Ore Deposits of the St. Lawrence County Magnetite District Northwest Adirondacks New York

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By B. F. LEONARD *and* A. F. BUDDINGTON

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The mineralogy, petrology, structure, and genesis of the ores in their geologic setting, with detailed geologic and dip-needle maps

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CONTENTS

ILLUSTRATIONS

[All plates are in separate volume]

TABLES

ORE DEPOSITS OF THE ST. LAWRENCE COUNTY MAGNETITE DISTRICT, NORTHWEST ADIRONDACKS, NEW YORK

By B. F. LEONARD and A. F. BUDDINGTON

ABSTRACT

The district is a major producer of magnetite and crystalline hematite concentrates from low-grade iron ores of Precambrian age. These hypogene iron oxide deposits are the chief subject of the report. One of the nation's largest iron mines, Benson Mines, is in the district. As the result of dip-needle and aeromagnetic prospecting during the 1940's, several promising new deposits were found. These have not yet been developed. The district also contains deposits of ilmenite-magnetite, iron sulfides, supergene hematite, tremolite, and feldspar, none of which was being exploited during this study. The important zinc deposits of the Edwards-Balmat district, also in St. Lawrence County, are west of the area mapped.

The iron ore reserves of the district are mostly in the form of low-grade ores whose average grade is about 25 percent recoverable Fe in the form of magnetite or hematite, or as a mixture of the two minerals. A few small or medium-sized deposits have ore of higher grade, but virtually no lump ore is known to remain in the district. The district's reserves of low-grade ore, though large, are indeterminate, owing to the limited exploration of most of the deposits. Reserves of iron-bearing material containing less than 20 percent Fe are very large. While material of this grade did not constitute ore during the 1940's, it is a potential future source of iron. Some of the material containing less than 20 percent Fe is found as wallrock at the more promising magnetite deposits; some of it is associated with deposits that contain little material of ore grade.

The district extends from the border of the Adirondack massif inward toward the anorthositic core. Northwest of the district but partly within the mapped area are the Grenville lowlands, underlain chiefly by metasedimentary rocks of the Grenville series. Within the district, granitic rocks predominate; metasedimentary rocks and migmatites are subordinate, making up about 15 percent of the bedrock. Anorthosite, metagabbro and amphibolite, diorite, metadiabase, basaltic dike rocks, and Potsdam sandstone (Cambrian) are sparingly represented. The bedrock, though locally well exposed, is mantled in many places by Pleistocene glacial debris. The granitic rocks belong to two series, an older quartz syenitic series and a younger series of granite and granite gneiss. Except for some of the younger granites and the basaltic dikes, all the Precambrian rocks have in some measure been dynamothermally metamorphosed, yielding by solid flow to form foliated and lineated rocks of complex structure and somewhat variable but high metamorphic grade. Part of the region was uplifted and block-faulted during Silurian (?) time.

Two principal types of hypogene iron oxide deposit are present: magnetite deposits in skarn or marble (skarn ores), and magnetite deposits—with or without hematite—in microcline granite gneiss (granite gneiss ores). Locally, the host rock of both types of deposit has been modified in a characteristic way to given two subtypes: modified skarn ores, and modified granite gneiss ores. Magnetite, virtually nontitaniferous, is the only significant iron oxide in skarn ores. The principal gangue mineral is green pyroxene of the diopside–hedenbergite series, though biotite or dark amphibole is prominent locally. The 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. We regard them as pyrometasomatic.

Granite gneiss ores are accompanied by quartz, barium-bearing potassic feldspar, biotite, manganiferous garnet, and sillimanite. Magnetite in these ores is often accompanied by slightly titaniferous primary crystalline hematite. Locally this hematite, with or without martite (hematite pseudomorphous after magnetite), forms sizeable ore bodies. Granite gneiss ores form disseminated deposits of uniformly low grade, remarkable continuity, and moderate to very large size. Ores of this type have few known parallels.

The host rocks of modified skarn and granite gneiss ores have been partly replaced by quartz, potassic feldspar, fluorite, barite, and scapolite (in skarn) or spessartite (in granite gneiss) before introduction of the iron oxides. Only one major deposit is known for each of the subtypes, but small parts of some skarn and granite gneiss ores are assignable to the subtypes. Modified skarn ore contains magnetite and has features transitional between the parent ore types. Modified granite gneiss ore contains both magnetite and hematite. Its character approaches that of some hypothermal vein material, and the gangue of the single known deposit closely resembles that of the base-metal deposits at Broken Hill, New South Wales.

The ore bodies are restricted to a complex structural knot of metasedimentary and younger granitic rocks (members of the granite and granite gneiss series). Within this knot, developed at the intersection of two dominant regional structural trends, the metasedimentary rocks and the sheets of younger granitic rock have been pressed into variously oriented isoclinal folds against buttresses of older granitic rocks (members of the quartz syenitic series). Other major structural controls have been recognized. All the major deposits are (1) on or within a mile of the borders of great 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. All the deposits are within a mile of the axes of major synclinal folds. Moreover, all the deposits are within 500 feet of at least one facies of the younger granite and granite gneiss.

The magnetite deposits are mainly concordant with the complex structure of the country rock which they replace. At very large and very small scales, the deposits are nevertheless discordant. In shape, the deposits are simple(?) tabular, fishhook, linear, multiple drag fold, and "complex". Deposits of complex shape are controlled by sets of intersecting fold axes of two or more distinct generations. Most deposits of the district, though imperfectly known, seem to have two long dimensions and one short one; that is, they are sheetlike rather than

lath- or rod-shaped. Many deposits have a fishhook shape, the ore having replaced the nose and part of one limb of a syncline. Faulting of ore bodies is not known to be of much consequence in the district.

Though magnetite and hematite are the only ore minerals in the deposits, a suite of metallic minerals is associated with them. Arranged roughly according to frequency and quantity, these metallic minerals are pyrite, pyrrhotite, chalcopyrite, sphalerite, molybdenite, bornite, ilmenite, marcasite(?), chalcocite, covellite, vonsenite, loellingite, graphite, unidentified minerals, and valleriite(?). Vonsenite is an important mineral in one deposit. A little maghemite, very likely supergene, has been found in a single deposit. The rarer metallic minerals are detectable only under the microscope. There seems to be no systematic distribution of sulfides according to type of magnetite deposit, though concentrations of pyrrhotite are usually associated with skarn ores and concentrations of pyrite with granite gneiss ores. Most of the metallic minerals associated with magnetite and hematite are related to the main mineralizing episode that yielded the iron oxides. Some metallic minerals, however, belong to a later stage, and a few are referred to a stage of late hydrothermal mineralization that yielded minerals of the epidote group, zeolites, fluorite, quartz, calcite, clay minerals, and others.

The evidence of supergene alteration of the deposits is generally slight, for the bedrock of the region was thoroughly scraped by Pleistocene glaciers. "Rotting," leaching, limonitization, and clay-mineral alteration are apparent at the suboutcrop of some deposits deeply mantled by glacial debris. Earthy hematite and chlorite form streaks and small masses along joints, faults, mica-rich zones ("sköls"), and marble layers at a few deposits. This type of alteration is comparable to that which affected the sulfide-bearing schists of the nearby Grenville lowlands and resulted in the development of scattered bodies of supergene hematite in marble. Some of these bodies were mined before World War I.

The magnetite deposits and the hematite bodies locally associated with them are closely related in space, time, and origin. They are thought to be high-temperature replacement deposits effected 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 first major deformation of the Adirondack rocks took place after the consolidation of the quartz syenitic rocks and their intrusion by scattered dikes of hypersthene metadiabase. Subsequently, younger granite magma was intruded into the metasedimentary rocks 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 metasedimentary rocks. Probably the same fundamental magma also yielded a high-potassium, volatile-enriched phase that intruded the metasedimentary rocks as thin sheets, reacted with the country rock, and in places metasomatized it extensively, yielding heterogeneous microcline granite gneiss. Locally, the younger granitic rocks were deformed. An advance wave of metasomatism by volatile emanations rich in F, OH, and Si, locally accompanied by Cl and P, preceded the intrusion of some of the granite and formed skarn. Once the skarns had been developed (and partly enriched in iron) and the heterogeneous microcline granite gneiss had formed, both rocks were locally modified by introduction of quartz, potassic feldspar, fluorite, barite, and scapolite or spessartite. Such modification, appreciable only where the skarn was enclosed in microcline granite gneiss, represents a continuation (or renewal) of the same process that developed all the microcline granite gneisses of the district.

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}, or 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. Progressive decrease in the concentration of Fe, decreasing temperature, and a change in the character of the metasomatizing solutions toward a dilute water-rich fluid led to local alteration and partial leaching of ore and wallrocks, followed by deposition of hydrous silicates, calcite, and sporadic base-metal sulfides. At some later date—possibly in late Precambrian time, possibly in Silurian or younger 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 and chlorite.

INTRODUCTION

Buddington and Leonard studied the geology and ore deposits of the St. Lawrence County magnetite district from 1943 through 1950. Though we did the fieldwork jointly, we divided the responsibility for preparing a report. Because the report was large, it has been issued as two separate publications. The section on regional geology and geography of the district and the history of its settlement, written by Buddington in 1950, has already been published as U.S. Geological Survey Professional Paper 376.

The report at hand contains several chapters prepared by Buddington and separately credited to him. Buddington also wrote most of the section on discovery and development of the iron ores. The rest of the report was written by Leonard. The general section on magnetite deposits was written in 1951, and descriptions of the ore deposits were completed in 1953. A few references to recent geologic literature give the work some appearance of being up to date. However, nearly all our ideas were developed and recorded by March 1951. Since then, one mine (Clifton) has been closed, and two private drilling campaigns have been conducted in the district, in 1953 and 1957. The results of the drilling, so far as they are known to us, require no drastic revision of our views on grade, gross structure, or genesis of the deposits. As one would expect, some of our guesses about structural details have been punctured by the new drill holes, of which there are now roughly five times as many as there were when we made our interpretations in the 1940's. Our report was already in process of publication when the information on recent drilling became available in preliminary form. This new information could not be incorporated in our report.

SCOPE OF THE TWO REPORTS

The study of the district was begun in 1943, during World War II, as one of many such investigations to further the development of strategic mineral resources in the United States. A primary objective was the discovery and development of magnetite deposits.

The St. Lawrence County district was selected for study for the following reasons: Substantial quantities of magnetite had been proved at the Clifton and Benson Mines deposits; the number of known magnetite deposits in the district was very much smaller than should be expected in the vicinity of a major deposit such as Benson; the area was relatively inaccessible and mining had gone on for only a few years during the entire period since settlement, so that there had been little incentive for intensive prospecting; and, finally, no geological survey had ever been made of the Childwold, Cranberry Lake, Stark, or Tupper Lake quadrangles. The area was therefore thought to have good possibilities of yielding additional discoveries.

Previous studies of the Adirondack magnetite deposits by Newland (1908, p. 29–30) and by Buddington had indicated that new magnetite deposits would more likely be found in belts where metasedimentary rocks of the Grenville series and granite were associated than elsewhere. Consequently, it was deemed of first importance to find such belts as quickly as possible, and to outline their general position. During the war period, our main objective was the making of an areal geologic map as expeditiously as possible.

Three "Preliminary Reports" were issued (Buddington and Leonard, 1944, 1945a, 1945b). In these, we indicated the belts that were potentially more favorable for the occurrence of magnetite.

The dependability and detail of geologic mapping in the district are somewhat uneven, being related to the immediate importance of an area for magnetite mineralization, its accessibility, and the number of outcrops. The last is an important factor. Outcrops are relatively abundant along the northwest border of the district but are relatively scarce in the area as a whole. Except for the northwest border, the district is almost wholly forested. Outcrops in the woods are mostly obscured by lichens, forest litter, and moss. Often the geologist must remove these in order to determine rock type and structure. Indeed, on many outcrops, satisfactory structural observations are practically unobtainable. Consequently, in the time and with the resources available, fewer detailed lithologic or structural studies have been made than would normally be called for.

The area shown on the geologic map includes about 1300 square miles, on which are located—so far as we know—all the major magnetite deposits of St. Lawrence County. However, complete geologic maps for the Nicholville and Childwold quadrangles are not available; possibly there are small magnetite deposits in the unmapped parts of these areas.

In addition to mapping the areal geology, we needed to make a detailed study of the known magnetite deposits, in order to determine their mode of occurrence. Such information could then be used as a guide for the further development and exploitation of the deposits, and for the search for new ones.

The U.S. Geological Survey cooperated with the U.S. Bureau of Mines in a program of dip-needle surveys and diamond drilling to discover additional magnetite deposits. The Survey recommended potentially favorable areas for dip-needle surveys, interpreted the geology of the areas in which magnetic anomalies were found, and logged 18,580 feet of core from 45 holes drilled by the Bureau of Mines. Mr. J. D. Bardill was district engineer in general charge of the Bureau of Mines work, and Messrs. P. Ryan, N. A. Eilertsen, W. T. Millar, and D. F. Reed, in association with C. J. Cohen, were successively local project engineers with headquarters at Star Lake. We are grateful for the cordial cooperation and courtesies afforded us by these engineers in connection with the prospecting program. About 33,000 feet of core from the Parish and Clifton deposits of the Hanna Coal and Ore Corporation, and 1603 feet from 8 holes drilled by the Newton Falls Paper Mill, were also logged.

An aeromagnetic survey of the district was carried out by the Geological Survey in 1945, and results for a large part of the district were subsequently published (Balsley, Buddington, and others, 1954a, 1954b, 1959a, 1959b). An aeromagnetic survey of peripheral areas was made in 1946.

IMPORTANCE OF THE DISTRICT

The St. Lawrence County magnetite district is one of three districts in New York that produce nontitaniferous magnetite. (The term "nontitaniferous magnetite," commonly used to designate magnetite ores and concentrates containing less than about 1 percent TiO_2, is convenient, if somewhat inaccurate, and serves to distinguish such ores and concentrates from those containing sufficient TiO_2 to be less desirable furnace products and from ilmenite-magnetite mixtures suitable for use as titanium ores but not as iron ores.) The other districts that produce nontitaniferous magnetite are the Clinton County (Chateaugay, or Lyon Mountain-Dannemora) district and the Mineville–Port Henry district. (See fig. 1.) A fourth area, near Tahawus, produces ilmenite concentrates as its primary product and titaniferous magnetite concentrates as a byproduct.

One magnetite mine is now being operated in the St. Lawrence County district. This is Benson Mines, near Star Lake, owned by the Jones & Laughlin Ore Co. In terms of crude ore production, the mine has at times been the 5th, 6th, or 7th largest iron mine in the United States (Minerals Yearbook 1949–52), and the operation is said to be the largest open-pit magnetite mine in the world (Iron Age, 1951). A second mine, the Clifton mine of the Hanna Coal and Ore Corp., near Degrasse, ceased operations in 1951. The Benson Mines and Clifton deposits, known for a century or more and operated from time to time, were reopened in 1941. From 1943 to 1951, inclusive, the two mines produced about 7 million long tons of shipping product, mostly sintered magnetite averaging 62.5–63.5 percent Fe. Their combined annual production for that period was two-fifths that of New York State, or about 1 percent of the country's total annual production of iron ore. (For production figures, see Crockett, 1943, 1944; J. R. Linney, 1945, 1946, 1947, 1948, 1949; Minerals Yearbook 1944–46, 1948–1951; Webb and Fleck, 1950.)

The production of magnetite concentrates at Benson Mines has been supplemented since 1952 by the production of sintered hematite concentrates averaging 61 to 62 percent Fe and 1.5 to 2.5 percent TiO_2. The capacity of the plant that treats the hematite ore (trade name "martite ore") was rated in 1953 at 300,000 tons per year. In that year, however, the Jones & Laughlin Ore Co. was already planning to expand its Benson Mines facilities to produce 1.9 million tons of iron-oxide concentrates per year, of which more than 500,000 tons will be hematite (Eng. Mining Jour., 1953a, p. 127; 1953b, p. 186; Mining World, 1953, p. 40).

The Benson Mines deposit has been studied in detail by the company geologists since 1941, and publication of a description of the property is contemplated by them. The Geological Survey therefore made no study of this deposit, though it is the largest in the district.

In addition to its magnetite deposits, St. Lawrence County contains the productive Edwards-Balmat zinc district (Brown, 1942) and the Gouverneur talc district (Engel, 1949a). Before World War I, the pyrite and pyrrhotite deposits of the county were an important source of sulfur (Buddington, 1917, 1934). The increasing demand for sulfur after World War II led to a restudy of these deposits by Prucha (1953, 1956). 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 lead veins at Rossie, Bigelow, and Macomb have been explored by diamond drilling (Neumann, 1952).

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, A. E. J., 1947, personal communication) and new finds of minerals in a new environment.

DISCOVERY OF ADIRONDACK IRON ORES (1796–1830)

The earliest development of iron ores in the Adirondack region was in the eastern part. J. R. Linney (1943, p. 480–481) describes it as follows:

> Although much earlier developments in the iron industry in northern New York may have existed, the earliest development of record seems to be the construction and operation of the first Catalan forge in 1798 on the Saranac River in Plattsburg, by Platt. Soon after, in 1801, Levi Highbey and George Throop erected an ironworks at Willsboro Falls on the Boquet River near Lake Champlain * * *. In 1802, an ironworks was established in New Russia, Essex County, some of the ore used coming from the Fisher Hill ore bed, Mineville. In 1803, William Bailey erected a Catalan forge on the Chateaugay River about five miles below the outlet of Lower Chateaugay Lake. He obtained his ore from the old opening known as the Prall vein, which is identical with the present 81 mine of the Chateaugay ore bed. At Ferronia, in the town of Ausable, what is known as the Arnold Hill ore mine was discovered * * * in 1806 * * *. In 1822, Major James Dalliba, in connection with John D. Dickenson, of Troy, erected the first furnace at Port Henry * * *. In 1823, the Chateaugay ore body, at what is now Lyon Mountain, was said to have been discovered by a trapper named Collins * * *. Presence of large quantities of iron ore in Mineville * * * was indicated during the earliest compass surveys in 1810 * * *. Another deposit, the Cheever ore bed, had been discovered in 1804 * * *.

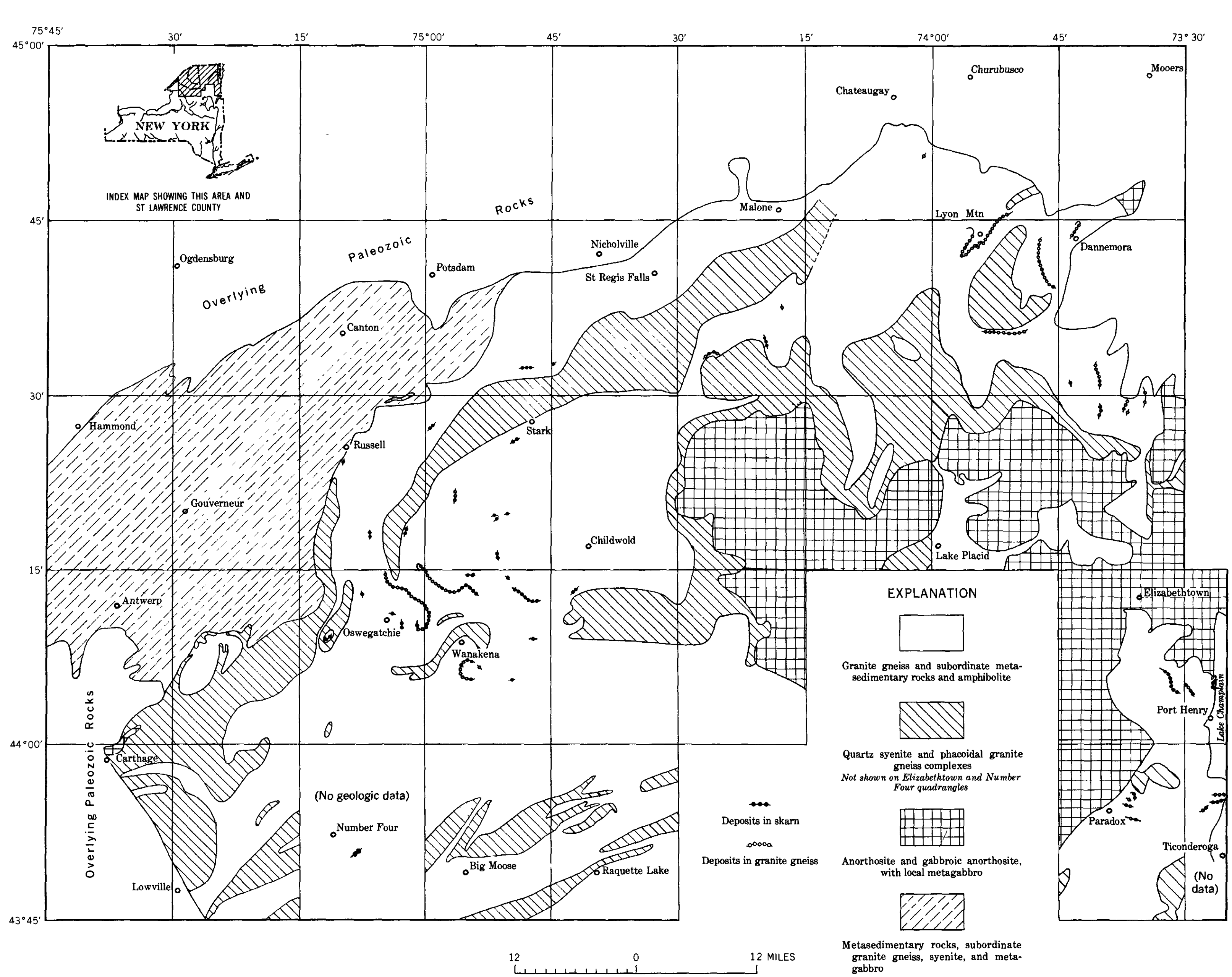

FIGURE 1.—Magnetite concentrations in the north half of the Adirondacks in relation to certain major geologic features, with inset map showing location of the area.

The early discovery of most of the iron ore deposits now known in the eastern Adirondacks is paralleled by the history of discovery of the iron deposits in the St. Lawrence County district.

The southeastern part of St. Lawrence County belonged within two great tracts of land (numbered II and III) purchased by Alexander Macomb from the State of New York in 1791 (Hough, 1853, p. 252–253). In the winter of 1798–99, Benjamin Wright obtained from the owners a contract for surveying these lands into townships, and the survey was carried out during 1799 and 1800. The notebooks of this survey, "A Survey of Sixty Townships, Being a Part of the Tract of Macomb's Purchase, A.D. 1799 and 1800 for Benjamin Wright" are on file at the St. Lawrence County Clerk's office. They contain descriptions of several localities at which magnetic disturbances of the compass were noted, and some optimistic references to the probable occurrence of magnetic iron ore in the towns surveyed.

The date of discovery of the Clifton and Benson Mines ore bodies is not known. It seems highly probable, however, that both ore bodies were found in the earliest period of settlement. Spafford (1813) says:

> A road from Caldwell at the head of Lake George, to Canton in St. Lawrence County, will be opened in 1810; iron-ore is found in Chaumont, and many indications in other parts of this tract; an arsenal is now building in the Town of Russell.

The road referred to is thought to have followed the course indicated on the map (pl. 1) as the Albany trail, thus passing near the ore outcrops at Benson Mines, which is in the old township of Chaumont. Similarly, this old road passed near the outcrops of ore at the Clifton deposit and by way of Degrasse to the village of Russell, where the arsenal was located. Emmons (1842, p. 347–350) refers to what must have been the Benson Mines deposits as follows:

> In the township of Chaumont, one of the largest ore beds in the county has long been known. Twenty-two or three years since, a large quantity of ore was transported to Canton for reduction. It is situated upon the Oswegatchie River, near the crossing of the Albany Road. The vein is in a hill or rocky eminence running east and west, about one hundred feet high on the south side, and fifty on the north side. The top of the ridge is a naked and smooth rock, about fifty rods wide and near a mile in length. The whole of the hill is magnetic iron and quartz or white flint, a pepper and salt mixture * * *. As to the quantity of ore in the southeastern townships [of St. Lawrence County], I have no hesitation in saying it is inexhaustible. The Chaumont ores in particular, together with other veins in the immediate vicinity, are remarkably developed.

The bog limonites and the red hematite ores of the belt between Keene and Antwerp (Jefferson County) were, however, the first iron ores to be worked. They supplied the first blast furnace in northern New York, which was erected at Rossie in 1813 and put into use in 1815.

Probably all the ore deposits worked up to the present time had been found by 1840.

DEVELOPMENT OF MAGNETITE ORES (1830–1939)

The following notes on the development of the iron ores and iron industry near Colton, at Fullerville, at Russell, and at the Clifton mine are taken from a history of St. Lawrence County by Curtis (1894).

A forge with two fires was built by Samuel Partridge at the head of the falls on the Raquette River near Colton in 1828. It was operated until 1830, chiefly in the production of iron from magnetic ore. No information is given as to the source of this ore.

A blast furnace was erected by the Fuller brothers in 1833 at Fullerville, southwest of Edwards. It operated on ores from the Little York red hematite mine, bog ores from Edwards, and magnetic ores from Jayville and Clifton. It was closed down in 1882.

A forge was built by Samuel Smith at the village of Russell on the Grass River in 1846. It had two fires capable of producing about 400 pounds of bar iron a day. It was worked with bog ore, magnetic ore, and scrap iron. The ores were obtained in unlimited quantities from beds (at Clifton?) about 11 miles from the forge. The business was abandoned a few years later.

The Clifton Iron Co. in 1866 built an iron furnace at the falls of the South Branch of the Grass River at Clarksboro. This village is no longer in existence. The ores were brought from the Clifton mine. The company built about 20 miles of wooden railroad from East DeKalb to their mines, completing it in 1865. Silliman (1873) writes, "A few months' use served to reduce the wooden rails to splinters, and render quite useless an expenditure of several hundred thousand dollars." The furnace operated until 1870.

Tyler and Wilcox[1] noted that, according to old reports, the Jayville property was first opened for iron ore in 1854 by Z. H. Benton.

Durant and Pierce (1878, p. 453) note:

> J. D. Judson of Ogdensburg owns a good iron mine in the town of Clifton known as the Parish ore-bed.

Smock (1889, p. 43) gives a few notes on the history of the Jayville mines:

> The existence of iron ore at this locality was known at least 20 years ago, and some mining was done before the present company took possession. In 1886 the Carthage and Adirondack railroad was built as far as the mines and they were worked

[1] Tyler, S. A., and Wilcox, R. E. [1942], Geological report on the Jayville area, Pitcairn Township, St. Lawrence County, New York: Private rept. to Jones & Laughlin Steel Corp.

until the autumn of 1888, producing 25,000 tons of ore in that time. The proprietors are the Magnetic Iron Ore Company.

Newland (1908, p. 132) writes of Benson Mines as follows:

> Systematic mining was not started until the extension of the railroad into the region in 1889. A mill was then erected on the property for the purpose of concentrating the ore into a commercial material, and was run until 1893 when, owing to a depression in the iron trade, the operations became unprofitable. Mining was again resumed in 1900 but only for a short period. The total production subsequent to 1889 has been estimated at 370,000 tons crude ore, or 150,000 tons mill concentrates of above 60 per cent iron. The mines were developed and worked by the Magnetic Iron Ore Co., who have recently been succeeded by the Benson Mines Co. Mining operations were resumed in the fall of 1907.

Work at these mines ceased for a long period after the first world war. The State of New York expressed its continuing interest in the district in the form of geologic reports by Dale (1934, 1935).

RECENT DISCOVERIES AND DEVELOPMENT OF MAGNETITE ORES (1940–46)

The recent development of iron ores in the Adirondacks was brought about by the entrance of large ore and steel producers into this field. In 1938 the Republic Steel Corp. leased the Witherbee-Sherman properties at Mineville, in 1939 arranged with the Delaware and Hudson Co. to operate the Chateaugay properties (Gillies, 1943), and in July 1944 purchased them outright. In 1940 the Jones & Laughlin Ore Co. began an investigation of the potentialities of the Jayville and Benson Mines deposits, and in the latter part of 1943 a mill was completed at Benson Mines and production of concentrates was begun. In 1940 the M. A. Hanna Co., of Cleveland, had proved by dip-needle surveys and diamond drilling an ore body at the Clifton mine warranting a sizable operation and in 1941 began the development of the property, which was brought into production in June 1942 (Hunner, 1943, p. 517). In addition, the Hanna Ore Co. made a dip-needle survey of the Parish deposit in the fall of 1940 and diamond-drilled it in 1941 and 1942. The company made a detailed dip-needle survey of the Granshue anomaly in 1943 and of the VanOrnum (Sweet Pond) anomaly in 1944, drilling the latter in 1946. The Shenango Furnace Co., of Pittsburgh, drilled the Walker Farm (Hughesville School) deposit in 1943. During the same year Mr. A. M. Ross, of the Newton Falls Paper Mill, initiated and supervised detailed dip-needle surveys of the Hardwood Mill, Trembley Mountain, and Jarvis Bridge magnetic anomalies, which are on the property of the company. The Newton Falls Paper Mill had several diamond-drill holes put down on each of the first two properties. In the spring of 1943 the Twin Lakes magnetic anomaly was found, a magnetic survey was made in 1944, and the vein was drilled in 1944 by the Jones & Laughlin Ore Co. In 1945, Mr. Ross had dip-needle surveys made of the Spruce Mountain area, northwest of Newton Falls. The U.S. Bureau of Mines drilled this property in 1946.

The conditions of World War II resulted in the need for greatly increased production of iron ore and raised some problems regarding the adequacy and vulnerability of boat transportation of ore from the Lake Superior district. Because of the strategic importance of iron ore in the Eastern States, the U.S. Geological Survey entered the St. Lawrence County district in July 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, Congress having made special appropriations to the Bureau to further the development of iron ore in the Eastern States. Cooperative exploration by private interests and the two Federal agencies continued in the district through 1946.

The U.S. Bureau of Mines made a detailed dip-needle survey of the Jarvis Bridge anomaly late in 1943 and began a program of diamond drilling on this property that lasted until 1945. During the period 1943–45 the Bureau of Mines also made detailed dip-needle surveys of the Brandy Brook, Silver Pond, Granshue, Jarvis Bridge Extension, Little Mountain, Trembley Mountain, and part of the Spruce Mountain magnetic anomalies, and diamond-drilled part of the Brandy Brook, Silver Pond, and Trembley Mountain deposits. In 1944 the Bureau of Mines began to make systematic reconnaissance dip-needle surveys of areas where the geology was thought to be favorable for magnetite deposits. A report on the results of the magnetic surveys and diamond drilling was published by Millar (1947).

In 1943 the U.S. Geological Survey party discovered an outcrop of low-grade magnetite ore near Dead Creek, Cranberry Lake quadrangle, and in 1944 made a detailed geologic study and reconnaissance dip-needle survey of the area. This work resulted in the finding of two intense magnetic anomalies (Dead Creek 1 and 2) and several subordinate ones (Buddington and Leonard, 1945a).

In 1944 J. R. Balsley and H. E. Hawkes called Buddington's attention to the development of new instruments that made aeromagnetic surveys possible. In April 1945, plans were made by Balsley, in cooperation with Buddington, for a test aeromagnetic survey of part of the Adirondacks. This survey was made by the Geological Survey under the direction of Balsley

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and Hawkes during the months of May and June 1945. A total area of 3,170 square miles in the northern and northwestern Adirondacks was flown on traverse lines spaced, for the most part, at quarter-mile intervals. This survey included the main part of the St. Lawrence County magnetite district. Marginal parts of the district were included in the aeromagnetic survey made in 1946. The magnetic anomalies revealed by the aeromagnetic surveys were studied in the light of their geologic setting. A group of anomalies thought to be most likely indicators of concealed magnetite ore deposits was selected for reconnaissance dip-needle surveys. From July 15 to October 20, 1945, H. E. Hawkes scouted these anomalies, at times assisted by Buddington, P. E. Hotz, Leonard, and C. L. Rogers. Additional ground checks of aeromagnetic anomalies were made by the Survey in subsequent years, and certain anomalies were investigated by field parties of the New York State Science Service under the direction of Shaub (1949).

The results of the aeromagnetic survey, pertaining to St. Lawrence County, can be summarized as follows: Fifteen bands of strong magnetic anomaly were previously known. All were readily recognized from the aeromagnetic records. Seven anomalies found by the aeromagnetic survey in St. Lawrence County were not previously known to us and proved to have both sufficient magnetic intensity and lateral extent to warrant further prospecting. These are the Brunner Hill, Burntbridge Pond, Deerlick Rapids, Outafit, Rock Pond, Skate Creek, and Twin Lake Stream anomalies. A preliminary report on the results of a ground survey was issued in the spring of 1946 (Hawkes and Balsley, 1946). A preliminary, detailed report on the aeromagnetic data for the area around Benson Mines was issued in the summer of 1946 (Hawkes, Balsley, and others, 1946). The aeromagnetic data for a large part of the district were placed in open file in 1950 and were subsequently published (Balsley, Buddington, and others, 1954a, 1954b, 1959a, 1959b).

In 1946, the Bureau of Mines made detailed dip-needle surveys of the Deerlick Rapids, Outafit, and Skate Creek anomalies, completed a detailed dip-needle survey of the Spruce Mountain tract, and diamond-drilled all four properties under the supervision of W. T. Millar, project engineer. That same season, the Bureau of Mines, acting upon recommendation of the Geological Survey, made detailed dip-needle surveys of the Brunner Hill, Hollywood Club, and Wilson Mountain anomalies, drilled one hole at the Brunner Hill deposit, and checked by dip needle several other magnetic anomalies in the district. A report on this work was issued in 1947 by the Bureau of Mines (Reed and Cohen, 1947).

The Outafit deposit is uniquely significant with regard to its mode of discovery. It is the first magnetite deposit, subsequently proved by drilling, discovered in the Western Hemisphere by means of the airborne magnetometer. The ore body itself is completely masked by glacial debris about 50 feet thick, and the existence of a strong magnetic anomaly at that place had never been reported. The sole outcrop found anywhere in the neighborhood is barren wallrock, and that outcrop was found only by intensive search made after the aeromagnetic anomaly had been detected. Since 1945, many notable discoveries of "blind" magnetite deposits have been made in the United States and Canada by means of the airborne magnetometer. Of these, the Outafit is not the largest deposit, but it is the first so found.

At least 47 deposits representing appreciable concentrations of magnetite are now known in the St. Lawrence County district. If, however, discussion is restricted to deposits having a length of at least 500 feet, as indicated by rather uniform dip-needle inclinations of at least moderate amount (+40°), there are 31 deposits whose aggregate length is about 80,000 feet. Very likely, part of this 80,000 feet represents magnetite deposits too small or too low in grade to be mined in the near future.

Of these 31 deposits, 8 (or about 25 percent) were discovered before 1840, in the first 40 years of settlement; 23 deposits (or about 75 percent) were discovered in the recent 7-year period of intensive search, from 1940 to 1946, inclusive. So far as we know, no deposits were found during the intervening period of 100 years.

All eight deposits discovered in the early period have outcrops of ore and were, presumably, found by outcrop. In contrast, all but 2 of the 23 recently discovered deposits were completely obscured by overburden whose thickness ranged from a few feet to 150 feet. And only one of the two was actually discovered by outcrop; the other deposit was indicated by dip needle before the exposed ore was found. The remaining 21 deposits, constituting two-thirds of the 31 known veins in the district, were blind veins, 12 deposits having been found by reconnaissance dip-needle surveys and 9 by aeromagnetic survey. Thus the importance of geophysical aids in finding new ore in the district was overwhelming.

Another aspect of the picture, that of "aggregate vein length," must be viewed. The southern part of the Benson Mines vein was discovered early by outcrop. The vein, which was later found to extend as far north as the Oswegatchie River, has a length equal to about 30 percent of the aggregate length of magnetite-bearing veins now known in the district. Indeed, its length is greater than that of any other five deposits, taken together, and its thickness exceeds that of any other vein. The Benson deposit is, therefore, unique and dominant in the district. Because of the great length of this one vein, it results that, if aggregate vein length rather than number of deposits is considered, 55 percent of the total vein length in the district was discovered by outcrop, and 45 percent is blind veins found by geophysical methods. In numbers of deposits, about 28 percent of all known deposits, or 80 percent of the deposits discoverable by means of ore outcrops, were found during the first 40 years of settlement. In aggregate length of deposits, however, roughly half was discovered in the early period and the other half a century later.

The results of exploration in the district may be reckoned as follows:

Mines and prospects known in 1941[2]

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

Discoveries, 1941–46[3]

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

The following tabulation shows by whom and by what means the discoveries were made.

Magnetite deposits and magnetic anomalies discovered in the St. Lawrence County district, 1941–46

[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	1–O 1–D 1–A	----	1–D
Minor but promising prospects or anomalies (12)	1–D	9–A	1–D	1–D
Small anomalies or prospects, all of questionable worth (25)	5–D	13–A 6–D	1–D	

NOTE: This tabulation 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.

The relative success of the airborne magnetometer and the dip needle as primary detectors of anomalies cannot be gauged from the tabulation above. 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 the tabulation 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 1943. When exploration was at its peak, four companies had crews at work in the district: Jones & Laughlin, Hanna, Shenango Furnace Co., 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.

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. 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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[2] Classification based on data available in 1950, not in 1941.
[3] Includes several found in delayed ground checks of aeromagnetic anomalies, after 1946.

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GEOLOGIC SETTING OF THE ORE DEPOSITS

We interpret the magnetite deposits as high-temperature replacement deposits genetically related to the granite and granite gneiss series, loosely termed "younger granite." Because the deposits are intimately related to the complex rocks in which they occur, knowledge of the regional bedrock geology is indispensable to an understanding of the nature and genesis of the ore deposits. The publishing of the detailed description of the regional geology as a separate report makes it desirable to include in this report on ore deposits a summary of some fundamental geologic features of the region. The summary, while calling attention to many earlier sources, is based on Buddington's report (1962) on the regional geology of this area.

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 80 percent has geologic maps on a scale of 1:62,500.

Topographically, the region is varied. High mountains (maximum elevation about 5,300 feet) make up the eastern part. Successively northwest are the swampy Childwold rock terrace, a "fall-zone slope," and the Grenville lowlands, which adjoin the St. Lawrence lowlands. A peneplain cut on Precambrian rocks is well preserved in the lowland area.

Geologically, the Adirondacks are a complex of igneous rocks and metamorphosed igneous and sedimentary rocks of Precambrian age constituting an extension 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 metasedimentary rocks predominate. Externally, the Adirondack Precambrian rocks as a whole are an asymmetric 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 résumé 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, 1945a, 1945b). Metasedimentary rocks and members of the quartz syenite gneiss 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 syenite gneiss series and a younger granite and granite gneiss 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 mineral for the quartz syenite gneiss 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 orthopyroxene 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 sodic 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 been discussed by Buddington (1948, 1957, 1962).

The metasedimentary rocks that flank the dominantly igneous massif have for many years been assigned to the Grenville series. Interleaved with the metasedimentary rocks are sheets and phacolithic bodies of granite, as well as sheets of gabbroic rocks. The ratio of igneous rocks to metasedimentary rocks is about 30:70, roughly the inverse of the ratio that obtains within the massif. The metasedimentary rocks comprise calcitic and dolomitic marbles, biotite-quartz-plagioclase gneiss (locally garnetiferous), quartzitic rocks, quartz-feldspar gneiss (granulite), amphibolites of uncertain origin, rare metaevaporites, and minor but economically important tremolite schist and related rocks. Most of the metasedimentary rocks have their counterpart in the interior of the Adirondacks. The southern tract of metasedimentary rocks is rather casually known, but the northwest tract has been the subject of repeated investigation, the most recent and thorough being those of Engel and Engel (1953; 1958) and Brown and Engel (1956). Nowhere in this vast and well-exposed tract of dominantly metasedimentary rocks is there evidence of bedded iron formations (Engel and Engel, 1953, p. 1038).

SEQUENCE OF INTRUSION, DEFORMATION, AND METAMORPHISM

The oldest rocks of the Adirondacks are metasedimentary rocks 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 metasedimentary rocks were later intruded by dikes, sheets, lenses, and funnellike 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 metasedimentary rocks 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 gneiss 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 before intrusion of minor hypersthene metadiabase dikes. Perhaps the younger granite made its appearance in the waning stages of this deformation. Granite veins, dikes, and crosscutting 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 or basalt was the last igneous rock intruded in the Adirondacks. Its distribution is rather local (Buddington, 1939, p. 191). In places, diabase dikes have entered fault zones along which postdike movement has also occurred (Cannon, 1937). Except for faulting, the dikes are undeformed and locally glassy. This and other evidence has led to the belief that the dikes are of late Precambrian age. Alkalic dikes are found locally in the eastern Adirondacks, but these are of post-Ordovician age (Buddington and Whitcomb, 1941, p. 85–86). A single alkalic dike of indeterminate age is present in the St. Lawrence County district.

Much of the rock of the Adirondack area, except parts of the younger granites, certain mineral deposits, and the diabase dikes, has in some degree been affected by a regional dynamothermal metamorphism. Its effect on the igneous rocks has been investigated by Buddington (1939, 1952, 1962), and a detailed study of its effect on the metasedimentary rocks of the Grenville lowlands is being made by Engel and Engel (1953, 1958). 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.

ROCKS OF THE ST. LAWRENCE COUNTY MAGNETITE DISTRICT

The bedrock geology of the St. Lawrence County magnetite district (pl. 1) comes close to being a truly representative sample of Adirondack geology. All the major rock types and most of the typical structural features 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 metasedimentary rocks are subordinate. 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 rocks formed in the geologically hotter interior.

Of the bedrock, 80 to 85 percent is granitic igneous or quasi-igneous rock, about 15 percent is metasedimentary or migmatitic, and the remaining 5 percent is mostly metagabbro and amphibolite. Anorthositic gabbro, diorite gneiss, metadiabase, and dike rocks of basaltic composition are found locally, but their areal extent is negligible. A few patches of Potsdam sandstone of Late Cambrian age are present in the northwestern part of the district.

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

METASEDIMENTARY ROCKS AND MIGMATITES

The metasedimentary rocks 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, 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 metasedimentary rocks and migmatites generally parallels the contacts of the rock units. On a small scale, intersecting foliations of the type described by Engel (1949b) have developed locally, but this feature seems generally to be limited to the northwest border of the area.

All the metasedimentary rocks 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, northeast rectangle, 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, southeast rectangle, 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 syenite 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 METASEDIMENTARY ROCKS

Metasedimentary rocks 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 where the exposed area is small, appears massive. Most of the marble shows several percent of disseminated silicates, chiefly phlogopite and diopside. Quartz, potassic feldspar, and epidote may also be present. Sulfide specks are common in some areas; graphite is relatively rare. The silicates may be disseminated, clotlike, 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 decription 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 in many places the only feldspar, but microcline and untwinned potassic feldspar may also be present, dominant, or excluding oligoclase. A little interstitial carbonate is commonly 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 mafic constituents 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. It consists mainly 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. Sillimanite-biotite gneiss occurs locally in the wallrocks 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 potassic feldspar in the groundmass. A detailed study of biotite gneiss from the Colton and Pierrepont areas has recently been made by Engel and Engel (1958).

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; and 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 seen more frequently 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, where present, is commonly biotite, but amphibole or pyroxene occurs locally. The mafic content is rarely as much as 10 percent; 2 to 5 percent is a representative amount. The gneisses may form thick, homogeneous units, or they may be intimately interlayered with thin films, lenticles, and layers of other metasedimentary rocks. Migmatization of quartz-feldspar gneisses seem to be less common than migmatization of other metasedimentary rocks. 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 rock. 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.

In the later years of the preparation of this report, some Fennoscandian geologists seem to have used leptite with less of a genetic implication, and the leptite formation of earlier years has come to be known by other names.

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. A detailed study of the pyrite- or pyrrhotite-rich "rusty gneiss" has been made by Prucha (1953, 1956).

METAGABBRO AND AMPHIBOLITE

Dikes of metagabbro cut the metasedimentary rocks at several places in the northern part of the district, and lenses and sheets of metagabbro are present within the metasedimentary 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 relics of ophitic structure. It consists principally of andesine (locally, sodic labradorite), hornblende, and augite, with or without hypersthene. Reconstituted varieties usually have considerable hornblende. Ilmenite-magnetite is invariably present, in quantities ranging from a few percent of disseminated metallic minerals to local concentrations of rather pure metallic minerals, including some iron sulfide. Some 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, commonly containing some brown mica or clinopyroxene. Garnet is present in one facies. The rock is dark, speckled, medium grained, and generally well foliated. A conspicuous 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 metasedimentary rocks 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 metaigneous rock.

QUARTZ SYENITE GNEISS SERIES

The quartz syenite gneiss 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, 1952, 1962).

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 syenite magma through gravity stratification. Rocks of the quartz syenite gneiss 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 metasedimentary rock 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 distinctions are noted in a subsequent section (p. 83–84). The ilmentite-magnetite deposits in the feldspathic ultramafic lenses are discussed separately (p. 230–231).

HYPERSTHENE METADIABASE

Small, scattered dikes of fine-grained hypersthene metadiabase are present in or close to members of the quartz syenite gneiss series. The dikes range in composition from hypersthene-augite-andesine diabase to garnetiferous hypersthene-augite "granulite" and micaceous amphibolite (Buddington, 1939, p. 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 wallrocks 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 gneiss 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 includes hornblende-microperthite granite, biotite alaskite, microcline granite gneiss, and very minor amounts of sodic 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 gneiss series and forms dikes within certain members of the series. It never forms thin sheets or small, arcuate bodies in the metasedimentary rocks. 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 metasedimentary rocks, and as a local border facies of the hornblende-microperthite granite. Wherever alaskite is found as a border facies, it appears as a "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 1 or 2 percent 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 metasedimentary rocks 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 ghostlike 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, migmatitic, or metasomatic, and its varietal mafic minerals are in a sense xenocrysts or modifications thereof.

In another facies of the gneiss, sillimanite is the 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, ghostlike relics of which are visible locally in the granite gneiss.

The relative age of the microcline granite gneiss is uncertain. Screens of metasedimentary rock always intervene between it and the hornblende granite, and the microcline granite gneiss is not found adjacent to members of the quartz syenite gneiss 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.

Sodic 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 potassic feldspar. Except for the presence of striated plagioclase, hand specimens of the pink sodic granite look exactly like alaskite. It is remarkable that sodic 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 gneiss of the lowlands area (Buddington, 1939, 1948, 1962; Prucha,[4] 1949; Engel and Engel, 1953, p. 1067).

ORIGIN OF THE GRANITE AND GRANITE GNEISS SERIES

Buddington (1948; 1962, p. 85–96) 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 rep-

[4] Prucha, J. J., 1949, A petrogenetic study of the Hermon granite in a part of the northwest Adirondacks. Princeton Univ., Ph. D. dissert. (unpub.).

resents 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 metasedimentary rocks, 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 potassium-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 potassium-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 sodic 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.

REGIONAL GEOLOGIC RELATIONS OF THE ORE DEPOSITS

OCCURRENCE OF THE MAGNETITE DEPOSITS

The magnetite deposits occur in narrow belts of metasedimentary rocks 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 metasedimentary rocks. One major deposit (Clifton) has replaced skarn that is part of a thin wedge of metasedimentary rocks 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 uncertain derivation. The restriction of magnetite deposits to areas in which granitic and metasedimentary rocks are intimately associated was pointed out many years ago by Newland (1908, p. 23–24) and repeated by Alling (1939, p. 143) and 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.

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. The distances cited represent the minimum horizontal distance, measured on plate 1, from a magnetite deposit to a point in a given facies of younger granite. For a very few deposits, the distances were measured on large-scale geologic maps. However, all bodies of granite referred to in this discussion are mappable units whose thickness is perhaps more than 50 feet but less than 1000 feet. Some bodies so thick, known from limited diamond drilling, have not been shown on plate 1 because their orientation and strike-length were uncertain. The figures so obtained, representing apparent distance uncorrected for dip of the structure, therefore exceed the true or "stratigraphic" distance by an unknown amount. Closer computation seems unwarranted, as no one deposit is completely known or its environs completely free of overburden. Because the bedrock units generally have an intricate pattern, it makes little difference whether the apparent distance is measured from the near or far end of a given deposit; the difference is, at most, about 0.2 mile.

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 from 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 from 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 to 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 from 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 to 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.

Parenthetically, we may note that most magnetite bodies are 1 to 3 miles from some facies of quartz syenite; that they are rarely less than 1 mile, and generally much more than 5 miles, from mappable bodies of hornblende-microcline granite gneiss or similar rock interlayered with amphibolite; and that their distance from mappable bodies of amphibolite is variable and usually measurable in miles, rather than in tenths of a mile. These data bearing on hypotheses of origin of the magnetite deposits are discussed subsequently.

ZONING OF MINERAL DEPOSITS

The distribution of northwest Adirondack deposits of oxides and sulfides is shown in figure 2. The distribution is asymmetric (see also fig. 1), perhaps in part because of the asymmetric exposure of the Precambrian rocks and in part because of more fundamental geologic factors. Major deposits of ilmenite-magnetite (not shown on figs. 1 and 2), bearing an epimagmatic relation to the anorthosite, are within the core of the Adirondacks (Stephenson, 1945a, 1945b). Major deposits of nontitaniferous magnetite are farther outward, closest in the eastern area, farthest in the northwest (fig. 1). 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 metasedimentary rocks. 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 metasedimentary rocks. 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. In places, as at the Hyatt mine, one finds deposits whose features are transitional between those of typical zinc deposits and those of typical iron sulfide deposits of the northwest Adirondacks (A. E. J. Engel, 1951, personal communication). 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 quantities of 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, 1936a, p. 251). Significant amounts of copper, lead, and zinc have never been found in the typical pyrite and pyrrhotite deposits (Buddington, 1951, personal communication), though Engel suggests that the Hyatt deposit might be regarded as a transitional variety of zinc-bearing iron sulfide deposit.

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. (See table 1 for analyses of these garnets.) Manganiferous silicates have long been known to occur in the Edwards-Balmat and Gouverneur districts. Engel (1951, personal communication) has noted that the manganese in the talc deposits may conceivably have come from dissociation of very weakly manganiferous dolomitic marble, and that some almandites from garnet-biotite-quartz-oligoclase gneiss of the lowlands contain as much as 6.4 percent MnO. However, an analyzed almandite presumably representative of much of the garnet from garnet-biotite-quartz-plagioclase gneiss contains only 1.57 percent MnO (Engel and Engel, 1953, p. 1073). So far as Leonard knows, concentrations of manganese comparable in extent to the Parish deposit have not been found in the lowlands belt of Grenville rocks.

Barium in significant amounts is common to deposits in several zones. In certain magnetite deposits, it is present as accessory barite and as a slightly barian variety of potassic feldspar. In the pyritic sphalerite deposit at Balmat, it occurs as barite, a common and locally abundant associate of the ore (Brown, 1936a, p. 252). In the talc belt, it is found in feldspar (Engel, 1951, personal communication [5]). The obvious con-

[5] Engel reports 3.95 percent BaO in a perthite from quartz-perthite pegmatite in marble associated with the talc deposit at Talcville.

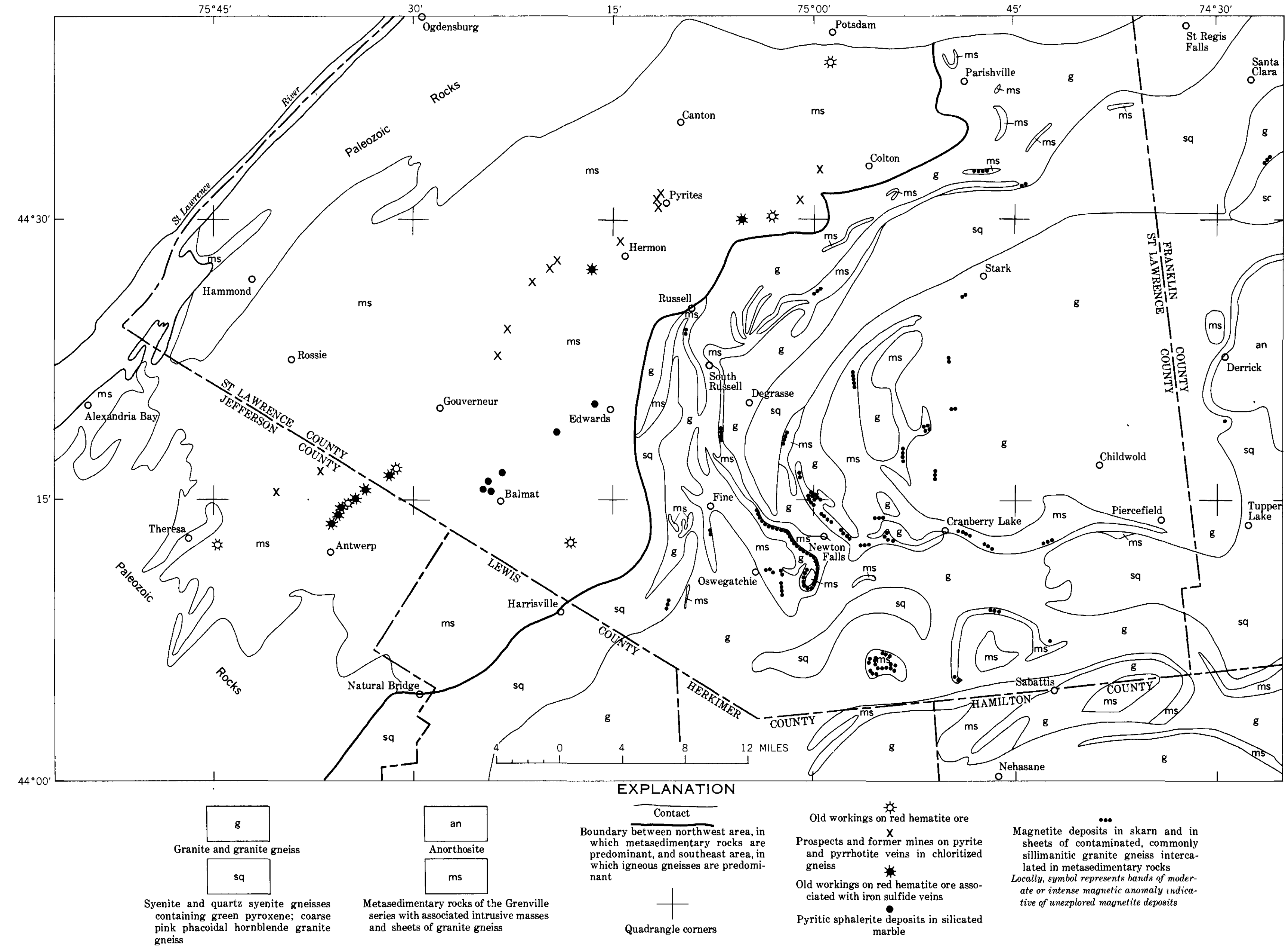

FIGURE 2.—Sketch map showing geologic relations of iron concentrations in St. Lawrence County.

centration of barium minerals is, as one might expect, in the deposits having concentrations of zinc with some lead and silver.

With regard to the argument for regional hypogene zoning of ore deposits, 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 apparently iron-rich) chlorite with the pyrite-pyrrhotite deposits suggested deposition at somewhat lower temperature than that of the marmatite (Buddington, 1939). The iron sulfides, in Buddington's opinion 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. In part this opinion is borne out by the detailed studies of Engel, who notes (Engel, 1951, personal communication) that though tremolite is clearly older than sphalerite, galena, and pyrite, and though anthophyllite is essentially contemporaneous with the zinc-lead sulfides, the serpentine and talc are largely younger than the zinc-lead sulfides and probably formed at lower temperatures.

The examples of iron oxide–iron sulfide zoning, regional in scale, seem to have their small-scale counterpart within two mineralized zones at the Skate Creek magnetite deposit (see p. 184). Here, pyrite, chlorite, epidote, and associated late-hydrothermal alteration products are so distributed longitudinally in the deposit that the quantity of late-hydrothermal alteration products is in a rough way inversely proportional to the quantity of magnetite present on a given cross section of the deposit. The pattern is similar to that found in many hypogene vein deposits whose mineral constituents or wallrock alteration products have a marked zonal arrangement.

Many details of the regional 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 a lower temperature of deposition than that of sphalerite) is at odds with the general paragenetic sequence of sulfides found in the magnetite deposits: pyrite ⟶ pyrrhotite ⟶ sphalerite. Of course, similar anomalies are known from districts regarded as classic examples of hypogene zoning, where for example a copper-bearing zone lies inside a zinc-bearing zone, yet according to microscopic studies of ore suites sphalerite is placed before chalcopyrite in the inferred paragenetic sequence. In St. Lawrence County, the recognition of at least two generations of pyrite in the magnetite deposits poses an additional question: which pyrite, if any, corresponds to the pyrite found in the pyrite-pyrrhotite deposits? Or, if we could actually have been omnipresent at the times and places of mineralization, would we have "seen" iron depositing as oxides in one environment and as sulfides in the other, each aspect of the mineralization process having its own train of mineral phases determined by temperature, pressure, composition, concentration, and a host of other variables?

In spite of these troublesome problems, the areal distribution of the different types of ore deposit (fig. 2) is a fact; the interpretation is debatable. Other hypotheses of origin of the ore deposits would, for example, attribute the regional zonal pattern to ultrametamorphic processes. These hypotheses are discussed on pages 85–92.

STRUCTURE

STRUCTURAL FRAMEWORK

Two great structural trends dominate the western half of the Adirondack region. (See Prof. Paper 376, pl. 4.) 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; Prof. Paper 376, figs. 4, 13, 14, 15.) Within the framework of these rigid masses, the metasedimentary 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 or subparallelism of foliation, preexisting planar structures, and rock contacts is the rule. The folding has affected both metasedimentary and igneous rocks, except for the diabases of late Precambrian age. The magnetite deposits themselves show no sign of internal deformation; they are younger than the folds, and their intimate relation to folds results from replacement of selected structures by magnetite and associated minerals. In size, folds in the district range from many miles in length and 2 or 3 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 the heading "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.

LINEAR STRUCTURES

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*,[6] 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. (For details, see Prof. Paper 376, p. 126–130, and fig. 15.) Thus some large areas characteristically have lineations parallel or subparallel 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 syenite gneiss 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. Buddington (Prof. Paper 376, p. 126) has described the two sets of lineation as parallel or subparallel, and perpendicular or subperpendicular. "Perpendicular" is too often misread as "vertical."

Discordant lineations include such features as small folds, braided pegmatite veins, and mineral streaks; the term is not intended to be synonymous merely with slickensides or grooving.

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 to perhaps 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 Precambrian age and largely of post-Taconian age. (For details, see Prof. Paper 376, p. 130–133.) 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

[6] *b* is the major fold axis; *a* is perpendicular to *b* in the movement plane; *c* is perpendicular to *ab*.

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 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 (a) on or within 1 mile of the borders of areas of discordant lineations, or (b) well within the central zone of concordant lineations at places where lineations culminate, diverge, or change markedly in trend. All but two of the minor deposits are (a) on or within 2 miles of the borders of areas of discordant lineations, or (b) 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. The setting of two minor deposits is indeterminate, owing to the lack of outcrops showing critical structures.

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 2 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 1 mile of the same structural boundary noted above; Edwards is about 3 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 regional 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 1 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.

Several 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, Burntbridge Pond).

GENERAL FEATURES OF 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").

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. Sometimes, Swedish geologists distinguish "skarn-iron ores" (magnetite deposits in skarn) from "skarn-sulfide ores" (sulfide deposits in skarn); they then use the term "skarn ores" to include both classes. Almost no significant sulfide deposits in skarn are now known in St. Lawrence County; therefore, the writer prefers to avoid the cumbersome terms "skarn-iron ores" and "skarn-iron deposits," using instead the terms "skarn ore," etc. The distinction between magnetite and sulfide deposits of skarn type is made clear in the text, when necessary, by some direct statement. The same practice is sometimes followed by Swedish geologists.

Locally in the district, the host rock of both skarn ores and granite gneiss ores 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, sizable concentrations of magnetite are found only in skarn, microcline granite gneiss, and their modified equivalents.

The magnetite deposits are closely related in space, time, and origin. All, 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 Precambrian. 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 Precambrian.

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 also 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 general definition of skarn is based chiefly on discussions of Eskola (1939, p. 383–384) and Geijer and Magnusson (1944, p. 119). Manganiferous skarn silicates have not yet been recognized in the St. Lawrence County magnetite district, and minerals of the olivine and humite groups have not been positively identified there.

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. Note, however, that primary hematite is *not* present in skarns of this district.

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 means of emanations from cooling granitic intrusives. Emanations from intrusive bodies 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 skarns;" skarns formed by metamorphic hybridism; and others. (See, for example, Geijer and Magnusson, 1944; Hjelmqvist, 1942; Korzhinsky, 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) concerning replaced liparites and banded tuffs at Laver, northern Sweden; and Holmes and Reynolds (1947) concerning 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 the Fennoscandian geologists' name for 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." [7] 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 largely 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 virtually 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, on the whole, poor in iron; tremolite is common; and garnet is rare.

Engel (1951, written communication) pointed out that he knows of only 3 or 4 occurrences of iron-rich garnetiferous skarn in the lowlands. However, he adds that certain amphibolites of the area are relatively iron rich and garnetiferous; they may be pyrometasomatic.

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 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. Engel and Engel (1953, p. 1032) have deduced that much of this marble was dolomitic in the early stages of metamorphism, or prior to the known metamorphism. Impure types of marble are less abundant; these are mainly siliceous, rather than aluminous, so far as the composition of the primary sedimentary rocks may be inferred (Engel and Engel, 1953, p. 1033). 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

[7] Translated from the Swedish text by Leonard.

discussion below of garnet skarn, modified skarn, and sköls.) Isolated skarnlike 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 felsic minerals have been introduced subsequently), either because they never had the appropriate chemical composition or because they have lost the impurities 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. Conversely, Paul Niggli (Engel, 1951, personal communication) interpreted the ferrosalite-andradite skarn lenses of the Alps as simply reconstituted magnesian marls. While these interpretations may have merit locally, they are 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 recirculated) and in part 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 (clino-) [table 1, Nos. 4–L through 8–L]
garnet [table 1, No. 1]

Subordinate or infrequent (but locally abundant) skarn minerals.—

amphibole (clino- and ortho-) [table 2, Nos. 11, 12, and 13]
mica (phlogopite and biotite) [table 1, Nos. 9 and 10]
scapolite [table 2, No. 17]
fluorite

Accessory skarn minerals (listed alphabetically).—

allanite [table 2, No. 7]
apatite [table 2, No. 4]
barite (locally prominent in one variety of "modified skarn")
bytownite (and labradorite)
calcite
datolite [table 2, No. 18]
epidote
goethite (?)
graphite
helvite (?)
hypersthene
ilvaite
monazite (?)
potassic feldspar } locally prominent in "modified
quartz } skarns"
sphene
spinel
unknowns (several)
zircon

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

Metallic minerals in the skarn.—

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

The chemical composition and optical properties of certain skarn minerals are given in table 1. Optical properties of some unanalyzed skarn minerals are given in table 2. These data are discussed or referred to in subsequent sections of the report. The bracketed references, above, are provided for the convenience of readers who may wish to see the tabular material at once.

TABLE 1.—*Chemical composition and optical properties of some silicate minerals from the St. Lawrence County magnetite district*

	Garnets				Pyroxenes					Micas		Feldspars	
	1	2	2a	3	4-L	5-L	6-L	7-L	8-L	9	10	11	12
	CN28.14	L-2148	Recalculated	P	BH1.23	CN20.4	CN69.1	CN28.14	OF1.15	CN47-1.10	J1.10	P23.4	P26.4
Chemical composition													
SiO_2	36.70	37.00	37.45	37.06	54.51	53.79	50.19	48.40	45.80	42.33	37.00		63.51
TiO_2	.84	.12	.12	.20	.04	.04	.20	.27	.31	.18	.41		.04
Al_2O_3	9.10	21.54	21.80	17.51	.40	1.41	2.73	3.95	5.11	13.00	17.16		18.88
Fe_2O_3	17.51	.57	.58	4.52	.40	.33	2.98	3.90	5.87	.91	2.40	[0.14]	.12
FeO	3.47	19.38	19.17	6.93	2.35	2.60	7.54	10.52	12.44	2.77	9.27		.12
MnO	.57	11.51	11.65	25.66	.16	.22	.40	.39	.41	.10	.11		$.00_2$
MgO	.78	2.10	2.13	.97	17.19	16.64	12.38	8.92	6.86	25.64	18.45		.05
CaO	30.98	6.71	6.79	6.54	24.90	24.66	23.58	23.20	22.54	.16	.07	[.04]	.08
SrO												[.01]	Tr.
BaO				.04						.08	.00	[.9]	1.18
Na_2O	None	None			.06	.14	.45	.46	.35	.43	.77	1.31	.75
K_2O	None	None			.05	.04	None	.02	.02	9.87	8.77	14.45	15.03
H_2O+(110°C)	.03	.26	.26		.03	.10	None	.16	.17	1.21	4.00		[1].17
H_2O-(110°C)	None	.04	.04		.03	.04	.02	.04	.09	.44	.80		.14
P_2O_5													.09
CO_2													
Cl										[2].11	[2].11		
F										4.54	.53		
S		.39											
ZrO_2		None		(3)									
Sum	99.98	99.62	99.99	99.43	100.12	100.01	100.47	100.23	99.97	101.77	99.85		100.16
Less O for F+Cl										1.93	.24		
Sum (corrected)										99.84	99.61		
Analyst	Norman Davidson 1948	Norman Davidson 1948		J. G. Fairchild 1945	Lee C. Peck	Lee C. Peck	Norman Davidson 1948	Lee C. Peck	Lee C. Peck	Eileen K. Oslund 1950	Eileen K. Oslund 1950	Harry Bastron and P. W. Scott 1955	Eileen K. Oslund 1950

Optical properties

[Optical properties and densities of the pyroxenes were determined by H. H. Hess (1949). Unless otherwise indicated, optical properties of the other minerals were determined by Leonard. The refractive indices determined in sodium light, are accurate to 0.001–0.0005, unless otherwise indicated]

n_{meas}	[4] 1.832±0.003	1.801		[4,5] 1.810±0.005									
n_{calc} [6]	1.832	1.790–1.795		1.811									
nX					1.6718	1.6740	1.6915	1.7073	1.7210	[7] 1.540	[8] 1.567	1.519	1.519_0
nY					1.6785	1.6805	1.6980	1.7140	1.7290	1.571_5	1.609_0	1.523	1.523_9
nZ					1.7013	1.7029	1.7185	1.7328	1.7463	1.572_0	1.609_0	1.526	1.525_0
$nZ-nX$					.0295	.0289	.0270	.0255	.0253	.032	.042	.007	.006
Optic sign					(+)	(+)	(+)	(+)	(+)	(−)	(−)	(−)	(−)
$2V_{meas}$					56½°	56¾°	59°	62°	67½°	10°	12°	73°±2°	[9] 48¼°±2°
Dispersion					B: $r>v$, weak. A: $r>v$, weak.	B: $r>v$, weak. A: $r>v$, weak.	B: $r>v$, weak. A: $r>v$, weak.	B: $r>v$, moderate. A: $r>v$; strong.	B: $r<v$, moderate. A: $r>v$, strong.	$r<v$, very weak.	$r<v$, very weak.	Very weak; indeterminate.	$r>v$, weak.
$Z\wedge c$					39½°	39½°	45° (?)	47½° (?)	51° (?)				
X					Colorless	Colorless	Pale bluish green.	Dark blue green.	Dark blue green.	Colorless	Pale greenish yellow.	Colorless.	Colorless
Y					Colorless	Colorless	Brownish green.	Dark blue green.	Blue green	Pale greenish yellow.	Dark green	Colorless.	Colorless
Z					Colorless	Colorless	Yellow green	[10] Yellowish blue green	[11] Yellow green.	Pale greenish yellow.	Dark green	Colorless.	Colorless
Absorption										Z≧Y>X	Z≧Y>X		
Orientation					Y=b	Y=b	Y=b	Y=b	Y=b	Length-slow.	Length-slow.		
G_{meas} [12]	3.70	4.03		4.07									
G_{calc} [6]	3.72	4.01		4.07									
Density					$D\frac{19°}{4°}$ 3.281		$D\frac{22°}{4°}$ 3.373		$D\frac{25°}{4°}$ 3.44	[7] $D\frac{26°}{4°}$ 2.879	[12] $D\frac{27°}{4°}$ 2.925	[13] $D\frac{25°}{4°}$ 2.58	[12] $D\frac{28°}{4°}$ 2.592
a_0 [14]	12.12±0.01	11.55±0.01											

[1] Ignition loss.
[2] Determined by K. Kuroda.
[3] See text.
[4] Due to variation in composition.
[5] Determined by H. H. Hess.
[6] Refractive index and specific gravity were calculated from those of the "end-member" garnets; values for n and G. of the "end-member" garnets are those listed by Larsen and Berman (1934, p. 234).
[7] ±0.002.
[8] ±0.001.
[9] $2V_{calc.}=51°$.
[10] $\perp bc$, yellow brown.
[11] $\perp bc$, brownish yellow.
[12] Determined by analyst.
[13] Determined by Lucile N. Tarrant.
[14] Determined by H. H. Hess.

No. in table	Specimen No.	Description
1	CN28.14	Andradite ($An_{55.2}Gr_{31.9}Al_{8.1}Py_{3.4}Sp_{1.4}$; $RO:R_2O_3:SiO_2=3.19:1:3.14$) from pyroxenic garnet skarn, Clifton mine, about 10 ft above the main ore zone, hole 28, depth 296 ft. The barren garnet-rich skarn forms a layer 20 to 30 ft thick within the main skarn zone, which is about 200 ft thick at this latitude. Megascopically, the garnet-rich skarn is dark brown (Ridgway color blackish brown, according to Miss Jewell J. Glass), fine grained, and massive; it contains disseminated dark-green pyroxene. Microscopically, the garnet is brown (Ridgway color grayish cinnamon, according to Miss Glass); it replaces strongly pleochroic salite. Apatite is the accessory mineral. For chemical analysis and optical properties of the salite, see this table, No. 7. The analyzed sample of andradite was pure. For a skarn andradite of similar composition, see Eskola (1914, p. 231). This garnet has the composition $An_{63.6}Gr_{26.2}Al_{8.5}Sp_{1.7}$; $RO:R_2O_3:SiO_2=3.06:1:3.00$.
2	L-2148	Spessartitic almandite (roughly $Al_{43}Sp_{26}Gr_{23}Py_8$; $RO:R_2O_3:SiO_2=2.9:1:2.9$) from magnetite ore in garnet-biotite granite gneiss, Benson Mines, southern pit (main pit in 1946). Grab sample, picked up from pit floor where it had fallen after blasting of larger blocks of ore, was selected because of its relative abundance of disseminated garnet. Otherwise, sample is representative of ore then mined in that part of pit. Megascopically, the ore sample is dark gray to black, medium grained, and virtually massive; it is studded with dark red garnets a few mm in diameter. According to Miss Jewell J. Glass, Ridgway color of garnet is morocco red in hand specimen and flesh ocher in thin section. For mode of this specimen, see table 6. For data on its untwinned potassic feldspar, see table 2. The analysis of the garnet concentrate is not of first quality, though it is adequate to show the character of the garnet. A pure concentrate of the garnet was not obtainable by the means then available; the −80, +200 fraction submitted for analysis contained about 1 percent of impurities (crystalline hematite, pyrite, and trace of limonite; hematite 3 times as abundant as pyrite). The garnet was thought to be highly manganiferous; therefore, treatment of the garnet concentrate with HCl seemed inadvisable. The value for S reported by the analyst seems much too high, in view of the microscopically estimated pyrite content ≈0.3 volume percent. Therefore, Leonard has recalculated the analysis (column 2a) as pyrite free but has made no attempt to deduct Fe_2O_3 for hematite.
3	P	Spessartite ($Sp_{60.3}Al_{16.1}An_{13.7}Gr_{5.9}Py_{4.0}$; $RO:R_2O_3:SiO_2=2.99:1:3.10$) from garnet-rich magnetite ore, Parish deposit. Grab sample of ore from long trench is a granular aggregate rich in magnetite and reddish-brown to yellowish-brown spessartite. Microscopically, one finds in addition quartz, untwinned potassic feldspar, crystalline hematite, barite, a little biotite, and traces of apatite, fluorite, and zircon. Minute inclusions, thought to be zircon and perhaps quartz, are common in the garnet; the inclusions were too small to be freed from the powdered garnet submitted for analysis. The garnet concentrate made in heavy liquids was cleaned by dilute HCl and purified electromagnetically. Analyst reports that "ZrO_2 could not be determined accurately because of small sample. It may be present in the sample to the extent of about 0.1 percent." Low sum is perhaps due to difficulty of determining so large a quantity of MnO. However, excellent agreement between observed and calculated physical properties suggests that the analysis is satisfactory. Some variation in the composition of individual garnet granules in the analyzed sample is indicated by the range in refractive index; for bulk of material, $n\approx1.810$, however. For analysis of similar spessartite, see Andrews and others (1922, p. 161). His analysis No. 1461, when recalculated free from galena, has a poor $RO:R_2O_3:SiO_2$ ratio; very roughly, however, this spessartite from the Junction North mine, Broken Hill district, has the composition $Sp_{66.5}Al_{23}Gr_{10}Py_{0.5}$.
4-L	BH1.23	Diopside ($Ca_{48.8}Mg_{46.8}Fe_{4.4}$; 0.6 percent Al in Z) from pale-green diopside skarn containing a little calcite and about 5 percent phlogopite, Brunner Hill deposit, hole 1, depth 226 ft. Sample is from middle of diopside skarn layer 12.0 ft thick, near top of skarn zone. "Stratigraphic" sequence of drill core, from top downward, is: hornblende syenite of hanging wall, highly contaminated; coarse green salite skarn with 5 percent feldspar, 7.3 ft; the diopside skarn, 12.0 ft; dark red-brown, fine- to coarse-grained andradite skarn and andradite-ferrosalite skarn with a 2-ft zone of magnetite-bearing pyroxenic marble, 8.3 ft; pale green diopside skarn, slightly phlogopitic, calcareous at top and bottom, 7.4 ft (last 0.1 foot is coarse, dark, calcareous salite skarn); ore zone—magnetite replacing green, medium-grained, calcareous salite skarn, 13.5 ft (see text description, Brunner Hill deposit); green, medium-grained salite skarn, 2.2 feet; calcareous andradite-ferrosalite skarn with abundant epidote, allanite, and sphene, 4.8 ft; highly calcareous salite skarn, 0.9 ft; contaminated hornblende syenite of footwall.
5-L	CN20.4	Diopside ($Ca_{49}Mg_{46}Fe_5$; 1.5 percent Al in Z) from graphitic diopside skarn, Clifton mine, hole 20, depth 174 ft. The skarn is white, rather fine grained, and granulose, with numerous black folia of graphite. The sample is from the upper part of a 20-foot zone of graphitic diopside skarn that is probably in the lower third of the main skarn zone, here more than 120 ft thick. The skarn zone at this latitude and elevation is highly varied, contains magnetite stringers too thin to exploit, and is cut in places by metadiabase dikes. Below the graphitic skarn is green salite skarn that becomes finer and darker downward; the dark skarn contains sporadic andradite and iron sulfides. Neither footwall nor hanging wall of the main skarn zone was cut by this drill hole.
6-L	CN69.1	Salite ($Ca_{48}Mg_{35}Fe_{17}$; 4.6 percent Al in Z) from medium-grained, medium-green, slightly biotitic salite skarn, Clifton mine, hole 69, depth 125 ft. Sample was taken from 10-ft layer of salite skarn in middle of upper skarn zone, here more than 150 ft thick, consisting mainly of barren ferrosalite and salite skarn, locally andradite rich, with scattered thin layers of calcitic marble. Thin section of the sampled salite skarn contains, in addition to pyroxene, less than 1 percent of blue-green hastingsitic amphibole and primary calcite, plus traces of brown biotite, allanite, sphene(?), and magnetite. Powdered salite submitted for chemical analysis contained a mere trace of amphibole and mica. Though this salite skarn is barren, pyroxene of this composition is very commonly the host of magnetite ore in the district.
7-L	CN28.14	Salite ($Ca_{49.5}Mg_{26.5}Fe_{24}$; 6.9 percent Al in Z) from pyroxenic garnet skarn, Clifton mine, about 10 ft above the main ore zone, hole 28, depth 296 ft. Compositionally, the pyroxene is almost on the border between salite and ferrosalite. For description of the occurrence of this skarn and analysis of the andradite, see this table, No. 1.
8-L	OF1.15	Ferrosalite ($Ca_{48.6}Mg_{20.8}Fe_{30.6}$; 10 percent Al in Z) from dark-green medium-grained pyroxene skarn, Outafit deposit, hole 1, depth 317 ft. Sample was taken 1 foot below contact of ore zone and barren footwall skarn. Much of the footwall skarn contains andradite, with or without calcite or feldspar; however, the sampled zone is almost pure pyroxene skarn containing a trace of hastingsitic amphibole.
9	CN47-1.10	Phlogopite $[(OH_{.55}F_{.99}Cl_{.01})(K_{.86}Na_{.06}Ca_{.01})(Mg_{2.62}Mn_{.01}Fe^{+2}_{.16}Fe^{3}_{.05}Ti_{.01})(Si_{2.90}Al_{1.04})O_{1}]$ from diopsidic fluor-phlogopite sköl, Clifton mine, hole 47-1, depth 392 ft. The sampled sköl, a 2-ft zone consisting largely of pale silvery mica, is the thickest of several sköls in a 45-foot layer of calcareous pyroxene skarn found in the lower third of the southeasternmost skarn zone. The latter zone, 200 ft thick, is mainly dark pyroxene skarn, in part garnetiferous, containing at and near its borders white-speckled layers in which potassic feldspar or scapolite is conspicuous. Dark pyroxene skarn in the upper third of the zone contains several magnetite-rich layers too thin to exploit. Thus the sampled sköl is nearly 100 ft removed, stratigraphically, from any appreciable concentration of magnetite. Pyroxene in the calcareous skarn containing the sköls ranges from medium green salite at the outer borders to gray-green or pale-green diopside near the sampled sköl. The sköl is medium to coarse grained, with randomly arranged fluor-phlogopite flakes surrounding and enclosing subhedral and euhedral diopside ($nX\sim1.675$, or $\sim Ca_{50}Mg_{44}Fe_6$). The approximate mode is: phlogopite, 62 percent; diopside, 35 percent; primary calcite, 2 percent; magnetite, 1 percent; pyrite and fluorite, trace. The sample submitted for analysis was pure, except for a trace of very fine included diopside and a trace of biotite J1.10 (this table, No. 10), the latter from contamination of the screens. A few of the larger phlogopite flakes in the analyzed −100, +200 mesh fraction contained prominent liquid inclusions; these are qualitatively startling but quantitatively insignificant, as they constitute 10–15 percent of only a few flakes. The sum of the (OH, F, Cl) group is 1.55. This may mean that the reported values for H_2O and halogens are a bit low, or it may mean that this fluor-phlogopite, like many carefully analyzed micas, has (OH, F, Cl)<2.0 The X-ray pattern is that of an octophyllite mica, slightly different from the Geological Survey's standard for phlogopite (J. M. Axelrod, 1950, written communication). Qualitative spectrographic analysis (K. J. Murata, analyst, 1950) showed: 10 percent range—Si, Al, Mg; 1 to 10 percent—K, Fe; 0.x percent—Ba, Na, Ti, Ca; 0.0x percent—Mn, Zr; 0.00x percent—B, Cr, Sr, Ni; 0.000x percent—Ga, Cu, Co. For a fluor-phlogopite of similar composition, see Von Eckermann (1922, p. 384). This fluor-phlogopite from the chondrodite-bearing limestone of Mansjö Mountain has 3.15 percent F. Cf. similar micas cited by Von Eckermann (1922, p. 384).
10	J1.10	Biotite $[(OH_{1.94}F_{.12}Cl_{.02})(K_{.86}Na_{.10})(Mg_{2.01}Mn_{.01}Fe^{+2}_{.56}Fe^{+3}_{.13}Al_{.18}Ti_{.02})(Si_{2.70}Al_{1.30})O_{10}]$ from dark green biotite sköl, Jayville deposit, hole 1, depth 196 ft. This sköl, a few feet thick, is one of many sköls found just below the middle of the main dark skarn zone, varibable in composition, here 150 ft thick. These sköls, interlayered with dark green pyribole skarn, are roughly midway between the upper and lower magnetite-rich zones. Microscopically, the sampled biotite sköl is a medium-grained, strongly foliated rock in which folia of rigidly parallel biotite flakes alternate with folia in which the biotite flakes form a herringbone pattern. A few biotite flakes are athwart the strong schistosity. The approximate mode is: biotite, more than 95 percent; pyrite and magnetite, several percent; allanite, apatite, and zircon, nearly 1 percent; unidentified nonopaque mineral, trace. Apatite, as well as allanite and zircon, shows pleochroic halos against biotite. The analyzed sample of biotite was pure, except for a trace of dust. Biotite J1.10 is flexible but inelastic. In addition to the perfect {001} cleavage, the biotite has two other cleavages or partings, one perfect and one imperfect. These two surfaces, subperpendicular to the {001} cleavage, intersect on {001} at angles of 88°±1° and 92°±1°. Thus some of the biotite breaks into short laths, lying on {001}, whose sides are bounded by the perfect parting(?) and whose ends are bounded by the imperfect parting(?). Because of the difficulty of obtaining percussion figures on this biotite, the writer has been unable to determine the crystallographic orientation of the partings(?). The laths of biotite are in some respects similar to the mica "rulers" described by Ferguson (1943). The X-ray pattern is that of an ocotophyllite mica, slightly different from the Geological Survey's standard for biotite (J. M. Axelrod, 1950, communication). Qualitative spectrographic analysis (K. J. Murata, analyst, 1950) showed: 10 percent range—Si, Al, Mg; 1-10 percent—K, Fe; 0.x percent—Ti, Na; 0.0x percent—Ba, Ca, Mn; 0.00x percent—Ga, Cr; 0.000x percent—Sr, Ni, Co.
11	P 23.4	Microcline ($Or_{86.3}Ab_{11.8}Cn_{1.7}An_{0.2}$) from weakly mineralized biotite-microcline granite gneiss, Parish deposit, hole 23, depth 167 ft. Megascopically, the rock is pink, rather fine grained, and very fresh in appearance; foliation is given by slight differences in concentration of magnetite, which is sparse. Microscopically, the rock is medium to fine grained and granoblastic, with a moderate foliation given by slight differences in magnetite content, by elongation of microcline grains, and by preferred orientation of microcline. Principal minerals are quartz, microcline, biotite, and magnetite, but the dark minerals form little more than 5 percent of the rock. Plagioclase (altered) and myrmekite occur as traces. Accessories are pyrite, sphene, apatite, fluorite, zircon, xenotime(?), and tourmaline. A little

No. in table	Specimen No.	Description
		microcline is altered to dusty, radiating, weakly birefringent material; traces of pumpellyite are also present. The sparse biotite flakes are partly or wholly altered to chlorite, epidote(?), and a feldsparlike product.
		The analyzed sample of flesh-pink microcline contained a trace of quartz and chlorite. About 25 percent of the microcline grains showed a yellow stain or some murky alteration products thought to be clay minerals colored reddish by adsorbed(?) iron oxide. A few microcline grains had small inclusions of apatite and minute inclusions of zircon.
		Chemical analysis: quantitative spectrographic analysis (Harry Bastron, analyst, 1955) showed, in percent, Ca 0.03, Ba 0.8, Sr 0.009, and Fe 0.1. Calculated values for the equivalent oxides are given in the table, above. Na_2O and K_2O were determined by flame photometer (P. W. Scott, analyst, 1955).
		Qualitative spectrographic analysis (Harry Bastron, analyst, 1955) showed: 0.x percent—Fe, Ba; 0.0x percent—Ca; 0.00x percent—B, Mg, Sr, Zr, Ti; 0.000x percent—Yb, Cr.
12	P26.4	Untwinned potassic feldspar ($Or_{90.5}Ab_{6.9}Cn_{2.2}An_{0.4}$) from magnetite ore in sillimanite-biotite granite gneiss, Parish deposit, hole 26, depth 138 ft. This sample represents the feldspar-rich variety of modified granite gneiss ore. (For data on similar untwinned potassic feldspars from unmodified granite gneiss ores, see table 2.) Megascopically, sample P26.4 is gray to faintly pinkish, fine to medium grained, and very crudely foliated. Much of the feldspar looks glassy. Microscopically, the sample is fine to medium grained and highly inequigranular; grain boundaries are straight or curved but not interlocking. A faint foliation is given by subparallel biotite and sillimanite, and by slight planar concentrations of magnetite. For mode of this rock, see table 8.
		The analyzed sample of untwinned potassic feldspar was more than 99.5 percent pure. As for the rest, some feldspar grains were slightly cloudy; some showed a trace of zoisitic(?) alteration; some were faintly shadowy. Some grains had minute inclusions of apatite and zircon. A few grains had attached plagioclase blebs. A trace of quartz was present. A few mg of the sample was allowed to stand in 20 percent HCl for half an hour. The washed grains of the microsample showed a similar proportion of cloudy grains, and the HCl was not discolored.
		According to J. M. Axelrod (1950, written communication), the X-ray powder diffraction pattern he obtained for feldspar P26.4 is nearest that of anorthoclase, the Geological Survey's standard being moonstone from Ceylon. According to J. R. Goldsmith (1954, written communication), two X-ray powder photographs made by him indicate that the feldspar is monoclinic and homogeneous; no exsolved plagioclase is evidenced by the X-ray patterns. A single-crystal study of the feldspar has not been made by either investigator.
		Qualitative spectrographic analysis (K. J. Murata, analyst, 1950) showed: 10 percent range—Si, Al, K; 1-10 percent—Ba; 0.x percent—Na, Fe; 0.0x percent—Ca, Mg, Ti; 0.00x percent—Mn, Cr, Sr; 0.000x percent—Pb, Ni.
		For the problem of a similar untwinned potassic feldspar, from Broken Hill, New South Wales, see Andrews and others (1922). For data on barian adularia, see Yoshimura and Shirozo (1953).

TABLE 2.—*Optical properties of some unanalyzed nonopaque minerals from*

	Sulfates			Phosphates						
	Barite			Apatite			Epidotelike minerals			Pyroxene
	1 R–9AT	2 P15.1	3 L–2911–2.4	4 CN–49–19	5 SM3.13	6 L–2503D	7 T3–195	8 SM8.13	9 TM	10 T11.9
nX (or nE) [1]	1.634	~1.636	1.635	1.629	1.633	~1.631		1.696		
nY (or nO)	1.636	~1.638	1.637	1.633	1.636	~1.635	~1.728 generally. ~1.706 locally.	1.703		
nZ	1.646	~1.648	1.647			~1.635		1.718		1.721–1.724
nZ–nX (or nO–nE).	.012	~ .012	.012	.004	.003	~ .004	~ .015	.022	~.028	[2] .014
Optic sign	(+)	(+)	(+)	(−)	(−)	(−)	(+)	(+)	(+)	(−)
$2V_{meas}$	45½°±½°									53½°–57½°±½°.
$2V_{est}$		Moderate	48°±3°			30°–35°	80°–85°	~65°	55°±15°	
Dispersion	$r<v$, very weak.	$r<v$, weak	$r<v$, distinct.			$r<v$, very weak.	$r<v$, very strong.	$r<v$, extreme	$r<v$, very strong.	$r>v$, strong
X∧ best cleav										
Y∧ elongation								0°	0°	
Z∧ cleav										
Z∧ c										0°
X	Colorless	Colorless	Colorless	Colorless	Colorless	Colorless	Light brown	Colorless	Colorless (faint pink or yellow?).	Pink
Y	do	do	do	do	do	do	Green	Intense blue-green.	Intense blue-green.	Yellow
Z	do	do	do	do	do	do	Brown	Colorless	Colorless (faint pink?).	Green
Absorption							Y>Z>X	Y>>Z=X	$Y>>Z \overset{?}{=} X$	X>Z>Y
Orientation	X=c Y=b Z=a			Length-fast.	Length-fast.	Length-fast.				Length-slow.
Density							~3.6; variable			

[1] Refractive indices of almost all minerals reported in this table were determined in sodium light and are generally correct to 0.001.
[2] Total birefringence determined by Berek compensator on grains of known thickness; by the same method, nZ–nY=0.003, and nY–nX≈0.012⁻; indices nX and nY show range of 0.003–0.004.
[3] Absorption indicatrix may not coincide with optical indicatrix.

No. in table	*Specimen No.*	*Description*
1	R–9AT	Barite associated with magnetite and pyrite in biotite-quartz schist, Trembley Mountain area, small prospect pit near west border of mapped area.
2	P15.1	Barite from mixed magnetite-hematite ore in spessartite-quartz-fluorite-barite gangue, Parish deposit, hole 15, depth 53 ft. Barite replaces quartz and yellowish spessartite, shows mutual relations with colorless fluorite or appears to be slightly younger than fluorite, and is replaced by magnetite, by hematite, and by late carbonate and semiopaque material.
3	L–2911–2.4	Barite occurring as coarse euhedral blades, about, 2 cm long, in matrix of fine-grained crystalline hematite, McCollum Farm deposit, dump of southernmost pit.
4	CN–49–19	Apatite (common fluorapatite) from pyroxene skarn layer in subore, Clifton mine, northeast end of pit floor. Apatite occurs as small, granular, light-green patches in dark-green ferrosalite(?) skarn, in part hornblendic, cut by pyrrhotite stringers.
5	SM3.13	Apatite (common fluorapatite) from low-grade ore in garnet-biotite granite gneiss, Spruce Mountain Northwest deposit, hole 3, depth 350 ft. Apatite is an abundant accessory, zircon is also present, and all the potassic feldspar is of the untwinned variety.
6	L–2503D	Apatite from titaniferous iron deposit, Port Leyden, Lewis County, N.Y. One specimen, from a loose block 20 ft east of the old pit, has the following mode: opaque minerals, 75.6 percent; apatite, 21.5 percent; biotite, 0.2 percent; sphene, 0.7 percent; secondary chlorite, 1.9 percent; secondary carbonate, trace. Another thin section of the same specimen shows, in addition, accessory garnet and monazite(?). Opaque minerals are chiefly magnetite and ilmenite and their exsolution bodies; pyrite and pyrrhotite are subordinate; a trace of chalcopyrite is present locally. Identification of the material as apatite was confirmed by ammonium molybdate test.
7	T3–195	Allanite as an abundant accessory mineral in magnetite-rich modified pyroxene skarn, Trembley Mountain South deposit, hole, T3, depth 195 ft. Allanite, constituting O.x percent of this sample of ore, looks uniform in thin section but in oils shows variable nY, faint zoning, and slight "flocculent" alteration. Density of allanite in heavy liquids is variable.
8	SM8.13	Pumpellyite as minute green fibers lining vugs in mineralized garnetiferous microcline granite gneiss, Spruce Mountain Northwest deposit, hole 8, depth 310 ft. Rock contains several percent garnet, about 5 percent magnetite, a little yellow epidote, and pyrite specks. Vugs, which cut magnetite, are circular or slightly elongate, ranging in larger diameter from 0.5 to 3.0 mm; most of them are 1–2 mm in diameter. They constitute roughly 3 percent of the volume of the rock. Some pumpellyite is disseminated as 1-mm patches in the groundmass; the rest of it is present as tufts of radiating fibers or as mats of fibers partly or completely lining the vugs. The fibers, less than 1 mm long, are dark green and have a vitreous to silky luster. Epidote granules are disseminated in the groundmass and rarely come in contact with the pumpellyite-lined vugs; where they do, epidote appears to be older than pumpellyite. Some pumpellyite shows pair twins; twinning plane trace is parallel to prismatic cleavage trace.
9	TM	Pumpellyite formed by partial, local alteration of untwinned potassic feldspar in modified skarn, Trembley Mountain deposit. Data from many thin sections of modified skarn, in large part ore bearing.
10	T11.9	Hypersthene (near En_{47} Fs_{53}) from magnetite ore in quartz rock, Trembley Mountain South deposit, hole 11, depth 285 ft. Rock, which is medium grained and inequigranular, with elongate sutured quartz grains, consists chiefly of magnetite, hypersthene, and quartz. Pale-green mica, bluish amphibole, altered potassic feldspar, and a trace of myrmekite are present. Accessory minerals are sulfides, allanite, apatite, and zircon. Quartz, hypersthene, and sparse euhedral magnetite seem contemporaneous; mica, amphibole, amoeboid magnetite, and sulfides are younger. This variety of quartz rock may represent a reconstituted impure quartzite, subsequently mineralized. Composition of hypersthene, estimated from curves of Walls (1935), ranges from Fs_{50} to Fs_{56}; according to Winchell (1951), range would be from Fs_{35} to Fs_{50}.
11	J1.23	Anthophyllite (ferroan anthophyllite) from rather dark gray-green skarn, Jayville deposit, hole 1, depth 378 ft. The skarn consists almost wholly of prismatic anthophyllite; magnetite forms less than 1 percent of the rock, and a trace of monazite(?) is present. Optical identification of the anthophyllite was subsequently confirmed by an X-ray powder diffraction pattern made by J. M. Axelrod. Cf. Rabbitt (1948), whose analyzed anthophyllite No. 8 contains 2.94 percent Fe_2O_3 and 15.32 percent FeO.
12	T11.13	Actinolite from calcareous diopside-actinolite skarn, Trembley Mountain South deposit, hole 11, depth 331 ft. See text for rock description. Actinolite contains about 18 mole percent ferrotremolite (estimated from Winchell, 1933, p. 246); diopside—almost a salite—has the approximate composition Ca_{50} $Fe_{9.5}$ $Mg_{40.5}$ (estimated from Hess, 1949, p. 641).
13	L–1871	Hornblende (ferrohastingsite) from hornblende-fluorite skarn, Jayville deposit, dump of Benson No. 1 shaft. Skarn is a medium- to coarse-grained aggregate of black subhedral hornblende and subordinate flesh-colored to orange anhe-

iron deposits of the St. Lawrence County magnetite district and environs

Silicates										
Amphiboles			Mica	Feldspars		Scapolite	Zeolites and zeolitelike mineral			
11 J1.23	12 T11.13	13 L-1871	14 P	15 JB3.5	16 L-2148	17 T13.40	18 SM3.2	19 T11.7	20 SM9.13	21 SM2.5
1.659	1.618	1.695	1.563	1.518	1.521	1.564	1.625	1.489	1.485	1.494.
1.666		1.714	1.595	1.523	1.525	1.589	1.654	1.489	1.485	1.502 .
1.675	1.645	1.717	1.596	1.524	1.526		1.671	1.494	1.487	1.505.
.016	.027	.022	.033	.006	.005	.025	.046	.005	.002	.011.
(+)	(−)	(−)	(−)	(−)	(−)	(−)	(−)	(+)	(+)	(−)
			29°±1½°	39½°	47°±3°					
80°–85°	~87°	30°–40°					~70°	<40°; variable.	~30°?	~45°.
$r<v$, very weak.	$r>v$, very weak.	1st axis: $r<v$, very strong; 2d axis: $r>v$, strong.	$r<v$, weak	$r>v$, distinct.	$r>v$, perceptible.		None observed.		Perceptible	$r<v$, weak.
										4°.
			0°				~0°			
0°	18°±½°	13°								
Pale yellow	Pale yellow	Yellowish olive	Colorless	Colorless	Colorless	Colorless	Colorless	Colorless	Colorless	Colorless.
Pale smoky green.	Pale yellow-green.	[3] Dark olive	do	do	do	do	do	do	do	Do.
Pale gray-blue	Pale green	Dark blue-green.	do	do	do	do	do	do	do	Do.
Z≧Y>X	Z>Y>X	Z=Y>>X								
	Y=b	Y=b	Length-slow.							Length-fast.
D $^{24°}_{4°}$ 3.245±0.01										

No. in table	Specimen No.	Description
		dral fluorite. Magnetite and biotite are rare. Cf. Buddington and Leonard (1953), analyzed hornblende No. 7, tables 1 and 2, containing 4.34 percent Fe_2O_3 and 23.28 percent FeO.
14	P	Sericite enclosing granules of spessartite and magnetite and flakes of biotite, Parish deposit, dump of test pit. Sericite forms soft, pale greenish to buff, fine-grained, massive aggregate having earthy or resinous luster and showing occasional flashing "faces" interpreted as cleavage. The aggregate is neither micaceous nor fibrous. The spessartite seems identical with that analyzed, table No. 3. Sample of the pale micaceous material purified by magnetic and heavy liquid separation and containing only a trace of spessartite gave C. S. Hurlbut an X-ray powder diffraction pattern identical with that of sericite. Qualitative spectrographic analysis by H. C. Harrison showed the following concentrations: >10 percent—Si, Al; 10–1 percent—K (+), Fe, Mg (−); 1–0.1 percent—Mn (+), Na (+), Ca, Zr, Ba (−); 0.1–0.01 percent—Ti, V, Sr, Ga; 0.01–0.001 percent—Hf, Cr, Cu, Ag, Be, Li, Rb, Ni, B, Y, Yb; 0.001–0.0001 percent—Not reported. (+) indicates upper half of concentration range; (−) indicates lower half of concentration range. Data of Hurlbut and Harrison were transmitted by E. S. Larsen, Jr. (personal communication, June 1947).
15	JB3.5	Untwinned potassic feldspar from typical low-grade ore, Jarvis Bridge deposit, hole 3, depth 355 ft. (See modes of ore of granite gneiss type, table 6.)
16	L-2148	Untwinned potassic feldspar from low-grade ore, Benson Mines deposit, southern pit (main pit in year 1946). For analysis of garnet from this specimen, and notes on its occurrence, see table 1, No. 2. X-ray powder diffraction pattern of feldspar by J. M. Axelrod is identical with that of analyzed feldspar P26.4, table 1, and is nearest that of anorthoclase standard of moonstone from Ceylon, according to Axelrod. Qualitative spectrographic analysis by K. J. Murata shows: 10 percent range—Si, Al, K; 1–10 percent—Ba; 0.x percent—Fe; 0.0x percent—Ca, Mg, Na, Mn; 0.00x percent—Cr, Ti, Sr; 0.000x percent—Be, Ni, Co.
17	T13.40	Scapolite (perhaps near Ma_{30} Me_{70}) from magnetite-bearing modified pyroxene skarn, Trembley Mountain South deposit, hole 13. The rather low birefringence for a scapolite with this nO was confirmed by direct measurement by means of a Berek compensator. The scapolite seems representative for magnetite-bearing scapolite-rich pyroxene skarn from this deposit. However, scapolites of varied occurrence in the district range in composition from Me_{36} to Me_{74}.
18	SM3.2	Datolite from slightly modified pyroxene skarn, Spruce Mountain Northwest deposit, hole 3, depth 60 ft. Megascopically, the datolite is cream colored, very fine grained, and of somewhat greasy luster; it is interstitial to pyroxene and is not vuggy or veinlike. Microscopically, the host rock has a medium-grained, slightly inequigranular polygonal-crystalloblastic fabric. The estimated mode is: pyroxene (salite or ferrosalite), 70 percent; datolite, 23 percent; untwinned potassic feldspar, 5 percent; minor constituents (mainly sphene and interstitial calcite, with traces of allanite, epidote, and sulfides), 2 percent. Datolite is definitely part of the groundmass, rather than a late-hydrothermal mineral. It occurs as irregular subangular or curving areas consisting of radiating small blades and fine fibers. The nucleus of the radiating datolite is generally at or near the side of the aggregate. In part, the datolite aggregates are interstitial; in part, they replace pyroxene, potassic feldspar, and allanite.
19	T11.7	Chabazite(?) as colorless, nearly cubic rhombohedra in vugs in magnetite-bearing feldspathic pyroxene skarn, Trembley Mountain South deposit, hole 11, depth 222 ft. Chabazite(?) also occurs as small grains and patches replacing the skarn silicates, and as veinlets cutting magnetite. Many of the small chabazite(?) grains are completely rimmed with fine pyrite. Much of the chabazite(?) is accompanied by roundish aggregates of fine green chlorite; some chabazite(?) is slightly replaced by pumpellyite. The chabazite(?) shows crude twinning, local zonal growth, and variable optical properties. Some optically negative material having nY=1.489 and total birefringence of 0.001 is present. It may also be chabazite(?), or it may be some other zeolite, perhaps analcime.
20	SM9.13	Chabazite(?) as colorless, nearly cubic rhombohedra in small vugs in modified pyroxene skarn, Spruce Mountain Northwest deposit, hole 9, depth 329 ft. Vug minerals, arranged according to tentatively inferred paragenetic sequence, are: (1) unidentified earthy pink mineral, (2) epidote, (3) chabazite(?), and (4) pyrite. The chabazite(?) has 3 good cleavages; two of these intersect at 80°–85°. Pair, quadrille, and irregular complex twinning are present in some grains Host rock is a pyroxene-quartz-scapolite skarn containing abundant pyrite and epidote. Pyroxene is partly altered to actinolite. Scapolite is partly or wholly altered to one or more of these minerals: zoisite, epidote, pumpellyite, carbonates, quartz, and sericite(?). A little hydrogarnet(?) is present locally. Accessory minerals are sphene, apatite, and epidotized allanite.
21	SM2.5	Stilbite from veinlet 1 mm wide cutting pyroxene-quartz-feldspar gneiss that contains sporadic epidote and pyrite, Spruce Mountain Southeast deposit, hole 2, depth 120 ft. Stilbite forms colorless to faintly greenish clusters of radiating fibers 2–5 mm long. Some fibers have an opaque powdery white coating.

NOMENCLATURE AND QUANTITATIVE LIMITS

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

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 to 95 percent Ca-Fe-Mg silicates, usually somewhat colored. When such minerals form less than that amount but more than 50 to 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, potassic feldspar, or both—with or without scapolite) constitute 10 to 60 percent of the material, the rest being chiefly pyroxene, amphibole, or both these mafic minerals. Where 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 (minor), Mn (minor), 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.

PYROXENE SKARN

Pyroxene skarn is 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, analysis 35–39, p. 662–664; this report, table 1, Nos. 4–L through 8–L.) 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

TABLE 3.—*Approximate mineralogic composition of skarns, St. Lawrence County magnetite district*

Mineral	Diopside skarn	Salite and ferrosalite skarn	Garnet-pyroxene skarn and variants	Pyroxene-amphibole skarn	Clinoamphibole skarn	Anthophyllite skarn	Sköls	
							Light	Dark
Pyroxene	Diopside	Salite or ferrosalite	Ferrosalite; 5–90 percent (30–50 percent is common).	Salite or ferrosalite	Ferrosalite (generally); absent or present.		Diopside relics	
Amphibole	Actinolite; rare	Actinolite to ferrohastingsite; absent, little or 10–15 percent.		Actinolite to ferrohastingsite.	Ferrohastingsite (generally).	Anthophyllite; ferrohastingsite (trace).	Tremolite in crush zones; local.	Anthophyllite only; little.
Mica	Phlogopite; trace to 15 percent	Phlogopite to biotite; absent, little or 10–15 percent.		Absent; present in one actinolite skarn.	Absent; rare trace.	Biotite; locally absent but generally abundant.	Phlogopite	Biotite; >50–99 percent.
Calcite (primary).	Almost ubiquitous; <5–15	Sporadic; locally 15 percent.	Trace to 15 percent or>.	Absent or little			Sporadic	
Garnet		Sporadic accessory.	Andradite; >90–5 percent (50–80 percent is common).					
Quartz		Little; not common.		Absent, trace, or little.				Generally absent.
Potassic feldspar.		Trace; extremely rare.	Trace; extremely rare.					
Plagioclase								Generally absent; An_{27} in one biotite-anthophyllite sköl.
Magnetite	Few percent locally.	Absent or few percent.	Generally absent	Absent or trace		Absent or little	Sporadic	Sporadic.
Sulfides	Trace to several percent.	Trace to several percent; locally absent, locally abundant.	Absent to several percent; locally abundant.	Absent, trace, or little.	Sporadic	Absent?	do	Do.
Graphite	Sporadic						Trace	Generally absent; rarely present.
Apatite	Very rare	Absent or trace	Absent or little	Absent or trace	Generally absent.	Absent or trace.		Sporadic.
Zircon		Trace; very rare	Absent?	Absent; rare trace		Sporadic		Generally present.
Sphene		do	Absent, trace, or little.					
Allanite		Sporadic	Generally absent	Absent, trace, or very little.		Sporadic		Generally present.
Fluorite	Trace; very rare	Absent, trace, or abundant.	Generally absent; rare trace.	Absent or trace	Sporadic; locally very abundant.			Absent or locally abundant.
Unknowns	Trace; sporadic	Trace; sporadic	Absent or trace	do	Absent; rare trace.	Absent or trace	Talc(?); local	Helvite(?); rare trace.
Spinel						Absent; locally abundant.		
Chlorite	Very rare	Sporadic		Absent or little			Little	Generally absent.
Calcite (secondary).	(?)	Generally present		do				Do.
Monazite						Sporadic	(?)	

layerlike; 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 ($Andr_{56}Gross_{32}Alm_8Pyr_3Spess_1$). (See table 1, No. 1.) Among skarn minerals of the district it is unique, for it does not show the customary variation from an Fe-poor toward an Fe-rich type. Yellowish-brown grossularitic skarn has been found with salite skarn in just one place out of scores of garnetiferous skarns in the district. Andradite is associated with ferrosalite ($Ca_{49.5}Mg_{26.5}Fe_{24}$—table 1, No. 7–L), which it replaces initially as minute vermicular intergrowths, beadlike 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 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 nowhere 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, clinoamphibole skarn alone has been recognized, and that only in minor amounts. The clinoamphiboles range in composition from actinolite through "hornblende" to ferrohastingsite. (See table 2, Nos. 11–13, for optical properties of ortho- and clinoamphiboles.) 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 clinoamphiboles 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 orthoamphibole.

Clinoamphibole 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, mnemonically, shell) often noted by Fennoscandian geologists (see Bugge, 1940, p. 16; Eskola, 1914, p. 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 biotite sköls are more abundant and more favored by magnetite.

The 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 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.[8] 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 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: $(OH_{.55}F_{.99}Cl_{.01})$ $(K_{.86}Na_{.06}Ca_{.01})$ $(Mg_{2.62}Mn_{.01}Fe^{+2}_{.16}Fe^{+3}_{.05}Ti_{.01})$ $(Si_{2.90}Al_{1.04})O_{10}$. [See table 1, No. 9.]

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 resem-

[8] Buddington disagrees. He believes that the scattered biotite present in these thick ore layers represents some mineral (such as pyroxene) that was 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.

bles a chlorite more than it does a mica, but it has the composition of an iron-rich biotite: $(OH_{1.94}F_{.12}Cl_{.02})$ $(K_{.86}Na_{.10})$ $(Mg_{2.01}Mn_{.01}Fe^{+2}_{.56}Fe^{+3}_{.13}Al_{.18}Ti_{.02})$ $(Si_{2.70}Al_{1.30})O_{10}$. [See table 1, No. 10.]

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 clinoamphibole 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 potassic 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, potassic 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 microfabric of the rock may be uniformly polygonal-crystalloblastic or highly irregular. Its mineral composition is variable. (See table 4.)

TABLE 4.—*Approximate mineralogic composition of modified skarns, St. Lawrence County magnetite district*

[Numerals, where given, are volume percent]

Mineral	Generally scapolitic skarn	Modified skarn, Trembley Mountain [1]		
	1	2	3	4
Pyroxene	Generally dominant	[2] 18	[2] 26	60–33.
Amphibole	Generally absent; little, locally.	[3] 30	[3] 20	~2.
Mica		Present [4]	Present [4]	Present.[4]
Calcite (primary)	Sporadic			
Garnet	Absent, trace, little, or dominant.			
Quartz	Absent; rarely present.	35	21	
Potassic feldspar	Generally present		20	22–46.
Plagioclase	Extremely rare; An_{60}	Present	Trace	0–1.
Scapolite	Major; rarely absent			
Magnetite	Generally absent	15	12	11–17.
Sulfides	Sporadic		Present	Present.[5]
Graphite				
Apatite	Generally present; locally abundant; locally absent.	Trace	Present	Present.
Zircon	Sporadic	Present	Present	Trace.
Sphene	Common accessory, locally abundant.	Present		
Allanite	Sporadic; locally abundant.	Present	Present	Present.
Fluorite	Sporadic; locally abundant.			
Unknowns	(?)			
Spinel		Present	Present	
Chlorite	Rare			Present.
Calcite (secondary)	Little; sporadic			Present.
Zoisite	(?)		Present	Present.
Epidote	Sporadic; trace		Trace	Trace.
Pumpellyite	Sporadic; trace		Present	Present.

[1] All are mineralized.
[2] Salite or ferrosalite.
[3] Hornblende?
[4] Biotite.
[5] Maximum~6 percent.

Dark-green ferrosalite is always present, almost always accompanied by scapolite, which ranges in composition from Me_{49} to Me_{74}. (See table 2, No. 17.) Some potassic 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 nonpolygonal, pyroxene is seen to be early; garnet, scapolite, feldspar, and sphene are later, replacing the pyroxene, part of which 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 skarnlike 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, where apparent, is diagnostic. It is important to recognize these rocks for what they are, as the scapolitic skarn represents a significant concentration of halogens, not only in the scapolite itself but also in the fluorite and apatite that are locally abundant in it.

MODIFIED SKARN WITH QUARTZ OR POTASSIC FELDSPAR

Modified skarn with quartz, potassic feldspar, and subordinate scapolite is the principal host for ore at the Trembley Mountain magnetite deposit. Similar rock occurs very sparingly in the wallrocks of several magnetite deposits of the granite gneiss type, notably at the Spruce Mountain Northwest deposit, in the same general belt of metasedimentary rocks as Trembley Mountain. A less usual type of modified skarn, containing much barite and local fluorite, is the host for magnetite at the north end of the Trembley Mountain deposit.

At Trembley Mountain, biotite-microcline granite gneiss and associated metasedimentary rocks 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 felsic minerals. Light-green skarn may show threadlike, 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 felsic minerals. Wispy and patchy areas of light minerals appear. Occasionally, the felsic minerals 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 felsic minerals spread out between adjacent layers, transecting, engulfing, and displacing the intervening mafic minerals. 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 felsic minerals 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. Potassic feldspar 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 gradational nature 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 mineral, whereas intermediate members have varying proportions of the two. No attempt is made in table 4 to give maximum and minimum values for the felsic minerals, 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 rare.

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 potassic 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.

Another type of modified skarn carries the magnetite at the north end of the Trembley Mountain deposit. Here pyroxene skarn showing slight local replacement by quartz and untwinned potassic feldspar has been extensively replaced by barite. Locally, the barite is accompanied by fluorite. The baritic modified skarn has been replaced by magnetite and subordinate amounts of pyrite, pyrrhotite, chalcopyrite, and sphalerite.

SKARN ORES

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. (See p. 54–68.) The associated nonmetallic minerals 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 *within* 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 wallrock 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 wallrock, 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 1 percent silicates) made up of diversely oriented polygons of magnetite ½ to ¾ 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 (35–50 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 (20–35 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 most of the scapolitic skarn with pyroxene or pyroxene and garnet. However, scapolitic pyroxene skarn serves locally as the host of ore at the modified skarn deposit of Trembley Mountain. Though not a host for iron ore, andradite skarn, and highly ferrosalitic pyroxene 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 Precambrian skarn deposits of magnetite. (See, for example, Bugge, 1940, p. 100; Geijer, 1936, p. 153; Geijer and Magnusson, 1944, p. 144–145.) A notable exception familiar to the writer is the Sulphur Hill deposit, Andover district, New Jersey (Sims and Leonard, 1952), where magnetite replaces andradite skarn in preference to other skarn varieties.

SPACE RELATIONS OF SKARN ORES

The wallrocks 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 sodic 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 wallrock, 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 75–90 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 zone 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 wall 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 "minable" 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 minerals such as pyroxene and mica, 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, owing 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 that at ore deposits in contact-metamorphic aureoles, the ore lies on the limestone side of the silicate zones (see, for example, 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. In many places 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 generally 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 that has been 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 to 2,000 feet or less. The average thickness of ore ranges from 30 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 was wholly or partly stoped through a vertical distance of more than 400 feet, equivalent to a "mining

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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 to its closing in January 1952 was slightly more than 2 million tons (estimate by the writer from figures published by Linney and others).

The skarn-magnetite deposits of central Sweden have so many geologic features in common with like deposits in St. Lawrence County that the following notes may be of interest. Many of the Swedish mines now have a vertical depth of more than 650 feet. The deeper mines are 1,000 to 1,400 feet deep. The deepest one on which the writer could find information is Ställberg mines, Ljusnarsberg field. The Ställberg ore body, in manganiferous skarn, has been mined to a depth of 1,650 feet and partly developed to about 2,000 feet. As the dip of this ore body is 70° to 80°, the "mining height" or "dip length" is about the same as the vertical depth. The data cited above are from maps, sections, and text of Geijer and Magnusson (1944).

Besides the most promising deposits, the St. Lawrence County 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.

Table 5, based on rather scanty data, gives some idea of the character of the crude skarn ores. Data on ore from the Clifton mine are given separately in table 15.

TABLE 5.—*Assays (weight percent) for several representative skarn ores*[1]

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 (locally 0.07–0.19)

[1] Data principally from Millar (1947) and Reed and Cohen (1947). For details, see tables accompanying descriptions of individual skarn deposits, following general section of present report.

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. Hematite, found sporadically in several skarn deposits, is a secondary mineral developed by supergene(?) alteration. A little soluble, nonmagnetic 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.

Comparative chemical data on high-grade ore (in this instance, lump ore), medium-grade ore, sintered magnetite concentrates, and tailings are available for only one deposit, at the Clifton mine. (See table 15.) Minor elements in the Clifton tailings have been determined spectrographically. (See table 16.) The chemical composition of skarn minerals occurring in or near ore is given in table 1, Nos. 1, 5–L, 6–L, 7–L, and 9.

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

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. For some years, magnetite was the only recoverable ore mineral, but with the successful completion of experimental work at Benson Mines (Webb and Fleck, 1950) hematite also became 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. In many places the mineralized zones are not sharply limited; material of ore grade passes outward through subore 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 (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 wallrocks, leaching them and depositing pyrite, chlorite, apatite, epidote, pumpellyite, and zeolites.

The mineralogy and paragenesis of the metallic and the 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 1/8 to 1 inch or more in thickness. In many places they are 1/4 to 1/2 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 metallic minerals are distributed as clots or gobs, trailed by finer grains, in a very crudely layered pattern. In some places, the metallic minerals form distinct crosscutting veinlets or irregular networks in the host.

Where magnetite predominates, the metallic material is black. Where hematite predominates, the metallic material has a dark bluish cast. Sporadic grains, small clusters, or veinlets of sulfides may accompany the iron oxides. Pyrite and pyrrhotite are common, chalcopyrite is sparse, molybdenite is infrequenty seen, 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 metallic minerals and only sparingly present in the felsic layers. The felsic minerals 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 subore 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 this report, materials are said to be fine grained if most of their component minerals have diameters less than 1 mm, medium grained if 1 mm to 5 mm, and coarse grained if more than 5 mm. At Benson Mines, the ore minerals are liberated by grinding to 20 mesh for magnetite ore and 14 mesh for hematite ore (Mining World, 1953c, p. 41). Thus the relatively fine grained ores of this district should not be confused with "taconites," whose milling presents quite a different problem.

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. Minable material has about 20 to 32 percent Fe and averages close to 25 percent or slightly less. At Benson Mines, magnetite ore averages 24 percent soluble Fe; hematite ore averages 23 percent soluble Fe (Mining World, 1953c, p. 41). As previously noted, the ore in deposits of granite gneiss type is transitional through subore (about 12–20 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 at 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. (See table 6 for modes of granite gneiss ores and their mineralized associates.) As a rule, hematite ore is closely associated with the sillimanitic facies, though that rock may contain magnetite as well. Ore very rarely occurs in hornblende-microcline granite gneiss. A description of the several facies of microcline granite gneiss was given

TABLE 6.—*Modes of granite gneiss ores and their mineralized associates, St. Lawrence County district* [1]

[Tr., trace]

	Biotitic				Garnetiferous						Sillimanitic		Pyroxenic	Hornblendic		
	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16
	JB1.43	JB1.44	JB10.12	JB10.16	SM6.8	SM6.10	SM6.12	JB3.6	JB9.42	L-2148	BM117.2	BM-49-5	SC1.18	JB3.5	JB8A.3	B-1205
Quartz	54.3	30.0	10.2	11.0	33.2	11.4	18.0	41.2	38.6	38.9	18.3	13.5	14.6	39.5	16.3	11.5
Microcline	-----	-----	68.1	65.4	56.2	-----	-----	-----	-----	-----	-----	-----	[2] 37.0	-----	-----	[2] 15.4
Untwinned potassic feldspar	28.1	28.4	-----	-----	5.2	[3] 63.0	43.2	12.0	30.5	8.0	7.0	[4] 37.4	[4] 10.3	24.4	42.4	[4] 23.4
Plagioclase	.2	Tr.	7.9	3.1	-----	[5] .2	[6] 4.2	Tr.	.1	-----	-----	Tr.	[7] 1.8	([8])	3.7	-----
Scapolite	-----	-----	-----	-----	-----	-----	5.0	-----	-----	-----	-----	-----	-----	-----	-----	-----
Muscovite	-----	.8	-----	Tr.	-----	Tr.	-----	.1	-----	-----	-----	[9] 4.8	-----	[8] 1.8	Tr.	-----
Biotite	1.2	5.0	9.8	3.2	2.3	2.1	6.3	7.5	4.7	3.5	5.8	3.4	-----	[10] 5.1	.5	Tr.
Sillimanite	-----	-----	-----	-----	-----	Tr.	-----	-----	-----	Tr.	24.1	Tr.	-----	-----	-----	-----
Garnet [11]	-----	-----	-----	-----	Tr.	.4	1.9	5.4	9.8	28.0	Tr.	-----	-----	-----	-----	-----
Pyroxene [12]	-----	-----	-----	-----	-----	-----	-----	-----	-----	-----	-----	-----	17.3	-----	Tr.	-----
Amphibole [13]	-----	[14] .7	-----	-----	-----	-----	-----	.3	-----	-----	-----	-----	Tr.	1.4	1.8	13.9
Magnetite	16.0	29.5	3.4	14.2	2.7	22.5	18.4	14.1	15.8	19.4	29.0	19.1	11.5	23.9	32.8	33.2
Hematite [15]	-----	-----	-----	-----	-----	-----	-----	-----	-----	-----	1.2	20.4	Tr.	-----	-----	.8
Pyrite	Tr.	} 3.9	Tr.	2.0	Tr.	.2	1.5	8.1	Tr.	Tr.	3.5	-----	.2	} 2.4	.7	-----
Other sulfides	Tr.		Tr.	Tr.	Tr.	-----	-----	Tr.	-----	Tr.	Tr.	-----	Tr.		Tr.	-----
Fluorite	-----	Tr.	-----	-----	-----	-----	Tr.?	Tr.	Tr.	Tr.	-----	-----	-----	-----	-----	-----
Barite	-----	-----	-----	-----	-----	-----	-----	-----	-----	-----	.9	Tr.	-----	-----	-----	-----
Apatite	.1	.4	.3	.2	Tr.	.1	1.1	1.2	.4	1.4	.7	1.1	.9	1.5	1.3	1.6
Zircon [16]	Tr.	Tr.	.3	.2	Tr.	.1	.1	-----	Tr.	Tr.	Tr.	Tr.	-----	-----	Tr.	.2
Allanite	.1	.5	-----	Tr.	Tr.	Tr.	.3	.5	Tr.	.7	-----	-----	Tr.	Tr.	.5	Tr.
Spinel [17]	-----	-----	-----	-----	-----	-----	-----	Tr.	-----	Tr.	Tr.	-----	-----	-----	-----	-----
Epidote	-----	-----	-----	[18] .7	[19] .4	-----	([6])	-----	-----	-----	-----	-----	6.4	-----	-----	-----
Pumpellyite	-----	-----	-----	Tr.	Tr.	-----	Tr.	Tr.	-----	-----	-----	Tr.	-----	-----	Tr.	-----
Chlorite	-----	Tr.	-----	Tr.	([19])	Tr.	Tr.	[20] 5.1	-----	Tr.	-----	Tr.	-----	Tr.	Tr.	-----
Carbonates	-----	.8	-----	Tr.	-----	-----	Tr.	[20] 4.5	.1	-----	-----	-----	-----	Tr.	-----	-----
Clay minerals	-----	-----	-----	-----	-----	-----	-----	-----	-----	-----	[9] 9.5	[9] .3	-----	-----	-----	-----
	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	99.9	100.0	100.0	100.0	100.0	100.0	100.0

[1] The specimens represent the major types of granite gneiss ore found in the district, but the selection does not show all the variation found within a given type. Moreover, the suite of specimens is neither complete nor quantitatively representative of all the ore types found in a given deposit.
[2] Twinning in microcline is not sharp.
[3] About 20 percent of the potassic feldspar is slightly shadowy, but well-twinned microcline is absent.
[4] Includes some shadowy potassic feldspar.
[5] Altered to sericite and zoisite.
[6] Plagioclase is altered to epidote and zoisite.
[7] Zoisitized.
[8] Sericitized.
[9] Largely derived from sillimanite.
[10] Includes 0.6 percent green mica; rest is normal brown biotite.
[11] Presumably slightly spessartitic almandite similar to analyzed garnet L-2148, table 1.
[12] Strongly pleochroic salite or ferrosalite.
[13] Hastingsitic.
[14] Wormy intergrowth of quartz and amphibole.
[15] Primary crystalline hematite; locally, a very little martite.
[16] Monazite may also be present in a few specimens.
[17] Granular green inclusions in magnetite.
[18] Zoisite.
[19] Combined epidote and chlorite.
[20] Largely as replacement of potassic feldspar

No. in table	*Specimen No.*	*Description*
1	JB1.43	Marginal magnetite ore in biotite granite gneiss, Jarvis Bridge deposit, hole 1, depth 355 ft.
2	JB1.44	Low-grade (almost medium-grade) magnetite ore in biotite granite gneiss, Jarvis Bridge deposit, hole 1, depth 360 ft. (The average of Nos. 1 and 2 is close to the average grade and mineralogical composition of the ore zone cut by hole 1.)
3	JB10.12	Weakly mineralized biotite-microcline granite gneiss, Jarvis Bridge deposit, hole 10, depth 340 ft.
4	JB10.16	Subore in biotite-microcline granite gneiss, Jarvis Bridge deposit, hole 10, depth 362 ft. (Specimen is unusual in containing microcline, without untwinned potassic feldspar; commonly, a rock with this much magnetite contains untwinned potassic feldspar, with or without microcline. However, Nos. 3, 4, 1, and 2, in that order, illustrate the customary changes in granite gneiss attendant on progressive replacement by magnetite.)
5	SM6.8	Weakly mineralized biotite-microcline granite gneiss containing a trace of garnet, Spruce Mountain Northwest deposit, hole 6, depth 108 ft.
6	SM6.10	Low-grade magnetite ore in slightly garnetiferous biotite granite gneiss, Spruce Mountain Northwest deposit, hole 6, depth 132 ft.
7	SM6.12	Low-grade magnetite ore in garnet-biotite granite gneiss, Spruce Mountain Northwest deposit, hole 6, depth 181 ft. (Nos. 5, 6, and 7 illustrate the customary changes in granite gneiss attendant on progressive replacement by magnetite.)
8	JB3.6	Subore in garnet-biotite granite gneiss, Jarvis Bridge deposit, hole 3, depth 366 ft. (Pyrite content of this specimen is abnormally high.)
9	JB9.42	Marginal ore in garnet-biotite granite gneiss, Jarvis Bridge deposit, hole 9, depth about 320 ft.
10	L-2148	Low-grade magnetite ore in garnet-biotite granite gneiss, Benson Mines deposit, southern pit (main pit in year 1946). Specimen is garnet rich. For chemical analysis of garnet, see table 1, No. 2. For data on untwinned potassic feldspar, see table 2, No. 16.
11	BM117.2	Low-grade magnetite ore, slightly hematitic, in sillimanite-biotite granite gneiss, Benson Mines deposit, hole 117. (Specimen is sillimanite rich.)
12	BM-49-5	Medium-grade magnetite-hematite ore in sillimanite-biotite granite gneiss, Benson Mines deposit, "hematite pit" of year 1949. (Nearly all the sillimanite has been altered to muscovite and clay minerals.)
13	SC1.18	Subore in pyroxene-microcline granite gneiss, Skate Creek deposit, hole 1, depth 385 ft. (The introduction of epidote, commonly with considerable pyrite and apatite, is a typical feature of certain mineralized pyroxenic zones in several ore deposits of granite gneiss type.)
14	JB3.5	Low-grade magnetite ore in hornblende-biotite granite gneiss, Jarvis Bridge deposit, hole 3, depth 355 ft. (Specimen contains considerable pyrrhotite.)
15	JB8A.3	Medium-grade magnetite ore in hornblende-biotite granite gneiss, Jarvis Bridge deposit, hole 8, depth 663 ft.
16	B-1205	Medium-grade magnetite ore in hornblende granite gneiss, Skate Creek deposit, outcrop 350 ft southeast of head of Skate Creek Flow.

in the section on rocks of the district. The highly varied 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 metasedimentary rocks.

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 potassic 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, potassic 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 no 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 magnetite 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 subore.

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 wallrocks and much of the remaining metasedimentary sequence are dominantly pyroxenic. Thin zones of biotite gneiss may or may not separate the pyroxenic metasedimentary rock from microcline granite gneiss. 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 from the reconstitution of pyroxene—is entirely unsupported by field or petrographic evidence.

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

Locally, as at Skate Creek, 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 discretely lenticular. In several deposits, the variable position of the ore gives rise to an undulating ore body within the granite gneiss sheet, as in the Jarvis Bridge deposit (text and pls. 13, 14), and the Skate Creek deposit (text and pl. 17). 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 vary 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 netlike 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 metallic and nonmetallic minerals, though the parallelism is more pronounced where there is increased concentration of ore minerals.

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 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 bodies.

At Skate Creek (see pl. 17), 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 0.5° to 1°.

At the Spruce Mountain Northwest ore body (see pl. 18), 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, oral 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 might also be explained by lensing, 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 wallrocks. 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.

One of the easily recognized outward signs of change is the bleaching of the host rock brought about mainly by a lightening of the color of the feldspar. 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[9]; Cushing and others, 1910; and Laitakari, 1920.) However, the change of pink potassic 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 potassic feldspar (see below), though locally the microcline itself merely changes color. The bleaching may take place without any sig-

[9] Prucha, J. J., 1949, A petrogenetic study of the Hermon granite in a part of the northwest Adirondacks: Princeton Univ. Ph. D. dissert. (unpub.).

nificant change in the mineralogic composition of the host rock, but changes in mineral content are likely to accompany the bleaching.

CHANGE IN CONCENTRATION OF MINERALS

The principal silicates in the host are quartz and potassic 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. Where present, it is usually highly sericitized or altered to chlorite and clay minerals, even where the accompanying potassic feldspar is entirely fresh. 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 varietal mineral are extremely rare, so far as the writer knows. 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 nearly the same in both rocks, but it is likely to be different. This applies especially to the accessories. (Cf. table 6; and Prof. Paper 376, table 28.) In the mineralized granite gneiss, apatite and zircon are slightly more abundant, and apatite is concentrated locally to the extent of 1 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 POTASSIC FELDSPAR

A phase change in the potassic 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. (Cf. table 6. However, the data on progressive change of microcline to untwinned potassic feldspar are derived from dozens of specimens; the few representative specimens whose modes are given in table 6 are only a small sample of all those examined for type of potassic feldspar.) Instead we find a clear, untwinned potassic feldspar that very rarely shows a "shadowy" or undulatory extinction and never shows grid structure. Microperthitic intergrowths are absent. 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°. The size of 2V, together with the known chemical composition, makes the material classifiable as a member of the orthoclase-cryptoperthite series of Tuttle (1952). (See table 1, Nos. 11 and 12, for chemical analyses and optical properties of microcline and untwinned potassic feldspar from the Parish deposit; and table 2, Nos. 15 and 16, for optical properties of untwinned potassic feldspar from the Jarvis Bridge and Benson Mines deposits.) Preliminary U-stage study of the cleavages with respect to the optical indicatrix is inconclusive but suggests that the untwinned potassic feldspar is triclinic, rather than monoclinic. X-ray diffraction powder studies are also inconclusive. J. M. Axelrod (1950, written communication) reports that the pattern is nearest that of anorthoclase, a triclinic K-Na feldspar. For the same sample, J.R. Goldsmith (1954, written communication) reports that two X-ray powder photographs made by him indicate that the feldspar is monoclinic and homogeneous. Using MacKenzie's data (1954, p. 356) as a guide, the writer would interpret the X-ray powder diffraction pattern of the untwinned potassic feldspar, table 1, No. 12, as that of a feldspar not precisely monoclinic but showing a minimum departure from monoclinic symmetry, comparable to that of MacKenzie's pattern B. A single-crystal study of the feldspar has not been made.

Both U-stage and single-crystal X-ray studies of the material are hampered by the difficulty of establishing a crystallographic reference direction. All the feldspar grains are small, nearly equant anhedra. Crushed fragments show multiple cleavages, partings, or straight, smooth fractures. For example, the number and quality of such identifiable separate surfaces seen in mounted fragments of feldspar No. 12, table 1, is: 1 excellent, 1 very good, 8 good, and 13 fair or poor. Which of these represents {010}, and which {001}?

The untwinned potassic feldspar has the composition $Or_{90.5}Ab_{6.9}Cn_{2.2}An_{0.4}$; it contains 1.18 percent BaO. For complete analysis, see table 1, No. 12.

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. "Cn" of the formula equals celsian. A feldspar very similar in composition, though not in occurrence, has been described by Yuan (1953, p. 307–308).

Some evidence suggests that the feldspar is merely a microcline that has lost its twinning. Other evidence suggests that the feldspar is indeed orthoclase. Further work on this and other feldspars will be necessary before a definite statement can be made. Perhaps the feldspar is of a sort structurally intermediate between well-developed microcline (triclinic) and sanidine or orthoclase (monoclinic); such feldspars have been described by Goldsmith and Laves (1954), MacKenzie (1954), Harker (1954), and others.

Certainly the original potassic feldspar (microcline) of the host has suffered a change of phase. In addition, it has acquired Ba. However, the presence of Ba in the untwinned potassic feldspar is not, as it once appeared, responsible for or correlated with the change of phase, for microcline and untwinned potassic feldspar from the Parish deposit have nearly identical compositions (cf. table 1, Nos. 11 and 12). The similarity is emphasized by the following comparison (calculations based on atomic weights of 1952):

Microcline P23.4

$Or_{86.2}Ab_{11.9}Cn_{1.7}An_{0.2}$ (mol. percent)

$Or_{87.5}Ab_{11.3}Cn_{1.1}An_{0.1}$ (wt. percent)

Untwinned potassic feldspar P26.4

$Or_{90.5}Ab_{6.9}Cn_{2.2}An_{0.4}$ (mol. percent)

$Or_{91.8}Ab_{6.5}Cn_{1.5}An_{0.2}$ (wt. percent)

Barium in feldspars from unmineralized country rock is sparse. 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 potassic feldspar in the rocks indicates that the potassic feldspars in typical microcline granite gneiss have only ~ 0.1 to 0.15 percent BaO, assuming all BaO to be present exclusively in the feldspars. Thus representative feldspars from magnetite deposits have roughly a tenfold concentration of BaO relative to feldspars from equivalent rocks that have not been mineralized.

The change in phase from microcline to untwinned potassic 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 potassic feldspar, some weakly mineralized gneiss (~ 5 percent magnetite) has nothing but untwinned feldspar, and one even finds rare specimens of ore (~ 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 untwinned potassic feldspar is especially well shown in continuous series of drill cores from the Spruce Mountain and Parish deposits. Zimmer (1947)[10] 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 potassic feldspar 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 potassic feldspar is analogous to the development of adularia in vein deposits, and thereby infer a decreasing temperature concomitant with mineralization.

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 potassic feldspar rather than microcline. In addition, some amphibolites, pyroxene gneisses, and biotite gneisses have untwinned potassic feldspar locally, though their potassic 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. (Cf. 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

[10] Zimmer, P. W., 1947, Phase petrology of the Lyon Mountain magnetite deposits in the northeastern Adirondacks: State College of Washington, M.S. thesis (unpub.).

typical ore. However, one garnet from typical "hanging-wall ore" at Benson Mines has the composition $Alm_{43}Spess_{26}Gross_{23}Pyr_8$; it contains 11.51 percent MnO. (For complete analysis, see table 1, No. 2.) An analyzed garnet from garnet-biotite-quartz-plagioclase gneiss of the Grenville lowlands has the composition $Alm_{65.9}Pyr_{25.5}Spess_{3.5}Gross_{3.2}Andr_{1.9}$; it contains 1.57 percent MnO. (For complete analysis, see Engel and Engel, 1953, p. 1073.) The analyzed garnet containing 1.57 percent MnO is presumably typical of garnets from much biotite-quartz-plagioclase gneiss of the lowlands, according to Engel and Engel (1953, p. 1073, 1075).

Garnet-biotite-quartz-plagioclase gneiss from the massif is very similar to corresponding rock in the lowlands; so it seems likely that the Engels' almandite containing 1.57 percent MnO is at least crudely similar to almandite in metasedimentary biotite gneiss layers near the magnetite deposits.

However, Engel (1951, personal communication) has stated that garnets from gneiss near Trout Lake contain 6.4 percent MnO. Relative to the almandites from the metasedimentary gneiss, the Benson garnet shows a notable increase in Mn, roughly 10-fold relative to the presumably typical almandite and roughly 2-fold relative to garnet from the neighborhood of Trout Lake. 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. These inferences are based on optical data, chiefly on determination of nY in single samples of material from the walls and ore zones of different magnetite deposits. In biotites from metasedimentary biotite gneiss found in the wall rocks of four deposits, nY ranges from less than 1.60 to more than 1.66. A similar but narrower range of nY for biotites from metasedimentary biotite gneiss of the lowlands is reported by the Engels (1953, p. 1070). Biotites from the ore zones of several magnetite deposits of granite gneiss type have nY ranging from $\sim$1.615 to $\sim$1.630 and clustering near 1.625. 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. The change is apparently related to the ore-depositing process, rather than to the granitization of biotite gneiss to form biotite-microcline granite gneiss. Biotites from a few, scattered, single samples of the latter rock have nY ranging from $\sim$1.604 to $\sim$1.634. This range is distinctly greater than that so far known to exist in biotites from the ore zones. No significance can be attached to the fact that the reported range of nY for biotites in the granite gneiss is less than that for biotites from biotite gneiss; too few samples of granite gneiss were studied.

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, possibly owing to admixed lepidocrocite and "limonite." Intergrowths of goethite(?) with spinel in magnetite, and the common arrangement of goethite(?) prisms in triangular patterns parallel to the {111} planes in magnetite, suggests that the goethite(?), like spinel, might have been exsolved from magnetite. However, the presence of goethite(?) in biotite suggests that some goethite(?) is an alteration product. There is no indication that the goethite(?) is supergene, or that it belongs with the late hydrothermal minerals referred to below.

The identification of the mineral as goethite(?) is tentative and very questionable. Actually, several different minerals may be present, but the prisms are too small and sparse to permit positive identification. Commonly, the prisms are biaxial with small 2V; optic sign questionably negative; dispersion very strong; relief and birefringence very high. Because they are biaxial, they cannot be rutile. Because 2V seems to be negative in white light, the mineral should not be brookite. The fragmentary data on optical properties and habit would suit many minerals, including some high-temperature borates. Goethite(?) represents an educated guess, permissive at best.

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. Several unidentified accessory minerals, too sparse to warrant study, have been noted in some specimens of ore.

In addition to containing distinctive accessory minerals, 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 and, locally, by spessartite. Titanium is present as sparse exsolution blades of ilmenite in the iron oxides, and in solid solution in primary hematite. (For data on the Ti and Mn content of hematite, see section on ore mineralogy of the Parish deposit, p. 53.) Barium is found in the potassic 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 deposit now being worked in the district. Several promising deposits of similar ore 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 nearly 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[11] is about 2 miles long and 80 to 400 feet wide, according to Tillinghast (1948, p. 28).[12] The ore reserve within the open pit area is about 65 million tons of magnetite ore and 38 million tons of hematite ore, equivalent respectively to 22 million tons of magnetite concentrates and 12 million tons of [nonmagnetic] iron oxide concentrates (Tillinghast, 1948, p. 28). 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 concentrating 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).

Three types of ore are found in the deposit: magnetite ore (less than 17 percent of the contained Fe is nonmagnetic), hematite ore (more than 17 percent of the contained Fe is nonmagnetic), and mixed magnetite-hematite ore. The ore types occur in predictable zones. Until the recovery of hematite was made feasible, using spiral concentrators, only the magnetite ore was mined. The magnetite concentrate is of Bessemer grade; the hematite concentrate is not (Mining World, 1953, p. 41–42). Production figures are given on page 42 of that report.

Typical analyses of magnetite ore and hematite ore are given in table 7. The analyses of magnetite ore and concentrate are not necessarily "paired analyses" carried out on equivalent ore and concentrate. Nevertheless, 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. (See p. 57–58.) The rather high Ti content of the hematite concentrate convinces Buddington (1954, personal communication) that the bulk of the iron oxide mineral in the Benson "nonmagnetic iron oxide concentrate" (table 7, No. 4) is indeed primary crystalline hematite, rather than martite. The proportions of hematite to martite in the magnetite ore, where nonmagnetic iron oxide minerals are relatively sparse, has not been investigated in detail, so far as Leonard knows.

TABLE 7.—*Typical analyses (weight percent) of crude ore and concentrates, Benson Mines*

	Magnetite ore [1] (crude)	Hematite ore [2] (crude)	Magnetite concentrate [1]	Hematite concentrate [2]
Fe				61–62
Soluble Fe	[3] 26.75	~23	63.75	
Magnetic Fe	24.00	[4] ~1	62.54	
P	.192		.024	.18
S	.48		.21	.25
TiO_2	.68		.77	1.5–2.5
Mn	.25		.25	
SiO_2	40.80		5.66	5.0

[1] Webb and Fleck (1950, p. 447). Note that these are not necessarily "paired analyses" carried out on equivalent ore and concentrate.
[2] Mining World, 1953, p. 40, 41.
[3] Average grade of magnetite ore is 24 percent soluble Fe, according to Mining World (1953, p. 41).
[4] Calculated by Leonard; very rough approximation.

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 to 25 percent magnetic Fe. Some of the bodies also have substantial thicknesses of subore

[11] The Jones & 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.

[12] Webb and Fleck (1950) give the width as 400–1,400 ft, and another writer (Mining World, 1953, p. 41) says that the "main ore structure" is about 2½ miles long and 400–1,600 ft wide.

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, N.Y., is several thousand feet long at the surface, has an average minable 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 the Lyon Mountain belt are replacements of pyroxene-microcline-microperthite granite gneiss and pyroxene-plagioclase granite gneiss (Postel, 1952, p. 78–79), but as ore deposits, rather than petrographic types, they resemble the granite gneiss ores of St. Lawrence County.

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, potassic 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, potassic 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 P 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.

The optical properties of some skarn minerals, minerals that modify skarn, accessory minerals, and late hydrothermal minerals of the Trembley Mountain deposits are given in table 2, Nos. 1, 7, 9, 10, 12, 17, and 19.

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 metasedimentary rocks and magnetite deposits. (See pl. 19.) A sheet of contaminated microcline syenite gneiss (and granite?), estimated to be 150 to 200 feet thick, separates the two belts; and rock of similar type appears to warp around the north end of the area. The folded complex of metasedimentary rocks 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 pyroxene skarn and modified 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 of modified pyroxene skarn occurring at the top of the zone of once-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 of the ore that has replaced it.

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, metallic minerals, and sporadic untwinned potassic 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 here, but the main features of its occurrence will be pointed out. (See description of Trembley Mountain deposit, p. 203, for a fuller account.) 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, in many places 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 to 50 percent quartz. The quartz rock itself everywhere contains disseminated magnetite or pyrite (in some places, 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 by having a fine-granulose texture and a regular, fine-scale lamination. These are interpreted as metaquartzites. Locally the coarser type of quartz rock accompanies them, and the microfabric of the two types is similar. Why not, then, interpret all the quartz rock as metamorphosed quartzite?

First, because 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 in many places 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.

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, p. 256–257); from the Falun district, Sweden, by Geijer (1917, p. 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, p. 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 atttributed 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 subore. A little ore is found in quartz-feldspar rock (aggregate of quartz and untwinned potassic feldspar) and contaminated granite gneiss and syenite gneiss, both of which now have untwinned potassic feldspar instead of microcline.

Although the modified skarn is so conspicuously preferred as a host rock, 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 1 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 (1951, personal communication) 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 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 potassic 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 is very similar to the sphalerite-galena deposits of Broken Hill, New South Wales (Andrews and others, 1922; Gustafson and others, 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. (See pl. 20.) 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 and slightly silicated marble 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 mafic minerals, 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. (See table 2, No. 14, for optical properties and spectrographic analysis of the sericite.) 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 millimeters in diameter, coated with epidote, zeolites, and other minerals. These leached zones are confined exclusively to the granite gneiss sheet; they are not found in the metasedimentary wallrocks. 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 subore and even some low-grade ore. Granite pegmatite and highly biotitic granite gneiss are seldom mineralized. Most material of ore grade replaces gangue consisting of medium-grained to coarse aggregates of reddish-brown or yellowish-brown spessartite; quartz; untwinned potassic feldspar, colorless and glassy; pink, green, or purple fluorite; and white barite. Typical untwinned potassic feldspar from the Parish deposit has the composition $Or_{90.9}Ab_{6.8}Cn_{2.3}$; it contains 1.18 percent BaO. (See table 1, No. 12, for chemical analysis of feldspar; table 2, No. 2, for optical properties of the barite.) All these gangue minerals are highly variable in proportion. Small quantities both of iron oxides and of gangue minerals have been found replacing recognizable granite gneiss whose microcline shows progressive transformation to untwinned potassic feldspar. However, where either ore minerals or spessartite and fluorite are concentrated, the "host" is quartz, untwinned potassic 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 reworked or replaced by quartz and feldspar before introduction of the other gangue and metallic minerals. 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.

The writer regards the spessartite as a hydrothermal product introduced as part of the ore-forming process but deposited before the iron oxides. Spessartites have been synthesized, in the presence of water vapor, at temperatures as low as 500° C (Michel-Lévy, 1951). In that series of experiments, higher temperatures were required for the production of manganiferous almandites. In a dry system, the melting point of pure synthetic spessartite is 1,195°±2°C (Yoder and Keith, 1951, p. 524). Deductions made from the local geologic evidence are consistent with the experimental evidence that spessartite can be produced in a watery environment at temperatures near the upper limit of the hypothermal range.

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 $Spess_{61}Alm_{16}Andr_{14}Gross_5Pyr_4$; it contains 25.66 percent MnO. (See table 1, No. 3, for chemical analysis.) The only known spessartites closely approximating the Parish garnet in composition come from Broken Hill, New South Wales. To date, no other Mn mineral has been positively identified from the Parish 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.

Modes of some common mineralogical types of ore are given in table 8, and chemical analyses of ore are given in table 9. Comparable analyses of low-grade or marginal ore, averaging about 22 percent Fe, are not available.

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.

TABLE 8.—*Modes of modified granite gneiss ores and mineralized gangue, Parish deposit* [1]

	P 26.4	P 15.1	P 3.1	P 15.12	P 15.3
Quartz	7.5	43.5	14.8		29.8
Microcline	Tr.				
Untwinned potassic feldspar	60.9	.5			1.7
Plagioclase	3.6				
Muscovite	.4	Tr.	Tr.		.5
Biotite	2.5	.2		Tr.	3.8
Sillimanite	2.2				
Spessartite	Tr.	16.8	Tr.	.6	50.5
Magnetite	22.6	18.4	75.4	52.5	3.1
Hematite [2]	.2	6.5		44.5	4.1
Pyrite	Tr.		Tr.		Tr.
Fluorite	Tr.	8.1	3.6	Tr.	.2
Barite	Tr.?	[3] 6.0	6.2	2.4	.3
Apatite	Tr.				Tr.
Zircon	Tr.	Tr.	Tr.		
Sphene	Tr.				
Allanite	Tr.?			Tr.	
Tourmaline	Tr.	Tr.			
Chlorite			Tr.	Tr.	[4] 6.0
Carbonates		(3)		Tr.	
Clay minerals	Tr.		Tr.		(4)
Total	100.0	100.0	100.0	100.0	100.0

[1] The specimens represent common types, but the selection does not include all major types for this deposit or show the enormous variation within a given type.
[2] Primary crystalline hematite; rare traces of martite in some specimens.
[3] Includes a little dark, highly birefringent material thought to be siderite.
[4] Includes a little unidentified clay mineral.

Specimen No.	*Description*
P 26.4	Low-grade magnetite ore in granite gneiss; host virtually unmodified(?) except for development of untwinned potassic feldspar. Hole 26, depth 138 ft. (For analysis of potassic feldspar and description of host rock, see table 1.)
P 15.1	Low-grade magnetite ore with subordinate hematite in quartz-rich gangue containing considerable spessartite, fluorite, and barite. Hole 15, depth 53 ft.
P 3.1	High-grade magnetite ore with relics of quartz, fluorite, and barite. Hole 3, depth 66 ft.
P 15.12	High-grade magnetite-hematite ore with relics of spessartite and barite. Hole 15, depth 99 ft.
P 15.3	Spessartite-quartz rock weakly mineralized by magnetite and hematite. Hole 15, depth 62 ft.

TABLE 9.—*Analyses of ore, Parish deposit*

[Weight percent. Data from Walker, 1943, p. 519.]

	Mill ore	Manganiferous ore
Fe	39.70	55.80
P	.065	.026
S	.030	.261
TiO_2	.24	.20
Mn	4.70	9.92
SiO_2	23.30	7.80

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. Parts 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. Many magnetite and hematite grains show no exsolved ilmenite. 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.

Chemical analyses of two varieties of hematite are given in table 10. The discussion that follows is contributed mainly by Buddington (1955, written communication). The first variety, subtitanhematite, shows no exsolution bodies, yet it contains 4.77 percent TiO_2. Subtitanhematite is Fe_2O_3 with less than 5 percent but generally more than 1.5 percent TiO_2 in solid solution. In the analyzed Parish hematite (table 10, No. 1), the TiO_2 is presumably present as dissolved ilmenite, rather than as dissolved rutile, for the quantity of FeO is slightly more than sufficient to combine with TiO_2 to form ilmenite. (But note again that the

TABLE 10.—*Chemical composition of hematite, Parish deposit*

[Analyst, J. J. Fahey, 1955]

	(1)	(2)
	P 25.1	P 15.2
Fe_2O_3	87.49	
FeO	4.76	
TiO_2	4.77	10.93
MnO	.16	
Insoluble		6.0
Sum	97.18	

No. in table	*Specimen No.*	*Description*
1	P 25.1	Hematite (variety subtitanhematite) from marginal ore in modified granite gneiss, hole 25, depth 241 ft. Analyzed material represents nonmagnetite fraction of iron oxides. Hematite ~ 15 percent by volume of rock, is the only opaque mineral visible in polished section; it shows rare twin lamellae but no exsolution bodies. A small number of hematite grains have abundant minute pits, regularly alined, but no foreign bodies that might produce these pits are detectable with a high-power oil immersion objective. The other hematite grains are unpitted. In thin section, rock is medium grained and foliated; grains are irregular in shape. In addition to hematite, rock consists dominantly of quartz, untwinned potassic feldspar, and microcline. Spessartite <5 percent. Several percent pale green mica. Accessory minerals: apatite, fluorite, and traces of sillimanite(?), muscovite, and rutile [possibly goethite(?)]. Secondary minerals: chlorite (very sparse), pumpellyite (trace), and a clay mineral (trace).
2	P 15.2	Hematite (variety titanhematite), very slightly ilmenitic, from modified granite gneiss, hole 15, depth 58 ft. Analyzed material (weight 0.1939 g.) represents nonmagnetite fraction of iron oxides. Opaque minerals (only a few percent of the rock): hematite, subordinate magnetite, and an unidentified semiopaque mineral. Much of the hematite shows no exsolution bodies; twin lamellae are rare. Some hematite contains blades and hairs of exsolved ilmenite, and a few grains contain minute bodies of ilmenite(?) visible only with a high-power oil immersion objective. There is a suggestion of yellowish internal reflection in some of these very fine exsolution bodies; such bodies may be rutile. The magnetite shows slight local alteration to veinlets, patches, and crude blades of martite. The unidentified semiopaque mineral is isotropic, has a brown internal reflection, forms crusts on zeolite(?) crystals, and partly surrounds some hematite grains. It might be goethite or rutile; the former seems more likely. In thin section, the rock is a fine-grained, nearly equigranular aggregate of polygonal and irregular quartz and untwinned potassic feldspar. A little spessartite and brown biotite (generally chloritized) are present, as well as the opaque minerals described above. Apatite is the chief accessory. Traces of zircon, allanite, muscovite, pumpellyite, and a zeolite are also present.

"chemical" ilmenite is accommodated within the hematite lattice; it has not been exsolved to form the discrete mineral ilmenite.) Subtitanhematite is the principal nonmagnetic iron oxide in the Benson ore. The low MnO content of the Parish hematite confirms our other chemical and mineralogical data that show spessartite, rather than an iron oxide, to be the source of nearly all the Mn in the Parish ore. Yet low as the MnO content of the Parish hematite is, it is highly concentrated relative to the MnO found in two samples of accessory hematite from microcline granite gneisses of the district. Each of these samples had less than 0.01 percent MnO.

The second variety of Parish hematite, titanhematite (defined as Fe_2O_3 wtih 5 to 10 percent TiO_2 in solid solution), contains sparse bodies of exsolved ilmenite and perhaps a trace of rutile. The known bulk of the exsolution bodies is far too slight to account for the high TiO_2 content of the analyzed hematite; therefore, most of the TiO_2 must be present in solid solution. Titanhematite is characteristic of the least titaniferous hematites occurring as accessory minerals in microcline granite gneisses of the St. Lawrence County district.

Pure magnetite from the Parish deposit has not been analyzed. However, the analyzed magnetic fractions of Parish ore are known to be low in Mn and Ti.

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 are occasionally found with bornite and chalcopyrite. Several discrete grains of valleriite(?) are present in one specimen.

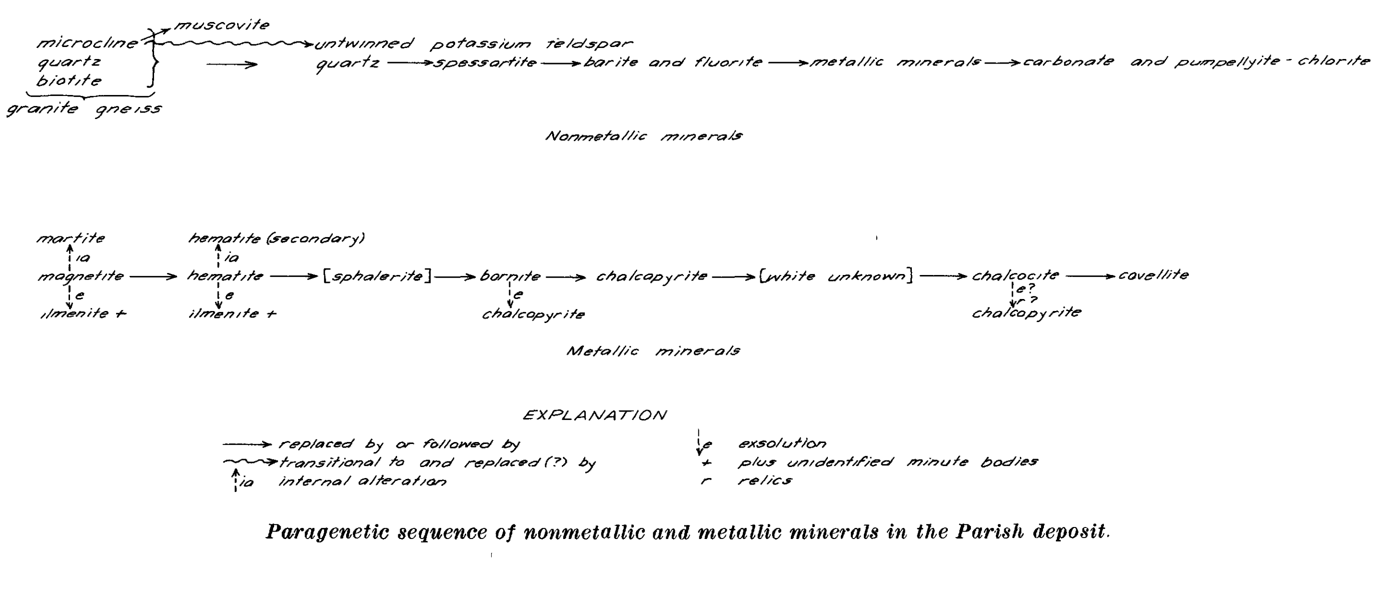
The paragenetic sequence of the nonmetallic and metallic minerals is shown below.

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 metallic minerals are given in a separate section.

MINERALOGY AND PARAGENETIC SEQUENCE OF THE METALLIC MINERALS

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, the opaque, generally metallic minerals are magnetite, hematite, pyrite, pyrrhotite, chalcopyrite, sphalerite, molybdenite, and bornite—all recognizable megascopically; and ilmenite, marcasite(?), chalcocite, covellite, vonsenite, ilvaite, loellingite, colloform graphite, unidentified minerals, valleriite(?), and cuprite(?)—all generally detectable only under the reflecting microscope. The opaque but nonmetallic mineral graphite is present locally in diopside skarn and silicated marble. 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 num-



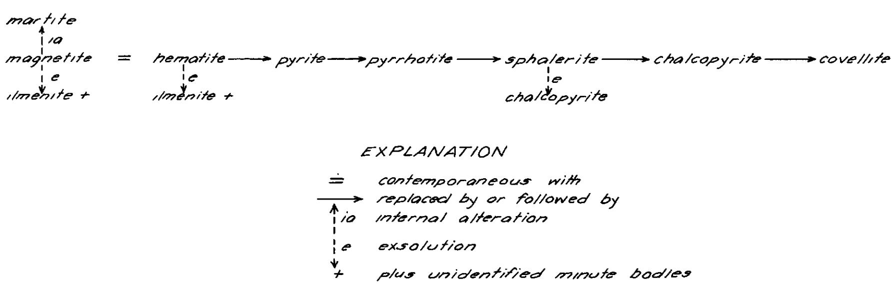
Paragenetic sequence of nonmetallic and metallic minerals in the Parish deposit.

ber 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 in substantial amounts has been found only at the Parish deposit. The other metallic minerals 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 a small mass of chalcopyrite found in one magnetite deposit (Millar, 1947, p. 12), but discrete Au- or Ag-bearing minerals have not been detected. Maghemite, very likely supergene, has been found in just one deposit.

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 an 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 and chalcocite (except for traces) 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, in the accessory sulfides.

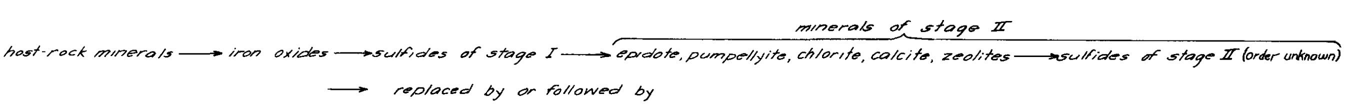
The paragenetic sequence of metallic minerals for most deposits of the district is shown below.



Paragenetic sequence of metallic minerals in the granite gneiss ores.

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 metallic minerals occur in every deposit of the appropriate type, but usually only one of them is missing. The position of several rare metallic minerals is discussed following the description of those minerals. The paragenetic sequence for the Parish deposit has been given separately.

Some deposits show slight additional mineralization after the bulk of the sulfides was formed. This second generation of minerals (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 to 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 shown below.



Position of sulfides of stages I and II in the paragenetic sequence

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 metallic minerals 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 veinlike patterns for certain sulfides of stage I are more common than for oxides. The evidence of metasomatic replacement improves with increasing quantity of the metallic minerals; small quantities of metallic minerals 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 mineral 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 mineral 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 mainly 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 METALLIC MINERALS

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 metallic minerals have been given in descriptions of ore types. A detailed account of microscopic features of the metallic minerals in the Trembley Mountain deposit is given separately. (See p. 212–216.)

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 ($\sim$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 metallic minerals, magnetite seems to have developed first. Whereas 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 preexisting 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—for example, a salitic pyroxene skarn with calcite relics, or a granite gneiss with biotite and garnet—the magnetite commonly 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, felsic minerals, apatite, fluorite, barite, or other minerals appropriate to the particular host. These included relics, particularly if they are very small, may give a badly pitted surface to individual grains in a polished section.

In addition to inclusions, many magnetite grains 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 mineral thought to be spinel. These minute bodies are regularly alined along either the octahedral or cubic direction in magnetite; locally, both sets are present. Similar bodies are noted in Schneiderhöhn and Ramdohr (1931, p. 581–582) and have been described briefly by Hotz (1953), Sims (1953), and others.

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 hypogene. (For tentative identification of goethite, see p. 47.)

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 very small quantities of Ti, Al, Mn, and Mg are present within the crystal lattice of some "pure" magnetite.

Rare grains of magnetite—containing no detectable zircon, monazite, or other normally radioactive accessory minerals—are radioactive, as indicated by pleochroic halos developed in biotite that touches this kind of magnetite.

Narten and McKeown (1952, p. 13) independently observed that accessory hematite from "red granite" (alaskite?) 1.5 miles east of Benson Mines is weakly radioactive. They also note (1952, p. 23 and table 2) that ilmenite and pyrite from certain pegmatites of the Adirondacks are radioactive.

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 potassium-rich rocks. Sillimanite is a very characteristic but not ubiquitous associate of the hematite.

It is undesirable to call the primary crystalline hematite "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 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; 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 magnetite; and hematite grains lie at random athwart magnetite aggregates having a definite planar orientation. However, Buddington does not believe the relations convincingly demonstrate that hematite is younger than magnetite in this deposit. Everywhere else in the district, 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 primary hematite is very rare. This notable feature at once sets apart the primary crystalline hematite from martite. (See below.) The 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. In the Dover district, New Jersey, primary crystalline hematite is concentrated locally in the Scrub Oaks magnetite deposit (Sims, 1953, p. 278). Hotz (1953, p. 206–207) noted hematite, apparently primary, in certain magnetite ores of the Sterling-Ringwood area, New York and New Jersey.

Primary crystalline hematite can accommodate considerable TiO_2 in solid solution, as dissolved ilmenite, dissolved rutile, or other ideal minerals. In places, part of this TiO_2 has been exsolved to form discrete mineral phases, but the data at hand suggest that even here a major part of the TiO_2 has been retained within the crystal lattice of the hematite itself. (Cf. table 10 and discussion following.)

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 at that 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. More data are required before this apparently distinctive difference can be stated as a rule. Moreover, the martite found among accessory dark oxides in rocks of the district contains very little dissolved TiO_2 (Buddington, 1955, personal communication). Thus there appear to be distinct chemical as well as physical differences between martite and primary crystalline hematite.

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, p. 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 each other.

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, 1952, p. 41). It is present in certain magnetite ores of the Dover district, New Jersey, but is not regarded as an economically important constituent (Sims, 1953, p. 278). Hotz (1953, p. 207) mentions its occurrence in the Sterling-Ringwood area, on the New York–New Jersey border; and Sims and Leonard (1952, p. 27–28) have described and figured martite from the Andover district, New Jersey. These are but a few of the many occurrences that come to mind. Some other iron deposits, commonly said to have abundant martite, probably contain much primary crystalline hematite along with, or perhaps instead of, true martite.

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 Precambrian 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 occurence 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} plane 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 mineral 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 quantity 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 about 2.8 percent, averaging about 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 wrote:

> 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 (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, and Postel (1952, p. 41) observed ilmenite in samples of ore from neighboring magnetite deposits of the Clinton County district. Colony (1923, p. 114 and plate 11, fig. 3) reported ilmenite from the Canopus mine, southeastern New York. More than 50 years ago, Bayley (1910, p. 115) inferred that ilmenite, either modal or normative, was present in some ores from the Highlands of New Jersey; later, Sims (1953, p. 277–278) and Hotz (1953, p. 206; 1954, p. 212) found ilmenite in polished sections of magnetite from a number of magnetite deposits in the Highlands.

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. By 1951, little had 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 with an ultramafic differentiate of the quartz syenite gneiss series. These titaniferous ores are considered very briefly in a subsequent section.

VONSENITE

Vonsenite (ferrous ferric borate)[13] is found with magnetite in ore from the Jayville deposit, Oswegatchie quadrangle, and from the Clifton deposit, Russell quadrangle. Jayville is a deposit of the skarn type, differing from others in the district by having abundant fluorite-rich hornblende skarn and considerable anthophyllite skarn, as well as the usual pyroxene skarn. Biotite sköls are prominent at Jayville, where they carry much of the ore. Vonsenite has been identified in several specimens from the main ore zone. Chemical analysis of drill core from hole J2, 293.5 to 311 feet, shows that the crude ore contains 4.80 weight percent B_2O_3. This is equivalent to about 36 weight percent vonsenite, based on the B_2O_3 content of analyzed vonsenite from the deposit. From the known mineralogy of the cores, their high B_2O_3 content, and the high soluble Fe content of the nonmagnetic tails, the writer infers that vonsenite is a major constituent of this part of the ore zone.

Vonsenite from Jayville is fully described by Leonard and Vlisidis (1961). The pure vonsenite contains about 62 weight percent Fe and 13.4 weight percent B_2O_3, after deducting for slight impurities in the analyzed sample. Physical properties of the vonsenite are summarized below.

In hand specimen, the vonsenite looks the same as the fine-grained magnetite with which it occurs. The vonsenite forms black, granular aggregates with metallic (locally adamantine) luster. Powdered vonsenite will move when an alnico magnet is run along a glass tube containing the pure mineral. However, a piece of drill core containing 25 volume percent vonsenite and no observable magnetite is not attracted to an alnico horseshoe magnet whose poles are roughly 1 inch apart. The same piece of core, suspended on a string, swings readily in the field of the magnet but does not cling to the magnet. Therefore, a piece of core relatively rich in vonsenite can be distinguished from one having appreciable magnetite by the relatively weak magnetism of the vonsenite. The black streak of vonsenite will distinguish that mineral from crystalline hematite, for which the vonsenite might otherwise be mistaken in the Jayville deposit. Vonsenite admixed with magnetite cannot be recognized in hand specimen by the simple magnetic test; and if one were to have both vonsenite and hematite in a hand specimen, the crude streak test might well be ambiguous. Flame test, X-ray, or careful polished-section study would then be required to confirm the presence of vonsenite.

Vonsenite is completely opaque in thin section. If the thin section is viewed in incident light, one sees that vonsenite is darker than the associated magnetite. If one examines polished specimens of ore with the naked eye, one sees that vonsenite is bluish relative to magnetite, and its reflectivity or "brightness" relative to magnetite is about the same as the reflectivity of magnetite relative to crystalline hematite. In polished sections examined with the reflecting microscope, vonsenite shows the following properties:

Habit.—Generally as aggregates of polygonal grains; locally as lath-, wedge-, and diamond-shaped crystals, commonly with rounded corners. Apparently contemporaneous with, or slightly younger than, magnetite; cut by veinlets of crystalline hematite, and locally replaced by ragged blades of hematite.

Color.—Reflection-pleochroism very strong, light gray or light brown to light blue-gray.

Reflectivity.—Less than that of magnetite, but much greater than that of polished iron-rich pyribole.

Anisotropism.—Very strong or extreme, from black to fiery orange. Extinction parallel to long edges of laths, but symmetrical in wedges and diamonds. No internal reflection.

Hardness.—Scratching hardness F or greater, or about 6 on Mohs' scale. Polishing hardness less than that of hematite, and about equal to that of magnetite.

Magnetism.—Pure mineral powder readily attracted to magnetized sewing needle.

Etch tests.—HCl positive; aqua regia weakly positive; HNO_3, KCN, $FeCl_3$, KOH, $HgCl_2$, H_2O_2 negative.

X-ray powder data (*d*-values in Å; FeKα radiation).—5.16 strong, 2.58 very strong, 2.37 to 2.38 strong (G. G. L. Henderson, 1950, written communication; J. M. Axelrod, 1951 and 1958, written communication).

[13] Vonsenite and ludwigite form an isostructural series whose major chemical features may be expressed by the formula $(Mg, Fe^{+2})_2Fe^{+3}BO_5$. Members having Mg $>$ Fe^{+2} are termed ludwigite; those having Fe^{+2} $>$ Mg are termed vonsenite. Cf. Palache, Berman, and Frondel (1951, p. 321–324), where the Fe^{+2} end member is, according to W. T. Schaller (1955, written communication), inappropriately called paigeite.

Without X-ray and chemical data, the vonsenite would easily be mistaken for ilvaite. (Cf. Ramdohr, 1955, p. 811–812.)

Vonsenite, first described, analyzed, and named by A. S. Eakle (1920), has remained till recently a relatively obscure mineral. The original vonsenite occurred 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). The better known related borate, ludwigite, occurs in a number of pyrometasomatic deposits in Sweden, central Europe, the U.S.S.R., Korea, and the western United States, and is found sporadically elsewhere. Geijer (1939) has presented an excellent review of the Swedish occurrences, summarizing their petrologic significance in relation to the development of the iron ores of central Sweden. (Cf. this report, p. 83, 86.)

In St. Lawrence County, Jayville is the only magnetite deposit known to contain much vonsenite. Its occurrence there makes one wonder if other magnetite-bearing sköls of the district also carry this borate. 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 or a related borate accompanies the magnetite in some of this ore.

ILVAITE

Ilvaite is present in modified skarn at the Trembley Mountain deposit. The same mineral has been found in very small quantities in pyrrhotite-rich skarn from the Clifton mine. The ilvaite was earlier thought to be a borate, though obviously it was not vonsenite. The original small sample was used up in a vain attempt to identify the supposed borate, but ilvaite was positively identified in a check sample collected in 1958 by C. L. Rogers. The following account is based on the occurrence at Trembley Mountain. Here, ilvaite constitutes 15 to 20 percent of one specimen from hole T3, depth 137 feet, and is locally abundant at a depth of 137.6 feet. Specimens collected nearby fail to show ilvaite. For the deposit as a whole, ilvaite seems to be a rarity.

At 137 feet, the host is a quartz-bearing salite skarn that contains a little actinolite, biotite, and apatite. This modified skarn has been partly replaced by magnetite, ilvaite, and pyrite. The pyrite belongs to stages I and II. Ilvaite and magnetite seem contemporaneous; pyrite of stage II cuts the ilvaite; the relation between ilvaite and pyrite of stage I is indeterminate. The ilvaite commonly forms a parallel intergrowth with an unidentified brown micaceous mineral. In places, the relations suggest that ilvaite replaces the micaceous mineral.

Two generations of ilvaite are present in the specimen taken at 137.6 feet. The gross mineralogy is similar, except that a little chalcopyrite replaces pyrite I. Ilvaite I is molded on magnetite. Pyrite I locally replaces ilvaite I. Ilvaite II is threadlike, winding about the borders of earlier silicate grains and in places cutting across a number of grains. Some of the ilvaite veinlets are dotted with pyrite specks tentatively referred to stage II. Locally, ilvaite II cuts across magnetite grains, or surrounds and locally replaces pyrite I.

The ilvaite is normally opaque in thin section, though exceptionally thin parts are translucent. Here the mineral is distinctly prismatic and shows parallel extinction. The long direction of the prisms is very dark brown; the short direction is opaque. Presumably, such prisms are not elongate parallel to *c*. Absorption and birefringence are so strong that the orientation of fast and slow rays cannot be determined. The mineral is definitely biaxial; acute bisectrix sections are virtually opaque; the optic sign is very questionably positive.

In polished section, ilvaite from Trembley Mountain shows the following properties:

Habit.—Generally, polygonal grains that are aggregates of very small individuals differing in optical orientation; locally, acicular.

Color.—Reflection-pleochroism strong, pinkish gray to blue gray or greenish blue; dark gray to purplish red in oil.

Anisotropism.—Extreme, from fiery pinkish to blue and black; intensified in oil. No internal reflection. Extinction parallel to elongation of fibers. Where fibers are best developed, their long direction is blue and has a blue interference color; their short direction is pinkish gray and has a fiery pinkish interference color. However, many elongate grains are intergrowths of individuals of varying orientation.

Reflectivity.—Slightly greater than that of polished salitic pyroxene but much less than that of magnetite.

Hardness.—F or F+.

Magnetism.—Powdered mineral is not attracted to magnetized sewing needle.

Etch tests.—Negative to all standard etch reagents except HCl and aqua regia. With HCl, some grains are slightly etched by minute pitting or by accentuation of fibrous structure; other grains show a brown stain that does not wipe off. With aqua regia, there is a slight etch. With NaN_2–I_2 solution, negative.

The above properties, while suggestive of ilvaite, are not fully diagnostic. Therefore, the apparent angle of rotation and its dispersion were measured by the methods of Cameron and Green (1950). The values obtained were so like those of known ilvaites that the

Trembley Mountain mineral was pronounced ilvaite—almost with assurance. As a check, material collected by Rogers was separated for X-ray analysis. The writer's cursory measurements of some principal interplanar spacings (film No. D–5119, taken by A. J. Gude 3d) are:

	d (A)	I
	7.31	s
	6.56	m
	~3.87	m–w
	3.24	w
(broad)	2.84	vs
	2.72	vs
	2.68	s
	2.43	m–w
	2.18	s
	2.11	s

CuKα radiation; Ni filter. vs—very strong; s—strong; m—moderate; m-w—moderate to weak; w—weak.

The X-ray data show conclusively that the mineral is ilvaite. In the same specimens, a few platy grains look like ilvaite but behave differently when studied with the oil-immersion objective. In oil, the blue direction of these grains becomes dark blue, not purplish red. The grains are too small for measurement of the apparent angle of rotation. A chance fragment of the original ilvaite-rich specimen was crushed and taken up in concentrated HCl for a flame test. A faint, evanescent yellow-green flame suggested boron. Subsequently, W. T. Schaller (1955, personal communication) tested for boron but found none in a fragment of impure material submitted to him, and Fred A. Hildebrand (1955, personal communication) was unable to find the pattern of vonsenite, paigeite, or hulsite in an X-ray film of impure material. If there is a borate mineral with the ilvaite at Trembley Mountain, its presence has not been proved.

At Clifton, ilvaite is present in pyrrhotite-rich garnet-pyroxene skarn from hole 58, depth about 60 feet. The rare ilvaite occurs as grains near pyrrhotite and as thin areas partly rimming pyrrhotite. Locally, ilvaite is veined by pyrrhotite. Chalcopyrite, sphalerite, and colloform graphite are also present. For paragenetic sequence, see the description of colloform graphite.

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. Some of the "moderately anisotropic pyrite" may really be marcasite, but this seems doubtful in view of the crystal form assumed by some of the material.

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 to occur in substantial amounts only at the Parish deposit, 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.

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 metallic minerals. In parts of some deposits (Trembley Mountain North), disseminated chalcopyrite is present to the extent of 2 percent. 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, lea˙ing 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.

A trace of a soft blue metallic mineral, possibly chalcocite, is present with colloform graphite in one specimen from the Clifton mine.

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 flecklike alteration of chalcopyrite. All the covellite shows strong reflection-pleochroism 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

Discovery and identification

Traces of cobaltian loellingite ($FeAs_2$) have been found in a single specimen from the Clifton mine, hole

51, depth about 143 feet. Brilliant silvery-white loellingite occurs as specks with magnetite, abundant pyrrhotite, and colloform graphite in gray phlogopite-pyroxene-bearing calcitic marble from a weakly mineralized zone of light-green diopside skarn. The white metallic mineral, noted by the writer and P. E. Hotz in 1948, was identified by X-ray by J. M. Axelrod. Miss Janet D. Fletcher confirmed the identification by spectrographic analysis. The relatively pure samples used by Axelrod and Fletcher were prepared at Princeton University by Dr. G. G. L. Henderson under the direction of Professor Edward Sampson. Henderson (1950, personal communication) found Fe and As by microchemical methods and made a preliminary X-ray study of the material. He thought the mineral might be loellingite, but this possibility was rejected because the white metallic mineral seemed to be easily scratched in hand specimen. Because of the risk of losing the sparse and crumbly material in sawing, a polished section was not prepared till the microchemical, X-ray, and spectrographic work had been completed.

In hand specimen, loellingite shows conspicuous conchoidal fracture. The mineral powder, showing no evidence of steel filings, is somewhat magnetic. Loellingite is the only arsenide found to date in the magnetite deposits.

Optical properties and etch tests

The loellingite shows the following properties in polished section:

Color.—Galena-white, highly reflecting; little changed in oil. Reflection-pleochroism faint, from whitish to yellowish; intensified in oil.

Anisotropism.—Strong, from bluish gray to brownish gray; greatly decreased in oil.

Hardness.—F or F—, or about 6 on Mohs' scale.

Magnetism.—Pure powder is attracted to a magnetized sewing needle.

Twinning.—Locally present; very thin lamellae, two or three to a grain; some lamellae are crossed.

Etch tests.—HNO_3 forms brown stain and etches the mineral, accentuating the twinning. HCl, KCN, $FeCl_3$, KOH, and $HgCl_2$ are negative.

X-ray and spectrographic analysis

An X-ray powder diffraction pattern made by Axelrod (1951, written communication) indicated that the white metallic mineral was loellingite. Henderson (1950, personal communication) previously reported the following data obtained by X-ray spectrometer, using FeKα radiation:

Peak	$2\theta°$	*d* (A)
1	44.48	2.56
2	48.60	2.35
3	43.90	2.59
4	62.76	1.86

Spectrographic analysis of the loellingite by Miss Fletcher (1951, written communication), based on a special set of standards for Fe, As, and Co, showed:

Element	*Quantitative analysis (percent)*	*Semiquantitative analysis (percent)*
As	70	>5
Fe	26	
Co	3	1–5
Ca, Pb, Bi, Ni, Mg, Si, Al	--------	.x
Cu, Zn, Mn, Ti, Zr, Ba, B	--------	.0x
Cr	--------	.00x
Ag	--------	.000x

The following elements were looked for but not found: Sn, Be, Au, Pt, Mo, W, Ge, Sb, Cd, Tl, Ga, V, Y, La, Th, Cb, Ta, U, Sr, Na, and P.

Microscopic occurrence

In order of abundance, metallic minerals in the polished part of the loellingite-bearing specimen are: magnetite, loellingite, vonsenite(?), colloform graphite, pyrrhotite, sphalerite, chalcopyrite, and chalcocite(?). Relatively few grains of each mineral are present.

Loellingite replaces the gangue as specks and veinlets; locally, loellingite veins the magnetite and vonsenite(?).

The magnetite is extensively altered to martite, which occurs as irregular, ragged, unoriented or roughly alined patches. Locally, the martite is bladed or acicular.

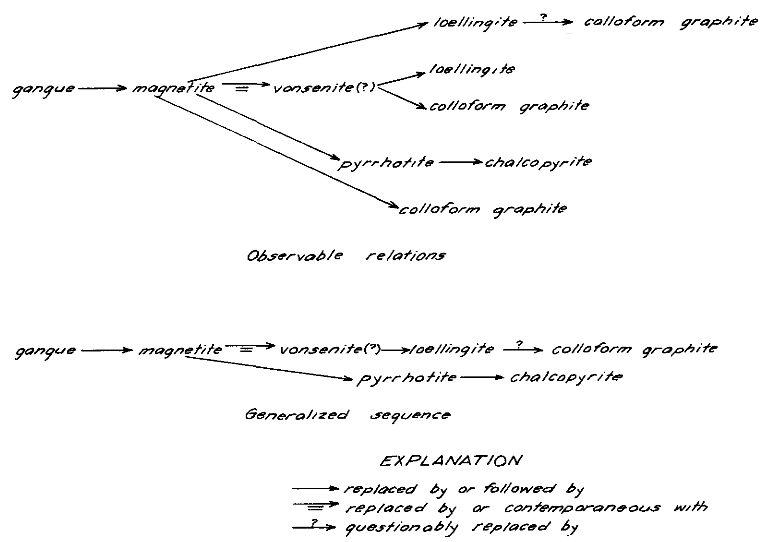
Vonsenite(?), generally granular, is crudely bladed in places. Relations between vonsenite (?) and magnetite are: (a) straight boundaries; (b) irregular, ragged intergrowths, each mineral having islands and embayments of the other; (c) veins of each cutting(?) the other. In general, the vonsenite(?) seems younger, but the two minerals may be essentially contemporaneous.

Pyrrhotite, locally with a little chalcopyrite, replaces the gangue and veins the magnetite.

The sphalerite occurs as networks and shreds in gangue.

Colloform graphite replaces gangue, veins and includes magnetite, and locally seems to replace the borders of loellingite grains.

The observable relations and generalized paragenetic sequence are shown below.



Paragenetic sequence in the loellingite-bearing specimen, Clifton mine

The positions of sphalerite and chalcocite(?) are unknown. The chalcocite(?) is present merely as a trace, intergrown with colloform graphite.

COLLOFORM GRAPHITE

Megascopic features

Colloform graphite, associated with loellingite, was found in 1948 by Leonard and P. E. Hotz in silicated marble at the Clifton mine, hole 51, depth about 143 feet. (For further details, see description of loellingite.) The graphite forms gray-black "microbotryoidal" masses of dull metallic luster, a few millimeters across, embedded in calcite and forming a rim between pyrrhotite and calcite. The graphite is sectile; it is rather whitish where scratched; and it leaves a gray mark on white paper. Identification of the mineral as graphite is based on optical and X-ray data given below.

The colloform graphite is distinct in habit and association from the disseminated, flaky graphite found in some metasedimentary rocks of the district. The colloform graphite, whatever its origin, is part of the sequence of ore minerals; the flaky graphite in the metasedimentary rocks is not.

Optical properties and etch tests

In polished section, the graphite consists almost wholly of aggregates of minute fibers showing the following properties:

Color.—Reflection-pleochroism very strong, bronze to gray; in oil, bronze (slightly pinker than in air) to black. Fibers are gray parallel to their length and bronze perpendicular to it. In conventional optical and crystallographic notation, $E=c=$gray, $O=a=$bronze; optically (—); elongation ‖ c. The bronze is distinctly darker and grayer than that of pyrrhotite. The gray is almost like that of magnetite.

Anisotropism.—Extreme, from burnished bronze to gray and black; in oil, orange-bronze to black. Extinction: with nicols set at 90°, four positions per 360° rotation (parallel and perpendicular to the elongation of the fibers); with nicols almost crossed, at an angle smaller than 78°, two positions (parallel or subparallel to fiber elongation). Thus with nicols nearly crossed, the extinction position corresponds to the "gray" position of fibers as seen in light from one nicol only.

Reflectivity.—Lower than that of pyrrhotite (and sphalerite?); higher than that of calcite. In "gray" position, nearly equal to that of basal plate of phlogopite; but in "bronze" position, much higher. Preliminary measurements made in 1959 with a Hallimond microphotometer give, for green light, $R{\sim}6$ percent for the "gray" position and $R{\sim}18$ percent for the "bronze" position. The values are only approximate, owing to the extremely small size of the graphite fibers and the difficulty of matching colors for the "bronze" direction in light of this wave length.

Hardness.—Very easily scratched by needle. Polishing hardness is equal to or less than that of pyrrhotite.

Etch tests.—Results are spurious.

X-ray data

Dr. G. G. L. Henderson, who undertook for the writer a preliminary investigation of the mineral under the direction of Professor Edward Sampson, reports (1950, personal communication) the following results obtained by X-ray spectrometer, using Fe $K\alpha$ radiation:

Peak	2θ°	d(A)
1	33. 45	3. 37
2?	30. 20	3. 72 (β of 3. 37)
3?	70. 20	1. 68

The X-ray powder diffraction pattern obtained by J. M. Axelrod (1951, personal communication) suggested graphite slightly contaminated by magnetite and a carbonate.

The samples X-rayed by Henderson and Axelrod were selected megascopically, before a polished section had been made. Certain apparently anomalous properties of the alleged graphite made a further check desirable. Therefore, a third sample for X-ray was carefully gouged from a typical spherulite of fibrous graphite, well shown in the polished section. This sample was removed and X-rayed by Fred A. Hildebrand, who identified the unknown as graphite with traces of calcite(?) and magnetite. He reported (1959, written communication) the following cursory measurements for graphite lines in film no. D–5480:

d (A)	I	d (A)	I
(β of 3.36)3. 71	vvw	(broad)1. 728e	w
3. 36	vvs	1. 675	ms
2.67e	vvw	1. 484e	vvw
2. 28e	vvw	1.230	mw
2. 13	vw	1. 157	w
2. 04	vw	1. 117	vvw

CuKα radiation; Ni filter. vvs—very, very strong; ms—moderately strong; w—weak; etc.

Leonard thinks that the lines marked "e" are extraneous and due to unavoidable contamination of the sample by a little pyrrhotite(?), calcite(?), and magnetite (?).

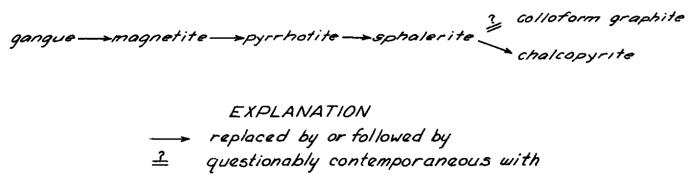
Microscopic occurrence

The graphite, almost everywhere minutely fibrous, occurs as veinlets; as rims on pyrrhotite; as polygonal or rounded, zoned, spherulites and aggregates of spherulites; and very commonly as colloform masses. The spherulites are about 0.2 to 1 mm in diameter, the larger ones having two or more concentric shells of graphite. Colloform aggregates in which shells of graphite alternate with those of another mineral, commonly sphalerite, are mostly about 0.05 to 0.5 mm wide. Individual graphite shells within the colloform aggregates are generally less than 0.1 mm wide; some of them are only a few microns wide. The graphite fibers are generally subperpendicular to the borders of the veinlets, rims, spherulites, or colloform masses; but in places the fibers are inclined as much as 45° to the borders. Some colloform masses are so arranged that layers of fibers seen edge on, alternate with layers of fibers seen end on. Upon slight rotation of the microscope stage, the first layers look chatoyant; the second layers have a moiré appearance. Most of the fibers are only a micron or two in width. The thickest are about 10 microns. In a few places, one finds a small, ragged-edged area of graphite that consists of tens of slightly curved, parallel plates, instead of fibers.

In specimen CN–51.1, where the mineral was first seen, the associates are mainly pyrrhotite, magnetite, and sphalerite. Some flecks of chalcopyrite are present. Loellingite (previously described) is found in part of the specimen, along with vonsenite(?).

The graphite veins and replaces magnetite, cuts pyrrhotite locally, and replaces calcite and phlogopite. Graphite is intergrown with sphalerite; the two minerals may be contemporaneous.

The paragenetic sequence observed for the main part of the specimen is:



Paragenetic sequence in the graphite-rich specimen, Clifton mine

(For additional data, see paragenetic sequence inferred in discussion of loellingite.)

Colloform graphite was found microscopically in another specimen from Clifton mine, in garnet-pyroxene skarn from hole 58, depth about 60 feet. This specimen, containing more than 50 percent pyrrhotite, has sparse chalcopyrite, rare ilvaite, a trace of sphalerite, and a little graphite. The paragenetic sequence for opaque minerals in this specimen is:

ilvaite⟶pyrrhotite⟶sphalerite⟶chalcopyrite

The position of graphite in the sequence is indeterminate.

Comparison with other graphites

Colloform and spherulitic graphite from the Clifton mine has many features likely to embarrass a mineragrapher. These are: gross habit, position in the paragenetic sequence, anomalous elongation, apparent magnetism, apparent, tarnish, and spurious etch reactions.

A colloform habit for natural, terrestrial graphite is probably very rare, according to Ramdohr (1955, p. 299). One such occurrence, from the Aurora mine, Kaveltorp, central Sweden, was earlier figured by Schneiderhöhn and Ramdohr (1931, p. 19, fig. 2). Botryoidal, almost colloform graphite formed in cracks in the masonry of coke ovens is described and illustrated in Ramdohr (1955, p. 299–301); this is the so-called "retort graphite," to which the Kaveltorp occurrence was likened. Ontoyev (1955) described and figured graphite occurring as thin crusts of radiating individuals on fragments of stony coal enclosed in quartz-tourmaline veinlets.

Natural, terrestrial graphite as spheres, spheroids, and spherulites of various sizes is mentioned in a number of reports. Wesselowski and Wassiliew (1934, p. 163) give a general classification and discussion of such forms, though mineragraphic data are lacking. Spherulitic graphite also occurs in iron meteorites (cf. Ramdohr, 1955; and Perry, 1944, esp. p. 21) and in foundry products such as cast iron and pig iron (see, for example, Morrogh, 1941).

The colloform graphite at Clifton was deposited relatively late in the main stage of mineralization. A search of the literature on a wide variety of graphite occurrences suggests that late crystallization of graphite is rather common, except for graphite directly attributable to the metamorphism of sedimentary carbonaceous material.

A fibrous habit for graphite is not rare, though mineragraphic studies of it are. Clear and beautifully illustrated accounts of fibrous, spherulitic graphite in artificial irons are given by Morrogh (1941) and by Morrogh and Williams (1947, 1948). Morrogh and Williams (1947, p. 326) find that graphite fibers in the spherulites generally are elongate parallel to *c*, and that the *c*-axes of the fibers coincide with the radii of the spherulites. In the Smithville meteoritic iron, graphite fibers are also elongate parallel to *c* (Ramdohr, 1955, p. 299). Fibers of Clifton graphite are likewise elongate parallel to *c*, since E is parallel to the fiber axis. These relations are the reverse of those found in platy or scaly graphite, and in "retort graphite," where the elongation of graphite individuals is perpendicular to *c*.

Some samples of apparently pure colloform graphite from Clifton were attracted to a magnetized sewing needle. The samples were removed with a sharp, unmagnetized needle while the observer looked through the microscope, and the magnetism of the powder was then checked microscopically. The writer concludes that, in spite of their apparent purity, such samples contained magnetite or pyrrhotite gouged from foreign grains concealed beneath the graphite.

Scratches made on colloform graphite in unpolished drill core grow dark after a few hours. Since a polished section showed no tarnishing of graphite or its associates after several years' exposure to the air, the darkening of unsmeared scratches on unpolished core is interpreted as a "healing" due to the slow return of graphite fibers to a less disturbed position.

Etch reactions observed on the polished graphite are spurious. The reactions are appropriate for the associated sulfides, some of which were unavoidably in contact with the very small drops of reagent placed on areas of graphite.

MISCELLANEOUS METALLIC MINERALS

Maghemite (ferromagnetic Fe_2O_3) has been recognized in magnetite-rich skarn from the Green Farm deposit; the mineral is most likely supergene. (See p. 129 for details.) The writer has not found maghemite anywhere else in the district.

Several metallic minerals of uncertain identity, including valleriite(?), cuprite(?), and others, have been found in polished sections of the ores. These minerals are present only in very small amounts, but some of them are interesting for their genetic significance. Notes on several of the minerals follow. The information is presented in the hope that other workers may find these minerals in greater abundance, identify them conclusively, and fit them into the paragenetic sequence.

VALLERIITE

Pale-yellow valleriite(?) is present in one specimen from the Clifton mine and one from the Parish deposit. Valleriite is a Cu-Fe sulfide of uncertain composition, perhaps $Cu_2Fe_4S_7$ or $Cu_3Fe_4S_7$ (Palache, Berman, and Frondel, 1944, p. 235). In the Clifton specimen (CN 44/12.1, hole 44/12, depth 12 feet) of pyrite-rich garnet-pyroxene skarn, valleriite(?) occurs as traces in gangue surrounded by pyrite.

Habit.—Minute shardlike and angular grains.
Color.—Reflection-pleochroism strong, pyrite-yellow to gray.
Anisotropism.—Strong, from black to light steel-gray.
Reflectivity.—In "yellow" position, greater or equal to that of pyrite; in "gray" position, less than that of pyrite.
Hardness.—Polishing hardness much less than that of pyrite; grains too small to scratch.
Twinning.—Polysynthetic; resembles the albite twinning of plagioclase; "pericline" twins present locally.

The valleriite(?) is confined to one small area of the polished section. The abundant pyrite, generally occurring as granular aggregates, locally has the form of elliptical, zoned grains. Small grains of chalcopyrite are common. Sphalerite is sparse.

Valleriite(?) from the Parish deposit (specimen P15.3, hole 15, depth 62 feet) is found in hematite-bearing spessartite rock that contains a little magnetite. Other parts of the core from this drill hole contain several percent of bornite locally.

UNIDENTIFIED YELLOW METALLIC MINERAL

Traces of an unidentified yellow metallic mineral are found in several sulfide-rich specimens of skarn from the Clifton mine.

Habit.—Blades, shreds, and elongate grains in pyrrhotite.
Color.—Pale yellow to cream. Reflection-pleochroism nil.

Anisotropism.—Similar to that of pyrrhotite, but less rose.

Hardness.—Scratching hardness perhaps E. Polishing hardness greater than that of pyrrhotite.

Twinning.—None observed.

Etch tests.—HNO_3 forms slight brown stain and etches the mineral slightly accentuating scratches; no effervescence. $HgCl_2$ forms light-brown stain, removed by hard rubbing; no etching. HCl, KCN, $FeCl_3$, KOH, aqua regia, H_2O_2—all negative.

Microchemical tests.—Reaction with NaN_3–I_2 solution uncertain but probably positive, indicating the presence of sulfur. Uncertainty is due to contamination of unknown by adjacent pyrrhotite. Test for Ni inconclusive, owing to difficulty of powdering minute grains of the relatively hard mineral.

From the observed properties, one suspects millerite or perhaps marcasite, but the mineral remains unidentified.

Metallic associates of the unknown are pyrrhotite (dominant), chalcopyrite (sparse), pyrite (very sparse), and sphalerite (rare). A trace of this unidentified yellow metallic mineral may be present in the sample that contains loellingite and colloform graphite.

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:[14] analcime, apophyllite(?), calcite*, chabazite*, chalcopyrite, chlorite*, datolite(?), epidote*, fluorite, kaolinite(?), pectolite(?), prehnite(?), pumpellyite*, pyrite*, quartz, sericite, specularite(?), sphalerite, stilbite, and uninvestigated minerals throught to be zeolites. Optical properties of some of these minerals are given in table 2, Nos. 8, 9, 19, 20, and 21.

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, whereas calcite, pyrite, and traces of other metallic minerals are late. (See preceding section on paragenetic sequence of metallic minerals.) 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) 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: (a) as sporadic or pervasive internal alteration products and replacements of earlier minerals; (b) as linings of vugs; and (c) 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 wallrock, 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 parageneses 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 wallrocks at a number of deposits. The vugs are usually more abundant in the ore and its immediate wallrocks, but they

[14] (?) The mineral has been identified megascopically but has not yet been checked microscopically.

*The mineral is common and frequently prominent.

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 at most a few millimeters long and a millmeter or two wide, roughly circular, irregular, or slightly elongate, and are oriented with complete disregard for the foliation or any other structure in the host. The adjacent rock may be substantially altered for a few millimeters 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 (1947) to be leached cavities post diabase in age. The St. Lawrence County vugs cannot be miarolites:[15] 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 in many places singularly "clean," without earthy hematite and with only local, sporadic chlorite; they are not distributed in any manner that suggests supergene alteration; many vugs 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 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. p. 469–473); the Brödtorp zinc mine, Orijärvi district, Finland (Borgström, 1908, p. 334–337); and mines in the central Swedish district (Geijer and Magnusson, 1944—zeolites and other minerals, 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, 1953, p. 280). Traces of a mineral that may be pumpellyite were found in the Andover district, New Jersey (Sims and Leonard, 1952). 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 are given in table 2, Nos. 8 and 9. Aside from the New York and New Jersey magnetites, pumpellyite has been found (a) in amygdules and veinlets, and as an alteration product, in basic igneous rocks; (b) in pyroclastics and tuffaceous graywackes; (c) in chlorite schists, glaucophane schists, and other metamorphic rocks; and (d) in modified skarns. Many of the significant references are cited by Winchell (1951, p. 519). An extensive bibliography on pumpellyite is given by Coombs (1953, p. 133–135).

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), some kaolinite(?) (from sillimanite of granite gneiss ores, where a clay mineral locally veins the sillimanite), zoisite (from feldspar), dark mica (from pyriboles), and hydrogarnet(?) (rare—in skarns). One has the impression that such minerals are roughly 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 for the most part comformable 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

[15] 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.

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 lathlike or otherwise elongate habit of certain New Jersey and New York magnetite deposits is inconspicuous, though perhaps not lacking, in the St .Lawrence County magnetite district. Magnetite bodies of elongate habit have been described and figured by Postel (1952, p. 34—Clinton County, New York), Kemp (*in* Kemp and Ruedemann, 1910, p. 105–122, 128–129—Mineville-Port Henry district, New York), Colony (1923, p. 55–57—southeastern New York), Hotz (1953, p. 201; 1954, p. 212—northern New Jersey), Sims (1953, p. 274—Dover, New Jersey), Bayley (1910, p. 135–140—New Jersey Highlands), and others.

SIMPLE(?) TABULAR BODIES

One may expect to find relatively simple tabular or sheetlike bodies on the limbs of folds. The long, rather 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 downdip to an average distance of 250 to 400 feet. The dip of the Jarvis Bridge deposit changes gradually along strike from about 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 in unknown.

The writer has described these deposits as relatively simple tabular bodies. However, in the detailed accounts of the deposits, the possible existence of shoot structure related to plunging folds has been pointed out. The need for considering such structural "jokers" is borne out by the changing model of the magnetite ore body in granite gneiss at Lyon Mountain, Clinton County district, New York. The Chateaugay ore body at first appeared to be relatively simple for some distance along strike and downdip, 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 the Benson Mines, Dead Creek, Deerlick Rapids, Spruce Mountain Northwest, Trembley Mountain, Twin Lake Stream, and Wilson Mountain deposits. (See pls. 11, 12, 18, 19, and figs. 24, 26 for details.) The Hardwood Mill deposit seems to have a "double fishhook" or "flowing S" pattern in plan view (see pl. 8). Possibly the Outafit deposit has a "triple fishhook" shape (see pl. 10), though that interpretation is highly speculative.

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,

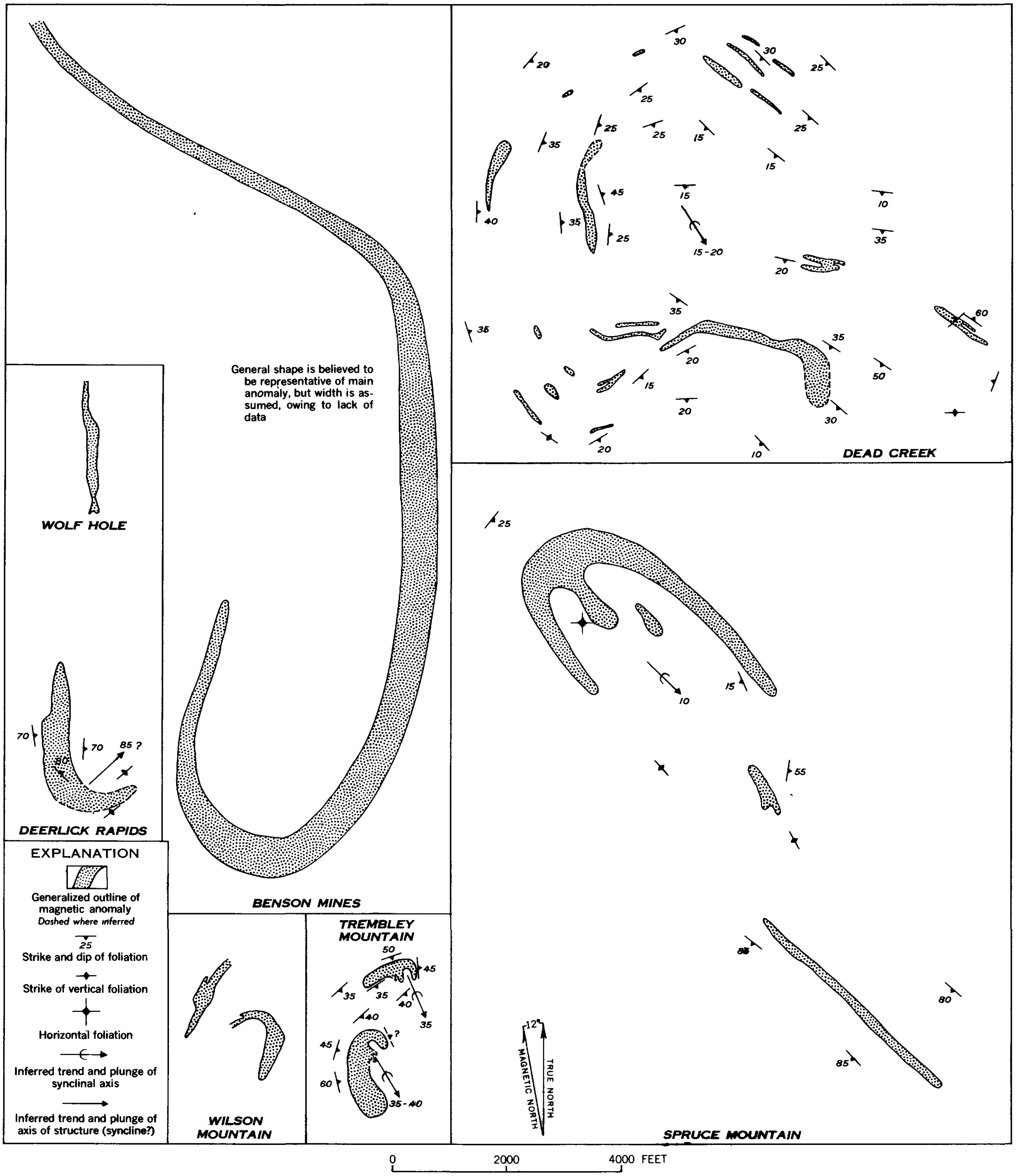


FIGURE 3.—Fishhook shape of magnetite deposits as shown by magnetic anomalies.

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 and others, 1950); Camlaren, Northwest Territories (Henderson and Fraser, 1948); Sherritt Gordon, Manitoba (Farley, 1948); Sullivan, British Columbia (Swanson, 1950); Arendal, Norway (Kjerulf and Dahll, 1861; Bugge, 1940); central Swedish district (Geijer and Magnusson, 1944); Dover district, New Jersey (Sims, 1953); Ducktown, Tennessee (Emmons and Laney, 1926); Edwards-Balmat district, New York (Brown, 1936a); Franklin, New Jersey (Spencer and others, 1908); 81 mine, Lyon Mountain, New York (Postel, 1952, p. 34, 81).

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. (Cf. Postel, 1952, p. 34.) Elsewhere, anticlinal noses may be favorable sites for magnetite deposition. For example, in the Sterling-Ringwood district, New York–New Jersey, Hotz (1953) reports the occurrence of ore on the noses of overturned anticlines as well as synclines. Sims (1953, p. 272) states that a mineralized zone wraps around the plunging 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 (1952, p. 34) found that in the Clinton County district, New York, several major magnetite deposits were on the overturned limb of isoclinal synclines. He suggested that greater shearing in the rocks of the overturned limb might have given 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? The question, as phrased, presupposes that both limbs have survived deformation in something like complete form. As Engel (1951, personal communication) pointed out, the two limbs have not been so preserved at such places as Balmat and Edwards, New York; Dover, New Jersey; and Broken Hill, New South Wales. Instead, the folds have been sheared out and pulled apart, or one limb has been squeezed nearly to extinction. Similar relations are inferred for many magnetite deposits of fishhook shape in the St. Lawrence County district, but lack of outcrops and scarcity of drill holes keeps the geologist from establishing those relations firmly. Engel (1951, personal communication) further remarked that the more intensely deformed, sheared-out limb is not the one replaced by ore at Balmat and Edwards. Yet the sheared-out limb, or the surface of discontinuity left where a limb has been torn off, would seem to approximate the "shear zones," locally richly mineralized, of so many other mining districts.

LINEAR BODIES

The apparent absence of distinct lathlike or pencillike 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. (See pls. 2, 9.) At both deposits, a prominent *b* lineation, parallel to the axes of local folds, is seen in some wallrocks. 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 Spruce Mountain Northwest, the presence of elongate lobes of magnetic "highs," parallel to the axis of the syncline, suggests linear ore shoots beneath. (See pl. 18.) At Dead Creek No. 1, an elongate area of low positive readings at the southeast end of the magnetic anomaly suggests the edge(?) of a pitching ore shoot. (See pl. 11.) At the Hardwood Mill, Outafit, Parish, and Skate Creek deposits, very scanty information suggests the presence of shoot structure. The suggestive evidence is noted and evaluated in the detailed accounts of these deposits.

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. (See Postel, 1952, p. 34, for ore shoots in Clinton County district.) 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 metasedimentary rocks. At the Green Farm (fig. 20) several pits were sunk in ore on the axes of minor folds. These folds, having an amplitude of only 20 to 30 feet, are found at intervals of 25 to several hundred feet along the N. 40° W. strike of the "vein." The folds trend about S. 15° W. and plunge about 20° SSW. The magnetic anomaly is continuous along the covered belt for about 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 Marian magnetite prospects, Tupper Lake quadrangle. (See p. 147–149.)

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 had accessible underground workings, and the only one the Geological Survey was able to map in detail (40 feet=1 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 grows.

The Clifton deposit lies within a narrow belt of metasedimentary rocks, several hundred feet thick, on the extreme eastern flank of the north-trending Stark anticline of phacoidal hornblende granite gneiss. (See pl. 1; Prof. Paper 376, fig. 14.) 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 metasedimentary rocks 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 metasedimentary rocks 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 geniss 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. (See fig. 4.) There is considerable local evidence indicating that the two systems belong to two different periods of deformation. One system has axes trending about 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 about N. 58° E. Let us designate the axes of this system as b_E. The axes of b_E folds plunge gently east-northeast or west-southwest, 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_{SR}. 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 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 system. The body was elongate in a northeast to east-northeast direction. The

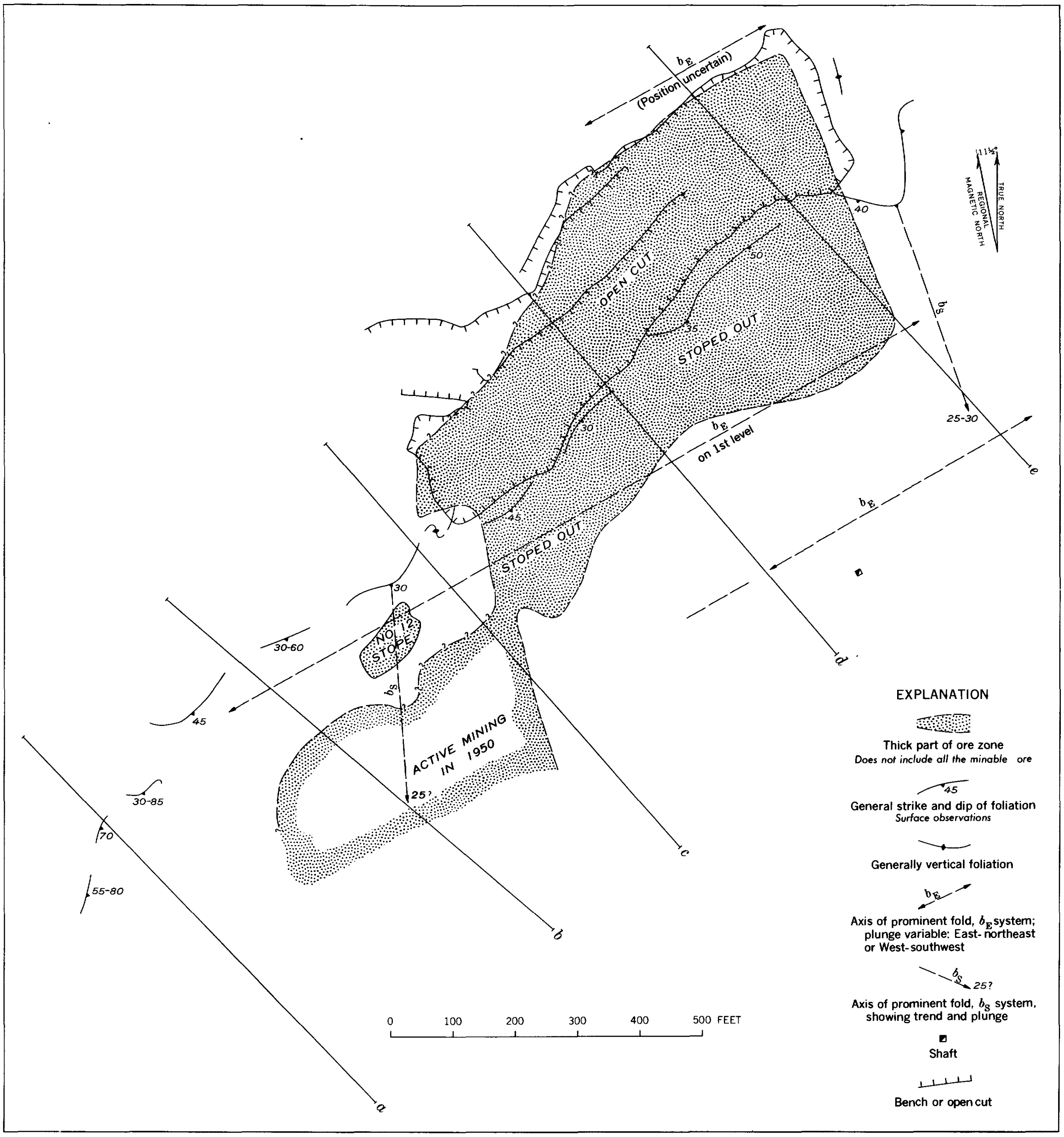


FIGURE 4.—Key structural features and horizontal projection of main ore shoots, Clifton mine. For skeleton block diagram containing sections *a* to *e*, see figure 18.

northeast edge of the ore body ended rather sharply against a relatively narrow anticline of the b_S system. This anticline, asymmetric but essentially upright at the surface, was overturned at depth. The southwest edge thinned and frayed out over the top of a broader anticline of the b_{SR} system. In detail, this anticline was modified by many minor folds of the b_S and b_{SR} systems. 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 the structural relations were not observable at the time this study was made. 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 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. Mining in 1950 was 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. Extensions of the southwestern ore-bearing zone had been only slightly explored in 1950, but the results were promising. No attempt had been made to prospect for deep ore bodies down the plunge of the b_S folds, below the main (northeastern) 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_{SR}, and b_E 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_E system are relatively open, they are rolls. Locally, folds of the b_S and b_{SR} systems are tight, isoclinal crumples and "drags"; and folds of the b_E system are overturned isoclinal folds whose under sides approximate low-angle overthrusts. A few representative cross sections are given in figures 4 and 18. Details are shown in plates 6 and 7.

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, p. 105–110) and others. Magnusson's description of the structure of ore bodies in the Kaveltorp field has been abstracted by the writer (Sims and Leonard, 1952, p. 35–36).

All the ore formed after the folding. Some of the very coarsest magnetite is in fact found within 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 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 descriptions 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 deposits 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. (See p. 39 and p. 44.)

Small discontinuities between ore and wall rocks are well shown at several places in the Clifton mine. The footwall contact of ore against pyroxene gneiss and garnet skarn is well exposed for 5 or 6 feet in the face of a small crosscut (430N, 195W) off the mining sublevel below the first level. (See pl. 5.) 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.

Trangression 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 wallrocks 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 trend (Sims, 1953, p. 267–268). Zimmer[16] (1947) and Postel (1952, p. 31, 81) have 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. The multiple lineations include rodding, fluting, and pencil structure—elements whose orientation is in some districts a highly reliable guide to the course of the ore shoots. One of the major structural problems, still not entirely solved, is the relation between certain sets of lineation and the probable direction of elongation of the ore bodies in the district. The problem is not: 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 on p. 21.)

[16] Zimmer, P. W., 1947, Phase petrology of the Lyon Mountain magnetite deposits in the northeastern Adirondacks: State College of Washington, M.S. thesis (unpub.).

LOCALIZATION OF ORE

The importance of regional and local structural features 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 replacements of carbonate rocks that were 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 (1952, p. 39–40) found that many ore specimens from the Clinton County magnetite district showed prominent zones of microgranulation serving as sites for replacement by magnetite. These contrasting relations are also reported from magnetite districts in New Jersey. Sims (1953, p. 277) described microbrecciated zones as sites favorable for magnetite deposition in the Dover district, whereas Hotz (1953, p. 215) did not recognize direct evidence of microbrecciation in the Ringwood–Sterling Lake area. 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) considered at length a similar problem relating to the sphalerite deposits. Perhaps a porosity study 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. (These shear zones are due to late, discontinuous shearing; they are not to be confused with foliations due to solid flow.) All the shear zones are small. Most of them strike east-northeast and dip very gently southward, but one in the open pit has a steep northward dip. These shears were formed later than the skarn, pegmatite, and ore. The writer believes they represent late movement induced by the same forces that earlier produced the prominent b_E 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 (see pl. 2). Inasmuch as local faulting after ore deposition 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 glaciers, which removed most signs of earlier weathering. However, some deposits show a characteristic type of alteration believed to be supergene. 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. 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.

Alteration is conspicuous 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. (See, for example, cross section through hole B1, pl. 2.)

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 (pl. 2, cross section through hole B9) 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.

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. (Cf. 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 wallrock alteration are a minor feature indeed.

The dumps at the Jayville deposit show magnetite-bearing biotite sköl fragments 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 suboutcrop 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 and other favorable structures 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 (1936b) from the Balmat zinc mine. Supergene alteration at Balmat extends to depths of 1,000 feet and more (Brown, 1936b, 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, p. 194–202; this report p. 231–233). The age of this type of alteration in the northwest Adirondacks is not certain. Buddington thinks that most of it took place in 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, N.J., has been described by Sims and Leonard (1952, p. 26–30). Similar alteration is known to occur in the Arnold Hill group of mines, Clinton County, N.Y. (Postel, 1952, p. 50–52). Limonitic bodies formed about magnetite in marble have been described by Hotz (1954, p. 213–217) and Bayley (1941, p. 62, 68–69) at the Ahles mine, Warren County, N.J.

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 merit, 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 before 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 metasedimentary rocks 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 metasedimentary rocks. Probably the same fundamental magma also yielded a high-potassium, volatile-enriched phase that intruded the metasedimentary rocks as thin sheets, reacted with the country rock, and in places metasomatized it extensively, yielding heterogeneous microcline granite gneiss.[17] 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, OH, and Si, locally accompanied by Cl and P, 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: (a) 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; or (b) that the rare cutting of skarn by

[17] The possibility that the microcline granite gneiss is relatively old must also be borne in mind.

granite is a purely local phenomenon, not entirely in phase with the main consolidation of the granite masses; or (c) that the skarns have indeed been deformed, recrystallized, or even reconstituted without leaving any recognizable sign of these happenings. At present, 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, potassic 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.[18] 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 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}, or 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 toward a dilute water-rich fluid led to local alteration of ore and wallrocks. 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 wallrocks, followed by the deposition of hydrous silicates, calcite, and sporadic base-metal sulfides.

Engel (1951, personal communication) notes that broadly similar processes, operating on rocks of the lowlands, locally converted carbonate rocks to amphibolitelike skarns by introduction of Fe and Si, emplaced small bodies of granite, and subsequently modified the amphibolitelike skarns by introduction of alkali silicates (in part hydrous) and silica.

At some later date—possibly in late Precambrian time, possibly in Silurian or younger 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 of postpegmatite emplacement and postdeformation. Their absolute ages are unknown. 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 metallic minerals 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

[18] 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.

can be presented in orderly fashion. Table 11 is a rough, semiquantitative expression of the elements introduced in minerals of the two types of magnetite deposit. It would be extremely difficult to reach even a very crude semiquantitative expression of the elements introduced to form the bulk chemical composition of the ores, principally because metasomatism affected two completely different host 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 potassic feldspar. Ba, relatively insignificant in unmodified skarns, concentrated with K and Al in the granite gneiss ores and in modified ores of both types.

TABLE 11.—*Elements introduced in minerals of skarn ores and granite gneiss ores*

[For a very crude semiquantitative estimate of elements present in the bulk chemical composition of skarn ores and granite gneiss ores, compare the data of this table with those in tables 3 and 6. With few exceptions, minor elements (such as Sr in feldspar or Ga in mica) are omitted, as are the cations in sulfides and sulfarsenides. Data for analyzed minerals taken from tables 1 and 10. Data for unanalyzed minerals approximated from chemical analyses in Dana (1892), due allowance having been made for composition estimated from optical properties.]

Element or ion	Common percent range in mineral	Skarn ores	Granite gneiss ores
Si	xx.	Pyroxene, garnet, amphibole, mica; scapolite and feldspar in local modified skarn ores. (Initially introduced; subsequently removed in part.)	Quartz, potassic feldspar.
Ti	.x	Dark skarn minerals, and exsolution product or solid solution in magnetite (.x); sphene in local modified skarn (xx.)	Biotite, and exsolution product in iron oxides (.x); solid solution in hematite (x.). Possibly, net loss of Ti in iron oxide minerals.
Al	x.	Dark skarn silicates; feldspar in local modified skarn ores.	Potassic feldspar, silimanite, biotite.
Fe^{+3}	xx.	Magnetite; dark skarn silicates, especially andradite.	Iron oxides, especially hematite; biotite (x.).
Fe^{+2}	xx. to x.	Magnetite; dark skarn silicates.	Magnetite; biotite.
Mn	.x	Dark skarn silicates; magnetite?	Hematite—and magnetite? (.x); spessartitic almandite (x.); spessartite in modified granite gneiss ores and locally in granite gneiss ores (xx.).
Mg	----------	Initially introduced?	LARGELY REMOVED
Ca	----------		LARGELY REMOVED
Ba	.0x to xx.	Mica (.0x); barite found sporadically in modified skarn ores (xx.).	Potassic feldspar (.x); barite (xx.).
Na	.x	Dark skarn silicates; scapolite locally present in modified skarn ores (x.).	Potassic feldspar. LARGELY REMOVED.
K	x. to xx.	Dark skarn silicates, esp. ferrohastingsite and biotite (x.); potassic feldspar in local modified skarn ores (xx.).	Potassic feldspar, biotite.
OH	x.	Amphibole, mica. Late hydrothermal minerals, sparingly present locally.	Biotite, sericite. Late hydrothermal minerals, locally conspicuous.
CO_2	----------	LARGELY REMOVED	
P	xx.	Apatite; sporadic monazite?	Apatite; sporadic monazite.
S	xx.	Sulfides; barite found sporadically in modified skarn ores. Possibly some S in scapolite locally present in modified skarn ores.	Sulfides; barite.
F	x. to xx.	Fluorite (xx.); amphibole, mica, apatite (x.)	Apatite, biotite, fluorite; tourmaline (.x?).
Cl	.0x to x.	Mainly(?) as scapolite locally present in modified skarn ores (possibly .0x to x.); amphibole, mica (.x); chlorapatite?	Mainly as local scapolite; biotite; chlorapatite?
B	x.	Vonsenite, datolite, other boron minerals?—all local.	Tourmaline, other boron minerals?—local.
Ce	xx.	Allanite; sporadic monazite?	Allanite; sporadic monazite.

RELATION OF MAGNETITE DEPOSITS TO YOUNGER GRANITES

The space relations of magnetite deposits to younger granites have been described above (p. 17, 18). It seems reasonable to rule out hornblende-microperthite granite magma as an immediate source of iron-rich emanations. Concentrations of magnetite have not been found in hornblende-microperthite granite in this district or, so far as we know, elsewhere in the Adirondacks or in the Highlands or New Jersey. Moreover, magnetite deposits have not been found in scattered skarn layers, unaccompanied by alaskite or microcline granite gneiss, enclosed in large masses of hornblende-microperthite granite. One might argue that magnetite concentrations are lacking merely because such environments were too hot to permit deposition of magnetite in concentrations. (The data of Buddington, Fahey, and Vlisidis (1955, tables 4 and 14 and schematic diagram on p. 514) indicate that the hornblende-microperthite granite magma probably crystallized at high temperature, roughly near 750°–800° C.) The one ore-bearing area of metasedimentary rocks isolated within hornblende-microperthite granite is the Dead Creek syncline, central part of Cranberry Lake quadrangle. Here, sheets of microcline granite gneiss are interleaved with the metasedimentary rocks. The microcline granite gneiss itself contains most of the iron oxide bodies, though minor deposits of magnetite in metasedimentary rocks are also present. Microcline granite gneiss was a prerequisite here for the occurrence of iron oxide ore bodies, and there is a very strong presumption that the iron oxide bodies were derived from the microcline granite gneiss, not from hornblende-microperthite granite.

The relation of unmodified magnetite 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 seem to be genetically related to their host, rather than occurring in it adventitiously, though one cannot rule out the possibility that alaskite (nearly always occurring near the deposits) 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, potassium-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, and in hornblende-microcline granite gneiss of the sort widespread in the Childwold quadrangle; 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, B (locally), 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 emanations from a granitic magma concentrated Fe, 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 conceivable, of course, that the emanations that yielded magnetite deposits came rather directly 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,[19] the St. Lawrence 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 about 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 yield fluorine sufficient for transporting 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 one 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. That more F-bearing minerals are found in skarns than in granite gneisses cannot be due solely to the greater quantity of Ca available to "fix" F in the skarns. Ca from some source was available to "fix" P for the apatite that is locally an abundant accessory in granite gneiss ores.

Perhaps at the magnetite-depositing stage, the solutions had already become substantially hydrous. Iron-poor phlogopites from skarns are rich in F and rela-

[19] 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.

tively 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, p. 372, 384), Geijer (1925, p. 687), Shand (1947, p. 635; but see Holser, 1950, p. 1080), Postel (1952, p. 44–45), 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 (1950, personal communication) finds that the disseminated accessory iron oxide of much of the microcline granite gneisses of the district is hematite, not magnetite. Conceivably, the development of hematite in these rocks is a consequence of their K-rich environment. However, an abundance of K alone could hardly be responsible for the oxidation, since some magnetite deposits in microcline granite gneiss contain relatively little hematite; and primary hematite is absent from the Trembley Mountain deposit, in which skarn was modified by K-rich emanations before magnetite was deposited.

The writer suggests that the development of hematite in the K-rich rocks of the district is more likely due to a change in character of the ore-depositing solutions, presumably from acid to alkaline. Mason (1943, 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.

Recently, Baker (1955)[20] has emphasized the importance of the water-magnetite reaction in causing oxidation of magnetite to hematite during cooling of an ideal system that, at some pressure and temperature, contained water and magnetite at equilibrium. Conceivably, water played a dominant role in the development of hematite in the St. Lawrence County district.

Several otherwise puzzling features of the magnetite deposits seem to be in accord with the hypothesis of decreasing acidity of the ore-depositing solutions. Skarn ores, which lack primary hematite, may have come from solutions too acid—that is, 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.

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

The role of boron, possibly important throughout the mineralizing process, is especially difficult to assess. Major concentrations of boron minerals in the district are known only at Jayville, though certain small concentrations have in their day been famous mineral localities. All the boron minerals are found in or close to members of the granite and granite gneiss series, and the writer regards the granitic magma or magmas as a likely source of this boron.

Tourmaline occurs locally in marble, biotite gneiss, microline granite gneiss, and granite pegmatite. The tourmaline of marble in the Sabattis area, Tupper Lake quadrangle, is the brown magnesian variety, dravite. The tourmaline finely disseminated in marble at the Green Farm magnetite deposit, Oswegatchie quadangle, is the ferroan variety, schorl. Large and showy crystals of schorl are still collected from marble near Pierrepont, Canton quadrangle (Slocum, 1948). Schorl is found as an accessory in microcline granite gneiss and granite pegmatite, commonly in association with sillimanite gneisses, as in the Russell belt of sillimanite gneiss. Clots of fine granular schorl are present in some biotite gneiss, as in the metasedimentary rocks of the Dead Creek syncline. *Danburite* is found in pyroxenic metasediments west of Dana Hill, Russell quadrangle, on the former VanBuskirk farm (Agar, 1921, 1923; Clark, 1949; Slocum, 1948). *Datolite* occurs in modified skarn at the Spruce Mountain Northwest magnetite deposit; Dana (1892, p. 1065) reports it from the VanBuskirk farm. *Vonsenite* is found in the Jayville and Clifton magnetite deposits. Other boron minerals, not positively identified, are suspected in certain granite gneiss ores. The boron minerals are thus widely distributed though at most places quantitatively slight; their iron content ranges from nil or very low (danburite, datolite, dravite) to extremely high (vonsenite); and their apparent position in the paragenetic sequence probably ranges from the skarn-forming stage (some tourmaline), through the stage of skarn modification (datolite at Spruce Mountain Northwest), to the magnetite-depositing stage (vonsenite at Jayville). Datolite may even be present among the late hydrothermal minerals, though its tentative identification has not yet been confirmed microscopically.

[20] Baker, D. R., 1955, Geology of the Edison area, Sussex County, New Jersey: Princeton Univ., Ph. D. dissert. (unpub.).

Geijer (1927a, 1939), Tilley (1951), Watanabe (1943, 1953), and others have presented evidence that boron, commonly in conjunction with fluorine, has been important in metasomatic processes resulting in the development of skarns, with or without concentrations of ore minerals. How significant boron may have been, here, as a mineralizing agent, one cannot say. It was available within the magnetite district; it was also available in many parts of the Grenville lowlands near small masses of granite. Within the magnetite district, the quantity of OH, F, and subordinate Cl and P fixed as minerals greatly exceeds that of B. But the quantity of a volatile so fixed is, one supposes, no reliable measure of the quantity passed, or of the effectiveness of the volatile in transporting Si, Fe, and other major elements now found in the magnetite deposits.

METASOMATISM AND IRON CONCENTRATION OF GRANITES AND QUARTZ SYENITES CONTRASTED

The quartz syenite series, genetically unrelated to the metasomatic magnetite deposits, has certain features of metasomatism and iron concentration that contrast strongly with those related to the younger granites. Three main features are summarized below; they point up the importance of nearly contemporaneous availability of volatiles and iron to permit the formation of metasomatic magnetite deposits in the district.

First is the very slight development of pegmatites from the quartz syenite magma. Second is the limited extent of mineralization and metasomatism adjacent to the felsic members of the series. 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 gneiss 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 (Buddington and Leonard, 1953, p. 894, table 1, no. 6) 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 Prof. Paper 376, p. 46, 48–49, 85.) 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 metallic minerals 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 manyfold 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 development of extensive pyrometasomatic alteration zones and extensive zones of migmatites is restricted to contact zones of the younger granites. The development of granite pegmatite, locally syenitic, is also a characteristic accompaniment of the younger granites, though 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 of the Adirondack massif, including anorthosite, but their extensive and characteristic development is with the younger granites.

TEMPERATURE OF FORMATION OF MAGNETITE DEPOSITS

Some evidence is at hand for estimating the temperature at which the magnetite formed in the metasomatic 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 occurring as disseminated accessory minerals in the country rocks. Moreover, iron oxides from the ore deposits generally contain much less TiO_2 in solid solution. These facts suggest that the iron oxides in the ore deposits formed at somewhat lower temperature than accessory iron oxides in country rock. The approximate total TiO_2 content of magnetites from skarn deposits ranges from 0.10 to 0.50 weight percent and averages 0.26 percent. For magnetites from granite gneiss deposits, the range is from 0.34 to 0.80 percent and the average is 0.60 percent. Magnetites so poor in TiO_2 are inferred by Buddington, Fahey, and Vlisidis (1955, p. 514, 526) to have formed at temperatures approaching 500° C. How great an effect the complex pneumatolytic or hydrothermal solutions may have exerted on the crystallization temperature of the magnetite, no one can at present assess. The inferred temperature of about 500° C, though tentative, is consistent with evidence that (a) magnetite in the ore deposits formed at lower temperatures than magnetite found as an accessory in the country rocks, and (b) magnetite in the ore deposits formed before the accompanying sulfides and presumably at slightly higher temperature.

The approximate total TiO_2 content of the magnetites is estimated mainly from the analyses of magnetic fractions (Davis tube concentrates) of composite samples of ore-bearing drill cores. Skarn deposits represented are Brunner Hill and Outafit. Granite gneiss deposits represented are Benson Mines (commercial magnetic concentrate from mill), Deerlick Rapids, Skate Creek, and Spruce Mountain. The magnetic concentrates are likely to include a few attached grains of ilmenite and perhaps a few free grains as well. The reported values for TiO_2 in concentrates may thus be very slightly higher than the TiO_2 actually present in magnetite alone.

Exsolution features in the sulfides give some clue regarding 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 the Clifton mine (skarn ore), 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 the Parish deposit is unknown. Sphalerite is very sparse in that 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 little significance.

Any extrapolation from the data of Schwartz (1931) and Buerger (1934) requires the minimum assumption that the intergrowths to be used as geologic thermometers have essentially the same composition as the respective intergrowths investigated by those workers, for—pressure being constant—exsolution at a given temperature depends on the chemical composition of the system yielding the two or more coexisting phases. Alternatively, one might make the minimum assumption that the respective exsolution curves are of such shape that a vast change in composition takes place in a relatively narrow temperature interval. At the time of writing, the phase relations in the systems Cu–Fe–S and Cu–Zn–Fe–S are not completely known. However, if we are permitted to make either of the assumptions noted above, the minimum temperatures inferred from intergrowths in bornite and sphalerite from the magnetite deposits agree very well with a hypothetical 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 more than 475° C for the initial deposition of magnetite.

For sphalerites from the skarn deposits at Orijärvi, Finland, and Kaveltorp, Sweden, Kullerud (1953, p. 129, 131, 141) reports pressure-corrected temperatures of formation of $560° \pm 25°C$. Presumably, similar skarn deposits in St. Lawrence County formed at roughly comparable temperatures. In contrast, the temperatures of formation of sphalerites from Balmat and Edwards, N.Y., are reported as $375° \pm 25°C$ and $385° \pm 25°C$, corrected for pressures of 2000 ± 1000 atmospheres (Kullerud, 1953, p. 130, 138). Because of its relative scarcity in the St. Lawrence County magnetite deposits, sphalerite has not yet been investigated by Kullerud's methods.

Sphalerite of stage II never shows exsolved chalcopyrite. This fact by itself cannot be taken to indicate that this generation of sphalerite formed at a temperature below 350°–400°C; chalcopyrite assignable to stage

II is not everywhere present with sphalerite of stage II, a relation suggesting that the solutions which deposited minerals of stage II were locally very poor or lacking in Cu. The white internal reflection of some sphalerite of stage II suggests a variety very poor in Fe; Fe-poor sphalerites are known to have formed at relatively low temperature (Kullerud, 1953, p. 65). The association of sphalerite II with the zeolites and calcite suggests a temperature considerably below 350° C for this waning stage of mineralization, and such a relation is consistent with that inferred from the nature of the sphalerite.

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.

Supplementary information on these inferred temperatures comes from measurements made by means of an F. Gordon Smith "pyrite geothermometer" (Smith, 1947). Using this device, W. R. Griffitts and the writer obtained the following results:

1 The "indicated temperature" for five polished specimens of pyrite I ranged from 500° to 560°C, and averaged 530°C. The range and average value (±20°) were the same for all specimens, regardless of the type of deposit in which the pyrite occurred. The specimens came from the Brandy Brook and Clifton deposits (skarn type, 1 specimen from each); Spruce Mountain deposit (granite gneiss type, 2 specimens); and Trembley Mountain deposit (modified skarn type, 1 specimen). Seven other specimens, variously distributed, gave anomalously high readings, presumably because the individual pyrite grains were too small to accommodate both probes of the geothermometer.

2 The "indicated temperature" for two specimens of pyrite II was 330° and 350°C, respectively. The specimens came from Brandy Brook and Spruce Mountain. The latter specimen contained pyrite of stages I and II, separately measurable.

Out of curiosity, without changing the original calibration of the pyrite geothermometer, we made readings on magnetite from 14 separate polished specimens. The "indicated temperature" ranged from 560° to 600°C and averaged 570° C. Eleven magnetites from the four types of ore deposit had the same range and average value (±20°) for their "indicated temperature." Two magnetite samples from ore deposits were nonconducting, though other samples from one of the deposits had "normal" magnetite. The 14th magnetite sample gave an anomalously high reading of more than 700°C. This sample was from the Parishville magnetite-ilmenite deposit, whose origin is uncertain.

The "indicated temperature" of pyrrhotite from 3 specimens (1 from Brandy Brook and 2 from Trembley Mountain) ranged from 430° to 460°C and averaged 440°C.

Magnetite from ilmenite-magnetite concentrations in rocks of the Diana complex, and pyrite thought perhaps to be supergene, could not be tested, owing to the small size of the individual grains or aggregates of these metallic minerals.

The average "indicated temperatures" of magnetite (570°C), pyrite I (530°C), pyrhotite (440°C), and pyrite II (340°C), read from the pyrite geothermometer, cannot be regarded as accurate by absolute standards. Nevertheless, the relative values are consistent with relations inferred from the paragenetic sequence of these metallic minerals. The close agreement with temperatures inferred from data on chemical composition and exsolution features may or may not be coincidental.

OTHER HYPOTHESES

This discussion of hypotheses of origin was written in 1951. A few references were added later, but it has not been possible to publish an up-to-date résumé of ideas on the genesis of magnetite deposits. Postel (1952, p. 43) and Alling (1939) have reviewed the changing opinions on the origin of eastern and northern Adirondack iron ores.

IGNEOUS SOURCE OF THE IRON; NONHYPOTHERMAL 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 *in* Kemp and Ruedemann, 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, 1954, and other papers) has inferred a complex origin. The skarns are closely associated with the leptite formation, now regarded as supracrustal volcanic rocks with interbedded metasedimentary rocks. 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 remineralized, so that they now appear as high-temperature deposits. Reaction metamorphism has played a dominant role in giving the deposits their present character. Nevertheless, the initial source of the iron was apparently regarded as igneous—volcanic. Geijer (1923, 1927a, 1936, 1939) has in the past argued convincingly that the skarn ores are essentially pyometasomatic deposits formed by volatile-rich emanations stemming from granites that intrude the leptite formation. In his earlier held view, reaction metamorphism played a minor part in developing these ore deposits. The writer is very hesitant to comment on the current status of the problem, for the most thorough exposition by both men (Geijer and Magnusson, 1944) is entirely in Swedish. A brief exposition of their later views (Geijer and Magnusson, 1952) became available during the revision of the present report. Geijer and Magnusson (1952) emphasize the role of reaction metamorphism (see below) in giving many of the nonapatite ores of central Sweden their present character and, relating the ores very intimately to the volcanic leptites, attribute the primary ore-forming process to sedimentation or to replacement at relatively shallow depth.

Backlund (1950) rejects the view that leptites are volcanic and interprets the entire complex of leptites, ores, and "urgranites" (old granites—the oldest Precambrian granites of central Sweden) as sedimentary rocks deformed, granitized, and subjected in varying degree to successive small- and large-scale basic fronts and alkali culminations.

As there is no evidence of the former existence of supracrustal volcanic rocks in the 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, p. 85–86), R. J. Linney (1943, p. 488), and Walker (1943, p. 520, 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, whose average grade (about 25 percent Fe) is very similar to that of unenriched iron formation, or taconite. It is especially appealing because of the wide distribution in the Precambrian 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.

SEDIMENTARY NONPYRITIC ORES

The nonpyritic ores are of three main types: banded siliceous ores, banded carbonate ores (both widely represented in the low-grade iron formations of the Precambrian), and generally oolitic ores (chiefly of Ordovician, Silurian, and Jurassic age). In terms of primary sedimentary facies, James (1954) classifies the banded types mineralogically as carbonate facies, silicate facies, and oxide facies, all of which may contain thin layers of chert. He (James, 1954, p. 247–248, 256–263) presents evidence that magnetite, as well as hematite, is present in the primary sedimentary oxide facies.

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 nonpyritic iron formation (Engel and Engel, 1953, p. 1038). 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, p. 194–202; Brown, 1936b, p. 331–354; this report, p. 230–233).

The primary sedimentary facies of iron ores are described and discussed by James (1954), whose paper is documented by 100 references. Parts of the iron formation in the Lake Superior region have been metamorphosed. James (1954, p. 238–239) notes that the grade of metamorphism of the iron formation is indicated by progressive coarsening of chert grains (from about 0.01 mm average diameter to 1 mm or more) and by development of such minerals as chlorite, mica, amphibole, and garnet. A full treatment of the problem of metamorphism is reserved for publication in the future.

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, p. 7–36). In an area where metamorphosed volcanic rocks predominate, a banded-silica iron formation of Keewatin (Precambrian) 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 Precambrian rocks (or highly metamorphosed lower 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 Håfjell 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 early Precambrian age. The age of the Sydvaranger ores is currently regarded as middle Precambrian (Bugge, 1953, p. 52). The character and composition of the ores is strongly suggestive of a derivation from siliceous iron formations of the Lake Superior type. (Cf. analyses of taconites given by Gruner, 1946, and by Bugge, 1953.)

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 about 2 mm in thickness. They are remarkably continuous along the strike; layers 0.5 mm thick can be followed for 10 meters. Contacts between ore and barren, unstratified wallrock are sharp and concordant. The wallrocks 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 (1911) and Bugge (1953) interpreted 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 to 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.

Quartz-banded hematite ores similar to those at Striberg are found at Stripa (Geijer, 1938) at Stråssa and Blanka (Geijer, 1927b), and in parts of the Norberg district (Geijer, 1936)—all in central Sweden. The geologic setting for these deposits resembles that at Striberg, though potassium-rich leptites are dominant over sodic leptites at Stråssa and Blanka and at Norberg. "Urgranites," generally carrying roughly equal amounts of potassium and sodium but locally highly sodic, are important rock units at Norberg.

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

688-628 O—63——7

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 by 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, in many places closely associated with K-rich and Na-rich rocks of granitic composition and varied 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. (See analyses of ores in papers cited.) 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, though in places—as at Stripa (Geijer, 1938, p. 15)—some magnetite porphyroblasts have subsequently been partly martitized. 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 wallrocks 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 wallrocks. 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 metasomatized 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.

SEDIMENTARY PYRITIC ORES

From time to time, a sedimentary origin has been postulated for the pyrite and pyrrhotite deposits of the Grenville lowlands, as well as for the disseminated pyrite widely distributed in accessory amounts in the metasedimentary rocks. A brief review of earlier ideas is given by Buddington (1934, p. 214), who concurred with Smyth (1912) that the deposits were hypogene, though some of the disseminated, accessory pyrite might indeed be of sedimentary origin. (See also p. 225–226, this report.) 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 similar magmatic sources, affected rocks in two different environments. (See regional and local zoning of iron oxides and sulfides, p. 18–20.) 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, 1926a) 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 it 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 negligible 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 metasedimentary rocks can scarcely have served as a source of Fe for the development of magnetite.

A variant of Miller's hypothesis has recently been advanced by Hagner and Collins (1955) to explain the magnetite deposit at the Scott mine, Sterling Lake area, southeastern New York.

REACTION METAMORPHISM AND METAMORPHIC HYBRIDISM

Over the years, Magnusson (1925, 1929, 1930, 1940a, 1940b, 1954) has developed the concept of reaction metamorphism in its relation to the origin of Swedish skarn ores. He has believed 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 period of mineralization. The manganese in the Swedish ores is chiefly in knebelite, an olivine; but it is also found in pyroxenes, pyroxenoids, amphiboles, and (in some deposits) garnets. Quartz-banded iron ores accompany some of the skarn ores. The primary source of iron for both ore types is the leptite formation. The present character of the deposits is due mainly to regional metamorphism of later date, attendant on the folding of the leptite formation and emplacement of the "urgranites." 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," 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. Lately, he (Magnusson, 1954) has attributed the reconstitution chiefly to reaction in the solid state, and in addition has noted without elaboration that he regards the skarn ores as sedimentary.

Geijer (1923, 1927a, 1936, 1939) long 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." Subsequently, Geijer and Magnusson (1952) presented a very brief synthesis of their formerly divergent views. The impact of this synthesis is that, though both reaction metamorphism and pyrometasomatism have formed skarn ores, reaction metamorphism has been dominant in the production of the bulk of these deposits in central Sweden. Perhaps the last word on the quantitative aspects of this problem has not been spoken.

Small bodies of magnetite with a small amount of 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 hydothermal 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 and basic behinds

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; Miles, 1947; Backlund, 1950; Friedman, 1953.) 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 of 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, Al, and subordinate Na 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 resisters of their heated attack.

Ramberg (1948; 1951; 1952) argued that K, Na, Si, O, and H_2O are "squeezed out" chemically from once-deeper areas occupied by rocks of the granulite facies. These chemical constituents have concentrated at higher levels, among the rocks belonging to lower temperature metamorphic facies, leaving a concentration of Ti, Fe, Mn, Mg, and Ca within the granulite facies. The fundamental aggregates of mafic minerals so derived have appropriately been termed "basic behinds." Not only are independent bodies of rock rich in mafic minerals or Fe-Ti oxides developed (Ramberg, 1948), but typical gneiss of the granulite facies also becomes enriched in Ti, Fe, Mn, Mg, and Ca (Ramberg, 1951, p. 30, 32; 1952, p. 264).

Though this freeing of Fe and Ti, among other elements, depends on *advancing* metamorphism or increasing depth within the crust (Ramberg, 1948), DeVore (1955) has argued that Ti and perhaps Fe, as well as other elements, are liberated during *retrograde* metamorphism from the granulite facies to the epidote-amphibolite facies.

To the present writer, it appears that the arguments of both men are at the moment speculative in nature when the attempt is made to explain the formation of ore deposits by these processes.

1. In St. Lawrence County the magnetite deposits are postmetamorphic, not synmetamorphic.
2. The development of skarn at the magnetite deposits predates the younger granite, but the development of ore postdates the younger granite and the granitization.
3. As one would expect from the first fact, the distribution of the magnetite deposits is not systematically related to the position of the metamorphic zones or subzones recognized in the district. (Cf. pl. 1, and fig. 12 of Prof. Paper 376. The Rock Pond, Sterling Pond, Brunner Hill, and White Place deposits are within the inferred hottest area—inside the line *B–B′* of that figure 12. The Long Pond anomaly is on or close to line *B–B′*. The bulk of the ore deposits are within the intermediate area, between lines *B–B′* and *A–A′*. The Lampsons Falls anomaly and the VanOrnum and Green Farm deposits are on or near line *A–A′*. The Jayville deposit is west of line *A–A′*.)
4. The inferred temperature of metamorphism was considerably higher than the inferred temperature of formation of the ore deposits. (See data and discussion, below.)
5. Metamorphism appears to have been progressive, not retrogressive, thereby eliminating DeVore's (1955) suggested mechanism for liberating certain metals.
6. Metamorphism of granites in the district has been attended by a slight increase in the amount of Fe_2O_3 and TiO_2 (Prof. Paper 376, p. 102); the process could not simultaneously have released large quantities of these compounds. Inadequate data on progressive metamorphism of amphibolites in the district suggest a similar relation. (Cf. Buddington, 1952, data in table 2.)
7. Ti has not been greatly concentrated with Fe in the ore deposits; this anomaly raises problems if the elements are to be regarded as derived by some metamorphic process operating on a common source material. (See discussion, below.)

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 (1952, p. 362), 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.) Indeed, as Barth suggests, a fraction of these same rejected ions might readily move forward in water rejected from the country rock.

Let us assume for the moment that such an ionic exchange might have occurred in the St. Lawrence County district. Pyroxene gneiss is initially Fe poor and can therefore be eliminated as a source of Fe. Biotite-quartz-plagioclase gneiss is a possible contributor among the metasediments. It seems a reasonable source of Fe, for it is intimately associated with all the granite gneiss ores. Amphibolite and metagabbro ought to be ruled out; they are sparse in the neighborhood of the magnetite deposits, and their granitized equivalent—hornblende-microcline granite gneiss—is extremely rare near the deposits. However, if one admits the possibility of long-range transport of Fe, the hornblende-microcline granite gneiss concentrated on the Childwold quadrangle requires discussion.

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 per standard cell: [21]

Biotite-quartz-plagioclase gneiss:

$$K_{3.0}Na_{5.9}Ca_{1.9}Mg_{2.5}Fe^{+2}_{2.8}Fe^{+3}_{0.7}Al_{13.4}Ti_{0.3}Si_{62.5}P_{0.1}O_{157.9}OH_{2.1}$$

Microcline granite gneiss:

$$K_{7.1}Na_{3.5}Ca_{0.4}Mg_{0.4}Fe^{+2}_{0.7}Fe^{+3}_{2.2}Al_{14.7}Ti_{0.4}Si_{62.8}P_{0.1}O_{158.1}OH_{1.9}$$

Granite gneiss ore:

$$K_{4.8}Na_{2.4}Ca_{0.3}Mg_{0.3}Fe^{+2}_{13.7}Fe^{+3}_{27.9}Al_{9.9}Ti_{0.3}Si_{42.0}P_{0.1}O_{158.7}OH_{1.3}$$

[21] The biotite-quartz-plagioclase gneiss is averaged from three analyses (Nos. 2, 3, 4, table 6, Prof. Paper 376; data in part from Engel). The microcline granite gneiss is averaged from two analyses (Nos. 17, 18, table 30, Prof. Paper 376; sillimanite-microcline granite gneiss and biotite-microcline granite gneiss, respectively). The granite gneiss ore is hypothetical. It consists of two parts average microcline granite gneiss, above, plus one part pure magnetite; these proportions are equivalent to a granite gneiss ore containing 24 weight percent magnetic Fe.

The 0.6 Fe, presumably as Fe^{+2} ion, would necessarily have to undergo partial oxidation to yield the granite gneiss ore, and roughly 70 standard cells of biotite-quartz-plagioclase gneiss would have to be converted to microcline granite gneiss in order to yield enough Fe to form one standard cell of granite gneiss ore. The metasomatic exchange of Ti is volume for volume, or nearly so.

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 reintroduced 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 1 mile at most, and generally less than 1,000 to 1,500 feet, to pass from the magma to its locus of deposition.

There remains the possibility that Fe removed from biotite-quartz-plagioclase gneiss may have served as a local, supplementary source of Fe for the formation of ore deposits of granite gneiss type. A volatile-rich granite magma may have selectively concentrated Fe from the gneiss, incorporated, fractionated, and transported it in volatile-rich fluids, together with magmatic Fe, to the sites now occupied by ore deposits.

Consider the possibility of deriving Fe for the ore deposits by converting amphibolite to hornblende-microcline granite gneiss. The standard cells are, respectively: [22]

Amphibolite:

$$K_{1.3}Na_{5.1}Ca_{10.0}Mg_{9.8}Mn_{0.2}Fe^{+2}_{7.2}Fe^{+3}_{2.1}Al_{18.3}Ti_{1.2}Si_{46.3}P_{0.2}O_{153.9}OH_{6.1}$$

Hornblende-microcline granite gneiss:

$$K_{9.2}Na_{3.3}Ca_{1.4}Mg_{0.8}Mn_{0.05}Fe^{+2}_{1.7}Fe^{+3}_{2.3}Al_{13.9}Ti_{0.5}Si_{61.8}P_{0.1}O_{158.4}OH_{1.6}$$

[22] The amphibolite is averaged from three analyses (see Prof. Paper 376, table 11). The hornblende-microcline granite gneiss corresponds to analysis No. 16, table 30, Prof. Paper 376. The granite gneiss ore is the hypothetical mixture explained in the preceding footnote.

Granite gneiss ore:

$$K_{4.8}Na_{2.4}Ca_{0.3}Mg_{0.3}Fe^{+2}_{13.7}Fe^{+3}_{27.9}Al_{9.9}Ti_{0.3}Si_{42.0}P_{0.1}O_{158.7}OH_{1.3}$$

The conversion of one standard cell of amphibolite to one of hornblende-microcline granite gneiss liberates 5.5 Fe^{+2} ions and requires 0.2 Fe^{+3} ion. The net amount of Fe to be injected to form a standard cell of granite gneiss ore is 38.6 Fe. Thus there must have been about seven times as much amphibolite, originally, as there now is granite gneiss ore, and roughly half the Fe^{+2} rejected in the granitization of amphibolite must have been oxidized in order to form typical granite gneiss ore. The postulated volume relations are reasonable for the district, if we assume that all or most of the hornblende-microcline granite gneiss of the Childwold quadrangle resulted from the granitization of amphibolite. However, the major concentrations of Fe in known ore deposits are miles and tens of miles distant from the exposed major areas of hornblende-microcline granite gneiss. Unless very large volumes of hornblende-microcline granite gneiss are concealed at relatively shallow depth beneath the known ore deposits, Fe ions rejected from amphibolite would have had to migrate miles and tens of miles to collect at favorable sites now represented by ore deposits. Such long-distance transfer of Fe does not, to this writer, seem reasonable. Once again, it is important to recall that a variety of hornblende-microcline granite gneiss is almost never the host of granite gneiss ore, that hornblende-microcline granite gneiss is exceedingly sparse near the ore deposits, that amphibolite—though common—is also sparse near the ore deposits, and that short-distance transfer of Fe by rejection during granitization of amphibolite is therefore an untenable hypothesis for the derivation of ore deposits in this district.

With regard to long-distance migration of Fe ions rejected during granitization of amphibolite, the role of Ti requires scrutiny. Relations between the standard cells indicate that 0.7 Ti ion is rejected through granitization of amphibolite. Ti must also be rejected if hornblende-microcline granite gneiss is converted directly to granite gneiss ore. (See standard cells.) The total Ti rejected in the two processes is 0.9 ion. If roughly 7 volumes of amphibolite must be transformed to yield enough Fe for 1 volume of granite gneiss ore, then 6.3 Ti ions are also rejected in the act. Do Ti and Fe move forward together? If so, the magnetite and primary hematite deposits, their immediate wallrocks, or the general vicinity of such deposits, should contain major concentrations of Ti whose total volume must be roughly one-sixth that of the Fe in the magnetite deposits. Such concentrations are not known, and the total amount of Ti disseminated as accessory sphene, ilmenite, titanhematite, rutile, or other minerals in the wallrocks seems inadequate to account for the quantity of Ti supposedly rejected during granitization of amphibolite. Do the rejected Ti and Fe ions move in opposite directions, the one downward and the other forward, or Fe forward and Ti upward to regions now eroded? If so, the behavior of the respective ions here constitutes a major geochemical anomaly.

Finally, it no longer seems reasonable to regard the large masses of hornblende-microcline granite gneiss on the Childwold quadrangle as the products of granitization of amphibolite. Though Buddington (1948, p. 36) previously regarded a substantial part of the hornblende-microcline granite gneiss as metasomatic, he later found (Prof. Paper 376, p. 83, 89) that the TiO_2 content of accessory magnetite in this gneiss is relatively high, about equal to that found in accessory magnetite of the microperthite alaskite, and suggestive of a magmatic temperature of formation for the gneiss. In contrast, microcline granite gneisses demonstrably of metasomatic origin, such as the biotitic and sillimanitic facies found in the district, probably formed at considerably lower temperatures—less than 600°C (Buddington, Fahey, and Vlisidis, 1955, p. 508, 514, 524). Temperature and mode of origin, as well as distance and other considerations, effectively rule out the large masses of hornblende-microcline granite gneiss as a source of rejected Fe ions later made available to form magnetite deposits.

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. It is true that equivalent rocks of the two areas may exhibit some differences in metamorphic facies, or grade of regional metamorphism. The metamorphosed Precambrian rocks of the lowlands appear to belong to the amphibolite facies (Buddington, 1952), whereas metasedimentary rocks within the magnetite district range in grade from the low part of the amphibolite facies to the granulite facies (Buddington, Fahey, and Vlisidis, 1955, p. 508; Prof. Paper 376, p. 98). However, the existence of rocks belonging to the low part of the amphibolite facies within, as well as west of, the magnetite-bearing area suggests that there is no profound difference in depth represented by exposed rocks of the lowlands and of the Adirondack massif. Moreover, sporadic sphalerite with exsolved chalcopyrite occurs in the magnetite deposits themselves, as well as in the pyritic sphalerite deposits, suggesting that no profound difference in depth separates the magnetite and sphalerite deposits. Differences between the dominantly igneous complex and the Grenville lowlands can generally 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, though rocks now exposed in part of the massif may indeed have formed at slightly greater depth than those now exposed in the lowlands. (See Prof. Paper 376, p. 98, 104.)

The most convincing evidence that the formation of iron oxide versus iron sulfide deposits was not determined by a profound depth difference is found at the Skate Creek deposit (p. 184–185), where one magnetite-bearing zone changes laterally, in a few thousand feet, into a zone of pyrite, chlorite, and epidote.

NOTES ON THE ORE TYPES

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 Precambrian. At present, the writer knows of the following deposits very similar to the granite gneiss ores of St. Lawrence County: Lyon Mountain, Clinton County, N.Y. (Postel, 1952); Edison mine, Ogdensburg, N.J. (Baker, 1955);[23] deposits in the Pequest district, New Jersey (Hotz, 1954, p. 219–223); Solberg belt, near Näs, southern Norway (Kjerulf and Dahll, 1862); Llano-Burnet area, Texas (Paige, 1911; Barnes and others, 1949). These deposits appear to be partly or wholly in potassic granite gneiss of several varieties, though the kind of feldspar accompanying the Solberg deposits, Pequest deposits, and some of the Llano-Burnet deposits is not clearly stated. Barite has been found in one of the Llano-Burnet deposits (Paige, 1911, p. 33). Deposits broadly referable to the granite gneiss type are found at Mount Olive, N.J. (Hotz, 1954, p. 234–238), and perhaps at several places in southeastern New York (Colony, 1923).

Deposits in sodic granite and related rocks are absent from the St. Lawrence County district, but they are found in the Clinton County district, New York (Postel, 1952); the Mineville-Port Henry district, New York; and the Dover district, New Jersey (Sims, 1953).

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 and others, 1922; Gustafson and others, 1950). In addition spessartite, quartz, and magnetite are found in nonproductive deposits associated with the lead-zinc lodes. Andrews (Andrews and others, 1922) ascribed a common replacement origin to the lead-zinc lodes and the nonproductive magnetite deposits. Gustafson and his associates (1950) consider that the lead-zinc 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 magnetite district should have a long and healthy life as a modest producer of iron ore from low-grade and medium-grade deposits. Greatly extended life and increased production may follow from development of deposits now submarginal in grade, especially since it has become feasible to recover crystalline hematite as well as magnetite. Gillies (1950) and Webb and Fleck (1950) expressed optimism concerning the reserves of Adirondack magnetite and the economic advantages that improvements in mining and milling may bring. The Adirondack magnetites may become still more prominent with the general development of the region after the opening of the St. Lawrence seaway in 1959.

The quantity of lump ore in the district appears to be so small that its production even in times of national emergency could be only a few thousand tons a year.

As the region develops, it may be appropriate to consider mineral deposits and possible byproducts not now of economic interest. For example, the ilmenite-magnetite-bearing lenses in feldspathic ultramafic zones of the Diana complex represent a very large tonnage of low-grade material which, if extractable en masse,

[23] Baker, D. R., 1955, Geology of the Edison area, Sussex County, New Jersey: Princeton Univ., Ph. D. dissert. (unpub.).

could yield concentrates of ilmenite, magnetite, zircon, and other accessory minerals. Intermittent shortages of sulfur might warrant the recovery of byproduct pyrite from certain magnetite deposits, such as the upper pyritic magnetite zone at the Spruce Mountain Northwest deposit, presuming the deposit to have been developed for its magnetite content. Some pyroxene that forms much of the gangue of skarn magnetite deposits seems to be suitable for the manufacture of rock wool, and the garnet locally abundant in the same skarns might make a suitable abrasive. Manganese garnets, such as the spessartite of the Parish deposit, are at times of interest in the welding trade. Limited work done on byproduct sillimanite from one magnetite deposit suggests that this sillimanite is not suitable for use as a refractory, owing to the considerable amount of alteration products in the sillimanite itself. Yet some sillimanite of the district, being relatively unaltered, might be more suitable for commercial use. Fluorite, abundant locally in the Parish deposit, might be recoverable in marketable form. Ordinary mill tailings may also have their uses. In other parts of the country, tailings comparable to those at Benson Mines and Clifton have been marketed in large quantities for roofing granules, poultry grit, and concrete aggregate.

Data on the radioactivity of some rocks and ores of the district are reported by Narten and McKeown (1952) and by McKeown and Klemic (1953). Though many of the rocks are regarded as abnormally radioactive (Narten and McKeown, 1952), containing 0.003 percent equivalent uranium or more, no economically significant concentrations of radioactive material were found in the district as the result of the two reconnaissance surveys.

So far, minerals in known deposits have been considered. Several other types of ore deposit might be sought, using geophysical and geochemical techniques to prospect favorable geologic sites for "blind" ore bodies of hematite, copper sulfides, and sphalerite. Many magnetite deposits are demonstrably localized by major structural features, especially intersecting structural trends and boundaries between areas having different lineation patterns. Not all such favorable structural sites are occupied by known magnetite deposits. Have other valuable minerals, undetected because they do not and cannot yield an intense positive magnetic anomaly, been deposited at these sites?

The intense negative magnetic anomaly south of the Parish magnetite-hematite deposit is an enigma. Does it represent a complement to the positive Parish anomaly, a large area containing disseminated accessory hematite—suitably polarized, or a relative concentration of hematite of ore grade?

Copper minerals in commercial quantity have not been found in the district. However, a considerable quantity of disseminated copper sulfides is present at the Parish deposit, and a concentration of chalcopyrite was cut by one drill hole at Brandy Brook Northwest. The available data on mineral zoning of the region indicate that copper has been concentrated with the magnetite deposits of the massif, rather than with the sphalerite deposits of the lowlands. Might one search for significant copper deposits in favorable structural sites between the main magnetite-producing areas and the main sphalerite-producing areas?

The known sphalerite deposits of the northwest Adirondacks are of the Edwards-Balmat type. Within the limits of plate 1, areas in many respects geologically similar to the Edwards-Balmat belt include the Loon Pond syncline, Tupper Lake quadrangle; area between Carr Pond, Russell quadrangle, and Fine, Oswegatchie quadrangle; Van Rensselaer Creek area, Canton quadrangle; and Browns School area, Potsdam quadrangle. All but the first are known to have been prospected for commercial talc. These areas might also contain "blind" deposits of sphalerite.

Yet another type of zinc deposit is of speculative interest. The Parish magnetite-hematite deposit is, in geologic setting and nature of its gangue, remarkably similar to the Broken Hill lode, New South Wales. Is there reason to suppose that, at some favorable structural site near the major magnetite deposits, there may be a sphalerite deposit of Broken Hill type?

If a systematic search for base-metal deposits is ever made within the district, the possibility of finding ore-bearing glacial cobbles and boulders should be kept in mind. In Fennoscandia, for example, careful mapping of the dispersal patterns of ore boulders has led to the discovery of major ore deposits. This very simple prospecting technique is commonly used in conjunction with geophysical and geochemical methods.

INDIVIDUAL MAGNETITE DEPOSITS AND MAGNETIC ANOMALIES

DEFINITIONS AND CAUTIONS

ASSAYS AND GRADE

The iron content of magnetite- or hematite-bearing material is conventionally expressed, in weight percent, in three ways.

Total iron.—Total Fe, determined after treating the sample with HCl and HF, includes not only iron in magnetite and hematite, but also iron in sulfides, silicates, or other compounds. Total Fe, as an indication of the quality of low-grade and medium-grade iron ores of the type occurring in St. Lawrence County, is worthless for evaluating skarn ores and of very limited value for evaluating granite gneiss ores. Its determinataion is a carryover from analytical practice followed on the Lake Superior iron ranges.

Soluble iron.—Soluble Fe is iron soluble in HCl; it includes iron in magnetite, hematite, vonsenite, pyrrhotite, and other minerals readily attacked by HCl. Pyrite is not soluble in HCl, and neither are most gangue minerals in this district (for example, iron-rich biotite). However, some silicates, it seems, may be slightly attacked by HCl. Ilvaite, wihch seems to be rare in the district, is readily decomposed by HCl.

Magnetic iron. Magnetic Fe is HCl-soluble Fe determined on a magnetic concentrate made in a Davis tube. The magnetic concentrate (magnetite) may be very slightly contaminated by free or attached nonmagnetic particles trapped in the water-washed network of magnetite particles suspended in the tube between the poles of a powerful electromagnet. Pyrrhotite may be concentrated magnetically with the magnetite. Martite may also be present, provided the individual martite grains have enough magnetite relics to make them appreciably magnetic. In any event, magnetic Fe determined by the Davis tube method corresponds very closely to the iron content of magnetic concentrates milled from the same ore.

Ore.—The writer has tried to use the term *ore* consistently for a mineral aggregate whose grade and thickness were such that a prudent operator could extract the material at a profit, during the 1940's, if the material were in an operating mine. Departures from such usage are usually indicated in this report by placing the term in quotation marks ("*ore*"), or by inserting parenthetic statements to qualify the customary meaning. Those acquainted with the mineral industry will realize that the working definition of *ore*, at an operating mine, may vary with the mill superintendent's digestion. Therefore, rigorous consistency should not be expected in a geologic report. More serious, however, is the fact that rapid advances in mill practice, especially in the recovery of crystalline hematite from mixed magnetite-hematite ores, make it feasible to mill material whose soluble Fe content is only 10 or 12 percent (Cf. Webb and Fleck, 1950). The grade limits stated below were historically useful, but it is already clear that too little attention was devoted to deposits of potential interest whose grade, during the mid-40's, would have made a mill man snicker.

Grade.—The following definitions of grade were adopted by the Geological Survey in classifying magnetite-bearing material. *High-grade ore* contains 50 percent or more magnetic Fe. *Medium-grade ore* contains 35 to 50 percent magnetic Fe. *Low-grade ore* contains 20 to 35 percent magnetic Fe. *Marginal ore* contains roughly 20 percent magnetic Fe. *Subore* contains less than 20 percent, but generally more than 12 percent, magnetic Fe, the lower limit being flexible. *Mineralized rock* contains appreciably more than accessory amounts of magnetite. When we use the term in the sense of grade, the upper limit is roughly 10 or 12 percent and the lower limit 5 or 6 percent magnetic Fe. Slight departures from these definitions of subore and mineralized rock can be noted on some cross sections illustrating this report.

By substituting soluble Fe for magnetic Fe in the above definitions, they may be applied to hypogene and supergene hematite ores.

RESERVES

Estimates of the reserve of indicated[24] and inferred ore are given in qualitative terms for certain deposits. The data on which these estimates are based are usually presented in such a way that the reader can assess the reliability and degree of completeness of the individual bits of information, and from them make his own reserve estimate. The qualitative estimates provide only a rough measure of the potential importance of the individual deposits.

Assumed tonnage factors are 8 or 9 cubic feet per long ton for medium-grade ores and 10 or 11 cubic feet per long ton for low-grade and marginal ores.

MAGNETIC ANOMALIES

For the interpretation of magnetic anomalies surveyed by dip needle, the writer has leaned heavily on the work of H. E. Hawkes, Jr. (written communication, 1944). The data and the approach set forth in that report have been supplemented by experience. A program of research on the interpretation of dip-needle anomalies was beyond the scope of our investigation, especially as it appeared that the geologists at Benson Mines were already undertaking such an investigation, basing it on an unmatchable wealth of data.

[24] The following definitions are in use by the Geological Survey:

Indicated ore is ore for which tonnage and grade are computed partly from specific measurements, samples, or production data and partly from projection for a reasonable distance on geologic evidence. The sites available for inspection, measurement, and sampling are too widely or otherwise inappropriately spaced to outline the ore completely or to establish its grade throughout.

Inferred ore is ore for which quantitative estimates are based largely on broad knowledge of the geologic character of the deposit and for which there are few, if any, samples or measurements. The estimates are based on an assumed continuity or repetition for which there is geologic evidence; this evidence may include comparison with deposits of similar type. Bodies that are completely concealed may be included if there is specific geologic evidence of their presence. Estimates of inferred ore should include a statement of the spatial limits within which the inferred ore may lie.

Readers are warned not to ascribe great accuracy to the magnetic profiles shown on cross sections. Data are reliable for exploration purposes, but probably not for measurement of the characteristics of the curves in relation to ore at depth. After a given area was surveyed by dip needle, the same grid was used to locate diamond drill holes. For the reliable interpretation of dip-needle results, each magnetic profile should be rerun after the field position of the drill hole has been established. This ties together the magnetic data and drilling results, thereby eliminating errors due to faulty position of magnetic contours, faulty extrapolation from the contours, and shifts in drill sites occasioned by drillers who prefer their own judgment to that of the engineer.

The technique of dip-needling is clearly described by S. A. Tyler (McKinstry, 1948, p. 121–122). The general interpretation of magnetic anomalies due to magnetite deposits of the Dover district, New Jersey, is briefly considered by Sims (1958). The interpretation of magnetic anomalies due to disseminated accessory oxide minerals in country rock is summarized by Buddington (Prof. Paper 376, p. 79–83).

STRUCTURE

The writer's interpretation of the structure of most deposits is highly speculative. It represents a synthesis of available and all-too-sparse surface and underground data and is presented not as canon but as a reasonable basis for undertaking exploration.

MAGNETITE DEPOSITS AND MAGNETIC ANOMALIES OF SKARN TYPE

BRANDY BROOK AND SILVER POND BELTS

The Brandy Brook and Silver Pond belts of magnetic anomalies are near Cranberry Lake village, northeast rectangle, Cranberry Lake quadrangle. (See pl. 1.) Each belt contains a known magnetite deposit, partly explored through diamond drilling by the U.S. Bureau of Mines. The Brandy Brook Northwest deposit is a body of low-grade to medium-grade ore, possibly of lathlike shape. The Silver Pond deposit is of no commercial interest at present. One or more promising magnetic anomalies remain to be tested by drilling.

Millar (1947) has summarized the exploratory work of the Bureau of Mines. Members of the U.S. Geological Survey logged drill core and mapped critical parts of the area. Most of the mapping was done by B. F. Leonard and C. L. Rogers in the spring of 1945. The following account is taken, with slight modifications, from a previously published report (Leonard, 1952).

The Brandy Brook and Silver Pond belts of magnetic anomalies, though not continuous, are closely related spatially and geologically. For that reason, the general geology of the belts is discussed before details of the separate belts are given. The map of the area (pl. 2) was compiled from unpublished magnetic maps (scale, 100 ft = 1 in.) of the Bureau of Mines, to which were tied pace-compass and plane-table traverses of the Geological Survey. The coordinate system of the Bureau of Mines has been extended from the central portion to cover the entire mapped area. The Silver Pond and Brandy Brook Southeast anomalies have been adjusted to this net to give a reasonably accurate picture of the belts. Topographic profiles along the drill sections were not available for all the holes.

GENERAL GEOLOGY OF THE BELTS

The map (pl. 2) shows three belts of metasedimentary rocks and two of granitic rocks trending west-northwest across the northwest part of the area. Owing to the scarcity of outcrops, the geologic mapping is sketchy and the number, position, extent, and relations of the belts are uncertain.

The metasedimentary rocks and granitic rocks are a segment of the great Clare-Clifton-Colton (CCC) belt, in this neighborhood flanked on the northeast by microcline granite gneiss and on the southwest by hornblende-microperthite granite. (See pl. 1. For definition of the belt, see Prof. Paper 376, p. 117.) The broader structural features of the neighborhood, though imperfectly understood, suggest that the principal magnetite deposit (Brandy Brook Northwest) lies at the intersection of two major structural elements of very different trend. These are the easternmost segment of the Jarvis Bridge syncline(?), which trends east-northeast where last mappable on the west shore of the north arm of Cranberry Lake, and the Brandy Brook syncline, which trends southeast in the area between Silver Pond and Brandy Brook. (See pl. 1 and fig. 2; Prof. Paper 376, figs. 14, 15.)

The Brandy Brook magnetic anomalies lie in the southern belt of metasedimentary rocks (pl. 2) which comprise skarn, quartzose gneisses and biotite gneiss, subordinate amounts of marble and amphibolite, and thin sheets of granite. This belt, of unknown width, extends through Bear Mountain Flow and Aldrich Swamp, crosses Silver Pond near its south end, and underlies the abandoned Emporium mill on the west side of the pond. Only the more resistant rocks—quartzose gneisses, amphibolite, and biotite gneiss—are exposed.

Rocks of the middle belt do not crop out within the mapped area but were cut in drill holes on the Brody Tract (Silver Pond anomaly). They are of the same general type as those in the southern belt. Their extension to the southeast is inferred from topographic evidence in the form of a swamp, about 700 feet wide, that crosses N.Y. Highway 3 about 1,800 feet northeast of the bulldozed road leading to the holes on the Gilbert Tract.

The northern belt, consisting of migmatitic biotite gneiss, is exposed along the highway half a mile northeast of the tote road. This belt has a minimum width of 300 feet; it may be 1,000 feet or more wide. The northern and central belts may be parts of a single belt, as the pink granite shown separting them on the map is of unknown thickness and may be rather thin.

The granite between the Brandy Brook and Silver Pond belts of Grenville rocks is pink, weakly foliated alaskite, in part with small nests of red garnet. East of Silver Pond it is medium grained; west of the pond it is predominantly fine grained and has several percent of disseminated magnetite. Locally, very thin partings of biotite gneiss, in some places accompanied by pink sodic granite, are present in the alaskite.

The structural relations of the various belts are uncertain. Though the same types of metasedimentary rocks are present in both the Brandy Brook and Silver Pond belts, no single unit or succession of units can be recognized to establish the identity of the two sequences. If the sequences are structurally equivalent, the most reasonable interpretation of the meager data is that the Brandy Brook belt represents the southwest (upright) limb, and the Silver Pond belt the northeast (overturned) limb, of a syncline that is overturned to the southwest. Alaskite occupies the axial zone of this structure. If the lineation is a fair guide, the axis of the inferred syncline is roughly parallel to the trace of the limbs and (in the neighborhood of Silver Pond) plunges 10° to 25° ESE. The interpretation of the structure as an overturned syncline is suggested by the somewhat gentler dip of the Brandy Brook "limb" as compared with the steeper dip of the Silver Pond "limb." It seems to be confirmed by the downward flattening of the dip in rocks of the Brandy Brook belt (see pl. 2).

Additional geologic mapping must be done before the structure enclosing the Brandy Brook Southeast anomaly can be understood.

BRANDY BROOK BELT

LOCATION, ACCESS, AND OWNERSHIP

The Brandy Brook belt of magnetic anomalies begins at the mouth of Brandy Brook, in the northeast rectangle of the Cranberry Lake quadrangle, and extends roughly N. 55° W. for about 12,000 feet to a point 500 feet southeast of a prominent curve in N.Y. Highway 3. The curve is 0.7 mile by airline east of Cranberry Lake village. The belt may be divided for convenience into three parts, with two principal areas of magnetic highs at the ends and an almost neutral area in the middle.

The Brandy Brook Northwest anomaly lies on the Gilbert Tract of the Newton Falls Paper Mill. A bulldozed road, leaving N.Y. Highway 3 at the curve noted above, runs along the north side of the anomaly. Only the west end of this road appears on plate 2. The anomaly passes through a saddle on hill 1660 (see pl. 1) and continues southeastward down the hill to Aldrich Swamp.

The Brandy Brook Southeast anomaly is on low ground east and west of the Clifton-Colton town line. It may be reached by an old tote road running through Aldrich Swamp, or by boat from Cranberry Lake. The State of New York owns this block of land and holds the mineral rights for the part east of the town line. The Shenango Furnace Co. holds the mineral rights for the Abbot Tract west of the town line. The block lies within the Adirondack State Park. When the State of New York holds the mineral rights for any land within the Park, the existing law forbids exploration or development of the minerals on that land.

The central third of the belt is on the Abbot Tract in Aldrich Swamp. Here the magnetic anomaly is very weak and of no commercial interest.

DISCOVERY AND EXPLORATION

The Brandy Brook belt was found in the winter of 1944 during a reconnaissance dip-needle survey made by the U.S. Bureau of Mines along the Clifton-Colton town line. Compass deflections along the line were noted by Benjamin Wright in 1803 during his original subdivision of Macomb's Great Tract into townships. The U.S. Geological Survey recommended reconnaissance dip-needle traversing along the line and directed its continuation southward to the belt of metasedimentary rocks, where the Brandy Brook belt of anomalies was then discovered. A detailed dip-needle and dial-compass survey of the belt was made by the Bureau of Mines in 1944, and nine diamond-drill holes were bored by that organization in 1944 and 1945. The neighboring Silver Pond (Brody Tract) anomaly was found by the Bureau of Mines, and 3 holes were drilled on it in 1945. The Bureau's work has been summarized by Millar (1947). The results of geologic mapping and core studies by the Geological Survey are presented with the magnetic data in plate 2, as are graphic logs of the drill holes.

MAGNETIC ANOMALIES

The Brandy Brook Northwest anomaly has three substantial strips, arranged tandem, within the 20° isoclinal. The most important strip is 2,200 feet long and generally 100 feet wide. The closed 40° isoclinal within it is 1,300 feet long and 40 to 50 feet wide. This contains four elongate areas within 60° isoclinals. Their aggregate length is 950 feet. Two shorter strips within the 20° isoclinal, each 400 to 450 feet long, are present between grid lines 12400E and 13150 E (pl. 2). Each of these has a length of 200 to 250 feet within the 40° isoclinal. An isolated "hot spot" is found at (10400N, 10100E), and an insignificant strip of +20° readings, 200 feet long, centers at (8400N, 13330E). Overburden above the Brandy Brook Northwest anomaly is relatively shallow, perhaps averaging 20 feet. Five of the eight peaks on the anomaly were drilled.

The Brandy Brook Southeast anomaly contains two long strips within the 20° isoclinal and an insignificant "bead" between them. The strips are arranged tandem. The northwest strip is 1,300 feet long, 30 to 110 feet wide, and has two small areas within the 40° isoclinal. The peak of this anomaly was tested by hole B8. Here the overburden is probably shallow, though it increases rapidly to the southwest. The southeast strip is 1,000 feet long and ranges in width from 40 to 200 feet. It contains an elliptical area of +40° readings, 300 by 100 feet in maximum dimensions. The character of the magnetic profiles across this part of the anomaly suggests that the overburden is thick—possibly about 100 feet. If this is so, the economic possibilities of this part of the Brandy Brook Southeast anomaly may be considerably more favorable than those indicated for the part drilled by hole B8. Indeed, the entire length of the southeast strip within the 20° isoclinal, from 18800E to 19700E, is worthy of prospecting.

Throughout the length of the Brandy Brook belt of anomalies, the magnetic profile is steeper on the northeast side than on the southwest. The steepness of the northeast slope of the Brandy Brook Northwest anomaly is accentuated by a negative anomaly, of intensity generally less than −10°, that runs along part of the northeast side of the positive anomaly. The negative anomaly does not appear on plate 2 because of the 20° contour interval adopted for that map. The asymmetry of the magnetic profile has been taken by some to indicate that the ore zone dips southwest. This view is totally unsupported by the drilling done so far.

On the aeromagnetic map (Balsley, Buddington, and others, 1954a), the Brandy Brook Northwest anomaly shows as a peak of 5,660 gammas. The Brandy Brook Southeast anomaly appears as a local embayment in the magnetic contours and as a magnetic "hill" of maximum value 5,340 gammas.

ROCKS

Four main rock units or zones were cut in drilling the Brandy Brook belt (see pl. 2). These are, from the surface downward: alaskite, felsic gneisses, skarn (the host of the ore), and felsic gneisses with contaminated granite. All the units are continuous except the felsic gneisses above the skarn. These form a lens cut by holes B4, B5, and B6.

The alaskite is pink, medium grained, and homogeneous, with an extremely faint foliation. It contains sporadic nests of red garnet and a few thin, scattered layers or schlieren of metasedimentary rocks. A pink sodic granite that looks much like alaskite has been developed in some places where alaskite comes against biotite gneiss.

The lens of felsic gneisses found locally above the skarn is 12 to 40 feet thick. These rocks are generally gray or greenish, fine grained, regularly foliated or laminated, and granulose. They include quartz-feldspar gneisses (feldspathic quartzites) with a little biotite or pyroxene, biotite quartzite, aplitic granite, and dark biotite gneiss. Some types are slightly migmatitic.

The skarn zone ranges in thickness from 65 to 175 feet and averages about 120 feet. It is thickest at the ends of the belt and thinnest in the neighborhood of holes B4, B5, and B6. The zone consists principally of green pyroxene skarn and some reddish-brown garnet skarn, both of which are massive and medium to fine grained. In addition, subordinate layers of silicated marble, feldspathic skarn, green amphibolite, quartzose gneisses, scapolitic rocks, and contaminated granite are present. Purplish fluorite is visible locally in the feldspathic skarn. Alteration of skarn and some marble to "greenstone"—a soft aggregate of very fine serpentine, chlorite, or both— is prominent in some cores. Hematite grains or stains, and veinlets of calcite, accompany the greenstone or occur independently in the wallrocks. The present evidence suggests that the alteration is mostly due to downward-percolating waters whose course was controlled by faults, joints, marble layers, and certain mica-rich zones.

The lowermost unit, of unknown thickness, is the most heterogeneous. Various quartzose gneisses and contaminated granite predominate, but biotite gneiss, migmatitic pyroxene skarn, and contaminated pegmatite are also present. A little red garnet occurs in some pyroxenic and biotitic quartzites. Generally, the unit is gray or greenish and characterized by an abundance of light-colored minerals—quartz and feldspar.

ORE

Occurrence

All the ore occurs in pyroxene skarn. Minor concentrations of magnetite—subore or mineralized rock—

are present as thin, scattered zones in silicated marble, skarn, feldspathic skarn, and pyroxene-contaminated granite or pegmatite. The magnetite itself has the form of disseminated grains, small blotches or amoeboid masses, and irregular, coalescing layers of variable thickness. Mineralization has been restricted to two zones well inside the skarn unit. In general, the lower zone is thicker, has a higher concentration of magnetite, and is the one of commercial interest. In holes B2 and B8, however, the lower zone is represented only by sporadic grains of magnetite. In hole B5, the upper zone is similarly faint. (These weakly mineralized zones have been omitted from the cross sections, pl. 2.) Magnetite in the lower zone is confined to pyroxene skarn, whereas in the upper zone (still within the skarn unit) magnetite replaces a variety of rocks—skarn, marble, and the pyroxene aggregates in contaminated granite—apparently preferring marble wherever relics of that rock were present. In hole B8, relations are reversed: the upper zone is in skarn and the lower one is in mixed rocks consisting of feldspathic skarn with granite or granite pegmatite veins.

Mineralogy

Magnetite is usually accompanied by some sulfides, chiefly pyrite with a very little chalcopyrite. Pyrrhotite is present locally. The quantity of sulfides ranges from nil to about 10 percent; it averages perhaps 3 or 4 percent. Some thin layers in hole B3 carry as much as 75 percent sulfides. One portion was rich in chalcopyrite (see table 13, footage 213.1–219.8; actually, most of the chalcopyrite was concentrated in less than 1 foot of this sample interval). Ten percent of black hematite accompanies magnetite in some places, particularly where subore is present in marble. This hematite is probably secondary.

Pyroxene is the principal gangue mineral. In places, small amounts of calcite, feldspar, scapolite, apatite, and fluorite are also present.

Grade

Details of the grade of ore and magnetite-bearing rock are given by Millar (1947, p. 10–12). The more important information is summarized in tables 12, 13, and 14.

As tables 13 and 14 show, several sulfide-rich portions of the core from holes B3 and B6 were assayed for gold and silver. The value for Au (0.14 oz per ton) in the sample from 146.2 to 147.7 feet, table 14, may be high. It was not possible to obtain a duplicate of the whole sample; but part of it, assayed for the Geological Survey by Ledoux & Co., gave 0.40 oz Ag per ton and only a trace of Au. There is no check assay for the high value for Ag reported in table 13, 213.1–219.8 feet.

Structure

Planar and linear structure.—The structure of the magnetite bodies is still a matter of conjecture. The upper surface of the higher magnetite-bearing zone in skarn dips about 55° to 60° NE. at hole B1 and about 45° NE. in hole B6. Thus the dip in this neighborhood flattens toward the southeast. In addition, the dip flattens with depth. These relations are shown in plate

TABLE 12.—*Assays of core from drill holes B1 to B9, Brandy Brook Northwest deposit*

[Assays by U.S. Bureau of Mines; data condensed from Millar, 1947, p. 10–12. The table below omits all layers whose thickness is less than 1 ft or which assay less than 12 percent Fe]

Hole	Footage	Apparent thickness (feet)	True thickness (feet)	Weight percent					
				Soluble Fe	Magnetic Fe	P	S	SiO_2	TiO_2
B1	481.5–497.3	15.8	13.5	35.2	33.5	0.018	0.38	10.4	0.06
B2	405.9–408.2	2.3	2.0	28.5	24.9	----	----	----	----
B3	[1] 126.5–153.0	26.5	23.6	27.1	21.8	[2] .040	4.05	17.3	.50
	197.0–213.1	16.1	14.4	42.2	35.9	.038	4.47	10.0	.15
B4	[3] 246.2–253.1	6.9	6.3	26.3	22.2	.033	6.83	20.7	.30
	275.1–283.6	8.5	7.7	28.8	26.7	.020	2.40	14.4	.44
B5	393.8–395.2	1.4	1.2	22.0	19.3	----	----	----	----
	404.8–410.4	5.6	4.9	38.4	36.4	.038	2.06	18.8	.18
B6	127.1–130.9	3.8	3.6	26.5	22.0	----	----	----	----
	140.0–141.0	1.0	.9	29.1	25.0	----	----	----	----
	146.2–147.7	1.5	1.4	19.8	14.5	----	----	----	----
	152.3–153.7	1.4	1.3	16.7	15.6	----	----	----	----
	[4] 163.3–165.3	2.0	1.9	25.3	1.2	----	----	----	----
	165.3–169.2	3.9	3.7	23.7	20.3	----	----	----	----
B7 [5]	----	----	----	----	----	----	----	----	----
B8	88.0–91.2	3.2	3.0	----	28.3	----	----	----	----
	115.6–120.8	5.2	4.9	----	29.1	----	----	----	----
B9	168.2–171.5	3.3	2.8	----	17.8	----	----	----	----
	171.5–188.2	16.7	14.2	----	35.7	----	----	----	----
	188.2–190.4	2.2	1.9	----	18.4	----	----	----	----

[1] Includes 126.5–127.5 assaying 10.3 percent soluble Fe and 5.3 percent magnetic Fe. If this foot of core is excluded, the interval 127.5–153.0 assays 28.3 percent soluble Fe and 23.7 percent magnetic Fe.

[2] Values for P, S, SiO_2, and TiO_2. from unpublished assays made by U.S. Bureau of Mines.

[3] Includes 246.2–247.7 assaying 13.4 percent soluble Fe, 10.3 percent magnetic Fe; 247.7–253.1 assays 30.1 percent soluble Fe, 25.8 percent magnetic Fe.

[4] This interval is hematite rich.

[5] Abandoned before ore zone was cut.

TABLE 13.—*Assays for copper, gold, and silver in ore zones and selected sulfide-rich zones from hole B3, Brandy Brook Northwest deposit*

[Unpublished assays by U.S. Bureau of Mines. See table 12 for apparent and true thicknesses, or multiply apparent thickness by 0.89 to get true thickness]

Footage	Weight percent Cu	Ounces per ton	
		Au	Ag
126.5–153.0	0.06	nil	0.68
153.0–170.0	.01	nil	nil
197.0–213.1	.04	0.02	1.13
213.1–219.8	4.91	.01	1.68

TABLE 14.—*Assays for precious metals in sulfide-rich core from hole B6, Brandy Brook Northwest deposit*

[Assays by U.S. Bureau of Mines; data from Millar, 1947, p. 12. See table 12 for apparent and true thicknesses]

Footage	Ounces per ton	
	Au	Ag
127.1–130.9	0.02	0.14
140.0–141.0	.02	.30
146.2–147.7	.14	.90

2. In general, the magnetite bodies are parallel to the planar structures (foliation) in the wallrocks; but the drilling undertaken so far indicates that the bodies themselves are not continuous and sheetlike in shape. For example, in the main or lower mineralized zone hole B1 cut 13.5 feet of ore; hole B2 cut thin, scattered stringers of magnetite-bearing rock; hole B3 cut two layers of ore, 23.6 and 14.4 feet thick. The limited drilling fails to determine whether the magnetite occurs as small pockets, commercially unattractive, or as shoots of a size warranting development. The apparently discontinuous distribution of magnetite, together with the presence of a pronounced lineation in certain wallrocks, suggests the hypothesis that the ore occurs as shoots parallel to the linear structure (pl. 2). Plunging ore shoots of this type are typical of magnetite deposits of the New Jersey Highlands and are also found at the 81 mine, Lyon Mountain, New York (Zimmer, P. W., 1948, oral communication; Postel, 1952, p. 81). However, these shoots are not in skarn, and at present we do not know if ore bodies in skarn behave the same way. (But see this report, p. 73–75; also Sims and Leonard, 1952, p. 22, 35.)

Near the center of the Brandy Brook Northwest anomaly, where the foliation strikes about N. 62° W. and dips about 60° NE., the lineation trends S. 70° E. and plunges 17° ESE. Where the dip flattens to 45°, the plunge decreases to 10°. Following this trend two hypothetical ore shoots are shown schematically in horizontal projection in plate 2. Of course, the dimensions of the shoots are unknown. The hypothesis of pitching ore shoots parallel to the lineation helps to explain the irregular distribution of ore cut in holes B1 through B6 and suggests where additional drilling might be done to follow the good showings of ore.

Detailed geologic mapping of the few outcrops in the neighborhood of the Brandy Brook Southeast anomaly must be done before the structure is clearly understood there. Present information indicates that foliation in outcrops closest to the anomaly, north of it, dips about 10° SW. (locally 30° SW.), whereas the drilling results from holes B7 and B8 indicate that the rocks underlying the anomaly dip northeast, probably at about 25°. The lineation in the wallrocks is horizontal. Thus the anomaly appears to lie very close to the axis of an open syncline, either a minor one or the prolongation of the major structure inferred to be an overturned syncline in the neighborhood of Silver Pond.

Faults.—The map (pl. 2) shows a small, isolated magnetic "hot spot," centered at (10400N, 10100E), detached from the main Brandy Brook Northwest anomaly. The "hot spot" was drilled by hole B9, which passed successively through alaskite, an ore-bearing skarn zone, and a thick sequence of footwall gneisses. The apparent thickness (and probably the true thickness) of skarn and ore in hole B9 is about equal to the true thickness of corresponding members cut in hole B1. The dip of the foliation ranges from 70° N. to vertical (see section through hole B9). From 481 to 511.5 feet, hole B9 cut broken, chloritized, hematite-stained, calcite-seamed rock that had to be grouted repeatedly.

The facts can be interpreted in several ways: (a) sharp swing in foliation from northwest to north and extreme thinning of rock units beneath the "hot spot"; (b) sharp right-angle or zigzag fold that still maintains the same thickness for the units and gives them a northwesterly strike beneath the "hot spot"; or (c) fault of unknown direction and displacement passing through the broken zone in footwall rocks and giving the ore zone an apparent horizontal displacement of approximately 200 feet. The hypotheses cannot be developed in detail in this report. The assumption of a fault fits more facts and raises fewer questions. For that reason, it has been favored in preparing the geologic map (pl. 2).

To satisfy the positions of the anomalies and the results of drilling, the fault must strike just a little east of north. This raises the question why a very thick sequence of footwall gneisses was cut by hole B9—that is, why the ore zone was not cut twice. If the fault has a moderate eastward dip, all relations appear to be satisfied. However, a vertical dip seems more likely, judging from the nature of known Adirondack faults. This will not explain the failure to cut the ore zone twice, unless the drill hole was deflected westward after intersecting the broken zone. With the bearings of the hole and the hypothetical fault meeting at an angle of 20° or less, such a horizontal deflection seems not only reasonable but necessary. As far as the writer knows, the inclination of the bottom of hole B9 was not checked, and none of the holes was surveyed for azimuth after drilling. Thus there is no check on the supposition of a deflected drill hole.

One way of getting additional structural information was proposed at the time of drilling. Overburden between the "hot spot" and the main anomaly is certainly shallow—probably 10 feet or less. A bulldozer could clear the critical area without difficulty, permitting examination of the bedrock for evidence of faulting or sharp folding.

Another fault is inferred from zones of altered, broken rock found in holes B1 and B2. The magnitude and significance of this fault are also unknown. The horizontal projection of broken zones in holes B9, B1, and B2 falls on the same line. However, a single fault, striking east-southeast, will not by itself explain the problems outlined in discussing hole B9 above.

Other zones of broken rock are present in the cores, but so little is known of their relations that speculation about them seems futile at this time.

Many major faults in the Precambrian show little evidence of breccia, mylonite, or alteration. Elsewhere, these features are present where movement has been negligible. For these reasons, we can only call attention to the evidence, note its implications, and wait for additional work on which to base reliable conclusions.

ECONOMIC POSSIBILITIES

At present, the most favorable area for additional drilling lies between 10000E and 12000 E (pl. 2), particularly between 10000E and 11200E. The possibility that the ore occurs in shoots parallel to a lineation in the wallrocks has been mentioned. The hypothetical shoots could be tested by a series of vertical holes laid out along their inferred trend. Though the present drilling data show the dip flattening somewhat with depth, the possibility that it may steepen again—giving rolls—is always present. Until more is known about the structural habit of the ore, no intelligent guess can be made about reserves. There is a good chance that ore bodies of commercial interest may be developed by intelligent drilling and geologic study of the area.

The 1,000-foot strip of +20° readings with a low peak at (4250N, 19300E) in the Brandy Brook Southeast anomaly deserves to be drilled. The character of the magnetic profiles suggests that the overburden may be thick—perhaps 100 feet—and may be masking a promising ore body. Such an interpretation is obviously not certain but is worth checking. Mineral and surface rights on this block are owned by the State of New York.

Reconnaissance dip-needle work along the projection of the Brody Tract anomaly has not been done, so far as the writer knows, except for a small area in the northern part of Silver Pond. Drilling on the Brody Tract showed nothing of immediate commercial interest, and the aeromagnetic survey (Balsley, Buddington, and others, 1954a) detected no sizable anomalies southeast or northwest of the Brody Tract anomaly. Nevertheless, reconnaissance traverses are warranted, as they might at least yield information of value in interpreting geologic structure.

SILVER POND (BRODY TRACT) BELT

LOCATION, OWNERSHIP, AND MAGNETIC ANOMALIES

The magnetic anomaly on the Brody Tract, now or formerly owned by the Emporium Forestry Co., Conifer, N.Y., lies northwest of the north end of Silver Pond, northeast rectangle, Cranberry Lake quadrangle. It begins 400 feet northwest of the "Y" on the recently abandoned Grasse River Railroad and runs N. 58° W. Three very short strips of +20° readings, arranged tandem, are present in the belt. (See pl. 2.) Two of these, each about 250 feet long and 20 to 50 feet wide, were drilled by the Bureau of Mines in the summer of 1945. The anomaly, sometimes called the Brandy Brook Flow Extension, was found by the Bureau of Mines when it extended its survey of the Brandy Brook belt to the northwest. The area between the Brody Tract anomaly and the Brandy Brook Northwest anomaly, including much of Silver Pond, was also surveyed by dip needle. No significant anomalies were found there. The Geophysical Division of the Bureau of Mines traversed part of the same area with an Askania magnetometer. As far as the writer knows, nothing of consequence was found. The work was done to check one interpretation, advanced by the Bureau, that the two discontinuous magnetic belts (Silver Pond and Brandy Brook) reflect parts of a single (canoe-shaped?) ore shoot that trends northwest and is too far beneath the surface to give a readily detected anomaly in the area between 7500E and 10000E. Geologic mapping, admittedly sketchy because of the scarcity of outcrops, lends no support to that interpretation. Geologic relations of the two belts are summarized in a preceding section of this report.

The Silver Pond anomaly shows on the aeromagnetic map as a magnetic "ridge" whose maximum value is 5,230 gammas (Balsley, Buddington, and others, 1954a).

ROCKS AND STRUCTURE

No rocks are exposed within 1,000 feet of the Brody Tract anomaly. Drilling of holes B10, B11, and B12 indicates that the rocks at the anomaly are, from the top downward: interlayered metasediments and amphibolite; hornblende granite gneiss, locally biotitic; skarn with magnetite; pyroxenic and biotitic gneisses and quartzites; and alaskite. (See pl. 2.)

The uppermost unit, of unknown thickness, consists of many relatively thin layers of biotite gneiss (locally garnetiferous), pyroxene gneiss, amphibolite, subordinate migmatitic mica-pyroxene skarn, and a little silicated marble or diopside skarn. The upper part of the unit is mostly biotitic, the lower part pyroxenic. Locally the biotite gneiss has 5 percent of disseminated magnetite; and a 0.4-foot layer rich in magnetite was observed in hole B11 in the skarn just above the granite gneiss sheet.

The granite gneiss sheet is 30 to 60 feet thick. It is pink or pinkish gray, well foliated, fine or medium grained, and has 10 to 20 percent of mafic minerals. Hornblende predominates, but some of the rock is biotitic. Though zones of "clean" granite are present, much of the rock is contaminated, with the mafic minerals rather evenly distributed or else concentrated in schlieren. A very little disseminated magnetite is present locally.

The principal skarn zone is 12 to 30 feet thick. It is mostly of dark-green, medium-grained pyroxene skarn. In places garnet, feldspar, scapolite, and calcite accompany some of the pyroxene. A 10-foot parting of gneissic rocks splits the skarn in hole B11. (This has been omitted from the cross section, pl. 2, in order to emphasize the presence of a zone of skarn.)

The unit below the skarn is 20 to 40(?) feet thick. Like the uppermost unit, it is heterogeneous. Biotite quartzite, in part feldspathic, predominates. Biotite gneiss, pyroxene gneiss, amphibolite, a little skarn, and contaminated granite are also present, though not all in the same hole.

The alaskite sheet, cut only by hole B11, is a fine- to medium-grained, pink, faintly foliated rock that is identical in composition with the hanging-wall alaskite of the Brandy Brook belt.

The repeated sequences of biotite quartzite-biotite gneiss, amphibolite, and marble (or its metasomatized equivalent, skarn) in the Silver Pond drill holes are suggestive of cyclic sedimentation from sandstone to shale to limestone. Not knowing the phase in which any given cycle began, we cannot use the cycles to determine tops and bottoms for the sequence. If the phase were known, it might be possible to prove or disprove the hypothesis that the metasedimentary rocks of the Silver Pond belt are on the overturned limb of a syncline.

The rocks strike N. 58° W. and dip 65° to 75° NE. The dip is steeper than in the Brandy Brook belt, and dips measured on the cores are consistent with those recorded on the first outcrops south of the Silver Pond anomaly.

"ORE"

Magnetite concentrations of immediate commercial interest were not found in drilling the anomaly. Several thin zones relatively rich in magnetite are present in the main skarn zone. These are described below. In addition, a 0.4-foot layer of high-grade "ore" was cut just above the contaminated granite sheet in hole B11. Disseminated grains and layers of magnetite, with or without a little sulfide, replace pyroxene skarn, forming mineralized layers 2 to 6 feet thick. In holes B11 and B12, which cut the main skarn zone about 100 feet below the topographic surface, the magnetite occurs as two layers separated by 8 to 10 feet of barren skarn. Only one layer rich in magnetite was cut in hole B10, and that was at a depth of 300 feet below the surface. Assays by the Bureau of Mines (Millar, 1947, p. 13) gave:

Hole	*Footage*	*Magnetic Fe (percent)*
B10	356. 3–358. 7	33. 6
B11	141. 0–146. 0	28. 7
	[1] 150. 0–158. 0	22. 9
B12	161. 6–164. 0	36. 6
	[2] 170. 6–181. 5	20. 1

[1] Includes some very lean material from hanging wall.
[2] Includes some very lean material from footwall.

ECONOMIC POSSIBILITIES

Drilling results were better than expected, considering the feeble anomaly. No one can say whether the magnetite-bearing layers are continuous between the two magnetic "highs" that were drilled. Under present conditions, the anomaly scarcely warrants additional drilling.

BRUNNER HILL PROSPECT

The Brunner Hill prospect is in the northeast rectangle, Stark quadrangle, at approximate lat 44°26′ N., long 74° 49′ W., 1.5 miles south of Huggards, a farm on N.Y. Highway 56. A graveled logging road, running from Huggards to the Granshue Club, crosses the center of the magnetic anomaly, which is 1.8 miles by road from Huggard's gate. According to Reed and Cohen (1947, p. 9), the property is in the west half of Lot 12, Township 8, Town of Colton, and is owned by the Northern Development Corp., Syracuse, N.Y. (R. B. Large, superintendent of lands).

The magnetic anomaly was discovered by the aeromagnetic survey made in 1945 by the U.S. Geological Survey. In 1946, a "ground check" was made by the Survey and a detailed dip-needle map prepared by the Bureau of Mines. The Bureau drilled one hole, 329 feet deep, on the northeastern half of the anomaly in November 1946.

The anomaly trends northeast. Within the 40° isoclinal contour, it is 1,000 feet long and 100 to 200 feet wide, terminating sharply at the northeast end. Two peaks with intensity greater than 60° are centered 600 feet apart within the 40° contour. (See fig. 5.)

The prospect lies north of the Brunner Hill mass of hornblende granite and south of a broad drift-covered area. (See pl. 1.) Fine-grained pink granite crops out 200 feet northwest of the Bureau of Mines base line. Immediately south of the fine granite is a belt of gray medium-grained hornblende syenite gneiss. Foliation in the outcrops strikes N. 50° E. and dips 55°–80° NW. The average dip measured on drill cores is 68° NW.

A layer of ore 13.5 feet thick was cut near the base of a skarn zone 56 feet thick. Neither ore nor host rock was exposed at the surface. Analysis of a composite sample of the ore, interval 251.0 to 266.3 feet, gave 43.6 percent magnetic Fe; the magnetic portion of the sample assayed 69.8 percent Fe, 0.06 percent S, 0.10 percent TiO_2, and 0.01 percent P (Reed and Cohen, 1947, p. 34). Magnetite with sporadic sulfides replaces green medium-grained calcareous pyroxene skarn. Within the 13.5-foot layer of ore, high-grade zones alternate with slightly leaner ones and a few that are almost barren, yet the overall grade is fairly uniform. The unmineralized skarn is mostly pale-green pyroxene skarn with several layers of calcareous garnet-rich skarn and silicated marble. Locally, the garnetiferous skarn is streaked with pink feldspar and yellow-green epidote. Allanite is abundant in some of the epidotized rock.

A simplified graphic log is given in figure 5, which also shows the magnetic anomaly, outcrops, and projected rock units.

BURNTBRIDGE POND ANOMALY

The Burntbridge Pond anomaly was discovered by the aeromagnetic survey of 1945, on which the anomaly was registered as a peak of 5,500 gammas. A reconnaissance dip-needle survey of the anomaly was made that same year by H. E. Hawkes (see fig. 6), whose published account (Hawkes and Balsley, 1946, p. 4) is the basis of the following description. The anomaly is along the north shore of Burntbridge Pond, northwest rectangle, Tupper Lake quadrangle.

The road shown on the topographic map (pl. 1) is passable to automobiles as far as the Grass River Club. An old woods road leads west from there, as shown accurately on the topographic map. At the southwest end of the second large clearing, about 1.5 miles from the Club, a very poorly defined trail leads south to within a few hundred feet of Burntbridge Pond. Because of the confusion of trails and landmarks, a person unfamiliar with the country might do well to follow a direct compass course from the Grass River Bridge.

The magnetic map of the Burntbridge Pond anomaly (see fig. 6) is incomplete, as no boat was available to make possible a survey of the part of the anomaly which extends into the lake. A complete magnetic survey of the anomaly would require either work from a boat or work in the winter over the ice. As surveyed, the strike of the axis of the anomaly is about N. 60° E. The anomaly terminates abruptly at the northeast end; toward the southwest, the maxima increase progressively as far as it was possible to take readings, suggesting that the most intense part of the anomaly lies offshore.

No outcrops were found in the immediate vicinity of the anomaly. Geologically, the anomaly is in the synclinorial Clare-Clifton-Colton belt of metasedimentary rocks and granite sheets, close to its inferred contact with a large mass of hornblende granite. (See pl. 1.) The Brandy Brook Southeast magnetite deposit is in the same belt, 4 miles west-southwest of the Burntbridge Pond anomaly. Skarn is known to be abundantly present in the metasedimentary rocks east and west of Burntbridge Pond. It is reasonable to infer from this fact, as well as from the characteristics of the dip-needle anomaly, that the source of the Burntbridge Pond anomaly is a magnetite body of skarn type, mantled by glacial debris and water. Foliation in outcrops closest to the pond, but still half a mile distant, strikes N. 65° E. and dips 12°–15° NW. (See pl. 1.) The anomaly is near an area in which lineations of similar or slightly divergent trend plunge toward one another. Thus the structure of the inferred magnetite body at Burntbridge Pond cannot be predicted with assurance, though the trend of the body is almost certainly northeast to east.

Since the anomaly is on State land within the Adirondack Forest Preserve, the prospects for exploration are slight.

CLAFLIN SCHOOL (ENSLOW) DEPOSIT

LOCATION AND HISTORY

The Claflin School, or Enslow, deposit is 0.65 mile northwest of Claflin school, southwest rectangle, Potsdam quadrangle. Floyd M. Enslow, Colton, N.Y., owned the deposit in 1948.

To reach the deposit, drive 0.6 mile northwest from Claflin school on the gravel road that runs to Pierrepont. Walk north 500 feet through the fields to the prospect, which lies in a small north-trending valley.

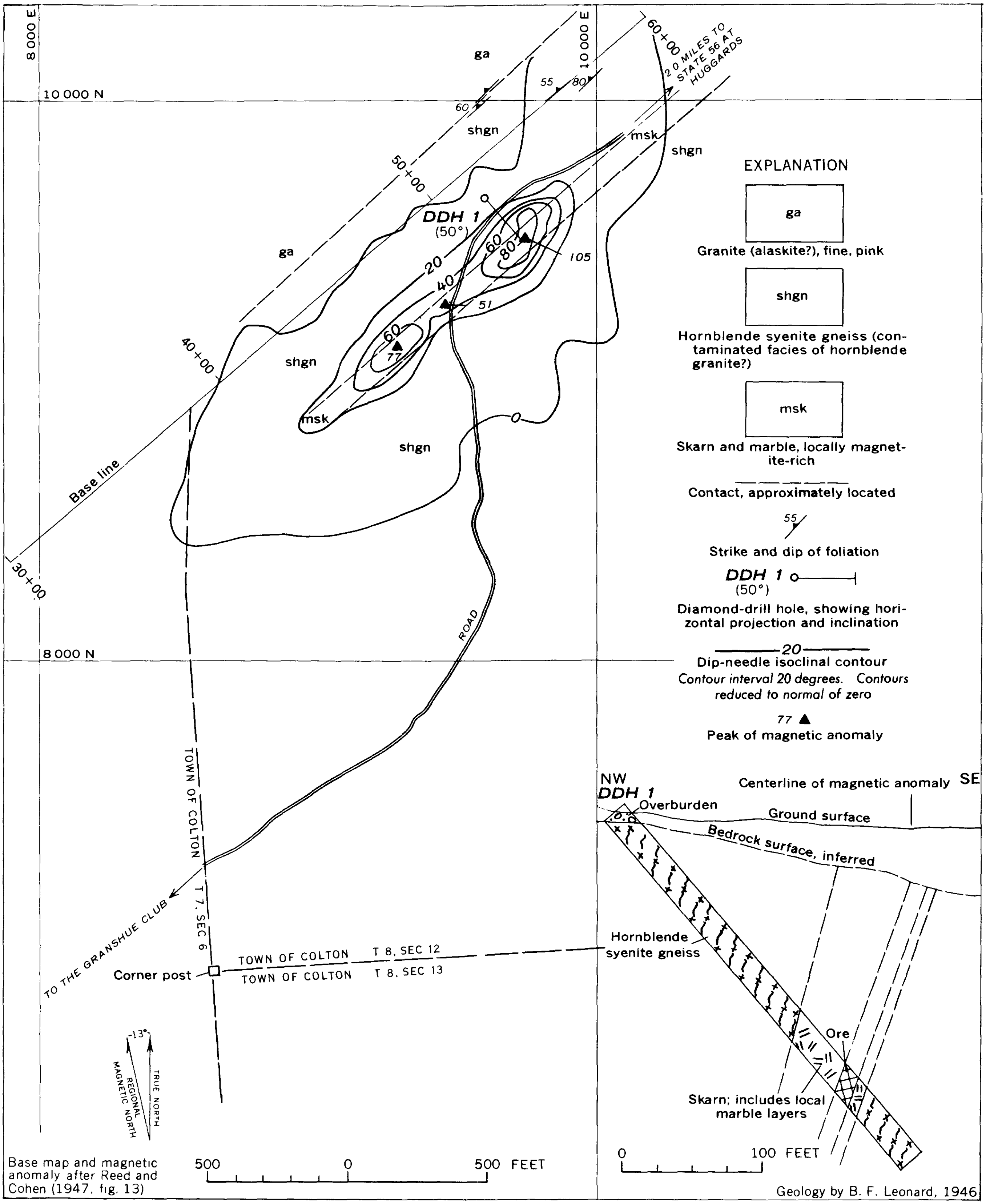


Figure 5.—Bedrock geologic and magnetic map of the Brunner Hill deposit and cross section through diamond-drill hole 1, northeast rectangle, Stark quadrangle.

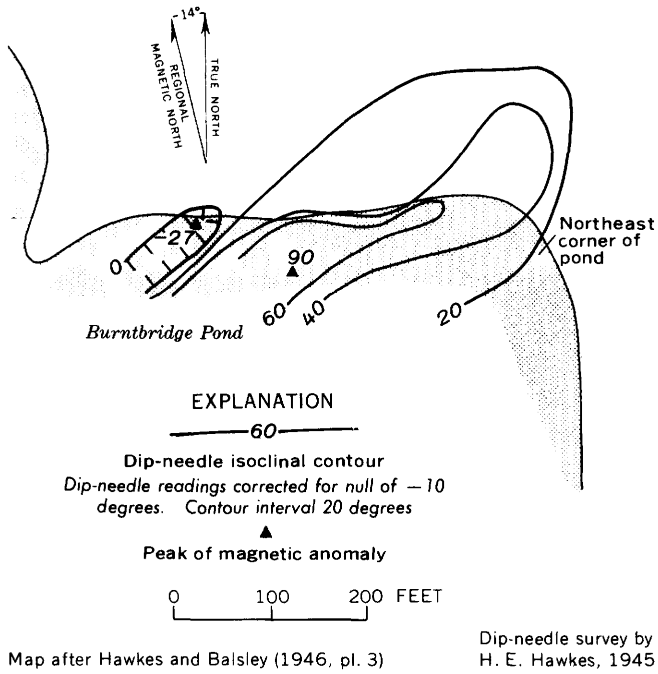


Figure 6.—Reconnaissance magnetic map of the Burntbridge Pond anomaly, northwest rectangle, Tupper Lake quadrangle.

The position of the workings is shown in figure 7. All workings and dumps are badly overgrown with brush. The trenches are shallow, and their walls are for the most part concealed by slumped surficial debris. The northwest shaft is completely choked with debris. The southeast shaft is full of water. The attitude of the shafts could not be determined.

The following notes were provided by the owner in an interview with Mr. C. L. Rogers, of the Geological Survey, on September 14, 1948:

Very little is known of the history of the deposit, which was supposedly opened up by a man named Vanderhool in the late 1880's or early 1890's. Some ore was shipped—perhaps in the 90's—but the quantity and destination are unknown.[25] The two shafts are said to be 185 feet deep, coming together at the bottom. A diamond-drill hole extended from the bottom of the shafts to an unknown depth.

GEOLOGY, MAGNETIC ANOMALY, AND MAGNETITE DEPOSIT

The magnetic anomaly given by the deposit is intense (+90°) right around the northwest shaft. The length of the anomaly within the zero isoclinal is less than 100 feet, and its maximum width is about 60 feet. The pattern of intense highs and lows within this anomaly is erratic, and the magnetic declination changes rapidly from place to place. It was not possible to determine

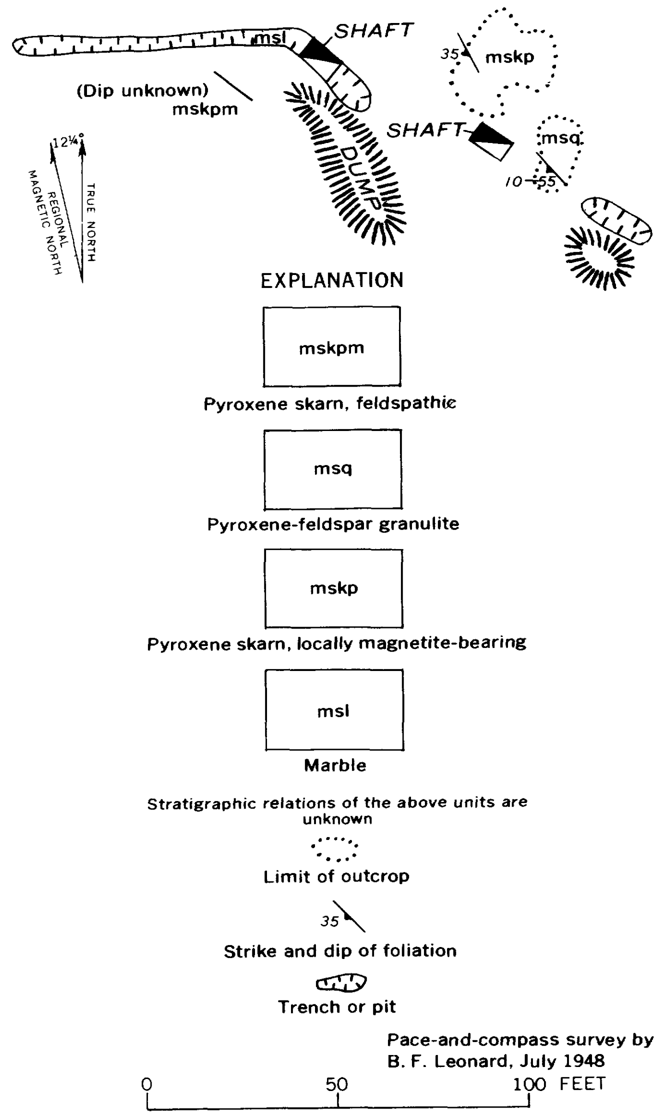


Figure 7.—Sketch map of the Claflin School (Enslow) deposit, southwest rectangle, Potsdam quadrangle.

how much of the erratic pattern was due to thinly concealed magnetite in place, and how much to magnetite on the grown-over dumps.

The Claflin School deposit failed to register as a significant anomaly on the aeromagnetic survey.

Material of ore grade was not seen in place. The only rocks exposed are, from the apparent bottom upward, (a) dark-green fine-grained pyroxene skarn, locally carrying specks and stringers of magnetite; (b) fine-grained friable pyroxene-feldspar granulite containing about 2 percent pyroxene; (c) coarse-grained blue calcitic marble containing a trace of pyrite and pyrrhotite; and (d) feldspathic pyroxene skarn. Possibly the marble and feldspathic pyroxene skarn are not in place. A few fragments of skarn on one dump contained enough magnetite to constitute subore.

[25] From 1828 to 1830, a forge with two fires was operated near Colton, N.Y., producing iron mainly from magnetic ore. It is possible that magnetite from the Claflin School deposit was treated at this forge. If that were so, the inferred and reported dates of the opening of the deposit would not agree.

The exposed rocks strike northwest and dip southwest. A roll is suggested by the variable dip of the foliation in the outcrop southeast of the southeast shaft. It is possible that the feldspathic pyroxene skarn exposed west of the northwest shaft dips northeast.

The grade and size of the deposit are unknown. The small size of the magnetic anomaly, the apparent absence of workings except prospect trenches and prospect shafts, and the absence of a significant aeromagnetic anomaly suggest that the magnetite deposit was small. The magnetite may have been deposited on or close to a minor fold associated with the southwest nose of the nearby Colton anticline. (See pl. 1.) No other magnetite deposits are known in this neighborhood. Presumably the Claflin School deposit represents weak magnetite mineralization of the type found close to the border between the dominantly igneous Adirondack massif and the dominantly metasedimentary lowlands. A small pyrrhotite prospect lies 0.9 mile north-northwest of the Claflin School deposit (Reed, 1934, p. 62).

With respect to the regional geology, the Claflin School deposit is in a major belt of metasedimentary rocks, just south of a thin sheet of alaskite gneiss and close to the projected axis of the Colton anticline. (See Prof. Paper 376, fig. 14.) The boundary between areas of parallel lineation and areas of discordant lineation is very close to the deposit. (See Prof. Paper 376, fig. 15.)

CLIFTON DEPOSIT

LOCATION AND ACCESS

The Clifton mine of the Hanna Coal and Ore Corp. is 3.5 miles southeast of the hamlet of Degrasse, southeast rectangle, Russell quadrangle. (See pl. 1.)

To reach the mine, turn south at the Degrasse church and drive 4.3 miles along the dirt road to the mine gate. Follow the old tote road, the railroad spur, or the main company roads to the opencut.

HISTORY, MINING METHODS, AND PRODUCTION

The Clifton deposit was probably discovered before 1840. The first systematic mining began about 1865, from small pits and shafts. In 1865, the Clifton Iron Co. built a 20-mile wooden railway from East DeKalb to the mine. A few months' use wrecked the railway. In 1866, the Clifton Iron Co. erected a furnace at Clarksboro, 2 miles north-northeast of the mine. In 1868, a furnace was erected at the mine. Mining presumably ceased in 1870, for in that year the Clarksboro forge was closed down. Soon after, the hamlet of Clarksboro was deserted.

Judge Northrup, who spent a night at Clarksboro in the summer of 1878 while on the way to a fishing expedition on Cranberry Lake, wrote (Northrup, 1880, p. 218):

> * * * we enjoyed and employed a part of [the early evening] in strolling about the deserted iron works, and inspecting the large buildings filled with slowly decaying charcoal, the heaps of valuable ore, the *disjecta membra* of heavy and costly machinery, and the falls, on the very brink of which the dam for the iron works had been erected. Here was another of those wrecks of great business enterprises where "somebody blundered,"—the blunder in this case being in forgetting the cost of transporting a ton of iron over a crazy, wooden rail-way out to civilization. The iron is there in abundance and of excellent quality, wood is plenty enough for charcoal to reduce the ore—but the ruin of the "Clinton [*sic*, for Clifton] Iron Works" tells the rest.

For 70 years the Clifton mine lay idle, several schemes for its further development having aborted. In 1940, the M. A. Hanna Co. made a detailed dip-needle survey of the area and began diamond drilling the main deposit. In July 1941, the company decided to develop the property. A road was built from Degrasse, and powerlines and telephone lines were extended to the mine. The Carthage and Adirondack Branch of the New York Central Railroad was extended 12 miles from Newton Falls. In December 1941 the first train arrived at the mine. Stripping for an open pit (figs. 8, 9) was begun that winter, and in June 1942 the first lump ore was shipped from the pit. Shipments of sinter followed. The sinking of a 13- by 15-foot, 6-compartment vertical shaft was begun in July 1942, and development of the first level (270 feet below the shaft collar) was undertaken in 1943. The shaft was deepened in 1948, permitting the second level to be developed, 200 feet below the first, in 1949.

The moderately dipping ore body was mined to a vertical depth of 120 feet in the open pit (fig. 8), ore being trucked about 3,000 feet to the primary crusher. Underground mining was by sublevel stoping. The six sublevels above the first level were spaced generally 25 feet apart, and most of the resulting stopes were holed through into the open pit. Sublevels between the first and second levels were fewer in number and less regularly spaced. Ore was drawn at the main levels through cross-haul scram drifts set 7 feet above the sill of the level. Some of the scram drifts were in ore, or just in the footwall; others were some distance below the stope floors, to which they were connected by one or more transfer raises. Where the ore body had a moderate or steep dip, broken ore ran by gravity to the scram drift, or to the transfer raise above the scram drift. Where the ore body flattened, broken ore had to be scraped from the stope floors. Electric trams on the main levels hauled ore to the shaft, where it was raised in balanced skips, dumped into trucks, and hauled 600 feet to the primary crusher. By the combination of open-pit and underground mining, the ore

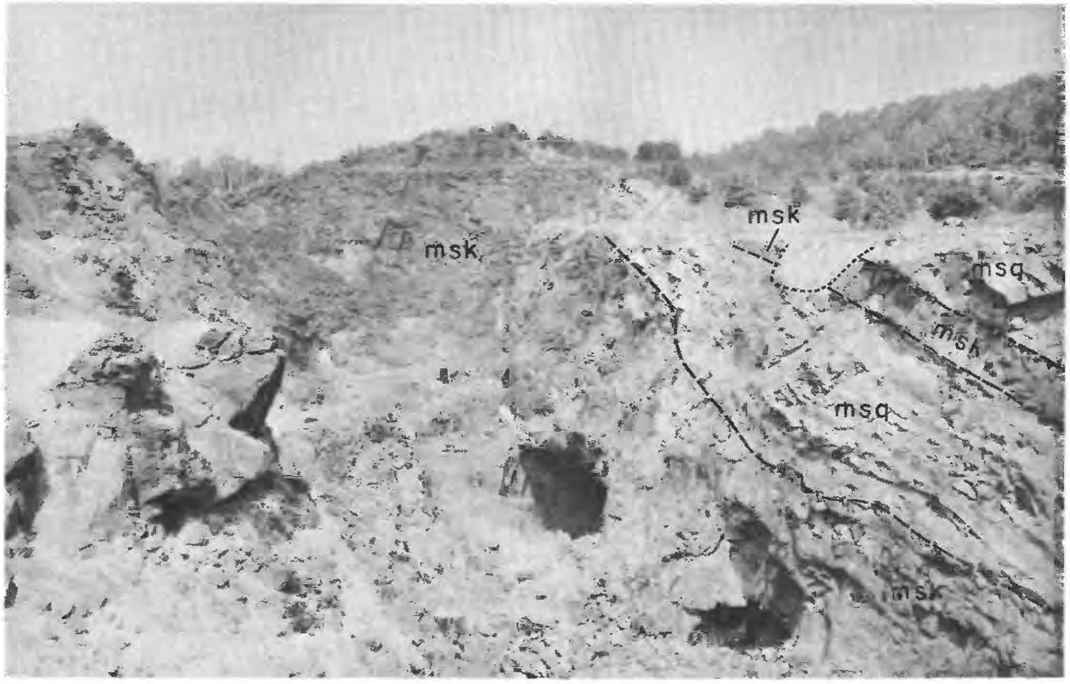


FIGURE 8.—Open pit of Clifton mine, looking northeast. Note sag in hanging-wall quartz-feldspar granulite (msq). Subordinate skarn lens (msk) in hanging wall is at right, just below bench.

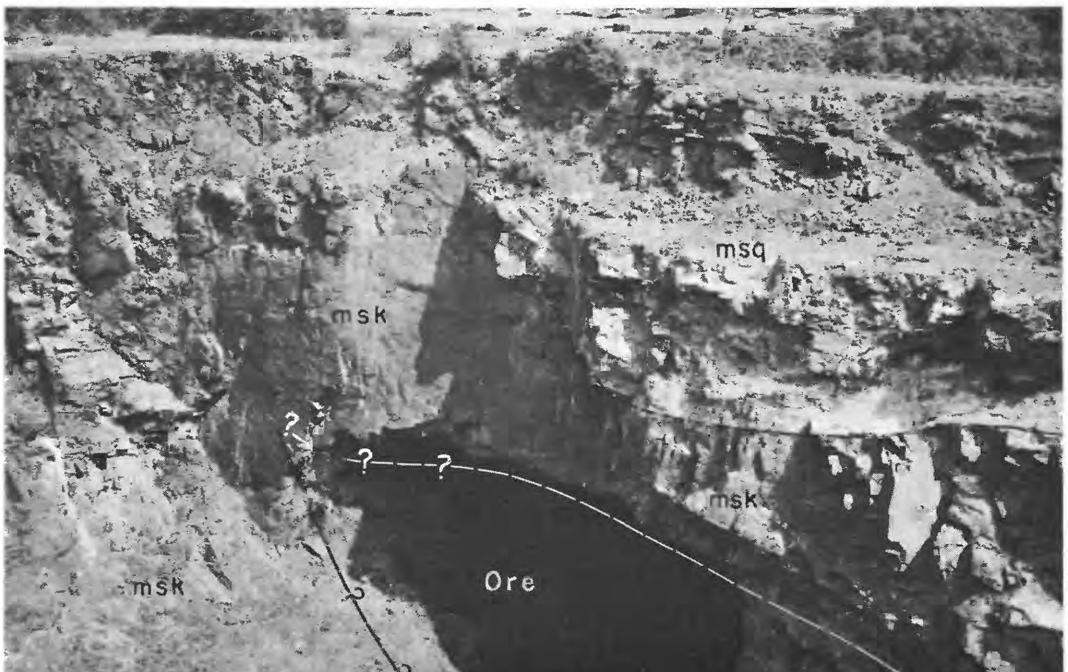


FIGURE 9.—Skarn pods (msk) above ore body and below hanging-wall quartz-feldspar granulite (msq). Left pod is on west flank of sharp anticline that marks eastern edge of main ore body (Ore). Pods and anticlinal axis trend south-southeast (toward the right) and plunge 20°–25° SSE. Clifton mine, northeast end of open pit, above No. 15 stope.

body was worked down the dip for about 700 feet, equivalent to a vertical depth of 500 feet.

The Clifton mine was closed in January 1952. According to one account (Eng. Mining Jour., 1952, p. 176), developed ore had been exhausted and reserves were not sufficient to warrant further development. Another account (Mining World, 1952, p. 95) cited the exhaustion of developed ore and the prohibitive cost of deeper development.

During the period 1942–51, 2 to 2½ million long tons of crude ore was mined; the grade of this ore probably averaged about 38 percent magnetic Fe. The crude ore yielded 1 to 1¼ million long tons of sintered magnetite concentrates averaging 62.5 to 63 percent Fe, and 20,000 to 25,000 long tons of lump ore averaging 58 to 59 percent Fe. Lump ore was shipped from 1942 to 1945, after which time it was not feasible to separate potential lump ore from the mill feed. Because the quantity of lump ore shipped was relatively small, the average grade of the mill feed for 9½ years must have been very close to 38 percent magnetic Fe. The grade of crude ore and shipping products, and the composition of the tailings, are given in tables 15 and 16.

TABLE 15.—*Chemical analysis (weight percent) of milling ore, shipping products, and tailings, Clifton mine*

	Milling ore		Lump ore		Sinter		Tailings
	1	2	3	4	5	6	7
Total Fe	42.25	42	58.16	58.96	62.45	62.55	(1)
Magnetic Fe	35.9						
P	.170	.127	.175	.124	.021	.027	[2] (.58)
S	.50	.4	.24	.327	.010	.040	.43
TiO_2	.18		Tr.				.20
Mn	.16		.41	.31		.27	[3](.31)
SiO_2	19.50		8.26	8.9	4.97	5.54	44.87
Ignition gain or loss				+2.13		−1.60	
Al_2O_3	1.75		1.86	1.73		2.41	7.61
Fe_2O_3							[4] 4.93
FeO							[4] 7.74
MgO	5.86		2.16	2.13		1.56	9.80
CaO	9.12		5.55	3.25		2.05	19.97
Na_2O							.52
K_2O							1.31
H_2O+							.55
H_2O-							.05
CO_2							.98
F							.50
Total							100.35
Less O							[5] .43
Total (corrected)							99.92

1 ~9.5 percent. Calculated from Fe_2O_3+FeO reported by analyst.
2 0.58 percent P_2O_5. Equivalent to 0.25 percent P.
3 0.31 percent MnO. Equivalent to 0.24 percent Mn.
4 Because of the presence of acid-soluble sulfides, the ratio FeO:Fe_2O_3 is not reliable [analyst's note].
5 Correction for S based on the assumption that all S is present as FeS (pyrrhotite). Actually, pyrite, pyrrhotite, and other sulfides were present in the sample, but the existing correction is the only one feasible for the analyst to make.

1. Walker (1943, p. 519); supplemented by Durfee (1943, p. 521).
2. Oliver (1945, table 4).
3. Walker (1943, p. 519).
4. Oliver (1945, table 5).
5. Durfee (1943, p. 522).
6. Oliver (1945, table 5).
7. Sample CN–49–41, collected by B. F. Leonard. Sample is a composite of samples collected on 75-ft centers over southwest half of tailings pile; it represents uppermost 1-ft layer of tailings as of September 21, 1949. Analyst, Lucille M. Kehl, U.S. Geological Survey, 1953.

TABLE 16.—*Semiquantitative spectrographic analysis for minor elements in tailings, Clifton mine*[1]

[Analyst, R. G. Havens, U.S. Geological Survey, 1953. For major constituents and description of sample, see table 15, No. 7]

	Percent		*Percent*
B	[1] 0.0x⁻	La	0.00x⁺
Ba	.0x⁻	Ni	.00x⁻
Be	.000x	Sn	.00x
Co	.00x⁻	Sr	.00x⁺
Cr	[1] .000x⁺	V	.00x⁻
Cu	.00x⁺	Y	.000x⁺
Ga	.000x	Zr	.00x⁺

Looked for but not found: Ag, As, Au, Bi, Cd, Ce, Ge, Hf, Hg., In, Ir, Li, Mo, Nb, Nd, Os, P, Pb, Pd, Pt, Re, Rh, Ru, Sb, Sc, Sm, Ta, Te, Th, Tl, U, W. Zn.

1 Example of plus-and-minus notation:
0.00x⁺=0.00464 to 0.01 percent
.00x = .00215 to .00464 percent
.00x⁻= .001 to .00215 percent

NOTE: "Geochemical conclusions" should not be drawn, for the tailings are inevitably contaminated by artifical materials introduced during mining, milling, and tailings disposal.

Hunner (1943, p. 517) estimated that not more than 10,000 tons of ore was removed from Clifton during the mining operations in the 19th century.

Hunner (1943) gave a detailed account of the M. A. Hanna Company's development of the Clifton mine. Sutter (1943) described the mining methods initially employed by the company. Regrettably, a full account of the underground development and mining has not been published, though a brief account of mining above the first level is available (Eng. Mining Jour., 1947, p. 126). Durfee (1943) described the milling and sintering of the ore.

GEOLOGIC FIELDWORK

Newland (1908), Dale (1934), and several private geologists made geologic observations at Clifton before the M. A. Hanna Co. arrived in 1940. The geologic work, magnetic surveying (see pl. 3), and exploratory diamond drilling done by the company were directed by A. E. Walker, who has published a brief, valuable account of the main features of the ore deposit (Walker, 1943). The magnetic work and drilling were extensive, but geologic work by the company was limited mainly to the spotting of drill holes and routine logging of drill core. The mine never employed a full-time resident geologist; for that reason and others, geologic work subsequent to the initial exploration was necessarily fragmentary. The results of this work were generously made available to the Geological Survey.

Until 1948, the Survey's work was limited to mapping the area on a scale of 1:62,500, checking scattered magnetic anomalies, and logging enough representative drill holes to give a general picture of the ore body and its environs. From 1948 to 1950, the Survey spent about 16 weeks in detailed mapping and study of the deposit. The geology and topography of the main Clifton area were mapped on a scale of 1:480, peripheral areas were mapped on a scale of 1:1,200, and the results were compiled for publication on a scale of

1:1,200. (See pl. 4.) The base map, prepared by planetable and telescopic alidade, was controlled by two transit traverses previously made by the company. The grid system in use by the company in 1950 was adopted. In this report, all directions are nevertheless referred to true north, unless otherwise indicated.

Because of the local magnetic attraction, all observations of foliation and lineation in surface exposures were made by sun compass. Time did not permit the mapping of joint systems by this method. The task of making the structure observations was eventually speeded up as follows: the geologist outlined structural features on the rock with yellow keel; his assistant later set up the sun compass, read the bearings, and marked them in yellow keel on the outcrop; the geologist subsequently plotted this structural information when he mapped each outcrop.

Underground, the Survey mapped on a scale of 1:480 the first and second levels (pl. 5); intervening sublevels (pl. 5), raises, and stopes (pl. 6); and several small adits. Stopes above the first level were mostly inaccessible; they were not mapped. Because the geologic work was intermittent, it was not possible to get into all the workings below the first level. Base maps kindly made available by the company were, in the final stages of the Survey's work, excellent. The direction of structural features was determined on the map sheets by visually orienting the plan of the workings, lining up a hammer handle parallel to the structural feature, and plotting the relative position of the feature as carefully as possible. The errors resulting from this indirect method apparently did not exceed 10° and were usually less than 5°—well within the limits of error of the geologist's estimate of the orientation of structural features exposed on dirty walls. Incidentally, the means for washing down walls was seldom at hand. In naturally wet places, the walls were quite clean. Elsewhere, they were exceedingly dirty. Everywhere, structure within the dark skarns and light marbles was difficult to map. Structure in ore (except the massive type) was moderately easy to map, and structure in the wallrocks of the main skarn zone was plain indeed. (Cf. figs. 8, 9, 15–17.)

The Survey's surface and underground geologic mapping was supplemented by the logging of 30,000 feet of diamond-drill core. This represented perhaps two-thirds of all the footage drilled at Clifton. The company had drilled cross sections at intervals of 50 to 100 feet along the main ore body, later augmenting their information by drilling a number of holes from the underground workings. In addition, a few test holes were put down on scattered small magnetite deposits in the neighborhood. The cores gave excellent data on the hanging wall and ore body, as well as rather meager information on the footwall. (See pl. 7.) Cores from certain critical holes were missing at the time of this survey, and a few holes had to be passed by because the days were too short.

During several weeks in 1949, E. F. Dressner[26] worked independently at the mine.

MAGNETIC ANOMALIES

The magnetic map (pl. 3) was compiled by the Geological Survey from several different dip-needle surveys made by A. E. Walker and his associates, of the M. A. Hanna Co. The dip-needle surveys were based on the company's original true-north grid. They have been transferred to the grid later adopted as the company's standard. As tie points were sparse, the transferred compilation (pl. 3) is less accurate than the original maps. Moreover, different dip needles were used for some of the surveys, and the data were not everywhere referred to the same null value. However, the main features shown on plate 3 are thought to be correct.

Bear in mind that the dip-needle surveys were made chiefly in 1940 and 1941, before the recent mining operation began. Magnetic features shown on plate 3 would thus have to be interpreted in accordance with the topography and ore distribution existing before the recent mining operation. The map is valuable principally for comparison with magnetic maps of other, unexploited magnetite deposits of skarn type found in the district.

A magnetic map is not available for the magnetite-bearing zone tested by the cut at 75S, 730W.

The Clifton magnetite deposit was registered as a peak of 6160 gammas on the aeromagnetic survey of 1945 (Balsley, Buddington, and others, 1954b).

MAIN MAGNETIC ANOMALY

The main magnetic anomaly is roughly 2,000 feet long and 150 to 300 feet wide within the +20° isoclinal, extending from 1500W to 500E. The anomaly is 1,900 feet long and 50 to 200 feet wide effectively within the +60° isoclinal. The segment between 450W and 500E corresponds to the thick part of the main ore body.

An average axis for the anomaly forms a gentle curve, more or less circular, opening toward grid south. The tangent to this arc bears approximately N. 55° E. At 00E, the anomaly branches into two strands, the southern one corresponding to the upper magnetite-bearing zone, the northern one corresponding to the main ore zone.

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[26] Dressner, E. F., 1950, Genesis of the Clifton Mine magnetite deposit, DeGrasse, St. Lawrence County, New York: Cornell Univ. M.S. thesis (unpub.).

A strong negative anomaly flanks the main positive anomaly on the northwest, giving the profile of the whole anomaly a pronounced scarp and dip slope whose orientation corresponds to the scarp and dip slope of the main ore body.

In detail, the pattern of the magnetic anomaly is complex, owing partly to the complex bedrock structure and partly to the varied topography and relatively shallow depth of the ore body. A prominent saddle in the main magnetic ridge is present between 750W and 450W. This corresponds to the thinned part of the ore body found on the crest of the complex anticline that bounds the west edge of the thick part of the main ore body. (Cf. pl. 4.)

OTHER MAGNETIC ANOMALIES CLOSE TO MAIN AREA

Three isolated magnetic highs are found south of 1500W. Each of these has been explored by diamond drilling. Only the southernmost high (corresponding to the south half of the company's Area G) was thought to be underlain at shallow depth by a substantial thickness of medium-grade ore.

An isolated intense magnetic anomaly, trending eastward, centers at (1180N, 300E), in the "Chimney Area" or Area N of the company. The anomaly is only 100 feet long and 30 to 40 feet wide within the +20° isoclinal. This anomaly was tested by five diamond-drill holes. The results were not encouraging. Only 1 or 2 feet of ore was cut by the one hole, thought to be representative, whose core was logged by the Geological Survey.

A long linear magnetic anomaly, trending northeast, extends from 1000E to 2000E, its axis being roughly at latitude 300N. This anomaly, designated as areas L and M by the company, is at the southeast edge of the swamp northeast of the open pit. For want of adequately contoured magnetic data, the anomaly is not shown on plate 3. This anomaly, 1,000 to 1,100 feet long and about 100 feet wide within the +20° isoclinal, consists of a number of small, isolated magnetic peaks. The anomaly was tested by several drill holes, and the ore layer was found for the most part to be thin. The only significant concentration of magnetite was limited to a steeply dipping zone 300 feet long and 10 feet thick, not worthy of exploitation.

SCATTERED MINOR MAGNETIC ANOMALIES

A dozen very small magnetic anomalies were found in a mile-wide belt extending about 1.5 miles northeast and 2 miles south-southwest of the mine. Detailed dip-needle surveys of these anomalies were made by the company in 1942 upon completion of a rather thorough magnetic reconnaissance of the Clifton tract and adjoining lands of the St. Regis Paper Co. The more significant of these very minor anomalies were found (a) 0.2 mile south of Area G, (b) in Big Marsh and 0.3 mile southwest of the marsh, and (c) 0.7 mile west of Camp Flat Rock—all in the southeast rectangle of the Russell quadrangle. (For location of the topographic features, see pl. 1.)

The small anomalies were 10 to 160 feet long within the +20° isoclinal. Most of them showed one or more intense magnetic anomalies within the limits of the +20° isoclinal. Garnet-pyroxene skarn containing very small local concentrations of magnetite cropped out or was exposed in shallow test pits at several of these anomalies. Clearly, none of them warranted drilling. From the results of these dip-needle surveys and the subsequent aeromagnetic survey of 1945 (Balsley, Buddington, and others, 1954b), it seemed certain that the Clifton deposit was the only magnetite body of consequence in that belt.

GEOLOGY

GEOLOGIC SETTING

The Clifton magnetite deposit is in a narrow belt of metasedimentary rocks jammed into a northeast-trending embayment on the east flank of the Stark anticline of phacoidal hornblende granite gneiss. (See pl. 1.) The gross structure of the belt of metasedimentary rocks here is synclinal, for an anticlinal mass of younger hornblende granite flanks the metasedimentary rocks on the southeast.

The metasedimentary rocks contain one major sheet of phacoidal granite gneiss and innumerable thin sheets of that rock. This complex is intruded locally by dikes and sheets of hypersthene metadiabase. Small sheets of alaskite and alaskite gneiss of the younger granite series are found in and near the metasedimentary rocks. Undeformed granite pegmatite dikes and sheets are sparingly present within the belt of metasedimentary rocks.

The dominant structural feature of the neighborhood is the Stark anticline. At the latitude of the Clifton mine, the strike of the axial plane of the Stark anticline changes from N. to N. 30° E., the dip remaining 35° to 40° E. The trend of the fold axis remains constant, at about S. 10° W., but the plunge ranges from about 5° S. to 20° S. Within a mile or so southwest of Clifton, the axial plane of the Stark anticline becomes vertical, the axis continuing to plunge southward at 5° to 10°, locally flattening to horizontal. Thus there is a profound change in the attitude of the axial plane of the Stark anticline at or near the latitude of the Clifton mine. The trend of the major anticlinal axis shifts only slightly, and the plunge is slightly undulatory, showing axial culminations and depressions. (See Prof. Paper 376, p. 117 and fig. 14, supplemented by data on pl. 1.)

With regard to the major lineation patterns of the district, the Clifton deposit is in a transitional zone (or virtually on the boundary) between areas of concordant lineations and areas of discordant lineations. (See Prof. Paper 376, fig. 15.)

The embayment of metasedimentary rocks at Clifton is at a local irregularity in the strike of the east flank of the Stark anticline. Here the flank strikes N. 58° E. This direction is roughly the same as the dominant trend (N. 65° E.) of the whole Stark anticline across the northern third of the Stark quadrangle. (See pl. 1.) The local change in the strike of the anticline's flank at the latitude of Clifton presumably served to concentrate later deformation there, giving rise to the b_E fold system which subsequently became one of the major structural controls of the Clifton ore body.

In summary, the salient features of the geologic setting are: presence of favorable host rocks in the Grenville series; proximity to members of the granite and granite gneiss series; and favorable structural environment. At Clifton, this structural environment is determined by: the presence of a relatively rigid anticlinal mass of phacoidal hornblende granite gneiss, a marked change in the attitude of the anticline's axial plane, and the existence of a local irregularity in the anticline's east flank. Precisely the same combination of geologic features is not found anywhere else along either flank of the Stark anticline, though a very close parallel is afforded by geologic features in the neighborhood of the Sterling Pond anomaly. Presumably the unique geologic environment at Clifton accounts for some of the unusual features of the Clifton ore body.

ROCKS

The principal rock types found at Clifton are skarn, quartz-feldspar granulites and related gneisses, phacoidal hornblende granite gneiss and related rocks of the quartz syenite gneiss series, metadiabase (mainly hypersthene metadiabase), and rocks of the granite and granite gneiss series.

The following descriptions of rock types are not uniform in scope. The object of these descriptions is: (a) to give enough data on each rock type to permit its identification in the field; (b) to present pertinent data supplementing the general descriptions of certain rock types; and (c) to cite local features of special interest.

QUARTZ-FELDSPAR GRANULITES AND RELATED GNEISSES

Megascopic features

Quartz-feldspar granulites are major rock units in the hanging wall at Clifton. Rocks similar in composition, but not necessarily equivalent stratigraphically, are present locally in the footwall of the main skarn zone. The granulites are gray, greenish, white, buff, or pink. Mafic minerals are sparse, typically amounting to about 5 percent of the rock but ranging from 1 percent to as much as 15 percent. Pyroxene, hornblende, and biotite are the varietal mafic minerals. In some places, a 1 percent or so of magnetite is the only dark mineral present. The dark minerals may occur singly or in combination. The granulites are fine grained, granulose, and megascopically uniform, weathering to a sugary surface that is nevertheless coherent. Their planar structure is faint to distinct, locally resembling bedding. In a few places, slight variations in grain size, accompanied by changes in the proportion or kind of mafic minerals, yield features strongly suggestive of graded bedding. Such features are unfortunately rare. Much of the quartz-feldspar granulite, regardless of the proportions of quartz and feldspar it contains, looks in hand specimen like a clean or slightly dirty metaquartzite.

In part, the quartz-feldspar granulites are homogeneous throughout a thickness of tens of feet, or they show only a slight color change, a slight change in grain size, a change in the varietal mafic mineral, or a change in the proportions of quartz and feldspar. In part, however, the granulites are interlayered with other metasedimentary rocks—commonly with skarn or feldspathic skarn, but locally with biotite-quartz-plagioclase gneiss, pyroxene gneiss, or hornblende gneiss. Interlayering with skarn is well shown on the hanging-wall bench at the open pit. Here, quartz-feldspar granulite layers several feet thick alternate with skarn layers a few inches or (rarely) a few feet thick. Some of these quartz-feldspar granulite layers have local pyroxene-rich films, and a little granitic material has been introduced in places. The interlayering of quartz-feldspar granulites with various metasedimentary rocks may be partly the result of primary sedimentation. Some of the variation is almost certainly the result of intense local deformation.

The relations between quartz-feldspar granulites and the several types of granitic rock (quartz syenite gneiss series, granite and granite gneiss series) are discussed below (p. 114, 118).

Certain layers within a given sequence of quartz-feldspar granulite are distinctive; for example, a white quartz-feldspar granulite with sparse brown biotite. However, such markers are rare. For this reason, precise correlation is seldom possible.

Locally associated with the quartz-feldspar granulites are rocks of similar composition and somewhat coarser texture, the quartz-feldspar gneisses. The gneisses are medium grained, less homogeneous, and perhaps slightly more mafic; in other respects, they resemble the granulites. Presumably the gneisses are mainly metasedimentary, closely related to the granulites. It is tempting to view them as an intermediate product between the metasedimentary granulites and the partly metasomatic microcline granite gneiss; however, the field evidence scarcely supports such a view. There is not, for example, a transitional zone of the gneisses between granulites and microcline granite gneiss, nor is there any suggestion that the quartz-feldspar gneiss in a given part of the sequence changes to microcline granite gneiss along strike or downdip.

In addition to the quartz-feldspar granulites and gneisses just described, biotite-quartz-plagioclase gneiss and several varieties of pyroxene gneiss are present at Clifton. Such metasedimentary rocks are widely distributed but not abundant. As they are common in the district, they require no separate description here.

Some pyroxene gneiss is commonly present at or near the footwall of the main skarn zone. This rock ranges from a normal, fine-grained, uniformly foliated or laminated pyroxene-quartz-plagioclase gneiss to a medium-grained (locally coarse), poorly foliated or blotchy pyroxene-microcline gneiss. Quartz is present sporadically. It has not been possible to study in detail the relation between the two principal varieties of pyroxene gneiss. Some of the blotchy pyroxene-microcline gneiss represents skarn partly disintegrated by pegmatite, alaskite, and the syenitic facies of phacoidal granite gneiss. Part of it, however, appears to be coarsened, feldspathized pyroxene-quartz-plagioclase gneiss.

Microscopic features

The quartz-feldspar granulites are fine grained or very fine grained, generally equigranular, and homogeneous. Grain outlines are almost always polygonal. Foliation may be absent, weak, or pronounced. When present, it is indicated by slight differences in quantity of mafic minerals, by subparallel mica or amphibole, by trains of opaque minerals, by differences in grain size, by the presence of quartz leaves, or by combinations of several of these features.

The principal mafic mineral may be clinopyroxene, amphibole, or biotite. It is frequently accompanied by one or both of the others, though biotite often occurs alone. Magnetite is the sole dark mineral in some specimens. Normally, the mafic mineral constitutes only 2 to 5 percent of the rock. Quartz and microcline predominate; plagioclase is subordinate or absent. The quartz content ranges from 15 to 75 percent, averaging perhaps 35 percent but showing little tendency to cluster about any single value. Usually the microcline is sharply twinned, and no specimens having potassic feldspar completely devoid of twinning or shadows have been noted. The plagioclase averages about An_{10}; exceptionally, plagioclase as calcic as An_{25} is observed. Commonly it forms grains somewhat larger than either quartz or microcline, rarely it is slightly myrmekitic, and locally it seems to replace microcline. It is poorly twinned, often cloudy, and occasionally faintly zoned, having nearly pure albitic borders. Rounded, interstitial calcite is an important light mineral in some specimens.

Typical accessories are sphene, magnetite, pyrite, and pyrrhotite. A little apatite and zircon are fairly common. Fluorite is the dominant accessory in one specimen, where it forms polygonal grains similar in size and shape to those of the quartz and feldspar.

Alteration is slight or absent. The mafic minerals commonly show slight alteration without any accompanying change in the felsic minerals: some clouding of feldspars occurs, in places a little zoisite(?) has developed, and a few prisms of pumpellyite are present in some specimens.

The quartz-feldspar gneisses are similar to the granulites, except that the gneisses normally have coarser, irregular, granoblastic grains, instead of polygonal ones.

PHACOIDAL HORNBLENDE GRANITE GNEISS AND RELATED ROCKS

Typical phacoidal hornblende granite gneiss contains phacoids of fine, granulated, recrystallized feldspar set in a matrix of fine feldspar, quartz, and hornblende. The size and shape of the phacoids are variable. The color of the rock is dominantly pink, with white or greenish mottling given by aggregates of plagioclase granules. The rock is well exposed just northeast of the headframe.

A feature of some phacoidal hornblende granite gneiss is the presence of rapakivi structure—oligoclase rims or mantles about potassic feldspar phacoids. The rapakivi structure is well shown in outcrops near 1000N, 900W. The same phacoidal granite gneiss locally contains patchy areas of plagioclase, suggesting that this plagioclase may have been moved about, perhaps late in the deformational history of the rock. A few coarse quartz-feldspar areas, alined along cracks in the gneiss, suggest the penetration of hot solutions along fractures, with resultant recrystallization of material bordering the fractures. Such quartz-feldspar aggregates seem to be the closest approach to the development of a pegmatitic facies of the phacoidal granite gneiss. In contrast, small dikes of aplite and alaskite occur sporadically in the gneiss.

In places, the phacoidal hornblende granite gneiss is strongly schistose. The phacoids are flattened or even obliterated, giving a very fine grained pink-and-gray schistose granite gneiss that is weakly biotitic locally. Schistose phacoidal granite gneiss is common in drill cores. A convenient exposure of the rock is found 45 feet southeast of the shaft.

Locally, phacoidal granite gneiss is crinkled or crenulated. The rock is pink and gray, fine grained, but slightly coarser than the schistose variety described above. Abundant minor folds having an amplitude of less than an inch to several inches are present. Remnants of greatly flattened phacoids are occasionally found. In 1950, the crenulated rock, locally encountered in drilling, was well exposed in the footwall of the contorted main skarn zone along the west side of stope 2–2.

The normally hornblendic phacoidal granite gneiss develops an alaskitic facies in places. This rock is commonly a pink leaf gneiss, very slightly biotitic or entirely devoid of mafic minerals. The phacoids of feldspar are greatly flattened; quartz leaves or plates are abundant. The texture of the alaskite is fine or almost medium, and the quartz leaves are likely to be coarser than the associated feldspar. Locally, the rock is very fine grained and schistose. In a few places near wollastonite-pyroxene syenite gneiss, the alaskite is medium grained, nonphacoidal, and megascopically only moderately deformed. Leonard believes that some or all of this medium-grained alaskite may represent the alaskitic facies of the granite-granite gneiss series, rather than a facies of phacoidal hornblende granite gneiss. Alternatively, the rock might represent "phacoidal" alaskite reworked by solutions active in the contact zone between phacoidal hornblende granite gneiss and certain layers of skarn.

Against certain layers of skarn (and perhaps pyroxene gneiss as well), the phacoidal hornblende granite gneiss becomes syenitic. Two distinctive types of rock are developed. The first is essentially a normal, contaminated syenite. The second is a distinctive wollastonite-pyroxene syenite gneiss. Both rocks may have been modified, subsequent to their main development, by introduction of material from younger granite magma.

The contaminated syenite, locally granitic, is white, greenish, or rarely pink. Normally, clots or xenocrysts of pyroxene are the mafic constituent, but in a few places hornblende is present. The pyroxenic variety is medium grained, almost coarse, and indistinctly or moderately gneissic. Definite phacoidal structure is seldom apparent in hand specimens. The rock, which appears to grade into wollastonite-pyroxene syenite gneiss, does not crop out, but it is locally abundant in drill cores from the belt northwest of the shaft. In the adit northwest of the main tailings pile, contaminated syenite and wollastonite-pyroxene syenite gneiss are exposed. At their conformable contact with the main skarn zone, a zone of skarn and marble boudins is present.

The composition and origin of the wollastonite-pyroxene syenite gneiss are discussed by Buddington (Prof. Paper 376, p. 52–53). Our knowledge of this rock comes chiefly from drill cores. However, in addition to the exposure in the adit (noted above), the rock is found along the southwest side of the headframe. At one time during the survey, removal of an ore pile exposed a narrow outcrop of wollastonite-pyroxene syenite gneiss 140 feet north-northwest of the headframe. At the headframe, the wollastonite-bearing rock is distinctly phacoidal. Considerable scapolite is present. The wollastonite and scapolite, which have weathered dull, woody, and flesh colored, occur as networks in the rock, suggesting late-stage replacement or recirculation of the material. Just southeast of the headframe, rock of similar composition more nearly resembles a migmatized metasediment. The relations are consistent with those observed in drill cores, though scapolite is perhaps more abundant in the outcrops.

The intrusive character of some contaminated syenite is clearly shown on the bench near the east corner of the open pit. Here, a thin dike of faintly phacoidal, medium-grained syenite, ⅛ to 3 inches thick, cuts across layers of quartz-feldspar granulite and feldspathic skarn at an oblique angle. The dike is exposed for a length of 5 feet; the "stratigraphic" distance transected by the dike is 10 inches. Locally, foliation in the dike is essentially parallel to the dike walls, but mostly it is parallel to the foliation in the metasedimentary rocks. Elsewhere, it represents a combination of the two surfaces (dike wall and country-rock foliation). Dike and country rock dip 30° SE., but the angle between the strike of the two planes is 45°. The writer interprets the evidence to indicate that the dike cut a layered rock sequence that was subsequently deformed; the earlier planar structure (bedding?) of the country rock was accentuated, and the secondary foliation was imposed on the dike. Both dike and country rock were subsequently cut, diagonally, by a thin pyrite vein.

Though much of the phacoidal granite gneiss is devoid of inclusions, slivers of quartz-feldspar granulite, a few inches to a few feet thick and of variable continuity along strike and downdip, are present in some sheets of phacoidal granite gneiss. Rarely, included layers a few feet thick are continuous for many feet, as in the footwall phacoidal granite gneiss found just below the main skarn zone. Inclusions of quartz-feldspar granulite characteristically show little or no evidence of material introduced from the phacoidal granite gneiss. On the second level, at several places, a few elliptical feldspar augen, roughly a centimeter long, were seen in quartz-feldspar granulite at its contact with phacoidal granite gneiss. The feldspar augen are aggregates of very fine feldspar grains, suggesting that if the augen were originally porphyroblasts, they have since been crushed down and recrystallized.

Several zones on the hanging-wall side of the main skarn zone consist of alternating thin layers of phacoidal granite gneiss and quartz-feldspar granulite. Individual layers of each rock type range in thickness from an inch or less to several feet. Such mixed zones appear to be stromatolitic, rather than migmatitic. The continuity of individual layers could not be determined. Studies of drill cores indicate that the foliation in the phacoidal granite gneiss locally is at a slight angle to the contacts between the gneiss and the quartz-feldspar granulite. There is more than a suspicion that the layered zones represent country rock spalled off and disrupted by the intrusion of granitic magma, though the zones might be mainly tectonic in origin, induced by differential flowage under stress. However, where phacoidal granite gneiss and quartz-feldspar granulite can be studied underground, the relations suggest that the two rock types possessed similar competence during deformation: major structural discordance of the sort found at contacts between skarn and granulite was not found between granulite and phacoidal granite gneiss. That the layered zones of phacoidal granite gneiss and quartz-feldspar granulite represent extensive granitization of granulite seems unlikely. The rarity of porphyroblasts (subsequently deformed) in the contact zones has already been noted, the presence of true lit-par-lit migmatites has not been observed, and the evidence for intrusion of quartz-feldspar granulite by members of the quartz syenite gneiss series has been presented in detail by Buddington (Prof. Paper 376, p. 45–46, 84).

The presence of aplite and alaskite dikes in phacoidal granite gneiss has previously been noted. At 985 N, 970 W, a dike of unusual composition cuts a slightly pyroxenic zone in phacoidal granite gneiss. The dike, 4 to 6 inches wide, is pink, fine grained, and foliated. Superficially it resembles alaskite and was so identified in the field, but it consists mainly of quartz, pinkish garnet, and an unidentified fibrous mineral resembling wollastonite. Plagioclase(?)—altered to scapolite, brownish material, and the fibrous unknown—may once have been a major constituent of the dike. Pyroxene is sparse and sporadic. Sphene is an abundant accessory, and apatite is common. Some secondary carbonate is present. Mineralogically, the dike rock resembles some metasedimentary rocks of the Greenville series. However, the dike strikes N. 5° W. and dips 75° E., whereas the foliation in the phacoidal granite gneiss strikes N. 35°–55° E. and dips 55° SE. Locally, the dike is interfingered with the phacoidal granite gneiss, and the dike itself has good foliation given by alternating garnet-rich and quartz-rich laminae. Some quartz grains are seen in thin section to be highly elongate, though the fabric of the rock is dominantly equigranular-granoblastic. Foliation in the dike seems more nearly parallel to foliation in the phacoidal granite gneiss than to the dike walls. The features described above suggest that the dike may have been emplaced when the phacoidal granite gneiss was already strongly foliated, but that subsequent deformation accentuated the earlier foliation and impressed a foliation on the dike rock as well. Alternate interpretations of the "intrusive dike" as a clastic dike, or a complexly infolded remnant of Grenville rocks, are suggested by the unusual composition of the dike. It has not been possible to check these alternate interpretations in the field.

In places, the foliation of phacoidal granite gneiss is transected by short, discontinuous shear zones a few inches wide. One such shear zone is exposed 45 feet northeast of the headframe. The vertical shear zone strikes N. 30°–35° W.; foliation in the host rock strikes N. 40° E. and dips steeply southeast.

METADIABASE

The metadiabase problem is discussed rather fully because, as a rule, the ore becomes thin and of low grade wherever metadiabase is abundant. (See also p. 39 and 75.) From exposures elsewhere in the district, rock of this kind is known to be intrusive: it cuts rocks of the quartz syenite complexes, is generally strongly metamorphosed, and is cut by rocks of the granite and granite gneiss series (Prof. Paper 376, p. 59). At Clifton, the relations are less clear. The observable habit of the metadiabase is generally concordant, but its dominant character is discordant, as described below. Because the metadiabase at Clifton is deformed and completely reconstituted, possibly incorporated in places by hornblende granite gneiss, and locally injected by pegmatite, the metadiabase is thought to be definitely older than the magnetite mineralization. However, the metadiabase is conspicuously barren and appears to cut across the magnetite deposit in several places. The writer interprets the accumulated evidence to mean that the magnetite, of replacement origin, has "inherited" certain dikes of metadiabase, selectively replacing once-calcareous rocks cut by the metadiabase but failing to replace the metadiabase itself.

Distribution

Hypersthene metadiabase crops out at (700N, 670–750W). Here garnet-hypersthene metadiabase forms thin wedges in diopsidic skarn that is slightly graphitic in places. Locally, the metadiabase has a narrow selvage of amphibolite (simple hornblende-plagioclase gneiss). In the northern exposure, the metadiabase seems to transect the planar structure of the skarn at a very slight angle, but the relation is not clear enough to state with much emphasis. At the time of this survey, no other outcrops of metadiabase were found.

The results of diamond drilling show that hypersthene metadiabase occurs sporadically, at depth, at the longitude of the southwest end of the pit bottom. It continues southwestward sporadically, appearing in force at 450W. Metadiabase continues in force approximately to 850W, beyond which it occurs sporadically to 1200W. From this longitude southwestward, drilling information is less adequate. Metadiabase was definitely cut in hole 68 (1635W), but we do not know whether the rock is absent, or present sporadically, from 1200W to 1635W. Metadiabase was not observed in cores from representative holes drilled southwest of 2000W.

Much of our information on the metadiabase comes from four drilled cross sections between 450W and 650 W. Because this information necessarily carries a large element of interpretation, as well as observation, it cannot be given as much weight as it probably deserves. On these cross sections, the apparent thickness of the metadiabase ranges from a few inches to more than 100 feet. Where metadiabase is abundant along a given cross section, its probable true thickness is 20 to 40 feet. The metadiabase is mainly on the hanging-wall side of the magnetite deposit, but locally it is in the magnetite body, forms a number of "splits" in the magnetite body, or constitutes the local footwall of the body. Commonly, the metadiabase is in part concordant with the structure of the enclosing rocks, or transects these rocks at an average dip that is less than the dip of the wallrocks. Locally, however, the metadiabase dips more steeply than the enclosing rocks.

At 450W, the evidence from close-spaced drill holes and a stope (inaccessible at the time of this survey, but well known to the company staff) strongly suggests that metadiabase cuts from the hanging wall to the footwall of the magnetite deposit, leaving a thick "horse" of barren metadiabase just above the first level. The stope—initially promising—had to be abandoned because of the abrupt appearance of a "thick horse of rock," accompanied by the disappearance of a thick section of ore.

At 600W, metadiabase apparently cuts across a thickness of hanging-wall skarn exceeding 100 feet, perhaps "transects" the magnetite body, and at 700N reaches a thickness of more than 100 feet in footwall skarn. This thick mass of metadiabase must come within 20 feet of the earth's surface, but neither thickened metadiabase nor its immediate wallrocks crop out.

Underground workings accessible at the time of this survey showed little metadiabase. The southwest limits of the rock could not be established, but the northeast limits can be closely defined. Metadiabase appears on the first level at 448W; the attitude of its contact with skarn is N. 18° W., 45° NE. On the sublevel below the first level, metadiabase appears at 450W. Here metadiabase comes against ore, their contact being crudely conformable, but locally discordant, along the axis of a minor anticline. The general strike of the metadiabase body is N. 30° W.; its dip is variable but gentle. In the back of the drift, metadiabase appears to cut across the planar structure of the ore at 45°, using strike directions as criteria for discordance. Narrow "apophyses" of metadiabase are present in the ore. On the south rib of the drift, metadiabase and ore have parallel strikes, but the contact of the two units dips 15° SW. where planar structure in the ore dips 35° SW. Neither metadiabase nor ore shows any textural change at the contact. Using indirect but presumably reliable reasoning, summarized elsewhere, the writer interprets the relations observed on the sublevel to mean that the metadiabase intruded and injected skarn (or perhaps a pre-skarn marble, later silicated) that subsequently was partly or wholly replaced by magnetite. The discordant relation of skarn and metadiabase was thus inherited by the ore.

Metadiabase is not exposed on the second level, though several flat holes drilled from the level cut a narrow dike(?) of metadiabase. Above the level, along the west side of stope 2–1 (418W), a sheet of metadiabase was found adjacent to a skarn "parting" in the thin, contorted ore zone. In this folded area, the metadiabase strikes N. 35°–75° W. and dips 25°–35° SW.

Though incomplete, the data from outcrops, drilling, and underground workings indicate that the bulk of the metadiabase is effectively confined to a northwest-trending zone extending approximately from 450W to 850W. Metadiabase is present sporadically for an unknown but generally slight distance beyond the limits noted above. Where observable, the contacts between metadiabase and its host rocks are usually concordant but locally discordant. Nevertheless, the general distribution of the zone of abundant metadiabase is distinctly athwart the trend of other rock units in the area. The writer interprets these relations to mean that the main zone of metadiabase is dikelike, though the relations viewed on a smaller scale are more often concordant than discordant. The metadiabase body would thus correspond to a stunted Christmas tree, most of whose branches on two opposite sides have been lopped off. The trunk is not exposed, or not recognized; the remaining branches correspond to the sheet-like, conformable or moderately discordant bodies of metadiabase described above.

With regard to the fold systems at Clifton, the zone of metadiabase is present along the major b_S anticline that determines the southwest edge of the main ore body. Possibly the emplacement of the metadiabase was controlled by planes of weakness related to one or more of the sets of north- to northwest-trending joints and local shear zones present in the neighborhood. Prominent vertical joints of that general orientation were mapped regionally, and the northwest-striking vertical shear zones have been noted above.

Megascopic description

Typically, the metadiabase is a fine-grained dark greenish-gray rock of homogeneous composition and uniform fabric. The foliation, given by alternating flat lenticles of light and dark minerals, is regular but fuzzy. Of the dark minerals, only brown biotite, dark-green pyribole, magnetite, and a little reddish-brown garnet can be identified. In places, either garnet or biotite may be lacking, the general aspect of the rock remaining the same. Thicker bodies of metadiabase may show scattered porphyroclasts of bluish plagioclase. Gray color, fine grain, and "ordinariness" of appearance make it easy to recognize the rock—so easy, indeed, that during the Survey's initial work at the mine the rock was tentatively adopted as a promising marker to use in working out the structure of the deposit. Evidence subsequently showed that the metadiabase is intrusive, and in no sense a stratigraphic marker.

The garnet-biotite variety just described is that normally found where metadiabase is enclosed by quartz-feldspar granulite, pyroxene gneiss, pyroxene skarn, or phacoidal granite gneiss. In some places, the normal variety has a slightly coarser border zone of simple amphibolite consisting essentially of plagioclase and green hornblende. This border zone is commonly only a few inches thick, but in the thicker metadiabase bodies it reaches several feet in thickness. Metamorphic reconstitution of homogeneous metadiabase may be solely responsible for these amphibolitic border zones, or the zones may be a reconstituted marginal chill facies of the metadiabase. Chemical analyses of the border zones are not available.

Where metadiabase is in contact with medium-grained alaskite, garnet and biotite are absent from the metadiabase, the rock is slightly coarser though still relatively fine grained, and a simple amphibolite (hornblende-plagioclase gneiss) is developed. A similar effect is observable where pegmatite seams are found in the metadiabase, though here the hornblende-plagioclase gneiss is medium-grained—almost coarse. The pegmatite seams in this environment are commonly syenitic, rather than granitic, and they usually contain clots of coarse, recrystallized pyribole derived from the metadiabase.

Microscopic description

Only the common variety of metadiabase will be described in detail. This rock is essentially a pyroxene-biotite-plagioclase gneiss with a very fine, equigranular, crystalloblastic fabric. All the nonmetallic mineral grains except biotite are polygonal, with an average diameter of 0.2–0.3 mm. Many grains 0.1 mm in diameter are also present. Thin layers rich in plagioclase (andesine, An_{34-38}) alternate with flat lenticles rich in pyroxenes, garnet, biotite, or a combination of these dark minerals.

The andesine is fresh, well twinned, and in places very slightly antiperthitic. Clear orthoclase occurs as strings of polygonal grains and as scattered anhedra.

Both monoclinic and orthorhombic pyroxenes are present. The dominant pyroxene is diopsidic: (+) $2V \sim 60°$; $r>v$, weak; $Z \wedge c=37°$; pale green, generally nonpleochroic. A few coarser grains, 0.5 mm in diameter, are filled with opaque dustlike inclusions, presumably of magnetite. These inclusions are thought to indicate unmixing of the pyroxene during reconstitution and before the attainment of complete equilibrium.

Subordinate hypersthene is present. It is weakly pleochroic, with X=very pale pink and Z=pale green; $(-)2V \sim 60°$; $r>v$, weak. It is readily distinguished from the monoclinic type by its pleochroism and parallel extinction, though these properties may be almost imperceptible on very small grains.

The hornblende is a common green variety forming almost equidimensional grains, like the pyroxenes. Its optical properties are: $(-)2V$ near 60°?; $r<v$; $Z \wedge c=$ 15°–16°; strongly pleochroic, with X=pale yellow, Y=olive, Z=slightly brownish green.

Fresh biotite, pleochroic from yellow to deep reddish brown, forms small laths throughout the rock. In places, it grows athwart pyroxene; elsewhere it is scattered hit-or-miss through amoeboid garnet masses that also enclose magnetite.

Very pale pink to brownish garnet occurs as anhedral individuals, trains of polygonal grains, and small amoeboid masses. The last enclose magnetite and biotite, and occasionally plagioclase and pyroxene, though they lack typical poikiloblastic structure. The weak birefringence that characterizes this garnet is visible only conoscopically.

Magnetite of three types is present. The first is the dustlike material included in some large clinopyroxene crystals. The second has the form of minute, rounded blebs scattered through the plagioclase groundmass. The third, which has the greatest volume, consists of smooth, irregular, often elongate grains that locally replace the mafic minerals and even plagioclase. Some magnetite has partial rims of garnet; other grains are embedded in larger masses of garnet. The amoeboid garnet and elongate magnetite appear to have grown later than the other rock minerals.

Apatite as blebs and euhedral prisms is scattered through the rock, and a few minute zircons are also present. A curious feature of the metadiabase is the near absence of sphene, a common accessory in amphibolitic rocks of the district. So far, sphene has been found in just one specimen of metadiabase from the Clifton mine.

Modes of the metadiabase are given in table 24 of Professional Paper 376. The uniform bulk chemical composition of the rock is emphasized by its nearly constant total Fe content, averaging 12.8 percent throughout an apparent thickness of 80 feet in hole 69. (Chemical data from laboratory of Hanna Coal and Ore Corp.)

QUESTIONABLE METADIABASE

The possibility that metadiabase occurs northeast of the open pit deserves consideration. No rock unequivocally of that distinctive type was found in representative drill holes studied by the Survey. However, amphibolite of unknown origin was cut by those holes. The amphibolite layers range in apparent thickness from 4 feet to 14 feet. One of these amphibolites is a fine, well foliated, polygonal-crystalloblastic, locally cataclastic rock consisting mainly of hornblende, clinopyroxene, biotite, and andesine. Magnetite containing exsolved ilmenite is an abundant accessory, apatite is plentiful, and pyrite fracture fillings are common. The coarser clinopyroxene grains are crowded with inclusions of opaque dust, presumably magnetite. The finer clinopyroxene grains are mainly present as clusters in porphyroblastic and poikiloblastic hornblende. The presence of dust-filled clinopyroxenes, the fabric of the rock, and the megascopic appearance of the rock strongly suggest that it is a facies of the metadiabase. Its mineral composition alone, however, is inconclusive, for amphibolites of diverse origin may show this same assemblage. If the amphibolite were truly a facies of the metadiabase, its distribution would be notable: there would be a zone of metadiabase close to, or on the outer flanks of, each of the major b_S anticlines that determine the northeast and southwest limits of the main ore body. Such speculation is interesting but not capable of firm support.

GRANITE AND GRANITE GNEISS SERIES

Four members of the granite and granite gneiss series, or so-called younger granites, are present at Clifton. They are: hornblende granite gneiss (gneissic facies of hornblende-microperthite granite), alaskite and alaskite gneiss, microcline granite gneiss (biotitic and pyroxenic varieties), and pegmatite (normally granitic but locally syenitic). The members are listed according to decreasing probable age; that order also corresponds roughly to decreasing abundance of each rock type. Criteria definitely establishing the age relations among members of the series have not been found at Clifton. For such data, we are dependent on the information derived from regional mapping.

Hornblende granite gneiss

Hornblende granite gneiss apparently underlies much of the area southeast of the railroad siding, including the sand plain about Orebed Pond.[27] Outcrops of hornblende granite gneiss are indicated on plate 4. In addition, the holes drilled southeast of 450S entered bedrock in hornblende granite gneiss or cut a substantial thickness of that rock beneath a thin zone of modified pyroxene skarn. The rock has not been found in the area of metasedimentary rocks and phacoidal granite gneiss generally present northwest of 100S. The hornblende granite gneiss is part of a generally anticlinal mass, perhaps several hundred feet thick, thrown into folds that are rather gentle relative to most folds in the Clifton area. The gneissic facies of hornblende-microperthite granite is a local development at Clifton, resulting from intense local deformation in the area about the mine. Southeast of Orebed Pond, the first outcrops are normal, undeformed hornblende-microperthite granite.

The bulk of the hornblende granite gneiss, both in outcrops and in drill cores, is uncontaminated. The rock is pink, medium- to fine-grained, distinctly foliated, and petrographically identical with the normal hornblende granite gneiss of the district. Here and there, a little granite pegmatite is present as thin seams or narrow ribbons. Ribbons of granite pegmatite are well shown in outcrops at (400–500S, 300E). The granite pegmatite forming these ribbons is highly inequigranular but apparently only slightly deformed. In contrast, the granite gneiss host has been completely deformed and recrystallized, as indicated by the grano-

[27] This Orebed Pond was the eastern of the two formerly present (see pl. 1); the western pond was drained by an adit to make way for tailings from the Clifton mill.

blastic, foliated structure and the presence of microcline and sodic oligoclase derived from the unmixing of preexisting microperthite.

Locally, the hornblende granite gneiss is contaminated with several percent of biotite derived from biotitic metasedimentary rocks sporadically present between the hornblende granite gneiss and a thick zone of underlying skarn.

Where hornblende granite gneiss comes directly against the underlying skarn zone, the gneiss may be essentially unchanged, or it may become white, be streaked with pyroxene skarn schlieren, and contain xenocrysts of pyroxene accompanied by abundant accessory sphene. The white, contaminated border zone is generally less than 5 feet thick.

Within the hornblende granite gneiss are local zones of amphibolite schlieren. Near these schlieren, the hornblende granite gneiss is strongly foliated, greenish, and medium grained or locally pegmatitic. The hornblende normally present is supplemented by iron-rich hypersthene and a little clinopyroxene (ferroaugite?). A little sphene is also present. The two contaminated facies just described have been studied in drill cores. They do not crop out within the area shown in plate 4.

Alaskite and alaskite gneiss

Alaskite gneiss is more abundant though less widely distributed than alaskite at Clifton. All the alaskite gneiss observed to date is slightly or moderately contaminated. Such rock is well exposed east and southwest of the main building. (See pl. 4.) The contaminated alaskite gneiss appears to be in part a marginal facies of the hornblende granite gneiss, and in part a separate sheet intercalated within the metasedimentary rocks close to their contact with the hornblende granite gneiss. Two varieties of contaminated alaskite gneiss are present. The first contains 5 to 10 percent of biotite; the second, considerably less mafic, contains a little biotite, hornblende, and acessory sphene. Both types are generally pinkish, fine- to medium- grained, and slightly or distinctly foliated. Small schlieren and pods of metasedimentary material are present locally in the alaskite gneiss.

A medium-grained alaskite, possibly a facies of phacoidal granite gneiss, has already been described as a local associate of the contaminated syenitic facies of phacoidal granite gneiss.

Occasionally one finds dump specimens of white, greenish, or pink alaskite—medium grained, apparently undeformed, and uncontaminated or only slightly so. Alaskite of this type was brought up at the time the second level was developed, but subsequent geologic mapping of the workings failed to disclose the source of the rock.

Thin layers of alaskite, half a foot to several feet thick, were found close to the margins of the main skarn zone in cores from a number of drill holes.

Microcline granite gneiss

Microcline granite gneiss does not crop out in the Clifton area. However, the rock has been mapped at the east end of the second level (see pl. 5) and found in cores from drill holes 45–4, 17, and 46–4. These sporadic occurrences appear to be part of a single zone, not necessarily continuous, in footwall rocks beneath the main skarn zone. Where measurable, the apparent thickness of the microcline granite gneiss zone ranges from 20 to 90 feet; its true thickness cannot be determined with certainty. The microcline granite gneiss is associated with quartz-feldspar granulites or occurs at or near the "stratigraphic" position normally occupied by the granulites. Locally, bodies of microcline granite gneiss, 5 to 20 feet thick and less than 100 feet long, are associated with biotite-quartz-feldspar granulites close to the hanging wall of the main skarn zone.

Microcline granite gneiss elsewhere in the district has been interpreted in part as a metasomatic rock mainly derived from metasedimentary gneisses and amphibolites, rather than from quartz-feldspar granulites. Tentatively, the microcline granite gneiss at Clifton is thought to represent migmatized and metasomatized quartz-feldspar granulite. At present, the data for establishing its origin are inadequate and inconclusive.

The microcline granite gneiss at Clifton is megascopically more homogeneous than rock of similar composition occurring elsewhere in the district. Such homogeneity would be expectable if the microcline granite gneiss at Clifton had indeed been derived from the rather homogeneous quartz-feldspar granulites.

Megascopically, the microcline granite gneiss is pink or pinkish gray, fine grained though slightly coarser than the typical quartz-feldspar granulites, and poor in mafic minerals. Locally, the gneiss contains thin and rather widely spaced seams of granite pegmatite; aside from this feature, the rock is very weakly foliated. Microscopically, the rock is a fine, nearly equigranular, granoblastic aggregate of quartz, microcline, a very little plagioclase, hornblende, biotite, and the accessories magnetite, apatite, zircon, and (occasionally) sphene. Pyroxene is present locally. The quartz content ranges from perhaps 5 percent to about 25 percent. The proportions of hornblende and biotite vary, but the mafic minerals seldom exceed 5 or 10 percent. Commonly the mafic minerals are partly or wholly chloritized.

Pegmatite

Granitic and syenitic pegmatite is fairly widespread but quantitatively unimportant. One small body is exposed in the open pit near the collar of the ventilation raise in the pit floor. This quartz-perthite pegmatite, containing many small euhedral green zircons and several irregular masses of schorl, is locally replaced by magnetite. Allanite is present in quartz-rich parts of the pegmatite. The pegmatite has been sheared and crackled subsequent to the introduction of magnetite. Several drill holes cut bodies of pegmatite 10 to 20 feet thick and less than 100 feet long.

Aside from these occurrences, pegmatite at Clifton is confined to lenses a few inches to a few feet thick, of undermined but short dimensions along strike and downdip, distributed at random through the various rock units. Much of the pegmatite, especially in the thinner lenses, is contaminated with pyroxene—less commonly, hornblende—derived from the country rock and commonly recrystallized into coarse single crystals or aggregates of grains. Such pegmatite is normally quartz free or quartz poor, undeformed, and bleached.

In a number of places, thin zones of pyroxene gneiss close to the margins of the main skarn zone have been migmatized or merely coarsened by pegmatite. Elsewhere, similar zones of coarsened pyroxene gneiss, commonly quartz poor, are found without pegmatite. How much of this latter pyroxene gneiss has been modified by pegmatite, we cannot say, for in some places phacoidal granite gneiss and alaskite have apparently induced comparable changes in the pyroxene gneiss.

Migmatites of pegmatite and country rock (chiefly biotite gneiss, pyroxene gneiss, and pyroxene skarn) are present sporadically, but they are sparse.

Ribbons of granite pegmatite are a prominent feature of hornblende granite gneiss at (400–500S, 300E). These linear pegmatites—a few inches wide, less than an inch thick, and of undetermined length downdip—in cross section resemble old-fashioned metal spectacle cases. The ribbons are dominantly hornblende-quartz-microperthite pegmatite, slightly biotitic, with considerable accessory zircon. Though the pegmatite ribbons are highly inequigranular, ranging in texture from fine to coarse, they do not seem to have been deformed. In contrast, their host is a completely recrystallized hornblende-microcline-oligoclase granite gneiss.

STRUCTURE

FOLDS

The complex structure of the Clifton ore body has already been summarized (p. 73–75). That summary is based on the geologic maps (pls. 4, 5), supplemented by data from drill cores (cf. pl. 7) and mapped sections in the accessible stopes (pl. 6). Plots of the orientation of linear structures are shown in figures 10–13. Examples of the minor structures used in interpreting the gross structure of the ore bodies are shown in figures 14–17. The structure of the main ore body is partly exposed in the pit (figs. 8, 9), and a schematic presentation of the ore bodies is shown in figure 18.

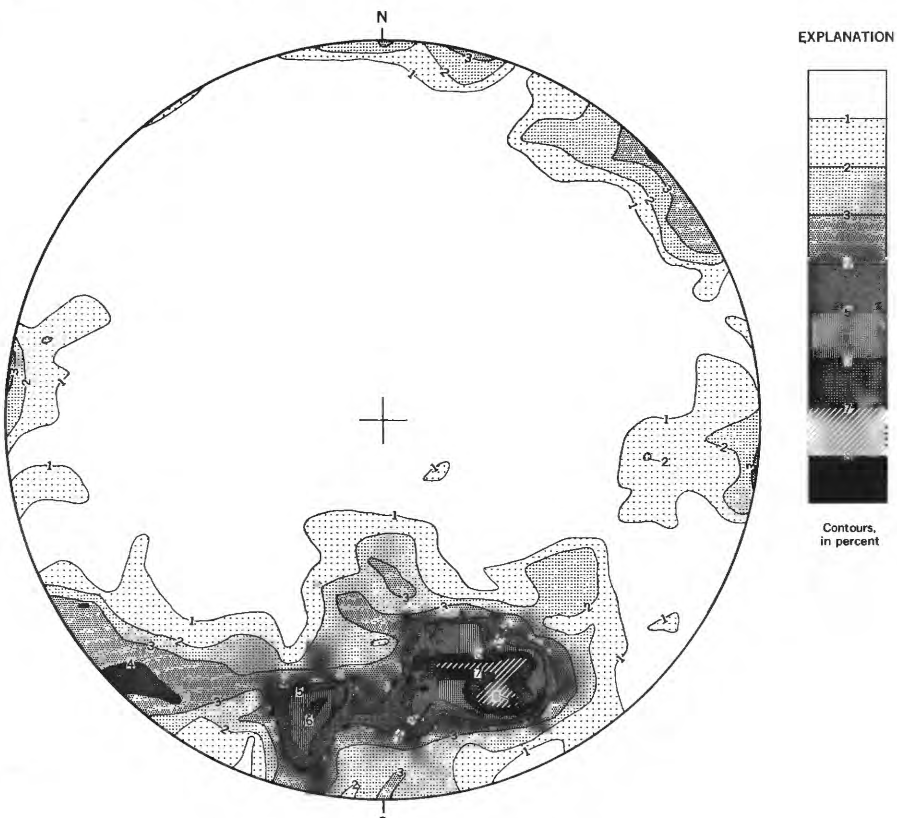

FIGURE 10.—Contoured equal-area projection of all lineations, Clifton mine. 247 observations plotted on lower hemisphere.

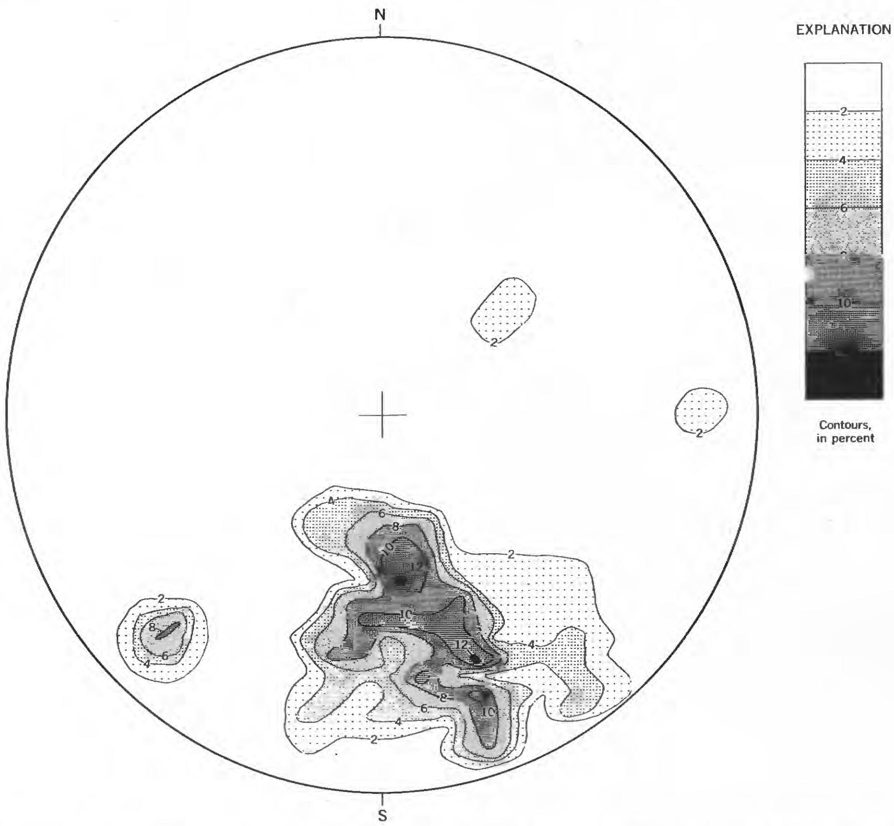

FIGURE 11.—Contoured equal-area projection of axes of tight folds and crumples, Clifton mine. 37 observations plotted on lower hemisphere.

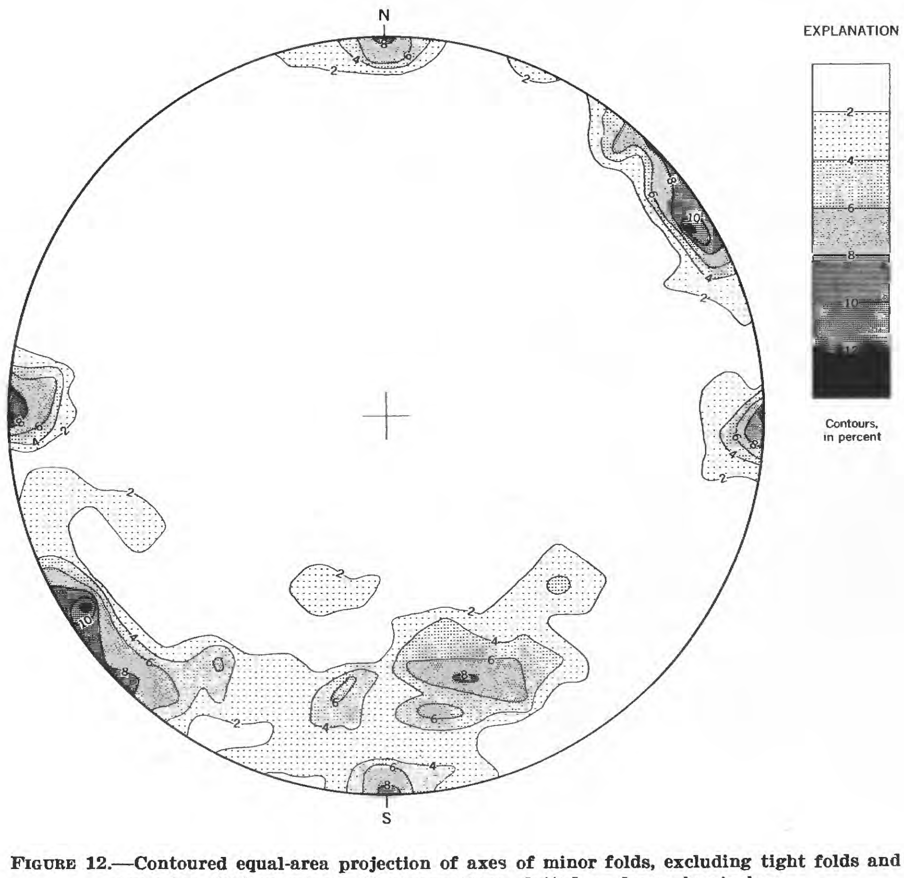

FIGURE 12.—Contoured equal-area projection of axes of minor folds, excluding tight folds and crumples, Clifton mine. 86 observations plotted on lower hemisphere.

Figure 13.—Contoured equal-area projection of lineations, excluding fold axes, Clifton mine. 123 observations plotted on lower hemisphere.

Figure 14.—Contorted zone of pyroxene skarn and syenite underlain by uncontorted quartz-feldspar granulite. Along contact diagonally upward from hammer, the border of the quartz-feldspar granulite has been folded into the contorted zone. Skarn, heavily stippled; syenite, unstippled; quartz-feldspar granulite, lightly stippled. Exposure is on bench at southeast side of open pit, Clifton mine.

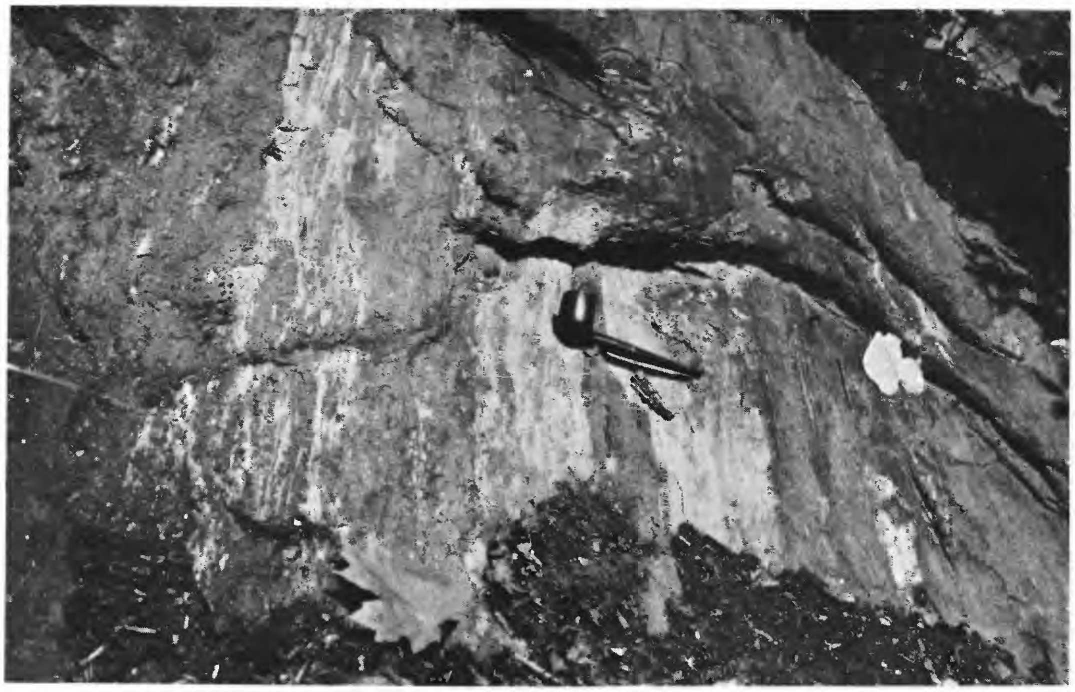

FIGURE 15.—Lineation in pyroxene skarn. Rods of skarn (dark), streaks in scapolitic skarn (light), and axes of gentle warps trend S. 30° E. and plunge 36° SE. (toward observer) on foliation striking N. 20° E. (roughly parallel to pipe stem) and dipping 45° SE. Clifton mine, coordinates 410N, 615E.

FAULTS

Three holes cut chloritized, slickensided, granulated, calcite-veined material interpreted as fault breccia. The fault or faults indicated by this breccia seem to be minor, though one cannot discount the possibility that one or more major faults may be present. The possibilities are discussed below, for major faults, if they existed here, might displace or eliminate ground otherwise favorable for the occurrence of a mineralized skarn zone.

The position of the fault breccia is as follows: hole 42, at 144 to 148 feet; hole 46–8, at 485 to 490 feet; hole 47–1, at 784 to 786 feet. The angle between the brecciated zone and the core axis was 15° to 30° in hole 46–8; elsewhere it was indeterminate. Let us assume that the three breccia zones are on the same fault. The three "points" determine a plane that strikes N. 46° W. (true) and dips 32° SW. The dip of this plane agrees moderately well with the gentler set of possible dips inferred from the core-breccia angle measured in hole 46–8. The plane, projected to the surface, passes through several outcrops that show no fault breccia. Fault breccia was not found in cores from six critical drill holes northwest of hole 46–8. Therefore, if a fault of such attitude exists, it must die out, or change its attitude radically, less than 400 feet northwest of hole 46–8.

Assume that the breccia in holes 46–8 and 47–1 determines a fault, and that the measured core-breccia angle indicates a steep dip. The fault must strike northward. No fault breccia was found in cores from holes 47–2 and 46–3. Therefore, the existence of a steep fault seems very doubtful. The absence of fault breccia in those cores also eliminates consideration of a fault of moderate dip, either northwest or southeast. Faults determined by the breccia zone of hole 42, paired with corresponding zones in either of the other holes, are likewise eliminated by absence of breccia in these and other critical holes.

Let us consider the breccia zones singly. Almost the only possible direction for a major fault passing through the zone in hole 47–1 is N. 60°–75° W. Geologic data are too sparse to eliminate or confirm this possibility.

Almost the only possible direction for a major fault through the breccia zone in hole 46–8 is N. 37° W. However, the lack of evidence for such a fault in the mapped mine workings makes the existence of the fault very doubtful.

A steep northeast-striking fault passing through the breccia zone of hole 42 is a possibility, but the lack of fault breccia in hole 47–2 suggests that such a fault could not continue much farther southwest than 1500W. Data are unavailable to check the possible continuation of the inferred fault northeastward from hole 42.

FIGURE 16.—Lineation in pyroxene-feldspar granulite. Streaks, twisted clots, and tight crumples of garnet-pyroxene skarn (dark) trend S. 10° E. and plunge 25° S. toward observer. Clifton mine, coordinates 395N, 560E.

FIGURE 17.—Lathlike body of contorted and slightly warped pyroxene skarn (dark) on dip surface of pyroxene-feldspar granulite. Foliation strikes N. 20° E. (roughly parallel to pipe stem) and dips 40° SE. (toward observer). Lineations—given by skarn lath, skarn streaks, and axes of crenulations and gentle warps—trend S. 30° E. and plunge 32° SE. Clifton mine, coordinates 425N, 635E.

A steep west-northwest-striking fault passing through the breccia zone of hole 42 would perhaps explain the apparent offset of rock units in the area centering at (100S, 400W). A northwest limit for such a fault is roughly 00N, for drill holes begun at 50–100N cut no fault breccia at the appropriate elevation. An alternative explanation for the apparent offset is, of course, the interaction of folds of the b_S and b_E systems.

FERN ANOMALY

The Fern anomaly is 1.1 miles east-southeast of Deerlick Rapids and 0.9 mile N. 60° W. of the Outafit cabin, south-central rectangle, Stark quadrangle. The south end of the Deerlick Rapids magnetite deposit is 3,500 feet west of the Fern anomaly. The anomaly is at the head of a small, unforested, fern-covered valley. The rest of the area is forested.

The anomaly was registered on the aeromagnetic survey of 1945 as a peak with a value of 6,260 gammas (Balsley, Buddington, and others, 1954b). The dip-needle anomaly is torpedo shaped, trending about N. 30° W. and having its blunt end on the north-northwest. The anomaly is about 550 feet long within the +20° isoclinal and has a maximum width of about 300 feet near the north end. The anomaly is about 350 feet long within the +30° isoclinal, 200 feet long within the +40° isoclinal, and 125 feet long within the +50° isoclinal.

A small outcrop of migmatitic pyroxene gneiss and pyroxene skarn is present at the north-northwest end of the anomaly. Alaskite crops out on the hill northeast of the anomaly.

The magnetite body that produced the anomaly is presumably small. It might be of skarn type (as the Outafit), or of granite gneiss type (as the Deerlick Rapids). The first seems more likely.

GRASS POND ANOMALY

The Grass Pond anomaly was discovered by the aeromagnetic survey of 1945, on which the anomaly was registered as a peak of 5,260 gammas (Balsley, Buddington, and others, 1954a). The following account is

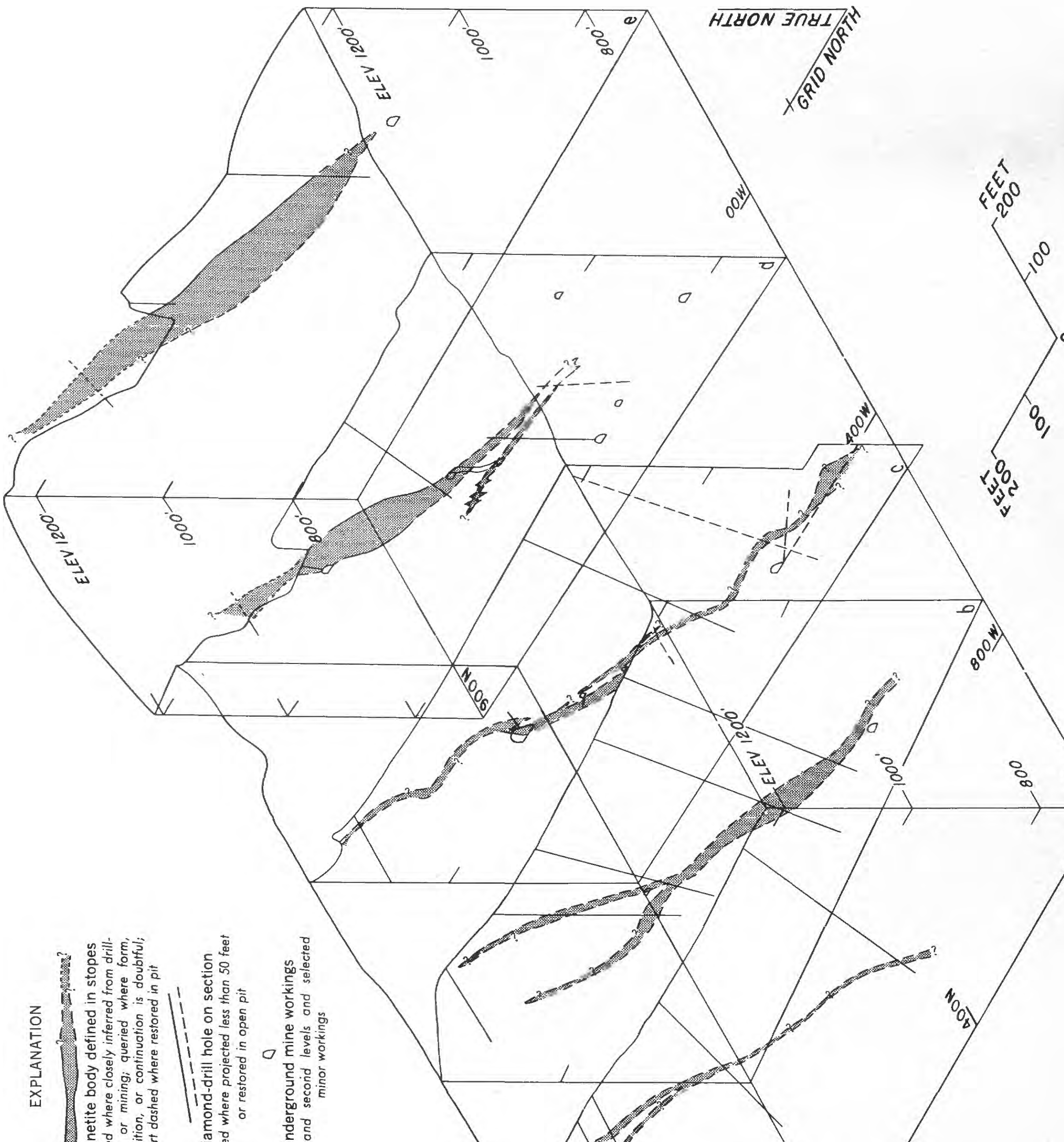

FIGURE 18.—Skeleton block diagram of the Clifton ore body. For location of cross sections, see figure 4.

slightly modified from that given by Hawkes (Hawkes and Balsley, 1946, p. 5).

The Grass Pond anomaly lies along the low ridge a few hundred feet south of the creek that flows into the north end of Grass Pond from the west, east-central rectangle, Cranberry Lake quadrangle. (See pl. 1.)

The anomaly is best reached from Wanakena over State Conservation Department trails leading from Wanakena to Cowhorn Pond. Beyond Cowhorn Pond, it is necessary to follow a compass course eastward for about 2 miles to the anomaly.

The axis of the anomaly trends about N. 65° W. (See fig. 19.) The overburden is generally very shallow; therefore, dip-needle readings of the intensity shown on figure 19 are not so significant as they would be where bedrock is covered by an appreciable thickness of overburden. The anomaly apparently follows a magnetic "contact" of unknown significance between neutral to slightly negative values on the northeast and moderately positive values on the southwest. The terrain between the Grass Pond anomaly and Cowhorn Pond was scouted for other anomalies along profile line spaced at 500-foot intervals, but no anomalous values greater than +10° were found.

Geologically, the Grass Pond anomaly lies close to the inferred Wolf Mountain fault (Prof. Paper 376, p. 133), presumably on the north side of the fault in or close to an area of alaskite contaminated with amphibolite. (See pl. 1.) The belt of Grenville rocks along Sixmile Creek is less than 500 feet west of the anomaly. Skarn is abundant in that belt. Contaminated granite and migmatitic amphibolite crop out close to the anomaly (see fig. 19), but the magnetic source of the anomaly is not exposed. Possibly the source is a weakly or moderately mineralized layer of skarn included in the contaminated granite. Alternatively, the anomaly might be due merely to contact effects between amphibolite and contaminated granite.

The Grass Pond anomaly is on State land within the Adirondack Forest Preserve.

GREEN FARM DEPOSIT

SUMMARY

The Green Farm deposit, of skarn type, has been known for at least 75 years. The deposit has been investigated many times, drilled twice, and partly explored by two shallow shafts and several test pits. A small quantity of ore—certainly less than 1,700 to 1,800 tons — has been shipped from the property. The known magnetite body is small but generally high in grade. Some lump ore is present. An intense magnetic anomaly is confined to the line of pits. The maximum strike length of the magnetite body is probably 500 feet; the thickness of ore now or formerly exposed ranges from 3 to 12 feet. The ore appears to be confined to a single zone that may or may not be continuous. Small folds having an amplitude of 20 to 30 feet and plunging generally southwestward are present in this zone; ore seems to be localized along the plunging folds. The small size of the deposit will probably discourage further prospecting.

LOCATION, OWNERSHIP, AND PRODUCTION

A small magnetite deposit is exposed in openings on the Everet Green farm, Town of Fine, 1.75 miles by air line south-southeast of the hamlet of Fine and 0.3 mile north of School No. 15, north-central rectangle, Oswegatchie quadrangle.

To reach the deposit, turn off N.Y. Highway 3 at its junction with N.Y. Highway 58, just east of Fine. Take the first left turn south of the bridge and drive southward. The road distance to the Green farm from the junction of routes 3 and 58 is 1.7 miles.

The openings are in a pasture 200 feet east of the farmhouse, which is on the east side of the graveled road that runs from Fine to the hamlet of Briggs. Newland (1908, p. 138) refers to the openings as the "mines on Vrooman ridge, Fine." "Vrooman Ridge" is still the local name for the broad shelf on which School No. 15 sits, though the Vrooman Ridge of the topographic map lies 4 miles south-southeast. An old report by Newberry[28] (1890) refers to the deposit as the "Fine ore bed."

Newland (1908, p. 138) thought no active mining had taken place here, but according to Mr. S. C. Green (1949, oral communication)[29] ore was drawn to Jayville more than 50 years ago by way of a winter road running toward Pitcairn; a smaller quantity of ore was drawn to Briggs and shipped down the railroad. A pile of ore, presumably from the Green Farm deposit, is still standing by the siding at Briggs; one may infer that the ore was hauled between 1886 and 1888, when the Jayville mines were being operated. The total quantity of ore extracted from the Green Farm deposit cannot have been more than 1,700 to 1,800 tons.

DISCOVERY, EXPLORATION, AND DEVELOPMENT

The discovery date and early history of the deposit are unknown. According to Mr. S. C. Green, the deposit was drilled twice—once about 1875–80, and again about 1885–90. Ore was shipped during the second period of exploration. At least part of the exploratory work was done by the Magnetic Iron Ore Company, then

[28] Newberry, S. B., 1890, Fine ore bed: Private mining report lent by J. G. Broughton, State Geologist of New York.

[29] Mr. S. C. Green, father of Everet Green, was a keen and well-informed gentleman of some 85 years in 1949.

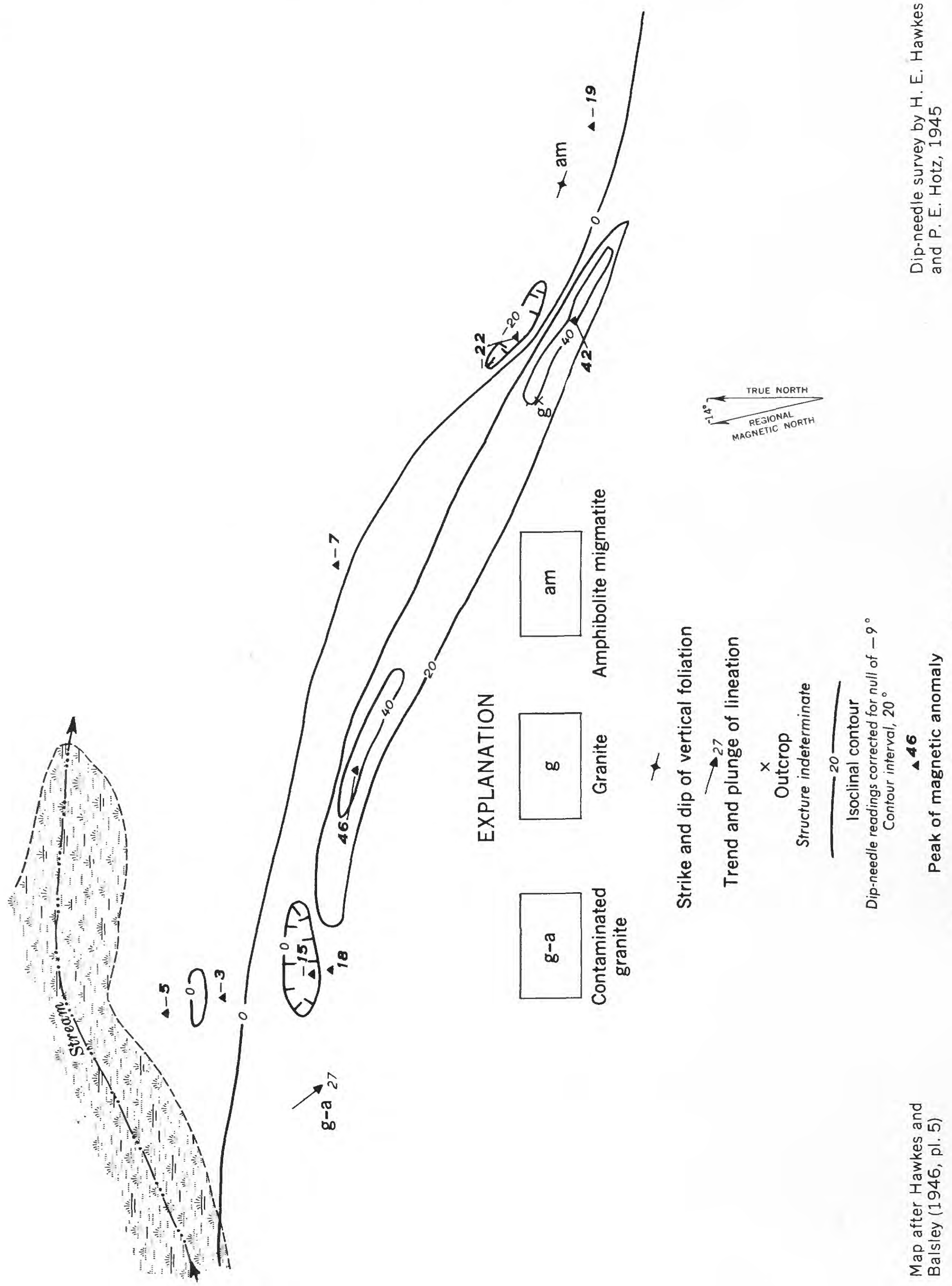

FIGURE 19.—Reconnaissance magnetic map of the Grass Pond anomaly, east-central rectangle, Cranberry Lake quadrangle.

operators of the Jayville mines. George D. Grannis[30] stated that he carried on drilling and sinking operations between 1884 and 1889. Grannis drilled more than four holes, but his description of their location, inclination, depth, and ore intersections is very sketchy. The approximate sites of four of Grannis' drill holes are shown on figure 20. These locations are based solely on Grannis' descriptions.

The workings include 6 pits. (See fig. 20.) Pit 1 is caved and filled with trash. Pit 2 is a very shallow depression. Pits 3 and 4 are now about 6 feet deep; according to Newberry (op. cit., p. 4) these pits were said to be about 12 feet deep. Pit 5 is 6 feet deep. This pit contains a water-filled shaft—commonly called the Main shaft or shaft No. 1—said to be 34 or 35 feet deep. According to Newberry (op. cit., p. 2), drifts at the foot of the shaft extend 10 feet north and south along the vein. Pit 6 is the old Upper shaft, or shaft No. 2. According to Newberry, this shaft, now water filled, is 24 feet deep. The inclination of the two shafts could not be determined in 1949, but we may assume that the shafts follow the vein.

Since the early exploration and development, the pits and shafts have been unwatered several times, though not for the last 35 years (Green, oral communication, 1949). The property has been examined repeatedly by mining engineers and geologists. The area has been surveyed by dip needle at least four times, but none of the available magnetic maps can be properly tied to the workings. The writer's examination of the property was confined to 1 day in October 1945, and 1½ days in July 1949. The accessibility, good exposures, and high grade of ore have made the Green Farm deposit one of the "most visited" prospects in the district.

GEOLOGY

The Green Farm deposit lies in a belt of metasedimentary rocks, 0.1–0.4 mile wide, known as the School No. 15 syncline (Prof. Paper 376, p. 115). This belt, running south along the road to Briggs, is on the west flank of an overturned anticline (Colton Hill anticline) of alaskite gneiss. The alaskite gneiss crops out 60–100 feet east of the Green Farm pits. (See fig. 20.) The belt of metasedimentary rocks is overlain on the west by hornblende granite gneiss. (See pl. 1.) Prominent lineations in rocks north and south of the Green Farm trend N. 80°–85° W. and plunge 35°–45° W. These lineations are roughly parallel to the constructed axis of the major overturned anticline (Colton Hill anticline). The constructed axis has an average orientation of S. 80° W., 42° W., though individual determinations range from S. 60° W., 42° SW., to N. 80° W., 42° W. This axial direction is not readily apparent from inspection of the geologic map (pl. 1), for the dominant feature of the anticline—the axial trace, or trace of the axial plane—trends N. 25° W.

Below, the writer develops the interpretation that the magnetite-bearing zone at the Green Farm follows a series of small, southwest-plunging folds whose amplitude is roughly 20 to 30 feet. These small folds are not simple drag folds directly referable to the Colton Hill anticline. Rather, the orientation of the small folds suggests that they belong mainly to a different stage of deformation than that which produced the Colton Hill anticline. However, we cannot rule out the possibility that both fold systems are represented in the Green Farm deposit.

The reason for emphasizing the regional background is this: dominant lineations on the west flank of the Colton Hill anticline plunge westward, generally down dip of the planar structure of the rocks. Indiscriminate application of this pattern might suggest that the Green Farm deposit continues for a considerable distance down the dip. Yet the local evidence, indicating a separate fold system, suggests that the shape of the deposit may be controlled by structures plunging south-southwest. The whole story is not known; perhaps both systems exercise control, as at Clifton (p. 73–75).

ROCKS

The ore zone is in pyroxene skarn and partly silicated marble. The partial thickness of these host rocks is known to be 9 to 12 feet; the maximum thickness is probably 20 feet, though it might be as much as 50 feet locally. The hanging wall of the skarn zone is pyroxene-microcline gneiss, in places carrying 20 to 50 percent quartz. The gneiss is yellowish or pinkish and fine grained, with a foliation that ranges from fine lamination to a crude, contorted structure. Thin layers or lenses of pyroxene skarn and feldspathic pyroxene skarn are present locally. In part, the gneiss is slightly migmatitic. South-southeast of pit 5, the rock is broken by small faults, spaced a few inches or feet apart, that show apparent horizontal displacement of inches or fractions of an inch.

The footwall of the skarn zone is little exposed, but it seems to be similar to the hanging wall. At pit 3, the footwall is pyroxene-feldspar gneiss with thin interlayers of rusty pyroxene skarn. In the outcrop 50 feet northwest of pit 6, the footwall of the ore itself is pyroxene-feldspar gneiss.

The skarn zone consists of skarn and marble. The skarn is medium to dark green, medium grained, and locally flecked with pyrrhotite or pyrite. The dominant

[30] Grannis, G. D. [no date]. Report on the Vrooman Ridge iron mine: Private mining report lent by J. G. Broughton, State Geologist of New York.

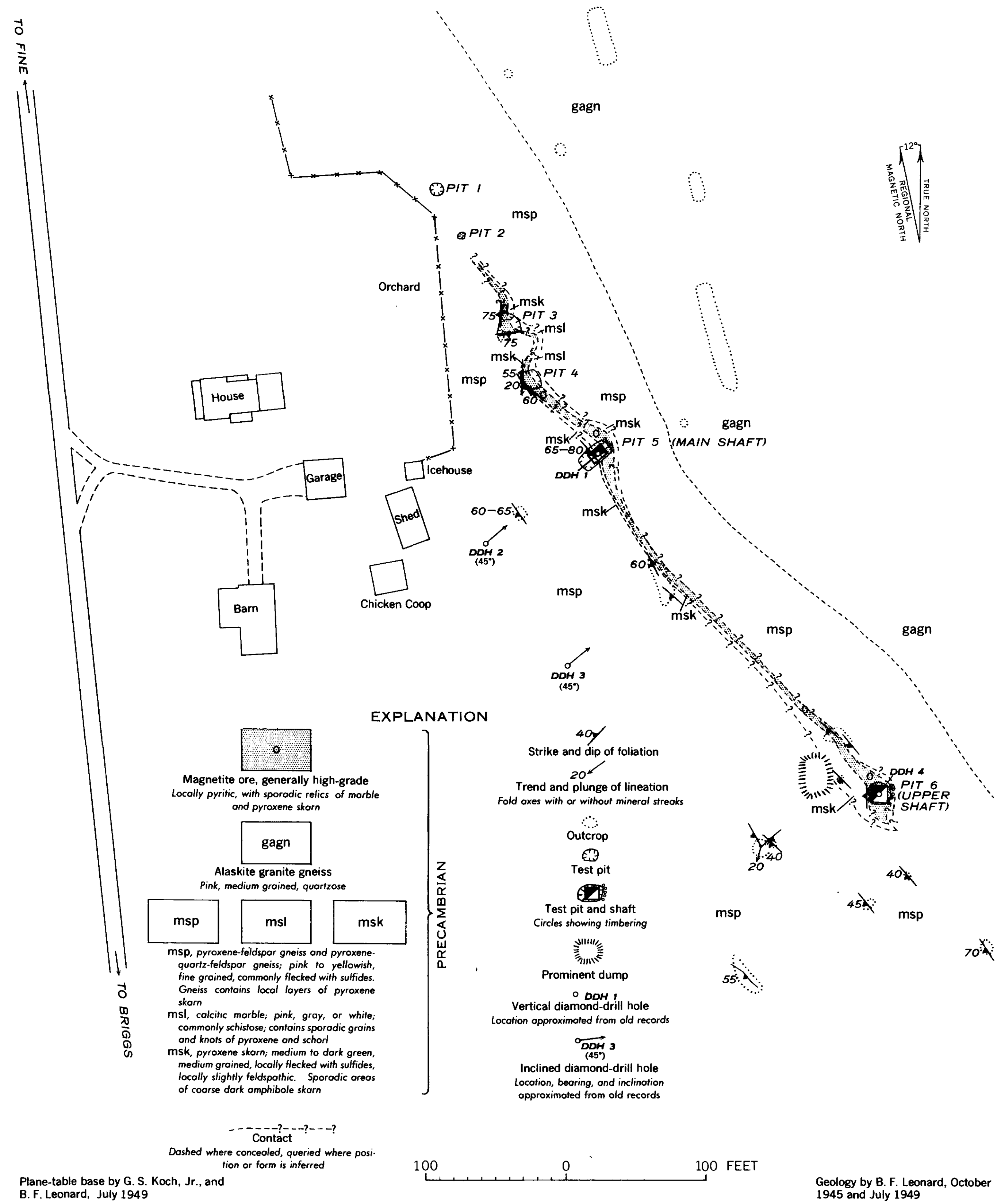

FIGURE 20.—Geologic map of the Green Farm deposit, north-central rectangle, Oswegatchie quadrangle.

skarn mineral is pyroxene, though coarse amphibole is present at the hanging wall in pit 4. In places, the pyroxene skarn is slightly feldspathic or calcitic. The marble is pink, gray, or white. It is commonly fine grained and schistose, showing dark-green pyroxene granules and pale-brown mica flakes disseminated through the fine calcitic matrix. Locally, the marble contains knots of pyroxene and schorl.

All the rocks except marble are more or less stained by limonite derived from oxidized specks of iron sulfides. Many of the rocks show microgranulation, and alteration to epidote, calcite, and (rarely) schorl. In places, the skarn is leached. Whether this leaching was hypogene or supergene, we do not know.

ORE

Occurrence and mineralogy

The ore is in marble at the north end of the deposit; elsewhere it is chiefly in pyroxene skarn ("hornblende schist" of Newland, 1908). Though the ore is essentially conformable to the outline of the skarn zone, the ore does not have a constant "stratigraphic" position within the skarn but weaves back and forth. For example: at pits 3 and 4, the ore is close to the hanging-wall gneiss; at pit 5, several feet of skarn separates the ore from the hanging-wall gneiss; and in the outcrop 50 feet northwest of pit 6, the ore comes against footwall gneiss. Locally, stringers of magnetite are present in the granitic gneisses and in the pegmatite seams parallel to their foliation. Magnetite forms thin veinlets, knots, and massive layers in marble and skarn. Grains and veinlets of pyrite may or may not accompany the magnetite.

Only one specimen of ore was examined under the ore microscope. This specimen, from pit 3, was massive ore showing a faint limonitic boxwork. Microscopically, this specimen consists chiefly of rather coarse magnetite with 5 to 10 percent relict pyroxene and other nonopaque minerals. The magnetite contains many minute "gashes" of exsolved spinel(?). Exsolved hairs of ilmenite in magnetite are sparse. A very little pyrite and traces of sphalerite are present locally. These minerals belong to the normal hypogene suite. Dusty limonite was lost in polishing the specimen, but there remained a few patchy areas and thin veinlets of a secondary iron oxide. The mineral appears to be maghemite, the ferromagnetic ferric oxide. The maghemite veinlets show local colloform banding, contain rare angular specks of magnetite, and seem to have replaced the magnetite host. Patchy areas of maghemite have also replaced magnetite but seem to have left the intergrown pyrite untouched.

The maghemite was too sparse for thorough investigation; however, its identification is fairly certain. The powdered maghemite, free of magnetite inclusions, jumps quickly to a magnetized needle. The maghemite has about the same hardness as magnetite. Optically, the maghemite is commonly blue gray and isotropic; locally it is very weakly pleochroic and very weakly anisotropic. It shows a very weak pinkish-yellow internal reflection along cracks; elsewhere, there is no internal reflection. The somewhat weathered appearance of the hand specimen, and the association of maghemite with powdery limonite, strongly indicate that the maghemite is supergene.

Grade and thickness

The grade of ore well exposed in pits and outcrops is medium to high; the grade of ore poorly exposed or beyond reach in the pits could not be estimated. The grade of ore on the dumps ranges from low to high, and some lump ore is present.

Table 17 gives analyses of lump ore from the deposit. Obviously, the samples were carefully selected to show the best ore available, in a day when hand sorting was practical.

The thickness of ore is variable. The maximum exposed thickness in pits 3 and 4 is 2.5 and 1.5 feet, respectively. The maximum possible thickness (assuming ore where the pit walls have caved) is about 4 to 5 feet. Newberry [31] (1890, p. 4) reported the ore in pits 3 and 4 to be about 5 feet thick.

TABLE 17.—*Chemical analyses (weight percent) of ore from the Green Farm deposit*

[From Newberry [31] (1890). Analyses also quoted by Newland (1908, p. 139). Analyst, S. B. Newberry]

	1 [1]	1a	2	2a	3	3a
Iron	71.12	70.95	70.74	70.86	--------	--------
Phosphorus	.049	.048	.080	.077	--------	--------
Sulfur	.005	.004	.019	.018	0.56	0.61
Titanic oxide	Tr.	--------	--------	--------	--------	--------
Silica	.860	--------	--------	--------	--------	--------
Manganese	Tr.	--------	--------	--------	--------	--------
Magnesia	Tr.	--------	--------	--------	--------	--------
Lime	.051	--------	--------	--------	--------	--------

[1] Fe calculated by Newberry from 98.22 percent Fe_3O_4. Mn, SiO_2, MgO, and CaO determined by Newberry on mixed sample of ores from Main shaft and Upper shaft. Sum of oxides in complete analysis is 99.185 percent.

1. Composite sample of ore from Upper shaft (pit 6).
1a. Duplicate of sample 1.
2. Composite sample of ore from Main Shaft (pit 5).
2a. Duplicate of sample 2.
3. Composite sample of ore from bottom of Main Shaft.
3a. Duplicate of sample 3.

The maximum inferred thickness of ore at pit 5 (Main shaft) is 8 to 9 feet. Newberry reported 9 feet of ore at the bottom of the shaft.

The ore is at least 2.5 feet thick in the outcrop 100 feet southeast of pit 5. The ore is only 3 to 4 feet thick in the outcrop 50 feet northwest of pit 6.

At pit 6 (Upper shaft), the maximum inferred thickness of ore is 12 feet. Newberry reported 12 feet of ore at the bottom of the shaft.

[31] Newberry, S. B., 1890, Fine ore bed: Private mining report lent by J. G. Broughton, State Geologist of New York.

Thus the ore now or formerly exposed ranges in thickness from about 3 to 12 feet. The maximum strike length of the ore zone is about 500 feet. The ore zone may or may not be single and continuous. This problem is discussed below, in the section on structure.

STRUCTURE

The writer believes that the Green Farm deposit is confined to a single, northwest-trending zone that shows a series of small folds plunging southwestward at various angles.

The general strike of the immediate wallrocks is more westerly than the trend of the whole belt of metasedimentary rocks (School No. 15 syncline)—more westerly, even, than the trend of nearby outcrops of alaskite gneiss. The trend of the pits agrees with the general strike of the wallrocks; it is N. 40° W. The dip of wallrocks and ore is highly varied, ranging from 20° SW. to 80° SW.

The axes of minor folds having an amplitude of 5 to 10 feet are measurable at two places. In the curving hanging wall at pit 4, a small roll has its axis trending about S. 15° W.; the plunge of this fold could not be measured directly, but it must be south-southwestward at a variable though moderate angle. A small anticline in the outcrop 90 feet southwest of pit 6 trends S. 14°–15° W. and plunges 20° SW. Thus the approximate orientation of one set of folds is well established by direct observation.

At other places, folds are clearly indicated by divergent foliations, though the orientation of these folds cannot be measured directly in the field. One can determine the orientation of the fold axes by plotting the foliations on a stereonet, for example. The values so determined are only approximate; they have all the errors inherent in the field measurement of the attitude of the individual foliation surfaces. Using this indirect method, we get the following data. At pit 3, the inferred fold axis trends S. 45° W. and plunges 69° SW. At pit 5, the inferred fold axis trends S. 28° W. and plunges 72° SW. At pit 6, where observations are very sketchy indeed, the inferred fold axis trends S. 48° W. and plunges southwestward. Thus the orientation of fold axes determined indirectly agrees roughly with the measured orientation of fold axes at pit 4 and at the outcrop southwest of pit 6: all folds plunge in the southwest quadrant.

Does the ore follow the axes of these small folds? Indirect evidence suggests that it does. First, all the pits are on or near folds and bends in the structure. Some of the early mining men were shrewd fellows: they usually opened up the most favorable segments of a deposit, guiding their work by dip needle, float, and outcrops. Second, the results of Grannis' diamond drilling are consistent with the pattern of southwest-plunging folds. The evidence and the earlier interpretation of this drilling are summarized below.

Newland (1908, p. 138) stated that there appeared to be two veins, an eastern and a western, with two pits on each. Newberry (1890, p. 4) thought that the ore exposed in pits 3 and 4 belonged to a vein roughly parallel to the main vein opened up by the two shafts, but that the western vein bore off more to the southwest. (Cf. bearing of a line joining pits 3 and 4, fig. 20.) This interpretation seemed to be supported by his dip-needle survey (Newberry, 1890, p. 1). Perhaps the observations made by Newland and Newberry were colored by the work of Grannis,[32] done between 1884 and 1889. According to Grannis, hole 2 (see fig. 20) cut 4 feet of ore, 4 feet of rock ("horse"), and 10 feet of ore. "These veins seemed to be the same veins that were opened by two pits [3 and 4] about 100 feet north of the main shaft." Grannis stated that hole 3 cut 4 feet of ore, 6 feet of rock ("horse"), and 6 feet of ore. He makes no correlation here, but the implication is that hole 3 was testing the two veins cut in hole 2. Remember that we do not know the exact location of the holes, the depths at which they cut ore, or the true thickness (as opposed to apparent thickness) of the ore.

The apparent existence of two principal vein systems—parallel, overlapping, and less than 150 feet apart, as Grannis later noted—has an alternate explanation. Consider the structural data shown on figure 20. If the ore exposed in pit 4 should continue down the axis of the small fold, hole 2 would cut this ore some distance west of the downdip projection of ore exposed in pit 5. The ore in pit 5 also seems to lie on a small fold. The trough of this synclinal fold might lie well above the extension of hole 2, assuming that hole 2 continued far into the supposed footwall of the ore reported by Grannis. The ore cut by hole 3 might be the down-plunge extension of ore exposed in pit 5. At present, the number of veins remains undetermined, but the available evidence favors the existence of a single magnetite-bearing zone thrown into a series of small southwest-plunging folds.

ECONOMIC POSSIBILITIES

Cohen's dip-needle survey (Reed and Cohen, 1947, p. 12), giving a length of 500 feet for the intense anomaly, indicates that the deposit is short. Little is now known of its average thickness or extension in depth. The grade of ore is attractive, but the inferred small size will probably discourage further prospecting.

[32] Grannis, G. D. [no date]. Report on the Vrooman Ridge iron mine: Private mining report lent by J. G. Broughton, State Geologist of New York.

On the aeromagnetic survey of 1945, the Green Farm deposit registered only as a small nose in the 300-gamma contour (Balsley, Hawkes, Buddington, and others, 1946). Relative to the datum adopted for magnetic maps of adjoining areas (Balsley, Buddington, and others, 1954a, 1954b), this nose is in the 5,500-gamma contour. The aeromagnetic data are consistent with other evidence suggesting that the deposit is relatively small.

If additional drilling is ever undertaken, one should keep in mind the possibility that the ore follows southwestward-plunging folds. Known folds trend S. 15° W. and have a variable southwest plunge. The axes of these folds might trend farther southwest—even so far west at S. 70° W. Conceivably, two dominant fold systems could each affect the distribution of ore. The best one could do, in guiding the exploration, would be to study each bit of structural information as it became available.

As a preliminary to further drilling, a careful dip-needle survey should be made. The survey should be adequately tied to the base map of the pits. It might be desirable to supplement the dip-needle work with a superdip survey of the covered area south-southwest of pits 3, 4, and 5.

HARDWOOD MILL DEPOSIT

LOCATION AND ACCESS

The Hardwood Mill deposit is 4 miles northeast of the village of Newton Falls, 0.85 mile northwest of the summit of Buck Mountain, and about 1,000 feet south of the Oswegatchie River, northwest and north-central rectangles, Cranberry Lake quadrangle. The east part of the deposit is on high ground above the river; the west part is in a swamp. Meridian 74°55′ west longitude passes through the deposit. The deposit, also known as the Ross Oswegatchie prospect and Area H, takes its name from a hardwood mill that formerly stood a short distance upstream.

To reach the deposit, turn off the Newton Falls-Cook Corners road 1 mile east of the bridge that crosses Tooley Pond Outlet and take a boat upstream 1,000 feet to the south bank. Walk 1,000 feet southward through the brush to the site of the magnetic anomaly. A system of abandoned logging roads runs from Jarvis Bridge through Chaumont Swamp to the neighborhood of the Hardwood Mill deposit; in 1952, access by this means was not advisable.

DISCOVERY AND EXPLORATION

The existence of local magnetic attraction in the area had been know for some time, as foresters and surveyors had had difficulty in running compass lines there. The deposit was discovered by A. M. Ross in 1941 as the result of reconnaissance dip-needle work in this area of local attraction. Several shallow test pits were sunk on the eastern part of the deposit in 1942. A detailed dip-needle survey of the area was made under Ross's direction in July 1934, and 4 short diamond-drill holes were put down on the deposit between July and November of that year. The holes were drilled only to test the quality and thickness of the ore, not to prove up tonnage.

MAGNETIC ANOMALY

The main part of the magnetic anomaly (see pl. 8) begins about 200 feet west of the boundary between the Newton Falls Tract and lands purchased from the Tri River Power Corp. This part of the anomaly extends westward for 1,500 feet before it ends rather sharply. The east end of the anomaly fades gradually into low positive readings (not shown on pl. 8) traceable for some distance east of the old boundary line. The significance of this part of the anomaly is at present unknown.

The main part of the anomaly consists of two magnetic "ridges," 700 and 500 feet long, effectively within the +40° isoclinal contour. These ridges are separated by a "saddle" of lower positive readings 180 feet long. The western magnetic ridge is outlined by relatively smooth isoclinal contours and is paralleled on the north by an area of low negative readings. The eastern magnetic ridge is much less regular and is paralleled on the north by a sharp magnetic "trough," or negative anomaly. For both ridges, the magnetic profile is steeper on the north than on the south. The different character of the west and east magnetic ridges is due in large part to the difference in depth of overburden—50 to 60 feet in the west part and less than 15 feet in the drilled segment of the east part. (The character of the magnetic profiles suggests that the depth of overburden may increase eastward from hole H103, reaching a depth of 30 to 35 feet at the east end of the east ridge.) However, inferred tight folds in the area between 50W and 250W also complicate the magnetic pattern of the east ridge, and it is possible that a local change in the grade of ore has further roughened the magnetic "relief."

The Hardwood Mill anomaly registered as a prominent peak of 6,030 gammas on the aeromagnetic survey of 1945 (Balsley, Buddington, and others, 1954a).

GEOLOGY

The Hardwood Mill deposit is in an east-trending belt of metasedimentary rocks interlayered with sheets of younger granite. The belt, most of which is obscured by glacial debris, lies between two sizeable masses of granite. The northern mass is alaskite with included layers and schlieren of amphibolite; the southern mass is chiefly hornblende granite that separates the Hardwood Mill and Trembley Mountain areas. Rocks of the belt of mixed metasedimentary and granitic rocks are tightly folded and steeply dipping. In plan, the folds seem to run roughly east and west, but the attitude of their axes is indeterminate. The prevailing lineation for several square miles around the deposit trends roughly southeast and plunges gently or moderately southeast.

ROCKS

There are no outcrops on or near the magnetic anomaly. When the writer examined the area in 1944, the test pit 75 feet north-northeast of the collar of hole H103 showed pink alaskitic granite with schlieren of pyroxene skarn partly replaced by magnetite. The strike of the schlieren was east. The other test pits were muddied and useless. Thus our knowledge of the Hardwood Mill deposit comes almost solely from drill cores. (See pl. 8.) The Geological Survey logged core from holes H101 and H103, and the company's logs of the other two holes were subsequently made available to the writer.

The magnetite deposit is a partial replacement of a layer of skarn enclosed within alaskite. The layer is generally about 35 feet thick. The unreplaced skarn consists largely of dark-green pyroxene (salite), in places accompanied by an indeterminate but substantial quantity of dark-green amphibole (hornblende?), a little dark-green biotite, or a little dark-red-brown garnet (andradite). Locally the skarn is very slightly feldspathic. Recrystallized primary calcite is lacking, but in places the skarn is cut by thin calcite veinlets adjoined by chloritic alteration.

The alaskite is pink and for the most part medium grained, though some of it is aplitic. Where the alaskite comes against skarn, it tends to be syenitic. This syenitic facies carries xenocrysts of pyroxene, or shreds and schlieren of the skarn itself. Coarse, contaminated syenite pegmatite is present locally.

A thin zone of biotite gneiss, in part migmatitic, was cut in the hanging wall of the ore zone by holes H102 and H103. This rock does not appear in holes H100 and H101, probably because the drill would have entered bedrock just a few feet north (on the footwall side) of the point where the biotite gneiss should appear. In addition, hole H102 is said to have cut several zones of biotite gneiss, in part mineralized, interlayered with skarn and pyroxene gneiss in the lower third of the hole. It is possible, but by no means certain, that this northern biotitic zone is the highly contorted equivalent of the southern biotitic zone, repeated by folding.

According to the company log, pyroxene-feldspar gneiss, with or without quartz or garnet, is abundant in the core from hole H102. Whether this material is metasedimentary, or merely a facies of alaskite contaminated by skarn, the writer does not know.

STRUCTURE

The rocks enclosing the Hardwood Mill deposit strike east and dip steeply south at angles ranging from 50° to 90°. The average dip seems to be about 70°–80° S., certainly varying along the strike and probably varying down dip as well. An alternate interpretation of the core-foliation angles would indicate dips of 10°–25° S. However, the latter interpretation is inconsistent with the known magnetic profiles and the inferred tightly folded structure of the belt, both of which suggest a steep dip for the magnetite body. No attempt has yet been made to drill a fan of holes to determine the cross-sectional shape of the body.

At the longitude of hole H102, the wide variation in core-foliation angles within a few feet of drill core, the repetition of biotite gneiss zones and mineralized skarn zones, and the swirled pattern of the magnetic anomaly suggest the presence of one or more tight isoclinal folds in the bedrock. A line connecting the "pinched" negative anomalies centering at (225N, 305W) and (125N, 205W) bears S. 45° E. The observed lineation (including axes of minor folds) in rocks of this general neighborhood is known from regional mapping to trend east-southeast to south-southwest and plunge 10°–45° SE. Thus it is possible that the inferred fold or folds at the longitude of hole H102 also trends southeast, through in rocks striking east and dipping 80°–90° S. the plunge of the inferred fold would have to be much steeper than 45°. The effect that such a fold might have on the structure of the ore body is discussed in a separate section, below.

ORE

Occurrence

Material of ore grade is generally confined to two zones within the limits of the main skarn layer. The upper (southern) zone is 3 to 5 feet thick. It is well defined in all holes except hole H102, where the zone is only weakly mineralized. A parting of skarn and pyroxene gneiss, 6 to 15 feet thick, separates the upper zone from the lower (northern) zone. This lower zone is 21 to 28 feet thick beneath the western "ridge" of the anomaly, and 9 to 10 feet thick beneath the drilled segment of the eastern magnetic ridge. Magnetite replaces the dark silicates preferentially, avoiding areas of pegmatite, pyroxene gneiss, and feldspathic skarn within the general limits of the main skarn layer. Most of the material of ore grade is compact magnetite with scattered granules and thin, irregular relict layers of skarn, but some of the material of ore grade consists of thinner magnetite streaks and layers interspersed with skarn. The position of the magnetite-rich zones is slightly undulatory within the limits of the main skarn layer: in places, ore and footwall alaskite are in contact, but elsewhere the two are separated by a parting of unreplaced skarn.

The shape of the magnetite body is unknown. It may be crudely tabular, extending downdip and along strike in a series of gentle pinches and swells. However, it may consist of one or more principal shoots controlled by the axes of inferred tight folds having their suboutcrop[33] at the longitude of hole H102. Ross[34] suggests that, inasmuch as the magnetite-bearing material cut in hole H102 is much thinner and lower in grade than the corresponding "stratigraphic" zone cut in hole H103, hole H102 may not have intersected the shoot cut by hole H103. This is a reasonable interpretation of the evidence if, for example, the upper edge of a hypothetical shoot (cut by hole H103) plunging S. 45° E. has its suboutcrop halfway between holes H103 and H102. Then the point where hole H102 intersected the mineralized skarn zone would be about 320 feet vertically (essentially the same distance updip) above the expected edge of the plunging ore shoot.

Mineralogy

Magnetite is the only ore mineral. Locally it is accompanied by a little pyrite and chalcopyrite, but as a rule sulfides are sparse. Dark silicates in the form of disseminated granules and thin, irregular films and layers are present even in the high-grade ore. The ore from this deposit has not been studied mineragraphically. However, it is likely that the small quantity of soluble nonmagnetic Fe reported in the assays (see table 18) is due chiefly to partial solution of Fe in the dark silicates, rather than to the presence of hematite.

Grade

Table 18 summarizes the information on grade of ore from the Hardwood Mill deposit. The grade of the upper and lower zones is given separately, and in addition the grade of the maximum thickness of material of ore grade is noted. Assays for soluble Fe and magnetic Fe for individual 5-foot samples of magnetite-bearing core are on record with the Newton Falls Paper Mill. The ore is low in P, S, and TiO_2. Roughly 5 feet of high-grade ore is present in the lower zone cut by holes H100, H101, and H103. Under especially favorable economic conditions, it might be feasible to recover this lump ore separately.

Extent

From the character of the magnetic anomaly, the results of the very limited diamond drilling, and a knowledge of similar deposits in the district, one might reasonably infer that the Hardwood Mill deposit has a strike length of about 1000 feet of medium-grade ore. An additional 400 feet of magnetic anomaly is favorable for prospecting. The rest of the anomaly (including the eastward extension) cannot be properly evaluated at this time.

The complex structure of skarn deposits makes it hazardous to project a known thickness of ore downward. The thickness of the ore is almost sure to change—abruptly, in places. This does not mean that all surprises would necessarily be unpleasant ones. It does mean, however, that much drilling would be required to determine the size, shape, and attitude of a deposit of this character. The results of preliminary drilling have been exceptionally favorable.

HOLLYWOOD CLUB ANOMALY

The Hollywood Club anomaly is 1 mile west of the south end of Clear Pond, southeast rectangle, Stark quadrangle. The anomaly was discovered by the aeromagnetic survey of 1945, on which it was registered as a peak of 6,450 gammas (Balsley, Buddington, and others, 1954b). A detailed dip-needle survey of the anomaly was made in 1946 by the U.S. Bureau of Mines (Reed and Cohen, 1947, p. 11 and fig. 17). The following account is slightly modified from that given by Cohen.

The anomaly, on land posted by the Massawepie Club, may be reached by driving half a mile northwest of the buildings of the Hollywood Club and then walking west-southwest about 3,000 feet. (See pl. 1.) The dip-needle anomaly is 1,050 feet long within the +20° isoclinal, and 470 feet long within +50° isoclinal. (See fig. 21.) The axis of the anomaly trends northeast. Pyroxene skarn crops out near the intense part of the anomaly, but the magnetic source of the anomaly is not exposed. Foliation in a skarn outcrop near the anomaly strikes N. 65° E. and dips nearly vertically (Cohen, C. J., personal communication). The presence of wall-rock skarn and the character of the anomaly suggest that a magnetite body of skarn type yields the dip-needle anomaly.

Except for the large mass of hornblende granite to the north, the area surrounding the anomaly is obscured by drift. (See pl. 1.) Structural data from neighboring areas suggest that the Hollywood Club anomaly

[33] "Intersection of a vein (or other structural feature) with an unconformity. What would be the outcrop if the overlying unconformable formation were removed" (McKinstry, 1948, p. 656). In this instance, the unconformable formation is glacial debris.

[34] Ross, A. M. [no date]. Report on core drilling project, Magnetic Area H: private report for use of Newton Falls Paper Mill, 2 p., logs of 4 drill holes, and detailed magnetic map.

TABLE 18.—*Assays of core from holes H100–H103, Hardwood Mill deposit*

[Assays by Jones & Laughlin Ore Co. for Newton Falls Paper Mill. Data condensed from Ross (written communication), published by permission. U, upper zone; L, lower zone; Max., maximum thickness of material of ore grade]

Hole	Zone	Footage	Apparent thickness (feet)	True thickness (feet)	Weight percent in crude ore					
					Soluble Fe	Magnetic Fe	P	S	TiO_2	SiO_2
H100	U	79.5– 83.5	4.0	3.3	36.25	33.88	--------	--------	--------	--------
	L	91.0–116.5	25.5	20.9	45.89	44.32	0.018	0.309	0.19	--------
	Max.	79.5–119.5	40.0	32.8	34.0	32.6	--------	--------	--------	--------
H101	U	74.4– 80.0	5.6	4.9	34.18	30.03	--------	--------	--------	--------
	L	89.0–120.8	31.8	27.7	34.6	32.5	--------	--------	--------	--------
	Max.	74.4–120.8	46.4	40.3	29.93	26.64	.026	.323	.07	28.00
H102	U	[1] 61.0– 70.0	9.0	8.2	([1])	([1])	([1])	--------	--------	--------
	L, Max.	96.0–106.0	10.0	9.1	31.40	29.14	--------	--------	--------	--------
	--------	[1] 170.0–193.0	23.0	20.9	([1])	([1])	([1])	--------	--------	--------
H103	U	43.0– 49.0	6.0	4.9	27.24	23.61	--------	--------	--------	--------
	L	67.0– 79.0	12.0	9.8	43.4	40.2	--------	--------	--------	--------
	Max.	43.0– 79.0	36.0	29.5	21.2	[2] 19.1	--------	--------	--------	--------

[1] Weakly mineralized; not assayed.
[2] Subore or marginal ore.

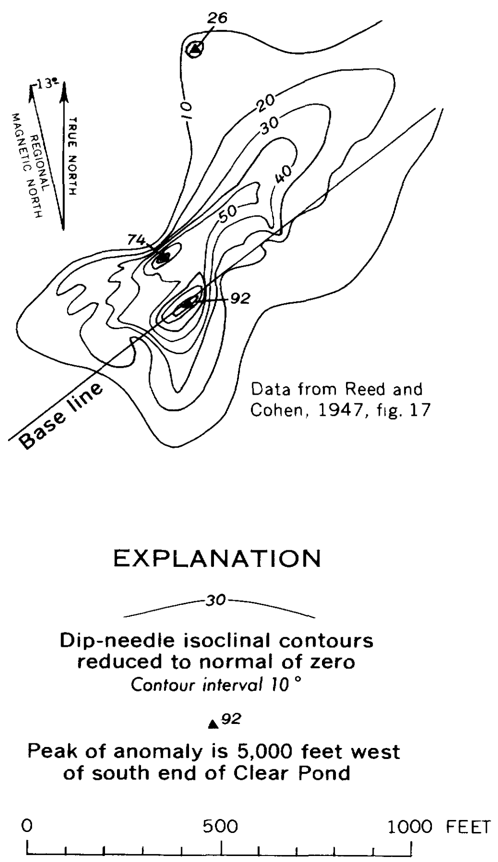

FIGURE 21. Magnetic map of the Hollywood Club anomaly, southeast rectangle, Stark quadrangle.

is close to the center of an area of diverging lineations. (See Prof. Paper 376, fig. 15.)

JAYVILLE DEPOSIT

SUMMARY

The Jayville deposit, opened in 1854 but abandoned since 1888, yielded between 25,000 and 200,000 tons of ore from scattered pits and underground workings. The ore shipped ranged in grade from 40 to 60 percent Fe. The deposit was explored in 1941 by the Jones & Laughlin Steel Corp., whose eight drill holes tested parts of the deposit to a vertical depth of 200 to 250 feet. Drilling showed that the main ore body was a short lens, 6 to 17 feet thick, averaging 43 to 46 percent soluble Fe. Phosphorus, sulfur, and titanium are low. Other magnetite-bearing zones of similar grade were encountered, but these are smaller and thinner. The possibility of finding commercially significant tonnages of medium-grade or low-grade iron ore seems slight. The main ore body appears to be a shoot, plunging moderately northward in steeply dipping rocks on the northwest limb of a compound, overturned, isoclinal syncline. Magnetite is accompanied in places by vonsenite (ferrous ferric borate) and by supergene hematite. The nonmagnetic tails from one core interval assaying 32 percent magnetic Fe ran 38 percent soluble Fe. Part of the nonmagnetic iron mineral in this interval is hematite, but much of it may be vonsenite. The ore minerals selectively replace dark mica sköls, pyroxene-amphibole skarn, and quartz-bearing amphibole skarn in dark skarn zones, predominantly pyroxenic, flanked by contaminated alaskite gneiss and sodic granite. Parts of the skarn and ore are rich in fluorite. Anthophyllite skarn is locally prominent.

LOCATION, OWNERSHIP, AND SOURCES OF INFORMATION

The Jayville magnetite deposit, west-central rectangle, Oswegatchie quadrangle, is 1.25 miles east of the hamlet of Kalurah. In 1948, one could drive a passenger car from Kalurah to a point within half a mile of the north end of the deposit. The Carthage and Adirondack Division of the New York Central Railroad runs past the deposit. The workings are on the east and west flanks of a low north-trending hill, just west of the former Jayville railroad station. (See pl. 1.)

The Jayville property, comprising Lots 111, 112, 130, and 131, Pitcairn Township, was formerly owned by the Benson Iron Ore Corp., New York, N.Y., and Morristown, N.J.

The following report is based in part on the work of Tyler and Wilcox,[35] who investigated the property in 1941 for the Jones & Laughlin Steel Corp. The Geological Survey subsequently mapped the areal geology, examined the outcrops and the dumps on Jayville hill, logged the drill core, and studied about 60 thin and polished sections of the cores. The Survey party did not repeat the company's detailed mapping. Data on history, production, and recent exploration are taken from Tyler and Wilcox. Most of their main conclusions on the structure, size, and economic significance of the deposit are accepted. The inferred contacts on their map have been modified (pl. 9) to include the drilling data and the writer's interpretation thereof. Leonard, while accepting responsibility for the report, maps, and sections in their present form, acknowledges his debt to Messrs. Tyler and Wilcox and trusts that his synthesis of new and old data is as soundly based as theirs.

HISTORY, PRODUCTION, AND RECENT EXPLORATION

According to old reports,[36] the Jayville property was opened for iron ore in 1854 by Z. H. Benton. Little ore was removed at that time. In 1886, upon completion of the Carthage and Adirondack Railroad to Jayville, the property was leased from Benton by the Byron D. Benson interests. Operation by the Bensons, under the name of the Magnetic Iron Ore Co., continued into 1888, in which year the property was abandoned in favor of the deposit at Little River (now Benson Mines).

Reliable estimates of production are not available. Smock (1889, p. 43) estimated that about 25,000 tons of ore was shipped before the mine was closed in 1888. Prospectus reports estimate the total amount of ore shipped as high as 200,000 tons. Ore for shipping was evidently hand sorted; low-grade ore was left on the dumps. The miners spoke of this ore as "concentrate," and the plan probably was to subject it to concentration processes at a later date, according to Tyler and Wilcox.

When lump ore was at a premium, just before World War II, the Jayville property seemed promising enough to warrant exploration. Accordingly, in 1941 the Jones & Laughlin Steel Corp. made geologic, dip-needle, and Hotchkiss superdip surveys of the area on a scale of 1:600 and drilled eight inclined holes whose total length was 2,336 feet. Most of the diamond-drill holes were 200 to 250 feet deep, but several were about 350 to 450 feet deep. As a result of this exploration, the company concluded that the property was of little interest as a potential producer of lump ore.

PRINCIPAL WORKINGS

The principal workings (see pl. 9), described by Tyler and Wilcox, are listed counterclockwise, beginning at the north.

New Find (*Hustler mine*).—Small pit, evidently in an isolated body of magnetite; old analyses gave 52 and 58 percent Fe. (Not shown on pl. 9.)

Benson No. 1.—Shaft, inclined about 65° NW., reported by Smock (1889) to have a slope length of 350 feet; other sources report the depth as perhaps not more than 200 feet. On 100-foot level, drifts are said to run 150 feet south and 50 feet north of the shaft. Smock (1889) reports that the ore zone was 4 to 20 feet wide for a length of 400 feet. Old analyses gave 25 to 49 percent Fe and averaged about 38 percent. This opening is said to have supplied most of the shipping ore from Jayville.

Adit.—Said to have opened up a lens of ore 60 or 70 feet long and 20 feet wide; no analyses available.

Fuller No. 1 (*Open Cut*).—Said to be 100 feet deep. Two old analyses gave 47 and 52 percent Fe.

Benson No. 2.—Shaft inclined 55° W., said to be 60 feet deep and to have penetrated 12 feet of ore; ore ranged in grade from 37 to 49 percent Fe and averaged about 40 percent, according to old analyses.

Fuller No. 2 (*Essler pit*).—Opening said to be about 50 feet deep, reached by an inclined shaft; ore shipped averaged about 54 percent Fe.

Fuller No. 3.—Opening 20 by 30 feet, reached by an adit; no analyses available.

Hart No. 1 (not shown on pl. 9).—Shaft, just west of old railroad station, said to be 300 feet deep with a 50-foot drift at bottom. Ore shoot said to be 10 feet thick; old analyses gave 51 to 61 percent Fe and averaged about 55 percent.

Hart No. 2.—Shaft said to be about 60 feet deep; no analyses available.

New York No. 1.—Shaft, 10 by 20 feet, inclined northwest, said to be 300 feet deep with short drifts at bottom. Old analyses gave 45 to 52 percent Fe and averaged about 47 percent.

[35] Tyler, S. A., and Wilcox, R. E. [1942]. Geological report on the Jayville area, Pitcairn Township, St. Lawrence County, New York: Private report to Jones & Laughlin Steel Corp.

[36] Tyler and Wilcox based their account of the early history and workings of the mine mainly on the following sources:
- A summary of testimony given to Mr. L. M. Southern in 1915 by men who worked in the mines when they were operating.
- Chance, H. N. [private rept.].
- Hobbs, P. S., Jayville Mines [private rept.].
- Knox, C. L., 1913 [private rept.].
- Menoher, J. C., 1915, The Jayville magnetic iron ores [private rept.].
- Newland (1908).

New York No. 2 (*Leary*).—Shaft said to be 50 feet deep; old analyses gave 22 to 38 percent Fe.

MAGNETIC ANOMALIES

A series of short, intense positive magnetic anomalies, trending north-northeast, was shown by the dip-needle survey made by the Jones & Laughlin Steel Corp. (See pl. 9.) Commonly, these anomalies are rather blunt at the south and somewhat more drawn out at the north. The longest continuous positive anomaly, between Benson shafts Nos. 1 and 2, measures only 350 feet within the +25° isoclinal. Since the null for the dip needle used here was about −15°, the +25° isoclinal is roughly equivalent to one of +40° corrected to a null of 0°. Negative anomalies are not prominent, except at the south end of the area, east of the Fuller No. 3 opening. All but one of the anomalies has been prospected by drilling or exploited at some depth. The single exception, the anomaly 300 feet northwest of Hart No. 2 shaft, has several test pits on small boudins of skarn. Whether these skarn boudins were joined at depth in a more nearly continuous and perhaps "ore"-bearing layer, no one knows. Because of the small size of the known magnetite bodies at Jayville, this undrilled anomaly is not likely to be worth further testing for magnetite.

A few very small, isolated "hot spots" and less intense magnetic anomalies, found during the survey, are omitted from plate 9.

The company surveyed the north part of the area by means of a superdip, a magnetic detector more sensitive than the dip needle. This survey was made to search for buried ore reported to have been found by drilling east of the branch of Little Bear Creek. (Note old drill hole, date unknown, pl. 9.) According to old reports, the ore had not been exploited because pumps were inadequate to cope with the water in that swampy place. The dip needle had showed no abnormal magnetic attraction here, the superdip failed to show abnormal attraction, and two holes—J6 and J7—drilled by the Jones & Laughlin Steel Corp. cut no magnetite or skarn. Tyler and Wilcox concluded that the old reports were unreliable.

The Jayville deposit was scarcely detected in the aeromagnetic survey of the region. It shows only as a slight displacement of the 200-gamma contour on the map of Balsley, Hawkes, Buddington, and others (1946). The 200-gamma contour of that map is equivalent to the 5,400-gamma contour relative to the datum adopted for aeromagnetic maps of adjoining areas.

GEOLOGY

The Jayville deposit is in a synclinal area of metasedimentary and younger granitic rocks that forms an "island" within the older granitic (quartz syenitic) rocks of the Diana complex. (See pl. 1.) This "island," geographically near the east margin of the complex, is perhaps close to the main axis of the great overturned syncline that involves the complex. In addition, the "island" is near a marked bend in the trend of the complex, is close to the boundary between areas of different lineation type, and may be on or near the axis of an east-northeast-trending cross fold. (See Prof. Paper 376, figs. 14, 15, 16.) The gross structural setting of the Jayville deposit thus resembles that of the Clifton deposit (p. 110–111), whose enclosing rocks are marginal to the Stark complex.

Foliation in the area about Jayville dips 35° to 50° NW., though on Jayville hill the dips are steeper. (Cf. pls. 1, 9.) The strong lineations within a 2-mile radius of Jayville have two distinct trends: (a) northwest, generally downdip of foliations belonging to northeast-trending folds; and (b) north-northwest, subparallel to the axes of north-trending folds. From the asymmetry of minor folds a few feet to a few hundred feet across, Tyler and Wilcox inferred that Jayville hill lies on the east flank of a major overturned anticlinal structure, or on the west flank of a major synclinal structure. That inference presupposes that the major anticlinal structure is of the socalled "normal" type, in which the axial planes of minor folds converge downward. Folds developed by irregular flowage or by shearing would not be so disposed. All three types of fold occur in the district. Because the axis of the major synclinal structure involving the Diana complex cannot be precisely located at this latitude, the relative structural position of Jayville hill is at best indefinite. Unfortunately, the relative structural position is important, for the assumptions made about it determine the way cross sections are drawn; these in turn govern one's interpretation of favorable sites for occurrence of magnetite at depth. (See discussion of local structures, p. 143–145.)

ROCKS

The granitic and metasedimentary rocks at Jayville are, in several respects, unique for the district. Most of the granitic rocks here are interpreted as contaminated alaskite and alaskite gneiss of the younger granite series, though a very little phacoidal hornblende granite gneiss of the Diana complex is also present. Much of the contaminated alaskitic material is an augen granite gneiss of a sort not known to occur in force anywhere else in the district. This granite gneiss yields the principal exposures in the Jayville area, encloses the local belts of metasedimentary rocks, and occurs as sheets more than 60 feet thick within the metasedimentary rocks. Some medium-grained, contaminated granite gneiss and uncontaminated alaskite gneiss are associated with the augen granite gneiss.

Another unusual facies of younger granite, highly sodic, is prominent at just one other place in the district (0.5 mile west of Crane Pond, northwest rectangle, Cranberry Lake quadrangle); a similar rock occurs within alaskite at the Brandy Brook magnetite deposit.

The metasedimentary rocks are mainly skarns, though some biotite-plagioclase gneiss and amphibolite are also present. The skarns are remarkable for their abundance of clinoamphibole (much of it ferrohastingsite), orthoamphibole (ferroan anthophyllite), and fluorite, and for the thick biotite sköls—generally mineralized—that make up the skarn zones. Anthophyllite skarns are not known to occur elsewhere in the district.

The map prepared in 1941 by Tyler and Wilcox (op. cit.) showed three rock units: granite gneiss, amphibolite (including magnetite bodies), and pegmatitic amphibolite. In modifying their map, the writer has subdivided parts of their granite gneiss unit and has lumped their other two units into one consisting predominantly of skarn with associated metasedimentary rocks and magnetite bodies. (See pl. 9.) The following rock descriptions are based on data from drill cores which, even for the relatively well exposed area of Jayville hill, give a better record of the whole rock sequence than one finds in the outcrops.

GRANITIC ROCKS

Phacoidal granite gneiss

Phacoidal granite gneiss of the Diana complex underlies much of the area near Jayville hill (see pl. 1), but the writer has found this rock in just one place in the drill core. A 2-foot layer of granulated, phacoidal hornblende granite gneiss occurs in the middle of the skarn zone cut by hole J3. This phacoidal granite gneiss may represent material injected as a sheetlike apophysis into skarn or preexisting marble long before the skarn and parent mass of phacoidal granite gneiss were invaded by younger granitic magma. However, the highly deformed state of the gneiss is not typical of phacoidal granitic rock found elsewhere intrusive into skarns. It seems more likely that the thin layer of phacoidal granite gneiss was torn off some larger mass of parent rock, folded or sheared out in mobile carbonate rocks before their major episode of silication, and preserved as a highly deformed relic in rock that became skarn. Plausible as this explanation may be with regard to the facts of regional geology, there is no means of demonstrating it from the meager data provided by the drill core.

The phacoidal granite gneiss is a pink-and-gray, medium- and fine-grained, strongly foliated, almost schistose rock consisting of flattened lenses or phacoids of pink potassic feldspar in a streaky matrix of gray plagioclase and black hornblende. The phacoids are commonly 1.5 to 2 cm long and a few millimeters thick, though the range is from about half to twice that size. Elongate streaks of potassic feldspar and hornblende give the rock a strong lineation.

Microscopically, the rock is inequigranular, fine to medium grained, with a few coarse, highly elongate microcline grains. Grain boundaries are mainly polygonal or gently curved, but some poorly sutured grains are present. A strong foliation is given by trains of subparallel hornblende prisms, locally accompanied by subparallel biotite flakes; by elongate microcline; and by varying proportions of quartz, microcline, and plagioclase.

An approximate mode is given in table 19, No. 1. Microcline is sharply twinned. Only the coarse grains are slightly microperthitic; the rest are not. Partial rims of albite on microcline are common. Plagioclase, locally slightly cloudy, is less variable than microcline in grain size. The hornblende, a blue-green ferrohastingsite, is slightly altered in places to biotite, chlorite, or mixtures of these minerals. Locally, quartz blebs give the hornblende a poikiloblastic structure. The biotite is pleochroic from light greenish brown to black. Allanite is absent. Carbonate and pumpellyite are very sparse alteration products.

Alaskite gneiss

Sheets of alaskite ranging in thickness from 2 to 15 feet were seen in four places in the drill cores. The alaskite occurs in coarse augen granite gneiss, in medium-grained hornblende granite gneiss between coarse augen granite gneiss and skarn, or a few feet from hornblende granite gneiss, just within skarn. At one place, the alaskite carries a few schlieren of amphibolite. At another place, the alaskite grades into granite pegmatite. The alaskite is pink, medium grained, and outwardly massive. Quartz and pink feldspar constitute more than 95 percent of the rock. Generally one can see a few very small grains of magnetite, or a few small red garnets.

In thin section, the "alaskite" is a medium- to fine-grained, granoblastic gneiss with a foliation given by subparallel biotite flakes, slightly elongate quartz and feldspar grains, and alternating more quartzose and more feldspathic layers. The approximate mode of one specimen is given in table 19, No. 2. The biotite is partly chloritized and contains local stringlets of zoisite(?). Allanite is the most abundant accessory. The opaque minerals, studied only in thin section, seem to include both magnetite and rutilohematite. The alaskite seen in thin section appears to have been deformed.

TABLE 19.—*Approximate modes of representative granitic rocks, Jayville deposit*

[Unless otherwise indicated, mineral composition was estimated by careful inspection, rather than by Rosiwal analysis]

	Phacoidal granite gneiss	Alaskite gneiss	Medium-grained contaminated alaskite gneiss		Schlieren-type contaminated alaskite gneiss	Faint augen-type contaminated alaskite gneiss		Coarse augen-type contaminated alaskite gneiss	Sodic granite			
	1	2	3	4	5	6	7	8	9	10	11	12
Quartz	15	30	31	13	10	29	17	19	19	9	45	18
Microcline	35	33	40	33	30	21	53	48	2	3	3	3
Plagioclase	44	35	24	42	55	38	23	28	73	81	51	73
Other minerals	1	2	5	12	5	12	7	5	6	7	1	6
Total	100	100	100	100	100	100	100	100	100	100	100	100
Myrmekite	X				Tr.	X	X					
Biotite	X	X	Tr.	X	Tr.	X	X	X	[1] X	[2] X	X	
Hornblende	X		X	X	X	X	X	X				
Pyroxene									[2] X	[1] X		
Unidentified varietal mineral									[2] X			
Magnetite	X	X	Tr.	X	X	X	X	X	X	[3] X	X	
Ilmenite		([4])		[5] (?)								
Sulfides					X	X						
Apatite	X	X	X	X	X	X	X	X	X	X		
Zircon	X	X	X	X	X	X	X	X	X	X	X	
Allanite		X	X	Tr.	X	X	X	X	X	[1] X	X	
Sphene		[5] (?)	Tr.		X					[2] X		
Unidentified accessory mineral		X	X						[2] X	[2] X		
Fluorite				Tr.		Tr.						
Secondary minerals [6]	X	X	X		X		X	X				

[1] Present in two specimens.
[2] Present in one specimen.
[3] Hematite also present.
[4] Some rutilohematite?
[5] Identification of mineral is questionable.
[6] Including one or more of these: chlorite, zoisite, pumpellyite, carbonates—commonly as partial alteration products of the varietal mafic mineral but locally as replacements of feldspars.

No. in table	*Specimen No.*	*Description*
1	J3.13	Phacoidal hornblende granite gneiss, hole 3, depth 287 ft.
2	J1.26	Biotite alaskite gneiss, hole 1, depth 399 ft.
3	--------	Average of two (J2.2, hole 2, depth 68 ft; J3.2, hole 3, depth 68 ft). Range in percent: quartz, 30–32; microcline, 43–38; plagioclase, 22–25; other minerals, 5.
4	J3.3	Contaminated hornblende-biotite alaskite gneiss, quartz poor, hornblende much more abundant than biotite; hole 3, depth 117 ft.
5	J1.19	Contaminated hornblende alaskite gneiss, hole 1, depth 340 ft. Shows folded mafic relics.
6	J2.1	Contaminated hornblende-biotite alaskite gneiss, hole 2, depth 47 ft.
7	--------	Average of two (J1.27, hole 1, depth 406 ft; J3.6, hole 3, depth 232 ft). Range in percent: quartz, 17; microcline, 65–41; plagioclase, 16–30; other minerals, 2–12.
8	--------	Average of six thin sections of four rocks (J1.29, hole 1, depth 437 ft; J2.16, hole 2, depth 230 ft; J3.1, hole 3, depth 68 ft; J3.7, hole 3, depth 243 ft). Range, in percent, for the four rocks: quartz, 12–28; microcline, 35–63; plagioclase, 22–32; other minerals, 4–8. Range in two representative thin sections for same rock sample may be almost as great as range for the group of four rocks.
9	--------	Average of three rocks of intermediate quartz content (J1.18, hole 1, depth 312 ft; J3.5, hole 3, depth 198 ft; J3.15, hole 3, depth 341 ft). Range in percent: quartz, 15–24; microcline, tr.–4; plagioclase, 81–64; other minerals, 2–9. Includes one sample whose mode was determined by point-counter.
10	--------	Average of three rocks poor in quartz (J2.30, hole 2, depth 332 ft; J2.36, hole 2, depth 369 ft; J5.4, hole 5, depth 200 ft). Range in percent: quartz, 8–10; microcline, 1–5; plagioclase, 86–76; other minerals, 4–9.
11	J1.25	Quartz-rich variety, hole 1, depth 388 ft. Mode determined by point-counter.
12	--------	Average of seven specimens, previously listed as numbers 9, 10, and 11.

Whether all the outwardly massive alaskite has been deformed, the writer does not know.

Medium-grained contaminated alaskite gneiss

Medium-grained granite is present in the drill core and may occasionally be seen in outcrops north of holes J6 and J7, where it cuts old phacoidal granite gneiss. From the drilling data, we know that the medium-grained granite forms sheets 10 to 20 feet thick within, or at the borders of, augen granite gneiss. Thicker zones of medium-grained granite interlayered on a small scale with augen granite gneiss are also present. The two rock types appear to grade into each other.

The medium-grained granite is pink to grayish, massive to foliated, locally with a faint augen structure. The texture is medium grained, locally with some finer, granulated material. Quartz, potassic feldspar, hornblende, and (occasionally) biotite are readily recognized in hand specimens. Locally, the granite contains shreds of amphibolite.

Microscopically, these rocks are inequigranular, fine to medium grained; or fine to coarse grained but dominantly nearly equigranular and medium grained. Grain boundaries are commonly polygonal, curved, and sutured, though the coarser rocks contain some interlocking grains. The foliation is usually due to clustering of mafic minerals, whose individuals may or may not be subparallel. One rock has some large elongate but not leafy quartz grains. Most rocks have a moderately granoblastic fabric, but the coarse variety seems only slightly granoblastic.

Approximate modes of three specimens of medium-grained, contaminated alaskite are given in table 19, Nos. 3, 4, and 5. Some of the microcline in these rocks is slightly microperthitic; the rest is a single, homogeneous phase. Partial rims of albite on more calcic plagioclase are fairly common in two specimens; the albitic material is present only where plagioclase is in contact with microcline. The hornblende is a blue-green variety, presumably femaghastingsite or ferrohastingsite. Biotite, where present, is distinctly subordinate to hornblende. Opaque minerals, studied in thin section only, are platy ilmenite(?), with or without magnetite. Apatite and zircon are consistently present, generally with an unidentified zirconlike mineral. Allanite is so sparse it could not be positively identified. A trace of fluorite, rutile, and sphene was found in one specimen apiece.

Rarely, a white medium-grained granite is present as layers and laminae in skarn, against which its contacts may be either sharp or gradational. Sphene is megascopically an abundant accessory, and epidote is present locally. In a crude way, this rock resembles the white border facies of phacoidal granite gneiss occurring against some skarn at the Clifton mine.

Augen granite gneiss

The augen granite gneiss is pink, mottled pink and white, or pink and gray. The augen may be coarse or medium in size (commonly about 2 by 3 cm), roundish, ellipsoidal, or flattened. Feldspars within the augen look generally medium grained or coarse, though some finer granulated material was noted in places. In the relatively small rock column given by 1⅛-inch drill core, the structure of the coarsest rock looks almost massive, yet outcrops of the same rock have an obvious planar structure. Rarely, faint minor folds are visible in the drill core, as well as in outcrops. The coarsest material is apparently gradational into a strongly foliated rock with much smaller, bluntly flattened augen. Locally, these augen disappear almost entirely, and the granite gneiss shows a layered or "banded" structure. Quartz, pink potassic feldspar, white or yellow plagioclase, dark-green or black hornblende, and black biotite can usually be identified, even without a hand lens. In a few places, thin zones of granite show mafic schlieren that make up perhaps 40 percent of the rock. Near some of these schlieren, the augen granite gneiss contains a little garnet.

While the augen granite gneiss is superficially similar in appearance to the phacoidal granite gneiss of the Diana complex, it lacks the highly sheared and strongly granulated character of the true phacoidal granite gneiss that is also present in the Jayville area. Some phases of the augen granite gneiss look much like the Hermon granite gneiss of the Grenville lowlands. Remnant flowage folds are rather common in some areas of the augen granite gneiss; they are entirely lacking in the old phacoidal granite gneiss. Drill cores show no evidence that augen granite gneiss is cut by medium-grained, equigranular granite. In fact, the two seem to grade into each other. Petrographic data, given below, are entirely consistent with the field relations. In particular, the presence of accessory allanite in the augen granite gneiss may be regarded as diagnostic of the younger series of granites; allanite has not been observed in phacoidal granite gneisses of the quartz syenite series. Conceivably, the augen granite gneiss represents old, phacoidal granite gneiss modified by action of younger granite magma or its emanations, but the case for such an interpretation is still to be made.

In spite of the obvious differences among hand specimens of rocks lumped as augen granite gneiss, their microfabrics and gross mineralogy are very similar. All are moderately to highly inequigranular, generally fine to medium grained but locally fine to coarse grained. The coarse grains in the rock are quartz or, less commonly, microcline. Grain boundaries are polygonal, curved, or sutured. A few interlocking grains are present. Foliation is given by faint to marked clustering of mafic minerals and by subparallelism of the individuals; by subparallelism of elongate quartz and microcline grains; and locally by alternation of quartz-rich and feldspar-rich layers. These features may occur separately or in combination. Where large augen are present in the rock, giving the hand specimens a coarse fabric, the thin sections show that the augen are made up of medium-grained, nearly equigranular microcline and plagioclase, accompanied by a little fine-grained quartz. The fabric of all types appears generally to be granoblastic.

Modes of these rocks are given in table 19, Nos. 6, 7, and 8. The range in mineral composition is virtually the same for all types, and mineralogical details are similar to those noted for the medium-grained granites.

Sodic granite

A considerable quantity of sodic granite is present in the Jayville cores. It forms layers and lenses 2 feet to more than 20 feet thick and is always associated with either biotite-plagioclase gneiss or amphibolite. The gneiss and amphibolite may occur as distinct layers or as thin schlieren, and the sodic granite itself varies from a schlieren granite or phantom gneiss to a massive, homogeneous rock. The largest body of sodic granite separates the northwest and middle belts of metasedimentary rocks. Medium- to fine-grained, contaminated pink granite may or may not be intimately associated with sodic granite. Locally, a fine, contaminated hornblende-biotite alaskite is interlayered with sodic granite, the layers being less than a centimeter thick.

In hole J8, an 18-foot layer of sodic granite occurs within pyroxene skarn, close to an 8-foot layer of coarse pink granite. Although amphibolite and biotite gneiss are lacking here, their former presence in the zone now occupied by sodic granite may be inferred from the stratigraphic succession of holes J3 and J5.

Locally, a coarse-grained syenite consisting of plagioclase and amphibole occurs at the contact of pegmatite with amphibolite or amphibole skarn.

The sodic granite is gray, greenish, yellowish, or pinkish, medium or fine grained, foliated or massive. Plagioclase feldspar with conspicuous twinning is abundant in the hand specimens; the quartz content is variable. The principal mafic component is biotite, which makes up about 5 to 10 percent of the rock. A little pyroxene was noted in one place. Except for its abundance of plagioclase, the pink variety of sodic granite is indistinguishable from some pink alaskite.

Microscopically, all the sodic granites are inequigranular, generally fine to medium grained but locally fine to coarse. For most rocks, the felsic groundmass is chiefly medium-grained plagioclase sprinkled with fine quartz blebs. Grain boundaries are generally polygonal and curved, though sutured grains are common in some specimens. Interlocking grains are sparse and are confined to a few specimens. Foliation ranges from nil to rather strong; it is usually due to slight variation in biotite content and to subparallelism of the biotite flakes. These features may be accompanied by one or more of the following: trains of magnetite, subparallel plates of ilmenite(?), and elongate grains of plagioclase.

The sodic granites range from quartz poor to quartz rich. (See table 19, Nos. 9, 10, 11, and 12.) Sodic plagioclase, variable in composition, is the predominant mineral. Microcline is sparse, ranging from a trace to about 5 percent. Commonly it occurs as partial rims and networks about quartz that is included in, or bordered by, plagioclase. A few discrete polygonal grains of microcline may also be present. Much of the microcline seems to have replaced the plagioclase, though the rimlike material might be interpreted as potassic feldspar exsolved from plagioclase and concentrated at the margins of its former host. Only one specimen of sodic granite has plagioclase that is slightly microantiperthitic; the rest is devoid of such intergrowths. Biotite is the usual varietal mineral, though pyroxene and an unidentified mineral have been found in one specimen each. Accessory minerals commonly include magnetite (locally accompanied by crystalline hematite or leucoxene), apatite, zircon, and allanite. Sphene and an unidentified zirconlike mineral are present in one allanite-free, apatite-rich specimen.

The origin of the sodic granite is obscure. At Jayville, sodic granite seems to have been derived mainly from biotite-plagioclase gneiss (or amphibolite) by action of younger granite magma (represented by contaminated alaskitic rocks) or its emanations. Wherever alaskitic rocks are found in contact with biotite-plagioclase gneiss or amphibolite, some sodic granite is also present. The sodic granite may be homogeneous, intervening between the alaskitic rocks and the old country rocks or occurring as thin, isolated lenses and sheets within them; or it may be interleaved, and apparently gradational into, alaskitic rocks—the two forming a compound sheet within the old country rocks. Some sodic granite may have developed mainly by direct reconstitution and replacement of biotite gneiss (or amphibolite), but the relations of some of the rocks suggest that locally small quantities of sodic magma may have developed by action of heat and volatiles on the old country rock, the active agents having been supplied by alaskitic magma. The amphibolite probably contributed nothing in the process except its proximity; amphibolite sheets were merely present in biotite gneiss zones that were partly converted to sodic granite.

Plagioclase in the granitic rocks

The composition of the core plagioclase in all the granitic rocks except the sodic granites is An_{22-24}. The sharply marked partial rims of sodic plagioclase, present only at contacts between plagioclase and microcline, range in composition from An_2 to An_{10}. Some of the plagioclase cores in these rocks show selective sericitization—generally very slight—of alternate lamellae in albite twins, a fact previously figured by Emmons and Gates (1943, pl. 4, figs. 3–6). Other plagioclase cores show very slight, patchy altered areas that extend across a number of the twin lamellae.

The plagioclases in the three varieties of sodic granite are variable and unrimmed. The quartz-rich variety has the most sodic plagioclase—An_{5-10}; the variety having a moderate quartz content has An_{12-17}; the quartz-poor variety has An_{13-15}.

Pegmatite, quartz, and carbonate veins

Veins, patches, and small bodies of pegmatite, at most a few feet thick, occur sporadically in granitic rocks, gneiss, and skarn. Pegmatite in the alaskitic rocks is normal quartz-microcline-plagioclase pegmatite, some of it quartz rich, some of it contaminated with biotite or hornblende. The granite pegmatite locally present in sodic granite is a sodic variety. Pegmatitic areas about a foot long, rich in plagioclase, are sparingly present in some biotite-plagioclase gneiss. Pegmatite veins and patches are present here and there in pyroxene skarn, some of which is a feldspathic variety. This pegmatite, either granitic or syenitic, coarsens the skarn pyroxenes, locally incorporates very coarse pyroxene xenocrysts, and in places develops hornblende from the skarn pyroxenes. Sphene is abundant near some of these pegmatites. Quartz veins, occasionally seen in skarn, have similar relations to their host rock. Quartz veins a few feet thick are found locally in augen granite gneiss. At one place, a narrow sphalerite-pyrite-calcite-fluorite veinlet cuts granite pegmatite. This veinlet is thought to belong to the late hydrothermal suite. Narrow calcite-hematite-pyrite-jasper veinlets, visible here and there in chloritized, hematite-impregnated sköls, are regarded as supergene.

METASEDIMENTARY ROCKS

Biotite-plagioclase gneiss

Biotite-plagioclase gneiss occurs in layers 1 to 25 feet thick at the margins of the skarn masses or enclosed by granite. Occasionally, a layer less than 5 feet thick is present within the skarn. The thicker zones may be parts of a once-continuous bed that has been isoclinally folded, pulled apart, and locally transformed to sodic granite. The rock is gray, medium grained, and speckled. Unlike most rocks of similar composition in the district, it lacks a pronounced planar structure. Biotite and well-twinned plagioclase are obvious constituents, the biotite making up roughly 20 to 25 percent of the rock. Even without a hand lens one can see that the two main constituents have a hit-or-miss arrangement.

Microscopically, the gneiss is medium grained and crystalloblastic, with a moderately good foliation given by slight, irregular compositional layering and locally by subparallel brown biotite flakes. The most striking mineralogic feature is the almost complete lack of quartz. A little microcline is present, replacing plagioclase. A streakier, somewhat granitized variety of biotite gneiss, occurring as partly disintegrated relics in augen granite gneiss, has a fabric very similar to that of the sodic granites. This biotite gneiss contains less biotite, and more quartz and microcline, than the ungranitized biotite-plagioclase gneiss. Allanite, absent from the ungranitized gneiss, is present as an accessory mineral in the streaky variety. Approximate modes of the two varieties are given in table 20, Nos. 1 and 2. Plagioclase has the composition An_{23-27} in the ungranitized variety, and An_{22-25} in the granitized variety.

Amphibolite

Layers of amphibolite 1 to 3 feet thick are present in the granitic rocks that flank the northwest skarn zone. The thickest amphibolite layer (10 ft.) was cut by hole J8, just at the hanging wall of the skarn zone. (See pl. 9.) Locally, thin amphibolite layers are found within the skarn zone, close to its walls.

Megascopically, the amphibolite is commonly a fine-grained, nearly massive, speckled black-and-white, hornblende-plagioclase rock, slightly biotitic, very dark in general appearance, with a barely distinguishable foliation. Locally, the rock is gray, slightly coarser, and better foliated, though the planar structure remains indistinct. In a few places, the rock is coarse

TABLE 20.—*Approximate modes of biotite-plagioclase gneiss and amphibolite, Jayville deposit*

	Biotite-plagioclase gneiss		Amphibolite	
	1	2	3	4
Quartz	Tr.	6		
Microcline	1	12	Tr.	4
Plagioclase	76	72	59	50
Biotite and accessories	23	10	6	4
Hornblende			35	42
Total	100	100	100	100
Composition of plagioclase	An_{23-27}	An_{22-25}		An_{23-24}

No. in table	*Specimen No.*	*Description*
1	J2.3	Unmodified biotite-plagioclase gneiss, hole J2, depth 78 ft.
2	J1.12	Biotite-quartz-plagioclase gneiss, slightly granitized, hole J1, depth 232 ft. In fabric and composition, this rock approaches sodic granite.
3	J5.7	Biotite amphibolite (metagabbro?), hole J5, depth 127 ft.
4	J4.1	Biotite amphibolite (metagabbro?), hole J4, depth 16 ft.

grained and massive, recrystallized by action of the associated granitic material.

The thin amphibolite layers found on both sides of the northwest skarn zone may or may not be equivalent in stratigraphic position. They are commonly associated with biotite-plagioclase gneiss and sodic granite. The writer thinks it reasonable that the upper and lower sequences are equivalent, and that the duplication has resulted from isoclinal folding. In any event, the thin sheets of amphibolite may be useful markers, together with biotite-plagioclase gneiss, for working out the structure of the area, in case that is ever undertaken in detail. However, one should be alert to the possibility that some of the amphibolite bodies may be discordant. (Cf. metadiabase sheets and dikes at Clifton mine, p. 114–116.)

Microscopically, the amphibolite is a fine-grained, roughly equigranular, crystalloblastic rock whose distinct but slightly irregular foliation is given by crude compositional layering and by subparallel biotite flakes. Hornblende may or may not show a preferred orientation. Quartz is absent. Microcline, rare or minor, is slightly microperthitic. The hornblende is a common green variety; the biotite is brown. The plagioclase has the composition An_{23-24}; locally, the grains have slight, curving areas of more sodic plagioclase. Among the accessories, sphene is rare to common; platy ilmenite is present; a little pyrite is ubiquitous; zircon is rare. Approximate modes of two specimens are given in table 20, Nos. 3 and 4.

Skarn

Skarn makes up the bulk of the metasedimentary rocks, of which two main belts are thought to be present. (See pl. 9 and subsequent account of structure.) The following description of skarns applies to those in the northern belt. The southern belt of metasedimentary rocks, where drilled, is only a few feet thick; its skarn is partly recrystallized by the action of granite pegmatite, and the rest consists of ferrohastingsite-fluorite skarn of a type common in the northern belt.

The skarns of the northern belt are highly varied, generally dark, and—so far as we know—unaccompanied by marble layers or large, conspicuous patches of relict carbonate minerals. Pyroxene is the principal skarn mineral, in places accompanied by a dark, coarse clinoamphibole. Locally, discrete bodies of clinoamphibole skarn and orthoamphibole (anthophyllite) skarn are present. Thick mica sköls are prominent, commonly at the borders of the skarn zone. These sköls contain much of the ore. Some small areas of clinoamphibole skarn have been formed from pyroxene by the direct action of quartz-rich pegmatites, but most of the amphibole- and biotite-rich skarns evolved from pyroxene skarn by progressive introduction of Fe, Na, K, OH, F, and Cl in the manner that is normal in the district but at Jayville is highly intensified. The broader features of the evolution of the Jayville skarns are well established by data from mineral associations, paragenetic sequence, and optical properties of skarn minerals corresponding to those of analyzed minerals found elsewhere in the district. However, a detailed study of the mineral chemistry of the Jayville skarns has not been completed, though it is essential to our knowledge of skarn evolution.

Pyroxene skarn and its varieties.—Most of the pyroxene skarn is a medium-grained, medium- to dark-green salitic to ferrosalitic variety. Light green diopside skarn, almost salitic, is a subordinate variety. Locally, the latter rock contains phlogopite, actinolite, interstitial calcite, and (rarely) graphite. Dark andradite skarn, slightly ferrosalitic, forms several layers more than five feet thick in the zone cut by hole J1. This variety is rare for the skarn zone as a whole, though a little andradite occurs sporadically in several zones of ferrosalite skarn. The pyroxene skarns, including the garnet-rich variety, are scapolitic, plagioclase bearing, or quartzose in places. These modified skarns usually contain abundant accessory sphene. Accessory epidote and allanite are present locally in assorted dark skarns; fluorite is rare. The above varieties of skarn are identical with those forming most of the skarn deposits in the district. Varieties transitional to amphibole- and mica-rich rocks are described below.

Amphibole skarn and its varieties.—Dark skarn having clinoamphibole as a major constituent is distributed erratically through the major skarn bodies in many discontinuous bodies a few inches to a few feet thick. Five feet is probably a maximum thickness for any single body rich in hornblende.

The term "clinoamphibole" is so cumbersome that "hornblende" will generally be used in the following account, in spite of the fact that not all the clinoamphibole is hornblende. Where precision is required, the general term or a varietal name will be used.

Hornblende-bearing skarns are mostly very dark green or black, and somewhat coarser than the neighboring pyroxene skarn; they may be foliated, with or without a good lineation; or their hornblende may have a hit-or-miss arrangement. The distribution of the hornblendic zones shows no direct correlation with the larger masses of sodic granite or with any other principal rock units in the area, excluding the skarn itself. None of the skarn consists exclusively of hornblende; pyroxene, biotite, or fluorite is always a major constituent. With regard to mafic components alone, the following varieties occur: hornblende-pyroxene skarn, hornblende skarn, hornblende-biotite sköl. A very little ferrosalite is present, though rare, in skarns that contain both hornblende and biotite. The existence of the skarn varieties just noted, and the absence of certain other varieties, is determined by the quantity and probable order of introduction of iron, alkalis, and volatiles, and by the oxidation state of the iron. (See p. 32.)

Diopside-actinolite-phlogopite skarn, a light skarn poor in iron and relatively poor in alkalis, is an apparent but chemically rational exception to the rule.

Two distinct varieties of "hornblende" are present, each with its own paragenesis. The two amphiboles are not known to occur together. The first variety is a ferrohastingsite (cf. table 2, No. 13), or blue-green amphibole. The second is a blue amphibole, some of which could be termed pargasite. Diagnostic optical properties of the two amphiboles are given in table 21. The two amphiboles are indistinguishable in hand specimen, but color alone usually suffices to tell them apart in thin section. The most convincing difference, readily checked by immersion, is the refractive index nX. Skarn hornblendes whose properties differ markedly from those listed in table 21 are sparse at Jayville.

Ferrohastingsite skarns generally contain some fluorite, and many of them are fluorite rich, containing 20 to 60 percent fluorite. Those poor in fluorite are rich in ferrosalite. Scapolite is present in some ferrohastingsite-ferrosalite skarn. Skarns with blue-green

TABLE 21.—*Diagnostic optical properties of dark clinoamphiboles in skarn, Jayville deposit*

	Ferrohastingsite	Blue amphibole
Z	Dark greenish blue or blue-green	Deep blue.
$2V_x$	30°–40°; locally large	80°–85°; locally 97°.
nX	1.695±0.005	1.652±0.002.
$Z \wedge c$	13°–20°	22°–26°.

amphibole are barren or only weakly mineralized with magnetite.

Skarns containing blue amphibole are either barren or ore bearing. Fluorite is absent or very sparse. Most of these skarns contain biotite. Some of them contain anthophyllite as well. Clinopyroxene is sparse or absent. Quartz is present in some specimens. A single specimen of pale-green diopside-phlogopite skarn showed some blue amphibole.

Accessory minerals in the two varieties of hornblende-bearing skarn are inconstant and not diagnostic. Allanite and magnetite are commonly present; in places, the magnetite is accompanied by hematite. Other accessories, all sporadic and generally very sparse, include quartz, apatite, zircon, monazite, sphene, one or more unidentified minerals of high relief, and an unidentified platy or prismatic opaque mineral (not graphite or ilmenite).

Anthophyllite skarn is present in three or more different zones, each at most about 15 feet thick, within the skarn. The anthophyllite skarn may be megascopically pure, moderately biotitic, or intricately intergrown with biotite in sköls. The "pure" layers are enclosed in dark pyroxene skarn. The anthophyllite skarn is barren or only weakly mineralized, though some of it occurs within a few feet of ore.

Megascopically, the "pure" anthophyllite skarn is gray green, prismatic, almost twice as coarse as the neighboring pyroxene skarn, and locally lineated. The anthophyllite prisms, commonly about 0.5 by 1.0 or 1.5 cm in size, looks fibrous in places, not because individual fibers are present but because the prismatic cleavage is closely spaced and remarkably sharp in appearance. Anthophyllite skarn interwoven with mica in sköls looks black and pitchy to vitreous against the silvery, somewhat silky mica. Anthophyllite prisms having dimensions similar to those just cited show prominent prismatic cleavage and basal parting. The rock as a whole has a very crude, wavy planar structure but no lineation, except one locally given by overlapping mica flakes.

Microscopically, the "purest" anthophyllite skarn is seen to consist of anthophyllite and less than one percent magnetite. Other anthophyllite skarns contain 20 percent biotite. Dark-green iron-bearing spinel (pleonaste) forms 6 percent of one specimen. Accessory minerals include apatite, zircon, and allanite(?). In addition, one or more unidentified accessories are present. The biotitic anthophyllite skarns may or may not be slightly mineralized. Anthophyllite skarns transitional to sköls and occurring therein are roughly half anthophyllite and half biotite. One unusual variety is supplemented with plagioclase (An_{30}, about 20 percent) and greenish blue clinoamphibole (about 2 percent). The accessories are much the same as those found in moderately biotitic anthophyllite skarn.

The anthophyllite, slightly variable in composition from rock to rock, is an aluminous, ferroan variety (cf. table 2, No. 11). The biotite is commonly pale green to colorless in thin section, comparable to that shown in table 1, No. 10.

Mica sköls.—Mica sköls, 5 to 30 feet thick, occur sporadically in the skarn. Those in the northwest belt are generally thicker and more abundant where the skarn zone thickens, at the latitude of holes J1 and J2, near the nose of an inferred syncline. (See pl. 9.) Here, sköls are prominent at and near the walls of the skarn zone, though micaceous layers are not confined to the margins of the skarn. Megascopically, the sköls at Jayville range in color from silvery or very pale green to dark greenish black. In part, they are well foliated; in part, they have a random fabric. Parts of them are fluorite rich, and parts are ore bearing or appreciably mineralized with magnetite or other metallic minerals. Structurally and mineralogically the sköls grade into several types of skarn, as previously noted. For an account of the microscopic features of some sköls, see description in table 1, No. 10, and Leonard and Vlisidis (1961, p. 792–796).

Though dark sköls predominate, there seems to be a gradual change in the composition of their mica from phlogopite to biotite. This is indicated by the nearly straight-line plot of nY, whose values range from 1.577 to 1.614 on micas from a dozen representative specimens of sköls and highly micaceous skarns. Cleavage flakes of the phlogopites are pale yellow, tan, or pale green in oils. Cleavage flakes of all biotites but one are pale green. One biotite is brownish.

STRUCTURE

The gross structure of the area is that of an overturned, compound, isoclinal syncline of metasedimentary and younger granitic rocks. The attitude of the axial plane of this structure must be about N. 30° E., 60° NW. The axial line trends about N. 30° E. The axis of the structure is skewed to the axial line. Detailed knowledge of the compound synclinal structure is important, for there is a strong likelihood that the various magnetite-bearing bodies are plunging lenticular shoots whose long dimensions are parallel to the axes of folds of intermediate size.

Two roughly parallel belts of metasedimentary rocks outline the main synclinal structure. The outer or southern belt, presumed to be stratigraphically lower, contains the Fuller No. 2 and No. 3 workings and perhaps the Hart No. 2 shaft as well. The inner or northern belt contains the other workings shown on plate 9. Conceivably, the two belts are parts of a single belt, though they are not demonstrably connected. A third belt of metasedimentary rocks, not shown separately on plate 9, may be present southeast of the collar of hole J8. The following description is based on the simpler interpretation that just two belts are present, a northern and a southern one.

The southern belt is thickest near the Fuller No. 2 and No. 3 workings, at the keel of the syncline. Toward the east-northeast, this belt is thinned and disrupted, locally consisting of skarn boudins and relics of folds strung out in the large granite outcrop northwest of Hart No. 2 shaft. The inferred V-shaped body of metasedimentary rocks between New York No. 1 and Hart No. 2 shafts is pictorial and imaginary, shown on plate 9 only to suggest a possible link between the southern belt and the isolated magnetite-bearing skarn pod opened by Hart No. 2 shaft. Northwestward from the Fuller No. 2 shaft, the southern belt thins rapidly. It may pinch out completely, or it may be represented by thin layers of amphibolite and mestasedimentary rocks cut from 70 to 160 feet in hole J2, or by rocks underlying the isolated magnetic high west of the mapped area.

The northern or inner belt comprises three elongate, locally bulging masses of metasedimentary rocks, presumably connected at the latitude of Benson No. 2 shaft by irregular areas of skarn, thought to represent the thickened keels of small isoclinal synclines. The individual masses are discussed below, from northwest to southeast.

The position and shape of the *northwest* belt, which contains the main ore body, are established by outcrops and drill holes. The position and shape of the *middle* belt are uncertain. Rocks assigned to this belt crop out in one place west of New York No. 2 shaft and are exposed in a small pit near Little Bear Creek. At both places, the downward extension of the metasedimentary rocks has been confirmed by drilling. However, no connection between the two areas has been demonstrated; they might be parts of two separate masses. The position and general shape of the *southeast* belt are established by outcrops, and in the neighborhood of New York No. 1 shaft the downward extension of the mass has been confirmed by mining and drilling. The shape of the irregular connection between the middle and southeast belts is indeterminate; indeed, such a connection may not exist.

The interpretation of the inner belt as a compound syncline is consistent with regional and local structural features. However, details of the structure are unknown. For example, the northwest, middle, and southeast belts may be a single stratigraphic unit. On the other hand, some thin layers of biotite-plagioclase gneiss and amphibolite, found locally above the skarn layer in the northwest belt, seem to be repeated below the skarn. (See pl. 9.) If the repetition is a fact and is due to isoclinal folding, the metasedimentary rocks of the northwest belt may have been isoclinally doubled and subsequently refolded. If the other two masses of metasedimentary rocks have also been doubled, the inner belt would represent a fivefold repetition of a single stratigraphic unit.

The attitude of the axes of the folds of intermediate size (limbs roughly a hundred feet apart) is, for purposes of exploration, the critical structural problem. Several possible attitudes for these fold axes may be inferred from the fragmentary data (1, 2) at hand.

(1) Measured axes of minor folds in the outer belt of metasedimentary rocks are oriented N. 50° E., 55° NE., and N., 50° N. (See pl. 9, outcrops northeast of Fuller No. 3 tunnel and northwest of Hart No. 2 shaft.)

(2) Axes constructed from the attitude of foliation at what one subjectively takes to be corresponding points on the opposite limbs of folds belong to two different sets. Axes of the first set trend and plunge about N. 10° W., 40° N. Their attitude is essentially parallel to that of lineations, including a major synclinal axis, mapped in the area southeast of Jayville. (See pl. 1.) Axes of the second set trend and plunge about S. 80° W., 55° W. These may parallel the axes of poorly defined "cross folds" suspected to disturb the major north- and northeast-trending folds of the region.

Determining the attitude of the axes of larger folds at the Jayville deposit is further complicated by (a) the steep southwest dip recorded by Tyler and Wilcox southwest of Benson No. 2 shaft and by (b) the lineation they measured at Fuller No. 2 shaft. The first feature suggests that, even if the axes of larger folds have a fairly regular trend, their plunge may change abruptly. (Cf. Clifton mine.) The second feature is best ignored for the moment, as hole J4 demonstrated that the ore body exploited by the Fuller No. 2 workings does not continue downward parallel to that observed lineation.

The writer regards an attitude of N. 0°–10° W., 40°–55° N., as most probable for the axes of synclines outlined by the belts of metasedimentary rocks. This direction is followed by some minor folds of the area; it parallels the axis of the next major syncline southeast of Jayville; it is essentially the attitude, in longitudinal vertical projection, of the sloping contacts between skarn and enclosing rocks in holes J1, J2, and

J3; it is consistent with Platt's observation in 1881[37] regarding the habit of ore bodies at Jayville (see section on occurrence of ore); and it provides one explanation for the failure of holes J8 and J4 to cut ore.

Faults of unknown but probably slight displacement may be present in the area. Tyler and Wilcox mapped a small shear zone southeast of Benson No. 1 shaft. Some cores and dump specimens show black chloritic slickensided zones suggestive of faults. Some of these zones cut the foliation; others are essentially parallel to it and may have resulted from interlaminar slippage during folding.

ORE

OCCURRENCE

Ore occurs in biotite sköl, in ferrosalite-ferrohastingsite skarn, and in quartz-bearing ferrohastingsite skarn. Some of the biotite sköl and quartz-bearing amphibole skarn is rich in fluorite. Scattered thin zones containing some magnetite are common in salite and ferrosalite skarn, and other dark skarns are mineralized locally. The ore is generally streaky or massive, though disseminated material grading into streaks or well defined thin layers is present in places. Most specimens of ore, regardless of habit, show some planar arrangement of the ore minerals.

The magnetite-bearing bodies are mostly at or near the walls of the main skarn zone. (See pl. 9.) The principal magnetite-bearing body occurs directly against, or within a few feet of, the hanging wall of the northwest skarn zone. A zone of subore or mineralized rock, or a very thin layer of medium-grade "ore," is present at the footwall of one or more skarn zones. In addition, hole J2 cut a thin layer of "ore" about 30 feet below the hanging wall of the skarn zone, just above a layer of gray sodic granite. The walls of the main skarn zone are generally granite, but the footwall at the latitude of hole J1 is biotite-plagioclase gneiss.

In gross habit, the ore bodies at Jayville were, according to Platt, elongate shoots whose trend was roughly parallel to the strike of the wall rocks. Tyler and Wilcox regard the ore body in the northwest belt (developed by Fuller No. 1 and Benson No. 1 openings; see pl. 9) as a short lens. They place a similar interpretation on the ore bodies in the southeastern and southern belts. The results of exploration by the Jones & Laughlin Steel Corp. are consistent with an interpretation, based on meager structural data and for at least part of the area confirmed by Platt's report, that all the ore bodies were northward-plunging lenticular shoots. The pattern of magnetic anomalies (pl. 9) also accords with the interpretation that the ore bodies plunge northward. For the northwestern (main) ore body, results of drilling indicate that the inferred plunge must be somewhere between 30° and 60° N. The plunge of the ore body or bodies in the southeastern belt is unknown but probably northward. Only one hole (J1) was drilled deep enough to test the downward extension of ore in this belt. That hole cut skarn in the appropriate position, but the skarn was barren. The plunge of the ore body developed by the Fuller No. 2 and No. 3 workings is not northwest, for hole J4 cut no ore and only a very thin zone of skarn. However, the ore body might plunge northward or northeastward, and such is our interpretation.

Hole J4 was drilled on the assumption that the ore body worked from the Fuller No. 2 openings might be a shoot that plunged about N. 20° W., 50° NW., parallel to a lineation visible in the wallrock exposed in the northern opening. If the ore body were elongate parallel to the axes of the local folds of intermediate size, it would trend about N. 0°–10° W. and plunge about 50° N. According to this interpretation, hole J4 did not adequately test the possible extension in depth of the ore body worked from the Fuller No. 2 and No. 3 openings.

MINERALOGY

The mineralogy of the host rocks differs little from that of the unmineralized equivalents. However, quartz is extremely rare in any but ore-bearing skarns. Hypersthene is abundant in one specimen of ore-bearing sköl. A few ore-bearing rocks contain chlorite, in part as small rosettes. Accessory minerals include one or more unidentified species; otherwise, they represent the same species, roughly in the same abundance, present in unmineralized skarns and sköls from this deposit.

The ore minerals include magnetite, vonsenite (ferrous ferric borate), and hematite. Of these, the first two are regarded as hypogene and the last as supergene. Sulfides are very sparse in the deposit. One ore specimen contained a little pyrite; another showed a trace of chalcopyrite. An unidentified sulfur-bearing mineral is sparingly present in one specimen of supergene hematite ore. A very little goethite, presumably supergene, is present in one specimen. Tin has been found spectrographically as a minor element in the Jayville vonsenite, but no minerals having Sn as a major element have been recognized. According to old analyses, Mn is relatively high in ore from the Benson No. 1 shaft. (See table 23.) No discrete manganese minerals have been recognized by the writer. The Mn may be present in dark skarn silicates.

[37] Platt, Franklin, 1881 [letter to Hon. D. J. Morrell concerning exploration of the Clifton deposit, with references to the Green Farm and Jayville deposits]: typescript in files of J. G. Broughton, New York State Geologist.

The magnetite so far examined is devoid of exsolved ilmenite blades. Some specimens show many minute spinel "gashes" in some of the magnetite. Martite has been observed only in a specimen of supergene hematite ore.

The vonsenite has been analyzed and fully described elsewhere (Leonard and Vlisidis, 1961). This borate contains about 62 percent Fe and 13.4 percent B_2O_3, after deducting for impurities in the analyzed sample. The vonsenite is no mere mineralogic curiosity in this deposit: out of seven representative specimens of ore, one had 25 volume percent vonsenite, no magnetite, about 0.5 percent goethite, and a trace of chalcopyrite; another had 35 percent vonsenite, 27 percent magnetite, and less than 1 percent hematite; and the third had 58 percent magnetite, 1 percent vonsenite, and 1 percent pyrite.

Tyler and Wilcox reported that, after magnetic separation, the nonmagnetic tails of the composite core sample from hole J1, 293.5–311 feet, ran 38.61 percent soluble Fe. [Elsewhere in their report, the figure 38.16 is given.] They attributed this remarkably high iron content to hematite. However, the hematite noted by the writer in logging this interval of drill core was very small in quantity and restricted to the upper few feet of the ore zone. Leonard and Vlisidis (1961, p. 791) later found that a duplicate sample of crude ore from this interval contained 4.80 percent B_2O_3, presumably in the form of vonsenite.

After vonsenite was identified in the core, the writer had no opportunity to collect additional core samples. Moreover, the soluble Fe and B_2O_3 content of nonmagnetic tails has been determined only for the composite sample cited. Thus a reliable estimate of the quantity of vonsenite present in the cores, let alone the deposit, cannot be made at this time. Vonsenite is now known to be a major metallic constituent of two out of seven representative samples of ore. One of these samples is from a B_2O_3-rich zone. This suggests that vonsenite may be quantitatively important in the Jayville deposit. In case the core is reexamined for vonsenite, the notes on simple field tests for that mineral may be helpful (p. 60).

Medium- to high-grade crystalline hematite ore, thought to be supergene, is restricted to the sköls and to their immediate wallrocks. A little earthy hematite, and narrow calcite-jasper-hematite veinlets, have been seen in dump specimens. A polished section of typical medium-grade hematite ore in biotite sköl contains hematite, an unidentified metallic mineral—very sparse, and a trace of magnetite. The magnetite occurs as one or two relict grains, largely replaced by hematite. The hematite occurs as (a) wavy or bladed, locally plumose, generally foliated aggregates of micron-sized individual grains; (b) polygonal "grains," continuous with the above, also consisting of minute individuals; (c) single veinlets and fine networks connecting any of the above forms; and (d) polygonal to irregular "grains" containing colloform and spherulitic aggregates of minute individuals. The unidentified metallic mineral, very sparingly present, occurs as patches in hematite and as replacement veinlets and fracture fillings in both hematite and gangue. The mineral is magnetic, contains sulfur, and resembles pyrrhotite in appearance, but its hardness is comparable to that of hematite. It does not look like marcasite. If the hematite host is supergene, the unknown must also be. It may be the low-temperature polymorph of pyrrhotite. The microscopic features of hematite in this specimen are unlike those of primary crystalline hematite found in the district, but they are similar to or identical with those of crystalline hematite from the sub-Potsdam supergene hematite deposits.

GRADE AND EXTENT

The grade and thickness of the main magnetite-bearing zones and a few weakly mineralized zones, as found in drill cores, are listed in table 22. Analyses of ore from the old workings are given in table 23.

Holes J1, J2, and J3 cut the ore zone exploited by the Benson No. 1 and Fuller No. 1 openings, in the northwest belt of skarn. Where drilled, this zone has a true thickness of 6.5 to 15 feet. Hole J5 intersected only 5 feet of sparsely mineralized skarn in this belt, and hole J8 cut a 25-foot thickness of barren skarn. The results of drilling indicate that the ore zone exploited by the Benson No. 1 and Fuller No. 1 openings has a maximum horizontal length of roughly 300 feet, a maximum thickness of 15 feet, and a grade of 43 to 46 percent soluble Fe. Phosphorus and sulfur are low. The height of the lens, assuming it to be a shoot plunging northward, may be roughly 150 to 300 feet; its maximum height probably does not exceed 450 feet. Tyler and Wilcox noted that the ore encountered in drilling agrees very closely in grade with the average ore shipped from nearby mine workings. Vonsenite, present in some cores from this zone, accounts for part of the iron.

The middle belt of skarn was tested by holes J1, J2, and J8. None of the skarn was mineralized. The core interval 246–263.5 feet, hole J1, showed marginal ore and subore. (Cf. table 22.) This presumably represents the downward extension of the irregular skarn

TABLE 22.—*Grade and thickness of magnetite-bearing zones assayed by Jones & Laughlin Steel Corp., Jayville deposit*

[Unpublished data from Tyler and Wilcox, 1942; some figures have been averaged by Leonard; published by permission]

Hole	Footage	Apparent thickness (feet)	True thickness (feet)	Weight percent			
				Soluble Fe [1]	Magnetic Fe	P	S
J1	113 –115	2	2	13.36			
	*115 –121.5	6.5	6.5	42.87			
	130 –165	35	34	4.6			
J2	246 –252	6	6	20.2			
	252 –263.5	11.5	11	11.1			
	[2] 293.5–311*	17.5	14.5	45.73	32.05	0.013	0.110
	311 –331.5	20.5	17	6.4			
J3	342 –362.5	20.5	17	.9–2.0			
	245 –248	3	2.5	21.14			
	[3] 248 –266*	18	15	43.80			
	266 –268	2	2	23.06			
	268 –270	2	2	5.21			
	333.5–335	1.5	1.5	34.01			
	335 –337	2	2	14.86			

[1] Assays are stated merely as "—% Fe," but S. A. Tyler (1956, written communication) states that the assays are indeed for soluble Fe, not for total Fe.
[2] Ore in this interval is known to contain vonsenite. The magnetic fraction was 70.09 weight percent of the composite sample, and the nonmagnetic fraction assayed 38.61 [38.16?] percent soluble Fe. Leonard and Vlisidis (1961, p. 791) report the B_2O_3 content of crude ore, magnetic fraction, and nonmagnetic fraction as 4.80, 0.32, and 5.66 weight percent, respectively.
[3] Vonsenite is known to be present in this zone.
*Main ore zone.

body that centers 120 feet northwest of the New York No. 1 shaft.

The southeastern skarn belt was tested only by hole J1, which encountered a 30-foot skarn zone devoid of magnetite. One may interpret hole J8 as having bottomed a little northwest of the downward extension of the skarn, or as having passed through the appropriate horizontal interval but over the top of a north-plunging skarn mass.

The unfavorable results of exploration at the northeast and southwest ends of the mapped area have previously been noted.

The ore bodies formerly worked appear to be small, thin lenses of medium grade. Though the writer thinks the possible extension of ore down the inferred plunge has not been adequately tested, the results of exploration by the Jones & Laughlin Steel Corp. show that the bodies are, in several places, so close to the limit of minable thickness that they are not now of economic interest as sources of iron ore, no matter what their pitch length.

If the vonsenite at Jayville should ever be required as a local and supplementary source of B_2O_3, or as a special flux for iron ore, the deposit and its environs might warrant additional exploration. In that event, the following features deserve consideration:

1. The trend of known magnetite bodies, some of them vonsenite bearing, is probably northward, but the plunge may be variable.
2. The pitch length of the magnetite bodies is still unknown.
3. Additional magnetite bodies, not exposed at the surface, may conceivably occur within the mineralized zones.
4. Though the known vonsenite is intimately associated with magnetite or occurs within magnetite-bearing zones, it is conceivable that vonsenite bodies containing little or no magnetite are present in the skarn. Such bodies would presumably not yield a strong positive magnetic anomaly.

LAKE MARIAN (RAMPART MOUNTAIN) PROSPECTS

LOCATION, ACCESS, AND HISTORY

A group of small magnetite prospects centers 0.3 mile north of the camps on the north shore of Lake Marian, west-central rectangle, Tupper Lake quad-

TABLE 23.—*Chemical analyses (weight percent) of ore from the workings, Jayville deposit*

[Data from records of the Magnetic Iron Ore Company; cited by Tyler and Wilcox, 1942; published by permission]

Source of shipment	Date of shipment	Fe	P	S	TiO_2	MnO	SiO_2	Al_2O_3	CaO	MgO
Hart No. 1 shaft	12/8/1886	60.61	0.019	0.946	None	[1] 0.74	9.17	0.79	3.04	2.08
	4/1887–11/1887	49.93–58.03	0.020–0.030	0.75–0.98						
New York No. 1 shaft	12/9/1886	54.10	0.006	0.612	None	0.50	14.42	1.52	4.69	3.72
	4/1887–10/1887	41.96–53.50	0.001–0.018	0.12–0.45						
New York No. 2 shaft, selected sample	7/22/1886	38.72	[2] 27.23	0.012						
Other selected samples	1886	22.61–33.59								
Benson No. 1 shaft	12/8/1886	46.31	0.002	0.251	None	2.52	20.70	2.14	5.71	1.58
	2/1887–12/1887	41.72–49.66	0.007–0.029	0.16–0.27						
	1/1888	47.19	0.018	0.178	None	1.25	16.52	2.30	5.38	3.07
	5/30/1888	51.93	0.024	0.147	None	2.14	12.65	1.83	3.56	3.82
	6/1888–10/1888	47.60–51.45								
Benson No. 2 shaft	2/1888–10/1888	42.00–49.31								
Fuller No. 1 shaft	8/4/1886	51.49		0.005			16.96			
	1/1/1887	45.91	0.017	0.015	0.60	0.40	17.50	2.24	5.80	4.39
Fuller No. 2 shaft	12/8/1886	53.49	0.007	0.097	None	0.99	12.86	3.79	5.91	2.37
	1/1/1887	59.25	0.015	0.010	0.25	0.60	8.30	1.56	3.40	1.58
	2/28/1887	51.80	0.005	0.061			14.73			
	4/1887–11/1887	41.35–48.58	0.009–0.022	0.01–0.16						

[1] Mn.
[2] Reported as value for P; error for SiO_2?

rangle. The test pits, on the Paul Moore estate, were put down in the early years of the Depression, possibly about 1931. The magnetite-bearing zones tested by these pits proved to be too thin to warrant development.

To reach the area, drive to the American Legion camp at Horseshoe, central rectangle, Tupper Lake quadrangle. Get a key to the gate of the Moore estate, and drive west to the end of the road at Lake Marian. Walk north about 2,500 feet along the tote road to the saddle between Rampart Mountain and the ridge to the east. One prospect pit (area 1) is in the saddle. The main group of pits (area 2) is about 700 feet east. To reach the third area tested (area 3), follow the south side of the eastern ridge about 2,000 feet southeastward from the main group of pits, or start at the camps on Lake Marian and walk 500 feet northeast, up the steep slope.

GENERAL GEOLOGIC FEATURES

Geologically, the prospects are near the eastern spur of the Darning Needle Pond syncline, in the southeastward extension of the Sucker Brook belt of Grenville rocks. (See pl. 1.) Here, metasedimentary rocks of the belt comprise, from the bottom upward: (a) quartzite, migmatitic pyroxene gneiss, and amphibolite migmatite, with subordinate granite and pegmatite—thickness probably less than 100 feet; (b) pyroxene skarn, locally magnetite bearing, with subordinate granite and pegmatite—thickness probably less than 50 feet; (c) quartzite with subordinate garnet-biotite gneiss, granite, and pegmatite—thickness several hundred feet. This sequence is overlain successively by (d) alaskite—thickness several hundred feet; (e) alaskite with abundant schlieren of amphibolite, locally magnetite bearing—thickness perhaps 100 to 200 feet; and finally by the somewhat contaminated alaskite (f) that underlies Rampart Mountain. The pits at area 2 are in skarn of unit (b); the pit at area 1 is in unit (e). This subdivision of rock units is based solely on brief reconnaissance.

Foliation in this neighborhood generally strikes northwest and dips 15°–25° SW. · (See pl. 1.) However, the average trend of the foliation is interrupted by a number of minor synclines and anticlines whose axes trend S. 40°–45° W. These minor folds, ranging from fairly open to rather tight, have amplitudes of a few feet to about 100 feet. The magnetite at area 2 is localized in one of the small, southwest-plunging synclines.

With regard to larger structural features, the pits at areas 1 and 2 are about on a synclinal axis of higher order that passes through Panther Pond, while the sulfide-bearing skarn at area 3 is approximately on an anticlinal axis that passes between Panther Pond and Triangle Pond. (See pattern of foliation north of Lake Marian, pl. 1.)

THE PROSPECTS

The pit at area 1 shows thin layers of low-grade magnetite "ore" in contaminated granite gneiss (alaskite containing amphibolite layers and schlieren). A few pieces of high-grade magnetite "ore" are piled on the small dump.

The pits at area 2 show a 15-inch zone of medium- to high-grade magnetite "ore" in pyroxene skarn. The magnetite occurs in layers, small masses, and grains, and in vertical veinlets that transect the dominant planar structure of the rock. The skarn, of which about 15 feet is exposed at the pits, is dark green to black; locally it contains a little quartz, feldspar, and sulfides. The magnetite-bearing zone is apparently in the trough of a minor syncline. As a dip needle was not available when these pits were examined, the zone could not be traced out.

The pit at area 3 shows sporadic sulfides, but no magnetite, in dark pyroxene skarn.

According to old newspaper accounts, the area about Lake Marian was once prospected for copper, but copper minerals were not observed in the pits described above.

MAGNETIC ANOMALIES IN THE AREA

The isointensity map based on the aeromagnetic survey of 1945 showed a magnetic peak centered over Rampart Mountain. Reconnaissance dip-needle traverses were subsequently run to locate the "ground anomaly." No anomalous dip-needle readings were recorded on Rampart Mountain, but a strong, narrow dip-needle anomaly was found in the valley northeast of Rampart Mountain, in the Sucker Brook belt of metasedimentary rocks. The negative anomaly was traced northwestward from area 1 (described above) for a distance of 2,000 feet. Intense positive readings (+75°) were found at the pit. The peak value on a traverse 600 feet northwest of the pit was +35°. This positive anomaly apparently ceases within a few hundred feet to the northwest. A weak negative anomaly locally flanks the positive anomaly on the northeast. The strong negative anomaly (readings ranging from −25° to −37°) that lies 200 to 400 feet southwest of the positive anomaly continues northwestward for at least 1,400 feet beyond the northwest end of the positive anomaly that was traced from the test pit at area 1. This negative anomaly is locally flanked on the southwest by a weak positive anomaly (+3° to +10°); elsewhere it is simply a trough in the prevailing magnetic datum.

Another aeromagnetic peak was registered over Iron Pond and the area just south of the pond. Reconnaissance dip-needle traverses showed half a dozen small, discontinuous "hot spots" south and southeast of Iron Pond. The source of these anomalies is not generally exposed. In one place, low positive readings were recorded over ribbed garnetiferous migmatite whose thin pegmatite lenses contained about 15 percent magnetite. Elsewhere, the presence of skarn float suggests that very minor concentrations of magnetite in skarn may be present in the area. In any event, the anomalies have no commercial significance.

Dip-needle readings taken on boat traverses along the center line of Iron Pond were neutral.

OUTAFIT DEPOSIT

SUMMARY

The Outafit deposit is the first new magnetite ore body found in the Western Hemisphere by means of the airborne magnetometer. The deposit, of skarn type, has been explored to an average depth of 200 feet by two diamond-drill holes. The ore zone, 65 feet thick, averages 28.9 percent magnetic Fe. Drilling tested only the most favorable segment of the anomaly, two of whose segments remain to be explored. One outcrop of wallrock has been found in the area, the rest of which is mantled with glacial deposits and alluvium having a thickness (where drilled) of 17 to 55 feet. The ore-bearing skarn zone is flanked by microcline granite gneiss that carries thin layers of metasedimentary gneisses. From the single outcrop and the information from the drill cores, it appears that the skarn zone may outline a series of nearly isoclinal folds whose axial planes are essentially vertical and whose axes plunge about 40° southwestward. If this structural interpretation is correct, the magnetite bodies indicated by strong positive anomalies lie on the east limbs of the synclines. The quantity of magnetite on the noses of folds and on the west limbs of the synclines is unknown. The drilled ore body may be a thick shoot whose long axis parallels the axes of the plunging folds.

LOCATION AND ACCESS

The Outafit deposit, owned by the Newton Falls Paper Mill, is in the southwest corner of the northeast quarter of Township 4 (Harewood) of Great Tract No. 2 of Macomb's Purchase, in the Town of Clifton (Reed and Cohen, 1947, p. 8), south-central rectangle, Stark quadrangle. The center of the deposit is 900 feet east of the camp of the Out-a-fit Club, on the east side of the South Branch of the Grass River, 3.8 miles north-northwest of Cranberry Lake village.

To reach the deposit, turn north on the dirt road that begins 4.0 miles east of Cranberry Lake post office, just east of the abandoned right of way of the Grasse River Railroad. Drive 2.9 miles on dirt road, taking left forks, to grade of abandoned lumber railroad. Drive straight ahead 2.0 miles to camp of Out-a-fit Club. In 1946, one could drive all the way to the camp. In 1952, the trip was still possible by jeep (Ross, A. M., 1952, written communication).

DISCOVERY AND EXPLORATION

The Outafit anomaly was discovered by the aeromagnetic survey of 1945. The anomaly registered as a prominent peak of 7,180 gammas (Balsley, Buddington, and others, 1954b). Hawkes made a reconnaissance dip-needle survey of the anomaly in 1945 (Hawkes and Balsley, 1946, p. 6 and plate 6). In 1946, the U.S. Bureau of Mines made a detailed dip-needle survey of the anomaly and drilled 2 diamond-drill holes. The results of this work have been summarized by Reed and Cohen (1947, p. 8, 31–32). The Outafit deposit is the first new magnetite ore body ever found by aeromagnetic means in the Western Hemisphere. Like most of the magnetite deposits recently discovered in the St. Lawrence County district, the Outafit deposit was completely blind, covered by many feet of glacial deposits and alluvium. Since 1945, many notable discoveries of iron ore have been made by means of the airborne magnetometer. Of these, the Outafit is scarcely the largest, but it is the first.

Results of private exploration of this deposit were made available to the Geological Survey in 1958. It has not been possible to incorporate the new data or their interpretation in this report.

MAGNETIC ANOMALY

The magnetic anomaly mapped by dip needle (Reed and Cohen, 1947, fig. 10; this report, pl. 10) consists of two parts, each crudely V-shaped, outlined by closed +10° isoclinal contours. The general trend of the anomaly is northeastward. Each part is flanked on the north-northwest by a low negative anomaly. The smaller, northeastern part of the positive anomaly is 300 feet long within the +20° isoclinal and has a single peak of +44°. This part of the anomaly has not been drilled. The larger, southwestern part of the anomaly, enclosed by a sinuous +20° isoclinal, has two areas (800 and 300 ft long, respectively) within the +50° isoclinal. The main area within the +50° isoclinal was tested by two drill holes. It is possible that the V shape of the positive anomalies reflects plunging folds whose axes trend southwestward.

From the shape of the magnetic profiles, Hawkes estimated that the depth of overburden was probably not greater than 50 feet in most places (Hawkes and Balsley, 1946, p. 6). Hole 2 subsequently cut 17 feet of overburden, and hole 1 cut 55 feet (pl. 10).

GEOLOGY

The Outafit deposit is isolated in a northeast-trending area of glacial deposits nearly a mile wide. (See pl. 1.) Presumably much of this area of glacial deposits is underlain by metasedimentary rocks with sheets of alaskite, microcline granite gneiss, or both types of granitic rock. Though diligent search turned up one outcrop near the anomaly, the nearest sizable masses of exposed bedrock are microcline granite gneiss, locally carrying layers of amphibolite, half a mile away. A small body of alaskite is exposed three-quarters of a mile northwest of the magnetite deposit.

The general structure of the area is indeterminate. The deposit is on the projection of north, northeast, and east foliation trends, somewhere near the place where lineations shift their course from southwest to southeast. Possibly the deposit is on a pair of minor folds near the nose, but on the east limb, of a tightly appressed syncline whose axial plane is nearly vertical.

ROCKS

The principal rock types are skarn, which carries the ore; and microcline granite gneiss, which flanks the skarn zone and presumably encloses it. Minor amounts of metasedimentary gneisses are present in the granite gneiss and in the skarn zone.

SKARN

The skarn zone ranges in thickness from 165 to 210 feet and may exceed 310 feet if the easternmost skarn cut in hole 2 is not a duplication of the western skarn zone (see pl. 10). The zone consists of 85 to 95 percent skarn and 5 to 15 percent metasedimentary gneisses, syenite pegmatite, and microcline granite gneiss (in part syenitic). The gneisses and pegmatite occur as layers 1 to 12 feet thick, apparently distributed at random through the skarn. The skarn is highly varied in character, comprising calcite-rich skarn or silicated marble (4–5 percent), diopside skarn (4–10 percent), salite and ferrosalite skarn (54–60 percent), garnet-pyroxene skarn—generally calcareous (10–19 percent), mica sköls (11–14 percent), and modified skarn—commonly scapolitic (4–5 percent). Only the dark pyroxene skarn forms thick, discrete layers (some of them 25–50 feet thick). The other varieties of skarn are present in sporadic layers generally less than 10 feet thick.

From 25 to 30 percent of the skarn zone has been partly or very largely replaced by magnetite. The preferred host for magnetite is pyroxene skarn (commonly somewhat micaceous or calcareous) and pyroxenic sköls. (See section on ore.)

The varieties of skarn have been adequately described in the general section of this report. Because of the local importance of mica sköls, additional comments on these rocks are presented here.

The sköls range in color from light or silvery to dark and greenish. One seems talcose. Individual sköls are 1 to 12 feet thick—mostly 3 to 5 feet—and not confined to the borders of the skarn zone. The main sköls are all in the ore-bearing part of the skarn zone. While some of them are barren, most of them are not. Some of the richest ore is in dark sköls. Many of the sköls have visible relics of pyroxene and calcite; one or two are especially rich in sulfides. Rarely the sköls show a little supergene alteration to hematite, earthy green chlorite, and calcite.

MICROCLINE GRANITE GNEISS

The microcline granite gneiss that flanks the skarn zone is chiefly a biotitic variety. Locally, a few small red garnets are present. Much of the rock is bleached, weathered, limonite or hematite stained, and somewhat porous. The shape of some of the small, lenticular, leached cavities suggests that sillimanite was formerly present in parts of the granite gneiss. Where the rock is least weathered, it is pink or pinkish gray, fine to medium grained, rarely pegmatitic, and locally streaked with thin layers and schlieren of biotite gneiss or garnet-biotite gneiss. In places, layers of granite gneiss alternate with 3- to 10-foot layers of migmatitic garnet-biotite gneiss, sillimanite-biotite gneiss, amphibolite, or pyroxene gneiss. Within the skarn, or against skarn or pyroxene gneiss, the microcline granite gneiss is pyroxenic. The pyroxenic variety commonly has abundant sphene and apatite, and a little blue-green amphibole. The quartz content is variable but generally low, and in places the gneiss is syenitic. Locally, the microcline granite gneiss has developed a thin selvage of amphibole skarn at the contact with pyroxene skarn.

METASEDIMENTARY GNEISSES

Metasedimentary gneisses in the flanking granitic rocks have already been mentioned. Within the skarn zone, pyroxene gneiss predominates among these gneisses. The rock is pinkish or grayish and consists chiefly of pyroxene and abundant microcline. Quartz and plagioclase are sparse or absent. Brown garnet and scapolite are present locally. Other gneisses, rarely represented within the skarn zone, are amphibolite, hornblende gneiss, and quartz-feldspar granulite.

STRUCTURE

The rocks enclosing the drilled part of the Outafit deposit strike N. 30°–35° E. and dip 90°±15. (See pl. 10.) The single outcrop so far found in the area lies 800 feet northeast of the collar of hole 1; foliation in it strikes N. 30° E., dips vertically, and shows a prominent lineation trending S. 30° W. and plunging 40° SW. The scanty information on strike and dip suggests a simple structure for the area. However, that suggestion is not borne out by the shape and distribution of the magnetic anomalies (cf. Cohen, *in* Reed and Cohen, 1947, p. 8). The V shape of the anomalies suggests the presence of plunging folds, and the orientation of the axes of the V's agrees closely with the trend of the strong lineation observed in the single outcrop.

Locally, mica flakes in drill cores of skarn and sköls give a lineation that plunges down the dip of the foliation (that is, rakes 90°). Presumably this lineation is in *a* relative to the inferred *b* direction of lineation measured in the outcrop.

Accordingly (see pl. 10), the writer infers a series of V-shaped folds plunging southwestward at a moderate angle. As shown schematically on plate 10, the peaks of the magnetic anomalies reflect magnetite concentrations on the east limbs of three plunging synclines. At present, there is no way of knowing whether the northeastern, undrilled anomaly is part of the same "stratigraphic" zone cut by holes 1 and 2, or part of an entirely different zone. The entire reconstruction is hypothetical, and the continuity of ore zones and skarn is not demonstrable from the meager data at hand. Nevertheless, the main outlines of the structural picture are plausible and perhaps in large measure correct.

Several other interpretations are possible. Folding may be present, but the ore zones may be wrongly connected. Or the magnetite may occur as boudins or lumps more or less randomly scattered through an irregular zone of skarn and granite gneiss. A great deal of well-planned drilling would be required to test the different hypotheses sketched above.

ORE

OCCURRENCE

Though some magnetite is present sporadically in almost all varieties of skarn, material of ore grade is virtually confined to mica sköls and pyroxene skarn. Concentrations of magnetite are absent from microcline granite gneiss and metasedimentary gneisses. The ore-bearing sköls may be light or dark; the dark, biotitic sköls are generally preferred by magnetite. Ore-bearing pyroxene skarn commonly carries some biotite or interstitial calcite. The pyroxene from ore-bearing skarn is mainly medium-green salite, though ferrosalite and a variety close to the diopside-salite boundary are represented locally.

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The bulk of the ore is in the middle of the skarn zone. (If the thick skarn zone cut in hole 2 is part of a single zone—not one duplicated by folding—the ore in that hole lies in the lower half of the skarn mass.) Some magnetite is also present near the borders of the skarn zone, but here the magnetite-bearing material is either too thin or too low in grade to constitute ore.

The several hypotheses outlined in the section on structure will guide one's conclusions regarding the shape of the Outafit magnetite body. While it is conceivable that the drilled body might be a simple tabular one, or in contrast highly irregular, the writer suggests that the ore may be confined to one or more shoots paralleling the southwestward plunge of fold axes and lying on the east limb of a southwest-plunging syncline. Two other magnetic highs (centering 300 feet south-southeast of hole 2 and 500 feet northeast of hole 1, respectively) may indicate buried magnetite bodies similarly disposed. The continuation of significant magnetite mineralization along both limbs and around the nose of the synclines inferred on plate 10 is very doubtful. That is not the habit of magnetite bodies in the district. However, significant concentrations of magnetite might be present on the nose, as well as on one limb, of a given fold, yielding the familiar fishhook pattern.

If the ore is concentrated in shoots, it seems unlikely that the shoots would parallel the inferred *a*-lineation given by mica flakes in some drill cores. This statement is based more on opinion than on fact. Nevertheless, deductions from studies at Clifton and from one drill hole at Jayville strongly support that opinion.

MINERALOGY

The gangue of the ore is simply the unreplaced relics of preexisting skarn. Ore from the Outafit deposit has received very little mineragraphic study. Therefore, the data summarized here are by no means complete. The metallic minerals are magnetite, ilmenite, hematite, pyrite, marcasite(?), chalcopyrite, and an unidentified mineral resembling cuprite. Of these metallic minerals, magnetite is the only one of commercial interest. The magnetite always contains a very little ilmenite as microscopic blades and hairs exsolved from the host. The average TiO_2 content of the magnetic fraction of the ore is, of course, low. (See assays, table 24.) Another exsolution product of the magnetite is spinel, which is present as minute spindles and "gashes" within the individual magnetite grains. Hematite, where present at all, occurs as hair-thin veinlets; it is probably supergene, related to the hematite-chlorite-calcite alteration that has locally affected sköls and granite gneiss. The total quantity of soluble nonmagnetic Fe yielded by the quantity of hematite so far detected, is almost infinitesimal.

Sulfides are distributed sporadically. Some of the ore is almost sulfide free; a little of it has abundant sulfides. Perhaps 1 to 4 percent sulfides would represent a crude average. Unfortunately, this "eyeball" estimate is uncontrolled by assays, for the sulfur content of crude ore was never determined chemically. Some of the highest concentrations of pyrite are in skarn or sköls devoid of magnetite. Pyrite seems to be more abundant than pyrrhotite in the ore; hence the sulfide content of a magnetic concentrate should be low (cf. assays of Davis tube concentrate, Reed and Cohen, 1947, p. 8; this report, table 24).

GRADE AND EXTENT

The Outafit deposit is one of the more promising deposits of skarn type discovered in the district. The deposit contains two zones of high-grade ore about 5 to 10 feet thick. Under exceptionally favorable economic conditions, selective mining of the high-grade ore might yield a small tonnage of lump ore. However, if the deposit were developed, it would probably have to be treated as a single, thick, low-grade ore body. The two drill holes show an average thickness of 65 feet of ore averaging 28.9 percent magnetic Fe. All the available analytical data have been presented by Reed and Cohen (1947, p. 8, 31–32). These data are summarized in table 24. Unfortunately, "complete" analyses of the composite samples of crude ore were never made. Nonrecoverable Fe in the dark silicates seems to account for most of the difference between total Fe and magnetic Fe in the assays.

The two holes drilled at the Outafit deposit cut ore averaging 65 feet in thickness at a vertical depth of about 200 feet. The length of this block of ore may be taken as at least 800 feet, the length of the north-south axis of the 50° isoclinal contour.

We do not know whether ore continues downdip or along a raking course parallel to the axes of the inferred folds. If the ore body is a shoot with a plunge of about 40° SW. (the plunge of lineation observed in the one outcrop near by), the available data roughly suggest its limits in two dimensions. The ore cut at a depth of 200 feet was thick; therefore, one may conservatively infer that the ore extends at least 50 feet deeper at hole 1 and 50 feet higher at hole 2. On this assumption, the minimum height of the raking

TABLE 24.—*Grade and thickness of magnetite-bearing layers, Outafit deposit*

[Assays by U.S. Bureau of Mines. Data condensed from Reed and Cohen, 1947, p. 31–32]

Hole	Footage	Apparent thickness (feet)	True thickness (feet)	Weight percent							
				Crude ore		Magnetic fraction only					
				Total Fe	Magnetic Fe[1]	P	S	TiO_2	SiO_2	Mn	Fe
1	179.0–183.0	4.0	3.4	27.1	24.3	--------	--------	--------	--------	--------	--------
	183.0–215.0	32.0	27.5			--------	--------	--------	--------	--------	--------
	215.0–216.5	1.5	1.3	34.0	30.4	--------	--------	--------	--------	--------	--------
	216.5–221.0	4.5	3.9			--------	--------	--------	--------	--------	--------
	221.0–226.0	5.0	4.3	9.4	8.4	--------	--------	--------	--------	--------	--------
	226.0–236.0	10.0	8.6			--------	--------	--------	--------	--------	--------
	236.0–246.0	10.0	8.6	24.4	21.8	--------	--------	--------	--------	--------	--------
	[2] 246.0–281.0	35.0	30.1	39.1	35.0	--------	--------	--------	--------	--------	--------
	281.0–290.0	9.0	7.8	19.2	17.2	--------	--------	--------	--------	--------	--------
	[3] 290.0–316.0	26.0	22.4	37.0	33.1	--------	--------	--------	--------	--------	--------
	316.0–353.0	37.0	31.8	6.4	5.7	--------	--------	--------	--------	--------	--------
	353.0–359.0	6.0	5.2	37.6	33.7	--------	--------	--------	--------	--------	--------
	236.0–316.0	80.0	68.8	[4] 37.4	30.7*	0.004	0.20	0.22	0.74	0.16	70.6
2	235.0–252.8	17.8	14.3	39.2	35.1	--------	--------	--------	--------	--------	--------
	252.8–275.0	22.2	17.8	22.6	20.2	--------	--------	--------	--------	--------	--------
	275.0–290.0	15.0	12.0	34.7	31.1	--------	--------	--------	--------	--------	--------
	290.0–310.0	20.0	16.0	25.4	22.8	--------	--------	--------	--------	--------	--------
	310.0–326.6	16.6	13.3	9.2	8.2	--------	--------	--------	--------	--------	--------
	[5] 326.6–352.5	25.9	20.7	18.8	~14	--------	--------	--------	--------	--------	--------
	352.5–376.7	24.2	19.4	2.7	2.4	--------	--------	--------	--------	--------	--------
	376.7–381.0	4.3	3.4	45.1	40.3	--------	--------	--------	--------	--------	--------
	381.0–399.7	18.7	15.0	4.7	4.2	--------	--------	--------	--------	--------	--------
	399.7–404.5	4.8	3.8	20.8	18.6	--------	--------	--------	--------	--------	--------
	404.5–410.3	5.8	4.6	11.4	10.2	--------	--------	--------	--------	--------	--------
	235.0–310.0	75.0	60.0	29.4	27.1*	0.01	0.02	0.22	0.86	0.12	70.7

[1] Magnetic Fe computed by the writer, using a conservative factor to convert total Fe to magnetic Fe. Starred values from Reed and Cohen (1947).

[2] Includes high-grade ore, 256–266 feet, 50.0 percent total Fe.

[3] Includes high-grade ore, 307–316 feet, 52.2 percent total Fe.

[4] Value probably too high. Weighted average for individual samples (Reed and Cohen, 1947, p. 31) is 34.3 percent total Fe.

[5] Additional data are available for the interval 345.0–352.5 feet (Reed and Cohen, 1947, p. 32):

Crude ore

	Percent
Total Fe	25.0
Magnetic Fe	13.0

Magnetic fraction only

	Percent
P	0.01
S	.05
TiO_2	.50
SiO_2	1.80
Mn	.15
Fe	68.6

The low value of magnetic Fe relative to total Fe is presumably due to the presence of considerable pyrite, as well as local supergene hematite, in the sample interval.

shoot is 400 feet. And if so, the suboutcrop of the lower edge of the shoot must lie at least 300 feet north of the horizontal projection of ore cut in hole 1, or roughly 100 feet north of the north end of the 0° isoclinal contour. The pattern of anomalies in this immediate area lends mild support to the possibility.

Actually, the inference that the upper edge of the hypothetical shoot lies only 50 feet above the ore intersection in hole 2 is not supported by the pattern of the strong positive anomaly. This pattern suggests that the suboutcrop of the shoot's upper edge is at least as far south as the south end of the +50° isoclinal. If the surface projection of the shoot is thus about 900 feet, the height of the shoot is about 600 feet.

Presumably the two undrilled segments of the positive anomaly indicate buried magnetite bodies. If the known deposit is explored by additional drill holes, these two undrilled segments will surely deserve attention.

ROCK POND ANOMALY

The west end of the Rock Pond anomaly is 225 feet north of the northeast corner of Rock Pond, southeast rectangle, Potsdam quadrangle. The anomaly was registered as a peak of 6,440 gammas on the aeromagnetic survey of 1946.

The western half of the dip-needle anomaly trends N. 60° W.; the eastern half trends about east. The anomaly is about 1,800 feet long, and about 150 feet wide, within the +20° isoclinal. It is about 1,400 feet long within the +40° isoclinal. Magnetic intensity greater than +60° is found only in the eastern half of the anomaly. (See fig. 22.) Eight traverses, roughly 200 feet apart, were made across the anomaly.

No outcrops were found in the area of the more intense anomaly. Migmatitic biotite-quartz-feldspar gneiss, locally garnetiferous, crops out within the less intense anomaly. Granite gneiss is also present. Feldspar-veined pyroxene gneiss crops out about 200 yards east-northeast of the northeast corner of Rock Pond. Foliation in the gneisses strikes west-northwest, parallel to the trend of the anomaly, and dips 50°–60° NNE.

The source of the anomaly is presumably a magnetite body of skarn type.

STERLING POND ANOMALY

LOCATION AND DESCRIPTION

The Sterling Pond anomaly is on the southeast slope of the conical hill southwest of Sterling Pond, southwest rectangle, Nicholville quadrangle. With respect to the dirt road west of the pond, the anomaly is about 1,200 feet east of the house foundations 1.1 miles south along that road from the culvert over Sterling Pond outlet. The axis of the anomaly is about 600 feet northwest of the edge of the swamp south of Sterling Pond.

The anomaly was found during the aeromagnetic survey of 1946. It was registered as a peak of 6,070 gammas. The anomaly, mapped by dip needle, trends N. 60° E. It is 700 feet long and 200 feet wide within the +20° isoclinal, and 600 feet long within the +40° isoclinal. An intense negative anomaly is found at the southwest end of the positive anomaly. (See fig. 23.)

There are no outcrops within the area of the positive anomaly. Traces of skarn and white granite are present at the negative anomaly. The country rock 500 feet northwest of the anomaly is medium-grained granite gneiss whose foliation dips 60° or more northwest. White granite gneiss with included lenses and layers of metasedimentary pyroxene gneiss crops out west of the anomaly. The granite gneiss is contaminated locally with xenocrysts of pyroxene. Pyroxene gneiss crops out southwest of the anomaly. Both granite gneiss and pyroxene gneiss are rusty in places. Presumably the magnetic anomaly is caused by a local replacement body of magnetite in skarn. The size of the anomaly suggests that the magnetite body is small. Perhaps it is a pod plunging northeast, though this inference from the nature of the anomaly and the general structure of the area is tenuous indeed.

GEOLOGIC SETTING

The geologic setting of this anomaly is analogous to, though not identical with, that of the Clifton deposit. Near Sterling Pond, the strike of the northwest flank of the Stark anticline changes abruptly from east to northeast. (See pl. 1; and Prof. Paper 376, fig. 14.) Data on the attitude of the axial plane of the anticline northeast of Sterling Pond are not available. However, from sketchy observations it is known that the dip of the axial plane changes from ~60° NW. to vertical somewhere northeast of the pond. The traces of axial planes of folds north of Sterling Pond converge near the pond. The orientation of the dominant lineation shifts from a position parallel to the major fold axes to a position transverse to the fold axes. (See Prof. Paper 376, fig. 15.) A mixed belt of metasedimentary rocks and younger granite gneiss comes directly against the anticline's flank in this neighborhood.

Here, as at Clifton, we find an area of major structural discordance, presence of favorable host rock, and proximity of younger granite. Again, the combination has yielded a magnetite deposit.

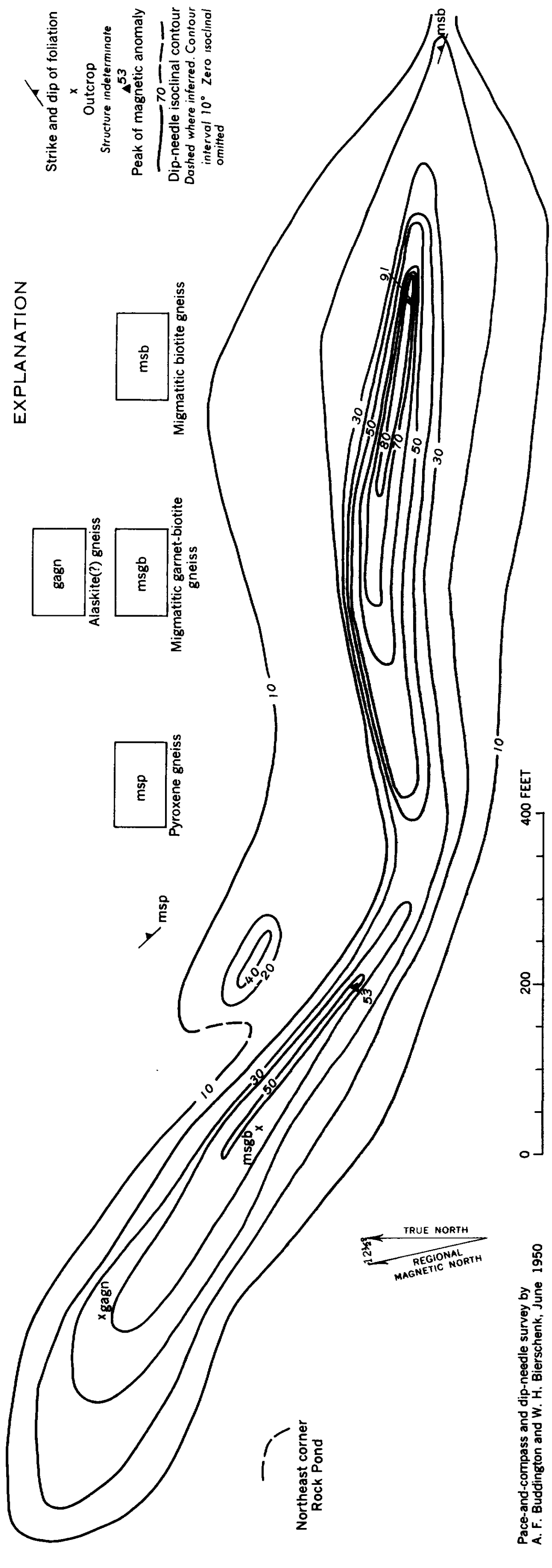

FIGURE 22.—Reconnaissance magnetic map and geologic sketch map of the Rock Pond anomaly, southeast rectangle, Potsdam quadrangle.

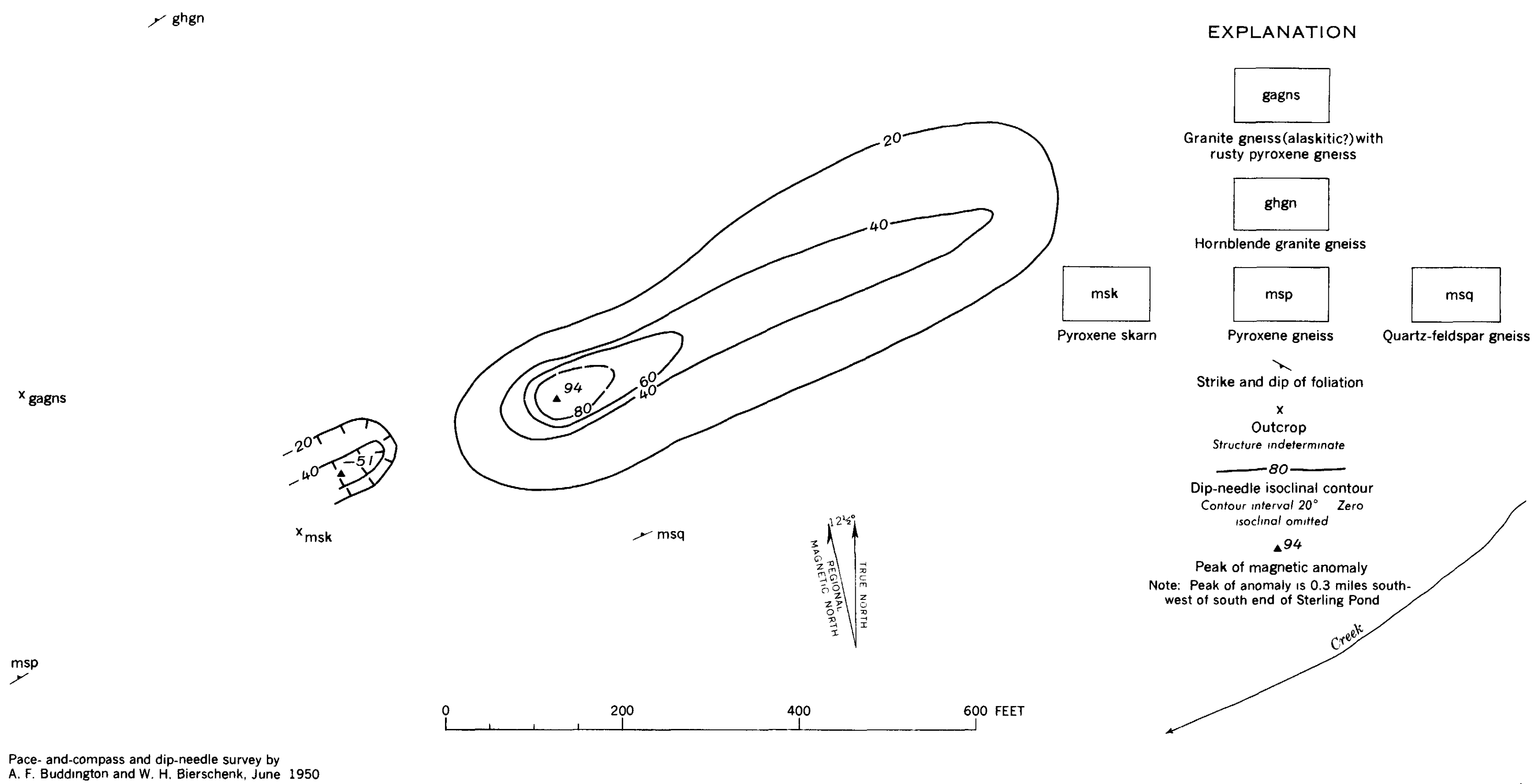

FIGURE 23.—Reconnaissance magnetic map and geologic sketch map of the Sterling Pond anomaly, southwest rectangle, Nicholville quadrangle.

Possibly there are other small magnetite deposits in the zone of mixed granite gneiss and metasedimentary rocks extending from Rock Pond (southeast rectangle, Potsdam quadrangle) through Sterling Pond to Sylvan Falls (southwest rectangle, Nicholville quadrangle).

SUCKER BROOK OCCURRENCES

Buddington and Leonard (1945b, p. 7–8) described several occurrences of magnetite in the Sucker Brook belt of Grenville rocks, east-central rectangle, Cranberry Lake quadrangle. Pyroxene skarn is a prominent member of the metasedimentary rocks in the belt. Several occurrences of magnetite are in the skarn, though others are in gneissic rocks. The following account is based largely on the published reference.

The Clifton-Colton town line crosses an outcrop of magnetite-bearing rock about 275 yards south of Sucker Brook. (See pl. 1.) The host rock is a granitic gneiss with thin layers of biotite schist and schlieren of pyroxene skarn. The exposure of low-grade "ore," about 10 feet thick, contains disseminated magnetite and thin seams ($\frac{1}{16}$–6 in. thick) of magnetite with quartz. Two seams of nearly pure magnetite, 1 inch thick, were observed. The maximum dip-needle readings across the mineralized zone were only moderate. About 400 yards eastward along the strike, loose blocks of biotite gneiss contain enough magnetite to constitute low-grade "ore." Here, too, the dip-needle readings are moderate. A little magnetite was also noted in pyroxene skarn on the narrow ridge on the south side of Sucker Brook, 1.2 miles northwest of Rampart Mountain, west-central rectangle, Tupper Lake quadrangle. The Lake Marian prospect pits are 1.5 miles to the southeast.

The existence of old prospect pits for magnetite in the area east of the east branch of Sixmile Creek has been reported, but the pits were not found by the writers.

Early in the spring of 1945, the U.S. Bureau of Mines began a detailed dip-needle survey of the Sucker Book belt of metasedimentary rocks, but the work had to be abandoned before the ice went out of Cranberry Lake. Subsequently, the aeromagnetic survey of the area failed to detect significant anomalies suggestive of major concentrations of magnetite. The magnetite-bearing zone crossed by the Clifton-Colton town line was perhaps registered as a slight magnetic "hill" of 5,230 gammas. (See Balsley, Buddington, and others, 1954a.)

TWIN LAKE STREAM ANOMALIES

Two local but intense magnetic anomalies lie north of the New York Central Railroad and east of the road from Star Lake to Browns Fall, about 1.5 miles north of New York Highway 3 at Star Lake, northeast rectangle, Oswegatchie quadrangle. Both anomalies are within a short walking distance of the road. The anomalies were registered on the aeromagnetic survey of 1945 as a peak of 900 gammas (Balsley, Hawkes, Buddington, and others, 1946). This is equivalent to 6,100 gammas, relative to the datum adopted for aeromagnetic maps of surrounding areas (Balsley, Buddington, and others, 1954a, 1954b). The following account is slightly modified from one originally prepared by Buddington (Hawkes and Balsley, 1946, p. 8).

The two anomalies (see fig. 24) are apparently not connected. The western anomaly consists primarily of an irregular area of high magnetic intensity on the hill crest immediately southwest of the elbow in Twin Lake Stream. A narrow tongue of less intense anomalous values extends for about 400 feet westward to the road, where it apparently dies out. Between the hill crest and the stream to the north, a relatively intense negative anomaly was observed. Another zone of magnetic highs extends a short distance south of the principal anomaly, but it is not so well defined or intense. Both zones are roughly conformable with the geologic structure.

The eastern anomaly is essentially a linear zone of magnetic highs trending northward. The anomaly ends abruptly on the north and tapers off to low values at the south.

The rocks exposed in the vicinity of the western anomaly are predominantly skarn, quartzite, and migmatite of the Grenville series, with subordinate pink granite. The most intense part of the anomaly is at an abrupt bend in the geologic structure, where the foliation changes strike from north to west. (See fig. 24.) Probably this minor structure served to localize the deposition of magnetite.

No outcrops were found in the immediate vicinity of the eastern anomaly.

Several old prospect pits were found in the area of the western anomaly. The magnetite in these pits is apparently a replacement of medium- to coarse-grained pyroxene skarn, as judged from dump specimens. Some of the material is rusty and decomposed, indicating an appreciable content of iron sulfides. All the magnetite-bearing specimens were low-grade "ore," and the general tenor of the dump material suggests that the distribution of the magnetite is erratic. In addition to the magnetite in the old prospect pits, an

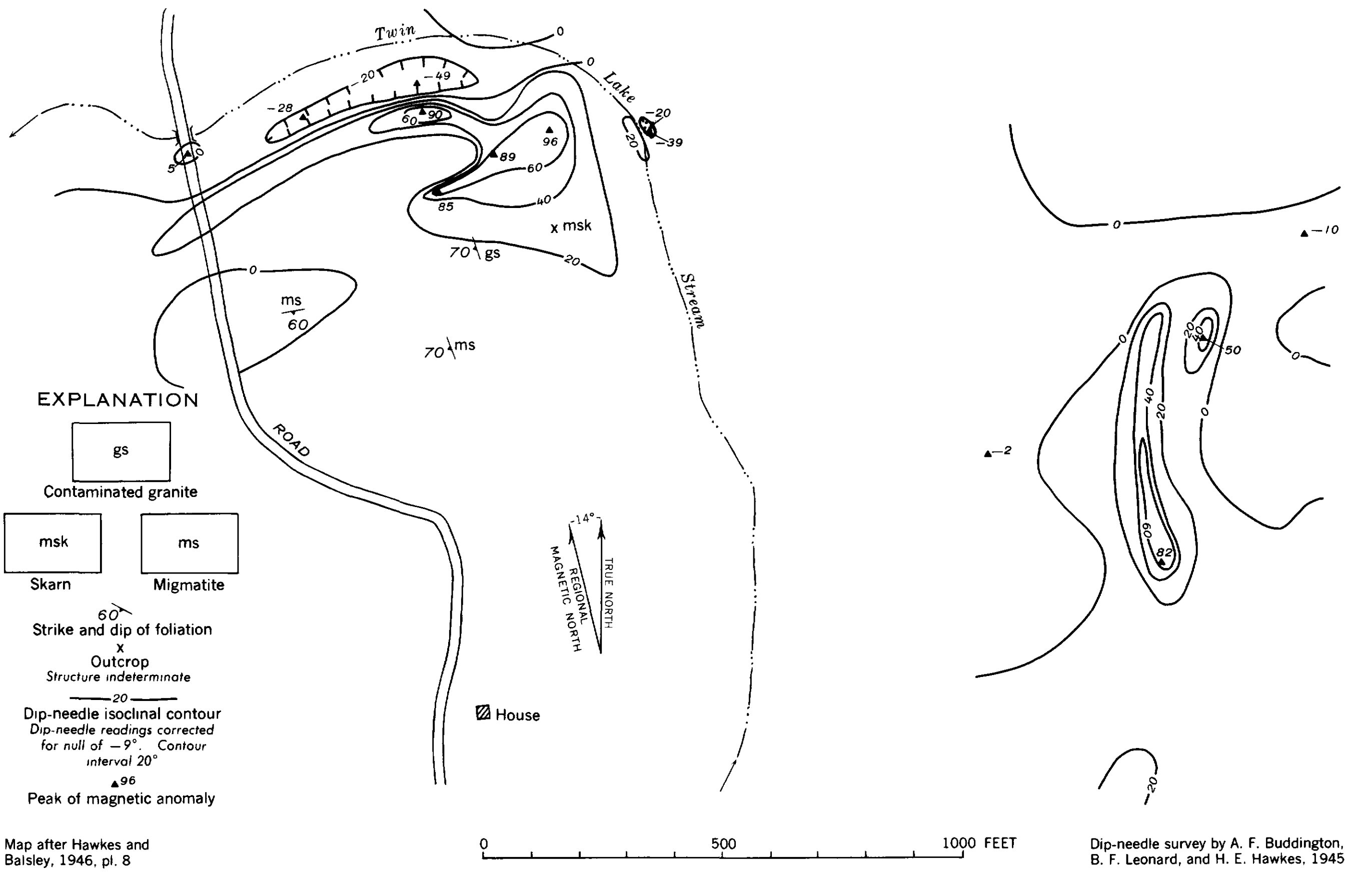

Figure 24.—Reconnaissance magnetic map and geologic sketch map of the Twin Lake Stream anomaly, northeast rectangle, Oswegatchie quadrangle.

outcrop of magnetite-bearing skarn was found in the creek bed east of the center of the western anomaly.

The Twin Lake Stream anomalies are in an area of metasedimentary rocks effectively isolated from the main bedrock areas by glacial debris. (See pl. 1.) Tentatively, it appears that the magnetite-bearing rocks are on the limb and nose of a minor syncline whose axis plunges steeply southwestward.

WALKER FARM (HUGHESVILLE SCHOOL) DEPOSIT

The Walker Farm (Hughesville School) deposit is 1,500 feet northwest of the abandoned Hughesville School, central rectangle, Russell quadrangle. The school is accessible from the macadam road connecting Edwards and Russell via Whippoorwill Corners. An old trail leads from the school north and west to the prospect pits on the Walker Farm. Other prospect pits are said to lie some distance to the west. According to one account, some magnetite was shipped from the deposit about 1870.

The Shenango Furnace Co. explored the property intensively in 1943 by means of 14 closely spaced diamond-drill holes. The results of the drilling were discouraging. A number of stringers of magnetite were encountered, but the small quantity of magnetite-bearing material did not warrant further work. Logs of the drill cores and a report on exploration were kindly made available to the Survey by Messrs. George L. Collord and W. Guy Srodes, of the Shenango Furnace Co. Unfortunately, the map showing drill-hole sites could not be found; therefore, the writer has been unable to piece together information obtained from the drilling.

A crude geologic sketch (fig. 25) shows the general relations of rock units in the neighborhood of the pits. The regional setting of the deposit may be seen on plate 1.

The quantity of observable magnetite is very small. Only a few thin stringers were exposed in the northern pits, though several small, loose blocks of good ore were present nearby. Some magnetite was piled beside the southern pit, but very little could be seen in the pit walls. The magnetite, accompanied by pyrite, pyrrhotite, and a little chalcopyrite, occurs in dark-green to black pyroxene-garnet skarn. The very dark andraditic garnet is readily mistaken for magnetite. Garnet is more abundant than pyroxene at the northern pits; some diopside skarn is also present, as are quartz-microcline-pegmatite veins. The southern pit has mainly pyroxene skarn, locally calcareous, with dark mica. Medium- to coarse-grained microcline aggregates are present near the southern pit, south of which is a mass of colorless diopside skarn.

A partial chemical analysis of a sample considered by the company to be representative is given below:

Analysis of grab sample of ore from Walker Farm deposit

[Sample collected by W. G. Srodes, 1943. Analysis by H. Davis, chemist, Shenango Furnace Co.]

	Weight percent
Total Fe	47.00
Magnetic Fe[1]	36.5
P	0.003
S	0.93
Mn	0.09
SiO_2	18.98
Al_2O_3	1.46
MgO	3.70
CaO	7.35
Loss on ignition	nil
Moisture	nil

[1] Calculated by Leonard; value crudely approximate. 77.77 percent of ore sample was magnetic, according to H. Davis.

The drill holes cut mainly skarn interlayered with silicated marble. The apparent thickness of multiple magnetite-bearing layers in the skarn was generally a few inches to a few feet.

The only reliable planar structure is shown by the granite. The skarn fails to show a good planar structure (layering of component minerals) except at the south pit, where there is a nearly horizontal mica sköl in the skarn. Farther south, a faint planar structure in the diopside skarn appears to strike east and dip 20° N. At the north pits, joints dip steeply eastward. Some of the barren drill holes were apparently put down on the assumption that this jointing represented "bedding" in the skarn. Such holes may well have been started on the footwall side of the magnetite-bearing zone. Perhaps very careful mapping of every available structural detail would have permitted somewhat wiser drilling of the deposit.

The fact that the deposit failed to show as a significant anomaly on the aeromagnetic survey of 1945 (Balsley, Buddington, and others, 1954b) bears out the evidence in pits and drill cores: the Walker Farm deposit is commercially unattractive because of its small size.

WHITE PLACE ANOMALY

The White Place magnetic anomaly, northwest rectangle, Stark quadrangle, centers 1,800 feet east-southeast of the White place, an abandoned farm 2,500 feet west-southwest of the south end of Cranberry Pond. The anomaly lies in a small valley about 200 feet south of the old road running from Twin Ponds (northwest rectangle, Stark quadrangle) past Clear Lake to Lower District School (northeast rectangle, Russell quadrangle).

The anomaly was registered as a peak of 6,080 gammas on the aeromagnetic survey of 1945 (Balsley, Buddington, and others, 1954b).

The dip-needle anomaly trends N. 75° E., following the valley. The +40° isoclinal of the anomaly is about 500 feet long and only 10 to 40 feet wide (generally 20

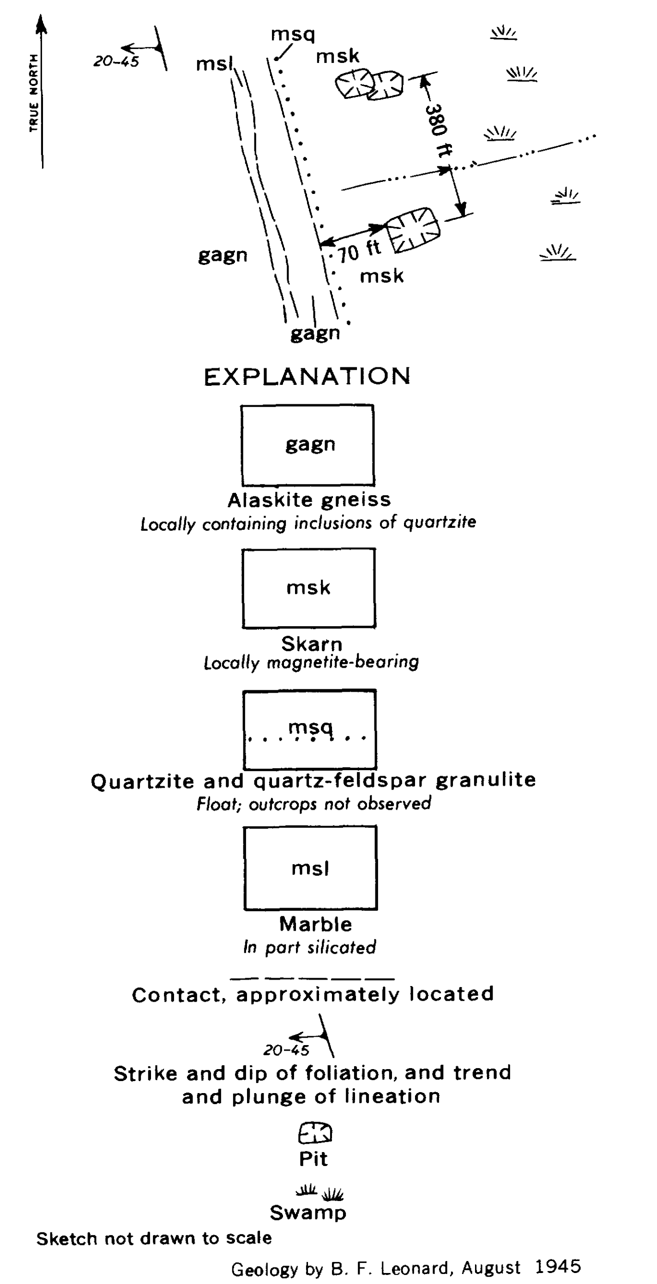

FIGURE 25.—Geologic sketch of the Walker Farm (Hughesville School) prospect, central rectangle, Russell quadrangle.

feet). This isoclinal encloses several peaks of +65° to +75°. The +20° isoclinal extends a few tens of feet beyond the +40° isoclinal.

Outcrops of wallrock are relatively abundant in the area, but the source of the anomaly is not exposed. The foliation strikes northeast and dips 40° NW. A lineation trending northwest plunges downdip at 40°. Granite, marble, feldspathic skarn, pyroxene gneiss, and migmatitic biotite gneiss crop out on the footwall (southeast) side of the anomaly. Alaskite and pyroxene gneiss, the latter locally migmatitic, crop out on the hanging-wall (northwest) side.

One may guess that the anomaly is due to narrow concentrations of magnetite in a thin layer of pyroxene skarn veiled by very little overburden. The anomaly appears to have no economic significance at this time.

WILSON MOUNTAIN ANOMALY

The Wilson Mountain anomaly centers 1 mile north-northwest of the summit of Wilson Mountain, south-central rectangle, Stark quadrangle. The anomaly was discovered by the Geological Survey in the fall of 1944, when extreme compass deflections were noted there in the course of regional geologic mapping. The anomaly was registered as two sharp peaks of 6,160 and 6,095 gammas on the aeromagnetic survey of 1945 (Balsley, Buddington, and others, 1954b). The Geological Survey made a reconnaissance dip-needle survey of the anomaly in the summer of 1946. Later that year, the U.S. Bureau of Mines made a detailed dip-needle survey of the area (Reed and Cohen, 1947, p. 11; fig. 16). The following account is modified from that given by Cohen (Reed and Cohen, 1947, p. 11).

To reach the anomaly, drive 2.32 miles northwest along the abandoned lumber railroad grade from its intersection with the road to the Massawepie Club. Traverse 3,050 feet, S. 70° W., from the railroad grade to a point on the Bureau of Mines base line. (See fig. 26.) The road to the Massawepie Club could be reached from N.Y. Highway 56 via the Hollywood Club cutoff, from N.Y. Highway 3 via the Buckhorn Ridge road. (See instructions for reaching Outafit deposit.) All these side roads, as well as the abandoned railroad grade, were accessible by light truck in 1950.

The magnetic anomaly consists of two main parts. (See fig. 26.) The eastern part has a distinctive fishhook shape. This part of the anomaly is 1,500 feet long within the +20° isoclinal and about 1,000 feet long within the +50° isoclinal, measuring around the U. The western part of the anomaly is nearly linear but slightly hooked. It is about 650 feet long within the +20° isoclinal and about 550 feet long effectively within the +40° isoclinal.

The magnetic anomaly is near the south end of the Webb Creek anticline (Prof. Paper 376, p. 119), in a drift-covered area thought to be underlain by Grenville rocks. (See pl. 1.) Hornblende granite of the Webb Creek anticlinal belt crops out 1,000 feet north of the anomaly, and a body of alaskite lies 1,500 feet south of the anomaly. Metasedimentary rocks, including pyroxene gneiss and skarn, are exposed about 3,500 feet southwest of the anomaly. Cohen (personal communication) found granite containing amphibolite layers close to the base line (see fig. 26), but he could not be sure that this rock was in place. It seems reasonable to infer that the source of the Wilson Mountain anomaly is a magnetite body of skarn type, perhaps a promising one.

The hookshaped pattern of the anomaly strongly suggests the presence of folds at depth. The axes of minor folds in the rocks a mile or so from the anomaly trend S. 20°–25° W. and plunge 40°–65° SW. The constructed axis of the Webb Creek anticline, at its south end, must lie between S. 5° E., 25° S., and S. 45° W., 40° SW.; its trend is very probably close to S. 20°–25° W. Thus it seems likely that the inferred folds underlying the Wilson Mountain anomaly trend about S. 20°–25° W. and plunge moderately to steeply southwestward. These folds would be interpretable as minor, satellitic features related to the major structure of the Webb Creek anticline.

MAGNETITE DEPOSITS AND MAGNETIC ANOMALIES OF GRANITE GNEISS TYPE

DEPOSITS IN THE DEAD CREEK AREA

By A. F. Buddington

SUMMARY

The Dead Creek area, south-southeast of Wanakena, Cranberry Lake quadrangle, contains unexplored magnetite deposits whose inferred extent is considerable. An area of about 6,800 by 7,600 feet contains a few outcrops of low-grade ore, many exposures of mineralized rock, and several long lines of magnetic anomaly. (See pl. 11.) Two belts of magnetic anomaly have particularly good characteristics indicative of magnetite concentrations. The first belt, Dead Creek No. 1, has a length of about 3,800 feet within the +20° isoclinal contour; 2,500 feet of this has dip-needle readings of +60° and greater. Several outcrops of magnetite gneiss are exposed along 500 feet of one part of the anomaly, and in one section a true thickness of 30 feet of ore may be conservatively inferred, with neither footwall nor hanging wall exposed. The estimated grade of ore exposed in this section is 20 to 40 percent Fe. The second belt with strong anomalies, Dead Creek No. 2, is about 2,000 feet long within the +20° isoclinal. This compound belt, 425 to 500 feet wide, contains several bands of magnetic anomaly. One band showing maximum values of +60° or greater is about 1,000 feet long. There are two parallel bands

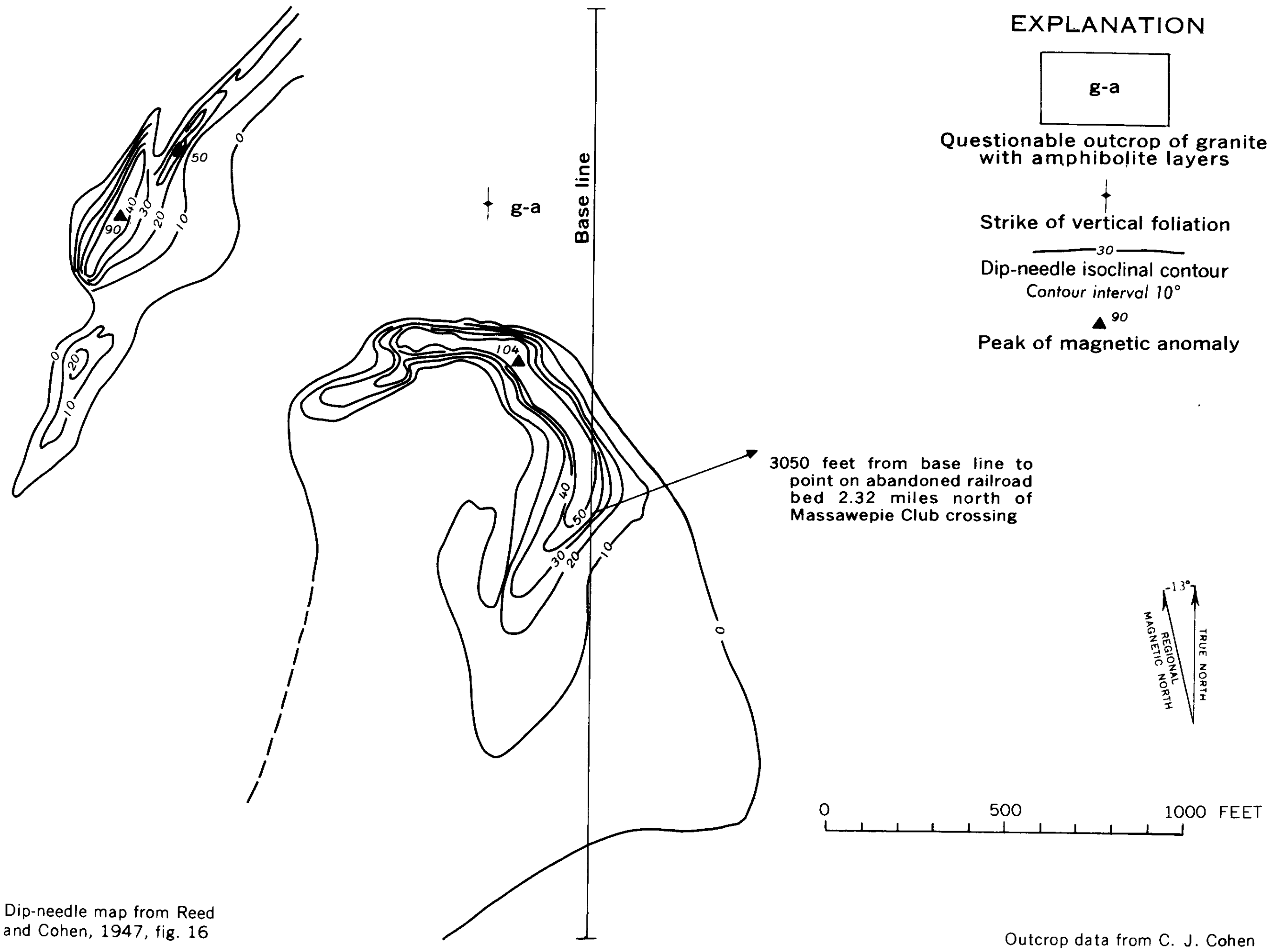

FIGURE 26.—Magnetic map of the Wilson Mountain anomaly, south-central rectangle, Stark quadrangle.

of anomaly, each 800 to 900 feet long within the +30° isoclinal, and another anomaly perhaps 900 feet long within the −30° isoclinal. A number of small bands of anomaly of moderate to high intensity are also present. The belt as a whole is bordered on the west by an additional zone, perhaps 200 feet wide, with low magnetite impregnation (about 10 percent by volume). We infer from these data that veins of low-grade magnetite ore with an aggregate length of over 3,500 feet are present along the belts of Dead Creek anomalies Nos. 1 and 2.

In addition to these two magnetic anomalies, there are several others, generally with low to moderate intensity, that have an aggregate length exceeding 6,000 feet. Local short sections of these, totaling more than a thousand feet, have a high intensity and may indicate additional small, local magnetite ore shoots.

For the whole Dead Creek area, the aggregate length of magnetic anomaly effectively within the +40° isoclinal exceeds 6,500 feet.

Some negative anomalies are not complementary to magnetite concentrations but are due to small or perhaps moderate amounts of TiO_2-bearing hematite, with or without some magnetite.

LOCATION AND ACCESSIBILITY

The area of mineralization (pls. 1 and 11) lies between Glasby and Skate creeks, southwest of the head of Dead Creek Flow, 1½ to 3 miles south-southeast of the village of Wanakena and 6 to 7 miles southeast of Benson Mines. The area is within the Adirondack Park and is part of lands owned by New York State. Wanakena is about a mile by macadam road from a main concrete highway (N.Y. Highway 3). The New York State Conservation Commission maintains two roads which extend from Wanakena to High Falls and Dead Creek Flow, respectively. These are not open to the public for use as highways, but they are available as foot trails. Three foot trails afford direct access to the area shown on plate 11. These are the Glasby

Creek trail on the east, the Dobson trail through the center, and the Leary trail on the west.

DISCOVERY AND EXPLORATION

In October 1943, while engaged in geologic mapping, Leonard found a small outcrop of magnetite-bearing gneiss about 2½ miles south-southeast of Wanakena, just west of the Dobson trail. During the succeeding field season, Buddington and Leonard spent five weeks making pace-compass reconnaissance dip-needle traverses to outline the area of mineralization and to determine the approximate position and trend of magnetic anomalies. They also studied the rock outcrops to obtain an understanding of the geology. The results of this work were published shortly (Buddington and Leonard, 1945a). The area lies within the Adirondack State Park, and therefore no development work has been undertaken or can be done at the present time. The Dead Creek deposits were registered on the aeromagnetic map (Balsley, Buddington, and others, 1954a) as a group of prominent peaks of which the maximum is 6,590 gammas.

Any magnetic exploration and interpretation must take into account the fact that widespread accessory disseminations or localized concentrations of titaniferous hematite, either alone or with magnetite subordinate, give negative magnetic anomalies; and that where magnetite is the major oxide, the presence of associated hematite will have the effect of reducing the intensity of the magnetic anomaly that would otherwise be obtained from the magnetite. Hematite from one sample of mineralized rock from the Dead Creek area was found, upon analysis, to contain about 10 percent ilmenite in solid solution. It is therefore the variety titanhematite. The mineralized rock had 10 percent hematite and a little magnetite. By analogy with the deposits at Benson Mines, the hematite of more intense concentrations will presumably run lower in TiO_2. Hematite certainly plays a substantial role as one of the oxides in the Dead Creek area, as it does in the Benson Mines deposits. Whether it is entirely an accessory mineral or is locally contentrated is wholly unknown.

GEOLOGY

DEAD CREEK SYNCLINE

The bedrock of the area (pl. 11) consists in general of metasedimentary rocks with two intercalated sheets of microcline granite gneiss, in large part contaminated with biotite, sillimanite, or garnet. These rocks form a small syncline that is completely surrounded by and lies in a mass of generally uniform hornblende-microperthite granite, which is the predominant rock of the region.

The foliation and bedding of the Grenville rocks are parallel. There is, so far as noted, complete parallelism also in the foliation of the Grenville rocks, the intercalated granite sheets, and the surrounding gneissoid granite. The magnetite deposits, as much as may be seen of them, are likewise tabular bodies conformable to the foliation of the rocks in which they occur.

The syncline (pls. 11 and 1) is elongated in a generally west-northwesterly direction with the longer axis about 2⅓ miles long and the shorter axis 1⅓ to 1⅔ miles long. The south limb of the structure has been forced over toward the north to form an overturned isoclinal fold whose layers dip generally between 10° and 30° south (pl. 11). The southeastern part of the south limb of the syncline has been forced northward relative to the extreme western part, narrowing the eastern part of the synclinal structure and bending and dragging northeastward the formations along the southeast side of Dead Creek. About two-thirds of a mile west-northwest of the summit of Roundtop Mountain the dips, seen as one comes south and east along the strike, change successively from 30° E. through vertical to 25° S. as a result of the overturning of the south limb.

Several gentle minor folds are present in the west-central part of the syncline, just north and west of the crossing of the Dobson trail and the north branch of Dead Creek. In addition, small isoclinal puckers are found in the Grenville rocks and in some of the sillimanite-microcline granite gneiss.

Most of the rocks in the western part of the syncline have a strong linear structure, as a result of the parallel orientation of certain of their minerals with their longer axes trending generally S. 20° E. to S. 35° E. and plunging gently south-southeast. In some gneisses of the Grenville series, a linear structure of similar orientation is given by strong ribbing in the foliation surface, or by elongate pegmatite stringers.

METASEDIMENTARY ROCKS OF THE GRENVILLE SERIES

The metasedimentary rocks of the Grenville series comprise biotite gneiss, locally sillimanitic and garnetiferous; feldspathic and scapolitic pyroxenic quartzose gneisses or granulites; pyroxene-scapolite gneisses; and a subordinate amount of pyroxene skarn. The biotitic gneisses, which are interpreted as originating through the metamorphism of shales or graywacke, commonly have a variable amount of thin granite pegmatite in layers parallel to their foliation and may therefore be called migmatites, mixed rocks, or veined gneisses. The quartzose gneisses are thought to be metamorphosed feldspathic or calcareous sandstones. The skarn is the result of recrystallization and hydrothermal replacement of original limestone beds to yield pyroxene aggregates.

Locally, the metasedimentary rocks of the Grenville series carry a little disseminated pyrite, which causes them to weather a rusty brown.

In places throughout the area, small granite pegmatite veins contain black tourmaline. Nodules of black tourmaline aggregates are also found locally in the biotite gneisses.

GRANITE GNEISS

The granite gneiss of the intercalated sheets is commonly a fine-grained, pink, granulose rock, heterogeneous in character and usually crumbly at the weathered surface. At many places throughout its extent the gneiss carries small partings, schlieren, layers, or laminae of partly granitized biotite- or garnet-biotite gneiss of the Grenville series. It is possible that some of the layers of Grenville are sufficiently continuous so that they separate the granite gneiss into discrete sheets, and that at least locally these constitute a composite sheet of two or more parts. Outcrops are not sufficiently continuous to permit proof of this possibility.

Microcline (occasionally another potassic feldspar) is by far the most abundant feldspar in the granite gneiss of the intercalated sheets, whereas microperthite preponderates in the hornblende granite, the country rock surrounding the syncline. Hornblende has not been observed in the granite gneiss of the intercalated sheets, except where the granite gneiss is associated with pyroxene gneisses of the Grenville series. Biotite, garnet, or sillimanite—singly or combined—occur in much of the microcline granite gneiss; in places they are abundant. These minerals have been derived, it is thought, from portions of metasedimentary rocks of the Grenville series. The sillimanite occurs as small, elongate bundles of fibers whose parallel orientation defines the linear structure. Locally the sillimanite is sufficiently concentrated to form small, flattened, elongated, fibrous discs or platy lenses, which appear as projecting white fins on a weathered surface. Small red garnets are present at many places within the granite gneiss and are especially noticeable in the sheets forming the north limb of the eastern part of the syncline. Garnet is conspicuously more abundant than sillimanite southeast of Dead Creek, in the granite gneiss of the north limb of the syncline, but the relative amount of sillimanite is much greater northwest of Dead Creek. In contrast, the granite gneiss of the south limb of the syncline, southeast of Dead Creek, contains little or no garnet or sillimanite.

SUPERPOSITION OF ROCK LAYERS

The superposition of rock layers in the syncline is given in table 25.

TABLE 25.—*Superposition of rock layers in Dead Creek syncline*

Rock layers	Thickness, in feet		
	Western part	Eastern part	
		North limb	South limb
Metasedimentary (including pyroxene skarn) and mixed rocks of the Grenville series	----------	----------	----------
Granite gneiss, in part sillimanitic or garnetiferous. The two major anomalies (Nos. 1 and 2) lie within this unit	575+	~400	150–225
Metasedimentary and mixed rocks of the Grenville series	175–275	~150	~50
Granite gneiss, in part sillimanitic or garnetiferous	350–425	~225	300–400
Metasedimentary (including subordinate pyroxene skarn) and mixed rocks of the Grenville series	~200	~125	~50–100
Hornblende granite, major country rock	----------	----------	----------

The aggregate thickness of the two granite gneiss sheets northwest of Dead Creek is 925 to 1,000 feet, whereas the aggregate thickness southeast of Dead Creek is only 450 to 625 feet. The greater thickness of granite gneiss in the west, resulting in additional strength, may have buttressed the western part of the syncline against squeezing, thus permitting it to retain a greater north-south width. The Grenville rocks likewise are thicker on the northwest than on the southeast. This may be in part the result of differential flow toward the keel of the syncline.

MAGNETIC ANOMALIES AND MAGNETITE DEPOSITS

Many outcrops in the western half of the Dead Creek syncline show a distinct but relatively low percent of magnetite, although there are only a very few that carry as much as 20 percent magnetite or that are located along zones of intense magnetic anomaly. Any ore veins that are present, therefore, can be delineated only by a systematic survey for magnetic anomalies. Much of the area of the syncline, especially the central and western part, yields low magnetic anomalies. The entire valley of Dead Creek and the lowland between its headwaters and the Leary trail fail to show any but minor anomalies. This low ground is almost completely covered with surficial material of unknown thickness. The presence of overburden may partly account for the low readings here.

Dip-needle readings less than $+20°$ but greater than $0°$ were found (wherever outcrops were available for observation) to be directly over rock carrying magnetite in relatively small quantities, or over rocks of too low grade to be considered ore. Only lines of anomaly with readings of intensity of $+20°$ or greater are therefore shown on the map (pl. 11). This is a reconnaissance map based on pace-and-compass traverses supplemented by rough picket lines across the strong anomalies. The traverses across all anomalies except No. 5 range from 100 to 500 feet apart and are commonly 200 to 300 feet apart. Alternate interpretations might have been made in connecting the lower magnetic readings, but it is believed that the map gives a generally correct picture of the moderate to strong anomalies and affords a basis for planning the future detailed magnetic survey that would be necessary before undertaking exploration.

The magnetite deposits indicated by Dead Creek anomalies Nos. 1, 2, 5, and 6 occur within sheets of microcline granitic gneiss intercalated in beds of the Grenville series. They are therefore deposits of granite gneiss type and may be expected to show the features appropriate to that type.

The country rock underlying Dead Creek anomalies No. 3 and No. 4 is uncertain; No. 7 is in the contact zone between granite gneiss and Grenville, and No. 8 is in the Grenville. There is too little evidence now at hand to predict what the nature of these deposits may be. At anomaly No. 7, magnetite was seen to occur in thin sills of granite gneiss, in feldspathic quartzose gneiss of the Grenville, and in pyroxene skarn.

ANOMALY NO. 1

Dead Creek anomaly No. 1 is crossed by the Dobson trail 0.6 mile northeast of the summit of Roundtop Mountain, near BM 1901. It is 3,800 feet long within the +20° isoclinal. Of this, 2,500 feet is effectively within the +60° isoclinal and yields values up to 90° or more, both positive and negative. The width of the anomaly with intensity greater than +20° ranges from 100 to 440 feet. A strong negative anomaly along the north border is presumably due to the moderate southerly dip of the ore layer. The south part of this negative anomaly was observed to be directly over ore in one section. Very high negative readings are also found locally within the general zone of positive anomalies and in two places were observed to be directly over ore. The strike of the anomaly appears to be conformable with the trend of the geologic formations, except at the east end. Here the anomaly turns south and transects the strike of the foliation and the different belts of rock. No indications of magnetite adequate to cause an anomaly of the magnitude of this eastern segment were seen in the several outcrops within its area. It is possible that this part of the anomaly marks the trace of the east edge of the magnetite ore body where it dips south beneath the surface. A rather uniform low positive anomaly (about +10°) extends for more than 1,000 feet continuously to the south of the main anomaly and may in part be due to gently dipping ore beneath this area.

The only bedrock exposed within the belt of the intense anomaly is low-grade ore that crops out on the Dobson trail and may be traced west-northwestward for more than 500 feet by means of several exposures and float. At one place a true thickness of about 30 feet of ore may be conservatively inferred, neither hanging wall nor footwall being exposed. The estimated grade of the exposed ore is 20 to 40 percent Fe in magnetite, which occurs as disseminated grains and clots and as sporadic thin seams and layers.

The ore vein and the anomaly lie within a sheet of microcline granite gneiss intercalated in gneisses of the Grenville series. The granite gneiss sheet is about 150 to 225 feet thick here, and the ore occurs in the uppermost part of the sheet. The ore-bearing zone is actually the lower part of the granite sheet if its proper stratigraphic position is considered, for the sheet is overturned. The dip probably averages about 25° S.

ANOMALY NO. 2

Dead Creek anomaly No. 2 lies on the south slope of the hill northwest of Dead Creek, about 1,300 to 1,800 feet west of the Dobson trail. A line of low anomaly can be traced from the west side of Dead Creek, opposite the end of Dead Creek anomaly No. 1, to the belt of Dead Creek anomaly No. 2. The evidence, however, is thought to be insufficiently conclusive to warrant treating No. 1 and No. 2 as parts of the same anomaly, though they may be. If the latter is true, the dip of the magnetite-bearing rock may be expected to change from gentle southeast (east of Dead Creek) through vertical (a short distance northwest of Dead Creek) to 30°–40° E. (where the strike is north). Dead Creek anomaly No. 2 is not a single, simple anomaly but a belt of anomalies. The belt within which values of inclination for the major anomalies are +20° or greater is about 2,000 feet long and 425 to 500 feet wide. The band of most intense anomaly within this belt is about 950 feet long effectively within the +60° isoclinal. The width within the +20° isoclinal for this anomaly is 100 to 200 feet. Parallel to this anomaly on the west is another anomaly, 800 feet long effectively within the +30° isoclinal and about 75 feet wide within the +10° isoclinal. (The +10° and +30° isoclinals are not shown on pl. 11.) Again to the west there is a parallel band of equally intense magnetic anomaly 900 feet long and 100 to 175 feet wide within the +10° isoclinal. North of these last two positive anomalies is a belt of negative anomaly 1,700 feet long effectively within the −20° isoclinal and about 600 feet long within the −40° isoclinal. In addition, a number of small, scattered anomalies of moderate to strong intensity are present.

The source of the moderate negative anomaly may be rock in which titaniferous hematite is the predominant iron oxide, the quantity of iron oxides being indeterminate but perhaps appreciable.

Within the bands of moderate to strong positive anomaly, no bedrock was observed. At one place in the hanging-wall zone, several boulders of lump ore were seen, and at another place an outcrop exposed several feet of low-grade magnetite ore. One outcrop of low-grade magnetite ore was observed in a local zone of moderate negative anomaly.

West of the belt of anomalies under discussion, along the east side of a small creek valley, sillimanitic granite gneiss is well exposed in an area 200 feet or more wide and 1,000 feet long. The gneiss carries disseminated iron oxides, together with iron oxide-bearing quartz veins or feldspathic quartz veins, and iron oxide-bearing granite pegmatite seams. The quantity of oxides varies from 5 to 20 percent but in general is probably about 10 percent. Low negative dip-needle readings ($-10°$) are commonly obtained in this zone; they are due to the predominance of a titaniferous hematite in the oxides, though a little magnetite is also present.

The magnetite-bearing outcrops and the belt of anomalies are all within the same sheet of granite gneiss that underlies Dead Creek anomaly No. 1. However, the sheet of granite gneiss is much thicker ($\sim$575 feet) in the vicinity of anomaly No. 2 and usually carries some sillimanite. Locally, thin layers and schlieren of biotite gneiss are present. The foliation and anomalies strike about north, and the structure dips 30°–40° E. There may be a true thickness of 350 to 400 feet of gneiss here, as layers carrying iron oxides (in part constituting lean impregnations, in part constituting low-grade ore) interleaved with some barren material, for a length of 1,800 feet or more.

ANOMALY NO. 3

Just west of the head of Dead Creek is a zone of low to moderate magnetic anomaly. The zone is about 400 feet wide and perhaps 700 feet long. There are no outcrops; hence, the nature and significance of the anomaly can be evaluated only after detailed work has been done.

ANOMALY NO. 4

On a gentle rise within a swampy area in the outermost part of the southwest bend of the syncline is a narrow belt of anomaly about 650 feet long with maximum inclination rather consistently between $+30°$ and $+40°$ but occasionally reaching $+60°$. The width within the $+15°$ contour (not shown on pl. 11) is in general probably less than 100 feet. There are no outcrops.

ANOMALY NO. 5

In the lower granite sheet in the western part of the syncline is an anomaly which, on one traverse, yielded several high readings. The intense anomaly, however, does not appear to extend very far. Outcrops along this general band show both pink sillimanite granite gneiss and biotitic migmatite. Some of the layers are rusty-weathering pyritic gneiss.

ANOMALY NO. 6

In the northern part of the lower granite gneiss sheet, about a quarter to half a mile east of the Dobson trail, is a narrow band of magnetic anomaly, about 1,750 feet long, of moderate to high intensity. All but the northwestern 600 feet of this anomaly has a width within the $+10°$ isoclinal (not shown on pl. 11) of only 50 to 60 feet. The greatest intensity is at the northwestern part, where the anomaly is about 100 feet wide within the $+10°$ isoclinal. Only one outcrop was seen within this anomaly. It is near the northwest end and consists of sillimanite granite gneiss with about 20 to 30 percent magnetite by volume as disseminated grains, discs, and lumps.

Some 400 feet northeast is another narrow anomaly, several hundred feet long but of only low to moderate intensity.

ANOMALY NO. 7

East of the Dobson trail, a narrow positive anomaly, succeeded on the west by a narrow negative anomaly, is present on the north border of the lower granite sheet in the contact zone between granite gneiss and metasedimentary rocks of the Grenville series. The zone of positive anomalies is shown on the map for a length of 1,100 feet. Outcrops showing some magnetite indicate an extension of the mineralized zone for another thousand feet southeast. The anomaly will have to be surveyed in detail to evaluate its significance. Outcrops of granite interlayered with pyroxene skarn and with gneiss of the Grenville series, in part pyritic, form the edge of the hill south of the mapped anomaly. Layers in which the content of magnetite varies from a few percent to perhaps 30 percent are visible in skarn and gneiss of the Grenville and in granite gneiss. No outcrop, however, exposes sufficient width to permit determination of the thickness of the mineralized zone or the relative amount of lean to barren rock in the zone. The maximum dip-needle values for the most part appear to be of moderate intensity.

The negative anomaly at the west is probably due to the presence of titaniferous hematite.

ANOMALY NO. 8

About a third of a mile northeast of bench mark 1901 on the Dobson trail, a compound magnetic anomaly 200 to 400 feet wide and 700 to 800 feet long is present. Parts of the anomaly are of high intensity. Only one outcrop was noted within the area of the anomaly. It shows coarse green pyroxene skarn and rusty-weathering, pyritic quartzose gneiss. The gneiss in part carries thin layers, veinlets, and disseminations of magnetite. Outcrops bordering the anomaly are of skarn, locally garnetiferous, and of quartzose feldspathic gneisses, locally pyroxenic. A little pink granite is present, but the dominant country rock belongs to the Grenville series. An outcrop on the border of the anomaly is in part garnet-pyroxene skarn with sporadic magnetite. Locally there is abundant float of rusty gneiss containing sparse to moderate amounts of magnetite.

ANOMALY NO. 9

Dead Creek anomaly No. 9 is the easternmost anomaly found in the area. It was located by dip-needle survey in 1949, in consequence of noting a peak of 5,540 gammas on the aeromagnetic map. The zone of magnetic anomaly extends for about 1,200 feet N. 55° W. within the +20° contour. The anomaly is divided in two by a creek valley. The northwest part of the anomaly is on a low ridge. The southeast part is the longer, about 700 feet within the +20° contour and 500 feet within the +40° contour. This anomaly has a fishhook shape, as if it occurred on the two limbs and nose of an isoclinal fold. The relation appears to be confirmed by such meager structural data as can be obtained from the few outcrops. At the northwest end of the anomaly, a layer of microcline granite gneiss striking northeast and dipping 25° NW. is in the appropriate position for the nose of a northwest-plunging anticlinal fold. Outcrops of ore on the northeast limb have a foliation dipping steeply northeast, opposite to the prevailing southwest dip of the major isoclinal fold. There are several outcrops of ore. They seem very rich in quartz and weather rusty brown. Some of the ore is rich in garnet.

DEERLICK RAPIDS PROSPECT

LOCATION, ACCESS, AND OWNERSHIP

The Deerlick Rapids anomaly centers 0.3 mile southeast of Deerlick Rapids, a ford in the South Branch of the Grass River, south-central rectangle, Stark quadrangle. (See pl. 1.) About 2 miles of the old trail running from the Cook Corners-Newbridge road northeast and north to Deerlick Rapids was made passable to jeeps in 1946 by the Bureau of Mines. The anomaly lies just east of the north end of the good trail, on Lots 11 and 12, Township 4 (Harewood) of Great Tract No. 2 of Macomb's Purchase, Town of Clifton, St. Lawrence County, N.Y. (Reed and Cohen, 1947, p. 9). The property is owned by the St. Regis Paper Co., and the mineral rights are under option to the M. A. Hanna Corp.

DISCOVERY AND EXPLORATION

The Deerlick Rapids magnetic anomaly and the Wolf Hole anomaly to the north (see p. 167) were registered on the aeromagnetic survey in 1945 (Balsley, Buddington, and others, 1954b). During the preceding winter, a party of the Bureau of Mines had put in a base line preparatory to making a reconnaissance dip-needle survey of a large block extending northward from Cook Corners (north-central rectangle, Cranberry Lake quadrangle) to the central Stark area, but the reconnaissance was never carried out. A ground check of the two anomalies was made by Hawkes in the fall of 1945 (Hawkes and Balsley, 1946, p. 4–5 and pl. 4). The following summer, the Bureau of Mines made detailed dip-needle surveys of both anomalies and drilled two holes on the Deerlick Rapids anomaly (Reed and Cohen, 1947, p. 9 and figs. 11 and 12). In 1950, the core was in the custody of the Hanna Ore Co.

MAGNETIC ANOMALIES AND OTHER INDICATIONS OF MINERALIZATION

The Deerlick Rapids anomaly is shaped like a fishhook (see pl. 12), with the "shank" beginning at the river and running south for 1,100 feet to the "hook," which curves eastward and extends almost to the river's edge. Negative anomalies flank all or part of the "hook" and form a "barb" at the east end. The aggregate length within the +40° isoclinal is 1,500 feet, of which the "shank" constitutes 950 feet. Here the anomaly is broad and simple and the overburden thick (50 feet at the site of hole 1). The complex pattern of highs and lows at the bend of the hook corresponds to an area with several exposures of magnetite-bearing gneiss, in part with some hematite.

GEOLOGY

ROCKS

Outcrops in the area were mapped by C. J. Cohen, geologist for the Bureau of Mines, during the dip-needle survey. The writer is indebted to Cohen for permission to use his unpublished data.

Outside the fishhook of intense magnetic anomalies is a belt of pyroxene-mica gneiss. Inside the hook is an area of sillimanite granite gneiss and garnet-biotite gneiss. Between the two and beneath the strongest anomalies is a series of granite gneiss sheets, parts of which contain appreciably more magnetite than is found in the weakly mineralized wallrocks. The granite gneiss is generally not a "clean" rock, for it contains leaves and layers of metasedimentary rocks. In the mineralized part of the granite gneiss, these layers are always a facies of the biotite gneiss, rarely carrying enough magnetite to be called low-grade ore.

The pyroxene-mica gneiss is gray green, fine to medium grained, poorly foliated, and massive or slightly blotchy. The mafic minerals, brownish mica and gray-green pyroxene, total 30 to 50 percent and vary in their proportion. Either may dominate, or the two may be roughly equal in amount. Certain mica-rich parts have sporadic garnet. Part of the rock is slightly migmatitic.

The biotite gneiss is gray, fine grained, and in places migmatitic, with dark minerals forming 10 to 40 percent of the rock. Some red garnet is almost always present; locally it is abundant. Certain zones are sillimanitic. Several percent of magnetite is usually visible.

The sillimanite granite gneiss is pink or bleached, and fine grained. In the cores, sillimanite is an abundant constituent. A little biotite may also be present. Garnet occurs sporadically. It is usually noted in conjunction with mineralized zones, and its aspect and association suggest that at least some of it is a manganiferous variety (spessartite). Here and there the rock is slightly reddened by hematite. It is likely that some of the disseminated black iron oxide is crystalline hematite.

The granite gneiss sheets beneath the main anomaly are fine grained, generally pink, but locally gray or white. Bleaching is more often seen in the magnetite-bearing parts, where also the grain size may be a bit coarser. The nonmetallic mafic mineral, where present, is biotite. Accessory garnet, magnetite, and sulfides occur sporadically. A few zones in the granite gneiss are garnet rich.

Relatively thin layers of granite pegmatite are infrequently met with in some of the rocks, and Cohen (personal communication) noted small hematite-quartz veins in a few outcrops. Drill cores often show vugs, a millimeter or two in diameter, scattered through the sequence. Chlorite, with or without pyrite, is found locally.

STRUCTURE

The rocks of the Deerlick Rapids area outline a fold that is inferred to have steeply dipping, closely appressed limbs. Only the western limb and the south end are exposed; fortunately these are also the only parts of the fold that are mineralized. The west limb dips about 70°–80° E. where it is cut by hole 1, and the south end of the structure is almost vertical. Lineation at the southeast end of the fold trends northeast and plunges 82°–90° NE. It is not possible to define the axis or the axial plane of the fold, or to say whether the fold is synclinal or anticlinal.

"ORE"

Disseminated magnetite partially replaces a number of thin layers in granite gneiss. Some included layers of biotite gneiss, and certain zones of the wallrock, also contain a substantial quantity of magnetite.

A little crystalline hematite accompanies the magnetite. The greater part of the hematite occurs as discrete grains that would be lost in magnetic separation. A variable but small quantity of hematite is present as martite within some magnetite grains; much or all of this would probably be recovered magnetically, as the martite forms only a minor part of a given magnetite grain. Accurate estimates of the proportion of hematite cannot be made from the specimens at hand. However, it appears that 5 to 10 percent of the iron oxide is hematite; of this, one-half or two-thirds would not be recovered during magnetic separation.

A very few ilmenite blades, in magnetite or in hematite, were observed in one sample.

Only a trace of pyrite is present in typical ore and subore. One or two thin zones of pyrite-chlorite alteration were cut in drilling, and this concentrated pyrite could contaminate otherwise sulfide-free material.

Almost all the core from both drill holes carries some magnetite and was assayed for total Fe by the Bureau of Mines. Reed and Cohen (1947, p. 33–34) give all the analytical data available.

Hole 1 cut one zone of marginal ore and several zones of subore. (See pl. 12.) Assays of composite samples of the marginal ore and one zone of subore are given in table 26 (data from Reed and Cohen, 1947, p. 33).

TABLE 26.—*Assays (weight percent) of marginal ore and one zone of subore, hole 1, Deerlick Rapids deposit*

Footage	Crude ore		Magnetic fraction only					
	Total Fe	Magnetic Fe	P	S	TiO_2	SiO_2	Mn	Fe
265–300	23.0	21.4	0.01	0.01	0.34	1.26	0.04	70.7
225–240	17.3	15.0	.01	.01	.54	2.44	.04	68.6

Hole 2 cut nothing of present commercial interest, though marginal ore was encountered at 240 to 244 feet, and subore at 257.5 to 264 and 314.5 to 320.5 feet. The interval 237–290 averaged roughly 17 percent total Fe (from data of Reed and Cohen, 1947, p. 34), most of which is in the form of iron oxides.

The relation between the mineralized cores of holes 1 and 2 is difficult to interpret because the holes are about 1,200 feet apart along strike. If we assume that the top of the pyroxene-mica gneiss cut at 354 feet in hole 1 is the same horizon as the top of the pyroxene-mica gneiss cut at 163 feet in hole 2, then the mineralized part of hole 1 is "stratigraphically" above the mineralized part of hole 2 (see pl. 12). On this assumption, hole 1 fails by about 10 feet to cut the granite sheet that is mineralized in hole 2. (At hole 1, however, there is no evidence in the magnetic anomaly that suggests magnetite concentrations any farther west than the projected bottom of hole 1.) Likewise, hole 2 must have been collared less than 10 feet "stratigraphically" below the sillimanite granite gneiss; hole 2 did not cut sillimanite granite gneiss, but outcrops of that rock are found 300 feet north of the hole.

One interpretation is that the mineralized zone cuts across the formational boundary represented by the top of the pyroxene gneiss. This is the interpretation adopted by Reed and Cohen (1947, p. 9) and by the present writer (cf. pl. 12). Transgression of rock boundaries by magnetite-bearing zones is well established at Skate Creek (p. 183) and has been observed at Benson Mines (McKee, 1944, oral communication).

A second interpretation is that mineralization has affected two "stratigraphically" different zones, in each case dying out along strike and giving two short, lenticular mineralized zones.

An equally valid interpretation is that the pyroxene-mica gneiss of hole 2, 163 to 233 feet, together with part of the biotite gneiss above it, is a lens that pinches out northward before reaching the latitude of hole 1. Obviously, we cannot know which interpretation is correct until holes are drilled between holes 1 and 2. The problem is raised here because of its bearing on the continuity of the mineralized zone.

ECONOMIC POSSIBILITIES

At present the property is of little interest commercially, owing to the marginal or submarginal grade of the magnetite-bearing rock. It will certainly deserve further investigation if the time comes when material averaging 10 to 15 percent magnetic iron can be extracted at a profit and the few percent of iron in hematite recovered as well, particularly if the structure proves to be a syncline. The possibility that the negative anomalies flanking the hook are given by hematite-bearing rocks must also be kept in mind.

WOLF HOLE ANOMALY

The mapped part of the Wolf Hole anomaly begins about 1,800 feet east of Wolf Hole, south-central rectangle, Stark quadrangle, and extends southward for 2,000 feet within the +10° isoclinal. It is essentially the northward extension of the Deerlick Rapids anomaly. A hunter's trail, running north from Deerlick Rapids, lies west of the Wolf Hole anomaly (see pl.12). Length within the +10° isoclinal is more than 2,000 feet (not closed at north end), and the width is 80 to 200 feet, narrowing toward the north. A strip of +20° readings, 300 feet long, is present in the central part of the anomaly, and the highest reading is +25°.

The economic possibilities of the anomaly are difficult to assess. The magnetic profiles suggest that the overburden may be as much as 100 feet thick, and a similar estimate may be made from a study of the topographic map. It seems likely that the mineralized zone causing the Wolf Hole anomaly consists of granite gneiss or sillimanite granite gneiss with disseminated magnetite. If the overburden is as much as 100 feet, the mineralized zone may be as good as, or even considerably better than, the north part of the Deerlick Rapids deposit. If the overburden is only 20 or 30 feet, the anomaly can scarcely have any economic interest.

GRIFFIN PARK (BENSON MINES EXTENSION) DEPOSIT

The Griffin Park (Benson Mines Extension) deposit, northeast rectangle, Oswegatchie quadrangle, is the northwest extension of the east limb of the Benson Mines deposit. It is an intermediate segment of the great mineralized zone that extends from Benson Mines through the Skate Creek deposit. (See pl. 1.)

To reach the deposit, turn north off the Browns Fall-Scotts Bridge road 0.2 mile northwest of Browns Bridge. Take the first right-hand road, leading through Griffin Park, and continue on foot or by jeep eastward about 1 mile. Here the tote road crosses the Griffin Park anomaly. Another "loop" of the tote road lies about 0.3 mile to the north.

Parts of the Griffin Park deposit were diamond-drilled by the Jones & Laughlin Steel Corp. in 1943. The deposit subsequently was registered as a long magnetic ridge, intensity 800 gammas, on the aeromagnetic survey of 1945 (Balsley, Hawkes, Buddington, and others, 1946). Relative to the datum subsequently adopted for aeromagnetic maps of adjoining quadrangles, the intensity of this magnetic ridge is 6,000 gammas.

Dip-needle traverses were run to check the character of the anomaly, which is everywhere well defined but variable in intensity. In part, this variation is due to the varying thickness of overburden, though variations in grade of the concealed magnetite-bearing zone would be expected. Peak dip-needle values on traverses spaced 500 to 1,000 feet apart ranged from +16° to +48°. The strongest segment of the dip-needle anomaly appears to be one extending about 2,000 feet northwest of the Oswegatchie River. (See pl. 1.) A negative anomaly, ranging from weak but distinct (−4°) to strong (−55°) is consistently present along the northeast and north side of the positive anomaly—that is, on the scarp side of the bedrock foliation. Where the anomaly crosses the dirt road at a point 0.75 mile north of Browns Bridge, the values are slight (+17° peak), but the axis of the anomaly remains clearly defined. Incidentally, blacksand (magnetite and other heavy minerals) is locally abundant along the road.

The source of the anomaly is well exposed where the Oswegatchie River cuts across the Benson Mines Extension. (See pl. 1.) An outcrop on the north shore shows subore or low-grade ore in sillimanite-biotite granite gneiss. The magnetite content generally ranges from 5 to 15 percent by volume, with some 1-foot zones of low-grade to medium-grade ore. A little gray hematite is present locally, as are sulfide-bearing zones. A small mass of schorl-biotite granite pegmatite is barren. At one place, a narrow stringer of magnetite cuts across the strike of the foliation at 45°, also transecting a minor drag-folded slip surface. Elsewhere, the magnetite-bearing zones are essentially conformable to the foliation of the gneiss.

The wallrocks of the magnetite-bearing zone are exposed at several places along the anomaly. Footwall rocks include pyroxene-quartz-feldspar gneiss, in part migmatitic, and microcline granite gneiss. Hangingwall rocks include weakly mineralized sillimanite-microcline granite gneiss with garnetiferous pegmatite seams; migmatitic, schorl-rich biotite gneiss; and migmatitic, quartz-rich garnet-biotite gneiss, locally sillimanitic.

The general structure of the magnetite-bearing zone is shown on plate 1.

This account attempts merely to give a brief general description of the anomaly, its source, and its wallrock. The geologic relations at the deposit are very similar to those found at Skate Creek.

JARVIS BRIDGE DEPOSIT

SUMMARY

The Jarvis Bridge deposit, explored to an average depth of 350 feet by 9 drill holes, lies beneath the Oswegatchie River northeast of Newton Falls. Overburden is generally 40 to 80 feet thick southwest of the linear magnetic anomaly given by the magnetite deposit; above the suboutcrop of the ore zone, the overburden is inferred to be thinner. The reserve of ore averaging 23.8 percent magnetic Fe is inferred to be sizable. If the average grade of ore is reduced to 19.7 percent magnetic Fe, the inferred reserve is almost doubled. The ore body, crudely tabular but undulating, ranges in thickness from 9.2 to 34.7 feet, averaging 20.5 feet, and has a strike length of 4,500 feet or more. The body, which is confined to a single zone within a sheet of heterogeneous microcline granite gneiss, is inferred to lie on the southwest limb of a slightly overturned isoclinal anticline. An exceptionally well defined sequence of wallrocks (part of the Clare-Clifton-Colton belt of metasedimentary rocks and granite sheets) may ultimately be useful in working out the structure of the deposit.

LOCATION AND ACCESS

The Jarvis Bridge deposit centers 1.5 miles east-northeast of Newton Falls, northwest rectangle, Cranberry Lake quadrangle. Most of the deposit is under the Oswegatchie River. A road bridge, locally known as Jarvis Bridge, crosses the river near the northwest end of the main part of the ore deposit and serves to name the deposit. To reach the bridge, drive about 1.4 miles east-northeast from Newton Falls along the gravel road that runs to Cook Corners and Cranberry Lake.

DISCOVERY AND EXPLORATION

The Jarvis Bridge deposit, owned by the Newton Falls Paper Mill, was discovered by A. M. Ross in 1942. Ross was attempting to demonstrate the effect that the ironwork of the bridge exerted on a dip needle when he inadvertently detected the magnetic anomaly produced by the underlying magnetite deposit. A reconnaissance dip-needle survey by Ross was followed in 1943 and 1944 by detailed surveys by the U.S. Bureau of Mines. Most of the detailed work was done while the river was frozen over. In 1944 and 1945, the Bureau of Mines explored the deposit to an average depth of about 350 feet by means of 9 diamond-drill holes. A tenth hole (DDH 6) was abandoned in rotten bedrock just beneath the overburden. The drilling was designed to test the character and thickness of the magnetite deposit, not to prove up tonnage. Results of the Bureau's work have been summarized by Millar (1947).

Drilling the initial holes through overburden was so slow that the Bureau of Mines, under the direction of J. D. Bardill, applied the technique of freezing the unconsolidated material with kerosene cooled by dry ice (Millar, 1947, p. 7). Coring the frozen ground, one drill hole pierced 68 feet of overburden in 4 hours of running time (5 hours, counting delays). Drilling an equivalent depth of overburden by conventional methods had required about 3 weeks for the initial holes.

MAGNETIC ANOMALIES

The only "surface" indication of the Jarvis Bridge deposit is the magnetic anomaly. (See pl. 13.) The anomaly was registered as a prominent peak of 6,790 gammas on the aeromagnetic survey of 1945 (Balsley, Buddington, and others, 1954a), 3 years after the preliminary dip-needle survey. Neither ore nor wallrock is normally exposed to view. However, it is said that when the river is very low one can look down through clear water and see ore on the river bottom southeast of the bridge, approximately at (9000N, 10000E). It is also said that one outcrop of footwall rock is visible on the island 100 feet northeast of the bridge, and that in exceptionally dry seasons footwall rocks are exposed at (9150N, 10800E) (Ross, 1952, written communication). Aside from these ephemeral signs of bedrock, no outcrops have been found within several thousand feet of the Jarvis Bridge deposit.

Geophysicists sometimes make a rough estimate of the depth of overburden by using the half-width of the half-anomaly. Using this approach, we find that the overburden beneath the magnetic anomaly appears to be thinner than it is at the drill sites, southwest of the anomaly. (See cross sections, pl. 13.)

The main Jarvis Bridge magnetic anomaly and its northwest extension resemble a narrow crotched stick or boy's slingshot (see pl. 13). The main anomaly, corresponding to the handle, begins 350 feet northwest of Jarvis Bridge and extends southeastward for 4,400 feet. The northwest extension, corresponding to the forks of the slingshot, extends northwestward more than 3,000 feet from the crotch. The pattern is given by the narrow area within the +10° isoclinal contour, which becomes discontinuous toward the northwest end. The left (southwest) fork of the anomaly seems to be continuous with the handle, or main anomaly, whereas the right (northeast) fork is not. The possibility that the forking anomaly reflects two limbs of an isoclinal fold is discussed in the section on structure.

Within the handle and left fork of the anomaly, a belt of readings effectively within the +20° isoclinal is 4,500 feet long. This belt is roughly bracketed by holes 10 and 8. A narrow belt of readings effectively within the +40° isoclinal begins at the crotch of the anomaly and extends about 2,500 feet southeastward to coordinate 11000E. This belt corresponds closely to the limits of the Jarvis Bridge ore deposit. Within the +40° isoclinal are discontinuous strips and beads outlined by the +60° and +80° isoclinals. The magnetic profile of the main anomaly is asymmetric, showing steep northeast and gentler southwest slopes. (See cross sections, pl. 13.) The steepness of the northeast slope of the magnetic profile is accentuated by the presence of low negative readings along the northeast edge of the main anomaly. These negative readings range from −1° to −9°; hence they do not appear on the contoured magnetic map (pl. 13).

An area of low positive readings extends several hundred feet southwestward from the southwest edge of the anomaly. The zero isoclinal that should bound these readings lies beyond the limits of the area covered by the dip-needle survey.

The right (northeast) fork of the anomaly shows 4 small beads of +20° readings between coordinates 8000E and 8850E. This fork of the anomaly appears to hold no economic interest. The anomaly is produced by scattered thin layers of magnetite in biotite gneiss that is locally garnetiferous. The magnetite-bearing zone, parts of which might run 6 to 10 percent magnetic Fe, lies about 200 feet stratigraphically deeper than the main mineralized zone of the Jarvis Bridge deposit.

The northwest extension of the Jarvis Bridge anomaly appears to be virtually continuous with the Spruce Mountain Southeast anomaly (see p. 188). The coordinate systems for the two separate surveys are different, and the means of relating them are not at hand. The plot of the anomalies on the regional map (pl. 1) is the best that could be made from the available data. It shows that the projection of the northeast fork of the Jarvis Bridge anomaly would meet the Spruce Mountain Southeast anomaly. This relation may not exist at all; reduction of the magnetic maps to a scale of 1:62,500 may appear to have eliminated some very real discrepancies. The effective continuity of the Jarvis Bridge and Spruce Mountain Southeast anomalies seems clear, but the details are unfortunately lacking.

The swampy area between the Jarvis Bridge anomaly and the Jarvis Bridge Extension anomaly has been adequately surveyed by dip needle. Low positive readings (less than +10°) extend between these two major anomalies, effectively linking them with each other. As no outcrops have been found in this part of Chaumont swamp, and the Jarvis Bridge Extension has not been drilled, a demonstration of geologic continuity between the two anomalies is not now possible.

GEOLOGY

The Jarvis Bridge deposit and its associated magnetic anomalies lie in the Clare-Clifton-Colton belt of metasedimentary rocks with interleaved sheets of alaskite and microcline granite gneiss. (See pl. 1.) An elongate body of albite-oligoclase granite, half a mile northeast of Jarvis Bridge, parallels the magnetite deposit on the footwall side. In the Jarvis Bridge area, the CCC belt trends southeastward, flanked by masses of hornblende-microperthite granite. The gross structure of the belt is synclinal, but minor anticlines are almost certainly present. The dip of axial planes is unknown; at the latitude of Jarvis Bridge, the dip may be $90° \pm 25°$. Minor fold axes and other lineations trends southeastward, mostly parallel to the borders of the CCC belt, and plunge southeastward at various angles. The summary of structural data is necessarily vague and unsatisfactory because of the lack of outcrops in this critical area.

ROCKS

A definite and readily recognized sequence of rock units is present at Jarvis Bridge. The continuity of most of these units for 4,000 feet along the strike suggests that the sequence, as found in the area drilled, is on the limb of a large fold, rather than in the apical zone of a fold. A knowledge of the stratigraphic sequence is helpful in avoiding unnecessary drilling into footwall rocks in places where the normally ore-bearing zone is only weakly mineralized, as in hole 10. Knowledge of the sequence can also be used to insure that a given drill hole is not stopped short in the hanging wall. While these two problems are relatively insignificant at the shallow depth to which the deposit has been explored, they may become extremely important in finding ore and working out structure if deeper drilling is undertaken.

Three principal rock units serve as a framework within which the rest may be described. These principal units are an upper marker (pyroxene-brown garnet-microcline granite gneiss), a lower marker (biotite gneisses and granulites in which red garnet is characteristic), and the mineralized zone. All rock units will be described in order from the apparent top downward. (If the sequence has been overturned, this order might be the reverse order of deposition.) The units are shown on plate 13, in plan and section. In more detail, their correlation is shown in a separate chart (pl. 14). Only the major units are discussed; actually, many of these are capable of much finer subdivision.

Granite pegmatite, locally syenitic, is present as thin seams and minor layers in rocks of most of the units. Quartz veins are rare.

HANGING-WALL ROCKS ABOVE UPPER MARKER

This unit comprises biotite gneiss (locally pyroxenic) and quartz-feldspar granulite, a 30-foot zone of diopsidic marble, and some interlayered pyroxene gneiss and marble. A 25-foot sheet of pink microcline granite gneiss is also present. Near the base of the unit is a sheet of sillimanite-microcline granite gneiss, 15 to 31 feet thick. At the base of the unit is a zone of calcareous pyroxene-brown garnet gneiss, in part scapolitic, locally with granite pegmatite and subordinate biotite gneiss. This basal member is 9 to 29 feet thick. Rocks of this hanging-wall unit were cut only in holes 5 and 8. Perhaps the basal member was also cut in hole 7; if so, it was not recognized, owing to the extreme alteration and leaching of the "rotted" bedrock beneath the overburden.

UPPER MARKER

The upper marker is a pink, fine- to medium-grained, indistinctly or crudely foliated microcline granite gneiss carrying green pyroxene and brown garnet. The garnet (a member of the grossularite-andradite series; $n = 1.822 \pm .002$, suggesting a composition of $Gr_{46}An_{54}$ with an unknown amount of other garnet molecules) and pyroxene commonly occur together as shreds, clots and blotches of recrystallized but still relatively fine-grained skarn; in part, however, the garnet and pyroxene are rather homogeneously incorporated in the granite gneiss. Both mafic clots and granitic host rock are calcareous in places; occasionally, a little calcite is found throughout. Epidote is present sporadically. Minor layers of granite pegmatite are found here and there.

Two types of calcite occur in the granite gneiss. One is interstitial recrystallized carbonate, more or less polygonal in shape, presumably inherited from metasedimentary rocks. The other is secondary calcite in the form of pervasive alteration, minute replacements and veinlets, and rare coarse blotches and veins. A good part of the secondary carbonate may be supergene, leached out of the overlying marble layer or the upper part of the granite gneiss itself. Some of the secondary carbonate may be hypogene. The problem is further complicated by the "rotting" and leaching of the suboutcrop of the rock.

The origin of the pyroxene-brown garnet-microcline granite gneiss poses a problem that cannot be solved at present. The rock has been seen only in drill cores; hence the broader features of field occurrence are unknown. The rock may have been derived—through injection or potassium metasomatism—from metasedimentary rocks similar to the pyroxene-brown garnet gneiss immediately above the upper marker, from skarn, or from a metasedimentary gneiss that had thin layers, lenses, or concretions of skarn.

The upper marker, cut in all holes except 10, 9, and 4, is 55 to 75 feet thick. Its value as a marker is twofold: it is unique and readily identified; and it has a nearly constant position above the top of the ore zone. The stratigraphic distance between the bottom of the upper marker and the top of the ore zone averages 180±20 feet. The distance ranges from 214 feet to 132 feet, decreasing southeastward along the strike.

PYROXENE-BIOTITE UNIT BELOW UPPER MARKER

The unit below the upper marker consists largely of rocks carrying pyroxene and biotite, or pyroxene alone as the characteristic mafic mineral. The rocks comprise pyroxene-biotite gneiss, slightly to moderately migmatitic; modified pyroxene-biotite skarn; subordinate pyroxene skarn; and local quartz-feldspar granulite, feldspathic biotite quartzite, and light-green fine-grained pyroxene gneiss. The pyroxene and pyroxene-biotite rocks are greenish, gray green, or brownish green, generally medium grained, and except for the skarn, distinctly foliated. Though the feldspar in the gneisses is a potassic variety, it is commonly white or gray rather than pink. The quartz content is variable. It is possible that some or all of the migmatitic pyroxene-biotite rocks represent migmatized pyroxene skarn; such rocks may have replacement augen of feldspar, instead of thin layers of felsic minerals. Some unmigmatized skarn has been modified by the introduction of quartz and scapolite. Brown sphene is locally prominent in fine pyroxene gneiss and quartz-feldspar granulite; megascopically, it is easily mistaken for brown garnet.

The unit ranges in thickness from 40 to 73 feet; its average thickness is 50 feet. The thickness of the unit decreases southeastward along the strike.

LOWER MARKER (RED GARNET UNIT)

Disseminated small red garnets (presumably an almandite with an unknown but appreciable mixture of other garnet molecules), accompanied by biotite, are characteristic of the lower marker.

The principal rock type is garnet-biotite gneiss, gray, fine to medium grained, and well foliated. Both quartz and feldspar are present; the feldspar may be altered plagioclase (a sodic oligoclase), microcline, or untwinned potassic feldspar. In places, all three feldspars are present. Locally, considerable very fine pumpellyite is present. Chiefly in hole 7, the rock is highly quartzose—actually a feldspathic quartzite.

A discontinuous lens of once-calcareous metasedimentary rocks splits the garnet-biotite gneiss of the lower marker in holes 9, 4, and 3. Red almandite is generally absent from rocks of the lens. The once-calcareous rocks are represented chiefly by pyroxene gneiss containing a little biotite, or a little brown garnet, calcite, and scapolite. Locally, other pyroxenic rocks are also present. These include pyroxene-quartz-feldspar granulite, pyroxene skarn, modified pyroxene skarn, and pyroxene-microcline granite gneiss.

The lower marker is highly variable in thickness, but its stratigraphic distance above the ore zone is relatively constant. The lower marker thins and thickens along the strike, depending on the absence or presence of lenses of pyroxenic rocks. The marker has an average thickness of 34 feet, a maixmum of 70 feet (hole 9), and a minimum of 9 feet (hole 8). Where present, the lenses of pyroxenic rocks have an average thickness of 23 feet and a maximum thickness of 40 feet (hole 3).

The stratigraphic distance from the base of the lower marker to the top of the ore zone averages 95 feet. The maximum is 118 feet (hole 9); the minimum, 74 feet (hole 3). Except for minor variations, the decrease is rather uniform from northwest to southeast along the strike.

PYROXENE-BIOTITE UNIT BELOW LOWER MARKER

Rocks of this unit are similar to those between the upper and lower markers, except that some amphibole commonly accompanies the pyroxene or proxies for it. In addition, there are present discrete layers of gray, fine- to medium-grained, slightly migmatitic biotite gneiss devoid of pyroxene, and layers of migmatitic pyroxene gneiss devoid of biotite. A variable amount of microcline granite gneiss, carrying pyroxene, biotite, hornblende, or combinations of these mafic minerals, is also found in this unit.

The unit is highly variable in thickness, ranging from 24 to 98 feet and averaging 68 feet. Most of this variation is displayed northwest of hole 7; southeast of hole 1, the unit has a thickness of 71±10 feet.

GRANITE GNEISS SHEET, GENERALLY ORE-BEARING

A sheet of microcline granite gneiss, generally ore bearing, underlies the pyroxene-biotite unit just decribed. As one usually finds in the district, the granite gneiss is heterogeneous. The central part of the sheet (holes 1 to 5) has biotite or garnet and biotite as mafic minerals; the ends of the sheet (holes 10, 9, 7 at the northwest, and hole 8 at the southeast) have pyroxene, amphibole, and subordinate biotite, with or without almanditic garnet accompanying the biotite. This variation in the type of mafic mineral is due to the difference in the metasediments invaded and replaced by granitic material, yielding the inhomogeneous microcline granite gneiss. A fairly thick layer of biotite gneiss, present in the central area, was converted to biotite-microcline granite gneiss, commonly garnetiferous, without appreciable attack on rocks of the overlying pyroxene-biotite unit. Locally, some quartz-feldspar granulite may also have been converted to microcline granite gneiss. Apparently the layer of biotite gneiss thinned and frayed out toward the northwest and southeast, for in these areas only the middle of the granite gneiss sheet is biotitic; the rest has pyroxene or amphibole, with subordinate biotite, as mafic constituents. However, at the ends of the thinned sheet a few relics of biotite gneiss, locally garnetiferous, have not been converted to microcline granite gneiss. This suggests that some feature other than simple lensing out of biotite gneiss controlled the emplacement of the granite gneiss sheet. The effect of the unknown processes has been to yield a sheet that is gently undulating and distinctly discordant with respect to the known stratigraphic sequence.

The microcline granite gneiss is fine to medium grained, generally distinctly foliated, and commonly pinkish (except where the rock has been replaced by magnetite). The facies carrying pyroxene and amphibole may be blotchy, rather than foliated. Seams and thin layers of contaminated granite pegmatite are sparsely developed. Here and there, streaky layers and "schlieren" of metasediments are preserved within the granite gneiss.

The known thickness of the granite gneiss sheet ranges from 30 to 134 feet. Because several drill holes bottomed in granite gneiss, it is not possible to give a reliable figure for the average and maximum thicknesses of the sheet. However, the sheet is generally more than 50 feet thick; it definitely thins northwestward, and it probably thins southeastward as well.

Features diretly relating to ore and mineralization are described in the section on ore.

FOOTWALL UNIT

Information on rocks underlying the microcline granite gneiss sheet is spotty. In all holes that went deep enough, the upper member of the footwall unit is a layer of quartz-feldspar granulite, 7 to 17 feet thick. (This member was not recognized in cores from hole 1, the first to be logged. Perhaps the member was overlooked, or misidentified as a facies of the granite gneiss.) Locally, a thin layer of pyroxene gneiss or modified pyroxene skarn separates the quartz-feldspar granulite from the overlying microcline granite gneiss. The quartz-feldspar granulite is pinkish, very fine grained, and outwardly massive. It contains abundant quartz, a very little biotite, and usually a little garnet.

Also present in the footwall unit are rather thick layers of biotite gneiss and pyroxene gneiss, in part hornblendic, showing a variable but generally slight degree of migmatization. Locally the biotite gneiss carries garnet or sillimanite. Amphibolite, either micaceous or pyroxenic, occurs sporadically. A little skarn is present locally. Microcline granite gneiss and granite pegmatite are sporadic and not abundant; all layers are thin. Relatively less microcline is present in the metasedimentary gneisses, and the plagioclase (dominantly An_{35}) is mostly fresh.

In general, the rocks of the footwall unit are very similar to those found above the granite gneiss sheet. From the evidence now at hand, it is impossible to say whether any of the upper units are equivalent to those in the footwall, indicating isoclinal folding. That possibility exists, and its significance is discussed under structure, below.

CORRELATION OF ROCK UNITS

The correlation of all major rock units is shown on plate 14. For all holes except 10, the datum for correlation is the top of the ore zone. Hole 10 carries only 5 percent magnetite, or less, in the "stratigraphic" position that carries ore in the other holes. Consequently, the footwall quartz-feldspar granulite has been used to correlate hole 10 with hole 9. The reader will see (pl. 14) that other members can also be correlated from hole to hole; still finer subdivision is possible but has not been attempted here.

Correlation of the Jarvis Bridge sequence with sequences in adjacent areas is an unsolved problem. The sequence of rock units in the Spruce Mountain area is shown on plate 18. It is possible that the base of the footwall biotite gneiss in hole 9, Jarvis Bridge, corresponds to the base of the lower mineralized zone (top of lower, generally pyroxenic, unit) at Spruce Mountain; that is, the Spruce Mountain middle unit corresponds to all Jarvis Bridge rocks between the top of the lower marker (almandite zone) and the base of the thick layer of footwall biotite gneiss. If so, the main ore zones at the two deposits are by no means stratigraphically equivalent.

Several other correlations are possible. For this reason, the writer doubts the value of any one of them. Extremely careful comparisons of subunits and members in the sequences from Jarvis Bridge and Spruce Mountain might yield a more reliable correlation. An intensive study of this sort, though not possible at present, could be made from the drill logs and reexamination of all drill core, if the economic significance of the correlation problem should become acute.

STRUCTURE

The rocks enclosing the Jarvis Bridge deposit strike N. 50°–55° W. and dip 55°–70° SW. The uniformity of strike, and the apparent uniformity of dip, suggest that the rocks in the area explored by diamond drilling are on the limb of a fold, rather than in the apical zone of a fold. Dips shown on the cross sections (pl. 13) have been projected from measured angles between foliation and axis of drill cores. Where the core-foliation angles vary significantly in cores from a single hole, some compromise has been necessary in projecting the dip of rock units upward to the inferred bedrock surface. The compromise is based on the position and dip of nearby rock units, assuming that each unit maintains a nearly constant thickness up the dip. (This may not be true at all, but it is one of the few permissible assumptions.) The axis of the magnetic anomaly has not been used (except at hole 4) to approximate the suboutcrop of the ore zone; projected dips have been used instead. Thus the dips shown on cross sections are accurate only for a few feet above and below the drill hole; elsewhere they are schematic but portrayed as objectively as possible. Other interpretations are possible; these are discussed below.

We know that isoclinal folding is common in the district. Several features suggest that one or more isoclinal folds, in part disrupted by flowage and rupture, may be present even in the limited area explored by drilling. These features are: the possible duplication of hanging-wall and footwall sequences, the thickening of the microcline granite gneiss sheet northeast of holes 7 and 1, the forking of the magnetic anomaly, apparent variations in the dip of rock units, and the lensing of certain members of the sequence. Unfortunately, these features, alone or in combination, are merely suggestive, not definitive.

Holes 9 and 10 are the only ones that continued far into the footwall. Here, the presence of biotite gneiss, lenses of pyroxene gneiss, and thin zones of biotite gneiss with garnet or sillimanite, suggests a sequence similar to that in the hanging wall, especially as shown in hole 9. The writer has already emphasized that the equivalence of any unit or units in hanging wall and footwall cannot be demonstrated; the equivalence is suggestive. If holes 9 and 10 had been drilled about 150 feet farther, it might have been possible to demonstrate or disprove this possible equivalence. This assumes that a unit similar to the upper marker might have been encountered by deeper drilling.

Northeast of holes 7 and 1, the sheet of microcline granite gneiss thickens markedly. (This assumes that the inferred contact between granite gneiss and footwall gneisses has been correctly placed in interpreting data from the other holes, several of which bottomed in the granite gneiss or went only a short distance into the footwall gneisses. Cf. pl. 13.) The thickening might be due to introduction of an abnormal quantity of granitic material; if so, the angle of discordance between granite gneiss and its footwall would be very large in the short space between holes 7 and 1. Alternatively, the abnormal thickening might be due to folding (duplication) of metasediments before, during, or after emplacement of the granite gneiss sheet. Such duplication is tentatively assumed and shown on the plan (pl. 13). This assumption is consistent with the forking of the magnetic anomaly northeast of hole 1. However, the forking of the anomaly may merely reflect a weakly mineralized shear(?) zone of footwall rocks making a slight angle to the main zone. Much drilling, with no expectable economic reward, would be necessary to check the various assumptions.

If isoclinal folding exists at Jarvis Bridge, apparent horizontal displacement of units by minor folds could wipe out the simpler "bend folds" now shown on the cross sections (pl. 13).

Several members of the rock sequence show as lenses on the map (pl. 13). This is a conventional and objective treatment of the data on distribution of those members. If the entire sequence has been isoclinally folded and locally disrupted, the lenses of pyroxenic gneiss may represent large boudins; the thinner intervening biotite gneiss layers may represent the attenuated zigzag limbs of folds.

If a large isoclinal fold is present at Jarvis Bridge, where does the ore zone lie with respect to the fold? The ore zone is very probably on the limb, not in the apical zone. Judging from the possible duplication of the rock sequence, the inferred fold axis should lie just northeast of the ore zone and parallel to it. If the forking magnetic anomaly indicates the nose of a fold, and if the axis of the fold plunges southeastward (see regional setting), the fold is an anticline. Only a reversal of the regional plunge would make the structure a syncline. (Drill cores yielded no *b* lineations; hence there is no check on the inferred southeastward plunge of the major fold axis. Lineation was observed on a single piece of core. This lineation, trending down the dip of the foliation, is presumably in the *a* direction). Unfortunately, the variation in the dip of the foliation is so slight and unsystematic throughout cores from holes 10 and 9 that one cannot use the foliation data to check the inference that the ore zone lies on the southwest flank of a slightly overturned isoclinal anticline.

ORE

OCCURRENCE

The ore at Jarvis Bridge is typical of the granite gneiss type: low grade, disseminated or streaky, uniform, and persistent. Its host is the main sheet of heterogeneous microcline granite gneiss. A little magnetite, locally as much as 5 or 10 percent, is found sporadically in many other rock types in the hanging wall and footwall, but all the material of ore grade is confined to a single and apparently continuous zone in the granite gneiss. The walls of the ore zone are not sharp: a bit of subore and weakly mineralized rock normally fringes the ore.

So far as we can infer at present, the ore body is essentially sheetlike or tabular in form—generally conformable to the structure of the enclosing rocks, but distinctly undulatory within the granite gneiss sheet. In plan, proceeding from northwest to southeast along strike, the ore zone migrates from the footwall side of the granite gneiss sheet to the middle and then to the hanging-wall side, with very slight undulations along its course. Unavoidable errors in projecting ore from drill holes to the inferred bedrock surface could conceivably accentuate this wavy pattern, but they cannot explain it away. Precisely the same undulation is shown on the correlation diagram (pl. 14), where the angle of dip enters only into calculations of thickness. Bear in mind that the datum for correlation—top of the ore zone—is shown as a straight line for convenience only; if the middle of the granite gneiss sheet were adjusted to fit a horizontal line, the undulation of the ore zone would be emphasized.

The orientation of the axes of these warps in the ore zone is unknown. Conceivably, the warps might determine stronger and weaker shoots within the ore zone. Whether the warp axes are parallel to, or at a large angle to, the axis of the inferred major anticline, we cannot say. The warp axes might parallel the *a* lineation observed on one piece of drill core. If so, the ore body could plunge essentially downdip. The beadlike character of highs in the magnetic anomaly has been taken by some to indicate a shoot structure in the ore. However, effectively tabular bodies of magnetic material may yield similar beadlike anomalies.

MINERALOGY

The ore is chiefly a mixture of magnetite, subordinate sulfides, and the unreplaced minerals of the granite gneiss host. Modes of typical specimens of ore are given in table 6, Nos. 1, 2, 9, 14, and 15. Regarding the gangue, we may note that the proportion of quartz to feldspar varies widely, that plagioclase is sparse or absent, and that the potassic feldspar is almost exclusively the untwinned variety in all specimens of ore grade. Among the accessories, allanite seems to be more widespread in ore than in unmineralized rock, and apatite is distinctly more abundant in ore. Though biotite, commonly accompanied by red garnet, is the usual mafic mineral, the ore and subore from the ends of the granite gneiss sheet carry pyroxene, amphibole, and biotite—a reflection of the mafic minerals inherited by the granite gneiss sheet. In hole 8, the granite gneiss sheet contains thin layers of garnet skarn, garnet-pyroxene skarn, and modified skarn. The skarns carry very little magnetite, though granite gneiss contaminated with skarn minerals contains subore locally.

Metallic minerals replace the nonmetallic irregularly. The only metallic minerals recognizable in hand specimen are magnetite, the iron sulfides, and chalcopyrite. Among these, magnetite is the dominant mineral and the only one of economic value. Exsolution blades and minute grains of ilmenite are commonly present in the magnetite, but the total quantity of ilmenite is negligible. (Cf. assays for TiO_2, table 28.) Neither martite nor primary crystalline hematite has yet been found, though rarely one finds a trace of hematite exsolved from ilmenite blades. Pyrite and pyrrhotite, singly or together, are common and locally prominent. Chalcopyrite is common but sparse, and sphalerite is present as mere traces. Very rarely one finds a trace of covellite in chalcopyrite.

Alteration of nonmetallic minerals to chlorite, carbonate, zoisite, pumpellyite, and clay minerals is slightly more evident than is normal in the country rock. However, even this alteration is weak and sporadic.

GRADE

Table 27 gives the grade and thickness of all assayed zones of ore, subore, and mineralized rock. On some cross sections (pl. 13), the intervals for ore and subore differ slightly from those given in table 27. The slight discrepancy is due to the difference between the intervals as they show in the cores, and the convenient intervals adopted for sampling by the Bureau of Mines. The ore zone (23.8 percent magnetic Fe) ranges in thickness from 9.2 to 34.7 feet and averages 20.5 feet. If the grade is reduced to 19.7 percent magnetic Fe, the resulting zone of subore (marginal ore) ranges in thickness from 16.4 to 48.0 feet and averages 29.9 feet.

The average composition of material of ore grade is very similar to that of Benson Mines ore (cf. table 7). At Jarvis Bridge, the ratio of magnetic Fe to soluble Fe averages 0.906; the ratio of magnetic Fe to total Fe averages 0.868. Though the Benson Mines magnetite ore normally carries a little soluble nonmagnetic Fe in the form of crystalline hematite, the soluble non-

TABLE 27.—*Grade and thickness of ore zones and principal zones of subore and mineralized rock, Jarvis Bridge deposit*

[Assays marked with an asterisk (*) are from Millar (1947, p. 7-10). Other assays are from unpublished data of U.S. Bureau of Mines]

Hole	Footage	Apparent thickness (feet)	True thickness (feet)	Weight percent						
				Total Fe	Soluble Fe	Magnetic Fe	P	S	TiO_2	SiO_2
1	345–365	20	19.4	31.4	*29.9	*25.9	0.254	1.10	0.63	37.5
	365–380	15	14.5	14.4	*13.1	*8.9	.120	.55	.54	57.5
2	332–355	23	21.8	29.3	28.1	[1]25.5	.33	1.63	.52	39.0
3	350–368	18	16.7	----	26.8	----	----	----	----	----
	368–370	2	1.9	----	9.2	----	----	----	----	----
	350–370	20	18.6	----	25.0	22.3	.28	1.52	.61	47.5
4	235–239.1	4.1	3.2	----	*8.4	----	----	----	----	----
	239.1–265.8	26.7	20.5	----	*24.8	23.1	.28	1.48	.42	44.7
	265.8–270	4.2	3.2	----	*6.9	----	----	----	----	----
5	547–548.7	1.7	1.2	----	*9.7	----	----	----	----	----
	548.7–591.5	42.8	30.4	----	23.2	[1]21.0	.29	1.39	[2].16	44.0
	591.5–595	3.5	2.5	----	*12.4	----	----	----	----	----
6	([3])									
7	497–500	3.0	2.1	*5.6	----	----	----	----	----	----
	500–548.9	48.9	34.7	*27.7	26.3	{[4]22.3 [5]23.2}	.33	1.19	.35	42.3
	548.9–553	4.1	2.9	*17.2	----	----	----	----	----	----
8	657.4–658.8	1.4	1.2	*10.7	*10.5	*7.9	----	----	----	----
	658.8–669.8	11.0	9.7	27.6	26.4	25.4	.26	1.57	.35	38.9
	669.8–670.6	.8	.7	*2.6	2.2	----	----	----	----	----
	670.6–712	41.4	36.4	18.5	16.5	13.5	.30	.97	.40	42.3
9	315–317	2.0	1.6	*12.9	----	*9.3	----	----	----	----
	317–328.2	11.2	9.2	*27.3	----	*24.5	----	----	----	----
	328.2–335	6.8	5.6	*18.4	----	*14.2	----	----	----	----
10	376.3–386.3	10.0	7.7	*10.8	----	*8.5	----	----	----	----
Average composition of material of ore grade, holes 1–5, 7 and 8				<29.0	26.2	23.8	.29	1.41	[6].43	42.0

[1] Estimated by Leonard. Value based on average magnetic Fe : soluble Fe ratio of 0.906.
[2] Value seems much too low.
[3] Abandoned just beneath overburden.
[4] Weighted average for 10 individual samples whose magnetic Fe content is reported by Millar (1947, p. 8-9), who erroneously gives 21 percent magnetic Fe as the value for the composite sample, 500–548.9 feet.
[5] Value determined directly on composite sample, 500–548.9 feet.
[6] Value may be a little low.

magnetic Fe in the Jarvis Bridge ore probably cannot be attributed to hematite. Significant minor amounts of that mineral have not been found in polished sections of Jarvis Bridge ore. (Sillimanite granite gneiss in the hanging wall unit carries a few percent iron oxides—both hematite and magnetite—but this rock does not constitute ore and is not near the ore zone.) This suggests that the 2.4 percent of soluble nonmagnetic Fe in average ore comes from pyrrhotite and from partial solution of biotite(?), pyriboles, and perhaps garnet and other nonmetallic minerals. Unless it is subsequently demonstrated that hematite is present in the ore, we may assume that 2.4 percent soluble nonmagnetic Fe cannot be recovered.

Table 28 gives analyses of the magnetic fraction, or Davis tube concentrate, of ore and subore. As we might expect, the magnetic fraction is essentially uniform, regardless of the grade of the crude material. Sulfur varies, depending on the ratio of pyrrhotite to pyrite in the crude material. Titania is higher in the magnetic fraction than in the crude material, confirming our belief that most of the TiO_2 is in the magnetite, as exsolved ilmenite. The grade of the Davis tube concentrate is somewhat higher than the expectable grade of a mill concentrate; nevertheless, the general similarities are striking. (Cf. Benson Mines concentrate, table 7.)

TABLE 28.—*Chemical analyses of magnetic fraction of ore and subore, Jarvis Bridge deposit*

[Cf. table 27. Assays marked with an asterisk (*) are from Millar (1947, p. 9; data erroneously given for analysis of crude ore); other assays are from unpublished data of U.S. Bureau of Mines. Composition of magnetic fraction, determined on Davis tube concentrate, would correspond closely to composition of a mill concentrate of the ore.]

Hole	Footage	Apparent thickness (feet)	True thickness (feet)	Weight percent				
				Fe	P	S	TiO_2	SiO_2
1	345–365	20	19.4	68.9	0.014	0.35	0.80	1.5
	365–380	15	14.5	69.2	.013	.15	.69	2.9
2–5	No data.							
6	([1])							
7	500–548.9	48.9	34.7	68.2	.032	.11	[2].19	2.9
8	658.8–669.8	11.0	9.7	68.8	*.02	*.09	*.63	*1.5
	670.6–712	41.4	36.4	68.1	*.02	*.27	*.71	*2.5
9, 10	No data.							
Average composition of magnetic fraction of ore and subore, holes 1, 7, and 8				68.6	.02	.19	[2].60	2.3

[1] Abandoned just beneath overburden.
[2] Value seems low.

EXTENT

The Jarvis Bridge deposit was drilled to test the character and thickness of ore, not to prove up tonnage. Any tentative estimate of the ore reserve, based on limited drilling to an average depth of 350 feet, would be very rough indeed. A crude estimate of the reserve of indicated ore, averaging 23.8 percent magnetic Fe and suggestive of the size and possible significance of the deposit, might be based on the following assumptions:

1. Ore similar in grade and thickness to that cut in hole 9 continues 300 feet northwestward from hole 9. (300 feet is less than half the distance to hole 10 and is well within the +20° isoclinal of the magnetic anomaly.)
2. Ore similar in grade and thickness to that cut in hole 8 continues 500 feet southeastward from hole 8. (Actually, the character of the magnetic profiles suggests that the magnetic "source" is essentially uniform for 1,000 feet southeastward from hole 8.)
3. The ore on all sections continues, with uniform grade and thickness, for 100 feet downdip from each drilled intersection and updip to the inferred bedrock surface.
4. A "mining back" of at least 50 feet downward from the inferred bedrock surface is preserved at each drilled section and excluded from the available reserve.
5. The tonnage factor for ore of this grade is 10 cubic feet per ton.

Using similar assumptions, but a tonnage factor of 11 cubic feet per ton, one may arrive at a crude estimate of the indicated reserve of subore averaging 19.7 percent magnetic Fe (range, 15.6 to 25.5 percent magnetic Fe).

The absence of deep drill holes means that data are lacking on the downdip extension of the ore body. Known deposits of granite gneiss ore are remarkably persistent and continuous; the Lyon Mountain deposit, for example, was years ago being mined 2,000 feet below its outcrop (W. J. Linney, 1943). If it is safe to draw an analogy between these two deposits, the inferred ore reserve at Jarvis Bridge might be sizable.

If the Jarvis Bridge deposit has a prominent shoot structure, the inferred ore reserve might be considerably larger or smaller than that based on the assumption of a nearly tabular ore body.

If the deposit should pass downward (and southwestward?) into a synclinal body of uniform, low-grade ore, the reserve might be greatly increased.

JARVIS BRIDGE EXTENSION

SUMMARY

The Jarvis Bridge Extension anomaly, 5,500 feet long, lies between the Jarvis Bridge and Trembley Mountain deposits, almost touching the latter. The anomaly has not been drilled, and no outcrops have been found in the area. The character of the magnetic profiles suggests that overburden here is generally 50 to 150 feet thick, perhaps as much as 250 feet locally. Though the source of the anomaly is at present unknown, the anomaly may reflect a promising magnetite deposit deeply buried under swamp and valley fill.

LOCATION, ACCESS, DISCOVERY, AND EXPLORATION

The Jarvis Bridge Extension anomaly centers 3.2 miles east of Newton Falls and 1.1 miles northeast of Chaumont Pond, northwest rectangle, Cranberry Lake quadrangle. The anomaly, on land owned by the Newton Falls Paper Mill, is in an indentation of Chaumont Swamp and on low ground east of the swamp.

To reach the anomaly, take any of the tote roads running into the Trembley Mountain area; then continue westward about half a mile in the valley at the south end of the Trembley Mountain deposit. Ross (1952, written communication) suggests this alternative: Find the north-trending western boundary of New York State lands at the marker on N.Y. Highway 3, 0.5 mile east of Heath Pond. Follow the boundary line northward 185 chains (12,210 ft) to the foot of the south slope of Trembley Mountain (local name for the hill 0.75 mile southwest of Buck Mountain). Turn westward, following the valley that drains the swamp south of Trembley Mountain, and continue for half a mile to the east end of the Jarvis Bridge Extension anomaly. The anomaly is close to, and roughly parallel to, the brook and log road running westward through the valley.

The anomaly, discovered in 1942 by the scout work of A. M. Ross, was dip-needled in detail by the U.S. Bureau of Mines in 1944 and 1945. (See pls. 15, 19. Cf. Millar,[38] 1947, figs. 6–8.) In spite of its promise, the anomaly has not been tested by drilling.

MAGNETIC ANOMALY

The Jarvis Bridge Extension anomaly strikes east and has a total length of 5,500 feet within the +10° isoclinal contour (pl. 15). In the aeromagnetic survey of 1945 (Balsley, Buddington, and others, 1954a), the anomaly registered as a prominent peak of 6,370 gammas. The dip-needle anomaly is roughly divisible into 3 parts. The western segment shows 2 areas within the +20° isoclinal; the middle segment shows a peak of +80° on the 1,100-foot-long magnetic "ridge" of +20° readings; the eastern segment shows 3 areas within the +20° isoclinal. The axis of the western segment is horizontally offset 500 to 600 feet from the axis of the rest of the anomaly. Whether this offset is due to faulting, folding, or distribution of magnetic material in two separate zones, we do not know.

The character of the magnetic profiles suggests that overburden is 50 to 150 feet thick (possibly greater) at the western segment of the anomaly. A similar thickness of overburden is inferred to be present along the middle segment of the anomaly, except in the area between coordinates 18800E and 19000E. Here the overburden seems to thin to 20 to 40 feet (possibly less).

Individual magnetic profiles for the eastern segment were not available to the writer. From the breadth of this segment of the anomaly, it seems likely that the depth of overburden may range from 100 to 250 feet or more. Such a depth of overburden is consistent with that found at the south end of the neighboring Trembley Mountain deposits.

No outcrops or drill holes being present, it is hazardous to draw conclusions from the dip-needle data alone. However, rough calculations from these data suggest that, for the western and middle segments, the magnetic material yielding the Jarvis Bridge Extension anomaly may be similar in character to that found between holes 5 and 8, Jarvis Bridge deposit. (Of course, one cannot distinguish grade or thickness of magnetite-bearing material, or type of host rock, merely by inspecting a contoured magnetic anomaly.) The eastern segment of the Jarvis Bridge Extension

[38] Apparently the coordinates were improperly numbered on some of Millar's published illustrations. The location of the Jarvis Bridge Extension anomaly shown on plates 15 and 19 is thought to be essentially correct; it has been checked by A. M. Ross, who had access to the original surveys.

anomaly may be at least as favorable as the other segments; in the absence of the individual magnetic profiles, that is all one can say.

The Jarvis Bridge Extension anomaly might be produced by a thick zone of rocks carrying only a little magnetite. Because of the anomaly's intermediate (but geologically indeterminate) position between the Jarvis Bridge and Trembley Mountain deposits, the likelihood is greater that the Jarvis Bridge Extension may be a significant deposit of magnetite, perhaps an ore body, just as deserving of exploration as its neighbors have proved to be. Ore, if it exists, may be similar to that at Jarvis Bridge (granite gneiss type) or Trembley Mountain (chiefly modified skarn type, but locally baritic quartz-biotite gneiss). The structural relation between the Jarvis Bridge Extension and Trembley Mountain South is enigmatic (see pl. 19) and especially intriguing. If the southern magnetite body at Trembley Mountain is developed, exploratory drilling of the Jarvis Bridge Extension will be almost mandatory.

SKATE CREEK DEPOSIT

SUMMARY

The Skate Creek deposit, indicated by a narrow linear anomaly 9,000 feet long, represents the northwest extension of the Benson Mines ore body. Six diamond-drill holes, spaced 900 to 1,600 feet apart, have tested the Skate Creek deposit to an average distance of 200 feet down dip. Two parallel, closely spaced mineralized zones are present. Exploration has shown that only the southeast half of the main mineralized zone is of economic interest at present. This southeast segment has an indicated strike length of 3,200 feet, an average thickness of 41.7 feet, and a modest reserve of indicated and inferred ore averaging 23.8 percent magnetic Fe.

Geologically, the Skate Creek deposit is of granite gneiss type. The rock sequence, from the top downward, is dominantly sillimanite granite gneiss; pyroxenic zone; heterogeneous unit consisting of microcline granite gneisses and biotitic metasedimentary rocks; and hornblende granite gneiss, the gneissic equivalent of hornblende-microperthite granite. The main (lower) mineralized zone, which is undulatory but pseudoconformable, straddles the contact between sillimanite granite gneiss and rocks of the pyroxenic zone. The upper mineralized zone, present only in the northwest half of the deposit, is about 40 feet "stratigraphically" above the middle of the lower mineralized zone. Parts of both zones show late hydrothermal alteration to pyrite, chlorite, epidote, and other minerals. The intensity of this alteration is in a rough way inversely proportional to the intensity of the magnetite mineralization.

Rocks enclosing the Skate Creek deposit have an average strike of N. 38° W. and an average dip of 50° SW. The deposit may be on the under (upright) limb of a major overturned isoclinal syncline whose axis is nearly horizontal. Two fold systems, each related to a dominant lineation of regional extent, may affect the deposit. Because of the uncertainty of the structural interpretation, we do not know whether the magnetite deposit is essentially sheetlike or shootlike. However, the broader structural features are clear. Additional exploration could be based, for the most part, on the assumption that the gross form of the deposit is sheetlike.

LOCATION AND ACCESS

The Skate Creek deposit centers 3.0 miles north of Oswegatchie (Lower Oswegatchie), 3.7 miles east-southeast of Fine, and 1.7 miles northwest of the Browns Fall powerplant, northeast rectangle, Oswegatchie quadrangle. Skate Creek, a tributary of the Oswegatchie River, crosses the deposit near its southeast end. This intersection is just north of the head of Skate Creek Flow, the flooded part of the creek. According to Reed and Cohen (1947, p. 7), the deposit extends across lots 2, 9, and 12, Township 12, Town of Fine. Land ownership in this area is complicated.

Those unfamiliar with the back roads of the area may reach the deposit most readily by turning off N.Y. Highway 3 at the Scotts Bridge school (north side of the Oswegatchie River, just north of Scotts Bridge) and driving 1.9 miles southeast on the gravel road to the bridge over Skate Creek Flow. Walk north about 3,700 feet to the head of the flow. To reach the north end of the deposit, take the same gravel road but stop your car at the creek 0.25 mile southeast of the Scotts Bridge school. Walk about 1 mile northeast along the old winter road that follows the creek.

DISCOVERY AND EXPLORATION

The writer understands that the deposit was discovered and surveyed by geologists of the Jones & Laughlin Ore Co. about 1943, in following out the Griffin Park magnetic anomaly. Perhaps the deposit was known long before this, for the magnetite-bearing zone crops out in at least three places where hunters or farmers would be likely to see it. The Skate Creek deposit registered as an anomaly of 900 gammas[39] on the aeromagnetic survey of 1945 (Balsley, Hawkes, Buddington, and others, 1946). Fourteen reconnaissance dip-needle traverses were run across the deposit by Buddington and Leonard in 1945, and the geology

[39] 6,100 gammas relative to the datum adopted for aeromagnetic maps of adjoining areas (Balsley, Buddington, and others, 1954a, 1954b).

and ore outcrops were briefly described by them (Hawkes and Balsley, 1946, p. 6–8 and pl. 7). In 1946, the U.S. Bureau of Mines made a detailed dip-needle survey of the anomaly and put down six diamond-drill holes. These holes, spaced 900 to 1,600 feet apart, explored the deposit to an average distance of 200 feet downdip.

MAGNETIC ANOMALY

The Skate Creek anomaly is the northwestern continuation of the Benson Mines anomaly. The Benson Mines anomaly extends north across the Oswegatchie River into the Griffin Park anomaly (Benson Mines Extension). This, in turn, is separated from the Skate Creek anomaly by almost 1 mile of very weak magnetic indications (less than +20°). These relations, clearly shown on the aeromagnetic map (Balsley, Hawkes, Buddington, and others, 1946), have been checked by many reconnaissance dip-needle traverses.

The dip-needle anomaly given by the Skate Creek deposit is a simple linear one extending N. 38° W. for 9,000 feet. (See pl. 16.) The anomaly is 7,750 feet long within the +10° isoclinal, but only 100 to 150 feet wide. Four narrow, elongate strips of +20° readings have a total effective length of 6,200 feet within the closed +10° isoclinal. Reconnaissance dip-needle profiles (Hawkes and Balsley, 1946, pl. 7) show the sharpness of the magnetic ridge—much of it greater than +60°—and the presence of a belt of weak negative readings paralleling the positive anomaly on its northeast side. The narrowness and sharpness of the magnetic profiles reflect the thinness of the magnetite-bearing zone and the shallow depth of cover that prevails throughout most of the length of the anomaly. Overburden at the drill sites was 0 to 10 feet deep northwest of latitude 14000N, and 40 to 50 feet deep southeast of that latitude. (Actually, the overburden appears to be thinner along the center line of the anomaly than at the drill sites, hole 5 excepted, and the magnetite-bearing zone crops out in at least three places.) This southeastward thickening of the overburden apparently accounts for the decreasing intensity and increasing width of the anomaly between holes 3 and 5 (see pl. 16). A very weak positive anomaly continues southeastward across the sand plain toward the Griffin Park anomaly (Hawkes and Balsley, 1946, p. 7). Presumably this part of the anomaly has no commercial interest. It is not shown on plate 16.

Two small areas of low positive readings are found northwest of hole 6, near the northwest end of the main anomaly. (See pl. 16.) Possibly these areas indicate the presence of a little additional disseminated magnetite in hanging-wall rocks well above the main mineralized zone. Alternatively, the areas may represent an intermittently mineralized segment of the main mineralized zone, here perhaps duplicated by isoclinal folding.

The Skate Creek anomaly has been described as a simple linear one. In detail, the anomaly shows a series of small bends or offsets in its center line. All but one of these are left-hand bends. Two well-defined left-hand bends are at (15100N, 5300E) and (14050N, 6200E). Two questionable left-hand bends are at (16250N, 4550E) and (12750N, 7150E).[40] The single right-hand bend is at (11250N, 8250E). The cause of these bends is not definitely known, but the evidence now available suggests that certain left-hand bends reflect isoclinal folds trending about S. 34½° E. and plunging about 7° SE., whereas the single right-hand bend may reflect an isoclinal fold whose axis trends S. 75° W. and plunges about 42° SW.

GEOLOGY

The Skate Creek deposit is in the northwest-trending segment of the Benson Mines belt of metasedimentary rocks and granite gneiss sheets. (See pl. 1.) The deposit roughly follows the contact between a sheet of sillimanite granite gneiss and an underlying thin zone of metasedimentary rocks. The metasedimentary rocks, in turn, are underlain by hornblende granite gneiss, the gneissic equivalent of hornblende-microperthite granite. These rocks dip 35°–60°SW., averaging about 50° SW. Alaskite gneiss appears less than half a mile northwest of the northwest end of the deposit, and the nose of the Stark anticline of phacoidal granite gneiss is about half a mile northeast of the deposit.

More precise definition of the local structural setting is difficult, owing to the lack of outcrops in critical areas. Certainly the Skate Creek deposit lies very close to the axis of an overturned isoclinal syncline (prolongation of the Benson Mines syncline) whose axial plane strikes northwest and dips about 50° SW. Possibly the deposit is on the lower, or upright, limb of this syncline, though that inference is a flimsy one. The synclinal axis seems to trend approximately parallel to the strike of the foliation.

With regard to regional structure, the deposit is in the junction zone where a broad area of overturned isoclinal folds comes against the rigid mass of the Stark anticline. (See Prof. Paper 376, p. 117 and figs.

[40] The bend at (12750N, 7150E) probably results from the effect of the creek valley on the southwest-dipping ore zone. The expected "outcrop pattern"—here reflected by the magnetic anomaly—would be a weak V pointing downstream.

13, 14.) Lineations in the area of overturned isoclinal folds are essentially down the dip of the planar structure; near the Skate Creek deposit, these lineations trend S. 60°–90° W. and plunge southwestward at moderate angles. These southwest-plunging lineations are given mainly by mineral streaking and elongation, especially by bundles of sillimanite fibers. However, some tight, rodlike minor folds also plunge southwestward. The southwest-plunging lineations are presumably in *a* with respect to the axis (*b*) of the major overturned isoclinal fold. In contrast, lineations in the Stark anticline trend south and plunge southward at gentle to moderate angles. (See Prof. Paper 376, fig. 15.) These lineations are in *b* with respect to the axis of the Stark anticline.

The Skate Creek deposit is thus localized at the border between areas of contrasting structural type and "discordant" lineation.

ROCKS

Except for three outcrops of ore, the rocks along the Skate Creek anomaly are poorly exposed. Our knowledge of the sequence comes from drill cores and from regional geologic mapping. Because we have so little precise information on the local structure, it is not possible to make an adequate geologic map (pl. 16); cross sections are schematic and incomplete. Detailed correlation of mineralized zones and adjacent rock units is shown on pl. 17. It is unfortunate that some of the minutiae of correlation are fairly well established whereas certain large features of local structure are poorly known or merely guessed at.

The bedrock sequence, from top to bottom, is: (a) sillimanite-microcline granite gneiss and associated rocks; (b) pyroxenic zone; (c) microcline granite gneiss, sodic syenite gneiss, biotite gneiss, and hornblende-biotite granite gneiss; (d) hornblende granite gneiss—the gneissic equivalent of hornblende-microperthite granite.

Two mineralized zones are present. The lower (main) mineralized zone generally straddles the contact between the pyroxenic zone and the overlying granitic rocks. This lower mineralized zone is an undulating, sheetlike(?) replacement body that transects rock units. The upper mineralized zone, about 40 feet above the main mineralized zone, is entirely in granitic rocks and associated biotite gneiss. (See pl. 17.)

SILLIMANITE GRANITE GNEISS UNIT

Hanging-wall rocks above the pyroxenic zone are for the most part sillimanite-microcline granite gneiss, locally garnetiferous and biotitic. Some biotite-microcline granite gneiss is present, as well as garnet-microcline granite gneiss. These two rock types, as defined, normally lack sillimanite, though that mineral is very sparingly present here and there. Granite pegmatite layers of variable thickness appear sporadically. These pegmatites consist chiefly of quartz and microcline, in places accompanied by mafic minerals derived from the country rock. Rarely, a little schorl is visible. Interlayered with the several facies of microcline granite gneiss is biotite gneiss, much of which contains almanditic garnet, sillimanite, or both these distinctive minerals. Locally, the biotite gneiss contains a little hornblende. Thin amphibolite layers are present in this facies. This heterogeneous assemblage, distinctive enough en masse, is bewildering in detail. Significant variations are shown on the cross sections (pl. 16) and, more fully, on the correlation diagram (pl. 17).

The sillimanite granite gneiss, together with its associated rocks, is 750 to 800 feet thick. Only the lowermost 300 feet of this unit was cut in the Skate Creek drill holes.

PYROXENIC ZONE

The pyroxenic zone ranges in thickness from 5 to 39 feet. It thins northwestward and southeastward from hole 3. The zone was cut in all holes except hole 4, which was stopped 29 feet above the top of the pyroxenic zone.

Rocks of the pyroxenic zone are mostly pyroxene gneiss and pyroxene-microcline granite gneiss. Granite pegmatite and biotite gneiss are rare and sporadic. At hole 3, where the pyroxenic zone is thickest, scattered layers of pyroxene skarn and garnet-pyroxene skarn are present. Late hydrothermal alteration has affected parts of the pyroxenic zone so intensely that the nature of the host rock is difficult to determine.

UNIT BELOW PYROXENIC ZONE

A heterogeneous group of rocks underlies the pyroxenic zone. Chief among these rocks are biotite-microcline granite gneiss and sodic biotite syenite gneiss (a rock derived from the partly granitized biotite-quartz-plagioclase gneiss with which it is associated). In addition, sillimanite granite gneiss, granite pegmatite, hornblende-biotite granite gneiss, and partly granitized biotite-quartz-plagioclase gneiss are present locally. Presumably the unit at one time consisted largely of biotite-quartz-plagioclase gneiss, in part hornblendic. Layers and schlieren of the gneiss have been preserved locally, and the rock serves as a screen between assorted granitic gneisses (already noted) and the underlying hornblende granite gneiss. The unit ranges in thickness from 20 to 45 feet or more.

In hole 2, sodic syenite gneiss was the only rock cored below the pyroxenic zone. Most of the sodic syenite gneiss is chloritized and brecciated. The significance of this breccia is unknown. It might indicate merely local fracturing and mylonitization, or it might point to a small fault, other evidence for which is not at hand.

HORNBLENDE GRANITE GNEISS

Hornblende granite gneiss—the gneissic equivalent of hornblende-microperthite granite—underlies the heterogeneous unit just described. The hornblende granite gneiss is a moderately deformed and recrystallized rock whose uniformity contrasts sharply with the varied character of the overlying rocks. The maximum thickness of hornblende granite gneiss cut in the drill holes is 20 feet (hole 5). Presumably this rock is the uppermost part of the mass of hornblende granite gneiss that borders the west flank of the Stark anticline of older, phacoidal granite gneiss.

STRUCTURE

The rocks enclosing the Skate Creek deposit have an average strike of N. 38° W. and an average dip of 50° SW. The known dip ranges from 30° to 70°. In general the average dip steepens as one proceeds northwestward along the strike of the deposit. Measurements of the dip of foliation in drill cores agree closely with measurements made on outcrops. The local structure, then, appears to be simple, and perhaps it is so. Several different kinds of data consistently suggest that this is not the case—that instead the local strucure is complicated by one or possibly two fold systems whose presence might have a very important bearing on future exploration and development of the magnetite body. Regrettably, the meager data now at hand do not permit us to define the local structure fully and precisely. Nevertheless, it is more desirable to point out possible complications (even at the risk of embellishing nature) than to bury these complications in an oversimplified account. If subsequent exploration shows that the structural complications do not exist, that will be so much the better.

An additional structural feature is presented by the nature of the magnetite deposit itself. Though the deposit is essentially a conformable replacement body, it is not restricted to one rock unit. Rather, it transects well-defined "stratigraphic" units at a very slight angle. This discordance, discussed in a subsequent section, is so slight that it can be disregarded in exploring and developing the deposit. However, the discordance has considerable genetic significance. (See discussion of habit of granite gneiss ore bodies, p. 44.)

FOLDS

Four independent sets of evidence suggest that local folds may be present in the Skate Creek deposit.

1. Bends are present in the anomaly—1 right-hand bend and 3 left-hand bends. Many features other than folding could account for these bends.
2. Regionally, 2 fold systems and 2 patterns of lineation are recognizable. The Skate Creek deposit is at the junction of areas characterized by these separate features (see p. 178–179).
3. Very tight crumpling, resulting in many small V-shaped or nearly isoclinal folds, is shown by cores of sillimanite granite gneiss from hole 2 (90–113 feet) and hole 1 (333–345 feet). In each place, the feature is of restricted range. Much sillimanite granite gneiss is present above and below the crumpled zones, yet none of it shows this crumpling; instead, it is rather regularly foliated. Within the district, similar areas of intense crumpling are commonly restricted to the axial zones of tight folds, usually isoclinal in character. If we assume that the crumpled zones in core from hole 2 and hole 1 are likewise on or near a fold axis, we can calculate the orientation of that axis. It is is S. 34½° E., about 7° SE. (The orientation is, of course, only approximate, for the two points used to determine it may not have exactly equivalent positions on the inferred fold.) This direction, almost parallel to the strike of the local foliation, is also parallel to the axis of the major overturned isoclinal syncline on the under limb of which the Skate Creek deposit presumably lies. (See Prof. Paper 376, p. 116 and fig. 14.)
4. The updip projection of magnetite-bearing zones cut in drill holes 4, 2, 1, and 5 is displaced laterally from the centerline of the magnetic anomaly. Where ore crops out in this area, the centerline of the anomaly is directly above the outcrop. Presumably the centerline of the anomaly would be almost vertically above the suboutcrop of the rest of the deposit where the overburden is shallow. Of course, the lateral displacement of the projected suboutcrop relative to the inferred suboutcrop might be explained by local features of surface or bedrock topography, by errors in surveyed position of drill holes or anomaly, or by other features, including simple rolls in the dip of the foliation. However, these explanations are not all equally plausible. They cannot be considered here in detail.

The map, cross sections, and composite illustration on plate 16 show these inferred folds schematically. Based as it is on very limited data, the reconstruction cannot be accurate in all details. It is suggestive, not definitive. Two points require emphasis. First, in gently plunging folds, a small change in plunge produces a relatively large shift in the position of the suboutcrop of the fold. Though we can only assume a fairly regular plunge, variations no doubt exist. Second, the exact pattern of folds—in plan and in section—cannot be determined from the available data. Where overlapping isoclines are shown on the sections (pl. 16), the thickening or thinning might be very different from what is shown.

Because we do not know the attitude of the axial planes of minor folds relative to the limbs of the major overturned isocline, it is impossible to show the pattern for all major rock units in plan and section.

Paired isoclines A

The magnetic anomaly shows a left-hand bend at (16250N, 4550E). The inferred suboutcrop of the ore cut in hole 6 agrees with the centerline of the magnetic anomaly if the measured core-foliation angles are used to project the ore up the dip from the drill hole. This is not true at hole 4, where the suboutcrop of projected "ore" would lie about 60 feet northeast of the centerline of the magnetic anomaly. Therefore, it seems reasonable to infer a pair of left-hand isoclinal folds (right-hand, in plan) whose suboutcrop would correspond to the left-hand bend in the magnetic anomaly. If the axes of these paired isoclines plunge 7° SE., their trend must be about S. 31½° E.—essentially parallel to the S. 34½° E. direction but shifted very slightly to match the steeper average dip of foliation along this segment of the deposit. (Keeping the trend constant at S. 34½° E. and shifting the plunge slightly gives a poorer fit among the suboutcrop of the isoclines, the position of the bend in the anomaly, and the inferred suboutcrop of ore cut in hole 6.)

Paired isoclines B

Three converging lines of evidence suggest the existence and position of paired isoclines B.

1. Local crumpling of sillimanite granite gneiss in cores from holes 2 and 1 suggests the presence of one or more tight isoclinal folds whose axes have the approximate orientation S. 34½° E., 7° SE. (See point 3 of section on folds.)
2. The suboutcrop of the mineralized zone cut by holes 2 and 1 is offset from the centerline of the magnetic anomaly if the mineralized zone is projected up dip along the measured core-foliation angles. Yet the dip measured in the ore outcrop between holes 2 and 1 agrees closely with the dip of foliation measured on the drill cores. Folds such as paired isoclines B would eliminate the discrepancy between the position of the suboutcrop projected from drilling data and the position inferred from the magnetic anomaly.
3. The magnetic anomaly shows a left-hand bend at (15100N, 5300E). Paired isoclines B—inferred on the cross sections, plate 16—would project up the plunge at N. 34½° W., 7° NW., to have a suboutcrop near this left-hand bend in the anomaly.

The evidence for the existence and position of paired isoclines B is rather good. Obviously it is the keystone in the argument for the other sets of paired isoclines, A, C, and D.

Paired isoclines C

A prominent left-hand bend in the centerline of the magnetic anomaly is at (14050N, 6200E). There is no independent evidence in this immediate neighborhood to suggest the presence of tight isoclinal folds. However, if the left-hand bend in the magnetic anomaly is to be interpreted in the same manner as the left-hand bends at (16250N, 4550E) and (15100 N, 5300E), what are the consequences?

First, a left-hand pair of isoclines seems capable of accounting for the bend; a right-hand pair of isoclines will not. [Isoclines that are right-hand in plan are required to explain the combination of features involved in the relations at holes 4, 2, and 1, and in the left-hand bends in the anomaly at (16250N, 4550E) and (15100N, 5300E).] As a consequence, paired isoclines C would show, in cross section, an isoclinal anticline overlying an isoclinal syncline. This is a complete reversal of the pattern exhibited by paired isoclines A and B. It is inconsistent with the pattern of true drag folds. However, in an occasional outcrop in the district one sees such reversals of pattern in the minor folds whose strike length is 1 or 2 feet. Here the folds are not true drag folds, but folds resulting from a complex interaction of flowage and shearing. Consequently, the reversed pattern of paired isoclines C may seem to spoil the symmetry for southeast-plunging folds at Skate Creek, yet the reversal is consistent with observable details of some much smaller folds in the district.

The second consequence of postulating paired isoclines C, having a suboutcrop at (14050N, 6200E) and an approximate orientation S. 34½° E., 7° SE., is that these isoclines would not necessarily be cut by any of the drill holes southeast of (14050N, 6200E).

The downplunge extension of paired isoclines C would fall very close to the inferred suboutcrop of the ore zone cut by hole 1. This pair of isoclines would lie about 200 feet updip from paired isoclines B. There is no independent evidence, especially magnetic evidence, that paired isoclines C should appear on the cross section through hole 1 (pl. 16); the logical and convenient assumption is that the isoclines may have been eroded.

Hole 3, a short hole, would also fail to cut paired isoclines C, which at the latitude of hole 3 should lie almost vertically beneath the collar of the hole but at an elevation roughly equal to the elevation of the bottom of the hole. Because all the evidence for paired isoclines C is extremely tenuous, these isoclines are not sketched in on the cross section through hole 3 (pl. 16).

Hole 5 might or might not cut paired isoclines C.

Paired isoclines D

The interpretation of the structure at hole 5 is uncertain, for the data are conflicting. This is the problem: ore and subore were cut throughout most of the interval from 110.1 to 151.5 feet in hole 5. Core-foliation angles consistently indicate a dip of 43°±5° SW. in rocks above, within, and just below this zone. An accurate topographic profile along the projection of hole 5 is not available, but the relief is relatively slight and the inferred profile shown on plate 16 is thought to be substantially correct. The bearing of the drill hole makes an angle of about 70° with the center line of the magnetic anomaly—not perpendicular, but nearly so. If the magnetite-bearing zone cut from 110.1 to 151.5 feet in hole 5 is projected up the dip at 43°±5°, the "outcrop" of this zone coincides with the centerline of the magnetic anomaly. However, the ore is not known to crop out; in fact, it is probably rather deeply buried by glacial debris. Hole 5 cut 50 feet of overburden, and the character of the magnetic profile here suggests that, if anything, the overburden is even deeper beneath the centerline of the anomaly. This means that the suboutcrop of the ore should lie somewhere between the centerline of the anomaly and the collar of hole 5, assuming a dip of 43°±5°. The thicker the overburden, the greater the apparent horizontal displacement of the projected trace of the ore relative to the centerline of the anomaly. This being so, we find a considerable discrepancy in the position of the suboutcrop inferred from magnetic data and the position inferred from drilling data. How can we explain this discrepancy?

The data presented above seem to be reliable. Thus we cannot beg the question by assuming serious errors in inclination of drill hole, observed dip of foliation, topographic profile, position of anomaly, or (we hope) inferred depth of overburden. Some structural reason may account for the discrepancy. Four possibilities are considered.

1. The horizontal projection of hole 5 makes an angle of 70° with the center line of the anomaly. Therefore, the measured core-foliation angles indicate a dip slightly greater than the dip existing in a vertical plane passing through the drill hole. However, a true dip of 43°, when projected to a plane oriented at 70° to the strike of the foliation, is reduced only to 40°. This alone will not explain the discrepancy.
2. There is no evidence here for faulting; so we can hardly resort to a normal fault to explain the discrepancy.
3. A simple bend fold, or roll, might be postulated. However, deformation resulting in that structure is not known to be characteristic of this particular area.
4. A tight fold or pair of folds, presumably isoclinal, seems to be the last resort. Isoclinal folds with axes trending S. 31½°–34½° E. have been inferred in the area between holes 1 and 6. However, it is not possible to account for the right-hand bend in the magnetic anomaly, and the apparent horizontal offset of the projected ore, by any kind of tight isoclinal fold plunging gently or moderately southward or southeastward—that is, by a fold whose axis is roughly parallel to the strike of the rock units and parallel to one of the dominant lineations in the area.

One can account for the observed relations by postulating a pair of isoclines (D) whose axis trends S. 75° W. and plunges about 42° SW. The axis of the postulated paired folds would parallel the dominant downdip lineation that prevails over large areas west and southwest of the Skate Creek deposit. Such paired isoclines, plunging almost at right angles to folds postulated in the area between holes 6 and 1, would explain the shift of the magnetic axis from a pattern of left-hand bends (prevailing northwest of hole 1) to a right-hand bend north of hole 5. It would also explain the troublesome horizontal offset of the ore zone.

The evidence from the core of hole 5 neither supports nor denies the existence of southwest-plunging paired isoclines near hole 5. Therefore, the hypothesis presented above is highly tentative. So are the cross section for hole 5, and the geologic map for that area (pl. 16). The hypothesis should be kept in mind if the deposit is further explored, but it should be abandoned as promptly as a new and consistent body of contradictory facts is developed.

What happens where paired isoclines A, B, and C meet paired isoclines D? This is anyone's guess. On plate 16 paired isoclines A, B, and C have been extended many thousand feet down the plunge, almost to the intersection with paired isoclines D. That is simply for purposes of illustrating the relations that might obtain. Actually, one or all the isoclines may die out downplunge. Alternatively, the two main sets of folds might still be present near hole 5, interacting in complex fashion as the fold systems do at the Clifton mine.

ORE

OCCURRENCE

Outcrops of typical low-grade ore were seen on three separate traverses across the Skate Creek anomaly. The first outcrop, 850 feet north-northwest of the head of Skate Creek Flow, is an exposure of 600 square feet in the dry floor of a small valley descending to the southeast. A Rosiwal analysis of part of the outcrop surface shows 27 percent magnetite by volume (30 percent Fe in the form of iron oxides). The true thickness of exposed ore, which dips 45° SW., is about 25 feet. The contacts between ore and walls are not shown but are probably very close to the limits of the outcrop.

The second outcrop, 250 feet north-northwest of the head of the flow, is a small weathered exposure beneath an overturned tree at the east edge of a low, sloping bench. Magnetite, broken by closely spaced rectangular joints, forms 20 to 25 percent of the exposure. A 1½ foot layer of rusty pyritic material is also present. The exposed and inferred thickness of ore is not less than 12 feet. The dip seems to be about 30° to 35° SW.

The magnetite in these two outcrops occurs as grains, blotches, and coalescing layers one-eighth inch to several inches thick. Magnetite-rich layers of varying purity alternate with layers of quartz-feldspar aggregate, or layers of rather coarse pink feldspar. These granitic layers have little or no magnetite but contain a little sillimanite, red garnet, biotite, and apatite. In places, magnetite occurs in thin zones rich in mica; a little pyrite is present with the magnetite. One- to two-inch pegmatite seams with very little magnetite commonly parallel the foliation, and thin veins of barren quartz also occur. Magnetite in the wall rocks ranges from a trace to about 15 percent.

The third outcrop, 350 feet south-southeast of the head of the flow, is similar in grade to the other outcrops, but the ore is in pyroxene gneiss. The true thickness of ore, exposed and conservatively inferred, is about 25 feet.[41]

The Skate Creek deposit is of granite gneiss type. Significant magnetite mineralization is confined to two zones (see pl. 17). The lower, or main, zone straddles the contact between the pyroxenic zone and the overlying granitic rocks. The zone is distinctly undulatory. It is not confined to a single rock unit; rather, it weaves back and forth among three major units. At hole 5, the zone is partly in granitic rocks above the pyroxenic zone, partly in the pyroxenic zone, and partly below the pyroxenic zone. The relations are similar at holes 3 and 1, except that the mineralized zone does not pass below the base of the pyroxenic zone. At hole 2, where the lower mineralized zone is thinnest, the mineralized zone is just above the pyroxenic zone. At hole 6, relations are similar to those at hole 5, though much more of the granitic rocks below the pyroxenic zone have been mineralized. According to the best evidence now available, hole 4 was stopped short at a true distance of 5 to 15 feet above top of the main mineralized zone.

The manner in which the lower (main) mineralized zone transects the host rock, including the well-defined marker or pyroxenic zone, indicates very clearly that the mineralized zone is not, and never has been, a metamorphosed bedded iron deposit. Rather, it is a replacement deposit presumably occupying a zone of pervasive shearing that was controlled by contacts between different rock units essentially stratigraphic in character. However, the angle at which the main mineralized zone transects "stratigraphic" units is very small, averaging about ½° to 1° over strike distances of 1,300 to 2,500 feet. Consequently, the dominant feature recognizable at any one place along the zone is the apparent concordance of the mineralized zone and its host rocks.

Two other significant features are coupled with this gently undulating, discordant character. The lower (main) mineralized zone has a fairly uniform thickness of 40 to 50 feet in holes 6, 1, 3, and 5, but it thins to 4 feet in hole 2. Moreover, within the broad limits of the lower mineralized zone the grade, thickness, and relative position of individual mineralized layers varies along the strike. Plate 17 shows that, in general, material exclusively of ore grade describes a great asymmetric arc within the mineralized zone. Going northwestward from hole 5, the material of ore grade passes from the pyroxenic zone into granitic rocks above the pyroxenic zone, continues here for much of the distance, and then (at hole 6) descends again into the pyroxenic zone.

Details of grade and extent of ore within the lower (main) mineralized zone are given in a subsequent section.

The upper mineralized zone (see pl. 17) is 6 to 19 feet thick in holes 6, 4, and 2. The zone pinches out somewhere between holes 2 and 1. The middle of the upper mineralized zone is "stratigraphically"

[41] The foregoing description of outcrops is from the report of Hawkes and Balsley (1946, p. 7–8).

about 40 feet above the middle of the lower (main) mineralized zone. The succession or rock units lends no support to the hypothesis that the two mineralized zones are a single zone duplicated by folding.

A sort of en echelon distribution of magnetite is suggested by relations of the two zones in holes 2 and 6. Hole 2 cut only 4 feet of ore in the lower zone, but at this hole the upper zone is thicker and higher in grade than it is to the northwest. At hole 6, where the upper zone is thin and very weakly mineralized, the lower zone carries a substantial thickness of subore and marginal ore.

Economically, the upper mineralized zone is of little interest, owing to the prevailing combination of poor grade and slight thickness of magnetite-bearing material.

Is the ore-bearing part of the lower mineralized zone essentially sheetlike, though showing inherited isoclinal folds of moderate dimensions, or is it in the form of one or more plunging shoots controlled by these folds? These questions cannot be answered at present, but they require discussion. The longitudinal vertical projection (pl. 16) shows 4 inferred sets of overlapping isoclines. With regard to the vertical position of these isoclines, the drill holes show a random distribution; that is, the holes did not all cut the ore zone at structurally equivalent positions. This means that, if the structural interpretation is essentially correct, the holes did not cut a single, shootlike magnetite body, severely limited in height, whose grade and thickness might be significantly better or poorer than the average for the entire ore-bearing part of the deposit. It does not mean that shoot structure is definitely lacking. On this point the data are inconclusive. Conceivably, the ore cut by holes 1, 3, and 5 might be part of one vast shoot, underlain (in the plane of the dip) by another, leaner shoot tested by holes 6, 4, and 2.

What about local thickening of ore along the axes of the inferred overlapping isoclines? This feature is shown on the cross sections (pl. 16), but—as emphasized before—these sections are schematic. There is no evidence at the Skate Creek deposit that ore in the schematic overlapping isoclines should be thicker or thinner, richer or leaner, than the average. Naturally, one would want to rule out one possibility or the other, but this cannot be done without additional exploration.

MINERALOGY

The ore is a medium- to fine-grained, crudely foliated mixture of magnetite, sporadic sulfides (both pyrite and chalcopyrite), unreplaced remnants of the host rock, and minerals developed by late hydrothermal alteration. Because of the great variation of the host rock (see pl. 17), the gangue minerals constitute a highly varied suite. Ore from the pyroxenic zone commonly contains quartz; potassic feldspar; sporadic myrmekite; sparse, altered plagioclase; ferrosalitic pyroxene, blue-green amphibole, or both; little or no biotite; magnetite; sulfides; accessory apatite, allanite, zircon, and unknowns; and late hydrothermal minerals. Ore from granitic rocks above or below the pyroxenic zone commonly contains quartz; potassic feldspar; little or no plagioclase; biotite and sporadic muscovite; garnet; sporadic sillimanite; magnetite; sulfides; accessory apatite, allanite, and zircon; and late hydrothermal minerals. The quartz-feldspar ratio is variable. Potassic feldspar may or may not show good microcline twinning. Apatite is abundant locally. Magnetite in some ore contains exsolved granules of bright green spinel.

Pegmatite, either granitic or syenitic, commonly carries a little magnetite, but it seldom has enough magnetite to constitute ore.

The late hydrothermal minerals are discussed elsewhere.

It has not been possible to make an extensive microscopic study of the ore minerals. Ore in pyroxene gneiss from the outcrop 350 feet south-southeast of the head of Skate Creek Flow consists of a network of magnetite grains replacing silicates. The magnetite, which forms 25 to 30 percent of the specimen examined, contains abundant spinel "gashes," rather sparse ilmenite hairs and thin blades, and a little hematite. This hematite is of two types: primary crystalline hematite as straight-sided grains in magnetite; and martite as ragged replacement veinlets, patchy replacements at grain boundaries, and rare well-defined blades—all in magnetite. The total quantity of hematite is less than 1 percent of the magnetite, or a few tenths percent of the entire rock. A few granules of pyrite are also present, and the silicates and magnetite contain local networks of supergene limonite.

Presumably the rest of the Skate Creek ore will be similar in the character of its metallic minerals, except for the presence of local chalcopyrite, very abundant local pyrite, and (in the sillimanitic facies of the granite gneiss) perhaps more hematite.

LATE HYDROTHERMAL ALTERATION

The products of late hydrothermal alteration are a prominent feature of parts of the Skate Creek deposit. Chief among these products are pyrite of stage II, earthy chlorite, and epidote. (See also p. 68–69.) In or within a few feet of zones of heavy pyrite-chlorite-epidote alteration, little vugs—1 or 2 mm in diameter—stud the rock. Few of the minerals lining the vugs can be identified megascopically, but here and there one can observe fibrous crusts of pumpellyite, in places accompanied by minute crystals of calcite and unidentified zeolites. Rocks sheathing these zones of visible alteration are seemingly unchanged, yet under the microscope they show very slight, sporadic development of quartz, chlorite, pumpellyite, fringelike pyrite, epidote, zoisite(?), pinkish zeolite(?), carbonate, and kaolin(?). The suite varies somewhat from place to place, and its total volume is commonly less than 1or 2 percent of the rock.

The distribution of the altered zones bears a remarkable relation to the two magnetite-bearing zones (upper and lower mineralized zones). First, visible alteration of the type described above is effectively confined to the mineralized zones and to a sheath of rock, generally 10 to 15 feet thick, above (but not below?) these zones. Second, the intensity of alteration increases northwestward along the strike of the deposit. In a roughly quantitative way, the intensity of alteration is inversely proportional to the intensity of the magnetite mineralization. This relation should be clear from the following description, in which the evidence is presented, hole by hole, from southeast to northwest along the strike of the deposit. (Cf. data on magnetite layers, pl. 17.)

Alteration in the lower (main) mineralized zone

At hole 5, chloritization is intense, but extremely limited. At hole 3, the zone shows local leaching, accompanied by mild development of chlorite and epidote. At hole 1, the zone shows mild development of pyrite, epidote, and subordinate chlorite. At hole 2 there is intense local pyritization and chloritization, as well as some weak, pervasive chloritization. Hole 4 failed to cut the lower zone. At hole 6, the entire zone shows considerable pervasive alteration to pyrite and chlorite, as well as local leaching.

Alteration in the upper mineralized zone

An upper magnetite-bearing zone was not cut in holes 5, 3, and 1; however, at the appropriate "horizon," hole 5 shows a 2-foot zone carrying a little chlorite and pyrite. The zone is absent from hole 3, but in hole 1 a little chlorite is present at the appropriate "horizon." Hole 2, the first to cut the upper mineralized zone, shows weak pyritic alteration of that zone. Hole 4 shows moderate pyritic alteration. Hole 6 shows moderate pyritic alteration of the zone itself, development of vugs just above the zone, and considerable chlorite-pyrite alteration between the upper zone and the altered lower zone.

The relations described above are consistent with the regional distribution of magnetite deposits and chlorite-pyrite (or pyrrhotite) deposits in the northwest Adirondacks. (See p. 18–20.)

Thin sections of ore from the outcrop 350 feet south-southeast of the head of Skate Creek Flow show limonite stains but no chlorite, pumpellyite, pyrite of stage II, or related secondary minerals. This fact alone suggests that the type of alteration described above as late hydrothermal cannot possibly be supergene. The nature of the chlorite in the fault breccia at the base of hole 2 is unknown. The age of the fault is also unknown; it is almost certainly post mineralization (by analogy with known faults in the district), and it may well be post-Precambrian.

GRADE AND EXTENT

Table 29 gives the grade and thickness of the principal magnetite-bearing layers. Reed and Cohen (1947, p. 26–30) give the total Fe content of individual 5-foot samples. The reader is referred to Reed and Cohen's report for this supplementary information. A "complete" analysis of the composite heads is not available, and we do not know the ratio of soluble Fe to magnetic Fe. From the known presence of hematite in the Skate Creek ore, and by analogy with other deposits in sillimanite granite gneiss, we may infer the presence of 1 to 2 percent soluble nonmagnetic Fe in suitable parts of the Skate Creek deposit. The sulfur content of a magnetic concentrate of Skate Creek ore would probably be considerably less than the sulfur content of the Davis tube concentrate for Reed and Cohen's "composite of composites." (See table 29, this report, especially footnote 8. Cf. discussion of distribution of pyrite-chlorite-epidote alteration.)

Only holes 1, 3, and 5 cut material of suitable grade and thickness to constitute ore by present standards. The strike length tested by these widely spaced drill holes is 2,700 feet. An additional 500 feet is effectively indicated by the character of the magnetic anomaly and the presence of outcropping ore northwest of hole 1. The ore has an average thickness of 41.7 feet running 23.8 percent magnetic Fe. The holes cut ore at an average distance of 210 feet down the dip. An additional downdip extension of 100 feet seems probable. Assuming a rudely tabular body with dimensions of 3,200×41.7×310 feet, and further assuming a tonnage factor of 10 cubic feet per long ton, we find that the indicated reserve of ore averaging 23.8 percent magnetic Fe is of modest size.

If the ore body cut by holes 1, 3, and 5 is shootlike, plunging 7° SE., the indicated reserve might be about the same, or it might be doubled. The random

TABLE 29.—*Grade and thickness of principal magnetite-bearing layers, Skate Creek deposit*

[Assays by U.S. Bureau of Mines. Data condensed from Reed and Cohen (1947, p. 26–30)]

Hole	Footage	Apparent thickness (feet)	True thickness (feet)	Grade (weight percent)							
				Crude ore		Magnetic fraction only					
				Total Fe	Magnetic Fe	P	S	TiO_2	SiO_2	Mn	Fe
6	150.8–159.0	8.2	6.4	17.0							
	159.0–180.0	21.0	16.4								
	180.0–222.0	40.0	31.2	18.3	[1] ~13.7						
	220.0–231.7	11.7	9.1	[2] 23.1	19.8						
4	185.0–206.0	21.0	15.4	8.4							
	206.0–221.0	15.0	11.0	[3] 17.2	14.7						
2	185.0–200.0	15.0	12.2	16.4	[4] ~13.5						
	200.0–240.0	40.0	32.4	10.3							
	240.0–245.0	5.0	4.1	23.1	[4] ~19.0						
1	353.0–357.8	4.8	3.5	11.3							
	357.8–384.0	26.2	19.4	32.8	[5] 30.9						
	384.0–408.0	24.0	17.8	23.7	[5] 22.1						
	408.0–411.0	3.0	2.2	10.5							
3	84.5–142.8 [6]	58.3	50.7	26.4	24.3						
5	88.0–111.0	23.0	21.4	9.1							
	111.0–131.0	20.0	18.6	16.0	[7] 12.8						
	131.0–151.0	20.0	18.6	29.7	[7] 27.8						
	151.0–152.0	1.0	0.9	15.6							
"Composite of composites" [8]				23.5	20.6	0.01	1.26	0.50	2.40	0.14	68.7

[1] Computed by Leonard, using magnetic Fe: soluble Fe ratio given for combined sample, 150.8–159.0 and 180.0–220.0 feet.

[2] Weighted average of individual samples is 22.1.

[3] Average of 3 individual 5-foot samples is 16.3.

[4] Computed by Leonard, using magnetic Fe: soluble Fe ratio given for composite sample, 190.0–200.0 feet.

[5] Weighted average, 357.8–408.0 feet, is 26.7 percent magnetic Fe.

[6] Actually, the material of ore grade is confined to the interval 84.5–118.0 feet. The interval 118.0–142.8 feet is rather weakly mineralized skarn and pyroxene gneiss. Thus the ore-bearing interval is thinner but higher in grade than this sample suggests.

[7] Average, 111.0–151.0 feet, is 20.3 percent magnetic Fe.

[8] Represents, in a crude way, the character of ore from this deposit. Reed and Cohen (1947, p. 26) state that this is a composite of the composite samples for holes 1–6, but actually it seems to exclude hole 2 and include the following intervals:

Hole 6—150.8–159.0, 180.0–231.7
4—206.0–221.0
1—357.8–408.0
3— 84.5–142.8
5—111.0–151.0

Thus it indicates the character of the ore only in a highly generalized way. Reed and Cohen (1947, p. 26) give 0.10 percent sulfur for the magnetic fraction of their composite samples for holes 1, 3, and 5. This is a much more representative value for the part of the Skate Creek deposit that constitutes ore. Other constituents (P, TiO_2, etc.) for this analyzed magnetic fraction were not reported.

"structural" position of the drill holes does not lead one to suspect large barren areas within the hypothetical shoot; hence the indicated reserve in a shootlike body would not be significantly less than in a crudely tabular body.

An estimate of the reserve of inferred ore, assuming a crudely tabular body, may be approached as follows.

Ore crops out near baseline station 50+00 (see pl. 16), 400 feet northwest of the section drilled by hole 1. This ore, visually estimated as carrying about 30 percent Fe, has an inferred true thickness of 25 feet. The distance from this outcrop northwest to the section drilled by hole 2 is 1,000 feet. Inasmuch as hole 2 cut some ore at the base of a thick zone of mineralized rock, it seems fair to infer that the ore cropping out near baseline station 50+00 continues northwestward for 500 feet, or half the distance to hole 2. The total strike length of ore inferred northwest of hole 1 is therefore 900 feet.

The character of the southeast end of the magnetic anomaly suggests that the zone of ore and subore cut by hole 5 continues southeastward at least 400 feet to the closure of the +10° isoclinal.[42] The total straight-line distance between this point and the point previously adopted for the northwest limit of ore is 3,950 feet. The distance measured along the centerline of the anomaly is greater, but negligibly so. Let us call it 4,000 feet, for convenience.

The average thickness of ore cut by drill holes and estimated in 3 outcrops is 37.5 feet. Let us assume that the ore extends 500 feet downdip. Then the dimensions of the inferred ore body are 4,000×37.5×500 feet. This body would include both indicated and conservatively inferred ore, nearly equal in quantity, averaging 23.8 percent magnetic Fe. Roughly the same quantity of inferred ore is calculable if we assume the ore body to be a continuous shoot plunging 7° SE.

At present, the northwest half of the anomaly (northwest of latitude 14100N) seems to hold no economic

[42] How inferred isocline D might affect the continuation of ore, we do not know. Judging from the anomaly, we should say "not much." Judging from the schematic representation in pl. 16, we should say "indeterminate."

interest. However, it cannot be fairly evaluated from the data at hand. Hole 6 cut 9 feet of marginal ore. Hole 2 cut 4 feet of marginal ore. But hole 4, between holes 2 and 6, was stopped short 5 to 15 feet above the main mineralized zone, according to the best correlation of available evidence. Consequently, we do not know the average grade and thickness of the zone as a whole. Moreover, if magnetite concentrations in the main mineralized zone have an en echelon shoot structure, we cannot rule out the possibility of finding ore of suitable grade and thickness at vertical depths greater than about 300 feet along this northwest half of the mineralized zone.

In concluding, it is prudent to emphasize again that the broader features of structure of the Skate Creek deposit are well known, but that the details are inferential. We do not know whether two fold systems definitely affect the deposit; we do not know the precise orientation of these folds, if they exist; and we cannot possibly know how the interaction of the fold systems localized the concentrations of magnetite. We do know that the inferences regarding local structure are consistent with the regional pattern. Since structure is the clue to the size and shape—and perhaps the grade—of ore deposits in the district, those who may further explore the Skate Creek deposit ought to pay close attention to every structural detail available.

SPRUCE MOUNTAIN DEPOSITS

The Spruce Mountain tract, north and northeast of Newton Falls, consists of two magnetically anomalous areas separated by 1 mile of weak magnetic indications. The tract is interpreted as a syncline of metasedimentary rocks and granite gneiss sheets plunging gently southeast. Much of the bedrock is obscured by a thin veneer of glacial debris. The exact structural relations of the ends of the tract are therefore uncertain. The known ore body is confined to the northwest part of the tract, where the magnetic anomalies have a fishhook pattern. Very limited drilling suggests that this ore body contains a sizable quantity of ore averaging 24 to 25 percent magnetic Fe. In addition, a considerable tonnage of subore and mineralized rock is present. Magnetite has partly replaced two zones, stratigraphically about 30 feet apart, within a continuous sheet of microcline granite gneiss. The upper, pyritic zone is not included in the tonnage estimate. The Northwest deposit might support a modest open-pit operation. The Southeast area is of no direct economic interest. The following account is a slight modification of a previously published report (Leonard, 1953).

LOCATION, OWNERSHIP, AND ACCESS

The Spruce Mountain group of anomalies can be divided for convenience into two principal areas: Spruce Mountain Northwest and Spruce Mountain Southeast. Both are on land owned by the Newton Falls Paper Mill and take their name from Spruce Mountain,[43] an elongate ridge 1.8 miles north-northeast of Newton Falls village, northwest rectangle, Cranberry Lake quadrangle.

Spruce Mountain Northwest is the area of commercial interest. It lies on the intersection of the Russell, Stark, Cranberry Lake, and Oswegatchie quadrangles, principally on the first two, with its center about 3 miles north-northwest of Newton Falls village. A bulldozed access road extends northwest to it from Jarvis Bridge, northwest rectangle, Cranberry Lake quadrangle, crossing the Clifton extension of the New York Central Railroad at a point 1.3 miles northeast of Newton Falls village. An old hunters' trail from the neighborhood of Browns Bridge, northeast rectangle, Oswegatchie quadrangle, runs northeast to the area.

Spruce Mountain Southeast anomaly, northwest rectangle, Cranberry Lake quadrangle, centers 1.5 miles north-northeast of Newton Falls village. It lies southwest of Spruce Mountain, between the ridge and Big Creek. The bulldozed access road from Jarvis Bridge runs past the anomaly.

DISCOVERY AND EXPLORATION

The anomalies were discovered by A. M. Ross, Woods Superintendent of the Newton Falls Paper Mill. Ross made reconnaissance dip-needle surveys of both areas in 1943, 1944, and 1945 (Ross, personal communication, 1949). The U.S. Bureau of Mines made detailed dip-needle surveys of the anomalies during 1944, 1945, and 1946, and drilled 10 holes on the property in 1946. Both groups of anomalies were partly explored by this diamond drilling, and a sizable deposit of low-grade ore was found at Spruce Mountain Northwest. The results of work by the Bureau of Mines have been summarized by Reed and Cohen (1947, p. 6–7, 17–25).

Members of the U.S. Geological Survey logged drill core and mapped the area. Most of the mapping was done by B. F. Leonard and A. E. J. Engel in the spring of 1946. Some structural observations made by C. J. Cohen, formerly an engineer of the Bureau of Mines, have been incorporated in the geologic map (pl. 18).

[43] This Spruce Mountain, locally known as Big Creek Mountain (Ross, personal communication), is not to be confused with Spruce Mountain, south-central rectangle, Stark quadrangle, near the Deerlick Rapids and Wolf Hole anomalies.

Results of private exploration of this deposit were made available to the Geological Survey in 1958. It has not been possible to incorporate the new data or their interpretation in this report.

MAGNETIC ANOMALIES AND FLOAT ORE

The Spruce Mountain Southeast anomaly begins 500 feet north of the railroad and runs northwest for 4,800 feet within the +8° isoclinal (see pl. 18). It is a simple strip or belt averaging 100 to 150 feet in width and enclosing several peaks of moderate to high intensity. Float of low-grade ore was found about 1,200 feet northwest of hole 1. Presumably this float was derived from the Spruce Mountain Northwest area, as the four outcrops along the Southeast anomaly show only a little disseminated magnetite in granitic garnet-biotite gneiss. Elsewhere along the ridge on which the anomaly lies, the overburden is probably very thin. The fact that the magnetite-bearing gneiss is at or very close to the surface presumably accounts for the relative strength of the anomaly.

The Spruce Mountain Northwest anomaly is a U-shaped group of anomalies with a total length (following around the U) of 7,000 feet. The individual members of the group are disconnected strips of low positive anomalies with two prominent negatives along the northeast half. The U, which opens toward the southeast, is divided almost symmetrically by a 1,500-foot strip or lobe of readings within the +10° isoclinal. A shorter lobe, 1,000 feet east of the first, is 700 feet long within the +10° isoclinal. The ore that causes the Spruce Mountain Northwest anomaly does not crop out.

If the Spruce Mountain Southeast anomaly were projected northwest, it would join the southwest limb of the U-shaped Spruce Mountain Northwest anomaly to form a fishhook-shaped structure. However, the two anomalous areas are now separated by 5,000 feet of neutral or weak magnetic readings. Toward the southeast, the Spruce Mountain Southeast anomaly is almost continuous with the northwest end of the Jarvis Bridge anomaly. (See pl. 1.)

The relation of the Spruce Mountain area to the regional magnetic picture may be seen on the aeromagnetic maps of Balsley, Buddington, and others (1954a, 1954b, 1959a). The Spruce Mountain Northwest anomaly was registered as a peak of about 6,400 gammas. The Spruce Mountain Southeast anomaly appears as a magnetic saddle between the magnetic peaks at Jarvis Bridge and Spruce Mountain Northwest.

GEOLOGY

ROCKS

Three main rock units can be distinguished in the Spruce Mountain area and used for correlation. These units are composite, comprising a number of rock types. Attempts to correlate individual members of the units, except ore zones, succeed only in obscuring a relation that is otherwise reasonably clear. For that reason, an admittedly subjective presentation is adopted here. Exceptions and other interpretations will be noted below.

All the rocks are of early Precambrian age. The metasedimentary rocks, including such general types as pyroxene gneiss, skarn, biotite gneiss, and amphibolite, belong to the Grenville series. The granitic rocks, chiefly microcline granite gneiss, are of complex origin.

UPPER (PYROXENIC) UNIT

The uppermost unit exposed at the surface or cut in drill holes may be termed "pyroxene gneiss," a catchall designation for interlayered pyroxene-quartz-feldspar gneiss, pyroxenic migmatites, and pyroxene skarn. The proportion of these types varies widely. In holes 1 and 2, considerable brown mica is present in pyroxene-mica gneiss or in biotite gneiss. Dark-brown garnet is present sporadically in some cores from the Northwest area. Scapolite occurs locally. A sheet of slightly pyroxenic or biotitic microcline granite gneiss, 45 feet thick, is present in the unit at hole 8. Some migmatitic amphibolite is present in holes from the Southeast area. Holes 1, 2, 3, 5, and 8 cut part of the upper (pyroxenic) unit, whose thickness is unknown but may be 800 feet or more.

MIDDLE UNIT

The middle unit, 150 to 255 feet thick, comprises microcline granite gneiss and biotite gneiss, with their variants. Pyroxene gneiss is present locally. In detail, the unit is a most heterogeneous one, owing to differences in quantity of injected granite magma, degree of granitization, and effect of mineralization. This unit contains the ore, a description of which is given in a separate section.

The rock type broadly termed "microcline granite gneiss" is pink, white, gray, or greenish, generally fine grained but commonly medium and locally coarse. Its foliation may be well or poorly marked. In some places, a foliation is given by thin seams of granite pegmatite. The rock ranges in composition from syenite to quartz-rich granite, though "granite" is appropriate for the bulk of it. In addition to quartz and microcline, biotite is usually present. It is very frequently accompanied by red garnet, less often by sillimanite. Garnet and sillimanite also occur singly or together without biotite; or magnetite may be the only visible accessory. Distribution of the sillimanite is discussed below. A striking feature of some of the granite gneiss is the presence of thin laminae (one to several millimeters thick) of magnetite and red garnet granules. Some of this particular garnet may be spessartite.

The biotite gneiss is gray to slightly pinkish, fine-grained, and locally migmatitic. A number of layers are garnetiferous, and large porphyroblasts of red garnet are characteristic of the weakly mineralized zone at Spruce Mountain Southeast. Rarely, the rock is slightly hornblendic or pyroxenic. Certain zones are sillimanitic; these are interlayered with quartz-rich sillimanite-muscovite gneiss or are adjacent to sillimanite-microcline granite gneiss. By an increase in the amount of microcline, shown by the pinker color of the rock, the biotite gneiss loses its character and comes to resemble the microcline granite gneiss already described.

Pyroxene gneiss appears at two horizons within the unit. The upper layer, 13 feet thick, was cut by holes 3 and 5 at a distance of 26 to 30 feet below the top of the middle unit. Except for hole 8, where the layer is missing, the other holes in the Northwest area entered the unit below this layer. Little of the original character of the gneiss is preserved, for it has been replaced by magnetite, pyrite, chlorite, epidote, and apatite to yield a highly pyritic low-grade ore. The second pyroxene gneiss layer is a lens cut only by hole 9. It occurs within the principal ore zone, centered about 60 feet above the bottom of the unit, and has a total thickness of 28 feet. This pyroxene gneiss is barren, weakly mineralized, or ore bearing. Part of it is metasedimentary, with thin skarn laminae; part of it represents a contaminated granite.

LOWER (GENERALLY PYROXENIC) UNIT

The lowermost unit, like the uppermost, may be termed "pyroxene gneiss," as that rock type is the one most frequently encountered, if not the most abundant. Also present are pyroxene skarn, feldspathic pyroxene skarn, garnet skarn (hole 5), and sporadic layers of "clean" granite and granite gneiss contaminated with pyroxene or hornblende. The pyroxene gneiss itself is fine to medium grained, foliated or laminated, commonly interlayered with skarn, and locally migmatitic. One or two zones of subore are present in holes 3 and 6.

The maximum thickness of the lower unit is more than 300 feet. However, most holes cut only the top 50 or 100 feet. If the correlation discussed below is correct, an upper pyroxenic zone, 45 to 85 feet thick, thins to 5 feet at hole 8 and is absent at Spruce Mountain Southeast (see pl. 18). As this pyroxenic zone thins, there is a corresponding thickening of the underlying zones comprising garnet-biotite gneiss, biotite gneiss, biotite granite gneiss (locally garnetiferous), and sillimanite granite gneiss. At hole 9, these zones are enclosed within the pyroxene gneiss unit. Perhaps the lower pyroxenic rocks also lens out on the northeast limb of the syncline. At Spruce Mountain Southeast, no trace of them was found at the surface or in drill cores.

Finally, there are within the rocks of the several units veins of granite pegmatite, syenite pegmatite, and quartz. In places, these veins carry small amounts of magnetite or sulfides, or both. At a number of places, the core shows small vugs, 1 or 2 mm in diameter, lined with pumpellyite, chlorite, pyrite, or chabazite(?). (See table 2, Nos. 8 and 20.) Elsewhere, thin veinlets of calcite, calcite-epidote, chlorite-pyrite, or stilbite have been observed. (See table 2, No. 21.) The vugs and their coating, and most or all of the vein material, have formed from hydrothermal solutions. Only in the uppermost few feet of drill core and along a few micaceous zones is there any evidence of supergene alteration.

CORRELATION OF MAJOR UNITS

The main units described above can, with a few exceptions, be correlated with assurance from hole to hole in the Northwest area. Similar units are present in the Southeast area, but their equivalence to those in the Northwest area is open to doubt. The units recognized in both areas have been correlated as shown on plate 18, with the understanding that the correlation is only tentative.

The datum selected for correlation of the holes at Spruce Mountain Northwest is the top of the lower (generally pyroxenic) unit—that is, the first part of the pyroxene gneiss unit appearing below the main ore-bearing zone. This horizon is readily identified in all holes except hole 8. There the writer has assumed that the horizon is the top of 5 feet of garnet-biotite-pyroxene-feldspar gneiss with thin layers of garnet skarn and pyroxene skarn, overlying more than 100 feet of sillimanite granite gneiss. The only reason for this assumption is that one principal mineralized zone is present in hole 8, and rocks originally of calcareous composition occur at and near the base of it, as in other holes of the Northwest area.

Other ways of correlating hole 8 deserve mention:

1. The hole may have entered bedrock below the top of the horizon used for a datum in the other holes. One would, in effect, assume that the main ore zone of the other holes had been eroded at hole 8. Then the main mineralized zone of hole 8 represents material stratigraphically 230 feet or more below the bottom of the principal mineralized zone of the other holes.

2. The pyroxene gneiss stratigraphically 54 feet below the bedrock surface in hole 8 may correspond to the datum. The granite gneiss just above it should correspond to the principal mineralized zone, but it is barren and perhaps greatly thinned. The main mineralized zone of hole 8 represents material stratigraphically 175 feet below the bottom of the principal mineralized zone of the other holes.
3. The thick zone of "hanging-wall pyroxene gneiss" (upper, pyroxenic unit) in hole 8 is just that. The 5-foot layer of garnet-biotite-pyroxene-feldspar gneiss below the main mineralized zone is an incidental feature: the true "footwall pyroxene gneiss" (lower, generally pyroxenic unit) lies somewhere below the bottom of hole 8. This means that the granite gneiss and biotite gneiss of the middle unit have been thickened enormously. The main mineralized zone of hole 8 may or may not correspond to the main zone cut in the other holes.

Other horizons that might have been used for correlation are the bottom of the upper (pyroxenic) unit and the top or bottom of the principal mineralized zone. The first would be extremely useful in guiding depth of drilling to ore. Unfortunately, only 3 of 6 holes in the Northwest area entered bedrock above this horizon. Owing to the replacement origin of the ore, the position of the top of the principal mineralized zone varies greatly with respect to the recognizable rock units, and for this reason it has not been used as a datum. Its use would, of course, emphasize the possible equivalence of units in the Northwest and Southeast areas. The bottom of the principal mineralized zone is almost the same horizon as the top of the lower (generally pyroxenic) unit.

The correlation adopted in this report suggests the following features:

1. Two significant mineralized zones are generally recognizable: a subordinate one in the upper part of the middle unit, and the principal one in the lower part of the same unit. (The mineralized zones are not everywhere of sufficient grade or thickness to constitute ore, but they contain conspicuously more magnetite than the rest of the middle unit.)
2. There is a progressive eastward thinning (holes 3, 5, 8) of the rocks between the upper and lower pyroxenic units. The thickness of the middle unit decreases from 256 to 144 feet. Moreover, the percentage of granite to metasedimentary rocks decreases from west to east: hole 3, 30; hole 5, 60; hole 8, 45. Also, the metamorphosed sediments of the western part of the unit are more granitized than their eastern counterparts. These observations are consistent with the concept that more granite magma was injected in the western part.
3. The composition of the footwall rocks changes northeastward, with more (originally) argillaceous material interlayered with the (originally) calcareous material in the upper part of the unit.
4. Southeastward along the northeast limb of the syncline, the "pyroxene gneiss" part of the footwall unit thins out almost completely, leaving only a few feet in hole 8.
5. If hole 8 is properly correlated, the site of granitization has become inverted at that place, with more metasedimentary biotitic material than usual preserved above the datum and less below.

The stratigraphic sequence at Spruce Mountain Southeast is similar to that found in the Northwest area, except that there is no "pyroxene gneiss" unit beneath the mineralized zone. Instead, only biotite granite gneiss, locally garnetiferous, is present. It is possible that the "pyroxene gneiss" has pinched out or has been squeezed out, just as it apparently thinned greatly along the northeast limb at Spruce Mountain Northwest. Because we have no surface or subsurface data on the geology of the area between the Northwest and Southeast areas, we can only infer that the upper (pyroxenic) units, the intermediate rocks, and the principal mineralized zones are equivalent in the two areas. Support for this inference is given by the remarkably close agreement between the thickness of rocks between the bottom of the upper (pyroxenic) unit and the bottom of the principal mineralized zone, and the thickness of the principal mineralized zones, in holes 1 and 2 (Southeast area) and hole 3 (Northwest area).

STRUCTURE

The tract as a whole constitutes a syncline of metasedimentary rocks and granite gneiss, plunging gently southeast from a culmination north of the northwest corner of the property. The syncline is flanked by medium-grained granite or granite gneiss. (See pl. 1; and Prof. Paper 376, fig. 14.) The Spruce Mountain Northwest area contains the nose and parts of both limbs of the syncline where it is relatively open. The structure is outlined by the curving pattern of the magnetic anomalies. The Spruce Mountain Southeast area probably represents the limb of the same syncline, at this latitude tightly and isoclinally folded and dipping steeply southwestward. Though the Southeast segment can be projected through a swamp to meet the southwest limb of the Northwest segment, the relation of the two areas is still uncertain. If, as it appears, Spruce Mountain Southeast corresponds to the southwest limb of an appressed syncline, where is the northeast limb? Neither outcrops nor float of metasedimentary rocks could be found for 2,000 feet northeast of the Southeast belt.[44]

Perhaps the northeast limb has been severed during deformation, or cut out by granite. If the southeast belt corresponds to the northeast limb of the appressed syncline, the only place for the missing southwest limb is in the bed of Big Creek Flow; the rocks farther southwest are granites. If this is so, the fold axis must have been offset between Spruce Mountain Northwest and Spruce Mountain Southeast. Such an offset might be explained by transverse faulting (for which there is no evidence in the immediate area), or by very tight folding. Prospecting for the missing limb would be desirable if there were some assurance of finding ore in it. However, the fishhook pattern of magnetite deposits is so well established, and the missing limb so elusive, that the problem is of interest for its bearing on the structure of the known bodies at Spruce Mountain, rather than for the discovery of new deposits.

SPRUCE MOUNTAIN NORTHWEST

The U shape of the Spruce Mountain Northwest anomaly reflects a syncline, plunging gently southeast, whose nose fits neatly into the northwest corner of the Newton Falls Paper Mill property. Foliation in outcrops near the corner post strikes north-northeast to northeast and dips 20°–25° SE., suggesting that the axis plunges 25° SE. Mineral lineation in an outcrop near the collar of hole 3 trends S. 55° E., parallel to the fold axis, and is horizontal. The apparent plunge of recognizable units between holes 6 and 3 is 8° to 8½° SE.; between holes 9 and 5 it is 6½° to 10½° SE. Thus the inferred plunge of the major synclinal axis is about 8° to 10° SE., steepening toward the nose (see pl. 18). There are no outcrops or drill holes along the southwest limb; hence the amount of dip of the limb is unknown. The northeast limb dips either gently or steeply southwest (see discussion below). Drag folds with an amplitude of tens or a hundred feet may exist on these limbs. The dip in the axial zone is gently rolling but essentially horizontal. One may expect that the axes of these gentle warps will parallel the major synclinal axis in trend and perhaps in plunge. The attitude of the axial plane is unknown; that plane may be vertical, or it may dip steeply southwest.

SPRUCE MOUNTAIN SOUTHEAST

Outcrops are fairly abundant in the Southeast area. They indicate a series of steeply dipping minor isoclinal folds that undoubtedly reflect a major isoclinal fold, presumably a syncline. Many outcrops of pyroxenic migmatite show minor isoclinal folds, less than 2 feet in height and with limbs a few inches apart, having axes horizontal or plunging less than 30° SE. The strike of the foliation is quite uniformly N. 50°–55° W., but in cross section the dip is seen to vary. Recorded dips range through vertical from about 60° SW. to 75° NE., and variations of that magnitude are often observed in a single small outcrop; an average value is 80°–85° SW. This dip, moreover, is most commonly seen in individual exposures.

A belt of "pyroxene gneiss" about 1,000 feet wide crops out along the northeast side of Big Creek, between the creek and the anomaly. The unit includes pyroxene-quartz-feldspar gneiss, pyroxenic migmatites, and subordinate pyroxene skarn. It is succeeded northeastward by a 700-foot-wide belt of granite gneiss and granitic garnet-biotite gneiss. Some of the granite gneiss has disseminated red garnets or carries granite pegmatite veins blotched with garnet. The magnetic anomaly lies along the middle of this belt. An area of drift, 1,000 feet wide, separates these scattered outcrops from the hornblende granite mass to the northeast.

STRUCTURE AT HOLE 8

The dip of the foliation at hole 8 must be either 10° SW. or 80° SW., according to alternative interpretations of the angle between foliation and core axis. The gentler dip agrees better with the very meager information now at hand, and for that reason it has been shown on the cross section, on plate 18. The pattern of the anomaly, however, suggests a steep dip rather than a gentle one. The outcrop nearest hole 8 is folded pyroxene gneiss, 900 feet southeast of the hole, dipping about 15° SW. The collar of hole 8 is almost on the strike of this outcrop. A second outcrop, 1,050 feet northeast of the hole, is sillimanite-biotite granite gneiss, striking north and dipping 10° E. This rock may correspond to some sillimanite-bearing biotite gneiss cut just below the hanging-wall pyroxene gneiss and skarn of hole 8. If so, a minor anticlinal fold is present between the hole and the second outcrop.

SECTION THROUGH HOLES 7 AND 9

Holes 7 and 9 provide the only section through the ore body (see pl. 18). This section is far from satis-

[44] Two other and more remote possibilities deserve mention: the valley east-northeast of Spruce Mountain, and the valley southwest of Moosehead Pond (see pl. 1). Both valleys are filled with glacial debris. The first valley lies 500 to 600 feet beyond the last outcrops that could be found when the area was mapped. The second valley is a possible site for the missing limb only if the structure of Spruce Mountain Northwest is highly convolute.

factory, owing to the ambiguity in interpreting structure from the drill cores. The foliation at hole 9 dips about 25°, but whether northeast or southwest we do not know. Because the hole lies a bit northeast of the axis of the syncline, the southwest dip has been shown on the cross section. At hole 7, the dip must be southwest; but we do not know if it is 10° SW. or 80° SW. (A short hole, 7A, started at the site of hole 7 to check the amount of dip of the foliation, was lost at 12 feet. The core from hole 7A was useless; part of it may have been from a boulder, and the foliation was indistinct. Trenching through the shallow overburden was not attempted.) The gentler dip has been selected for the cross section for two reasons:

1. Apparently, three horizons can be correlated between holes 9 and 7. They are the top of the mineralized zone, the bottom of the zone, and the top of the footwall pyroxene gneiss and related rocks. The structural picture is simpler if these horizons are joined by a gentle southwest dip.
2. Holes 7 and 8, at roughly the same elevation, appear to be on the strike of the foliation, roughly the same distance from the anomaly, and directed toward the same mineralized zone. An imaginary line connecting the points at which the two holes entered the mineralized zone plunges 4° SE.; that is, it is essentially horizontal. This suggests, but does not prove, that the dip of the foliation is in effect the same at both holes. The weight of evidence favors a gentle southwest dip at hole 8, and therefore a gentle dip at hole 7 seems likely.

With one gentle and one steep dip, with steep dips at both holes, or even with gentle dips, there remains the possibility that a sharp but small isoclinal anticline exists between holes 9 and 7.

The correct interpretation of structure along this section is particularly important because of its bearing on additional exploratory drilling, both here and elsewhere in the syncline. If the dip is gentle at hole 9 and steep at hole 7, it is possible that the mineralized zone cut by hole 7 represents rocks that lie stratigraphically far below those carrying the ore in hole 9. To cut this lower zone, if it exists, hole 9 should have been drilled to a depth of 800 or 900 feet. The existence of two separate mineralized zones, 700 feet apart stratigraphically, seems doubtful in view of the reasonably good correlation of rock units and mineralized zones in the various drill holes. However, the possibility of finding such a lower zone should be borne in mind during further prospecting.

ORE

OCCURRENCE

The ore at Spruce Mountain is confined to the Northwest area, where it occurs chiefly in a facies of the microcline granite gneiss of the middle unit. In addition, biotite gneiss, pyroxene-feldspar gneiss, and granitized or migmatized pyroxene skarn serve locally as the host for ore. The ore is generally disseminated or streaky in habit. With two exceptions (holes 5 and 8), the better grade ore is in the lower half of the middle unit. With one exception (hole 6), the mineralization has avoided—or is weakest in—the middle of the unit. Though granite gneiss is the preferred host, the volume of ore does not depend solely on the volume of granite gneiss in the middle unit. However, where granite gneiss is conspicuously less abundant and metasedimentary rocks predominate (hole 8), there is no true ore, only some layers of ore grade too thin to mine.

ORE TYPES AND MINERALOGY

Two principal ore types are present: low-sulfur and high-sulfur. The low-sulfur ore consists of magnetite, with little or no pyrite, replacing various facies of the granite gneiss. This type occurs more frequently, forms thicker layers, has a higher content of magnetic Fe, and constitutes the bulk of the deposit. Assays suggest that a very little ilmenite and a variable but small quantity of hematite accompany the magnetite. The gangue minerals are mostly those of the country rock: quartz, untwinned potassic feldspar commonly rather glassy in appearance, biotite, garnet, and sporadic sillimanite. Apatite is present in microscopic amount. (See table 2, No. 5.) Some of the garnet may well be spessartite. Where the ore has replaced pyroxene gneiss (hole 9), remnants of pyroxene are visible. A little scapolite is present locally in both the granite gneiss and pyroxene gneiss varieties. (See table 6, Nos. 5, 6, and 7, for modes of weakly mineralized granite gneiss and ore-bearing granite gneiss from Spruce Mountain Northwest.)

The high-sulfur ore contains abundant pyrite—10 to 15 percent by volume. Rare chalcopyrite accompanies pyrite or is present in nearby layers of nonpyritic subore. Granular yellow epidote is frequently present and is very abundant in places. Chlorite is often visible with the epidote, and it may occur as extensive replacement areas in the host rock. The potassic feldspar of the host is reddened, or earthy and pinkish. The mafic mineral, pyroxene, is either fresh or chloritized. Brown or pale-green apatite is conspicuous in hand specimens; it forms 5 to 10 percent of the ore in some places. This pyritic ore, 12 to 13 feet thick, was cut only in holes 3 and 5 at a stratigraphic distance of 26 to 30 feet below the upper (pyroxenic) unit. (Holes 6, 7, and 9 entered bedrock below the pyritic zone.) It is limited to a layer of pyroxene-feldspar gneiss enclosed in granite gneiss, and its boundaries are rather sharp, in spite of the fact that it is adjacent to subore and mineralized granite gneiss. Of course, scattered zones with magnetite and considerable sulfides are found at other places in the Spruce Mountain rocks, but they are either too thin or too lean to be called ore.

SPECIAL STRUCTURES OF ORE BODIES

The drilling data now at hand indicate that the ore bodies are more or less sheetlike, generally conformable to the foliation of the enclosing rocks, and confined to two significant zones within a single, continuous middle unit that is overlain and underlain by "pyroxene gneiss." At hole 9, however, the lower (main) ore zone either replaces a lens of pyroxenic rocks now forming part of the middle unit or transects a "formational contact" and replaces the upper part of what is elsewhere the lower (generally pyroxenic) unit. The first interpretation has been adopted for purposes of emphasizing correlation of all units, ore included, but the second may be nearer the truth (cf. the magnetite deposits at Benson Mines, Deerlick Rapids, and Skate Creek).

Linear structures have already been described under general structure. Whether lineation trending southeast and plunging about 10° SE., essentially parallel to the synclinal axis, has given a shoot structure to the ore bodies, we cannot tell until the deposit has been further explored. Certainly the two lobes or strands of magnetic anomaly within the U suggest it, and the possibility should be kept in mind during exploration.

GRADE

The ore is consistently low grade. One or two 5-foot samples of core assayed 37 to 40 percent total Fe (perhaps 35 to 38 percent magnetic Fe), only to be balanced by samples in the same ore zone assaying about 17 percent magnetic Fe. Much of the material now classed as "waste" carries perhaps 10 percent magnetic Fe. Reed and Cohen (1947, p. 17–25) give all the available analytical data. This information has been summarized in tables 30 and 31 of the present report. Total Fe alone was determined on much of the core. In places, at least a rough value for magnetic Fe was desirable; here the writer converted total Fe to magnetic Fe by using variable but conservative factors based on knowledge of the mineralogy of the sampled core.

Table 31 points out the grade and thickness of the two distinct ore zones. In this recalculation, thickness has been increased at the expense of grade, particularly at hole 7, to give the maximum thickness of ore averaging 23 to 26 percent magnetic Fe. It is at once apparent that the lower zone is generally thicker and slightly higher in grade.

Unfortunately, there is no analysis of ore from the upper zone for P, S, TiO_2, etc. The pyritic, apatite-rich type of ore forms much of this zone in holes 3 and 5. However, the bulk of the sulfur from pyrite and phosphorus from apatite would be removed by magnetic separation. Analyses of the magnetic fraction of ore from the lower zone were made for holes 7 and 9 only. These may be roughly representative of the zone as a whole. The "composite of all composites" (Reed and Cohen, 1947, p. 19) is not particularly meaningful, as it represents subore as well as ore and includes material from two different ore zones.

In the absence of mineragraphic studies and systematic assays for both soluble Fe and magnetic Fe, it is impossible to say how much iron oxide is in the form of hematite. Comparison of the analyzed total Fe-magnetic Fe pairs (table 30) with the logs suggests that there is roughly 1 percent Fe in hematite for every 10 percent Fe in magnetite.

ADDITIONAL PROSPECTING

The main area for prospecting is on and within the U of the Spruce Mountain Northwest anomaly. In addition, part of the central area between the Northwest and Southeast anomalies deserves attention. An outline and discussion of certain features of the prospecting is presented below.

DIAMOND DRILLING

Further diamond drilling at Spruce Mountain Northwest should be directed toward solving structural problems, as well as toward proving up tonnage. Several of these structural problems might be solved incidentally in the search for ore, but keeping them in mind as separate problems should permit more intelligent exploration and prevent their being disregarded as economically unimportant. Such problems include:

1. The identity of the principal ore zone in holes 7 and 9, a problem directly related to number 2.
2. The dip of the northeast limb—10° SW. or 80° SW.
3. The dip and existence of ore on the southwest limb of the syncline.
4. The existence of ore shoots parallel to the synclinal axis.
5. The structural relation of the Northwest and Southeast areas. This problem is not of immediate

TABLE 30.—*Grade and thickness of the more important magnetite-bearing layers, Spruce Mountain Northwest deposit*

[Includes all layers of ore grade having an apparent thickness of 5 feet or more, as well as certain thick layers of subore and mineralized rock. All analytical data from Reed and Cohen (1947, p. 17–25); some figures represent averages of their data]

Hole	Footage [1]	Apparent thickness (feet)	True thickness (feet)	Grade (weight percent)							
				Crude ore		Magnetic fraction only					
				Total Fe	Magnetic Fe	P	S	TiO_2	SiO_2	Mn	Fe
3	155–170	15	12.9	21.5	14.5						
	(2)										
	270–290	20	17.2	17.4	15.2						
	(2)										
	320–335	15	12.9	20.1	[3] 18						
	335–365	30	25.8	26.4	23.5						
	365–390	25	21.5	34.5	32.5						
	390–410	20	17.2	24.2	22.0						
	410–416	6	5.2	20.3	[3] 18						
5	80–105	25	20.2	18.1	13.3						
	105–150	45	36.4	26.6	21.5						
	150–179	29	23.5	16.3	11.0						
	179–197	18	14.6	~13.5	[3] 9?						
	(2)										
	240–270	30	24.3	13.0	[3] 8?						
	(2)										
	290–363	73	59.1	~12.7	[3] 8?						
6	28.5–38	9.5	9.2	26.8	25.4						
	38–59	21	20.4	18.8	17.1						
	59–129	70	67.9	13.6	[3] 9?						
	129–194	65	63.0	28.9	26.6						
	194–204	10	9.7	16.5	[3] 14?						
	(2)										
	244–249	5	4.9	21.4	20.1						
	(2)										
	254–259	5	4.9	16.6	[3] 14?						
7	56.7–69	12.3	10.1	13.6	[3] 9?						
	69–104	35	28.7	33.9	30.6	0.014	0.21	0.80	2.29	0.32	67.2
	104–124	20	16.4	15.4	[3] 11.5?						
8	[4] 314–318.1	4.1	3.3		Low-grade						
9	11.7–22	10.3	9.3	14.8	[3] 10.5?						
	(2)										
	92–122	30	27.0	21.1	[3] 17?						
	(2)										
	152–172	20	18.0	28.8	25.9	0.004	0.14	0.74	1.73	0.21	68.2

[1] Footages correspond to those sampled and assayed by the Bureau of Mines. They may differ slightly from those given in the logs prepared by the Geological Survey.
[2] No data.
[3] Calculated from value for total Fe where value for magnetic Fe had not been determined chemically. The conversion factors are variable and conservative.
[4] This is the thickest layer of ore grade in hole 8.

TABLE 31.—*Grade and thickness of ore zones, Spruce Mountain Northwest deposit*

[Based on data from table 30. Grade and thickness of ore have been balanced to give maximum thickness of ore averaging 23–26 percent magnetic Fe]

Hole	Footage	Apparent thickness (feet)	True thickness (feet)	Grade (weight percent magnetic Fe)
Upper zone				
3	155–170	15	13	[1] 14.5
5	105–150	45	36	21.5
6	28.5–42	13.5	13	23
7	(2)			
8	(3)			
9	(4)			
Lower zone				
3	320–416	96	83	~24
5	(5)			
6	129–204	75	73	~25
7	69–124	55	45	~24
8	(3)			
9	152–172	20	18	25.9

[1] Value seems low. Visually estimated as low-grade ore in logging.
[2] Collar is below zone.
[3] Zone is too thin to constitute ore.
[4] Collar is below most of zone?
[5] Zone is only weakly mineralized.

concern but might become important if ore were found in the area between latitudes 6000N and 8000N.

Bulldozing and trenching could supplement drilling to advantage in 2 and perhaps 3, above.

MAGNETIC WORK

Additional magnetic work in this area warrants discussion.

1. *Detailed work with sensitive instruments.*—A large area between 4000N and 9000N was covered by reconnaissance dip-needle traverses, followed by detailed work in one small area. The readings were low, generally less than +10°. In spite of the low readings, at least the northern half of this central area may be favorable prospecting ground, provided the syncline has maintained its gentle southeastward plunge. A plunge of 8°–10° SE. will cause the rock units to decline southeastward about 16 feet vertically for every 100 feet of hori-

zontal distance along the plunge. Thus the top of the ore shoot cut at 129 feet in hole 6 would, if continuous, occur at a depth of 529 feet at latitude 8000N, 2,500 feet southeast of hole 6, assuming ground elevations to be about the same. A magnetite body with 500 feet of cover is considerably beyond the effective "reach" of the dip needle, and its detection would require a more sensitive magnetic instrument, such as a vertical magnetometer. A magnetometer survey might be of particular aid in detecting and outlining ore shoots as opposed to sheets, if they exist. The considerable cost and uncertain prospect of such a survey would have to be balanced against the cost of wildcat drilling that might still leave a favorable area inadequately prospected. Possibly a Hotchkiss superdip could be used instead of a vertical magnetometer. Regardless of the instrument used, the magnetic pattern will almost certainly be complex, owing to multiple mineralized zones, shoots, structural complications, termination of ore, or the combined effect of these features.

2. *Reconnaissance dip-needle work.*—Two areas deserve attention in the future. Scattered magnetic highs have been found east of the Northwest area by A. M. Ross. Systematic reconnaissance of these highs and determination of their relation to the main Spruce Mountain anomaly may disclose other magnetite deposits of interest.

The problem of the "missing limb" of the inferred tight fold at Spruce Mountain Southeast was discussed in the section on structure. Whether the undiscovered limb has any economic importance in its own right, we obviously cannot tell. It has an indirect importance that may become great if ore is found in the central area, for it may be necessary to know something about the "missing limb" in order to understand the structure of the central area. There are two reasonable sites for the limb: southwest of the Southeast anomaly, in the bed of Big Creek Flow; and within 1,000 feet northeast of the anomaly, in an area of drift. If the limb carries more magnetite than the enclosing granite gneiss does, it should be possible to detect the limb by reconnaissance dip-needle traverses, in the hope of putting it to geologic or economic use.

ECONOMIC POSSIBILITIES OF SPRUCE MOUNTAIN SOUTHEAST

The economic possibilities of the Southeast area do not seem good. The weakly mineralized garnet-biotite gneiss underlying the anomaly averages 6 to 7 percent total Fe (Reed and Cohen, 1947, p. 17–18), of which one-fifth to one-fourth would be in the silicates garnet and biotite. Unless microcline granite gneiss takes the place of the garnet-biotite gneiss at greater depth, there is little likelihood that thc weakly mineralized rock is masking a deeply buried ore body. In the district, ore in garnet-biotite gneiss is restricted to thin layers forming parts of granite gneiss sheets; garnet-biotite gneiss alone never is the host for an ore body.

VANORNUM (SWEET POND) DEPOSIT

The south end of the VanOrnum (Sweet Pond) deposit is 2,000 feet west-southwest along the Russell-Fine town line from the south end of Sweet Pond, south-central rectangle, Russell quadrangle. A sand road leads to Sweet Pond from the macadam road connecting Degrasse with N.Y. Highway 3, east of Fine.

To reach the deposit, follow the blazed trail from the north end of Sweet Pond west and south to the anomaly. Alternatively, walk west along the town line from the point where it crosses the macadam road noted above. The first route is less swampy.

Magnetic attraction on the VanOrnum tract, west of Sweet Pond, was observed many years ago. A few very old test pits, showing magnetite in silicated marble, lie 1000 feet north-northeast of the north end of the VanOrnum anomaly, or 2,500 feet west-northwest of the north end of Sweet Pond, close to the blazed trail leading westward from the pond. (See pl. 1.) The VanOrnum anomaly and the northern magnetite-bearing area were registered as slight "hills" on the aeromagnetic survey of 1945 (Balsley, Buddington, and others, 1954b).

The M. A. Hanna Co. made a detailed dip-needle survey of the anomaly and tested it by two diamond-drill holes in 1946. (See fig. 27.) The holes, about 200 feet deep, cut two zones of subore, 5 to 10 feet thick, separated by 3 to 15 feet of weakly mineralized rock. Parts of the wallrocks were also weakly mineralized with magnetite. Overburden at the collars of the drill holes was only 3 or 4 feet thick.

The host rock of the subore is heterogeneous, including quartz-rich garnet-biotite-microcline granite gneiss, pyroxene-microcline granite gneiss, and quartzose pyroxene skarn. To judge from the analyses of the subore, some of the garnet is spessartitic.

The wallrocks are mainly biotite-quartz-plagioclase gneiss, pyribole gneiss (locally scapolitic), and several facies of microcline granite gneiss, including biotite-microcline granite gneiss (in part garnetiferous) and sillimanite-biotite granite gneiss. The metasedimentary rocks are locally migmatitic; schorl is abundant in some of the biotitic migmatites. The foliation in wallrocks and some of the subore is commonly transected by close-spaced fractures and microscopic shear zones, in part filled with calcite. Slight to moderate

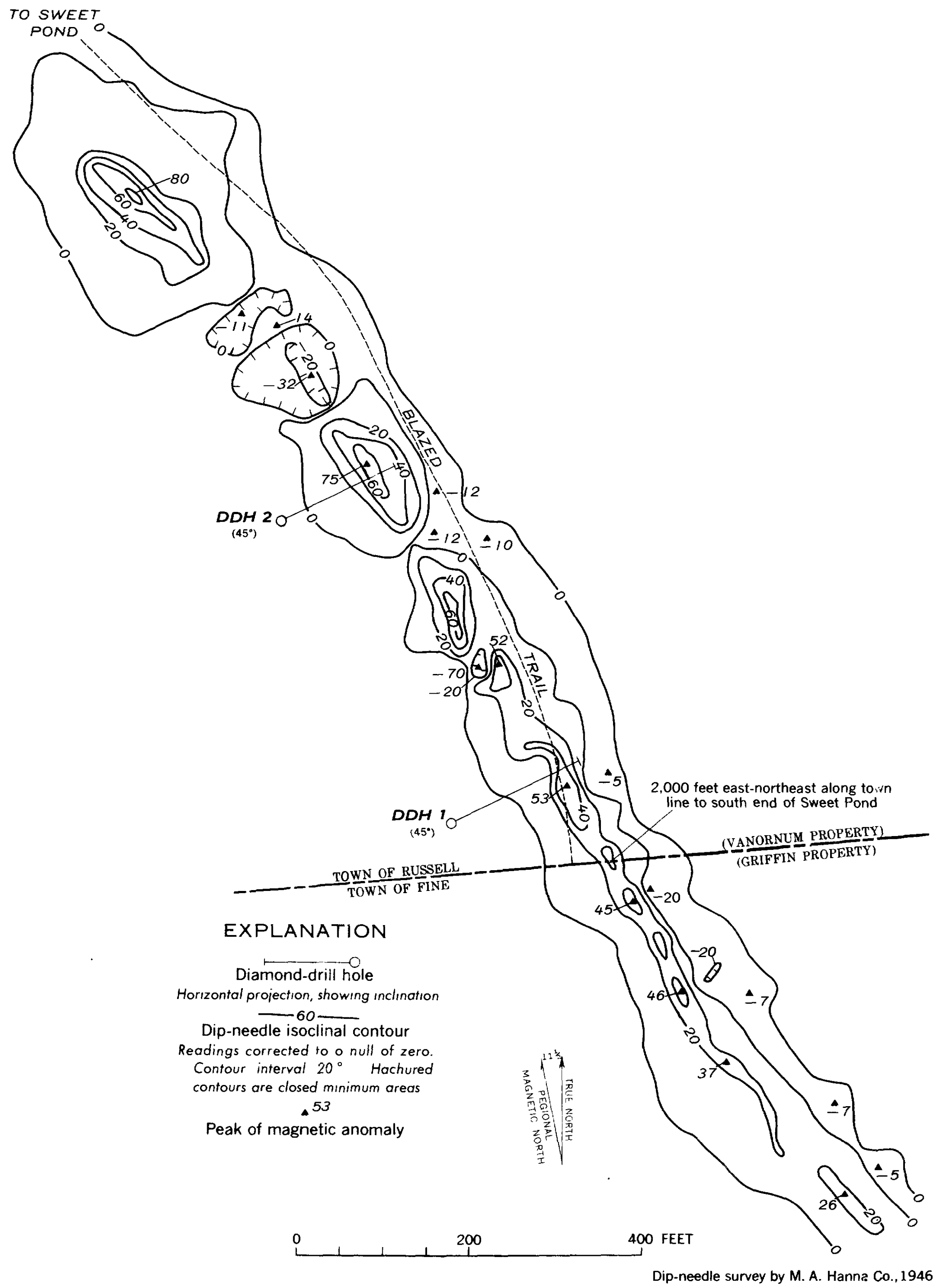

Figure 27.—Magnetic map of the VanOrnum (Sweet Pond) anomaly, south-central rectangle, Russell quadrangle.

alteration of wallrocks to chlorite, calcite, hematite, epidote, and pumpellyite is characteristic of specimens from the drill cores.

The zone of subore and mineralized rock strikes N. 30° W. and dips 55° SW. The trend of the anomaly is slightly athwart the dominant structural trend of rocks in this neighborhood (see pl. 1), suggesting that the mineralized zone may be on one limb of a local, nearly isoclinal fold whose amplitude is several hundred feet.

The belt of metasedimentary rocks and granite gneiss in which the VanOrnum deposit lies, is flanked on the west by a thin body of alaskite gneiss (a few feet of which was cut just beneath the overburden in hole 1), and on the east by hornblende granite gneiss. (See pl. 1.) The deposit is close to a major mylonite zone, in an area of diverging lineations. (See Prof. Paper 376, fig. 15.)

WIER PROSPECT

The Wier prospect, noted by Shaub (1949, p. 3 and pl. 1), is about 2,300 feet north of Little Lake, near South Russell, central rectangle, Russell quadrangle. As shown on our plate 1, this area is underlain chiefly by sillimanite-microcline granite gneiss. Shaub (1949, p. 3) reports that the "ore" occurs as streaks in a feldspathic rock. The small positive anomaly is presumably due to slight local concentrations of magnetite.

MAGNETITE DEPOSITS IN MODIFIED ROCKS

TREMBLEY MOUNTAIN DEPOSIT

Because of its geologic and potential economic significance, the Trembley Mountain area is treated in considerable detail in this report. The magnetite deposits of the area bridge the gap between the typical skarn ores and the typical granite gneiss ores. In this regard, then, the Trembley Mountain deposits are a freak, but petrologically a highly instructive freak. Individual rock types are varied and their relations complex; hence descriptions are necessarily longer than usual. The area is complex structurally. The structural interpretation advanced may not be correct in detail, but the general pattern seems clear. A knowledge of it is essential for the continued exploration of the area. Finally, the deposits are promising enough economically to warrant the study we have been able to give them.

Geologic field work extended over parts of the seasons 1943, 1944, and 1945, requiring about 13 man-days. A good part of this time was spent in hunting for outcrops and cleaning them. The few fortuitous outcrops are small and mostly covered with moss, humus, or a gray film that makes accurate observation of structural data difficult or even impossible. Some of the observations are based on exposures showing not more than a 2-inch length of foliation. Moreover, lineation is so strong in some outcrops that it is easily confused with the planar structure. Certain key outcrops were revisited a year after the first hand-stripping (done during a wet season), and the original observations were confirmed or corrected.

The magnetite deposits of the Trembley Mountain area, explored by 9 drill holes, may contain a very large reserve of low-grade ore averaging about 25 percent magnetic Fe. The soluble Fe content of the deposits is about 27 percent. Two percent of soluble nonmagnetic Fe appears to come from partial solution of pyroxene in the host rock; soluble nonmagnetic Fe from this source would not be recoverable.

The deposits are of modified skarn type. They are part of a structurally complex tract of metasedimentary rocks and microcline granite gneiss sheets that is enclosed by hornblende-microperthite granite. The magnetite deposits are conformable replacements confined to a southeastward-plunging overturned isoclinal syncline whose gross structure is complicated by the presence of many minor, satellitic folds. (See p. 49–51 for a detailed geologic summary of the deposits.)

The magnetite deposits and their associated magnetic anomalies (Trembley Mountain North and Trembley Mountain South, respectively) are in parts of the same overturned syncline. They may be parts of a single folded zone, or they may lie at two separate stratigraphic positions. Both overburden and ore are relatively thin at Trembley Mountain North and in the northern part of Trembley Mountain South. Ore at the south end of the drilled area is perhaps 80 feet thick, thickened by repeated folding of its host rock; but it is covered by 150 feet of unconsolidated overburden. The character of the dip-needle and aeromagnetic anomalies suggests that this thick zone of ore may continue southeastward for a considerable distance beneath 150 to 250 feet of valley fill.

LOCATION, OWNERSHIP, AND ACCESS

The Trembley Mountain area is in the north-central rectangle, Cranberry Lake quadrangle, 3.8 miles east of the hamlet of Newton Falls. The property and mineral rights are owned by the Newton Falls Paper Mill, Newton Falls, N.Y.

During the fall of 1945, the Bureau of Mines built a temporary bridge across the Oswegatchie River about 2 miles below the outlet of Cranberry Lake and approached the area on the old tote road running southwest from the river. The road and bridge have since been abandoned. The usual method of reaching the area is to cross the Oswegatchie by boat at a point 1 mile east of Tooley Pond Outlet and walk south along the eastern boundary of the company's land.

DISCOVERY AND EXPLORATION

The Trembley Mountain deposits were discovered in 1942 by the reconnaissance dip-needle work of A. M. Ross. In the summer of 1943, the Newton Falls Paper Mill made a systematic dip-needle and dial-compass survey of the Trembley Mountain area. At the same time, the Geological Survey made a geologic sketch map (scale 1 : 3600) of the area, using the company's grid for control. Structural information was the first concern of the geologic work. The initial mapping was supplemented by additional work in 1944 and 1945. The company drilled 6 relatively short holes (T1 through T6) during the winter of 1943–44. The results of this drilling were promising enough to warrant further prospecting, which was undertaken by the Bureau of Mines in the summer and fall of 1945. At that time a substantial part of the area was resurveyed by dip needle, and 3 deeper holes (11, 12, and 13—in this report prefixed by the letter "T") were drilled. The Geological Survey cooperated with the Bureau of Mines in planning the drilling, and the Survey logged the core and presented its interpretation of the results. Results of more recent private exploration of these deposits were made available to the Geological Survey in 1958. It has not been possible to incorporate the new data or their interpretation in this report.

MAGNETIC ANOMALIES

The pattern of magnetic anomalies is remarkably complex. (See pl. 19.) This complexity reflects two significant controls: the complex structure of the ore bodies, and the varying depth of overburden. Structural control, briefly discussed in this section, is treated more fully elsewhere.

The main hook-shaped areas of anomalous magnetic attraction (designated on pl. 19 as Trembley Mountain North and Trembley Mountain South anomalies) outline the noses of one or more inferred overturned isoclinal synclines that plunge south-southeastward. It is possible that some of the individual highs and lows in these hook-shaped belts represent shoots of magnetite along the axes of minor folds satellitic to the major structure.

The contrasting effects of shallow and deep overburden are illustrated by the patterns of the anomalies north and south of latitude 7600N (see pl. 19). North of that latitude, the "magnetic relief" is very great: close-spaced highs and lows dot the hook-shaped belts of anomalies. All these anomalies lie near the crest or on the south flank of a hill. Outcrops are relatively abundant, and information from drilling indicates that the maximum thickness of overburden immediately above the magnetite-bearing rocks is perhaps 25 feet.

South of latitude 7600N, the clutter of magnetic "hot spots" gives way to a broad lobe or fan of low positive readings extending far to the south-southeast. This part of the anomaly is in an east-trending valley filled with glacial debris. The known depth of overburden is 95 feet at hole T3, 141 feet at hole T13, and 149 feet at hole T12. From magnetic data lent by A. M. Ross, the writer infers that overburden may be as much as 250 to 300 feet deep in parts of the area south of latitude 7000N.

Actually, the thicker zones of better grade ore were cut in holes T3, T12, and T13. Here we have an instructive example of thick ore, very deeply buried, giving a broad area of relatively low positive readings. None of the showier, thinly covered "hot spots" north of latitude 7600N has yet been found to overlie a zone of ore comparable in thickness to that cut in holes T3, T12, and T13.

The Trembley Mountain deposits registered as a prominent peak of 7,070 gammas on the aeromagnetic survey of 1945 (Balsley, Buddington, and others, 1954a). The marked southeastward extension of the aeromagnetic anomaly confirms the south-southeastward extension of low positive dip-needle readings. Both suggest that the thick zone of ore cut in holes T3 and T13 may continue south-southeastward for a very considerable distance. (The westward-projecting lobe of the aeromagnetic anomaly probably reflects the presence of the adjoining Jarvis Bridge Extension deposit.)

Plate 19 shows that the Trembley Mountain South adjoins the Jarvis Bridge Extension. The structural relation between these two areas of magnetic anomalies is unknown. No outcrops have been found in the critical area; in fact, overburden is probably more than 150 feet deep (the depth found in hole T12) and possibly 250 feet deep. Structurally, it is possible that Trembley Mountain South and Jarvis Bridge Extension are separated by a closely appressed anticline whose axis trends south-southeastward. (See also description of Jarvis Bridge Extension, p. 176–177.)

GEOLOGY

The Trembley Mountain area is a small structural knot—a sort of embayment—in the much larger Clare-Clifton-Colton belt of metasedimentary rocks and intercalated granite sheets. (See pl. 1.) In this neighborhood, the main belt of metasedimentary rocks, generally synclinal in structure, trends eastward and is flanked by masses of hornblende-microperthite granite. This rock shows an alaskitic facies about the north and east ends of the tract of metasedimentary rocks at Trembley Mountain. The area is a structural node about which lineations are highly variable in trend, ranging from dominantly southward to dominantly eastward in the space of a few square miles. (See Prof. Paper 376, fig. 15.) Within the node, however, the major lineation trends rather consistently south-southeastward.

ROCKS

BIOTITE GNEISS AND GARNET-BIOTITE GNEISS

Biotite gneiss forms the footwall of the ore zone in the limited area tested by holes T5 and T6. Locally, the gneiss is slightly mineralized, and at one place (T5) it encloses a 2-foot layer of sulfide-rich ore in leached, chloritized gangue. Sheets of microcline granite gneiss and pegmatite accompany the biotite gneiss. The hanging wall of the southern area contains much biotite gneiss with microcline granite gneiss and granitized gneisses carrying pyroxene, hornblende, or biotite. (See drill section T12–T13, pl. 19.) Locally, a slightly pyroxenic biotite gneiss occurs within the skarn (T11). Thin hornblendic zones are present in some biotite gneiss. These seem more resistant to erosion and probably account for the presence of "amphibolite," rather than biotite gneiss, in many outcrops of the wall rocks. The hornblendic zones may be accompanied by blotchy areas of hornblende-bearing white granite or pegmatite. Generally, pegmatite veins in biotite gneiss carry biotite, but in some places a gray-green pyroxene is present instead.

Garnet-biotite gneiss, a distinctive variety of the biotite gneiss, forms the footwall of the zone of modified skarn in holes T12 and T13. Garnet-biotite gneiss is also present in footwall rocks cut by holes T1 and T2. Traces of garnet-biotite gneiss are present locally within the zone of modified skarn.

The biotite gneiss is commonly gray, locally pinkish, fine grained, and weakly or indistinctly foliated. In places, the rock is slightly migmatitic, showing flat lenses, augen, or thin layers of pink feldspar. Biotite and gray or pink feldspar are readily identified in all specimens, and in some of these one may see pyroxene, magnetite, a little hornblende, or pyrite, singly or together. Weathered surfaces of the rock are neutral or brownish and somewhat sugary, closely resembling exposures of the pink microcline granite gneiss.

Microscopically, the rock is generally fine grained, inequigranular, and crystalloblastic, with a foliation given by subparallel biotite flakes and by variations in mineral composition. Curved grain boundaries for quartz and feldspar are usually predominant, though some specimens show a polygonal mosaic. Rarely the structure is crisscross, or granoblastic with elongate quartz.

Biotite, quartz, microcline, and plagioclase are essential minerals. Of these, the feldspars are most abundant, commonly forming 65 to 70 percent of the rock. The quartz content is about 15 percent or less, locally negligible. Biotite, too, is relatively sparse, 10 percent or a bit more. A little amphibole is present in some specimens; pyroxene is rare.

Ubiquitous accessories are magnetite, apatite, and zircon; with pyrite and sphene fairly common. Allanite is an important accessory in some specimens. A little secondary epidote, chlorite, and carbonate occur sporadically.

Three subtypes of the rock may be distinguished according to the ratio of plagioclase to microcline. Where plagioclase dominates, its composition is An_{23-37}. The plagioclase of one specimen ranged from An_{27} to An_{37}; elsewhere the An content was uniform. In rocks having roughly equal proportions of plagioclase and microcline, the plagioclase is near An_{15}. Where microcline was dominant, the plagioclase could not be determined. It was associated with a little myrmekite.

Microcline in the biotite gneisses is often slightly microperthitic, and the sharpness of its "grid structure" is variable.

A mode of the biotite gneiss is given in table 32, No. 1.

Garnet-biotite gneiss that forms the footwall of the zone of modified skarn is a fine- to medium-grained, foliated migmatite containing 20 to 40 percent of injected whitish granite or granite pegmatite. Foliation in the garnet-biotite gneiss may be distinct or indistinct; commonly it is somewhat swirled. In places, the gneiss is slightly pyroxenic or hornblendic. Pegmatite seams within the gneiss may contain pyroxene instead of the more usual biotite; locally, these pyroxene xenocrysts show coronas or replacement areas of garnet. The garnet-biotite gneiss commonly carries several percent of magnetite, with or without sporadic sulfides. Only one specimen of typical garnet-biotite gneiss has been examined in thin section. This specimen, except for the presence of garnet, is mineralogically similar to the biotite gneiss previously described. Potassic feldspar (untwinned) is much more abundant than plagioclase, which is almost wholly altered to aggregates of sericite, zoisite, and pumpellyite. These aggregates are the bluish or greenish "pseudomorphous chlorite" noted in drill logs.

In a few places, the garnet-biotite gneiss is highly quartzose, essentially a garnet-biotite quartzite with very sparse potassic feldspar.

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PYROXENE AND AMPHIBOLE GNEISSES

Though two rather distinct mineralogic types exist, one with amphibole as the dominant mafic, the other with pyroxene, the gneisses of this group have a similar distribution and show similar variations in composition. As the rocks seem to have no special significance genetically, or value as markers, mistaking one type for the other while logging core is not serious.

Both types occur in the hanging-wall sequence of granitized gneisses and within the zone of modified skarn. Pyroxene gneiss is abundant beneath the zone of subore in hole T4, where it is interlayered with skarn and granite pegmatite. Pyroxene gneiss also occurs as thin layers in the footwall granite and garnet-biotite gneiss. Amphibole gneiss is less often observed within the modified skarn. However, it has been noted in the quartz rock, and in hole T6 it forms a 6-foot layer (highly granitized) between the ore and silicated marble.

The gneisses are gray or greenish, locally pinkish, fine grained, foliated, and locally massive. Much of the pyroxene gneiss has a laminated appearance. A greenish mafic mineral, often too fine to identify except as "pyribole," and gray or greenish feldspar are the only recognizable minerals in most specimens. Occasionally, quartz, biotite, or specks of sphene and pyrite can be distinguished. Many specimens look like granite gneiss, syenite gneiss, or aplite. The problem of their classification is considered in a later section.

Microscopically, the gneisses are fine grained, somewhat inequigranular, foliated, crystalloblastic aggregates of polygonal individuals. The foliation is given by differences in composition, subparallelism of hornblende or biotite, or elongation of feldspars. These features may occur singly or together. Rarely, the biotite may be decussate in a rock that has compositional layering. Slight deformation of the polygonal fabric, perhaps with introduction of elongate quartz and potassic feldspar, is suggested by the evidence in some slides. No differences exist in the fabric of the two mineralogic types.

Essential minerals are feldspar, usually quartz, and either pyroxene or amphibole. The mafic content ranges from 3 to 25 percent but is ordinarily less than 10 percent. Magnetite, apatite, and zircon, usually with sphene, are the accessories. Biotite is sometimes an abundant associate, though it may be lacking altogether. Sulfide grains and rims are present in half the specimens studied; allanite, myrmekite, and blue-green secondary pumpellyite in a few; and garnet, scapolite, and clinozoisite in one each. Chlorite and carbonate are sporadic secondary products.

Quartz, locally absent, is generally present to the extent of 10 to 15 percent; rarely, it may reach 20 to 35 percent.

Plagioclase and potassic feldspar may occur separately or in combination. Potassic feldspar usually predominates. It exists as two types: microcline, and untwinned potassic feldspar with $(-)2V$ near 40°, $r>v$, distinct. Either one may be very slightly microperthitic. Microcline is the principal potassic feldspar in the hanging-wall gneisses, whereas untwinned potassic feldspar is the principal one for similar rocks occurring within the zone of modified skarn. Though the two generally occur separately, some specimens show both types.

Plagioclase in the hanging-wall gneisses is An_{25-27} (locally, An_{30}). Within the zone of modified skarn, it is usually sparse, somewhat altered, and apparently variable in composition. Locally, it is almost completely scapolitized.

There is no relation between the potassic feldspar-plagioclase ratio and the kind of mafic mineral (pyroxene or amphibole), or between the type of potassic feldspar and the kind of mafic mineral.

The pyroxene has not been studied in detail. The color varies from specimen to specimen, both in the hanging-wall gneisses and within the zone of modified skarn. It is always at least faintly greenish, and the darker types show distinct pleochroism. Probably it ranges from a diopsidic variety through salite to ferrosalite. In some specimens the pyroxene is partly altered to yellow chlorite, chlorite-calcite, epidote, or bluish amphibole (not the same as the primary amphibole described below).

Two types of amphibole are present in the pyroxene and amphibole gneisses. Without exception, the hanging-wall gneisses carry "common" green hornblende. The same type of rock within the modified skarn zone contains bluish-green amphibole. The break between the two types is as sharp as the demarcation between hanging wall and skarn: a specimen (T12.15 at 412 feet) one foot above the contact still contained the green hornblende found throughout several hundred feet of wallrocks. Rarely, the amphibole shows slight alteration to yellow-brown mica.

Because of the highly variable composition, especially with regard to feldspars, modes for these gneisses are not given.

MICROCLINE GRANITE GNEISS AND SYENITE GNEISS, WITH SUBORDINATE PEGMATITE

Microcline granite gneiss and syenite gneiss crop out in the belt separating the north and south areas of metasedimentary rocks. (See pl. 19.) Similar rock appears to form a thin shell separating the metamorphosed sediments from the surrounding mass of hornblende-microperthite granite, or from the alaskite (see section on structure). The microcline granite gneiss and syenite gneiss are not homogeneous units: numerous metasedimentary layers are present in the belts mapped as "granite gneiss," and thin sheets of granite gneiss are scattered through the metasedimentary rocks. The larger masses of "granite gneiss" are quartz poor or syenitic, whereas rock of similar appearance in the zone of modified skarn and its footwall gneisses is normal microcline granite gneiss and aplite.

Aplite is common in the footwall gneisses cut in hole T12. It occurs as sheets 1 to 5 feet thick throughout some 30 feet of garnet-biotite gneiss, quartz rock, and granite pegmatite. A few thin, isolated sheets are found in the skarn zone and in footwall pyroxene gneiss.

Except as thin seams in some microcline granite gneiss and migmatitic gneiss, granite (and syenite) pegmatite is sparse. A few thin layers are present in skarn, pyroxene gneiss, and garnet-biotite gneiss.

The microcline granite gneiss and syenite gneiss are pink or gray, fine grained (locally medium), and foliated (locally speckled or mottled). In part, they are homogeneous, though films, schlieren, or thin layers rich in biotite, hornblende, or pyroxene are frequently seen. Thin pegmatite seams, parallel to the foliation, may be present in otherwise homogeneous granite. Outcrops are buff or pinkish, with a sugary weathered surface.

The aplite is pink or gray, finer grained than the normal granite, and homogeneous.

The pegmatite is either pink or greenish and usually contains clots of recrystallized pyroxene. Greenish pegmatite is found in pyroxene gneiss or skarn, whereas pink pegmatite is present between skarn and the footwall gneisses, or within those gneisses.

Microscopically, the granite gneiss and syenite gneiss are fine grained (rarely medium grained), inequigranular, and granoblastic. Grain boundaries are irregular and nonpolygonal. Subparallel biotite flakes, locally with elongate grains of microcline, give a well-developed foliation.

The aplite also has a granoblastic appearance. However, a similar fabric is characteristic of aplites that are presumably undeformed (Johannsen, 1932, v. 2, p. 91–92).

The pegmatite is coarse and undeformed, with sutured grain boundaries and interstitial myrmekite.

Quartz, microcline, plagioclase, and either biotite or hornblende are essential minerals of the granite gneiss. Magnetite, apatite, and zircon are accessory. Allanite, sphene, and pyrite occur less frequently. Slight secondary alteration to carbonate and chlorite is observed in some places.

The quartz content of the normal granite gneiss is about 20 to 25 percent. A quartz-poor type (10–15 percent quartz) is probably greater in bulk here, but this cannot be proved at present, owing to the difficulty of sampling the poor outcrops of the central "granite gneiss" mass.

The microcline occurs with and without very slight perthitic intergrowth (tiny blebs and pinpricks). Both larger and smaller grains may show this feature, suggesting that deformation of primary, slightly microperthitic microcline took place at a temperature high enough to prevent unmixing during deformation except to a very limited extent. Infrequently, a few grains of "shadowy" microcline (grid structure poorly developed) and untwinned potassic feldspar are noted. The optic axial angle in the microcline is near 75°.

Plagioclase, a minor constituent in the granite, is either An_{12-15} or about An_{25}. It is often altered to a fine, unidentified aggregate. A few grains of myrmekite are present locally.

Much of the brownish biotite is slightly altered to chlorite and epidote. Common green hornblende occurs in granites of the hanging wall and footwall. It is generally poikiloblastic and more or less altered to chlorite, or to chlorite, biotite, and magnetite(?) dust.

The syenite gneiss and aplite have about the same minerals as the granite gneiss. Quartz is absent, or present only as an accessory, in the syenite gneiss. Plagioclase may be more abundant than in the granite gneiss. In one specimen, sodic plagioclase forms partial rims on some microcline grains. There is no plagioclase in the aplites.

Modes of typical granite gneiss, syenite gneiss, and aplite are given in table 32, Nos. 2, 3, 4, and 5.

The pegmatite was not studied in detail. It is granitic or syenitic, and almost always contaminated by aggregates of recrystallized pyroxene. A typical specimen of the syenitic variety contains shadowy, slightly perthitic microcline, a few partly altered plagioclase crystals, myrmekite, accessory quartz, pyroxene (in part, altered to green hornblende and colorless to golden-brown mica), magnetite within pyroxene and also interstitial to the felsic minerals, late sulfides, and a few apatite and zircon grains. The myrmekite, which is locally altered to carbonates, replaces microcline slightly

TABLE 32.—*Modes of representative biotite gneiss and granitic rocks, Trembley Mountain deposit*

	1	2	3	4	5	6	7
Quartz	13.9	22.5	25.0	5.9	26.6	25.2	5.6
Microcline	27.4	67.0	61.0	51.9	70.5	(1)	------
Microperthite	------	------	------	------	------	[1] 43.2	62.8
Plagioclase	44.1	4.5	7.1	36.6	------	22.3	19.4
Myrmekite	------	Tr.	Tr.	------	.6	------	------
Biotite	9.3	Tr.	6.4	4.6	1.2	.1	10.2
Hornblende	1.8	5.0	.4	------	------	8.6	------
Magnetite	1.9	.3	Tr.	.5	.6	.2	1.3
Sulfides	.6	------	------	------	------	------	.2
Apatite	.4	.4	Tr.	.1	------	.3	.4
Zircon	.1	Tr.	.1	.1	------	Tr.	Tr.
Sphene	Tr.	Tr.	------	------	------	------	------
Allanite	Tr.	Tr.	------	------	.1	------	------
Epidote	Tr.	------	------	------	------	------	------
Chlorite	} .5	{ ------	------	------	------	------	------
Carbonates	}	{ .3	------	------	------	------	------
Limonite	------	------	Tr.	.3	.4	.1	------
	100.0	100.0	100.0	100.0	100.0	100.0	99.9
Composition of plagioclase.	An_{23}; An_{27-37}.	------	------	Oligoclase.	------	An_{11-15}; rims An_4	An_{10}

[1] Includes a little microcline.

No. in table	*Specimen No.*	*Description*
1	---------	Biotite gneiss; average of two specimens of nearly identical composition (T12.4, hole T12, depth 207 ft; and T12.8, hole T12, depth 286 ft).
2	T12.23	Hornblende-microcline granite gneiss, hole T12, depth 696 ft.
3	T11.15	Biotite-microcline granite gneiss, hole T11, depth 377 ft.
4	H–133	Biotite-microeline syenite, 350 ft south of collar of hole T5.
5	T12.19	Aplite, hole T12, depth 632 ft.
6	---------	Hornblende-microperthite granite; average of two specimens of very similar composition (R–5–T and R–8–T) between 8000 N and 8600 N, east edge of mapped area.
7	T11.19	Biotite-microperthite syenite, hole T11, depth 542 ft.

DISTINCTION BETWEEN "MICROCLINE GRANITE GNEISS AND SYENITE GNEISS" AND "PYROXENE, HORNBLENDE, AND BIOTITE GNEISSES"

Two classes of rocks—"microcline granite gneiss and syenite gneiss" versus "pyroxene, hornblende, and biotite gneisses"—have been delimited and described above. In general, the two broad types were recognized in logging by differences in color, texture, or proportion of mafic minerals. However, there arises the problem of what to do with certain rocks whose mineral composition is alike ($\sim$10 percent quartz, high potassic feldspar, a little plagioclase, and 5–10 percent mafic minerals) but whose fabric is different. The undisputed metasedimentary gneisses of the Trembley Mountain area have a slightly inequigranular texture and a foliated, crystalloblastic structure in which the felsic minerals form a mosaic of polygonal grains with straight or gently curved sides. In contrast, the granite gneiss and syenite gneiss show a markedly inequigranular texture and a foliated, granoblastic structure in which the felsic minerals are irregular in shape, often elongate (with larger grains partly surrounded or "terminated" by aggregates of smaller ones), and nonpolygonal. The type of fabric, therefore, has been used to separate the doubtful cases for description.

MICROPERTHITE GRANITE AND SYENITE; SODIC SYENITE

Hornblende-microperthite granite, locally syenitic, encloses the metasedimentary rocks and microcline granite gneiss of the Trembley Mountain area. (See pl. 1.) The general characteristics of this hornblende-microperthite granite are described in the regional study (Prof. Paper 376). The only specimens studied from this area come from a northeast-trending belt of outcrops that begins at (8000N, 24250E). (See pl. 19.) The granite, which contains sporadic slabs of biotite gneiss, has a remarkable variation in quartz content in a small area.

Biotite-microperthite syenite forms a layer of undetermined thickness (possibly 30 ft or more) in the footwall cut by hole T11. The syenite carries layers of laminated pyroxene-plagioclase gneiss and in several places appears to have developed a massive, pyroxenic, sodic facies within the gneiss.

Hornblende-microperthite granite and syenite.—Megascopically, the hornblende-microperthite granite and syenite are medium to fine grained, granulated, pinkish, yellowish, or grayish, with a foliation varying from weak to strong. A few medium-sized or coarse potassic feldspar grains stand out in relief on the weathered surface. Microscopically, the same inequigranular, granoblastic fabric is evident, with foliation due to subparallel hornblende aggregates, elongation of quartz and microperthite grains, or prominent differences in grain size. The microperthite granite and syenite from this limited area is somewhat more deformed than the bulk of the microperthite granite that flanks the Clare-Clifton-Colton belt.

Quartz, microperthite, plagioclase, and hornblende are essential minerals, accompanied by accessory magnetite, apatite, and zircon. A little biotite, in part derived from hornblende, is sometimes present, and sphene is found in a few specimens. A mode is given in table 32, No. 6.

The base of the microperthite grains, both large and small, is "shadowy," and even the smaller grains have a slight microperthitic intergrowth. Plagioclase grains are usually "zoned," with a core of An_{15} and a very thin rim of An_{4-10}. One specimen with unzoned plagioclase had a few larger grains of composition An_{15} and many smaller grains of An_{11}.

The hornblende is similar to that found throughout large masses of the hornblende-microperthite granite. It is presumably a hastingsite.

Biotite-microperthite syenite.—Biotite-microperthite syenite, a homogeneous pink rock lacking visible foliation, is seen under the microscope to be fine, highly inequigranular, and granoblastic, with a foliation given by subparallel biotite flakes. Quartz, microperthite, plagioclase (An_{10}), and biotite are essential; and magnetite, pyrite, apatite, and zircon accessory. The mode is given in table 32, No. 7. Except for a few flakes of slightly greenish-brown biotite, the dark mica is pleochroic from pale-yellow to golden brown.

Pyroxene-plagioclase syenite (*sodic syenite*).—The pyroxene-plagioclase syenite (sodic syenite), present as layers grading into the biotite syenite just described, is thought to be pyroxene-plagioclase gneiss highly modified by action of the magma that also yielded the enclosing syenite. Hand specimens of the pyroxene-plagioclase syenite are pink, medium grained, and nearly massive, with subhedral pyroxene crystals scattered about in random orientation. The texture is much coarser than that of either the associated pyroxene gneisses or the biotite-microperthite syenite, and the laminated structure of the typical pyroxene gneiss is absent.

Microscopically, the pyroxene-plagioclase syenite is highly inequigranular, the texture varying from fine to coarse. The finer stuff occurs as polygonal grains, whereas the larger grains are sutured. These relations suggest reworking of an earlier crystalloblastic fabric.

Essential minerals are plagioclase, microcline (principally with the fine material), and 10 to 15 percent pyroxene with subordinate amphibole. A few large grains of microperthite are also present. Quartz, magnetite, apatite, zircon, very little sphene, and a trace of pyrite and chloritized biotite are accessory. Plagioclase (An_{15}) is the predominant component. Twinning lamellae in some of the larger grains are slightly bent. The pyroxene is a very pale green, iron-poor variety, partly altered to chlorite, amphibole, calcite, and specks of magnetite(?). The amphibole is rather light green and poikiloblastic. Slight alteration of some feldspar to calcite and epidote is visible locally. The instability of pyroxene and the formation of poikiloblastic hornblende support the mega- and microscopic evidence from the fabric that this is a modified rock.

QUARTZ ROCK

The presence of quartz rock is a constant and characteristic feature of the zone of modified skarn in the southern area of mineralization. If the individual layers are added together, one finds that quartz rock makes up 4 to 26 percent of the zone. Its proportion is roughly consistent (12–15 percent) in holes 1, 2, and 11 but highly variable in holes 3, 12, and 13 [4(?), 26, and 4 percent, respectively]. Hole T4, which cut part of the zone, showed 4 percent of quartz rock. Holes T5 and T6, on the northern area of mineralization, showed none.

At present, it seems impossible to correlate the quartz rock as "beds," though one zone about 20 feet below the "upper" garnet-biotite gneiss was cut in several holes.

The quartz rock forms layers a few inches to several feet thick. The thickest, cored in hole T12, is 27 feet. 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. In places, its contact with other rocks is sharp. Elsewhere, it grades on the one hand into "quartz-feldspar rock" and "feldspar rock" (greenish "syenite"), and on the other into pyroxene skarn that has patches, wisps, or thin layers of felsics. Locally, it is associated with thin zones of garnet-biotite gneiss present in the modified skarn.

Only one outcrop (at 7920N, 22840E) of quartz rock has been found in the area. It is fine- to medium-grained, pitted, glassy quartzite faced on the east side with 6 inches of subore.

Megascopically, the quartz rock is gray (locally white), medium grained, and always glassy rather than milky. Much of the medium grained variety has mafic aggregates that give it a rude planar structure. In a few places, particularly where it is associated with garnet-biotite gneiss, the quartz rock is fine grained and laminated. Such portions are interpreted as metasedimentary quartzite. A third variety of the quartz rock is coarse grained and generally contains large, scattered crystals or aggregates of pyroxene. The pyroxene—now fresh, now chloritized—increases in quantity toward the enclosing pyroxene skarn, which often shows a coarsened selvage against the quartz. This pattern is quite the same as that shown by many contacts between skarn and pegmatite. Rarely, the quartz (as well as the associated skarn) is vuggy. At one place, a late quartz vein was noted in a chloritized zone. Vugs in this vein were lined with euhedral quartz crystals.

Quartz rock almost always contains some extremely fine magnetite or pyrite, and it is this magnetite "dust" that accounts for the "smoky" aspect of much of the rock (as in hole T1). In addition, pyroxene or amphibole, or garnet and a little biotite, are commonly present. Less easy to detect is colorless or greenish potassic feldspar, locally present as grains or nests, especially where quartz rock grades into greenish "granite," "syenite," or highly modified skarn. Not only is the included feldspar similar in color to the quartz, it is also glassy and fails to show a prominent cleavage. The average quartz content is about 80 or 90 percent.

Regardless of their megascopic differences, all specimens of quartz rock show essentially the same fabric in thin section. All are markedly inequigranular and generally show a strong foliation. The large quartz grains are interlocking, slightly strained, and often elongated to form "leaves." Fine quartz, when present with the coarser variety, forms polygonal, equant grains. In general, the coarse quartz leaves are much larger than any of the mafic minerals, which tend to occur as trains passing through quartz and about parallel to its elongation. A noteworthy feature of certain specimens is the pronounced flattening (or elongation) of some grains of clinopyroxene, orthopyroxene, and magnetite. The fabric in some specimens looks completely recrystallized, except for late magnetite and sulfides. In others, quartz, a little garnet, sporadic amphibole and mica, magnetite, and sulfides appear to be "late," replacing the pyroxene. Even in these specimens, introduced quartz seems mostly to have recrystallized, and only subordinately replaced, pyroxene.

Two principal mineralogic types are recognized: pyroxenic, and garnetiferous.

Pyroxenic quartz rock.—This type contains clinopyroxene, commonly with considerable orthopyroxene. Amphibole and biotite or pale-green mica are generally present, sometimes with a little untwinned potassic feldspar. Magnetite is also an essential mineral.

The accessories are sulfides, apatite, and zircon. In addition, allanite or sphene is usually present.

Clinopyroxene (probably salite) forms pale-green equant or elongate grains. Locally, a few of these may be partly altered to blue amphibole.

Orthopyroxene has been observed only in quartz rock and quartz-bearing skarn. The equant, or frequently elongate, grains are readily recognized by their strong pink and green pleochroism. Some show partial alteration to chlorite. Exsolution lamellae have not been observed. The optical data (see table 2, No. 10) obtained from a typical specimen indicate that the mineral is an iron-rich hypersthene, possibly carrying much of the iron as Fe^{+3}. Hietanen (1938) has described hypersthene-grünerite quartzites from Vittinki and grünerite-magnetite quartzites from Laihia, Finland. The rocks are interpreted as metasedimentary with grünerite—at higher temperature, hypersthene—formed by reaction between SiO_2 and $FeCO_3$ (Hietanen, 1938, p. 105). At Vittinki, iron-manganese ores and pyrite ores occur in quartzite (Saxén, 1925, p. 25) associated with biotite-quartz-plagioclase gneiss, leptite, and amphibolite (Hietanen, 1938, p. 47, 49). At Laihia, local accumulations of magnetite and pyrrhotite are present in the quartzite, which is associated with amphibolite and migmatites (Hietanen, 1938, p. 60–61).

Primary amphibole (pale bluish-green) may or may not be poikiloblastic. In one specimen, it forms partial rims between quartz and magnetite, or pyroxene and magnetite.

Brown biotite occurs as scattered flakes in two specimens that are probably metaquartzites. In certain other specimens, pale greenish-yellow mica occurs as scattered flakes, or as blades that locally replace amphibole or (rarely) magnetite. Occasionally, the pale mica is slightly replaced by magnetite.

Two varieties of magnetite are commonly present: very small blebs and euhedral crystals disseminated through quartz; and large flattened or amoeboid grains that are molded on quartz, or slightly replace quartz and dark silicates. The larger areas of magnetite transect the foliation somewhat. Locally, large magnetite grains have partial rims of amphibole. Rarely, the magnetite has small inclusions of green spinel(?).

Garnetiferous quartz rock.—This type is composed of abundant quartz, a little potassic feldspar and plagioclase, and conspicuous garnet, biotite, and magnetite. A little amphibole is sometimes present, and one specimen shows a corroded xenocryst of pyroxene. Sulfides, apatite, zircon, monazite(?), and allanite have been noted as accessories, but their distribution is variable. A little chlorite-carbonate alteration is common.

The potassic feldspar is untwinned and unaltered, with $(-)2V \approx 45°$ or less and $r > v$, perceptible. The plagioclase, which occurs as grains or (rarely) as partial rims on magnetite, is sericitized, carbonatized, or altered to a radiating mass of pale-green chlorite and muscovite. It seems at least as calcic as andesine.

The poikiloblastic garnet is pink or yellowish pink and includes any or all the other minerals in the quartz rock. Its composition has not been determined. It may be spessartitic but more likely is almanditic.

The biotite occurs as single flakes and as clumps or books. In addition to primary biotite, a little secondary biotite (locally accompanied by fine green chlorite, calcite patches, and magnetite specks) has formed from some amphibole.

The amphibole, where fresh, is pleochroic from straw yellow to light greenish blue. Some of it is poikiloblastic.

Magnetite is usually present to the extent of several percent, though locally it may reach 20 to 25 percent. The bulk of the magnetite forms grains and amoeboid aggregates; but occasionally tiny blebs and euhedral crystals, or elongate—almost platy—grains, are present. Small inclusions of green spinel(?) and chlorite are rare.

In general, garnet, most magnetite, and sulfides are late. Some quartz, at least, was apparently introduced, as it replaces hornblende, biotite, and plagioclase slightly. Some plagioclase (rims) has been either introduced or reworked.

The origin of the Trembley Mountain quartz rock is discussed in the section of this report dealing with the petrology of the ore deposits, pages 49–50.

SKARN AND MODIFIED SKARN

In this report on the Trembley Mountain area, the term "skarn" is used for rocks in which 90 percent or more of the nonmetallic minerals are Ca-Mg-Fe silicates.[45] Pyroxene is the characteristic silicate, usually the principal one. The term "modified skarn" has been used for similar rocks in which felsic minerals (quartz, feldspar, or scapolite) constitute perhaps 10 to 60 percent of the nonmetallic minerals, the rest being chiefly pyroxene, amphibole, or both. Skarn, modified skarn, and directly associated quartz rock, granite gneiss, syenite gneiss, pegmatite, aplite, and metasedimentary gneisses are referred to en masse as the "zone of modified skarn."

Table 33 shows the volumes of skarn, modified skarn, and quartz rock relative to the rest of the rocks in the zone of modified skarn. Where a nearly complete section of the zone was cut by the drill, or can be inferred, the sum of skarn plus modified skarn constitutes about 40 to 65 percent of the total thickness.

Reference to the graphic sections (pl. 19) will show, roughly, what the rest of the zone is.

TABLE 33.—*Percentage distribution of rock types in zone of modified skarn, Trembley Mountain deposit*

Hole	T1	T2	T3	T13	T12	T11	T4
Modified skarn	43	17	57	46	17	20	8
Skarn	6	8	9	8	24	18	26
Quartz rock	12	12	4	4	26	12	4
Rest	39	63	30	42	33	50	62
	100	100	100	100	100	100	100

Skarn

Two types of skarn are present at Trembley Mountain: light-green diopsidic or salitic; and medium- or dark-green salitic or ferrosalitic. The darker type is more abundant by far. Both are homogeneous and fine or medium grained. Megascopically, most of the skarn appears to consist only of pyroxene, locally with sulfides or magnetite. Infrequently, one sees a little black amphibole in the dark skarn. Neither planar nor linear structure is detectable in most hand specimens. The whole outward appearance of the skarn is deceptively simple, even to the trained eye.

Light-green skarn.—Skarn that is light green occurs as layers 5 to 10 feet thick in a sequence of pyroxene-quartz-feldspar gneisses below the mineralized zone in hole T4, as thicker layers with quartz rock hanging wall and granite gneiss footwall in hole T11, and as interlayers within a thick zone of dark skarn in hole T13. It has been noted as layers 1 to 3 feet thick within the dark skarn cut by other drill holes. The light-green skarn is never the host of ore, but it may occur just beneath sparsely mineralized zones of salite skarn. A high proportion of light-green skarn occurs only where there is a thick zone of unmodified (or only slightly modified) dark skarn. Apparently, thin layers of skarn in granitic rocks were readily changed over to the darker, salitic type.

Typical light-colored "monomineralic" skarn is really a fine-grained, nearly equigranular, polygonal-crystalloblastic aggregate of pyroxene and amphibole, roughly equal in quantity, with subordinate altered(?) carbonate and a little quartz, apatite, magnetite, and sulfides. A faint planar structure is given by the alinement of some amphibole crystals.

The pyroxene is a diopside (almost a salite) with $nX = 1.679 \pm 0.001$. The nearly equant grains are pale-green in thin section. A few of them are twinned, and rare grains are partly altered to fibrous chlorite (actinolite?).

The amphibole is actinolite with about 18 mole percent ferrotremolite. (For its optical properties, see table 2, No. 12.) Crystals are slightly elongate (ratio$\sim$2:1) and distinctly pleochroic, with a somewhat patchy color distribution. Because the actinolite crystals are relatively stubby and have the same color as pyroxene in hand specimens, they were not identified when the core was logged.

Equant grains of altered(?) carbonate are scattered through the rock. They consist of a carbonate base with parallel or radiating fibers of yellow-green chlorite (possibly actinolite). One or two grains have relics(?) of a colorless unknown. The carbonate may represent residual calcite replaced by late chlorite(?), or pseudomorphs after an unknown mineral.

Quartz occurs as small, scattered, wormy grains making up less than 1 percent of the rock. A few elongate, anhedral grains of apatite are also present. In addition, there are a few grains of magnetite, hematite (?—altering to limonite; present in the calcite), and pyrite, as well as a trace of pyrrhotite and zircon(?).

[45] Masses of pyroxene or amphibole within pegmatite veins have usually been called "recrystallized pyroxene (or amphibole) aggregates," rather than skarn.

Dark-green skarn.—A darker skarn, either "pure" or very slightly modified, occurs as layers in modified skarn, in quartz rock, between modified skarn and quartz rock or granite gneiss, or between granite gneiss and quartz rock—all within the zone of modified skarn. Thin layers of pyroxene skarn are present sporadically in the granitic wallrocks, and one layer of pyroxene skarn, garnet-pyroxene skarn, and feldspar-veined skarn occurs in the footwall garnet-biotite gneiss of hole T12. In places, dark skarn is slightly mineralized or serves as the host of ore. Zones with disseminated or concentrated sulfides (chiefly pyrrhotite, with some pyrite and a little chalcopyrite) are common.

Typically, the rock is a medium-grained, roughly equigranular, polygonal - crystalloblastic aggregate without visible planar structure. Locally, the rock is inequigranular, fine and medium, or medium and coarse grained. Green, pleochroic pyroxene (salite or ferrosalite) makes up about 88 to 95 percent of typical skarn. Quartz ($\sim$ 2 percent) and magnetite (a trace to 3 percent) are always present, the former as scattered blebs or as small areas replacing pyroxene and often rimming magnetite. A little greenish-yellow mica or secondary green hornblende occurs sporadically, as do sulfides, apatite, and a trace of sphene. One specimen contains about 5 percent of apatite as clustered grains locally surrounded by quartz. This is a high concentration of apatite for any skarn of the St. Lawrence County magnetite district. Allanite has not been observed in the unmodified skarn, either light or dark.

Modified skarn

By a transition that is usually gradual, the 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.

Almost all the modified skarn is composed of dark-green pyribole and white, gray, or pale-green felsic minerals. Light-green skarn may show threadlike, 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 felsic minerals. Wispy and patchy areas of light minerals appear. Occasionally, the felsic minerals 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 felsic minerals spread out between adjacent layers, transecting, engulfing, and displacing the intervening mafic constituents. 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 "granite" are interspersed throughout the modified skarn. Their contamination effects have been described elsewhere. 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. Wherever the two were easily distinguishable, their relations toward the skarn were seen to be identical. In addition, it was clear that the quartz: feldspar ratio in the masses, thin layers, patches, and knots of felsic minerals was highly variable.

The writer interprets the megascopic evidence as indicating extensive 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. That evidence will be noted below.

There are incidental thin layers of skarn, and a few of laminated pyroxene gneiss, within the wallrocks; but the blotchy, heterogeneous, or strongly migmatitic material that characterizes modified skarn is restricted to the zone that bears its name.[46]

The rapid variation from skarn to modified skarn to granitic rocks or quartz rock seems chaotic, and the writer has been able to correlate only one unit of skarn from hole to hole. This unit, about 20 feet thick, appears in three holes (T2, T11, and T12) about 40 feet above the footwall gneisses.

The microscopic fabric of modified skarn is variable. Commonly, the rock has an inequigranular or a 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. Potassic feldspar, when present, may show a similar habit. The quartz (and feldspar?) show strain shadows. Both types of microfabric—polygonal-crystalloblastic and "replacement" leaves—are shown by rocks that look identical in hand specimen and show felsic minerals replacing skarn.

Estimates of the mineral composition of representative specimens of modified skarn are given in table 4, Nos. 2, 3, and 4. One may consider that the "end mem-

[46] A trace of skarn partly replaced by quartz is present with a little quartz rock in the footwall of hole T12. However, a correlation based on a lower part of the garnet-biotite gneiss would have placed the "anomalous" material in the zone of modified skarn.

bers" have either quartz or feldspar as the sole felsic mineral, while intermediate members have varying proportions of the two. No attempt is made in table 4 to give maximum and minimum values for the felsic minerals, as the gradational nature of modified skarn has already been emphasized. Scapolitic skarn, not studied in detail, is certainly less abundant than the quartz-feldspar type but not scarce.

Essential minerals of modified skarn are clinopyroxene and a felsic mineral. Amphibole is an 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 (cf. mineralogy of the quartz rock). The feldspar, where present, is the untwinned potassic variety, accompanied by a very little plagioclase. Optical properties of a representative scapolite are given in table 2, No. 17.

The 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, are chlorite, carbonate, sericite, zoisite, and pumpellyite.

The pyroxene is a pleochroic salite or ferrosalite. Preliminary work suggests that pyroxenes in the modified skarn are essentially the same as those in unmodified dark skarn and have about the same range of composition; nX, determined on two typical specimens, was 1.692 and 1.703, respectively. Locally, the pyroxene is partly altered to chlorite, or to carbonate and pale green amphibole.

The primary amphibole is generally blue green, though in a few specimens it is brownish green. The intensity of pleochroism of the blue-green variety, optic axial angle, and extinction angle $Z \wedge c$ vary somewhat, reflecting changes in chemical composition. The amphibole may be termed a hornblende; it is probably a hastingsite.

The potassic feldspar is exclusively the clear, untwinned variety with $(-)2V=40°-55°$, $r>v$ perceptible. Cleavage may be distinct, but more often it is imperceptible. In some grains, a few curved or radiating cracks are seen. Rarely, the feldspar is very slightly perthitic, the albite inclusions occurring as pinpricks dotted through the host. In one specimen, albite occurs as ragged replacements at the borders of potassic feldspar. In another, the potassic feldspar is partly replaced by quartz. Occasionally, some or all the grains in a specimen show wavy extinction.

A little plagioclase (normally less than 5 percent) is present in most of the modified skarn. Fresh material is seldom observed; in rocks otherwise unaltered, the plagioclase is almost completely converted to sericite and zoisite. The unaltered remnants appear to be at least as calcic as oliogoclase-andesine. A distinctive feature of the plagioclase is its occurrence as partial rims between magnetite and pyroxene, or less frequently between magnetite and quartz (cf. mineralogy of the garnetiferous quartz rock). In addition, isolated grains of plagioclase or myrmekite may be present.

Brown mica is present in trifling quantity in much of the modified skarn. Some of it is secondary after pyroxene or amphibole. More often, however, brown mica occurs as partial rims or fringes on magnetite. In this relation, it is partly older and partly younger than the magnetite.

General comments on allanite are found in a separate section (p. 219). The problem of the metallic minerals in modified skarn is discussed in the section on ore minerals.

A type of alteration seen in about half the specimens is replacement of potassic feldspar by pumpellyite, a strongly pleochroic mineral of the epidote(?) group. (Partial optical data for the pumpellyite are given in table 2, No. 9.) This mineral, which has high relief, occurs as isolated prisms, patchy areas (sometimes with calcite, sericite, or an unidentified weakly birefringent mineral of low relief), or fibrous, radiating aggregates. The alteration is usually related to cracks in the feldspar, to boundaries between quartz and feldspar, or less frequently to feldspar cleavage. In places, however, the pumpellyite appears at the center of unbroken feldspar grains. Plagioclase or mafic minerals adjacent to the pumpellyite rarely seem to be affected by the alteration, which varies in extent from a trace to 5 or 10 percent. Where alteration of potassic feldspar to pumpellyite plus calcite is most extensive, a good part of the pyroxene has been replaced by green amphibole, calcite, and minor chlorite.

In some specimens of quartz-bearing modified skarn, magnetite grains have discontinuous fringes of pumpellyite. These fringes project out into quartz from the magnetite-quartz borders, thereby showing a relation similar to that of pyrite rims on magnetite. This and other features suggest that pumpellyite and late (stage II) pyrite are genetically related.

Another type of modified skarn carries the magnetite at Trembley Mountain North. Here pyroxene skarn showing slight local replacement by quartz and untwinned potassic feldspar has been extensively replaced by barite. Locally, the barite is accompanied by fluorite.[47] The baritic modified skarn has been replaced by magnetite and subordinate amounts of pyrite, pyrrhotite, chalcopyrite, and sphalerite. Barite is also present in mineralized biotite-quartz schist from the test pit at (9065N, 21920E). (For the optical properties of this barite, see table 2, No. 1.)

VUGS

An impressive but local feature of the modified skarn zone is the presence of vugs lined with colorless chabazite cubes, quartz crystals, and pyrite crusts. Some chlorite is often present, too. The vugs are found in cores from several holes but are especially noteworthy in hole T13 at intervals from 328 to 523 feet. The vugs are independent of "stratigraphic" or topographic control. Rather, they are usually restricted to zones where several percent of sulfides are present. The vugs are of postmagnetite age and probably genetically related to pyrite and sphalerite of stage II. Whether or not the vugs were once filled with some soluble mineral, such as calcite, is unknown. Their occurrence at considerable depth suggest hydrothermal development rather than supergene alteration.

One thin section of modified skarn shows chabazite crystals of cubic habit projecting into a vug. In addition, small grains and patches of chabazite replace the silicates, and a stringer of chabazite cuts magnetite. Many of the chabazite grains are completely rimmed with pyrite of stage II. The chabazite is optically anomalous, showing both positive and negative optic sign, weak but variable birefringence, crude twinning, and local zonal growth. (See table 2, No. 19.) Chabazite is often accompanied by rounded aggregates of fine green chlorite that replaces feldspar and some pyroxene. Locally, chabazite is slightly replaced by pumpellyite.

CORRELATION OF ROCK UNITS, TREMBLEY MOUNTAIN AND JARVIS BRIDGE

Any attempt to correlate rock units at Trembley Mountain with those at Jarvis Bridge is extremely hazardous, owing mainly to the fact that the rock sequence at Trembley Mountain is not well defined. It is conceivable that the garnet-biotite quartzite within the modified skarn zone at Trembley Mountain (see pl. 19, holes T1, T12, and T13) is equivalent to the lower marker at Jarvis Bridge (see p. 171 and pl. 14). The garnet-biotite quartzite at Trembley Mountain is only 4 to 13 feet thick—much thinner than the lower marker at Jarvis Bridge usually is. However, this correlation seems to be the only reasonable one, if indeed the rock sequences are at all equivalent in detail. On the basis of this correlation, half the zone of modified skarn at Trembley Mountain would be equivalent to the pyroxene-biotite gneisses between the lower marker and the top of the thick "footwall" garnet-biotite gneiss at Jarvis Bridge. (The possibility that much of the pyroxene-biotite gneiss at Jarvis Bridge represents migmatized skarn was considered on p. 171.) It follows from this correlation that the ore zones at these two properties are not strictly equivalent, stratigraphically. At Trembley Mountain, the ore is in the modified skarn itself; at Jarvis Bridge, the ore is in part of a heterogeneous granite gneiss sheet within the zone of pyroxene-biotite gneiss (migmatized skarn?).

STRUCTURE

The difficulties of mapping and interpreting the structure of the area have already been mentioned, but they need to be reemphasized. A layer of glacial debris of variable thickness covers much of the area, including the parts underlain by ore bodies and the localities where structural data are critical.

In contrast to most of the deposits studied, the Trembley Mountain deposits lack a well-defined and easily recognized succession of rock units that can be used in working out the structure. Thus an initially complex structural puzzle is made more difficult to solve because some of the pieces are missing or especially hard to recognize. Fortunately, zones of biotite and garnet can be used to some extent in correlating the drilling results and working out the structure. Careful search for these "indicators" should be made when additional drilling is undertaken.

The two belts of magnetic anomalies shown on the map (pl. 19) are underlain by metasedimentary rocks and magnetite deposits. A sheet of contaminated microcline syenite gneiss (and granite gneiss?), estimated to be 150 to 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 metasedimentary rocks and "granite gneiss" sheets is flanked by hornblende-microperthite granite, locally alaskitic. (See pl. 1.)

The rocks of the northern belt consist of a 40- to 50-foot zone of modified pyroxene skarn, silicated marble, and pyroxene skarn underlain by biotite gneiss. Interleaved sheets of contaminated microcline granite gneiss and syenite gneiss are also present; these resistant rocks yield most of the outcrops of the area. Subordinate amphibolite occurs as schlieren or thin layers in these exposures. Ore is restricted to a 15- to 20-foot layer of modified skarn occurring at the top

[47] In hand specimen, the rock was mistaken for silicated marble, owing to its softness and effervescence with dilute HCl. Actually, the effervescence is due to secondary calcite. This calcite, accompanied by chlorite, is a local alteration product of the pyroxene.

of the zone of calcareous rocks, just beneath biotite granite gneiss and granitized biotite gneiss.

The 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 is locally abundant. Layers of laminated pyroxene gneiss occur sporadically, and very thin biotitic zones are also present. Contaminated granite gneiss sheets occur in variable frequency. Few drill holes cut the whole modified skarn zone and its wallrocks, but the combined data show the modified skarn zone to have a maximum thickness of approximately 300 feet, locally thinning to about 120 feet. The footwall is garnet-biotite gneiss and contaminated microcline granite gneiss; the hanging wall is biotite gneiss, granite gneiss, and granitized metasedimentary rock. The metasedimentary unit, as noted above, is enclosed by granite.

The relation of the northern and southern belts of metasedimentary rocks is discussed below.

Foliation in metasediments and sheets of granite and granite gneiss dips northeast, east, southeast, or south. There are two possible explanations for this dip toward the eastern quadrants: (a) eastward tilting of the metasedimentary layers and more or less conformable granite sheets, or (b) isoclinal folding of the complex with overturning toward the southwest.

Evidence favoring the second hypothesis is:

1. Curving of the magnetic anomalies, especially at the northern end of the two belts. This arcuation strongly suggests that the mineralized zone and its enclosing rocks are curved about the end of a fold.
2. Presence of a marked negative anomaly accompanying the strong positive anomaly at (8400N, 22800–23300E). At the west end of the anomaly, where the foliation dips 50°–65° E., the "dip slope" of the magnetic profile is east. The "dip slope" faces south, however, along the northern border of the same anomaly, suggesting that the foliation, in swinging from north-northeast to east, has changed from an east to a south dip. With considerably less assurance (because of the more complicated magnetic picture), one may reason that the negative and positive anomaly centered at (8250N, 23250E) reflects a north strike and east dip of the foliation there, thus giving the end of an isoclinal fold.
3. Curving of the foliation at the northeast end of the northern belt of anomalies.
4. Presence of a strong lineation (axes of minor crumples in the foliation, pegmatite braids, fluting, or dimensional orientation of elongate minerals) that becomes intense in a zone roughly following the company base line. The lineation, trending S. 25°–40°E. and plunging about 45° SE., approximates the keel of the inferred isoclinal fold. An intense lineation in the axial zone is characteristic of all the isoclinal folds studied in detail in the Adirondacks.
5. Consistency of drilling results from holes T12 and T13 with the hypothesis of isoclinal folding (see below).
6. Association of ore deposits with isoclinal folds is a distinctive feature throughout the St. Lawrence County magnetite district.

Evidence against the hypothesis is:

1. The writers feel they have demonstrated that 3 miles to the east the same belt of rocks is overturned toward the north, whereas the hypothesis considered above necessitates overturning toward the southwest. The existence of the opposed directions of overturning may not sound logical, but in an area of so great complexity it is not improbable.
2. Data that are really critical to prove the hypothesis of isoclinal folding have not been found. The areas where the foliation should strike northwest and dip vertically have no outcrops, and the interpretation of the "stratigraphic sequence" is subject to doubt.

If the demonstration of isoclinal folding at Trembley Mountain be granted, then the major fold is presumably an overturned syncline, rather than an anticline. Here again the "evidence" is disputable. If the southeast-plunging lineation is essentially parallel to the axis of the major fold, which closes on the north end, a syncline is indicated. The interpretation of the rock sequence in holes T12 and T13 is consistent with this view.

The map (pl. 19) shows the north and south belts of anomalies as belonging to two parallel zones of mineralization separated by microcline syenite gneiss and granite gneiss. An alternate interpretation would put them in a single belt separated by a pinched anticline of microcline "granite gneiss." Glacial debris covers the critical area where this hypothesis should be tested. The evidence from the rock sequence is inconclusive. In both tracts of metasedimentary rocks, a mineralized zone of pyroxenic rocks is enclosed by biotitic rocks (if sheets of granite gneiss and syenite gneiss are neglected). In the northern area, the pyroxenic zone is 40 to 50 feet thick. In the southern area, the maximum thickness appears to be about 300 feet. However, at hole T13 the unit may have been isoclinally folded twice, thus giving a fourfold thickness to a unit that may have been about 75 feet thick before deformation.

The area centering at (8750N, 21500E) may represent an overturned anticline with granite gneiss core, closing on the southwest and plunging generally south. Such an anticline would be complementary to the mineralized syncline. However, the scarcity of outcrops and other supporting evidence makes such an interpretation a tenuous one.

The cross section through holes T12 and T13 (pl. 19) illustrates the hypothesis of an isoclinal syncline. Hole T12 cut a true thickness of 120 feet of modified skarn, quartz rock, pyroxene gneiss, and associated granitic rocks. The principal mineralization is confined to the uppermost 37 feet. Hole T13 cut a true thickness of about 300 feet of pyroxenic rocks (assuming that the first fragments of cored rock are hanging wall), with mineralization throughout, though principally in the upper and middle portions. Both holes entered the well-defined footwall—migmatitic garnet-biotite gneiss—at approximately the same elevation, yet the foliation angle in the drill cores showed a rather uniform dip of 40°–45° NE. A high-angle fault between the two holes would account for the appearance of the inclined footwall at the same elevation but would not, of itself, explain the great difference in thickness of the pyroxenic zone. However, one might postulate a primary difference in thickness, or thickening updip due to plastic deformation before faulting.

The alternate interpretation of isoclinal folding gives a consistent explanation of the structure, not only for these two holes but for the entire block south of latitude 7750N. The footwall garnet-biotite gneiss is well exposed in holes T1, T12, and T13, and a remnant of it may be present at 110 feet (drilled distance) in T2. Hole T3 was not drilled deep enough to hit the footwall, though it probably penetrated the economically important part of the mineralized zone. The hanging-wall contact was cut only in hole T12 and presumably in T13. A thin but recognizable layer of garnet-biotite quartzite or garnet-biotite gneiss was cut just below one of the principal ore zones (generally the upper one) and 90 to 145 feet stratigraphically above the footwall garnet-biotite gneiss. It was not noted in hole T3, probably because that drill core was not logged in detail. The known relations among the two garnet zones, the ore, and the hanging wall have been used to project the limits of the mineralized zone at holes T1, T2, and T3.

The resulting isoclinal syncline shows a minor anticlinal roll between holes T12 and T13. Consideration of the projected trace of the hanging wall (inferred) at hole T3 and its relation to the anomaly suggests a change in strike 50 feet south of the hole. A minor fold so drawn projects downward at 45° SE. (the general plunge of the lineation) to hit nicely the anticlinal roll shown between holes T12 and T13. Extension of the isoclinal syncline northward to the section T1–T2 seems reasonable.

Structure in the neighborhood of holes T4 and T11 is more difficult to interpret in detail. The core-foliation angle in T4 is such that the rocks may dip either 35° SE. or 85° SE. Because of the 25° dip observed in an outcrop 150 feet north-northwest of the collar of T4, the gentler dip is shown on the cross section (pl. 19). Neither hanging wall nor typical footwall was cut by this hole, and the zone of magnetite-bearing rock is probably not of commercial grade. Moreover, the zone that is chiefly modified skarn in the block south of latitude 7750N is relatively pure skarn with associated laminated pyroxene-feldspar gneisses. The possibility that the foliation strikes northwest and dips southwest or is vertical, is suggested by the elongation of the high positive anomaly at (8000N, 23100E). It is not borne out by the core-foliation angle and succession of rock units found in hole T4. However, one is at a loss to explain how the zone of subore has given so strong a magnetic anomaly. Probably the virtual absence of overburden accounts for the high "kick" from the magnetite.

At hole T11 there is a similar ambiguity in interpreting the core-foliation angle. The dip must be steeply southeast in the lower part of the hole, for the foliation dips 63° SE. in the outcrop at (7920N, 22840E). Yet within 200 feet northeast and southwest of the collar, the rocks dip 35°–45° SE.

The writer has assumed that the dip is steep throughout hole T11, that three very thin zones of garnetiferous gneiss within the modified skarn are essentially the same unit, and that the structure can be represented schematically as a pair of isoclinal synclines whose axial planes have an apparent dip of 60°–80° SE. (See pl. 19.) This interpretation seems reasonably consistent with other structural data from the Trembley Mountain area. However, at least two other possible interpretations of the observed core-foliation angles should be pointed out.

1. The dip may be alternately steep and gentle in the upper part of hole T11 (to a depth of 250 feet), and steep in the lower part. This could indicate two or more synclinal folds, the axial plane of the upper syncline having an apparent dip of about 55° SE. and the axial plane of the lower syncline having an apparent dip of about 65° SE.
2. Either a steep dip or a gentle dip may prevail throughout hole T11; the resulting structure may

be part of a large roll or series of rolls between holes T4 and T1.

As the rock sequence is presumably isoclinally folded, another interpretation would be change in dip on a drag fold.

Another structural problem is exposed when the ore zone cut by holes T5 and T6 is projected to the surface. The dip indicated by the core-foliation angle is either 30° SE. or almost vertical. Because of the 30° dip of the foliation in outcrops east and west of the drill holes, the gentler dip is shown on the cross sections (pl. 19). Foliation in outcrops along the north side of the anomaly dips 40°–60° SE., and the inference from drilling data is that this dip flattens with depth. When the ore zone is projected to the surface according to these assumptions, its intersection lies farther north along the horizontal projection of hole T5 than along T6, so that the join of corresponding points appears to transect the strike of the foliation. One is strongly tempted to explain the apparent offset by inferring a small drag fold between the projections of T5 and T6. However, if the trend and plunge of the minor fold should correspond to those of the strong lineation present in outcrops just north of the anomaly, hole T5 would have intersected the downdip extension of the fold; that is, T5 would have cut the ore zone twice. Such was not the case. The difficulty can be obviated by assuming that the drag fold trends farther southeast, or plunges much more steeply, than the observed lineation north of the anomaly, but this is a case of special pleading. Similar objection may be raised against postulating a fault.

The small test pit in the projection of hole T5 was opened on the crest of the magnetic anomaly before any drilling was done. It showed disseminated magnetite and pyrite in granular pyroxene-hornblende-quartz-feldspar rock associated with thin layers of biotite schist. The updip projection of a 2-foot layer of mineralized gneiss present in the hanging-wall granite intersects this test pit nicely. Alternatively, the mineralized rock in the pit can be interpreted as part of the main ore zone, with which it could be connected by a sharp roll in the foliation. However, the 21-foot ore layer occurs in baritic pyroxene skarn, whereas the rock of the test pit is mineralized gneiss of the type found locally in the hanging-wall granite gneiss.

Joints and slip-surfaces are minor structural elements that appear here and there in the drill cores. Chlorite, often with a little calcite or pyrite, coats these surfaces, which cut both ore and country rock. The amount of displacement along slip-surfaces is unknown but probably very small.

ORE

The following opaque minerals occur in the Trembley Mountain deposit: magnetite, pyrite, pyrrhotite, chalcopyrite, sphalerite, molybdenite, ilvaite, ilmenite, and hematite. The last three were seen in polished sections only. The sulfide group averages perhaps 2 to 3 percent or less, and ilmenite averages perhaps 0.5 percent. Magnetite is the only mineral of commercial interest at this time.

Traces of the metallic minerals are found in almost all rocks of the area, but substantial concentrations are found only within the zone of modified skarn. The deposit is not a low-sulfide ore in the Swedish sense, but the sulfur content of typical ore is relatively low, ranging from 0.52 to 1.22 percent. Ore in baritic pyroxene skarn from hole T5 averaged 3.51 percent sulfur.

MAGNETITE

Occurrence

Magnetite mineralization has affected, in some place, 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 rocks as well, but here it may be syngenetic. The writer considers the extreme variation in the nature of the host rock to be indirect but strongly suggestive evidence that the Trembley Mountain ore deposit originated by replacement.

Not all rocks within the zone of modified skarn have been equally susceptible to mineralization by magnetite. The preference for skarn and modified skarn is strikingly shown in table 34, column 1. These two rock types contain almost 70 percent of the "mineralization," and almost 80 percent of the ore (col. 4). Modified skarn alone contains more than half of all the "mineralization" and has about 71 percent of the ore. Quartz rock contains relatively little ore but has a high proportion of subore. (See table 33 for proportions of major rock types in zone of modified skarn; and table 34, columns 2, 3, and 4, for proportions of ore, subore, and mineralized rock.)

The important role played by modified skarn in serving as host for ore is brought out in tables 34 and 35.

If one could neglect holes T1 and T2, the correlation between footage of ore and ratio of modified skarn to skarn would be very good. Obviously, one cannot neglect them but one may explain the discrepancy by the following facts (see pl. 19):

1. Hole T1 passed from overburden directly into ore. The inferred position of the hanging wall of the modified skarn is 50 to 60 feet "stratigraphically"

above the point where the drill entered bedrock.

2. Hole T2 entered bedrock at a point some distance "stratigraphically" above the bottom of hole T1.
3. In hole T2, the position of the footwall of the zone of modified skarn is uncertain.

TABLE 34.—*Percentage distribution of magnetite by rock type, Trembley Mountain deposit*

[Figures are based on drilled footage, not on true thickness. The error thereby introduced is slight, and the figures are probably correct to about 2 percent. Decimals are given only for convenience in addition]

Rock type	All material with >~5 percent magnetite[1]	Mineralized rock (~5–12 percent Fe in magnetite)	Subore (~12–20 percent Fe in magnetite)	Ore (>20 percent Fe in magnetite)
Skarn (dark)	7.4		28.0	7.7
Modified skarn	61.5	50.0	28.0	71.1
Quartz rock	3.7		29.3	1.8
Feldspar rock	1.2		14.0	
Quartz-feldspar rock	4.6			7.3
Pegmatite	.3			.2
Contaminated microcline granite gneiss and syenite gneiss	9.5	15.0		8.4
Garnet-biotite gneiss	<.1		.7	
Undivided (chiefly modified skarn and felsic rocks)	11.8	35.0		3.5
	100.0	100.0	100.0	100.0

[1] More than 98 percent of this comes from the zone of modified skarn.

Because of these gaps in our knowledge, it is impossible to say whether the ratios given in table 35 are completely misleading or reasonably reliable. If the missing sequences in the section T1–T2 were ore in modified skarn, the present ratios would be increased and satisfied. If the sequences were nearly barren skarn, a new, much lower, but consistent ratio would be established. We already know that some factor other than rock type controls the position of magnetite, but the preference of magnetite for modified skarn is strong.

Megascopic and microscopic features

Magnetite occurs as disseminated grains, small amoeboid blotches, coalescing blotches, irregular layers, and (rarely) streaks and homogeneous masses. The individual "grains" of disseminated magnetite are fine (<1 mm) or medium (1–5 mm) in size. In aggregates of magnetite, it is impossible to tell the size of the individual grains unless one examines a polished section that has been etched.

The individual layers of magnetite-bearing rock range in thickness from a few millimeters to many feet, but the thicker zones always contain admixed silicates.

Typical low-grade ore consists of disseminated grains, amoeboid blotches, and irregular layers of magnetite in silicates. 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. Lineation is seldom seen in the drill cores, but in one place the magnetite itself showed a lineation inherited from the elongate amphibole of the host rock.

TABLE 35.—*Comparison between quantity of ore and ratio of modified skarn to skarn for drill holes in southern area, Trembley Mountain deposit*

Hole	Ratio of modified skarn to skarn	Footage of ore (>20 percent magnetic Fe)
T3	6.7	Very high.
T13	5.8	Very high.
T11	1.1	Moderate.
T12	.7	Moderate.
T4	.3	Subore only.
T1	7.2	Low.
T2	2.1	Very low.

In all its aspects, the ore exhibits the features attributed to metasomatic replacement. The evidence is more easily seen in modified skarn than in skarn, because of the contrast in color among magnetite, pyribole, and felsics. For example, a rock having coarse aggregates of feldspar in the pyribole skarn showed some interstitial magnetite that had replaced both light and dark minerals. The magnetite had started at the crystal boundaries, worked part way around some of them, embayed certain grains, and left specks or small "islands" of feldspar in the larger magnetite masses. Most of the ore shows pyribole—especially pyroxene—replaced preferentially, with less replacement of the felsic minerals. However, as table 34 showed, layers of modified skarn are generally preferred to layers of "pure" skarn, and in places masses of felsic rocks have been extensively replaced.

Thin sections duplicate on a smaller scale the evidence seen in logging core. True, much of the magnetite in a given slide is molded on, or interstitial to, silicates; but the evidence for replacement is usually equally compelling. Embayment and isolation of single grains and patches of the host are common. The magnetite always conforms roughly to the foliation of the host rock but is not restricted to it: areas of magnetite cut across from one foliation plane of the rock to another. Grain boundaries and the preexisting planar structure have exercised primary control, with magnetite later enveloping the silicates without regard to foliation.

Pyroxene, amphibole, quartz, feldspar, and scapolite are replaced by magnetite, the first two preferentially. Any one or several of the chief rock-forming minerals may thus be included in the magnetite, occasionally with grains of the accessories apatite, zircon, and allanite. In addition, minute grains of pale-green isotropic spinel and ragged areas of distinctive brown or greenish, fine-grained chlorite(?) are often present as inclusions in magnetite. They have never been observed in the silicate groundmass. In a few places the spinel grains are alined as beads along a parting in the magnetite. They were probably exsolved as the magnetite crystallized. The origin of the brown chlorite(?) inclusions is unknown. Though they do not look it, they may be pseudomorphs without trace of their host. Or they may be late-stage replacements of magnetite, as their ragged borders suggest.

Where garnet and biotite are present, as in garnetiferous quartz rock, these silicates seem to have formed contemporaneously or after, as well as before, magnetite. Some platy and euhedral magnetite accompanies the replacement magnetite in garnetiferous and pyroxenic quartz rock.

The remarkable local development of silicate rims on magnetite has been described elsewhere.

Etching of polished sections shows that there is no very coarse magnetite and not much very fine magnetite. Except for scattered specks of magnetite, the smallest individuals are about 0.1 mm in diameter, the largest about 2 mm. Many are 0.3 to 0.5 mm in diameter. Most of the "grains" seen in hand specimens are really aggregates, several millimeters long, consisting of two or more individuals. The discrete particles of magnetite are often elliptical or anhedral. Boundaries between adjacent individuals in aggregate grains or masses are gently curved or subpolygonal. None of the etched individuals shows concentric growth.

The shape of the aggregates or little masses of magnetite varies from rudely lenticular and smooth walled to highly irregular and amoeboid. In the "amoebae," the magnetite wraps around some silicate grains, completely encloses others, sends off thin veinlets that end in a thicker "bead" or connect with other areas of magnetite, and shows small bootlike projections. In fact, the higher the grade of ore, the more evidence one sees of replacement, for the discrete magnetite grains are less abundant and the masses more so. The arrangement of these masses may be roughly parallel to the foliation of the host rock, or it may take the form of loose networks that cross the host rock at an angle to the foliation.

To the writer, the evidence that magnetite has grown by replacement, and not by recrystallization, is overwhelming.

Besides the inclusions of silicates, apatite, spinel grains, and ilmenite noted elsewhere, the magnetite locally exhibits parallel or rectangular arrangements of ultrafine "gashes." These have been interpreted by many writers as spinels formed by exsolution.

Magnetite in biotite-quartz schist

A unique occurrence for magnetite at Trembley Mountain is seen in the prospect pit at (9065N, 21920E). This small pit had been flooded before the writer examined it. Fragments of medium-grade "ore" beside the pite showed magnetite and sulfides in a biotite-rich schist dotted with prominent "eyes" of quartz and a soft white mineral easily scratched with a knife. The nearest outcrops are sparsely mineralized granite gneiss. The magnetic anomaly given by the "ore" indicates that the material has little commercial significance. Mineralogically and genetically, however, it is remarkably interesting.

The "ore" is shown by a thin section to consist of magnetite, quartz, barite, and biotite, with a little blue amphibole and pyrite, and a trace of pyroxene(?). The fabric is strongly foliated, and both quartz and barite have a preferred orientation that is readily observed when a gypsum plate is inserted into the microscope system. Large quartz leaves are accompanied by thin sheared zones of granulated quartz and amphibole, greatly elongate barite, and magnetite. Larger masses of ungranulated magnetite transect the veinlike, sheared quartz and contain inclusions of quartz, pyribole, and barite. In addition, filamentous and platy magnetite is present along zones of sheared quartz. Part of the granulated quartz and all the large quartz leaves have sutured boundaries, suggesting recrystallization after deformation of the quartz. Barite in zones of sheared quartz looks undeformed but essentially contemporaneous with quartz. It may have flowed and recrystallized readily. One reasonable interpretation of the rock's history would be: replacement of biotite schist by quartz and barite; granulation and some recrystallization; introduction of magnetite along the zones of granulation; introduction of pyrite as cubes and anhedra replacing quartz, barite, and biotite; slight movement; introduction of late pyrite as hair-thin veinlets.

The barite was identified optically (see table 2, No. 1) and checked by a flame test and X-ray spectrometer pattern. The barite gave the typical green flame of Ba with an initial flash of red. Whether the red indicated Sr or Ca could not be determined by the means available. Both Trembley Mountain barite and "optically normal" barite from Missouri were tested by X-ray spectrometer. The two graphs were essentially the same, the peaks being relatively displaced from those of "optically normal" celestite, which was also tested.

Barite, quartz, and magnetite are associated locally in the Parish and Benson Mines ore bodies; but the Trembley Mountain material appears to have replaced a biotite schist, rather than granite or granite gneiss. Moreover, barite is abundant in modified skarn at Trembley Mountain North. Its presence is another link between the granite gneiss ores and skarn ores, though the local introduction of abundant barite is a characteristic feature of modified ores of both types.

ILMENITE

Small quantities of ilmenite are present in much of the magnetite. The maximum is about 2.8 percent ilmenite, and the average perhaps 0.5 percent, in the ore. The ilmenite is always attached to, or included in, magnetite; it never occurs as discrete particles in the silicates. The ilmenite has the form of single grains, blades, and minute slivers. Both blades and grains may be present in a single area of magnetite, though they usually occur separately. The ilmenite shows distinct reflection-pleochroism, with the direction of minimum absorption the same shade of gray as the enclosing magnetite, and the direction of maximum absorption somewhat brownish. Anisotropism is strong, from light gray to brownish or purplish gray. Maximum absorption and the darker "polarization color" are observed when the ilmenite blade is oriented normal to the plane of polarization of the polarizer. A few grains show four extinction positions for a 360° rotation, instead of the usual two.

Single grains of ilmenite, about 0.1 to 0.2 mm square, are present in some specimens. Frequently this ilmenite is surrounded on three sides by magnetite, while the fourth touches a silicate. In places, elongate, curving grains of ilmenite serve as a bridge between adjacent areas of magnetite. Boundaries of the ilmenite grains are usually straight or gently curved. Locally, projecting tongues of ilmenite suggest that magnetite has been slightly replaced by ilmenite. Rarely, a few very thin blades of exsolution hematite occur in the ilmenite. One somewhat ragged ilmenite grain showed parallel blebs and "worms" of a nonopaque mineral, possibly exsolved spinel.

Blades of ilmenite in magnetite are present in almost every polished section. Commonly, a few magnetite grains will show one or two blades apiece, but in certain specimens more than half the magnetite grains have several blades of ilmenite. Ilmenite grains are always present where these blades are abundant. Ilmenite blades often occur hit-or-miss in the magnetite, though reticulate plates parallel to the {111} plane of magnetite are observed in places. 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 mineral. Slivers of ilmenite, a few microns long, are found in a few magnetite grains. These slivers are brought out only after the magnetite has been etched by HCl.

HEMATITE

Traces of hematite have been observed in several polished sections. The hematite occurs as tiny flecks in silicates, or (in one specimen) as an extremely fine network in magnetite and silicates. In one place the hematite network was associated with veinlets of pyrite II. The amount of hematite so far detected in polished sections can account for a mere fraction of the 1 to 2 percent of soluble nonmagnetic Fe characteristic of the ore zones in hole T3, for example.

ILVAITE

Ilvaite, a hydroxyl-bearing Ca-Fe^{+2}-Fe^{+3} silicate, soluble in HCl, is present locally in the ore. Though ilvaite forms 15 to 20 percent of one specimen, the mineral seems to be rare in the deposit as a whole. It is described on pages 61–62, in the section on ore minerals of the district.

SULFIDES

Occurrence

The habit and occurrence of sulfides, stated in general terms, are as follows:

1. Specks in any type of rock, including all grades of magnetite-bearing rock.
2. Euhedral crystals (rare; pyrite, molybdenite).
3. Small blotches a few millimeters in diameter.
4. Veinlets (pyrrhotite).
5. Concentrations. Pyrrhotite tends to form concentrations, whereas pyrite is more commonly observed as disseminated grains or blotches.
6. Joint coatings of pyrite with chlorite, quartz, and calcite. Fissure veinlets of sphalerite (exceedingly rare).
7. Coatings of minute pyrite crystals on some vugs found in holes T12 and T13. The pyrite is associated with chabazite cubes, quartz crystals, and chlorite.
8. Hair-thin veinlets and rims of pyrite (observed microscopically).

Relative concentrations (several percent) of sulfides may occur in ore, or in zones that have little or no magnetite. Nevertheless, magnetite and sulfides are very likely to occur together. Sulfides replace both felsic and mafic minerals of modified skarn or mixed skarn and pegmatite, but they usually replace the skarn schlieren or recrystallized aggregates preferentially, often to a marked degree. In the light of this statement, it is surprising that disseminated sulfides may be more abundant in a zone of contaminated granite gneiss (mineralized by magnetite) than in the overlying modified skarn, which is also magnetite-bearing (cf. hole T3, 217–237 ft).

Pyrrhotite, pyrite, or chalcopyrite may be the dominant sulfide disseminated through a given ore layer, or pyrrhotite and pyrite may be about equal and chalcopyrite quite subordinate. Chalcopyrite does not seem to favor pyrrhotite above pyrite as an associate. A very slight alteration of pyribole to chlorite is common wherever pyrite is abundant.

Molybdenite specks are extremely rare. They usually appear only where other sulfides are present, but rarely a speck of molybdenite is present as the sole sulfide.

Sphalerite was seen megascopically in one place (hole T12, 268 ft). It occurs with pyrite, chlorite, and epidote as the selvage of a cross-cutting fracture filled with calcite. It probably belongs to stage II (see below). In addition, minute but significant quantities of sphalerite were seen in polished sections viewed at high magnification.

Zoning

The writer did not carry out special studies to determine the presence of zoning for the deposit as a whole, but an incidental study of sulfide distribution was made on core from hole T3. A summary of this work is found on p. 217. Moreover, chemical data are limited, since core from the three holes drilled by the Bureau of Mines was not assayed for sulfur. The most one can say regarding zoning is that significant amounts of chalcopyrite were found only at the north end of the property (the ore zone in hole T5 contains several percent of chalcopyrite), and that the shoots of ore vary in their sulfur content. Ore zones with higher sulfur content also contain appreciably more phosphorus (and TiO_2?).[48]

Microscopic details

Study of thin sections confirms the relations visible on the surface of drill cores. The sulfides replace pyroxene, amphibole, biotite, quartz, and feldspar (but not garnet?). They occur as discrete grains, small masses, and fillings between silicate grains. A little chlorite in very fine-grained aggregates accompanies the sulfides. "Late" pyrite fills cleavages and fractures in feldspar, pyroxene, and (rarely) magnetite.

Study of polished sections checks the paragenetic sequence determined megascopically and, in addition, reveals the presence of two generations of pyrite and sphalerite.

Early pyrite (designated pyrite I, for convenience) forms anhedral and subhedral grains and rare cubes in the silicates. It replaces magnetite and is replaced by pyrrhotite (and locally chalcopyrite?). It is pyrite of this generation that one usually notes in logging the core.

Later pyrite (pyrite 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 seen only in polished sections. Some "filigree pyrite" may also belong to this later generation. All pyrite II is isotropic, with one questionable exception: a single rim of anisotropic marcasite(?) so fine that one can barely see the anisotropism, even with an oil immersion lens.

Microscopically, sphalerite of stage I replaces silicates, magnetite, pyrite, and pyrrhotite. The sphalerite, which is filled with minute exsolved blebs of chalcopyrite, commonly occurs near larger grains of chalcopyrite. The chalcopyrite of these larger grains locally replaces the sphalerite.

About half the ore specimens showed 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 grains about 50 microns in diameter.

The grains are gray (not so white as magnetite), softer than magnetite, and isotropic. At high magnification they show strong internal reflection in shades of white or brown. The grains were much too small to use for a microchemical test for zinc. Instead, several areas were etched for about 15 seconds with 1:1 $KMnO_4$ to which a drop of H_2SO_4 had been added. The resulting brown stain, locally red and blue, was wiped off with a moist cloth, exposing the etched surface. This showed that many "single" grains were aggregates of individuals showing parallel lines of cleavage (twinning?). In places, two sets of lines intersect at about 60°. These features are visible only with an oil immersion lens.

The 50-micron grains are 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. On a few grains, tiny projections cut the enclosing silicates.

Pyrrhotite in the ore appears as disseminated grains, locally elongate, in the silicates. The grains are about 0.1 to 0.3 mm across. Local protuberances

[48]

Hole	*Depth (feet)*	*Percent*	
		S	*P*
T5	82.5–109.7	3.51	0.10
T3	165–230	1.22	.117
	335–385	.52	.017
T1	24.5–55*	.64	.024

*Magnetite in this zone is of questionable commercial grade.
Assays by Jones & Laughlin Steel Corp. for the Newton Falls Paper Mill.

cut silicates and magnetite, though the magnetite and pyrrhotite often show "mutual boundaries." Pyrrhotite also replaces ilmenite and pyrite I. Small areas of chalcopyrite are visible in some of the pyrrhotite, and one pyrrhotite grain showed a trace of bladed molybdenite. Certain pyrrhotite veinlets in skarn tail out into veinlets of chalcopyrite. In the very few places where pyrrhotite is abundant, it clearly replaces magnetite as well as silicates. The aggregates of pyrrhotite are made up of individuals with curved or sutured borders, unlike the polygonal individuals found in magnetite.

Chalcopyrite is also present as small, discrete grains in silicates, as well as in pyrrhotite. Locally, it occurs in traces along the edge of pyrite I grains. In one pyrrhotite-rich specimen that had perhaps 2 percent of chalcopyrite (a considerable concentration for that sulfide), the chalcopyrite partly rimmed some magnetite and formed conspicuous replacement veinlets in other magnetite aggregates. A bit of this chalcopyrite showed twinning.

Stages of sulfides; paragenetic sequence

Two, or perhaps three, generations of sulfides are thus recognized. Pyrrhotite, chalcopyrite, most of the pyrite, the sphalerite with exsolved chalcopyrite, and molybdenite belong to stage I. These appear to be closely related in origin to the magnetite. To stage II belong the hair-thin veinlets and rims of pyrite, as well as the tiny, ragged specks of sphalerite. Probably the vug coatings of pyrite and the sphalerite fissure veinlet also belong to this stage. Pyrite coatings on joints may belong to stage II, or they may belong to a third stage, perhaps supergene in origin. Such coatings and veinlets of pyrite-chlorite-calcite occur sporadically in ore as well as in unmineralized rock. They cut silicates, magnetite, and the sulfides of stage I.

The paragenetic sequence of the main metallic minerals of stage I is well defined from field and microscopic study. It is:

magnetite→pyrite→pyrrhotite→sphalerite→chalcopyrite

The position of molybdenite is uncertain. It may be slightly younger than pyrrhotite. Magnetite and ilvaite seem to be contemporaneous, in part.

PATTERN OF DISTRIBUTION OF ILMENITE, APATITE, AND SULFIDES

The distribution of ilmenite, apatite, and sulfides in the two thicker ore zones cut in hole T3 was studied in thin and polished sections. A. M. Ross pointed out to the writer that the two ore zones had nearly the same content of soluble and magnetic Fe but very different values for P, S, and TiO_2. (See table 38, hole T3, 165–230 and 335–385 ft.)

A note regarding the general method of sampling will explain why this hole was selected for study and why it was not feasible to check the results with data from other holes on the property. By mutual agreement, the Geological Survey refrained from taking samples from ore zones cut by holes T11, T12, and T13 until the core had been split for assaying. A study of the distribution pattern of minor constituents requires that every piece of split core be returned to the core box in precisely the right sequence. The Newton Falls Paper Mill carefully filed all split core in proper sequence. Ordinarily, the transposition of a few pieces in a 5- or 10-foot sample length makes no difference in the record, but it does make the split core useless for detailed study. Only hole T3, then, provided suitable core for studying thick ore zones where some kind of distributional pattern might be expected.

For the present study, representative chips of the ore were taken at 10-foot intervals. Parallel slices were used, wherever possible, for the thin and polished sections. The ore has replaced modified skarn and quartz rock of highly variable composition. The grade of the ore itself is variable, as is shown by the Rosiwal analyses of polished sections, table 36.

Table 37 shows the distribution of ilmenite, sulfides, and apatite in the two ore zones. It is regrettable that the observations are given in qualitative terms, rather than in figures or in graphs representing the figures. However, the quantity of these minor constituents ranges from a trace to a maximum of 3 to 5 percent (10–12 percent, in the case of pyrrhotite). Thus, the task of determining the absolute quantity by Rosiwal analysis or by grain count would be very difficult. Comparisons within a column can be made with assurance, as the terms are relative to the abun-

TABLE 36.—*Modes (volume percent) of four specimens from ore zones, hole T3, Trembley Mountain deposit*

	"Average" ore [1]	Subore ("leanest" of 13 specimens)	High-grade ore ("richest" of 13 specimens)	Ore with highest ilmenite content
Depth (feet)	205	350	215	165
Magnetite	21.9	16.3	50.7	33.9
Ilmenite [2]	.2	$<<.2$	.4	2.8
Sulfides [2]	$<<.2$	0	.2	4.6
Nonopaque minerals	77.9	83.7	48.7	58.7
Total	100.0	100.0	100.0	100.0

[1] Specimen has less ilmenite and sulfides than the assays indicate for the whole zone, 165–230 ft, but magnetite content (24 percent magnetic Fe, assuming sp. gr. 3.0 for rock) agrees well with 24.22 percent magnetic Fe, the assayed value for the zone.
[2] Values approximate. Analyses were made to obtain a value for magnetite correct to ~ 2 percent. Values for ilmenite are probably too low, except in the last column.

TABLE 37.—*Distribution of ilmenite, sulfides, and apatite in two ore zones, hole T3, Trembley Mountain deposit*

Depth (feet)	Host rock	Ilmenite [1]		Pyrite I (grains)	Pyrrhotite	Chalcopyrite	Pyrite II (veinlets)	Sphalerite II	Apatite [2]
		Grains	Blades						
				Upper ore zone					
165	Feldspar-pyribole skarn	Abundant	Abundant	Abundant	Tr.	Common	Abundant	Tr.	Little.
175	Feldspar-pyroxene skarn	Common	Common	Tr.			Fairly common	Tr.	Very little.
185	Quartz-feldspar-pyroxene skarn	Few	Fairly common	Very little	Tr.		Fairly common	Tr.	Little.
195	Quartz-feldspar-pyroxene skarn	Common	Fairly common	Common			Common	Tr.	Common.
205	Quartz-feldspar-pyribole skarn	Very few	Fairly common	Little	Tr.		Few		Little.
215	Quartzose skarn	None	Few	Very little	Fairly common	Tr.	Few		Little.
225	Pyroxenic quartz rock	None	Few	Tr.	Common	Very little	Tr.		Extremely abundant.
230	Quartz-pyroxene skarn	None	None	Few	Abundant	Little	None		Extremely abundant.
				Lower ore zone					
340	Pyroxenic quartz-feldspar rock	None	Common	Very little			Very few		Common.
350	Quartz-feldspar-pyribole skarn	None	Common						Common.
360	Feldspathic skarn	None	Few	Tr.			Very few	Tr.	Little.
370	Feldspathic skarn	None	Few					Tr.	Extremely abundant.
380	Quartzose skarn	Abundant	Fairly common	Little			Common	Tr.	Extremely abundant.

[1] Maximum at 165 feet: ~2.8 percent.
[2] Maximum at 370 feet: ~0.7+ percent (probably low because of torn slide).

dance of a given mineral—for example, pyrite I. Comparisons between columns must not be made; the term "abundant" applied to pyrrhotite may mean about 10 percent, whereas "abundant" applied to ilmenite blades may mean several tens of blades that total less than 1 percent of the rock.

Much more serious than the lack of absolute values is the spacing of the samples. The discrepancy between grain counts of apatite and assays for phosphorus suggests that variations in minor constituents may be so rapid that 5- or 3-foot samples are required for proper study. The conclusions stated below are, therefore, tentative.

Ilmenite.—In the upper ore zone, ilmenite decreases progressively from top to bottom. A similar decrease takes place in the lower zone, except that a relative concentration appears at the base of this zone. Study of thin and polished sections shows that the quantity of ilmenite is independent of the rock type and, roughly, of the quantity of magnetite. Certainly the highest grade of ore is not consistently the one with most ilmenite, nor does the leanest ore always show the smallest quantity of ilmenite, at least in the 13 samples studied.

Almost all the TiO_2 present in the two zones is in the form of ilmenite. Sphene occurs sporadically, but only in traces. The upper zone assays twice as much TiO_2 as the lower because grains of ilmenite often accompany the blades present in the upper zone, whereas grains are usually lacking from the lower zone.

Sulfides.—Not much can be made of the variation in the sulfides, except that they are more abundant in the upper zone. They show in detail the same erratic distribution noted in routine logging of core. Pyrrhotite is important only in the basal part of the upper zone. Chalcopyrite commonly accompanies pyrrhotite, rather than pyrite, but may occur without it. The close association of "late" pyrite and "late" sphalerite is worth mentioning.

Apatite.—A remarkable concentration of apatite occurs at the base of both ore zones. The variation in quantity of apatite is not directly related to the kind or quantity of the felsic mineral present in the skarn, or to the quantity of magnetite. The variation in the apatite is also independent of the variation of ilmenite.

GRADE AND EXTENT

Table 38 shows the grade and thickness of the principal layers of ore.

The principal ore zones occur at or near the top of the zone of modified skarn in holes T5, T6, T11, T12, and probably T1. Hole T3 cut 3 ore zones roughly grouped about the central part of the zone of modified skarn. Hole T13 cut ore through most of its bedrock extent, with one zone at the top of the zone of modified skarn but with the principal ore zone slightly above the middle of the zone of modified skarn. In hole T2, ore of marginal grade occurs near the base of the modified skarn zone. The "stratigraphic" position of the subore cut in hole T4 is uncertain. In holes that cut multiple zones of ore, the thickness of waste rock between adjacent ore zones ranges from 5 to 60 feet and is very variable within those limits. The thickest and highest grade ore has been found at the south end of the property.

Obviously, the information obtained from magnetic and geologic mapping and the drilling of 9 holes has not proved either the structure or the extent of the ore zones. One may reasonably infer the thickness and strike length of several zones; but their extension

TABLE 38.—*Grade and thickness of principal magnetite-bearing layers, Trembley Mountain deposit*

[Includes all layers of ore grade (at least 20 percent magnetic Fe) having a minimum true thicknesss of 6 ft. Figures for soluble Fe and magnetic Fe are weighted averages (for assays of individual samples, see Miller, 1947, p. 13–14, and records of Newton Falls Paper Mill)]

Hole	Footage	Apparent thickness (feet)	True thickness (feet)	Weight percent				
				Soluble Fe	Magnetic Fe	P	S	TiO_2
T1 [1]	24.5– 55.0	30.5	21.1	26.7	25.6	0.024	0.64	0.39
T2 [2]	18.0– 54.0	36.0	29.5	------	~22	------	------	------
	62.7– 75.7	13.0	10.7	------	~21	------	------	------
T3 [1]	165.0–230.0	65.0	47.5	26.8	24.2	.117	1.22	.72
	270.0–290.0	20.0	14.6	31.6	30.0	------	------	------
	335.0–385.0	50.0	36.5	26.3	24.5	.017	.52	.37
T4 [3]	16.2– 33.5	17.3	15.7	------	<20	------	------	------
T5 [1]	82.5–109.7	27.2	21.0	27.9	25.0	.10	3.51	------
T6 [1]	97.0–118.5	21.5	18.7	28.1	26.2	------	------	------
T11 [4]	94.5–115.0	20.5	17.8	------	27.3	------	------	------
T12 [4]	413.0–440.5	27.5	21.5	------	28.2	------	------	------
	517.4–527.2	9.8	7.7	------	22.9	------	------	------
T13 [4]	146.5–189.0 [5]	42.5	31.0	------	~21.7	------	------	------
	259.0–360.8	101.8	74.3	------	27.7	------	------	------
	374.7–446.1	71.4	52.1	------	24.5	----	------	------
	453.3–465.3	12.0	8.8	------	27.4	------	------	------
	502.2–518.7	16.5	12.1	------	~19.5	------	------	------
	549.0–557.7	8.7	6.4	------	25.8	------	------	------

[1] Assays by Jones & Laughlin Steel Corp.
[2] No assays. Grade estimated by megascopic Rosiwal analysis.
[3] Data included only for completeness of record. Magnetite-bearing layer is probably not of ore grade.
[4] Assays by U.S. Bureau of Mines.
[5] First 10 ft is subore.

downdip, and the possible importance of shoot structure, remains unknown. If the major structure is that of an overturned isoclinal syncline, the ore cannot continue indefinitely to the east or southeast. One cannot now predict whether the axial portion of the syncline will be richer or poorer than the limbs. The section through holes T12 and T13 suggests that the concentration of ore may be "updip," rather than in the axial region. However, the existence of another shoot east of hole T12 is not ruled out. The concentration of ore in holes T3 and T13 may be related to a plunging drag fold on the limbs of the major structure. (See pl. 19 and section on structure, above.) The concept of plunging shoots related to similar drags, or to other linear elements, is worth retaining as a guide for additional exploration.

Table 39 seeks to classify what we know and may conservatively infer regarding the extent of ore in some of the blocks of ground. The tabulation should not be misleading; surely it will be provocative.

TABLE 39.—*Extent of ore (conservatively inferred) for several blocks at Trembley Mountain*

Block	Length (feet)	Average thickness of ore (feet)	Downtip extension (feet)
Area centering at (9350N, 23150E). Partly drilled.	[1] 240	20	300?
Area centering at (9600N, 23800E). Undrilled.	[1] 220	(?)	(?)
Area centering at (8400N, 23100E). Undrilled.	[1] 400	(?)	(?)
Area between latitudes 8000N and 7550N. Partly drilled.	450?	19	200?
Area between latitudes 7550N and 7150N. Partly drilled.	500–600	[2] 80?	450?
Area south of latitude 7150N. Undrilled.	(?)	(?)	(?)

[1] Within +40° isoclinal.
[2] Several layers.

ADDITIONAL EXPLORATION

A low positive anomaly extends for a considerable distance south of 7000N. Overburden is thick in this area (140–150 ft in holes T12 and T13), yet the area is most favorable for prospecting, judging from the results of hole T13 and the character of the dip-needle and aeromagnetic anomalies. A number of traverses run by vertical magnetometer or Hotchkiss superdip might develop a pattern for the weak dip-needle anomaly and thereby assist in spotting additional drill holes south of the T12–T13 section.

If additional exploratory drilling is undertaken, the sites listed in table 40 are logical ones to occupy, roughly in the order given. Obviously, any company that might drill the deposit would adopt a scheme fitting its particular needs. If preliminary work indicates that the overburden is thin, bulldozing might yield part of the information sought by proposed holes T21, T22, T23, T26, and T27.

MISCELLANEOUS ECONOMIC FEATURES

Soluble nonmagnetic iron

Soluble nonmagnetic Fe averages about 2 percent in the ore, according to the assays now available. As previously noted, hematite is sparse or absent in sections of representative ore from two zones cut by hole T3. Therefore, one must either assume local, undetected, concentrations of hematite or look to some other mineral as the source of soluble Fe in the Trembley Mountain ore. Unfortunately, the relatively few "complete" assays at hand are inadequate for a statistical study of the Fe-S-TiO_2 relations.

Pyrrhotite, which varies considerably in magnetic properties, may have contributed part of the soluble Fe. Chalcopyrite could, but the amount of this sulfide is negligible except in holes T5 and T6. Pyrite is relatively insoluble in hydrochloric acid used to determine souble Fe; so it should not be guilty. However, one wonders if the very fine pyrite of stage II might be more soluble than common pyrite. Of the dark silicates present (garnets, pyroxenes, amphiboles, micas; rarely, ilvaite), only the pyroxene is consistently present in substantial quantities in zones relatively high in soluble nonmagnetic Fe. Maghemite has not been detected in the ore. Ilmenite is generally sparse, and it is intimately bound to its magnetite host; moreover, the ilmenite itself is somewhat magnetic. Therefore, it should not be giving the iron in nonmagnetic fractions of crude ore. At present, it appears that pyroxene is the most likely major source of soluble nonmagnetic Fe at Trembley Mountain.

TABLE 40.—*Possible drill sites for additional exploration at Trembley Mountain*

Hole	Coordinates N	Coordinates E	Bearing	Inclination (degrees)	Approximate depth (feet)	Purpose
T14	7000	23400		90	600	Ore extension from T13.
T15	6700	23400		90	700	Ore extension from T13 (and T14).
Additional holes south of T15, pending magnetometer survey and favorable results from T15.						
T16	7100	23045		90	300	Structure and extension of T13 ore.
T17	7295	22988		90	400	Structure and extension of T3 ore.
T18	7600	23100		90	600	Continuity of ore between T3 and T1.
T19	WSW. of T18; location variable.			90?	?	Extension of T18 (if favorable).
T20	7508	23571		90	700	± Wildcat. Extension of T12 ore. Possible concentration along synclinal axis.
Additional holes SSE. and NNW. of T20, if that hole is especially good.						
T21	8320	23340	W.	45	200	Structure and ore.
T22	8100	23200	N. 18° W.	45	600	Structure and ore.
T23	8100	23200	N. 50° W.	45	450	Only if T22 is favorable.
T24	8090	23400		90	450	± Wildcat. Possible SE-plunging shoot centering at (8400N, 23220E).
T25	9100	23250		90	400	Extension of T5–T6 ore.
T26	9340	23170		90	150	Structure.
T27	9600	23950	W.	70	140	Testing anomaly.

The significance of the presence of 2 percent soluble Fe and the apparent absence of hematite is obvious: any attempt to recover the additional 2 percent Fe "in hematite" by means of Humphreys spirals, or similar devices, would be futile. The problem of what yields soluble nonmagnetic Fe could best be attacked by studying thin and polished sections of the magnetic and nonmagnetic fractions of the ore. The fractions should not be so fine as those used for assay, but should be about minus-150, plus-200 mesh. Assays of the magnetic and nonmagnetic portions for soluble Fe, TiO_2, and S would be a necessary aid.

Ilmenite-apatite-sulfides

Studies of zoning in the two ore zones of hole T3 show that the content of ilmenite, apatite, and sulfides is variable. While the average content of TiO_2, P, and S is low, these substances are concentrated locally. In general, the apatite would be removed in the magnetic concentration, but most of the ilmenite and part of the sulfides would not. The difficulty would not arise, of course, if a sufficient width of any ore zone were mined, for the concentration of impurities would be reduced to the average low level. Until the pattern of zoning is clearly established, it is desirable to assay the 5- or 10-foot samples for TiO_2, P, and S, instead of making a so-called complete assay on the composite sample only.

Copper

Several percent of chalcopyrite is present in the ore cut by holes T5 and T6. Conceivably, this might some day be worth recovering as a byproduct.

Virtual absence of precious metals

Because of the reported values for silver and gold in sulfides in hole B6, Brandy Brook deposit, several sulfide-rich zones at Trembley Mountain were sampled and assayed. The results are shown in table 41.

Allanite

Trace elements are increasing in importance, and allanite is a source for some of them. At present, accessory allanite in the Trembley Mountain ore has no commercial interest. The greatest concentration so far observed is only about 0.6 percent. However, local concentrations of allanite occur in the iron mines of the Port Henry district; and something similar may be looked for, if not predicted, at Trembley Mountain. (For partial optical data on allanite from Trembley Mountain, see table 2, No. 7.)

TABLE 41.—*Precious-metal assays of selected sulfide-rich ore, Trembley Mountain deposit*

[Assayed for the U.S. Geological Survey by W. W. Hommel, Ledoux & Co., Inc.]

Sample	Footage	Description of material	Assay Ag	Assay Au
T1-100	84′5″–85′	Pyroxene skarn with disseminated magnetite, pyrrhotite, and a little chalcopyrite. One quartz-pyrrhotite blotch.	Tr.	Tr.
T4-100	51′3″–51′6″	Quartz-pyrrhotite vein with pyroxene blotches.	Tr.	Tr.
T4-101	51′6″–51′10″	Pyrrhotite-bearing pyroxene skarn with a little quartz, pyrite, and chalcopyrite.	Tr.	Tr.

PARISH DEPOSIT

LOCATION AND ACCESS

The Parish deposit, also known as the Parish ore bed, crops out on a low ridge west of Tracy Pond Outlet, between Gooseberry Mountain and Tunkethandle Hill, west-central rectangle, Stark quadrangle (see plate 1).

To reach the deposit, turn east at the bridge over the Middle Branch of the Grass River, about 2 miles by macadam road north-northeast of Degrasse. Drive eastward on the graded sand road to Canton Farm. Continue eastward 1.6 miles on the ungraded road. Park your car, and walk south about 1,000 feet along the hunters' trail to the outcrops at the north end of the deposit. In dry seasons, the trip could be made by passenger car as far as the junction of the road and the hunters' trail, part of which could also be driven under favorable conditions.

The alternate route to the deposit, via the grade of the abandoned lumber railroad that runs northwest across the Stark quadrangle, is not recommended.

In 1949, the mineral rights for the Parish deposit were held by the W. H. Loomis Talc Corp., Gouverneur, N.Y.

EXPLORATION AND GEOLOGIC STUDY

The Parish deposit, known for many years, is one of the few outcropping magnetite deposits in the district. However, no ore has been produced from the Parish, and the only intensive exploration of the deposit was that undertaken by the M. A. Hanna Co. in 1941. The company made a detailed dip-needle survey of the deposit and drilled 30 holes. (See pl. 20.) Walker (1943) mentions this exploration and gives a brief account of the geology of the deposit.

In 1944, the Geological Survey logged in detail 2,500 feet of core from 8 representative drill holes on cross sections spaced 300 to 500 feet apart along the strike of the deposit. Cores totaling 500 feet from 5 other holes were briefly examined to locate the contacts of certain major rock units and to get additional data on the distribution of manganese in the ore. Structural data could not be obtained from the cores, because almost all these had been split for assay before any core-foliation angles were recorded. Some plane-table geologic work was done in 1945 on the northern end of the deposit, but the thick cover of brush and the scarcity of outcrops and reliable float in most critical areas quickly showed that little additional geologic information could then be gained by detailed mapping. The major geologic features of the deposit are well known (see section on magnetite deposit in modified granite gneiss), though certain details of engineering and economic interest, such as the presence or absence of shoot structure in the ore, are not.

MAGNETIC ANOMALY

The magnetic anomaly mapped in detail by the M. A. Hanna Co. is 5,000 feet long. (See pl. 20.) The main part of the anomaly, north of latitude 1800S, trends north and is 2,600 feet long within the +40° isoclinal. It contains two segments, 1,150 feet and 780 feet long, effectively within the +60° isoclinal. All but the northernmost 500 feet of the main part of the anomaly has been tested by diamond drilling.

An economically less promising part of the anomaly, trending generally north-northwest, lies south of latitude 1800S. This part has a northern segment, about 900 feet long, effectively within the +40° isoclinal; an intervening "magnetic saddle" of low positive readings; and a southern segment, 300 feet long, again within the +40° isoclinal. The northern segment of this part of the anomaly has been tested by three drill holes; the rest is unexplored.

In gross form, the Parish anomaly is a simple linear one. However, local V's in the anomaly, as at (450S, 150E) and (750S, 190E), are faintly suggestive of isoclinal folding. The offset of the anomaly's axis at latitude 250S, together with the bulge in the wallrock contact at (250S, 130W), might also be interpreted as effects due to folding. The interpretation, a highly tentative one, is further discussed in the section on structural features of the deposit.

Most of the Parish anomaly has strong local magnetic relief indicative of shallow overburden. The thickness of overburden at most drill sites for which we have data is commonly 10 feet or less, and part of the mineralized zone crops out between 400S and 50N. The character of the magnetic anomaly suggests that the overburden thickens considerably in the area between 500N and 1000N, possibly becoming 60 to 70 feet thick. A similar pattern of anomaly might be given by northward decrease in the quantity of magnetite. However, the fact that the overburden thickens to 30 feet at the sites of holes 15 and 16 lends support to the hypothesis that the lower intensity of the magnetic anomaly in the area between 500N and 1000N is due at least in part to progressive northward thickening of overburden. This segment of the anomaly, never explored by drilling, deserves at least one hole, if additional drilling is ever undertaken.

The results of drilling in the area between 2000S and 2400S, together with the pattern of the anomaly, suggest that the possibilities of finding considerable magnetite in the area south of 1800S are slight.

The Parish deposit was registered on the aeromagnetic survey of 1945 as a magnetic ridge having a peak value of 8,755 gammas (Balsley, Buddington, and others, 1954b). A strong negative anomaly having an intensity of 4,070 gammas centers near Tracy Pond, 1.85 miles south of the 8,755-gamma peak (see Balsley, Buddington, and others, 1954b), or about 1.5 miles south of latitude 4000S, plate 20. We do not know whether this negative anomaly is (1) the complement of the intense positive anomaly yielded by the Parish deposit, (2) the result of varying proportions of iron and titanium oxides occurring as accessory minerals in the country rock, or (3) the result of a substantial concentration of crystalline hematite similar to concentrations occurring locally in the Parish magnetite deposit. The first possibility seems remote, and the second is not readily susceptible of proof, for the Tracy Pond area is largely mantled by glacial debris (see pl. 1). A drill hole or two, placed to test the third possibility, might be warranted, especially if the Parish deposit is further explored or developed.

GEOLOGY

GEOLOGIC SETTING

The Parish deposit is in the Clare-Clifton-Colton belt of metasedimentary rocks and granite gneiss sheets. (See pl. 1.) Here rocks of the belt strike north, dip vertically, and are flanked by narrow belts of hornblende granite, in part containing layers of metasedimentary rocks. Because outcrops in this part of the belt are neither abundant nor widespread, the gross structure of the immediate area of the deposit is uncertain. Presumably the gross structure of the belt of metasedimentary rocks is synclinal or synclinorial, for the belt lies between the northward extension of the Spruce Mountain anticline (Stark quadrangle) and minor anticlines on the east flank of the Stark anticline. (See pl. 1; and Prof. Paper 376, p. 118–119, and fig. 14.) The Parish deposit may occupy the apical zone of a tight, southward-trending isoclinal fold: pyroxenic and biotitic gneisses like those found on the east side of the deposit are also found on the west (see pl. 20), suggesting duplication by folding. Small isoclinal folds are abundant in wallrocks near the deposit. The trend of these folds is roughly southward. We did not succeed in measuring the plunge of the fold axes or recognizing consistent asymmetry that might have been helpful in working out the larger structures.

There is some suggestion that, in addition, the Parish deposit is localized on an anticlinal cross structure that trends roughly eastward. Poorly exposed minor folds in pyroxenic quartzites on Gooseberry Mountain trend about S. 25° E. and plunge 33° SE. But 3 miles east, near the south end of the Granshue syncline, linear structures trend and plunge north-northeastward (Prof. Paper 376, fig. 15).

ORE

GENERAL FEATURES, GRADE, AND EXTENT

The main geologic features of this deposit, including the mineralogy of the ore and gangue, have been presented elsewhere (p. 51–54). The following account stresses features of potential economic interest.

Magnetite, in places accompanied by or proxied by crystalline hematite, occurs as lenticular bodies in a vertically dipping sheet of microcline granite gneiss that generally ranges in thickness from 100 to 140 feet. The ore is usually disseminated or streaky, but in places thin zones of massive high-grade ore are present. The main mineralized zone ranges in thickness from 11(?) to 134 feet; its average thickness is 55 feet. Its grade ranges from 14.6 to 38.5 percent Fe;[49] the weighted average is 21.7 percent Fe. The main mineralized zone, which nowhere occupies the full width of the granite gneiss sheet, is undulatory in plan (see pl. 20), weaving from side to side within the granite gneiss sheet. Within the limits of the main mineralized zone, individual ore-bearing zones commonly range in thickness from 3 to 35 feet; exceptionally, a single zone may be 70 to 80 feet thick. The grade of these individual zones commonly ranges from 24 to 40 percent Fe; exceptionally, it is as low as 20.3 percent Fe or as high as 57.6 percent Fe. Zones of high-grade ore are seldom more than 5 feet thick, and it is unlikely that lump ore could ever be recovered except during a national emergency. In a single drilled cross section, 1, 2, or 3 ore-bearing zones, in the strict sense of the term, may be present, the rest of the zone consisting at that latitude of weakly or moderately mineralized modified granite gneiss, in places with relict metasedimentary material. For much of the deposit, the combination of ore-bearing zones plus subore and mineralized rock constitutes marginal ore, as noted above. The available data are inadequate to show whether the individual ore-bearing zones form discrete subparallel lenses or merge to form an elongate, braided pattern. The latter interpretation has been adopted for plate 20; the alternate interpretation is discussed subsequently.

Measured and indicated ore, as determined by the M. A. Hanna Co., has an average thickness of 36 feet; its grade averages 38 percent Fe and 3 percent Mn. The company used 30 percent Fe as the lower limit for defining ore. We have used 20 percent Fe as the cutoff in this report and on the accompanying illustrations. By our definition of ore, much of the main mineralized zone constitutes marginal ore, if crystalline hematite as well as magnetite were recoverable.

Most drill holes tested the main mineralized zone at vertical depths of 50 to 150 feet. Six holes tested it at vertical depths of 250 to 600 feet. Data from the shallower holes have been used in projecting the main mineralized zone to the surface, on plate 20. On plate 20 are also shown two drilled cross sections through the deposit.

The grade and thickness of ore vary in depth, on a given cross section, much as they do along strike—from one section to the next. On the section at 000N,

[49] For the Parish deposit, percent Fe, determined by the M. A. Hanna Co., is percent total Fe. Data on percent soluble Fe and percent magnetic Fe are not available. The implications with regard to our knowledge of the distribution of magnetite and hematite are obvious. For all but garnet-rich parts of the ore, the value for soluble Fe (representing mainly iron in magnetite and hematite) should be very close to the value for total Fe.

there is almost no change. On sections at 500S and 1300S, the thickness of ore is increased two to six times at depth; the grade is reduced about 30 percent, but the iron-bearing material at depth still constitutes ore. On the section at 1000S, the thickness is drastically reduced, about nine times, but the grade is increased 10 percent. If for this section we consider marginal ore, however, the thickness of the zone is halved at depth and the grade is reduced 15 percent.

A similar comparison of the main mineralized zones, not merely the ore-bearing parts, cannot be made from the data at hand.

Outside the main mineralized zone, but still within the main granite gneiss sheet, are scattered weakly mineralized zones and rare 1- to 2-foot zones rich in iron oxides, not shown on plate 20. These mineralized zones, plus the main one, give the granite gneiss sheet as a whole a content of about 13 percent Fe. This figure is a crude approximation only. Iron analyses are available for cores from two holes (13 and 26, near the ends of the explored belt) that cut the entire thickness of the granite gneiss sheet. Similar data are at hand for cores from hole 16, which cut all but the eastern third of the granite gneiss sheet. The data are tabulated below (table 42).

TABLE 42.—*Approximate iron content of main granite gneiss sheet, Parish deposit*

Hole	*Footage*	*True thickness (feet)*	*Fe (percent)*
13	20–220	141	13.4
26	105–245	99	10.0
16	35–180	[1] 102	16.8
	Weighted average		13.4

[1] Partial.

Since two holes (13 and 16) sampled the richer, northern part of the granite gneiss sheet and just one hole (26) sampled the leaner, southern part, the weighted average grade of about 13 percent Fe is at best only crudely representative of the iron content of the whole granite gneiss sheet.

Though all the iron oxide mineralization of potential commercial interest is restricted to the main granite gneiss sheet, some iron oxides (apparently all magnetite) occur in the wallrocks, either in the metasedimentary rocks themselves or in minor thin sheets of microcline granite gneiss, a few feet thick, intercalated among the metasedimentary rocks. Magnetite commonly forms less than 5 percent of such zones, whose thickness is generally a few feet. Very rarely, a 1-foot layer of gneiss may have 10 to 12 percent magnetite. Such sporadic magnetite-bearing zones in wallrock account for the weak positive anomaly just south of the collar of hole 2, for example. Slightly higher concentrations of magnetite in wallrock may account for the strong positive anomaly just north of the collar of hole 24, though an abrupt change in the dip of the ore zones in the main granite gneiss sheet, or the presence of shoot structure, might instead be responsible for this anomaly.

POSSIBLE SHOOT STRUCTURE

Four features suggest that the ore may occur as plunging shoots within the main mineralized zone. These features are: (a) presence of local concentrations of garnet, hematite, or both minerals; (b) consistent parallel orientation of certain ore-bearing zones; (c) lenticular shape of certain ore zones; (d) pattern of the magnetic anomaly. If shoot structure is present, the plunge of the shoots is probably southward, perhaps at a very high angle. The problem is discussed, not because it has been solved, but because it is important for the effective additional exploration or development of the deposit, should such work ever be undertaken.

The irregular distribution of garnet and hematite is no more suggestive of shoot structure than the irregular distribution of any metallic or nonmetallic mineral would be. However, in some ore deposits nonuniform distribution of major constituents is due to their concentration in shoots, and for that reason the notes on garnet and hematite are presented here.

The thicker zones rich in spessartitic garnet are in the northern third of the deposit, perhaps chiefly in the northern sixth. Since the manganese content of the deposit is due almost wholly to the presence of spessartite, it follows that the thicker zones high in manganese (3–5 percent Mn for true thicknesses of 50 ft or more) are also in the northern part of the deposit, a fact made clear by the assays. Since some of the spessartite-rich zones are poor in iron oxides, while others have a moderate or large quantity of iron oxides, there is no direct correlation between high Fe and high Mn in parts of the mineralized zone that constitute ore.

A study of the distribution of hematite in the ore is handicapped by the lack of paired analyses for soluble Fe and magnetic Fe. Data from the cores we studied suggest that hematite is much more abundant at the north end of the deposit (segment tested by holes 15, 25, and 2) than in the middle or southern parts. The abundance is absolute, as well as relative to magnetite.

A second feature suggestive of shoot structure, the consistent parallel orientation of certain ore-bearing zones, is illustrated by plate 21. The zones were drawn up merely by connecting the horizontal projections of ore cut in the drill holes. This procedure has its concealed hazards, the chief one being the possibility that there are marked local changes in the dip of the ore. Whether the individual zones contain distinctive minerals or proportions thereof, we do not know; the holes whose core we logged are too widely spaced to give the data needed for a mineralogic correlation of these zones. The trend of the zones is N. 3°–10° E., which is more nearly parallel to local segments of the magnetic anomaly than to the axis of the anomaly as a whole. Moreover, the trend of the zones is in places at a considerable angle to the inferred strike of the wallrocks. So many uncertainties exist in the nature and quantity of data that the relations shown on plate 21 must be regarded as highly tentative and schematic.

A third feature suggestive of shoot structure is the lenticular shape of certain ore zones, well shown by the zone that apparently begins a short distance north of 800S. There is no indication in the magnetic anomaly that this lens extends north of 750S. Unfortunately, hole 7 was stopped too short to test the possible northward extension of this zone. The lens thins abruptly near 1500S and perhaps ends a short distance south of 1600S. Another lenticular body is found between 400N and 100S; the northward extension of this lens has not been explored.

Though the data are sparse, there is some suggestion that, as one goes southward, the deeper holes cut a greater number of ore zones, arranged en echelon, than were found at shallower depths farther north. This feature is not apparent on plate 20, for data from the shallower holes alone were used in preparing that illustration. The relations of the hypothetical shoots are such that their plunge, if southward, must be steep.

Some features of the dip-needle anomaly (pl. 20) and of the aeromagnetic anomaly (Balsley, Buddington, and others, 1954b) are consistent with an interpretation of ore occurring as shoots. The little V's is the dip-needle anomaly, previously mentioned as suggestive of tight folds, might reflect the presence of subparallel ore shoots, either related to folds or not so related. Such V's are visible at (800S, 200E), at (460S, 170E), at (300S, 120E), and at (100S, 100E). They occur at or near the north ends of what may be separate ore lenses or shoots. Their orientation with respect to the horizontal projections of the ore suggests that the shoots, if present, plunge southward. A similar plunge may be inferred from the pattern of the aeromagnetic anomaly, whose contours are closely spaced at the north end and more widely spaced at the south.

It should be plain that the foregoing account of possible shoot structure is highly speculative, yet the hypothesis is worthy of consideration if the deposit is further explored. Intensive study of local outcrops might yield information on the nature and orientation of local linear elements, such as fold axes or mineral elongation, likely to control the orientation of the ore shoots. Additional structural information could be obtained by stripping the very shallow overburden from critical parts of the mineralized zone. Conceivably, the hypothetical shoots might follow the *a* direction of the fold system, rather than the *b* or axial direction. If more than one fold system is present, the results would need especially careful interpretation. However, by drilling a relatively few holes one should be able to test the various possibilities for shoot structure, or at least to demonstrate that shoot structure is a negligible factor to be dealt with in developing the deposit.

ADDITIONAL ECONOMIC POSSIBILITIES

Three additional mineral products are worth keeping in mind in case the Parish deposit is developed for its magnetite content. These are crystalline hematite, spessartite, and copper sulfides.

When the Parish deposit was explored in 1941, the possibility of recovering crystalline hematite from Adirondack magnetite ores was recognized but not considered feasible under the existing wartime conditions. An appreciable tonnage of crystalline hematite concentrates is now produced annually at Benson Mines, concentration of the hematite from mixed magnetite-hematite ore being made by means of Humphreys spirals. The recovery of hematite at the Parish deposit should be equally feasible. In fact, since the proportion of hematite in the Parish ore is considerable though not precisely known, the possibility of recovering the hematite as well as the magnetite might make the difference between mere exploration and eventual development of the deposit.

In this connection, the possibility of testing the Tracy Pond negative anomaly for hematite concentrations again deserves mention.

Spessartite, locally abundant in the Parish deposit, is at times in demand for use as a welding flux. An analysis of the Parish spessartite is given elsewhere in this report (table 1, No. 3). The tonnage of spessartite in the explored part of this deposit may be sizable, if the ore averages 3 percent Mn, if all the spessartite contains 25 percent MnO, and if most of the Mn is assumed to be in the spessartite. A spessartite concentrate presumably could be made concurrently with the hematite concentrate, if a device such as the Humphreys spiral were employed. Fluorite, which in places constitutes several percent of the ore, might be concentrated by the same means.

If mass-mining of the entire mineralized zone were contemplated, it might be worth while to determine the copper content of all the core from the main granite gneiss sheet, or from selected parts thereof. The distribution of copper sulfides was not studied in detail by us. However, all specimens known to contain copper sulfides are from the main granite gneiss sheet; most of them were in the iron ore, but some of them were outside the intensely mineralized zone. Local zones, 1 foot or so thick, contain 5 to 10 volume percent copper sulfides, fine-grained and disseminated. For most of the core, this would represent an abnormally high concentration, the average presumably being less than 1 percent. However, since tarnished bornite closely resembles tarnished hematite, some copper sulfides may have been overlooked in logging the core. If a large enough tonnage of material were milled, the possibility of recovering the copper sulfides might be of interest.

HYPOGENE HEMATITE CONCENTRATIONS

By A. F. Buddington

Major hematite-bearing zones (Benson Mines, Parish) associated with magnetite deposits have already been described, and the occurrence of some primary crystalline hematite in all the granite gneiss ores of the district has been noted. Data on the composition and magnetic properties of accessory hematite disseminated as discrete grains in certain types of country rock have been summarized by Buddington (Prof. Paper 376, p. 79–82). Several lean concentrations of hematite are discussed below.

TITANIFEROUS HEMATITE AND TITANHEMATITE IN THE DEAD CREEK AREA

Three substantial negative anomalies in the Dead Creek area appear not to have a complementary relation to positive anomalies that are due to magnetite concentrations. The bedrock beneath all three negative anomalies is covered by overburden. They could be due either to localized vein concentrations of titaniferous hematite, or to accessory disseminations of titanhematite or ilmenohematite distributed through a broad zone of microcline granite gneiss. (Titaniferous hematite contains 1.5 to 5 percent TiO_2 in solid solution. Titanhematite contains 5 to 10 percent TiO_2 in solid solution. Ilmenohematite is hematite that contains exsolved ilmenite; the two-phase aggregate contains about 11 to 22 percent TiO_2.)

A lean concentration of titanhematite occurs in sillimanite-microcline granite gneiss in the Dead Creek area, on the east side of a small valley, west of Dead Creek anomaly No. 2. Exposures are inadequate to determine the extent of the deposit. A sample of the rock contained about 12 percent iron oxides, of which 80 percent was titanhematite and 20 percent magnetite. The titanhematite carries about 6 percent TiO_2 in solid solution. The magnetite is substantially replaced by martite. The rock gives a negative reading on the dip needle.

ILMENOHEMATITE AT THE LOUP PROSPECT

The only ilmenohematite concentration known to us is at the Loup prospect, 1.1 miles southwest of South Russell, central rectangle, Russell quadrangle, about halfway between Stockwell Pond and Little Lake. A small prospect pit was sunk here in a quartz-rich ilmenohematite aggregate in sillimanite-muscovite-microcline granite gneiss. Microcline pegmatite is associated with the "ore." The "ore" forms a small lens of no economic importance. Shaub (1949, p. 3) states that the ratio Ti : Fe is 0.134, which would be equivalent to 13.6 percent TiO_2 in the hematite. Examination of polished surfaces of the mineral shows it to be ilmenohematite. The "ore" gives a negative magnetic anomaly.

TITANIFEROUS HEMATITE AT THE GEORGIANA SLATE FARM

By B. F. Leonard

An unusual occurrence of hematite is found on the Georgiana Slate farm, Town of Russell, 0.4 mile south-southeast of Green Valley School, south-central rectangle, Russell quadrangle. The Slate farmhouse is 450 feet north of the intersection of Stammer Creek and the road from Fine to Green Valley School. Hematite crops out north of the creek and just north of the boundary between the towns of Fine and Russell, S. 52° E. from the Slate farmhouse. The hematite-bearing area is in a belt of metasedimentary rocks at or very close to the contact with muscovite-sillimanite-microcline granite gneiss. (See pl. 1.)

Banded crystalline hematite and quartz are traceable in outcrops and float for a maximum width of 22 feet and a maximum length of 90 feet. The area of best exposure measures only 20 by 30 feet. The outcrop is reddened and broken by joints and minute faults. Foliation given by alternating hematite-rich and quartz-rich layers strikes N. 5° E. and dips 15° W. The local dip is much gentler than the 30° to 35° dip generally prevailing in this neighborhood, suggesting that the hematite is localized along a roll in the foliation. A prominent lineation (grooving and axes of flat folds) trends S. 65° W. and plunges 13° SW. Joints strike N. 60° W. (locally N. 75° W.) and N. 85° E. The minute faults, spaced a fraction of an inch apart and having a throw of less than half an inch, strike N. 85° E. and dip 75° N.

The middle part of the exposure is a "ribbon rock" consisting of fine gray hematite-quartz layers one- to three-fourths of an inch thick alternating with layers of coarse, glassy quartz of similar thickness. Several quartz layers are 3 inches thick. The coarse quartz has only a trace of hematite. In contrast, the gray layers have 40 to 50 percent of fine, crystalline hematite disseminated in a fine quartz matrix. On the whole, these gray layers appear to be homogeneous, but locally their two mineral constituents are very finely lenticular. Locally, hematite from the gray layers cuts across the coarse quartz. The middle part of the exposure shows a pronounced grooving or fluting.

The upper part of the outcrop consists of vaguely banded hematite-rich and quartz-rich layers, generally about half an inch thick. Several thicker quartz-rich layers are present locally. The axes of very flat folds give a strong lineation parallel to the grooving observed in the middle part of the exposure.

The poorly exposed lower part of the outcrop is not banded but consists instead of small grains and specks of hematite disseminated in coarse quartz.

Hematite from the Slate farm is titaniferous, containing 1.9 percent TiO_2 in solid solution. A trace of magnetite accompanies the hematite. The iron oxide minerals and perhaps much of the quartz are thought to be of hypogene origin.

A separate exposure of hematite is found about 700 feet southwest of Stammer Creek Bridge. Here one finds a very poor outcrop of hematite-bearing quartz schist in a belt separate from that exposed on the Slate farm. Neither occurrence is of economic interest.

IRON SULFIDE DEPOSITS

By A. F. Buddington

DISTRIBUTION, CHARACTER, AND ORIGIN

Iron sulfides occur as accessory minerals in the skarns and locally in some of the gneisses within the metasedimentary rocks of the main igneous complex, but iron sulfide concentrations of a size and grade that warranted prospecting and mining are confined to metasedimentary rocks of the Grenville lowlands. A considerable number of these low-grade iron sulfide deposits are in a belt about 40 miles long and 3 to 4 miles wide that extends northeastward across the Antwerp, Gouverneur, Russell, and Canton quadrangles. (See fig. 2.) Only a short segment of this belt was mapped during the present study. (See pl. 1.) None of the deposits has been worked recently. Of these sulfide deposits, only the Stella mine, last worked during World War I, lies within the area of plate 1. Prucha (1953, p. 42) states that total production of the mine exceeded 500,000 tons of pyrite concentrates averaging 40 percent sulfur. A number of small deposits, most of them prospected, are also present in the mapped area.

The sulfide concentrations of the northwest Adirondacks occur almost exclusively as tabular vein deposits in belts of migmatitic garnetiferous and biotitic quartz-feldspar gneiss. The veins replace sheared zones parallel to the foliation of the country rock. Most of the pyrite deposits are near contacts of the gneiss and marble belts, or within relatively thin interbeds of gneiss in marble. Pyrrhotite may or may not accompany the pyrite. Chlorite is nearly everywhere present as a secondary mineral accompanying pyrite, and replacing the mafic minerals and also in part the quartz and feldspar. Some of the altered material is almost all chlorite and quartz.

The origin of the pyrite deposits has been thoroughly studied by Smyth (1912) and Prucha (1956) and is discussed by Miller (1926b, p. 65–67), Buddington (1934, p. 214), and others.

Smyth (1912, p. 180–181) sums up his views as follows:

> * * * it appears to the writer that four periods of pyrite formation are probable: (1) A primary precipitation of pyrite contemporaneous with the formation of the sediments; (2) a concentration of this pyrite by circulating ground waters, with the addition, perhaps, of pyrite of deep-seated origin, before the period of metamorphism; (3) but of minor importance, a recrystallization of all pyrite, accompanied perhaps by a certain amount of formation of new pyrite during metamorphism; (4) a further development and concentration of pyrite by magmatic agencies, perhaps working in combination with ground waters, * * * following the period of active igneous intrusion and metamorphism.

The rather widely developed pyritous gneisses would on this hypothesis be largely due to processes 1 to 3 and the concentrated pyrite deposits largely to process 4, which has involved addition of pyrite from external sources. Smyth has also noted the common concentration of graphite with the pyrite deposits. He believes (p. 176) that for the graphite

> * * * a history somewhat similar to that of the pyrite is indicated, but with the difference that most of the former is thought to be carbon that was original in the sediments, which has undergone some concentration and may have received minor additions from magmatic sources, while, in the case of the pyrite, the relative importance of these sources is reversed.

Goldschmidt (1922, p. 116–117) has suggested that carbon and sulfur may be introduced through the process of "carbon metasomatism" whereby CS_2 or

COS reacts with silicates rich in iron to yield graphite, pyrite, and quartz. However, iron has also been added in large quantities to these deposits. The ore is later than the granite pegmatite veins, and Smyth concludes that the agent effective in forming the pyrite concentrations (and the associated graphite) was a solution given off by granitic magma, now represented by associated granite masses and the granite pegmatite veins. We believe that Smyth's conclusions are appropriate and that the iron sulfide concentrations are mesothermal or possibly hypothermal replacement veins.

DESCRIPTION OF INDIVIDUAL DEPOSITS

STELLA MINES

The deposits at the Stella mines, about 1.3 miles northwest of Hermon, northwest rectangle, Russell quadrangle, have been described by Brinsmade (1905), Eckel (1905, p. 587–588), Vogel (1907), Smyth (1912, p. 153–182), Buddington (1917, p. 21–23), Newland (1917), Miller (1926b, p. 65–67), and Prucha (1953, p. 42–48). The sulfide deposits occur within two layers of migmatitic biotite-quartz-feldspar gneiss, locally garnetiferous. The migmatitic gneiss layers are separated from each other by a lens of metagabbro gneiss. The upper layer of migmatitic gneiss is overlain by a sheet of biotite granite gneiss, in turn overlain by an amphibolite layer and then by marble. The lower migmatitic gneiss layer is underlain by amphibolite and that, in turn, by marble. (See fig. 28.) The granite gneiss is a variety relatively rich in plagioclase. Below the amphibolite it carries porphyroblastic hornblende and a little epidote and sphene. The biotite is a green variety.

The mines are now abandoned. The following description is based on a visit to the mines by Buddington when they were working in 1918.

The upper ore vein, the Stella vein, was worked for a length of 1,100 feet along the strike and 900 feet down the dip and had an average thickness of 10 to 12 feet. The dip varied from 20° to 30°. The vein is in a migmatitic biotite-quartz-feldspar gneiss, but a short distance in the footwall there is several feet of serpentinized and tremolitic marble.

The lower zone was the one being worked in 1918. Operations were carried on through the Anna shaft. The ore here occurs as a series of veins parallel to each other and to the foliation of the enclosing gneiss; 4, or perhaps 5, such veins were worked from this shaft. They were called the Z, A, B, and C veins, going from bottom to top. Each of the four veins is from 75 to 100 feet above the other. In thickness, the Z vein is 12 to 14 feet, the A and B veins each average

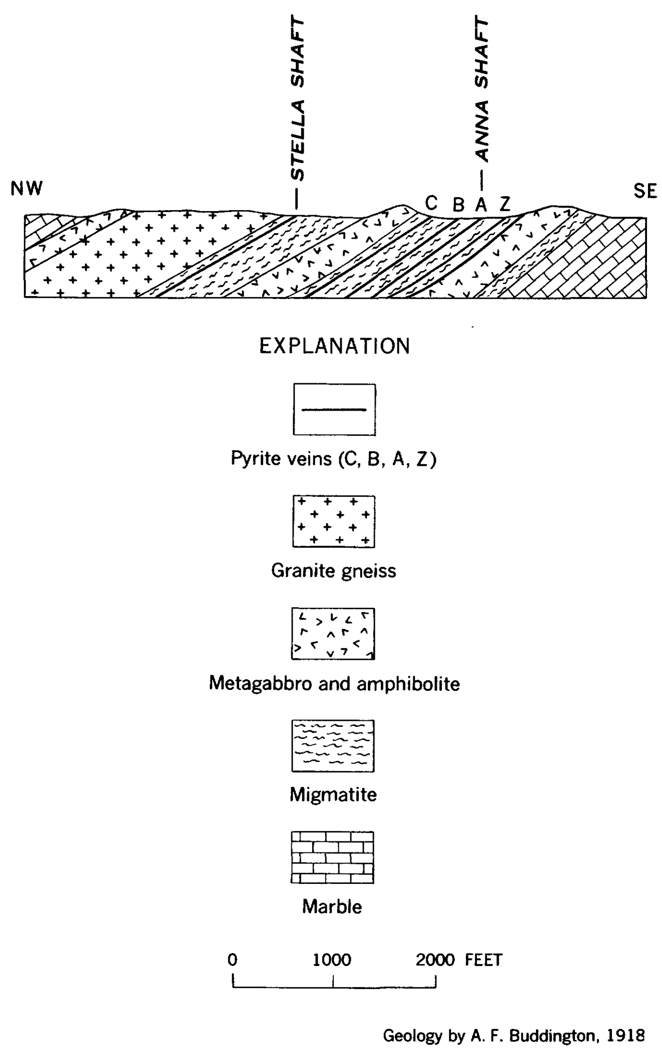

FIGURE 28.—Generalized structure section at the Stella mines, northwest rectangle, Russell quadrangle.

about 18 feet, and the C vein is 6 to 8 feet; but the individual veins vary somewhat, reaching a maximum thickness of 40 feet locally on the A and B veins.

The Anna shaft is sunk on the A vein, and the workings went down to 600 feet. The veins were developed along the strike for a length of 1,800 feet and were proved by diamond drilling to extend much farther. The strike is N. 45° E. at the southwest end of the vein, changing to N. 30° E. towards the northeast. The dip of the vein at the shaft is about 45° NW. At the northeast, the dip flattens in depth to 30°. In tracing the veins along the strike, one finds a sharp flexure that affects all the veins. The B vein at one place forks and encloses a horse of country rock. The veins diminish at the southwest, in part by becoming leaner and in part by thinning.

At the borders of the veins, there is a gradation from the typical ore to wall rock impregnated with a little pyrite; but as this gradation is usually abrupt, the veins are fairly well defined. The rock between the veins locally carries some pyrite but in general is fairly clean.

Locally, the metagabbro is pyritized and carries thin layers of granular pyrite a few inches thick, but no pyrite veins of economic significance have been found in this rock.

The ore consists of (a) a granular aggregate of disseminated pyrite, in part grown together in small bunches, in part as a close irregular network of veinlets; and (b) a soft gray alteration product, associated with considerable quartz, some feldspar, and disseminated flakes of fine graphite. Smyth has called the soft gray material that is the dominant gangue mineral a chloritic alteration product. Where exposed at the surface, this material changes to the typical green color of chlorite. Small bodies of pyrrhotite are encountered locally. The ore averaged about 20 percent sulfur and was concentrated to 40 percent sulfur.

The pyrite and chlorite for the most part have replacement relations towards the rock in which they occur.

OTHER OCCURRENCES

Pyritic zones a few inches to a foot or so thick are visible in many outcrops of garnet-biotite gneiss in the foothills between the northwest part of the Russell quadrangle and the central part of the Potsdam quadrangle.

Reed (1934, p. 62) refers to a highly pyritic layer [pyrrhotite-bearing diopside skarn], about 10 feet thick, which has been prospected on the Potsdam quadrangle, southwest rectangle, on the south side of the road, about 2.5 [2.8] miles west of Colton. He notes a similar occurrence in the river bank about halfway between Colton and Browns Bridge.

Pyrite-rich chloritic schist crops out at the north end of the hill at the Dillabaugh farm, southeast rectangle, Canton quadrangle, 1.75 miles south-southwest of Pierrepont.

Pyritic zones at Devils Elbow and the O'Brien Farm (Little River), more important for their oxidation products than for the primary sulfides, are described in the following section on supergene hematite deposits.

A molybdenite occurrence on the northwest edge of the village of Colton, southwest rectangle, Potsdam quadrangle, was trenched by the Shenango Furnace Co. during World War II. Some of the molybdenite-bearing rock contained 2 percent Mo, but the prospect was considered too small to warrant further exploration (Collord, G. L., 1947, personal communication).

ILMENITE-MAGNETITE DEPOSITS

By B. F. Leonard

DEPOSITS ASSOCIATED MAINLY WITH METAGABBRO OR AMPHIBOLITE

PARISHVILLE MAGNETITE-ILMENITE DEPOSIT

The Parishville deposit is 1.1 miles east-northeast of Parishville, east-central rectangle, Potsdam quadrangle. The northern pit is S. 58° E., 920 feet, from the intersection of the Parishville-Hopkinton road and a dirt road running north. The area about the pits is covered with thickets of small poplars and patches of briars. The deposit was briefly described by Reed (1934, p. 62), who notes that, according to an old resident, considerable quantities of iron ore were removed from this locality more than 50 years before the time of Reed's examination. According to W. G. Srodes (personal communication), the ore is titaniferous.

Magnetite, accompanied by minor amounts of ilmenite and pyrite, occurs in one or more small masses of biotitic amphibolite (metagabbro?) enclosed by pink, fine-grained hornblende-biotite granite gneiss (gneissic equivalent of hornblende-microperthite granite). The amphibolite is a normal hornblende-plagioclase gneiss, slightly biotitic, with considerable accessory sphene and apatite. Near the amphibolite, the granite gneiss also contains several percent of sphene.

Two small openings, about 60 feet apart, give poor exposures of magnetite-bearing amphibolite and the granite gneiss wall rocks. (See fig. 29.) Scattered trenches, test pits, and small dumps are found near the prospect pits. At the northern pit, a sheet of amphibolite 4 feet thick strikes N. 25° E. and dips 25° NW. Both hanging wall and footwall of the amphibolite are hornblende granite gneiss uncontaminated by anything except sphene and perhaps biotite. Most of the magnetite seems to be near the footwall of the amphibolite. The main mineralized zone was traced for about 150 feet by means of a dip needle. If the zone in which the larger pits were sunk is the same as the zone yielding a magnetic anomaly 300 feet southeast of the pits, the total length of the anomaly is about 400 feet. The anomaly has a width of 30 to 50 feet within the + 10° isoclinal. Though peak values of +90° were registered on magnetite-bearing amphibolite at the pits, the peak value for most of the anomaly ranged from +13° to +60°. The corrected null reading was +3° for areas where hornblende granite gneiss alone was present.

The deposit is reflected on the aeromagnetic map (survey of 1946) by a closer spacing of the isointensity contours, not by the presence of a separate, "closed" magnetic anomaly.

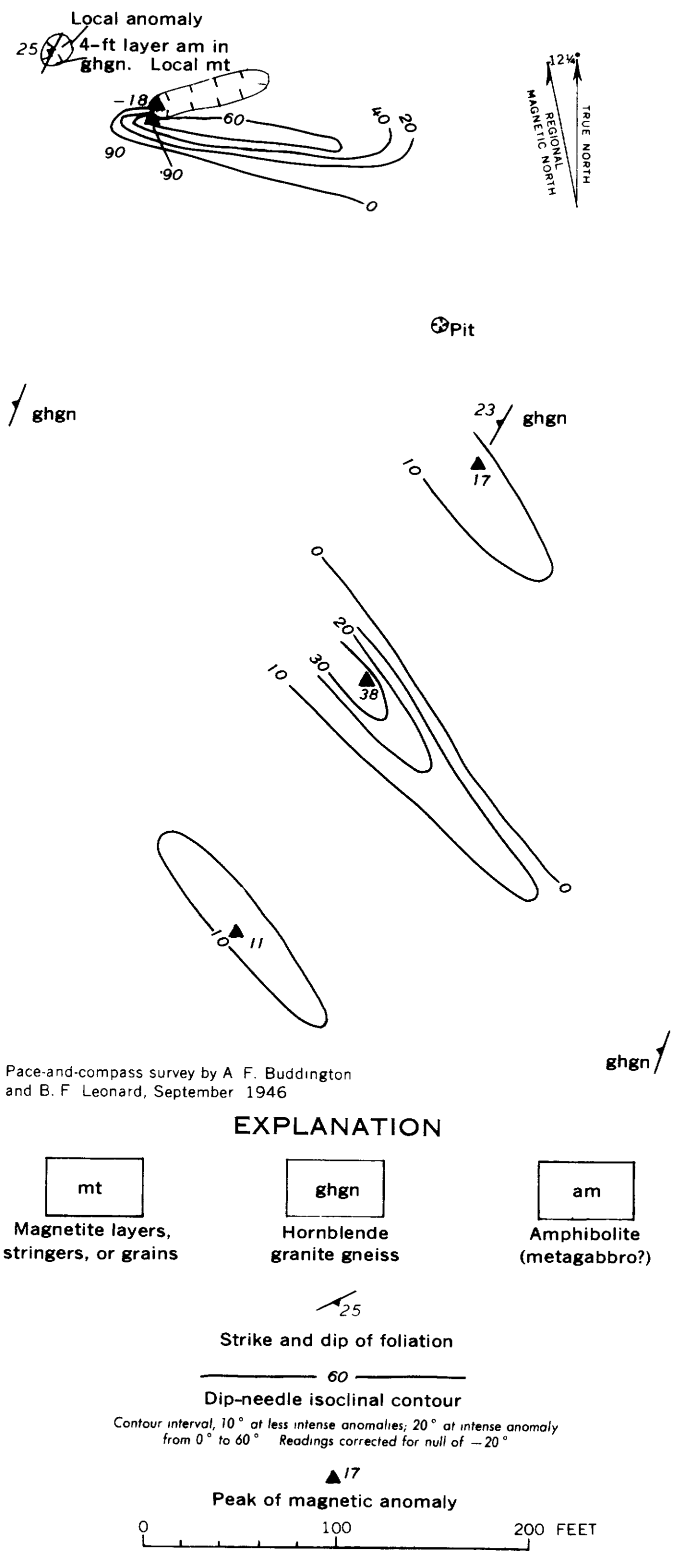

FIGURE 29.—Reconnaissance magnetic map and geologic sketch map of the Parishville deposit, east-central rectangle, Potsdam quadrangle.

The geologic and surface magnetic relations are ambiguous. Foliation in scattered outcrops of hornblende granite gneiss strikes consistently N. 20°–25° E. and dips 20°–25° NW. Magnetite-bearing amphibolite in the northern pit is enclosed conformably by the granite gneiss. Yet the trend of the magnetic anomaly is east-southeast and southeast from this pit. The amphibolite with its local concentrations of magnetite thus seems to be a dike or "pseudodike" in the granite gneiss. Nevertheless, dump specimens show granite pegmatite cutting amphibolite.

Two other interpretations of the disparate geologic and magnetic relations come to mind:

1. The magnetite-bearing amphibolite may occur as isolated lenses, each conformable with the foliation but arranged en echelon along an axis that transects the foliation.
2. The magnetite-bearing amphibolite may be a tightly folded layer, or layers, that may or may not be continuous. If the folds, whose amplitude would be 50 to 100 feet, plunged roughly north or north-northwest (parallel to the inferred axis of the Parishville anticline, pl. 1, and parallel to nearby lineations in the granite gneiss), the apparent transgressive relation of the amphibolite might be accounted for.

Clearly, the data necessary to support any one of these interpretations are not at hand and probably never will be. The small size and titaniferous character of the deposit make further prospecting very unlikely.

MICROSCOPIC FEATURES

Several polished sections of material containing more than 85 percent metallics were examined microscopically. Primary metallic minerals are magnetite, ilmenite, and pyrite. Magnetite is abundant as medium to coarse grains, some of which contain hair-thin blades of exsolved ilmenite; in terms of weight percent, this exsolved ilmenite is entirely negligible. Grains of ilmenite, either pure or containing exsolved blades of hematite, are sparingly present in some specimens. Pyrite, as scattered subhedral crystals and granules, constitutes perhaps 5 percent of the ore minerals. In these specimens, there is no evidence that the opaque minerals have been deformed.

Crystalline hematite and marcasite are clearly secondary. Hematite occurs as scattered roundish or subpolygonal aggregates, irregular areas, and veinlets. Most of the individual hematite grains in the three forms noted above are extremely fine grained; rarely, coarse blades of specularite are present in the fine-grained aggregates. Though the hematite is badly pitted, inclusions seem to be rare; they are commonly nonopaque and do not appear to be exsolution products. Hematite replaces magnetite preferentially, forming martite. Some silicates show limited replacement and "flecking" by hematite. Hematite-pyrite and hematite-marcasite relations are inconclusive.

Marcasite is usually confined to pyrite grains, which it replaces irregularly. A few marcasite veinlets fill cracks passing through magnetite, pyrite, and silicates.

Several thin sections of amphibolite and "ore" were examined. All these showed alteration of two types: (a) earthy or fine-grained crystalline hematite as flecks, patches, and veins in silicates and in magnetite; and (b) opaque white material ("leucoxene"), mainly in sphene but locally in amphibole and plagioclase. In places, the hematite is accompanied by veinlets of chlorite, or chlorite and chalcedonic quartz. Scattered flakes of stilpnomelane(?) are present.

The writer believes that the hematite is not merely secondary—younger than magnetite, ilmenite, and pyrite—but supergene. The marcasite may or may not be supergene.

The quantity of ilmenite so far observed in polished sections of typical "ore" is much too slight to account for the high titanium assays orally reported by one of the iron companies. Several percent of sphene, identified in the "ore," would contribute some Ti, as would the "leucoxene."

ORIGIN

The source of the magnetite, ilmenite, and pyrite is uncertain. If the amphibolite is indeed metagabbro, one might logically expect the rock to carry local, minor concentrations of these opaque minerals. However, no evidence has yet been found that the opaque minerals have been deformed. Moreover, the quantity of ilmenite so far observed is slight—much less than at the South Edwards prospect, for example. The Parishville deposit might, therefore, be a small, high-temperature replacement deposit related to the granite-granite gneiss series or to the pegmatites derived therefrom. Whatever its origin, the deposit has been somewhat altered by supergene processes, perhaps before the Potsdam sandstone was eroded from this area.

SOUTH EDWARDS ILMENITE-MAGNETITE PROSPECT

The South Edwards ilmenite-magnetite prospect is 0.8 mile west-northwest of South Edwards and 0.15 mile west-southwest of a right-angle bend in the Oswegatchie River, southwest rectangle, Russell quadrangle. The prospect pit is 15 feet south of the old road and 55 feet southeast of the brook. The water-filled crosscut pit is 15 feet long and 7 feet wide. The dump and nearby outcrops along the strike of the foliation show small aggregates of ilmenite-magnetite, accompanied by specks of pyrite, intergrown with the pyroxene, plagioclase, and biotite of the metagabbro. At the pit, the foliation strikes north and dips steeply west. Similar ilmenite-magnetite-bearing metagabbro crops out 350 feet east-southeast of the brook, but at this outcrop the foliation strikes N. 80°–90° W. and dips about 70° S. The metagabbro body that carries these local, slight concentrations of ilmenite-magnetite is very small. (See pl. 1.) It seems to be a south-southwest-plunging, overturned anticline enclosed by strongly foliated hornblende syenite gneiss (locally biotitic). The syenite gneiss contains sporadic layers of diopside skarn and diopsidic quartzite of the Grenville series.

No systematic dip-needle survey of the prospect was made. Maximum readings recorded on two traverses were +12° and +14°, respectively. The area was not covered by the aeromagnetic survey of 1945.

The ilmenite-magnetite deposit appears to be very small and of no consequence economically.

ORE MICROSCOPY

A typical specimen of ilmenite-magnetite-bearing metagabbro was examined under the ore microscope. About 10 percent of the rock was metallic minerals: ilmenite dominant, magnetite subordinate, and pyrite very minor. Ilmenite and magnetite appear to be contemporaneous, and pyrite slightly younger. Grains of metallic minerals range in size from medium to very fine; they are strung out in crush zones in the deformed and recrystallized rock.

The ilmenite almost everywhere contains two generations of exsolved hematite: first, relatively coarse spindles, and second, very fine spindles. Some of the coarser exsolved hematite in turn contains extremely fine bodies of exsolved ilmenite and, rarely, a nonmetallic—possibly a spinel.

Some magnetite grains are pure; others contain moderate to fine blades of exsolved ilmenite. Some of this ilmenite appears to contain traces of exsolved hematite(?).

Pyrite shows two habits: (a) coarse grains, thought to be essentially primary; and (b) very fine interstitial and veinlike material, thought to have been redistributed during or after regional metamorphism.

Very fine blades of hematite(?) appear to have been exsolved in regular patterns in some silicates.

PROSPECT NORTHWEST OF DILLON POND

A shallow shaft was sunk in rock 0.5 mile northwest of Dillon Pond, north-central rectangle, Cranberry Lake quadrangle, on the east side of hill 1445. The rock is a granitized amphibolite consisting of plagioclase (about 50 percent by weight), hornblende, augite, and a little hypersthene (totaling about 15 percent), potassic feldspar (about 20 percent), quartz (5–10 percent), accessory apatite, and disseminated magnetite and ilmenite (totaling 5–10 percent). No concentration of Fe-Ti oxides of possible economic significance was seen.

OCCURRENCES ASSOCIATED WITH ANORTHOSITIC GABBRO, CENTRAL PART OF RUSSELL QUADRANGLE

Shaub (1949, p. 7–9) and his fellow workers noted occurrences of ilmenite-magnetite associated with anorthositic gabbro in the central part of the Russell quadrangle. These occurrences are not of economic interest. The reader is referred to Shaub's report for details

DEPOSITS ASSOCIATED WITH MAFIC SYENITE GNEISS IN DIANA COMPLEX

A large reserve of low-grade ilmenite, magnetite, zircon, and zirconlike minerals is present in one part of the Diana complex. The material, of no economic interest at the time the district was mapped, was not investigated in detail. Because of its potential worth, should changing economic conditions warrant the exploitation of such deposits, the following account is presented.

DISTRIBUTION AND GENERAL CHARACTER

A belt or large lens of pyroxene syenite gneiss containing smaller lenses of shonkinite gneiss and feldspathic ultramafic gneiss occurs about one-third of the way upward, stratigraphically, from the base of the Diana complex. (See Buddington, 1948, pl. 3 and text; Prof. Paper 376, p. 46, 85; this report, p. 83.) The belt is about 12 miles long; its maximum "outcrop width" is about 1 mile. Only the northeast segment of the belt is within the area of plate 1. This segment is about 3 miles long and 1,200 to 1,500 feet wide. Within the segment, lenses of feldspathic ultramafic gneiss containing 20 to 40 percent metallic minerals (ilmenite and magnetite) range in thickness from mere films to several feet. Their length is commonly a few feet to as much as several hundred feet. Dip-needle readings over the feldspathic ultramafic gneiss have a maximum value of +45°, whereas a prevailing value over the syenite gneiss is about +6°. The northeast tip of the large positive aeromagnetic anomaly attributable to this general zone of the Diana complex shows on map 1 of Balsley, Hawkes, Buddington, and others (1946).

The feldspathic ultramafic gneiss is well exposed 0.3 mile east-southeast of the south end of Jenny Lake, northwest rectangle, Oswegatchie quadrangle; and 0.6 mile south-southwest of the railroad grade crossing 0.7 mile west-southwest of Jenny Lake, or 1.3 miles west of Kalurah. East of Jenny Lake, a 3-foot layer of feldspathic ultramafic gneiss containing 15 to 20 percent total ilmenite and magnetite is visible in porphyroclastic shonkinite gneiss, locally feldspar rich. In one place, the true thickness of the ultramafic layer is almost 6 feet. Ultramafic schlieren are present elsewhere in the outcrop. Locally, the shonkinite host has interstitial ilmenite and magnetite as its dark component, instead of pyroxene. Foliation here strikes N. 40° E. and dips northwest at a moderate angle. Part of the exposure is an agmatite of hornfels (an inclusion) cut by normal pyroxene syenite gneiss and by aplite. This agmatite separates pyroxene syenite gneiss from the overlying red hornblende syenite gneiss.

At the locality west of Kalurah, feldspathic ultramafic gneiss, overlain and underlain by mafic syenite gneiss, crops out along the southeast slope of a small, sharp hill. Here, a zone of shonkinite gneiss, 200 to 300 feet thick, contains layers of feldspathic ultramafic gneiss and normal pyroxene syenite gneiss. Individual layers of the ultramafic gneiss are generally less than 5 feet thick. The feldspathic ultramafic zones grade into footwall mafic syenite gneiss of similar texture through a distance of about 1 foot. At two places along this contact, graded "bedding" indicates that the layers are upside down, a conclusion consistent with larger features of the complex, which indicate that it is an overturned, isoclinally folded, gravity-stratified sheet.

The feldspathic ultramafic gneiss has a rusty surface. Outcrops are broken by many joints, commonly closely spaced. The foliation strikes N. 45° E. and dips 45°–50° NW. Hand specimens of the gneiss affect a compass needle weakly.

Megascopically, the fresh ultramafic gneiss is a highly deformed rock with prominent planar and linear structures given by small feldspar augen. Typically, the ultramafic gneiss is 80 to 90 percent mafic, but it contains zones that are only 40 to 60 percent mafic. The ultramafic gneiss consists chiefly of fine, granular, rather dark green pyroxene and granular, bluish metallic minerals (ilmenite and magnetite) studded with small feldspar augen showing various stages of crushing. Some augen are a single, elliptical feldspar grain, 4 to 5 mm long. Other augen consist of a relatively coarse feldspar core surrounded by a thin rim of granulated feldspar. Many augen are wholly deformed, consisting of myriad sugary white feldspar granules.

ORE MICROSCOPY, CHEMICAL COMPOSITION, AND ECONOMIC IMPLICATIONS

Polished sections of the feldspathic ultramafic gneiss show networks of highly inequigranular ilmenite and magnetite following and replacing the finer zones of granulated silicates. Coarser grains of metallic minerals are accompanied by finer grains strung out behind or sprinkled through the rock. The fabric relations are consistent with the petrographic interpretation that the whole rock has been granulated and recrystallized. The ilmenite of two samples is well twinned and devoid of intergrowths. Ilmenite in a third sample is untwinned and mainly free of intergrowths, but locally hemoilmenite is present. The hemoilmenite, which is untwinned, shows a very few regular blades of exsolved hematite in the ilmenite host. The magnetite of all three samples is generally free of intergrowths, but locally the variety ilmenomagnetite is present. The ilmenomagnetite shows a few very thin blades of exsolved ilmenite as one, two, or three sets parallel to {111} of magnetite, or two sets parallel to {100}.

Individual grains of metallic minerals are mostly between 0.3 and 0.6 mm in long dimension, though grains about 0.15 mm long are fairly abundant and grains less than 0.1 mm long are common. The grain size is less than that of magnetite from the skarn and granite gneiss ores.

The analyses in tables 43 and 44 show the chemical character of the feldspathic ultramafic gneiss and its metallic minerals. Most individual layers of such gneiss are barely of minable thickness. However, one may suppose that under favorable economic conditions mass extraction of a zone of shonkinite gneiss with its layers and lenses of ultramafic gneiss and some normal syenite gneiss might be feasible, with attendant dilution of the content of heavy minerals.

TABLE 43.—*Chemical analysis of ilmenite-magnetite-bearing feldspathic ultramafic gneiss, sample L–2186, 1.3 miles west of Kalurah, N.Y.*

[Analyst, Charlotte Warshaw, 1949]

	Weight percent		*Weight percent*
SiO_2	38.25	K_2O [1]	1.10
TiO_2	7.15	H_2O+	.24
Al_2O_3	4.31	H_2O-	.17
Fe_2O_3	11.91	ZrO_2	.46
FeO	17.71	V_2O_5	<0.05
MnO	.60	P_2O_5	1.89
MgO	3.62	CO_2	None
CaO	10.89		
Na_2O [1]	1.25		99.55

[1] Determined with flame photometer by S. M. Berthold.

Recalculation of the data given in table 44 show that the magnetic concentrate obtained in the laboratory contained about 4 percent TiO_2, and that the nonmagnetic oxide concentrate (ilmenite) was relatively high in TiO_2 for such material. Titanium dioxide in the magnetic concentrate is higher than in iron oxide concentrates now shipped from the district, but it is considerably lower than in magnetic concentrates shipped from Tahawus, N.Y. While the laboratory-made concentrates were no doubt purer than those expected under actual milling conditions, they are nevertheless a fair approximation of the character of corresponding mill products. One should bear in mind that magnetite (or ilmenomagnetite) and ilmenite (or hemoilmenite), perhaps with a trace of rutile, are the only oxide phases that have been identified microscopically. The Fe-Ti oxides of table 44, column 4, are calculated chemical constituents; only incidentally are they discrete mineral phases.

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TABLE 44.—*Chemical analyses and calculated ideal mineral composition of accessory oxide minerals from feldspathic ultramafic gneiss, sample 215, 1.3 miles west of Kalurah, N.Y.*

[Joseph J. Fahey and Angelina Vlisidis, analysts]

Chemical analyses (weight percent)			Analyses recalculated to 100 percent ideal minerals (weight percent)			
	Magnetic concentrate	Nonmagnetic concentrate		Magnetic concentrate	Nonmagnetic concentrate	Combined accessory oxide minerals
Fe_2O_3	62.86	15.57	Fe_3O_4 (magnetite)	87.12		53.28
FeO	28.09	28.58	Fe_2O_3 (hematite)	4.59	10.23	6.78
TiO_2	3.96	47.73	$FeO \cdot TiO_2$ (ilmenite)	8.29	79.30	35.87
MnO [1]	.12	n.d.	TiO_2 (rutile)		10.47	4.07
V_2O_5 [1]	.15	n.d.				
Total	95.18	91.88		100.00	100.00	100.00

Weight percent magnetic oxide concentrate in rock	12.77
Weight percent nonmagnetic oxide concentrate in rock	8.11
Total accessory oxide minerals	20.88

[1] Disregarded in recalculating ideal mineral composition.

Several minor constituents, potentially of economic interest if the Fe-Ti oxides were ever exploited, deserve mention. Two representative samples of the ultramafic gneiss contain 0.5 and 1.1 percent zircon, as well as 1.5 and 1.8 percent apatite, respectively. Buddington (1939, p. 82) reports that perovskite, though rare in the ultramafic gneiss, is an abundant accessory locally, forming 3 percent of one sample of sphene-rich ultramafic gneiss. The presence of monazite, suspected by the writer to occur in thin sections of gneiss from the locality southwest of Jenny Lake, has not been confirmed.

SUB-POTSDAM (SUPERGENE) HEMATITE DEPOSITS

By A. F. BUDDINGTON

DISTRIBUTION, CHARACTER, AND HISTORICAL IMPORTANCE

Supergene hematite bodies have been found and prospected at numerous localities in marble of the Grenville lowlands in the west half of the Russell quadrangle and on the Canton and Potsdam quadrangles. All such hematite bodies in this area are believed to be small and of no commercial value for iron ore. Indeed, with the exception of three minor deposits described below, none of the hematite veins was seen to be more than a few inches thick, individual veins were not concentrated in zones, and all appeared to die out downwards. The occurrences are in one segment of a belt of hematite deposits that extends northeastward through Jefferson and St. Lawrence counties from Antwerp and Theresa to Potsdam. This belt for the most part coincides with the belt of iron sulfide deposits previously described. (Cf. fig. 2.) All the hematite deposits are either directly associated with and found beneath a capping of Potsdam sandstone, or so situated in altitude and physiographic setting that one may reasonably infer that the Potsdam has been eroded from the vicinity very late in geologic time. In St. Lawrence County, the hematite deposits are nearly all in marble.

Hematite in the bulk of deposits of the entire belt is the red, earthy variety, and this was predominant in deposits that were mined. Massive, fine-grained, bluish, crystalline hematite and platy black specularite are present locally, with or without earthy red hematite. Much of the material is so pure and massive that it has repeatedly attracted the attention of prospectors.

Historically, hematite deposits of the lowlands are of interests because theirs were the first ores exploited for iron in the northwest Adirondacks. They were known as early as 1812 and were used in the first blast furnace built in northern New York, which was constructed at Rossie in 1813 and put into production in 1815. Most of the ore was obtained from the group of mines in the Keene-Antwerp belt and from the Shurtleff mine 3½ miles northwest of Philadelphia, N. Y. Newland (1921, p. 127) estimates the total production at roughly 2½ million tons. The last producing mine shut down in 1910. The deposits have been described by Smock (1889, p. 44–48), Winchell (1893, p. 104–108), Smyth (1894, p. 498–511), Chadwick (1920, p. 18–19), Newland (1921, p. 126–129), Buddington (1934, p. 194–209), and Smith.[50]

Within the area of plate 1, prospect pits on sub-Potsdam supergene hematite deposits have been noted at the following places:

Russell quadrangle, southwest rectangle: 1.85 miles southwest of Scotland School, on top of hill 0.25 mile northwest of road.

[50] Smith, N. L., 1943, Geological prospects in the Keene-Antwerp hematite zone, N. Y.: unpublished report prepared for New York State Dept. of Commerce.

Russell quadrangle—Continued

west-central rectangle: 1.75 miles southwest of Scotland School, west of the road; 1.2 miles west-southwest of the school; 1.6 miles northwest of the school.

northwest rectangle: 0.8 mile west of Pestle Street School, in the valley of Gibbons Brook (Devils Elbow deposit); 0.8 mile north-northwest of Hermon.

north-central rectangle: 1.15 miles northwest of Hamilton Corners, in the valley of Little River (O'Brien Farm deposit).

northeast rectangle: 0.4 mile northwest of West Pierrepont (Allen Farm).

east-central rectangle: 0.7 mile northwest of Upper District School, at the small rise on Burns Flat.

Canton quadrangle, southeast rectangle: 0.2 mile north of where West Pierrepont-Pierrepont road crosses Van Rensselaer Creek (McCollum Farm deposit); 1.9 miles south of Pierrepont, just east of old road (Dillabaugh place).

Cushing (1899, p. 26) mentions a spot, now covered with water, on the Potsdam quadrangle about 0.9 mile northwest of Browns Bridge, west-central rectangle. He writes:

> The east bank of the stream, which is here only a few rods wide, is composed of rotten pyritiferous, quartzose gneisses, much stained by hematite in their upper portion, which belong to the Grenville series. Drillings for hematite ore not far east of the river show apparent Potsdam conglomerate overlying the Grenville rocks * * *.

There are, in addition, a few occurrences of earthy hematite, too trivial to record separately, that have not been dignified by prospect pits.

ORIGIN

The following summary of the pertinent geologic relations of these ores is based on a previous report by Buddington (1934, p. 201–202). The major hematite deposits are uniformly associated with local layers of schist or gneiss of the Grenville series that have been strongly pyritized and in part strongly chloritized; the hematite occurs for the most part as a replacement of marble or of the chloritized pyritic gneiss, and in minor part as a ferruginous basal portion of the Potsdam sandstone (and breccia); relics of Potsdam sandstone (or breccia) rest as a cap over the ore deposits, or else may be inferred to have capped the deposits up to relatively recent time. It has previously been shown (Prof. Paper 376, p. 10) that the surface of the Precambrian deposits beneath the Potsdam sandstone is a low, rolling peneplain. This permits the inference that deep weathering may have preceded the deposition of the Potsdam, with the accompanying development of ferruginous solutions that penetrated deeply into the underlying bedrock. Subsequent erosion of the residues of weathering did not remove the deeper roots of the zone affected by the descending ferruginous solutions.

From the foregoing evidence and inferences, Buddington (1934, p. 201–202) followed Smyth (1894) in drawing the conclusion that the hematite deposits are the result of oxidation and weathering, in Precambrian time, of the associated pyritic schists to yield iron-rich solutions that were effective in replacing nearby limestone and chloritic schists to depths as great as 200 feet. At the time of deposition of the Potsdam sandstone, part of the iron ore is thought to have been eroded and incorporated in the basal portion of that formation. In post-Potsdam time, circulating ground waters have further accentuated some local redistribution of the iron and effected its further introduction into the basal part of the Potsdam. The action of ground water has also formed local veinlets of specularite, in part associated with quartz, calcite, or siderite; and vugs lined with quartz, siderite, and specularite. Since these may occur in the Potsdam sandstone as well as in the subjacent deposits, they must be in part of post-Potsdam age. The deposits as a whole, however, are interpreted as largely of pre-Potsdam age.

Smith (see footnote, p. 232) has advanced the hypothesis that both the pyritic mineralization of the gneiss and the hematite mineralization are of hydrothermal origin, formed by solutions rising from depth. This hypothesis appears to be based on an inferred relation between ore and a fissure system, since a few ore bodies locally appear to be oriented across the foliation of the enclosing rocks. Any fissure system present would, of course, affect the course of descending solutions just as much as ascending solutions, and the relation of ore to a fissure system certainly would not, of itself, appear to favor a hydrothermal origin. If the hematite deposits were formed by hydrothermal solutions, one might expect that they would be distributed through a considerable vertical interval. However, all the available data indicate that these hematite deposits in substantial concentration do not ever extend in depth more than a very few hundred feet below the base of the Potsdam sandstone or its equivalent peneplain level. Smith apparently also recognized this, for although he assumed that the hematite may have also been deposited at levels higher than those at which it is now found, he does not suggest that deep hematite deposits may occur; on the contrary he writes:

> Ore bodies have had such a limited extension in depth and are in localities where there is so much of the soft limestone that once the sandstone was removed the probability of ore escaping complete erosion would be slight.

The greatest depth to which hematite is known to extend in the northwest Adirondacks is at the Balmat zinc mine, where strong hematite mineralization is described by Brown (1936b, p. 341) as occurring down to the 400-foot level and spotty hematite mineralization down nearly to the 1,000-foot level.

The fact that the hematite deposits are so clearly related to the pre-Potsdam surface leads us to favor the hypothesis that they were formed by ferruginous waters descending from this surface, rather than by ascending hydrothermal solutions.

Local alteration of certain magnetite deposits (p. 77–78) has resulted in the development of hematite, chlorite, and other minerals. These small hematite bodies, resembling in many features the hematite deposits of the lowlands, are likewise thought to be supergene, though they are not directly referable to the pre-Potsdam erosion surface.

DESCRIPTION OF INDIVIDUAL DEPOSITS

By B. F. Leonard

DEVILS ELBOW DEPOSIT

The Devils Elbow hematite deposit is 2.5 miles southeast of Hermon, northwest rectangle, Russell quadrangle. The deposit was explored many years ago by means of two shallow pits on the west bank of Gibbons Brook, 750 feet S. 76° W. from the intersection of the brook and the Russell-Hermon road.

The hematite deposit is in a narrow belt of marble flanked by garnet-biotite gneiss. (See pl. 1.) A few hundred feet upslope to the northwest is a mass of pyrite, 6 feet thick and 10 to 12 feet long, on the nose of a minor fold in garnet-biotite-graphite-sillimanite gneiss. The pyrite is accompanied by quartz and chlorite; all three minerals appear to replace the gneiss.

Of the two pits that tested the hematite body, one is completely overgrown. The other, 15 feet long and about 6 feet deep, follows a joint oriented N. 40° W., 80° NE., in marble. The marble, in part serpentinized, is capped by Potsdam sandstone. The sandstone, locally calcareous, is cut by ramifying veinlets of fine crystalline hematite and chalcedony. The underlying marble shows some thicker, rootlike veinlets of hematite, but much of the marble is barren. The contact between marble and sandstone is irregular and largely obscured by debris. Some hematite blocks scattered about the pit contain abundant green chlorite.

Near the pit, float of marble contains clastic dikes of Potsdam sandstone.

A specimen of steely gray hematite streaked and flecked with earthy red hematite was examined under the ore microscope. It is a fine, foliated, distinctly colloform aggregate of extremely small grains of crystalline hematite accompanied by patches of reddish hematite. The material is vuggy in places. The folia are generally 0.01–0.02 mm to 0.1–0.2 mm thick, composed of hematite "dust" particles less than 4 microns in diameter. Locally one sees scattered crusts of coarser hematite, and rarely one finds skeletal specularite plates 0.2 to 0.6 mm long. The earthy red groundmass contains isolated minute specularite plates about 5 mm long.

The hematite was presumably derived from partial oxidation of sulfides occurring a few hundred feet upslope. The oxidation products were transported, deposited in part along strong joints in the soluble marble, and preserved until recently beneath a blanket of Potsdam sandstone. Some hematite subsequently diffused upward into the calcareous sandstone capping.

McCOLLUM FARM DEPOSIT

LOCATION, OWNERSHIP, AND EXPLORATION

The McCollum Farm hematite deposit is about 0.8 mile by road northeast of West Pierrepont and 500 feet N. 70° W. from the farmhouse. The farmhouse, just west of the road, is 0.1 mile north-northeast of the intersection of Van Rensselaer Creek and the West Pierrepont–Pierrepont road, southeast rectangle, Canton quadrangle. The farm is owned by the First National Bank, Canton, N.Y.

The deposit is apparently the one referred to by Winchell (1893, p. 104–105) as the iron location of Captain Wood. It has been explored at least three times. The earliest exploration, of unknown date, is witnessed by several old, tree-grown pits and dumps. In 1936, a Canadian firm put down a number of diamond-drill holes for the St. Lawrence Land Co. The total footage drilled at this time was 1,000 to 1,500 feet. Drilling was supplemented by test pitting. Perhaps two years later, Long and Caulfield drilled two(?) vertical holes on the deposit. Results of the successive explorations were unpromising. The accompanying map (fig. 30) shows the location of test pits and seven drill holes. An eighth drill hole is said to lie some distance southeast of the farmhouse.

J. S. Brown, geologist of the St. Joseph Lead Co., kindly contributed information on drilling done in 1936. H. E. Allen, a local resident, contributed additional information on exploration.

GEOLOGIC SETTING

The hematite deposit is in a southwest-trending belt of marble and quartzite almost surrounded by the garnet-biotite gneiss that outlines the east end of the "Pierrepont sigmoid." (See pl. 1.) The belt of marble and quartzite underlies the broad valley of VanRensselaer Creek. Perched here and there on marbles and quartzites of the valley are small, thin patches of Potsdam sandstone, in part conglomeratic. These scattered patches of sandstone are remnants of a once-continuous blanket that presumably covered the entire valley. Low hills of garnet-biotite gneiss, locally pyritic, rise northwest of the deposit. The hematite deposit is close to the contact between the marble belt and the garnet-biotite gneiss, and local relics of Potsdam remain as partial cappings directly above the hematite-bearing marble. (See fig. 30.)

HEMATITE BODIES

The McCollum Farm deposit is significant for three reasons: (a) Certain features bearing on the genesis of this and similar deposits are clearly shown here; (b) the erratic distribution and shallow depth of hematite concentrations is demonstrated by evidence in pits and drill cores, indicating that this deposit and other supergene hematite deposits of the district are likely to prove disappointing when explored; (c) hematite and barite specimens of museum quality are obtainable here.

Hematite is exposed in 1 outcrop and in 3 of the 8 pits. According to Brown (oral communication), core from holes drilled in 1936 showed leached, vuggy carbonate rocks with slight hematite stain below hematite exposed at the surface. Features observable at the surface are noted below.

An outcrop of slightly serpentinized dolomitic marble, 250 feet N. 35° E. from the main pit (No. 1, fig. 30), contains a few flecks of specularite and earthy hematite.

The main pit is about 10 feet deep. A trench 3 feet deep extends east-northeast from the pit. The only bedrock exposed in the pit is coarse white dolomitic marble, locally reddened by disseminated hematite dust. In one place, a group of barite blades replaces marble along a poorly developed fracture. The marble walls are partly slumped, partly obscured by debris, and locally channeled by solution. The trench, in large part soil covered, shows white marble. One small knot (1 foot by a few inches) of crystalline hematite and sparse barite replaces the marble.

Dump material from the trench is marble with a few pieces of hematite. Dump material from the pit is marble with blocks of hematite. The barren marble

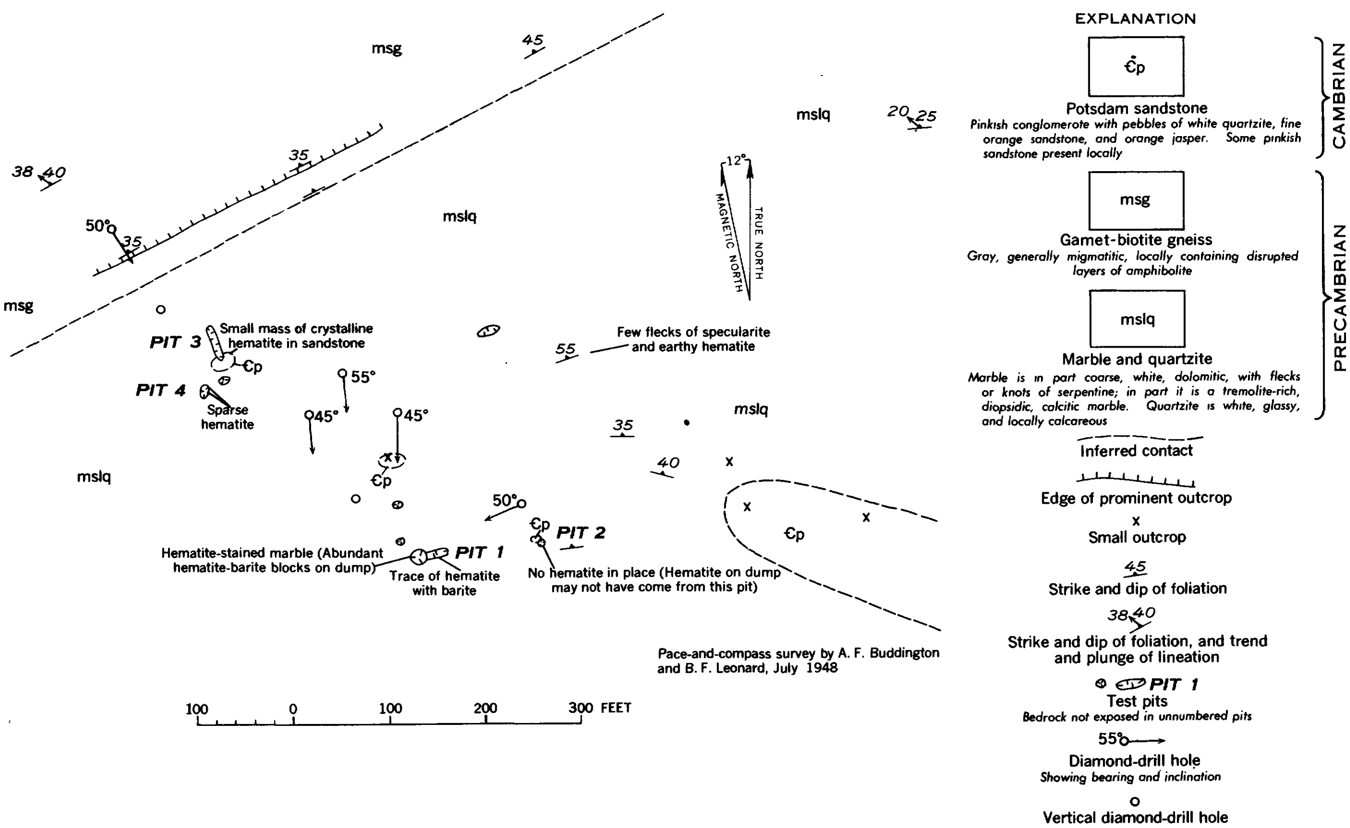

FIGURE 30.—Geologic sketch map of the McCollum Farm hematite deposit, southeast rectangle, Canton quadrangle.

is both dolomitic and calcitic. Knots of black serpentine are present locally. In places, a white talclike or claylike substance coats blocks of marble. Euhedral barite replaces the marble locally. Crystalline hematite occurs as fillings along cracks, as replacement veins apparently related to cracks, and as coarse masses replacing marble. Fine sprinklings of hematite are present in marble bordering the fracture fillings and replacement veins. Much hematite in the larger masses is accompanied by barite, which appears to be contemporaneous with it. The barite occurs as disseminated granules; as earthy knots; as euhedral blades, a few millimeters to about 3 cm long, arranged in crisscross, hit-or-miss, or slightly radiating fashion; and as fine "diabasic" intergrowths with hematite. Locally, leaching of carbonate and recrystallization(?) of barite has resulted in the formation of large euhedral barite crystals in vugs in hematite. Rarely, secondary carbonate veinlets are visible in the hematite. (For optical properties of the barite, see table 2, No. 3.)

Though the main pit now shows no crystalline hematite, most or all of the high-grade hematite blocks must have come from the pit, which was presumably picked clean in the prospecting.

Near by, an old pit (No. 2) and its dump show serpentinized marble. Several angular fragments of leached Potsdam sandstone are present on the dump, together with a few small pieces of hematite. The latter may, however, have been transported from one of the more recent pits.

Pit 3 shows closely jointed, altered marble and pink to gray, locally leached Potsdam sandstone. The sandstone, which encloses a small mass of vuggy crystalline hematite, seems to extend a bit below the level of the exposed marble, forming a "root" or incipient clastic dike. Dump specimens of marble and sandstone are veined by calcite, or contain small, rounded, more or less colloform bunches of calcite. Some leached sandstone is a mere boxwork of chalcedony veinlets and recemented sand grains. Crackling of the sandstone in place probably resulted from solution of underlying marble, followed by deposition of chalcedony and secondary calcite in the fractures.

Hematite is abundantly present on the dump, but barite is sparse. The hematite occurs as finely crystalline masses; as wavy leaves and coarse, radiating blades of specularite in a finely crystalline matrix; and as thin plates of specularite lining vugs in a fine hematite matrix. Locally, specimens showing marble in contact with sandstone are veined by specularite, the same veinlets penetrating both rocks; rosettes of specularite replace the marble. Some sandstone is merely hematite-stained and vuggy, the vugs being lined with very small quartz crystals. Elsewhere, the sandstone has fracture fillings of specularite and calcite, or of specularite and chalcedony.

An old pit (No. 4), 40 feet southwest of pit 3, shows coarse marble with a few thin veinlets and small knots of hematite, and thin veinlets of chalcedony. The crest of pit 4 is about 3 feet lower than the south end of pit 3. The topographic relations, together with the general absence of leaching and staining at pit 4, suggest that the marble here was barely affected by supergene alteration; that is, it was several feet lower than the once extensive sandstone capping (a remnant of which is exposed in pit 3), and below the local level to which hematite formation may once have extended.

The possible extension of hematite downward from the main exposures at pits 1 and 3 was tested by drill holes. (See fig. 30.) The group of inclined holes north and northeast of pit 1 should have intersected the potentially favorable zone 120 to 200 feet vertically below the pit, or at shallower depth if the downward extension of hematite were controlled by the northwest-dipping foliation of the bedrock. In cores from holes he examined, Brown (personal communication) noted hematite stains, abundant vugs, or traces of hematite and pyrite in marble at depths appropriately representing the downward projection of hematite found in pit 1. In addition, he stated that core from the 55° hole showed a foot of disseminated pyrite and hematite at a depth of 125 feet, and that core from the 50° hole northwest of pit 3 showed slight hematite stain.

It is clear from relations observable at the surface and in drill cores that the hematite bodies on the McCollum farm are individually very small, shallow, and erratically distributed. The hematite bodies, if they persist at all in depth, become progressively thinner and lower in grade. Singly and en masse, the bodies are too small to warrant further exploration.

O'BRIEN FARM (LITTLE RIVER) DEPOSIT

The O'Brien Farm hematite deposit is 1.0 mile east-southeast of North Russell and 1.15 miles northwest of Hamilton Corners, on the east bank of Little River, north-central rectangle, Russell quadrangle. The deposit is about 300 feet south of the border between the Canton and Russell quadrangles. There is no road to this deposit.

In the winter of 1942–43, the Shenango Furnace Co. explored the deposit by means of test pits and three diamond-drill holes. Their exploration showed that the deposit, though of high grade, was of very limited lateral and vertical extent. According to W. G. Srodes (personal communication), an average analysis of the ore showed the following weight percents: Fe, 68.99; P, 0.005; Mn, 0.08; SiO_2, 0.71; Al_2O_3, 0.25.

Massive gray crystalline hematite is found as stringers and masses in coarse white marble. Specularite plates are present locally in vuggy parts of the "ore"; some vugs contain parallel growths of doubly-terminated quartz crystals. Hematite is concentrated along the foliation of the marble and along a N. 40° W. joint.

Crystalline hematite is confined to a small area in white marble that may represent a small lenticular marble body in garnet-biotite gneiss, or an infolded part of a larger mass of marble and associated metasedimentary rocks. The area is shown as garnet-biotite gneiss on plate 1, but relations of the bedrock here are uncertain, owing to the thick mantle of glacial debris. Locally, Little River has cut through the debris, exposing bedrock at the hematite prospect and downstream from it.

About 200 feet downstream from the hematite prospect is an exposure of pyritic graphite-chlorite schist. This pyrite deposit has been described and sampled by Prucha (1953). A true thickness of more than 3 feet of schist, estimated to contain about 30 percent of pyrite, is visible. The footwall is weakly pyritic chlorite schist; the hanging wall is not exposed. Foliation in the schist strikes N. 40°–70° E. and dips 15°–25° NW. A distinct lineation, given by the axes of warps in the foliation, trends N. 25° W. Upstream at the hematite prospect, foliation in the marble strikes N. 75° E. and dips northward.

Between these separate exposures of Precambrian bedrock is a little thin-bedded, contorted, locally brecciated, iron-stained sandstone. The bedding strikes N. 0°–20° E. and dips about 45° W. This rock is thought to be Potsdam sandstone unconformably overlying the Precambrian. This interpretation is supported by evidence from thin sections, which show that, though most quartz of the "sandstone" is recrystallized, there are present a few well-rounded, detrital quartz grains with quartz overgrowths, and many shards of quartz. The microfabric is thus appropriate for quartzite of the Potsdam sandstone, rather than for quartzite of the Grenville series. Prucha (1953, p. 58) interpreted the sandstone as a brecciated quartz-feldspar gneiss underlying the pyritic schist but overlying the marble.

The present writers interpret the O'Brien Farm hematite deposit as a small high-grade body of crystalline hematite, economically of no consequence, formed from the supergene alteration products of the nearby pyritic schist and preserved till recently by a capping of Potsdam sandstone.

A polished section of crystalline hematite, examined by microscope, shows a tightly packed mass of tabular, wedge-shaped, and elongate diamond-shaped grains of hematite, generally 1.5–2.5 mm long but locally 4 mm long, oriented at random. Within this mass, some hematite occurs as aggregates of nearly equigranular, polygonal grains. There is no sign of exsolution products, or of any metallic except hematite. The presence of a hematite plate growing parallel to a crack in quartz, together with the projection of terminated hematite plates into the margin of a quartz grain, suggests that hematite replaces quartz.

ALLEN FARM PROSPECTS

Prospect pits in hematite are found on the H. E. Allen farm, northeast rectangle, Russell quadrangle, 0.4 mile northwest of West Pierrepont, just north of VanRensselaer Creek. (See pl. 1.) Some pits are old; others are relatively recent. Pits 1 through 6 (see fig. 31) are in the Potsdam. Pits 7 and 8 are in marble of the Grenville. All the pits are shallow, and none of them shows anything of economic value. The area is of interest for the type of hematite occurrence prospected so intensively here, and for the variety of Potsdam rocks found in narrow compass.

GEOLOGY

The geologic setting of the area is essentially the same as that of the McCollum Farm hematite deposit.

The Precambrian rocks shown on figure 31 are serpentinized marble, quartzite, and quartz-mesh silicated marble, the quartzose rocks possibly occurring as large boudins or disrupted layers in the serpentinized marble. The serpentinized marble consists of a white marble matrix studded with serpentine nodules of variable size. Disseminated serpentine granules are present locally, and here and there one finds some relics of diopside.

The quartzite is white, locally pitted, and either massive or foliated. In places, layers of fine, homogeneous quartzite, 1 to 3 feet thick, alternate with slightly coarser layers of pitted tremolite-bearing quartzite. This sequence contains some layers, a few inches thick, of crisscross. serpentinized tremolite. A few knots of coarse tremolite, about 3 inches across, are present sporadically in the massive quartzite. Rarely, a network of tremolite extends through the quartzite.

The quartz-mesh silicated marble consists of quartzose zones, 2 or 3 feet thick, alternating with irregular layers and lenses of tremolite-diopside skarn and tremolite schist generally 1 to 3 feet thick but locally more than 10 feet thick. The quartzose zones—locally medium grained, massive, and rather pure—are generally a rather thin-layered, schistose, pitted quartzite with films, patches, and knots of white tremolite, minor diopside, and local calcite. The thin-layered zones may show regular foliation, fine crumpling, or a highly irregular netlike pattern in a tremolite-rich matrix. In part, the material simulates bedded or uniformly sheared quartzite; in part, the quartz may represent isoclinally folded quartzite, later reworked; in part, the quartz may have been introduced hydrothermally, or at least extensively redistributed during metamorphism.

The silicate-rich zones in quartz-mesh silicated marble are white, massive, medium to coarse grained, consisting mainly of hit-or-miss and rosettelike tremolite, subordinate diopside, local quartz, and local relict carbonate (now calcite). In shape, these zones are generally irregular, twisted, and lenticular, commonly containing patches of crenulated quartz schist or cut by a network of quartz. Some coarse, younger tremolite has developed along fractures in the silicate masses; it is uniformly oriented at about 45° to the fracture walls.

Foliation in the Precambrian rocks strikes east-northeast to west-northwest and dips 25°–80° N. (See fig. 31.) Lineation given by crenulations in quartz schist, elongated lenses of tremolite schist, and knots of serpentine trends and plunges northwest. Its orientation is variable even in a single outcrop.

The Potsdam sandstone here is mostly breccia, though lenses of sandstone—some of mappable size—are also present. One outcrop of conglomerate was seen.

The breccia consists largely of angular quartzite fragments in a sandstone (silica-cemented quartz arenite) matrix. The fragments range in diameter from a few inches to about 2 feet. Some fragments are definitely foliated, pitted quartzite, identical in appearance with quartzite of the Grenville that crops out near by. Other fragments are of white quartzite that looks like typical massive Potsdam. Locally, one finds small fragments of jasper, or of pink-and-white bedded sandstone. The matrix for the breccia is usually sandstone, mostly brownish but locally pink or white. In places, this matrix is highly calcareous. Rarely, the matrix is finer breccia, or conglomerate. Lenses of brown sandstone, confined to single outcrops, range in thickness from a few feet to 10 feet.

The sandstone of the mappable lenses is white, pink, or interbedded pink and white. It is generally medium grained but is locally coarse. Here and there it con-

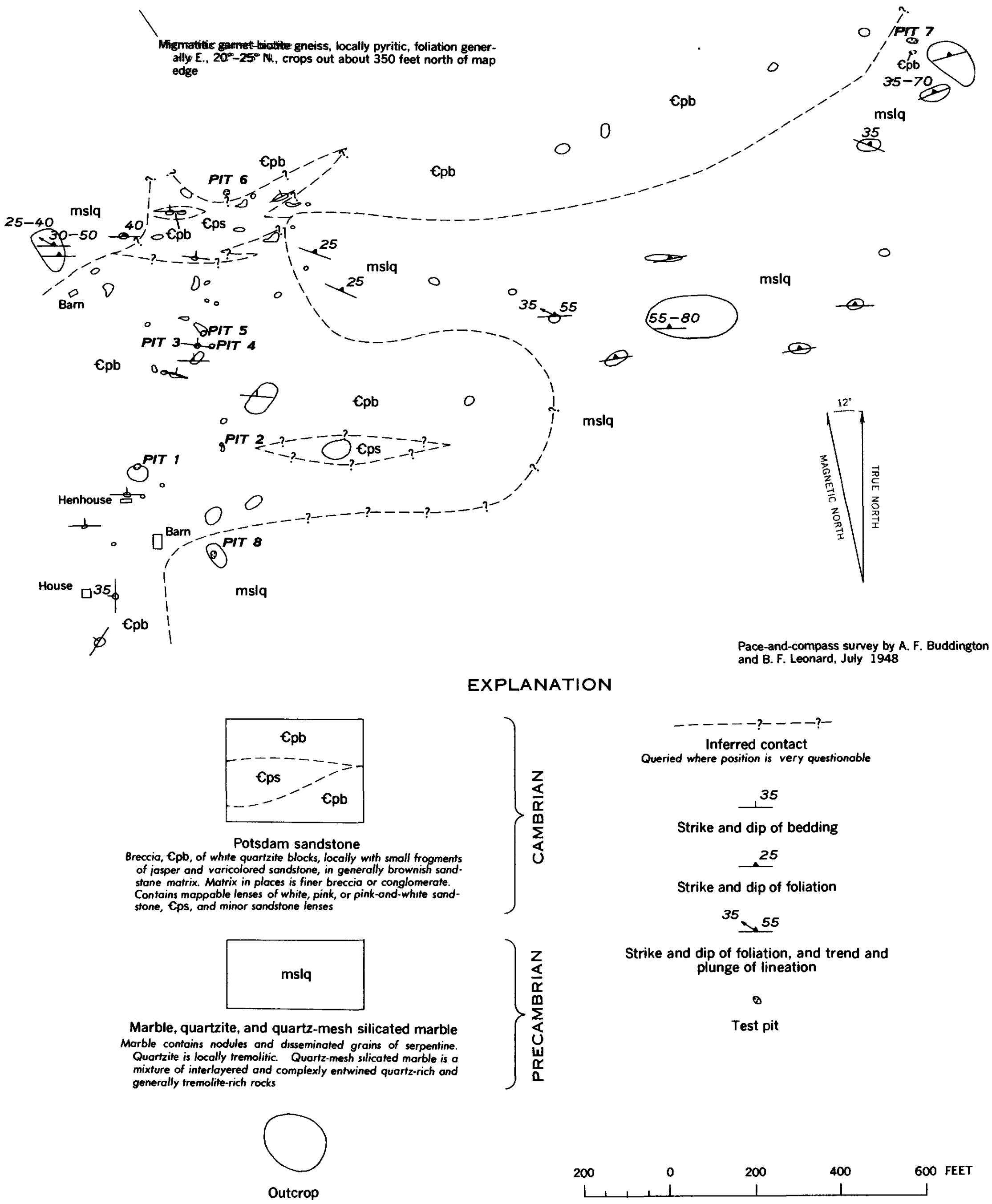

FIGURE 31.—Geologic sketch map of the Allen Farm hematite prospects, northeast rectangle, Russell quadrangle.

tains a few small fragments of jasper or white quartzite. Rarely, it encloses a thin lens of breccia.

The conglomerate consists of rounded and subangular fragments of sandstone similar to that in the Potsdam, and of quartzite similar to that in the Grenville, generally less than 2 inches across, in a purplish-pink sandstone matrix.

Most of the Potsdam, where its structure in determinable, strikes east and dips north. (See fig. 31.)

PROSPECT PITS

Pit 1, in breccia of the Potsdam containing a lens of fine brown sandstone, shows angular fragments of pitted white quartzite (presumably derived from the Grenville) in a matrix of earthy to finely crystalline hematite. One hematite veinlet, 2 mm wide, cuts a quartzite fragment. Small vugs of fine quartz crystals and limonite-stained dolomite crystals, locally present, indicate redistribution of material by ground water.

Pit 2 shows reddened sandstone as a matrix for small quartzite fragments.

Pit 3, blasted to a depth of 5 feet, shows breccia whose sandstone matrix is coated with a little powdery red hematite. Limonite stains and nests of euhedral quartz crystals are visible in places. The adjoining dump contains a few fragments of low-grade crystalline hematite "ore."

Pit 4, filled with broken blocks, shows only coarse breccia with a lens of brown sandstone.

Pit 5 shows thin films of hematite and crusts and parallel intergrowths of fine, glistening quartz in pinkish sandstone containing quartzite breccia.

Pit 7, about 5 feet deep, shows one patch of earthy red hematite coating a joint in serpentinized marble. Veinlets of fine gray crystalline hematite, 1 mm to 3 cm wide, cut marble found on the dump. Locally, these veinlets contain a little platy barite. Calcite veinlets parallel or cut the hematite. The calcite occurs as fibers perpendicular to the veinlet walls, and as slickensided plates. The slickensides perhaps developed through subsidence of the host as underlying marble was dissolved. Serpentine knots near hematite veinlets are bluish, instead of pale green.

Pit 8 shows a few facings of hematite in marble that contains serpentine nodules.

SOME NONMETALLIC MINERALS OF ECONOMIC INTEREST

By A. F. Buddington

No study has been made of the nonmetallic minerals or mineral aggregates that might have economic interest, and in 1950 there was no active mine or quarry for such materials. Only casual references can therefore be given for such rocks and minerals.

MINERAL SPECIMENS

St. Lawrence County was one of the famous areas from which mineral specimens were collected in the nineteenth century, and outstanding examples found their way into American and European museum collections. Nearly all these minerals came from the marble of the Grenville lowlands. A number of localities for good mineral specimens in marble are known in the Russell quadrangle. One of the famous old localities is that from which considerable quantities of danburite were obtained. The locality is about a third of a mile a little east and south of the intersection of the Edwards-Russell and the Dana Hill-Red School roads, west-central rectangle, Russell quadrangle. The locality has been described by Agar (1921, 1923), by Slocum (1948), and by Clark (1949). Crystals from the locality have been described by Brush and Dana (1880, p. 112). Slocum (1948) and others have described the Pierrepont tourmaline locality. Good tremolite crystals have been obtained just south of the road about two-fifths of a mile east of West Pierrepont; just east of Moores Corners (1¾ miles east of West Pierrepont); from the hillock half a mile northeast of Owens Corners; about half a mile southwest of Van House Corners; and half a mile southeast of East Road School. Diopsidic marble occurs about two-fifths of a mile south and southwest of Clarks Corners. The localities cited for tremolite and diopside are in the northeast part of the Russell quadrangle.

TREMOLITE SCHIST

Tremolite schist occurs in the marble just north of Fine on the Oswegatchie quadrangle and extends locally around the border of the granite mass to the northwest of Fine in the southern border of the Russell quadrangle. It is exposed about 0.1 mile southwest of Carr Pond. Tremolite-diopside schist was observed about 1 mile west of the Round Lake dam, southern part of Tupper Lake quadrangle, where it is associated with siliceous schist and white quartzite. Tremolite schist crops out in the Van Rennselaer Creek area, Russell and Canton quadrangles, and abundant float of coarse tremolite was seen half a mile south of Colton, southwest rectangle, Potsdam quadrangle. Tremolite, as well as talc associated with serpentine, is found within a 1-mile radius of Browns School, south-central rectangle, Potsdam quadrangle. Several of the localities cited above have been prospected for commercial talc. No tremolite deposits have been seen which are comparable to those mined in the Gouverneur talc belt.

FELDSPAR AND PEGMATITE

Dale (1935, p. 92–94) reports that three carloads of feldspar were produced from a pegmatite vein near Oswegatchie in 1923 and 1924. He has described a number of pegmatite bodies between Fine and Star Lake. An extension of this zone of pegmatite lenses is indicated by a large pegmatite lens that forms the hill 1.6 miles north of Fine, 0.6 mile northwest of Partlow Pond, south-central rectangle, Russell quadrangle. Dale (1934, p. 15) has also noted pegmatite lenses about a mile west of Hamilton Corners and near Scotts Corners, north-central rectangle, Russell quadrangle.

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INDEX

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(Continued on next card)

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