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Magnetite deposits
of the
St. Lawrence County district,
New York

by

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1921-

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Abstract

Two nontitaniferous magnetite deposits of St. Lawrence County, northwestern New York, were reopened in 1941. Between 1941 and 1946, 10 promising new deposits and 12 small prospects were found, chiefly by systematic geophysical prospecting based on local and regional geologic studies.

The magnetite deposits occur in old pre-Cambrian rocks. They belong to two geologic types, skarn ore and granite gneiss ore, each of which has been modified locally to form important subtypes. Skarn ores are variable in grade, complex in structure, and small to moderate in size, tending to yield massive ore bodies. They are strikingly similar to those found in Central Sweden. Granite gneiss ores are accompanied by quartz, potash feldspar, biotite, manganiferous garnet, and sillimanite. A modified type has abundant spessartite and fluorite, with barite as a prominent accessory. Granite gneiss ores form disseminated deposits of uniformly low grade, remarkable continuity, and moderate to very large size. Primary crystalline hematite occurs with magnetite in ores of this unusual type.

Ore is restricted to a complex structural knot of metasediments and younger granite sheets pressed into isoclinal folds against buttresses of older granitic rocks. Ore bodies lie on the limbs or noses of folds, often appearing as fishhooks in plan. The ore is younger than the deformation and metamorphism of the host rocks. It represents a high-temperature replacement effected by solutions derived from younger granite magma. Late hydrothermal leaching of the miner-

alized zones is a noteworthy feature of several deposits. Zoning of Adirondack mineral deposits is discussed.

Introduction

Location and importance of the district

The St. Lawrence County magnetite district is in northern New York, in the southeast part of St. Lawrence County. The district, which covers about 1,000 square miles, lies on the northwest flank of the Adirondacks, in the foothills between the high Adirondack massif and the lowlands of the St. Lawrence valley. It is one of the three districts in New York that produce nontitaniferous magnetite, the others being the Clinton County (Chateaugay) district and the Mineville-Port Henry district. (See fig. 1.)

There are at present two operating magnetite mines in the district: Benson Mines of the Jones and Laughlin Ore Company, near Star Lake; and the Clifton mine of the Hanna Coal and Ore Corporation, near Degrasse. These deposits, known for a century or more and operated from time to time in the past, were reopened in 1941. From 1943 to 1949, the two mines produced about 5 million long tons of shipping product, mostly sintered magnetite averaging 62.5 - 63.5 percent Fe. Their combined annual production is roughly half that of New York State, or about 1 percent of the country's total production of iron ore. (For production figures, see Crockett, 1943, 1944; J. R. Linney, 1945, 1946, 1947, 1948; R. J. Linney, 1949; Minerals Yearbook, 1944, 1945, 1946, 1948; Webb and Fleck, 1950.)

Between 1941 and 1946, 10 promising new deposits and 12 small prospects were found, chiefly by systematic geophysical prospecting based on local and regional geologic studies. In terms of "vein length", as much favorable ground was found in those five years as had been known for the preceding century. Unfortunately, this is not equivalent to saying that the ore reserve was doubled by the exploration of those five years, for much of the newly found ore is in smaller, scattered deposits. The value of geophysical prospecting for magnetite deposits has been recognized for many years, but the "pay-off" in the St. Lawrence County district has been startling indeed.

In addition to its magnetite deposits, St. Lawrence County contains the productive Edwards-Balmat zinc district (Brown, 1942) and the Gouverneur talc district (Engel, 1949). Before World War I, the pyrite and pyrrhotite deposits of the county were an importance source of sulfur (Buddington, 1917, 1934). The hematite deposits of the Keene-Antwerp belt were a modest source of iron ore before 1910 (Buddington, 1934). During and after World War II, interest in the hematite deposits was renewed, and for that reason a section of this report treats the subject very briefly. More recently, the old Rossie lead veins have been explored by diamond drilling.

A generation or two ago, another aspect of the mineral wealth of St. Lawrence County was known to mineralogists and collectors the world over. That was the celebrated store of crystals of pyroxene, scapolite, danburite, apatite, phlogopite, tourmaline, and associated contact minerals from Rossie, Russell, Gouverneur, Pierrepont, and other places in the county. Many of the famous mineral localities were described by Agar (1923). Though the county has lost its preeminence as a source of large and showy cabinet specimens, it can still boast of new minerals (Engel, 1947) and new finds of minerals in a new environment.

Purpose and scope of this report

The U. S. Geological Survey undertook a study of the geology and ore deposits of the St. Lawrence County magnetite district in 1943. This work was supervised by A. F. Buddington. Most of the regional geologic mapping and part of the field studies of the ore deposits were carried out jointly by Buddington and the writer. The field work was completed in 1950, and the results of the study will be presented in a forthcoming professional paper of the Geological Survey (Buddington and Leonard, in preparation). Buddington assumed responsibility for the general framework of the final report and for the areal geology of the district. The writer assumes responsibility for the general problem of the magnetite deposits and for the description of individual mines and prospects.

In the report at hand, the writer presents a general description of the magnetite deposits, considers their origin, and attempts to relate the deposits to their regional geologic setting. The report summarizes one part of the Survey's work. It supplements detailed descriptions of individual ore deposits (Leonard, in preparation).

Exploration and development of the district

The existence of magnetite deposits in the district has been known for almost a century and a half. Linney (1943) has given a brief account of Benson Mines, Jayville, and Clifton in his history of iron mining in the Adirondacks. Attempts to mine the deposits were intermittent until 1941, when intensive exploratory work by the Jones and Laughlin and Hanna corporations led to the development and active mining of the Benson Mines and Clifton deposits, respectively. During the long years between the several mining operations, the State of New York expressed its continuing interest in the district in the form of geologic reports by Newland (1908) and Dale (1934, 1935).

The strategic importance of iron ore in the Eastern States led the U. S. Geological Survey to enter the district in 1943. A program of geologic mapping, dip-needle reconnaissance, study of known deposits, and assistance in diamond-drill exploration was then begun. The U. S. Bureau of Mines commenced its exploratory work later in 1943, and cooperative exploration by private interests and the two Federal agencies continued through 1946. When exploration was at its peak, four companies had crews at work in the district: Jones and Laughlin, Hanna, Shenango Furnace Company, and Newton Falls Paper Mill. The last company, owner of several large blocks of timberland, led all other groups, public and private, in the number of discoveries of promising deposits. In 1945, the Survey's application of the airborne magnetometer to the search for ore increased the discovery rate considerably. Most of the district was flown that year, and the rest was flown in 1946. The most important results of exploratory work by the Geological Survey and Bureau of Mines are embodied in the preliminary reports of Buddington and Leonard (1944, 1945a, 1945b); Hawkes and Balsley (1946); Hawkes, Balsley, et al. (1946a, 1946b); Balsley, Buddington, et al. (1950); Millar (1947); and Reed and Cohen (1947). Certain anomalies found in the aeromagnetic survey were checked by dip-needle surveys made by the New York State Science Service under the direction of Shaub (1949).

The results of exploration in the district are given in the following tabulations:

<u>Mines and prospects known in 1941</u> _/

3 major deposits (Benson Mines, Clifton, Parish)
2 minor deposits, faintly promising (Green Farm, Jayville)
6 small and apparently worthless prospects

_/ Classification based on data available in 1950, not in 1941.

<u>Discoveries, 1941-1946</u> _/

10 promising deposits
12 minor but rather promising deposits or anomalies
25 small prospects or anomalies, all of very questionable worth

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_/ Includes several found in delayed ground checks of aeromagnetic anomalies, after 1946.

Table 1 (below) shows by whom and by what means the discoveries were made.

Table 1. Magnetite deposits and magnetic anomalies discovered in the St. Lawrence County district, 1941-1946 _/

A - anomaly detected in aeromagnetic survey; subsequently checked by dip needle

D - anomaly first found by dip needle

O - deposit discovered from outcrop

Class of deposit or anomaly	Companies	USGS	USBM	Joint USGS-USBM
Major deposits (10)	6-D 6	1-O 1-D 1-A 3		1-D 1
Minor but promising prospects or anomalies (12)	1-D 1	9-A 9	1-D 1	1-D 1
Small anomalies or prospects, all of questionable worth (25)	5-D 5	13-A 6-D 19	1-D 1	

_/ Includes several anomalies found in subsequent ground checks of aeromagnetic anomalies. May omit some anomalies found by companies but not known to writer. Omits many unnamed and inconsequential anomalies, as well as aeromagnetic anomalies produced by widespread disseminations of accessory iron oxides in country rock.

Of the new deposits, all but two were completely obscured by overburden whose thickness ranged from a few feet to 150 feet. However, only one of the two was discovered by outcrop; the other deposit was indicated by dip needle before the exposed ore was found.

Thus the importance of geophysical aids in finding new ore in this district was overwhelming.

The relative success of the airborne magnetometer and the dip needle as primary detectors of anomalies cannot be gauged from Table 1. The dip needle was diligently applied in the district for four years before the airborne magnetometer became available. All the known major deposits and dip-needle anomalies were later detected by the airborne magnetometer. The situation reaffirms the old saying that hindsight is better than foresight: if the airborne magnetometer had been available in 1941, it would have done what the dip needle did, but more quickly.

A second point brought out by Table 1 is the high score of the companies, two of which had particularly effective prospecting organizations. So far as the writer knows, five of the six major deposits discovered by the companies were found between 1941 and July of 1943.

The relatively high ratio of new-found promising deposits to minor ones (10/12), and generally promising deposits to unpromising ones (22/25), is encouraging to those searching for new ore bodies in other districts. This record of success is offset somewhat by the reasonable assurance that most of the promising deposits in the St. Lawrence County district have already been found (thanks to the coverage of the aeromagnetic survey), and that there is no second Benson Mines. Major deposits with an inferred reserve of several million to many million tons were found, but none of these appears to approach the huge size of the Benson deposit.

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Geologic setting of the district

General geology of the Adirondacks

The Adirondack region lies in northern New York within a triangle formed by the St. Lawrence, Black, Mohawk, and Hudson-Champlain valleys. If the geologically similar Grenville Lowlands are included, the Adirondacks cover nearly 10,000 square miles. Of this area about 60 percent has geologic maps on a scale of 1/62,500.

Topographically, the region is varied. High mountains (5,300 feet) make up the eastern part. Successively northwest are the swampy Childwold rock terrace, a "fall-zone slope", and the Grenville Lowlands bordering the St. Lawrence. A pre-Cambrian peneplain is well preserved in the lowland area.

Geologically, the Adirondacks are a complex of igneous rocks and metamorphosed igneous and sedimentary rocks of old pre-Cambrian age constituting an outlier of the Canadian Shield. Their gross structure may be divided into internal and external elements. Internally, the Adirondacks have a body of igneous rocks flanked on the northwest and south by tracts in which metasediments predominate. Externally, the Adirondack pre-Cambrian rocks as a whole are an asymmetrical dome overlapped on the north, west, and south by sedimentary rocks of Early Paleozoic age. Fault blocks capped with Lower Paleozoic rocks line the steep eastern face, and scattered outliers of similar rocks are found in the central Adirondacks. Doming is related to the Taconic orogeny, according to Kay (1942). The following resumé deals chiefly with the internal features of the region.

Major subdivisions and their rocks

The core of the Adirondacks is a large body of anorthosite with several subsidiary masses nearby. Closely associated with the anorthosite are small bodies of gabbro, in part genetically related to the anorthosite, in part of younger age. Commercially important bodies of ilmenite-magnetite accompany the anorthositic rocks in several areas and are mined extensively at Tahawus (Stephenson, 1945). Metasedimentary rocks and members of the quartz syenite series occur sparingly in the core area. The related problems of the anorthosite, gabbro, and quartz syenite have been treated extensively by Balk (1931, 1944), Buddington (1939), and others.

Surrounding the core and forming the bulk of the Adirondack massif is a complex of igneous rocks with subordinate metasedimentary rocks. The ratio of the two is about 85:15. The igneous rocks are dominantly granitic. They belong to two great groups: an older quartz syenitic series and a younger granitic series. Though the bulk chemical composition of the quartz syenitic rocks approaches that of a granite, the series shows a remarkably high degree of differentiation ranging from shonkinitic syenite with layers of feldspathic ultramafic rock, through pyroxene syenites and quartz syenites, to ferrohastingsite granite and alaskite. The differentiation has come about through gravity stratification. Clinopyroxene is the characteristic mafic for the quartz syenite series as a whole and several members of the series have a prominent phacoidal structure resulting from the deformation of an originally porphyritic structure. One quartz syenite complex is characterized by ortho-pyroxene as a major mafic mineral and therefore is of charnockitic character.

The younger granite series has hornblende-microperthite granite as its fundamental type, with alaskite, microcline granite, and soda granite (albite-oligoclase granite) as extreme differentiates. Microcline granite magma and its derivatives have, in places, affected certain metasedimentary rocks in a manner ascribed to granitization. Locally, microcline granite, alaskite, and granite or syenite pegmatite vein the country rock, giving rise to migmatites. Similar migmatites are widespread in parts of the dominantly metasedimentary tracts. The origin of all the granitic rocks has recently been discussed at length by

Buddington (1948).

The metasedimentary rocks that flank the dominantly igneous massif have for many years been assigned to the Grenville series. Interleaved with the metasediments are sheets and phacolithic bodies of granite, as well as sheets of gabbroic rocks. The ratio of igneous rocks to metasediments is about 30:70, roughly the inverse of the ratio that obtains within the massif. The metasediments comprise calcitic and dolomitic marbles, biotite-quartz-plagioclase gneiss (locally garnetiferous), quartzitic rocks, quartz-feldspar gneiss (granulites), amphibolites of uncertain origin, and minor but economically important tremolite schist and related rocks. Most of the metasediments have their counterpart in the interior of the Adirondacks. The southern tract of metasediments is rather casually known, but the northwest tract has been the subject of repeated investigation, the most recent and thorough being that of Engel (in preparation). Nowhere in this vast and well exposed tract of dominantly metasedimentary rocks is there evidence of bedded iron formations.

Sequence of intrusion, deformation, and metamorphism

The oldest rocks of the Adirondacks are metasediments of the Grenville series. Into these was intruded a gabbroic anorthosite magma that differentiated to yield a gabbroic anorthosite border facies (partly resulting from assimilation of skarn), an anorthosite core, and local deposits of ilmenite-magnetite. The anorthosite and metasediments were later intruded by dikes, sheets, lenses, and funnel-like bodies of gabbro, some of which have yielded small bodies of ilmenite-magnetite. Locally in the northwest Adirondacks, dioritic rocks were intruded subsequent to the emplacement of the gabbro. Later still, large volumes of quartz syenitic magma were injected into the metasediments as thick sheets and scattered lenses and sills. Some of the magma consolidated as dikes in anorthosite and gabbro. What the condition of the Grenville rocks may have been at this time, no one can say with assurance. Certainly the sedimentary rocks were virtually flat or only gently inclined, for the members of the quartz syenite series show strong evidence of origin by gravity stratification. Yet sheets of quartz syenitic rocks and their host rocks are now as a rule isoclinally folded, overturned, and in places pulled apart. In general they are gneissic, recrystallized, or reconstituted. In places they are cut by later granites which are much less deformed than the quartz syenites themselves. Thus the region was subjected to a profound deformation after solidification of the differentiated quartz syenite magma and intrusion of minor hypersthene metadiabase dikes. Perhaps the so-called *younger* granite made its appearance in the waning stages of this deformation. Granite veins, dikes, and cross-cutting bodies of batholithic dimensions transect the highly deformed quartz syenitic gneisses but are themselves undeformed or only slightly recrystallized. However, in some areas the sheets and phacoliths of "younger" granite are moderately or strongly deformed, indicating a second but less intense period of deformation subsequent to the emplacement of the granite magma.

Diabase was the last igneous rock intruded in the Adirondacks. Its distribution is rather local (Buddington, 1939). In places, diabase dikes have entered fault zones along which post-dike movement has also occurred (Cannon, 1937). Except for faulting, the dikes are undeformed. This and other evidence has led to the belief that the dikes are of late Proterozoic (Keweenawan?) age. Alkalic dikes are found locally in the eastern Adirondacks, but these are of post-Ordovician age (Buddington and Whitcomb, 1941, pp. 85-86).

Of the earlier, pre-kinematic intrusive rocks, only the anorthosite produced extensive pyrometasomatic alteration of its wall rocks, and none of the earlier intrusives developed extensive zones of migmatites. Such features are restricted to contact zones of the younger granites. The development of granite pegmatite, locally syenitic, is also a characteristic accompaniment of the younger granites, but the volume of pegmatite is very small indeed in comparison to the bulk of the granites themselves. Extensive mineralization taking the form of iron oxides, sulfides of the base metals, boron minerals, rare-earth minerals, fluorine- and chlorine-bearing minerals, manganese silicates, barium minerals, and certain hydrous silicates is associated only with the younger granites, and there only in certain definite zones. It is true that minor quantities of some of these minerals are found with one or another of the older intrusives, but their extensive and characteristic development is with the younger granites.

Much of the rock of the Adirondack area, except certain mineral deposits and the diabase dikes,have in some degree been affected by a regional dynamothermal metamorphism. Its effect on the igneous rocks has been investigated by Buddington (1939), and detailed study of its effect on the metasedimentary rocks of the Gouverneur area is being made by A. E. J. and Celeste Engel (1950). Buddington (1939) recognized successive metamorphic zones or facies based on progressive deformation, recrystallization, and reconstitution of the igneous rocks. Deformation took place mainly by plastic flow in the solid state. Directed pressure, heat, and local addition of new material and leaching of old material have all played a part in the regional metamorphism. While the principal variations in facies are of wide areal extent, local variations arise in particular geologic environments, where for example "mobile" marble may have cushioned sheets of quartz syenite against cataclastic deformation, or the sheer size and rigidity of a body of igneous rock may have localized deformation at its borders, leaving the interior relatively undeformed. In a broad way, however, the intensity of metamorphism increases inward toward the core of the Adirondacks.

Zoning of mineral deposits

The distribution of Adirondack deposits of oxides and sulfides is shown in figure 2. The distribution is asymmetric, perhaps in part because of the asymmetric exposure of the pre-Cambrian rocks and in part because of more fundamental geologic factors. Major deposits of ilmenite-magnetite, bearing an epimagmatic relation to the anorthosite, are in or near the core of the Adirondacks. Major deposits of nontitaniferous magnetite are farther outward, closest in the eastern area, farthest in the northwest. A belt of small magnetite prospects succeeds the major magnetite deposits in the northwest Adirondacks. The magnetite deposits are restricted to areas of granitic rocks with subordinate belts of metasediments. Beyond the dominantly igneous massif lie the pyritic sphalerite deposits of the Edwards-Balmat district. These deposits are replacements of partly silicated marble in one of the great tracts of metasediments. Still farther outward from the Adirondack core is a belt of pyrite-pyrrhotite replacement deposits in chloritized migmatitic biotite-quartz-feldspar gneiss of the Grenville series. The pattern of distribution of the various ore deposits is similar to the zonal arrangement characteristic of many other metalliferous districts where the relation is interpreted as the effect of decreasing temperature on outward-spreading hypogene ore carriers. The significance of the pattern was pointed out by Buddington (1939).

In connection with the regional distribution of iron oxides, zinc sulfides, and iron sulfides, the content of minor metals is also significant. The magnetite deposits show a relative concentration of copper in the form of disseminated chalcopyrite (locally, disseminated bornite) and rare veins of chalcopyrite. One of these chalcopyrite veins is known to carry minor values in gold and silver. Sporadic traces of molybdenite and sphalerite are common in the magnetite deposits, but galena has not been detected. On the other hand, chalcopyrite is extremely rare in the pyritic sphalerite deposits, and galena is present in small amounts (Brown, 1936, p. 251). Significant amounts of copper, lead, and zinc have never been found in the pyrite and pyrrhotite deposits (Buddington, personal communication).

Notable concentrations of manganese in the form of spessartite are found in one of the magnetite deposits (Parish), and accessory amounts of spessartitic almandite are characteristic of one entire class of magnetite deposit. Manganiferous silicates have long been known to occur in the Edwards-Balmat and Gouverneur districts, but concentrations of manganese comparable in extent to the Parish deposit are unknown to the writer.

For the purpose of this argument, the epimagmatic ilmenite-magnetite deposits are irrelevant. The nontitaniferous magnetite, sphalerite, and iron sulfide deposits are closely associated with granite. The writer, following Buddington and others, believes that the association is genetic. The deposits are bodies and veins of replacement origin. The magnetite deposits were formed at a high temperature, according to the thesis subsequently developed. The iron-rich primary sphalerite (marmatite) of the Edwards-Balmat zinc deposits, together with other features of the occurrence, suggested to Smyth (1918), Brown (1936a), and Buddington (1939) that these sulfide deposits belonged to the hypothermal or possibly mesothermal range. The abundance of primary (and apperently iron-rich) chlorite with the pyrite-pyrrhotite deposits suggested deposition at somewhat lower temperature than that of the marmatite (Buddington, 1939). The iron sulfides, clearly of replacement origin, have an iron content roughly equivalent to that of many magnetite deposits of the St. Lawrence County district. Some workers

(Smyth, Buddington, 1939) have thought that the talc-tremolite deposits of the Gouverneur district represented mineral deposition at still lower temperatures, but this opinion appears to require some modification after the work of Engel.

Many details of the zonal pattern of the replacement deposits are still to be worked out. Direct comparison of primary rock alteration accompanying the deposits is not possible because the host rocks are different (Buddington, 1939). Moreover, the zonal position of pyrite and pyrrhotite (suggestive of lower temperature with respect to sphalerite) is at odds with the general paragenetic sequence of sulfides found in the magnetite deposits: pyrite ⟶ pyrrhotite ⟶ sphalerite. Nevertheless, the areal distribution of the ore deposits is a fact; the interpretation is debatable.

Geology of the St. Lawrence County magnetite district

The bedrock geology of the St. Lawrence County magnetite district comes close to being a truly representative sample of Adirondack geology. All the major rock types and most of the typical structures are present in the district, though the proportions of the various rock types do not correspond to those of the region as a whole. Most of the district lies in the intermediate zone where granitic rocks predominate and metasediments are subordinate (about 15 percent). However, to understand the environment of the magnetite deposits it has been desirable to extend geologic mapping far enough northwest to take in the bordering tract where Grenville rocks predominate, and far enough southeast to cover part of the geologically hotter interior.

The quality of bedrock exposures ranges from excellent to hopeless, depending on the depth of glacial debris and the density of the forest cover. The whole district has been mapped on a scale of 1/62,500, except for detailed work around certain ore deposits.

Occurrence of the magnetite deposits

The magnetite deposits occur in narrow belts of metasediments enclosed by granite or granite gneiss. Within the metasedimentary belts, magnetite has replaced selected layers of skarn close to alaskite or microcline granite gneiss, or it has replaced sheets of microcline granite gneiss enwrapped by metasediments. One major deposit (Clifton) has replaced skarn that is part of a thin wedge of metasediments enclosed by phacoidal granite gneiss but closely bordered by younger hornblende granite gneiss. A minor deposit (Jayville) has replaced skarn that is enclosed by porphyroclastic hornblende granite gneiss of rather uncertain derivation. The restriction of magnetite deposits to areas in which granitic rocks and metasediments are intimately associated was pointed out many years ago by Newland (1908, pp. 23-24) and repeated by Buddington (1939, p. 178). Recent work has emphasized this association. In refining our knowledge of the deposits, it has been necessary to reject Newland's concept of the relation of magnetite to augite syenite, but his general interpretation of association and origin remains unchanged. It is especially fitting that this should be true for one whose work on geologic problems never received a full measure of deserving praise during his lifetime.

Principal rock types

Eighty to 85 percent of the bedrock is granitic igneous or quasi-igneous rock, about 15 percent is metasedimentary or migmatitic, and the remaining 5 percent is mostly metagabbro and amphibolite. (See fig. 2.) Anorthosite, diorite, metadiabase, and dike rocks of basaltic composition are found locally, but their areal extent is negligible. A few patches of Potsdam sandstone (late Cambrian) are present in the northwestern part of the district. The proportions of the more abundant rock types are given in Table 2.

Table 2. Areal extent of major rock types, St. Lawrence County magnetite district. _/

		Percent	
Granitic series (younger)	Microcline granite (including some associated alaskite) and equivalent gneiss	11	58
	Hornblende granite, biotite alaskite, and equivalent gneiss	47	
Quartz syenitic series (older)	Phacoidal hornblende quartz syenite gneiss and phacoidal granite gneiss	12	26
	Pyroxene syenite gneiss and pyroxene quartz syenite gneiss (including local feldspathic and ultramafic facies)	14	
	Metagabbro and amphibolite	3	
	Metasediments and migmatites (Grenville series)	13	
		100	

_/ Data from Buddington and Leonard (report in preparation)

Descriptions of the major rock types and a few highlights of their petrology are given below in chronological order, starting with the oldest rocks.

Metasediments and migmatites

The metasediments and migmatites of the Grenville series show the greatest diversity of type and variation within type of any major group of rocks. The Grenville rocks are confined principally to several belts (see fig. 2), many miles in length, having an inferred synclinal structure. Sheets of granitic rocks and some metagabbro are usually interleaved with the metasedimentary rocks of these belts. Most or all of the types of metasedimentary rocks commonly occur in a given belt, though certain types are more restricted. In addition, thinner layers and lenses of metasedimentary rocks are included in the masses of granitic rocks, particularly along the contacts of the principal Grenville belts.

Most of the Grenville rocks have a well developed planar structure or foliation. This results from compositional layering (variation in kind or proportion of mineral constituents), from the preferred orientation of platy and elongate minerals, or from a combination of these features. In the migmatites, this planar structure is accentuated by the presence of thin parallel seams of pegmatite, usually granitic but locally syenitic. The foliation in the metasediments and migmatites generally parallels the contacts of the rock units. On a small scale, intersecting foliations of the type described by Engel (1949) have developed locally, but this feature seems generally to be limited to the northwest border of the area.

All the metasediments have been completely recrystallized, most of them have been reconstituted, and some of them--notably the limestones--have been metasomatically replaced. Thus there are no relics of sedimentary textures. Compositional layering is the only primary sedimentary feature now visible, and in a number of places the rocks that show it have been so deformed by plastic flow that there is serious doubt whether the layering any longer reflects bedding. Even here, however, it appears that certain layers have taken up most of the local deformation by folding, leaving the others to shear and recrystallize to a minor or moderate extent. Thus the dispute between relict bedding and major extension of units by shearing and solid flow, or layering by metamorphic differentiation, is not so much one of which took place as how much took place at a particular locality.

Another feature that may be of sedimentary origin is the repeated alternation of layers of biotite quartzite (and biotite gneiss), amphibolite, and marble (or its equivalent, skarn) in rocks from the Silver Pond magnetite deposit, n.e.r., Cranberry Lake quadrangle. The layers are generally 5 to 10 feet thick (locally 50 feet thick), and the sequence is repeated seven times in rocks cut by one 400-foot drill hole. The repetition is suggestive of cyclic sedimentation from sandstone through shale to limestone, but the information available from drill cores is inadequate to test this inference.

Sedimentary facies change from graywacke (sodic shale?) to illitic sandstone is suggested by the apparent transition of biotite-quartz-plagioclase gneiss to biotite-quartz-microcline granulite in one rock unit at the Clifton mine, s.e.r., Russell quadrangle. As the rock unit is known largely from drill cores, rather than from underground or surface exposures, it is entirely possible that the apparent transition is due instead to deformation, either by rupture or by folding of two discrete units.

The lack of relict sedimentary features for distinguishing tops and bottoms of beds means that the metasedimentary rocks themselves cannot be used to determine the attitude of beds in folds, whether upright or overturned. This does not imply that one cannot work out the general character of folds from other data. However, the lack of criteria for recognizing tops and bottoms in metasedimentary layers leaves the general stratigraphic succession uncertain. The present distribution of facies in the gravity-stratified quartz syenitic complexes actually seems to be a more reliable guide to the attitude of the larger structures than is any feature in the metasedimentary rocks of the district.

Pyroxenic metasediments

Metasediments originally of calcareous or dolomitic affinities are especially significant because of their close association with much of the ore. Such rocks include marble and its metasomatized equivalent, skarn; various pyroxene gneisses; some "rusty gneiss"; and perhaps some amphibolite.

The marble is white or gray, locally bluish, and rarely pink. Its texture ranges from very fine to very coarse but is commonly medium to coarse. Its structure, particularly when the exposed area is small, appears massive. Most of the marble shows several percent of disseminated silicates, chiefly phlogopite and diopside. Quartz, potash feldspar, and epidote may also be present. Sulfide specks are common in some areas; graphite is relatively rare. The silicates may be disseminated, clot-like, or layered. In addition, continuous or disrupted layers of skarn, pyroxene gneiss, amphibolite, or quartzite may be present. In the main part of the district (Stark and northern Cranberry Lake quadrangles), outcrops of marble are extremely rare, yet diamond drilling at most magnetite deposits has disclosed layers of marble ranging in thickness from a few inches to tens of feet. All this marble is calcitic; dolomite is restricted to the northwest border of the district.

Skarn is abundant and widely distributed. Because so many of its features are intimately related to the occurrence of magnetite bodies, a description is given in the section on the ore deposits.

Pyroxene gneiss is green, gray-green, or green and pink, fine- to medium-grained, and well to crudely foliated. Locally it is finely laminated. It commonly consists of clinopyroxene, quartz, and feldspar, locally with subordinate mica, amphibole, or garnet. Oligoclase is often the only feldspar, but microcline and untwinned potash feldspar may also be present, dominant, or excluding oligoclase. A little interstitial carbonate is frequently present. The mineralogic composition ranges from pyroxenic quartzite to pyroxenic amphibolite, yet pyroxene-quartz-oligoclase gneiss is the distinct and fundamental type. The color of the mafics ranges from light to very dark, indicating a variable iron content. Migmatitic varieties are common in places. Pyroxene in the granitic veinings of the migmatites is likely to be corroded, and its place may be taken by amphibole or by aggregates of amphibole and brown mica.

"Rusty gneiss" and amphibolite are discussed below in more appropriate sections of this paper.

Biotite gneiss

Biotite gneiss is present as thick units in the northwest part of the district and is a prominent member of all the other belts of Grenville. Typically the rock is fine, gray, and well foliated, consisting of biotite, quartz, and plagioclase (usually oligoclase, occasionally andesine or labradorite, and rarely bytownite). In places the foliation is poorly developed and the rock almost resembles a dark hornfels. A little hornblende or pyroxene accompanies the biotite locally, giving a rock that is transitional to amphibolite. Small knots of schorl are present in some of the gneiss. Sillimanitic biotite gneiss occurs locally in the wall rocks of the magnetite deposits. Very often the biotite gneiss is migmatitic, with white or pinkish pegmatite veins. This migmatitic variety is likely to be garnetiferous. More or less granitized biotite gneiss has microcline or, less frequently, untwinned potash feldspar in the groundmass.

Quartzites and quartz-feldspar gneisses

Quartzites and quartz-feldspar gneisses ("granulites") are ubiquitous but quantitatively subordinate members of the Grenville. The main varieties of quartzite are: white, massive, medium-grained, slightly calcareous quartzite with a little disseminated pyroxene; fine, schistose, thin-layered, contorted quartzite or quartz schist found in certain marble belts; fine, sugary quartzite with scattered purplish-pink garnet metacrysts. Magnetite-bearing "quartzites" that appear to be of metasomatic origin are present at one of the ore deposits (Trembley Mountain). These "quartzites" carry hypersthene, or garnet and biotite, in addition to magnetite.

Quartz-feldspar gneisses or granulites are more frequently encountered than quartzites. The quartz content of these gneisses ranges from 75 percent to about 10 percent, yet the appearance of the rock remains almost unchanged. The gneisses are white, gray, greenish, or pink, generally fine and sugary though locally medium-grained, with a planar structure that is rather obscure in hand specimens. Some polished hand specimens show two or more intersecting foliations, one of which possibly represents the trace of bedding. The feldspar is almost always microcline. The mafic mineral, when present, is commonly biotite, but amphibole or pyroxene occurs locally. The mafic content seldom reaches 1 percent; two to five percent is a representative value. The gneisses may form thick, homogeneous units, or they may be intimately interlayered with thin films, lenticles, and layers of other metasediments. Migmatization of quartz-feldspar gneisses seems, if anything, to be less common than migmatization of other metasediments. Hence an origin by granitization of once-pure quartzite is unlikely. The gneisses may represent reconstituted illitic sandstones.

The question arises whether some of these quartz-feldspar gneisses ought to be called leptites. Leptite is a term used by Fennoscandian geologists for certain widespread, very old, fine-grained, generally felsic rocks which they interpret to be metamorphosed "supracrustal" volcanics. These are the principal members of the leptite formation, which also contains metamorphosed limestone, dolomite, skarn, amphibolite, quartzite, biotite gneiss, and the like. The leptite formation is the host for many deposits of iron ore and base-metal sulfides. The leptites range in composition from sodic to potassic. Though the leptites may be completely deformed and recrystallized, some of them show relict phenocrysts and other textures suggestive of volcanic origin. In some less metamorphosed areas, agglomerates accompany the leptites. The two classes of evidence have led to the belief that the leptites are tuffs and flows interstratified with normal sedimentary rocks, the whole sequence having been metamorphosed and intruded by several types of igneous rocks. Until similar relict textures or volcanic associates can be demonstrated in the St. Lawrence County district, it seems unwise to give the quartz-feldspar gneisses or granulites a name that now carries a definite genetic connotation.

The term "rusty gneiss" has been used by many Adirondack geologists for fine- to medium-grained felsic gneisses that weather rusty-brown upon oxidation of the disseminated pyrite or pyrrhotite that they contain. Much of the "rusty gneiss" is therefore a variety of the quartz-feldspar gneiss described above, but some of it is alaskite gneiss, biotite-quartz-plagioclase gneiss, and amphibolite.

Metagabbro and amphibolite

Dikes of metagabbro cut the metasediments at several places in the northern part of the district, and lenses and sheets of metagabbro are present within the metasediments and granitic rocks of a large part of the district. The granitic rocks intrude and inject the metagabbro, much of which is now amphibolite. The least deformed metagabbro is a dark, medium- to coarse-grained rock with an ophitic structure. It consists principally of andesine (locally, sodic labradorite) and clinopyroxene. Reconstituted varieties usually have hornblende as well. Ilmenite-magnetite is invariably present in quantities ranging from a few percent of disseminated metallics to local concentrations of rather pure metallics, including some iron sulfide. A number of these concentrations of ilmenite-magnetite have been prospected for iron ore, with disappointing results. The known bodies are small and of no commercial value as sources of magnetite or ilmenite concentrates.

Much more abundant than recognizable metagabbro is the rock termed amphibolite. This is essentially a hornblende-plagioclase gneiss, often with some brown mica or clinopyroxene. The rock is dark, speckled, medium-grained, and generally well foliated. A prominent lineation given by parallel hornblende crystals is present locally. Varieties having megascopically green hornblende, instead of the usual black, are easily confused with feldspathic or scapolitic pyroxene skarn. There is an apparent mineralogic transition from amphibolite to pyroxene-plagioclase gneiss on the one hand and biotite-plagioclase gneiss on the other. The close association of amphibolite with metasediments also leads one to suspect that some of the amphibolite is metasedimentary, yet the evidence in this district is never compelling. Some amphibolite is traceable into metagabbro, and the suspicion grows that much of the amphibolite of the district is a meta-igneous rock. Whether or not some of it was originally tuffaceous, one cannot say.

Quartz syenite series

The quartz syenite series is a group of older granitic rocks whose relation to the magnetite deposits is important but indirect, and structural rather than genetic. The origin, petrology, and structure of the series have been treated at length by Buddington (1939, 1948).

The average composition of the series is that of a pyroxene quartz syenite, a rock approaching granite in composition but possessing so many distinctive features that it deserves a name of its own. For convenience in distinguishing rocks of this series from younger and quite different granitic rocks, the name "quartz syenite", given by earlier Adirondack geologists, has been retained. Members of the series show a wide range in composition resulting from differentiation of a pyroxene quartz syenitic magma through gravity stratification. Rocks of the quartz syenite series now exist as several large complexes, of which the Stark complex and parts of the Diana and Tupper complexes are found in the district. These complexes of older granitic rocks appear to have acted as rigid blocks or buttresses against which parts of the younger granitic rocks and belts of metasediments were pressed in the last stages of major deformation in the area. The difference in degree of deformation between the older quartz syenites and the younger granites is as outstanding as their differences in composition.

Other less striking features of the quartz syenites also bear on the problem of ore deposits in the district. One of these is the very slight development of pegmatites from the quartz syenitic magma. A second is the limited extent of mineralization and metasomatism adjacent to the felsic members of the series. A third is the unusual manner in which iron and titanium have been concentrated in one member of the series.

The existence of contact metamorphic products between marble and members of the quartz syenite series has long been known. Agar (1923) studied several of the more important localities in the Northwest Adirondacks and noted that the metasomatic effects related to the quartz syenites resulted in a simpler suite of minerals, and one suggestive of higher temperature and drier conditions, than that produced by similar action of younger granites on marble. It is true, in addition, that the extent of the metasomatism is generally less. Moreover, no sizeable replacement deposits of sulfides or magnetite are referable to the quartz syenite series.

Certainly the quartz syenite magma was capable of concentrating small amounts of iron, fluorine, and water in its most felsic differentiates. This is evidenced by the ferrohastingsite (unpublished analysis) that constitutes the mafic in the phacoidal granite gneiss of the Stark complex (originally a porphyritic granite, in part with rapakivi structure). Locally, small bodies of dark, iron-rich skarn have been formed by the action of quartz syenitic magma on marble. However, the magma seems never to have been rich enough in iron and volatiles at the appropriate stage in its differentiation to yield extensive bodies of skarn, magnetite, or sulfides.

Perhaps a partial explanation of this circumstance is given by the manner and position in which iron and titanium were concentrated in the quartz syenite magma. Stratigraphically about a third of the way upward from the base of the Diana complex (now a deformed, overturned sheet) is a zone of pyroxene syenite gneiss with lenses of shonkinite gneiss and feldspathic ultramafic gneiss (see Buddington, 1948, pl. 3). The feldspathic ultramafic rock, consisting of small granulated feldspar augen in a matrix mainly of hypersthene and ferroaugite (locally with abundant zircon), contains 20-40 percent metallics in the form of ilmenite and magnetite. The feldspathic ultramafic lenses range in thickness from mere films to several feet, and their length is commonly a few feet to as much as several hundred feet. The dark silicates and oxides concentrated in the shonkinite and feldspathic ultramafic lenses represent a many-fold increase of iron and titanium relative to that found in other members of the quartz syenite series. The pyroxene syenite zone carrying the mafic concentrations is actually a large lens whose thickness varies directly with the total thickness of the Diana complex. Buddington (1939, 1948) has interpreted the evidence from composition and distribution of the zone to indicate that the more mafic portions are rhythmically interlayered crystal accumulates. It thus appears that much of the iron, together with titanium, was precipitated from the differentiating quartz syenite magma at an intermediate stage and hence was never available in the late stages when any volatiles that might have accumulated would have been available for its mobilization.

The potential economic value of ilmenite, magnetite, and zircon from the feldspathic ultramafic lenses will be discussed briefly at the end of this paper.

Hypersthene metadiabase

Small, scattered dikes of fine-grained hypersthene metadiabase are present in or close to members of the quartz syenite series. The dikes range in composition from hypersthene-augite-andesine diabase to garnetiferous hypersthene "granulite" and micaceous amphibolite (Buddington, 1939, pp. 133-134). Though they cut across the structure of their host, they usually have an internal secondary foliation parallel to the foliation of the host rock. At the Clifton mine, the appearance of metadiabase dikes signals additional complications in the structure of the ore body. The metadiabase transects wall rocks and the skarn that carries ore. In the latter case, it appears as if metadiabase cut ore; actually, however, ore has replaced skarn and avoided metadiabase, leaving "inherited dikes" in some parts of the ore body.

Granite and granite gneiss series

A series of granitic rocks younger than the quartz syenite series is thought to be genetically related to the magnetite deposits of the district. These rocks have been collectively termed "younger granite" by Buddington (1939). The term remains useful but has certain limitations; it is not to be considered a formation name. The series included hornblende-microperthite granite, biotite alaskite, microcline granite gneiss, and very minor amounts of soda granite (albite-oligoclase granite) and porphyritic granite. Metamorphism has affected these rocks in varying degrees so that members which are massive or gneissoid in some areas are gneissic in others. In addition, contamination with metasediments and amphibolite has modified the fundamental types locally.

Main types and their occurrence

Hornblende-microperthite granite forms large masses in the central part of the district. In places it transects the contacts of the quartz syenite series and forms dikes within certain members of the series. It never forms thin sheets or small, arcuate bodies in the metasediments. The rock is generally pink, medium-grained, and slightly to moderately foliated. It consists principally of hornblende (femaghastingsite), quartz, microperthite, and plagioclase, with accessory magnetite, apatite, and zircon. Locally, a facies carrying ferrohastingsite has been developed; in part, this facies is more quartzose than the normal hornblende granite.

Alaskite occurs as thin sheets and phacolithic masses in the metasediments, and as a local border facies of the hornblende-microperthite "roof rock" on the upper side of the associated mass of hornblende-microperthite granite. The alaskite is generally pink, medium- to fine-grained, and virtually lacking in megascopic foliation. It consists principally of quartz, microperthite, plagioclase, and a percent or two of biotite. Magnetite, fluorite, and zircon are the usual accessory minerals. Fluorite is especially characteristic of much of the alaskite. Generally it is visible only in thin section; rarely it is concentrated in vugs (miarolites?) in the alaskite, or forms thin coatings with quartz and calcite along joint planes in the alaskite.

Microcline granite gneiss is a heterogeneous rock forming broad belts and thin sheets in and along the belts of metasediments and amphibolite. The microcline granite gneiss is host to one of the two major types of magnetite deposit in the district. The unmineralized rock is generally pink, fine-grained, and moderately or prominently foliated. Locally it is sugary in texture and rather weakly foliated. Rarely the rock is pink, uniform, medium-grained, and apparently massive--a granite without gneissoid or gneissic structure.

One facies of the microcline granite gneiss consists of quartz, microcline, a little plagioclase, and a mafic mineral that may be pyroxene, hornblende, or biotite. Some of the biotitic variety is slightly garnetiferous. Common accessories are iron oxides, apatite, sphene, and zircon. Thin seams of pegmatite parallel to the foliation are a usual feature of the rock. Locally in the biotitic variety these pegmatite seams, or their quartz-rich counterpart, are schorl-bearing. The amounts of quartz and mafic minerals in the rock vary considerably. The varietal mafic mineral is commonly the same as that in the country rock in which the microcline granite gneiss is found. Moreover, layers, lenses, schlieren, and ghost-like remnants of country rock--partly or considerably modified--are almost ubiquitous in the microcline granite gneiss. Thus the rock appears in large part to be contaminated or migmatitic, and its varietal mafics are in a sense xenocrysts or modifications thereof.

Another facies of the gneiss has sillimanite as its characteristic mineral. A little biotite, with or without garnet, may also be present. Locally, muscovite takes the place of sillimanite. The sillimanite is concentrated in thin, discontinuous films or in lenticular aggregates consisting of quartz, sillimanite, and iron oxides. Locally these aggregates are twisted and contorted. Irregular masses of grayish or yellowish andesine pegmatite appear sporadically in the rock. The sillimanitic facies of the microcline granite gneiss is found in conjunction with belts of biotite-quartz-plagioclase gneiss, ghost-like relics of which are visible locally in the granite gneiss.

The relative age of the microcline granite gneiss is uncertain. Screens of metasediments always intervene between it and the hornblende granite, and the rock is not found adjacent to members of the quartz syenite series. If some or all of the amphibolite occurring within microcline granite gneiss represents metagabbro, the granite gneiss is younger than metagabbro, for the amphibolite is locally incorporated by microcline granite gneiss. A fine, granoblastic fabric is so typical of most of the gneiss that one might, on that basis, assume the rock to be older than the hornblende granite and alaskite, parts of which are relatively undeformed. However, the granulose fabric might be interpreted as aplitic and due to rapid loss of volatiles, rather than to deformation. The close association of microcline granite gneiss and one type of ore bespeaks a genetic relation and makes it difficult for the writer to believe that the granite gneiss is vastly older than the apparently undeformed ore.

Soda granite (albite-oligoclase granite) is found in subordinate amounts close to several magnetite deposits in the district. The rock is pink or gray, medium- to coarse-grained, and weakly foliated or massive. It consists essentially of biotite, considerable quartz, albite-oligoclase, and very little potash feldspar. Except for the presence of striated plagioclase, hand specimens of the pink soda granite look exactly like alaskite. It is remarkable that soda granite, so common in the Clinton County and Mineville-Port Henry magnetite districts of New York, the Dover district of New Jersey, and parts of the Central Swedish district, should be so rare in the St. Lawrence County district.

A few small bodies of porphyritic biotite granite are found in the foothills along the western border of the district. This granite may correspond to the Hermon granite of the lowlands area (Buddington, 1939, 1948; Prucha, 1949).

Origin of the granite and granite gneiss series

Buddington (1948) has considered in detail the problem of origin of the granite and granite gneiss series. He concludes that the hornblende-microperthite granite is magmatic and that the alaskite represents a volatile-enriched portion of the same magma concentrated in the upper part of the hornblende-microperthite granite masses. Locally this mobile magmatic material escaped into the metasediments, giving rise to thin sheets and phacolithic bodies of alaskite. Many of these are accompanied by aureoles of contact metamorphism. The origin of the microcline granite gneiss is much more complex. Part of the rock represents a potash-rich differentiate, pegmatitic in composition, charged with volatiles and injected as sheets into the metasediments. Much of it is a mixed rock produced by intimate penetration of the metasediments by this potash-rich material, reaction with metasedimentary material, and metasomatic replacement of the host rocks. The porphyritic biotite granite may be a manifestation of similar processes at work outside the main igneous complex of the Adirondacks. The rather rare soda granite has resulted from the action of granitic magma or its emanations on biotite-quartz-plagioclase gneiss.

Certain features of the types of mineralization associated with the granitic rocks have already been mentioned and will be discussed further in the section on ore deposits.

Position of magnetite deposits with respect to younger granites

All the magnetite deposits are close to at least one, and generally more than one, facies of the younger granite. Just how close the relation is, is brought out by the figures that follow. Several deposits far removed from any considerable area of outcrops have been omitted from the summary because of inadequate data. Figures for the rest of the deposits represent "maximum" distances, as no one deposit is completely known or its environs completely free of overburden.

Let us assume for the moment that the biotite- and sillimanite-microcline granite gneisses are facies of the younger granite series. Then all the magnetite deposits are within 500 feet of at least one facies of the younger granite. The range is 0 to 500 feet. Granite gneiss ores are, of course, in the microcline granite gneiss itself, and most of the skarn ores are within 200 or 300 feet of younger granite.

All magnetite deposits are within 0.5 mile of alaskite or its equivalent gneiss and generally within 0.2 mile. The range is 5 feet to 0.5 mile.

With two exceptions, all the deposits are within 0.5 mile of hornblende-microperthite granite or its equivalent gneiss, and most of them are within 0.2-0.3 mile. The exceptions are the Jayville deposit (0.7 mile) and the Hughesville School, or Walker Farm, prospect (0.9 mile). The range is 200 feet or less to 0.9 mile.

The distance from magnetite deposits to microcline granite gneiss is highly variable. Though 9 deposits are in it and 2 deposits are within 100 feet of it, 6 are within 1 mile, 5 are within 1 to 4 miles, and 4 are more than 4 miles from it.

Let us now assume that the microcline granite gneiss is not a facies of the younger granite. All deposits in, or possibly related to, microcline granite gneiss are also within 0.4 mile of alaskite or its equivalent gneiss, and most deposits are within 0.2 mile. The same deposits are within 0.3 mile (generally within 0.2 mile) of hornblende-microperthite granite or its equivalent gneiss.

Thus we find that most of the magnetite deposits, regardless of type, are within 0.2-0.3 mile of alaskite or hornblende-microperthite granite or their gneissic equivalents. All the deposits, including those isolated by large areas of overburden, are still within a mile of one of those rock types. In contrast, the distance to microcline granite gneiss is highly variable, ranging from 0 to 7.5 miles.

Regional metamorphism

In some degree, metamorphism of regional extent has affected all the pre-Cambrian rocks of the district except the ores and the late pre-Cambrian diabase dikes. However, this metamorphism has not affected the whole area equally. Rather, it has been progressive, leaving some of the rocks in certain areas entirely free of metamorphism, whereas rocks of similar type elsewhere show intense metamorphism. The general pattern for the district is the same as that displayed by the Adirondacks as a whole, where Buddington (1939, pp. 251-333) has recognized four principal metamorphic zones of regional extent. Facts having special significance in relation to the problem of mineralization are:

1. The rocks of the quartz syenite series are more highly metamorphosed than rocks of the granite series, even where members of the two series are found side by side.

2. The intensity of metamorphism in the two great series of granitic rocks does not always increase in the same geographic direction, supporting independent structural evidence of at least two distinct periods of deformation.

3. The magnetite deposits show no sign of deformation or regional metamorphism, even where some of the enclosing granitic rocks are appreciably metamorphosed. This absence of deformation and regional metamorphism in the ore, coupled with the intense metamorphism of quartz syenite rocks, effectively rules out the quartz syenitic magma as a potential source of the iron ores. (See also discussion of the quartz

syenitic rocks, above).

Regional metamorphism in the district has been essentially isochemical. It is manifest by changes in fabric (e.g., granulation, followed by progressive recrystallization), change in chemical composition of varietal minerals (e.g., hornblendes in younger granites), the appearance of new minerals (e.g., garnet in quartz syenitic rocks and metagabbros), and an increase in the oxidation state of iron present in the rocks. The above changes are exclusively phenomena of recrystallization and reconstitution. Locally, the effect of replacement is indicated by a very slight increase in the total iron content of metamorphosed members of the younger granite series. For example, total iron oxides in analyzed pairs of hornblende granite and equivalent gneiss are 3.38 percent and 4.60 percent,respectively, indicating an absolute increase of 1.22 percent of iron oxides (equivalent to 36.1 percent, relatively). Unmetamorphosed alaskite contains 1.49 percent iron oxides, whereas its equivalent gneiss contains 2.04 percent--an absolute increase of 0.55 percent of iron oxides or a relative increase of 36.9 percent. Thus it appears impossible to concentrate iron for the magnetite deposits by metamorphism of this character, though it might be argued that a regional source of iron contributed that element to the country rocks during metamorphism and by some process concentrated iron locally to form magnetite deposits. Inasmuch as the ores are distinctly of post-metamorphic origin, a regional introduction of iron seems unlikely to the writer.

Metamorphism will embrace, in the minds of some readers, the phenomena of granitization, development of skarns and ores, and zonal distribution of the mineral deposits of the region. Certain aspects of these problems are treated in a subsequent discussion of the origin of the ores.

Structure

Structural framework

Two great structural trends dominate the western half of the Adirondack region. (See Buddington, 1939, fig. 22, p. 238.) These trends are east-northeast in the southern segment and northeast in the northwest segment. Folds are overturned northward in the southern segment and southeastward in the northwest segment adjacent to the massif. The St. Lawrence County magnetite district lies approximately at the intersection of the two trends, in a "node" of extremely complex structure. Three rigid anticlinal masses of older quartz syenitic rocks dominate the structure of the district. (See fig. 1.) Within the framework of these rigid masses, the metasediments and younger granitic rocks have been squeezed into folds of diverse orientation, and overturning in several directions has taken place. The resulting knotted structure has afforded a locus of deposition for the more important magnetite ore bodies.

Structural elements

Folds

Folds of several orders are outlined by the internal planar structure and contacts of the rock units. Parallelism of foliation, preexisting planar structures, and rock contacts is the rule. The folding has affected both metasedimentary and igneous rocks, except for the late pre-Cambrian diabases. In size, the folds range from many miles in length and two or three miles in width (Stark anticline of quartz syenitic rocks) to crenulations a few centimeters long and a few millimeters wide (twisted lenses in sillimanite-microcline granite gneiss). Only the most rigid units preserve open folds; asymmetry, overturning, and isoclinal folding are the rule.

The mining geologist is directly concerned with ore bodies related to folds of intermediate size. These are discussed separately under "structure of the magnetite deposits". In order to map and interpret these intermediate structures, a knowledge of the minor folds, or "drag folds", is essential. The minor folds, however, are of many types. Often consistent among themselves, they do not always mimic the major structure with which they are associated. In type they may be open warps, rolls, asymmetric "drags", chevrons, symmetric isoclines, shear folds, or essentially ptygmatic folds. One of the few systematic features of the minor folds is their increase in number and complexity in the axial zones and noses of folds. Their axial trend seems generally to agree with that of the attendant larger fold, but the plunges of major and minor folds need not be the same. Moreover, the axial planes of minor folds may not parallel the axial plane of the larger fold, or give some clue to the fold's symmetry. The pattern is further complicated by the inferred existence of at least two periods of deformation with forces acting from slightly different directions.

Lineations

Linear structures of many types are found in rocks of the district. These lineations include mineral elongation, mineral streaks and ribbons, pencils, rods, fluting and grooving, pods and clots, pegmatitic braids, intersecting foliations, slickensides, axes of minor folds, and other features. Lineations commonly occur on foliation surfaces, but in many places a lineation becomes so strongly developed that no trace of planar structure can be detected. The lineations may lie in _a_, _b_, or some intermediate direction with respect to the associated fold axes, of which there may be two or more sets. (This statement is based on detailed surface and underground mapping at the Clifton mine. In areal geologic mapping of the district, one commonly finds that lineations are in _b_ if referred to two separate fold systems, the axes for one of which may be very roughly in _a_ with respect to the other.) For this reason, it is utterly fallacious to assume that a given lineation is in _b_, parallel to some fold axis; it is absolutely essential to work out the pattern of individual lineations before attempting to infer even the local pattern of fold systems. The relation of individual ore bodies to lineations is discussed separately.

While several sets of lineations are present in restricted areas near the ore bodies, Buddington has found that there is also a broader, regional distribution of the attitudes of lineations. Thus some large areas characteristically have lineations parallel to the axes of major folds, whereas other areas have lineations at a large angle to the trend of major structural elements. Rigidity of the older rock units--such as members of the quartz syenitic series and, possibly, buried anorthosite masses--seems to have been the controlling factor in the development of areas of "discordant" _/ lineations.

_/ The writer does not know a handy term for lineations at large angles to the trend of major structural elements. Provisionally--and with little liking--he has adopted the term "discordant", using "concordant" to denote lineations essentially parallel to major fold axes.

Faults

Faulting is more in evidence in the eastern part of the district than in the highly mineralized central part. The eastern area has been subjected to normal faulting along north and northeast lines. The known faults range in length from a few miles possibly to 15 or 20 miles. The inferred throw on some of the faults is several thousand feet. At least one prominent graben has been produced. These normal faults cannot be dated within the mapped area, but they may be partly of late pre-Cambrian age and largely of post-Taconian age. (See summary of Adirondack geology.) In several places, structural discordance along east or east-northeast lines suggests the presence of thrust faults, possibly of great age. A few minor faults are present in or near the magnetite deposits. These are discussed separately. There is no genetic relation between known faults and magnetite mineralization, but intense plastic deformation in rocks that now carry ore may have been a prerequisite for ore deposition.

Relation of ore deposits to regional structure

The location of the district within a node formed roughly at the intersection of two regional structural trends has already been pointed out. Inside this node, the principal magnetite deposits are found in an embayment between the three major complexes of rigid syenitic and quartz syenitic rocks. Another group of magnetite deposits flanks the major anticlines of quartz syenitic rocks constituting parts of the Stark and Tupper Lake complexes. (See fig. 1.) In general, magnetite deposits in granite gneiss are concentrated in the structurally most complex, central part of the district. Magnetite deposits in skarn are also found here, but many other representatives of this ore type have a peripheral relation to the anticlinal masses of quartz syenitic rocks. Additional structural features to which the magnetite deposits are spatially related are: areas showing "discordant" lineation--especially the borders of such areas; and the axes of major synclinal folds.

Areas of discordant lineation

All the major deposits of magnetite are (1) on or within a mile of the borders of areas of discordant lineations, or (2) well within the central zone of concordant lineations at places where lineations culminate, diverge, or change markedly in trend. With two exceptions, _/

_/ These may not be exceptions; the structural data for the surrounding area are inadequate.

the minor deposits are (1) on or within two miles of the borders of areas of discordant lineations, or (2) within one of several major lineation zones--concordant or discordant--some distance from their borders but at places where lineations culminate, converge, or change markedly in trend.

It seems highly significant that two other groups of mineral deposits bear a similar relation to the boundaries between belts of concordant and discordant lineations. Half of the known major deposits of iron sulfides are on or within two miles of the boundary between a belt of concordant lineations in Grenville rocks and a belt of discordant lineations in mixed rocks, predominantly igneous. Among the pyritic sphalerite deposits, the Balmat group and Hyatt mine are within a mile of the same structural boundary noted above; Edwards is about three miles from the boundary. All the sphalerite deposits are within a broad belt of discordant lineations.

Thus it appears that borders between areas of contrasting lineation, and sharp local variations in attitude of lineation, are significant structural controls for localizing ore deposition. The nature of the lineation characteristic of the area--whether concordant or discordant--does not seem to matter.

Major fold axes

All the magnetite deposits are within a mile, and generally within some hundreds of feet, of the axes of major synclinal folds; that is, synclines recognizable and mappable for distances of 2 to 15 miles. Some of the synclines are upright, but most of them are sharply overturned.

The direction of overturning varies with the position of the fold relative to the framework of rigid anticlinal elements of syenitic and quartz syenitic rocks. Usually the axial planes of the synclines dip away from the nearest mass of quartz syenitic rocks, though a few exceptions have been noted.

A number of magnetite deposits are found near the intersections of synclines with anticlines of markedly different axial trend (Dead Creek, Jarvis Bridge Extension, Benson Mines, Griffin Park, Skate Creek, Clifton?, Brunner Hill?). Not only do the axial trends differ; the attitudes of the axial planes may differ also: where the syncline is overturned, the transverse anticline is upright or nearly so. The converse does not seem to hold. The transverse anticlines are, in a sense, "barrier anticlines", as it may be impossible to trace the projected synclinal axis beyond the transverse anticline. Whether this is due to some inherent structural peculiarity, or merely to problems of mapping the structures and interpreting them, we do not know.

Other magnetite deposits occur at or near axial culminations of essentially upright synclines (Spruce Mountain Northwest). Still others lie close to the point where synclines change markedly in axial trend and in attitude of axial plane (Brandy Brook Northwest, Silver Pond, Trembley Mountain, Hardwood Mill?). One deposit (Jarvis Bridge) that is close to a synclinal axis is also close to an inferred anticlinal axial depression produced by the intersection of an upright anticline and an overturned anticline. A few minor deposits lie near major synclinal axes that are apparently uninterrupted by culminations or by anticlines (Sweet Pond, Walker Farm, White Place). Several promising deposits are in areas so obscured by glacial debris that their local structural setting cannot be inferred with assurance (Twin Lakes, Outafit, Bumtbridge Pond).

The magnetite deposits

Types of magnetite deposit

Two principal types of magnetite deposit are present in the district: deposits in skarn or marble (conveniently termed "skarn ores") _/, and deposits in microcline granite gneiss (conveniently termed "granite gneiss ores"). Locally, the host rock of both types of deposit has been modified in a characteristic way to give two important subtypes, hereafter termed "modified skarn ores" and "modified granite gneiss ores". Magnetite replaces, to some extent and in some places, almost every kind of metasedimentary rock found in the district. In addition, it cuts or replaces phacoidal granite gneiss, microcline granite gneiss, and pegmatite. However, sizeable concentrations of magnetite are found only in skarn, microcline granite gneiss, and their modified equivalents.

_/ The association of skarn and marble is so nearly universal, and the distribution of ore in the two rocks is so likely to be erratic in detail, that the term "skarn ore" is used for ore in marble as well as in the skarn itself, except when specific reference is required. This conforms to the usage of the term by some Fennoscandian geologists.

The magnetite deposits are closely related in space, time, and origin. All of them, regardless of type, are thought to be high-temperature replacement deposits effected by emanations from younger granite magma. The skarn ores are typical pyrometasomatic deposits of the kind widely distributed through geologic time but especially common in the pre-Cambrian. They are strikingly similar to those of Central Sweden. The granite gneiss ores, on the other hand, have few known parallels, and these are restricted to the pre-Cambrian.

Skarn

Because of the close genetic and spatial relations between skarn and one major type of magnetite deposit, it seems desirable to discuss certain features of the skarn in this section on the ore deposits. In a broad sense, skarn is both the host rock and the gangue of the skarn ores.

Definition and general discussion

Skarn is an aggregate of Ca, Mg, Fe (and sometimes Mn) silicates characteristically associated with certain metasomatic deposits of iron ore and sulfides. The skarn minerals are principally those of the pyroxene, garnet, and amphibole groups, though the epidote, olivine, humite, scapolite, and mica groups may also be represented. Fluorite is a frequent and sometimes abundant accessory constituent. The term skarn is generally restricted to dark or at least distinctly colored aggregates, though some writers apply it as well to pale or colorless masses of diopside, tremolite, or anthophyllite. Some carbonate almost invariably accompanies the skarn, as disseminated grains, knots, lenses, interbeds, or associated layers. This definition of skarn is based chiefly on discussions of Eskola (1939, pp. 383-384) and Geijer and Magnusson (1944, p. 119).

The oxides magnetite and hematite _/, and the common sulfides pyrrhotite, chalcopyrite, sphalerite, and (rarely) galena are often part of the "skarn-mineral association", though their presence is by no means essential for calling a silicate assemblage skarn.

Skarn generally occurs at or near the contacts between limestones or dolomites and granitic rocks, and the belief is widely held that the skarn has developed by metasomatism of the carbonate rocks by emanations from cooling granitic intrusives. Emanations from intrusives of alkaline, intermediate, or mafic character may also give rise to skarns, but these are not discussed in this article.

In addition, a number of other genetic types of skarn are recognized by some Fennoscandian geologists. These types, all embraced by the broad designation "reaction skarns", include those resulting from the metamorphism (and metamorphic differentiation?) of interlayered impure calcareous and siliceous beds; "low-temperature skarns" associated with the supracrustal leptites; "secondary skarns" formed by additional metasomatism of earlier-formed "low-temperature skarn"; skarns formed by metamorphic hybridism; and others. (See, for example, Geijer and

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_/ Note, however, that primary hematite is <u>not</u> present in skarns of this district.

Magnusson, 1944; Hjelmqvist, 1942; Korjinsky, 1945; Magnusson, 1928, 1930, 1940a, 1940b; Mikkola, 1947; and Ramberg, 1944.) Finally, one might mention the skarns that form part of the "basic fronts" ascribed to the advance wave of regional granitization.

Recently, some writers have departed widely from accepted usage and have applied the term to metasomatically-formed masses rich in Mg-Fe-Al silicates and comparatively poor in Ca silicates, showing no genetic relation to carbonate rocks. (See, for example, du Rietz (1945) -- replaced liparites and banded tuffs at Laver, northern Sweden; and Holmes and Reynolds (1947) -- biotite-rich rocks derived by "regional" metasomatism of quartzite and epidiorite, County Donegal, Ireland.)

The writer would suggest that the unqualified term skarn be restricted to the silicate masses having a demonstrable or inferred genetic connection with limestones or dolomites.

The relation between skarn and tactite deserves comment. Someone has facetiously remarked that skarn is what the Fennoscandian geologists call tactite. However, as the term tactite was originally applied by F. L. Hess (1919) to rocks of the innermost part of contact-metamorphic aureoles, it is more accurate to say that tactite is one variety of skarn. Moreover, as Geijer and Magnusson (1944, p. 119) have aptly noted: "The designation 'contact-metasomatic' is to a certain extent misleading [when applied to skarn deposits], for deposits of this kind may develop even at a considerable distance from the eruptive mother-rock." _/ The concept of proximity, with or without actual contact, is clear in Eskola's discussion of skarns (1939, p. 384). In good truth, Lindgren (1922, p. 293) was careful to note, in his definition of pyrometasomatic deposits, that these are found at or near intrusive contacts. However, we sometimes tend to overlook the broader definition and regard the deposits as strictly contact effects.

The strong general similarity that unites all pyrometasomatic deposits has been documented by many geologists, notably Lindgren (1933) and Knopf (1942). Nevertheless, there are significant differences even among closely related members of the broad class, and these differences--particularly the chemical ones--are worthy of future detailed investigation.

Mode of occurrence in the district

In the Adirondack region, the occurrence of skarn is restricted to the massif itself, that is, to the once-hotter, perhaps deeper area now characterized by the predominance of igneous rocks such as granites, quartz syenites, and anorthosites. The vast area of the Grenville Lowlands, underlain by calcitic and dolomitic marbles, quartzites, biotite-quartz-plagioclase gneiss, and subordinate gabbro sheets and phacolithic granite masses, is devoid of skarn, though small masses of diopside, tremolite, and other pyrometasomatic minerals have developed locally. Some marble is present within the massif, but in general marble is rather sparse and skarn is rather abundant. The pyrometasomatic minerals of the Lowlands are all poor in iron; tremolite

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_/ Translated from the Swedish text by Leonard.

is common; and garnet is unknown. In contrast, the skarn minerals within the massif are generally iron-bearing; tremolite is rare; and garnet is locally abundant.

Within the St. Lawrence County magnetite district, skarn occurs most widely as layers or masses in granite, either alone or associated with such rocks as pyroxene gneiss, amphibolite, quartz-feldspar granulites, and calcitic marble. (Dolomitic marble is absent from areas where skarn is found.) In addition to the sporadic layers or lenses of calcitic marble, calcite is present as knots, disseminations, or late (secondary) veinlets in the skarn itself. A common feature of the thicker skarn layers is the gradation of marble or calcareous diopside skarn through pale-green and medium-green skarn to a very dark green or almost black pyroxene skarn that may or may not contain appreciable amounts of dark reddish brown garnet rich in the andradite molecule.

The conclusion seems inescapable that the skarns of the district developed from calcitic or dolomitic limestones by progressive introduction of substance, mainly SiO_2 and Fe. The previous nature of the carbonate rocks, whether pure or impure, is unknown. Several tenuous bits of evidence point toward relatively pure carbonate rocks as the original host, but the argument built on these is far from compelling. It is still more difficult to determine whether the host rocks were originally dolomitic or calcitic. The marble now found with skarn is entirely calcitic. This proves nothing regarding its antecedents, as the calcitic marble might be recrystallized calcitic limestone or the expectable calcitic residue from the progressive but incomplete metamorphism of mixed Ca-Mg carbonates. Some facts and suggestions bearing on the two problems are given below.

Large areas of relatively pure dolomitic and calcitic marbles are found in the adjacent Grenville Lowlands. Impure types are less abundant. It is reasonable to suppose that similar ratios prevailed in what is now the Adirondack massif. Many layers of skarn are homogeneous and almost monomineralic for thicknesses of tens of feet, suggesting that the host rock was similarly homogeneous. Where the skarn itself is inhomogeneous, one usually finds independent evidence that new material has been introduced, not that originally different material has been reconstituted. (See discussion below of garnet skarn, modified skarn, and skōls.) Isolated skarn-like boudins in calcitic marble seem to represent disrupted beds of impure calcareous material. As many of these boudins contain considerable quartz and a little feldspar, in addition to pyroxene, their character is similar to that of certain pyroxene gneisses and very different from that of the major skarn bodies. Had the latter been derived in like manner from what were probably sandy and somewhat argillaceous calcareous rocks, they too might be expected to contain quartz, feldspars and micaceous minerals. Yet they do not (except where the felsics have been introduced subsequently), either because they never had the appropriate chemical composition or because they have lost the embarrassing components by metamorphic differentiation.

The structural behavior of some highly contorted skarn masses in the district suggests that of calcitic marbles plastically deformed and subsequently metasomatized. However, in places where skarn is intimately involved with what is now partly-silicated calcitic marble, the relations between the two rocks are susceptible of two interpretations: the skarn represents metasomatized dolomite that was originally more brittle than the calcitic marble with which it was associated, and was also more favorable to metasomatizing solutions; or some calcitic marble, converted to skarn, has remained cushioned during deformation of the more plastic, relict calcitic marble that partly surrounds it. Many more facts are required before this suggestive evidence can be correctly interpreted.

The present composition of the skarns throws little light on the nature of their antecedent carbonate rocks. Some Fennoscandian geologists interpret tremolitic and diopsidic skarns (Ca-Mg-rich skarns) as metasomatized dolomites and ferrosalite-andradite skarns as metasomatized calcitic marble. While this interpretation may have merit locally, it is perhaps more dogmatic than the facts often warrant. All the writer would say regarding the antecedents of the St. Lawrence County skarns is this: if dolomite was the source, little or no Mg had to be introduced during metasomatism--an economical arrangement indeed; if calcitic marble was the source, much Mg was introduced (or re-circulated) and subsequently removed by progressive replacement with Fe.

Mineralogy of skarns in the district

The component minerals of the skarn may be classified roughly as follows:

Principal skarn minerals

pyroxene
garnet

Subordinate or infrequent (but locally abundant) skarn minerals

amphibole (clino- and ortho-)
mica (phlogopite and biotite)
scapolite
fluorite

Accessory skarn minerals (listed alphabetically)

allanite
apatite
bytownite (and labradorite)
calcite
datolite
epidote
goethite (?)
graphite
helvite (?)
hypersthene (?)
ilvaite (?)
monazite (?)
potash feldspar } locally prominent in "modified skarns"
quartz }
sphene
spinel
unknowns (several)
zircon

(In addition, certain minerals of late hydrothermal origin appear adventitiously in some modified skarn.)

Metallics in the skarn

magnetite
vonsenite
pyrrhotite
pyrite
chalcopyrite
molybdenite
sphalerite
loellingite
valleriite (?)
cuprite (?)
hematite (always secondary)
unknowns

Nomenclature and quantitative limits

The skarn is named for its principal mineral or minerals: for example, pyroxene skarn or garnet skarn (when essentially monomineralic); pyroxene-garnet skarn (roughly equal parts of the two minerals); garnetiferous pyroxene skarn (pyroxene skarn with relatively little garnet); etc.

Probably Buddington and Leonard have not rigidly observed quantitative limits for defining skarn, but in general they have restricted the term to rocks consisting of 90-95 percent Ca-Fe-Mg silicates, usually somewhat colored. When such minerals form less than that amount but more than 50-60 percent of the rock, a qualifying term is used, such as calcareous pyroxene skarn, scapolitic garnet-pyroxene skarn, feldspathic pyroxene skarn, The writer has used the collective term *modified skarn* for rocks in which felsic minerals (quartz, potash feldspar, or both--with or without scapolite) constitute 10 to 60 percent of the material, the rest being chiefly pyroxene, amphibole, or both these mafics. When the felsic varieties are distinctly foliated, they may appropriately be termed gneisses; for example, pyroxene-quartz-scapolite gneiss. Mafic aggregates occurring in quartz or pegmatite veins have not been termed skarn unless their derivation from skarn is unquestionably indicated by the field relations.

Main varieties of skarn

Unmodified skarn

The two principal varieties of skarn (pyroxene skarn and garnet skarn), and two less abundant but important varieties (amphibole skarn and mica sköls), require description. Each of these may be essentially monomineralic, and the pyroxene skarn may be transitional into one or more other varieties. In all varieties except garnet skarn, there is a progressive change in the composition of the individual skarn minerals from Fe-poor to Fe-rich types, with concomitant increase in Al, (Ti), (Mn), and--where appropriate--in K, OH, and F. Within the district we find garnet-pyroxene skarn, mica-pyroxene skarn, and amphibole-pyroxene skarn. However, we do not find amphibole-mica skarn (minor exceptions), amphibole-garnet skarn, or garnet-mica skarn. The replacement relations, paragenetic sequence, and chemical composition of the constituent skarn minerals show that, with minor exceptions, all the types have probably developed by progressive introduction of substance into simple diopsidic skarn. However, the possibility of direct conversion of pure or impure carbonate rocks to some varieties of skarn cannot always be excluded. The lack of certain skarn types is a consequence of the chemical composition of the particular mineral assemblages and the sequence in which they developed. A fuller treatment of the chemistry of the skarns must be deferred for separate publication.

A summary of the average mineralogy of the varieties of skarn is given in Table 3.

Table 3. Approximate mineralogic composition of skarns. _/

Mineral	Diopside skarn	Salite and ferrosalite skarn	Garnet-pyroxene skarn, etc.	Pyroxene-amphibole skarn
Pyroxene	di	sal or fs	fs--5-90% (30-50% is c.)	sal or fs
Amphibole	act - r.	act to fhst 0, l., 10-15%	---	act to fhst
Mica	phlog--tr. to 15%	phlog. to bi 0, l., 10-15%	---	0 (present in one act skarn)
Calcite (prim.)	±ubiq.-<5-15%	p. or 0 - loc. 15%	tr. to 15% or >	0 or l.
Garnet	---	spor. access.	andr-->90-5 (50-80% is c.)	---
Quartz	---	l.-not c.	---	0, tr., l.
Potash feldspar	---	tr.-extr. r.	tr.-extr. r.	---
Plagioclase	---	---	---	---
Magnetite	few % loc.	0 or few %	gen. 0	0 or tr.
Sulfides	tr. to sev.%	tr. to sev.% loc. 0, loc. abund.	0 to sev.%; loc. abund.	0, tr., l.
Graphite	spor.	---	---	---
Apatite	vy. r.	0 or tr.	0 or l.	0 or tr.
Zircon	---	tr.-vy. r.	0 (?)	0; r. tr.
Sphene	---	tr.-vy. r.	0, tr., l.	---
Allanite	---	0 or p.	gen. 0; loc. p.	0, tr., vy. l.
Fluorite	tr.-vy. r.	0, tr., or abund.	gen. 0; r. tr.	0 or tr.
Unknowns	spor.-tr.	spor.-tr.	0 or tr.	0 or tr.
Chlorite	vy. r.	0 or p.	---	0 or l.
Calcite (sec.)	?	often p.	---	0 or l.
Spinel	---	---	---	---
Monazite	---	---	---	---

Table 3 (continued)

Mineral	Clinoamphibole skarn	Anthophyllite skarn	Sköle	
			Light	Dark
Pyroxene	gen. fs - 0 or p.	---	di relics	---
Amphibole	gen. fhst.	anthoph., r. tr. fhst	loc. trem in crushes	anthoph only - l.
Mica	0; r. tr.	bi, loc. 0 but gen. abund.	phlog	bi- >50-99%
Calcite (prim.)	---	---	spor.	---
Garnet	---	---	---	---
Quartz	---	---	---	gen. 0; loc. p.
Potash feldspar	---	---	---	---
Plagioclase	---	---	---	gen. 0; An_{27} in one bi-anthoph sköl
Magnetite	---	0 or l.	spor.	spor.
Sulfides	0 or p.	0 (?)	spor.	spor.
Graphite	---	loc. p.	tr.	gen. 0; r. p.
Apatite	gen. 0; loc. p.	0 or tr.	---	spor.
Zircon	---	spor.	---	gen. p.
Sphene	---	---	---	---
Allanite	---	spor.	---	gen. p.
Fluorite	0, p.; loc. vy. abund.	---	---	0, or loc. abund.
Unknowns	0; r. tr.	0 or tr.	loc. talc (?)	r. tr. helvite (?)
Chlorite	---	---	l.	gen. 0; loc. p.
Calcite (sec.)	---	---	---	gen. 0; loc. p.
Spinel	---	0; loc. abund.	---	---
Monazite	---	spor.	?	---

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Mineral abbreviations	
act	actinolite
am	amphibole
andr	andradite
anthoph	anthophyllite
bi	biotite
di	diopside
fs	ferrosalite
fhst	ferrohastingsite
hb	hornblende
phlog	phlogopite
px	pyroxene
sal	salite
trem	tremolite

Quantitative expressions	
abund.	abundant
access.	accessory
c.	common
ca.	circa, about
dom.	dominant
extr.	extremely
gen.	generally
l.	little
loc.	locally
0	zero, absent
p.	present
prim.	primary
r.	rare
sec.	secondary
sev.	several
spor.	sporadic
tr.	trace
± ubiq.	more or less ubiquitous
vy.	very
>	more than
<	less than
?	questionable, uncertain

Pyroxene skarn

Pyroxene skarn is, above all others, the predominant skarn of the district. It ranges in color from white, yellowish, or very pale green through medium green to dark green or black. Typically, the skarn is a fine- to medium-grained, homogeneous, massive, equigranular or inequigranular, crystalloblastic aggregate of subhedral pyroxene crystals. Rarely, an extremely faint foliation, sometimes visible in hand specimens, is given by alternating zones of finer and coarser pyroxene. Some specimens of apparently massive skarn show pyroxenes with preferred optical orientation but no dimensional orientation. In general, however, the skarn appears undeformed. Coarse skarn is present only near veins of pegmatite or quartz.

The color change in the skarn reflects a progressive change in composition of the pyroxene from diopside ($Ca_{48.8}Mg_{46.8}Fe_{4.4}$) through salite to ferrosalite ($Ca_{48.6}Mg_{20.8}Fe_{30.6}$). (See Hess, 1949, analyses 35-39, pp. 662-664). The change is effected by an increase mainly of Fe^{+2}, Fe^{+3}, and Al at the expense of Si and Mg, not merely by the substitution of Fe^{+2} for Mg^{+2}. Thus the trend is not directly toward hedenbergite, the Ca-Fe end-member, but toward a pyroxene high in Fe^{+3} and Al as well as Fe^{+2}. Owing to the marked pleochroism of the salite and ferrosalite, clinopyroxenes of this type have frequently been mistaken for hornblende or acmite.

In the magnetite-bearing part of the district, diopside skarn is found only in the thicker skarn zones; that is, those generally more than 50 feet thick. Its gradation into calcitic marble is usually evident in the field. Mineralogically, the diopside skarn is the simplest variety found in the district. Characteristic associates of the diopside are phlogopite and primary calcite, the latter as polygonal grains or interstitial aggregates of polygonal grains. Ore is not known to occur in this variety of skarn.

Salite and ferrosalite skarn are the most abundant and widespread varieties. Thin skarn zones (a few inches to 10 feet thick) are usually made up of salite or ferrosalite with little or no carbonate, and the two varieties commonly dominate the thicker skarn zones, where they may or may not be slightly calcareous. Though either variety may pass directly into calcitic marble, the change is usually gradational through diopsidic skarn. However, this transition is almost always layer-like; relics of lighter (diopsidic) skarn in darker (salitic or ferrosalitic) skarn, or knots of darker skarn in lighter skarn, are extremely rare. Mineralogically, the two varieties are more complex than diopside skarn (see Table 3), and they are the preferred host of ore.

Garnet skarn

Garnet skarn appears to form a complete gradational series with ferrosalite skarn, though the end-members are more common than intermediate varieties. All the garnet is dark, reddish-brown andradite ($Andr_{56}Gross_{32}Alm_8Pyr_3Spess_1$)._/ Among skarn minerals of the district

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_/ Unpublished chemical analysis by the late Norman Davidson, U. S. Geological Survey.

it is unique, for it does not show the customary variation from an Fe-poor toward an Fe-rich type. _/ The andradite is associated with ferrosalite ($Ca_{49.5}Mg_{26.5}Fe_{24}$--Hess, 1949, analysis 38, p. 663), which it replaces initially as minute vermicular intergrowths, bead-like strings, veinlets, and networks. This replacement of pyroxene by garnet is evident in all but the most garnet-rich variety, where the two minerals might be interpreted as contemporaneous. Most of the garnetiferous skarn has a regular or irregular planar structure, ranging from indistinct to rather pronounced, given by alternating layers richer and poorer in garnet. These dark reddish-brown and dark green layers are highly variable in thickness as well as in garnet-pyroxene content. The writer believes that slight structural inhomogeneities in pyroxene skarn were essential for its subsequent replacement by garnet. Moreover, replacement took place only after the pyroxene had approached or reached ferrosalite in composition. At that stage, some ferrosalite was converted to andradite by a greatly increased access of iron in the form of Fe^{+3} ions, accompanied by increasing Al with minor Ti and Mn.

A curious feature of some garnet skarn is the presence of irregular, fist-sized aggregates of coarse calcite, locally studded with very coarse andradite crystals. Similar aggregates of pure or impure

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_/ Yellowish-brown grossularitic skarn has been found with salite skarn in just one place out of scores of garnetiferous skarns in the district.

carbonate are seldom observed in other varieties of dark skarn.

Garnet skarn or garnet-pyroxene skarn contains local concentrations of sulfides and very rarely has a little magnetite, but it is never the host for magnetite ore.

Amphibole skarn

Amphibole skarn occurs sporadically in the district. It is abundant in one magnetite deposit (Jayville), where both orthorhombic and monoclinic amphiboles are present. Elsewhere, clino-amphibole skarn alone has been recognized, and that only in minor amounts. The clino-amphiboles range in composition from actinolite through "hornblende" to ferrohastingsite. However, actinolite is rather rare (occurring locally in diopsidic skarn), and ferrohastingsite seems predominant. The anthophyllites at Jayville show a rather limited range of aluminous, ferroan types. In consequence of their relatively high iron content, all amphiboles except the pale actinolite are distinctly green, very dark green, or black. The clino-amphiboles may form nearly equant grains, readily mistaken for pyroxene, or they may be moderately elongate. The anthophyllites are prismatic but not fibrous. _/

_/ X-ray studies by J. M. Axelrod confirm the writer's optical identification of the material as ortho-amphibole.

Clino-amphibole skarn, with or without some pyroxene, occurs as knots in calcite or in dark pyroxene skarn, as thin layers (millimeters to feet thick) in or with dark skarn, as small masses, and as thin selvages between pyroxene skarn and alaskitic (?) granite. Anthophyllite skarn forms distinct layers in the dark skarn at Jayville. The fabric of amphibole skarn may be random, foliated but not lineated, or distinctly lineated. As a rule, amphibole skarn is coarser than the pyroxene in it or associated with it. Moreover, where amphibole and pyroxene occur together, the amphibole is always the younger mineral. Though some amphibole skarn may have formed directly from carbonate rocks, most of it is a replacement of earlier-formed pyroxene skarn. Ferrosalite was converted to ferrohastingsite by addition of Fe^{+2}, Fe^{+3}, Al, K, OH, and F. It is noteworthy that the only deposit with abundant amphibole skarn (Jayville) also has abundant fluorite intergrown with the ferrohastingsite.

Amphibole skarn (both clino- and ortho-) is the host for some ore at Jayville, and ferrohastingsite is commonly present with ferrosalite in ore from the Hardwood Mill deposit. Elsewhere, magnetite is rarely found in amphibole skarn. Instead, the rock is likely to have minor amounts of sulfides.

Mica sköls

A feature of some skarns is the presence of mica as contorted, streaky masses, thick selvages, partings, or thin films in skarn or magnetite ore. The selvages and partings are analagous to one type of sköl, or shell, often noted by Fennoscandian geologists (e.g., Bugge, 1940, p. 16; Eskola, 1914, pp. 226, 259; Geijer and Magnusson, 1944, p. 140; Hjelmqvist, 1942; Magnusson, 1940b, p. 185).

Sköls are prominent in three magnetite deposits of the St. Lawrence County district (Clifton, Jayville, and Outafit), where they carry a considerable part of the ore. Very minor sköls are present at several other deposits. All the sköls are closely associated with ore; none has been found in thicker skarn zones altogether lacking in evidence of magnetite mineralization. The sköls are of two types, light and dark, corresponding to phlogopite and biotite, their dominant minerals. Both varieties of sköl are usually present in the major deposits, and both may be mineralized, but the biotite sköls are more abundant and more favored by magnetite.

Sköls range in thickness from mere films to layers several feet thick. Some sköls a few mm thick, occurring in ore, have their mica flakes perpendicular to the layering of the ore; and some thicker sköls have a hit-or-miss (or decussate) fabric. However, the mica flakes in many sköls are parallel to the layering of the rock. As biotite is sometimes the only visible silicate in high grade ore layers tens of feet thick, one may infer that these ore layers were formerly very thick sköls. _/ The sköls may be distributed sporadically through a thick zone of skarn and ore (Clifton, Outafit), or they may be best developed at the borders of the skarn zone (Jayville). At Clifton

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_/ Buddington (personal communication) disagrees. He believes that the scattered biotite present in these thick ore layers represents some mineral (such as pyroxene) left as sporadic relics that were made over into biotite by the same solutions that deposited magnetite. The quantity of biotite developed would thus represent only a fraction of the pyroxene formerly present.

there is some suggestion that sköls are best developed on the flat parts of prominent rolls, but this is not always the case.

The light sköls, like the light skarns, are simple in mineralogic composition. (See Table 3.) Some light sköls contain relics of diopside and calcite and are more or less transitional into phlogopitic diopside skarn. The phlogopite of the sköls has a silvery appearance, locally with a pale greenish tinge. Except for its association, one might mistake it for muscovite. However it has the composition of a fluor-phlogopite:

$(OH_{.55}F_{.99}Cl_{.01})(K_{.86}Na_{.06}Ca_{.01})(Mg_{2.62}Mn_{.01}Fe^{11}_{.16}Fe^{111}_{.05}Ti_{.01})(Si_{2.90}Al_{1.04})O_{10}\cdot-$

_/ Unpublished chemical analysis by E. K. Oslund; H_2O = 1.21 percent, F = 4.54 percent, Cl = 0.11 percent. X-ray study by J. M. Axelrod confirms the writer's optical identification of the mineral as a mica.

The dark sköls are more complex in mineralogic composition, and some from Jayville are fluorite-rich. The biotite is green to greenish-black, flexible but inelastic, and has two nearly perpendicular cleavages that cut the basal cleavage flakes. Megascopically, it resembles a chlorite more than it does a mica, but it has the composition of an iron-rich biotite:

$(OH_{1.94}F_{.12}Cl_{.02})(K_{.86}Na_{.10})(Mg_{2.01}Mn_{.01}Fe^{11}_{.56}Fe^{111}_{.13}Al_{.18}Ti_{.02})(Si_{2.70}Al_{1.30})O_{10}\cdot$

_/ Unpublished chemical analysis by E. K. Oslund; H_2O = 4.00 percent, F = 0.53 percent, Cl = 0.11 percent. X-ray study by J. M. Axelrod confirms the writer's optical identification of the mineral as a mica.

The two types of mica form an essentially continuous series of sköls whose intermediate members have not been investigated chemically. Comparison of the two analyzed end-members shows that the change from light to dark mica requires an increase in Fe^{+2}, Fe^{+3}, and Al, the same major ions increasing in the pyroxene and clino-amphibole series.

From their field relations, paragenetic sequence, and close relation to ore, the sköls appear to have developed "late" in the sequence of skarn formation. In this respect they agree with certain Fennoscandian sköls, many of which have been attributed to late hydrothermal alteration of skarn. The St. Lawrence County sköls may in part be reconstituted aluminous zones of the original carbonate rock, but in large part they probably represent sheared (?) skarn reconstituted by volatiles and developed slightly in advance of the magnetite mineralization.

Modified skarn

Two main types of modified skarn are present in the district. One is a widely-distributed scapolitic skarn that is essentially barren. The other, a local type characterized by quartz and potash feldspar in variable proportion, is the host of a very promising ore body. The writer believes that modified skarn represents earlier-formed skarn partly replaced by quartz, potash feldspar, or scapolite in variable combination.

Modified skarn, generally scapolitic

Modified skarn of this type is found in subordinate quantity in almost every thick skarn zone of the district, regardless of whether the zone carries magnetite bodies. The modified skarn occurs as layers a few inches to many feet thick within dark, unmodified skarn. Locally, it forms discrete thick layers near the major skarn zones, between younger granite and metasedimentary gneisses. Elsewhere it is present as wispy zones and thin interlayers in pyroxene gneiss or granulite, particularly where intense local deformation affects the sequence. Modified skarn tends to develop at or near contacts between skarn and granite or pegmatite, though this association is not universal.

The rock is generally dark-green to black, speckled, fine- to medium-grained, and massive. Locally, part or all of a given layer has a faint to pronounced planar structure. The micro-fabric of the rock may be uniformly polygonal-crystalloblastic or highly irregular. Its mineral composition is variable. (See Table 4.) Dark-green ferrosalite is always present, almost always accompanied by scapolite, which ranges in composition from Me_{49} to Me_{74}. Some potash feldspar generally accompanies the scapolite, and locally this feldspar is the only light mineral present. The feldspar is untwinned or slightly shadowy; rarely it displays the grid structure of microcline. (See discussion of feldspar in mineralogy of granite gneiss ores.) A little primary calcite may or may not be present. Andradite garnet is often an accessory or major constituent, though it may be altogether lacking. Sphene is the characteristic accessory, locally abundant. Where the fabric is non-polygonal, pyroxene is seen to be early; garnet, scapolite, feldspar, and sphene are later, replacing the pyroxene, part of which

Table 4. Approximate mineralogic composition of modified skarns. _/

Mineral	Generally scapolitic skarn	Modified skarn, Trembley Mountain _/		
Pyroxene	gen. dom.	sal or fs-18%	sal or fs-26%	60-33%
Amphibole	gen. 0; loc. l.	hb?-30%	hb?-20%	2% ±
Mica	---	bi-p.	bi-p.	bi-p.
Calcite	spor.	---	---	---
Garnet	0, tr., l., dom.	---	---	---
Quartz	0; r. pl	35%	21%	---
Potash feldspar	gen. p.	---	20%	22-46%
Plagioclase	extr. r. -An_{60}	p.	tr.	0-1%
Scapolite	major-r. 0	---	---	---
Magnetite	gen. 0	15%	12%	11-17%
Sulfides	p. or 0	---	p.	p.-6%
Graphite	---	---	---	---
Apatite	gen. p; loc. abund.; loc. 0	tr.	p.	p.
Zircon	p. or 0	p.	p.	tr.
Sphene	c. access., loc. abund.	p.	---	---
Allanite	p. or 0; loc. abund.	p.	p.	p.
Fluorite	0, p., loc. abund.	---	---	---
Unknowns	?	---	---	---
Chlorite	r.	---	---	p.
Calcite (sec.)	l.-spor.	---	---	p.

Table 4. (continued)

Mineral	Generally scapolitic skarn	Modified skarn, Trembley Mountain		
Spinel	---	p.	p.	---
Zoisite	?	---	p.	p.
Epidote	tr.--0 or p.	---	tr.	tr.
Pumpellyite	spor. tr.	---	p.	p.

_/ For abbreviations, see ftn., Table 3.

_/ All are mineralized.

may be altered to blue-green amphibole. Sulfides appear sporadically in the modified skarn, but magnetite is generally absent. Late alteration evidenced by small amounts of quartz, epidote, pumpellyite, chlorite, and calcite affects the rock locally. The accompanying unmodified skarn commonly shows little or no trace of this later alteration.

In the field, the scapolitic pyroxene skarn is readily mistaken for amphibolite. Even after the skarn-like character is recognized, one is likely to call the rock feldspathic, rather than scapolitic. Color is no guide, as both light minerals may be white, gray, yellowish, or green. The greasy luster of scapolite, when apparent, is diagnostic. It is important to recognize these rocks for what they are, as the scapolitic skarn represents a significant concentration of halides, not only in the scapolite itself but also in the fluorite and apatite that are locally abundant in it.

Modified skarn with quartz or potash feldspar

Modified skarn with quartz, potash feldspar, and subordinate scapolite is the principal host for ore at the Trembley Mountain magnetite deposit. Similar rock occurs very sparingly in the wall rocks of several magnetite deposits of the granite gneiss type, notably at Spruce Mountain Northwest, a deposit in the same general belt of metasediments as Trembley Mountain.

At Trembley Mountain, biotite-microcline granite gneiss and associated metasediments enclose a zone of heterogeneous rocks 120 to 300 feet thick. The zone, much of which is mineralized, consists of skarn, modified skarn, quartz rock, pegmatite, microcline granite and syenite, sodic syenite, aplite, and metasedimentary gneisses. Skarn and modified skarn constitute 40 to 65 percent of this zone, with modified skarn predominant. The unmodified skarn consists of salitic and ferrosalitic skarn with subordinate calcareous actinolite-diopside skarn. By a transition that is usually gradual, the unmodified skarn loses its homogeneous aspect and becomes a modified rock containing quartz, feldspar, or both minerals, to the extent of 10 to 60 percent. Scapolite is present locally, instead of quartz and feldspar.

The modified skarn is almost always composed of dark-green pyribole and white, gray or pale-green felsics. Light-green skarn may show thread-like, ramifying quartz veinlets with "hazy" borders, but that is the extent of the modification.

Dark skarn, in contrast, becomes veined, flecked, speckled, and blotched by felsics. Wispy and patchy areas of light minerals appear. Occasionally, the felsics are scattered rather evenly throughout the skarn, but typically the modified rock is distinctly heterogeneous. In many places, a crudely-foliated, migmatitic rock is developed. Sometimes the felsics spread out between adjacent layers, transecting, engulfing, and displacing the intervening mafics. Replacement augen develop. Elsewhere, the alternation of light and dark laminae is more regular and the boundaries sharp. Masses of quartz rock, pegmatite, and green granitic rocks--in part contaminated--are interspersed throughout the modified skarn. In many places, the modification of the skarn has proceeded outward from these layers of felsic rocks. However, the relation is not universal, for in other places skarn that borders the felsic rocks is affected slightly, or not at all.

Because the quartz and feldspar present in modified skarn are so similar in appearance, it is impracticable, or even impossible, to estimate their relative proportion in the field. Wherever the two are easily distinguishable, their relations toward the skarn are seen to be identical. In addition, it is clear that the quartz-feldspar ratio in the masses, thin layers, patches, and knots of felsics is highly variable.

The writer interprets the megascopic evidence as indicating metasomatic replacement of skarn by quartz and feldspar. Regrettably, the microscopic evidence does not always confirm such an interpretation, though it by no means denies it. The conflicting evidence is noted below.

The microfabric of modified skarn is variable. Commonly, the rock has an inequigranular or fine-grained and equigranular texture, a faint or crude foliation, and polygonal or rounded grains. In contrast, other specimens of the rock show coarse, interlocking leaves of quartz that appear to have replaced pyroxene. Potash feldspar, when present, may show a similar habit. The quartz and feldspar have shadows. Both types of microfabric--polygonal-crystalloblastic and "replacement" leaves--are shown by rocks that look identical in hand specimen and megascopically show felsic minerals replacing skarn.

Estimates of the mineral composition of representative specimens of modified skarn are given in Table 4, nos. 2, 3, 4. One may consider that the "end-members" have either quartz or feldspar as the sole felsic, while intermediate members have varying proportions of the two. No attempt is made in Table 4 to give maximum and minimum values for the felsics, as the gradational nature of modified skarn has already been emphasized. Scapolitic skarn, megascopically very difficult to recognize, is certainly less abundant than the quartz-feldspar type but not infrequent.

Essential minerals of modified skarn are salitic or ferrosalitic clinopyroxene and a felsic mineral. Blue-green or brown-green amphibole is an additional essential constituent in more than half the specimens studied, and some mineralization of the rock is so characteristic that magnetite might well be considered essential too. Some quartz-rich modified skarn has a little orthopyroxene. The feldspar, where present, is the untwinned potash variety, accompanied by a very little altered plagioclase. Accessories are brown mica, apatite, zircon, allanite, sphene, magnetite (where not present as an essential), and sulfides. Apatite, zircon, allanite, and magnetite are almost ubiquitous; the rest are sporadic. Secondary minerals, also sporadic, comprise chlorite, carbonate, sericite, zoisite, and pumpellyite. Locally, some clinopyroxene has been partly converted to secondary amphibole.

Skarn ores

General characteristics

The skarn ores are replacement bodies of magnetite in skarn or calcitic marble. Most of the ore in major deposits of this type is in the skarn itself, with subordinate magnetite in marble. Deposits exclusively in marble appear to be rare and commercially insignificant. Magnetite deposits in skarn are generally compact, small to moderate in size, highly variable in grade, and complex in structure. Often they contain layers of high-grade ore accompanied by medium- and low-grade ore, but as a class they are characterized by compact--rather than diffuse--mineralization.

Magnetite is the only ore mineral in the skarn deposits. It is accompanied by variable but generally small amounts of sulfides, chiefly pyrite and pyrrhotite. Details of the metallic minerals and their paragenetic sequence are discussed separately. The associated non-metallics that make up the gangue are merely the unreplaced minerals of the skarn and sköls described in the preceding section. Among these rocks the ore shows a marked preference for pyroxene skarn and biotite sköl. Chloritization is slightly more common and more extensive in mineralized skarn, but in most places even this alteration is still extremely mild. Rarely, a bit of stilpnomelane (?) accompanies magnetite in ore-bearing pyroxene skarn. However, the major changes attendant on mineralization have taken place <u>within</u> the lattice of the individual skarn minerals, developing more iron-rich and more aluminous varieties from initially simpler members of the species.

Contacts between ore and wall rock are generally sharp. In places, the skarn for several feet adjoining the main magnetite layer, or at a distance of several feet below it, may have thin parallel magnetite stringers or--very infrequently--a little uniformly disseminated magnetite, commonly with considerable sulfides. Nevertheless, "assay walls" are generally lacking: one is in ore or in wall rock, with no teasing transitions.

Description of typical ore

Three main grades of ore are recognizable, and each has a fairly distinct appearance. High-grade ore (more than 50 percent magnetic Fe) consists of magnetite with scattered grains or very small aggregates of green to dark-green pyroxene or white, gray, or yellowish calcite. Dark-green mica may be present as fine, scattered flakes or thin, continuous films. Rarely, very small aggregates or thin layers of dark-green or black amphiboles are seen. The magnetite and impurities are really aggregates of roughly equigranular, polygonal grains a few millimeters in diameter. Moreover, much of the megascopically pure magnetite has scattered crystals or fine aggregates of the same impurities, only the relative proportions being different. Very rarely, one finds almost pure magnetite (less than one percent silicates) madê up of diversely-oriented polygons of magnetite one-half to three-quarters cm in diameter. The coarse texture is readily visible in hand specimens, but the differing orientation of the grains can be seen only after the ore has been sawn or polished.

Though the individual magnetite grains are polygonal and locally rather fine, the ore is firm and compact. "Shot ore" of the type found in some magnetite deposits is rare in this district.

Medium-grade ore (35 to 50 percent magnetic Fe) consists of layers rich in magnetite alternating with layers rich in dark pyroxene, biotite, or--rarely--amphibole, locally with a little calcite. The individual layers are commonly a quarter of an inch to several inches in thickness, rather poorly defined, and highly variable within the limits noted. The layers are crudely tabular but irregular in detail.

Typical specimens of low-grade ore (20 to 35 percent magnetic Fe) are harder to find. They are similar in character to the medium-grade ore, but the magnetite layers tend to be thinner, more sharply defined, and more uniform in thickness. Some specimens with biotite are remarkably platy; others show small-scale crenulation.

Obviously, ore as mined may include the three types described and some barren partings as well. Two or more discrete ore layers separated by barren or weakly-mineralized rock may be considered for mining purposes as one "ore zone" carrying material of reduced grade. A point to be noted, however, is the almost complete absence of uniformly disseminated, relatively unlayered magnetite giving low-grade ore or mineralized rock. Locally one finds marble with small, scattered, amoeboid patches of magnetite; but skarn or sköls containing disseminated magnetite are rare indeed.

Sulfides may or may not accompany any of the ore types just described. Their occurrence seems entirely sporadic.

Preferred host of ore

For the district as a whole, salitic or ferrosalitic pyroxene skarn is the preferred host of ore. The range in composition of the pyroxene from ore layers is from Fs_{12} to Fs_{30}, but there is a pronounced clustering in the range from Fs_{15} to Fs_{25}. The host next preferred is a mica sköl, generally the dark, biotitic variety. The third host rock appears to be marble, though it may be that ore showing relict calcite granules has actually replaced a calcareous pyroxene skarn, selecting pyroxene in preference to carbonate. (The scarcity and small size of deposits exclusively in marble has previously been noted.) The fourth host is ferrohastingsite skarn, with or without relict ferrosalite. At a given deposit, the above order may or may not obtain. At Jayville, for example, mica sköls and amphibole skarn are preferred above pyroxene skarn. The general regional order of preference, however, is often the one that obtains in an individual thin section. Here, for example, magnetite replaces pyroxene and mica in preference to calcite; and pyroxene seems generally preferred above mica or amphibole.

Some varieties of skarn are definitely unfavorable hosts for ore. These include diopsidic skarn, highly ferrosalitic skarn, andradite skarn, and scapolitic skarn with pyroxene or pyroxene and garnet. _/

On the other hand, andradite skarn, and highly ferrosalitic pyroxene

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_/ Scapolitic pyroxene skarn serves locally as the host of ore at the modified skarn deposit of Trembley Mountain.

skarn with ferrohastingsite, may carry concentrations of sulfides.

Andradite skarn with some ferrosalite locally contains a very little magnetite. Thin sections of such material show magnetite to be younger than both pyroxene and garnet but definitely favoring pyroxene for replacement. However, no concentrations of magnetite occur in andradite skarn. This seems to be a rule among pre-Cambrian skarn deposits of magnetite. (See, for example, Bugge, 1940, p. 100; Geijer, 1936, p. 153; Geijer and Magnusson, 1944, pp. 144-145.) A notable exception familiar to the writer is the Sulphur Hill deposit, Andover district, New Jersey (Sims and Leonard, in preparation), where magnetite replaces andradite skarn in preference to other skarn varieties.

Space relations of skarn ores

The broader features of the distribution of magnetite deposits with respect to granite masses and regional structure have already been noted. Let us now consider some of the space relations of the individual skarn ore deposits.

The wall rocks immediately enclosing the main ore-bearing skarn zones fall into two groups: granitic rocks and metasedimentary rocks. The granitic rocks include alaskite, microcline granite gneiss, hornblende or hornblende-pyroxene syenite (a contaminated facies of hornblende-microperthite granite), medium-grained hornblende granite gneiss with local augen structure, phacoidal granite gneiss with metasedimentary interlayers, and rare soda granite. The metasedimentary rocks include pyroxene-quartz-feldspar gneiss and granulites (locally with thin marble interlayers), and rare biotite-quartz-plagioclase gneiss. The granitic rock may form the footwall, the hanging wall, or both walls of the skarn zone. There is no association of skarn with a particular wall rock, though the combination alaskite-pyroxene gneiss is rather common. One must also keep in mind that all the deposits are within 500 feet of at least one significant granite mass.

The average thickness of the main ore-bearing skarn zones ranges from 45 feet to more than 240 feet. The thickness of the main skarn zone is highly variable at some individual deposits: 12 to more than 200 feet at Clifton (average 100 feet or more), less than 5 to 150 feet or more at Jayville (average 75 to 90 feet), 65 to 175 feet at Brandy Brook (average 120 feet). At other deposits, less well known from drilling, the thickness seems less variable: 40 to 50 feet at Hardwood Mill, 166 to more than 310 feet at Outafit. Some of these deposits have minor skarn zones in addition to one large one that carries all or most of the ore. For each deposit as a whole, it is safe to say that the magnetite never picks a thin skarn zone to the exclusion of a thick one. However, one sometimes finds along a given drilled section that the main ore in the main zone is lacking or very thin, whereas a minor skarn zone in the hanging or footwall continues to carry magnetite, perhaps even increasing in quantity or grade.

The ore is always thinner than the skarn zone that carries it, but otherwise there is no correlation between the thickness of the main skarn zone and the ore it contains. The amount of skarn replaced by "mineable" ore ranges from 8 percent to 70 percent. Among the most promising deposits it is 25 percent or more.

The distribution of sköls within the skarn zones has been discussed in the section on "sköls". The distribution of marble requires comment. Marble layers, partly silicated with pyroxene, mica, etc., are found sporadically within most of the main skarn zones and as layers with or without skarn in minor zones above or below the main skarn zones. These marble layers range in thickness from a fraction of an inch to several feet, rarely reaching 40 feet. As previously noted, they often grade into skarn. They may or may not be partly mineralized. They have no constant stratigraphic position within skarn zones, and their distribution even in a given deposit is likely to be erratic, due to flowage of plastic marble or to varying intensity of silication. The main ore layers bear no constant relation to the marble. In contrast, it has frequently been noted for ore deposits in contact-metamorphic aureoles that the ore lies on the limestone side of the silicate zones. (e.g., Umpleby, 1916.)

Position of ore within the main skarn zone

The ore may be confined to a single layer, locally with one or more minor magnetite-bearing zones too thin or too poor in grade to constitute ore; or it may be present as two principal layers with subordinate layers of no economic importance. Commonly the two layers differ in grade. The position of the main ore layer or layers within the skarn is highly variable. If a single layer is present, it may be near the top, middle, or base of the main skarn zone. If two layers are present, one is likely to be in the upper half of the skarn zone and one in the lower half, but the two may be in the middle of the zone.

Locally, two layers may coalesce to give ore throughout the skarn zone, leaving a little unmineralized skarn at top and bottom.

The position of the ore varies not only from one deposit to another; it may also be highly variable within a given deposit. The Clifton ore body, part of which is well known from drilling and stoping, shows a high degree of complexity. Here the position of the ore varies along the strike and, locally, down the dip of the folded skarn zone. In very general terms, the ore is usually in the lower half and rarely at the top, but it may be anywhere. Locally, it replaces almost the whole skarn zone. Except in the area where metadiabase is abundant, the main ore layer tends to have the same general stratigraphic position within the skarn along a given cross section, but it varies (roughly serially) from section to section along the ore body. If we view these cross sections in sequence from northeast to southwest, we find that the ore moves from the middle to the lower half of the main skarn zone, continues roughly in the lower half for about 800 feet, rises to the middle of the skarn, and becomes highly erratic where metadiabase has intruded the skarn zone. Here prominent splits are commonly present in the ore. For example, a single ore layer near the bottom of the skarn zone at elevation 800 (second level) may lie in the middle of the zone from elevation 900 to elevation 950, split into two or more distinct layers at elevation 970 (below first level), and continue upward with one or two layers in the upper third of the skarn zone and one layer close to the base. Frequently the pattern is much more complex than this. Yet throughout the mine the continuity of any single layer of magnetite, even a very thin one, is remarkable. In addition to rising, falling, and splitting within the skarn zone, the ore thickens and thins along the strike, giving a complicated pattern indeed.

It is clear that the skarn ores, though essentially conformable to the outlines of the skarn zone, may show in detail a markedly transgressive relation within the zone to which they are confined. Locally, magnetite veinlets cut across the contact between skarn and other rocks. Near the floor of the Clifton pit, between open stopes 9 and 11, magnetite veinlets pass from the hanging-wall contact of the ore body upward into an irregular layer of contaminated granite that cuts the skarn. This minor feature is significant in showing that, on a very small scale, the magnetite locally passes outside its normal bounds.

Grade and size of ore bodies

The grade of a number of ore bodies has been determined by drilling, but no adequate information on reserves is available. Some idea of the expectable size of the deposits is indicated by the length of their magnetic anomalies and the thickness of ore in the exploratory drill holes. This information can be supplemented by our knowledge of the Clifton deposit, the only skarn magnetite body now being mined.

The inferred length of the most promising magnetite deposits in skarn is more than 800 feet to less than 4,000 feet--generally about 1,000 feet to 2,000 feet or less. The average thickness of ore ranges from 30 feet to 65 feet for ore assaying 30 to 44 percent magnetic Fe. Within a 40-foot zone of ore assaying 42 percent magnetic Fe, there may be several layers of 60-percent ore, some barren partings, and some

40-percent ore. Elsewhere in the same ore body, a similar thickness of ore may be essentially uniform in grade. At Outafit, within the 65-foot thickness of ore assaying 27.1-30.7 percent magnetic Fe, there are rare layers of high-grade ore 2 to 10 feet thick, some layers of medium-grade ore 10 to 15 feet thick, some low-grade ore, and a good deal of almost barren skarn (data from Reed and Cohen, 1947, and core logs of U. S. Geological Survey). The individual high-grade layers might not be worth separate extraction, but the thick, heterogeneous zone constitutes low-grade ore of some promise. Locally at Clifton, the ore zone consisted of more than 50 feet of 50-percent ore. The highest-grade parts of such layers were initially shipped as lump ore.

Little is known concerning the depth of the ore deposits, for none of them has been adequately explored. Again, if we look to Clifton for a clue, we find that the ore there has already been wholly or partly stoped through a vertical distance of more than 400 feet, equivalent to a "mining height" or "dip-length" of more than 600 feet. The total production of concentrates and subordinate lump ore from that mine from its opening late in 1942 through 1949 was about a million tons (estimate by the writer from figures published by Linney and others).

Besides the most promising deposits considered above, the district contains a number of presumably smaller magnetite bodies inferred to be 500 to 1,000 feet in length. A few of these have been partly explored by drilling, which shows 5 to 15 feet of ore generally assaying 40 to 44 percent magnetic Fe. Finally, the district has a number of apparently worthless prospects and magnetic anomalies less than 500 feet in length. The figures suffice to show the considerable range in size and grade for skarn ores in the district.

The following table, based on rather scanty data, _/ gives some idea of the character of the crude skarn ores.

Assays (weight percent) for several representative skarn ores

Total Fe	Magnetic Fe	P	S	TiO_2
29.4-59.7 (generally 37-46)	27.1-44 (generally 30-44)	0.01-0.15	0.11-3.37 (generally 0.1-0.6)	0.27-0.31

Chemical character of the ores

The skarn ores of the district are characterized by high Fe, low Mn, moderate SiO_2, low P and Ti, rather low but variable S, moderate Ca and Mg, low K, Na, and Al (Al is moderate in sköl ores), and locally high F. All the recoverable Fe is in the form of magnetite. _/ A little soluble, non-magnetic Fe is yielded by pyroxenes (Buddington, personal communication); this Fe is not recoverable. Iron contributed by pyrrhotite is generally negligible; it is concentrated magnetically, but the accompanying sulfur burns out as the concentrate is sintered. Mn and Ti are present in the dark silicates ferrosalite, ferrohastingsite, and biotite, and perhaps locally in the magnetite lattice itself. Rarely, Ti is present as exsolution blades in magnetite.

_/ Principally from Millar, 1947, and Reed and Cohen, 1947.

_/ Hematite, found sporadically in several skarn deposits, is a secondary mineral developed by supergene (?) alteration. See section on "alteration".

Several statements concerning the minor constituents, made rather baldly, are supported by the evidence now at hand. However, to the writer's knowledge, no thorough chemical, mineralogic, and metallographic study has been made on suites of ores, concentrates, and gangue. Until such a study is carried out, certain statements regarding Ti and Mn should be accepted with reservation.

Granite gneiss ores

General characteristics

The granite gneiss ores are replacement bodies of iron oxides in microcline granite gneiss and associated metasedimentary gneisses. Both magnetite and hematite are present, accompanied by sporadic and generally negligible amounts of sulfides. Until recently, magnetite was the only recoverable ore mineral, but with the successful completion of experimental work at Benson Mines (Webb and Fleck, 1950) hematite has also become an ore mineral. All the hematite is crystalline, rather than earthy, and much of it is primary (in the sense that the accompanying magnetite is primary). However, martite pseudomorphous after magnetite also occurs.

Granite gneiss ores constitute deposits of moderate, large, or very large size, uniformly low grade, and remarkable continuity. Though in detail they show many minor variations, they have a broad general uniformity and consistency that contrast strongly with features of the skarn ores. In particular, the mineralization is diffuse but persistent. The mineralized zones are generally conformable to the contacts between lithologic or "stratigraphic" units, but on scales that are extremely large or extremely small they transect these contacts. Often the mineralized zones are not sharply limited; material of ore grade passes outward through sub-ore into weakly mineralized rock. In places, one wall is well defined and the other is an "assay wall". (At Benson Mines, for example, the hanging wall is sharp and the footwall is gradational. See Tillinghast, 1948, p. 28.)

Introduction of the iron oxides has been accompanied by bleaching of the host rock, changes in phase or composition of the minerals of the host, and introduction or concentration of minerals containing Mn, Ba, P, F, and other elements. In places, late hydrothermal alteration has affected the ores and wall rocks, leaching them and depositing pyrite, chlorite, apatite, epidote, pumpellyite, and zeolites.

The mineralogy and paragenesis of the metallics and late hydrothermal minerals are treated separately. The section immediately following deals with general features of the granite gneiss ores.

Description of typical ore

Typical granite gneiss ore is a foliated rock in which thin layers rich in iron oxides alternate with thin layers rich in light-colored silicates. The layering is rather crude and irregular but generally recognizable. Locally it is sharply defined. The layers rich in iron oxides range from one-eighth of an inch to an inch or more in thickness. Often they are one-quarter to one-half inch thick, separated by slightly thicker or thinner layers rich in silicates. Within the recognizable layered structure there may be a less obvious netlike distribution of magnetite grains and small amoeboid patches, blurring and softening the outline of the layers. Some dark-gray, highly garnetiferous varieties appear homogeneous except for slight variations in the concentration of garnet. In general, both regularly laminated and regularly disseminated types of ore are rare. Locally the metallics are distributed as clots or gobs, trailed by finer grains, in a very crudely layered pattern. In some places, the metallics form distinct cross-cutting veinlets or irregular networks in the host.

Where magnetite predominates, the metallic is black. Where hematite predominates, the metallic has a dark bluish cast. Sporadic grains, small clusters, or veinlets of sulfides may accompany the iron oxides. Pyrite and pyrrhotite are commonly represented, chalcopyrite is sparse, molybdenite is infrequent, and bornite is extremely rare.

In addition to the iron oxides and incidental sulfides, one usually recognizes quartz, feldspar, and one or more of the following: black biotite, dark-red garnet, colorless, white, or yellowish sillimanite, and (rather seldom) dark-green pyroxene. Often the dark silicates are clustered with the metallics and only sparingly present in the felsic layers. The felsics themselves are commonly green, gray, yellowish, or white, but locally distinctly pinkish. In general, bleaching of the feldspar is more pronounced in low-grade ore and less pronounced or lacking in sub-ore or weakly mineralized rock. The bleached feldspar may be glassy in appearance and almost indistinguishable from quartz. Moreover, the proportions of quartz and feldspar are highly variable.

Most of the ore is fine- or medium-grained. In places, magnetite replaces coarse granite or syenite pegmatite, but commonly (?) there is less magnetite in thin pegmatite seams than in the body of the granite gneiss host.

While one may be able to find a hand specimen of typical ore, the grade of the material is so variable in detail that one usually needs to see a very large slab, a working face, or a series of drill cores to know what typical ore looks like. High-grade ore more than a few inches thick is virtually unknown from the granite gneiss ore deposits, and medium-grade ore is sparse or rare. The material now mineable has about 20 to 32 percent Fe and averages close to 25 percent or slightly less. As previously noted, the ore is transitional through sub-ore (perhaps 12 to 20 percent Fe) into weakly mineralized rock. Further details are given under "grade and size of ore bodies".

Magnetite is the dominant iron oxide, almost invariably accompanied by a little hematite. Hematite concentrations are so far known only at Benson Mines and the Parish deposit (a modified granite gneiss ore described below).

Preferred host of ore

In general, granite gneiss ores are confined to the biotitic and sillimanitic facies of microcline granite gneiss, though locally the pyroxenic facies serves as the host for part of the ore in a given deposit. As a rule, hematite ore is closely associated with the sillimanitic facies, though that rock may contain magnetite as well. Ore is not known to occur in hornblende-microcline granite gneiss. A description of the several facies of microcline granite gneiss was given in the section on rocks of the district. The highly variable character of the gneiss, and the presence of metasedimentary relics, was pointed out. Buddington (1948) developed the hypothesis that the microcline granite gneisses are in large part the result of granitization of metasediments and amphibolite.

Subordinate layers of metasedimentary gneisses within, adjacent to, or near the mineralized sheets of heterogeneous microcline granite gneiss may also be mineralized with iron oxides, giving a wide variety of possible host rocks. The metasedimentary gneisses are principally biotitic and pyroxenic gneisses somewhat modified by introduced potash feldspar. However, independent ore bodies exclusively in such gneisses have not been found in the district: to be significantly mineralized, the gneisses must be part of a heterogeneous microcline granite gneiss sheet.

Locally, pyroxene skarn or migmatitic pyroxene skarn adjacent to or near a mineralized granite gneiss sheet may be replaced by a minor or significant quantity of magnetite. Such skarn is likely to be a type slightly modified by quartz, potash feldspar, or both these minerals. However, a lack of magnetite mineralization in skarn is the general rule in the immediate neighborhood of granite gneiss ores. For example, at the Jarvis Bridge deposit layers of pyroxene skarn in the hanging and footwalls within 100 feet of the ore-bearing granite gneiss sheet are unmineralized except at Hole 8, where migmatitic skarn at the footwall of the ore zone carries some magnetite. At the Spruce Mountain Northwest ore body, a 33-foot layer of migmatitic pyroxene skarn 8 feet below ore in Hole 9 is completely barren. In Hole 5 at the same deposit, 30 feet of salitic pyroxene skarn 42 feet above the upper ore zone carries sporadic pyrite and chalcopyrite but not magnetite. The avoidance of skarn by magnetite at granite gneiss ore deposits is both startling and provocative.

Perhaps the following summary will give a clearer picture of the occurrence of granite gneiss ores. The ore deposits occur in sheets of heterogeneous microcline granite gneiss enclosed in a series of metasediments. They show a marked preference for the biotitic facies or the sillimanitic facies of the granite gneiss, both of which may occur together, with or without garnet. (Crystalline hematite often shows a strong preference for the sillimanitic facies.) Locally, a pyroxenic facies may carry part of the ore. Incidental layers and relics of metasedimentary rocks may also serve as the host of ore, but these are always subordinate. The mineralization has tended to avoid skarn or its derivatives.

The granite gneiss sheets that carry ore range in thickness from 38 feet to more than 575 feet. Commonly the thickness is about 100 to 300 feet or less. For a given deposit, the thickness may be roughly constant, but it is usually variable. The ore-bearing part of the sheets constitutes 4 to 30 percent or more of the total thickness. One or more distinct ore zones may be recognizable. In addition to ore averaging about 25 percent Fe, there is always some weakly mineralized rock and sub-ore.

The metasedimentary rocks enclosing the granite gneiss sheets include many varieties. However, at several deposits (Jarvis Bridge, Spruce Mountain, Parish, and perhaps others) the immediate wall rocks and much of the remaining metasedimentary sequence are dominantly pyroxenic. (Thin zones of biotite gneiss may or may not separate the pyroxenic metasediments from microcline granite gneiss.) Yet the granite gneiss sheets themselves contain relict layers mostly of biotite gneiss, together with varietal minerals derived therefrom. It seems clear that the granite gneiss sheets have been emplaced in biotitic zones enclosed by pyroxenic rocks. Perhaps this localization of the granite gneiss sheets was due to the greater fissility of the biotite gneiss, as well as to its more favorable chemical composition. An alternative hypothesis--that the biotite within granite gneiss sheets has resulted ffom the reconstitution of pyroxene--is entirely unsupported by field or petrographic evidence.

Scattered thin sheets of microcline granite gneiss are present in the wall rocks that enclose the main sheets. The subsidiary layers may or may not be appreciably mineralized. They have no economic significance.

Locally (at Skate Creek, for example), the main microcline granite gneiss sheets come within a few feet of hornblende-microperthite granite. Elsewhere the sheets are as much as 0.3 mile from the hornblende granite or its equivalent gneiss. Moreover, all the ore deposits in microcline granite gneiss are within 0.4 mile of alaskite or its equivalent gneiss (generally within 0.2 mile).

The position of ore within the granite gneiss sheet varies from one deposit to another and usually varies in any single deposit. The ore bodies appear to be continuous, not lenticular. In at least one deposit (Jarvis Bridge), the variable position of the ore gives rise to an undulating ore body within the granite gneiss sheet. In addition to, or along with, this internal variation, the mineralized zone may transect the borders of the granite gneiss sheet. (See below.) Possibly, undulation and transection of rock contacts are also characteristic features of the granite gneiss sheets themselves.

Discordant relations

Ore bodies in granite gneiss are in the main concordant, but they also exhibit local discordance. For purposes of prospecting, exploring, and developing the ore bodies, it is well to emphasize the dominant feature of concordance. Yet from the genetic viewpoint, the local discordance may be highly significant.

Discordant relations are apparent in very large or very small samples of the ore bodies. On samples of intermediate size, concordance is the more impressive feature. In some polished sections and small hand specimens, the ore minerals have a crude planar structure that makes an angle of 5 to 15° with the sharply-defined foliation of the host rock. Elsewhere, the ore minerals have a net-like or meshwork pattern partly controlled by foliation in the host and partly independent of that structure. Few drill cores or polished sections show a rigorous parallelism of the metallics and non-metallics, though the parallelism "improves" as the concentration of ore minerals increases.

Some outcrops show veinlets of magnetite, an inch wide and several inches long, splitting off from rather regular magnetite seams at an angle of 45° or so. These veinlets transect the planar structure of the host rock before they peter out or join up with other seams parallel to the dominant foliation.

Undulation of ore bodies within the enclosing granite gneiss has already been mentioned. In addition, real or apparent transection of the contacts of the granite gneiss sheets is a feature of several magnetite bodies, including Benson Mines, Deerlick Rapids, Jarvis Bridge, Skate Creek, and Spruce Mountain.

At Skate Creek, the ore-bearing part of the mineralized zone lies along the contact between biotite-microcline granite gneiss (locally sillimanitic and garnetiferous) and underlying pyroxenic rocks (including pyroxene-microcline granite gneiss, pyroxene gneiss, and skarn). If, as it appears, this contact between two recognizable and very different rock types is essentially a continuous surface, then the ore-bearing part of the zone warps gently back and forth across the contact. At both ends of the 7,000-foot belt, most of the ore is in pyroxenic rocks, with subordinate mineralization in the overlying granite gneiss; but in the middle of the belt, most of the ore is above or straddling the contact. The maximum "stratigraphic distance" transected by the ore zone is about 25 feet. However, the horizontal distance over which the transgression takes place is 1,300 to 2,500 feet. Thus the angle of transection is extremely small--about one-half to one degree.

At the Spruce Mountain Northwest ore body, Hole 9 cut ore in pyroxene gneiss 75 feet stratigraphically above the usual base of the granite gneiss unit that carries the ore. Unless the pyroxene gneiss represents an isolated lens within the granite gneiss, the top of the lower mineralized zone has cut across 75 feet of host rock in a horizontal distance of 500 to 600 feet (to adjacent Holes 6 and 7, respectively), giving an angle of discordance of about 8°. Somewhat similar relations obtain at Deerlick Rapids (Reed and Cohen, 1947, p. 9) and Jarvis Bridge.

John McKee (1944, personal communication) told the writer that the ore zone on the east limb of the Benson Mines structure migrates some feet from one lithologic unit to another over a strike distance of several thousand feet.

All the above examples of large-scale discordance can also be explained by lensing, boudinage, certain types of minor folding, and other means. Clearly, the discordance is slight (less than 10°); discrete, clearly-defined "shear zones" are lacking; and the data on which discordance is based are meager. Nevertheless, the mineralized zones do shift position within the enclosing host rocks, and they do give the appearance of transecting the contacts between their customary host and customary wall rocks. Whether these features are indicative of obscure but through-going zones of deformation, only detailed information can make clear.

Mineralogy of host rock and gangue

In the initial stages of mineralization, the microcline granite gneisses show no significant change in mineralogy except for a slight increase in their quantity of iron oxides. As mineralization increases in intensity, marked changes take place in the concentration, phase, or composition of minerals previously present, and, locally, new minerals make their appearance. The several facies of microcline granite gneiss have been described earlier, and additional mineralogic information is found in Buddington (1948).

One of the easily-recognized outward signs of change is the bleaching of the host rock. Roughly, this bleaching increases in intensity as the host rock acquires more iron oxides, but there are many local exceptions. Moreover, bleaching of any one of the granitic rocks of the district may also take place where the rocks come against amphibolite, skarn, or limestone. (See for example Prucha, 1949; Cushing, 1910; and Laitakari, 1920.) However, the change of pink potash feldspar to a gray, green yellowish, or colorless variety seems characteristically related to mineralization. It is generally effected by the conversion of microcline to an untwinned potash feldspar (see below), though locally the microcline itself merely changes color. The bleaching may take place without any significant change in the mineralogic composition of the host rock, but changes in mineral content are likely to accompany the bleaching.

Changes in concentration of minerals

The principal silicates in the host are quartz and potash feldspar. Their ratio is highly variable in ore-bearing rocks--much more variable than in the unmineralized equivalent. Generally, both minerals are present, but either may be found almost to the complete exclusion of the other. Plagioclase is rare or absent. When present, it is usually highly sericitized or altered to chlorite and clay minerals, even when the accompanying potash feldspar is entirely fresh. Possibly some of the altered "plagioclase" is cordierite (cf. Dale, 1935, p. 87), though this mineral has not been identified by the writer. Scapolite, generally altered, is a rare constituent of some ore.

Accompanying the felsic minerals are one or more of the following varietal minerals: garnet, biotite, sillimanite, pyroxene, or amphibole. Garnet, biotite, and sillimanite occur singly or together. Pyroxene (rather infrequent) may be accompanied by biotite or a little blue amphibole. Very small amounts of blue amphibole may be present locally in biotitic rocks that lack garnet and sillimanite; however, granite gneiss ores containing amphibole as the sole variotal mineral are unknown to the writer. Accessory minerals are apatite, allanite, and zircon, all of which are likely to be radioactive, as indicated by their accompanying haloes in biotite. Sphene is an extremely rare accessory. All the varietal and accessory minerals noted above are common to the unmineralized granite gneiss, as well as to ore. Their concentration may be essentially the same in both rocks, but it is likely to be different. This applies especially to the accessories. In the mineralized granite gneiss, apatite and zircon are slightly more abundant, and apatite is concentrated locally to the extent of one percent or more. Allanite, rare or entirely lacking in unmineralized granite gneiss, is a common accessory in the ores. The reverse is true of sphene.

Phase change in potash feldspar

A phase change in the potash feldspar is perhaps the most striking, widespread, and distinctive accompaniment of mineralization. In the initial stage of replacement of microcline granite gneiss by iron oxides, the microcline shows characteristic twinning (grid structure) as it does in the unmineralized rock. Where the quantity of iron oxides reaches about 10 to 12 percent, microcline disappears. Instead we find a clear, untwinned potash feldspar that very rarely shows a "shadowy" or undulatory extinction and never shows grid structure. This feldspar would at first glance be called orthoclase, if in cursory examination it were not mistaken for quartz. (Cleavage is often very difficult to see except in broken grains at the edge of a thin section, and cloudy alteration is extremely rare.) The feldspar characteristically has a negative optic axial angle of 35 to 55°--generally 40 to 48°. Preliminary U-stage study of the cleavages with respect to the optical indicatrix shows almost without doubt that the untwinned potash feldspar is triclinic, rather than monoclinic. The X-ray diffraction powder pattern is nearest that of anorthoclase, a triclinic K-Na feldspar (J. M. Axelrod, personal communication, 1950). However, the feldspar has the composition $Or_{90.9}Ab_{6.8}Cn_{2.3}$._/ The writer believes the feldspar

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_/ Unpublished chemical analysis by E. K. Oslund, 1950; BaO = 1.18 percent. The analyzed material came from the Parish deposit, a modified granite gneiss type. The identity of the analyzed material with feldspars in typical granite gneiss ore at Benson Mines was checked optically by the writer, spectrographically by K. J. Murata, and by X-ray by J. M. Axelrod. A single-crystal investigation of the analyzed sample is being undertaken by Axelrod. "Cn" of the formula equals celsian.

may be merely a microcline that has lost its twinning, but further work on this and other feldspars will be necessary before a definite statement can be made. Certainly the original potash feldspar (microcline) of the host has suffered a change of phase. In addition, it has acquired Ba. _/ Work on the relation of these two features is in progress.

The change in phase from microcline to untwinned potash feldspar is not so complete or so regular as was perhaps suggested in the foregoing account. The change does not always take place abruptly when the iron oxide content reaches 10 or 12 percent; in places, some well-twinned microcline may be retained until the iron oxides reach 15 percent, and some ore has a large proportion of shadowy feldspar grains with the completely untwinned ones. Furthermore, some unmineralized microcline granite gneiss has subordinate shadowy and untwinned potash feldspar, some weakly-mineralized gneiss (ca. 5 percent magnetite) has nothing but untwinned feldspar, and one even finds rare specimens of ore (ca. 20-25 percent magnetite) with well-twinned microcline. However, these instances are minor exceptions to a remarkably general rule. The transition from mineralized rock with microcline to ore with

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_/ The BaO content of representative St. Lawrence County granites, including two varieties of microcline granite gneiss, ranges from 0.02 to 0.10 percent. Comparison of the BaO content with modal potash feldspar in the rocks indicates that the potash feldspars in typical microcline granite gneiss have only ca. 0.1 to 0.15 percent BaO, assuming all BaO to be present exclusively in the feldspars.

untwinned potash feldspar is especially well shown in continuous series of drill cores from the Spruce Mountain and Parish deposits. Zimmer (1947b) has described a similar change of microcline to "orthoclase" in the ore at the Chateaugay magnetite mine, Lyon Mountain, Clinton County district, New York. From this "orthoclasization" he inferred a temperature increase during the period of ore deposition. There is so much uncertainty regarding the nature of the St. Lawrence County feldspar, and indeed the stability relations of potash feldspars as a group, that the writer does not believe the observed phase change is a reliable indicator of temperature increase in the St. Lawrence County ores. One might with equal force argue that the change from microcline to untwinned potash feldspar is analogous to the development of adularia in vein deposits, and thereby infer a decreasing temperature concomitant with mineralization.

It is interesting to note that the replacement feldspar in ore deposits in modified host rocks (modified granite gneiss at Parish, modified skarn at Trembley Mountain) is this same untwinned variety with a small or moderate optic angle. Most modified skarns, regardless of their location in the district, have untwinned potash feldspar rather than microcline. In addition, some amphibolites, pyroxene gneisses, and biotite gneisses have untwinned potash feldspar locally, though their potash feldspar is commonly microcline. Phase differences in the feldspar of these last three rocks appear to be entirely sporadic. (It is worth pointing out that normal pegmatites in the district have microcline, and so do most quartz-feldspar granulites.)

Change in composition of minerals

Some red garnet associated with granite gneiss ores seems to have been inherited from metasedimentary garnet-biotite gneiss. Much of the garnetiferous ore, however, contains more garnet than one normally finds in the metasedimentary gneiss. Also, some weakly-mineralized rock consists of thin laminae of the magnetite and cinnamon-brown garnet in an otherwise "clean" granite gneiss. This brown, spessartitic garnet appears to have been introduced with the magnetite. (See Parish deposit.) The relative importance of reconstitution versus replacement for garnet in much of the ore cannot be assessed, nor has it been possible to investigate the composition of a number of garnets from typical ore. However, one garnet from typical "hanging-wall ore" at Benson Mines has the composition $\mathrm{Alm_{43}Spess_{26}Gross_{23}Pyr_{8}}$. _/ An analyzed garnet from garnet-biotite-quartz-plagioclase gneiss of the Grenville Lowlands has the composition $\mathrm{Alm_{65.9}Pyr_{25.5}Spess_{3.5}Gross_{3.2}Andr_{1.9}}$. _/

Relative to almandite from the metasedimentary gneiss, the Benson garnet shows a notable increase in Mn. (Garnet-biotite-quartz-plagioclase gneiss from the massif is very similar in bulk chemical composition to

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_/ Unpublished chemical analysis by the late Norman Davidson, U. S. Geological Survey; MnO = 11.51 percent.

_/ Unpublished analysis by E. K. Oslund, 1950; MnO = 1.57 percent. Analysis from forthcoming report by A. E. J. Engel, U. S. Geological Survey.

corresponding rock in the Lowlands; so it seems likely that the analyzed almandite is representative of almandite in metasedimentary biotite gneiss layers near the magnetite deposits.) Vastly more Mn has been introduced directly as spessartite at the Parish deposit.

Biotites from the granite gneiss ores have not yet been analyzed. They appear to be rather uniform in composition, whereas biotites from metasedimentary biotite gneiss are highly variable (conclusion based on optical data, chiefly determination of n_β). If the biotite in granite gneiss ores has been largely inherited from metasedimentary biotite gneisses, as field evidence suggests, some process has brought the formerly variable biotite to a relatively uniform composition.

Introduction of new minerals

Locally the granite gneiss ores contain distinctive accessory minerals which are unknown in the unmineralized microcline granite gneiss or so exceedingly rare therein that they may be regarded as "new" minerals in the ores. These accessories are green spinel, goethite (?), fluorite, barite, monazite (?), and zoisite. They are not all found in every specimen of ore, but usually several of them are present as traces or as grains totalling a few tenths of one percent. The spinel is dark-green pleonaste or possibly hercynite. It usually occurs as irregular granules within or attached to the edge of magnetite grains and aggregates. Rarely, a free grain is present in the silicate groundmass. Goethite (?) forms small yellowish-brown prisms and reticulate aggregates of prisms in iron oxides or, less commonly, in biotite. In places the prisms are dark reddish brown, probably owing to admixed lepidocrocite and "limonite". Goethite (?), as well as spinel, often appears to be an exsolution product of the iron oxides, and there is no indication that the goethite (?) is supergene. Fluorite and barite occur as scattered granules or minute patches. Monazite (?) is a disseminated accessory accompanying zircon. Zoisite--rather rare--forms prisms in some magnetite grains.

In addition, the granite gneiss ores often show incipient or extensive alteration to late hydrothermal minerals, separately described.

Chemical character of the ore

Chemically, the granite gneiss ores are moderately high in Fe, and high in SiO_2, K, and Al. They contain a small but appreciable quantity of Mn, Ti, and Ba. They are generally low in S and P, though these elements are concentrated locally. The iron oxides that they carry have a higher oxidation state than iron oxides in skarn ores, for the granite gneiss ores invariably contain some hematite. Manganese is contributed by spessartitic almandite. Titanium is present as sparse exsolution blades of ilmenite in the iron oxides, and locally in solid solution in primary hematite. Barium is found in the potash feldspar and in sporadic grains of barite.

Grade and size of ore bodies

The known deposits in granite gneiss are moderate to very large in size. The Benson Mines deposit is one of the largest single magnetite deposits in the world. It is the only one of its kind now being worked in the district. Several promising deposits of similar type have been explored by a few shallow drill holes to an average depth of only 200 to 400 feet. The grade of these deposits is essentially the same as that of Benson, but the inferred size of the bodies is considerably smaller. The lack of deep drilling on these newly found deposits is a serious handicap to their evaluation. Their continuity and relative uniformity of grade, together with the persistence and depth of mined ore bodies of similar type, give them a potential value much greater than a few scattered drill holes can suggest.

The main ore body at Benson Mines _/ is about two miles long and 80 to 400 feet wide, according to Tillinghast (1948, p. 28. _/ The ore reserve within the open-pit area is about 65,000,000 tons of magnetite ore and 38,000,000 tons of hematite ore, equivalent respectively to 22,000,000 tons of magnetite concentrates and 12,000,000 tons of iron oxide concentrates. Reserves of ore for potential underground development have not been thoroughly explored; they may be very large (Tillinghast, 1948). If--or when--iron-bearing material with a con-

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_/ The Jones and Laughlin geologic staff has done very detailed magnetic and geologic work on the Benson Mines deposit, paying special attention to the correlation of magnetic and geologic information. For that reason, the Geological Survey has done only enough reconnaissance work on the deposit to get a "nodding acquaintance" with some of the geologic problems. It is hoped that the company memoir will soon be published.

_/ Webb and Fleck (1950) give the width as 400 to 1,400 feet.

centrating ratio of 4, 5, or 6 to 1 becomes of economic interest, the reserves within the Benson deposit alone would be increased many times (Webb and Fleck, 1950, p. 448).

Webb and Fleck (1950, p. 447) give the following typical analyses of crude magnetite ore and concentrate:

	Soluble Fe _/	Magnetic Fe	P	S	TiO_2	Mn	SiO_2
Crude ore	26.75	24.00	0.192	0.48	0.68	0.25	40.80
Concentrate	63.75	62.54	0.024	0.21	0.77	0.25	5.66

_/ All figures are in weight percent.

These are not necessarily "paired analyses" carried out on equivalent ore and concentrate. They show very clearly the presence of 2 to 3 percent of soluble but nonmagnetic Fe that is a common feature of all the granite gneiss ores. From our knowledge of other deposits of similar type, we may infer that part of the soluble nonmagnetic Fe is in primary crystalline hematite and part in martite.

Analyses of Benson hematite ore (with some associated magnetite?) are not available. Webb and Fleck (1950, p. 447) state that the concentrate to be obtained from such ore will be higher in P and TiO_2, and somewhat lower in Fe, than the magnetite concentrate now produced.

Several promising new deposits were investigated briefly by the U. S. Bureau of Mines during World War II (Millar, 1947; Reed and Cohen, 1947). These deposits contain magnetite bodies 3,500 to 7,000 feet long with average thicknesses of 20, 30, and perhaps 50 feet of ore averaging 24-25 percent magnetic Fe. Some of the bodies also have substantial thicknesses of sub-ore and mineralized rock. Magnetite ore from all the deposits carries several percent of soluble nonmagnetic Fe in the form of primary crystalline hematite and martite. No independent bodies of hematite ore were found with the magnetite.

It is worth noting that a magnetite deposit of similar type in granite gneiss at Lyon Mountain, Clinton County, New York, is several thousand feet long at the surface, has an average mineable thickness of about 15 feet of ore averaging 25 percent Fe, and is now being mined more than 2,000 feet below its outcrop (W. J. Linney, 1943).

Magnetite deposits in modified rocks

Magnetite deposits in modified skarn and in modified granite gneiss are intimately associated with areas of microcline granite gneiss. Indeed, the process by which the fundamental types of host rock were modified represents an extension (or possibly a renewal) of the same process that yielded the microcline granite gneiss itself. This process has been likened by Buddington (1948) to the development of a "pegmatitic" residual magma that intruded and substantially replaced the country rock. Modification of skarn and microcline granite gneiss took place before the introduction of magnetite and involved additions of quartz, potash feldspar, fluorite, barite, and local apatite. Scapolite developed locally in modified skarn. Its absence from modified granite gneiss may be explained in part by the relative scarcity of carbonate rocks at the site of intense modification. Large amounts of Mn in the form of spessartite were introduced into the modified granite gneiss. Equivalent Mn-rich minerals have not yet been identified in the modified skarn. Late hydrothermal alteration affected the modified host rocks and their mineral deposits more strongly than it affected either of the unmodified types.

Magnetite deposits in modified skarn

General characteristics

The Trembley Mountain deposits bridge the gap between ores in skarn and those in granite gneiss. Essentially, they are skarn deposits modified by the introduction of quartz, potash feldspar, and subordinate scapolite before the onset of magnetite mineralization. In addition to extensive K-Al-Si metasomatism, there has been local introduction of considerable Ba, F, and B in the form of barite, fluorite and apatite, as well as widespread introduction (?) of small quantities of rare earths in the form of abundant accessory allanite. Leaching and development of late hydrothermal minerals are a conspicuous feature of the area. The resulting complex deposits show evidence for the intensification of several processes that were already apparent in the simpler skarn and granite gneiss types. Structurally, the deposits have the intricate pattern of some skarn deposits, but the grade and habit of their ore are very similar to those of deposits in granite gneiss.

Some geologic features of the Trembley Mountain area

The Trembley Mountain area, north-central rectangle, Cranberry Lake quadrangle, contains two belts of magnetic anomalies underlain by metasediments and magnetite deposits. A sheet of contaminated microcline syenite gneiss (and granite?), estimated to be 150-200 feet thick, separates the two belts; and rock of similar type appears to wrap around the north end of the area. The folded complex of metasediments and "granitic" sheets is effectively enclosed by hornblende-microperthite granite. Alaskite comes against the northwest part of the area.

Rocks of the northern belt consist of a 40 to 50-foot zone of silicated marble and pyroxene skarn underlain by biotite gneiss. Sheets of contaminated microcline granite gneiss and syenite are also present, along with subordinate amphibolite. Ore is restricted to a 15- to 20-foot layer to silicated marble occurring at the top of the zone of calcareous rocks, just beneath biotite-microcline granite gneiss and granitized biotite gneiss.

Rocks of the southern belt are predominantly highly modified pyroxene skarn, most of it mineralized. The thicker zones of better-grade ore are usually close to the hanging wall of the skarn unit. "Quartz-rock" is ubiquitous and locally abundant. Layers of laminated pyroxene gneiss and very thin biotitic zones are present locally. The maximum thickness of the zone of modified skarn is estimated to be about 300 feet, though in places the zone thins to about 120 feet.

Structurally, the area is an isoclinal syncline--possibly compound--plunging southeastward. Many minor isoclinal folds are present, causing rapid variation in the thickness of the modified skarn zone and the ore it contains.

The modified skarn, described in a separate section, has been interpreted as a pyroxene skarn partly replaced by felsic minerals. The "quartz rock" may have had a similar metasomatic origin. "Quartz rock" is a glassy, gray to colorless, generally medium-grained, locally foliated rock consisting of quartz (80-90 percent) with accessory dark minerals, metallics, and sporadic untwinned potash feldspar (also glassy in appearance). In places, quartz rock is the host of ore, but more often it is present as thin, weakly-mineralized or barren layers in modified skarn. The rock cannot be described in detail in this work, but the main features of its occurrence will be pointed out. Quartz rock is limited to the zone of modified skarn where it apparently occurs as discontinuous layers of variable thickness (generally a few inches to a few feet; maximum of 27 feet). Drill cores show that quartz rock grades into modified skarn, often through a quartz-feldspar or feldspar-rich zone. Elsewhere, nearly "pure" skarn adjacent to quartz rock has thin layers or fine, wispy replacement areas of glassy quartz. At and near the contact, the quartz rock may show xenocrysts of pyroxene derived from skarn. Gradation, rather than sharp transition, is indicated by the study of thin sections: modified skarn adjacent to quartz rock may contain 40-50 percent quartz. The quartz rock itself always contains disseminated magnetite or pyrite (less often, pyrrhotite). Where it is enclosed by modified skarn, it is pyroxenic, carrying ferrosalite and local hypersthene; where it is intimately involved with remnants of garnet-biotite gneiss, it contains garnet and biotite. The microfabric is highly inequigranular and generally shows a strong foliation. In part, the rock appears to have been deformed and recrystallized; in part, its component minerals appear to have crystallized late, replacing or recrystallizing pyroxene from skarn.

Certain thin layers of quartz rock differ from the rest in having a fine-granulose texture and a regular, fine-scale lamination. These are interpreted as meta-quartzites. Locally the coarser type of quartz rock accompanies them, and the micro-fabric of the two types is similar. Why not, then, interpret all the quartz rock as metamorphosed quartzite?

First, because of the scientifically indefensible but geologically reasonable criterion that they do not look like metaquartzites--at least not like any other quartzites found in the St. Lawrence County district. Second, because the writer has been unable, with one exception, to correlate zones of quartz rock from drill hole to drill hole. Third, because the quartz rock is confined to the zone of modified skarn; it is entirely absent from the enclosing rocks, which include some metasedimentary gneisses. Fourth, because the contact relations between skarn and quartz rock often resemble those between skarn and granite pegmatite or demonstrable vein quartz. Fifth, because the quartz rock appears, in places, to grade into skarn by replacement. Sixth, because the quartz rock is comparable in general occurrence to introduced quartz present in the Parish magnetite deposit (q.v.).

From this interpretation of the available evidence, the writer concludes that much of the Trembley Mountain quartz rock has been introduced as a rather irregular replacement of skarn. The physical differences between skarn and thin layers of sedimentary quartzite or garnet-biotite gneiss may have localized some of the replacement. The quartz rock lacks the sharply-walled habit of quartz veins, and it seems doubtful that emplacement of the quartz was localized by fractures or accompanied by dilation of the host rock. Much or all of the quartz may have been contributed by the desilication of the surrounding microcline syenite gneiss, which is inferred to have been formerly a microcline granite gneiss with normal quartz content.

Metasomatic quartz-rich rocks very similar in occurrence to those at Trembley Mountain have been described from the Orijärvi region, Finland, by Eskola (1914, p. 209, pp. 256-257); from the Falun district, Sweden, by Geijer (1917, pp. 303-308); and from Attu, southwest Finland, by Pehrman (1931). Somewhat analogous rocks are perhaps represented by one phase of "Pochuck granite" closely associated with magnetite bodies in southeastern New York (Colony, 1923, pp. 52-53).

Description and occurrence of ore

Typical low-grade ore consists of disseminated grains, amoeboid blotches, and irregular layers of magnetite in a silicate groundmass. In ore of intermediate grade, magnetite forms coalescing layers and veinlets that in part conform to the color-banding (foliation) of the modified skarn, but in part transect and replace the foliation. Locally, magnetite serves as a matrix for angular remnants of coarse feldspar crystals. In a few places, the magnetite itself shows a lineation inherited from the elongate amphibole of the host rock. In all its aspects, the ore exhibits features customarily attributed to metasomatic replacement.

Magnetite mineralization has affected, in some places, every rock type in the area, with the possible exception of hornblende-microperthite granite, aplite, certain pyroxene and hornblende gneisses, and diopside skarn. Accessory magnetite is present in these four rock types as well, but here it may be syngenetic. However, not all rocks within the zone of modified skarn have been equally susceptible to replacement by magnetite. The preference for skarn and modified skarn--especially the latter--is striking. These two rock types contain almost 70 percent of the "mineralization", and almost 80 percent of the ore. Modified skarn alone contains more than 60 percent of all the "mineralization" and has about 70 percent of the ore. Quartz rock contains relatively little ore but has a high proportion of sub-ore. A little ore is found in quartz-feldspar rock (aggregate of quartz and untwinned potash feldspar) and contaminated granite gneiss and syenite gneiss (both of which now have untwinned potash feldspar instead of microcline).

When the modified skarn is so conspicuously preferred as a host rock, it is surprising to find that within this host the dark minerals--pyroxene and amphibole--are preferentially replaced in advance of the attack on felsic minerals. Accessory apatite and allanite in modified skarn are concentrated locally to the extent of one percent apiece.
In the northern deposit, the host of ore is a pyroxene skarn that has been extensively replaced by barite and, locally, by fluorite. In some places, more than half the gangue is barite. The rock shows slight local modification by quartz and feldspar, which antedate the barite and fluorite but precede late hydrothermal alteration. Ore from one small "pocket" of magnetite occurs in a biotite-quartz schist that was extensively replaced by barite before the introduction of magnetite and sulfides. No manganese minerals have been identified in any of the ore.

The overall grade of ore is low, averaging close to 25 percent magnetic Fe. Sulfides are sporadic. Several percent of chalcopyrite is present in the northern deposit; elsewhere, the mineral is insignificant in quantity. Much of the magnetite shows exsolution blades of ilmenite, but the average TiO_2 content of the ore is comparable to that of typical granite gneiss ores. The writer has not found any significant quantity of hematite in the ore. It seems likely that the 2 or 3 percent of soluble nonmagnetic Fe shown in the assays has come from pyroxene and is not recoverable. (Buddington, personal communication, 1951, has found that pyroxene in hematite-free pyroxene-microcline granite gneiss from the district is selectively attacked by HCl to yield an assay for soluble nonmagnetic Fe.)

The ore cut in 9 drill holes ranges in thickness from 7 feet to 135 feet. A substantial part of the zone is 20 feet thick, but at the southern deposit an area of unknown size is underlain by a very thick body of ore. The thicker zones of better-grade ore in the area are usually close to the hanging wall, though in places they are grouped in the middle of the modified skarn zone, found near the base, or (at the south end) distributed more or less throughout.

Magnetite deposit in modified granite gneiss

General characteristics

The Parish deposit closely approaches a hypothermal vein in many of its features, yet it is clearly transitional into the normal granite gneiss type of deposit. Magnetite, locally with much crystalline hematite, has replaced a sheet of microcline granite gneiss that has been extensively modified by the introduction of quartz, untwinned potash feldspar, spessartite, fluorite, and barite. The ore body is more compact and more sharply defined than other deposits in granite gneiss. The coarse and irregular texture of the introduced material, its irregular distribution, the mineral assemblage present as gangue, the nature of the accompanying sparse sulfides, the associated local alteration of wall rocks to chlorite, sericite, and clay minerals, and the conspicuous local leaching and zeolitization: all these are strongly reminiscent of the mineralization that yields hypothermal replacement veins. At the Parish deposit, this process included a thorough K-Al-Si metasomatism followed by the introduction of large quantities of Fe, Mn, and F, smaller quantities of Ba, S, OH, and CO_3, and minor quantities of Cu and traces of Zn. The pattern is the same as that found in all the granite gneiss deposits, but here the effects of the mineralization process are much more intense. Except for the difference in ore minerals concentrated, the Parish magnetite-hematite deposit has a remarkable similarity to the sphalerite-galena deposits of Broken Hill, New South Wales (Andrews, et al., 1922; Gustafson et al., 1950).

General geology of the Parish deposit

The Parish ore body replaces part of a sheet of microcline granite and granite gneiss found in a narrow belt of biotite gneiss flanked by pyroxene gneisses. The rocks of the belt dip almost vertically. They show many tight crumples and minor folds, and there is more than a chance that the ore body occupies the axial zone of a vertical isoclinal fold, either anticlinal or synclinal. The flanking pyroxene gneisses, locally quartzose, carry microcline as well as plagioclase. Some of them carry primary, polygonal calcite grains and a little grossularite. Thin layers of pyroxene skarn appear sporadically in the pyroxene gneiss. The biotite gneiss ranges from the normal biotite-quartz-plagioclase variety to one rich in microcline. Garnet and sillimanite are present locally. The biotite gneiss belt, whose inferred thickness is 40 to more than 200 feet, is preserved as a fairly continuous sheath between the flanking pyroxene gneisses and the microcline granite gneiss sheet that carries the ore. Remnants of granitized biotite gneiss are also found within the microcline granite gneiss sheet.

The central microcline granite gneiss sheet is generally 100 to 140 feet thick, thinning to 40 feet or less at the north and south ends of the 3,000-foot prospected belt. The microcline granite gneiss is highly variable in texture and composition. Much of the rock is a wispy or schlieren-banded, pinkish or grayish biotitic gneiss in which sillimanite is distributed sporadically, locally occurring in abundance. Part of the rock is aplitic; part of it, also poor in mafics, is medium-grained and extraordinarily clean-looking. Granite pegmatite veins, a few feet thick, are rather common; some of them contain schorl. While microcline is the usual feldspar in these pegmatites, coarse pinkish or yellowish plagioclase is present locally. The pegmatite veins generally appear to be conformable, but in places they are crosscutting. The granite gneiss sheet as a whole appears to be slightly transgressive within the biotite gneiss zone.

Irregularly scattered through the granite gneiss sheet are sericitized and kaolinized areas. Concentrations of biotite in such areas are commonly altered to a brilliant green chlorite. Accompanying or independent of the altered areas are zones in which granitic rocks and ore contain small vugs, a few mm in diameter, coated with epidote, zeolite, and other minerals. These leached zones are confined exclusively to the granite gneiss sheet; they are not found in the metasedimentary wall rocks. However, the latter may show local microscopic alteration to epidote, pumpellyite, chlorite, and carbonate, without the development of vugs or aluminous alteration products.

Description and occurrence of ore

Ore is restricted to the heterogeneous microcline granite gneiss sheet, which it replaces rather irregularly. Weakly-mineralized host rock is a streaky biotite-microcline granite gneiss, aplite, or medium-grained microcline granite. In places, these rocks carry sub-ore and even some low-grade ore. Granite pegmatite and highly biotitic granite gneiss are seldom mineralized. Most material of ore grade replaces medium- to coarse aggregates of reddish-brown or yellowish-brown spessartite; quartz; untwinned potash feldspar _/ -- colorless and

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_/ Typical untwinned potash feldspar from the Parish deposit has the composition $Or_{90.9}Ab_{6.8}Cn_{2.3}$, calculated from unpublished chemical analysis by E. K. Oslund; BaO = 1.18 percent. See discussion of feldspar problem in granite gneiss ores.

glassy; pink, green, or purple fluorite; and white barite. All these gangue minerals are highly variable in proportion. Small quantities both of iron oxides and gangue minerals have been found replacing recognizable granite gneiss whose microcline shows progressive transformation to untwinned potash feldspar. However, where either ore minerals or spessartite and fluorite are concentrated, the "host" is quartz, untwinned potash feldspar, or a mixture of the two; traces of recognizable granite gneiss are lacking. The inference is that the microcline granite gneiss has been completely replaced by quartz and feldspar before introduction of the other gangue minerals and metallics. Moreover, where iron oxides or spessartite appear in force, they are intimately associated with each other and definitely younger than the quartz-feldspar host. These features are borne out by the paragenetic sequence determined microscopically.

Two types of ore are found in the deposit: high-manganese ore (also as a rule high in iron) and low-manganese ore (generally a medium- to low-grade iron ore). The Mn in the ore is directly attributable to spessartite. Every portion of drill core giving a high Mn assay is rich in spessartite, and all cores assaying low in Mn have little or no spessartite. The composition of this Mn-rich garnet is $\text{Spess}_{61}\text{Alm}_{16}\text{Andr}_{14}\text{Gross}_{5}\text{Pyr}_{4}$. _/ (It is interesting to note that the only known spessartite[5] closely approximating the Parish garnet in

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_/ Unpublished chemical analysis by the late J. G. Fairchild, U. S. Geological Survey; MnO = 25.66 percent.

composition come from Broken Hill, New South Wales.) To date, no other Mn mineral has been positively identified from this deposit. Some spessartite-rich ore lying beside an old test pit has sooty coatings of what may be secondary Mn oxides, but similar oxides are entirely absent from all the Mn-rich drill core examined by the writer. The medium- and high-grade ore with abundant spessartite has an irregular distribution within the ore body.

Walker (1943, p. 519) gives the following analyses of ore from the Parish deposit:

	Mill ore (percent)	Manganiferous ore (percent)
Fe	39.70	55.80
P	0.065	0.026
Mn	4.70	9.92
SiO_2	23.30	7.80
S	0.030	0.261
TiO_2	0.24	0.20

The ore body has an undulatory outline within the microcline granite gneiss sheet, weaving from side to side. In addition, the ore body pinches and swells slightly. Generally, about 12 to 35 percent of the granite gneiss sheet constitutes ore. Toward the north end of the deposit, one segment of the ore body frays out into several thinner layers before resuming its usual average thickness. There is a suggestion that local concentrations of iron oxides and garnet occur as shoots within the mineralized zone, but additional data will be needed to confirm this suggested distribution.

Metallic minerals and paragenetic sequence

Some primary crystalline hematite and sporadic martite are present with the magnetite even in weakly-mineralized rock. Portions of the low-, medium-, and high-grade ore have abundant hematite. Exsolution blades of ilmenite are fine, minute, and rare in both magnetite and hematite. Internal alteration of magnetite to martite is generally slight, and many magnetite grains contain no martite. Magnetite, with or without martite, is replaced locally by primary crystalline hematite, though the hematite usually replaces gangue in preference to magnetite. There is a slight suggestion that the martitization of magnetite has, as a counterpart, the local internal replacement of primary crystalline hematite by traces of secondary crystalline hematite.

In general, sulfides are sparse in the Parish deposit, and much of the ore has none whatever. Where a sulfide is megascopically detectable, it is usually chalcopyrite, commonly occurring to the extent of 1 or 2 percent but locally forming 5 to 10 percent of layers a few feet thick. Pyrite and pyrrhotite are relatively rare minerals in the deposit. In polished sections, bornite is as common as chalcopyrite, and a few bornite grains were noted megascopically. Inasmuch as tarnished bornite looks very much like tarnished hematite, some bornite was undoubtedly overlooked in logging the drill core. Both chalcopyrite and bornite show slight local replacement by gray or blue chalcocite. The gray chalcocite (and bornite?) may be slightly replaced by covellite. Traces of sphalerite are present in about half the ore specimens. Traces of a soft white unknown ore are occasionally found with bornite and chalcopyrite. Several discrete grains of valleriite (?) are present in one specimen.

The paragenetic sequence of the non-metallics is: _/

_/ Symbols and abbreviations stand for:

→ -- replaced by or followed by

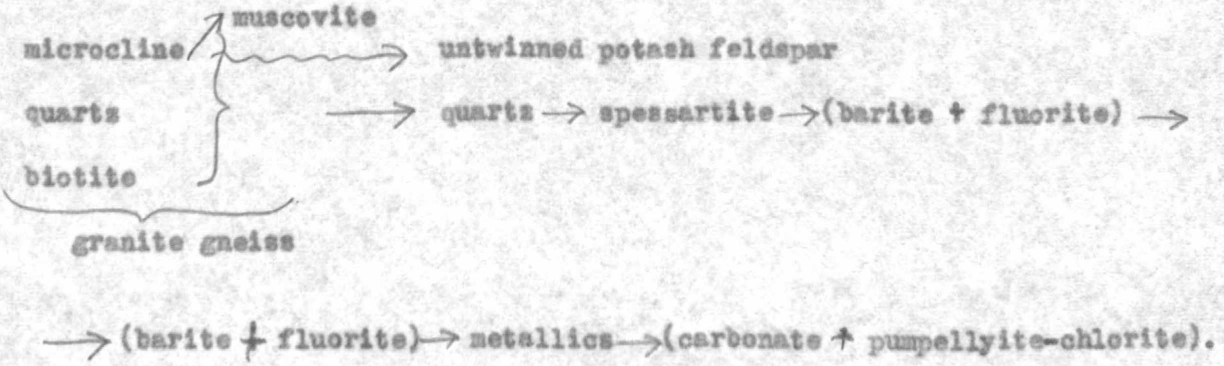
⇝ -- transitional to and replaced (?) by

ia -- internal alteration

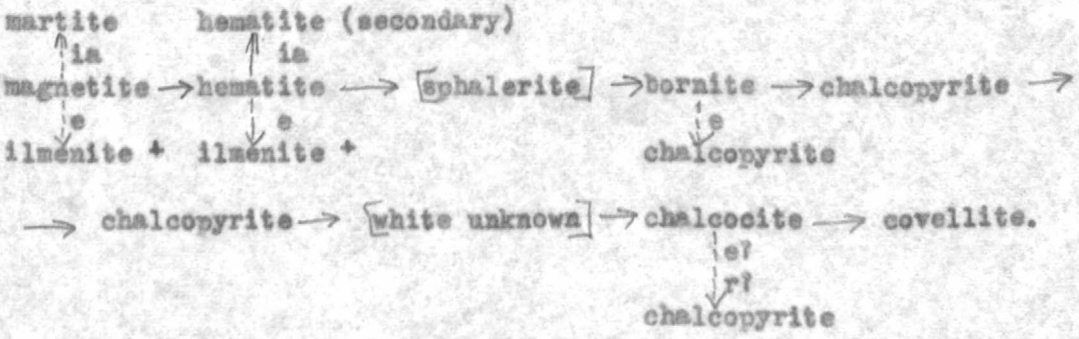
e -- exsolution of

+ -- plus unidentified minute bodies in magnetite and hematite

r -- relics



The paragenetic sequence of the metallic minerals is:



The position of pyrite and valleriite (?) is unknown. The position of sphalerite and the unknown white metallic is uncertain. Pyrrhotite has not been found in polished sections. Further details on the metallics are given in a separate section.

Mineralogy and paragenetic sequence of the metallics

General features and paragenetic sequence

A relatively small and simple suite of metallic minerals is found in the magnetite deposits of the district. Arranged roughly according to frequency and quantity, these metallics are magnetite, hematite, pyrite, pyrrhotite, chalcopyrite, sphalerite, molybdenite, and bornite--all recognizable megascopically; _/ and ilmenite, marcasite (?), chalcocite, covellite, vonsenite, loellingite, unidentified minerals, valleriite (?), and cuprite (?)--generally detectable only under the reflecting microscope. Primary crystalline hematite is restricted to the granite gneiss ores and their modified equivalent. Martite (hematite that has replaced magnetite) is present in traces or minor amounts in the magnetite from a number of deposits, regardless of type. Pyrite and pyrrhotite, with subordinate chalcopyrite, are ubiquitous. Sphalerite and molybdenite are sporadic and quantitatively insignificant. Ilmenite, as exsolution blades in iron oxides, is widespread but generally negligible in quantity. Bornite, chalcocite, and valleriite (?) have been found

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_/ The opaque but nonmetallic mineral graphite is present locally in diopside skarn and silicated marble.

only at the Parish deposit, where the last sulfide is extremely rare. The other metallics are found in small amounts at one or more deposits. Galena has not yet been found in the district. Minor values for Au and Ag have been reported from one magnetite deposit (Millar, 1947, p. 12).

A small part of the disseminated oxides in the metasedimentary gneisses and quartzites may represent reconstituted iron-bearing material present in the original sediments, and part of the iron oxides in granitic rocks and their pegmatites is undoubtedly syngenetic. An additional small part of the iron oxides present in both classes of rock appears to be the product of metamorphic reconstitution. However, the writer believes that the iron oxides present as concentrations represent material introduced from a magmatic source. The iron oxides were accompanied or closely followed by sulfides of Fe, Cu, Zn, and Mo. Locally, a second generation of Fe sulfides is represented in some deposits. The evidence now available strongly favors a hypogene origin for all the sulfides, though Fe sulfides of the inferred third generation might be supergene.

There seems to be no systematic distribution of sulfides according to type of magnetite deposit, though concentrations of pyrrhotite are usually associated with skarn deposits and concentrations of pyrite with granite gneiss deposits. Also, bornite, chalcocite, and valleriite (?) are restricted to the deposit (Parish) that most closely resembles a hypothermal vein. Zoning of sulfides within the district has not been detected. However, the magnetite district shows a relative concentration of Cu, and the Edwards-Balmat zinc district a relative concentration of Pb.

The paragenetic sequence of metallics for most deposits of the district is: _/

$$\begin{tikzcd}
\text{martite} & & & & & \\
\text{magnetite} \arrow[u, dashed, "\text{ia}"'] \arrow[d, dashed, "\text{e}"] \arrow[r, phantom, "="] & \text{hematite} \arrow[r] \arrow[d, dashed, "\text{e}"] & \text{pyrite} \arrow[r] & \text{pyrrhotite} \arrow[r] & \text{sphalerite} \arrow[r] \arrow[d, dashed, "\text{e}"] & \text{chalcopyrite —} \\
\text{ilmenite +} & \text{ilmenite +} & & & \text{chalcopyrite} &
\end{tikzcd}$$

$$\text{chalcopyrite} \rightarrow \text{covellite.}$$

Specifically, the sequence as written applies to the granite gneiss ores. Deleting hematite gives the sequence for skarn ores. The position of molybdenite is not known. Not all the metallics occur in every deposit of the appropriate type, but usually only one of them is missing. The position of several rare metallics is discussed following the description of those minerals. The paragenetic sequence for the Parish deposit has been given separately.

_/ Symbols and abbreviations stand for:

= -- contemporaneous with

-- replaced by or followed by

ia -- internal alteration

e -- exsolution of

+ -- plus unidentified minute bodies in magnetite and hematite

Some deposits show slight additional mineralization after the bulk of the sulfides. This second generation of metallics (conveniently referred to as minerals of stage II) occurs as local replacements of the host rock; as thin fissure veinlets of calcite accompanied by chlorite and epidote; and as small vugs lined with epidote, pumpellyite, chlorite, and zeolites. Pyrite (with marcasite?) is the common sulfide in such occurrences. Locally, sphalerite is also present. Rarely, a trace of chalcopyrite (and specularite?) has been found. Pyrite of this generation is known to fringe earlier pyrite and vein magnetite. It is also the youngest mineral in vugs that cut all the sulfides, magnetite, and silicates of low-grade ore. The minerals of stage II overlap somewhat, but their usual, generalized sequence is:

host-rock minerals→iron oxides→sulfides of stage I→(epidote, pumpellyite, chlorite, calcite, zeolites)→sulfides of stage II [order unknown].

Perhaps a third stage of mineralization is represented by the sporadic chlorite-calcite veinlets, locally carrying pyrite, that cut ore and country rocks. Some veinlets of similar habit contain quartz or fluorite. Signs of faulting, or of supergene alteration to chlorite, carbonate, and hematite, may or may not be found with the little veinlets. It is probable that some of them are supergene, though it is possible that they are all hypogene, belonging either to stage II or to a separate stage III. They are mineralogically simpler than aggregates of stage II minerals but have never been found cutting those minerals. Because of their simpler composition and close association in places with faults and alteration zones, it seems desirable to assign them provisionally to stage III after emphasizing the uncertainty of their position in the sequence.

A few general features of the habit of the metallics deserve mention. Throughout the whole of stage I and locally through stage II, the boundaries between grains served as primary sites of attack by metasomatizing solutions. Sulfides are likely to show more obvious replacement criteria than iron oxides, and coarse vein-like patterns for certain sulfides of stage I are more common than for oxides. The evidence of metasomatic replacement improves with increasing quantity of the metallics; small quantities of metallics may take the form of single grains or unfilamented aggregates. Sulfides of stage I may or may not replace gangue or other sulfides in preference to iron oxides (until sulfides become abundant). Sulfides of stage II may vein or rim both iron oxides and gangue, but they usually show a marked preference for replacement along cleavages and cracks in silicates.

Thus it appears that the habit of an individual metallic depends partly on the concentration (or effectiveness of precipitation) of its components in the ore-bearing solutions, and partly on the stage of mineralization during which the individual metallic was deposited. The habit assumed in a given stage largely depends, in turn, upon the deformational history of the host rock. The fabric of the host rocks subjected to iron oxide replacement is essentially granoblastic or crystalloblastic. Evidence of cataclasis is extremely rare; instead, the deformational fabric is a more subtle one evidenced by foliations, mineral elongation, and slight variations in grain size. The host appears to have been deformed and recrystallized, wholly or locally, before or at the time the iron oxides were deposited. The sulfides pyrite and pyrrhotite closely followed the deposition of iron oxides. Deposition of younger sulfides of the same stage I in part (or in places) continued without interruption, though in some places the host, iron oxides, and earlier sulfides were very slightly fractured before precipitation of chalcopyrite. Substantial local fracturing occurred before the deposition of sulfides of stage II; and in several granite gneiss deposits, leaching of the host rocks and development of vugs lined with epidote, zeolites, and other minerals, preceded or accompanied the sulfide deposition. Later, local crackling and fracturing of the now brittle rocks took place before the advent of stage III minerals.

Descriptive mineralogy of metallics

This section concerns chiefly the microscopy of the metallic minerals. The coverage is not complete, and no attempt is made to repeat common diagnostic properties already recorded in the literature. Larger-scale features of the more important metallics have been given in descriptions of ore types.

Oxides and borates

Magnetite

Magnetite occurs as single elliptical grains, irregular areas, amoeboid blotches, coalescing blotches, irregular layers, and (rarely) as streaks, crude veinlets, and homogeneous masses. All except isolated grains are aggregates of gently-curved, subpolygonal, or polygonal individuals usually of a size roughly equal to or slightly greater than that of the constituents of the host rock (i.e., roughly 0.5-5 mm). Their aggregated state is usually visible only in polished sections and is further accentuated by etching in HCl. There is no sign of concentric growth. Euhedral crystals of magnetite are very rare.

Among the metallics, magnetite always seems to have developed first. While some of it is interstitial to and molded on minerals of the host rock (particularly where the magnetite is sparse), most of it shows clear evidence that it has replaced pre-existing minerals by attack at the grain boundaries or at isolated points. Small projections of magnetite, finger- or U-shaped, commonly occur at the periphery of magnetite aggregates, and thinner "necks" connect single grains and aggregates, isolating and embaying the minerals of the host. Primary control is exercised by any planar structure within the rock, but with increasing quantity of magnetite this planar structure is transected and the minerals of the host are enveloped indiscriminately, either as individuals or as aggregates. In a rock containing both light and dark minerals (e.g., a salitic pyroxene skarn with calcite relics, or a granite gneiss with biotite and garnet), the magnetite often attacks the dark minerals first or more extensively, usually showing slight regard for cleavage. As a result of initially preferential and later indiscriminate replacement of host-rock minerals, the magnetite grains and aggregates contain a wide variety of subhedral, rounded, or corroded inclusions of dark silicates, felsics, apatite, fluorite, barite, or other minerals appropriate to the particular host. These include relics, particularly if they are very small, and may give a badly pitted surface to individual grains in a polished section.

In addition to inclusions, the magnetite grains often contain small and generally regular bodies exsolved from the magnetite on cooling. The commonest exsolution bodies are ilmenite (see below) and spinel. Dark-green spinel (hercynite or pleonaste?) forms small, scattered grains and larger areas at or near the borders of some magnetite grains from all types of deposit, though these recognizable spinel bodies are fairly common in granite gneiss ores and ores of modified type and fairly rare in skarn ores. Very rarely, a stray spinel is found outside the magnetite. Polished sections show that many magnetite grains have extremely minute "dashes" or flattened spindles of an apparently isotropic nonopaque thought to be spinel. These minute bodies are regularly aligned along either the octahedral or cubic direction in magnetite; locally, both sets are present. Similar bodies are noted in Schneiderhöhn and Ramdohr (1931, pp. 581-582) and have been described briefly by Sims (1950, p. 209).

Minute yellowish or brownish prisms of goethite (?), as isolated crystals or as reticulate aggregates along (111) in magnetite, are seen in many thin sections of granite gneiss ore. Very rarely, they also appear in skarn. Their euhedral shape and regular orientation suggest exsolution from magnetite, but the fact that rare prisms of similar material occur locally in biotite may mean that the goethite (?) is an alteration product. If so, it is almost certainly hopogene.

Another mineral sparingly present in some magnetite is olive-brown, weakly-birefringent chlorite (?) as minute irregular patches. This material may be an alteration product, or it may be an unusual exsolution product. Its identity and origin are uncertain. It does not resemble certain types of late chlorite occurring with carbonates as veinlets in the ore and country rock.

None of the magnetites has yet been analyzed, chiefly because most grains have inclusions or exsolution products. Inadequate data from certain assays and polished sections suggest that small quantities of Ti, Al, Mn, and Mg are present within the crystal lattice of some "pure" magnetite.

Hematite

Primary crystalline hematite

Primary crystalline hematite of hypogene origin occurs as a discrete phase in most of the granite gneiss ores and their modified equivalent. As an accessory or major mineral in a number of magnetite deposits, and as a disseminated accessory mineral in certain rock belts of the district (Buddington, personal communication), this primary crystalline hematite is associated exclusively with potash-rich rocks. Sillimanite is a very characteristic but not ubiquitous associate of the hematite.

It is undesirable to call the primary crystalline "specularite", for--though metallic--it lacks the platy habit and extremely high luster of specular hematite. It is quite wrong to call the mineral "martite", for that term is limited to hematite pseudomorphous after magnetite; this the primary crystalline hematite certainly is not. Where clarity _/ is essential, the cumbersome term "primary crystalline hematite" will be used to designate this discrete crystalline phase of iron oxide; elsewhere "hematite" or "primary hematite" will suffice.

Megascopically, primary crystalline hematite has a bluish cast relative to black or gray magnetite. In thin sections viewed by reflected light, the hematite is distinctly lighter than magnetite, and in polished sections the hematite is extremely bright and whitish or bluish-white relative to brownish magnetite. The hematite occurs in grains and aggregates of the same size, shape, and relation as those of magnetite. Locally it is somewhat elongate, but it is never platy. Sporadic grains--especially the larger ones--have one or two sets of twin lamellae. Inclusions and exsolution bodies are common but less conspicuous and less abundant than similar features in magnetite.

Hematite replaces host-rock minerals the same way magnetite does. However, in all deposits except the Parish, the two iron oxides appear to have developed contemporaneously as discrete phases. At the Parish deposit, where hematite is locally abundant, hematite looks slightly younger than magnetite. (Aggregates of hematite grains partly encircle grains of magnetite; subhedral hematite crystals project into anhedral magnetite grains; angular projections of hematite appear to cut

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_/ The long term itself is ambiguous, but the writer has not thought of a better one. Suggestions are in order.

magnetite; and hematite grains lie at random athwart magnetite aggregates having a definite planar orientation. (Buddington, personal communication, does not believe the relations convincingly demonstrate that hematite is younger than magnetite in this deposit.) Everywhere else, magnetite and primary crystalline hematite appear to be contemporaneous or "conjunctive".

Primary crystalline hematite never contains corroded relics of magnetite; in fact, magnetite completely included in hematite is very rare. This notable feature at once sets apart the primary crystalline hematite from martite. (See below.) Primary hematite is commonly accompanied by magnetite, but the latter may be entirely lacking. Within the district, economically important concentrations of primary crystalline hematite are limited to the Benson Mines and Parish deposits, so far as we now know.

Martite

Martite (hematite pseudomorphous after magnetite) occurs sporadically and in variable quantity in skarn ores, granite gneiss ores, and their modified equivalents. The martite consists of very fine aggregates of hematite that replace parts of individual magnetite grains. Martitization begins at grain boundaries and works inward, successively forming partial rims, patches, blades, stringlets, veinlets, lattices, dendrites, and networks--all with highly irregular, ragged boundaries. Control of replacement by definite crystallographic directions in the magnetite is often pronounced. In places, a magnetite grain will be almost wholly replaced by martite, leaving only a few scattered relics of magnetite to indicate that the "hematite" grain, as it now exists, developed by replacement of magnetite, instead of by replacement of silicates by primary crystalline hematite. The remarkable fact is that martitization is seldom complete; progressive replacement is observable in some magnetite grains in a given specimen, and relics or "islands" of magnetite are left in the grains that at low magnification appear to be all hematite. From the evidence now at hand, it appears that martite never contains exsolved bodies of ilmenite or spinel, whereas primary crystalline hematite may have such bodies. More data are required before this apparently distinctive difference can be stated as a rule.

Martite is more likely to be found in granite gneiss ores than in skarn ores. Its distribution in a given specimen is usually highly irregular, and the extent of martitization shown by the magnetite in a given deposit varies widely. Primary crystalline hematite may or may not be present in the same specimen with martitized magnetite. Ordinarily, there is nothing about its distribution to suggest that martite is supergene; and on the whole, martite is usually present in rocks that are entirely fresh. For these reasons, it seems that most martite in the district--particularly in granite gneiss ores--is hypogene. (Similar views have been expressed by Schneiderhöhn and Ramdohr, 1931, pp. 529-532, and by others.) However, the position of martite in the paragenetic sequence is uncertain. It is, of course, younger than magnetite; but its relation to sulfides of stage I is unknown, for sulfides and martitized magnetite have seldom been found touching one another.

Locally in the skarn deposits, some supergene martite is present. For example, a few magnetite-bearing sköls have been altered by supergene agents, and part of the magnetite they contain is martitized. The same sköls have thin hematite veinlets cutting indiscriminately through the rock. Also, some skarn magnetite near inferred faults shows partial martitization.

Martite is present and locally abundant in magnetite from the Clinton County district, New York (Postel and Rogers, in preparation), and the Dover district, New Jersey (Sims, 1950, p. 208).

Other varieties of hematite

Other varieties of hematite are present in the district. These include the veinlets of finely-crystalline or jaspery hematite cutting skarn and sköls locally; the earthy red hematite (accompanied by dark-green ferriferous chlorite, carbonate veinlets, and sporadic pyrite) in fault zones and along contacts of some marble and sköls; and the scattered deposits of hematite (earthy, finely crystalline, or specular) replacing pre-Cambrian marble beneath or close to cappings of Potsdam sandstone. All these occurrences are believed to be supergene.

Ilmenite

Ilmenite is widely distributed in very small amounts in almost all the "nontitaniferous" magnetite deposits of the district. Generally the ilmenite can be detected only by means of the reflecting microscope but some polished specimens of ore have blades of ilmenite visible to the unaided eye. The ilmenite occurs exclusively as an exsolution product in magnetite or--less often--in primary crystalline hematite. It is common in all the granite gneiss ores and in the modified skarn ores. It is sparse in modified granite gneiss ores, and its occurrence in skarn ores is sporadic and quantitatively insignificant. The ilmenite occurs as hairs, slivers, and blades, and (very sparingly) as small, angular grains. None of the material has enough ilmenite to be classed as titanomagnetite. Hairs and blades are always confined to individual grains of iron oxides, and this is usually true also of the ilmenite grains. However, the latter occur very rarely with one side touching silicates and the other three surrounded by magnetite; or as small, elongate, curving "bridges" connecting two larger aggregates of magnetite. Discrete ilmenite grains in silicates have never been found.

Blades or hairs of ilmenite in magnetite are present in almost every polished section of granite gneiss ores and modified skarn ores. Commonly, a few magnetite grains will show one or two blades apiece, but in some specimens more than half the magnetite grains have several blades of ilmenite. Included ilmenite grains are always present when blades are abundant. The blades often occur hit-or-miss in the magnetite, though reticulate plates parallel to the (111) plane of magnetite are observed in places. Infrequently, the (100) direction of magnetite also has a few ilmenite blades. Some blades fail to reach the borders of the magnetite host, but a great many have one end at the edge of the host and may even extend across the entire magnetite grain. Ilmenite blades have not been detected in the smallest magnetite grains present in the sections. Locally, ilmenite blades are skeletal and seem partly replaced by a nonopaque or by hematite. The secondary hematite may be present even when the enclosing magnetite shows no martite. Very rarely, the ilmenite blades contain minute exsolution bodies of hematite and spinel.

The quantity of ilmenite is variable but small. Perhaps an average value for a thick ore zone in granite gneiss is about 0.5 percent. Detailed studies of the distribution of ilmenite have been made only at the Trembley Mountain deposit (modified skarn type), where the quantity ranges from nil to a maximum of ca. 2.8 percent, averaging ca. 0.5 percent. A statistical study of cores from two ore zones cut by one drill hole (T3) at the deposit shows a relative concentration of ilmenite at the top of the upper ore zone and the bottom of the lower one. There is no direct correlation between the quantity of ilmenite and any one of the following: grade of ore, type of host rock, quantity of sulfides, quantity of apatite.

Newland (1908, p. 23) long ago pointed out that the "nontitaniferous" Adirondack magnetite deposits are only relatively free of TiO_2, as some TiO_2 has been found in almost all of them. He writes: "The titanium is traceable usually to the mineral titanite which is a common constituent of the wall rocks and is often intergrown with the magnetite. Its proportion is generally higher in crude ores than in concentrates, the titanite being removed to a greater or less extent by mill treatment." Newland's work was done long before the days of mineragraphy, and he was less familiar with the St. Lawrence County deposits than with others in the Adirondacks. We now know that sphene (or titanite) accounts for very little of the TiO_2 in the St. Lawrence County ores.

Gallagher (1937, p. 67) found ilmenite and sphene in ore from Lyon Mountain. Colony (1921, p. 114 and plate 11, fig. 3) reported ilmenite from the Canopus mine, southeastern New York. More than forty years ago, Bayley (1910, p. 115) inferred that ilmenite, either modal or normative, was present in some ores from the Highlands of New Jersey, and Sims (1950, pp. 207-208) has recently noted ilmenite in magnetite from several deposits in the Dover district.

Bugge (1943, p. 130) lists ilmenite among the minerals found in the skarn-iron ore deposits of Arendal, Norway, but gives no data on its occurrence. To date, a little has been published on the mineragraphy of the magnetites of central and northern Sweden. The writer believes that ilmenite has not been reported from the Central Swedish district, where the ores are characteristically very low in TiO_2. However, Geijer (1910, p. 102) found ilmenite as well as sphene occurring locally in ores from the Kiruna district, northern Sweden.

The above notes point up the fact that small amounts of ilmenite are present in many magnetite ores that are regarded metallurgically as nontitaniferous. This ilmenite may account for part or most of the TiO_2 present in the ores from several districts. (Additional comments are found in the section on economic applications.) True titaniferous magnetites are also found in the St. Lawrence County district, but these are associated with metagabbro and an ultramafic differentiate of the quartz syenitic series. These titaniferous ores are considered very briefly in a subsequent section.

Vonsenite

Introduction

The rare mineral vonsenite (iron-rich ludwigite) _/ is found with magnetite in ore from the Jayville deposit, w.c.r., Oswegatchie quadrangle, New York. Jayville is a deposit of the skarn type, differing from others in the district in having abundant fluorite-rich ferrohastingsite skarn and considerable anthophyllite skarn, as well as the usual pyroxene skarn. Biotite sköls are prominent at Jayville, and they carry much of the ore. Vonsenite has been found in abundance in "high-grade ore" (half vonsenite, half magnetite, with a little gangue) from one of these sköls. It also occurs as a subordinate metallic in ore that replaces quartz-bearing ferrohastingsite skarn.

Discovery and identification

In the winter of 1947, the writer found an unidentifiable metallic in ore from drill holes J2 and J3, Jayville deposit. The striking optical properties of the unknown set it apart from other metallics that had been investigated mineragraphically and described in the literature. It was thought for a while that the unknown might be ilvaite. Owing to the pressure of other work, the writer had no time for a thorough study of the material. In the spring of 1950, Mr. G. G. L. Henderson, of Princeton University, investigated the unknown for the writer under the

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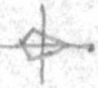
_/ Ludwigite is an Mg-Fe borate of the composition $(Mg,Fe)_2O_2Fe'''BO_3$?, according to Winchell (1951, p. 136).

direction of Professor Edward Sampson. Henderson made the usual etch tests and at great labor bored from the polished specimen enough "pure" material for spectrographic analysis and X-ray study. His X-ray work showed that the unknown was not ilvaite, but attempts to identify the mineral were fruitless. During January, 1951, samples isolated by Henderson were studied by J. M. Axelrod and Miss Janet D. Fletcher of the U. S. Geological Survey, under the direction of Michael Fleischer. X-ray work by Axelrod and spectrographic analysis by Miss Fletcher identified the unknown Jayville mineral as vonsenite.

Physical properties in hand specimen

In hand specimen, vonsenite is indistinguishable from the fine-grained magnetite with which it occurs. The vonsenite forms gray, granular aggregates with metallic (locally adamantine) luster. Powdered vonsenite is very weakly magnetic, but because of the intimate association of vonsenite and magnetite, this property is not of much use in distinguishing the two minerals where they constitute ore. If one examines polished specimens of ore with the naked eye, one sees that the vonsenite is definitely metallic. Relative to magnetite, the vonsenite appears bluish; and its reflectivity or "brightness" relative to magnetite is about the same as the reflectivity of magnetite relative to primary crystalline hematite.

Optical properties and etch tests

Vonsenite is completely opaque in thin section. In polished sections examined with the reflecting microscope, vonsenite shows the following properties. *Color*--pleochroic from brownish to bluish-gray; looks very gray with respect to magnetite, which appears brownish; euhedral vonsenite crystals appear brownish in this position, and bluish-gray in this position. *Reflectivity*--relatively low. *Anisotropism*--strong, from reddish-brown to blue; no internal reflection. *Hardness*--less than hematite; equal to magnetite (i.e., hardness of F on Short's scale or 6 on Mohs' scale).

Etch tests made by Henderson gave the following reactions: HCL--definite etch, appears to reveal cracks; soluble aureole formed around drop, possibly a "sweat aureole". $HgCl_2$, KOH, $FeCl_3$, KCN, HNO_3--negative.

Distinction from other metallics

When so little information on vonsenite is available, it is difficult to say with assurance how that mineral can be distinguished from others of similar appearance. Nevertheless, the following suggestions may be helpful--at least until the exceptions make themselves felt.

Vonsenite is most likely to be confused with ilvaite, and the distinction may be very difficult to make in polished sections alone. The occurrence, habit, hardness, etch reactions, pleochroism, and anisotropism of the two are almost identical. However, ilvaite is almost always at least slightly translucent in thin section; in polished section, it may show a red-brown internal reflection. Some ilvaite shows one good cleavage, and some shows zonal growth. (These data are compiled from Ramdohr, 1950, pp. 759-760; and from Hotz, 1949, pp. 34-35.) X-ray patterns of the two minerals are very different (Henderson, 1950).

In general, color and pleochroism alone are sufficient to distinguish vonsenite from bixbyite, cassiterite, hematite, lepidocrocite, and α- vredenburgite. (See Mason, 1942, 1943b, for data on the manganese minerals; Ramdohr, 1950, for data on the others.)

The properties of ludwigite in polished section (Schneiderhöhn and Ramdohr, 1931, pp. 567-568) are almost identical with those of vonsenite, given above by this writer. The distinction between ludwigite and vonsenite would probably be very difficult to make without a semiquantitative determination of the Fe content. However, the ludwigite so far described is at least slightly transparent in thin section or in thin splinters. For some varieties, prismatic habit and silky luster are characteristic though not diagnostic.

X-ray and spectrographic analysis

Henderson's data on the vonsenite, obtained by means of the X-ray spectrometer, are:

Peak	2θ	d
1	44.1	2.5802
2	33.70	3.3417
3	47.95	2.3837

A subsequent X-ray investigation by J. M. Axelrod identified the mineral as vonsenite. Axelrod notes that the pattern of the Jayville vonsenite is slightly different from that of the standard, vonsenite from Riverside, California.

A qualitative spectrographic analysis of Jayville vonsenite, made by Miss Janet D. Fletcher, confirmed Axelrod's identification of the mineral and showed the following elements to be present:

Element	Percent
Fe, B	5
Mg, Si, Al	1-5
Sn, Pb, Zn, Mn, Ti	.X
Cu, Cr, Zr, Ba, Ca	.0X
Ag, Co, Ni	.00X
-	.000X

(Looked for but not found: As, Bi, Be, Au, Pt, Mo, W, Ge, Sb, Cd, Tl, Ga, V, Y, La, Th, Cb, Ta, U, Sr, Na, P.)

Microscopic occurrence

A thin section of "high-grade ore" from a biotite sköl shows magnetite and vonsenite (both gray and opaque) replacing a crystalloblastic aggregate of pale-green biotite and strongly-pleochroic anthophyllite (? -- possibly hypersthene). A few grains of bluish ferrohastingsite and traces of zircon are also present, together with traces of one or more unidentified accessories. The anthophyllite (?) is slightly altered to greenish-brown chlorite. The polished section of this rock shows aggregates of magnetite and vonsenite, generally as polygonal grains, replacing the nonopaques. Vonsenite also occurs as scattered diamond- or lozenge-shaped crystals and as small subhedral flecks in the gangue. The magnetite and vonsenite appear to be contemporaneous. A few hair-thin veinlets of secondary hematite cut both gangue and vonsenite. Ragged, crudely-bladed patches of hematite replace the vonsenite locally.

A thin section of high-grade ore from quartz-bearing ferrohastingsite skarn that contains a little pale-green mica (biotite?), green chlorite rosettes, and traces of unknowns has abundant magnetite, subordinate vonsenite, pyrite (ca. 5 percent), and a trace of hematite. Magnetite and vonsenite replace amphibole, mica, and quartz. Subhedral vonsenite grains, usually attached to magnetite, appear to be contemporaneous with or slightly younger than the magnetite. Pyrite of stage I replaces magnetite and gangue. Sparse secondary hematite is present as flecks and as crack fillings in gangue.

Economic significance

Vonsenite and magnetite look so much alike to the unaided eye, and are so intimately associated in granular masses which in bulk are strongly magnetic, that the writer cannot estimate the quantity of vonsenite in the Jayville deposit. It may be substantial.

The finding of vonsenite at Jayville makes one wonder if other magnetite-bearing sköls of the district also carry this rare boron mineral. The number of polished sections of sköl ores in the writer's possession is too small to sample the sköls adequately. Some drill cores logged as high-grade magnetite ore showed medium-grade ore upon assay; possibly this merely reflects the difficulty of making a fair estimate of grade by eye, but possibly it means that vonsenite accompanies the magnetite in some of this ore. (The total Fe content of vonsenite may be ca. 55 percent, whereas the total Fe content of theoretically pure magnetite is 72.3 percent).

Other occurrences of vonsenite

Vonsenite was first described, analyzed, and named by A. S. Eakle (1920). The original vonsenite occurfed with magnetite, chlorite, green pyroxene, and a white talcose mineral in a large mass formed at the contact between limestone and granite at Riverside, California (Eakle, 1920, p. 141). Butler and Schaller (1917) had previously discussed the ludwigite series, ranging as they thought from one known end-member--magnesioludwigite--to an inferred iron-rich end-member--ferroludwigite. Geijer (1939, p. 21) mentioned vonsenite as an end-member of the ludwigite series and presented an excellent review of the Swedish occurrences of ludwigite, summarizing their petrologic significance. Ludwigite is, of course, known from a number of pyrometasomatic deposits in Sweden, central Europe, Korea, and the Western United States.

Additional work

Optical and chemical data for various magnesian and intermediate members of the ludwigite series are available in the literature, but the only published information specifically regarding vonsenite seems to be that of Eakle (1920). For that reason, an early account of the Jayville vonsenite is desirable, emphasizing optical properties (on which there is no published information) and occurrence. This account should, if possible, be accompanied by a chemical analysis of the vonsenite, provided enough material can be isolated for analysis. It is doubtful if the Jayville samples will provide suitable material for single-crystal X-ray study, but such an investigation needs to be undertaken. (Takeuchi, Watanabé, and Ito, 1950, have recently investigated the crystal structure of warwickite, ludwigite, and pinakiolite.) It would also be desirable to determine and present optical data for the original vonsenite of Eakle.

Sulfides and arsenides

The occurrence, distribution, and general features of the sulfides have already been summarized. The observations that follow are scattered and incomplete.

Pyrite and marcasite

Pyrite of stage I occurs as flecks, elliptical grains, irregular aggregates, partly or wholly developed cubes, arrays of partly developed pyritohedra, stout veinlets, and masses. It is pyrite of this generation that one usually sees in hand specimens. Most of the pyrite is isotropic, but some of it (including certain partly developed cubes and pyritohedra) is weakly to moderately anisotropic. Pyrite of stage I is molded on some silicates, replaces others, and locally replaces magnetite and primary hematite. Concentrations of pyrite with chlorite, epidote, and abundant apatite are found locally in granite gneiss ores. Much of this pyrite may belong to stage II, but some of it may belong to stage I.

Pyrite of stage II occurs as extremely fine veinlets that fill fractures and cleavages in silicates and, less often, fractures in magnetite. The veinlets pass into thin films and fringes of pyrite that partly or wholly envelop some grains of silicate, magnetite, and subhedral pyrite I. Because the veinlets and many films are exceedingly thin, they are usually seen only in polished section. Some "filigree" pyrite may also belong to this later generation. All pyrite of stage II is, with one questionable exception, isotropic. Pyrite assigned to stage III also appears to be isotropic.

Marcasite has not been positively identified from the nontitaniferous magnetite deposits. (It has been found sparingly in titaniferous magnetite deposits associated with metagabbro.) However, some of the "moderately anisotropic pyrite" may really be marcasite, though this seems doubtful in view of the crystal form assumed by some of the material. (See notes on pyrite of stage I.)

Pyrrhotite

Pyrrhotite occurs as grains, knots, stout veinlets, irregular aggregates and masses, and rare fracture fillings. Local protuberances cut silicates and magnetite, though when magnetite and pyrrhotite are sparse they tend to show "mutual boundaries". Pyrrhotite is known to replace silicates, carbonates, magnetite, ilmenite, and pyrite of stage I. Its relation toward molybdenite is indeterminate. Aggregates of pyrrhotite are made up of individuals with curved or sutured borders. Certain pyrrhotite veinlets in skarn tail out into veinlets of chalcopyrite. Pyrrhotite masses in skarn are likely to be rich in apatite.

Sphalerite

Sphalerite of stage I occurs as specks, grains, and small areas in silicates, as rims on magnetite, as elongate grains cutting magnetite and primary crystalline hematite, as intergrowths with pyrite, and as patches along cleavages and cracks in pyroxene. The sphalerite is found in small quantities in all types of magnetite deposit, but its concentration seldom reaches 1 percent. Often it is present only as a trace. It is always accompanied by another sulfide, though individual areas of sphalerite may or may not be intimately associated with the sulfide. Locally, sphalerite of stage I contains exsolved blebs of chalcopyrite.

Sphalerite of stage II has been identified in calcite veinlets from several deposits, as perched crystals with zeolites from the Parish deposit, and as a minute accessory in ore from Trembley Mountain. About half the specimens of ore from Trembley Mountain show minute grains of sphalerite tentatively assigned to stage II, restricted almost wholly to ore in which fringes and veinlets of pyrite II were present. This sphalerite II occurs as minute irregular aggregates about 50 microns in diameter, made up of individuals about 10 microns in diameter. The grains occur in trains near (but not on) the borders of magnetite grains, on the outer edges of distinct silicate rims. In addition, they are present as rare scattered specks in silicates. In a few grains, tiny projections of sphalerite cut the enclosing silicates.

Sphalerite of both stages is gray and isotropic, with whitish, yellow, or brown internal reflection. The identification of sphalerite of both generations was checked by etching the material with $KMnO_4$ reagent. After the brown stain (locally red and blue) was wiped off, single "grains" of sphalerite were found to be aggregates of smaller individuals showing prominent cleavage (or twinning).

Bornite

Bornite, so far known in this district only from the Parish deposit (q.v.), occurs as scattered flecks, small grains, and irregular small areas in the gangue of some ore and weakly-mineralized rock. The borders of the bornite are very ragged, and the mineral seldom comes against iron oxides. However, in places it replaces magnetite and cuts primary hematite. Its relations with sphalerite are uncertain; bornite seems younger. Against grains of chalcopyrite, the bornite usually displays "mutual boundaries". However, at high magnification some of the "mutual boundaries" are clearly seen to be irregular, showing tiny filaments of chalcopyrite projecting into the bornite. From this it seems certain that chalcopyrite is younger. Moreover, much of the bornite has exsolved chalcopyrite as regularly-oriented minute blades or shreds, and as rare "flames" and lenticular blebs. In places, the exsolved chalcopyrite blades form square or rhombic lattices where they lie along the cubic or octohedral directions of the bornite. Part of the bornite is usually replaced by chalcopyrite (q.v.).

Chalcopyrite

Chalcopyrite may be the only sulfide in an ore specimen, but it is usually accompanied by at least one other sulfide. Pyrite and pyrrhotite are its common associates. The chalcopyrite occurs as flecks, grains, and minute veinlets in silicates and earlier metallics. In parts of some deposits (Trembley Mountain North), disseminated chalcopyrite is present to the extent of 2 percent. Rarely (at Brandy Brook Northwest) it forms a vein as much as a foot thick. It is also found as replacement veinlets in magnetite, networks in pyrite, and blebs and veinlets in pyrrhotite. It is younger than sphalerite. Rarely, the chalcopyrite is twinned. Some chalcopyrite grains have corroded cores of pyrite. As an exsolution product, chalcopyrite is found in sphalerite of stage I and in bornite.

Chalcocite

Chalcocite, like bornite, is restricted to the Parish deposit, where it occurs as irregular rims on bornite and chalcopyrite, as ragged networks in bornite, and as sheaths about exsolved chalcopyrite blades in bornite. The chalcocite is blue-gray and anisotropic; some of the "sheath" type is a deeper blue. Locally, chalcocite replaces all but the centers of bornite and chalcopyrite grains, leaving ragged islands of the host. However, bornite is usually replaced more extensively than chalcopyrite. In a few places, the chalcocite contains regularly oriented blades of chalcopyrite that resemble an exsolution product but are probably unreplaced chalcopyrite exsolved from bornite that was later entirely replaced by chalcocite. There is no evidence to suggest that any of the chalcocite is supergene.

Covellite

Microscopic traces of covellite have been found in sulfides from the Parish, Benson Mines, and Jarvis Bridge deposits. Covellite from the Parish deposit occurs as local replacements of chalcocite and bornite. Covellite from Benson Mines forms a partial fringe on chalcopyrite grains that contain pyrite relics. Covellite from Jarvis Bridge occurs as a rare fleck-like alteration of chalcopyrite. All the covellite is strongly pleochroic from silver-white to deep blue; between crossed nicols, the mineral is a brilliant, fiery red. Were it not for its unmistakable optical properties, the mineral would pass unnoticed, for it is very sparse. The covellite may or may not be hypogene. It is found in fresh specimens collected at the surface and in fresh cores from deep drill holes.

Loellingite

Traces of loellingite have been found in a single specimen from the Clifton mine. Brilliant silvery-white loellingite occurs as specks with magnetite, abundant pyrrhotite, and an unidentified colloform metallic in silicated marble from a weakly-mineralized zone of skarn. The loellingite was noted by the writer and P. E. Hotz in 1948. The mineral seemed very easily scratched. G. G . L. Henderson investigated the mineral microchemically and by X-ray in the spring of 1950, under the direction of Professor Edward Sampson. _/ Henderson found Fe and As by microchemical tests and inferred the presence of other undetermined elements. His X-ray data are as follows:

Peak	2θ	d
1	44.48	2.5593
2	48.60	2.3538
3	43.90	2.5914
4	62.76	1.8602

A spectrographic analysis of the unknown seemed desirable. This was made by Miss Janet D. Fletcher on material that Henderson had bored

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_/ Polished sections of the sample were not made at that time. Because the unknowns were so sparse in the rather crumbly specimen, it seemed better to perform as many tests as possible before risking the loss of the material in sawing.

from the sample. Miss Fletcher prepared a special set of standards to get a rough quantitative analysis of the unknown. Her results are:

Element	Percent
As (70%) Fe (26%)	>5
Co (3%)	1 - 5
Ca, Pb, Bi Ni, Mg, Si Al	.X
Cu, Zn, Mn Ti, Zr, Ba B	.0X
Cr	.00X
Ag	.000X

(Many other elements were looked for but not found.) The spectrographic analysis confirms J. M. Axelrod's X-ray identification of the unknown as loellingite.

Polished section studies by the writer are in progress.

Miscellaneous metallics

Several metallics of uncertain identity, including valleriite (?), cuprite (?), and others, are still being investigated by the writer. These minerals are present only in very small amounts, but some of them are interesting for their genetic significance. As the observations are tentative and incomplete, they will not be given here.

Late hydrothermal minerals

Minerals of late hydrothermal origin are found sporadically in and close to the magnetite deposits of the district. The quantity of these minerals ranges from microscopic traces to heavy concentrations in certain zones a few feet thick. Some of the commoner minerals have been found as veinlets or local alteration products in rocks far removed from the magnetite deposits, but the suite as a whole shows a notable restriction to the immediate vicinity of the mineralized zones.

This suite of late hydrothermal minerals includes: _/ analcime, apophyllite (?), calcite*, chabazite*, chalcopyrite, chlorite*, datolite (?), epidote*, fluorite, kaolinite (?), pectolite (?), prehnite (?), pumpellyite*, pyrite*, quartz, sericite, specularite (?), sphalerite, stilbite, and uninvestigated minerals thought to be zeolites.

The predominance of hydrous minerals is striking. The paragenetic sequence has not been entirely worked out, but epidote and pumpellyite are known to be relatively early, while calcite, pyrite, and traces of other metallics are late. (See preceding section on paragenetic sequence of metallics.) The writer believes that the members of the suite were deposited from hydrothermal solutions representing the final stage of the connected process of mineralization in the district. Minerals of relatively higher temperature and high Fe content (epidote and pumpellyite)

_/ (?) means that the mineral has been identified megascopically but has not yet been checked microscopically.

* means that the mineral is common and frequently prominent.

formed first, a series of hydrous K-Na-Ca-Al silicates followed with falling temperature, and a feeble spurt of base-metal sulfides and calcite concluded the mineralization process.

The late hydrothermal minerals are found to some extent in association with all types of magnetite deposit in the district, but as a rule they are sparse in skarn ores, common in granite gneiss ores, and common--even abundant locally--in modified deposits of both types. The minerals occur in three ways: (1) as sporadic or pervasive internal alteration and replacement of earlier minerals; (2) as linings of vugs; and (3) as joint-coatings and fracture-fillings. The distribution of the minerals according to the first two modes of occurrence is entirely unsystematic; that is, there is no relation to land surface, depth within bedrock, inferred faults, rock type, stratigraphic position, grade of ore, kind of host or wall rock, or internal structure of the rock. Moreover, the distribution of joints and fractures locally occupied by minerals of the third group has no known relation to the features cited above, though more information is needed on this point. The one feature common to all occurrences is the restriction of the more complex minerals and paragenesis to the immediate vicinity of ore.

The writer cannot give a reliable quantitative estimate to show the relative significance of alteration by the late hydrothermal minerals. At most, perhaps 20 or 30 percent of all the rocks cored at a given deposit would show some sign of late hydrothermal minerals. However, if all the late hydrothermal minerals in the deposits most affected were packed together, they would probably not exceed 2 or 3 percent of the bulk of the rocks drilled. If vuggy zones alone are considered, their aggregate for the deposits most affected would be a very small part of 1 percent. Nevertheless, their "frequency" would be considerable.

The only minerals commonly occurring as internal alteration products are chlorite, epidote, pyrite of stage II, pumpellyite and zoisite. These are present, singly or together, as granules and aggregates in the form of flecks in silicates, replacements along cleavages of silicates, and veinlets in--or irregular rims on--any earlier-formed mineral. In skarn deposits, the modified skarns are often the only rocks showing pumpellyite, and in most deposits the felsic minerals of a given rock are more pumpellyitized than the other constituents. Kaolinite (?) and chabazite from vugs locally spread out as alterations of the host rock.

Vugs occur sporadically in ore and wall rocks at a number of deposits. The vugs are usually more abundant in the ore and its immediate wall rocks, but they may occur several hundred feet away from the magnetite-bearing zones. A few of the vugs are simply leached cavities, but most vugs show a lining of new minerals of the late hydrothermal suite. The vugs are small--at most a few mm long and a mm or two wide--; roughly circular, irregular, or slightly elongate; and oriented with complete disregard for the foliation or any other structure in the host. The adjacent rock may be substantially altered for a few mm about the vug, though in general the host is quite fresh; only traces of alteration are visible beyond the vug itself. The vugs are definitely younger than magnetite and pyrite I, for these minerals and their host have in places been dissolved and have received a coating of the vug minerals. Epidote, pumpellyite, chlorite, chabazite, and pyrite are the commonest minerals lining these vugs. (The chabazite is very easily mistaken for colorless fluorite.) Vugs from the Parish deposit show the greatest number of late minerals.

The vugs are similar in a very crude way to the "miarolites" at Lyon Mountain, shown by Zimmer (1947a) to be leached cavities post-diabase in age. The St. Lawrence County vugs cannot be miarolites: _/ they are not lined with projecting crystals of the same minerals found in the host, and they are not restricted to rocks of magmatic origin. The host rocks are known to include microcline granite and its gneissic equivalent, pegmatite, metasedimentary gneisses, modified skarn, mineralized and ore-bearing parts of all these preceding rocks, biotitic amphibolite, and skarn (very rare). Moreover, the vugs and their linings are almost certainly not supergene: their host is often singularly "clean", without earthy hematite and with only local, sporadic chlorite; they are not distributed in any manner that suggests supergene alteration; they often contain pumpellyite, so far known only as a hypogene mineral; and the linings consist of minerals typical elsewhere of late hydrothermal origin.

The late hydrothermal minerals cannot be genetically related to

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_/ Fluorite-bearing vugs found very rarely in alaskite in the district may be true miarolites. They are not the same as the vugs just described.

diabase, as that rock is rarely found in the district and is not known to occur near any of the magnetite deposits. The association of these late minerals is with mineralized rocks, principally those in which K, Al, Si, and OH had already effected some modification. Somewhat similar relations apparently exist in other districts, including Arendal, Norway (Bugge, 1940, p. 93); the Oslo region, Norway (Goldschmidt, 1911, pp. 469-473); the Brötorp zinc mine, Orijärvi district, Finland (Borgström, 1908, pp. 334-337); and mines in the Central Swedish district (Geijer and Magnusson, 1944--zeolites, etc., cited in mineralogical appendix).

Pumpellyite warrants special comment. So far as the writer knows, St. Lawrence County is the first magnetite district found to contain pumpellyite. The mineral was subsequently found by Sims and identified by the writer at the Scrub Oaks mine, Dover district, New Jersey (Sims, 1950, p. 210). Traces of a mineral that may be pumpellyite have recently been found in the Andover district, New Jersey (Sims and Leonard, in preparation). The St. Lawrence County pumpellyite appears to be a ferroan variety occurring as local pervasive replacements (minute to extensive), veinlets, and vug linings in a variety of wallrocks and ore types. Its optical properties and further details of occurrence will be presented in a subsequent publication. Aside from the New York and New Jersey magnetites, pumpellyite has been found (1) in amygdules and veinlets, and as an alteration product, in basic igneous rocks; (2) in chlorite schists, glaucophane schists, and other metamorphic rocks; (3) in quartz veins; and (4) in modified skarns. Most of the significant references are cited by Winchell (1951, p. 519).

Several minerals found as alteration products of host-rock silicates cannot yet be fitted into the paragenetic sequence. These include muscovite (from microcline, sillimanite, and combinations of silicates), zoisite (ffom feldspars), dark mica (from pyriboles), and hydrogarnet (?) (rare--in skarns). One has the impression that such minerals are more or less contemporaneous with magnetite and definitely older than the late hydrothermal minerals, but the relations so far observed are entirely inconclusive. The quantitative extent of these alterations is very small indeed.

Structure of the magnetite deposits

The magnetite deposits of the district are replacement deposits developed on the noses and limbs of folds after deformation of the host rocks by plastic flow. The deposits are essentially conformable to the structure of their host, but locally they are transgressive. The pattern of at least one deposit has been altered by post-ore faulting. The regional structural setting of the deposits has already been presented.

It is often difficult or impossible to make a smooth transition in relating major folds, miles in length, to the local folds, hundreds or thousands of feet long, in which the magnetite deposits occur. Much of this difficulty stems from the inadequacy of outcrops in the immediate area of the deposits, but part of it is due to the complexity of the structural pattern imposed by at least two major deformations. In general, the local folds in their present form are not simple drag folds or flowage folds accompanying the major anticlinal masses of quartz syenitic rocks, nor are they minor folds related to major folds in areas some miles from quartz syenitic masses. They are, in varying degree, the result of later deformation acting on an earlier pattern to give a composite effect.

In the account that follows, the writer summarizes the structural types of deposit before discussing some of the elements that make up those types. This departure from logical order seems justified because the shape and pattern of the ore bodies are of primary concern to the men who must work them.

Structural types of deposit

The structural type--or simply the shape--of the deposits can be classed as follows: simple (?) tabular, fishhook, linear, multiple drag fold, and complex. Before these types are discussed, two features must be pointed out. First, no single deposit--operating or inactive--is so well known that its structure can be described with comfortable assurance. Second, most deposits in the district seem to have two long dimensions and one relatively short one, rather than one long dimension and two relatively short ones. There are exceptions to this rule, but in general the remarkable lath-like habit of certain New Jersey and New York magnetite deposits is inconspicuous--perhaps even lacking--in the St. Lawrence County magnetite district.

Simple (?) tabular bodies

One may expect to find relatively simple tabular or sheet-like bodies on the limbs of folds. The long, straight magnetic anomalies given by the Jarvis Bridge and Skate Creek deposits, coupled with the continuity of the mineralized zones shown by initial diamond drilling, suggest that these two deposits are roughly tabular in shape. They extend for several thousand feet along the strike, are generally about 20 feet to less than 50 feet thick, and have been tested down the dip to a distance of less than 500 feet. The dip of the Jarvis Bridge deposit changes gradually along strike from ca. 50 SW. at the southeast end to 70 SW. at the northwest end. The dip of the Skate Creek deposit varies somewhat more. As neither deposit has been tested by a fan of drill holes, the possible variation in dip at increasing depth is unknown.

The writer has described these deposits as relatively simple tabular bodies. However, he would point out that the structure of the magnetite ore body in granite gneiss at Lyon Mountain, Clinton County district, New York, appeared to be relatively simple for some distance along strike and down dip, yet the body is now known to have inherited a very complex group of folds. The structure of the upper parts of the ore body was that of a steeply-dipping, crudely-tabular mass; mining to depths of 2,000 feet and more eventually showed a complex, plunging syncline with sharp anticlinal cross folds. A small sample of the structure of any one part of a magnetite deposit may be a poor guide to the structure of the deposit as a whole, if one optimistically thinks of "the whole" as embracing bodies that possibly extend to great depth.

Fishhook bodies

Magnetite bodies whose plan views have the shape of a fishhook are common in the district. (See fig. 3.) Whether these deposits are also fishhook-shaped in cross section, we do not know. Examples are Spruce Mountain Northwest, Trembley Mountain, Deerlick Rapids, Dead Creek, Wilson Mountain, and Benson Mines. The fishhook pattern results from the replacement of host rocks on the nose and one limb of a fold. In this district, there is reasonable evidence that most or all of the "fishhooks" are isoclinal synclines. The synclines may be relatively open (Spruce Mountain Northwest), or closely appressed (Dead Creek). The inferred axial plane of the syncline may be vertical, or it may be inclined at a moderate angle, giving an overturned isoclinal fold. The magnetite deposit may lie on the overturned or upright limb of the fold. The inferred plunge of the fold axis may be gentle or steep.

The fishhook shape of ore bodies in metamorphic rocks is very common, though many American geologists, recalling the unusual ore bodies of the Franklin district, New Jersey, are loath to think of this structural pattern as common in any way. However, a few references to other deposits will suggest at least the geographic range of "fishhook" deposits: Broken Hill, New South Wales (Gustafson et al., 1950); Camlaren, Northwest Territories (Henderson and Fraser, 1948); Sherritt Gordon, Manitoba (Farley, 1948); Sullivan, British Columbia (Swanson, 1948); Arendal, Norway (Kjerulf and Dahll, 1861; Bugge, 1940); Central Swedish district (Geijer and Magnusson, 1944); Dover district, New Jersey (Sims, 1950); Ducktown, Tennessee (Emmons and Laney, 1926); Edwards-Balmat district, New York (Brown, 1936); Franklin, New Jersey (Spencer et al., 1908); 81 mine, Lyon Mountain, New York (Zimmer, 1948).

Three points relating to the fishhook shape of the magnetite deposits of St. Lawrence County deserve mention. First is the apparent restriction of magnetite deposits to synclines. This does not mean that, within a given deposit, ore will be absent from minor anticlinal structures; it does seem to mean that the noses of anticlines of intermediate size (amplitude of several hundred feet to 1,000 feet or so) have been unfavorable sites for replacement by magnetite. Yet anticlines of appropriate size and rock type are known within the district; they simply do not carry magnetite deposits of significant size. /In the Sterling-Ringwood district, New York-New Jersey, Hotz (1945) found ore on the noses of overturned anticlines as well as synclines. Sims (1950, p. 152) states that a mineralized zone of dark gneisses can be followed around the nose of the Hibernia anticline, Dover district, New Jersey./

A second point is the occurrence of ore on the upright, as well as on the overturned, limb of the syncline. Postel (personal communication) found that in the Clinton County district, New York, magnetite deposits were usually related to the overturned limb of isoclinal synclines. This he attributed to greater shearing in the rocks of the overturned limb, giving a more favorable site for ore deposition.

A third point, of general application, is the prevalence of the fishhook structure in so many ore deposits in different districts: why fishhooks, instead of symmetrical U's or V's? The common explanation is that deformation on the nose and in the axial zone of the fold was most intense; this, coupled with the plunge of the fold, gave a favorable site for ore deposition in those areas. But what deformational feature of one limb caused replacement there, instead of on the nose and axial zone only, or on both limbs symmetrically?

Linear bodies

The apparent absence of distinct lath-like or pencil-like bodies from the district has already been noted. However, two deposits only partly explored by diamond drilling may have elongate ore shoots of analogous habit. These deposits are Brandy Brook Northwest and Jayville. At both deposits, a prominent b lineation (parallel to the axes of local folds) is seen in some wall rocks. This, together with the distribution of ore in the few scattered drill holes, suggests that any additional drilling ought to consider the possibility of linear shoots of ore, as opposed to continuous sheets. (At Jayville, a very pronounced lineation in a is also present locally. A hole drilled to test the assumption that ore might occur in shoots parallel to the a lineation was barren.)

At Spruce Mountain Northwest, the presence of elongate lobes of magnetic "highs", parallel to the axis of the syncline, suggests linear ore shoots beneath (Leonard, report in preparation). At Dead Creek No. 1 (Buddington and Leonard, 1945a), an elongate area of low positive readings at the southeast end of the magnetic anomaly suggests the edge (?) of a pitching ore shoot.

It must be emphasized that the existence of linear ore shoots in this district has not been demonstrated, though such shoots are known to occur in neighboring magnetite districts (e.g., 81 mine, Lyon Mountain, New York--Zimmer, 1948). However, the possibility that linear ore shoots are present in some deposits ought to be kept clearly in mind. The extent to which the "fishhooks" may persist as large linear features parallel to the axial plunge of their associated folds, can only be conjectured.

Deposits on minor folds

Several prospects, distinguished chiefly for their small size, are on minor folds that interrupt very slightly an otherwise continuous belt of skarn and associated metasediments. At the Green Farm (Newland, 1908, pp. 138-139) several pits were sunk in ore on the axes of minor folds. These folds, having an amplitude of only 15 to 20 feet, are found in intervals of 25 to 100 (?) feet along the N. 30 W. strike of the "vein". The folds trend ca. S. 15 W. and plunge ca. 20 SSW. The magnetic anomaly is continuous along the covered belt for ca. 450 feet, and one does not know if the pits are on the best part of the deposit or on "average ore". Somewhat similar structures are found in the Rampart Mountain-Lake Marion magnetite prospects, Tupper Lake quadrangle.

Ore is known to occur on minor folds, both anticlinal and synclinal, in many larger deposits of other structural types in the district. For that reason, it is perhaps unfair to place in a separate class the prospects noted above. However, the presence of magnetite concentrations on minor folds seems to be the only distinctive structural feature of these prospects.

Complex deposit

The Clifton deposit (magnetite in skarn) does not fit any of the simpler structural types found in the district. It is a deposit of complex habit. Perhaps this "relative complexity" is due mainly to our greater knowledge of the deposit--the only mine in the district that has underground workings, and the only one the Geological Survey has been able to map in detail (40 feet to the inch). However, there is also a sound geologic reason why the deposit is more complex: Clifton lies just within the anticlinal Stark complex of old, rigid granitic rocks, twice deformed by regional forces. Because of its unusual environment, the deposit may represent, compounded and accentuated, most or all of the structural features expectable in the rest of the deposits of the district. This is a comforting but probably unreliable speculation. Similar complexities are likely to appear as our knowledge of other deposits (particularly skarn ores) grows.

The Clifton deposit lies within a narrow belt of metasediments, several hundred feet thick, on the extreme eastern flank of the north-trending Stark anticline of phacoidal hornblende granite gneiss. At this latitude, the anticline is upright and its axis nearly horizontal or plunging gently southward. A very thin body of phacoidal granite gneiss, locally absent, intervenes between the metasediments and a large mass of younger hornblende granite on the east. In the immediate neighborhood of the mine, this hornblende granite is a completely deformed and recrystallized gneiss, locally having an alaskitic facies along its contact with the older rocks. The metasediments are intimately penetrated and locally brecciated by sheets of phacoidal granite gneiss. Both groups of rocks are cut, in one area, by dikes and sheets of hypersthene metadiabase, known from its occurrence elsewhere to be older than the hornblende granite gneiss and its alaskitic facies.

The average strike of the belt of metasediments is N. 45° E., and the usual dip is southeast. Two principal systems of folds are present in the mine area. There is considerable local evidence indicating that the two systems belong to two different periods of deformation. One system has axes trending ca. S. 10° E. and plunging 0-40° S. Let us designate the axes of this system as b_S. The b_S folds are essentially parallel to the axis of the great Stark anticline of phacoidal granite gneiss, the major structure of the area. Folds of this system appear to be relatively older; they may be interpreted as flanking drag folds subsidiary to the great Stark anticline. The second system of folds has axes trending ca. N. 58° E. Let us designate the axes of this system as b_E. The axes of b_E folds plunge gently ENE. or WSW., seemingly averaging about horizontal. The b_E folds appear to be relatively younger. Some members of this system are severely overturned toward the northwest, and in places the movement has approximated low-angle overthrusting. A third and subordinate fold system, intermediate between the first two in direction, apparently bears the relation of a resultant toward the first two. (Why this should be so, is not entirely clear.) Folds of this system are similar in character to those of the b_S system, and their present trend suggests that they are b_S folds rotated (or by some means moved) into a new trend by the forces that produced the b_E folds. Let us designate these rotated (?) resultant folds as b_{S_R}. Many other types of lineation accompany the three fold systems. Most of these lineations appear to be in b with respect to the individual fold system, but some are in a. In addition, there are a few anomalous lineations whose significance is at present unknown.

The thickest known area of high-grade ore at Clifton formed a large, blunt lens, rhomb-shaped in plan, whose known and inferred boundaries were clearly controlled by elements of the b_S, b_{S_R}, and b_E fold systems. (See fig. 4.) The stoped-out ore body lay in a synclinal sag between anticlinal folds of the b_S and b_{S_R} systems. The body was elongate in a NE. to ENE. direction. The northeast edge of the ore body ended rather sharply against an overturned anticline of the b_S system. The southwest edge thinned and frayed out over the top of a broader anticline made up of many minor folds of the b_{S_R} system. Relations at this edge are much less certain, owing to the appearance of dikes and sheets of hypersthene metadiabase. The thinned upper part of the main ore body cropped out at the surface. This area was mined out in the open cut, so that the structural relations are now unobservable. According to the writer's restoration based on field mapping and drilling data, the upper edge of the ore body had thinned along the crest of an anticlinal roll of the b_E system. From underground mapping and drilling data, we know that the lower edge of part of the thick lens thinned abruptly along an overturned fold of the b_E system. Present mining is largely from a part of the ore body lying southwest of the thick main lens, and at greater depth. This part may represent another crude rhomb arranged en echelon with respect to the main stoped-out ore body. Exploration in this area has been limited, but the results are promising. No attempt has been made to prospect for deep ore bodies down the plunge of the b_S folds, below the main lens.

The pattern outlined above gives only the general features of the ore body or bodies. This pattern is greatly complicated by many subordinate folds belonging to the b_S, b_{S_R}, and b_R systems. These result in subordinate but conspicuous pinching, swelling, and billowing of the ore. Where folds of the b_S system are relatively open, they are warps; where folds of the b_R system are relatively open, they are rolls. Locally, folds of the b_S and b_{S_R} systems are tight, isoclinal crumples and "drags"; and folds of the b_R system are overturned isoclinal folds whose under sides approximate low-angle overthrusts. A few representative cross sections are given in figure 4.

The most favorable places for ore deposition were the intersections of synclinal folds of the two main systems. Less favorable sites were the intersections of synclines of one fold system with anticlines of the other. Ore seems to be lacking where anticlines of the two fold systems intersect. (These tentative conclusions are based on grade-thickness contour maps prepared by consultants for the Hanna Coal and Ore Corporation and on the results of detailed mapping by the Geological Survey.)

The relative ages of the fold systems are not definitely established, but several lines of evidence point to the b_S system as the older one. The fact that the two deformations (or recurrent phases of the same deformation) yielded so complex a structural pattern can be attributed to the inhomogeneity of the rock units involved, especially to the presence of plastic carbonate rocks interlayered with more rigid metasedimentary granulites and sheets of phacoidal granite gneiss.

Very similar complex structures in magnetite-sulfide bodies in skarn have been described by Magnusson (1940b, pp. 105-110) and others, but space forbids the citing of comparisons here.

The ore is entirely post-folding. Some of the very coarsest magnetite is in fact found with a few inches of crushed, sheared, deformed, and recrystallized phacoidal granite gneiss. Part of the skarn was probably developed by the quartz syenitic magma that yielded what is now phacoidal granite gneiss; however, most of the original limestone was probably deformed as such and subsequently metasomatized slightly in advance of ore deposition by solutions emanating from the younger granites.

A few small normal faults cut the rocks in the mine area, but these faults are not yet known to produce major dislocations of the ore bodies.

Primary structural features

The main structural types of magnetite deposit have been described, and details of special structures and replacement relations have been given in description of the different ore types. The conformable, layered character of much of the ore, and the general regularity of many of the deposits, has been pointed out. All the deposits show local irregularities due to the sporadic effect of replacement; and some show, in addition, local irregularities such as pinching and swelling, or marked thickening and thinning due to folding and extreme flowage of the host rocks. Sköls have developed locally, and boudinage is known to occur in some places. Other special features, already noted, are considered below in more detail.

Local transgressive relations

All types of magnetite deposit in the district exhibit local transgression on several scales. Gentle undulation of mineralized zones within the host rock and local transection of "stratigraphic" contacts by these zones is evidence of transgression on the scale of hundreds or thousands of feet.

Small discontinuities between ore and wall rocks are well shown at several places in the Clifton mine. Five or 6 feet of the footwall contact of ore against pyroxene gneiss and garnet skarn is well exposed in the face of a small cross cut, 430 N, 195 W, off the mining sub below the first level. The contact is a surface of discontinuity, truncating drag folds in the underlying gneiss. The contact strikes N. 77° E. and dips 40° SE., whereas the drag folds trend S. 15° E. and plunge 5° SE. (If these drag folds were in the plane of the contact, their plunge would have to be 40°, for the angle between the strike of the surface and the trend of the folds is 88°.) These relations suggest incongruent flowage between limestone (now represented by ore that has replaced skarn) and more resistant zones of impure calcareous gneisses contained within the original major limestone unit. The discontinuity may have been emphasized (or used again) during late shearing that preceded ore deposition. Similar discontinuities along contacts between what were carbonate rocks and their less plastic adjoining rocks may have been an important means of access for the ore-depositing solutions.

Transgression of host rock by magnetite on the scale of a few inches or feet is indicated by the magnetite veinlets that cut across structural surfaces at small angles, and transgression on a microscopic scale is shown by fabric and replacement relations previously described.

The role of discontinuity and transgression is emphasized here because structural features of this sort are believed to have controlled the replacement by ore. For purposes of prospecting and developing the ore bodies, of course, the more usual and more obvious concordant relations of the bodies should be stressed.

Linear structures and their use

In New Jersey, it has long been known that the trend and plunge of lineations in the wall rocks could be used to predict the course of the elongate ore shoots characteristic of several districts in the state. The lineations, of several types, are all essentially parallel to b (fold axes) in an area where all fold axes have a remarkably uniform plunge (Sims, 1950). Zimmer (1947b) has noted a similar relation at the 81 mine, Lyon Mountain; and the relation is duplicated by many deposits in Sweden, Norway, the Appalachians, and elsewhere. Unfortunately, multiple lineations--that is, lineations of many types _/ with diverse but systematic orientation--are typical of several deposits in the St. Lawrence County district, notably the Clifton mine. One of the major

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_/ Including rodding, fluting, and pencil structure--elements whose orientation is in some districts a highly reliable guide to the course of the ore shoots.

structural problems, still not entirely solved, is the relation between certain sets of lineation and the probable direction of elongation of the ore bodies. The problem is not one of which set controls the elongation; several do. The problem is: how are the sets related, spatially and genetically, and what does that mean in terms of structural control of ore. The writer would urge extreme caution in the use of lineations to predict the course of ore bodies until the individual pattern is worked out for each deposit. (See discussion of "complex deposit".)

Localization of ore

The importance of regional and local structures in localizing ore has already been detailed, with considerable emphasis placed on the role of structural discontinuities. We still do not know why a particular part of a favorable host rock is replaced by ore. Detailed mapping at Clifton has shown that many apparently homogeneous portions of skarn are intensely drag-folded on the scale of inches and several feet, the tight folds and crumples frequently being outlined by fine films of micaceous minerals. Similar and much more obvious minor folds are present in some granite gneiss ores, and low-grade ores of all types are likely to be foliated to some extent. Yet signs of cataclasis of host minerals are usually lacking in thin sections of ore. In contrast, Postel (Postel and Rogers, in preparation) found that many ore specimens from the Clinton County magnetite district showed prominent zones of microgranulation serving as sites for replacement by magnetite. If the host rocks of the St. Lawrence County magnetite deposits were ever subjected to microgranulation, the evidence has largely been obliterated, either because the last deformation of the fabric occurred at temperatures high enough to induce complete recrystallization, or because the mineralization process itself induced a recrystallization of the unreplaced host minerals. Brown (1947) has considered at length a similar problem relating to the sphalerite deposits. Perhaps a study of porosity of the magnetites would be equally rewarding.

Secondary structural features

Shear zones, faults, and joints are secondary structural features affecting the magnetite deposits.

Shear zones a few inches thick have been found locally at the Clifton mine. All the shear zones are small. Most of them strike ENE. and dip very gently southward, but one in the open pit has a steep northward dip. These shears are post-skarn, post-pegmatite, and post-ore. The writer believes they represent late movement induced by the same forces that earlier produced the prominent b_2 rolls and overturned folds that show local evidence of limited overthrusting.

Minor faults at several deposits are inferred from drill cores showing mylonite and chloritized breccia. The extent and character of the faults are entirely unknown; displacement has probably been negligible. At one deposit (Brandy Brook Northwest), an inferred fault has displaced the northwest end of the ore body for an apparent horizontal distance of 200 feet (Leonard, report in preparation). Inasmuch as local post-ore faulting is a rather common feature of other magnetite districts in the Adirondacks, we may expect similar faults to "appear" in the St. Lawrence County district as knowledge of the individual deposits grows.

Joints in the magnetite deposits have not been studied or mapped in detail. Time did not permit the tedious measurement by sun compass of the orientation of joints when data on planar and linear structures held priority. Such a joint study might prove highly desirable, both as a means for inferring orientations of lineations (and ore shoots?) in areas where normally prominent lineations are locally obscure or absent, and as a possible means of relating the ages of fold systems.

Joints and inferred faults have controlled supergene rock alteration at several deposits in the district.

Supergene alteration

Supergene alteration of the magnetite deposits is generally slight, owing to the present rigorous climate and the erosive action of Pleistocene glaciation, which removed most signs of earlier weathering. However, some deposits show a characteristic type of alteration believed to be supergene. _/ Several skarn deposits show local intense alteration, but most of the deposits in granite gneiss are unaffected or merely show slight leaching and clay-mineral alteration in the rotten upper zone of bedrock covered by a thick mantle of glacial debris.

_/ This belief is based mainly on the similarity of the alteration products with those developed elsewhere in the district by supergene agencies. Until more data are at hand, the evidence from drill cores and dump specimens alone is inconclusive though strongly suggestive.

Alteration is prominent in certain rocks at the west end of the Brandy Brook Northwest deposit. The central and eastern parts of the belt show similar alteration, weakly developed, in a few places. Skarn and limestone are by far the most frequently and most extensively altered rocks; alaskite, mica sköls, and quartzose gneisses have been affected locally. The principal alteration product of skarn is "greenstone"--a soft, blue-gray or greenish aggregate of extremely fine chlorite or serpentine, or perhaps both. Usually accompanying the "greenstone" are specks or stained areas of red hematite, groups of thin calcite veinlets, or veinlets of calcite, chlorite, pyrite, and hematite. Veinlets are sometimes found where "greenstone" is absent. Some "greenstone" seems to have formed directly from limestone adjacent to altered skarn, and hematite within the altered limestone is black and granular. The rock alteration is not limited to a particular part of the skarn, though it is especially prominent at and near the top and bottom of the skarn zone.

At several places, the alaskite sheet is chloritized, veined by calcite (with or without pyrite), stained with hematite, or altered to clay minerals. Usually these features are combined. Where the alaskite is so affected, thin zones of sheared rock or mylonite may be visible.

The inferred fault zone _/ is chloritized and hematite-stained, as well as brecciated. Considerable carbonate may once have been present, for the zone is porous and had to be grouted repeatedly during drilling.

_/ See preceding section on secondary structural features.

The writer believes that the alteration described above is almost wholly supergene. Owing to their greater susceptibility to solution and alteration, skarn and limestone have been affected most. Sköls--now completely altered except for relics of mica, rarely seen--may have been channels especially favorable for downward movement of solutions. (See Eskola, 1914, p. 259). The inferred fault and minor zones of mylonite have also served as channelways, perhaps assisted by joints that are scarcely visible in the unweathered core. One cannot exclude the possibility that a small part of the alteration is hypogene, for calcite-fluorite-pyrite veinlets (in part with chloritized selvages) are found locally in the deposit. These veinlets are thought to be hypogene, but the veinlets and their chloritic wall-rock alteration are a minor feature indeed.

The dumps at the Jayville deposit show magnetite-bearing biotite sköls highly altered to earthy chlorite and cut by networks of thin calcite-hematite-pyrite-jasper veinlets. The same kind of alteration has been found in drill cores from that deposit.

The greatest known vertical depth of hematite-chlorite alteration in the magnetite deposits is 450 feet (Hole B9, Brandy Brook Northwest). It must be noted that this intense alteration is always localized along joints, faults, sköls, or marble layers, whereas the more pervasive "rotting", leaching, limonitization, and clay-mineral alteration found at the sub-outcrop of some deposits extends downward into the bedrock for only 40 or 50 feet. (However, the overburden itself may be 50 or 100 feet thick.) Only the better protected zones that extended deep into bedrock along faults, etc., have been preserved after glaciation; weathered surficial material was almost entirely scraped away.

Similar supergene alteration to chlorite, hematite, and secondary zinc minerals has been described by Brown (1936) from the Balmat zinc mine. Supergene alteration at Balmat extends to depths of 1,000 feet and more (Brown, 1936, p. 340). A neighboring belt of earthy hematite deposits (see fig. 2) derived from the weathering of pyritic gneisses was mined for iron ore during the last century (Buddington, 1934, pp. 194-202). The age of this type of alteration in the northwest Adirondacks is not certain. Buddington thinks that most of it took place in post-Archaean, pre-Potsdam time, while Brown believes some of the alteration may be much younger. There is no direct way of dating the alteration that accompanies the magnetite deposits.

Secondary hematite derived from weathering yields one type of "soft ore" associated with magnetite deposits of the Central Swedish district (Geijer and Magnusson, 1926). Hematite-chlorite alteration along a fault zone cutting a magnetite deposit at the Andover mine, Andover, New Jersey, has been described by Sims and Leonard (report in preparation). Similar alteration is known to occur in the Arnold Hill group of mines, Clinton County, New York (Postel and Rogers, in preparation).

Origin of the magnetite deposits

Metasomatism effected by younger granites

The writer believes that the magnetite deposits of the district are high-temperature replacement deposits produced by emanations from younger granite magma. The deposits represent one aspect of a process that, under slightly different conditions, yielded the neighboring pyritic sphalerite deposits, pyrite and pyrrhotite deposits, and perhaps also tremolite-talc deposits. The evidence on which such an interpretation can be based has already been presented. A brief summary of the more important geologic events is given below, together with a discussion of the origin of the ores. All thoughtful geologists will realize that this is but one reading of the record, that many other interpretations have merits, and that there is still no sure answer for some of the most pressing geologic problems found in this and related districts.

Summary of events

The first major deformation of the Adirondacks took place after the consolidation of the quartz syenitic rocks and their intrusion by scattered dikes of hypersthene metadiabase. The conclusion of that deformation marks the end of one great chapter of Adirondack geology. Subsequently, younger granite magma was intruded into the metasediments and partly metamorphosed older igneous rocks. This magma, which consolidated chiefly as hornblende-microperthite granite, differentiated to give a volatile-enriched phase that worked upward and outward, crystallizing as alaskite, in part as "roof rock", in part as satellitic sheets and phacolithic bodies in the metasediments. Probably the same fundamental magma also yielded a high-potash, volatile-enriched phase that intruded the metasediments as thin sheets, reacted with the country rock, and in places metasomatized it extensively, yielding heterogeneous microcline granite gneiss. _/ In places, or at times, the younger granite may have consolidated under directed pressure; part of it is deformed, part of it is not. Generally, deformation of the younger granite occurred after the rock was completely solid.

An advance wave of metasomatism by volatile emanations rich in F, (Cl), (P), OH, and Si preceded the intrusion of some of the granite, for in places contact zones of skarn are cut by alaskite or granite pegmatite (often syenitic in this relation), or replaced by microcline granite. The character of the skarn as it existed in this stage is entirely unknown. The silicates that composed it may have been very poor in iron, or they may have been slightly ferriferous. Moreover, the duration of the period of metasomatism is uncertain; though skarn is locally cut by granite, and some granite areas have been metamorphosed locally, none of the skarn seems to have been appreciably deformed or reconstituted. These facts suggest (1) that the metasomatism that yielded skarn began before the intrusion of the granites and continued (or was renewed) after most of the granite had largely consolidated;

_/ The possibility that the microcline granite gneiss is relatively old must also be borne in mind.

or (2) that the rare cutting of skarn by granite is a purely local phenomenon, not entirely in phase with the main consolidation of the granite masses; or (3) that the skarns have indeed been deformed, recrystallized, or even reconstituted without leaving any recognizable sign of these happenings. At the moment, the writer favors the first possibility.

Once the skarns had been developed (and partly enriched in iron) and the heterogeneous microcline granite gneiss had formed, both rocks were--in especially favorable sites--modified by introduction of quartz, potash feldspar, fluorite, barite, and (locally) scapolite or spessartite. This modification was highly localized and quantitatively insignificant, as it affected just two major deposits of magnetite and is found elsewhere only in traces. Extensive modification of skarn in this characteristic way took place only where the skarn was enclosed in microcline granite gneiss. _/ Initially, small masses and veinlets of granite pegmatite, syenite pegmatite, and quartz were introduced. From these sites, intense local replacement of skarn took place, giving modified skarns of the Trembley Mountain type. At one place (Parish deposit), the microcline granite gneiss itself was modified in a very similar way. The type of modification described above represents a continuation (or

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_/ Skarn close to alaskite and far removed from microcline granite gneiss has been modified locally, especially to scapolite-fluorite-garnet-pyroxene skarn. Magnetite is not known to occur in quantity in this type of skarn associated with alaskite.

renewal) of the same process that developed all the microcline granite gneisses of the district, but the small number of magnetite deposits in modified rocks must again be emphasized.

At sites favorable because of their structure and their proximity to the supply of metasomatizing solutions, the skarns were subjected to the progressive introduction of more iron. Initially, iron was substituted within the silicate lattice of diopside, producing salites and ferrosalites. Where the appropriate concentrations of volatiles existed, pyroxenes were locally replaced by amphiboles or by micas. Local access of Fe^{+3} (perhaps merely local oxidation of Fe^{+2}) permitted the development of andraditic skarn. At some appropriate but unknown pressure, temperature, and degree of concentration, the silicates could no longer accommodate all the Fe within their lattices; at that stage, magnetite was precipitated, closely followed by a series of simple sulfides, minor in quantity.

Very similar processes, locally affecting biotitic or sillimanitic microcline granite gneiss, resulted in the formation of magnetite deposits in those rocks. Perhaps the local development of hematite in the granite gneiss ores (representing an increased oxidation state of the Fe) is analogous to the local development of andradite in the skarns, where primary hematite is lacking. A comparison of the processes at work in skarns and granite gneisses is given below.

Progressive decrease in the concentration of Fe, decreasing temperature, and a change in the character of the metasomatizing solutions towards dilute water-rich fluid led to local alteration of ore and wall rocks. This alteration was generally weak in the skarn ores and marked in granite gneiss ores and modified ores of both types. Its most conspicuous result was the partial leaching of ores and immediate wall rocks, followed by the deposition of hydrous silicates, calcite, and sporadic base-metal sulfides.

At some later date (possibly in late pre-Cambrian time), a few of the deposits were faulted. Still later, faults, joints, and permeable rock units conveyed surface waters downward, yielding local masses of earthy hematite, dull-green chlorite, and minor calcite, jasper, and pyrite.

Comparison of skarn ores and granite gneiss ores

The skarn ores and granite gneiss ores, outwardly so dissimilar, have many features in common. Both are relatively nontitaniferous ores with magnetite as the principal ore mineral. They occur together not only in the same district but even in the same deposit, though in the latter case they do not usually carry equal concentrations of magnetite. Their temporal relations are the same: both are post-pegmatite and post-deformation. Their absolute ages are roughly the same: 1,300 million years for Jayville (skarn type) and 1,200 million years for Benson Mines (granite gneiss type). [Preliminary determinations of the age by P. M. Hurley; letter to A. F. Buddington, 1942. These ages agree closely with Pb-Th age determinations given by Shaub (1940) and Marble (1943) for Adirondack pegmatite minerals.] Both types of deposit are conformable replacements with local transgressive relations.

Magnetite from both types of deposit shows the same exsolution phenomena. Both types have essentially the same suite of accompanying sulfides, arranged in the same paragenetic sequence. The manner in which metallics replace the host rock is the same in both types of deposit.

If one considers a long enough span in the history of both types of deposit, one finds that the same chemical substances have been introduced into both. The absolute quantities and their proportions were very different, and the stage at which a given element appeared may have been different. Certainly the manner in which the elements were concentrated was often very different. These comments are speculative, and much more work needs to be done before even the preliminary results can be presented in orderly fashion. Table 5 below is entirely a qualitative expression. Actually, to reach a fair quantitative expression would be extremely difficult, principally because metasomatism affected two completely different "source rocks" in very different concentrations.

The behavior of K and Al is especially interesting. In the skarns, these elements entered more and more into the lattice of the pyroxenes and their successors, amphiboles and micas. (See discussion of skarns.) In the granite gneisses, however, K and Al developed preponderantly as potash feldspar. Ba, relatively insignificant in unmodified skarns, concentrated with K and Al in the granite gneiss ores and in modified ores of both types.

Relation of magnetite deposits to younger granites

The space relations of magnetite deposits to younger granites has been described above. (See section on general geology of the district.)

Table 5. Elements introduced in skarn ores and granite gneiss ores. (Comparison is only qualitative.)

Element or ion	Skarn ores	Granite gneiss ores
Si	Pyroxene, garnet, amphibole, mica; scapolite and feldspar in local modified skarn. (Initial introduction; subsequent partial removal.)	Quartz, potash feldspar
Ti	Dark skarn minerals; exsolution or solid solution in magnetite; sphene in local modified skarn	Biotite; exsolution in iron oxides; solid solution in hematite?
Al	Dark skarn silicates; feldspar in local modified skarn	Potash feldspar, sillimanite, biotite
Fe^{+3}	Magnetite; dark skarn silicates, especially andradite	Iron oxides, especially hematite; biotite
Fe^{+2}	Magnetite; dark skarn silicates	Iron oxides; biotite
Mn	Dark skarn silicates; magnetite?	Spessartitic almandite; iron oxides?
Mg	Initially introduced?	LARGELY REMOVED
Ca	LARGELY REMOVED	
Na	Scapolite in local modified skarn	Potash feldspar. LARGELY REMOVED
Ba	Micas (very slight)	Potash feldspar, barite
K	Dark skarn silicates, esp. ferrohastingsite and micas; potash feldspar in local modified skarns	Potash feldspar, biotite
OH	Amphiboles, micas. Late hydrothermal minerals (little)	Biotite, sericite. Late hydrothermal minerals (considerable)
CO_2	LARGELY REMOVED	
P	Apatite	Apatite
S	Sulfides; ?scapolite in local modified skarn	Sulfides, barite
F	Fluorite, amphiboles, mica, apatite	Apatite, biotite, fluorite
Cl	Mainly (?) as local scapolite	
B	Vonsenite, datolite (both local)	Tourmaline (local)
Ce	Allanite; monazite (?)	Allanite; monazite (?)

Inasmuch as concentrations of magnetite have not been found in hornblende-microperthite granite far removed from alaskite or microcline granite gneiss, it seems reasonable to rule out hornblende-microperthite granite as an immediate source of iron-rich emanations. However, the relation of unmodified deposits to alaskite and microcline granite gneiss is, in some places, equivocal. The writer believes that most skarn deposits (and perhaps all of them) are genetically related to alaskite. However, the granite gneiss deposits are related to their host, though one cannot rule out the possibility that alaskite (always occurring near them) was also essential for the occurrence of ore in the granite gneiss. The restriction of modified deposits of both types to close association with microcline granite gneiss strongly suggests a genetic connection with that rock, though again alaskite cannot be entirely excluded.

The writer believes that both alaskite and microcline granite gneiss were capable of contributing iron-rich emanations. Both appear to be the product of volatile-enriched magmas developed by differentiation of the magma that yielded hornblende-microperthite granite. It seems as reasonable to suppose that both volatile-enriched, siliceous, potash-rich residua were locally capable of emitting iron-rich emanations as to suppose that only one of them could do so. It is at present quite beyond the capabilities of our knowledge to prove that any granitic magma did yield magnetite deposits of the type found in this district. However, the development of ferrohastingsite (high in Fe, F, OH) in the microperthite granites, locally quartz-rich; the relative enrichment in Fe shown by analyzed alaskites and microcline granite gneisses of the district; the presence of biotite and accessory fluorite in the alaskites; the field relations of the alaskites and microcline granite gneisses; the concentration of Fe, Si, F, OH, and other elements in the magnetite deposits; and the areal distribution of the magnetite deposits themselves: all these are, to the writer, strongly suggestive that a granitic magma concentrated Fe, (local Mn), Si, alkalies, volatiles, and other constituents that escaped into structurally favorable sites and yielded ore deposits.

Perhaps the chief distinction between the material that crystallized as alaskite and the material that formed microcline granite gneiss was the relative concentration of individual volatile constituents (in particular, more F in alaskite, and more OH in microcline granite?), the order in which these escaped, and the extent to which the emanations were modified by what they came in contact with. So far as we now know, Ba and Mn were concentrated with the emanations given off by the material that yielded microcline granite gneiss deposits.

It is possible, of course, that the emanations that yielded magnetite deposits came from a deep source related to the magma that yielded all the younger granitic rocks. If this were so, the remarkable space relations of ores with alaskite or microcline granite gneiss would be even more difficult to explain.

Character of the metasomatizing solutions

Unlike some areas containing skarn deposits, _/ the St. Lawrence

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_/ In this regard, one sometimes wonders how much pale fluorite has been overlooked in other districts. Much of the fluorite in the St. Lawrence County skarns, sköls, and ores of all types is colorless, pale-green, or pinkish, and rather fine grained. Megascopically, the colorless, granular stuff is likely to be mistaken for quartz, the greenish fluorite for apatite, and the pinkish fluorite for microcline. Purple fluorite is not rare, but the other varieties are much more common.

County district has preserved ample evidence of the former existence of "mineralizers". Signs of these mineralizers are preserved at individual magnetite deposits of all types, as well as in the granitic rocks that are regarded as sources of the metasomatizing solutions. As is commonly the case, the quantity of mineralizers preserved at a given deposit is not always proportional to the quantity of Fe, Si and other elements that apparently were introduced. However, this does not mean that the mineralizers were always deficient; at several deposits, the preserved mineralizers now locked up in OH-, F-, Cl-, P-, SO_3-, and B-bearing minerals are more than adequate to have transported all the heavier introduced elements with which they are associated. Also, it is quite impossible to estimate the quantity of mineralizers (and heavier elements) that may have swept out of the granitic rocks and through the host rocks without being fixed. If, for example, we assume that as much F escaped from alaskite as is now fixed in the form of accessory fluorite (giving ca. 0.2 percent F in typical analyzed alaskites), and if we make very liberal assumptions regarding the total quantity of iron oxides in the district, we find that there is roughly two to three times as much alaskite present as would have been required to transport all the Fe in the form of fluorides alone. This does not mean that F was the only mineralizer or even necessarily the chief one, or that alaskite was the exclusive source of all elements introduced. It is merely a very crude yardstick to measure the quantitative significance of <u>one</u> mineralizer that has left its mark upon the district.

What the character and composition of the metasomatizing solutions were, we can only guess. The mineral associations and their sequence of development suggest that the solutions were initially "pneumatolytic", passing with decreasing temperature and changing concentration into dilute hydrothermal solutions. At first, the concentration of F was probably high; with it were Cl, P, SO_3, OH, and (locally) much B. The greater quantity of F-bearing minerals in the skarns suggests that the emanations (chiefly from alaskite) that formed the skarns and their ores were richer in F than emanations that yielded granite gneiss ores. The latter probably were higher in OH, P, and SO_3.

Perhaps at the magnetite-depositing stage, the solutions had already become substantially hydrous. Iron-poor phlogopites from skarns are rich in F and relatively poor in OH, whereas Fe-rich biotites intimately associated with magnetite are OH-rich and relatively poor in F.

The various mechanisms for transporting Fe as halides, hydroxide hydrosols, oxides, and the like, are discussed and illustrated by Eskola (1939, pp. 372, 384), Geijer (1925, p. 687), Shand (1947, p. 635) [but see Holser (1950, p. 1080)], and many others. Restating the equations and arguments serves no useful purpose here.

The presence of primary hematite in the granite gneiss ores requires oxidation of Fe relative to that required for the development of magnetite. Primary hematite in magnetite deposits of the district is confined exclusively to K-rich rocks--the granite gneiss ores and their modified equivalent. Moreover, Buddington (personal communication) finds that the disseminated, accessory iron oxide of all the microcline granite gneisses of the district is hematite, not magnetite. The writer believes that the development of hematite in these rocks is chiefly a consequence of their alkaline environment. Mason (1943a, p. 128) notes that the oxidation of Fe^{+2} to Fe^{+3} takes place much more readily in alkaline than in neutral or acid solutions, and it seems likely that the extreme concentration of alkalis in the forms of K, assisted by OH, was responsible for most or all of this oxidation.

Several otherwise puzzling features of the magnetite deposits are explicable in this light. Skarn ores, which lack primary hematite, may have come from solutions too acid (i.e., too high in F) to oxidize all the Fe required for hematite at the stage favorable for the precipitation of Fe oxides. (In some districts, however, oxidation of Fe by halides in the presence of water vapor has been inferred as the mechanism for developing pneumatolytic specularite deposits.) Very modest and local Fe oxidation in skarns, reflected in sporadic martite, took place well after the magnetite had formed; it may have resulted from increasing alkalinity of the solutions in the waning stages of mineralization.

The absence of primary hematite from the Trembley Mountain deposit (skarn modified by K-rich emanations before magnetite deposition), where one may infer an alkaline character for the solutions or for the host rock, must have required action by some reducing agent. If any CO were liberated during metasomatism of the enclosing marbles, and were still effective at the time of magnetite mineralization, that might have served as the reducing agent.

The prevalence of an alkaline environment for the granite gneiss ores makes one wonder why there are any magnetite deposits in microcline granite gneiss, particularly when these rocks typically have hematite as a "regional" accessory mineral. Perhaps the controlling factor was simply the concentration of Fe^{+2}. Or the P (represented by ubiquitous apatite) and F (in apatite, micas, and sporadic fluorite) may have been of an appropriate concentration at an appropriate stage to prevent complete oxidation of the Fe^{+2}.

For these and many other questions, there is now no certain answer. All the writer can offer is a few suggestions.

Temperature of formation of magnetite deposits

Little evidence is at hand for estimating the temperature at which the magnetite formed in the ore deposits. The suite of early-formed, associated minerals--both nonmetallic and metallic--is typical of high-temperature replacement deposits. Magnetite and primary hematite from the ore deposits characteristically show less exsolved ilmenite than one finds in magnetite and primary hematite disseminated in the country rocks. This suggests that the iron oxides in the ore deposits formed at somewhat lower temperature than accessory iron oxides in country rock.

Exsolution features in the sulfides give minimum temperatures for crystallization of bornite and sphalerite. Bornite from the Parish deposit (modified granite gneiss ore) has exsolution bodies of chalcopyrite. Schwartz (1931, p. 200) gives 475° C. as the minimum temperature of formation for such intergrowths. Some sphalerite I from Trembley Mountain (modified skarn ore) and Jarvis Bridge (granite gneiss ore) shows exsolution bodies of chalcopyrite. Buerger (1934, p. 530) gives 350-400° C. as a range for the minimum temperature of formation of such intergrowths. The temperature of formation of sphalerite from other types of deposit is unknown. Sphalerite is very sparse in skarn ores and in the Parish deposit; so far, none of the minute grains has been found to contain exsolved chalcopyrite. When so few grains of sphalerite are available for inspection, however, the apparent lack of exsolution phenomena has no significance.

The minimum temperatures indicated by intergrowths in bornite and sphalerite agree very well with a hypothermal temperature (400° C. or more) inferred from the parageneses. As the sulfides are distinctly younger than magnetite, it seems safe to infer a temperature of 500° C. or higher (at least more than 475° C.) for the initial deposition of magnetite.

Sphalerite of stage II never shows exsolved chalcopyrite, indicating that this generation of sphalerite formed at a temperature below 350-400° C. The association of sphalerite II with the zeolites and calcite actually suggests a considerably lower temperature for this waning stage of mineralization.

The depth at which the deposits formed is entirely unknown. It must have been considerable. Depths of several miles have been suggested for the initial stage of mineralization for deposits having similar characteristics.

Other hypotheses

Igneous source of the iron, non-hypothermal emplacement

A magmatic or epimagmatic source of Fe has been postulated for some nontitaniferous magnetite deposits, as well as for titaniferous ones. Vogt (1907, 1910), for example, regarded a number of Norwegian magnetite deposits as magmatic segregations, and at times some of the Adirondack deposits have been so regarded (Kemp, 1910). Geijer (1931) has postulated an epimagmatic origin for the Kiruna deposits and the apatite-rich (Grängesberg) type of nontitaniferous magnetite found locally in Central Sweden. (He included, in his stimulating discussion of the "Kiruna type", the Mineville-Port Henry deposits of the eastern Adirondacks.) The extensive evidence of replacement shown by all types of nontitaniferous magnetite deposits in the St. Lawrence County district clearly sets them apart from magmatic segregations and epimagmatic deposits, and such modes of origin cannot seriously be considered for these ores.

For certain skarn deposits of the Central Swedish district, Magnusson (1925, 1930, 1940a, 1940b, and others) has inferred a complex origin. The skarns are closely associated with the leptite formation, now regarded as supracrustal volcanic rocks with interbedded metasediments. This association of ores with the leptites, together with the occurrence of some fine-grained skarn magnetite suggestive of colloidal deposition, led Magnusson to believe that the skarn ores were initially deposited at low temperature from solutions derived from the volcanic members of the leptite formation. The ores were subsequently metamorphosed at high temperature and pressure and in places re-mineralized, so that they now appear as high-temperature deposits. Reaction metamorphism has played a dominant role in giving the deposits their present character. Nevertheless (and this is Leonard's interjection), the initial source was igneous--volcanic. Geijer (1923, 1927, 1936, 1939) has inthe past argued convincingly that the skarn ores are essentially pyrometasomatic deposits formed by volatile-rich emanations stemming from granites that intrude the leptite formation. In his view, reaction metamorphism has played a minor part in developing these ore deposits. The writer is very hestitant to comment on the current status of the problem, for the most recent and thorough exposition by both men is entirely in Swedish (Geijer and Magnusson, 1944), a language with which the writer is still not quite at home. As there is no evidence of the former existence of supracrustal volcanic rocks inthe St. Lawrence County district, it seems unnecessary to consider a volcanic source for the ore deposits of the district.

Metasedimentary origin

Derivation of the Adirondack magnetites by intense metamorphism of sedimentary iron ores has been advocated by some men, recently and publicly by Gillies (1950, pp. 85-86), R. J. Linney (1943, p. 488), and Walker (1943, p. 520--for Parish deposit), and privately by many others.

(See also Nason, 1922; and discussions by Miller, 1922, and Newland, 1923.) The argument applies particularly to the foliated granite gneiss ores. It is especially appealing because of the wide distribution in the late pre-Cambrian of iron formations of the Lake Superior type.

For convenience, we may divide the sedimentary iron ores into two broad types, nonpyritic and pyritic, and consider each separately.

The nonpyritic ores are of three main types: banded siliceous ores, banded carbonate ores (both represented in the low-grade "iron formations" of the late pre-Cambrian), and generally oölitic ores (chiefly of Ordovician, Silurian, and Jurassic age).

Two main lines of evidence bear against a metasedimentary origin for the magnetite deposits of the St. Lawrence County district. First is the complete absence from the entire northwest Adirondacks of anything resembling the known sedimentary iron ores. Second is the contrast between the chemical composition, paragenetic sequence, distribution, and fabric of the St. Lawrence County ores and the highly metamorphosed equivalent of demonstrably sedimentary iron ores.

Nowhere in the vast and well-exposed area of the Grenville Lowlands is there anything resembling a sedimentary iron formation. Metasedimentary rocks constitute about 75 percent of this area. All types represented there also occur in the St. Lawrence County magnetite district, within the dominantly igneous complex. It does not seem reasonable that highly metamorphosed iron formations should appear as magnetite deposits within the dominantly igneous complex when no sign of similar magnetite deposits or sedimentary iron formations has been found in the

Lowlands, where better exposures and a far greater proportion of originally sedimentary material are available for study. It is true that a belt of hematite deposits, in part carrying specularite, is found in the Lowlands and foothills close to the dominantly igneous massif. However, these deposits are supergene, derived from deep weathering of sulfide-bearing rocks (Buddington, 1934, pp. 194-202; Brown, 1936b, pp. 331-354).

Sedimentary iron ores that have been regionally metamorphosed to a slight or moderate degree are well exemplified by the Goulais River iron range, district of Algoma, Ontario (Moore and Armstrong, 1948, pp. 7-36). In an area where metamorphosed volcanic rocks predominate, a banded-silica iron formation of Keewatin (pre-Cambrian) age has been intensely folded, recrystallized, and partly reconstituted to a fine, laminated quartz-magnetite rock with a little actinolite, hornblende, or grünerite. A few layers of red jasper have been preserved. The iron formation is cut by dikes of diabase and by dikes and stocks of granite and porphyry.

An excellent petrologic study of banded iron formations metamorphosed under low-grade and high-grade conditions in Western Australia has been made by Miles (1946).

Few geologists would expect to find iron formations preserving their customary low-grade metamorphic and sedimentary features in an area such as the St. Lawrence County district, where granitic rocks predominate and regional dynamothermal metamorphism of rather high grade has affected much of the country rock. However, several areas of similar old pre-Cambrian rocks (or highly metamorphosed early Paleozoic rocks) contain magnetite or hematite deposits whose sedimentary origin seems well established. Notable among these are the Sydvaranger deposits, northern Norway (Geijer, 1911); deposits at the Rana mines, Dunderlandsdal, northern Norway (Bugge, 1948); deposits in the Båfjell syncline, Ofoten, northern Norway (Foslie, 1949); and deposits of the Striberg type, central Sweden (Hjelmqvist, 1942).

The Sydvaranger, Striberg, and parts of the Ofoten ores are very similar. All are low-grade quartz-banded ores found in highly-folded areas where granitic rocks are abundant and the degree of metamorphism is rather uniformly high. The Striberg ores are of old pre-Cambrian age. The current interpretation of the age of the Sydvaranger ores is unknown to this writer. The character and composition of the ores is strongly suggestive of a derivation from siliceous iron formations of the Lake Superior types. (Cf. analyses of taconites given by Gruner, 1946.)

The ore at Sydvaranger (Geijer, 1911) is a finely-banded or stratified quartz-magnetite rock with a little green hornblende and (locally) red garnet and "diopside". The magnetite layers are thin and sharp, averaging ca. 2 mm thick. They are remarkably continuous along the strike; layers 0.5 mm thick can be followed for 10 meters. Contacts between ore and barren, unstratified wall rock are sharp and concordant. The wall rocks are generally biotite-quartz-plagioclase gneiss ("leptite") and microcline granite. Dikes of aplite, pegmatite, and diabase cut the ore. The ores were earlier interpreted as magmatic by Vogt and by Sjögren. Geijer interprets them as metamorphosed silica-rich sedimentary ores.

The main ore type at Striberg (Hjelmqvist, 1942) is quartz-banded hematite ore with local andradite, epidote, apatite, and allanite. The ore is very regularly banded in layers 1-10 mm thick. Some magnetite is present more or less throughout; it develops from hematite near quartz and pegmatite veins, near sköls, in pinches, and at the borders of ore bodies. Locally, scaly hematite ore (Åsboberg type) and compact, granular hematite ore are present. In addition, subordinate skarn deposits of magnetite and sulfides are found in the district, locally in close association with the quartz-banded hematite ores. The surrounding terrain is made up of leptites (predominantly sodic), with subordinate limestone and amphibolite. "Urgranites" are not exposed. Hjelmqvist believes that the quartz-banded ores are metamorphosed silica-rich sedimentary ores. Local skarn ores have developed by reaction metamorphism, but some larger skarn deposits appear to have formed by metasomatism of limestone.

Metamorphosed sedimentary ores of marine origin and presumed Cambro-Silurian age are found at Ofoten (Foslie, 1949), where they occur in rocks metamorphosed in the high-temperature part of the epidote-amphibolite facies. The ores are of two types. The Håfjell type, partly rich in Mn, is a mixture of hematite and magnetite with manganocalcite, and (locally) grünerite and spessartite. The silicates have formed as local reaction skarns. Magnetite is always younger than the hematite. The Sjåfjell type contains magnetite but no hematite; ores are quartz-banded, ankerite-banded, or non-banded and grünerite-rich, with local garnet.

Somewhat similar ores, presumably of Cambrian or Ordovician age, are found at Rana, Dunderlandsdal (Bugge, 1948). These metasedimentary ores are nonmanganiferous and mineralogically rather more complex. Three types of ore are recognized: magnetite, hematite, and mixed magnetite-hematite. The iron oxides are usually accompanied by quartz, calcite, dark silicates, and (rarely) a little acid plagioclase. Chemically the ores are much more similar to ores of the Clinton type than to taconites. This seems entirely reasonable, in view of the early Paleozoic age inferred for the Rana ores.

In all the occurrences summarized above, low-grade ores of demonstrably sedimentary affinities are found in areas of rather high-grade metamorphic rocks, often in close association with K-rich or Na-rich rocks of granitic composition and variable origin. The ores are highly folded and their constituents have been partly reconstituted during metamorphism at elevated temperature and great depth, yet in these terrains of alkalic rocks the ores are uniformly low or lacking in alkalis. Their chemical composition is in no way comparable to that of ores from the St. Lawrence County district, though local quartzose ores (derived by modification of skarn or granite gneiss) are found in very subordinate amounts in the district. Where magnetite and hematite are found in the Swedish and Norwegian ores described above, magnetite is distinctly younger. Sulfides, if present at all, are commonly of later and unrelated introduction. Most of the ores are prominently, very regularly, and very thinly layered. Contacts with wall rocks are generally sharp, the iron oxides show no evidence of having replaced country rock, and the ore bodies do not transgress the structure of their host or wall rocks. The deposits give every evidence of being the expectable metamorphic equivalent of sedimentary iron ores, but only by the severest straining of analogy can these metasedimentary ores be compared to granite gneiss ores of the St. Lawrence County district.

The Ofoten ores are significant in still another way. Both ankerite-magnetite ores and manganocalcite-hematite-magnetite ores have preserved much of their primary carbonate. This carbonate has, of course, been recrystallized, but only a limited part of it has been reconstituted by reaction metamorphism even under the temperatures and pressures prevailing in the upper part of the epidote-amphibolite facies. These facts suggest to the writer that primary iron carbonate cannot have been the source of skarn ores in the St. Lawrence County district, where conditions of temperature and pressure were not a great deal higher than at Ofoten. One might argue, however, that a mixed iron oxide-iron carbonate sediment had been extensively metasomatised by K- and Si-rich emanations to yield the granite gneiss ores of the St. Lawrence County district. Such an argument would of necessity overlook the nonexistence of relics of the supposed sedimentary ore, the contemporaneity of magnetite and hematite in the granite gneiss ores, the present fabric of the ores, their transgressive relation, and the relation of the deposits to their regional setting.

In times past, a sedimentary origin has been postulated for the pyrite and pyrrhotite deposits of the Grenville Lowlands. A brief review of earlier ideas is given by Buddington (1934, p. 214), who concurred with Smyth (1912) that the deposits were hypogene. Much chlorite was introduced with the sulfides as replacements along sheared zones in the migmatitic biotite-quartz-feldspar gneiss. Some geologists, viewing the iron sulfide deposits as metasedimentary, might regard the granite gneiss ores of the St. Lawrence County district as the more highly metamorphosed, oxidized equivalent of the iron sulfide deposits. It is true that both types have roughly the same Fe content, but this might as easily be used as an argument that hypogene Fe-bearing solutions of roughly equivalent Fe content, coming from a common source, affected rocks in two different environments. Regardless of their origin, the iron sulfide deposits can hardly be viewed as the less metamorphosed equivalent of the granite gneiss ores; the iron sulfide deposits themselves are younger than the latest granite pegmatites of the district, and there is no evidence of subsequent metamorphism--particularly of a kind adequate to convert the iron sulfides to magnetite and "primary" hematite.

Transfer of material from the country rock

Local transfer of material

Miller (1919, 1921, 1922, 1926) developed the hypothesis that the Lyon Mountain magnetite deposits originated from the action of intrusive granite on older metagabbro. Highly fluid residua of the granite, similar to the material that yielded quartz and pegmatite veins, removed Fe from the gabbros that the granite intruded and incorporated. Some of the Fe came from disseminated magnetite in the gabbro, but most of the Fe came from the conversion of hypersthene and hornblende of the gabbro into slightly less ferriferous clinopyroxene. The Ti originally present in the gabbro did not enter the new-formed magnetite; it formed accessory sphene. Newland (1920) pointed out the quantitative inadequacy of the process envisioned by Miller. To Newland's argument might be added the statement that much of the rock termed "gabbro" by Miller is pyroxene skarn and related metasedimentary rocks. Miller's hypothesis is not applicable to the St. Lawrence County district, for amphibolite, metagabbro, and their granitized equivalent--hornblende-microcline granite gneiss--are found in negigible amounts near the ore bodies. Moreover, the pyroxene of the accompanying Grenville rocks has suffered an introduction of Fe, not a removal of that element, near the magnetite deposits. Hence pyroxene from the metasediments can scarcely have served as a source of Fe for the development of magnetite.

Reaction metamorphism and metamorphic hybridism

Over the years, Magnusson (1925, 1929, 1930, 1940a, 1940b) has developed the concept of reaction metamorphism in its relation to the origin of Swedish skarn ores. He believes that the skarn deposits of magnetite originally formed as low-temperature colloidal replacements of limestone by very fine-grained iron oxides. The replacing agents emanated from the volcanic members of the leptite formation, with which the skarn ores are closely associated. The manganese present in some ores _/ belongs to this same period of mineralization. Quartz-banded

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_/ The manganese in Swedish ores is chiefly in knebelite, an olivine; but it is also found in pyroxene, amphiboles, and (in some deposits) garnets.

iron ores accompany some of the skarn ores. The present character of the deposits is due mainly to regional metamorphism of later date. During this metamorphism, local (?) transfer of material took place among leptites, limestones, and early-formed iron ores, developing high-temperature mineral assemblages in skarns formed by reaction (reconstitution). Subsequently, Mg-F metasomatism has affected the skarns locally, developing tremolite, diopside, and phlogopite from earlier Fe-rich silicates. Sulfide mineralization is related to the period of Mg-metasomatism and intrusion of "urgranites" (old granites), rather than to the early stage of magnetite mineralization associated with leptites. The area has also been intruded by younger granites (and other rocks); pegmatites derived from the younger granites have in some places modified the older skarns and introduced wolframite, scheelite, and a new generation of sulfides. Magnusson himself stresses the importance of reaction metamorphism in developing the skarn ores. Geijer (1923, 1927, 1931, 1936, 1939) has maintained a very different view. According to him the skarn, iron ores, and sulfides are closely-related pyrometasomatic deposits produced by emanations from the "urgranites".

Small bodies of magnetite with a little sulfides are found in dolomite, tremolite skarn, and actinolite-chlorite schist at the contact between dolomite and phyllite in the Karelian rocks of North Bothnia, Finland. Mikkola (1947) interprets the magnetite and skarn as the result of metamorphic reaction between phyllite and dolomite. The absence of minerals suggestive of pyrometasomatic or hydrothermal processes, the absence of nearby intrusives, and the chemical character of the ore rule out a hypogene source, according to Mikkola. He believes the phyllites, locally iron-rich, acquired their iron by volcanic processes acting during the deposition of the phyllites. This iron was later concentrated in the dolomite-phyllite contact zone, possibly (?) in connection with the intrusion of post-Karelian granites.

Some laminae of dark silicates in metasedimentary pyroxene gneisses and granulites of the St. Lawrence County district may have formed by reaction metamorphism and metamorphic differentiation. Some of the quartz-rich rocks in modified skarn at Trembley Mountain may have had a similar origin. However, the writer does not believe that reaction metamorphism of the kind inferred by Magnusson has yielded any major skarn layer or magnetite deposit in the district.

Regional transfer of material

Basic fronts

The development of "basic fronts" as an essential forerunner of advancing granitization might be regarded by some geologists as a suitable mechanism for developing the skarns and magnetite deposits of the district. (See, for example, Reynolds, 1946.) However, it appears that the maximum concentration of granites and the maximum concentration of mafic material have both advanced to the same place in this area. Relative to the neighboring Grenville Lowlands, rocks of the magnetite district show a marked increase in Fe--in dark skarns, in magnetite deposits, and in disseminated accessory iron oxides. Relative to the Lowlands, the magnetite district also shows an abrupt increase in the proportion of granitic rocks. If advancing K, (Na), and Al ions must--to make room for themselves--push out Fe, Mn, and Mg ions, the alkalic and aluminous constituents of the St. Lawrence County district must now be quite uncomfortable in sharing the same menage with the erstwhile resistors of their heated attack.

Metasomatic exchange

While the coexistence of Fe-rich material and the major areas of granite is entirely inconsistent with the concept of the "basic front", it might be explained by Barth's development of the idea of metasomatic exchange. According to Barth (unpublished MS), this type of metasomatism by ionic exchange requires that, as the "injected ions" make their way into the country rock and transform it, the "rejected ions" move away and toward the source of the injected ions. The "rejected ions" thus make up the loss in ions suffered by the depletion of the ionic source. If so, according to Barth, the "rejected ions" of Fe, Mg, etc., will probably disappear downward, and, we might add, if the "rejected ions" disappear downward they would become available for subsequent pneumatolytic or hydrothermal transport and emplacement. (Cf. Landergren, 1948, on the need for exogene concentration of Fe.)

Let us assume for the moment that such an ionic exchange might have occurred in the St. Lawrence County district. Amphibolite and metagabbro will have to be ruled out as a source of Fe; they are sparse in the neighborhood of the magnetite deposits, and their granitized equivalent--hornblende-microcline granite gneiss--is extremely rare near the deposits. Pyroxene gneiss is initially Fe-poor and can therefore be eliminated. Biotite-quartz-plagioclase gneiss is the only possible contributor left among the metasediments. It seems a reasonable source of Fe, for it is intimately associated with all the granite gneiss ores.

If we calculate the "standard cells" (Barth, 1948) for biotite-quartz-plagioclase gneiss, its granitized equivalent (biotite-microcline granite gneiss and sillimanite-microcline granite gneiss), and typical granite gneiss ore, we find that the net change produced by converting the metasedimentary gneiss to granitic gneisses liberates a very small amount of Fe (0.6 Fe^{+2} ion per standard cell). For the district as a whole, the known volumes of biotite gneiss and its granitized equivalent are of the right order of magnitude to yield the inferred volume of ore. However, the quantity of Fe liberated in any one place is so small that the Fe would have to travel several miles in order to form magnetite deposits of moderate size, and tens of miles to form a deposit as large as Benson Mines. The likelihood is slight that such an extended migration could take place by diffusion in the solid state; the diffusion rate for Na_2O, a rapidly moving substance, is only 1,000 feet (300 m) in 3 billion years at a temperature of 700° C. (Jagitsch, 1949). Even if the return of Fe to a deep source could be aided by some fluid medium, and even if the Fe could then be re-introduced through the agency of volatiles, the total distance traveled by the rejected Fe ions would be very great. In contrast, Fe expelled from cooling magma and carried by volatiles emitted by that magma would have to move a mile at most--generally less than 1,000 to 1,500 feet--to pass from the magma to its locus of deposition.

Different levels of erosion

Some geologists have suggested that the magnetite deposits of the district are exposed at an entirely different, and much deeper, level of erosion than the pyritic sphalerite and iron sulfide deposits to the west. Processes of ore formation may have been entirely different at these two levels. However, equivalent rocks of the two areas belong to the same general regional metamorphic grade and exhibit similar patterns of deformation. There is no profound contrast, as between the green schist facies and the granulite facies, for example. Moreover, sporadic sphalerite with exsolved chalcopyrite occurs in the magnetite deposits themselves, as well as in the pyritic sphalerite deposits, suggesting that no profound depth separates the magnetite and sphalerite deposits. Differences between the dominantly igneous complex and the Grenville Lowlands can be explained by the differing extent to which the two areas have been intruded by granitic rocks and affected by their emanations; the differences need not be ascribed to profound difference in level of erosion.

Notes on the ore types

The writer had intended to compare the ore deposits of the St. Lawrence County district with those of similar type found elsewhere. It is not possible to present the comparison here. Instead, a few brief notes are given on similar deposits from other districts.

The writer has previously pointed out that the skarn ores are a common and widespread type, whereas the granite gneiss ores are apparently restricted to the pre-Cambrian. At present, the writer knows of the following deposits very similar to the granite gneiss ores of St. Lawrence County: Lyon Mountain, Clinton County, New York (Postel and Rogers, in preparation); Edison mine, Ogdensburg, New Jersey (Rogers, 1946); Solberg belt, near Nås, southern Norway (Kjerulf and Dahll, 1862); Llano-Burnett area, Texas (Paige, 1911; Barnes et al., 1949). These deposits appear to be partly or wholly in potash granite gneiss of several varieties, though the kind of feldspar accompanying the Solberg deposits and some of the Llano-Burnett deposits is not clearly stated. The Llano-Burnett deposits are especially interesting to the writer because barite has been found in one of them (Paige, 1911, p. 33).

Deposits in soda granite and related rocks are absent from the St. Lawrence County district, but they are found in the Clinton County district, New York (Postel and Rogers, in preparation); the Mineville-Port Henry district, New York; and the Dover district, New Jersey (Sims, 1950).

Modified deposits crudely similar to part of the Trembley Mountain and Parish deposits belong to the Källfallet and Myrback types of Geijer (1923). They are, respectively, cordierite-anthophyllite quartzites and biotite-almandite-quartz rocks of metasomatic origin, found in the Riddarhytte field, Västmanland, Sweden. Granite gneiss ores of the St. Lawrence County type seem to be lacking from the Central Swedish district.

Spessartite garnet is an abundant gangue mineral in the galena-sphalerite deposits of the Broken Hill district, New South Wales (Andrews, 1922, Gustafson et al., 1950). In addition, spessartite, quartz, and magnetite are found in non-productive deposits associated with the Pb-Zn lodes. Andrews (1922) ascribed a common replacement origin to the Pb-Zn lodes and the non-productive magnetite deposits. Gustafson and his associates (1950) consider that the Pb-Zn lodes are hypothermal but suggest that the magnetite deposits may be metamorphosed sedimentary rocks. The Broken Hill deposits are found in an area of complexly-folded sillimanitic rocks. The general geology of the district, and many features of the mineral deposits, are remarkably similar to those of the Parish magnetite deposit, St. Lawrence County. The association of magnetite, Mn minerals, and Ba minerals in the two districts is particularly interesting. The association of Ba with Mn in many deposits formed under a wide variety of conditions amounts almost to a geochemical rule and deserves serious study. The association of Ba with Fe is much less common.

Economic future of the district

The St. Lawrence County district should have a healthy future as a modest producer of iron ore from low-grade and medium-grade deposits. The ore from several small deposits of relatively high grade might, in times of national stress, serve as sweetener for lower grade ore. However, the quantity of lump ore in the district appears to be very small. Greatly extended life and increased production may follow from development of magnetite deposits now sub-marginal in grade. Gillies

(1950) and Webb and Fleck (1950) have expressed optimism concerning the reserves of Adirondack magnetite and the economic advantages that improvements in mining and milling may bring.

The Adirondack magnetites have a strategic position that might prove to be an invaluable asset or a considerable liability. If the two steel plants to be erected on the Delaware River at Paulsboro, New Jersey, and Morrisville, Pennsylvania, are completed before the heavy shipment of foreign ores begins, the Adirondack ores will be even more in demand than they are at present. They may become still more prominent with the general development of the northern region that should follow the construction of the St. Lawrence seaway.

Perhaps the development of the Adirondack area will focus attention on the very low grade magnetite-ilmenite-zircon deposits in the mafic layers of the Diana quartz syenite complex. The ilmenite from these metamorphic rocks has little exsolved iron oxides, and much of the magnetite appears to be free of exsolved ilmenite. (This is a rather uncommon feature among titaniferous magnetites; it is probably a consequence of thorough unmixing and recrystallization during deformation.) The relative purity of the metallics, the presence of abundant zircon, and the suspected occurrence of subordinate monazite in these ultramafic zones within the mafic syenite might offset the extremely low grade of the material.

The possibility of finding sizeable low-grade deposits of primary crystalline hematite in association with microcline granite gneisses in the St. Lawrence County district must also be kept in mind.

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DEPARTMENT OF THE INTERIOR

INFORMATION SERVICE

GEOLOGICAL SURVEY

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GEOLOGIC REPORT ON ST. LAWRENCE COUNTY, NEW YORK IRON ORE DEPOSITS RELEASED

A study by Geological Survey scientists of the geology of the magnetite deposits in St. Lawrence Co. New York has yielded information that will aid in the development of newly discovered iron deposits in the St. Lawrence County district, Secretary of the Interior Oscar L. Chapman announced today.

The two active mines in the district, the Benson Mines at Star Lake, and the Clifton Mine at Degrasse, now produce about 1 million tons of high quality magnetite concentrate annually, about half of the total iron ore production of New York State, Secretary Chapman said. Between 1941 and 1946, 10 promising new deposits and 12 small but favorable prospects were discovered. The discoveries, made by private industry, together with the Geological Survey and the Bureau of Mines, were based on systematic geologic and geophysical prospecting.

A preliminary report covering the geologic studies made between 1943 and 1951 by B. F. Leonard of the Geological Survey has been made available for public inspection in advance of formal publication because of the great interest shown in the region by the iron mining industry.

According to this report, the iron deposits occur in ancient rocks that are pre-Cambrian in age. Two types of deposits are recognized--those in skarn, a highly altered limestone, and those in granite gneiss. The deposits in skarn are variable in grade, complex in structure, and small to moderate in size. The deposits in granite gneiss are moderate to large-sized, relatively low-grade bodies of disseminated ore.

The ore bodies are restricted to a complex structural knot of metamorphosed sedimentary rocks and granitic sheets that have been pressed into tight folds. The ore bodies lie on the limbs and noses of the folds, and the outlines, when plotted on a map, resemble fishhooks.

The ore is younger than the folding and metamorphism of the host rocks. It represents a high temperature replacement effected by solutions derived from the molten granitic rock.

The report entitled, "Magnetite deposits of the St. Lawrence County district, New York," is accompanied by a map showing the relation of the principal iron deposits to certain major geologic features, a map showing some structural features at Clifton Mine, and 2 other illustrations. Copies of the report and illustrations are not available for distribution but may be examined at the Geological Survey, Room 1033 (Library), General Services Building, Washington, D. C.; at the Library, Princeton University, Princeton, N. J.; at the Office of the State Geologist, State Education Building, Albany, N. Y.; and at the Gouverneur Reading Room Association, Gouverneur, N. Y.

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