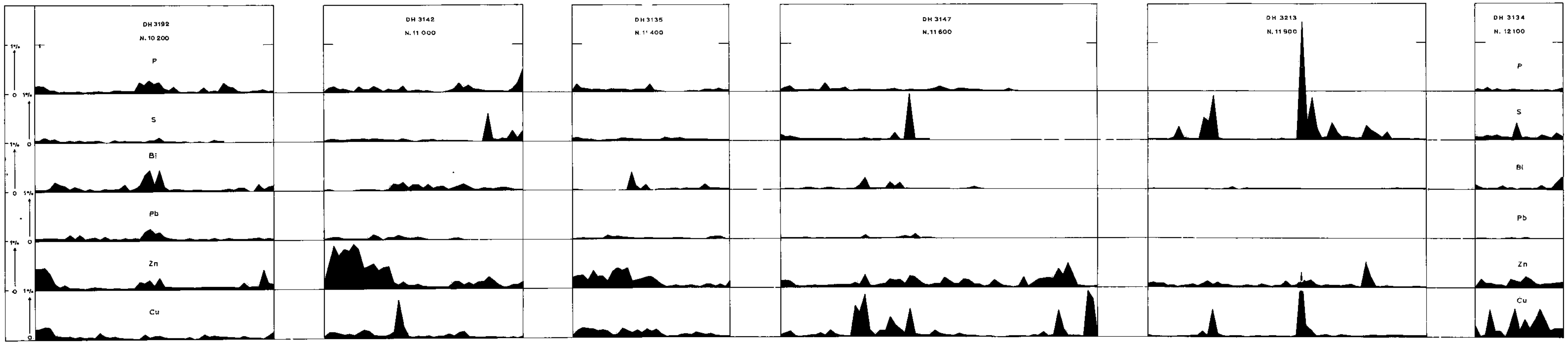

DH 3192
N. 10 200
DH 3142
N. 11 000
DH 3135
N. 11 400
DH 3147
N. 11 600
DH 3213
N. 11 900
DH 3134
N. 12 100
P
S
Bi
Pb
Zn
Cu
1%
0

Fig. [illegible] Diagram [illegible]

XV THE DEPOSITS AT BUKIT PESAGI AND BUKIT SANLONG

Introduction

These deposits lie close together and are similar in type, but only Bt. Pesagi was in production at the time of the writer's fieldwork. The Bt. Sanlong deposit consists of one small primary body with some detrital and secondary ore, and lies one and a half mile south-southwest of Bt. Ibam. The Bt. Pesagi deposit is named after the prominent hill of that name, which lies two and a half miles south-southwest of Bt. Ibam, but iron ore is scattered over an area of approximately one square mile. The locations are shown in Fig. 24. The deposits are the most southerly of the Ulu Rompin group, and from Bt. Pesagi, 750 ft, the ground drops away to the east, south and west. The area is dissected by several streams, and the ore deposits are found on the tops and upper flanks of the hills and ridges, extending down to about 300 ft elevation. Although all the deposits lie close together, for the convenience of mining operations the company recognizes nine deposits in Bt. Pesagi - Pesagi 1 to 9, and two in Bt. Sanlong - Sanlong A & B.

Country Rock

The surface exposures show only a quartz-bearing clay, which is usually stained or impregnated with secondary iron material. Even in the excavations left by mining no hard rocks can be found, but it is clear that the bedrock consists of highly decomposed acid volcanic rocks which have been sheared in a general north to south direction, and contain veinlets of quartz and kaolinite. The volcanic rocks contain disseminated pyrite, which also occurs as veins in some areas. Drilling throughout the area confirms the bedrock, and also proves the presence of some hornblende granodiorite below surface. There may have been some thin calcareous lenses within the volcanic sequence, but this is not certain.

Ore Deposits

General Description

The iron ore occurs in steeply dipping primary lenses, and in relatively thin sheets of detrital and secondary ore on, or close to, the surface. The primary bodies are aligned parallel with the regional foliation in the sheared volcanic rocks. The primary ore is usually of high grade, but the other is much more variable. In complete contrast to the Bt. Ibam deposit, primary haematite is the most important component in the ore bodies. The mine authorities believe that the steeply dipping ore bodies formed by oxidation of massive pyrite, but this is incorrect, and haematite is a primary crystallization product. The only important primary bodies are Pesagi 6, which is themost northerly of all the Bt. Pesagi deposits, and Sanlong B, which lies one mile to the north, but there are other small primary bodies throughout the Pesagi zone.

The superficial ore is a mixture which consists partly of detritus and partly of ore formed by secondary enrichment. Mining of this type of ore consists of stripping over wide areas, but only to shallow depths.

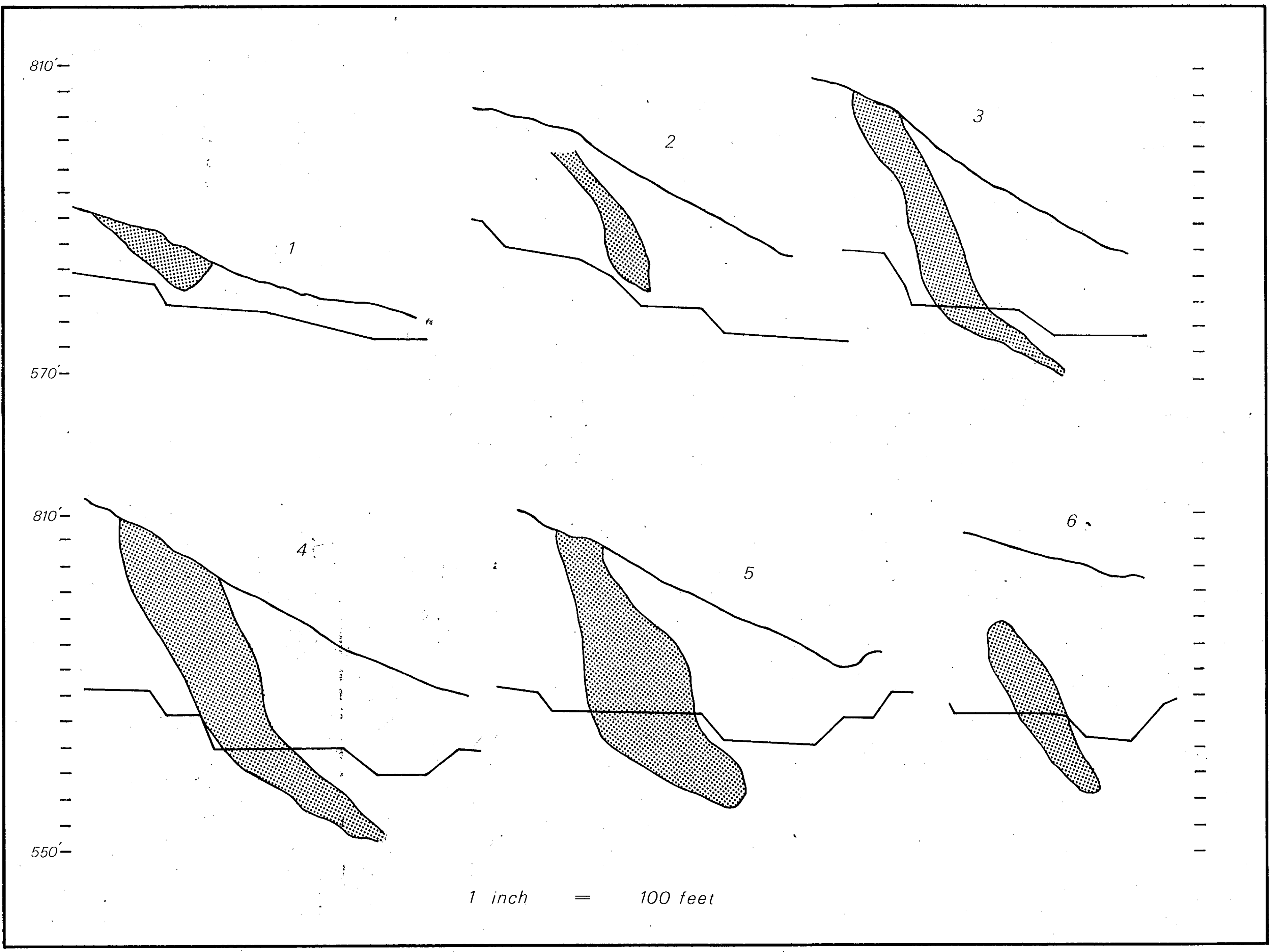
Primary Ore Bodies

Pesagi 6 is the most important single ore body, but Sanlong B, which is separated from it by a distance of one mile, is identical except for its smaller size. There are other very small primary bodies throughout the Bt. Pesagi area. Pesagi 6 is a tabular body whose strike varies from north to south to 20°, and dips east at an average of 55°-60°. The original outcrop was at an elevation of 800 ft and ore extends down to approximately 550 ft. Below this level mineralization continues, but pyrite becomes the predominant mineral. The rise in sulphur values in depth led the mine authorities to conclude that the iron ore had formed by oxidation of a massive pyrite body, but in fact the pyrite is a later phase. The ore body shows a gradual tapering in depth, except along one section towards the southern end where there is some thickening. Fig. 112 shows cross sections through Pesagi 6. In plan view the ore body is a narrow lens which has a distinctive stepped appearance due to the presence of small cross faults. Fig. 113 shows the surface plan view in September 1966. The maximum strike length is 650 ft, maximum thickness 90 ft, and maximum extent downdip 300 ft. As can be seen from the working levels plotted on Fig. 112, much of the ore body had been worked out at the time of the writer's visit. Sanlong B is exactly the same type of body, even to the presence of the cross faults, but is smaller.

The near-surface country rock is a clay which has been derived from acid volcanic rocks. Some of the clay is greyish-white, but it is typically coloured in reds, yellows, browns and greens, and has a marked banded appearance. The latter is due to strong shearing in the original volcanic rocks, and there is a general increase in the intensity of shearing as one approaches the ore bodies. Some of the immediate wall rocks contain small lenses and disseminations of haematite. A few of the cross faults which displace the ore body can be traced into the country rock, but others disappear in the clay.

The contacts between ore and country rock are sharp, and the haematite disseminations in the volcanic rocks are restricted in size and occurrence. Along the contact in Pesagi 6 there is usually a development of goethite, which is highly porous and forms cellular and ropy masses, boxworks and stalactites. One mass had a horizontal banded texture, which may have been connected with old levels of the water table. The goethite has replaced both the wallrock and theprimary ore. Within the ore body there are some small masses of goethite which are often developed along the cross faults and some joints.

The ore body contains haematite which is very variable in its physical appearance. There is a little hard, dense metallic ore, but most is soft and highly friable. The ore originally found close to the surface was hard due to the cementing effect of goethite.



profiles after ROMPINCO plan

Fig. 112 Cross sections through Pesagi No. 6 ore body. The working levels are as at September 1966. Lines of sections are shown at Fig. 113.

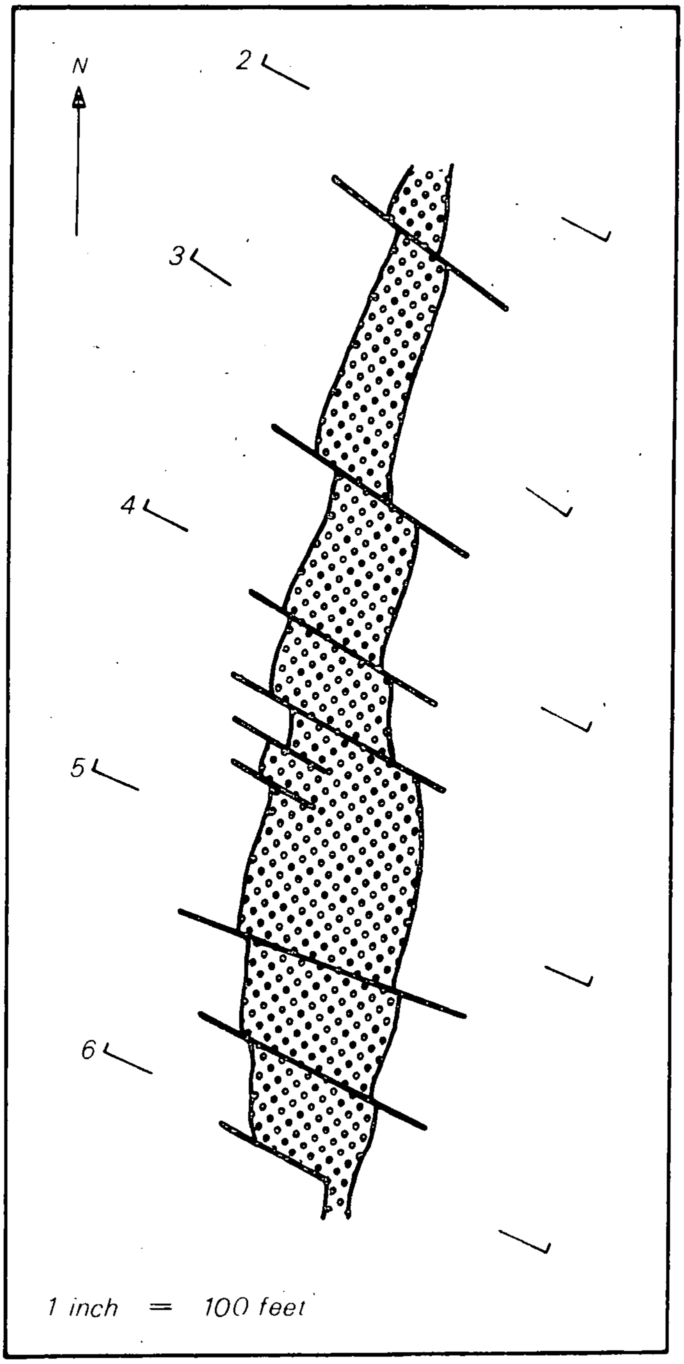


Fig. 113 Plan of Pesagi 6 ore body at Sept. 1966.

The friable ore is typically granular, but some is a fine dust. Mixed with the friable ore, and occasionally forming quite large masses, are coarse euhedral crystals of specularite and some micaceous haematite. The specularite usually forms thin plates, averaging one half by one eighth inch which are flattened parallel to (0001), but much larger crystals can be found. The largest encountered by the writer were two by one half inch. The larger crystals are usually brittle due to the development of very good parting planes, which are easily visible to the naked eye, and can be so perfect that they are virtually cleavages. The parting planes are developed parallel to $\{10\bar{1}1\}$ and are probably due to twinning. The smaller plates of specularite can be lifted with a hand magnet due to minute inclusions of magnetite. Much of the haematite is of high grade, but some patches contain quartz which contaminates the ore. Down to the working levels seen by the writer little pyrite is apparent in the ore body, but drill core shows that at the lower levels it becomes important, and the sulphur values rise sharply. Some drill holes have encountered massive pyrite below the haematite body. Although Sanlong B had not been worked at the time of the writer's visit, drill core shows that it is the same as Pesagi 6, and sulphur values rise in the deeper parts of the ore body. The other primary bodies throughout the Bt. Pesagi area are much smaller, with dimensions usually measurable in a few tens of feet. Their mineralogy is similar, but some consist largely of granular ore and lack the coarse specularite crystals.

Mineralogy

The Pesagi 6 ore body is made up of earthy and crystalline haematite, magnetite, martite and goethite. The texture is variable, even within a single polished section, and the majority of samples are porous to a greater or lesser degree. The typical ore contains granular haematite and martite with a few specularite crystals set in a porous mass of earthy haematite and goethite.

The most striking component is specularite, which can form very good euhedral crystals - Fig. 114. It may occur as isolated crystals, but some samples consist of interlocking aggregates of coarse grains, with voids or fine grained haematite and goethite between them. A few specularite crystals take a near perfect polish, but some pitting is usually present. Most of the pits are irregular, but they can follow definite crystallographic planes, which appear to be $\{10\bar{1}1\}$. Internal reflections are rarely seen unless there is some flaw in the polished surface, but the bireflectance and anisotropy are clearly visible, particularly in twinned crystals. Some specularites have an irregular rim of goethite, with rare internal patches of the same mineral. The crystals of specularite can project into cavities in the porous ore where they appear to be incorporating the more granular material, but in other places the earthy and granular ore have partially replaced the specularite.

The specularite crystals are frequently twinned. In cross section one sees two directions which intersect at an acute angle and are $\{10\bar{1}1\}$ planes, but they are occasionally accompanied by a third set of twin lamellae which appear to be following (0001). In basal sections the twin lamellae often display a complicated pattern, but are probably following $\{10\bar{1}1\}$ directions. The twin lamellae are thin with very sharp, straight edges, and they normally transect the crystal from edge to edge with a constant thickness.

Rounded relics of magnetite are found in many specularite crystals - Fig. 115. The magnetite is fresh and rarely shows any martite lamellae, and in only one or two grains have the $\{111\}$ planes in the magnetite controlled the replacement by specularite. Pyrite has crystallized in cavities in the specularite, and occasionally replaced the host, and one or two grains of chalcopyrite have also been identified.

A rare but striking texture is displayed by a number of large specularite crystals. In contrast to the usual good polish, a few grains have a very dull pitted appearance due to the presence of minute parallel bodies, which do not cut through twin planes or magnetite relics - Fig. 114. In a few grains the bodies are found in the thin twin lamellae, and not in the main body of the specularite crystal. Under very high magnification it can be seen that the bodies are thin plates with a definite orientation,

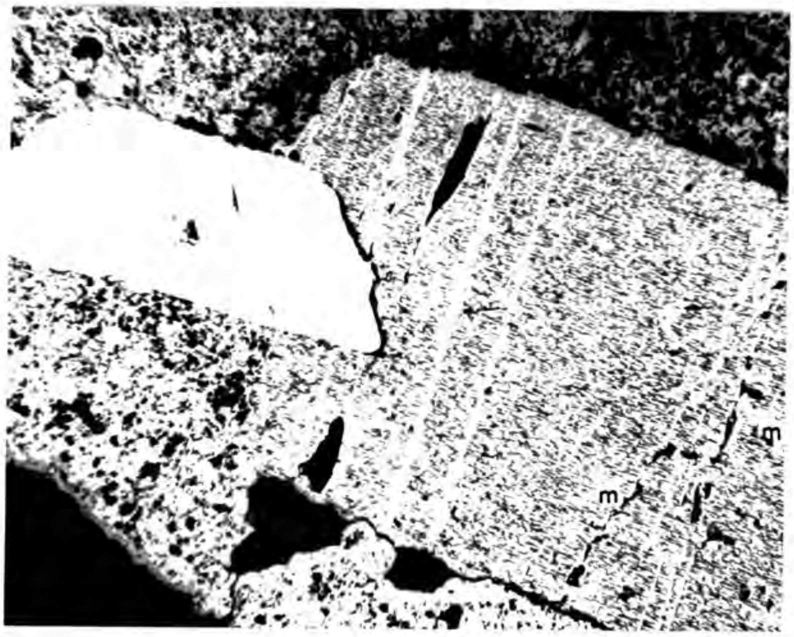


Fig. 114. Two specularite crystals, one of which has taken a very good polish while the other has a pitted appearance. Twin planes and two small magnetite grains can be seen in the crystal on the right The specularite is set in a highly porous goethite-haematite matrix. x 70.

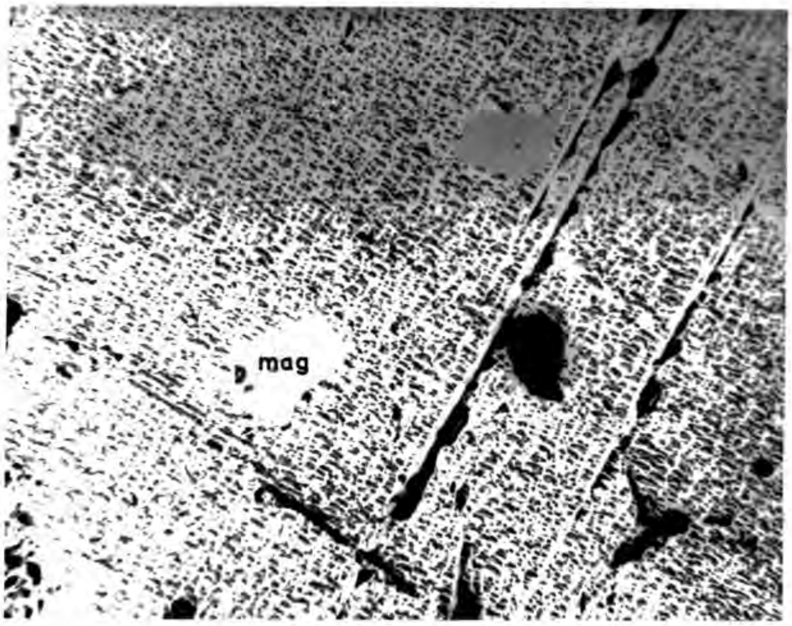


Fig. 115. Enlargement of part of the pitted crystal in Fig. 114. Myriads of small corundum (?) bodies can be clearly seen, many of which have an edge which is convex towards the top right. x 135.

and in non-basal sections the bodies have a crescent shape as they are dipping into the plane of the polished surface - Fig. 115. In basal sections of specularite crystals the plates are more equant in shape and appear to be parallel to the polished surface. The plates are formed from a hard transparent mineral, which is believed to be corundum. The corundum has possibly exsolved as small plates with its (0001) planes paralle with the (0001) in the specularite. Only a very few grains show this phenonemon.

A texture which is very common in the specularites from Bt. Pesagi and Bt. Sanlong, and has also been noted in rare grains of primary haematite from Bt. Ibam, can best be described as "pseudo perthitic" from its similarity to the texture seen in some alkali feldspars. Considering the abundance of haematite throughout the world published descriptions of this texture appear to be rare, and illustrations even rarer. The texture can be clearly seen under the microscope, but the writer found it very difficult to photograph satisfactorily. It is suspected that the texture may be much more common than references to it would indicate, because unless a very careful examination of the polished section is made the texture can be easily overlooked. It can only be observed under very high magnification, with the grain in a critical angular position under slightly uncrossed nicols. The only illustrations known to the writer are in Ramdohr (1960) and a photograph of a Venezuelan iron ore (R. Phillips, pers. comm. and Kalliokoski, 1965).

The texture is revealed by the anisotropy of haematite as the microscope stage is rotated, the most common variety of the texture being very similar to perthite - Fig. 116. The texture is developed by elongate irregular "bodies", which have a strong parallelism on a broad scale. Some of the "bodies" appear as isolated units, but many link up to form a type of network. Although the term "bodies" is used, they are in fact small areas of haematite in different orientation to the adjacent haematite, and are not composed of foreign material. The texture can be developed on such a fine, regular pattern, that in a certain angular position some of the specularite crystals look to have a micro ridge-and-valley surface - Fig. 117. In this variety of

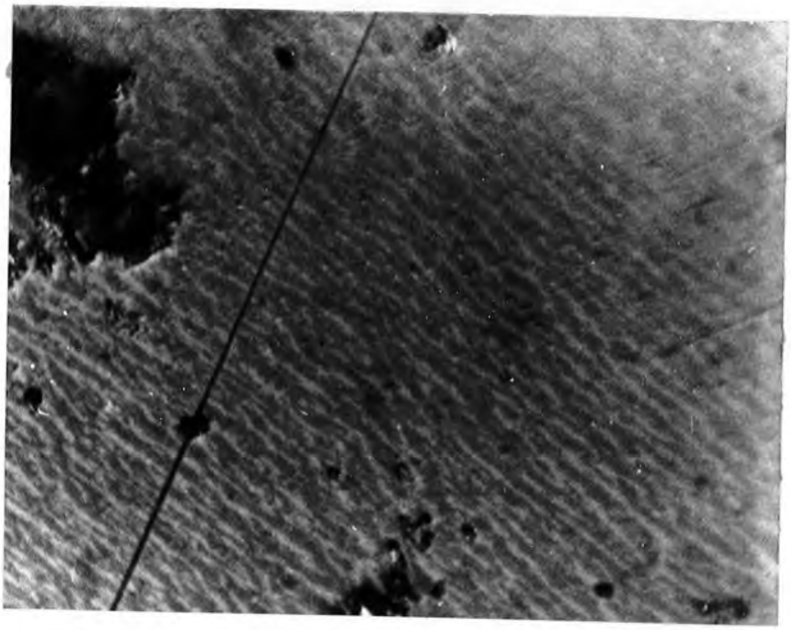


Fig. 116. Typical "pseudo perthitic" texture which is seen in many specularite crystals from Bt. Pesagi and Bt.Sanlong x 800.

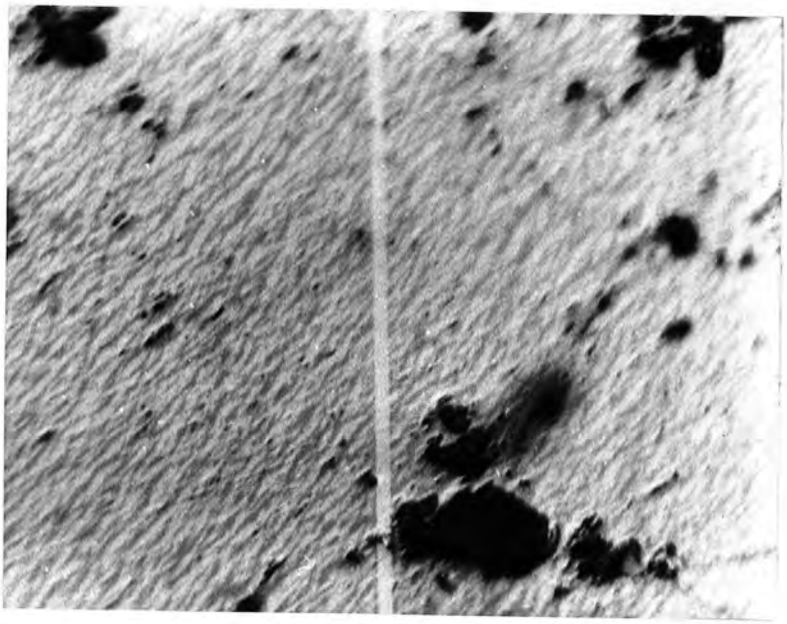


Fig. 117. "Pseudo perthitic" texture in grain of specularite. The are lower left looks to have some relief, although it has in fact a very high surface polish. x 800.

the texture it is impossible to say which are the "bodies" and which the host, as both are developed in equal proportion and show the same pattern. In rarer examples however, one set of haematite "bodies" is subordinate to the main mass of the crystal, and the texture changes slightly as the specimen is rotated. An example is shown in Fig. 118, where it can be seen that the light coloured lamellae do not display the same pattern as the darker material. In contrast, if the position of the dark and light areas in Fig. 117 were reversed, the texture would not be significantly altered. An even more extreme difference between "bodies" and host can be seen in some specularites where thin, hairlike lamellae are set in a much more massive host - Fig. 119.

The pseudo perthitic texture can be uniformly distributed throughout a whole crystal, or developed only in certain patches. It usually follows only one direction in a single crystal, and where a change in direction is present, there is often evidence that the whole crystal has changed its orientation e.g. $\{10\bar{1}1\}$ twin lamellae show a change in orientation. The angular relation of the pseudo perthitic texture relative to $\{10\bar{1}1\}$ planes is not consistent, as it can be parallel - Fig. 120, oblique - Fig. 117, or almost at right angles - Fig. 118. The optical orientation also shows a variable relationship to $\{10\bar{1}1\}$ lamellae, as the pseudo perthitic texture can extinguish with the main twin lamellae, but more often the positions are different. The texture normally terminates abruptly at a $\{10\bar{1}1\}$ lamella, to continue on the other side, but in one example the texture appeared to continue through the twins - Fig. 121.

The origin of this texture is problematic. As haematite is the only mineral involved and the texture is produced by differences in orientation, it must be a form of microtwinning. Unfortunately it is not known if the pseudo perthitic texture follows one consistent crystallographic direction in the haematite; the evidence from polished sections suggests that the direction may be variable. If the texture is caused by twinning it could be a growth phenomenon or a type of pressure twinning. If it were the former one would expect a consistent crystallographic direction, and it is concluded that the "perthitic" texture is

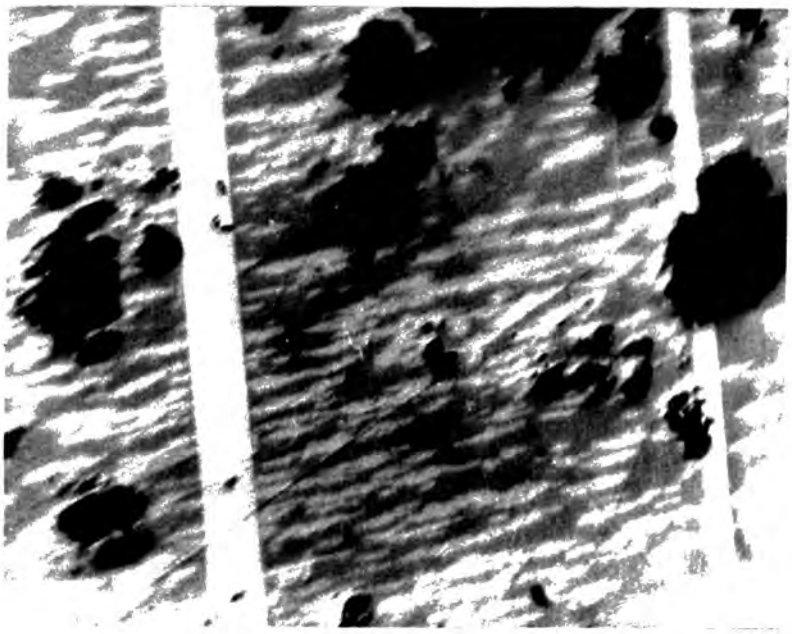


Fig. 118. A relatively coarse variety of "pseudo perthitic" texture in specularite. x 800.

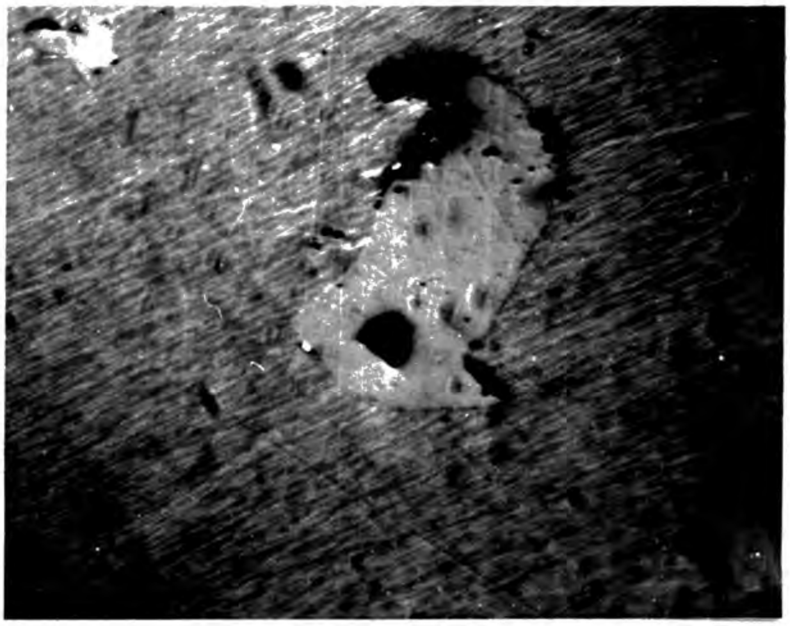


Fig. 119. A hair-like variety of "pseudo perthitic" texture. The grain in the centre is magnetite. x 800.

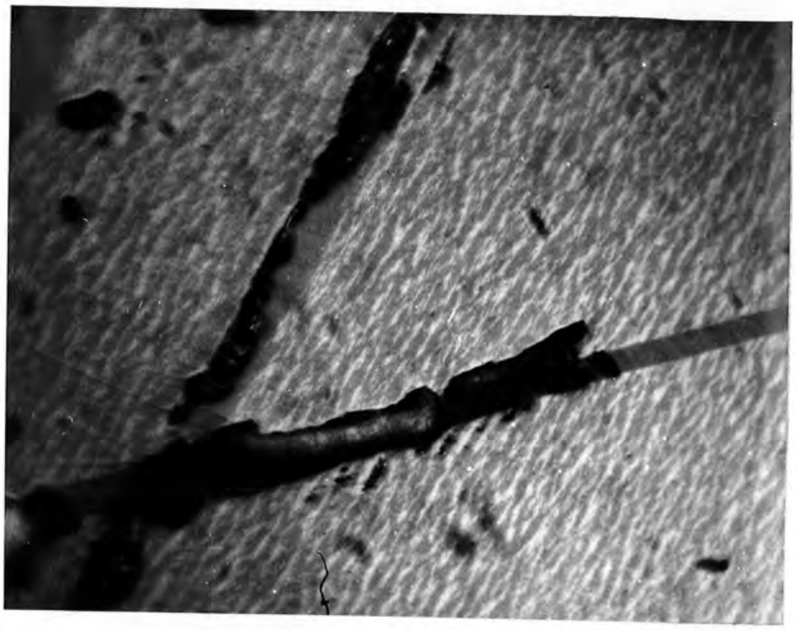


Fig. 120. "Pseudo perthitic" texture in specularite. The texture is parallel to one $10\bar{1}1$ twin plane and at an angle to another. x 800.

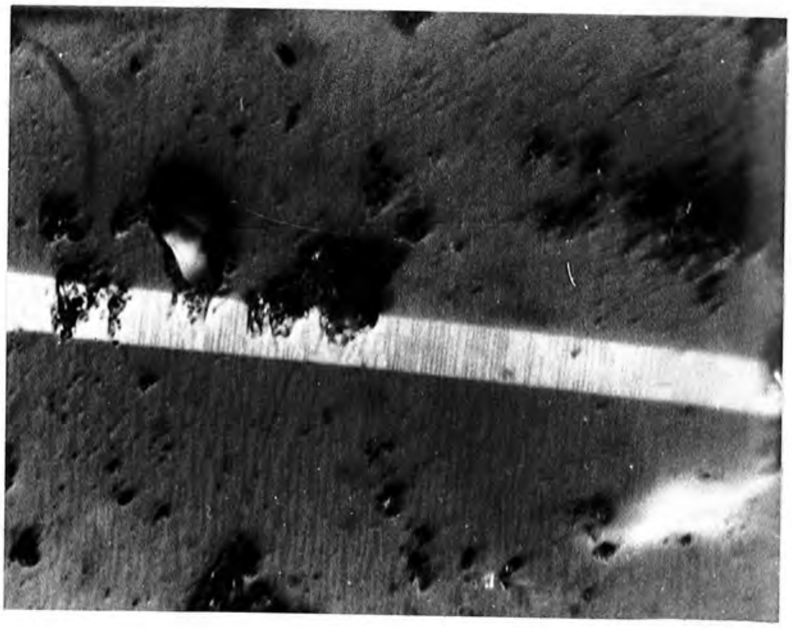


Fig. 121. Very fine "pseudo perthitic" texture which passes without deviation through a $10\bar{1}1$ twin plane. x 800.

the result of stress. The causative stresses must have been relatively light as there is no sign of dislocation or shattering in the polished sections studied, but the presence of cross faults in the ore body proves that stress has been operative at some period after mineralization. The iron ores of Yampi Sound, Western Australia, contain haematite which displays the perthitic texture, and this is thought by Canavan and Edwards (1938) to be a micro twinning caused by internal stresses, and the ores at El Pao, Venezuela, which have been dynamothermally metamorphosed show the same feature (Kalliokoski, 1965).

The granular component of the Bt. Pesagi type ore is partly martite and partly haematite. The ore which contains considerable martite usually takes a bad polish, but the typical lattice texture can be seen, while the granular primary haematite takes a much better polish. One can find magnetite grains which have been pseudomorphed by an aggregate of small grains of haematite, and some grains have a martite centre with granular haematite forming the edge, or vice versa. Relics of magnetite are usually present, and some maghemite has been identified. A diffraction trace of a sample from one of the small primary bodies showed a clear maghemite pattern. The granular ore has a very porous texture, with numerous irregular cavities. Specularite crystals have occasionally crystallized in the cavities, and some cavities have been partially filled with a goethite-haematite mixture.

Goethite is a variable component, but has crystallized in fractures and cavities, and replaced the other iron oxides. The veinlets and cavity fillings usually display a colloform texture, and the goethite is mixed with colloform haematite.

The main sulphide is pyrite, but small grains of chalcopyrite have been observed. Pyrite is found as small rounded bodies in some of the specularite crystals, and it also forms veinlets and intergrain fillings. One polished section showed a magnetite grain which had been peripherally pseudomorphed by pyrite, and the sulphide has also been observed to replace haematite. Pyrite has often crystallized in cavities in the porous type of ore. Evidence from numerous polished sections shows that pyrite is later than the iron oxides, and in one abandoned mine working pyrite veins in granular iron ore were clearly visible. Rare grains of bismuthinite and pyrrhotite have been identified in the pyrite.

All the primary bodies, from the large Pesagi 6 to the smallest lens, show the same sequence of mineralization, although their final percentage mineral composition can vary. In all the bodies magnetite was the first ore mineral to crystallize, and primary haematite deposition followed later. In the smaller bodies the haematite has partially replaced the magnetite but retained the fairly even, granular texture, while in the larger bodies the haematite has, in addition, frequently crystallized in large specularite crystals. The ore with abundant specularite superficially looks quite different from that in the smaller bodies, which has led some of the mine authorities to suggest that different types of ore bodies are involved, but the mineralogical evidence shows that all the bodies have followed the same genetic sequence.

Secondary Ore Bodies

The secondary ore forms relatively thin surface sheets over some of the hill tops and upper flanks of the ridges in the Bt. Pesagi and Bt. Sanlong areas, those in the former being much more extensive. Three sources have contributed to the development of these ores; detritus which has broken off from the primary ore bodies and moved downhill under gravity, material from the destruction in situ of small primary lenses, and an enrichment of original iron minerals in the bedrock. The ore which has formed by secondary enrichment has developed from pyrite in the volcanic rocks, and possibly from magnetite in the suboutcropping granodiorite. Some of the detrital ore has remained uncemented and is found as boulders and pebbles in a clay matrix, but it is not unusual to find that the detritus has been cemented by secondary ore into hard solid masses. The general picture is that of an extensive lateritic sheet, upgraded in places by the incorporation of material from the primary bodies. The laterite probably formed as a continuous sheet when the ground surface was flatter and less dissected, and the present river system has broken through and left the isolated pockets found on the hill tops and ridge slopes.

The bulk of this ore is goethite with secondary haematite, but relics of primary haematite and magnetite are found where primary ore bodies have contributed material. The secondary ore is dominated by colloform textures, and in hand specimen is usually highly porous.

Geochemistry (1) Iron Ores

The fact that these deposits contain ore ranging from high grade specularite to laterite means that there is a corresponding variation in composition. The best ore is found in the upper levels of Pesagi 6 and Sanlong B; at lower levels the sulphur values rise steeply. From company analyses the writer calculated the following average for an 85 ft drill core section through Pesagi 6:

Fe 66.8%; S 0.095%; P 0.013%; Cu 0.008%; Zn 0.012%

and the range of these elements in the same section is

Fe 62.8% - 69.3%; S 0.005% - 0.312%; P 0.002% - 0.021%;
Cu 0.005% - 0.014%; Zn 0.006% - 0.020%; As tr - 0.005%;
Pb $<$0.005%; Bi $<$0.010%.

The writer separated a sample of high grade ore from Pesagi 6 into its two components, large specularite crystals and fine haematite powder with dust, and analysed them for comparison purposes. The results are shown below, and reveal that the two fractions have almost identical compositions.

	Fe	SiO_2	Al_2O_3	P	CaO	TiO_2
1.	69.5	0.20	0.14	n.d.	0.07	0.03
2	69.3	0.21	0.10	n.d.	n.d.	0.01

1. Specularite ore. 2. Fine powder and dust.

For the Sanlong B ore body the following results are typical:

	Fe	S	P	Cu	Zn	Pb	Bi	As
1	66.2	0.098	0.019	0.008	<0.005	<0.005	<0.005	0.030
2	54.6	3.78	0.026	0.009	0.005	0.005	0.005	0.020
3	63.4	0.060	0.015	0.009	0.006	-	-	0.015
4	57.7	4.10	-	0.015	-	-	-	-

1. Drill core 0ft - 135 ft) calculated by the writer from
2. Drill core 135ft - 155 ft) company analyses.
3. Average for upper levels of the ore body.
4. Average for lower levels of the ore body.

There is a good agreement between these pairs of figures (1 + 2 and 3 + 4), which clearly show that sulphur is the only impurity of note, and it only becomes important in the lowest levels of the ore body. Cross sections through Pesagi 6 contructed by the company demonstrate the marked rise in sulphur values at depth, and led some of the company geologists to suggest that the iron oxides had formed by oxidation of a massive sulphide body. This is a convincing example of the danger of interpreting chemical data without supporting mineralogical evidence. Fig. 122 is a typical cross section through Pesagi 6 showing the iron and sulphur values.

Chemical data for the boulder ore are shown below:

	Fe	S	P	Cu	As	SiO_2	Mn
1	52.5	0.054	0.029	0.033	-	13.23	0.24
2	56.0	0.064	0.016	0.015	0.024	-	-
3	54.0	0.018	0.050	0.013	-	12.03	0.16

1. Bt. Pesagi, Japanese data.
2. Bt. Sanlong, company data.
3. Bt. Sanlong, source uncertain but probably company data.

The average composition of the Bt. Sanlong lateritic ore which as formed over the local pyritic volcanic rocks by residual concentration is given in company records as

Fe 54.0%; S 0.168%; P 0.040%; Cu 0.021%; As 0.045%.

The overall composition of all the ores, primary and secondary, found at Bt. Pesagi and Bt. Sanlong is reported to be

	Fe	S	P	Cu	Zn	Bi
1	58.0	0.05-0.5	0.03	0.03	0.03	0.02
2	60.0	0.02-0.15	0.02-0.05	0.01-0.02	0.02	0.02

1. Bt. Pesagi, company data.
2. Bt. Sanlong, UN (1970) and company data.

A number of trace elements in a selection of iron ore samples from Bt. Pesagi and Bt. Sanlong were determined by the writer, and the results are shown below (in ppm).

<u>Bt. Pesagi</u>

Sample No.	Cu	Zn	Pb	Bi	Ba	Mn	Cr
109	63	80	209	n.d.	n.d.	-	-
110	60	<10	n.d.	n.d.	12	-	-
111	39	90	36	n.d.	12	-	-
112	159	<10	84	<10	26	-	-
113A	38	<10	n.d.	<10	n.d.	62	<10
113B	48	48	34	25	n.d.	180	<10
298	93	100	36	n.d.	93	124	n.d.
302	36	68	78	n.d.	52	-	-
Average	67	50	60	4	24		

109. Granular haematite. 110. Specularite. 111. Granular haematite. 112. Specularite + granular haematite. 113A. Specularite. 113B. Haematite dust. 298. Granular haematite replacing magnetite. 302. Martite and granular haematite replacing magnetite.

The results show that the general level of these impurities is low, and that there is little variation from sample to sample.

The figures for the samples from Bt. Sanlong are shown below (in ppm).

Bt. Sanlong

Sample No.	Cu	Zn	Pb	Bi	Ba	Mn	Cr
280	54	58	93	n.d.	76	124	84
282	33	45	18	n.d.	18	-	-
283	39	31	≤10	n.d.	20	-	-
284	58	21	42	n.d.	21	93	n.d.
287	1460	12	≤10	n.d.	16	-	-
Average	545 (46)*	36	35	-	30		

* Average of first four samples only.

280. Granular haematite. 282. Granular haematite with specularite. 283. Granular haematite with specularite. 284. Granular haematite with some pyrite. 287. Granular haematite and specularite with some pyrite.

Samples No. 280 to No. 284 came from one drill core with depths ranging from 10 ft to 140 ft respectively. The results above are comparable with those obtained for the Bt. Pesagi ore except for the high copper in No. 287. This sample contains megascopic pyrite (as does No. 284), but there is no sign of any copper mineral, and it is thought that the result may be spurious due to instrumental error. It is perhaps significant that the writer's values are closely comparable with those obtained in the company's laboratories on similar, but not identical, material, except for the high copper in No. 287. The values obtained by the writer for the Bt. Pesagi and Bt. Sanlong ore are of the same order of magnitude as those from other sources quoted earlier in this section and confirm the overall low level of metallic impurities. Copper and zinc are lower than in the Bt. Ibam ore but lead is slightly higher.

Geochemistry (2) Country Rocks

The major element chemistry of the country rocks has been described earlier - chapters III, IV and V - and this section is restricted to a number of trace element determinations. Decomposed acid volcanic rocks are by far the commonest type and trace elements in these were determined as (in ppm):

Bt. Pesagi

Sample No.	Cu	Zn	Pb	Bi	Ba
105	26	18	n.d.	n.d.	1720
106	29	20	33	n.d.	770
170	14	56	n.d.	n.d.	875
172	17	30	18	n.d.	2358
300	69	16	n.d.	n.d.	2240
304	15	48	19	n.d.	1280
326	15	108	174	n.d.	303
Average	26	42	35	-	1363

Bt. Sanlong

286	14	27	<10	<10	692
288	24	<10	n.d.	n.d.	775
Average	19	19	-	-	733

Between Bt. Pesagi and Bt. Ibam

329A	17	26	n.d.	n.d.	1037
329B	43	800	45	n.d.	418
329C	45	960	<10	n.d.	1672

Samples Nos. 105, 106, 170 and 172 were collected from benches in the workings in the important No. 6 ore body, and No. 326 from near to No. 2 ore body which is a small primary lens. Sample No. 300 came from drill core at a depth of 200 ft and below No. 6 ore body, and No. 304 from a depth of 160 ft in country rock. No. 286 came from above Sanlong B ore body at a depth of 87 ft, while No. 288 came from the same drill hole at a depth of 184 ft and below the ore body. The three samples numbered 329 were collected from a cutting along the haulage road from Bt. Ibam to Bt. Pesagi and do not really belong with either body.

The barium shows a consistently high level, which agrees with other determinations made on the volcanic rocks throughout Ulu Rompin. Of the metallic impurities bismuth is of no consequence, but two samples show above average zinc. It is noticeable that the other elements have not increased sympathetically with the zinc. The two samples came from a road cutting which lies outside the zone of iron mineralization, and the writer believes that they may have been affected by the highly zincian zone found at the southern end of the Bt. Ibam ore body. The zinc has probably been introduced in carbonate-bearing solutions and incorporated into goethite when the carbonates were oxidized; the high zinc is very local as sample No 329A originated from within the same road cutting. Sample No. 326 shows slightly elevated lead and zinc values, but the writer has seen no minerals which would account for these values and they probably represent higher lead and zinc in the original volcanic rock.

Two samples of relatively fresh, hard granodiorite were analysed with the following results:

Sample No.	Cu	Zn	Pb	Bi	Ba
311	23	94	<10	n.d.	158
324	34	71	n.d.	n.d.	865
Average	29	83	-	-	512

Sample No. 311 came from a depth of 333 ft and sample No. 324 from a depth of 160 ft.

Three unusual samples taken from drill core which did not encounter primary ore but ended in hard granodiorite (sample No. 311) showed much higher than average zinc values, although no other impurities were significant. The values are shown below:

Sample No.	Depth ft.	Cu	Zn	Pb	Bi	Ba
308	156	17	662	n.d.	n.d.	117
309	228	61	2280	58	n.d.	182
310	231	68	1060	n.d.	n.d.	411

Sample No. 308 is a skarn-like rock with pyroxene and amphibole and is mentioned in chapter IV. The other two samples are types which the writer has not encountered elsewhere in Ulu Rompin, and they appear to be hybrids. They lie above granodiorite and are probably the result of the igneous rock encountering a small lens of calcareous material. The samples are unique among the Bt. Pesagi and Bt. Sanlong country rocks as they contain appreciable magnesia (ca 10% MgO), and in this respect show similarities to the gangue in the Bt. Ibam body. There is no sign of sphalerite or other discrete zinc mineral, and the writer believes that the zinc may have been present in the original carbonate rock and subsequently incorporated into the hybrid rocks. It is clear that these rocks are completely out of the normal range of types found in the area, and they occupy only a very small area.

XVI THE BUKIT HITAM DEPOSIT

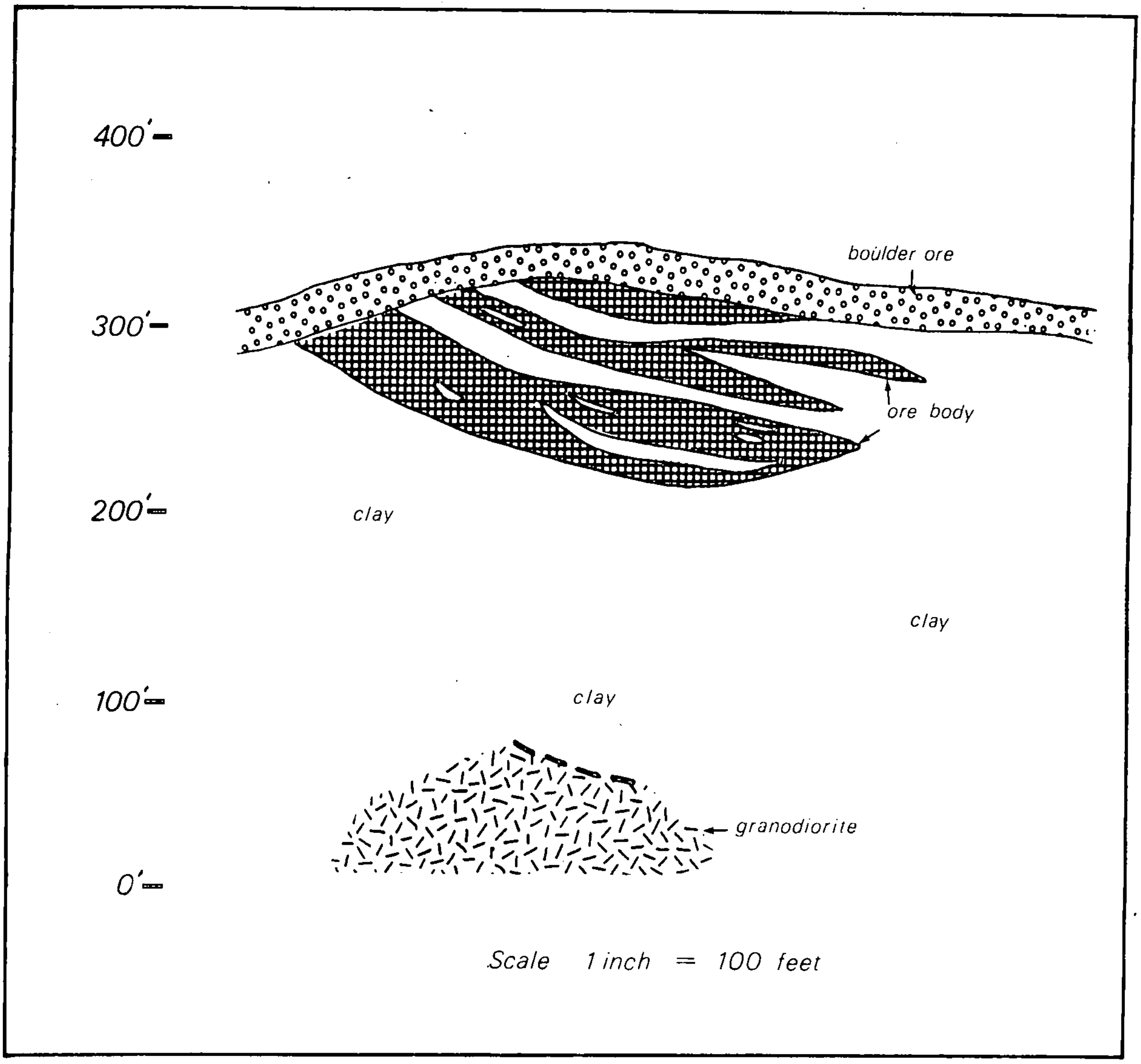
It was not possible to visit Bt. Hitam during the writer's final fieldwork as there was a dispute over the ownership of the land, but it has been examined during an earlier trip to Ulu Rompin. No mining had been undertaken up to September 1966, but the area had been prospected.

Bt. Hitam lies four and a half miles on a bearing of 10° from Bt. Ibam - Fig. 24, and is the most northerly of the deposits within the narrow belt of country which contains all the primary deposits, with the exception of Bt. Batu Puteh. The ore occurs on the crest and eastern and southern flanks of a horseshoe-shaped ridge, which is open to the north and rises to a peak of just over 350 ft. The main concentration of ore is found at the southern end of the ridge. The topography is much less rugged than that found along the main mass of the Bt. Sembilan ridge, with low rounded hills surrounded by flat areas of seasonal swamp. There are no outcrops anywhere near the ore deposits, and a company report states that no outcrops were found "within 3 miles" of the area held under title, but this must be a mistake, as exposures are found on Bt. Chepemak 664 ft, which lies within three miles of Bt. Hitam.

The local geology is virtually unknown. Drill core is not a great deal of assistance as the rocks are usually completely decomposed, but some of the drill holes pass into granodiorite at depth. It is not possible to locate the precise contact, but unequivocal granodiorite has been encountered at depths ranging from 200 ft to 300 ft from surface. Granodiorite also forms the small hill mass of Bt. Chepemak which lies two miles to the east-northeast, and where there are signs of iron mineralization. A company report dated 1962 states that "limonite and manganese-limonite mixtures pseudomorphing rhombdodecahedral garnet crystals are common", and suggests that these were originally andradites formed in a skarn, but no other evidence is presented to support the identification. While in the mine the writer tried to obtain samples of the pseudomorphs but they had all disappeared, having been used for assay purposes or thrown away. Sample No. 17683 in the University of Durham collection however, contains limonite pseudomorphs of rhombdodecahedral crystals and is almost certainly the type of material to which the company report refers, but there is no doubt that the euhedral crystals were originally magnetite and not garnet. The conclusion that the ore formed in a skarn the writer regards as unproven. As a corollary to the postulated skarn origin, a company report suggests that the primary ore formed in a thin calcareous horizon which occurs in a synclinal structure with a southeasterly plunge. The writer has no means of commenting on this suggestion due to lack of evidence, but as mentioned above, it would appear to be mere conjecture until the presence of calcareous rocks is positively established.

By analogy with the regional geology in surrounding areas it would appear that the local bedrock is dominantly altered acid volcanic rocks, and this is corroborated to some extent by the barium content. The volcanic rocks throughout Ulu Rompin show a higher barium content, >500 ppm, than the other rock types, while the skarns found at Bt. Ibam are very low in barium. The fact that the clay samples from Bt. Hitam contain a relatively high barium content strongly suggests that they have been derived from volcanic rocks, but the former presence of small calcareous lenses cannot be discounted.

The main deposit consists of primary lenses with associated detrital ore, and some ore formed by secondary enrichment. The main primary body is a stubby lensoidal mass with several clay intercalations. The dimensions are not known precisely, but the maximum thickness is 110 ft, with the length and breadth approximately 400 ft and 350 ft respectively. The body strikes northeast to southwest and dips southeast at about 10°. Towards the southwest the ore body splits into several thin lenses separated by clay. Fig. 123 shows a cross section through the thickest part of the ore body. The two deepest drill holes encountered granodiorite approximately 150 ft and 200 ft below the base of mineralization. The superficial ore contains detritus from the primary body, plus limonitic concretionery ore which can in part be manganiferous.



after ROMPINCO plan

Fig. 123 Diagrammatic cross section through the Bt. Hitam ore body

The typical primary ore is a very hard, fine grained, metallic grey magnetite, which often contains irregular cavities and cracks. Some of the cavities contain a hard siliceous goethite or a soft earthy limonite, or they can be filled with skeletal limonite boxworks. A few of the cavities contain hard to soft manganese oxides, which are later than the siliceous goethite. Where the magnetite has been extensively altered to martite the ore takes on a reddish colour. Some coarse crystals of magnetite can be found, and a company report refers to limonite pseudomorphs of "pseudocubic striated haematite" crystals, but to the writer these appear to be replacements of pyrite or magnetite - see below. Fresh sulphides have not been observed in the outcropping ore.

In polished section the primary ore is seen to be an interlocking granular mass of magnetite in various stages of alteration to martite. In the less altered ore the crystal outlines can be discerned, but when highly altered the original grain edges are often obscured. Martite shows the typical lattice texture and forms preferentially along the grain edges and cracks. Some of the magnetite has a patchy colour which is probably due to incipient oxidation, but maghemite lamellae as seen in the Bt. Ibam ores have not been observed. It is interesting that three samples taken at different levels within the ore body showed more advanced martitization in the deeper samples, which suggests that martite is not necessarily a weathering product but may also be a retrograde feature of the main mineralization.

Goethite veins and partially replaces the magnetite and martite, and veinlets of manganese oxides have been observed in polished sections. Rare specks of pyrite were seen in cavities in the primary ore, and one grain of native copper, although no other cupriferous minerals have been identified. A polished section of some limonite cubes with striated faces showed a cellular mass of goethite and, surprisingly, some magnetite and martite relics. No sulphides were present. The external form certainly suggests that the crystals were originally pyrite, but the polished section shows that magnetite was present. There could have been a pyrite overgrowth on a magnetite crystal or replacement of pyrite by magnetite, but the latter is most unlikely as the sulphides are normally later than the oxides. Cubic forms of magnetite are known but are reported to be rare (Dana & Ford, 1959).

In addition to the primary ore lenses there are surface concentrations of mixed iron and manganese oxides. These ores are lateritic in origin, and consist of hard to earthy goethite with a variable manganese content due to the presence of unidentified manganese oxides.

Geochemistry

The massive magnetite ore is usually of very good grade but it has a relatively high manganese content, and the zinc can also be high, but the other impurity elements are typically low. A company report gives the average composition as

Fe 65%; Mn 2.0%; Cu 0.05%; Zn 0.2%

and states that all the manganese is locked up in the magnetite lattice, but this is incorrect as part is contained in the veinlets of manganese oxides which are often seen in polished sections. Part of the zinc content is possibly present in the manganiferous veinlets while some is in the magnetite lattice. Two samples of hard, massive magnetite taken from surface exposures contained the following:

	Fe	Mn	Cu	Zn
1	67.2	2.0	0.015	0.35
2	66.0	1.8	0.009	0.30

A United Nations (1970) survey of the iron ore resources of the world gives the composition of the Bt. Hitam magnetite deposit as

Fe 61%; S 0.05%; P 0.07%; Cu 0.1%; Zn 0.13%

but no source is quoted, and the figure for copper looks too high; the figures may well include the detrital as well as the primary in situ ore. The limonitic material associated with the magnetite occurs in two forms (1) hard brown and (2) soft yellowish-brown, and typical compositions are

	Fe	SiO_2	Al_2O_3	Mn	Cu	Zn
1	58.6	2.1	1.9	0.07	0.075	0.14
2	61.1	-	-	0.36	0.057	0.08

No discrete copper or zinc minerals have been identified in these limonites and theimpurity elements must be contained in the lattice structure.

The average composition for the detrital and secondary ore in three sub-areas at Bt. Hitam is reported to be

	Fe	Mn
1	56	0.7
2	53	2.5
3	42	10.0

the samples in area 3 being manganiferous laterites.

For the deposit as a whole i.e. primary plus detrital and secondary ore the company state that the composition is

Fe 55%; Cu 0.07%; Zn 0.05%; Bi tr.

Fig. 124 shows a graph of the Fe, Zn, Cu, S and P values in drill core obtained from the centre of the primary ore body from surface to 120 ft. It can be seen that on a broad scale the impurity elements are antipathetic with iron. Zinc is the most abundant element and shows a wide variation in concentration, but drops below 0.07% in only one short section, whereas copper is usually low but rises in the central part of the ore body. Sulphur and phosphorous are typically below 0.05%, and cause no problems as impurities. Arsenic and bismuth are low and so is lead, although the latter is higher in the low grade ore. Manganese is over 1.0% in all but two or three samples.

Samples of ore and country rock collected and analysed by the writer gave the following results (in ppm).

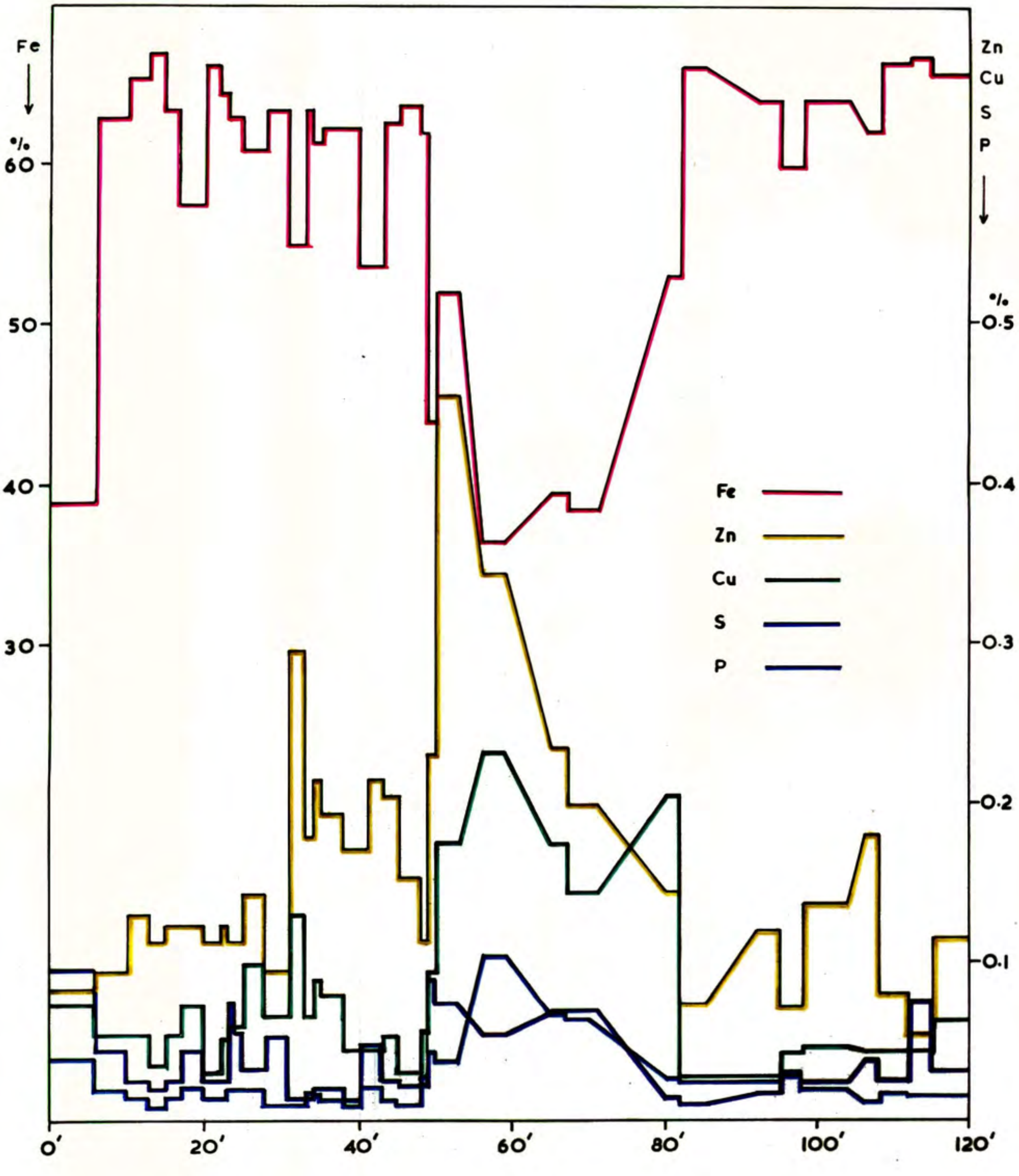


Fig. 124 Graph showing the variation of iron and four impurity elements in drill core from Bt. Hitam.

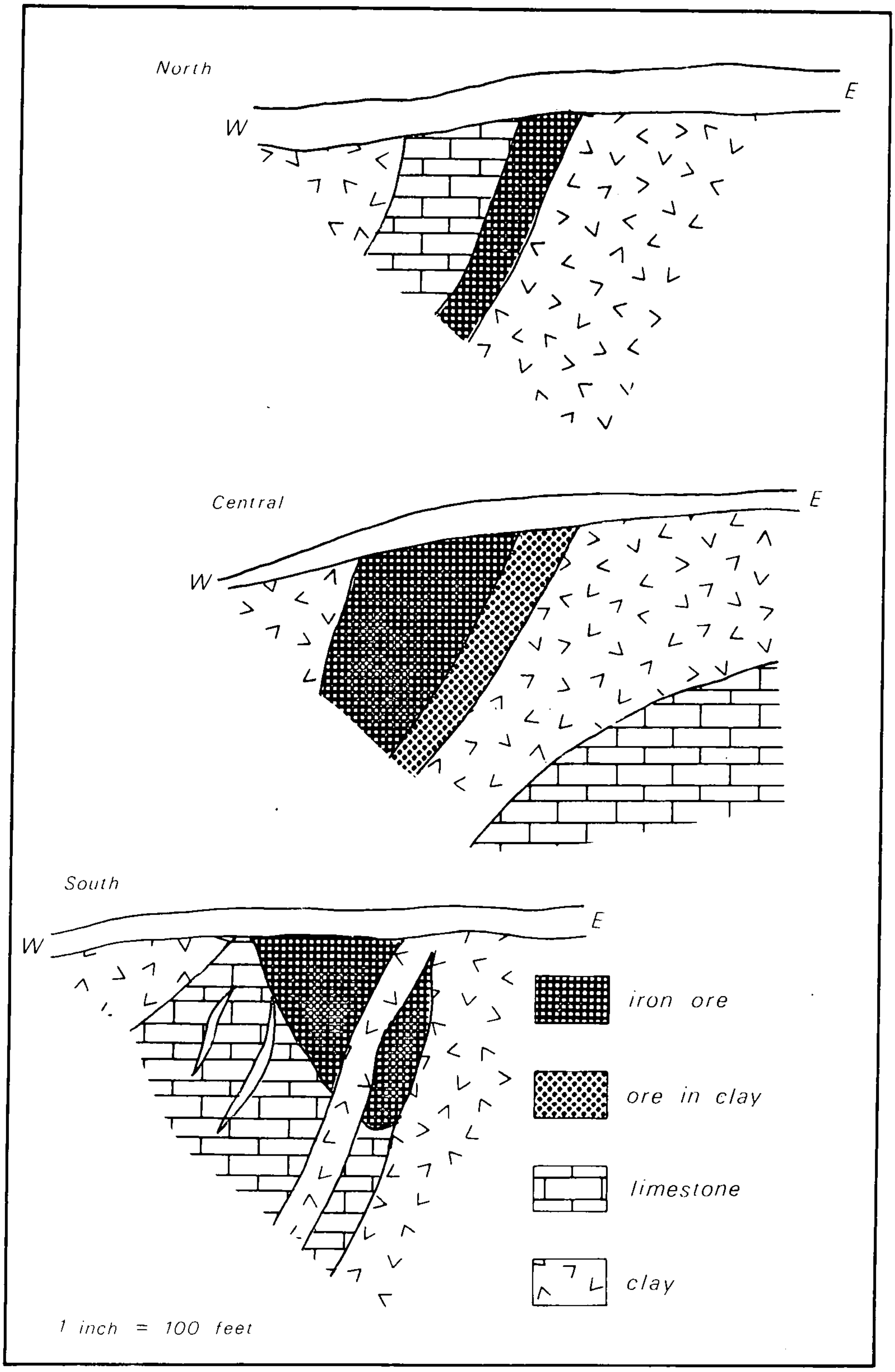
Sample No.	Depth ft.	Cu	Zn	Pb	Bi	Ba
317	70	1575	1950	1650	95	1222
323	126	33	134	57	<10	1275
318	352	38	60	151	n.d.	815
319	173	43	68	231	12	100
320	32	1700	3190	1482	58	2876
321	83	73	1030	43	n.d.	n.d.
322	119	122	401	27	206	92

317. Clay from volcanic rock. 323. Clay from volcanic ? rock. 318. Solid granodiorite. 319. Clay from granodiorite ?. 320. Manganiferous boulder ore. 321. Solid magnetite. 322. Solid magnetite.

XVII THE BUKIT BATU PUTEH DEPOSIT

Bt. Batu Puteh, which rises to a height of 250 ft, is located five miles northeast of Bt. Ibam, near the confluence of the Sg. Batu Puteh and the Sg. Trok - Fig. 24. It is the only significant primary deposit outside the narrow mineralized zone which runs from Bt. Pesagi to Bt. Hitam. The area had not been opened for mining at the time of the writer's fieldwork, and the only information was contained in drilling results. These are capable of more than one interpretation, but it is certain that the primary ore is contained in one or more tabular bodies, aligned roughly north to south and dipping west at 60°. The ore has a maximum thickness of 80 ft and a strike length of 900 ft, and ore has been encountered 120 ft downdip, but the full extent of the ore in depth has not been determined. In plan view the ore body appears to have a shallow "S" shape, which may be due to faulting, folding, or the fact that there are two bodies en echelon.

The Bt. Batu Puteh mineralization is interesting as it is the only body apart from Bt. Ibam where the ore is definitely associated with carbonate rocks. Most of the drill holes intersect a marble which is hard, medium to coarse grained (crystals up to 4 mm), and is usually white or grey in colour, but can be stained in shades of red and brown. No structure is discernible in the drill core samples, and it is clear that the rock has completely recrystallized, but no trace of calcsilicate minerals has been seen. Some parts of the marble contain cavities, usually filled with a brown clay. The marble occurs as one or more lenses in a reddish-brown clay, which has probably been derived from a parent volcanic rock by extreme alteration. No trace of any original texture or mineralogical composition is visible, but X-ray diffraction shows that the clay contains only quartz, kaolinite and sericite, which is typical of the altered volcanic rocks elsewhere in Ulu Rompin. In some core samples there is a suggestion that the original rock may have been brecciated. Fig. 125 shows three sections through the ore body.



after ROMPINCO plan

Fig. 125 Cross sections through the Bt. Batu Puteh deposit.

Drill intersections towards the centre of the ore body show a massive, good quality ore, but towards the northern and southern ends the ore is more fragmentary and enclosed in a brown clay. The writer has not seen any samples which show iron ore in marble, although marble fragments are occasionally found in the clay together with fragments of iron ore. The primary ore is crystalline magnetite, up to 5 mm diameter, now partially altered to martite and goethite. The magnetite is dodecahedral, some of the crystal faces being visible in polished sections. Alteration to martite and goethite can develop in a rough zonal pattern, although etching with HBr does not produce any marked texture. Small cavities sometimes occur in lines parallel to $\{110\}$ faces. The martite displays the typical lattice texture and pitted surface. Goethite preferentially replaces the magnetite and the martite is abviously much more resistent, as orientated lamellae of martite are frequently set in solid masses of goethite. Colloform goethite and haematite vein the earlier minerals, the veins often showing alternating parallel layers of goethite and haematite, and there is not infrequently a cavity along the centre of the vein.

Rare minute specks of pyrite, and even rarer chalcopyrite, have been seen in polished section.

Geochemistry

The iron ore is of good quality and the usual impurities are tolerable, although copper can be on the high side. The two analyses shown below for intersections of the ore body are for total split core samples, so all the clayey gangue is included. Simple washing will upgrade this ore, and the analyses for the detrital ore probably closely approximate to the composition of the final product. It can be seen that copper and zinc tend to be higher in the core samples than in the detrital ore, which indicates that some of the contamination is in the clayey gangue. The analyses were made in the company's laboratory.

—	Fe	P	S	Cu	Zn	Bi	As	SiO_2	Al_2O_3
1	63.0	.075	.055	.036	.045	tr	.005	2.4	1.2
2	41.3	.077	.043	.140	.093	.008	.008	-	-
3	64.3	.056	.036	.045	.020	.004	.059	-	-
4	56.2	.055	.040	.041	.043	tr	.024	-	-

1. 50 ft detrital ore overlying:- 2. 113 ft intersection of ore.
3. 15 ft detrital ore overlying:- 4. 95 ft intersection of ore.

Samples collected and analysed by the writer gave the following results (in ppm):

Sample No.	Cu	Zn	Pb	Bi	Ba
312	21	24	n.d.	n.d.	39
316	13	73	n.d.	n.d.	13
313	1139	205	<10	15	850
315	473	56	282	n.d.	113
314	342	60	n.d.	n.d.	<10

312 & 316.	Marble with no visible impurities.
313.	Clay - decomposed volcanic rock.
315.	Clay - probably decomposed volcanic rock.
314.	Solid iron ore.

These figures show a general low level for the impurity elements, except for sample No. 313 which contains some manganiferous material, and this has probably contributed towards the enhancement of the copper and zinc contents.

A company report assessing reserves gives the composition for the deposit as

Fe 61%; Cu 0.1%; Zn 0.13%; Bi 0.02%

but the figures for copper and zinc seem on the high side. The writer suspects that these figures do not allow for some of the copper and zinc being reduced when the clayey gangue is removed during treatment.

XVIII BUKIT SANAM AND OTHER MINOR DEPOSITS

Bt. Sanam

This is a small deposit which lies about one mile north of Bt. Ibam - Fig. 24. It was for many years thought to contain only secondary ore, but relics of magnetite have been found and diamond drilling located decomposed granodiorite at depth. There is a magnetic anomaly over the Bt. Sanam area. The country rock has completely decomposed to a clay and there are no outcrops in the vicinity, but company reports suggest that a small primary body may have formed at the contact of the granodiorite with calcareous rocks, and draw a comparison with the Bt. Hitam deposit. This suggestion may be correct, but there is no way of proving it at the present time.

The bulk of the ore is found as a surface and near-surface accumulation of boulders and fragments of secondary iron ore in a lateritic or bauxitic clay gangue. The ore is a mixture of limonite and haematite, and original Japanese records show that the composition of this ore is

Fe 54.3%; SiO_2 11.8%; Mn 0.99%; S 0.012%; P 0.023%; Cu 0.07%.

Other analyses show copper values up to 0.118%, and this high level suggests that some primary mineralization may well be present. A thin layer of bauxite which contains about 50% alumina is present at Bt. Sanam, but the silica and iron are high.

Other Minor Deposits

Throughout Ulu Rompin there are areas of superficial iron ore which has formed by a process of residual concentration. Some of these ores are extensive enough to be capable of exploitation, the most important being located at Bt. Pesagi and Bt. Sanlong (already dealt with in chapter XV), Bt. Merchong, Ulu Chepai and Bt. Guntong Raga - Fig. 24. All these deposits have a common mode of formation, but there are two main sources for the iron supply. In many areas the iron has been derived from minerals, notably disseminated pyrite, in the underlying bedrock, while in others the breakdown of small primary lenses of magnetite, haematite, and occasionally pyrite have supplied the iron. Both sources have contributed in some localities. The ores derived from the pyritic volcanic rocks which contain no primary iron oxides tend to be low grade and highly siliceous.

The most extensive areas of secondary ore are found on the relatively smooth and gentle hill tops and upper slopes of some ridges, but not on the steep hill slopes. This is partly due to the fact that lateritic iron ore is more easily formed on smooth undulating terrain, and partly that the present streams have possibly cut down through a more continuous pre-existing sheet. As the local geology and climate are the same throughout the area thepreferential development on smooth slopes is almost certainly connected with the position and oscillation of the water table. The latter will fluctuate through a greater vertical interval on the smooth slopes, so producing a more extensive zone of intense solution and precipitation. Some of the secondary ore forms very hard sheets, and is really a thickened iron pan. Where primary ore lenses were originally present in the bedrock the sheets of secondary ore contain fragments broken off from these primary bodies. Part of the secondary ore can be exploited as it stands, but some needs upgrading by blending with high grade primary ore. This process can be mutually beneficial as it usually reduces the level of deleterious impurities, such as copper, in the primary ore.

No analytical data are available for those deposits which contain only secondary ore, but reference to figures given in the chapter dealing with Bt. Sanlong (XV) will give a good idea of the type of ore formed over pyritic volcanic rocks. Any impurities in the secondary ore is a consequence of the composition of the source material.

XIX GENESIS OF THE ULU ROMPIN IRON ORE DEPOSITS

Introduction

Although all the primary ore bodies are almost certainly genetically related, it is convenient to consider them separately to bring out special features. Most of the chapter is devoted to the Bt. Ibam ore body, while for the other deposits points of difference are mentioned but common features are not considered at length. The source of the iron is examined in the following chapter which treats the Ulu Rompin deposits as one unit.

The Geological Environment Before Mineralization

Before the onset of metamorphic and metasomatic changes the geology of the Ulu Rompin mineralized zone consisted of a thick sequence of acid pyroclastic rocks and associated lavas, within which were a few lenses of carbonate. There were also rare bands of other sediments as arenaceous rocks have been identified in drill cores. The largest single carbonate mass occupied most of the space which is now filled by the Bt. Ibam ore body, but other thin lenses are now represented by calc silicate rocks and marble. In composition the carbonate rocks at Bt. Ibam were dolomites with some limestone bands. Elsewhere in Ulu Rompin calcareous rocks were definitely present at Bt. Batu Puteh as fairly pure calcite limestones, and very small lenses of dolomite occurred in the Bt. Pesagi area. No other carbonate rocks have been proved, but indirect evidence suggests that they may have been present at Bt. Hitam and Bt. Sanam.

The thick pile of volcanic rocks and sediments was tilted to the east, and at the same time a strong shearing stress produced a marked foliation in many of the volcanic rocks. Where calcareous rocks were present the carbonate responded to the stress field by plastic flow, so that the volcanic rocks immediately enclosing the carbonates are normally unsheared. In some areas the shearing stress was strong enough to produce fracture zones, the best example being found at Pesagi No. 6 where the intensity of the shearing increases markedly towards the ore body. The stresses in the volcanic rocks caused a rise in temperature which initiated mineralogical and textural changes; this phase can be correlated with the first recrystallization seen in the quartzite sample from Bt. Ibam - page 42. The changes in the volcanic rocks consisted of some breakdown of feldspars and recrystallization of the groundmass, with abundant sericite contributing to the foliation. The carbonate rocks probably recrystallized to form marbles, such as are found at Bt. Batu Puteh.

The Bukit Ibam Ore Body

Previous Interpretations

Speculation on the genesis of the Bt. Ibam deposit has continued since the early prospecting carried out by the Japanese, and as more detailed information became available various interpretations have been proposed. A major change in opinion can be dated from the onset of detailed prospecting in the 1960's; all the early workers ruled out replacement as a possible mechanism, but more recent and better data has shown that whatever the ultimate source of the iron, the ore minerals formed as a result of replacement reactions. The main objection to a replacement origin raised by the older workers was the apparent lack of material which was transitional between country rock and ore, but most of them thought that the porphyritic rocks only were important, and the skarns were either not recognized or else not allowed sufficient importance in the scheme or ore genesis. The early investigators postulated that the strongly sheared zones had acted as loci for the mineralization, but as has been shown in earlier chapters, the country rocks immediately surrounding the ore body are in fact only lightly sheared, if at all. This situation contrasts strongly with the Pesagi No. 6 ore body where the intensity of shearing increases towards the mineralization.

The ideas of the early Japanese mining company (Ishihara Sangyo Koshi) on the origin of the Bt. Ibam body have not been preserved in any records seen by the writer, but MacDonald (1970) implies that they did not interpret it any differently from the Bt. Pesagi and Bt. Sanlong bodies. These two bodies were thought to have had their origin in a segregation product from the acid magma which gave rise to the liparite (their identification) country rock. The Japanese were aware of the presence of granodiorite below the footwall of the Bt. Ibam body, and it seems unlikely that they did not bring this rock into their speculation on the origin of the ore.

Fitch (1941) rejected a replacement origin because of the lack of specimens intermediate between ore and country rock. He considered that "The most reasonable theory .... is that a body of quartz-porphyry was sheared and later invaded by epidote granite rich in migrant solutions which deposited their load in the overlying quartz-porphyry in locations governed by the shearing to form haematite-magnetite veins". Fitch could not enter any of the adits and had to rely on surface exposures and spoil heaps.

Macandie and Canavan (1948) briefly examined the deposit, and also had access to earlier records. They concluded that "The ore evidently has been introduced as magnetite in the form of veins, some of which remains unaltered, but most of which is altered to haematite or limonite".

Hitchen and Moss (1954) examined the Bt. Ibam body in their capacity as consultants to the Colonial Development Corporation, who had been approached as a source of development finance. Some inconclusive diamond drilling had been undertaken and a few old adits reopened by the time of their visit. They interpreted the ore as a segregation product from a basic magma which had also produced the serpentinous (their identification) and chloritic rocks. This idea is based on the premise that the serpentinous and chloritic rocks are magmatic in origin; there is no evidence that any basic magma has ever existed, and it is much more likely that the magnesia-rich rocks are the product of metamorphism.

Tillia, one of the earliest Rompin Mining Company geologists, reported in 1956 in similar terms to Hitchen and Moss. He wrote that "The origin of the ore is believed to have been the result of magmatic segregation in which solutions and gases carrying the hematitic ore penetrated a zone of weakness along a shear zone of the quartz porphyry. This action was quite likely associated with the basic peridotite and serpentine intrusion found at depth". This theory is open to the same objections as those which can be raised against that of Hitchen and Moss, and there is quite definitely no peridotite intrusion at depth. The writer believes that those geologists who refer to serpentine as an important mineral have mis-identified some of their material. While serpentine is undoubtedly present at Bt. Ibam, the writer detected it in small quantities only, even after the examination of numerous thin sections and diffractometer traces. It is highly probable that talc, which is extremely common, was often identified in hand specimen as serpentine. It is significant that neither Hitchen and Moss nor Tillia mention talc, which is definitely the most abundant gangue mineral.

MacDonald (1970) examined the Ulu Rompin area and proposed a radically different theory to explain the origin of the iron ore. He took note of the skarn rocks but did not think that they had been derived from a bed of limestone or dolomite because "There is no evidence, however, that any such bed of adequate size was present, nor does the ore body exhibit any typical internal replacement textures or structures". He thought that the skarns could have formed by lime metasomatism from a basic or ultrabasic magma, but he had previously stated elsewhere that there was no evidence that such a magma had existed. MacDonald was strongly influenced by Landergren's study of the Swedish iron ores (1948), and although there was little evidence with which to draw comparisons he concluded "The possibility is suggested therefore, that the Ulu Rompin deposits may be the result of partial granitization of original sedimentary horizons .....". MacDonald thought that granitization was not complete but had produced rocks "such as the so-called 'quartz-porphyry'". As a colleague of MacDonald for many years the writer knows that he was impressed by the similarities between the main country rock in the Ulu Rompin mineralized zone and the leptites of Central Sweden, a likeness which the writer's studies have confirmed in more detail, but there is no unanimously accepted theory to explain the genesis of the iron ores (Magnusson, 1970), and Landergren's ideas are not supported by all workers on the Swedish deposits. Ignoring the problem of what type of rock originally filled the space which the ore body now occupies, MacDonald's theory is based on the somewhat nebulous concept of partial granitization. The writer can find no evidence which supports this idea, and is convinced that the porphyritic rocks which MacDonald considers to be the product of partial granitization are normal acid lavas and tuffs which have suffered a considerable degree of alteration.

Since the late 1950's a great deal of work has been undertaken by company geologists, and the Bt. Ibam ore body is interpreted as a replacement mass which has been enriched by supergene activity. The company believes that the granodiorite was responsible for the mineralization, first converting a lens of carbonate rock to skarn and later partially replacing the skarn by magnetite. This original ore contained 30-40% Fe, but as it came within the zone of weathering some of the silicates were leached out leaving an enriched deposit. Whilst it is undoubtedly true that tropical weathering can cause profound changes in superficial rocks, and enrichment by selective leaching is a well attested process, the writer feels that it does not offer a full explanation of the evidence at Bt. Ibam.

The theory accepted by the company relates the genesis of the primary mineralization to the underlying granodiorite, but one geologist (Taylor) who worked on the mine for some years modified this aspect of the theory and has written "The orebody may have originated contemporaneously with the volcanic series as an exhalative - sedimentary deposit over an intrusive at shallow depth. Subsequent folding and fracturing with further rise of the granodiorite to close below the ore caused considerable thermal metamorphism of the foot wall rocks and shearing and recrystallization of the ore zone, with some redistribution of the copper and sulphur. The high grade ore was formed subsequently by supergene processes which upgraded the iron content by removal of magnesium, silicon and sulphur and oxidation of magnetite to haematite and pyrite to limonite". It can be seen that Taylor has in effect added the concept of a sedimentary origin for the iron to the official company theory, as he still has recourse to the granodiorite to explain the final form of the mineralization prior to the action of supergene alteration.

Background Details

It is worthwhile at the outset to list some pertinent facts which have relevance to any consideration of the genesis of the Bt. Ibam ore body and are incontrovertible, although their relative importance may be disputable; the list below is not in any suggested ore of importance.

1. The primary ore is virtually all magnetite, and what little primary haematite is present is later than the magnetite.
2. Most of the iron ore is intimately associated with magnesian silicates.
3. It is usually possible to demarcate precisely the limits of the ore body, which is found in one compact mass.
4. Granodiorite is present to the west of and below the ore body.
5. Calcic skarns occur below parts of the footwall, while massive chlorite is present along sections of the hangingwall.

One striking feature is the marked contrast between the chemistry of the silicate minerals in the ore zone and those in the normal country rocks, and an explanation of this fact must form an integral part of any theory of ore genesis. It can be postulated that the essential chemical features of the ore zone silicates were developed when the original rocks were formed, or alternatively an extensive transferrance of material can be invoked. The writer believes that the former is the correct interpretation, and does not accept that the present ore zone was originally occupied by acid volcanic rocks, which is a corollary of any hypothesis which suggests that the location of the ore was controlled by shear zones in the country rocks. Such a hypothesis demands the entry of large quantities of magnesia, lime, iron and manganese, and the removal of some silica and alumina and virtually all the potash and soda. An important objection to this hypothesis is that there is no apparent reason why one small, clearly defined zone should have suffered such a a profound chemical change compared with the surrounding volcanic rocks. The shape of the ore zone suggests that within the succession of volcanic rocks there was a lens of material which had a radically different composition. Although it is not possible to be absolutely certain of this composition, it is suggested that a mixture of argillaceous dolomite and limestone would require the addition of only a little iron, and possibly some silica, to be capable of producing the silicate minerals which are found in the skarn and gangue. To the writer, therefore, the prime control over the locus of mineralization has definitely been lithological, and it is probable that structural elements played very little or no part.

The composition of the lens within the volcanic rocks is an important question: both MacDonald and Taylor suggested that the iron which now forms the ore body was present in the original rocks, and that it recrystallized under the influence of granitization (MacDonald) or the intrusion of the granodiorite (Taylor). A syngenetic origin for the iron in the very much larger deposits found in central Sweden, and which have a number of features in common with the Bt. Ibam body, is held by a number of geologists including Magnusson (1970), an authority on these deposits. For the reasons given below a syngenetic origin for the Bt. Ibam iron ore is rejected:

1. There is no banding in the ore body such as might have been expected to be inherited from an original sedimentary ore.
2. There are no recognizable relics within the ore body of anything which can be even tentatively correlated with the primary source material.
3. Iron-rich sediments are typified by their large lateral extent.
4. Iron-bearing sediments are not rich in magnesia, which is abundant in the Bt. Ibam gangue minerals.

It is interesting that Magnusson uses the presence of the features described in (1) and (2) above as evidence for an original sedimentary genesis for the iron in the central Swedish ores. A corollary of (2) is that if the original iron was sedimentary then it has undergone a complete change to magnetite, some of it very coarse grained, which is a dubious proposition.

Neither MacDonald nor Taylor discuss in any detail the type of syngenetic iron deposit they envisage, but it could hardly have been one of the classic types of Iron-Formation e.g. Superior Type, Clinton Type, Minette Type or Algoma Type (Gross, 1965), as these are characterized by their very large areal extent, and the metamorphism of an Iron-Formation does not result in a deposit of the type seen at Bt. Ibam. If the iron at Bt. Ibam was deposited as a sediment then it accumulated over a very restricted area, and could just conceivably have been due to volcanic exhalations, but it is difficult to believe that the iron would not have been dispersed over a much wider zone. It is also difficult to explain why at Bt. Ibam the iron should have been changed to magnetite and at Bt. Pesagi and Bt. Sanlong to haematite, as the localities are very close together.

It is easier to explain the form and composition of the ore body by postulating that the iron has been introduced by metasomatism into metamorphosed carbonate rocks which contained both magnesium and calcium. It has to be admitted that this hypothesis, like all the others, still leaves some problems unresolved, but it offers a better explanation of the evidence.

Source of the Magnesia

The invariable association of magnetite with magnesian minerals suggests two possible causes; either the magnetite has been preferentially emplaced in pre-existing magnesian silicates because they proved to be a favourable medium for replacement, or the magnesia was introduced at the same time as the iron with the result that the two are always in close association. From what has been stated earlier in this chapter it is apparent that the writer favours a replacement origin, but it is necessary to consider some aspects of the other alternative. The fundamental question is how much magnesia was in the rocks before the deposition of the iron. If the original lens was largely dolomitic then it probably contained sufficient magnesia to account for what is now present in the gangue minerals. It is possible to make a rough calculation which bears on this point, by determining how much magnesia is present in a specified volume of the main minerals involved.

Actinolite

$Ca_2(Mg_3Fe_2)(Si_4O_{11})_2(OH)_2$

(874gm 282cc) contain 72gm Mg

∴ 100cc contain 25gm Mg

Chlorite

$Mg_5Al_2Si_3O_{10}(OH)_8$

(554gm 205cc) contain 120gm Mg

∴ 100cc contain 58gm Mg

Dolomite

$CaMg(CO_3)_2$

(184gm 65cc) contain 24gm Mg

∴ 100cc contain 37gm Mg

It is impossible to be certain what were the relative proportions of actinolite and chlorite in the gangue before the iron mineralization, but actinolite was more abundant than chlorite, and if one assumes that a mixture of actinolite and chlorite in the ratio of 2 : 1 replaced the same volume of dolomite the following figures are obtained.

	200cc actinolite	contain	50gm Mg
	100cc chlorite	contain	58gm Mg
∴	300cc mixture	contain	108gm Mg
	300cc dolomite	contain	111gm Mg

These calculations are obviously only rough approximations because of the number of indeterminate factors involved, but they show that dolomite by itself contains sufficient magnesia to form the new minerals, and magnesia metasomatism is not an essential process. The larger the proportion of actinolite in the gangue the easier would it be for dolomite to provide the necessary magnesia.

There is no visible evidence of magnesia metasomatism in the country rocks, as might possibly be expected following the passage of magnesia-rich solutions, but the same argument could also be applied to the iron which formed the magnetite. The passage of iron through the rocks at a later stage was marked by the widespread development of pyrite. Although there is no megascopic evidence of magnesia metasomatism, the chemical data could be interpreted as supporting such a hypothesis, although it is far from conclusive. The footwall rocks are slightly higher in MgO (2.68%) than the hangingwall rocks (1.19%), and both contain more magnesia than the country rock round Bt. Pesagi and Bt. Sanlong (0.63%) where there are no magnesian silicates associated with the iron ore. Although it is possible to use these figures to support a hypothesis of magnesia metasomatism, they could equally well indicate slight differences in original composition. It has been shown in chapter III that there are some differences in the whole rock chemistry between the rocks of the footwall and hangingwall at Bt. Ibam.

One group of iron ore deposits where magnesia metasomatism is widely invoked is in central Sweden (Geijer & Magnusson, 1952), and here the leptites contain minerals such as cordierite and gedrite which are thought to have formed as a result of the passage of magnesian solutions. At Bt. Ibam the only mineral which could have played a similar role is chlorite, but it is just as likely to have originated by internal mineralogical changes during metamorphism, or by "leakage" from the ore body during the iron mineralization.

After considering all the evidence it is concluded that the magnesia was present in an original lens of dolomite, and it did not accompany iron during the process of mineralization. To invoke magnesia metasomatism introduces an unnecessary complication, as the composition of the gangue rocks can be readily explained by straightforward mineralogical reactions. It is also a pertinent fact that the largest masses of magnesian rocks preserved at Bt. Ibam - the chlorite rocks along parts of the hangingwall contact - are virtually devoid of magnetite, a most unlikely situation if the iron and the magnesia were contemporaneous and genetically related.

The presence of these massive chlorite rocks provides an interesting problem, and although they probably formed as a consequence of the magnetite mineralization they can be conveniently considered further at this point. The location of the massive chlorite between the ore body and the volcanic rocks suggests to the writer that as the iron mineralization progressed magnesia was driven out towards the hangingwall, where, on encountering the volcanic rocks, it reacted to form chlorite. Some movement of magnesia obviously must have taken place, as there would not be sufficient in dolomite alone to form chlorite in a volume for volume replacement. Although this idea formulated itself from purely spatial considerations while studying the ore body in the field, it is interesting to find that an example of the type of reaction which could explain the formation of the chlorite has been proved in the Salton Sea geothermal area in California. Muffler and White (1968) have shown that under a temperature gradient beginning in the 150° to 200°C range, there is an inverse relationship between the amounts of dolomite + kaolinite and chlorite, which they ascribe to reaction between dolomite and kaolinite to form chlorite. At Bt. Ibam it is possible that the magnesia driven out of the actinolite-chlorite rock during the mineralization reacted with alumina and silica in the volcanic rocks to form more chlorite. It is also of interest to note that in the Salton Sea area dolomite reacts to form chlorite, but where calcite is present epidote is produced. This is somewhat analagous to the situation at Bt. Ibam, as the calcic rocks below the footwall are rich in epidote, but this mineral is extremely rare in the gangue of the main ore body.

Formation of the Skarns and Magnesian Rocks and Facies of Metamorphism

In response to rising temperatures caused by the intrusion of the granodiorite the argillaceous carbonate rocks became involved in mineralogical changes which, with the addition of some iron and silica, resulted in the formation of epidote, andradite, chlorite and actinolite, with a little mica, olivine, pyroxene and possibly anthophyllite. There was probably a small amount of carbonate left after the formation of the main mass of silicate minerals. The outcome of the mineralogical changes was to produce a rock unit with three distinct sections.

Upper	A thick sequence of magnesian rocks with abundant actinolite and chlorite. The space originally occupied by these rocks is where the main ore body is now located.
Middle	A very thin band of magnesian-calcic rocks, which is now located immediately below the footwall contact. This band contains a little disseminated magnetite.
Lower	A variable thickness of calcic skarn characterized by abundant epidote and andradite. There is no magnetite in these rocks, but they contain a little micaceous haematite (and pyrite).

This layered structure strongly suggests that the original carbonate lens was predominantly dolomitic in the upper part, with a thinner limestone band at the base.

It is unlikely that equilibrium conditions were always fully attained during the formation of the rocks, but the typical assemblage in the calcic skarns is epidote-andradite, in the thin calcic-magnesian band it is epidote-actinolite-chlorite, and the space now occupied by the ore body originally consisted very largely of chlorite-actinolite. The latter two assemblages belong to the albite-epidote-hornfels facies, which is shown in the triangular ACF diagram in Fig. 126. According to Turner and Verhoogen (1960) this facies is typically developed at the outer margins of zoned contact aureoles, and with proximity to the intrusion passes into the hornblende-hornfels facies. These authors further state that it is rare to find the albite-epidote-hornfels facies developed without some associated assemblages of higher grade, but it can happen in weakly developed aureoles. The albite-epidote-hornfels facies is developed in conditions of relatively low temperatures and pressures, and is the equivalent of the greenschist facies of regional metamorphism.

The assemblage epidote-andradite is ubiquitous in the calcic skarns, but its status in the facies of contact metamorphism seems to be in some doubt. F.J.Turner has apparently changed his views between the publication of the first and second editions of the

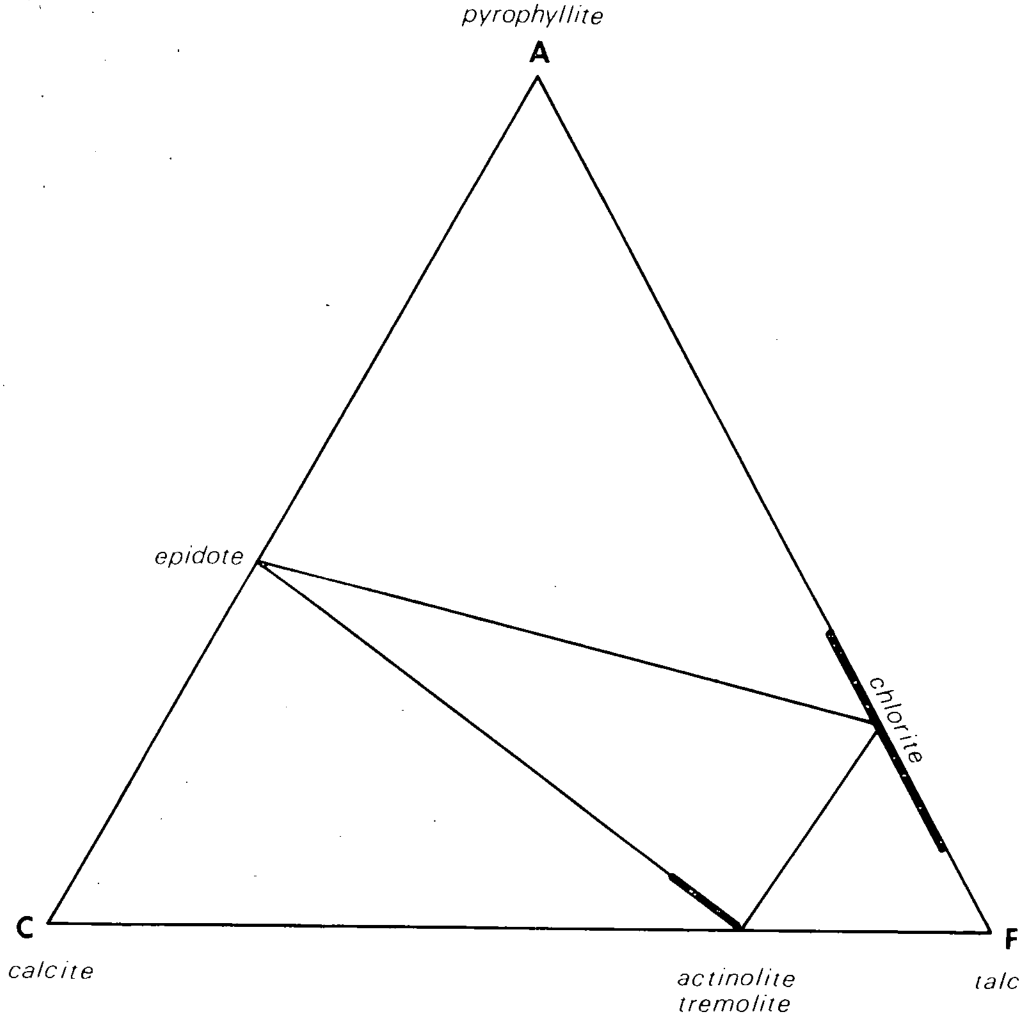


Fig.126 **ACF** *diagram for the albite — epidote — hornfels facies.*

well known textbook "Igneous and Metamorphic Petrology" (Turner & Verhoogen, 1951 and 1960). In the first edition he refers to grossularite associated with epidote in the actinolite-epidote-hornfels subfacies of the albite-epidote-amphibolite facies: these facies and subfacies were later incorporated into the albite-epidote-hornfels facies. In the second edition of the book garnets are not shown in the lower facies, but appear in the hornblende-hornfels facies. This is supported by Winkler (1967) who says "Grossularite/andradite garnet is stabilized at the beginning of the hornblende-hornfels facies", and he quotes a reversible reaction involving garnet and epidote.

$$\text{epidote} + \text{calcite} + \text{quartz} \rightleftharpoons \text{grossularite/andradite} + H_2O$$

It has been shown in chapter IV that petrographic evidence demonstrates the development of garnet from epidote and vice versa. Turner (1968) quotes a Japanese example in which epidote present in the lowest grade of contact metamorphism is replaced by grossularite in the next higher facies. It would appear therefore, that while epidote and andradite can coexist,the assemblage is probably maintained over a fairly narrow range of P-T conditions and marks the approximate meeting point of the albite-epidote-hornfels facies and the hornblende-hornfels facies. It is also highly probable that the original composition of the sediments had a marked influence, and in particular the presence or absence of alumina. Those skarn samples which consist of alternating monomineralic bands of garnet and epidote must have been subject to a compositional control. Garnets are stable over a wide range of temperatures, and andradite has been synthesized under conditions which in nature would produce rocks in the albite-epidote-hornfels facies, i.e. up to 500°C and at about 500 bars (Christophe-Michel-Levy, 1956).

Further evidence which suggests that low grade metamorphic conditions were operative is the almost total lack of pyroxenes. The Bt. Ibam skarns differ from the classical type of iron-rich skarn as they lack hedenbergite, and diopside is very rare. Abundant actinolite was available, and in the presence of calcium carbonate and silica and in response to rising temperature this mineral breaks down to form hedenbergite (Deer, Howie & Zussman vol.2, 1963). The absence of hedenbergite in the skarns is almost certainly due to the fact that the temperatures did not reach a sufficiently high level, as the chemical conditions were apparently favourable. According to Kranck (1961), hedenbergite can be unstable in the presence of excess carbon dioxide, and instability for this reason may also have contributed to the absence of hedenbergite in the Bt. Ibam rocks. In a description of the contact metasomatic magnetite deposits of British Colombia, Canada, Sangster (1969) discusses the temperatures of formation of various skarn silicates and comes to the conclusion that while pyroxene probably formed in the temperature range 700° to 800°C, epidote had an upper stability limit of 480°C. The Bt. Ibam deposit is of a similar type, and the abundance of epidote and the paucity of pyroxene is suggestive of relatively low temperatures. The presence of some actinolite and chlorite in the skarns is also indicative of low grade metamorphism.

The major part of what was the original carbonate lens is now occupied by the ore body, and the gangue lacks the two minerals typical of the calcic skarn, andradite and epidote. The dolomitic rocks have been replaced by actinolite and chlorite, and possibly some talc, although it is impossible to determine precisely how much of the latter mineral is of metamorphic origin: most of the talc now found in the ore body has undoubtedly formed by alteration of the other magnesian minerals, which have frequently been pseudomorphed. It is probable, however, that talc was one of the very first minerals to form in response to the rising temperature, but this original talc was replaced by other minerals as the metamorphism progressed. It is known that talc is the first mineral to form in the metamorphism of a siliceous dolomite provided that the partial pressure of water is high enough (Winkler, 1967). There is evidence that there has been a selective reaction involved in some of the chemical changes. At the extreme northern end of the ore body, where the iron mineralization is reduced to very small dimensions, it is not uncommon to find a gangue which consists of calcite and chlorite: the same association is found on a small scale elsewhere in the ore body, and also in some veins. The writer believes that the magnesia in the original dolomite has been more reactive and formed chlorite, while at the same time some of the lime has remained in the carbonate form. It is known that the thermal decmposition of dolomite begins with the preferential breakdown of the magnesium carbonate component

(Bowen, 1940; Goldsmith, 1959).

A possible influence on the skarn and magnesian rocks which must be mentioned is that of retrograde metamorphism. It would be possible for most of the minerals now present to have developed from pre-existing minerals of a higher metamorphic grade, but the writer can find no evidence that the effects of retrograde metamorphism have been widespread, apart from the production of talc. One would expect to find numerous relics of the higher grade minerals if they had been developed in any quantity, but these are not present. A later section discusses the probable range of temperatures during metamorphism and metasomatism, and this evidence indicates that the minerals found in the skarns and gangue are representative of the highest grade attained, and they have not been produced in response to falling temperatures.

The whole of the mineralogical evidence indicates that the metamorphic conditions were of albite-epidote-hornfels grade, and occasionally reached the lowest levels of the hornblende-hornfels grade.

Iron Oxide Mineralization

The introduction of large quantities of iron led to the widespread replacement of the existing magnesian silicate minerals by magnetite. The results of the polished section studies described earlier have shown that the ore was deposited by replacement, and where the age relationship between the iron ore and silicates can be determined the iron ore is usually later. No magnetite has been encountered by the writer, or apparently recorded by mine geologists, in the acid volcanic rocks or in the calcic skarn, but there is a little disseminated magnetite immediately below the footwall contact where the skarns are calc-magnesian. The presence of magnetite crystals with a complex zonal structure, plus the evidence of completely unzoned crystals among zoned ones, proves that there have been various phases in the mineralization. There is no discernible overall pattern to the distribution of the iron mineralization within the confines of the ore body, but there are rare examples of banded ore; in these ores the banding is on an extremely delicate and fine scale. The mineralization has typically produced an ore which is a fine grained mosaic of interlocking grains, but larger crystals do occur, both singly and in aggregates. There is no evidence that any other metallic minerals have accompanied the magnetite, but there are occasional signs of some local reaction with the gangue silicates.

The Bt. Ibam deposit is noteworthy for the persistence of magnetite until the very last stages of iron oxide mineralization, when a little haematite formed; once haematite started to crystallize there was no further magnetite mineralization. This is in marked contrast to the mineralization at Bt. Pesagi and Bt. Sanlong where the initial magnetite rapidly gave way to haematite, which is the predominant mineral. In the calcic skarns at Bt. Ibam there are some small patches of micaceous haematite, but the granular type of haematite is found in close association with magnetite ore, which it has often replaced, and the formation of the haematite indicates that there was a significant change in the conditions of mineralization. One possible cause was a drop in temperature, which could have promoted the change from magnetite to haematite crystallization; magnetite is often considered to be a high temperature mineral in comparison with haematite (Freitsch, 1967). It seems unlikely, however, that temperature was the sole factor controlling the change from magnetite to haematite deposition. The quantity of haematite is extremely small and it is found in a few localities only, which shows that the conditions favourable to haematite formation were of very short duration and developed in restricted areas, but the fall in temperature is likely to have been a fairly slow and gradual process which affected the whole area equally. A possible exception to this generalization is discussed later. There are geologists, particularly from the U.S.S.R., who do in fact relate the parageneses in hydrothermal ore deposits solely to falling temperatures, but the writer feels that such a mechanism must be subjected to modification by local conditions, otherwise all deposits from a similar genetic source should show the same invariable paragenesis, which is not the case.

It would appear that in addition to a fall in temperature there were other contributory factors which had a rapid effect on the conditions of crystallization, and hence the type of mineral deposited. The change from the crystallization of magnetite to that of haematite indicates the development of a higher degree of oxidation in the system, which can be described in terms of the partial pressure of oxygen. Some writers use fugacity, but at low pressure this is virtually identical to the partial pressure as the gases behave as perfect gases (Krauskopf, 1969). In a paper which discusses the relationship between magnetite and haematite Freitsch (1967) says, "The important factors determining whether magnetite or haematite will be formed in magmatic and metamorphic rocks are oxygen fugacity and temperature .....The role of pressure is .... not known".

Holland (1959) constructed a graph which shows the stability relationships in the Fe-O system at various temperatures and oxygen fugacities, and this is reproduced at Fig. 127. A drop in temperature at a constant oxygen fugacity or a rise in the fugacity at a constant temperature can produce the same result, a move from the stability field of magnetite to that of haematite. The graph confirms that a fall in temperature could control the change from magnetite to haematite crystallization, but it also shows that a simple drop in temperature does not readily explain the abrupt cessation in the deposition of haematite, as this mineral is stable over a wide range of temperatures and oxygen fugacities. Possible temperatures during ore deposition are discussed later, but if 500°C is taken as the approximate starting temperature of magnetite crystallization then the oxygen fugacities at this point are fixed as 10^{-20}atm maximum and 10^{-28}atm minimum. Krauskopf discusses oxygen fugacities in the same system but at 600°C, and he deduced $10^{-15.2}$atm as a maximum and 10^{-24}atm as a minimum. He further states that it is doubtful if O_2 pressures could approach so low a figure and says "a more reasonable lower limit for most magmatic gases would be about 10^{-21}atm". Lower pressures would obtain at 500°C, and at Bt. Ibam the lowest possible oxygen pressure was probably about 10^{-24}atm, but as at this low oxygen pressure magnetite is stable down to 380°C and it is highly probable that the haematite formed above this temperature, the evidence suggests that the lowest oxygen pressure was above 10^{-24}atm.

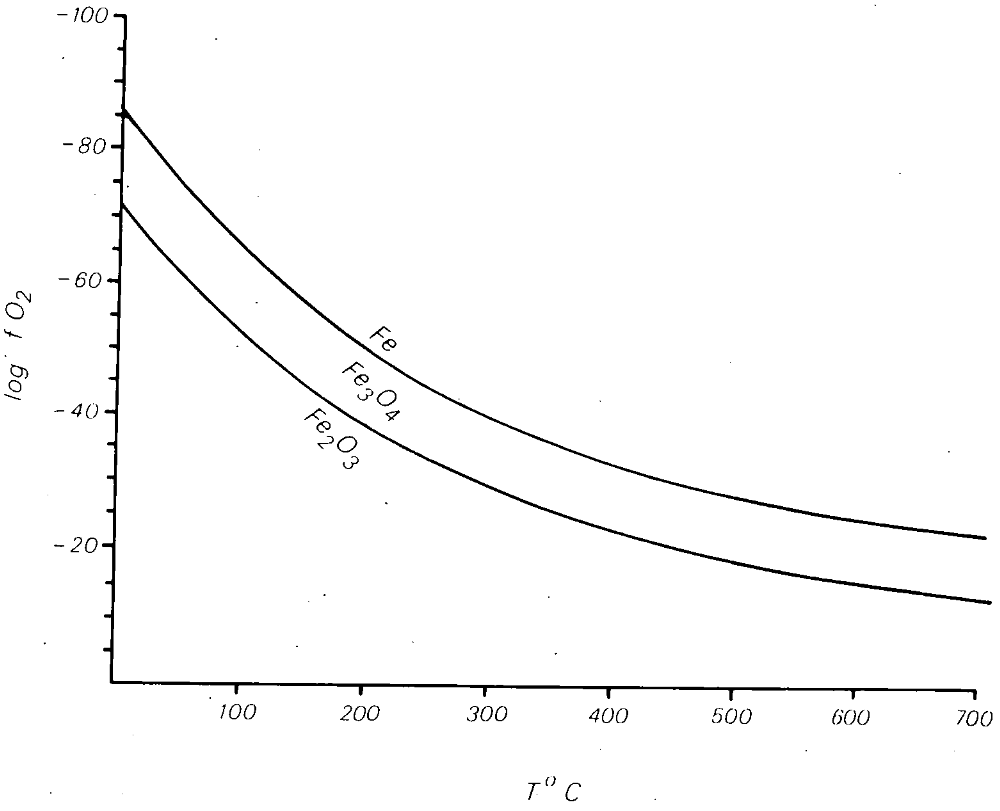


Fig.127 Stability fields of iron and iron oxides in the Fe - O system.

(after Holland 1959)

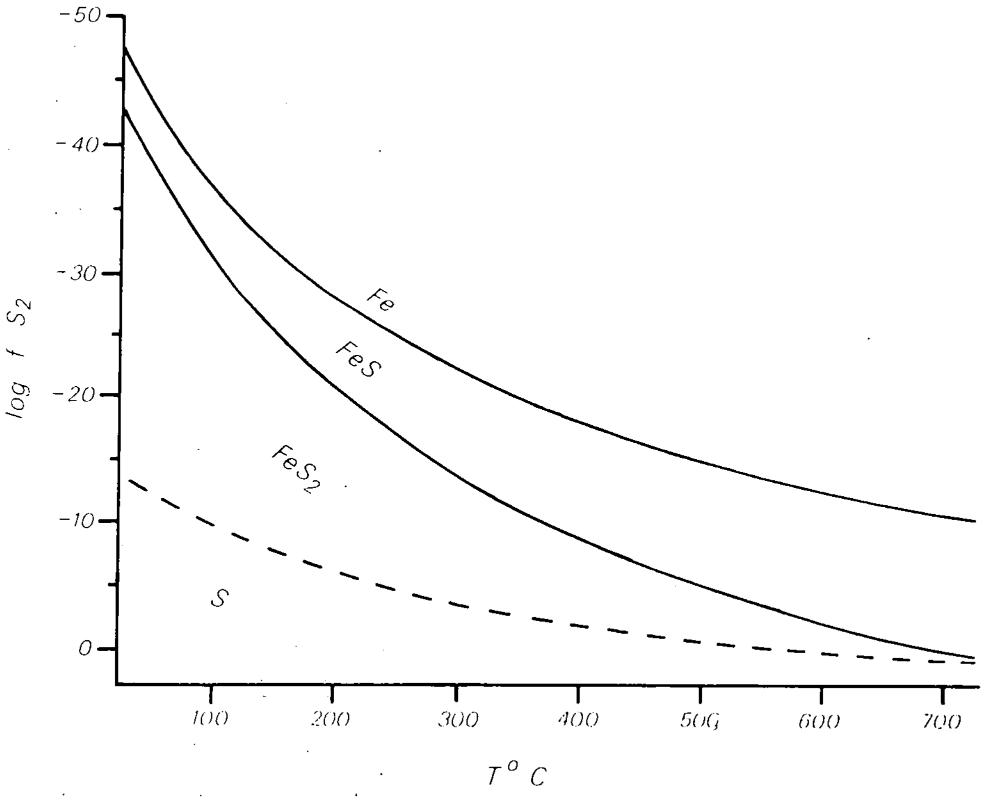


Fig.128 Stability field of iron and iron sulphides in the Fe-S system.

(after Holland 1959)

The precise action of any reducing agents within the system is not fully understood, but they would obviously help to promote the formation of magnetite rather than haematite. At Bt. Ibam sulphides were absent, but ferrous iron was present in actinolite, with probably a little in chlorite, and it may have exerted some influence over the reactions. Mueller (1961) has demonstrated that in magmatic and metamorphic rocks any ferromagnesian minerals which coexist with haematite are poor in ferrous iron, whereas those which are stable with magnetite are ferrous-rich. It appears that any increase in oxygen pressure which would potentially favour the formation of haematite rather than magnetite promotes reactions involving ferrous iron in the silicates, with the result that the oxygen pressure is held down and haematite does not form as might have been expected. The overall effect of ferrous iron is summed up by Freitsch (1967) as "This means that the oxygen available to convert magnetite to haematite in mineral assemblages is contolled by the amount of ferrous silicates and that they act as reducing agents in this connection". It is possible to suggest, therefore, that at Bt. Ibam the presence of ferrous iron contributed to the scarcity of haematite by holding down the oxygen fugacity.

A factor which may have been of major importance was the acidity or alkalinity of the solutions depositing the iron minerals. Mason (1943) has stated "Alkaline solutions therefore, even without the presence of atmospheric oxygen, probably have an oxidizing effect on magnetite, whereas acid or neutral solutions do not have this effect", and Korshinsky (1964), who has made a long and detailed study of metasomatic processes and deposits, believes that the acidity-alkalinity of the ore-forming fluids is a vitally important factor in oxidation-reduction reactions. The higher the acidity of the solutions the lower is the activity of oxygen, and vice versa. Korshinsky relates this to the dissociation of water under deep seated conditions.

$$2H_2O \rightleftharpoons 4H^+ + O_2 + 4e^-$$

from which is derived

$$\left[H^+\right]^4 \left[O_2\right]\left[e^-\right]^4 = K_T\left[H_2O\right]^2$$

where K_T is a constant at temperature T^o. In aqueous solutions the activity of H_2O varies very little, and Korshinsky believes that for a given fluid flow, $\left[e^{-}\right]$ is also held fairly constant. The effective result is that

$$\left[H^{+}\right]^{4}\left[O_{2}\right] = \text{Constant}$$

i.e. oxygen activity is inversely related to the hydrogen ion activity. Many geologists from the U.S.S.R. place great emphasis on the acidity-alkalinity of ore solutions, which they relate primarily to falling temperature, but also invoke other factors. Ovchinnikov (1968) says ".... one should suggest that probably this change of the acidity regime is connected with the decrease of temperature..." and "The significant raising of solution alkalinity probably coincides with the destruction of mass of complexes that simultaneously brings about the deposition of the ore parageneses, too": Ovchinnikov is referring to large complex ions which he believes enhance the acidity of the solutions.

It is suggested that the relation between the acidity of the mineralizing fluids and the oxygen activity offers an explanation of the genesis of the types of iron ore found in Ulu Rompin. At Bt. Ibam and Bt. Batu Puteh the only important primary iron oxide is magnetite, whereas at Bt. Pesagi and Bt. Sanlong it is haematite. The other obvious and important geological feature which differentiates these pairs of deposits is the original presence of carbonate rocks at one and their absence at the other, and there may have been a connection between the presence of the carbonate rocks and the deposition of magnetite. Such a relationship has been proposed by other workers; e.g. Eastwood (1965) described the replacement bodies of magnetite on Vancouver Island and said, "It may be that limestone is in some way necessary in the deposition of magnetite ..." but he does not elaborate on this statement.

The composition of most mineralizing fluids is complex, but it is widely believed that post-magmatic fluids are initially acid in composition (e.g. Korshinsky, 1964; Burnham, 1967; Sangster, 1969), and there is no reason to suppose that the fluids from the Bt. Ibam granodiorite were not acidic. Consequently the solutions possessed a low oxygen activity, and magnetite formed in preference to haematite; this applied to the early stages of all the primary deposits in Ulu Rompin. At Bt. Ibam the solutions appear to have maintained their acidity over a long period of time, and it is suggested that this may in some way have been related to the presence of carbonate rocks. Reaction with a carbonate would normally rapidly neutralize an acid solution, but as described earlier, by the time that the iron mineralization commenced most of the carbonate had been converted into calc and magnesian silicates. The carbon dioxide which was released during these reactions could have been retained within the system initially, provided that the overlying cover rock was sufficiently impervious, and it is this which helped to maintain the acidity of the solutions (probably as some form or complex of carbonic acid). The evidence provided by the alteration of the silicate gangue minerals also suggests the presence of carbon dioxide, as will be described later. It would be quite possible to sustain large fugacities of carbon dioxide without affecting the nature of the minerals produced. Holland (1959) has shown that even at such a modest temperature as $600^{o}K$ ($327^{o}C$) a fugacity of over 100atm is required before siderite forms a stable phase in the system Fe-O-S-C, and at higher temperatures even larger fugacities would be needed.

The sudden and total cessation of magnetite formation may have been due, at least in part, to the development of a fracture system which permitted the carbon dioxide and other volatiles to escape rapidly from the system. This would have contributed to the solutions losing their acid composition, with a consequent rise in oxygen activity. Any sudden development of an open fracture system would also help to promote a more rapid dispersion of heat from the zone, and there may have been a sharp, even if slight, drop in temperature; Fig. 127 shows that if this had any effect it would be to favour the formation of haematite at the expense of magnetite. Such a process could explain why the very small amount of primary haematite came right at the end of the oxide mineralization.

The same reasoning will hold for the other primary ore bodies. At Pesagi No. 6 and Sanlong B, where carbonate rocks were absent, magnetite very quickly gave way to haematite, but at Bt. Batu Puteh where marble is still preserved the only primary ore is magnetite. The only other primary body is at Bt. Hitam, but so little is known about the geology of the country rock that it is not possible to comment on any influence it may have exerted. It is not suggested that the presence or absence of carbonates was the only factor involved in determining whether magnetite or haematite was deposited, but it may have made an important contribution. There is no reason to suppose that the fluids which formed the main primary ore bodies differed significantly in composition, which means that any variation in the type of minerals produced must have been due to a control at the site of ore emplacement. It seems to be more than coincidental that where carbonates are known to have been abundant the iron oxide is magnetite, and where absent the ore is haematite.

The final phase of iron oxide mineralization was marked by an increase in the activity of oxygen, but this condition was maintained for a brief period only as the amount of primary haematite is extremely small. The increase in oxygen activity may have been connected with an increase in the alkalinity of the fluids, and a concomitant drop in temperature could have contributed. Even though there is very little primary haematite it is present in two forms, finely granular and fairly coarse micaceous, and the fact that these two types are never found together - the micaceous ore is only found in the calcic skarns - suggests some local form of control. As the micaceous ore is typically found with pyrite and quartz in irregular masses, it is possible that this type of haematite is not related to the main oxide phase of mineralization, but formed from quite different reactions.

Park and MacDiarmid (1970),following Kullerud, quote some interesting equations which show that in the presence of sulphur a silicate mineral (hedenbergite in the example quoted) can break down to form magnetite or haematite depending on the ratio of sulphur to silicate, plus pyrite, quartz and wollastonite. It is interesting that apart from wollastonite this is precisely the assemblage found in the calcic skarns. It is possible that similar reactions based on andradite, which is not unlike hedenbergite in composition, produced the micaceous haematite at Bt. Ibam. It is also possible that the micaceous haematite and pyrite crystallized together, or nearly so, and found the calcic rocks a receptive host. Holland (1959) has shown that haematite and pyrite are a stable pair, and they could form simultaneously if the fugacities of oxygen and sulphur were correctly balanced. Viewed in this light, the mineralized masses in the calcic skarns could be considered an intermediate stage between the periods of high oxygen fugacity and high sulphur fugacity.

The granular haematite is always found in magnetite ore, and has often replaced it. No other minerals are involved, and it seems that there has simply been a rise in the oxygen activity which has promoted the change from magnetite to haematite.

Sulphide Mineralization

The increase in oxygen fugacity was short-lived, and was replaced by a marked rise in the fugacity of sulphur which initiated the phase of sulphide mineralization. A number of sulphide and sulphosalt minerals were deposited in the ore body, but only pyrite formed in any quantity. There is an overall sequence for the formation of the various sulphide minerals which is apparently invariable, but the finer details cannot always be determined. Once the main pyrite mineralization commenced there is no evidence that any further iron oxide minerals were deposited, and the sulphide and oxide minerals belong to quite distinct episodes. This indicates that there was a marked change in the conditions of mineralization, as phase diagrams compiled by Kullerud (1957) and fugacity diagrams of Holland (1959) show that iron oxides and sulphides can coexist, except for the pair haematite-pyrrhotite. For all practical purposes pyrrhotite is absent from the Bt. Ibam ore body, which places it in a minority type of magnetite deposit. Schwartz and Ronbeck (1940) studied data on 130 deposits which contained magnetite and sulphides, and of these, six contained pyrrhotite but no pyrite, 77 had both pyrrhotite and pyrite, while 43 contained pyrite but no pyrrhotite. From a study of this data Kullerud (1959) concluded that the partial pressure of oxygen during sulphide formation is very low, usually below 10^{-25}atm and never exceeding 10^{-14}atm (up to 600°C). The partial pressure of sulphur on the other hand must have been much higher, and as pyrite is the sulphide to the virtual exclusion of pyrrhotite, it is possible to make some deductions regarding the sulphur fugacity during mineralization. Fig. 128 is taken from Holland (1959), and shows the stability fields of iron and iron sulphides in the Fe-S system. Possible temperatures during mineralization are discussed later, but if the range 400°C to 500°C is taken as covering the crystallization of most of the pyrite, it can be determined that the sulphur fugacity had maxima varying from 10^{-2} to 10^{-3}atm and minima between 10^{-6} and 10^{-9}atm. Higher pressures would have allowed sulphure to condense, while lower pressures would have resulted in the formation of pyrrhotite.

There has been very little interaction between the sulphides and the iron oxide minerals, but a few examples have been noted in polished sections. The most interesting is the reduction of haematite to magnetite, even though it is only on a microscopic scale. Pyrite has very occasionally replaced magnetite and haematite, but has more often filled cavities and fractures. Most of the other sulphide minerals have replaced magnetite, but it is a rare phenomenon.

There is no doubt that apart from minute quantities of molybdenite, pyrite was the first sulphide to crystallize. It has been deposited not only in the ore body and skarns but in the volcanic rocks, typically along joints, but also in the mass of the rock where it clearly post-dates some of the secondary muscovite. The pyrite has proved to be an important loci for replacement by later sulphides. The paragenetic sequence for the full suite of sulphide minerals is not certain, but Fig. 129 shows a synthesis of the evidence obtained from polished sections. Gold is clearly later than pyrite and chalcopyrite, but its relation to the other sulphides is unknown. The bismuth minerals seem to occur earlier in the sequence than is considered typical, but while bismuthinite is definitely later than pyrite its relation to the other sulphides is not fully proved. The little pyrrhotite in the deposit is later than pyrite but otherwise unplaced, and arsenopyrite has not been observed in contact with other sulphides.

The completion of the sulphide mineralization marked the end of the main primary ore genesis at Bt. Ibam.

Transfer of Materials

Throughout the period of formation of the skarns and the oxide and sulphide mineralization there was an extensive movement of

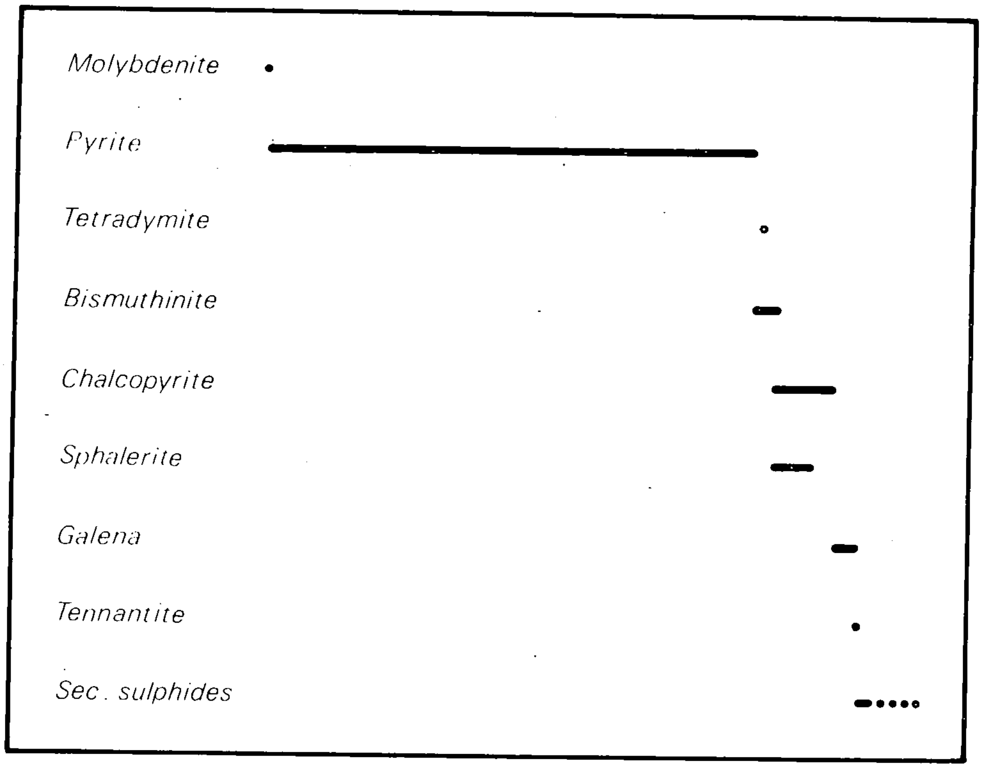


Fig. 129 Paragenetic sequence of the sulphides in the Bt. Ibam ore body. The lengths of the lines are roughly proportional to the abundances.

materials, and as stated earlier, the direction of "flow" must have been from the footwall side towards the hangingwall side. The elements introduced into the skarns and ore zone in important quantities were iron and silica and possible a little alumina, and the main components removed were lime and carbon dioxide. The passage of the fluids which carried these materials affected the enclosing volcanic rocks, the most widespread reaction being the formation of sericite. Chemical analyses show that most of the volcanic rocks now contain a high alumina content, which is probably due to the removal of some lime and alkalies. The passage of the silica and iron related to the oxide phase of mineralization left no enhancement of these elements in the volcanic rocks. There is a small number of examples of volcanic rock from immediately beneath the skarns which show higher than average values of lime and magnesia, and it is probable that these components were derived from the overlying ore zone during mineralization. Some of the original lime and carbon dioxide expelled from the dolomite may be represented in the calcite veins found throughout the area, but most has been completely removed.

Physical Conditions During Metamorphism and Metasomatism

As the writer relates the processes of metamorphism and metasomatism to the intrusion of the granodiorite it is necessary to make some estimate of the P-T conditions during the emplacement of the magma and subsequent mineralization.
As was stated in chapter V, the Ulu Rompin granodiorites are high level intrusions and can be placed in the Epizone of Buddington (1959), which extends from the surface to about six miles, with the normal level of intrusion being about four miles. Schneiderhohn (1961) estimated that the most frequent depth of granitic intrusions at the time of solidification was three to eight kilometres i.e. two to five miles, but Ovchinnikov (1968) asserts that ore-forming fluids associated with granitic rocks must have originated within five kilometres of the surface, as below this depth "the rocks greatly change their physical and chemical properties" and "become impermeable to solutions". There is no evidence at Bt. Ibam which allows the original depth of the intrusion to be determined precisely, largely due to poor exposure which makes it impossible to reconstruct the palaeostratigraphy and structure, but from the evidence which is available,

and a comparison with other Malayan intrusives, it is safe to assume that the granodiorite was not emplaced very deep below the surface. Buddington's average of four miles is probably in excess of the depth of the Ulu Rompin intrusions, and this figure can be taken as an extreme lower limit. Load pressure increases at a rate of approximately 250 to 300 bars/km (400 to 480 bars/mile) depending on the type of rocks involved, which means that the lithostatic pressure at a depth of four miles cannot have been more than 1920 bars, and was probably less. The nature of the overlying rocks, plus the fact that the granodiorite was probably emplaced less than four miles from the surface, leads the writer to the conclusion that the lithostatic or load pressure was within the range 1000 to 1500 bars, with 2000 bars as the extreme limit. The pressure conditions within the carbonate rocks would be the same.

Having made an estimate of the confining pressure it is then possible to make some predictions regarding the temperature of the magma during its emplacement. Fig. 130 shows a pressure-temperature graph, the basis of which has been taken from Kullerud (1959). He combined the beginning-of-melting curve of a natural tholeitic basalt as determined by Yoder and Tilley (1956), with the minimum melting curve of granite constructed by Tuttle and Bowen (1953). The granite depicted on the graph contained 72% silica and the basalt 49% silica. The melting temperature of igneous rocks increases with a composition change from acid to basic, and it seems not unreasonable to assume that the temperature can be roughly correlated with the silica content. On the graph shown in Fig. 130 the writer has plotted a third curve which lies one quarter of the distance from the granite to basalt curves, and this represents about 66% silica, which is very close to the average for the Ulu Rompin granodiorites. If the limits of the probable confining pressures are projected onto the 66% silica curve, a temperature range of 740°C to 780°C is obtained, which shows that the pressure variation does not greatly affect the temperatures. It is concluded that the temperature of the Ulu Rompin granodiorite at the time of its intrusion was approximately 760°C. Winkler (1967) gives the temperature of granitic intrusions as 700°C to 800°C, and Goranson (1931) melted granite in the laboratory at 700°C under 1000atm water pressure. Larsen (1945) used a value of 820°C for the quartz diorite magma in the Californian batholith, and Buseck (1966) accepted this temperature for the granodiorite which is related to

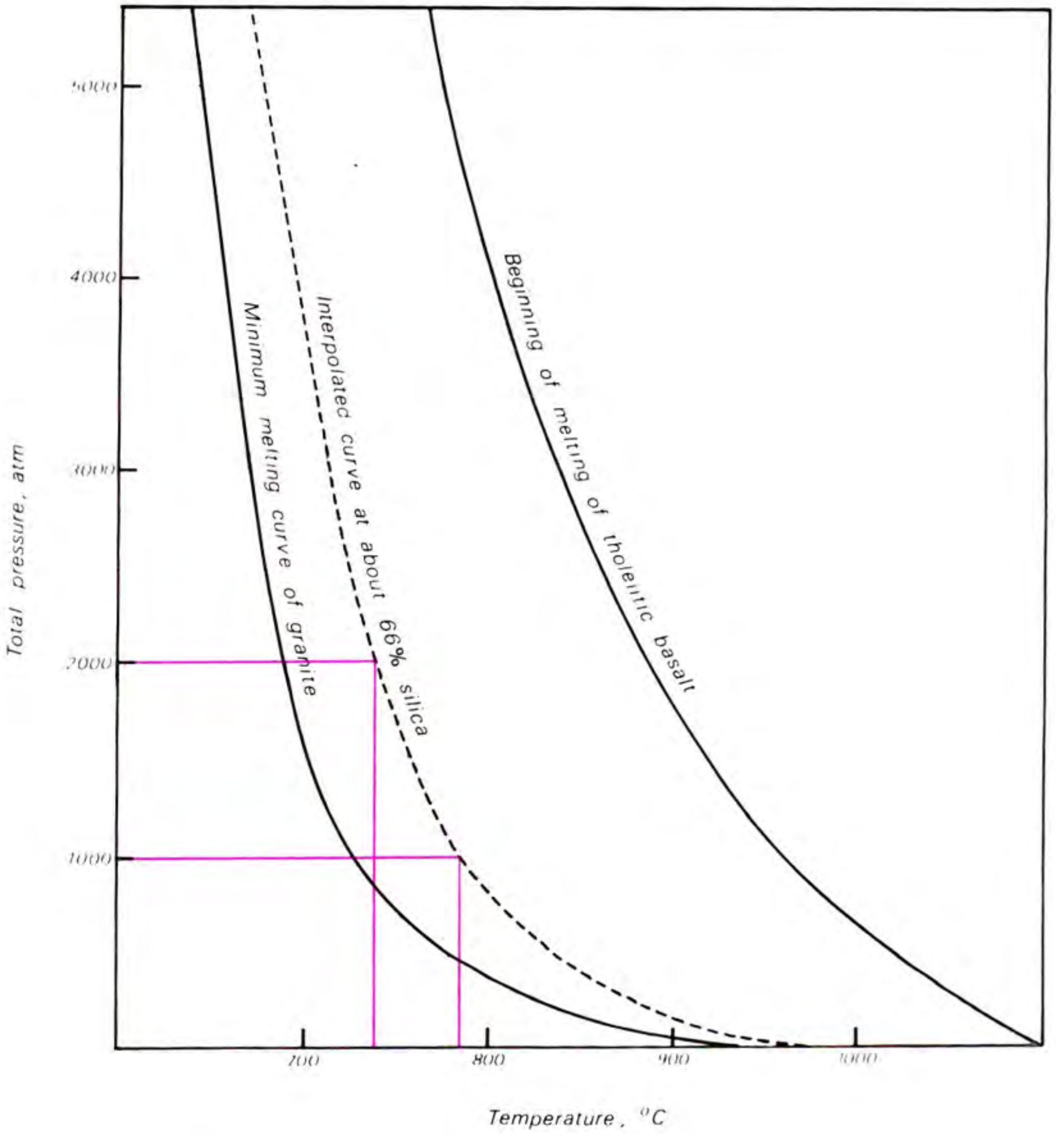


Fig. 130 Diagram which is used to make an estimate of the temperature of intrusion of Ulu Rompin granodiorites.

(Based on Kullerud 1959)

the contact deposits at Concepcion del Oro, Mexico. Sangster (1969) calculated that the intrusions associated with the magnetite deposits in southwestern British Colombia also had a temperature of 820°C: the Canadian rocks contain less silica (62.5%) than those found at Ulu Rompin, and are thought to have been subjected to confining pressures of less than 1000 atm, so one would expect the temperature to be higher than that in the magma at Ulu Rompin.

The other important factor which governs the rise in the rocks surrounding an intrusion is their original temperature. According to Winkler (1967) the minimum geothermal gradient is 10°C/km and "on the continent, beyond a depth of several kilometres, a geothermal gradient of 20°C/km is considered 'normal'". Turner and Verhoogen (1960) give an average figure of 30°C/km, while Barth (1962) says that 20°C/km is the average value, although wide variations have been recorded. Accepting an average value of 20°C/km gives a temperature of about 120°C at a depth of four miles, but allowing for a somewhat shallower depth of intrusion 100°C is taken as a fair estimate of the original temperature of the country rocks. Jaeger (1957) has made some theoretical calculations on the possible temperatures attained adjacent to intrusives, and his conclusion is that at the contact the country rock is heated to a temperature which is slightly greater than 60% of the temperature of the intrusive mass, plus the temperature of the invaded rocks. Taking a temperature of 760°C for the intrusion and 100°C for the country rocks, it is concluded that the temperature at the contact of the Bt. Ibam granodiorite was about 550°C. One important qualification for which allowance must be made is the effect of the transfer of heat by volatiles. Jaeger based his calculations on the heating of the country rocks by conduction, but if volatiles are important they will disperse the heat over a wider area and produce slightly lower temperatures at the contact. The figure of 550°C therefore, is taken as an upper limit for the contact rocks, and the maximum temperature was probably slightly below this and in the range 500°C to 550°C. The original carbonate rocks were very close to the intrusive and are classed as contact rocks. Other workers have given estimates of the temperatures developed adjacent to granitic intrusives, and many of these are in good agreement with the figures deduced for the Ulu Rompin granodiorites. Winkler (1967) quotes an intrusion temperature of 700°C which gives a temperature at the contact of 420°C plus the temperature of the invaded rocks, i.e. at Ulu Rompin the figure would be 520°C. Turner (1968) says "In summary, the respective cooling models of Lovering, Jaeger and Hori suggest the following inferences: Temperatures at contacts are normally 500°C to 550°C ...." Buseck (1966) calculated that the maximum contact temperature at Concepcion del Oro was about 500°C, but this seems to be slightly low in view of his assumption that the temperature of the intrusion was 820°C. The figures for the Ulu Rompin rocks are obviously estimates rather than precise calculations as a number of assumptions have been made, but they can, nevertheless, be critically examined to see if they are consistent with the mineral assemblages found in the rocks. It is gratifying to find that the assemblages are fully compatible with the suggested range of temperatures.

In an earlier section it was stated that the mineralogy of the calcic skarns and magnesian rocks showed that they fell in the albite-epidote-hornfels facies and occasionally passed into the lowest part of the hornblende-hornfels facies. According to Winkler (1967) the hornblende-hornfels facies begins at 535° ± 15°C at 1000 bars and 540° ± 20°C at 2000 bars, while Turner and Verhoogen (1960) say "we tentatively correlate the hornblende hornfels facies with temperatures of about 550° to 700°C in the pressure range P_{H_2O} = 1000-3000 bars ...." A few years later however, Turner (1968) came to a different conclusion, and after stating that "Even less satisfactory is the boundary between this facies" i.e. albite-epidote-hornfels "and the hornblende-hornfels facies", he reduced the temperature of the lowest part of the hornblende-hornfels facies to between 420° and 450°C, and included the whole of the albite-epidote-hornfels facies within a 50°C temperature interval. It is not altogether clear why he reduced the minimum temperature of the hornblende-hornfels facies by 100°C, but the higher figures seem to be in better accord with many of the known temperatures of formation of diagnostic minerals.

Some of the individual minerals in the skarn and gangue can be considered in the light of possible temperature conditions. According to Metz and Winkler (1964) diopside forms from tremolite + calcite + quartz at a temperature of around 500°C at confining pressures which vary from 1000 to 3000 bars. The temperature is also dependent on the carbon dioxide concentration in contact with the rocks, but the effect is only marked at very low concentrations. Turner (1968) states that where water is present and $P_{H_2O} = P_{CO_2}$ diopside forms at 540°C and 1400 bars. The scarcity of diopside strongly suggests that the necessary temperature of formation was rarely attained. Chlorite reacts with quartz to form either gedrite or anthophyllite + cordierite, and this reaction has been determined as occurring at 560° ± 10°C at 2000 bars and 530° ± 10°C at 500 bars, with an interpolated value of 540° ± 10°C at 1000 bars (Winkler, 1965). The presence of abundant chlorite shows that these temperatures were seldom, if ever, reached. The stability of the tremolite-actinolite series has been studied by Hellner et al (1965), and they report that actinolite (Fe 60% Mg 40%) has a lower stability limit of 360°C at 1000 atm. Anthophyllite was shown by the same workers to be stable above approximately 500°C, but at this temperature the mineral contained a large proportion of iron and the magnesian varieties were only stable at even higher temperatures. Epidote is reported to have a temperature of formation which varies from approximately 300°C at the lower end (Stringham, 1952) to between 400° and 500°C at the upper end (Ramberg, 1949 and Rosenqvist, 1952). Using data obtained by Merrin (1960), Sangster (1969) deduced that at 1000 bars the upper stability limit of epidote is 480°C. Andradite is stable over a wide range of temperatures, but because of the identity of its associated minerals it must have formed at moderate temperatures in the Bt. Ibam deposit, and that this is feasible is fully substantiated by synthetic studies. Andradite has been synthesized at 500°C and 500 bars by Christophe-Michel-Levy (1956), and at 480°C and only 150 atm by Jagitsch (1956).

All the above data is compatible with a suggested maximum temperature during the formation of the skarns and gangue of between 500° and 550°C.

It can be seen from the preceding paragraphs that the silicate minerals give some useful indications of temperature conditions, but unfortunately the metallic minerals, both oxide and sulphide, are not as helpful. There are a number of minerals and mineral associations which are frquently used as geothermometers, but they are not well developed in the Bt. Ibam area.

The only sample of sphalerite for which a chemical composition can be determined occurs in veins with no associated minerals, and for all practical purposes pyrrhotite is absent from the entire deposit. From the composition of the sphalerite as determined by a cell size calculation - $Zn_{.85}Fe_{.15}S$ - a temperature of formation of about 425°C is indicated, although this cannot be precise as there is no way of determining if there was sufficient FeS available within the system to promote equilibrium. Lovering (1958) pointed out that if it cannot be proved that equilibrium conditions prevailed during the crystallization of the sphalerite, then the apparent temperature of formation will be a minimum figure. Titaniferous magnetite can have some use as a temperature indicator, but the Bt. Ibam ore is extremely low in titanium. Magnetite is by far the most abundant ore mineral, but it can form under such widely varying conditions that by itself it reveals nothing about the temperatures prevailing at the time of its formation, although it indicates that the oxygen activity was low.

Chalcopyrite obtained from a small vein below the ore body showed well developed twinning in polished section, and this can be due to inversion of the high temperature cubic form to the low temperature tetragonal mineral. Yund and Kullerud (1966) have determined this inversion temperature as 547° ± 5°C, but the temperature is lowered with increasing pressure, although only at the rate of approximately 4°C/1000 bars. None of the other chalcopyrite samples encountered by the writer showed any sign of twinning, which suggests that the sample in the vein is atypical, and that the chalcopyrite in the ore body formed below the inversion temperature. There is some evidence (Shadlun, 1953) that stress can produce twinning in chalcopyrite which looks identical to inversion twinning, but at Bt. Ibam the chalcopyrite seems to be later than the main faulting movements, although local stresses could conceivable have caused the twinning in the vein material. All that can be deduced with reasonable onfidence is that most of the chalcopyrite formed below 540°C. Chalcopyrite is sometimes found as exsolution lamellae in sphalerite, showing that these particular grains crystallized at a temperature which permitted solid solution between the minerals.

Experimental studies have shown that sphalerite–chalcopyrite intergrowths are homogenized at about 400°C (Park & MacDiarmid, 1970), and Lovering (1958) states that exsolution blebs of chalcopyrite in sphalerite indicates a temperature of formation above 350°C. It is deduced that sphalerite and chalcopyrite formed above 400°C and below 540°C.

It is certain that the temperatures which prevailed during metallization were lower than the maximum reached during skarnification i.e. 500° to 550°C. If the temperatures had risen significantly during the deposition of the ore minerals, then the assemblages of silicate minerals would show more characteristics of the hornblende–hornfels facies.

Late–Stage Mineralization

Subsequent to the main oxide and sulphide mineralization solutions at moderate to low temperatures were still active, an important product being siderite. The late–stage minerals are typically found in veins, but along section N. 10,400 siderite may also have partially replaced the acid volcanic rocks. Other products found in veins are quartz and calcite, and rare barytes and zeolites. Both siderite and barytes have been encountered in veinlets cutting the granodiorite, so they were obviously intruded after the magma had solidified and the mass become jointed. Barytes was deposited later than the siderite. The solutions which deposited the siderite found in the acid volcanic rocks along N. 10,400 must have encountered sphalerite or other zinc mineralization during their circulation, and by reaction acquired the elevated zinc content which the siderite now contains; the siderite veinlets which cut the granodiorite contain no significant zinc. The bulk of the zincian siderite is confined to the southern end of the ore body, and its location is possibly controlled by a postulated fault on or near N. 10,400, and the proven zinc mineralization at depth. Quartz and calcite are plentiful in some areas, and in a few veins it is apparent that the calcite has followed and partially replaced the quartz. The location of a number of quartz veins has been controlled by faults and fractures. The largest masses of quartz are found in the skarns, where it is accompanied by micaceous haematite and pyrite in pods of mineralization. Quartz has been deposited over a long period of time; the preservation of fresh actinolite, chlorite and fragmented magnetite in silicified masses suggests that this particular quartz formed relatively early during the main phase of mineralization, but most is a late-stage product.

Siderite, calcite and barytes were deposited from solutions which must have contained CO_3'' and SO_4''. Some of the barium could have been leached from the acid volcanic rocks, as these usually contain a significant barium content. The iron and manganese in the siderite are possibly a residue from the original granodiorite magma, or they could have been leached from the solidified intrusive mass. Another possible source of the iron is pyrite, which is widely disseminated throughout the volcanic rocks.

Hydrothermal Alteration

There has been a long history of alteration, and in the volcanic rocks it is often extremely difficult or impossible to distinguish those products due to hypogene processes from those caused by supergene action. Many geologists have commented on this problem, as for example Grim (1953) who says "The study of the alteration products associated with many ore bodies is complicated by the difficulty of distinguishing between hypogene and supergene products". The alteration sequence must have started during the passage of the early mineralizing solutions, when the volcanic rocks suffered mineralogical and textural changes. Rare biotite was formed, and some of the quartz phenocrysts recrystallized to a fine quartz mosaic. The groundmass, which was probably ash or glass or a mixture of both, recrystallized to a fine granular aggregate. The only important alteration products are kaolinite and sericite, and while their presence can give some clues as to the nature of the generating solutions, any speculation has to be made with care. There is no doubt however, that some of the sericite and larger flakes of muscovite have resulted from hydrothermal processes, as pyrite can be found moulded round mica flakes.

It is significant that the only samples of really hard volcanic rock come from core in some of the deepest drill holes, while the nearer-surface rocks are usually soft, and most of the exposed rocks are extremely soft and often a pure clay. This strongly suggests that weathering has been responsible for the physical decomposition of the volcanic rocks, which in its turn appears to be largely due to the formation of kaolinite. It is concluded, therefore, that most of the sericite is of hydrothermal origin but much of the kaolinite is probably supergene.

At room temperature sericite is only stable within the vary narrow pH range of 9.5 to 10.5 (Garrels, 1957), but a rise in temperature and a change in the chemical conditions can profoundly modify this. Both a rise in temperature and an increase in the K^+ concentration can shift the stability field of sericite to lower pH values (Gruner, 1944), and a combination of the two would presumably enhance the effect. Gruner (1939) has shown that sericite can form from solutions with a pH of about one (determined when cold) at temperatures over 350°C. Sericite could well have formed as a result of the action of the solutions which deposited the magnetite and are thought to have been acidic, and as the volcanic rocks are rich in potash, K^+ would have been available in abundance. Sericite is stable to over 600°C (Yoder and Eugster, 1955), which is well above the temperature envisaged for the early solutions. Although kaolinite is formed preferentially in an acid environment the temperatures of the early solutions, and probably the K^+ concentration, would be too high to allow it to develop. Kaolinite is stable up to a temperature slightly over 400°C (Roy and Osborn, 1952). As the temperature dropped the alkalinity of the solutions probably rose, so the formation of kaolinite would still have been inhibited. By the time that pyritization was affecting the volcanic rocks it would appear that much of the sericite had been formed.

It is not always clear at what stage the chlorite formed, but some is definitely later than pyrite as it is devloped round grains of this mineral, presumably extracting some iron. A little chlorite has formed by the alteration of feldspar in situ, but it is also found in veinlets. The latter usually contain calcite, and these veins may have been filled by material expelled from the ore body.

There is no discernible zonation of the alteration products, and apart from a few samples immediately adjacent to the ore body, it is impossible to distinguish samples of volcanic rock from Bt. Ibam from others collected far away.

The changes within the ore body were largely confined to the replacement of actinolite and chlorite by talc. The latter mineral has been prepared at all temperatures under 800°C (Deer, Howie and Zussman vol. 3, 1962). Turner (1948) quotes equations which show that reaction between carbon dioxide and tremolite (actinolite) or chlorite in constant volume reactions can produce talc, and such reactions would readily explain the widespread occurrence of talc which pseudomorphs the fibrous amphibole and the flaky chlorite. These reactions are described as taking place "At the lower temperatures in the greenschist facies", which is in accord with the evidence at Bt. Ibam. If carbon dioxide was an important constituent in the formation of talc, then it provides additional support for the suggestion already advanced that the carbon dioxide was retained within the system until the end of magnetite deposition. At very low temperatures talc is unstable in the presence of excess lime and carbon dioxide and gives way to magnesite, and as this mineral has not been identified at Bt. Ibam, it indicates that the lime and carbon dioxide expelled from the original dolomite had been completely removed by the time that low temperature conditions prevailed.

It is possible that some of the magnetite was oxidized to martite during the phase of hydrothermal alteration, but the extent of the reaction must have been small as most martite is of supergene origin. There is a reasonably consistent pattern of increased alteration with proximity to the surface, but some rare reversals may indicate martite of hypogene origin.

Supergene Alteration

Alteration under conditions of low temperature and pressure has extensively modified the original Bt. Ibam ore body and country rocks, and made an important contribution to the geological features as found today. The most widespread changes have been kaolinization in the volcanic rocks, oxidation - with or without hydration - of the primary magnetite, and the development of laterite. The mine authorities attribute considerable economic benefits to the action of supergene alteration, but the writer is no in complete agreement with this assessment.

Within the volcanic rocks very large quantities of kaolinite have been formed, and some of the near-surface rocks are virtually a pure kaolinite-quartz mixture. The formation of kaolinite indicates an acid environment, and the effect of the carbonic acid normally found in rainwater would be greatly enhanced by the weak solutions of sulphuric acid produced by the decomposition of the widely disseminated pyrite. As Loughnan (1969) says, although sulphides "are not abundant in silicate rocks .... their influence on rock weathering" is "locally quite marked ....". Another factor which may have contributed to the development of kaolinite in the volcanic rocks is the virtual absence of calcium. Grim (1953) reports that "When present in the environment, calcium particularly appears to retard the formation of kaolinite", but as has been shown in chapter III, the volcanic rocks are extremely low in lime. The same argument is not applicable to the hydrothermal solutions which, although they were probably acidic, were too hot to permit the formation of kaolinite. The main source of kaolinite has been feldspar, but with advanced alteration both mica and chlorite have been decomposed. Much of the kaolinite has formed in situ, but some has been transported and deposited in veinlets. This is well displayed in the rocks close to the hangingwall contact which are cut by myriads of small joints, many of which are filled with kaolinite. It is likely that volume changes during mineralization caused tensions to develop in the brittle volcanic rocks which consequently fractured, the resulting joints forming ideal passageways for solutions when the rocks came within the zone of weathering.

Simultaneously with the formation of kaolinite, pyrite has been altered to iron hydroxides, and these are responsible for the widespread colouration in shades of red, brown, orange etc. found in the volcanic rocks. Most of the colouration is completely irregular, but occasionally a banded appearance is produced where closely spaced veinlets of pyrite were originally present. Limonitic material has been deposited along some fractures and joints in the volcanic rocks, and the main north to south fault which cuts both the ore body and country rock contains a mixture of iron and manganese oxides and hydroxides, in addition to quartz. Nodules and small concretionary masses of iron hydroxides have been important in the development of laterite.

Towards the southern end of the mine, and especially near to section N. 10,400, zincian siderite has been extensively altered to goethite, some of which has remained in situ as coronas on siderite grains, but most has been transported in solution and deposited elsewhere. It is largely to the action of these mobile solutions that the writer attributes the present distribution pattern of zinc.

Within the ore body weathering processes have attacked the ore minerals and gangue. Magnetite has been replaced by martite and goethite, and in general there is a gradation from the extensively altered magnetite near surface to fresher ore at depth. Martite has preferentially attacked grain peripheries and along cracks, and every gradation can be found from a few lamellae of martite to martite pseudomorphs with minute relics of magnetite. Some of the goethite which has replaced magnetite in situ is distinctly granular, but by far the greater part of the goethite is found in colloform masses. The circulating colloidal solutions have played an important role in the distribution pattern of zinc and, to a lesser extent, copper. Because of the very large surface area of colloidal particles they have a high surface energy (Barton, 1959), and the ability to absorb ions. The writer believes that most of the copper and zinc ions released by the breakdown of chalcopyrite and sphalerite-zincian siderite respectively were absorbed by the colloidal solutions, and incorporated in the structure of the goethite when it was eventually deposited. The fact that the concentrations of copper and zinc in the goethite rapidly vary is probably due to fluctuations in the rate at which the elements became available. There must have been some intermingling of solutions as primary sources of copper and zinc are not typically found together, but the zone through which the colloidal solutions moved was relatively restricted, as zinc has not been transported to the northern half of the ore body in large quantities, and the same applies to copper with respect to the southern half. The main movements of the colloidal solutions must have been in a vertical rather than a lateral direction.

Some of the goethite masses are intimately associated with colloidal haematite, which suggests that there has been a delicate control over the crystallization from colloidal solutions. The precise nature of the control is unknown, but Valeton (1972) discussed the system Fe_2O_3—H_2O and concluded "The factors governing crystallization of specific phases are: Eh, pH, concentration of the solution, other ions in solution, ageing conditions of the gels, and temperature, among others". According to Garrels (1959) the first precipitate is a hydrous oxide of indefinite water content, but whether this becomes goethite or haematite is apparently dictated by compositional differences in the solutions. The petrographic evidence shows that goethite and haematite did not coexist, and for most of the time the conditions favoured the formation of goethite.

An important function of goethite has been to act as a cementing agent. There is no doubt that if goethite had not been so widely distributed most of the high grade magnetite ore would have been highly friable. Following the breakdown of the grain to grain adhesion because of the development of peripheral martite, the goethite found entry along the intergrain channels and firmly cemented the whole mass. In the patches of friable ore which are still preserved there is virtually no goethite. Goethite has also formed solid masses along parts of the footwall and hangingwall contact zones, probably because they proved to be easy channels of movement, partly by a cementing action and partly by replacement.

Goethite has extensively replaced the gangue minerals, and the process of replacement can be traced in some polished sections. A number of samples showed a delicate network of gothite where it had followed the cleavages in amphibole grains, but as alteration advanced the delicate threads broadened and eventually coalesced to produce a solid mass. One of the few patches of banded ore in the Bt. Ibam ore body showed a gradual change from alternating bands of magnetite and talc to solid goethite in which the banding was clearly preserved. Close to the surface some of the massive chlorite rock has been altered to a tenacious limonitic clay.

It is in the treatment of the gangue minerals within the weathering zone that the writer disagrees with the mine authorities. They postulate that supergene activity actively leached out the silicate minerals, thereby leaving an enriched iron ore near the surface. A company report suggested that some 32% by volume of silicates in the 'protore' had been removed by leaching, of which 20% is now accounted for by intergranular space and 12% by collapse, particularly along the hangingwall side. It is also suggested that some magnetite formed from the circulating meteoric waters. The writer believes that the gangue minerals close to the ground surface have been replaced progressively by goethite, which requires that most of the original silica, alumina and magnesia was removed as part of the reaction. The difference between the two interpretations is that the mine geologists remove the silicates by simple leaching, thereby leaving a volume deficiency, whereas the writer believes they were released as a direct result of replacement by goethite. The company's theory implies that there should be an overall distribution pattern of iron values which is horizontal or roughly parallel with the ground surface, but there is no evidence for this, and in fact very high grade ore can be found in some of the deepest levels of the ore body. There must have been some volume changes as a result of all the mineralogical reactions which took place, but there is no evidence for the large changes postulated in some company reports. As a result of accomodation to small volume changes the ore body has progressively slipped down the footwall contact, producing the well-marked plane of movement described in chapter VI.

The secondary copper minerals have been formed as a direct result of supergene processes. The presence of chalcanthite shows that sulphate-bearing solutions have been active, and indeed are still in circulation, so the commonly accepted method of supergene copper formation which involves reaction between sulphide minerals and copper sulphate solutions (e.g. Batemen, 1955) has almost certainly been operative. The breakdown of pyrite would produce abundant sulphate ions to maintain the reactions. The crystallization of the chalcanthite can be interpreted as breaking the sequence of reactions involved in the formation of secondary copper sulphides. A large part of the secondary sulphides have formed by direct replacement of earlier sulphides, and it is rare to find the secondary copper minerals far removed from the primary chalcopyrite.

There is no clearly marked zone of copper enrichment, and much of the present distribution closely reflects the primary dispersion. There is some indication that the most favourable zone for secondary reactions has been just above the water table, but this must have fluctuated as is shown by the age relationships between the secondary sulphides and oxidized products. It is not uncommon to find evidence that cuprite, which must have formed above the water table in a zone of freely flowing oxidizing waters, is earlier than some of the secondary sulphides. The copper within the oxidizing zone has shown more mobility than that which formed the secondary sulphides. Native copper has always crystallized in open spaces where solutions could freely circulate, but most of the secondary sulphides are typically found in the body of the samples. Native copper seems to have acted as a catalyst for the precipitation of suprite in some zones. Where calcite was present within the oxidizing zone the acid cupriferous solutions have reacted to produce malachite and, very rarely, azurite. The secondary copper sulphides have attacked most of the earlier sulphides, and on rare occasions magnetite.

The weathering solutions have leached some silica from the near-surface rocks and deposited it at lower levels as opaline quartz, particularly along joints. Some of this quartz is intimately associated with cuprite. Just to the north of the Bt. Ibam ore body in the area round Bt. Sanam the leaching of silica has been very effective, and has left behind a small deposit of low grade ferriferous bauxite.

As weathering gradually broke down the volcanic rocks to a clay the ground surface was progressively lowered, with the result that along parts of its outcrop the iron ore formed striking cliffs protruding above the general level. The small cliffs of iron ore could not stand indefinitely, and they gradually collapsed to form the important accumulations of boulder ore found on the ridge slopes below the outcrops.

The Bukit Pesagi and Bukit Sanlong Ore Bodies

The original Japanese prospectors interpreted the iron ore as a segregation product from the acid magma which had produced the country rock, identified by them as liparite. During the intensive prospecting operations of Rompin Mining Co. ideas changed gradually. At one time it was thought that the primary ore bodies were situated just above dioritic stocks and had been derived by segregation from the original magma, but subsequent deep drilling showed the presence of substantial pyrite below the iron oxide bodies, and it was then suggested that the oxide bodies had formed by oxidation of pyrite.

The important factors which distinguish the primary ore bodies from the Bt. Ibam body are:

1. The only important primary iron mineral is haematite, although magnetite was the first to crystallize.
2. The gangue is decomposed country rock, and there are no calc or magnesian silicates.
3. The country rock is sheared, the intensity of shearing increasing towards the ore bodies.
4. There has been some scattered haematite mineralization in the country rock immediately adjacent to the main ore bodies.
5. Pyrite forms substantial masses below the haematite bodies.

Item one shows that initially the mineralizing fluids possessed a low oxygen activity, but this rapidly gave way to more oxidizing conditions. Items two, three and four taken together provide good evidence that the control over the loci of mineralization was structural and not lithological. There is little doubt that Pesagi 6 and Sanlong B ore bodies occupy the sites of shear zones, which must have provided easy access channels for the ore solutions and thereby determined the location of the mineralization.

The original fluids were probably similar in type to those which deposited the Bt. Ibam ore body, i.e. acidic and possessing a low oxygen activity. The first oxide mineral to crystallize was magnetite, but in the large ore bodies this soon gave way to haematite, which attacked the magnetite and left only the small relics seen in polished sections. By analogy with the Bt. Ibam deposit it is suggested that the change from magnetite to haematite deposition may have been connected with a rise in the alkalinity of the solutions. The fact that the mineralization is located in fracture zones probably means that any volatiles would have been able to escape fairly easily, as the cover rocks were unlikely to have formed an impervious trap. This would have two effects; firstly, because of the volatiles taking heat with them the temperature would be lowered which would help to promote the formation of haematite, and secondly, the acidic fluids would be neutralized by reaction with the country rock, and unlike the position at Bt. Ibam there were no special factors helping to sustain the acidity of the fluids. The lack of ferrous iron at the sites of mineralization could also have contributed to the early formation of haematite, as ferrous iron appears to act as a reducing agent by helping to maintain a low level of oxygen activity. In some of the very small pods of ore the mineralization has been restricted to magnetite, which has subsequently suffered extensive oxidation. In these bodies all primary oxide mineralization ceased before the start of haematite crystallization.

Much of the haematite shows abundant evidence of being a primary product; it occurs as specularite crystals with a good euhedral shape, as micaceous ores, and it is extensively twinned, none of these features being typical of supergene ore. It is usually quite easy to distinguish the secondary haematite, as it is in colloform and earthy masses associated with goethite.

The oxide phase was brought to an end by a rise in the sulphur fugacity, which initiated the formation of pyrite. Some of the fluids followed the same main entry channels, i.e. the shear zones, but were blocked when they encountered the recently deposited haematite masses, and so the massive pyrite is found immediately below the oxide bodies. In the Pesagi 6 and Sanlong B ore bodies the rise in sulphur values with depth is due solely to this fact. The evidence from polished sections, plus some macroscopic evidence, confirms that the pyrite is later than the haematite. The suggestion put forward by some company geologists that the oxide ores have formed by oxidation of massive pyrite is incorrect. Apart from the evidence that the sulphide is later, the ore bodies are not of a type which is formed in a supergene environment.

In addition to forming the massive bodies, pyrite has been very widely disseminated throughout the country rock as fine grains and veinlets.

The only important product of hydrothermal alteration has been sericite, which is abundant in some of the volcanic rocks.

Supergene alteration has reduced the near-surface volcanic rocks to a kaolinite-quartz mixture as at Bt. Ibam, but there has been one development which is not found at Bt. Ibam, and it is probably the result of differences in topography. Round Bt. Pesagi and Bt. Sanlong there are extensive superficial sheet-like bodies of ore, which are made up of secondary iron material and relics of small primary bodies which have completely disintegrated. The secondary ore is in effect a thickened iron pan, and is roughly parallel with the ground surface. It appears to have formed at or close to the water table on the rounded upper slopes of the hills; it is not found on steep slopes. The iron has been derived partly from the disseminated pyrite and partly from small primary bodies, the latter providing iron taken into solution and detritus of haematite. The fragments of the primary ore have been cemented into the solid masses of secondary material. The conditions essential to the formation of this type of ore appear to be a fairly gentle topography, and a periodic oscillation in the level of the water table caused by the seasonal monsoon rainfall.

The Bukit Hitam Ore Body

A company report states that the Bt. Hitam deposit formed in a skarn, but this conslusion is based on very flimsy evidence. There are some limonitic pseudomorphs of a dodecahedral mineral which the company report identifies as garnet, but the writer believes they were originally magnetite, which typically displays this crystal form throughout the Ulu Rompin mineralized zone. Also, analyses of rock samples made in the company's laboratories show low magnesia and lime while the writer has detected relatively high barium, all of which indicates that the rocks are acid volcanics and that skarn as found at Bt. Ibam is not present. Because of the extreme decomposition of the country rocks there is little concrete evidence regarding the genesis of the body. It is located just above a granodiorite stock and the ore mineral is magnetite, so it bears a general similarity to the Bt. Ibam body but lacks the skarn and magnesian rocks.

It is suggested that the magnetite may have completely replaced a small carbonate lens or lenses, which had not been subjected to skarnification prior to mineralization. A comparison may be drawn with the Bt. Batu Puteh deposit where coarse marble is still preserved, except that at Bt. Hitam the carbonates have been completely replaced. The hypothesis developed earlier for the Bt. Ibam deposit suggests that there is a connection between the crystallization of magnetite and the presence of carbonate rock, and this may be applicable to the Bt. Hitam deposit.

The Bukit Batu Puteh Ore Body

There are two features which distinguish this deposit from the others in Ulu Rompin; it does not appear to be closely associated with an intrusive body, and it lies outside the narrow belt of country in which the other primary bodies are located. The failure to locate an intrusive body is negative evidence only, but as discussed earlier in chapter V, the magnetic anomaly map suggests that there is no near-surface intrusion at Bt. Batu Puteh.

Fairly pure, coarse-grained marble is found in the decomposed acid volcanic rocks, and it is reasonable to assume that the control over the location of the mineralization has been lithological as there is no evidence which suggests a structural control. The iron ore is magnetite, which is typically associated with a brown clay, although it is also very close to marble. The writer has not seen any samples which actually show magnetite in marble, but the evidence is sparse, and it may be necessary to await large-scale excavations before a clear picture emerges. It is suggested that the magnetite has replaced marble and any unreplaced fragments of marble in the iron ore subsequently removed by hydrothermal alteration or weathering. The absence of any skarn minerals may be due to the absence of a nearby intrusive body.

Bukit Sanam and Other Minor Deposits

Diamond drilling has produced evidence which suggests that there may be a small primary ore body at Bt. Sanam, but the data is far from conclusive. Most of the ore in this and the other minor deposits is of secondary origin, and has formed by a process of enrichment. The details have been discussed previously in chapter XVIII.

XX ORIGIN AND TRANSPORT OF THE IRON

Source of the Iron

As discussed in the previous chapter, it is highly probable that the Ulu Rompin ore deposits formed as a consequence of the intrusion of granodiorite magma, but this does not necessarily mean that the iron and other elements had their origin within the cooling magma. The magma would have provided heat and active fluids, but there is no firm evidence as to the ultimate source of the iron and other elements involved in the mineralization. A sedimentary or volcanogenic origin is ruled out, but there are several alternative sources from which magmatic fluids could have obtained their useful constituents, and these can be listed under two main headings.

By segregation or fractionation from

1. The cooling magma which formed the granodiorite stocks.
2. The same magma, but at a deeper level in the crust.

By leaching from

3. The solid or partially solidified granodiorite.
4. The volcanic country rocks.
5. Rocks of unknown composition at a deep level in the crust.

A completely different mechanism which could have been operative is leaching and redeposition via meteoric waters heated by the granodiorite, but this is not a convincing alternative. Many workers have noted the close association between non-titaniferous magnetite deposits and intrusive rocks of a restricted compositional range - typically hornblende granite, adamellite, granodiorite and alaskite - which strongly suggests that the composition of the intrusive is an important factor, whereas an intrusive of any composition could supply the heat necessary to raise the temperature of meteoric waters. It is thought to be most unlikely that the Ulu Rompin ores formed from meteoric waters.

Due to very poor exposure and lack of drill core there is little information on the internal characteristics of the granodiorite, but those parts which have been examined invariably contain magnetite - company geologists have reported up to 10% - and the intrusive is the dominant influence on the regional magnetic pattern, Iron is also present in silicate minerals such as hornblende and biotite. It would appear the the granodiorite was, at least theoretically, fully capable of supplying the iron. One cubic mile of granodiorite with 2% Fe contains 220 million tons of iron, which is equivalent to about 300 million tons of magnetite. In contrast, it seems unlikely that the acid volcanic rocks would have provided a ready source of iron. Much of the iron in the analyses of volcanic rocks presented in chapter III is contained in pyrite, which is later than the magnetite mineralization, and goethitic material deposited from weathering solutions. The composition of the volcanic rocks at depth is unknown, but the evidence suggests that they are unlikely to be significantly different from those near the surface, and so will be poor in iron. Invoking the presence of other iron-rich rocks at depth is pure speculation. Immediately to the west of the Ulu Rompin mineralized zone there are large areas of highly ferriferous sediments, and although these are probably younger than the Ulu Rompin rocks, if similar types were present at depth they could supply an ample source of iron. It is concluded that the iron involved in the mineralization was derived from the granodiorite, but whether or not the iron was present from the time of the initial formation of the magma or acquired by assimilation during its rise through the crust is not known.

The granodiorite could have released the iron in hydrothermal fluids separated from the cooling magma, or fluids which were originally unmineralized may have leached iron from the solid or partially solidified rock. The latter mechanism has been demonstrated convincingly in the magnetite deposits of Iron Springs District, Utah, where Mackin (1968) showed that deuteric solutions leached iron from ferromagnesian minerals in the quartz monzonite or granodiorite porphyry intrusives. In a more generalized conclusion, Buddington (1966) has suggested that alaskites originate in a hornblende granite from which iron-bearing solutions had been extracted. He notes the association of both igneous rock types with magnetite concentrations in the states of New York and New Jersey, U.S.A., and the fact that hornblende granite grades into alaskite. It is unfortunate that there is not sufficient evidence on the Ulu Rompin granodiorites on which such hypotheses can be tested; information on the internal features is essential, but at Ulu Rompin only the peripheries have been exposed or drilled.

The known outer margins of the granodiorite stocks are rich in iron, which implies that if leaching has been an effective mechanism it has affected the internal and deeper parts of the intrusion. As the highest parts of the intrusion have been among the last to form, the quantity of iron in the margins can be interpreted as evidence of an increasing enrichment of iron in the final stages of crystallization, which could lead to the formation of iron-rich fluids. Although the evidence at present available is not sufficient to allow firm conclusions to be drawn, the writer favours the formation of iron-rich fluids from the granodiorite by a process of fractionation rather than by leaching.

Transport of the Iron

An ore-forming fluid is a highly complex system and its precise nature is still uncertain. It is clear, however, that water is invariably the most important single constituent, and is the medium in which the other products are normally carried. Although proof of the existence of an iron ore magma has been discovered in South America (Park, 1961), there is nothing to suggest that one was ever present in the Ulu Rompin area, and the mineralization should be explained in terms of an aqueous system.

The temperature and pressure conditions have been discussed in the previous chapter, where it was also stated that the fluids were initially acidic. The acid nature of the fluids was probably due to the presence of constituents such as chlorine, fluorine and sulphur dioxide. There have been a number of studies of solutions and gases in active volcanic areas, and the emanations are typically acidic. The classic studies of the Valley of Ten Thousand Smokes, Alaska (Zies, 1929; Fenner, 1933) showed that HCl, HF and H_2SO_4 were present in abundance. The acid nature of the fluids in Ulu Rompin probably contributed to their low oxygen activity.

Many of the minerals found in a hydrothermal ore deposit are virtually insoluble in water, so it is clear that the minerals were not transported in simple solution. Various mechanisms have been proposed to explain the transport of the constituents in an aqueous solution, and for the Ulu Rompin deposits chlorite and chloride complexes may provide the answer. There is a great deal of evidence that chloride ions play an extremely important role in ore fluids. According to Rosler and Lange (1972) "Most hydrothermal solutions are slightly dissociated electrolytic solutions rich in alkali chloride". The material found in many fluid inclusions is thought to be representative of the mineralizing solutions, and the evidence is summarized by Stanton (1972) who says that they "were highly saline, contained high concentrations of chloride ions and CO_2, and were of neutral to slightly acid pH". Modern solutions can also show high chloride; two samples from the Salton Sea geothermal area in California each contained 155,000 ppm chlorine, with about 50,000 ppm sodium and 17,000 ppm potassium (Skinner et al, 1967), while three of the brine pools in the Red Sea contain on average 156,200; 155,200 and 42,000 ppm chloride respectively (Degens & Ross, 1969).

In the Valley of Ten Thousand Smokes, where magnetite was the earliest of the ore minerals to be deposited from the fumaroles, Zies (1929) suggested that the iron was transported as a halide vapour and magnetite formed according to the reaction

$$Fe_2Cl_6 + FeCl_2 + 4H_2O \rightleftharpoons Fe_3O_4 + 8HCl$$

It is likely that the temperatures during the formation of this magnetite were considerably higher than those which obtained at Ulu Rompin, as temperatures up to 650°C were measured in the fumaroles.

It seems reasonable to suggest that at Ulu Rompin the iron could have been transported in the form of chlorides, and possibly some chloride complexes. Below 550°C (the maximum temperature at Ulu Rompin) ferric chloride can dissociate partially to form ferrous chloride and chlorine, so both ferric and ferrous ions would have been available. Holser and Schneer (1961) have prepared magnetite from acid solutions of ferric and ferrous chlorides at temperatures and pressures not significantly different from those which retained at Ulu Rompin. Ovchinnikov et al (1964) have stressed the importance of chloride complexes in the extraction, transportation and deposition of metals in ore deposits, and have some experimental data to support their suggestions. They point to the formation of complex compounds such as $Na(FeCl_4)$ and $K(FeCl_4)$ when weakly acidic solutions of sodium and potassium chlorides attack natural rocks and minerals which contain iron; other metals can react similarly and form chloride complexes.

Although the iron could possibly have been carried in the form of simple chlorides in solution, the copper, zinc, lead and bismuth which were deposited as sulphides may have been carried in chloride complexes. Helgeson (1964) has shown that many metallic elements can form complexes, and their order of stability coincides with the usual order of deposition of the ore minerals in nature. Ovchinnikov et al (1964) suggest that the complexes of alkali chlorides and heavy metal chlorides are quite stable, and the sulphides of the metal will not form until the complexes have broken down. This may have been connected with a change from acid to neutral or weakly alkaline conditions, which, as has already been discussed, possibly coincided with the cessation of magnetite crystallization at Ulu Rompin.

XXI MINING METHODS AND QUALITY CONTROL

Mining Methods

The primary ore bodies at Bt. Ibam and Pesagi 6 are worked in open pits with 20 ft terraced benches. The workings are kept free draining, an important factor as rainfall is very heavy, but in the later stages of mining in Bt. Ibam some pumping will be necessary to de-water the lowest workings. The ore bodies have been worked progressively downdip from their outcrops, and the nature of the wall rocks has been of vital importance in the design of the pits. Apart from isolated pockets, the country rocks have been decomposed and are often jointed and they have little inherent strength, so it has been necessary to leave the walls with a lower batter than would have been possible with stronger wall rocks. At Bt. Ibam a great deal of stripping has been required on the hangingwall side to follow the ore body in depth, while the footwall has been cut back to about 25°-30° i.e. a lower angle than the ore body. The decomposed nature of the wall rocks has its compensations, as much of the waste can be removed without recourse to blasting.

Mining continues 24 hours a day six days per week, Sunday being utilized for servicing and repairs. One evening blast per day is sufficient to keep the treatment plant supplied for the following 24 hours. Due to the variable physical characteristics of the ore the blast has to be carefully planned to ensure proper breakage. The broken ore is loaded by 4½ cubic yard shovels into rear-dump trucks which carry up to 70 tons of ore. Three shovels are adequate to operate the ore removal and stripping in the Bt. Ibam pit. There are four products from this pit;

1. Ore sent directly to the treatment plant.
2. Low grade magnetite ore which is stockpiled for magnetic separation.
3. Heavily contaminated ore which is stockpiled.
4. Waste which is dumped.

Mining continues throughout the year, but during the monsoon season the operations are occasionally halted, largely because of bad conditions along the haulage roads.

Treatment Plant

One main plant serves all the Ulu Rompin deposits, but a subsidiary magnetic separation plant is needed to treat some of the low-grade pyritic magnetite ore. The usual treatment is based on sizing and washing. The lump product, +¾" to -8", is washed and then stockpiled, while the fine ore is recovered using cyclones and spiral classifiers, and then stockpiled.

Transportation

The treated ore is taken from the mine to the coastal installations in 40 ton bottom-dump wagons via the company's narrow gauge railway. The ore-carrying ships, which have to stand out to sea because of shallow water conditions, are loaded from lighters towed by launches. During the period of the northeast monsoon from November to April the loading operations have to be suspended, and the ore is stockpiled to await the next shipping season.

Quality Control

As there is considerable variation in the chemical composition of the ore body, it is necessary to exercise a very close control over all stages of production from mining to shipping. The quality control system is based on the provision of a large number of chemical analyses on samples taken at different stages of production, the extensive use of stockpiles for the different grades of ore, and blending during the loading operations. To cope with the large numbers of analyses which are required, highly automated laboratories have been established at the mine and the loading centre on the coast. The bulk of the work is carried out with a direct reading spectrometer with tape machine, a cathode ray polarograph, and a spectrophotometer.

The ore is sampled at five separate stages;

1. During prospecting.
2. The debris from blast hole drilling.
3. On the conveyor belts which lead from the treatment plant.
4. At the rail loading point.
5. At the jetty loading point.

The results from (1) and (2) permit a close estimation to be made of the bulk composition of an individual blast, stage (3) checks the feed to the mine stockpiles, but stages (4) and (5) are the most important. Each ore train and lighter is sampled automatically during loading, a 200 kg sample being taken for every 75 tons loaded. The samples are crushed and split to 5 kg, dried and ground to -100 mesh and sent to the laboratory. Each train and lighter is additionally sampled for sizing tests. The rail samples are analysed for Fe, Cu, Zn, Bi, As and S, the results being ready within 90 minutes i.e. before the train reaches the coast, so the Quality Control Officer knows which stockpile is appropriate for each wagon load. The samples taken from the lighters are analysed for Fe, Cu, Zn and Bi, the results being ready before the lighter is unloaded. The lighter samples are later analysed for Fe, FeO, Cu, Zn, Bi, As, P, Mn, TiO_2, Al_2O_3, S, SnO_2, Pb, CaO, MgO and SiO_2, the composition of a complete shipment being calculated from weighted analyses. This final composition can be checked against analyses made by the purchaser at the unloading point, and in the event of a dispute an independent laboratory acts as arbitrator.

REFERENCES

Barth, T.F.W.

1962: Theoretical Petrology 2nd ed.; J.Wiley and Sons Inc., New York.

Barton, P.B.

1959: The chemical environment of ore deposition and the problem of low-temperature ore transport; in Researches in Geochemistry ed. Abelson; J.Wiley and Sons Inc., New York.

Basta, E.Z.

1957: Accurate determination of the cell dimensions of magnetite; Mining Mag., vol. 31, pp. 431-442.

1959: Some mineralogical relationships in the system Fe_2O_3-Fe_3O_4 and the composition of titanomaghemite; Econ. Geol., vol. 54, pp. 698-719.

Bateman, A.M.

1955: Economic Mineral Deposits 2nd ed.; J.Wiley and Sons Inc., New York.

Bean, J.H.

1966: Geology, Petrography and mineral resources of Pulau Tioman; Mem. Geol. Sur. Malaysia. (Manuscript in press)

Berry, L.G. and Thompson, R.M.

1962: X-Ray powder data for the ore minerals; Geol. Soc. Am., Mem. 85.

Black, J.M.

1952: Iron Hill; Brit. Col. Dept. Mines, Ann. Rep. 1952, pp.A221-228.

Bowen, N.L.

1940: Progressive metamorphism of siliceous dolomite and limestone; Jour. Geol., vol. 48, pp. 225-274.

Buddington, A.F.

1959: Granite emplacement with special reference to North America; Bull. Geol. Soc. Am., vol. 70, pp. 671-748.

1966: The Precambrian magnetite deposits of New York and New Jersey; Econ. Geol., vol. 61, pp. 484-510.

Burnham, C.W.

1967: Hydrothermal fluids at the magmatic stage; in Geochemistry of Hydrothermal Ore Deposits ed. Barnes; Holt, Rinehart and Winston, New York.

Buseck, P.R.

1966: Contact metasomatism and ore deposition: Concepcion del Oro, Mexico. Econ. Geol., vol. 61, pp. 97-137.

Callahan, W.H. and Newhouse, W.H.

1929: The ore deposits at Cornwall, Pa.; Econ. Geol., vol. 24, pp. 403-411.

Cameron, E.N.

1961: Ore Microscopy; J.Wiley and Sons Inc., New York.

Canavan, F. and Edwards, A.B.

1938: The iron ores of Yampi Sound, Western Australia; Austr. Inst. of Min. & Metall., N.S. No. 110, pp. 59-101.

Christophe-Michel-Levy, M.

* 1956: Reproduction artificielle des grenats calciques: grossulaire et andradite; Bull. Soc. Franc. Min. Crist., vol. 79, p. 124.

Cooke, S.R.B.

1936: Microscopic structure and concentratability of the important iron ores of the United States; U.S. Bur. Mines, Bull. 391.

Dana, E.S. and Ford, W.E.

1959: A Textbook of Mineralogy; Charles E. Tuttle Co., Tokyo. (Modern Asia Edition).

Davidson, A. and Wyllie, P.J.

1965: Zoned magnetite and platy magnetite in Cornwall type ore deposits; Econ. Geol., vol. 60, pp. 766-771.

Deer, W.A., Howie, R.A. and Zussman, J.

1962: Rock-Forming Minerals Vol.1 Ortho- and Ring Silicates.

1962: Rock-Forming Minerals Vol.3 Sheet Silicates.

1962: Rock-Forming Minerals Vol.5 Non-Silicates.

1963: Rock-Forming Minerals Vol.2 Chain Silicates. Longmans, Green and Co. Ltd., London.

Degens, E.T. and Ross, D.A. (eds.)

1969: Hot Brines and Recent Heavy Metal Deposits in the Red Sea; Springer-Verlag, New York.

Donnay, J.D.H. and Nowacki, W. (with Donnay, G.)

1954: Crystal data; Geol. Soc. Am., Mem. 60.

Donnay, G. and Kullerud, G.

1958: High-temperature chalcopyrite; Carnegie Inst. Wash., Ann. Rep. Direct. Geophys. Lab. 1957-1958, vol. 57, p. 246.

Eastwood, G.E.P.

1965: Replacement magnetite on Vancouver Island, British Colombia; Econ. Geol., vol. 60, pp. 124-148.

Fenner, C.N.

1933: Pneumatolytic processes in the formation of minerals and ores; in Ore Deposits of the Western States, Am. Inst. Min. Metall. Eng.

Fitch, F.H.

1941: Report on the iron ore deposits of Ulu Rompin, Pahang. Geol. Sur. Malaya; Unpublished report.

Freitsch, R.

1967: The relationship between magnetite and haematite in the iron ores of the Kiruna type and some other iron ore types. Sveriges Geol. Unders., C 625.

1970: Trace elements in magnetite and haematite mainly from northern Sweden; Sveriges Geol. Unders., C 646.

Garrels, R.M.

1957: Some free energy values from geologic relations; Am. Mineral., vol. 42, pp. 780-791.

1959: Rates of geochemical reactions at low temperatures and pressures; in Researches in Geochemistry ed. Abelson; J.Wiley and Sons Inc. New York.

Geijer, P. and Magnusson, N.H.

1952: The iron ores of Sweden; in Symposium sur les gisements de fer du monde, tome II; XIX Int. Geol. Cong., Algiers.

Gheith, M.A.

1952: Differential thermal analysis of certain iron oxides and hydroxides; Am. J. Sci., vol. 250, pp. 677-695.

Gilbert, G.

1925: Some magnetite-hematite relations; Econ. Geol., vol. 20, pp. 587-596.

Goldsmith, J.R.

1959: Some aspects of the geochemistry of carbonates; in Researches in Geochemistry ed. Abelson; J.Wiley and Sons Inc., New York.

Goranson, R.W.

1931: The solubility of water in granite magmas; Am. J. Sci., vol.222, pp. 481-502.

Grim, R.E.

1953: Clay Mineralogy; McGraw-Hill Book Co., London.

Gross, G.A.

1965: Geology of iron deposits in Canada Vol. 1 - General geology and evaluation of iron deposits; Geol. Sur. Can., Econ. Geol. Rep. No. 22.

Gruner, J.W.

1926: Magnetite-martite-hematite; Econ. Geol., vol. 21, pp. 375-393.

Gruner, J.W.

1939: Formation and stability of muscovite in acid solutions at elevated temperatures; Am. Mineral., vol. 24, pp. 624-628.

1944: The hydrothermal alteration of feldspars in acid solutions between 200°C and 400°C; Econ. Geol., vol. 34, pp. 578-589.

Hatch, F.H., Wells, A.K. and Wells, M.K.

1961: Petrology of the Igneous Rocks 12th ed.; Thomas Murby, London.

Hegmann, F. and Albrecht, F.

* 1954: Zur geochemie oxydischer eisenerze; Chenie d. Erda, vol. 17, pp. 81-103.

Helgeson, H.C.

1964: Complexing and Hydrothermal Ore Deposition; Pergammon Press, New York.

Hellner, E., Hinrichsen, Th. and Seifert,F.

1965: The study of mixed crystals of minerals in metamorphic rocks; in Controls of Metamorphism ed. Pitcher and Flinn; Oliver and Boyd, London and Edinburgh.

Hey, M.H.

1954: A new review of the chlorites; Min. Mag., vol. 30, pp. 277-292.

Hills, E.S.

1963; Elements of Structural Geology; Methuen & Co., London.

Hitchen, C.S. and Moss, F.

1954: (Title unknown) Colonial Development Corporation, London. Private Report.

Holland, H.D.

1959: Some applications of thermochemical data to problems of ore deposits - part 1; Econ. Geol., vol. 54, pp. 184-233.

Holser, W.T. and Schneer, C.J.

1961:Hydrothermal Magnetite; Bull. Geol. Soc. Am., vol. 72, pp. 369-386.

Jaeger, J.C.

1957: The temperature in the neighbourhood of a cooling intrusive sheet; Am. J. Sci., vol. 255, pp. 306-318.

Jagitsch, R.

* 1956: The synthesis of some skarn minerals from the powdered compnents; Arkiv. Kemi., vol. 9, p. 319.

Jones, C.R.

1970: The geology and mineral resources of Grik are, Upper Perak, Malaysia; Geol. Surv. Malays., Mem. 11.

Kalliokoski, J.

1965: The metamorphosed iron ore of El Pao, Venezuela; Econ. Geol., vol. 60, pp. 100-116.

Kerr, P.F., Kulp, J.L. and Hamilton, P.K.

1949: Differential thermal analysis of reference clay mineral specimens; Am. Petroleum Res. Inst., Res. Proj. No. 49; Colombia University, New York.

Koopmans, B.N.

1968: The Tembeling Formation - a lithostratigraphic description (West Malaysia); Geol. Soc. Malays., Bull. 1.

Korshinsky, D.S.

1964: Acidity conditions in post-magmatic processes; Report of XXII Int. Geol. Cong., Section 5, pp. 16-27. New Delhi.

Kranck, S.H.

1961: A study of phase equilibria in a metamorphic iron formation; Journ. Petrol., vol. 2, pp. 137-184.

Krauskopf, K.B.

1959: The use of equilibrium calculations in finding the composition of a magmatic gas phase; in Researches in Geochemistry ed. Abelson; J.Wiley and Sons Inc., New York.

1969: Thermodynamics used in geochemistry; in Handbook of Geochemistry ed. Wedepohl; Springer-Verlag, Berlin, Heidelberg, New York.

Kullerud, G.

1957: Annual Report of the Director of the Geophysical Laboratory 1956-1957; Carnegie Inst. Wash. Year Book, vol. 56, pp. 198-200.

1959: Sulphide systems as geological thermometers; in Researches in Geochemistry ed. Abelson; J.Wiley and Sons Inc., New York.

Landergren, S.

1948: On the geochemistry of Swedish iron ores and associated rocks; Sveriges Geol. Unders., C 496.

Larsen, E.S.

1945: Time required for the great batholith of southern and lower California; Am. J. Sci., vol. 243-A (Daly Volume), pp. 399-416.

Lepp, H.

1957: Stages in the oxidation of magnetite; Am. Mineral., vol. 42, pp. 679-681.

Loughnan, F.C.

1969: Chemical Weathering of the Silicate Minerals; American Elsevier Publishing Co. Inc., New York.

Lovering, T.G.

1958: Temperature and depth of formation of sulphide ore deposits at Gilman, Colorado; Econ. Geol., vol. 53, pp. 689-707.

Lundegardh, P.H.

1946: Geochemistry of magmatic iron ores; Nature, vol. 157, pp. 625-626.

Macandie, A.G. and Canavan, F.

1948: The iron-ore and manganese-ore deposits of Malaya; Broken Hill Propietary Ltd., Melbourne. (Private Report).

MacDonald, S.

1970: Geology and mineral resources of the Lake Chini-Sungei Bera-Sungei Jeram area of south-central Pahang. Geol. Surv. Malays., Map Bull. 1.

Mackin, J.H.

1968: Iron ore deposits of the Iron Springs District, southwestern Utah; in Ore Deposits in the United States (Graton-Sales Volume), Amer. Inst. Min. Engs., pp. 992-1019.

Magnusson, N.H.

1970: The origin of the iron ores of central Sweden and the history of their alterations; Sveriges Geol. Unders., C 643.

Mason, B.

* 1943: Mineralogical aspects of the system $FeO-Fe_2O_3-MnO-Mn_2O_3$; Geol. Foren. Forh., vol. 65, pp. 97-180.

1952: Principles of Geochemistry; J.Wiley and Sons Inc., New York.

McLeod, C.R.

1970: Some Canadian occurrences of maghemite; Geol. Surv. Can., Paper 70-7.

Merrin, S.

1960: Synthesis of epidote and its apparent P-T stability curve (abstr.); Bull. Geol. Soc. Am., vol. 65, pp. 1007-1032.

Metz, P.W. and Winkler, H.G.F.

* 1964: Experimentelle untersuchung der diopsidbildung aus tremolit, calcit, und quartz; Naturwissenschaft, Jahrg. 51, pt. 1o, pp. 1-3.

Milton, C., Axelrod, J.M. and Ingram, B.

1958: Bismutoferrite, chapmanite and "hypochlorite"; Am. Mineral., vol. 43, pp. 656-670.

Mueller, R.F.

1961: Oxidation in high temperature petrogenesis; Am. J. Sci., vol. 259, pp. 460-480.

Muffler, L.J.P. and White, D.E.

1968: Origin of CO_2 in the Salton Sea geothermal system, southeastern California, U.S.A.; Report of XXIII Int. Geol. Cong., Section 17, pp. 185-194; Prague.

Ovchinnikov, L.N.

1968: Geological and physico-chemical conditions of the formation of plutogenic hydrothermal deposits; Report of XXIII Int. Geol. Cong., Section 7, pp. 167-178; Prague.

Ovchinnikov, L.N., Shaliapnikov, D.S. and Shoor, A.S.

1964: Mobilization and transportation of matter in the endogenic ore-formation; Report of XXII Int. Geol. Cong., Part V, pp. 596-622; New Delhi.

Palache, C., Berman, H. and Frondel, C.

1944: System of Mineralogy, Vol. 1; J.Wiley and Sons Inc., New York.

1951: System of Mineralogy, Vol. 2; J.Wiley and Sons Inc., New York.

Park, C.F.

1961: A magnetite "flow" in northern Chile; Econ. Geol., vol. 56, pp. 431-436.

Park, C.F. and MacDiarmid, R.A.

1970: Ore Deposits 2nd Edition; W.H.Freeman & Co., San Francisco.

Ramberg, H.

1949: The facies classification of rocks, a clue to the origin of quartzo-felspathic massifs and veins; Jour. Geol., vol. 57, pp. 18-54.

Ramdohr, P.

1953: Ulvospinel and its significance in titaniferous iron ores; Econ. Geol., vol. 48, pp. 677-688.

1960: Die Erzmineralien und Ihre Verwachsungen; Akadamie Verlag, Berlin.

Rosenqvist, I.Th.

* 1952: The metamorphic facies and the feldspar minerals; Universitetet i Bergen Arbok, Naturr., No. 4.

Rosler, H.J. and Lange, H.

1972: Geochemical Tables; Elsevier Publishing Co., Amsterdam, London New York.

Roy, R. and Osborne, E.F.

1952: Studies in the system alumina-silica-water; in Problems of Clay and Laterite Genesis; Am. Inst. Min. Engrs., New York.

Sandell, E.B. and Goldich, S.S.

1943: The rarer metallic constituents of some American igneous rocks; Jour. Geol., vol. 51, pp. 99-115, 167-189.

Sangster, D.F.

1969: The contact metasomatic magnetite deposits of southwestern British Colombia; Geol. Sur. Can., Bull. 172.

Savage, H.E.F.S.

1937: The geology of the neighbourhood of Sungei Siput, Perak; Geol. Sur. Malaya, Mem. 1.

Schneiderhohn, H.

* 1961: Die Erzlagerstatten der Erden Volume II, Die Pegmatite; Stuttgart.

Schwartz, G.M.

1942: Progress in the study of exsolution in ore minerals; Econ. Geol., vol. 37, pp. 345-364.

Schwartz, G.M. and Ronbeck, A.C.

1940: Magnetite in sulphide ores; Econ. Geol., vol. 35, pp. 585-610.

Shadlun, T.N.

* 1953: The change in the structure of aggregates and in the internal structure of grains of chalcopyrite subjected to the influence of dynamic factors; Min. Sbornik, L'vov Geol. Soc., No. 7, pp. 75-80 (M.A. 13-36).

Shaw, D.M.

* 1964: Interpretation geochemique des elements en traces dans les roches cristallines; Masson, Paris.

Skinner, B.J., White, D.E., Rose, H.J. and Mays, R.E.

1967: Sulphides associated with the Salton Sea geothermal brines; Econ. Geol., vol. 62, pp. 316-330.

Stanton, R.L.

1972: Ore Petrology; McGraw-Hill Co., New York.

Stringham, B.

1952: Fields of formation of common hydrothermal alteration minerals; Econ. Geol., vol. 47, pp. 661-664.

Tillia, H.J.

1956: Geological, mining and economic study of Rompin iron ore deposit, Pahang, Malaya; Eastern Mining & Metals Co., Singapore. (Private Report).

Turekian, K.K. and Wedepohl, K.H.

1961: Distribution of the elements in some major units of the earth's crust; Bull. Geol. Soc. Am., vol. 72, pp.175-192.

Turner, F.J.

1948: Mineralogical and structural evolution of the metamorphic rocks; Geol. Soc. Am., Mem. 30.

1968: Metamorphic Petrology; McGraw-Hill Co., New York.

Turner, F.J. and Verhoogen, J.

1951: Igneous and Metamorphic Petrology; McGraw-Hill Co., New York.

1960: Ditto 2nd ed.

Tuttle, O.F. and Bowen, N.L.

1953: Annual Report of the Director of the Geophysical Laboratory 1952-1953; Carnegie Inst. Wash. Year Book, vol. 52, pp. 207-233.

United Nations

1970: Survey of world iron ore resources; Dept. of Econ. and Social Affairs, United Nations, New York.

Valeton, I.

1972: Bauxites; Elsevier Publishing Co., Amsterdam.

Van Rensburg, W.C.J.

1966: Supergene oxidation and hydration of magnetite of the Messina Formation, Soutpansberg District, Transvaal; Annals Geol. Sur. S. Africa, vol. 5, pp. 93-105.

Williams, H., Turner, F.J. and Gilbert, C.M.

1954: Petrography; W.H.Freeman & Co., San Francisco.

Winkler, H.G.F.

1965: Petrogenesis of Metamorphic Rocks; Springer-Verlag, New York.

1967: Ditto 2nd ed.

Yoder, H.S. and Eugster, H.P.

1955: Synthetic and natural muscovites; Geochim. Cosmochim. Acta, vol. 8, pp. 225-280.

Yoder, H.S. and Tilley, C.E.

1956: Annual Report of the Director of the Geophysical Laboratory 1955-1956; Carnegie Inst. Wash. Year Book, vol. 55, pp. 169-171.

Yund, R.A. and Kullerud, G.

1966: Thermal stability of assemblages in the Cu-Fe-S system; Jour. Petrol., vol. 7, pp. 454-488.

Zies, E.G.

1929: The Valley of Ten Thousand Smokes - I. The fumarolic incrustations and their bearing on ore deposition; Tech. Paper Nat. Geogr. Soc., vol. 1 No. 4, pp. 1-61.

Zussman, J.

1959: A re-examination of the structure of tremolite; Acta Crystallogr. vol. 12, p 309

* Not seen in the original publication.