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THE GEOLOGY OF THE BESSEMER MAGNETITE DEPOSIT

By

Ernest Henry Nickel, B. Sc.

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The author of this thesis holds the following degree:

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This thesis was prepared under the supervision of:

Professor H. S. Scott, Department of Geology.

Professor D. M. Shaw, Department of Geology.

Scope and contents of this thesis:

A field and laboratory study of the Bessemer magnetite deposit with a discussion of the possible origin of this deposit. Illustrated with maps and with photographs taken by the author.

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INTRODUCTION

General Comments

During the summer of 1950 while in the employment of the Ontario Department of Mines the writer spent some time in the examination of the Bessemer property. Using as a guide a base map kindly furnished by the Frobisher Exploration Company, Limited, the surface geology was closely studied, old information checked, and supplementary information added. Diamond drill core was examined and samples taken for laboratory study.

The laboratory work consisted of the compilation of information gained during the course of the field work, and an intensive microscopic study of thin and polished sections of ore, wallrock, and associated rocks. The Federov universal stage was used for the determination of the optics of a number of minerals.

History of the Bessemer Mine

The Bessemer property includes lots 2, 3, 4, and 5, concession VI; and lot 1, concession VII in Mayo Township, Hastings County, Ontario. A number of ore lenses of different sizes are known, but mining operations have been confined to four. This study is concerned with the largest and richest of these deposits, the Bessemer No. 4, situated on lots 4 and 5, concession VI.

This deposit, as well as others in the locality has been exploited intermittently since 1902 by a number of companies. The latest shipments from this mine were reported in 1914 when the total amount of ore shipped to that time was reported to be 92,413 tons (Lindeman and Bolton, 1917, p. 50). The deposit has been developed by a pit 275 feet in length (Plate

I figure 1) and averaging 50 feet in width, as well as by a shaft and three levels. During 1941 and 1942 the Frobisher Exploration Company Limited carried out an extensive surface and underground exploration programme including diamond drilling. Based on this work, the ore reserves were estimated to be 757,500 gross tons before dilution, averaging 42.18% iron, 0.86% sulphur, 0.018% phosphorus, and 15.9% silica. Including ten percent dilution from walls the tonnage is 833,250, averaging 39% iron and 0.78% sulphur (Dadson, 1942, p. 3).

At the present time the open pit and the underground workings are flooded and are inaccessible.

Acknowledgments

This work was made possible by the cooperation of the Frobisher Exploration Company who gave permission to undertake this study of their property and who gave the writer complete access to all information regarding it. Louis Moyd, geologist, was particularly helpful in this respect and in discussing some of the aspects of the problem. The writer's wife, Mrs. Muriel Nickel, was of great assistance in several phases of the work, especially in the preparation of the photomicrographs. The writer wishes to thank Dr. H. S. Armstrong, Professor H. S. Scott, and Professor D. M. Shaw of the Department of Geology, McMaster University for helpful suggestions and criticisms during the course of the laboratory work. Informal discussions with the writer's colleagues of the problems attending this study were both helpful and informative.

REGIONAL GEOLOGY

The Bessemer iron deposit lies within the Grenville sub-province of the Canadian Precambrian Shield. The rocks form part of the Grenville series underlying a large part of southeastern Ontario and extending into the Adirondack region of the United States of America.

The Grenville series consists of a highly altered complex of sediments, volcanics, and intrusives. The sediments include limestones and dolomites ranging from pure crystalline types to highly impure calcareous gneisses and amphibolites; paragneisses characterized by an abundance of secondary minerals; and small areas of conglomerate. Volcanics include basic flows and tuff beds, locally considerably altered to hornblende-chlorite schists. The sediments and volcanics are highly deformed and folded, and dips are commonly vertical or near-vertical. A large part of the area is underlain by granite gneiss, pegmatite and hybrid acid gneiss. Other intrusives include numerous gabbro and diorite stocks and dikes. Of particular interest to geologists have been the nepheline syenites and pegmatites occurring in the area.

BESSEMER NO. 4 DEPOSIT

SCALE OF FEET
0 100 200 300

LITTLE MULLET LAKE

ORE DUMP

LEGEND














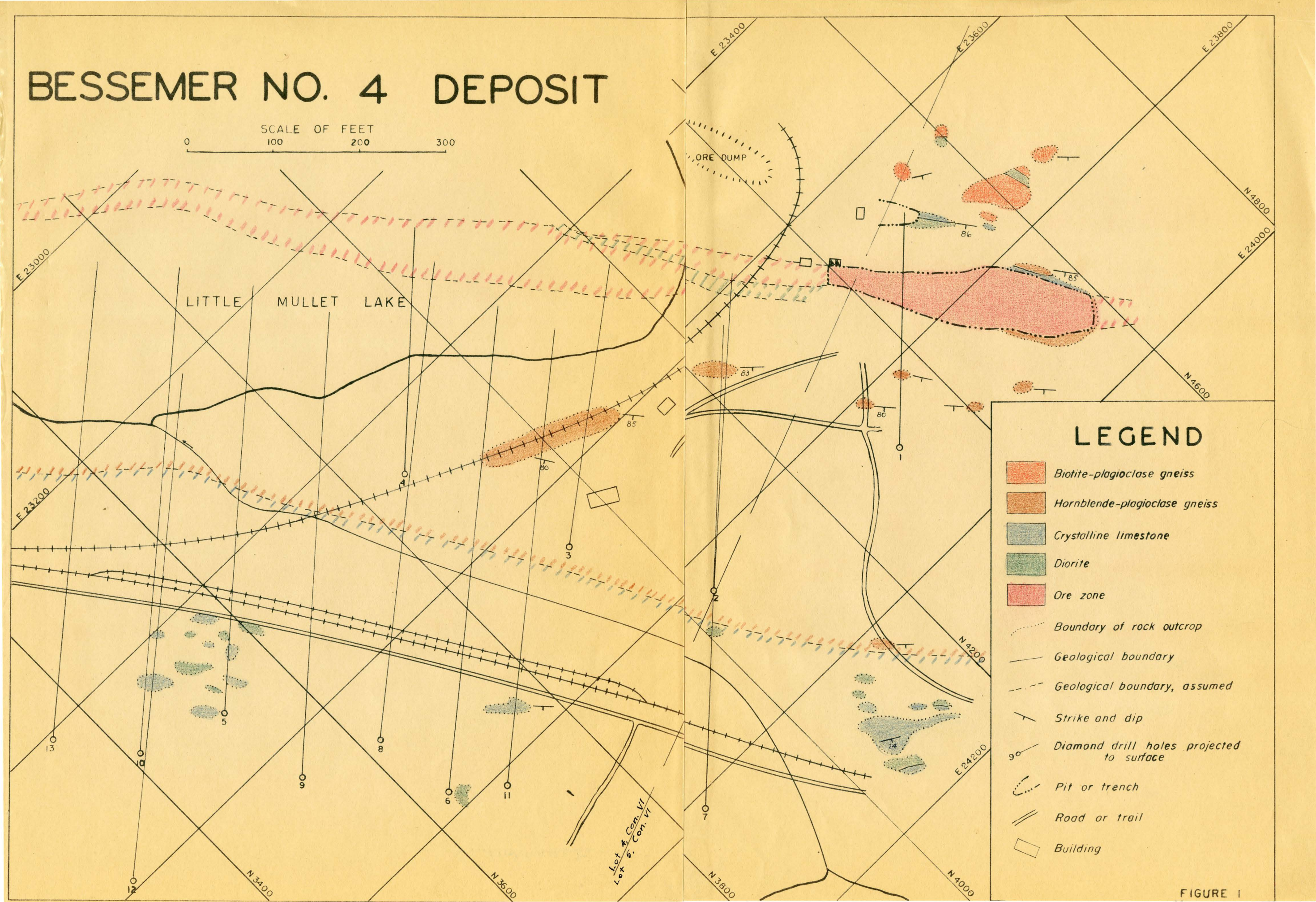
-  Biotite-plagioclase gneiss
-  Hornblende-plagioclase gneiss
-  Crystalline limestone
-  Diorite
-  Ore zone
-  Boundary of rock outcrop
-  Geological boundary
-  Geological boundary, assumed
-  Strike and dip
-  Diamond drill holes projected to surface
-  Pit or trench
-  Road or trail
-  Building

FIGURE 1



GEOLOGY OF THE BESSEMER AREA

The rocks of the area vary in strike from $N30^{\circ}E$ to $N60^{\circ}E$ and dip to the southeast at an average of seventy-five degrees. The rock types of the area consist of hornblende-plagioclase and biotite-plagioclase gneiss, crystalline limestone, a mineralized skarn zone, and diorite. The gneisses, limestone, and skarn zone are conformable, whereas the diorite dikes cut across these rocks. The field relationships are shown on the map page 4.

Hornblende-Plagioclase Gneiss

The hornblende-plagioclase gneiss forms a band from 300 to 400 feet wide and is bounded by crystalline limestone to the southeast and by the skarn to the northwest. The gneiss is strongly foliated parallel to its contacts.

The bulk of this gneiss is coarsely crystalline with a grain size averaging about 0.05 cm. in diameter and consists mainly of plagioclase, quartz, calcite, and hornblende. (Plate II, figure 3). The plagioclase is clear and unaltered; twinning is present in some grains, but lacking in others. Optical determinations indicated a $(-)$ 2V of eighty-eight degrees and an extinction angle (between Z and O10) of fifty-nine degrees, corresponding to $Ab_{28}An_{72}$. Quartz and calcite are present in varying amounts throughout the gneiss. The hornblende is dark in colour and shows strong pleochroism from light yellow-green to dark blue-green. It has variable optical properties with 2V ranging from $(-)$ sixty-seven degrees to $(-)$ seventy-seven degrees, and the extinction angle from ten to twenty-nine degrees. This corresponds roughly to an amphibole ana-

lyzed by A. M. Billings (1928, pp. 290, 291) which he terms "Femaghasting-site" and which has a composition of:

CaO.....	12.42%
MgO.....	7.26
Fe ₂ O ₃	6.21
FeO.....	15.85
Na ₂ O.....	3.21

Other minerals in the gneiss include a pale green diopside, some strongly pleochroic biotite, orthoclase and microcline, and scattered grains of pyrite and magnetite.

There are a few narrow limestone members in the hornblende-plagioclase gneiss which are entirely conformable and which exhibit sharp boundaries against the gneiss. An outcrop of one of these bands observed near the eastern corner of the large pit is shown in (Plate I, figure 2).

There is a fine-grained band near the center of the coarse gneiss exposed in the railroad cut (23575E; 4050N). This is composed largely of a fine-grained equigranular aggregate of mineral grains about 0.005 cm. in diameter. The minerals include quartz, plagioclase, sericite, and a dark opaque mineral believed to be ilmenite. (Plate II, figure 4). It is prismatic in outline and has a dense white alteration. These ilmenite grains are distributed fairly evenly among the other minerals, and have a common orientation. Small sericite flakes with the same orientation have developed a foliation in the fine-grained rock. Veinlets of quartz and plagioclase cross this fine-grained material.

It is believed from the evidence available that this gneiss is of sedimentary origin, hence a paragneiss. It lies conformably between limestone members of sedimentary origin. The banding caused by the ar-

arrangement of light and dark minerals is parallel to the contacts and may be relic bedding.

The original rock is believed to have been a graywacke with a fine-grained silt member and several thin limestone members.

Biotite-Plagioclase Gneiss

In the northwestern part of the map sheet is gneissic rock that has been mapped as granite by Lindeman and Bolton (1917) and by geologists of the Frobisher Exploration Company. Microscopic examination of several specimens, however, showed that the rock is composed largely of quartz, untwinned plagioclase feldspar of similar composition to the plagioclase in the paragneiss to the southeast, and biotite, with a few calcite and magnetite grains. The rock is quite equigranular and exhibits some foliation due to the arrangement of the biotite. The contact between this gneiss and the limestone is nowhere exposed, but the general field relations show that it is roughly conformable. The gneissosity is parallel to the regional structure. This is obviously not a normal granite, but may be a contaminated phase of granite. Unfortunately, because of lack of information regarding the rock to the northwest, it is not known whether this gneiss grades into a normal granite.

Crystalline Limestone

The limestone is present in two bands: a wide belt occupying the southeastern part of the area mapped, and a belt less than 100 feet wide just to the northwest of the skarn zone. All of the limestone is coarsely crystalline and is generally quite massive. Most has been converted to a pure white marble, but there are some dark impure bands consisting

mostly of lime silicates. Some of the limestone adjacent to the skarn zone has been silicated, with the formation of tremolite, but in several places along the contact the limestone is simply recrystallized. The limestone is cut in a number of places by small veinlets of pure white quartz; around these has been developed a zone of tremolite from one to two inches in width.

Diorite

A large number of diorite dikes are found in the area, and these cut at an angle all the rocks previously mentioned. The margins against the intruded rock are in general quite sharp, and almost without exception fine-grained. However, one contact exposed by a trench about 100 feet northwest of the main pit shows dark hornblende grains impregnating the adjoining crystalline limestone for a width of about one foot (Plate III, figure 7). The dikes are quite irregular in shape, frequently bifurcating and changing in strike. Hence it is not possible to obtain a clear idea of their over-all orientation. The diorite exhibits a slight foliation parallel with the walls of the dikes.

The diorite is made up chiefly of untwinned plagioclase and a highly coloured, pleochroic hornblende. Accessory minerals include brown biotite, quartz, titanite, magnetite, and rare pyrite. Some of the biotite is evidently secondary after hornblende. The magnetite is closely associated with the titanite, grains of the former in a number of cases being enclosed in the latter mineral. This is rather suggestive of ilmenite or titaniferous magnetite breaking down to form titanite and magnetite.

The diorite appears to represent the last stage of igneous ac-

tivity in the area, and must have been intruded considerably later than the metamorphism of the other rocks in the area. This is inferred by injection of the dioritic magma along fractures cutting the foliation of the other rock as well as the skarn zone.

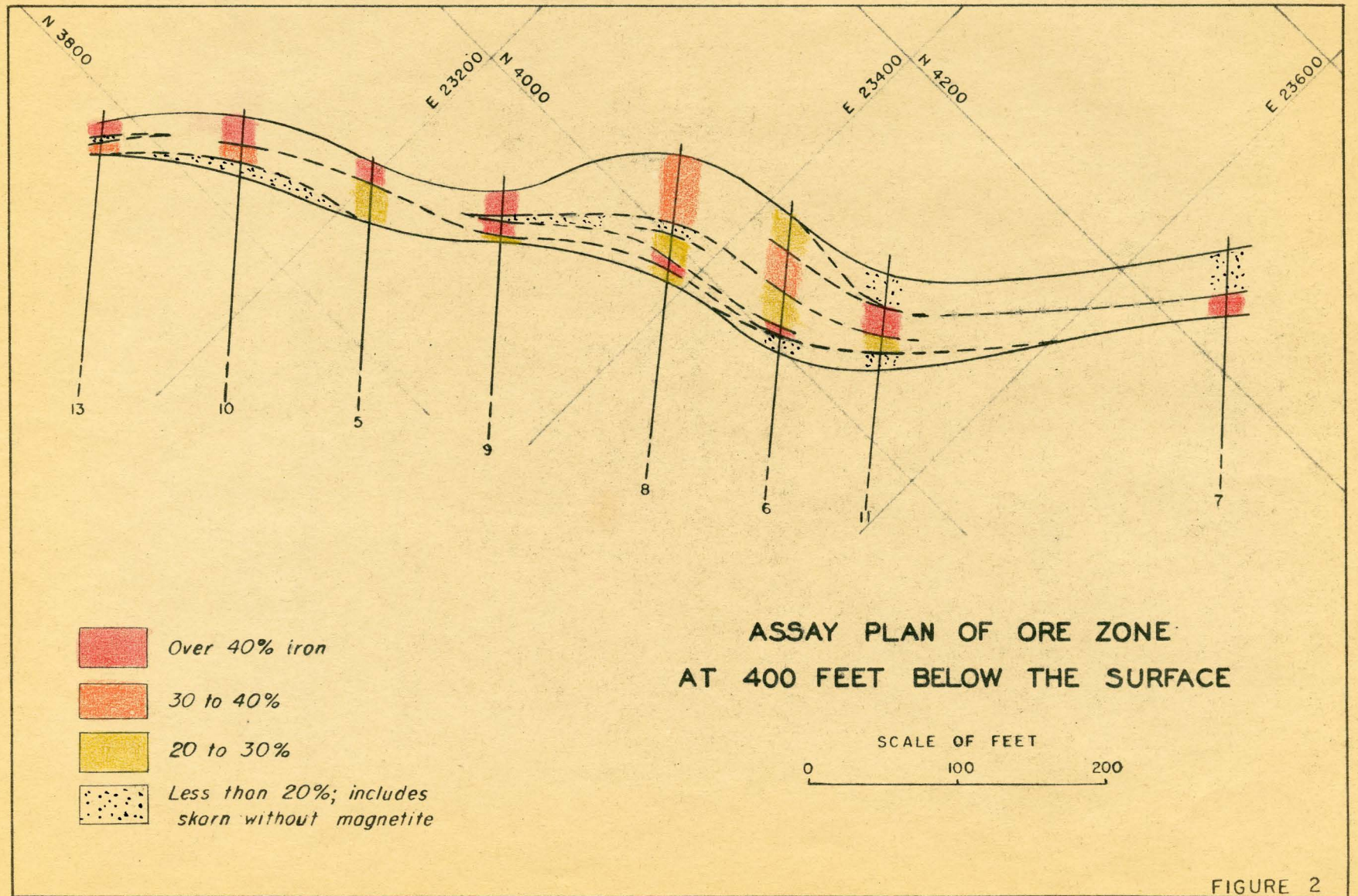


FIGURE 2

GEOLOGY OF THE ORE ZONE

Structure

The ore zone is here considered as the zone of skarn development. Most of the skarn zone contains abundant magnetite, although in some cases the skarn extends up to twenty-five feet beyond the margins of the magnetite. The skarn zone is sinuous in shape and lies conformably between crystalline limestone to the northwest and paragneiss to the southeast. It dips to the southeast at seventy-three to eighty-one degrees (see figure 2). Evidence of interbed movement was observed along the southeastern margin of the body. Although it has been exposed at the surface for only about 300 feet, the entire zone, as shown by diamond drilling and magnetometer survey has a total length of about 4200 feet. The width varies from about ten to ninety feet.

There is an irregular banding in the skarn zone produced by the arrangement of the minerals (Plate III, figure 8). This is more or less parallel to the walls of the skarn zone and may be a relic bedding. The magnetite itself is irregular in distribution both in horizontal and vertical section (See figure 3). At the southwestern end of the skarn zone the magnetite is immediately adjacent to the limestone lying to the northwest. At the northeastern end, however, the magnetite is separated from the limestone by a twenty-five foot width of skarn.

Mineralogy

The distribution of minerals in the skarn zone is very irregular. This is shown in part by figure 4 which represents a cross-section of the skarn zone taken along diamond drill hole number 9. The percen-

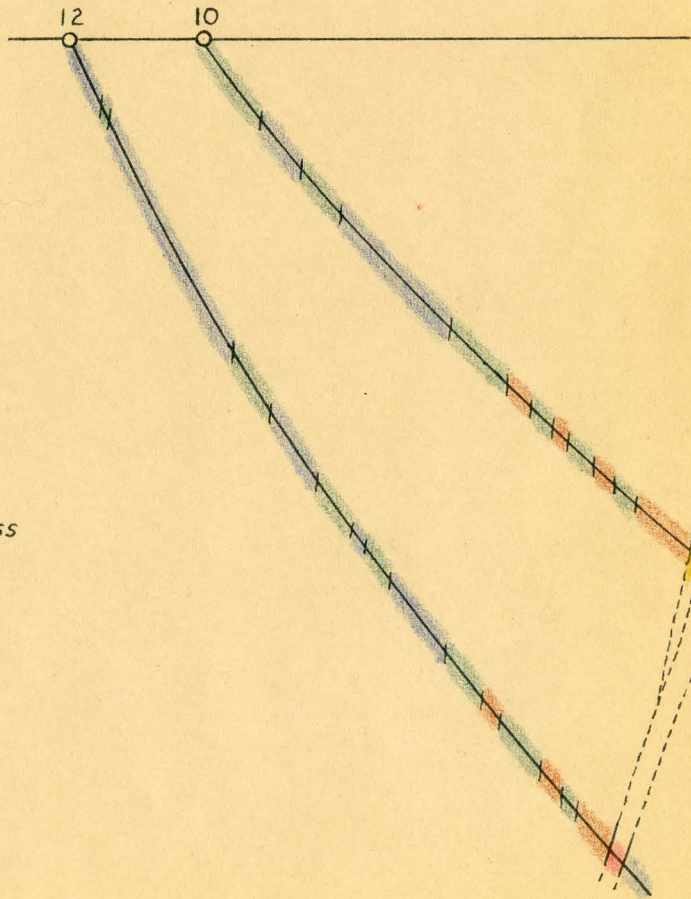
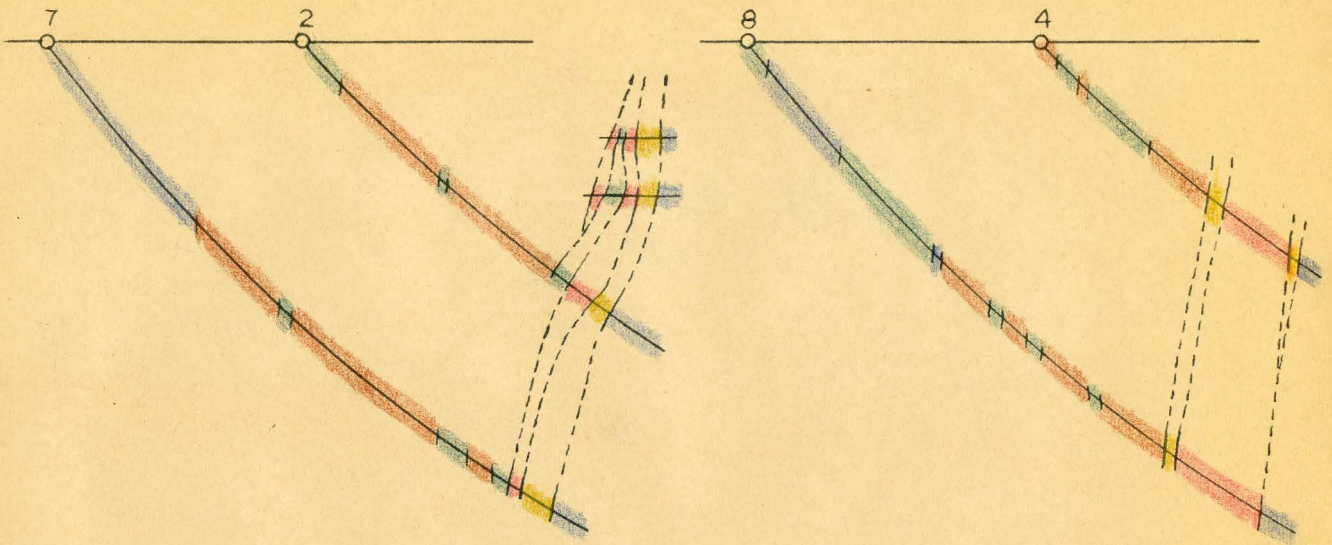







FIGURE 3

D.D.H. SECTIONS

-  *Crystalline limestone*
-  *Diorite*
-  *Hornblende-plagioclase gneiss*
-  *Skarn*
-  *Magnetite zone*

SCALE OF FEET
0 200 400

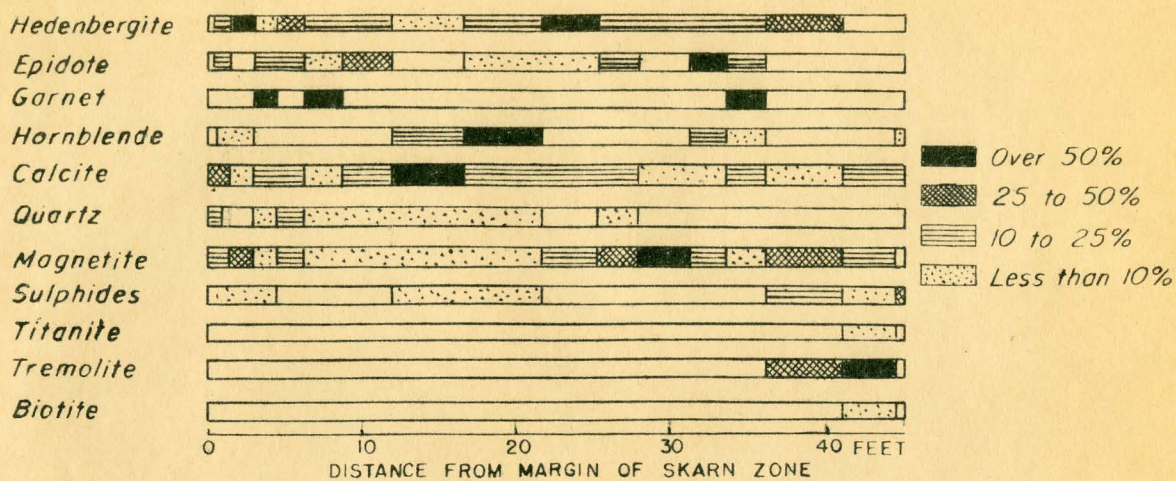


Figure 4: Mineral distribution along D.D.H. 9

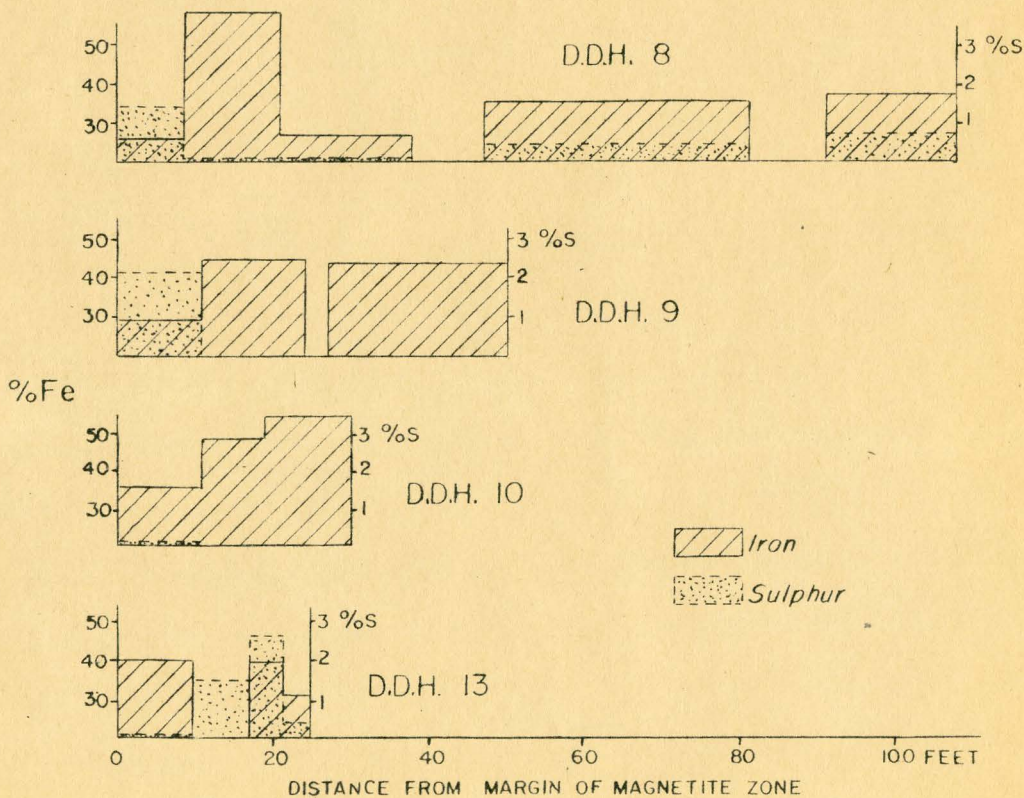


Figure 5: Relationship of iron and sulphur along four diamond drill hole sections

tages are approximate as they were estimated from thin sections made from specimens of the core taken at intervals, but they serve to illustrate the irregular distribution of the minerals.

Magnetite

As stated previously, magnetite is not found across the entire width of the skarn zone. Where the magnetite is rather sparsely disseminated it occurs as scattered grains commonly subhedral to euhedral (Plate IV, figure 9). Where it is more abundant, it occurs as ragged masses intricately intergrown with gangue minerals (Plate IV, figures 5 and 6). Where hornblende and pyroxene are both present the magnetite is found largely with the latter. Massive magnetite devoid of gangue minerals was nowhere found. The magnetite, in large part, at least, seems to be contemporaneous with the gangue minerals.

There is a late stage magnetite, present only in microscopic quantities, which occurs as tiny replacement veinlets cutting and replacing gangue minerals as well as sulphides (Plate IV, figure 10). Some sulphide grains are entirely surrounded by magnetite. (Plate IV, figures 11 and 12). A number of magnetite veinlets contain blebs of chalcopyrite and pyrrhotite, and one veinlet was observed having a margin of magnetite and a core of chalcopyrite (Plate V, figure 13). It appears, then, that the late magnetite is mostly later than the sulphides, but that some of it is contemporaneous with and even somewhat earlier than chalcopyrite. B. S. Butler (1927) suggests that the appearance of late magnetite represents a reversal of the normal order of deposition and is the result of a temporary or local increase in temperature (p. 235).

Sulphides

The sulphides constitute a minor part of the mineral assemblage of the skarn zone. The sulphur content of the ore is in most cases less than one percent, and only rarely and locally does it increase to over two percent.

Figure 5 attempts to show the quantitative relationship between iron and sulphur along several diamond drill holes. Most of the iron is present in magnetite and probably all the sulphur in sulphides, so that this chart may well be taken as representing the quantitative relationship of magnetite and sulphides. These values were obtained from ore assays made for the Frobisher Exploration Company, and make it evident that there is no uniform quantitative relationship between magnetite and sulphide.

Pyrite. This mineral is of common occurrence in the orebody, and is generally present as irregular grains. In most cases there is little evidence to suggest replacement by this mineral. However, in a few cases tiny pyrite veinlets cut across gangue minerals and occasionally surround them or embay them (Plate V, figure 14). Several veinlets were found cutting magnetite. Some rounded and angular grains of magnetite are in a number of places found entirely enclosed in large grains of pyrite.

Pyrite is commonly replaced by other sulphides, notably chalcopyrite and pyrrhotite, and the pyrite is observed to be veined by these sulphides and is frequently left as partly replaced remnants in them (Plate V, figure 15).

Pyrrhotite. Pyrrhotite is quite common in the ore zone and is present in irregular masses. It replaces gangue minerals, but its re-

relationship to magnetite is more difficult to determine. The contacts between the two are generally smooth and rounded, suggesting contemporaneous deposition. One section, however, shows an euhedral crystal of magnetite almost entirely surrounded by pyrrhotite. Also, where magnetite and gangue are adjacent, tongues of pyrrhotite are commonly found injected at the contact, suggesting replacement by pyrrhotite along the margins of the grains. Although the evidence is not conclusive, it appears that pyrrhotite is later than magnetite. This is shown indirectly by its relation to pyrite which it replaces (Plate V, figure 15), and which in turn cuts across magnetite. Pyrrhotite is replaced by later chalcopyrite (Plate V, figure 16).

Pentlandite. This mineral has not been positively identified, as it is only present in very small quantities, and is found in only one section. It is believed to be pentlandite because of its hardness, colour and etch test results. It is found only in very close association with pyrrhotite. It occurs as replacement rims around pyrrhotite and contains remnants of this mineral (Plate VI, figure 17). It is in turn cut and replaced by chalcopyrite.

Chalcopyrite. Chalcopyrite is present in veins and masses. Replacement veinlets of this mineral cut and replace gangue minerals (Plate VI, figures 18 and 19), magnetite (Plate VI, figure 20), pyrite, pyrrhotite and pentlandite. Chalcopyrite is more closely associated with pyrrhotite than with any other mineral and frequently surrounds and veins it (Plate V, figure 16). It commonly contains pyrrhotite remnants.

Chalcopyrite is the latest of the sulphides and the latest metal-

lic mineral with the exception of late stage magnetite.

Non-metallic Minerals

Garnet. Garnet is by no means found throughout the ore zone, but is restricted to a number of lenses. Hence in quantity it is a minor constituent of the skarn zone.

The garnet is dark and massive in hand specimen, red-brown in thin section, and completely isotropic in polarized light. It is darker in colour along some fractures, suggesting addition of iron. The garnet belongs to the ugrandite group (Winchell, 1933, p. 180), particularly rich in andradite. This was established by the high refractive index of the garnet, considerably above 1.85, and its magnetism. Blow-pipe tests showed that the garnet fuses easily to a black magnetic globule, characteristic of andradite.

The garnet is irregularly fractured and some of the fractures are in part filled by epidote and hedenbergite (Plate VII, figure 21). In some instances, however, veins of garnet cross a fine-grained aggregate of epidote and hedenbergite. This indicates more or less contemporaneous crystallization of these minerals. In a number of cases the garnet is cut by numerous calcite veinlets (Plate VII, figure 21).

Epidote. This mineral is, with few exceptions, found across the entire width of the skarn zone. It is usually present as small grains, but one section shows large recrystallized grains. Epidote is very fresh in appearance, bright yellow in colour, and quite pleochroic in thin section. Universal stage determinations show that it has a $(-)$ 2V of seventy-two degrees, and the angle between Z and OOL cleavage is twenty-nine

degrees. These optics correspond to an epidote containing about fourteen percent iron by weight (Winchell, 1933, p. 313), hence quite iron-rich.

The minerals most closely associated with epidote are calcite and hedenbergite; in fact, this association is almost universal in the skarn zone. These minerals show mutual boundaries. One section, however, shows a fine-grained veinlet of epidote and calcite cutting an aggregate of magnetite, hedenbergite, calcite, and epidote grains (Plate VII, figure 22). This suggests an epidote of slightly later age.

Pyroxene. Pyroxene is found across the entire skarn zone. It is commonly bright green in colour and in some cases faintly pleochroic. Universal stage determinations show that it has a 2V of fifty-nine degrees, and the direction Z makes an angle of fifty-two degrees with c. It is thus established as almost pure hedenbergite (Winchell, 1933). It is usually present as small scattered grains, although several sections show quite massive pyroxene aggregates.

In some cases the pyroxene has been partly replaced by a pleochroic hornblende (Plate VII, figure 23), but one section shows definite evidence of hornblende being replaced by pyroxene (Plate VII, figure 24). This indicates a reversal of the normal order of alteration.

The pyroxene of the skarn zone is distinguished from that outside the zone by its green colour and by its larger extinction angle.

Amphibole. The amphibole is a highly coloured, extremely pleochroic hornblende quite similar to that found in the adjoining paragneiss. It is not very common in the skarn zone, and its most common occurrences

is an alteration product of pyroxene. Some of the hornblende contains numerous round inclusions of calcite (Plate VIII, figure 25).

Light green amphibole is uncommon, but locally found. It lacks the bright colour of the common hornblende, but shows slight pleochroism. Its optics, $(-)$ 2V of seventy-seven degrees, and extinction angle of about twenty degrees designate it as a member of the tremolite-actinolite series. There appears to be a characteristic lack of magnetite associated with this amphibole; adjacent pyroxene rock contains abundant magnetite.

Quartz. Quartz is common as scattered grains throughout most parts of the skarn zone. It is fresh and unaltered in appearance.

Calcite. Calcite is present in all parts of the skarn zone. The grains are commonly large and exhibit glide twinning. Calcite is interstitial to all other minerals in the zone, and in a few cases occurs in veinlets cutting the silicates (Plate VIII, figure 21). In some sections it is present as rounded inclusions in hornblende (Plate VIII, figure 25). This suggests either partial replacement of calcite by hornblende, or the separation of calcite from hornblende.

Chlorite. Chlorite was found in only one section. It is rather fibrous, and the fibres are arranged more or less normal to the veinlet walls (Plate VIII, figure 26). The chlorite is commonly found at the boundary between hornblende and calcite grains, and is not found in the absence of either of these minerals. This suggests that the chlorite has been formed by reaction between calcite and hornblende.

Titanite. This mineral is rare in the skarn zone and was observed

only in one thin section. It occurs in irregular masses and is altered in part to calcite and a yellow-brown alteration.

Biotite. This mineral is also quite rare in the skarn zone. The biotite is light green and pleochroic, but is altered around the margins of grains to a dark green pleochroic mica (Plate VIII, figure 27). This may be due to hydrothermal alteration (Winchell, 1933, p. 275).

Summary of Ore Mineralization

The gangue minerals are characterized by richness in iron and calcium. Garnet and, to some extent, hedenbergite show enrichment in iron along fractures. Calcite and quartz have been recrystallized as shown by their large grain size and fresh appearance. They are intergrown with the ferromagnesian minerals, suggesting a contemporaneous relationship. Two ages of magnetite are present. The earlier magnetite appears to be largely contemporaneous with the gangue minerals, while the later magnetite occurs in veinlets cutting gangue and sulphides. The sulphides include pyrite, pyrrhotite, chalcopyrite, and possibly pentlandite. They are present as replacement masses and in veinlets cutting gangue and early magnetite. The following table illustrates the sequence of metallic minerals.

TABLE 1

Gangue minerals	
Magnetite	. .?.
Pyrite	
Pyrrhotite	
Pentlandite (?)	..	
Chalcopyrite	

The formation of the skarn and accompanying mineralization re-

sulted in the obscuring of original bedding, if any was present, and the development of an irregular banding roughly parallel to the long dimensions of the body. (Plate III, figure 8). This, and the presence of garnet veinlets cutting pyroxene and epidote; and epidote, pyroxene, and calcite veinlets cutting garnet, suggests that physical readjustment was taking place during development of the skarn body.

The sulphides were deposited after the recrystallization of gangue minerals and magnetite. The late-stage magnetite may have been introduced by iron-rich solutions, or by re-solution of pre-existing magnetite. This action is probably closely related to the alteration of garnet and hedenbergite along fractures by iron-rich solutions.

SUGGESTED ORIGIN OF THE BESSEMER MAGNETITE DEPOSIT

Possibly Hypotheses Reviewed

There is, in general, a sharp division between titaniferous and non-titaniferous magnetite deposits. The former are associated with basic intrusives; the latter, more commonly with processes of metamorphism or metasomatism caused by acid or intermediate intrusives. The Bessemer iron deposit contains no appreciable titanium, so in the discussion which follows, it will be compared with deposits of non-titaniferous magnetite.

A number of different causes have been ascribed to the non-titaniferous magnetite deposits which have been studied to date. The most important theories are the injection, metasomatic replacement, and syngenetic theories.

Injection of Magnetite as a Magmatic Differentiate

The proponents of this theory for the origin of non-titaniferous magnetite ores believe that the ore is a late differentiate of an igneous body. Geijer (1931) attributes this origin to the Kiruna iron ores, connecting them with igneous rocks of an intermediate or moderately siliceous character. Zavaritsky (1927) found fragments of wallrock suspended in the magnetite of some Ural Mountain deposits. In northern New Jersey there are a number of occurrences of magnetite in pods conformable to the enclosing gneiss. This magnetite, L. L. Smith (1933) believes, was formed by emplacement by end-phase aqueo-igneous differentiates.

This theory requires support by emplacement relations or similar

evidence. Such textures might include brecciation of wallrock, sharp contacts between magnetite and wallrock, fragments of wallrock in the magnetite, and cross-cutting relationship of magnetite. High temperature alteration would be probable. None of these features have been observed in the Bessemer deposit; hence the theory of injection is unlikely for this deposit.

Metasomatic Replacement

Metasomatic replacement or pyrometasomatism involves the "replacement of the enclosing rock with addition of substance. They (the deposits) are formed at high temperature by emanations issuing from the intrusive" (Lindgren, 1933, p. 696). The intrusive is generally of acid to intermediate composition, while the host rock must be one favourable for replacement, such as an impure limestone. Knopf (1933) states that the "distinguishing feature of pyrometasomatic deposits is the association of metalliferous minerals with the calcic silicates of the so-called contact-metamorphic class of minerals" (p. 538). The calcic silicates include andradite-rich garnet, pyroxene, amphibole, and epidote.

The general paragenesis of these minerals formulated by V. M. Goldschmidt in the Oslo region showed that magnetite and iron sulphides were partly older and partly younger than the skarn silicates (Knopf, 1933). Most later writers (Behre, Osborn and Rainwater, 1936; Schmidt, 1939; and Holser, 1950) found that the silicates were developed before the deposition of magnetite. Kihlstedt (1948), however, found that the skarn silicates were formed after deposition of the iron and sulphur.

The nature of the wallrock is of great importance in the process of metasomatism. H. Schmidt (1939) found that epidote invariably selects

aluminous rocks, shale, and shaly limestone; garnet avoids the igneous rock and shale and selects limestone or slightly shaly limestone; the best ore favours the purer limestone.

The problem of origin of the iron is a difficult one. Skarn development is caused almost entirely by acid to intermediate intrusives. Yet richness in iron is not characteristic of granite liquids (Reynolds, 1946). Neither Miller (1921) nor Holser (1950) found any magnetite in the intrusives near magnetite bodies in the Adirondack region and Montana respectively. Grout (1923), however, found what he considered primary magnetite to the extent of ten percent in otherwise normal Vermilion granite in Minnesota, and granite pegmatites containing considerably more magnetite. There is some doubt as to whether this magnetite is of primary origin (Miller, 1923). Some writers have obviated this difficulty by postulating the "leaching" of iron from iron-rich rock by igneous emanations and its concentration and redeposition by replacement farther along (Miller, 1919 and Moyd, 1949). This is supported in part by observations of Backlund (1936), Currier (1938) and Reynolds (1946) who discerned a "basic front" development of mafic minerals some distance in front of advancing granitic magma. But this theory is difficult to prove in the case of a specific iron deposit unless definite evidence of impoverishment in iron in the nearby area is established.

Many of the Adirondack non-titaniferous magnetite deposits have been attributed to a pyrometasomatic origin (Callagher, 1937; Colony, 1921; and Alling, 1939). To date no intensive study of the non-titaniferous magnetite deposits of southeastern Ontario has been made, but the consensus of opinion favours pyrometasomatic replacement (Thomson,

1943; R. E. Jones--personal communication--1950; and L. Moyd--personal communication--1951).

Syngenetic Origin

The basic assumption of this theory is that the iron was present in some form in the rock before metamorphism, and that the existing ore-body is this material in its original or reconstituted state. Nason (1922) believes that the magnetite deposits in gray gneisses in the Adirondack region are contemporaneous in origin with their host rocks which he considers metamorphosed sediments. Other writers, however, are opposed to this theory (Miller, 1922; and Newland, 1923). Magnusson (1936) found skarn iron ores showing a transition to banded hematite ores of sedimentary origin in central Sweden, and considers that granite emanations formed the skarn minerals and altered the pre-existing iron ores by a "regional thermo-metamorphism, through reaction between already existing substances such as oxide, hydroxide, and carbonate of iron (and manganese), silica, and carbonates of calcium and magnesium." Geochemical trace element analyses of Swedish iron ores by Landergren (1948) showed that there was a closer genetic relationship between the skarn magnetite and the surrounding wallrock than between it and the granitic intrusions. E. L. Bruce (1922) has described lenses of magnetite and hematite interbedded with gneisses. He considers that the iron was deposited as hydrated oxides or as carbonate, and altered to magnetite and hematite through regional disturbances. Ramberg (1948) suggests that the formation of a granulite from a hornblende-biotite-augite rock results in the release of iron and titanium, forming titaniferous magnetite from basic rocks. The same principle could conceivably

be applied to non-titaniferous deposits.

If the iron has been deposited as ferric oxide, regional metamorphism can effect the change to magnetite, for when a deposit of hydrated ferric oxide undergoes metamorphism, there is a reduction of limonite to hematite, and then of hematite to magnetite (Harker, 1939, p. 64). Buddington (1939) found a distinct increase of ferrous iron at the expense of ferric in passing from zones of lower to higher metamorphism in the Adirondack region. Kuhara, experimenting on the hydrothermal precipitation of magnetite and hematite found that magnetite is formed when siderite or other ferrous minerals are present either in the sediment or the igneous body. (Behrend and Berg, 1927, p. 167). Thus either ferric oxides or ferrous carbonate could be changed to magnetite through regional metamorphism.

Discussion of Origin of the Bessemer Magnetite Deposit

Any theory or group of theories taken to explain the origin of the Bessemer magnetite deposit should be able to account for all of the structural and mineralogical features which characterize it. These features include:

- Relationship with the adjoining rocks
- Shape of the magnetite body
- Variability of the iron content
- Internal structure of the orebody
- Relationship of skarn and magnetite
- Composition of the ore zone
- Evidence of hydrothermal activity
- Late-stage magnetite

The magnetite is conformable with the metamorphosed sediments on either side. Conformability would be expected in the case of syngenetic, replacement, or injection deposits. If the iron were present

as a sedimentary band it would be conformable. On the other hand, the conformability might represent a bed especially favourable, chemically or physically, for replacement. Or, if injected, the ore would tend to follow the previous foliation.

Another structural feature is the position of the magnetite band relative to the other rocks, that is, between gneiss and limestone. This may represent a change in sedimentation, perhaps due to a different source or a change in conditions of deposition. In the case of the Bessemer area, the order would probably be limestone, followed by ferruginous carbonate, succeeded in turn by graywacke, reflecting shallowing of the basin of deposition; if the beds are overturned, the order would be reversed, suggesting deepening of the basin.

Since the skarn zone is at the contact between two different rock types, there is a possibility that this contact afforded easy passage to altering solutions. However, this would not explain the position of the magnetite, as the magnetite zone is not adjacent to the limestone along its whole length, but diverges somewhat at its northeastern end. This slight divergence is possible when considering a sedimentary origin for the iron, as this might represent a thickening of a sedimentary wedge between the iron band and the limestone. The long, sinuous shape of the magnetite deposit can be attributed either to replacement or to sedimentation. This shape is common in the case of banded iron formations, yet a favourable horizon of this shape, when replaced, would have the same form. The deposit widens at flexures, which again could be equally well explained by either hypothesis.

The iron content is extremely variable across the entire zone

(Figure 3). The syngenetic theory explains this as sedimentary variation; the epigenetic, as differential replacement.

There is a rough banding throughout the magnetite deposit which is parallel to the margins. If the skarn minerals and the magnetite were formed from pre-existing sedimentary constituents, it would be expected that the original sedimentary foliation would be preserved, as the formation of each mineral band would depend on the composition of the original assemblage. However, the banding in the orebody is too irregular to represent a transformation in situ of a bedded sediment. A more plausible explanation is that material was added and that this process resulted in the partial obscuring of the original foliation.

The zone of magnetite concentration is entirely within the skarn zone, suggesting a close genetic relationship between the magnetite and skarn minerals. It is altogether likely that an iron-rich band, unstable under conditions of metamorphism, might be readily attacked by altering solutions. A more likely possibility is, however, that the metamorphic processes instrumental in the development of the skarn could conceivably have been those involved in the introduction of the iron if, indeed, the iron has been introduced.

Most of the magnetite appears to be contemporaneous with the gangue minerals. The grains are commonly euhedral, and only rarely do the magnetite grains exhibit replacement textures against the gangue. This suggests that the iron was present as a primary mineral or that, if the iron was added, recrystallization of the mineral assemblage took place after addition of the iron. Both explanations seem equally plausible.

The approximate composition across one part of the orebody (diamond drill hole no. 9) was obtained by estimating the percentages of the various minerals observed in the thin-sections and recalculating them to give values for the oxide percentages. The minerals used in the calculations include hedenbergite, epidote, garnet, hornblende, calcite, quartz, tremolite, and magnetite. Other minerals, present in smaller quantities were not used in the calculations because they would not appreciably affect the over-all composition. The margin of error probably does not exceed fifteen percent. In Table 2, column A gives the values for the composition across the orebody, while column B gives the values without considering the magnetite.

	A	B	C	D	E	F
SiO ₂	22	37	23.90	34.80	42.99	42.14
CaO	15	27	22.25	10.02	8.16	17.80
FeO	53	12	10.72	19.70	18.45	13.29
Fe ₂ O ₃		8	0.44	18.60	2.03	1.47
CO ₂	5	8	32.42		19.20	5.08
Al ₂ O ₃	3	4	0.07	3.82	0.12	3.72
MgO	1	2	8.52	4.34	7.07	0.72
H ₂ O	0.4	0.7	0.99	0.40	1.45	2.90
MnO			0.28	7.50	0.25	0.46
SO ₃			0.17			
TiO ₂				0.20		0.54
P ₂ O ₅				0.78	0.009	13.01
FeS					0.13	
Na ₂ O	0.6	1				
	<u>100.0</u>	<u>99.7</u>	<u>99.76</u>	<u>100.16</u>	<u>99.789</u>	<u>101.13</u>

Table 2. Approximate composition of a section across the Bessemer magnetite deposit compared with analyses of other ferruginous bodies. A. Approximate composition of section along D.D.H. no. 9 across Bessemer deposit; B. Composition of same section without considering magnetite; C. Iron carbonate, Gunflint Lake, Canada (Clarke, 1924, p. 583); D. Hedenbergite-garnet-magnetite rock, Druideig Lodge, Lock Duich, Ross-shire (Tilley, 1936, p. 338); E. Average Gunflint ferruginous carbonate (Zapffe, 1912, p. 162); and F. Parting rock in Wabana iron ore formation (Hayes, 1915, p. 52).

When the analyses in columns C and E are recalculated to equalize

the CO₂ content with that of the Bessemer ore, a more justified comparison can be made, as carbon dioxide would be driven off under metamorphism. These values are shown in Table 3.

	A	D	B	C ₁	E ₁	F
SiO ₂	22	34.80	37	53.6	50.5	42.14
CaO	15	10.02	27	31.3	9.58	17.80
FeO	} 53	19.70	12	15.1	21.7	13.29
Fe ₂ O ₃		18.60	8	0.62	2.40	1.47
CO ₂	5		8	5.0	5.00	5.08
Al ₂ O ₃	3	3.82	4	0.11	0.14	3.72
MgO	1	4.34	2	12.0	8.33	0.72
H ₂ O	0.4	0.40	0.7	1.39	1.71	2.90
MnO		7.50		0.40	0.30	0.46
SO ₃				0.24		
TiO ₂		0.20				0.54
P ₂ O ₅		0.78			0.01	13.01
FeS					0.15	
Na ₂ O	0.6					
	<u>100.0</u>	<u>100.16</u>	<u>100.7</u>	<u>99.86</u>	<u>99.82</u>	<u>101.13</u>

Table 3. Comparison of approximate composition across Bessemer magnetite deposit with analyses of other ferruginous bodies. Column headings are the same as used in Table 2, with the exception of C₁ and E₁ which correspond to C and E respectively, recalculated to reduce the CO₂ content to five percent.

The analyses of ferruginous carbonates (columns C₁ and E₁) and the Wabana parting rock (column F) are similar to the Bessemer assemblage when the magnetite is not considered. The Bessemer ore (column A) is similar to the hedenbergite-garnet-magnetite rock analyzed by Tilley (column D). He considers that rock as a metamorphosed iron carbonate sediment, but suggests that some additional iron may have been added from external sources. A consideration of these results suggests that iron in some form has been added to the Bessemer skarn zone to account for the magnetite, although it need not necessarily be the case.

The pyroxene, epidote, and garnet are particularly iron-rich. It is this iron which largely accounts for the twenty percent iron

oxides in addition to the magnetite found in the mineralized zone. The iron of the gangue may have been a primary constituent of the zone, or it may have been added later.

The quantitative distribution of the minerals in the ore zone is of interest. A study of figure 4 shows that hornblende and pyroxene are generally present in inverse proportions, and that magnetite and pyroxene are present in direct proportions. The former relationship suggests that one ferromagnesian mineral has been formed at the expense of the other. The comparatively high degree of metamorphism in the skarn zone favours a high-temperature assemblage; thus pyroxene has formed instead of hornblende. Pyroxene is found replacing hornblende, although later adjustments have probably caused the formation of some hornblende from pyroxene. It seems, then, that the pyroxene constituents were present before metamorphism. The direct quantitative relationship between magnetite and pyroxene may be accidental. However, the introduction of magnetite might be accompanied by increased metamorphism, hence a high-grade assemblage including pyroxene would develop.

There is no doubt that there has been some hydrothermal activity in the area, as the presence of minerals such as chlorite and epidote, and the alteration of biotite suggest hydrothermal action, but whether this process can be ascribed to the origin of the magnetite is questionable. It is probable that the sulphides were introduced hydrothermally. If the sulphides and magnetite were introduced by the same process, it is to be expected that they would have a similar distribution throughout the skarn zone. This is not the case (Figure 5), which suggests that they have different origins. On the other hand, it may be

that the magnetite was introduced before the sulphides, and that recrystallization of the skarn assemblage before introduction of the sulphides resulted in new channels being opened for the sulphides.

The late-stage magnetite and the enrichment of iron in the garnet is evidence for the addition of iron to the zone. This appears to be the final stage of mineralization, and the material may have been introduced hydrothermally. However, it may be argued that late-stage re-solution of existing magnetite may have given rise to the veinlets of magnetite and the enrichment of iron in the garnet.

Summary and Conclusions

From the foregoing discussion it can be seen that the origin of the Bessemer magnetite deposit can be ascribed to one of two causes--metamorphism of a sedimentary ferruginous bed, or metasomatic addition of iron to a calcareous sediment.

Considering the first possibility, the requirements can be briefly summarized: A bed of ferruginous carbonate, with perhaps additional iron oxides was laid down between graywacke and limestone. The area was regionally metamorphosed, causing deformation and upturning of the strata, as well as a change in the mineral assemblages. The graywacke was converted to a hornblende-plagioclase gneiss, and the limestone recrystallized. The reducing conditions produced by the regional metamorphism caused the conversion of ferric oxides and ferrous carbonates to magnetite. The presence of iron made this band particularly susceptible to metamorphism and to the entrance of solutions from a nearby igneous body, resulting in the assemblage of skarn minerals

and the emplacement of the sulphides. Re-solution of a small part of the magnetite produced the late-stage magnetite veinlets. The final phase of igneous activity resulted in the intrusion of diorite dikes.

Considering the second possibility, metasomatic introduction of the iron, the requirements are as follows: A band of rock favourable for replacement was present between graywacke and limestone. This rock was rich in lime and may have had some iron present. During metamorphism of the area, the adjoining rocks were regionally metamorphosed, while the calcareous band provided a suitable channelway for emanations from a nearby igneous body. The present position of this intrusion is in doubt, but it may have been the intermediate acid gneiss to the northwest, or a more basic body at depth. These emanations were rich in iron and probably deposited this metal as replacement of carbonates or other minerals. These conditions resulted in the formation of the skarn minerals and were followed by recrystallization of the magnetite. Sulphides were then introduced in fractures and as replacements. The final solutions carried some magnetite which filled late veinlets. The final stage was the intrusion of diorite dikes considerably later than the mineralization. The diorite dikes are considered as bearing no genetic relationship to the magnetite body. This is considered to be the case because the diorite intrusions are rich in titanium, which is practically absent in the magnetite body, and because the diorite dikes are obviously much younger than the skarn zone.

From the evidence available, both the syngenetic and metasomatic theories have much to recommend them, and it seems impossible definitely to ascribe either origin to the Bessemer magnetite deposit.

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Descriptions of Plates

PLATE I

- Figure 1. Bessemer pit looking northeast.
Figure 2. Limestone outcrop near eastern corner of pit.

PLATE II

- Figure 3. Hornblende-plagioclase gneiss (crossed nicols). x72
Figure 4. Fine-grained gneiss member. x150
Black minerals are ilmenite grains. Note the sub-parallel arrangement of sericite flakes.
Figure 5. Magnetite (black) with calcite and hedenbergite x72
Figure 6. Intergrowth of magnetite and hedenbergite x72

PLATE III

- Figure 7. Hornblende grains in calcite.
This was found at the contact between crystalline limestone and diorite exposed in the trench northwest of the pit.
Figure 8. Specimen of ore showing irregular banding.
Garnet (G), epidote (Ep), magnetite (M).

PLATE IV

- Figure 9. Scattered magnetite grains with hedenbergite and calcite x72
Figure 10. Magnetite veinlets (light gray) cutting gangue (dark gray) and partly surrounding chalcopyrite grains (white). x150
Figure 11. Triangular pyrite grain (light and pitted) partly surrounded by magnetite (light gray). x72
Figure 12. Chalcopyrite grains (white) entirely surrounded by thin margins of magnetite (light gray). x72

PLATE V

- Figure 13. Veinlet with borders of magnetite (light gray) and core of chalcopyrite (white) cutting gangue (dark gray). x72

Figure 14. Pyrite veinlets (white) along margin of gangue mineral grain (gray). x72

Figure 15. Pyrite remnants (Pr) surrounded by pyrrhotite (Ph) which is in turn surrounded and cut by chalcopyrite (Ch). x72

Figure 16. Pyrrhotite (Ph) replaced by chalcopyrite (Ch) x35

PLATE VI

Figure 17. Pyrrhotite (Ph) surrounded and embayed by pentlandite (Pn) x150

Figure 18. Veinlet of chalcopyrite (white) crossing gangue (gray). x72

Figure 19. Remnants of gangue mineral (gray) in chalcopyrite (white). x72

Figure 20. Chalcopyrite veinlets (Ch) crossing magnetite (M) x72

PLATE VII

Figure 21. Garnet (G) veined by calcite (Ct) and hedenbergite (H) x17½

Figure 22. Fine-grained epidote-calcite vein (left to right) crossing an aggregate of magnetite and hedenbergite. x35

Figure 23. Large hedenbergite mass (white) replaced around its margins by hornblende (dark). x35

Figure 24. Hornblende grains (black) rimmed by hedenbergite (gray). White mineral is calcite x49½

PLATE VIII

Figure 25. Hornblende (dark) containing rounded inclusions of calcite (white). x35

Figure 26. Chlorite (Cl) with calcite (Ct) and hornblende (Hb) x72

Figure 27. Biotite (white) altered around margins to a dark green mica. x35

PLATE I



Figure 1



Figure 2

PLATE II



Figure 3

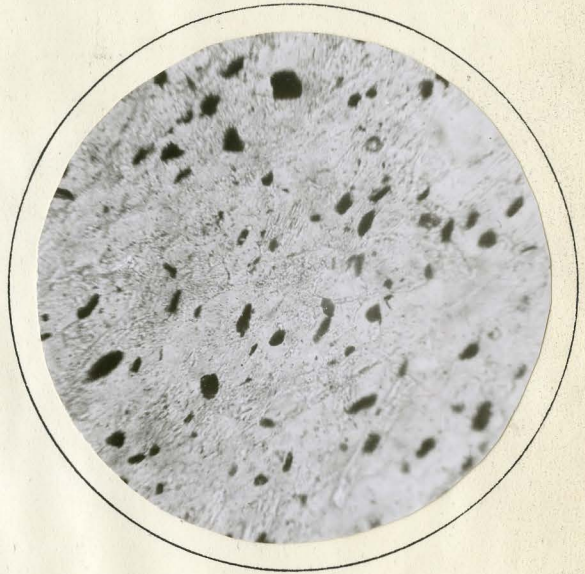


Figure 4

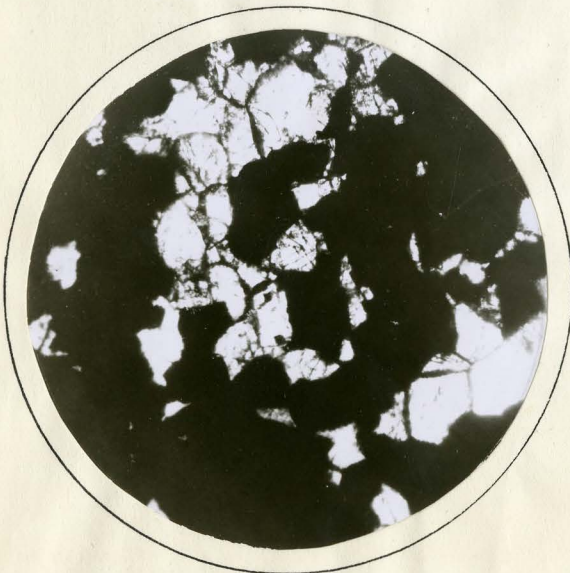


Figure 5

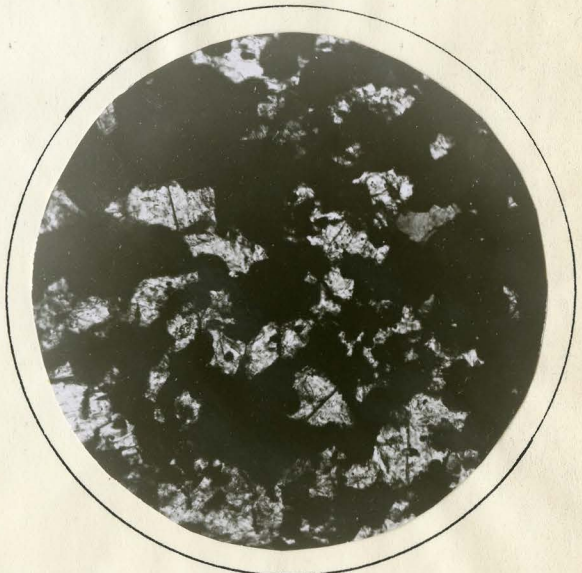


Figure 6

PLATE III



Figure 7

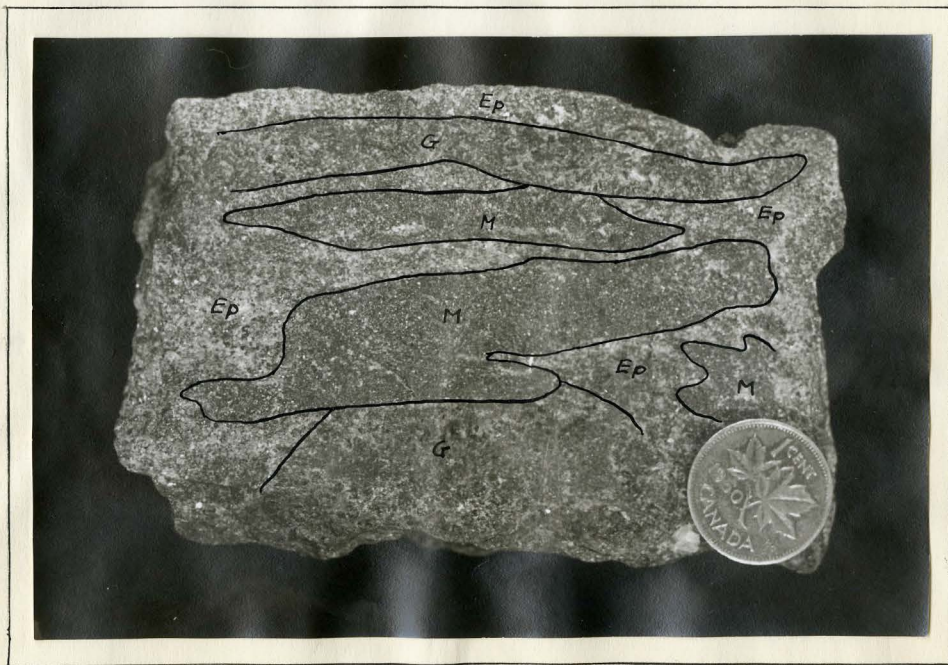


Figure 8

PLATE IV

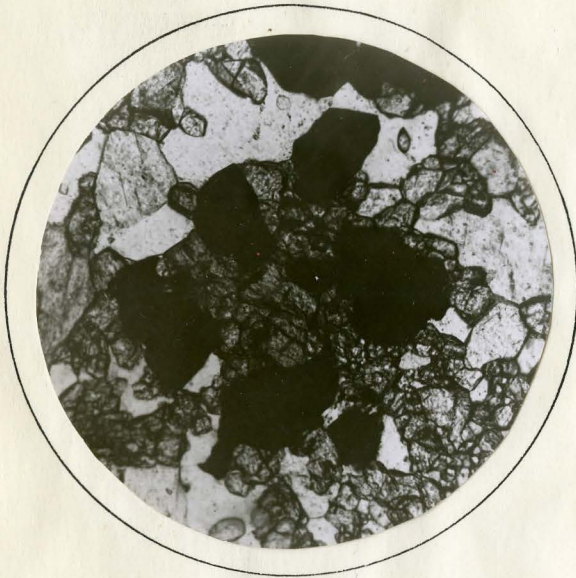


Figure 9

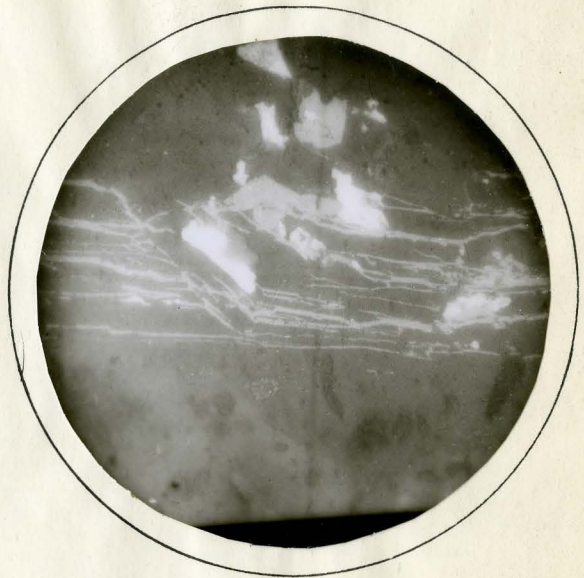


Figure 10

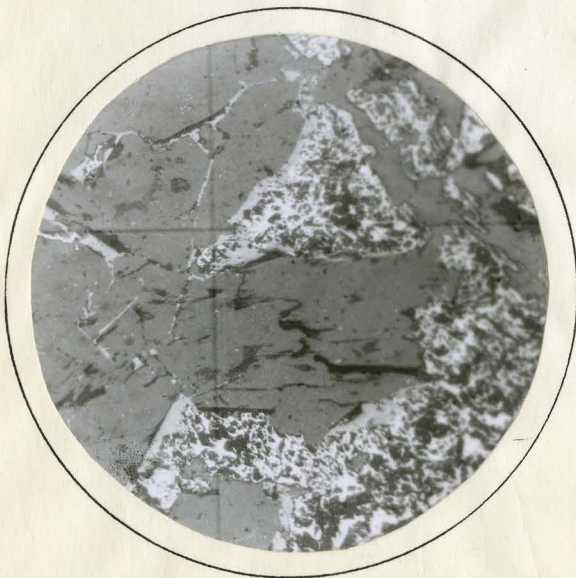


Figure 11

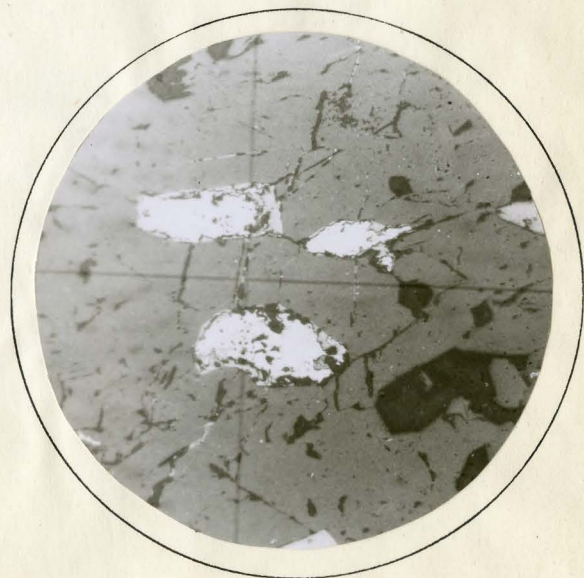


Figure 12

PLATE V



Figure 13



Figure 14

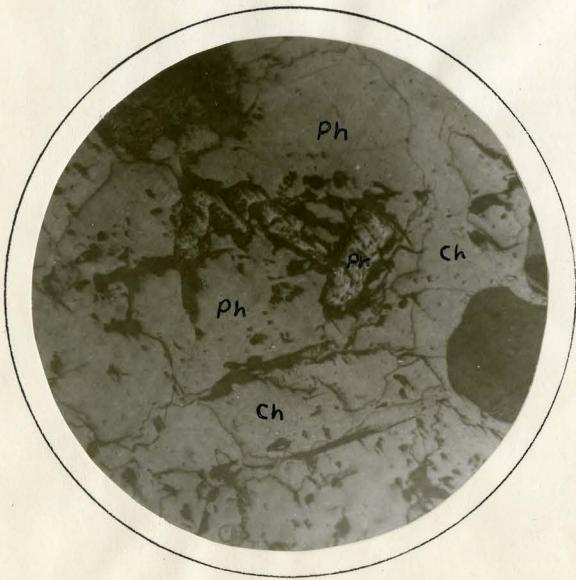


Figure 15



Figure 16

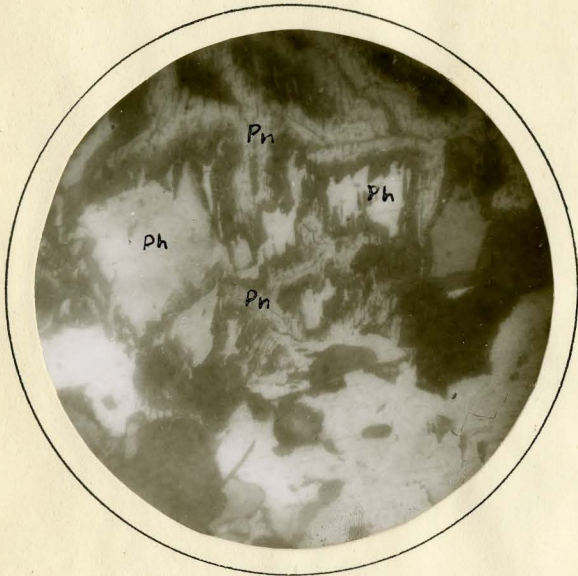


Figure 17



Figure 18

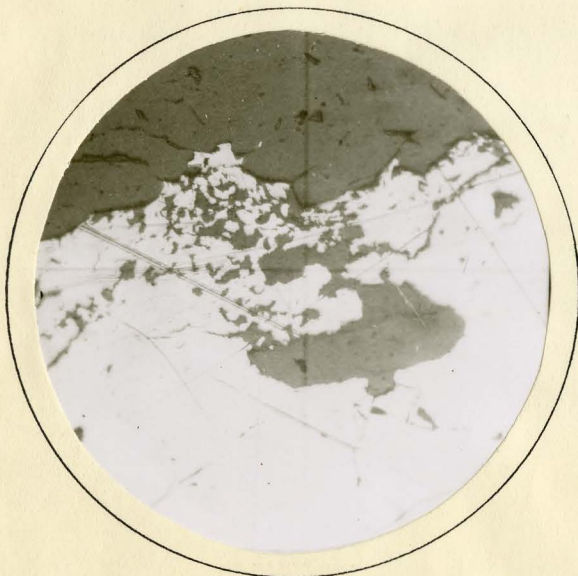


Figure 19

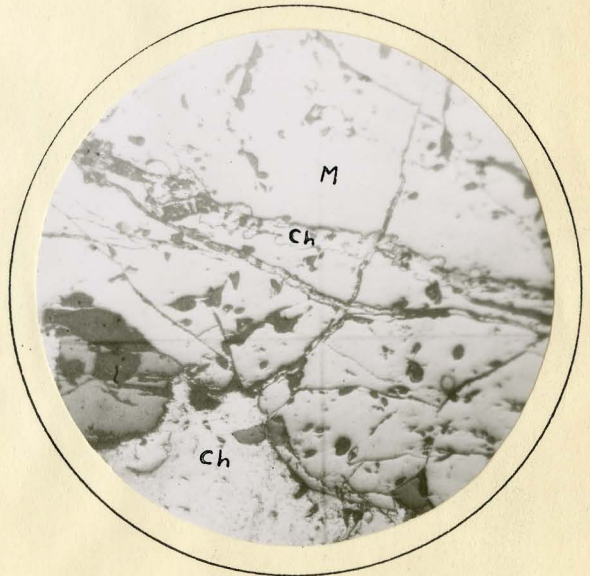


Figure 20

PLATE VII

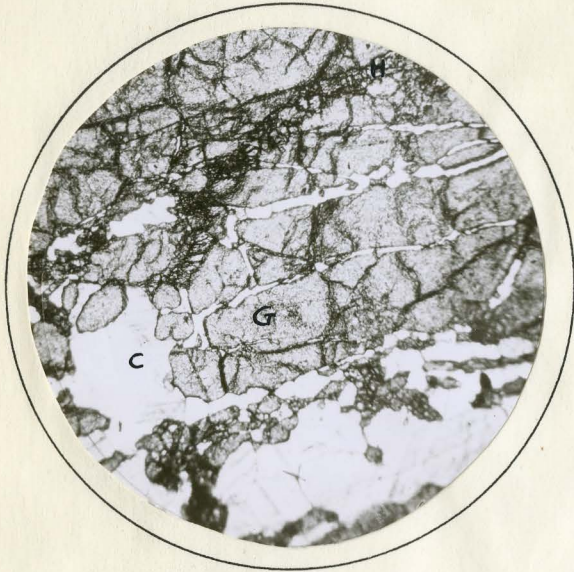


Figure 21

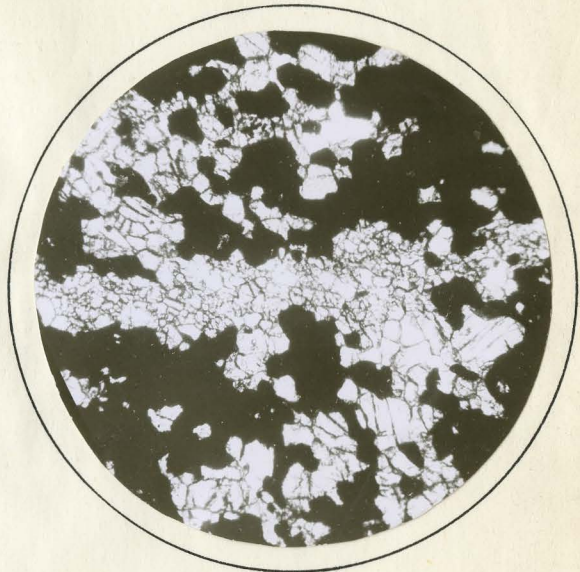


Figure 22



Figure 23

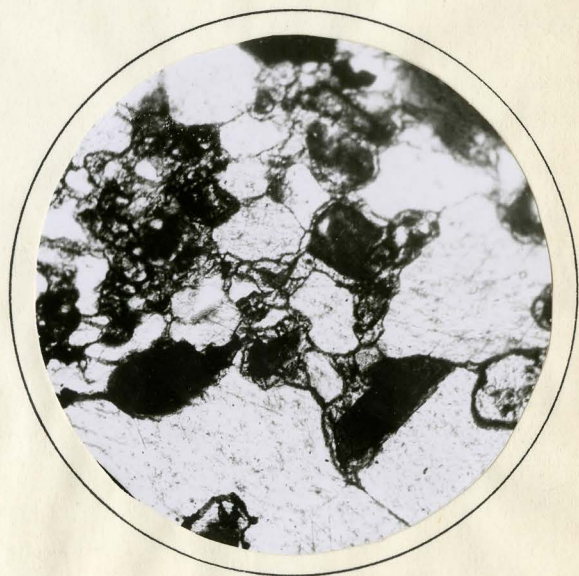


Figure 24

PLATE VIII



Figure 25



Figure 26



Figure 27