Geology and Magnetite Deposits of the Franklin Quadrangle and Part of the Hamburg Quadrangle, New Jersey

GEOLOGICAL SURVEY PROFESSIONAL PAPER 638

Geology and Magnetite Deposits of the Franklin Quadrangle and Part of the Hamburg Quadrangle, New Jersey

By DONALD R. BAKER *and* A. F. BUDDINGTON

GEOLOGICAL SURVEY PROFESSIONAL PAPER 638

A study of the rocks, structure, and magnetite deposits of an area in northern New Jersey

UNITED STATES GOVERNMENT PRINTING OFFICE, WASHINGTON : 1970

UNITED STATES DEPARTMENT OF THE INTERIOR

WALTER J. HICKEL, *Secretary*

GEOLOGICAL SURVEY

William T. Pecora, *Director*

Library of Congress catalog-card No. 76-606118

For sale by the Superintendent of Documents, U.S. Government Printing Office
Washington, D.C. 20402

CONTENTS

ILLUSTRATIONS

TABLES

GEOLOGY AND MAGNETITE DEPOSITS OF THE FRANKLIN QUADRANGLE AND PART OF THE HAMBURG QUADRANGLE, NEW JERSEY

By Donald R. Baker and A. F. Buddington

ABSTRACT

This report deals with the geology and magnetite deposits of the Precambrian rocks in the Franklin quadrangle and the southeast corner of the Hamburg quadrangle in the New Jersey Highlands. The area covered is about 65 square miles. A substantial amount of iron ore was produced in past years from mines in this area, but no mines have been operated for many years.

The Precambrian rocks include marble and other metasedimentary rocks, modified by the introduction of foreign material and the injection of granitic pegmatite. The origin of amphibolite, quartz-oligoclase gneiss, biotite-quartz-oligoclase gneiss, albite alaskite, and quartz-microcline gneiss is uncertain. All these rocks were orogenically deformed and metamorphosed before the emplacement of a pyroxene-microantiperthite series of igneous rocks, ranging in composition from syenite through granite to alaskite, and another series of igneous rocks ranging from hornblende granite to fluorite alaskite. All were emplaced as conformable sheets and phacoliths, probably during the late stages of a period of orogeny. The youngest intrusive rocks are a few diabase dikes, inferred to be of Triassic age, and rare lamprophyre dikes.

Gently folded beds of Cambrian and Ordovician age overlie the Precambrian rocks unconformably in the northwest, and folded Silurian and Devonian beds overlie the Precambrian rocks unconformably in the southeastern part of the Franklin quadrangle. These beds have not been studied for this report.

The general structure of the Precambrian rocks consists of a series of tight anticlinal and synclinal isoclinal folds slightly overturned to the northwest and broken into longitudinal blocks by great faults approximately parallel to the northeast trend of the formations. Blocks of Paleozoic formations have been locally downdropped into the Precambrian along these faults. There are a few subordinate oblique and cross faults.

Magnetite deposits occur as tabular layers in quartz-potassium feldspar gneiss, amphibolite, and, rarely, in marble or skarn. Most of the magnetite mineralization consists of low-grade disseminations. High-grade ore shoots occur within the zones of low-grade ore. Most of the ore is inferred to have replacement relationships to the country rock. Part of the iron may have resulted from metamorphic differentiation of metasedimentary rocks, but much of the initial concentration of the iron was probably directly related to igneous activity. Mining in the past has been largely confined to the high-grade ore bodies. An early attempt in the last decade of the 19th century to produce from the low-grade ore near Edison was not successful. Increasingly improved methods of mining and mineral separation may, however, eventually permit production from these low-grade ores under favorable economic conditions.

INTRODUCTION AND ACKNOWLEDGMENTS

The study of the geology and magnetite deposits of the Franklin quadrangle and part of the Hamburg quadrangle began with an investigation of the magnetite deposits near Edison, N.J., during World War II, as one of many such investigations in the United States to aid in the development of strategic minerals. After the war, the mapping was extended to include the geology of the Precambrian rocks of the Franklin quadrangle and part of the Hamburg quadrangle within which the Edison magnetite deposits lie.

The primary objective of the work was to prepare a detailed geologic map which would serve as a basis for a better understanding of the relationship of the magnetite deposits to their regional petrologic and structural setting and which would provide a more comprehensive understanding of the geology of the Precambrian highlands of New Jersey and adjacent States. This report represents the third such detailed study made by the U.S. Geological Survey in the New Jersey Highlands. The first was by Hotz (1953) on the Sterling-Ringwood magnetite district (New York and New Jersey); the second was by Sims (1953, 1958) on the Dover magnetite district.

This report is based on about 21 weeks' fieldwork during the field seasons of 1951–53. The detailed description and interpretation of the geology of the Edison area is by Baker, based on 6 weeks' fieldwork in 1952–53. The authors are indebted to Cleaves Rogers who began the study of the Edison magnetite deposits and prepared a map and preliminary report based on several weeks' fieldwork in 1945 and 1946. He also logged the drill core of several diamond-drill holes put

down by the Pittsburgh Coal & Iron Co. in the Edison district. John Howe ably assisted Buddington for about 6 weeks in 1951 in preparing a planetable map of the Edison district.

GEOGRAPHY

The area discussed (fig. 1; pl. 1) in this report is in northern New Jersey between long 74°30′ and 74°37′30″ W. and lat 41°00′ and 41°12′ N. It includes all of the Franklin quadrangle, which is underlain by Precambrian rocks, and a small part of the southeast corner of the Hamburg quadrangle. The area comprises a total of about 65 square miles and lies within the townships of Sparta, Hardiston, and Vernon in Sussex County, and Jefferson Township in Morris County. The towns of Franklin, Ogdensburg, and Stockholm are within the area; New York City is about 40 miles to the southeast.

The area is within the New Jersey Highlands, a part of the Reading Prong of the New England physiographic province. It is characterized by northeast-trending ridges separated by broad or narrow valleys. Altitudes range from 500 to 1,400 feet. The topography was formed by stream erosion and is controlled by the structure trends and lithology of the bedrock. Pleistocene glaciation has modified the preexisting fluvial erosion pattern. The area is north of the terminal moraine of the Wisconsin Glaciation. Bedrock exposures are in general very numerous in the uplands, although local areas are covered with drift. The lowlands, which are farmed, have relatively few bedrock exposures.

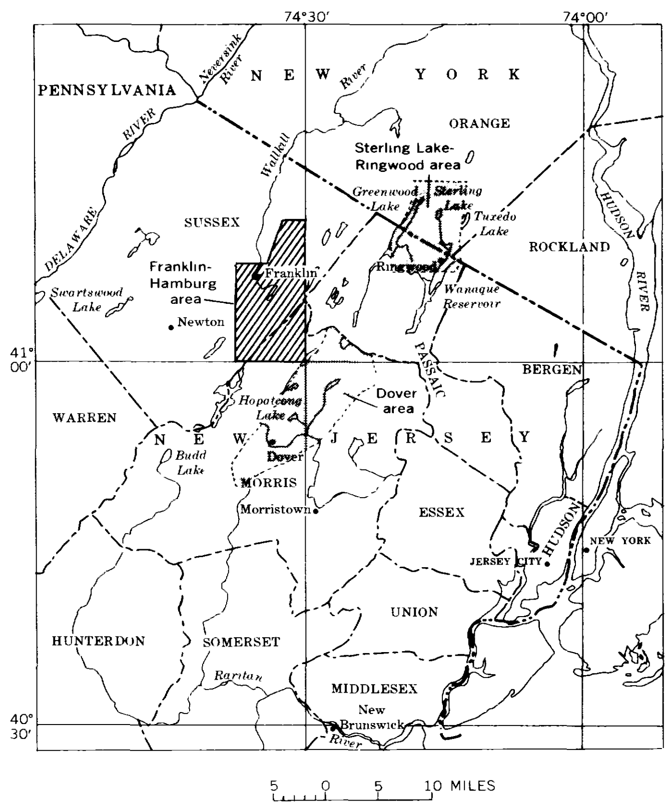

Figure 1.—Index map showing location of the Franklin-Hamburg area, Sussex and Morris Counties, N.J. The Dover district (Sims, 1953) and the Sterling Lake-Ringwood district (Hotz, 1953) are also shown. Modified from Sims, 1953.

An interesting topographic feature of the quadrangle is the "lake belt" (fig. 2) which extends southwest from Silver Lake and includes Lake Girard, Beaver Lake, Heaters Pond, Hawthorne Lake, Glen Lake, Sunset Lake, and Lake Saginaw. The depressions occupied by the lakes have been eroded in a belt of rock less resistant to weathering than that on either side. The resistant, generally slightly higher, rock on the southeast is a sheet of syenite gneiss; that to the northwest of the belt is a more homogeneous facies similar to that underlying the lake belt itself. This latter rock appears to have more schlieren, leaves, and layers of amphibolite than that to the northwest. The heterogeneity has made the zone weaker to erosion. Irregular deposition of drift, in some places aided by manmade works, has resulted in the formation of lakes within this belt.

The region as a whole is inferred to have been intensively block faulted in Triassic time and reduced to a peneplain by Cretaceous time. The present topography is the result of differential erosion of the uplifted peneplain. Many steep slopes still correspond to the locus of fault planes—as, for example, where Paleozoic rocks are faulted against the Precambrian gneisses. Where Precambrian gneisses are faulted against one another, there is generally no difference in elevation on opposite sides of the fault. The faulting has locally produced shatter zones in the rocks as much as 1,200 feet wide. These are often more easily eroded, and lowlands are formed on them. The main part of Morris Lake basin is underlain by such a shatter zone. Similarly, much of the Vernon fault zone is underlain by a drift-filled lowland.

At the beginning of the 20th century there were many farms in the highlands, but practically all of them have been abandoned. The area of the Precambrian rocks is now almost wholly covered with young forest, except around some of the lakes where there are extensive housing developments, largely for a summer population.

Nearly all parts of the area are very accessible by a network of primary or secondary roads. Few parts of the district are more than a mile from an automobile

FIGURE 2.—Aerial photograph of Beaver Lake and Lake Girard in the north-central part of the Franklin quadrangle, New Jersey. The lakes are in basins in quartz-oligoclase gneiss that trends northeast. See plate 1.

road. The New York, Susquehanna and Western [Railroad] crosses the district by way of Stockholm and Beaver Lake and serves Franklin and Ogdensburg.

PREVIOUS WORK

Geologists and engineers have made observations on the geology and magnetite deposits of the New Jersey Highlands over a long period of time. Magnetite mining began at the Mount Hope mine in the Dover district as early as 1710 and at the Ogden mine in the Edison district of the Franklin quadrangle as early as 1772. Papers were published on these mining areas as early as 1822. An excellent survey of the literature to 1909 may be found in the report on the iron mines of New Jersey by Bayley (1910). Geologic maps of the Franklin Furnace (Spencer and others, 1908) and Passaic (Darton and others, 1908) quadrangles were published in 1908 and of the Raritan (Bayley and others, 1914) quadrangle in 1914. A geologic map of

373-565 O—70——2

New Jersey on a scale of 4 miles to the inch was published in 1910–1912 (Lewis and Kümmel, 1912). Practically no systematic geologic mapping was done in the New Jersey Highlands for 30 years after the publication of these geologic folios.

In 1944 the U.S. Geological Survey began a series of detailed studies of the more important magnetite mining districts in the New Jersey Highlands. These studies included the Sterling Lake-Ringwood district (Hotz, 1953), the Dover district (Sims, 1953, 1958), the Andover mine area (Sims and Leonard, 1952), and several other areas containing magnetite bodies (Hotz, 1954). The detailed geologic mapping that has characterized most of this more recent work has revealed a complex series of folds in the metasediments and has established certain systematic relationships between younger intrusive rocks and the folds. These findings marked a distinct advance in knowledge of the geology of the region.

In addition, the geological staff of the New Jersey Zinc Co. has made a detailed study of a belt 30 miles long and 5 miles wide through the Sterling and Franklin Furnace mines (Hague and others, 1956). This work provides an excellent analysis of the type of complex isoclinal folding and faulting to which the Precambrian rocks have been subjected.

The first systematic geologic map of any part of the district was that of Wolff and Brooks (1898). This map covered much of the northern half of the area described in this report. They differentiated and mapped six rock units: (1) Edison Gneiss, characterized by richness in disseminated magnetite; (2) Losee Pond Granite, a greenish-white gneissoid binary granite with local layers of hornblendic gneiss; (3) Sand Pond Gneiss, a coarse foliated syenitic gneiss; (4) Hamburg Mountain Gneiss, a complex of gneisses of which the most prominent member is a coarse layered hornblendic gneiss which remembles phases of the Edison Gneiss, and which contains granitic facies that are probably intrusive layers; (5) undifferentiated granular and layered gneisses such as occur north and southwest of Stockholm; and (6) Franklin White Limestone. They further write (p. 439):

> As will be seen by the map, the easternmost gneisses, including the Edison belt, curve westward just north of the termination of the Losee Pond granite, so as to nearly join the Hamburg Mountain gneiss, which fact, together with their lithological identity suggests a possible anticlinal arch.

The Losee Pond Granite corresponds to the oligoclase-quartz gneiss of the present map. The Sand Pond Gneiss corresponds to the hornblende syenite gneiss of our map. The syenite gneiss also forms part of the Hamburg Mountain and Edison Gneisses as mapped by Wolff and Brooks. The anticlinal structure suggested by them is confirmed by the present work.

Wolff had earlier (1894) published the first systematic geologic map of a fold structure (the Hibernia anticline) in the New Jersey Highlands. Very little amplification of structural data was made, however, in the numerous publications on the geology of the Precambrian of the New Jersey Highlands for the next half century.

In 1908 the geologic folios of the Passaic quadrangle by Darton and others and of the Franklin furnace quadrangle by Spencer and others were published. Most of the area herein described lies within the Franklin Furnace quadrangle. In both quadrangles the following units were defined and mapped: Franklin Limestone, a white marble, locally containing disseminated silicates or silicate nodules, amphibolite lenses, pegmatite veins, and other intrusive bodies; Pochuck Gneiss, a dark granular gneiss composed of hornblende, pyroxene, oligoclase, and magnetite; Losee Gneiss, a white granitoid gneiss composed of oligoclase, quartz, and in places orthoclase, pyroxene, hornblende and biotite; Byram Gneiss, a gray granitoid gneiss composed of microcline, microperthite, quartz, hornblende or pyroxene, and in places mica; and pegmatite, coarse, obscurely foliated granite, mainly quartz and feldspar. Spencer and Bayley considered part of the Pochuck Gneiss, the Losee and Byram Gneisses, and the pegmatites as of primary intrusive igneous origin. In 1934 the U.S. Geological Survey (Wilmarth, 1938, p. 1653) adopted the name Pickering Gneiss for the Precambrian sedimentary rocks associated with Franklin Limestone. The Pickering Gneiss is described as a medium-grained quartz-feldspar-mica rock, usually light colored, with graphite-bearing beds an important feature in the Phoenixville and Honeybrook quadrangles of Pennsylvania. Bayley (1941) included dark gneiss layers, formerly mapped with the Pochuck Gneiss, as part of the Pickering Gneiss where these dark layers were thought to be of metasedimentary origin. This assignment by Bayley restricted the true Pochuck Gneiss to metamorphosed mafic igneous rocks.

The report by Hague and others (1956) retains the names Losee and Byram Gneiss, but, the former is restricted to quartz-oligoclase gneiss and the latter, to hornblende granite and related facies. Their results are noteworthy in that they give for the first time (p. 457) some specific evidence that part of the amphibolite of the Precambrian gneisses may be of volcanic origin.

Detailed areal mapping has permitted closer discrimination of rock units than had been possible on the older, smaller scale maps. It was found that the Pochuck, Losee, and Byram Formations, as previously mapped in the Franklin Furnace and Raritan quadrangles (Spencer and others, 1908; Bayley and others, 1914) each contained mappable units of diverse character, including rocks of metasedimentary, igneous, and mixed origin. Thus, Hotz (1953), Sims and Leonard (1952), and Sims (1953) found the terms Pochuck, Losee, Byram, and Pickering Gneisses inadequate and unsatisfactory and used petrographic names for the rock units. Similarly, rocks previously described and mapped in the Franklin Furnace quadrangle as the Byram, Losee, and Pochuck Gneisses have been found by us to include members of diverse origins. Consequently, we have also used descriptive petrographic names for the rock units.

ROCKS OF METASEDIMENTARY ORIGIN

Metasedimentary rocks are the oldest in the district; in the highlands they generally constitute 10–25 percent of the bedrock. They occur as layers and lenses, as complex folded isoclinal masses, and as concordant inclusions in igneous rocks or orthogneiss. They are so intensely metamorphosed that primary textures and structures have been obliterated. One cannot be certain whether some gneiss is of metasedimentary, metasomatic, or metaigneous origin. This is particularly true for certain quartz-oligoclase gneiss and amphibolites. The interpretation of metasedimentary origin is based on analogy with the composition of known sediments and with rocks demonstrably of metasedimentary origin.

MARBLE OR METALIMESTONE

There are two belts of metalimestone or marble (pl. 1). One passes through the Sterling and Franklin mines; the other is a narrower belt to the northwest.

The metalimestone has been termed "white limestone" in some of the older literature to distinguish it from the blue limestone of the Paleozoic Kittatinny Formation. At one time it was thought that the white limestone was a metamorphosed facies of the blue limestone, but subsequent work has proven conclusively that the white limestone is of Precambrian age, and that Hardyston Quartzite rests unconformably upon it. Where the limestone of the Kittatinny Formation is downfaulted against the Precambrian metalimestone, there may be locally a pseudotransition zone in which the Paleozoic limestone is sheared and bleached so as to resemble the similarly sheared metalimestone. This may be seen in the fault zone about 0.7 mile east of the high school at Franklin.

The marble (or metalimestone) has not been studied in detail by us, and the following description is abstracted from the report of Hague and others (1956, p. 437–441). The metalimestone of both belts is mapped as Franklin Limestone in the Franklin Furnace folio (Spencer and others, 1908). Hague and others, however, refer to the larger belt as the Franklin band of marble and the narrower northwestern belt, as the Wildcat band of marble. The latter is inferred to lie about 800–1000 feet stratigraphically above the Franklin marble band in this district.

They describe the rock as a white to gray coarse to locally fine-grained crystalline marble, which varies from a slightly impure calcite rock to almost pure dolomite. Layers of low-magnesia rock commonly alternate with layers of dolomite or dolomite marble. Conformable dolomite layers are present locally and may represent original bedding, but the usual irregularity of the dolomite distribution strongly suggests hydrothermal dolomitization. The marble usually carries a little disseminated graphite and silicate. Phlogopite, clinopyroxene, and tremolite are quite common. Locally, chondrodite or norbergite, pyrrhotite, pyrite, sphalerite, scapolite, and quartz are also found.

The marble belts contain many lenses of pegmatite and granite, and, locally, lenses of quartz-rich gneiss are inferred to be metamorphosed sandy beds.

In the Franklin marble belt is a layer of gneiss which Hague and others (1956) have traced discontinuously from near the Sterling mine to about 5 miles northeast of Franklin. It consists of feldspathic quartzite, biotite gneiss, microcline gneiss, and local hornblende gneiss. The layer is reported to be 100 feet thick or less at Franklin and 800 feet thick in the area to the northeast. It has been called the Median Gneiss by Hague and others (1956) because it lies within the marble belt.

Extensive plastic flow of the marble is indicated by isolated small thin lenses or fragments of dark gneiss which, though they may be a few feet to many feet apart, outline a skeletal fabric of flowage folds and are themselves boudins.

QUARTZ-POTASSIUM FELDSPAR AND ASSOCIATED GNEISS

Quartz-potassium feldspar gneiss forms two belts (pl. 1) on opposite flanks of the Beaver Lake anticline. The eastern belt starts at the fault plane about 2 miles southwest of Edison and extends northeast to a little beyond Summit Lake, where it is offset by a fault. The belt reappears on the other side of the fault about 1 mile northeast and extends north for about 2 miles, where it passes into quartz-microcline gneiss. The western, or Sparta Mountain, belt starts at the fault north of Lake Saginaw and extends southwest to beyond the border of the area. These two belts were probably originally continuous around the nose of the anticline north of Lake Wildwood, where they are now either cut out, or replaced by, the quartz-microcline gneiss. The epidotic gneiss, which occurs as a lens in the eastern belt of quartz-potassium feldspar gneiss, indicates the general zone where the quartz-potassium feldspar gneiss should be located along the northwest flank of the anticline. Whether the quartz-potassium feldspar gneiss is absent on the northwest flank of the anticline because of thinning by flowage deformation, because of cutout or replacement by granitic or syenitic intrusions, or because of nondeposition of original sedimentary facies is unknown.

The quartz-potassium feldspar gneiss is of economic importance because it is the host rock for the magnetite mineralization in the Edison area and the magnetite concentrations at the Bunker and Sherman mines.

The quartz-potassium feldspar gneiss consists of garnetiferous, sillimanitic, and biotitic quartz-microcline varieties. The gneiss is often seamed with quartz-microcline granite pegmatite and quartz-microcline gneiss layers. Coarse granite pegmatite lenses are also present. Where magnetite mineralization exceeds about 10 percent, the microcline is usually recrystallized to an untwinned variety of potassium feldspar. In the footwall zone of the Edison deposits is a biotitic quartz-microcline gneiss that has a schlierenlike structure, as though resulting from granitization of biotite-rich quartz-feldspar gneiss. This could be a metamorphic structure, however. A detailed description of the magnetite-enriched gneiss will be found in a discussion of the ore deposits. The modes of representative samples of the gneisses are given in table 1. Accessory minerals include apatite, zircon, and, rarely, monazite, spinel, or corundum. A noteworthy feature of the gneisses is the relative paucity of plagioclase and the unusual richness in quartz.

TABLE 1.—*Modes of quartz–potassium feldspar gneiss*

	419	421	146	154	271	276
Microcline	38.5	52	40.0	17.1	50.0	26.4
Quartz	43	44	49.3	66.5	35.1	52.0
Plagioclase	9	3				
Oxides	6	Trace	5.0	3.3	6.9	5.5
Sillimanite	1.5		1.0	7.8		
Garnet	2	1		5.1		12.2
Biotite			.2		7.6	2.0
Sericite			4.5			
Accessory				.2	.4	1.9

419. Garnetiferous sillimanitic quartz-microline gneiss, sillimanite partly altered to chlorite. East side of northeast end of Sparta Mountain, Newton East quadrangle.
421. Garnetiferous quartz-microcline gneiss, plagioclase almost wholly sericitized. Just east of triangulation station, Sparta Mountain, Newton East quadrangle.
146. Sillimanitic quartz-microcline gneiss, about one-half mile northeast of Edison. Sericite secondary after plagioclase.
154. Garnetiferous sillimanitic feldspathic quartz gneiss 1,070 ft north of road forks at Edison.
271. Biotitic quartz-microcline gneiss, 2,000 ft north-northeast of Mahola.
276. Garnetiferous quartz-microcline gneiss, west of Mahola.

PYROXENIC GNEISS

Pyroxenic gneiss forms the keel and trough of a long synclinal belt from the southwest corner of the Franklin quadrangle through Pine Swamp, Ryker Lake, and Lake Stockholm. The same series of rocks is continuous northeast a little east of Stockholm as the southeast limb of a syncline. About a mile northeast of Stockholm (just beyond the limits of the area), the formation continues around the nose of a tight anticline, then extends southwest along the southeast flank of the anticline to the Reservoir fault. Another belt occurs in the eastern part of Bowling Green Mountain and extends southwest.

The pyroxenic gneiss is prevailingly green to gray in dark hues. The mineral compositions of representative samples are given in table 2. The predominant rock is a pyroxenic quartz-plagioclase gneiss. The pyroxene is a clinopyroxene and the amount varies widely and may give rise to layering; in much of the rock, however, pyroxene forms only a few percent. Quartz may range from 0 to 40 percent, but commonly is from 20 to 40 percent. Apatite and iron oxides are nearly always present, and sphene is common. Zircon is sporadically present. Microcline is slightly perthitic.

The pyroxenic series of rocks that form the keel of the pinched syncline southwest from Ryker Lake for 6.5 miles through Pine Swamp to beyond the border of the area include rocks that have the appearance of schlieren gneiss consisting of granitic material contaminated with pyroxene gneiss layers. In addition, pyroxene-plagioclase gneiss, in part seamed with pegmatite, and local layers of pyroxene-plagioclase granulite and amphibolite are present. A common gneiss of this belt has 50–70 percent plagioclase, 20–40 percent pyroxene, and accessory hornblende, oxides, sphene, and apatite. In other facies, quartz may range from a few to 30 percent. Apatite often forms as much as 1 percent of the gneiss. Granitic facies may have microcline in substantial amounts, and hornblende, in place of pyroxene. Much of this belt is obscured by drift, and the amount of true granitic material present is not known. There are sparse layers of migmatitic biotite gneiss.

The pyroxenic gneiss north of Ryker Lake occupies the trough of a syncline. Pyroxene-plagioclase and pyroxene-scapolite gneisses are the most important types in that area. Layered character may be well developed locally because of a variation in the percent of

TABLE 2.—*Modes of pyroxenic gneiss*

	1	2	3	4	5	6	7	8	9	10	11	12	13
Clinopyroxene	42.6	10.8	24.0	26.8	5.3	1.8	14.4	4.5	1.1	4.6	0.4	5	9
Hornblende	1.9		2.8	1.2		12.0	6.4				10.7		
Magnetite	1.5	1.3	1.3		.9	1.0	4.0		3.8	1.0	1.7		
Sphene		4.9	.9	.6				.6	.3	1.2			
Apatite	1.0	.4	1.3	.2	1.0	.5	.9	.6	.2	.2	.7	1	
Plagioclase	53.0		69.7	55.0	65.8	39.6	74.3	58.1	51.7	57.7	55.0		
Microcline				1.0		24.8		7.2	2.6	1.4	10.0	40	18
Quartz				14.8	27.0	20.2		29.0	40.0	33.9	21.5	50	49
Scapolite		82.6											
Mica				.4									
Zircon						0.1			.3				
Chlorite and epidote alteration												4	4
Sericitized plagioclase													20

1. 0.5 mile east of Ryker Lake.
2. North of Ryker Lake.
3. 0.6 mile northwest of Camp Mogiska.
4. 1,100 ft southeast of Sparta Lake.
5. 2,000 ft west of Woodport triangulation station.
6. 2,200 ft west of Woodport triangulation station.
7. 0.85 mile east of Stockholm, Newfoundland quadrangle.
8. 100 yd northwest of north end of Oak Ridge Reservoir.
9. 0.3 mile south-southwest of north end of Oak Ridge Reservoir, Newfoundland quadrangle.
10. 0.8 mile west-southwest of south end of Dunker Pond, Newfoundland quadrangle.
11. 0.6 mile north-northeast of Russia, west of road.
12. 4,700 ft southwest of Petersburg.
13. 1,400 ft west of Petersburg.

pyroxene, or it may result from the occurrence of scapolite (giving rise to a lighter color) in place of plagioclase. The accessory potassium feldspar is in part untwinned. Green pyroxene granite, pink hornblende granite, and pegmatite intrude these rocks.

Pyroxene skarn is also present in the hills north of Ryker Lake, and very locally elsewhere within or near the pyroxene gneiss. There is a little feldspathic pyroxene skarn about 0.55 mile north of the west end of Silver Lake. A few small magnetite concentrations are known in the skarns and pyroxene gneiss.

Adjacent to the southwest arm of Oak Ridge Reservoir is a wedge of rocks in which pyroxenic gneiss is characteristic. The pyroxenic rocks include clinopyroxene-plagioclase gneiss, local beds of feldspathic pyroxene skarn, clinopyroxene-clinozoisite-plagioclase granulite, and biotitic hornblende-pyroxene-plagioclase gneiss. There are also local masses of actinolitic clinozoisite granulite with accessory plagioclase, sphene, apatite, and calcite. Another local facies is an intensely saussuritized plagioclase granulite that contains epidote porphyroblasts. Some of the hornblende-plagioclase gneiss or amphibolite has accessory quartz. Alaskite and granite sheets and granite pegmatite veins also occur.

Pyroxene and hornblendic rocks occur in a minor synclinal structure about a mile northeast of Edison. These rocks are largely granitic, but are strongly contaminated with schlieren of pyroxenic and hornblendic gneiss and disseminated pyroxene or hornblende. Amphibolite and pyroxene gneiss layers and granite pegmatic veins are also present.

The pyroxenic gneiss near Bowling Green Mountain is a clinopyroxenic quartz-feldspar gneiss of medium grain. The feldspars may be exclusively microcline, or both plagioclase and microcline. In the latter, the plagioclase is largely altered to sericitic aggregates. The pyroxene of both types is partly altered to aggregates of chlorite and epidote with scattered relics of amphibole. Locally, there is biotitic hornblende-quartz-plagioclase gneiss. Sphene, apatite, and magnetite are usually present as accessory minerals.

GARNETIFEROUS GNEISS AND AMPHIBOLITE

Northwest of Lake Stockholm is a small crescent-shaped area of heterogeneous gneiss, consisting of interbedded garnetiferous biotite-quartz-feldspar gneiss and amphibolite, sheets of granite, and locally, pyroxene-plagioclase gneiss. Some of the amphibolite has porphyroblastic plagioclase, and some of the garnetiferous biotitic quartz-feldspar gneiss has several percent magnetite.

HYPERSTHENE-QUARTZ-OLIGOCLASE GNEISS

Hypersthene-quartz-oligoclase gneiss is the predominant rock in a large belt northwest of Stockholm and northeast through Canistear Reservoir (Newfoundland quadrangle). The belt occupies part of the Ryker Lake syncline. Rock of this character also occurs as a part of the metasedimentary gneiss shown as biotite-quartz-feldspar and associated gneiss on the nose of a northeast-plunging anticline through Mount Paul.

The gneiss is white to light buff or green and commonly medium grained, though locally coarse. The hypersthene and biotite may be homogeneously distributed through the gneiss, but often are concentrated on foliation surfaces.

The hypersthene-quartz-oligoclase gneiss of the Ryker Lake syncline is much more uniform than is

TABLE 3.—*Modes of hypersthene-quartz-plagioclase gneiss*

	1663	1591	1627	1651	1675	1635	1315	1350
Quartz	18.0	22	21	24	23		30.8	15.6
Plagioclase	67.0	63.6	66	64	63	55.5	58.7	65.4
Potassium feldspar	6.3	3.5		4.5	6			5.5
Hypersthene	4.3	4.5	9	7	3.5	25	4.7	9.7
Biotite	1.4	0.5	3		Trace	19	4.4	.7
Hornblende	2.1	4.5			4.5			
Magnetite	.4	1.0	.6		Trace		1.3	2.6
Apatite	.3	.5	.6		Trace	.5	.1	.2
Graphite	.2			.3				.3

1663. East border of Franklin quad., 1.4 miles north of Pacock Brook.
1591. 2,100 ft south of southeastern part of Canistear Reservoir, Newfoundland quadrangle.
1627. 1.1 miles north of Stockholm.
1651. About 2.0 miles north-northeast of Stockholm, 1,350 ft south of old road junction.
1675. West of north end of Canistear Reservoir, Wawayanda quadrangle.
1635. About 1 mile northeast of Stockholm on west side of Pacock Brook.
1315. 1,100 ft northwest of lake southwest of Mount Paul.
1350. 3,000 ft north of Russia.

normal for rocks inferred to be of metasedimentary origin. It contains a few local sheets of hornblende granite and alaskite gneiss. Granite pegmatite veins are common, and, locally, there are intercalated amphibolites.

Representative mineral compositions of the gneiss are given in table 3. All the gneiss carries a few percent hypersthene (3–5 percent). Additional mafic minerals may be brown biotite (0.5–3 percent) or green hornblende (2–5 percent). Magnetite and apatite are always present to the extent of a few tenths of a percent. Oligoclase usually forms 63–70 percent of the gneiss, and slightly microperthitic microcline commonly forms 3–6 percent, but may be absent locally. Quartz is normally about 20 ±3 percent. Locally, though rarely, there are mafic beds of plagioclase schist in which hypersthene and biotite may each form about 20 percent. The hypersthene is usually somewhat altered. A distinctive feature of much of the gneiss is the presence of disseminated flakes of crystalline graphite. These flakes may form only a very few tenths of a percent of the rock, but they are easily recognized on careful inspection.

The chemical analysis of a specimen of typical hypersthene-quartz-oligoclase gneiss is given in table 4. The potassium feldspar is slightly microperthitic and has an indistinct microcline twinning. A few grains of oligoclase show a little exsolved intergrowth of potassium feldspar. The hypersthene is partly altered to a shreddy unidentified aggregate. The carbon is present as disseminated crystalline flakes of graphite. The rock is medium grained and greenish to buff.

BIOTITE-QUARTZ-FELDSPAR AND ASSOCIATED GNEISS

Quartz-microcline gneiss in which microcline is the predominant feldspar and quartz-oligoclase gneiss in which oligoclase is the predominant feldspar have been previously described. There is also quartz-feldspar gneiss in which potassium feldspar and oligoclase are each present as major constituents.

Typically, these rocks are grayish white to reddish buff on fresh surfaces; they are medium or coarse to medium grained, but fine- to medium-grained facies are present. The plagioclase is oligoclase and is commonly partly sericitized. The potassium feldspar is usually microcline but may be a monoclinic variety. There may be a slight microperthitic intergrowth of plagioclase. Subordinate accessory minerals include almandite, hornblende, ilmenomagnetite, ilmenite,

TABLE 4.—*Chemical analysis, norm, and approximate mode of hypersthene-quartz-oligoclase gneiss*

[Analyst, E. H. Oslund. Collected from just north of the old road 0.5 mile northwest of where Pacock Brook enters Canistear Reservoir, Newfoundland quadrangle]

Chemical analysis			
SiO_2	67.69	TiO_2	0.48
Al_2O_3	15.99	P_2O_5	.16
Fe_2O_3	.64	MnO	.05
FeO	2.42	BaO	.16
MgO	1.16	S	.04
CaO	2.88	C	.14
Na_2O	4.64		
K_2O	2.86		99.84
H_2O+	.35	Less O=S	.01
H_2O-	.14		
CO_2	.04	Total	99.83
Norm			
Quartz	16.86	Hypersthene	4.94
Orthoclase	22.80	Magnetite	.93
Albite	39.30	Ilmenite	.91
Anorthite	10.84	Apatite	.34
Diopside	2.29		
Approximate mode			
Plagioclase	67.0	Biotite	1.4
Microcline	6.3	Magnetite and ilmenite	.4
Quartz	18.0	Apatite	.3
Hypersthene	4.3	Graphite	.2
Hornblende	2.1		

rutile, apatite, zircon, hypersthene, and allanite. The fabric is crystalloblastic.

A belt of such gneiss extends from the Vernon fault, just north of Pine Swamp, northeast to just south of Edison. This belt is described in detail in the discussion of the magnetite deposits of the Edison area. The mean compositions of two facies near Edison are given in table 5.

Rusty-weathering pyritic graphitic varieties of the biotite-quartz-feldspar gneiss were occasionally observed. Such varieties are characterized by a richness in quartz. Typical mineral compositions are given in table 6. Similar graphitic gneiss near Oak Ridge Reservoir has been prospected by a few pits. This gneiss is associated with pegmatite seams that also carry coarse graphite. A similar graphite-bearing bed in biotite quartz-feldspar gneiss occurs about 700 yards south-southeast of Big Springs (northwest corner of Franklin quadrangle). This zone can be traced for 1,000 yards southwest.

Garnetiferous quartz-plagioclase gneiss, often of migmatitic character with granite pegmatite seams, is also associated with the biotite-quartz-feldspar gneiss. This lithologic type is interlayered in the biotitic quartz-oligoclase gneiss in the area of Mount Paul and is the major rock type of a belt about 500 yards southeast of Big Springs. It also occurs with the

TABLE 5.—*Modes of biotite-quartz-feldspar gneiss*

	A	B
Quartz	27.4	23.9
Plagioclase	38.9	31.9
Potassium feldspar	18.5	27.3
Biotite	11.3	10.9
Garnet	Local trace	Present
Accessories	3.9	6.0

A. Mean of 7 thin sections of biotite-quartz-feldspar gneiss in southwestern part of belt southwest of Edison.
B. Mean of 11 thin sections of biotite-quartz-feldspar gneiss (usually with a few garnets) from northeastern part of belt southwest of Edison where there is interdigitation with quartz-microcline gneiss.

TABLE. 6—*Modes of graphitic biotite-quartz-plagioclase gneiss*

	1236	1309	2503
Quartz	46.7	63.6	42
Plagioclase	38.6	16.6	41
Potassium feldspar	6.0	11.2	------
Graphite	1.6	4.4	.7
Biotite	1.7	2.0	16
Pyrite	3.2	1.1	------
Apatite		1.1	.3

1236. 3,200 ft northeast of southwest end of Newark Reservoir, about 250 yd from water.
1309. 0.6 mile northwest of Russia.
2503. About 700 yd south-southeast of Big Springs.

TABLE 7.—*Modes of garnetiferous biotitic quartz-feldspar gneiss*

	1314	1313	2482–a	2504
Quartz	34.5	56.5	40	24
Plagioclase	52.4	12.2	43	53
Potassium feldspar	5.9	23.1	1.5	2
Biotite	4.7	2.4	5	15
Garnet	1.9	4.9	7	5
Magnetite	.6	.4	3.5	1
Sillimanite		.5		
Apatite			Trace	Trace

biotite-quartz-plagioclase gneiss northwest of the southwest arm of Newark Reservoir. The gneiss, where not modified by pegmatitic veinings, appears to have plagioclase as almost the only feldspar. The modes of several specimens are given in table 7. Locally, the feldspar of the gneiss may be largely potassium feldspar.

EPIDOTE-SCAPOLITE-QUARTZ GNEISS AND ASSOCIATED ROCKS

Epidote-scapolite-quartz gneiss is a characteristic member in two rock units on opposite flanks of the Beaver Lake anticline. One unit originates about 750 yards northeast of Edison and extends northeast through Tamarack and Summit Lakes; the other starts just south of Franklin Pond Creek, about 950 yards east-southeast of the inlet of Franklin Pond, and extends northeast to the limits of the area.

Typically, the epidote-scapolite-quartz gneiss is interlayered with pyroxenic and hornblendic quartz-microcline gneiss and with pegmatite-seamed, locally garnetiferous, biotitic quartz-feldspar gneiss. The zone on the east flank contains local layers of plagioclase-microcline granitic gneiss contaminated with disseminated pyroxene and hornblende and relics of metasedimentary gneiss. There are narrow sheets of hornblende granite in the gneiss of the west limb. Layers of gneiss consisting of yellow-brown garnet (grossularite(?)), scapolite, and quartz are also present. Locally, the epidotic gneiss is slightly calcareous. The rocks weather with a characteristic thin-layered or ribbed appearance resembling a sedimentary series. The quartz-microcline gneiss layers resemble migmatitic layers, but their origin is uncertain. The epidote occurs both as primary grains and as part of a saussuritic alteration of the plagioclase. The scapolite is in part a product of alteration of plagioclase, as indicated by relics of the latter, and in part occurs as discrete uniform primary grains. Accessory minerals present are sericite, a phlogopitic mica, clinozoisite, allanite, apatite, zircon, sphene, and iron-titanium oxide minerals. A more detailed description of this rock type is presented in the section on the Edison magnetite deposits.

MISCELLANEOUS GNEISS

The rock unit west of the zone of marble through Franklin Pond is a heterogeneous mixture of gneiss and does not fit into any of the previous descriptions. This body consists predominantly of amphibolite (in part migmatitic and with gneissic lamination), biotitic quartz-oligoclase and biotitic quartz-feldspar gneiss (in part with pegmatite veinings), and pink hornblende granite, alaskite, and pegmatite. Locally, there are quartz-feldspar granulites, pyroxene-plagioclase gneiss, and garnetiferous or sillimanitic quartz-microcline gneiss. The rocks of this belt belong to what Hague and others (1956, p. 468) have called the Cork Hill Gneiss.

STRATIGRAPHY OF METASEDIMENTARY GNEISS

Hague and others (1956, p. 468) have proposed a stratigraphic sequence for parts of the Franklin Furnace, Hamburg, and Stanhope quadrangles. As applied with modification to the Franklin Furnace quadrangle, the arrangement for the rocks between Hamburg Mountain and the Pimple Hills would be as indicated in the sequence below. Known intrusive granitic materials are present in all metasedimentary units but are not referred to here. The gneiss west of Ogdensburg uniformly dips steeply southeast but is inferred to be overturned.

Inferred stratigraphic sequence, northwestern part of Franklin quadrangle

Biotite-quartz-oligoclase gneiss (gnb)[1] (intrusive?) rock unit underlying the Pimple Hills.

Wildcat Marble; northwesternmost marble unit (mls) shown on map.

Cork Hill Gneiss (mixed gneiss, gnph); amphibolite, biotite-quartz-feldspar gneiss; local pyroxene-plagioclase and garnetiferous and sillimanitic quartz-microcline gneiss.

Franklin Marble (mls); marble units adjacent to East fault and Zero fault; contains an interlayer (median gneiss) of gneiss.

Hamburg Mountain Gneiss Series, Hamburg Mountain. Biotitic, garnetiferous and sillimanitic quartz-potassium feldspar gneiss (gnm) with epidote-scapolite-quartz gneiss (gne) overlain by biotite-quartz-feldspar gneiss (gnbg) in the Edison area and by biotite-quartz-oligoclase gneiss, in part garnetiferous or biotite-rich, on the northeast nose of the Beaver Lake anticline.

Mafic gneiss (amg); amphibolite and local pyroxene gneiss.

Quartz-oligoclase gneiss (gno); (intrusive?); core of Beaver Lake anticline.

[1] Letter symbols refer to plate 1, this report.

If the igneous rocks are disregarded, the structure of the area between the Vernon and Reservoir faults permits the following stratigraphic section to be inferred:

Inferred stratigraphic sequence, southeastern part of Franklin quadrangle

Hypersthene-quartz-oligoclase gneiss (gnh);[1] in synclinal trough northwest of Stockholm.

Amphibolite and garnetiferous biotite-quartz-feldspar gneiss (gna); northwest of Lake Stockholm.

Pyroxenic gneiss (gnp); southwest of Lake Stockholm.

Correlation between the rock units of the northwestern and southeastern parts of Franklin Furnace quadrangle can be little more than gross speculation. Perhaps the biotite-quartz-feldspar gneiss of the Mount Paul anticline corresponds to the upper part of the gneiss series of Hamburg Mountain, and the hypersthene-quartz-oligoclase gneiss northwest of Stockholm in the Ryker Lake syncline corresponds to the biotite-quartz-oligoclase gneiss of the Pimple Hills syncline. If so, the mafic pyroxenic and amphibolitic gneiss southwest of Stockholm may correlate with the Franklin and Wildcat Marble and Cork Hill Gneiss of Hague and others (1956). This would indicate that the marble must have undergone intensive metasomatism to yield the silicates now found in the mafic correlatives. Doubt as to whether the gneiss of the Pimple Hills is metaigneous or metasedimentary in origin renders the basis for the correlation most speculative. In addition, questions regarding this correlation are accentuated by our ignorance as to the amount and nature of the faulting and by the fact that quartz-oligoclase gneiss is such a common rock type.

ORTHOGNEISS

METAGABBRO GNEISS

Gneiss or amphibolite which can positively be identified as derived from igneous gabbro is very rare in the studied area. A very small exposure of rock about 1,650 yards southeast of the southwest corner of Holland appears to be metagabbro gneiss. Outside the area of this report, excellent exposures of metagabbro gneiss are found on the hill about 1,650 yards N. 73° E. of the dam at Weown Lake in the Dover quadrangle. The rock can certainly be identified as derived from igneous gabbro, because of relics of primary texture and the occurrence of thin differentiated anorthositic layers. The mineral composition of samples of the metagabbro gneisses are given in table 8.

TABLE 8.—*Modes of metagabbro gneiss*

	1239	2239
Plagioclase	46.0	62
Hornblende (brown)	22.7	24.5
Clinopyroxene	22.5	12
Orthopyroxene		
Biotite	7.4	
Magnetite and Ilmenite		.5
Apatite	1.4	1.0

1239. About 1,650 yds southeast of the southeast corner of Holland.
2239. About 1,650 yds N. 73° E. of dam at Weown Lake, Dover quadrangle.

SYENITE GNEISS

A belt of syenite gneiss extends northeast from the southwest corner of the Franklin quadrangle through Mud Pond (Hamburg quadrangle), arcs around through Lake Wildwood and Hamburg Mountain, and passes southwest to just north of Ogdensburg. This belt is inferred to be a deformed sheet or phacolithic mass on the flanks and nose of the Beaver Lake anticline. The mass is 700–800 feet thick in the Edison area but is nearly 10 times this thick on the nose of the anticline.

The gneiss is a very uniform rock. The gneissic structure is due to the parallel lenticular form of the mineral grains and aggregates. There is very little compositional layering, and, except for rare amphibolite inclusions, the gneiss is homogeneous. Generally, it is uniformly medium grained; however, finer grained facies and pegmatitic seams are present. The texture is largely granoblastic. The color of a fresh surface is usually green, though locally, pale yellowish green.

The northwest limb of the syenite gneiss mass, the nose through Hamburg Mountain, and the southeast limb as far southwest as Beaver Lake, is a gray hornblende syenite gneiss. In contrast, the southeast limb southwest of Beaver Lake is a green-gray pyroxene syenite gneiss. In a zone about a mile northeast of Edison, the two facies appear to be interlayered.

The hornblende syenite gneiss of the Lake Wildwood area was described by Wolff and Brooks (1898) as the Sand Pond Gneiss. It also forms part of the Hamburg Mountain and Edison Gneisses as mapped by them.

Syenite gneiss occurs as an outlying sheet 1.1 miles east of Lake Wildwood, and another sheet is 0.8 mile west of Lake Wildwood. Granoblastic hornblende syenite gneiss is also exposed east of the railroad on the dirt road about 2,100 feet north-northwest of the north end of Morris Lake. The gneiss here has numerous pegmatitic veinings, a few inches thick, parallel to the foliation.

Another lens of hornblende syenite gneiss is exposed within the Morris Lake fault zone. The north end of the lens crops out on the hill on the northwest side of Morris Lake, but the best exposures are in the hills southwest of Morris Lake. The syenite is much fractured throughout, and slickensided surfaces coated with chlorite and epidote are present. The hornblende of the rock is commonly altered to chlorite or to fine-grained reddish-brown aggregates.

Approximate modes of characteristic hornblende syenite gneiss and pyroxene gneiss are given in table 9. The chemical analysis, norm, and mode of two specimens of the pyroxene syenite gneiss are presented in table 10.

The hornblende syenite gneiss is usually granoblastic. Two specimens were found, however, that had escaped crushing and recrystallization. In these were preserved the original structure of magmatic crystallization and primary mesoperthite feldspar—that is, very fine textured intimate intergrowth of subequal amounts of potassium and plagioclase feldspar (Michot, 1951). Much of the gneiss retains relics of mesoperthite grains set in a finer grained matrix of recrystallized slightly

TABLE 9.—*Approximate modes of facies of syenite gneiss*

	1	2	3	4	5	6	7	8	9
Microperthite and microcline	41.3	35.8	74.5	42.4	47.3	47.6	44.6	49.1	75.2
Oligoclase	44.9	44.2	7.2	45.2	40.0	44.8	48.7	36.8	8.9
Quartz	2.5	4.8	2.1	3.9	2.7	Trace	2.1	1.5	4.8
Augite	2.1		2.3	3.9	4.2	3.8	1.6	12.1	2.0
Hornblende	7.4	13.7	12.0	.5		.8			7.3
Magnetite and ilmenite	1.3	.8	1.5	1.2	3.7	1.6	2.1	.1	1.2
Apatite	.3	.4		.4	.6	.2	.4	.2	.4
Zircon	.1	.25	.4	.1		.1	Trace		
Sphene	.1				1.5		.5		.2
Epidote and chlorite				2.4		1.1			

1. Average of 7 sections of hornblende syenite gneiss from part of syenite gneiss belt between 1.2 miles north-northeast of Edison and 1 mile north-northeast of summit Lake.
2. Average of 2 sections of hornblende syenite gneiss from nose of syenite gneiss belt in Wildwood Lake area.
3. Hornblende microperthite syenite, 750 yd southeast of Lake Wildwood.
4. Average of 2 specimens of pyroxene syenite gneiss within 1 mile northeast of Edison.
5. Pyroxene syenite gneiss, on highway 0.8 mile northeast of Sunset Lake.
6. Pyroxene syenite gneiss, average of 2 sections from part of belt in northeast corner of Stanhope quadrangle.
7. Pyroxene syenite gneiss, 0.5 mile south-southeast of Sunset Lake.
8. Hornblende syenite gneiss, from sheet about 1.5 miles north-northwest of Lake Wildwood.
9. Hornblende microperthite syenite, from sheet 1.1 miles east-southeast of southeastern part of Lake Wildwood.

TABLE 10.—*Chemical analyses, norms, and modes of pyroxene syenite gneiss*

[Analyst, E. H. Oslund]

	1726	1922
Chemical analyses		
SiO_2	59.88	59.38
Al_2O_3	17.12	18.44
Fe_2O_3	3.88	3.16
FeO	3.09	2.69
MgO	0.33	0.65
CaO	3.07	2.60
Na_2O	4.72	5.33
K_2O	5.72	5.36
H_2O+	.25	.64
H_2O-	.06	.14
CO_2	.40	.04
TiO_2	.69	.70
P_2O_5	.18	.25
MnO	.18	.13
BaO	.20	.24
S	.08	.03
	99.70	99.85
Less O=S	.04	.01
Total	99.66	99.84
Norms		
Quartz	4.56	1.02
Orthoclase	35.03	33.36
Albite	39.82	44.54
Anorthite	8.06	9.87
Diopside	3.07	1.27
Hypersthene	.83	2.36
Magnetite	5.57	4.53
Ilmenite	1.37	1.37
Apatite	.47	.60
Calcite	.90	--------
Modes		
Quartz	2.7	0.9
Microcline and microperthite	47.3	45.7
Oligoclase	40.0	46.3
Augite	4.2	1.4
Hornblende		1.5
Magnetite and ilmenite	3.7	--------
Apatite	.6	.8
Sphene	1.5	.6
Veinlets of secondary mineral		[1] 2.8

[1] Chlorite and epidote on slickensided surfaces.

1726. Pyroxene syenite gneiss; roadcut on highway 0.8 mile northeast of Sunset Lake.
1922. Pyroxene syenite gneiss; on road about 2,000 ft north-northwest of Edison. Sample has a few fractures coated with epidote and chlorite.

micoperthitic microcline and oligoclase. The potassium feldspar of the wholly recrystallized gneiss is slightly perthitic microcline. The plagioclase is oligoclase. The data of table 11 show that, whereas the percentage of total feldspar remains about the same throughout the rock, the ratio of oligoclase to potassium feldspar increases with increasing intensity of metamorphism and resulting recrystallization of the mesoperthite to independent grains of microcline and oligoclase. The relics of mesoperthite do not show microcline twinning, whereas the unmixed potassium feldspar does. The hornblende of the gneiss is pleochroic from a deep brownish or yellowish green to a very dark olive green. The hornblende is probably a ferrohastingsite variety. Hinds (1921, p. 358) gives a partial analysis made by J. E. Wolff of a hornblende from the "Byram Gneiss" of Hamburg Mountain, which probably came from the hornblende syenite gneiss. The analysis shows 28.08 percent FeO, 0.22 percent Fe_2O_3 and 0.58 percent MgO. The very high percentage of FeO relative to magnesia indicates a ferrohastingsite. No evidence was seen to indicate that the hornblende was derived by alteration of pyroxene. Each occurs as independent grains. The accessory pyroxene is all a monoclinic variety. Accessory minerals include titaniferous magnetite, ilmenite, zircon, apatite, and sphene.

TABLE 11.—*Increase in percentage of oligoclase with degree of metamorphism in syenite gneiss*

	1	2	3	4
Microperthite and microcline	74.7	53.0	45.0	31.3
Oligoclase	8.0	32.9	38.0	52.5

1. Average of 2 undeformed hornblende microperthite syenite.
2. Average of 4 partly deformed and recrystallized hornblende syenite gneiss.
3. Average of 4 partly deformed and recrystallized hornblende syenite gneiss.
4. Average of 4 wholly granoblastic hornblende syenite gneiss.

The pyroxene syenite gneiss is less mafic than the hornblende syenite gneiss. It is granoblastic. The potassium feldspar is slightly microperthitic microcline, and the plagioclase is oligoclase. The pyroxene is ferroaugite. Sphene is nearly always present in the pyroxene syenite gneiss, in contrast to the hornblende syenite gneiss. It occurs as coronas around ilmenomagnetite and ilmenite and as separate grains. Quartz, titaniferous magnetite, ilmenite, zircon, and apatite are always present as accessory minerals.

IGNEOUS ROCKS

PYROXENE SYENITE, PYROXENE GRANITE, AND PYROXENE ALASKITE SERIES

Pyroxene syenite and pyroxene granite occur in association in several areas. These rocks have a microstructure inferred to be due to magmatic crystallization, and a gneissoid fabric resulting from magmatic flowage (as distinguished from gneissic structure resulting from plastic flow of solids). The areas occupied by these rocks are designated as the Sparta Lake, Holland, and Ford belts.

PETROLOGY

The fresh surface of the pyroxene syenite, and of much of the pyroxene granite, is green. In part, the pyroxene granite and alaskite are a pale greenish buff.

The weathered surface is usually gray, but, locally, the granite and alaskite may be pale pink. The rock is generally uniformly medium grained, locally containing nests and seams of pegmatite. It usually has a gneissoid structure but locally is almost massive.

The mutual relationships of the syenite and the granite and of the granite and alaskite have nowhere been observed. Lenses of syenite occur within the granite.

Chemical analyses, norms, and modes of two syenites and one granite are given in table 12. The syenite is very similar to the granite except that, by definition, the syenite has 10 percent or less quartz. The syenite also usually has more pyroxene, apatite, zircon, titaniferous magnetite, ilmenite, and sphene than does the granite, although there are exceptions. Often the mesoperthite feldspar is modified by the introduction of Na_2O and local recrystallization of the potassium feldspar intergrowth into coarser blebs and patches. The syenite analysis (table 12, No. 1923) is of a facies significantly modified by the introduction of Na_2O with extensive recrystallization of the potassium feldspar intergrowth. The other syenite (table 12, No. 1924) is representative of some of the least modified syenite. The granite (table 12, No. 175) seems to be almost wholly primary, that is, completely unmodified.

The variation in the character of the feldspars is the most noteworthy feature of these rocks. The principal feldspar in each of the major facies is a film microperthite in which the host is plagioclase—that is, the amount of plagioclase slightly exceeds that of the potassium feldspar. The norm ratio of the two feldspars of the pyroxene granite for which a chemical analysis is given in table 12 (No. 175) is representative. In such rock, pure plagioclase is restricted to a thin local exsolution border or rim of grains and to a little fine-grained intergranular aggregate partly mixed with microcline. The primary feldspar in all the pyroxene granite and syenite was mesoperthite. Except for the Sparta Lake belt, however, the feldspars are now very heterogeneous. Some are primary mesoperthites; others may be a single feldspar consisting of an inner core of mesoperthite, grading through a bleb microperthite outward to a zone in which there is a patch intergrowth of microcline in plagioclase; still others may consist of bleb microperthite or patch microperthite only. The microcline of the patch perthite commonly is in grains with a breadth of 0.2–0.4 mm (millimeters) though locally, the microcline has exsolved largely to the borders of the plagioclase, where it may form grains as much as 0.8 wide. Intergranular porphyroblastic grains of plagioclase are occasionally found. The rocks with such modified feldspar carry a higher ratio of

TABLE 12.—*Chemical analyses, norms, and modes of pyroxene syenite and pyroxene alaskite*

	1923	1924	175
Chemical analyses [N.d., not determined]			
SiO_2	61.62	61.55	72.43
Al_2O_3	14.54	14.36	12.94
Fe_2O_3	6.57	3.68	1.44
FeO	4.06	5.35	1.38
MgO	.46	.41	.34
CaO	2.13	3.08	1.70
Na_2O	5.21	4.46	3.83
K_2O	3.50	5.11	4.74
H_2O+	.20	.36	.37
H_2O-	.08	.13	.04
CO_2	.23	.15	.15
TiO_2	.96	.89	.24
P_2O_5	.16	.21	.11
MnO	.09	.21	.07
S	.09	.07	.02
BaO	.11	.08	.25
ZrO_2	.24	N.d.	N.d.
	100.25	100.10	100.05
Less O=S	.04	.03	.01
Total	100.21	100.07	100.04
Norms			
Quartz	13.71	8.82	28.29
Orthoclase	21.13	30.58	29.07
Albite	44.01	37.73	31.96
Anorthite	5.70	3.75	3.75
Diopside	2.13	8.94	3.25
Hypersthene	.71	2.03	.30
Magnetite	9.51	5.34	2.09
Ilmenite	1.82	1.67	.46
Apatite	.34	.50	.27
Calcite	.50		
Approximate modes			
Quartz	10.3	9.2	30.0
Mesoperthite		70.4	54.4
Antiperthite } Plagioclase }	74.5	7.2	
Plagioclase			9.9
Clinopyroxene	7.0	9.8	5.3
Hornblende		.4	
Magnetite and ilmenite	6.0	2.4	.2
Apatite	.2	.6	.2
Zircon	.2	Trace	
Sphene	1.8		

Chemical analysis of ferrohedenbergite from No. 1924

SiO_2	48.18	K_2O	.17
Al_2O_3	1.76	H_2O+	.80
Fe_2O_3	4.07	H_2O-	.35
FeO	23.71	TiO_2	.28
MgO	1.90	MnO	.97
CaO	16.59		
Na_2O	.61	Total	99.39

Chemical analysis recalculates to $Ca_{38.8}\ Mg_{6.2}\ Fe_{55.0}$, or hedenbergite 74.8, ferrosilite 12.9, and diopside 12.3.

1923. Pyroxene syenite; roadcut on State Highway 6A at Woodport 200 yd south of lake cove, Dover quadrangle. Analyst, E. H. Oslund.
1924. Ferrohedenbergite syenite, roadcut on State Highway 6A at road junction about 1.4 miles south of inlet of Saginaw Lake, extreme southwest corner of Franklin quadrangle. Analyst, E. H. Oslund.
175. Ferrohendenbergite granite, 6,200 ft east of outlet of Sickle Pond, 1,000 ft north of transmission line, and 1,700 ft south of road junction, Stanhope quad. Analyst, Doris Thaemlitz.

Na_2O to K_2O than rocks with mesoperthite only. This may be seen by comparing chemical analyses and norms of the syenites (table 12, Nos. 1923, 1924) and has been previously demonstrated by a series of chemical analyses (Buddington, in Sims, 1958 p. 155–56).

Most of the modified feldspar grains seem to be grains that maintain their original size and shape. The development of the bleb and patch antiperthite is attributed to the activity of sodic solutions that have irregularly seeped through the rock, introduced Na_2O at the expense of K_2O, and effected a varying degree of aggregation of the primary film intergrowth of potassium feldspar into coarser blebs and patches of microcline.

The feldspar of the syenites has been modified more widely than that of the granite and alaskite. The modification does not seem to be related to cracks, fractures, or faults or to any major crushing. It is as though the sodic fluids had moved inhomogeneously and intergranularly throughout the rock. However, the intergranular fine-grained feldspar aggregate, which occurs in most of the rocks, may have resulted from movement between the feldspar grains that facilitated the permeation of the sodic fluids.

Rock in which the feldspars have been largely albitized were mapped as Losee Gneiss (Ford and Holland belts) in the older geologic work, whereas the primary and slightly modified facies were mapped as Byram Gneiss (Sparta belt).

The clinopyroxene of the syenite was separated and analyzed (table 12, No. 1924). It is a ferrohedenbergite.

Locally, there are granite pegmatite veins a foot or so thick containing disseminated titaniferous magnetite concentrations and a few large complex granite pegmatites that have titaniferous magnetite concentrations in the core.

Titaniferous magnetite is the major opaque oxide mineral in all the pyroxenic rocks and is always associated with independent grains of ilmenite. The titaniferous magnetite uniformly has a microintergrowth of ilmenite; it is, therefore, the variety ilmenomagnetite. Partial chemical analyses of several ilmenomagnetites are shown in table 13, together with recalculation in terms of mineral molecules. Chemical analyses of coexistent ilmenomagnetite and ilmenite from two different rocks are given in table 14.

The analyses of the coexistent ilmenite and ilmenomagnetite, according to the data presented by Buddington and Lindsley (1964, p. 332), indicate a temperature of 675°C and log f_{O_2} of -18.4 atm (oxygen fugacity of $10^{-18.4}$ atm) for the conditions of formation of these two minerals in the syenite and 600°C and log f_{O_2} of -20.1

TABLE 13.—*Chemical analyses and recalculations of ilmenomagnetite concentrates from pyroxene syenite-pyroxene alaskite series*

[N.d., not determined]

	1924	173	169	401	131	949	170	1923	147
Fe_2O_3	55.34	58.91		61.69	68.25	66.07	65.18	64.74	
FeO	32.31	29.24		31.84	27.07	29.71	29.93	30.23	
TiO_2	7.97	3.60	2.76	3.67	2.75	1.97	1.52	1.96	0.97
SiO_2	1.69			.94	.63				
Al_2O_3	1.98			1.59	1.50				
CaO	.30			.11	.00				
MgO	.07			.14	.05				
MnO	.62			.14	.10				
	100.28	91.75		100.12	100.35	97.75	96.63	96.93	
Analyses recalculated to 100 percent for Fe_3O_4 $FeTiO_3$ Fe_2O_3 and Fe_2TiO_4									
Magnetite	83.6	93.2		92.1	92.0	92.1	95.4	94.7	
Ilmenite	15.1	6.1		5.7	5.5	3.8	3.0	3.8	
Hematite					2.5	4.1	1.6	1.5	
Ulvöspinel	1.3			2.2					
Excess TiO_2		.7							
Percent ilmenomagnetite in rock	N.d.	3.01	1.37	N.d.	N.d.	N.d.	20.64	N.d.	5.67

1924. Ilmenomagnetite concentrate from ferrohedenbergite mesoperthite syenite. Analyst, J. A. Maxwell.

173. Ilmenomagnetite concentrate from clinopyroxene mesoperthite granite, 1,000 ft southwest of Woodport triangulation station, beneath transmission line, Franklin quadrangle. Analyst, J. J. Fahey and Angelina Vlisidis.

169. Ilmenomagnetite concentrate from albite-oligoclase alaskite, Hibernia mine, Dover quadrangle. Analysts, J. J. Fahey and Angelina Vlisidis.

401. Ilmenomagnetite concentrate from ilmenomagnetite-rich segregation associated with pegmatite in pyroxene-quartz syenite. Goble prospect, 1.3 miles south-southeast of Saginaw Lake, and 0.2 mile northeast of road junction with State Highway 6A, extreme southwest corner of Franklin quadrangle. Analyst, J. A. Maxwell.

131. Ilmenomagnetite concentrate from ilmenomagnetite-rich segregation associated with granite pegmatite in pyroxene granite. Woods mine, 0.65 mile northeast of Stockholm, on extreme east edge of Franklin quadrangle. Analyst, J. A. Maxwell.

949. Ilmenomagnetite concentrate from ilmenomagnetite-rich granite pegmatite in clinopyroxene mesoperthite granite, 0.7 mile north of Sparta Lake, Franklin quadrangle. Analyst, J. A. Maxwell.

170. Ilmenomagnetite concentrate from ilmenomagnetite-rich albite granite pegmatite lens, Hibernia mine, Dover quadrangle. Magnetite is in knots or small elongate gobs with irregular radiating tentaclelike veinlets. Analysts, J. J. Fahey and Angelina Vlisidis.

1923. Ilmenomagnetite concentrate from clinopyroxene syenite with feldspars much modified and recrystallized accompanying introduction of Na_2O. Roadcut on State Highway 6A at Woodport, 200 yd south of lake cove, Dover quadrangle. Analyst, J. A. Maxwell.

147. Ilmenomagnetite concentrate from augite syenite gneiss, on highway 0.8 mile northeast of Sunset Lake, Franklin quadrangle. Analyst, J. J. Fahey and Angelina Vlisidis.

TABLE 14.—*Chemical analyses of coexisting ilmenite and ilmenomagnetite*

	1924		401	
	1	2	3	4
Fe_2O_3	3. 65	55. 34	4. 90	61. 69
FeO	41. 12	32. 31	42. 25	31. 84
TiO_2	49. 25	7. 97	48. 25	3. 67
SiO_2	1. 13	1. 69	. 70	. 94
Al_2O_3	. 93	1. 98	1. 94	1. 59
MnO	3. 15	. 62	1. 74	. 14
MgO	. 43	. 07	. 24	. 14
CaO	. 00	. 30	. 00	. 11
	99. 66	100. 28	100. 02	100. 12

1. Ilmenite concentrate from ferrohedenbergite mesoperthite syenite. Roadcut on State Highway 6A at road junction about 1.4 miles south of inlet of Saginaw Lake, Franklin quadra gle.
2. Ilmenomagnetite concentrate from same rock as No. 1.
3. Ilmenite concentrate from ilmenomagnetite-rich segregation associated with pegmatite in pyroxene-quartz syenite. Goble prospect, 1.3 miles south-southeast of Saginaw Lake and 0.2 mile northeast of road junction with State Highway 6A, extreme southwest corner of Franklin quadrangle.
4. Ilmenomagnetite concentrate from same rock as No. 3. Analyst, J. A. Maxwell.

for those in the segregation within the granite pegmatite. The percent of TiO_2 in the ilmenomagnetite from the alaskite is similar to that from the pegmatite and possibly indicates similar temperatures of formation. The percent of TiO_2 in the titaniferous magnetite from the pegmatite is, in general, lower than that from the syenite and granite. The ilmenomagnetite from the modified syenite has about the same percent of TiO_2 as that of some pegmatites. The TiO_2 of the titaniferous magnetite of the pyroxene syenite gneiss (table 13, No. 147) has the lowest TiO_2 and probably formed below 600°C. The ilmenomagnetite of all the pyroxene syenite, granite, alaskite, and pegmatite rocks carries more TiO_2 than the magnetite of the replacement concentrations in paragneiss and skarn.

There is a marked concentration of MnO and MgO in the ilmenite as compared with the ilmenomagnetite (table 14). This is a normal relationship (cf. Buddington and Lindsley, 1964, 352–353).

SPARTA LAKE BELT

The area of pyroxene granite and syenite that lies southeast of the Vernon fault zone and extends southwest from the outlet of Ryker Lake, through Sparta and Foxtail Lakes to the southwest border of the area is referred to as the Sparta Lake belt. It is separated from the Ford belt on the southeast by the area of pyroxenic gneiss. That part west of the included zone of quartz-potassium feldspar gneiss will be called the western belt and that to the east, the eastern belt.

Representative modes of the rocks are given in table 15. The composition varies from pyroxene syenite through pyroxene granite to pyroxene alaskite. The chemical composition, norm, and mode of an alaskite and a syenite from this belt are given in table 12.

The feldspar is usually a mesoperthite, plagioclase being slightly in excess of potassium feldspar. Generally, the mesoperthite is a homogeneous fine intergrowth of the two feldspars, but local modification resulting from the recrystallization of the potassium feldspar to rods, blebs, and patches is also observed. A subordinate part of the mesoperthite is inhomogeneous, individual grains being composed of relic parts of primary mesoperthite and adjacent areas of recrystallized blebs or patches. The mesoperthite grains usually have a thin halo of exsolved pure plagioclase on their borders. There is commonly a little fine-grained intergranular clear plagioclase. The preferred orientation of inequidimensional feldspar grains gives rise to the gneissoid structure.

Quartz ranges from a minor accessory mineral in the syenite to more than 30 percent in the alaskite. It

TABLE 15.—*Representative modes of pyroxene granite and pyroxene syenite, Sparta Lake and Holland belts*

	Sparta Lake belt						Holland belt	
	Western part				Eastern part			
	1	2	3	4	5	7	8	9
Mesoperthite	58. 9	76. 0	70. 4	60. 1	53. 7	70. 1	57. 7	41. 0
Plagioclase	8. 2		7. 2	7. 3	7. 9	2. 6	12. 2	22. 7
Antiperthite	----		----	----	----	----	----	
Quartz	30. 1	15. 0	9. 2	18. 4	32. 6	19. 6	24. 6	30. 0
Pyroxene	2. 4	4. 9	9. 8	11. 0	2. 0	5. 7	3. 4	4. 2
Hornblende	----	0. 4	. 4	----	----	. 3	. 2	----
Magnetite and ilmenite	. 2	3. 4	2. 4	2. 6	3. 3	1. 3	1. 5	1. 8
Apatite	. 1	. 2	. 6	. 2	. 1	. 2	. 1	. 2
Zircon	. 1	. 2	Trace	. 4	Trace	. 1	. 1	. 1
Sphene	----	----	----	----	. 1	. 1	. 2	----

1. Pyroxene mesoperthite alaskite, average of 4 (Nos. 175, 398, 407, 1426).
2. Pyroxene granite with some modified feldspar, average of 3 (Nos. 365, 401, 2303).
3. Hedenbergite mesoperthite syenite (No. 1924).
4. Clinopyroxene mesoperthite granite (No. 1429). 1.65 miles a little east of due north of Henderson Cove, Stanhope quadrangle.
5. Quartz-rich pyroxene granites, average of 5 (Nos. 330, 339, 940, 943, 1405).
7. Pyroxene mesoperthite granite, average of 5 (Nos. 343, 438, 440, 1115, 1429).
8. Pyroxene mesoperthite granite, average of 2 (Nos. 1702, 1706), from northeastern part of Holland belt.
9. Quartz-rich pyroxene granite with some modified feldspar, average of 3 (Nos. 1306, 1699, 1822), southwestern part of Holland belt.

occurs both as small rounded droplike grains and as elongate rounded grains oriented parallel to the gneissoid structure. The pyroxene is exclusively clinopyroxene, is free of any exsolved intergrowth, is generally medium to dark green, and is locally slightly altered to hornblende or chlorite(?). The pyroxene may be weathered to residual iron hydrate. Accessory minerals are ilmenomagnetite, ilmenite, apatite, and zircon. Sphene is locally present. The amount of titaniferous magnetite and ilmenite varies from a trace to several percent. Alaskite of the western part of the Sparta Lake belt has only a few tenths of a percent of ilmenomagnetite and ilmenite, whereas the quartz-rich granite of the eastern part has an average of 3.3 percent.

A narrow sheet of pyroxene syenite at the southwest end of the Sparta Lake belt was mapped. The syenite is very similar to the granite, except that it is darker green, has less than 10 percent quartz, and generally contains more pyroxene, apatite, and zircon. The two rocks are very difficult to tell apart in the field. Narrow layers of syenite occur within the granite adjacent to the northwest border of the syenite sheet. No contacts were seen, but the granite and alaskite may be younger than the syenite.

A similar narrow lens of pyroxene syenite was mapped along the southeast margin of the Sparta Lake belt. This syenite is locally contaminated with small lenses and schlieren of pyroxene gneiss and amphibolite.

FORD BELT

The area of pyroxene granite and alaskite that extends from the cross fault south of Mount Paul southwest to beyond the border of the area is referred to as the Ford belt. The Reservoir fault forms the southeast boundary of the belt. A central zone consists of alaskite with lenses of amphibolite; the rest of the belt is granite with alaskitic facies and, locally, syenitic facies. Representative modes of the various facies of the rocks are given in table 16.

The northwestern part of the Ford belt (the zone through Woodport triangulation station) consists of rocks that are very similar to those of the Sparta Lake belt. The feldspar is usually mesoperthite and is only locally modified by partial recrystallization. In contrast, the feldspar in most of the alaskite and the granite and syenite to the southeast shows varying but substantial modification by recrystallization with introduction of sodium oxide. Relics of the primary mesoperthite are nearly always present, however, and there are local facies of the granite in which mesoperthite is still virtually the only feldspar (table 16, No. 5). The feldspar in recrystallized and modified rocks grades from unmodified mesoperthite to almost clear plagioclase; intermediate stages consist of blebs and patches of potassium feldspar intergrown in plagioclase. Locally, there is alaskite in which the feldspar is exclusively albite-oligoclase.

The rocks of the Ford belt contrast with those of the Sparta Lake belt in commonly having accessory sphene, which occurs mostly as rounded grains but also as coronas around the iron and iron-titanium oxides. Accessory zircons are elongate and euhedral.

The sheet of pyroxene syenite northwest of the Reservoir fault extends southwestward through Woodport in the Dover quadrangle. Although the sample of pyroxene syenite (table 12, No. 1923) came from this southwest extension, it is representative of the syenites of the Ford belt. The ratio of albite to orthoclase in the norm of the Ford belt syenite (table 12, No. 1923) is greater than the ratio in the norm of the syenite

TABLE 16.—*Representative modes of pyroxene granite, alaskite, and syenite, Ford belt*

	1	2	3	4	5	6	7	8	9	10
Mesoperthite	57.0	67.3			64.9					67.4
Plagioclase	3.9	5.8	65.3	76.2	4.6	61.4	68.2	74.5	82.3	11.6
Antiperthite										
Quartz	34.9	20.1	29.5	17.4	25.8	35.8	30.0	10.3	7.4	7.8
Pyroxene	2.1	3.6	1.1	3.7	2.3			7.0	6.0	5.6
Hornblende				.1			.9			
Magnetite and ilmenite	1.7	2.2	2.2	1.9	1.7	2.4	.9	6.0	2.7	5.8
Apatite	.4	.4	.4	.3	.2			.2	.4	.9
Zircon	.2	.2	.2	.1	.2	.1		Trace	.2	Trace
Sphene	.3	.5	1.3	.4	.3	Trace		1.8	1.0	.9
Chlorite						.3				

1. Pyroxene mesoperthite, alaskite, average of 2 (Nos. 1152, 1292); near Woodport triangulation station.
2. Pyroxene mesoperthite granite, average of 2 (Nos. 355, 1297); near Woodport triangulation station.
3. Pyroxene alaskite, average of 2 (Nos. 1263, 1265); about 0.7 mile west of Milton.
4. Pyroxene granite, average of 14, from area between alaskite and syenite masses southeast of Ford Lake.
5. Pyroxene mesoperthite granite, average of 2, about 0.4 mile southeast of Ford Lake.
6. Alaskite, average of 4 (Nos. 1104, 1105, 1150, 1201); belt through Ford Lake.
7. Albite-oligoclase alaskite, (No. 1101); about 550 yd southwest of Ford Lake.
8. Pyroxene syenite, (No. 1923); roadcut on State Highway 6A at Woodport, 200 yd south of lake cove, Dover quadrangle.
9. Pyroxene syenite, average of 9, southeastern part of Ford belt.
10. Pyroxene mesoperthite syenite, (No. 2273); Woodport, Dover quadrangle. Plagioclase is fine grained intergranular.

(table 12, No. 1924) from the Sparta Lake belt. This is related to the greater modification of the feldspar in the more sodium-rich syenite. The original nature of the feldspar in both syenites must have been the same—that is, a mesoperthite, as still exists in the syenite of table 16, No. 10; but most of the feldspar in the syenites of the Ford belt has been modified by the introduction of Na_2O, accompanied by recrystallization of the initial potassium feldspar films into coarser blebs and patches. However, some original unaltered grains of mesoperthite still remain, and there are relics of the primary mesoperthite at the core of many grains. The introduction of Na_2O at the expense of K_2O, accompanied by recrystallization, has worked into the feldspar grains from the intergranular boundaries. The introduction of the Na_2O is not related to cracks or crushing and is therefore inferred to be an autometasomatic or deuteric process arising from the activity of residual sodium-rich magmatic fluids. The primary texture of the rock appears to have been little affected by the introduction of the Na_2O.

Due east of the Woodport triangulation station is a thin layer of mixed rock between the pyroxene granite and the alaskite. This belt of mixed rock consists of amphibolite and biotitic gneiss and some associated contaminated alaskite. There has been a little local magnetite mineralization. The amphibolite layers within the alaskite have locally been replaced by magnetite to form ore shoots at the Ford-Schofield mine and the Duffee mine (pl. 1). The alaskite locally forms pegmatite.

HOLLAND BELT

A belt of pyroxene granite extends from a little southwest of Holland to about a mile northeast of Stockholm (just beyond the east margin of the area). At the southwest the belt fingers out into the metasedimentary pyroxene gneiss. The rock is similar to that of the Sparta Lake belt. It is notably uniform, except for scattered local lenses of amphibolite. It is even grained and generally gneissoid, and has sporadic quartz seams. Mesoperthite feldspar that has plagioclase as host and that is only slightly modified (table 15, No. 8) is common. Other facies have feldspars that are modified (table 15, No. 9).

HYPERSTHENE GRANITE

A narrow band of hypersthene-bearing granite occurs just west of Camp Acquackanonk. It is similar to the normal green pyroxene granite, except that it has a more prominent foliation, contains hypersthene instead of clinopyroxene, and a little biotite.

The feldspar is a mesoperthite of the microantiperthite type. The microantiperthite to a slight extent is albitized with an accompanying coarser recrystallization of the potassium feldspar microintergrowth. The quartz is in amoeboid grains, the long diameter being oriented parallel to the foliation. The hypersthene also is in elongate grains oriented parallel to the foliation. Accessory minerals are biotite, ilmenomagnetite, apatite, rounded grains of sphene, and a few zircons. There has been no crushing.

The hypersthene granite is a border facies of the normal clinopyroxene granite; its different composition may have arisen from a slight assimilation of metasedimentary aluminous material.

HORNBLENDE GRANITE AND ALASKITE

Hornblende granite and related alaskite occur as rather narrow sheets at a number of localities. The largest mass is on the flanks and nose of the northeastern part of the Beaver Lake anticline. The alaskitic granite occurs as local facies within the hornblende granite, as independent narrow sheets in the metasedimentary gneisses, or as roof facies in phacolithic masses, as on the nose of the Beaver Lake anticline at the extreme northeast end of the area. There are locally included lenses or schlieren of amphibolite.

The hornblende granite and the alaskite are usually pinkish buff but may grade to buff. Most of the hornblende granite has a gneissoid structure. Microscopic examination shows grains of mesoperthite with a mortar of plagioclase and microcline. Locally, the plagioclase-microcline aggregates are thin granoblastic bands having a grain size of 1 mm or less, whereas the mesoperthite grains are 2–4 mm. Other facies have a more uniform medium-grained texture. The microcline of the groundmass is usually slightly perthitic. There are all gradations between perthite grains in which the plagioclase and potassium feldspar are about equal, to microcline that is only slightly perthitic. Grains of intermediate composition are subordinate, however. The approximate modes of representative samples are given in table 17. In massive facies of the granite the amount of plagioclase independent of perthitic intergrowth is small (table 17, No. 1). The feldspar in such rock is mostly mesoperthite. The more common type with more plagioclase (table 17, Nos. 2–4) is inferred to result from flowage at a late stage and lower temperature, causing mesoperthite grains to recrystallize and yield aggregates of microcline and plagioclase. Some larger mesoperthite grains remain as unstable relics. Locally, as in the hornblende granite mass just west of Petersburg on Bowling Green Mountain, deformation has been so extensive that the whole rock is modified, and all the mesoperthite is recrystallized to

TABLE 17.—*Approximate modes of representative types of hornblende granite and of alaskite*

	1	2	3	4	5	6	7	8
Quartz	23.0	12.8	18.7	19.8	28.6	31.6	34.2	31.5
Mesoperthite	57.1	47.9	49.2	56.3		56.6	47.1	
Microcline, perthitic					34.0			39.2
Plagioclase	7.9	29.4	21.4	21.1	32.4	9.2	15.7	26.6
Hornblende	9.3	7.5	9.6	1.6	3.8	.4		
Biotite	.4		.1	.3	.4	.6		1.7
Magnetite and ilmenite	1.8	2.2	.7	.6	.6	.9	1.7	.7
Apatite	.1	.1	.1	.1	.2			.2
Zircon	.2	.1	.1	.1		.1		.1
Fluorite	.2					.6	1.3	
Sphene			.1	.1				

1. Massive hornblende mesoperthite granite, about 0.4 mile southeast of road junction at Sand Hills, Hamburg quadrangle.
2. Hornblende granite; about 1.4 miles south of Sand Hills, Hamburg quadrangle.
3. Hornblende granite; southwest of Morris Lake.
4. Granite; about 1.6 miles south of Sand Hills, Hamburg quadrangle.
5. Hornblende granite gneiss; 1 mile west of Petersburg on south scarp of Bowling Green Mountain, Franklin quadrangle.
6. Mesoperthite alakite; average of 4 samples from different localities.
7. Mesoperthite alaskite; 0.7 mile south-southeast of road junction at Sand Hills, Hamburg quadrangle.
8. Perthitic microcline-plagioclase alaskite; 1 mile northeast of the head of Weown Lake, Dover quadrangle.

a granoblastic aggregate of slightly perthitic microcline and plagioclase (table 17, No. 5).

Quartz usually forms about 20 percent of the rock. Locally, however, the quartz may be as low as 10 percent; the rock then approaches a syenite (table 17, No. 2). As quartz increases, the percentage of mafic minerals decreases, and the rock passes into the alaskitic facies.

Hornblende is the principal mafic mineral and commonly forms nearly 10 percent of the rock. It is probably a femaghastingsite or a ferrohastingsite. A little biotite is usually present, as are accessory ilmenomagnetite, ilmenite, apatite, and zircon.

The alaskite (table 17, Nos. 6, 7) is typically less deformed and less recrystallized than the granite facies. The feldspar in the alaskite is mostly mesoperthite with a thin border of plagioclase exsolved to the borders of the grains. Locally, however, the alaskite contains highly recrystallized feldspar (table 17, No. 8). Quartz commonly forms more than 30 percent of alaskite, and mafic minerals, less than 5 percent. Biotite, ilmenomagnetite, and ilmenite are usually present. Alaskite may also carry a little sphene as well as a trace of zircon and apatite. Fluorite is present as an accessory mineral in much of the alaskite; it is found only rarely in the hornblende granite. Locally, the alaskite is contaminated with hornblende from amphibolite schlieren.

ROCKS OF UNCERTAIN ORIGIN

AMPHIBOLITE

Amphibolite is a general name for rocks in which hornblende and plagioclase are the major minerals. Pyroxene and biotite may also be present in substantial amounts. Amphibolite occurs in small and varying amounts within every unit of metasediments. In addition, it occurs as layers and schlieren in pyroxene granite, hornblende alaskite, pyroxene syenite, hornblende granite and associated alaskite, and the quartz-oligoclase gneiss. Mappable bodies of amphibolite occur in the extreme northwest of the area, on the nose of the Beaver Lake anticline, as layers in the pyroxene alaskite of the Ford belt, and on the nose of the Weldon Brook anticline just south of Bowling Green Mountain.

The amphibolite is usually intruded by granite or syenite. Veinlets and lenses of granite pegmatite and local granitization are associated with the intrusions. The introduction of quartz into the amphibolite may accompany the emplacement of the granite and pegmatite.

The amphibolite is typically dark gray to black and medium grained; where pyroxene is abundant, the amphibolite is more greenish gray. The pegmatitic veinings are light in color and may form a migmatitic facies with the amphibolite.

The variable composition of the amphibolite is shown by the modes given in table 18. The plagioclase is andesine (An_{30-40}) and is often much altered to sericite and an unidentified material. The hornblende is pleochroic from yellowish green to dark green or brown. The texture is crystalloblastic.

Part of the rock of these belts is derived from gabbro. The amphibolite body that defines the crest or nose of the Weldon Brook anticline at the south margin of the area may have been derived from igneous gabbro, that is, metagabbro. Evidence for this was observed in outcrops of the amphibolite to the southwest on the adjacent Dover quadrangle. Specifically, about 1,650 yards N. 73° E. of the dam at Weown Lake (Dover quadrangle), there are excellent exposures of meta-

TABLE 18.—*Modes of amphibolites*

	542	543	722	1059	1126	1221	1361	1854	1964-b
Plagioclase	49. 5	47. 2	34. 7	30. 6	25. 0	61. 3	55. 1	53. 5	39. 5
Hornblende	21. 0	16. 8	54. 2	64. 9	55. 4	11. 8	40. 3	17	53
Clinopyroxene	17. 0	25. 7	7. 4	3. 5	10. 4			19	
Orthopyroxene	11. 3	6. 7	1. 7		3. 0	17. 5		7	
Biotite									7. 5
Magnetite and ilmenite	1. 2	3. 4	1. 7	. 5	3. 2	Trace	1. 0	2	
Apatite	Trace	. 2	. 3	. 3	Trace	Trace	. 7	1. 5	
Quartz				. 2	3. 1	8. 5			
Epidote and chlorite							2. 9		

542. Just south of east end of Silver Lake, northeast Franklin quadrangle; green hornblende.
543. West of No. 542; green hornblende.
722. 2,100 ft east of Summit Lake, eastern Franklin quadrangle; green hornblende.
1059. 2,300 ft east of McAfee School, Hamburg quadrangle; brown hornblende.
1126. 4,100 ft east of Edison, brown hornblende.
1221. 5,300 ft east-southeast of Pine Swamp, green hornblende.
1361. 4,700 ft northeast of Russia, green hornblende.
1854. West side of southwestern part of Lake Stockholm, green hornblende.
1964-b. 5,500 ft west of north end of Canistear Reservoir, Wawayanda quandrangle; a 1-ft bed of amphibolite intercalated in biotite quartz-oligoclase greiss; green hornblende in amphibolite.

gabbro with differentiated anorthositic layers. Both the gabbro and anorthosite have primary texture with interlocking plagioclase laths.

QUARTZ-OLIGOCLASE GNEISS

A mass of quartz-oligoclase gneiss forms the core of the Beaver Lake anticline and extends southwest from Silver Lake through Beaver Lake and Sunset Lake to beyond the border of the area. Beaver Lake was originally called Losee Pond. This is the derivation of the name Losee Gneiss, which was originally applied to the rocks of this zone.

The weathered surfaces of the gneiss are generally white. On fresh surfaces the rock is also white but is spotted with a few small greenish blotches due to chlorite and epidote aggregates. The rock is generally medium and even grained, but fine- and coarse-grained facies are present. It usually has included layers of pyroxenic amphibolite and a well-developed foliation. The latter is caused by the preferred orientation of lenticular-shaped grains and aggregates of quartz. Foliation is also evidenced by interlayers of slightly variable composition and texture. Much of the gneiss exposed along the axial zone of the Beaver Lake anticline has only indistinct planar structure but a very strong linear structure that plunges northeast.

Two varieties of granitic pegmatite have been observed in the quartz-oligoclase gneiss. The most abundant is a white gneissic pegmatite that is conformably interlayered with the gneiss. The white pegmatite appears very similar to leucocratic facies of the quartz-oligoclase gneiss, and is inferred to be simply a coarse-grained variant. About 0.2 mile southeast of the power substation north of Ogdensburg, a shaft has been sunk in one of these pegmatites in which there is a small concentration of magnetite.

The less abundant type of pegmatite is pink and massive (undeformed), and cuts across the foliation of the gneiss.

Thin layers or veins, pods, and lenses of white massive milky quartz are in places conformably interlayered with the quartz-oligoclase gneiss.

A garnetiferous zone about 1,000 feet wide extends northeast from Morris Lake, through the west side of Hawthorne Lake and Heaters Pond, and northeast beyond the Ogdensburg-Edison road. Within this zone, layers of the gneiss as much as 100 feet wide locally contain disseminated garnet.

The principal minerals of the rock are quartz and oligoclase. Representative modes are given in table 19. Usually some of the plagioclase is cloudy and brownish white because of numerous microinclusions of secondary epidote. The principal mafic mineral in fresh rock is biotite. It is generally altered to a shreddy aggregate of chlorite and epidote. Other aggregates of chlorite, epidote, and a light-brown unidentified mineral appear to be the alteration product of some other mafic min-

TABLE 19.—*Representative modes of quartz-oligoclase gneiss*

	1	2	3	4	5
Quartz	36	32. 2	23. 7	35. 9	31. 4
Plagioclase	59	64. 8	72. 2	61. 3	65. 8
Orthoclase		. 5	. 4	. 2	
Hornblende	2. 6				
Biotite	1. 5		2. 1	0. 8	1. 6
Magnetite and ilmenite		. 3	. 4	. 1	
Apatite	. 9		. 2	. 1	
Zircon				Trace	
Chlorite and epidote		2. 2	1. 0	1. 6	. 4
Garnet					. 8

1. Average of 12 thin sections from Losee Pond belt of Losee gneiss (Spencer and others, 1908, p. 5).
2. Average of samples 0.5 and 0.75 mile east of north end of Sickle Pond, Starhope quadrangle. Most of the plagioclase is very cloudy from alteration; biotite is altered.
3. Average of four samples, southwest and west of Hawthorne Lake; fresh plagioclase and biotite.
4. Average of three sections from area west of Beaver Lake and Lake Girard; biotite altered to epidote and chlorite.
5. Northeast end of Beaver Lake (northwest of northeast arm); much of plagioclase is cloudy from alteration, and biotite is altered to chlorite and epidote.

eral. Chloritic "fingers" may radiate out from areas of alteration into the quartz and plagioclase. The average composition given by Spencer and others (1908, p. 5) shows 2.6 percent hornblende. Neither hornblende nor pyroxene was observed in any of the quartz-oligoclase gneiss examined during this study. Nevertheless, some of the alteration aggregates may have been derived from these minerals. Biotite, however, is the only major mafic mineral now identifiable.

The gneiss commonly consist of intermingled coarse and fine facies. The plagioclase of the coarser part is 1.5–2.5 mm in diameter; that of the finer part is a granoblastic aggregate of plagioclase grains 0.3–0.6 mm in diameter. Rounded quartz grains of similar size are associated with the plagioclase of the finer part. Larger quartz grains are about the same size as the larger plagioclases and are elongate parallel to the foliation. They have rounded borders and small amoeboid protrusions. Plagioclase and quartz grains of intermediate sizes are also present.

BIOTITE-QUARTZ-OLIGOCLASE GNEISS

Biotite-quartz-oligoclase gneiss occurs as the predominant rock in certain zones, such as that through the Pimple Hills. In addition, it is a major component interlayered with biotite-quartz-feldspar gneiss, as in the hills just southeast of Big Springs, in the belts through Mount Paul, and in the pyroxenic gneiss northwest of the southwest arm of Oak Ridge Reservoir.

The biotite-quartz-oligoclase gneiss of the Pimple Hills contains local layers of amphibolite. In general, the gneiss weathers buff. The rock prevailingly has biotite as the accessory mafic mineral, but locally may have a little hornblende. Some of the gneiss has an unidentified secondary mafic mineral aggregate. The modal analyses of biotite-quartz-oligoclase gneiss samples are presented in table 20.

ALBITE ALASKITE

In the extreme northwestern part of the Pimple Hills are two narrow belts of gneissoid alaskitelike rock separated by a sheet of hornblende granite. The alaskite is white to pink, even grained, medium grained, and uniform except for some layers of amphibolite. It is composed almost entirely of albite and quartz and a little accessory hornblende. The plagioclase is exceptionally clear and unaltered. A little biotite is often present, and microcline may be an accessory. There is a trace of apatite. The modal composition of representative samples is given in table 21.

QUARTZ-MICROCLINE GNEISS

Mappable bodies of quartz-microcline gneiss occur northeast of Edison and at the northeast end of the Beaver Lake anticline. Rock of this type is also associated with the quartz-potassium feldspar gneiss and occurs as local layers in biotite-quartz-feldspar gneiss. The garnetiferous quartz-microcline gneiss of the lens northeast of Edison interdigitates on the southwest with a biotitic (locally garnetiferous) quartz-plagioclase gneiss. There is an increase in the potassium feldspar of the plagioclase gneiss where the interdigitation occurs.

The rocks grade from nearly homogeneous quartz-microcline granulite to quartz-microcline granitic gneiss with well-developed foliation. The rock is characteristically pinkish buff and is fine to medium grained. It is composed almost exclusively of a granular aggregate of quartz and microcline with a mosaic texture. Garnet is a common accessory mineral. The garnet may be homogeneously distributed, or it may be concentrated in layers. The garnet from a sample northeast of Edison was partially analyzed (table 26). Other accessory minerals include plagioclase, sillimanite, biotite, muscovite, magnetite, and ilmenite.

TABLE 20. *Modes of biotite-quartz-oligoclase gneiss*

	2396	2342	2494	V-847-F	1359	1229	1085	1314	1333
Oligoclase	66	71	60	62.2	46.4	59.3	54.8	52.4	51.1
Quartz	30	19	36	33.4	34.6	35.2	30.4	34.5	27.6
Potassium feldspar	1			1.0	11.9	4.5	9.5	5.9	18.4
Biotite		9	4	3.3	6.1	.4	5.0	4.7	2.9
Hornblende	1								
Magnetite		1	Trace		1.0	.3		.6	
Zircon	Trace						.1		
Garnet								1.9	
Mafic alteration	2								

2396. 6,300 ft a little south of west of South Ogdensburg, Pimple Hills area.
2342. 4,500 ft northwest of Sterling Hill, Pimple Hills area.
2494. 3,700 ft northwest of Sterling Hill, Pimple Hills area.
V-847-F. 7,800 ft S. 85° W. of Ogdensburg (Hague and others, 1956, p. 450).
1359. 4,700 ft northeast of Russia.
1229. About 1.5 miles northeast of Russia.
1085. About 3,500 ft southwest of Vernon, Wawayanda quadrangle.
1314. About 1.0 mile west of Russia.
1333. About 900 ft northwest of Russia.

TABLE 21.—*Modal composition of albite alaskite*

	2385	2392
Albite	63. 8	62. 8
Quartz	32. 7	31. 4
Hornblende	3. 0	3. 1
Biotite	. 5	------
Apatite		. 1
Microcline		2. 6

2385. 0.5 mile southwest of No. 2392, on Newton East quadrangle.
2392. North of road, about 1.1 miles northwest of Sterling Hill, in Pimple Hills area.

Plagioclase and sillimanite, where present, are nearly always partly sericitized.

The modes of several representative samples are given in table 22. Quartz ranges from 30 to about 42 percent. This is lower than the quartz content of most of the metasedimentary quartz-potassium feldspar gneiss. Quartz in the latter ranges from 40 to 60 percent. Also, the quartz-microcline gneiss is more homogeneous than the metasedimentary quartz-potassium feldspar gneiss.

DIKES

Several diabase dikes intrude the Precambrian rocks. Except for fracturing near faults, the diabase is undeformed. It is inferred that the dikes are of Triassic age and that they are related to the basaltic igneous activity that took place in the Triassic lowlands to the southeast. Some dikes were not large enough to show on the map.

Rare dikes of lamprophyric character also occur. The dike which crosscuts the ore body at Franklin is well known; it was called camptonite by Spencer and others (1908, p. 13) and kugel-minette by Milton (1947, p. 523). Another minette dike occurs in granite about 0.8 mile southeast of McAfee schoolhouse (Hamburg quadrangle). The rock has phenocrysts of biotite and brown hornblende. The groundmass is plagioclase, interstitial alkali feldspar, and a highly altered mafic mineral that appears to have been pyroxene and (or) hornblende.

A diabase dike near the townsite of Russia, has been described in detail by Milton (1947, p. 526). The dike is fine grained and has veinlets of quartz and epidote. It is somewhat schistose and slickensided. The chemical analysis and norm are given in table 23. The plagioclase crystals are altered, and much of the ilmenite has been transformed to leucoxene. The dike is adjacent to the west side of the Reservoir fault. The shattering and alteration of the dike may be related to the faulting, and if the dike is of Triassic age, these relationships indicate that some movement on the fault occurred in post-Triassic time.

The largest diabase dike observed is just southwest of Beaver Lake. It strikes about N. 30°E. and dips about 50°SE., parallel to the foliation of the enclosing gneiss. It is about half a mile long and 40 feet thick. The rock is medium grained with diabasic texture. The two major minerals are cloudy plagioclase and clinopyroxene. Titaniferous magnetite and ilmenite are accessories; interstitial quartz, long, slender apatite needles, and a little secondary chlorite are also present.

At the northeast end of Beaver Lake is another dike of diabase. It has a diabasic texture, and the principal minerals are cloudy plagioclase and pyroxene (clinopyroxene and orthopyroxene). Interstitial quartz and granophyre, secondary hornblende and biotite, titaniferous magnetite, ilmenite, and long, slender apatite needles are subordinate constituents.

TABLE 22.—*Modes of quartz-microcline gneiss*

	A	115b	447b	B	794	809
Microcline	50. 8	43. 1	44	42. 2	57	58
Quartz	40. 9	32. 8	32	39. 0	30	36
Plagioclase	. 6	9. 2	21	15. 4	8. 5	----
Oxides	3. 1	4. 6	. 5	2. 8	4	2. 5
Sillimanite	----	----	1. 0	Trace	----	----
Garent	2. 8	----	----	----	----	1
Biotite	. 8	7. 8	----	. 6	. 5	----
Chlorite	. 7	----	1. 5	----	----	. 5
Sericite and muscovite	. 2	----	----	----	----	2
Hypersthene	----	2. 4	----	----	----	----

A. Garnetiferous quartz-microcline gneiss; from northeast of Edison; average of 16 thin sections.
115-b. Biotite-quartz-microcline gneiss (contains pegmatite seams;) about 1,000 ft north of the west end of Foxtail Lake.
447-b. Sillimanitic quartz-microcline gneiss (plagioclase is largely sericitized); about 500 ft north of north end of Lake Saginaw.
B. Biotite quartz-microcline gneiss; 1,200-1,400 yd east of McAfee School; average of four thin sections; 1 specimen has sillimanitic nodules not cut in the thin section.
794. Biotite quartz-microcline gneiss; 1,400 yd northeast of Lake Wildwood.
809. Muscovite-garnet-quartz-microcline gneiss; 1,400 yd south-southeast of McAfee School.

TABLE 23.—*Chemical analysis and norm of diabase*

[From Milton, 1947, p. 526]

Chemical analysis			
SiO_2	47. 42	CO_2	Trace
Al_2O_3	13. 99	TiO_2	2. 30
Fe_2O_3	1. 97	P_2O_5	. 30
FeO	11. 49	S	. 26
MgO	6. 62	MnO	. 15
CaO	7. 81	BaO	. 00
Na_2O	3. 30	ZrO_2	. 05
K_2O	1. 54		
H_2O+	2. 71	Total	100 02
H_2O-	. 11	Density	3 06

Norm			
Orthoclase	8. 9	Magnetite	2. 8
Albite	27. 8	Ilmenite	4. 4
Anorthite	18. 9	Water	3. 9
Diopside	15. 8		
Olivine	17. 3	Total	99. 8

On the hilltop 0.3 mile northeast of Beaver Lake is a pluglike mass of diabase that is quite different from the other diabase. The plagioclase displays crude flow structure. It is fresh and clear and is locally recrystallized to a granoblastic aggregate. The major mafic mineral is hornblende, which appears to be secondary. Hypersthene, pleochroic from pale green to pale red, is associated with the hornblende. Titaniferous magnetite is sparse. Ilmenite is the principal opaque oxide. The magnetite has presumably been absorbed by the hornblende during the secondary alteration that has apparently affected this rock.

PETROGENESIS

PYROXENIC GNEISS

There is general agreement by geologists that many pyroxenic gneisses are derived from carbonate-bearing sedimentary rocks. Carbonate-bearing sedimentary rock, except for carbonate-bearing graywacke, do not have a sodic content comparable to these gneisses. Neither do they have adequate chlorine to yield the scapolitic facies. Therefore, it seems probable that the primary carbonate-bearing sedimentary rock from which the pyroxenic gneiss was presumably derived was modified, at least to some degree, by metasomatizing solutions. If the original sedimentary rock was a limestone, such as the Franklin Limestone, clearly there has been intense introduction of silica, alumina, soda, magnesia, and, locally, iron and chlorine. The requisite amount of metasomatism to form pyroxenic gneiss would, of course, depend on the chemical nature of the original material.

QUARTZ-POTASSIUM FELDSPAR GNEISS AND ASSOCIATED EPIDOTIC GNEISS

The generally quartz-rich composition, the occurrence of sillimanite, almandite, and biotite, the bulk chemical composition, the heterogeneous character, and the interlensing of epidotic gneiss (locally calcareous) are evidence that the quartz-potassium feldspar gneiss was derived from a sedimentary rock sequence. Epidotic gneiss is inferred to have a chemical composition derived from a calcareous shale or arkose. In some of these gneisses there is a still excess carbonate, that is, calcite uncombined with silicates. In the Edison area epidotic gneiss lenses into the quartz-potassium feldspar gneiss to the southwest. A very small isolated lens occurs farther to the southwest. These features suggest a primary facies change within a series of bedded sedimentary rocks. These rocks have been modified to varying degrees by granitic metasomatism, that is, granitization.

QUARTZ-OLIGOCLASE ROCKS

Quartz-oligoclase rocks form major belts, such as the core of the Beaver Lake anticline, the trough of the Ryker Lake syncline, and the zone through the Pimple Hills. They are also abundant as interlayers in heterogeneous metasedimentary gneisses, such as the Mount Paul belt, the area southeast of Big Springs, and the zone half a mile south of Edison.

The origin of these rocks has not been solved. They may represent simple metamorphic rocks formed by recrystallization of sedimentary rocks or tuffs, metasomatic rocks resulting from the replacement by oligoclase of quite different rocks through the activity of sodic solutions, or orthogneisses representing metamorphosed igneous rocks. Several origins may be, and probably are, represented by them.

HYPERSTHENE-QUARTZ-OLIGOCLASE GNEISS

The hypersthene-quartz-oligoclase gneiss of the Ryker Lake syncline has been grouped with the metasedimentary gneiss because of the widespread distribution of disseminated graphite, the local interlayering of materials of different composition, and the occurrence of similar layers of hypersthenic biotitic quartz-plagioclase gneiss intercalcated with metasedimentary garnetiferous gneiss of the Mount Paul area. Despite evidence for a metasedimentary origin for the gneiss, there are difficulties with the hypothesis. The only common sedimentary rocks of such sodic composition are graywacke or tuff. Graywacke may occur as thick uniform beds, but there are always some intercalcated normal shale layers. Equivalents for such shale in the hypersthenic biotitic quartz-oligoclase gneiss were not found. If the material were recrystallized sodic tuff, metamorphosed lava flows would also be expected. These have not been found, but metamorphism has been so intense that their recognition would be difficult. Some geologists believe that some quartz-oligoclase granite would, if metamorphosed, form an orthogneiss similar in composition to the hypersthenic gneiss. However, it is difficult to explain the graphite if this hypothesis is used. Others would doubtless explain this gneiss as the product of metasomatism of normal sedimentary shale in which the original high ratio of K_2O to Na_2O had been reversed by sodic metasomatism. No relics of original rock that might have served as the raw material for replacement have been found. Thus, the metasomatic hypothesis is without supporting evidence.

QUARTZ-OLIGOCLASE GNEISS (LOSEE GNEISS)

The quartz-oligoclase gneiss (Losee Gneiss) of the Beaver Lake belt was interpreted as of igneous origin by Spencer and others (1908, p. 5) and as intrusive into Pochuck Gneiss and the Franklin Limestone with unknown relationship to the Byram Gneiss. Hague and others (1956, p. 458) infer that the Losee Gneiss was formed as a predominantly concordant magmatic intrusion resembling a phacolith. In detail they describe small layers of hornblende gneiss which "are actually rectangular blocks (inclusions) which sometimes have coarse pyroxene borders," and they refer to small-scale discordant relationships between the Losee Gneiss and some of its included hornblende gneiss layers. Baker (1955, p. 30) has described similar discordant relations between the Losee Gneiss and metaquartzite.

The metaquartzite layers are less than a foot thick and are strongly deformed into tight minor folds and crenulations. The limbs of these minor folds are often broken up and pulled apart and surrounded by quartz-oligoclase gneiss in a boudinage fashion.

One such locality is by the roadside about 0.8 mile west of Mahola. North of Silver Lake is a zone of amphibolite between the quartz-oligoclase gneiss and the hornblende syenite gneiss (pl. 1). A small amount of feldspathic pyroxene skarn has formed between the amphibolite and each of these rocks. This suggests a common origin for the quartz-oligoclase gneiss and syenite gneiss, that is, igneous, with development of skarn as a reaction product at the intrusive contact. The general uniformity in composition of the quartz-oligoclase gneiss is consistent with an origin as orthogneiss. Of 15 samples studied, none showed more than 40 percent quartz or less than 21 percent. The associated albite granite pegmatite veins, one of which carries titaniferous magnetite enrichment, are also consistent with a magmatic hypothesis. If the gneiss is inferred to have originated initially from magma, then at least part of the biotite-quartz-plagioclase gneiss layers in the areas mapped as metasedimentary may be of similar origin. The field relationships in many places show sharp contacts between the biotite-quartz-plagioclase gneiss and adjacent rocks, which would be consistent with this possibility.

A metasedimentary origin for the gneiss could be argued for the following reasons. In general, there is conformity between the quartz-oligoclase gneiss and the amphibolite layers. The presence of disseminated garnets in a narrow central belt is consistent with a metasedimentary origin. The Losee Gneiss as mapped by Hague and others (1956, pl. 1) crops out in the core of (Pochuck Mountain and Stag Pond) anticlines as it does in the core of the Beaver Lake anticline. If reasonable allowances are made for complicating effects arising from granitic intrusion and plastic flowage, then in all three anticlinal structures the quartz-oligoclase gneiss may be inferred to be overlain by hornblende gneiss and amphibolite, and these, in turn, overlain by potassium feldspar gneiss, as though all belonged to a similar stratigraphic series of metasediments.

The hypothesis of simple metamorphic recrystallization of metasediments meets the same objection as that raised in connection with the hypersthene-quartz-oligoclase gneiss of the Ryker Lake syncline—that is, the peculiar composition and relative lack of clear-cut bedded structure. Furthermore, no graphite has been found in the quartz-oligoclase gneiss. Finally, no evidence for derivation of the gneiss by metasomatism from some other rock has been found. Although evidence slightly favors magmatic origin, it is so conflicting that, for the present, this rock is regarded of uncertain origin.

AMPHIBOLITE

Amphibolite may have several modes of origin. The primary gabbroic texture and differentiation into anorthositic and gabbroic layers observed on the east flank of the Weldon anticline (Dover quadrangle) indicates a metagabbro origin. Hague and others (1956, p. 443, 457) have found what appears to be "pillow structure" in certain amphibolites of Pochuck Mountain, suggesting origin by metamorphism of basaltic or andesitic lavas. Some quartz- or calcite-bearing amphibolite which is interlayered and gradational with metasedimentary gneiss is inferred to be metamorphosed dolomitic argillaceous or argillaceous siliceous carbonate-bearing rocks. For example, in the hill just north of Weown Lake (Dover quadrangle) is thin-layered (on the scale of a fraction of an inch) gneiss with light-colored clinopyroxene-plagioclase layers and dark-colored hornblende-clinopyroxene plagioclase layers. These rocks, which are adjacent to biotite-quartz-oligoclase gneiss, appear to be metasediments. Elsewhere (Adams, 1909), amphibolites have been interpreted as replacements of carbonate rocks.

As a result of intensive studies, Hague and others (1956, p. 443–457) have defined four types of hornblende gneiss, that is, amphibolite. Type I is banded on a scale of 1/8 inch to 1 inch. The dark bands are of hornblende and pyroxene with minor feldspar and some biotite; the light bands are mostly of feldspar with minor quartz and dark minerals. Masses of pyroxene-scapolite rock may also occur within "pillows"; locally, calcite is present and weathers out to form irregular cavities in the center of the "pillows." In many places type I hornblende gneiss grades into dark biotite gneiss. Type II hornblende gneiss is described as a fine- to medium-grained gneiss with an irregular striped appearance. The plagioclase in type I is An_{75}, whereas in type II it is An_{32}. Type III hornblende gneiss is like type I, but it also has lenses of calcite and layers or lenses rich in pyroxene and accessory sphene. Type IV hornblende gneiss is dark green to black and contains clusters or augen of medium- to coarse-grained hornblende in a fine-grained biotite-hornblende-plagioclase groundmass. Types I and II hornblende gneiss are stated (Hague, and others, 1956, p. 457) to be similar to type III and, locally, to grade into each other. Type III calcitic hornblende gneiss is inferred to result from metamorphism of very impure limestone or a sequence of argillaceous and calcareous sediments. Hague and others (1956) suggest that type I hornblende gneiss was formed by metamorphism of two different kinds of rocks, namely, volcanic flows and argillaceous sediments of similar bulk composition. Alternatively, they indicate that types I and II hornblende gneiss are volcanic in origin, and type III was formed by interfingering of volcanic flows and calcareous sediments. Type IV is assumed to be metaigneous because of its texture and its numerous discordant contacts with quartz-rich microcline gneiss.

Evidence that would permit definite decisions on the origin of the amphibolite in the studied area was not obtained.

QUARTZ-MICROCLINE GNEISS

The quartz-microcline gneiss commonly has accessory minerals, such as almandite or sillimanite, which indicate a metasedimentary origin. The composition is also comparable to the composition of some shale. The uniformity and structural relations, however, suggest that the rock might have been of magmatic origin. The uniformity is in contrast to the heterogeneity of the quartz-potassium feldspar gneiss. The quartz content of the quartz-microcline rock is usually between 30 and 40 percent, and does not exceed 42 percent, whereas in the quartz-potassium feldspar metasedimentary gneiss the quartz commonly exceeds 40 percent. The quartz-microcline rock also interdigitates with the biotitic quartz-plagioclase gneiss south of Edison in a fabric comparable to that of an intrusive rock. Fingers of the quartz-microcline rock an inch wide have sharp walls against the biotitic quartz-plagioclase gneiss. The rock is exceptionally high in potassium and low in sodium for a normal igneous rock. Nevertheless, it would be similar in composition to potassium-rich sodium-poor Tertiary vitric tuff (magmatic) beds described by Swineford, Frye, and Leonard (1955). The garnet and sillimanite could be explained as contamination products from reaction with metasediments.

Similar rock, which occurs extensively in the Precambrian of the Adirondack area in New York, has been interpreted (Buddington, 1957, p. 297–302) as partly metasomatic (metasomatism of biotite-quartz-plagioclase gneiss by potassium-rich fluid) and partly magmatic. However, the relative importance of metamorphic recrystallization, metasomatic replacement, and magmatic injection is so difficult to assess that the rock is classified as of uncertain origin.

HORNBLENDE AND PYROXENE SYENITE ORTHOGNEISS

The hornblende and augite syenite gneiss are definitely metamorphosed igneous syenite. Although the feldspar of the gneiss is mostly slightly microperthitic microcline and oligoclase, there are relics of mesoperthite and grains transitional between mesoperthite and granoblastic aggregates of microcline and oligoclase. This occurrence indicates that the transitional grains are derived by crushing and recrystallization of the mesoperthite.

GRANITIC AND SYENITIC ROCKS

The Byram Gneiss, which includes the augite and hornblende syenite gneiss, the pyroxene syenite, pyroxene granite and related alaskite, and the hornblende granite and related alaskite, was interpreted as of magmatic origin by Spencer and others (1908). Sims (1953, p. 258–260) similarly interpreted the hornblende granite and related alaskite of the Dover district as igneous. Further, Hotz (1953, p. 186–187) stated that in the Ringwood-Sterling district the hornblende granite, related alaskite, and pegmatite probably formed by crystallization from a magma. Hague and others (1956, p. 459) concluded from their study of the Franklin-Sterling area that:

> Most of the field and microscopic evidence points to an igneous origin for the Byram gneiss. However, some of the evidence contradicts the igneous origin, and hence the Byram gneiss may have a complex history of igneous intrusion coupled with partial replacement.

The hornblende and augite syenite gneiss, the pyroxene granite and related alaskite, and the hornblende granite and related alaskite in the Franklin-Hamburg area are interpreted as of magmatic origin. This conclusion is based on the following: (1) The analogy in composition and texture with other igneous rocks, (2) the relative uniformity and slight variation of composition of each, (3) the evidence that the primary feldspar of all these rock was anorthoclase (now exsolved to mesoperthite), (4) the TiO_2 content of the unaltered and unrecrystallized titaniferous magnetite (Buddington and others, 1955; Buddington, 1956), (5) the euhedral and elongate character of the zircons, (6) the development of appropriately correlated pegmatites, and (7) the field relationships.

The pyroxene syenite commonly is intimately associated with amphibolite and pyroxene gneisses. This association indicates that it is a reaction product of the pyroxene granite magma with mafic country rock. This applies in particular to the narrow syenite lenses within the Sparta belt and to the small lenses in the Ford belt. Pyroxene syenite facies were found by Buddington (in Sims, 1958) to occur at the Hibernia mine exclusively within zones of amphibolite intruded by alaskite. The larger syenitic masses, such as the sheet through Weldon and Woodport in the Dover quadrangle, may be a primary differentiate or could be a facies of the pyroxene granite magma modified at depth by reaction with mafic wallrocks.

The hypersthene granite (and locally, hypersthene syenite) just southeast of Edison is inferred to be a hybrid facies resulting from reaction of the pyroxene granite magma with aluminous biotitic metasedimentary rock. This is in accord with the reaction whereby alumina reacts with pyroxene to yield anorthite (dissolved in plagioclase) and hypersthene.

The fact that part of the mesoperthite in the pyroxene syenite, the pyroxene granite, and associated hornblende alaskite has been modified to a more albitic feldspar has been referred to in a description of these rocks. The sodic fluids effecting this transformation are inferred to have permeated upward through effectively consolidated, but still hot, rock along intergranular boundaries. Very locally, the pyroxene-quartz-mesoperthite granite has had its feldspar modified so that a pyroxene quartz-oligoclase granite of metasomatic origin has formed. However, the oligoclase of such rock has at least traces of relics of the original antiperthitic texture. At the Hibernia mine (Buddington in Sims, 1958), albite granite is known as a member of the pyroxene granite series. It is thus certain that sodium-rich fluids were an end product of the pyroxene granite magmatic series. No evidence was seen that the wall (country) rocks of the pyroxenic igneous series were similarly modified by the sodic solutions, although they may have been.

METAMORPHIC FACIES

Any attempt to assign the rocks of this area to current classification of metamorphic facies is fraught with many difficulties. Unrecognized unconformities may exist, and two or more periods of metamorphism may have occurred, so that the degree of equilibrium recrystallization may differ for different rocks and different geographic belts. Faulting has complicated interpretation, because continuity of systematic changes in regional intensity of metamorphism cannot be mapped. Furthermore, there is the problem of distinguishing contact metamorphism induced by intrusive igneous rocks from regional metamorphism. Finally, certain critical types of rock do not occur in all three units so that comparison can be made, and there is widespread lack of equilibrium in many of the gneisses. Also, a knowledge of the mineralogy of the rocks of bordering regions is necessary before any dependable interpretation can be made, and this is not available. Nevertheless, a discussion of possible correlation in the light of available knowledge is given.

The classification of metamorphic facies presented by Turner and Verhoogen (1960, p. 531–560) is used. Three units are discussed: (1) the Franklin block between the Hamburg and Zero faults, (2) the Beaver Lake horst between the East and Vernon faults, and (3) the Ryker Lake block between the Vernon and Reservoir faults. The rocks of the block southwest of Bowling Green Mountain probably also belong to the same facies.

Sillimanitic and garnetiferous biotite-quartz-microcline gneiss is common throughout the area. Muscovite may occur in the gneiss as an alteration product, but it is not contemporaneous with the associated minerals. The potassium feldspar is predominantly microcline instead of orthoclase, but this is common in certain terranes of the high-amphibolite facies. A grade of metamorphism equivalent to, or higher than, the sillimanite-almandine-orthoclase subfacies of the amphibolite facies is thus indicated for the whole region. There appears to be a systematic increase in the grade of metamorphism of the units from the northwest to the southeast.

FRANKLIN BLOCK

The gneiss and marble of the Franklin block are here referred to the sillimanite-almandine-orthoclase subfacies of the amphibolite facies. Hague and others (1956, p. 443) state that the microcline of the gneiss in the Pimple Hills contains as much as 20 percent albite blebs, 0–5 percent southwest of the Pimple Hills, and less than 1 percent to the northeast. Quartz-microcline gneiss has an average of about 14 percent plagioclase, which varies from An_{5-15}. Marble beds contain diopside, tremolite, and chondrodite or norbergite. Hornblende is the dominant mafic mineral of the amphibolite gneiss but may be associated with biotite or clinopyroxene. More systematic study of the hornblendic gneiss is needed, however, with respect to the problem of assignment to a specific metamorphic facies.

BEAVER LAKE BLOCK

The gneiss of the Beaver Lake horst is in part similar to that of the Franklin block but in part is different and may be of a higher metamorphic grade. The evidence for assignment to a specific subfacies is not conclusive.

The amphibolitic gneiss characteristically has the mafic assemblage green hornblende + clinopyroxene + subordinate orthopyroxene with locally a little biotite, whereas in the Franklin block, orthopyroxene is generally absent and clinopyroxene is local. The assemblage with orthopyroxene is characteristic of the hornblende granulite subfacies.

The potassium feldspar of the biotite-quartz-plagioclase gneiss carries about 10–15 percent perthitic plagioclase and is associated with oligoclase (An_{20}). Some of the potassium feldspars have relatively low triclinicity. These data suggest development at a higher temperature than similar gneiss of the Franklin block.

The epidote-scapolite-quartz gneiss and associated beds, taken by themselves, pose a problem for assignment to a specific metamorphic subfacies. The plagioclase is calcic oligoclase (An_{20-30}), the potassium feldspar is nonperthitic microcline, the scapolite has meionite$_{40-60}$, the pyroxene is salite, and the hornblende is pleochroic green with a faint bluish hue for *Z*. Sphene is present, as are clinozoisite and locally grossularite. The oxide mineral is ilmenohematite. This assemblage must have formed under conditions of high volatile pressure of H_2O, CO_2 and Cl, and could be classified in the upper amphibolite facies.

Locally, a quartz-oligoclase-microcline gneiss layer in the quartz-oligoclase gneiss of the core of the Beaver Lake anticline carries blue-green hornblende and sphene.

The accessory minerals epidote and chlorite in the quartz-oligoclase gneiss of the core of the Beaver Lake anticline are widespread; they are products of alteration in the greenschist facies. Why this moderate grade of alteration should be of so much more uniform occurrence in this belt of gneiss than elsewhere is not understood.

RYKER LAKE BLOCK

The regional metamorphic rocks of the Ryker Lake block clearly belong to the hornblende granulite subfacies. The amphibolite gneiss characteristically has the mafic assemblage brown hornblende + clinopyroxene + orthopyroxene, though local facies may have green hornblende instead of brown, or hornblende ± biotite may be the only mafic minerals.

The calcareous rocks have the typical assemblages cited by Turner and Verhoogen (1960, p. 556), including the combination andesine + clinozoisite + clinopyroxene.

The hypersthene-quartz-oligoclase gneiss with disseminated graphite also has a mineral assemblage appropriate for the hornblende granulite facies: hornblende (2–5 percent) + hypersthene (3–5 percent) + brown- to reddish-brown biotite (0.5–3 percent), and a little microcline with a slightly microperthitic intergrowth of plagioclase that is locally antiperthitic. The hornblende, in part, is pleochroic from a straw color to brown.

COMPARISON WITH NEARBY AREAS

The older metamorphic rocks of the Dover area, southeast of the Franklin quadrangle, may also belong to the hornblende granulite facies. Sims (1958, p. 12, 16) reports that all the amphibolite carries pyroxene, in part with both clinopyroxene and orthopyroxene, and plagioclase An_{37-48}. Metasedimentary gneiss includes sillimanitic, garnetiferous biotite-quartz-feldspar gneiss. The plagioclase is sodic oligoclase. Another member of the gneiss is described by Sims (1958, p. 28–29) as hypersthene-quartz diorite. Hypersthene ranges from 6 to 18 percent, and the plagioclase is mainly andesine, though, locally, oligoclase. Most of it is antiperthitic.

Hornblende granite gneiss, of younger age than the gneiss referred to above, must belong to the upper amphibolite facies because it is composed of quartz, microcline, albite-oligoclase, and accessory blue-green hornblende and sphene (Sims, 1958, p. 34). The accessory mineral assemblage is characteristic of the amphibolite facies.

The older metamorphic rocks of the Sterling Lake-Ringwood area, which is about 10 miles east of the Hamburg quadrangle, also seem to belong to the hornblende granulite facies (Hotz, 1953). Part of the amphibolites carry both clinopyroxene and orthopyroxene (Hotz, 1953, p. 170). The hornblende ranges from brownish green to dark green and the plagioclase from An_{38-48}. Hotz also reports (1953, p. 173) that both hypersthene and diopsidic augite may accompany biotite in quartz-biotite gneisses and in quartz-oligoclase gneiss (1953, p. 175–177). The plagioclase is commonly antiperthitic, and, locally, is accompanied by perthitic microcline. The hornblende is green.

HARDYSTON QUARTZITE AND UNCONFORMITY ON PRECAMBRIAN ROCKS

A great unconformity occurs between the Precambrian rocks and the Lower Cambrian Hardyston Quartzite which forms the basal member of the Paleozoic formations. The relationships of the quartzite to the Franklin Limestone have been previously described in detail by Wolff and Brooks (1897). The Hardyston has been called a quartzite, but it is really a well-cemented sandstone in which the original rounded character of the grains is still well preserved.

The Paleozoic rocks have a northeast strike and northwest dip and lie unconformably across the Precambrian rocks of the Hamburg and Wawayanda Mountains where the Precambrian structures strike N. 10–15° E.

Two contacts between beds of the Hardyston and the Precambrian gneisses were observed. One contact is exposed in the bed of the brook about 600 yards southeast of the road junction south of the Sand Hills. Here, alaskite is overlain by Hardyston sandstone and conglomerate with a N. 50° E. strike and 30° NW. dip. The sandstone also extends downward like sandstone dikelets into joint-filled fissures in the alaskite. The sandstone is overlain by Kittatinny Limestone with an east-northeast strike and a dip of 15° NNW. These beds are part of continuous formations whose strike cuts across the foliation and layering of the underlying Precambrian rocks. Great jumbled blocks of Kittatinny Limestone and sandstone from the Hardyston occur on the ridgetop (1,040 ft) 1 mile southwest of Vernon and on the spur at 1,000–1,050 feet altitude about 0.9 mile a little south of east of McAfee School. The blocks are out of place but are present in such size and quantity that they may actually represent outliers disturbed more or less in place by the overriding Pleistocene ice. If so, then the block of Precambrian rocks has been tilted, probably by faulting, so that the surface of unconformity dips northwest.

At the junction of the farm roads about 2,000 feet south-southeast of Vernon School (Wawayanda quadrangle), patches of quartz conglomerate rest unconformably on the Precambrian gneisses.

A third locality is in the town of Franklin on the west side of the main business street. Here, Hardyston Quartzite with a northeast strike and northwest dip overlies a coarse granite pegmatite vein and granitic gneiss with a more northerly strike and a steep southeast dip. The basal part of the quartzite is conglomeratic in part and contains abundant feldspar. About 25 feet higher up, it grades into Kittatinny Limestone.

Several localities have also been observed where the Hardyston sandstone rests unconformably on the Franklin Limestone.

At the abandoned limestone quarry just south of the southwest end of the mine at Franklin Furnace, there is a complex consisting of a feldspathic quartz-rich rock, an aphanitic material and a breccia in the Franklin Limestone. The complex has both a dike and a sill-like form. The origin of this complex has long puzzled geologists and has been discussed by Milton (1939), who interpreted the feldspathic quartz-rich rock as a modified granite intrusion and the aphanitic rock as a partial sodic replacement of the granite. The chemical and mineralogical composition of these two rocks, as determined by Milton, is given in table 24.

We here follow Wolff and Brooks (1897) in interpreting the feldspathic quartz-rich rock as a sandstone related to the Hardyston Quartzite which has worked down into solution fissures in the Franklin

TABLE 24.—*Chemical and mineralogic compositions of feldspathic quartz-rich rock and aphanite*

[After Milton, 1939, p. 171]

	1	2
Chemical composition		
SiO_2	82.00	77.44
Al_2O_3	5.19	8.33
Fe_2O_3	.13	1.22
FeO		Trace
MgO	2.05	1.50
CaO	2.19	.52
Na_2O	.28	3.84
K_2O	3.56	1.58
H_2O-	.05	1.07
H_2O+	.64	
TiO_2	.16	.20
CO_2	3.25	.51
P_2O_5		Trace
F	.47	
S	.43	2.05
Fe (as pyrite)	.40	1.36
	100.80	99.62
(O correction for F)	.20	
Total	100.60	
Mineralogic composition computed from analyses		
Quartz	66.5	46.9
Microcline	18.6	8.9
Albite	2.6	32.5
Muscovite	4.1	.9
Magnesian calcite	6.7	1.2
Fluorite	1.1	
Pyrite	1.0	3.4
Rutile (with sphene)	.2	0.2
Chlorite	.3	5.1
Sphalerite		Trace
Total	101.1	99.1

1. Felspathic quartz-rich rock; (?)Hardystone Quartzite.
2. Aphanite; (?)dike or sill of quartz keratophyre in (1); slightly contaminated by incorporation of sandstone and carbonate.

Limestone. The aphanitic rock we consider a quartz keratophyre dike and sill slightly contaminated by incorporation of sandstone and carbonate. The fluorite, pyrite, and sphalerite are due to hydrothermal vein introduction after formation as the quartz keratophyre. The feldspathic quartz-rich rock, examined with the microscope, consists of rounded grains of quartz with associated microcline and microperthite and a secondary dusty cement. Thin sections of normal Hardyston Quartzite overlying granite cannot be distinguished from those of the feldspathic quartz-rich rocks of the Franklin quarry. The feldspathic quartz-rich rock also locally contains angular fragments of Franklin Limestone.

Milton quotes (1939, p. 167–168) as follows from a letter to him by Allen Pinger of the New Jersey Zinc Co. giving supplementary data on the subsurface relationships of these rocks.

Drill holes were drilled a number of years ago, along an east-west line about 1,500 feet north of the quarry. In several of these holes "quartzitic" material was found * * * of character similar to that found in the quarry. There is a pronounced east-west slip, or fault, traceable for a distance of about 400 feet across the quarry. The slip dips to the north at an observed angle of about 25° * * *. All the occurrences of "quartzite," "arkose," "breccia" and dark aphanitic material were found to be directly associated with this slip * * *. The (aphanitic material) is very irregular in shape and has distinctly intrusive-border characteristics * * *. The occurrence of the quartzitic material in the drill holes is found at about the same depth in each hole, and the depth correlates well with the observed dip of the slip in the quarry with which the material appears to be associated.

Pinger thus writes of the aphanitic material as being irregular in shape and having distinctly intrusive-border characteristics. He has also implied a similarity in character between the feldspathic quartz-rich rock and the Hardyston Quartzite; he ruled out acceptance of this latter concept, however, because of the depth to which the feldspathic quartz-rich rock extends and the distance below the projected surface of the base of the known Hardyston. Neither of these objections, however, seems valid to us. There is some evidence to question whether the dip of the Hardyston Quartzite as found in the town at Franklin if projected upward to the southeast will truly give its former probable position. Residual relics of the Hardyston Quartzite are actually still found on the present surface of the Franklin marble belt, and one locality has been described by Wolff and Brooks (1898) at the southwest end of hill 70 about 1,100 yards south of Hardistonville. Again, at a quarry about 900 yards northwest of Rudeville, there is a relic of conglomerate and quartzite from the Hardyston on the surface of a marble within the Franklin Limestone. In the adjacent quarry face is a foot-wide tabular sheet of quartzite that extends for a depth of 40 feet down into the marble and has a strike of N. 75° W. and a dip of 60° E. The quartzite is at an angle to the banding of the marble. The layering of the quartzite is parallel to the walls. It is thus possible that the present surface of the Franklin Limestone is at, or but little below, the base of the Hardyston. Potsdam Sandstone fissure fillings in marble are known to extend to depths of several hundred feet, as has been shown by diamond drilling in the Gouverneur quadrangle of northwestern New York. There, also, solution holes in marble have been found to depths of 1,100 feet at the Balmat mine. In view of these data, the depth of occurrence of the feldspathic quartz rock in the Franklin Pond quarry seems consistent with its being a root or fissure filling of Hardyston Quartzite.

Wolff and Brooks (1898) also described a feldspathic quartz-rich rock layer in the marble at the Franklin Pond quarry and interpreted it as a sand filling of a cavity. This dikelike mass is not now exposed. They also noted that the stratification was parallel to the steep walls of the band of quartzite. Milton raises this point as an argument against the quartzite dike being derived from sand filling of a fissure. It may be noted that the quartzite layer at Rudeville also had a stratification parallel to the steep walls. This is admittedly a puzzling phenomenon but cannot offset the positive sedimentary character of the rock. One can think of the stratification as flow planes developed in a sludge of sand that had sufficient water for mobility and abruptly entered an open fissure. This might be a reasonable explanation if the phenomenon were not too common.

STRUCTURE

The major structures of the Precambrian rocks are a group of northeast-trending folds and faults. The principal folds are from northwest to southeast, the Pimple Hills syncline, Beaver Lake anticline, Ryker Lake syncline, St. Paul anticline, and the northeast end of the Weldon Brook anticline. All the folds plunge about 10°–30° NE., commonly 20°–25° NE. From northwest to southeast the longitudinal faults are the Hamburg, Zero, and East; the Morris Lake and Sand Hills fault zones; and the Vernon and Reservoir faults. All strike northeast about parallel to the trend of the formations, but, in detail, cross the layering and foliation of the Precambrian rocks at a small angle. The position of the Zero fault has been taken largely from the map of Hague and others (1956).

There are also cross and oblique faults. Generally, these are shorter and have displacements of smaller magnitude than the longitudinal faults. Other longitudinal, cross, and oblique faults within the Precambrian rocks probably occur but have not been recognized. The longitudinal faults are characterized by a brecciated zone several hundred feet to 1,500 feet wide. The rocks in these zones are shattered, fractured, slickensided, and full of epidote veinlets. Although the epidote veins are more abundant within the fault zones, some secondary epidote occurs as much as a mile from the major faults.

The longitudinal faults of the New Jersey Highlands have been classified as thrust faults of Paleozoic age, as normal faults of late Triassic age, and as of uncertain nature and age. Unequivocal longitudinal thrust faults of post-Ordovician age are represented by the Jenny Jump Mountain fault described by Bayley, Salisbury, and Kümmel (1914, p. 19) and the Storm King-Breakneck fault described by Berkey and Rice (1921, p. 77, 114). The Jenny Jump Mountain fault involves a mass of gneiss overthrust from the southeast upon a faulted anticline of Kittatinny Limestone. The Storm King-Breakneck fault dips 45° SE. Granite and gneiss have been overthrust on slate and limestone. Longitudinal faults such as these are known to be characteristic of the Paleozoic periods of deformation (Taconic, Acadian, or Appalachian; King, 1951, p. 104–115) and not of earlier periods.

Longitudinal normal faults of post-Newark age form the southeast border of the New Jersey Highlands and bring rocks of the Newark Group in contact with the Precambrian gneiss. On the geologic map of the Raritan quadrangle, north of Gladstone, one of these border faults extends into the gneiss of the highlands. The rocks on the southeast side of faults of this type and age are downdropped relative to those on the northwest. Within the highlands, however, the direction of dip and the nature and age of most of the longitudinal faults have not been determined.

PIMPLE HILLS SYNCLINE

The structure of the Precambrian rocks in the area northwest of the Zero fault is based, in part, on the geology of the adjacent quadrangle (Hague and others, 1956, p. 464). This work shows that the Pimple Hill syncline is an overturned isoclinal fold. Thus, the rocks southeast of the axis of the Pimple Hills syncline are in reverse stratigraphic succession, that is, overturned.

BEAVER LAKE ANTICLINE

The core of the Beaver Lake anticline is quartz-oligoclase gneiss, which is overlain successively by the syenite gneiss and metasedimentary quartz-potassium feldspar gneiss. The southwestern part of the anticline is complicated by the Morris Lake fault, but both syenite gneiss and quartz-potassium feldspar gneiss occur in appropriate positions southwest of Morris Lake. The northeast nose of the anticline is characterized by phacolithic masses of quartz-microcline granite and hornblende granite and alaskite.

The anticline is slightly overturned to the northwest. The northwest flank dips 70°–85° SE. (commonly 80°), and the southeast limb dips 45°–70° SE. (commonly 60°–65°). Planar foliation is well developed southwest of Heaters Pond and on both flanks of the anticline. However, foliation is elusive in the core of the anticline northeast of Heaters Pond where lineation is more strongly developed. Lineation is especially dominant along the crestal zone of the long, gently plunging nose of the anticline. The lineation plunges in general 20°–30° NE. in the area northeast of Silver Lake, and 10°–20° NE., southwest of the lake.

EDISON SYNCLINE

The axis of a syncline with a northeast plunge of 30°–40° extends southwest from the swamp east of Edison. This syncline is cut off at both ends by the Vernon fault. The simplicity of its structure is complicated by intrusive hornblende granite, alaskite, and pyroxene granite southwest of Lake Acquackanock (pl. 2).

RYKER LAKE SYNCLINE

A major synclinal structure extends from Ryker Lake to the northeast. The belt of pyroxenic gneiss through Pine Swamp probably represents the keel of this syncline. The northeastern part of the northwest limb of the syncline is cut off by the Vernon fault. The southwestern part of the syncline is disturbed by the intrusive pyroxenic granite.

The metasedimentary gneiss on the southeast flank of the Beaver Lake anticline and northwest of the Vernon fault may originally have been continuous with the Ryker Lake syncline. This possibility is suggested by the way in which the metasedimentary gneiss turns east to meet the fault southwest of Edison.

The syncline shows slight overturning to the northwest. The northwest limb generally dips 50°–60° SE., whereas the southeast limb dips 70°–85° SE. (generally about 80°–85°). Lineation is strong in the trough of the syncline. Generally, it plunges about 30° NE. in the area to the southwest and about 10°–15° NE. at the extreme northeast.

MOUNT PAUL ANTICLINE

The rocks in the area of Mount Paul are poorly exposed, but enough strikes and dips of foliation were obtained to clearly indicate an anticline with a northeast strike. The metasedimentary gneiss of the anticline are in abrupt contact with pyroxene granite on the southwest. A transverse fault is inferred to account for this abrupt discontinuity in lithology and structure. The axis of the Mount Paul anticline has been traced to the northeast for at least 1 mile beyond the limits of the area—that is, to the area just south of Canistear Reservoir in the Newfoundland quadrangle. The zones of pyroxene gneiss northwest and southeast of Holland outline the northeast-plunging nose of the fold along the east edge of the area. The pyroxene gneiss on both limbs of the anticline has a dip of about 85° SE., indicating a symmetrical tight fold slightly overturned to the northwest. The axis of the fold plunges about 25° NE.

The belt of pyroxene granite through Holland along the axis of the Mount Paul anticline is a conformable sheet which terminates just northeast of the east boundary of the Franklin quadrangle.

WELDON BROOK ANTICLINE

The northeast-plunging nose of an anticline occurs on the west side of Bowling Green Mountain. The core of the anticline is a phacolithic mass of hornblende granite. A zone of migmatitic amphibolite clearly defines the northeast nose of the fold. The fold plunges about 30° NE. The northwest limb of the fold is cut off by the Reservoir fault.

LINEATION

Lineation results from dimensional orientation of minerals or mineral aggregates. It is exceptionally well developed along the axes and plunging noses of the folds. Planar foliation may be elusive or indeterminate where lineation is strongly developed. In general, the lineation is parallel to the trend of the major folds, and plunges 10°–30° NE. Very locally and rarely it plunges to the southwest.

Hague and others (1956, p. 461) refer to local areas where the lineation deviates from the general trend. For example, in the Sterling mine and the area just to the west, the lineation plunges 25°–50°, S. 80° E. Similarly, the lineation in the Edison belt of mineralized rocks between the Davenport and Iron Hill mines has an abnormally steep plunge of 50°–55° NE., whereas 20°–30° is normal.

EMPLACEMENT OF IGNEOUS ROCKS

The major mass of syenite orthogneiss has the structural form of a phacolith. It is also reasonable to visualize it as a folded laccolithic sheet. The occurrence of a hornblende facies in the anticlinal nose and pyroxenic facies on the flanks is consistent with either hypothesis. The outlying layers of syenite gneiss might be either folded sills or satellite phacolithic emplacements. Lack of evidence of a fine-grained border facies, such as is characteristic of laccoliths and sills, makes a phacolithic mechanism seem more probable. In any case, the syenite bodies were deformed and metamorphosed after their emplacement.

The microantiperthite series of rocks including pyroxene syenite, pyroxene granite, and pyroxene alaskite have a texture resulting from magmatic crystallization. They must have been emplaced after the period of orogenic deformation which folded and deformed the metasedimentary rocks and syenite sheets of the Beaver Lake anticline. They occur both as conformable sheets and lenses. The large bodies extend into other quadrangles where mapping has not been completed, so that data are inadequate to determine their structural relationships. A phacolithic mass of pyroxene granite and syenite was observed, however, along the extension of the Ryker Lake syncline to the northeast in the Wawayanda quadrangle. Insofar as mapped, these bodies are conformable with the country rock. The albite-oligoclase granite mapped by Sims (1953, pl. 24) in the Dover district is associated with microantiperthite alaskite and is believed to be a facies of the pyroxene microantiperthite granite series. It occurs both as conformable sheets and as phacolithic masses. The pyroxene microantiperthite series of rocks of the Franklin-Hamburg area were probably also emplaced in previously folded rocks, both as conformable sheets and as phacoliths. Their emplacement probably occurred during, or at the end of, a renewed period of deformation.

The age relationship of the hornblende granite and related rocks to the pyroxene syenite, granite, and alaskite series is unknown. The hornblende granite and alaskite are believed to be younger than the syenite gneiss. The hornblende granite and associated alaskite have been emplaced as a compound phacolith in the plunging nose of the Beaver Lake anticline and as a synclinal phacolith in the nose of the Edison syncline. The southern part of the granite south of Edison syncline may cut across the syncline at a small angle. The hornblende granite of Bowling Green Mountain is the northeast end of an anticlinal phacolith. The hornblende granite and alaskite also occur as conformable sheets. The emplacement of this series of granitic rocks as conformable sheets and phacoliths is common throughout the rocks of the highlands (Lowe, 1950; Hotz, 1953, p. 186, 191–192; Sims, 1953, 265–266;

Hague and others, 1956, p. 472). Some of the alaskite shows no evidence of postconsolidation deformation and must have been intruded under relatively quiet tectonic conditions. Thus, the hornblende granite and related rocks were probably emplaced in folded rocks syntectonic with very late stages of deformation.

EAST FAULT AND MORRIS LAKE FAULT ZONE

The fault just east of Ogdensburg (pl. 1) has been mapped as the East fault by Hague and others (1956). Southwest of Ogdensburg, the fault splits into a wide fault zone which is here called the Morris Lake fault zone. About a mile northeast of Franklin the East fault has displaced Paleozoic limestone against the Franklin Limestone. Southwestward, the fault zone cuts across the layering of the Precambrian rocks at a slight angle; it leaves the Franklin Limestone, enters gneiss just northeast of Ogdensburg, and then crosses the successive rock layers forming the west limb of the Beaver Lake anticline until it enters the oligoclase-quartz gneiss of the anticlinal core. North of Lake Saginaw, sillimanitic microcline-quartz gneiss, similar to that on the east limb of the anticline at Edison, is faulted against the oligoclase-quartz gneiss. Farther southwest across the Raritan quadrangle (Bayley and others, 1914), the fault zone changes into two well-defined faults with a broad (as much as 2 miles wide) grabenlike trough of Paleozoic rocks downfaulted between them. As shown on the State geologic map (Lewis and Kümmel, 1910–1912), this block of Paleozoic rocks and the bounding faults is continuous southwest almost completely across the State.

The Morris Lake fault zone—that is, the zone within which the rocks are fractured and slickensided, is about 1,000–1,200 feet wide. The zone is best exposed in the area between Lake Saginaw and Morris Lake. Although within this zone some of the rock remains uncrushed, most of it is intensely crackled, slickensided, and mylonitized. Secondary epidote has formed along the slickenside fractures, and epidote and chlorite are common alteration products in adjacent rocks. Adjacent to the junction of the Sunset Lake and Sparta Glen Brooks is a zone about 30 feet wide of white quartz containing breccia fragments of granite and a few druses.

The rocks on the northwest side of the fault zone have been downfaulted relative to those on the southeast. This is indicated by the downfaulting of Paleozoic beds northeast of Franklin and by the absence of the syenite sheet between the sillimanite-microcline-quartz gneiss and the oligoclase-quartz gneiss north of Lake Saginaw.

SAND HILLS FAULT ZONE

Just south of Sand Hills in the Hamburg quadrangle, a fault striking southwest has offset the unconformity between the Paleozoic beds and the underlying Precambrian rocks. The Cambrian and Ordovician beds dip gently northwest. This may be part of a much longer fault zone. Breccia and mylonite were observed in the hornblende syenite gneiss near the south boundary of the Hamburg quadrangle in line with the southwest projection of the Sand Hills fault. The sharp turn to the northeast along Franklin Pond Creek in the Franklin quadrangle may be due to a northeast-striking fault. The steep slope northeast of Ogdensburg within quartz-oligoclase gneiss may mark the southwest trend of this fault. The rocks along the deeply incised tributary creek about 0.5 mile west of the community of Beaver Lake are much sheared. These inferred fault zones are in the general trend of the Sand Hills fault and may represent parts of the same fault zone.

VERNON FAULT

The Vernon fault starts just northeast of Vernon, N.J. (in the Wawayanda quadrangle), and extends south-southwest along the line of the Vernon-Stockholm road, across the Wawayanda quadrangle into the Franklin quadrangle. The east side is relatively downfaulted as indicated by the offsetting of the Paleozoic beds at Vernon. Throughout its length from Vernon across the Franklin quadrangle, the fault is very poorly exposed. The presence of some shattering and slickensiding within the zone, and the abrupt changes in lithology and discordances of structure across the zone, provide the best evidence for the continuation of the Vernon fault across the studied area. The actual magnitude of displacement may be moderate or small on the fault.

The local discordance of structure and sharp diminution of width of the sheet of syenite gneiss north of Tamarack Lake seems best explained as the result of a subordinate oblique fault that offsets the major Vernon fault.

Southeast of Edison the Vernon fault is bordered by intrusive pyroxene granite. The intrusion of the latter may have been controlled by a preexisting structural break. Structural relationships along this zone are uncertain, but they do suggest that both the emplacement of igneous rocks and post-Paleozoic faulting may have been controlled to some extent by older Precambrian structures.

RESERVOIR AND LONGWOOD VALLEY FAULTS

The Reservoir fault is a major longitudinal fault along the west side of the Oak Ridge Reservoir and extending southwest to Bowling Green Mountain and the south border of the area. It has been traced northeast into New York State. The fault has been traced to the southwest to the adjacent Dover quadrangle. Where exposed, the fault zone is characterized by shattered rock with epidotic veinlets, slickensides, mylonite, and some calcite. The fault transects the foliation and layering of the Precambrian rocks at a slight angle.

The Longwood Valley fault forms the southeast side of Bowling Green Mountain and extends southwest for many miles.

The geologic relationships of the Paleozoic beds to the Precambrian along the Reservoir fault have been mapped in detail by Kümmel and Weller (1902). The following discussion is based on their work. The fault transects at a small angle the northwest limb of a major syncline in the Paleozoic rocks. The amount of relative downthrow increases so that successively younger formations abut against the Precambrian to the northeast. The major structure on Bowling Green Mountain (fig. 3) is an anticline in beds of the Green Pond Conglomerate of Silurian age overlying the Precambrian and plunging 35°–50° NE. At the western end of the mountain, the conglomerates are folded up against the gneiss in a sharp syncline. The western limb of the syncline dips 75°–90° SE., although in the hill

FIGURE 3.—Bowling Green Mountain (lower left) and Green Pond Mountain (lower right); both underlain by Green Pond Conglomerate. The cultivated lowlands are underlain by shale and limestone of Paleozoic age.

southwest of Russia it is overturned for a length of 0.5 mile with a dip of 80° NW. Kümmel and Weller interpret the Reservoir fault as reverse type with a dip to the northwest. This interpretation would be consistent with the overturned relationship of the Green Pond Conglomerate in the hill southwest of Russia. A diabase dike near Russia similar to the Triassic diabase (Milton, 1947) is slickensided and altered. This suggests that the Reservoir fault is post-Triassic.

TRANSVERSE FAULTS

Subsurface exploration and development of the magnetite ore bodies in the highlands have shown that cross faults are quite common, but direct evidence for them is rarely exposed at the surface. Their presence may be indicated or inferred by apparent horizontal offsets in the gneiss or ore deposits, by abrupt changes in lithology along the line of strike, or by geomorphic features. Detailed data for the workings on the magnetite bodies are not available for this area. The discussion by Sims (1953) on transverse faults in the Dover district provides an indication of the kind of phenomena that probably also occur in the Franklin area. According to Sims, the transverse faults are high-angle normal faults that generally dip to the south. They are grouped into two classes. The first class includes the major transverse breaks that trend N. 70°–80° W. and dip 60°–70° S. The relative movement along the surface of dislocation is always to the right as one faces the fault plane. The second class of transverse faults strikes more northwest and generally has a steeper dip, commonly being near the vertical. With rare exceptions, these breaks also have an apparent horizontal displacement to the right.

Outcrops showing many slickensided shear planes or fault breccia with a west-northwest strike have been found in several of the stream valleys along the northwest escarpment of the Precambrian rocks between Ogdensburg and Sand Hills. The courses of some of these streams probably have been localized by cross faults. In addition, intensely slickensided gneiss is present in outcrops along the road northwest of Heaters Pond; slickensided shatter zones with a N. 60° W. strike and 50° S. dip were noted on the north side of Franklin Pond Creek 0.6–0.7 mile southwest of Beaver Lake road junction; a fault breccia, 1 foot wide, with a N. 65° W. strike and steep dip, and crushed and mylonitized gneiss occur in the creek valley 1.0 mile northwest of Silver Lake.

A fault oblique to the trend of the formations occurs just south of Mount Paul. The fault strikes about S. 80° W. There are no outcrops along the line of the fault itself; its position is based on the contrast in the nature of the rocks and the structure on opposite sides of the valley.

MAGNETITE DEPOSITS

Deposits of economic value in the area studied include nonmetallic materials, such as marble, gneiss, granite, sand and gravel, and metalliferous ore bodies, such as the zinc deposits of the Franklin and Sterling mines and numerous magnetite deposits. Brief descriptions of the zinc and nonmetallic deposits may be found in the reports by Spencer and others (1908, p. 25–27) and by Hague and others (1956, p. 469–473). A description of the granite pegmatite lenses southwest of Franklin as a potential source of feldspar is given by Lodding (1951, p. 31–44). Graphite occurs in minor amounts in much of the hypersthene-quartz-oligoclase gneiss of the Ryker Lake syncline. It is present in large amounts, perhaps locally as much as a few percent, in a pyrite-bearing gneiss layer just northwest of the southwest arm of Oak Ridge Reservoir. Prospect pits have been sunk in this area. Similar gneiss crops out on the roadside about 0.8 mile northwest of the road forks at Russia. Hague and others (1956, p. 447–448) describe several localities in the northwestern part of the Franklin quadrangle where graphitic gneiss layers are found, but no graphite production has been reported from the Franklin quadrangle.

The magnetite deposits near Edison where the largest amount of ore has been produced are the principal subject of this report. Only brief reference will be made to the other magnetite ore bodies. No work has been done on them for many years, the openings are all flooded and partly caved, and no more information is available than that previously presented by Bayley (1910).

The magnetite deposits are in quartz-potassium feldspar gneiss, amphibolite and pyroxenic gneiss, and to a minor extent in marble and skarn. A minor occurrence, of different origin, are the local concentrations of titaniferous magnetite and ilmenite in granite pegmatite.

The area has been flown with an airborne magnetometer, and a total-intensity magnetic-contour map has been published (Henderson and others, 1957a, b). There is an exceptionally intense positive magnetic anomaly over the belt of magnetite mineralization through Edison. This anomaly reemphasizes the large volume of low-grade magnetite that must underlie this zone. The magnetic data do not indicate the location of any magnetite deposit not previously known. This area was thoroughly prospected by dip needle in the last half of the 19th century.

The pyroxene granite in the contact zone with the belt of metasedimentary gneisses about 0.35 mile southwest of Russia is slightly higher than normal in magnetite and ilmenite, that is, about 4 percent (by volume) of these oxides. The belt gives rise to a positive magnetic anomaly on the aeromagnetic map. The anomaly is probably due to the few percent of disseminated magnetite in this wide zone, rather than to a narrow zone with an economic concentration of magnetite. Similarly, a facies of the pyroxene syenite in a belt striking northeast from Woodport (Dover quadrangle), which has a higher-than-normal amount (± 6 percent by volume) of magnetite and ilmenite, also gives rise to a positive magnetic anomaly on the aeromagnetic map.

ECONOMIC POSSIBILITIES

Only high-grade ore shoots have been economically worked in this area. No new information on these shoots, in addition to that given by Bayley (1910), is available. Bayley (1910, p. 305) concluded, from his brief descriptions of the Dodge, Ford, and Schofield mines, that these mines were not exhausted and that under favorable conditions they may again be made to yield a great quantity of ore. None of the other deposits are suggested by him to afford favorable possibilities for substantial production.

There is a substantial volume of low-grade magnetite ore in the belt of gneiss through the old mines northwest and southwest of Edison. The grade has been too low to work economically in the past, except for local high-grade shoots. Increasingly improved methods of mining and mineral separation may, however, eventually permit production from these deposits under favorable economic conditions.

MAGNETITE IN QUARTZ–POTASSIUM FELDSPAR GNEISS

SHERMAN-BUNKER DEPOSITS

The Sherman-Bunker magnetite mine is in the southwestern part of the Franklin quadrangle, about 2.0 miles southwest of Mahola, just north of Lake Saginaw (pl. 1). The deposits were worked from four small opencuts from which an indeterminable amount of ore was removed.

The Sherman-Bunker deposits are in quartz-potassium feldspar gneiss, similar to the host lithologic type at the Edison mine. Detailed plane-table mapping and a dip-needle survey showed (fig. 4) that they are confined to a north-south-trending zone about 1,000 feet long and 150–200 feet wide. The zone coincides with the axis and west limb of a gently south-plunging syncline, whose east limb has been faulted off by a major north-trending steep-angle fault. The synclinal zone is characterized by numerous minor folds and rolls. Along the fault, both the host rock and wallrock are severely fractured and slickensided. Epidote fills many of the fractures.

Medium-grained magnetite-quartz-potassium feldspar gneiss, which is virtually identical with the principal variety of gneiss in the Edison deposits, is the host rock (see table 28, Nos. 154, 153, 151). By an increase in the proportion of magnetite, the gneiss grades into ore in which the magnetite is generally disseminated but may occur in layers ¼–½ inch thick. Other minerals in the gneiss, in order of their volumetric importance, include sillimanite, chlorite, biotite, rutile, ilmenohematite, plagioclase, sericite, epidote, apatite, zircon, garnet, and hemoilmenite.

The magnetite-quartz-potassium feldspar gneiss is interlayered with chlorite-quartz-feldspar gneiss, hornblende granite, fluorite-bearing alaskite, and seams and nests of pegmatite. These rock types make up the principal wallrocks of the deposits.

Assay data showed that three samples of ore ranged from less than 10 to 17 percent and averaged 12 percent Fe in magnetite. The deposit seems to be very low grade and probably does not warrant further consideration.

OTHER OCCURENCES

Many magnetite concentrations occur in the belt of quartz-potassium feldspar gneiss on the southeast flank of the Beaver Lake anticline. This belt includes the zone of mines near Edison. The latter will be described in detail in the section on the Edison magnetite deposits.

Quartz-potassium feldspar gneiss also occurs locally as included lenses between granite sheets, as just southwest of Lake Acquackanonk and south-southwest of Edison Pond (pl. 1). Prospect pits have been sunk in some of these outcrops. The magnetite concentrations appear to have been small and lean.

MAGNETITE IN AMPHIBOLITE AND PYROXENIC GNEISS

Magnetite ore shoots in amphibolite and other mafic gneiss have been mined at a locality indicated by the shaft symbol 0.5 mile south of Edison; in the Dodge-Ford-Schofield mine belt which extends northeast from the lake about 2,000 feet southeast of Woodport triangulation station; and in the Duffee belt at the south border of the Franklin quadrangle about 0.6 mile south southwest of the lake (pl. 1). The amphibolite com-

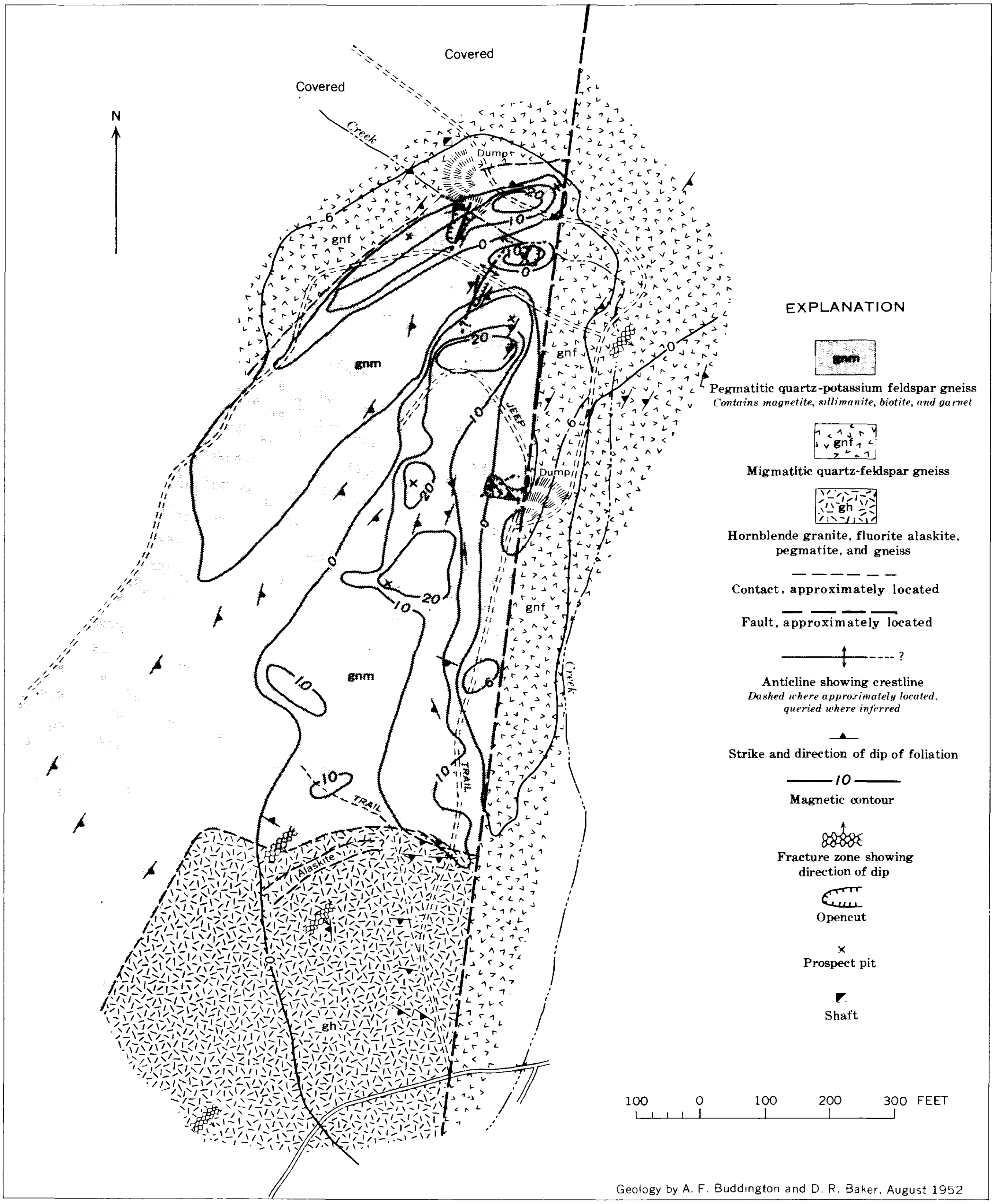

FIGURE 4.—Geologic map of the Sherman-Bunker magnetite deposits, Sussex County, N.J.

monly contains granite pegmatite lenses and is locally bordered by, or associated with, contaminated granite. The workings of the Dodge-Ford-Schofield belt have been described by Bayley (1910, p. 301–307), and one of the Duffee workings is mentioned by Bayley (1910, p. 300).

The immediate host rock for many magnetite deposits in the New Jersey Highlands is amphibolite with associated granite pegmatite veins. The Ringwood and Sterling Lake group of mines, described by Hotz (1953, p. 216–238), and the Hibernia mine, described by Buddington (in Sims, 1958), are examples. The Weldon belt of mines, just south of the Franklin quadrangle, along Weldon Brook in the Dover quadrangle is also in amphibolite.

The amphibolite lens with ore south of Edison is enclosed in pink alaskite, and the Weldon group of mines is in a belt of amphibolite that is adjacent to pink alaskite. The amphibolite layers containing the Dodge-Ford-Schofield and the Duffee ore bodies are enclosed in pyroxene microantiperthite alaskite. The Hibernia ore body (Boonton quadrangle) is enclosed in albite alaskite related to pyroxene microantiperthite alaskite. The ore body at the Williams mine (Wawayanda quadrangle), 1.5 miles east-northeast of Mud Pond (Hamburg quadrangle), and at the Hurd mine (west of Weown Lake on the Dover quadrangle) are in pyroxene granite or alaskite.

FORD-SCHOFIELD-DODGE MINES

The Ford, Schofield, and Dodge mines are in the same zone of magnetite mineralization in the southwest corner of the Franklin quadrangle, Morris County, N.J. The mineralized zone crops out, dips steeply southeast, and trends northeast conformably with the foliation of the country rocks for a distance of about half a mile. A dip-needle survey indicates the richest part of the mineral zone to be only about 50 feet wide on outcrop, whereas the entire mineralized zone is about 200 feet in outcrop breadth. The Dodge mine, which was worked from a single shaft, is at the extreme southwest limit of the mineralized zone. The Schofield mine is at the extreme northeast limit of the mineralized zone and was worked from a single shaft and a long opencut. The Ford mine is in the middle of the mineralized zone about 800 feet southwest of the Schofield mine.

Bayley (1910) reports that each of the three mines was worked on two main veins. In the Dodge mine a footwall vein, 12 feet thick, is separated by 10–12 feet of lean wallrock from a hanging-wall vein 20 feet in maximum thickness. Bayley speculated that these veins connect with the Ford and Glendon veins in the Ford mine. The Glendon, or footwall, vein is 9–12 feet thick and 400 feet long. Furthermore, Bayley indicated that production in the Schofield mine was from a large lens 110 feet high and 4–6 feet thick in the hanging wall, and from an 18-foot-thick vein in the footwall, both of which are an extension of the Glendon vein from the Ford mine. In summary, Bayley's description indicates that the mineralized zone consists of two principal veins which pinch and swell along their strike and plunge, giving rise to tabular, lath-shaped ore bodies separated by lean or unmineralized bottom and caprocks as well as wallrocks.

According to Bayley, the Ford mine produced at least 95,000 tons of ore. The average grade was about 50 percent Fe. The last active mine was the Ford in 1896.

The country rock which conformably surrounds the Ford-Schofield-Dodge mineralized zone is green pyroxene granite. The host rock for the ore is a magnetite-pyroxene-quartz-feldspar gneiss. Magnetite occurs as both disseminated grains and 1/8 to 4-inch-thick layers in the gneiss. Typically, the ore consists of layers of gneiss with disseminated magnetite, alternating with layers of magnetite-rich gneiss. Both these layers contain the same minerals but in considerably different proportions. Principal gangue minerals include quartz, pale- to medium-green clinopyroxene, microcline, plagioclase, microperthite, and antiperthite in both film and patchlike intergrowths, and apatite. Magnetite is the only significant ore mineral; rutile and pyrite are minor consituents. The magnetite encloses a very small proportion of ilmenite both in the form of irregularly shaped grains along the borders and as thin blades oriented parallel to the octahedral plane of the host magnetite. No martite is intergrown with the magnetite.

In general, the ore is medium and fairly even grained, and has distinct gneissic structure in both lean and rich ore. The magnetite-rich layers are generally parallel to the foliation of the lean, adjacent pyroxene-quartz-feldspar gneiss host rock; in some places, however, thin seams of magnetite cut across the gneissic structure of the adjacent rocks at very small angles. This indicates that at least some of the magnetite mineralization (fixation) was subsequent to the formation of the foliation.

A small amount of amphibolite (hornblende-plagioclase gneiss) and pink pegmatite which contain microperthite in film and patchlike intergrowths is associated with the ore. The pegmatite definitely seems to intrude the amphibolite; however, the age relationship between the pegmatite and the ore is not apparent.

In summary, the mineralogy of the typical ore of the Ford-Schofield-Dodge mineralized zone suggests a genetic relationship with the large surrounding volume of pyroxene granite country rock. Specifically, the magnetite, pyroxene, and characteristic microperthitic and antiperthitic intergrowths of both the ore and the country rock indicate a common genesis. Therefore, it is suggested that during the emplacement and crystallization of the pyroxene granite an iron-rich hydrothermal solution formed, probably as a result of magmatic differentiation. It is believed that this ore solution entered into a replacement reaction with the parent pyroxene granite to yield the alternating layers of magnetite-pyroxene-quartz-feldspar gneiss (modified pyroxene granite) and heavy magnetite-rich layers (nearly completely replaced pyroxene granite). The magnetite deposits are, therefore, believed to be a hydrothermal replacement of the parent ore-forming rock.

ORIGIN OF MAGNETITE DEPOSITS IN AMPHIBOLITE

The magnetite ore deposits in amphibolite (and skarn) in the Ringwood-Sterling district and in the Dover district have been interpreted by Hotz (1953, p. 215) and by Sims (1953, p. 282–286), respectively, to have been formed by solutions emanating from the magmas that gave rise to the hornblende granite and alaskite. A similar origin is applicable to the ores in these host rocks in the Franklin-Hamburg district, except that the solutions in part must also have come from the magma that yielded the pyroxene granite and associated alaskite.

Hagner and Collins (1955) have interpreted the ore in the amphibolite at the Scott mine of the Sterling Lake group to be the result of metamorphic differentiation of the iron from the amphibolite and of metasomatism accompanying regional metamorphism. They assume that associated granitic rocks and pegmatite are replacements of amphibolite. On the contrary, the compositions of the titaniferous magnetite and the original feldspar (anorthoclase) of the granite, their structural relationships, and the development of metasomatic rocks in contact zones with marble, indicate their magmatic origin. In addition, the potentiality for granitic magma to yield iron solutions competent to form extensive magnetite deposits has been conclusively proven by Mackin (1947). It may be that some of the iron for the ore bodies was derived from amphibolite at considerable depths below present mine workings. On the contrary, however, if the amphibolite is derived by modification of sedimentary material, as seems probable, then iron may have been introduced into the amphibolite itself, analogous to the iron enrichment in the formation of pyroxene- and garnet-skarn replacements of marble. The hypothesis of derivation of the iron for the magnetite ore bodies, at least in large part, from fluids emanating from granitic magma has much in its favor.

MAGNETITE IN MARBLE

A vein of disseminated magnetite in marble occurs in the footwall marble of the Franklin ore (zinc) body. The vein in part followed around the synclinal trough beneath the zinc ore and in part it extends intermittently to the southwest in a zone near the contact with the northwest border of the marble. It has been described by Spencer and others (1908, p. 22–23 and economic geology map). No evidence of more detailed relationships of the iron ore and the zinc ore has been found. Another magnetite vein west of that just referred to and west of Franklin Pond occurs along the contact between a syenite pegmatite vein and marble; the gangue minerals are of skarn type. Hornblende is abundant, and a variety of garnets are reported. Aggregates of hornblende, epidote, and axinite were found on one dump.

TITANIFEROUS MAGNETITE IN GRANITE PEGMATITE

Several prospect pits have been sunk in granite pegmatite containing magnetite. The size of ore shoots in pegmatite are all small, and the ore is titaniferous. The magnetite is actually ilmenomagnetite (table 13) and that at the Goble and Woods prospects carry 3.67 and 2.75 percent TiO_2, respectively. The ilmenomagnetite of a granite pegmatite vein 0.2 mile east of the north end of Lake Acquackanonk has 1.97 percent TiO_2. All three of these granite pegmatites are in pyroxene granite or its quartz syenitic facies. Ilmenite also occurs as independent grains in association with the ilmenomagnetite.

EDISON DEPOSITS

Most of the known magnetite deposits of the Franklin quadrangle lie within the outcrop breadth of quartz-potassium feldspar gneiss (pl. 1; fig. 5) passing through Edison.

C. L. Rogers, of the U.S. Geological Survey, had previously made a reconnaissance of the outcrops near Edison in 1945, had logged the diamond-drill cores of the Pittsburgh Coke & Iron Co., and had prepared a memorandum on the work. We are under obligations to Rogers for the use of these results.

We are also indebted to W. Hildebrand of the Thomas A. Edison Co., Inc., for making available maps of the Edison area on which were shown roads, many

FIGURE 5.—Aerial photograph of the Edison magnetite deposit area. Lowlands (left) are underlain by Cambrian and Ordovician rocks; forested highlands are underlain by Precambrian gneiss. The Edison area extends southwest from the pond in the upper right part of the photograph. Ogdensburg is located at upper left.

of the surface workings and prospects, location of diamond-drilled holes, and the results of a detailed dip-needle survey, together with the results of the diamond-drill holes made by the Bethlehem Steel Co. in 1920 and the Pittsburgh Coke & Iron Co. in 1943.

The rock complex mapped as quartz-potassium feldspar and associated gneiss (pls. 1, 2) for the sake of brevity is informally referred to as the mixed-gneiss complex throughout the following discussion of the Edison magnetite deposits. The mixed-gneiss complex is a tabular sheet, about 4 miles long, which contains most of the magnetite deposits of the Franklin quadrangle. Most of the magnetite production has come from the part of the complex just northeast and southwest of Edison. To the northeast near Tamarack Lake, the mineralized zone is displaced by an oblique fault. Similar rocks occur north of the fault and extend northward. However, these have only local magnetite mineralization. At the southwest, the mixed-gneiss complex is cut out by a longitudinal fault (Vernon fault). Just northeast of Edison the mixed-gneiss complex is split by a lens of epidote-scapolite-quartz-feldspar gneiss. There, a narrow layer of the mixed-gneiss complex strikes off to the northeast where it, too, is cut out by the Vernon fault. The mixed-gneiss complex is on the southeast limb of the Beaver Lake anticline and is underlain by the sheet of pyroxene syenite gneiss which serves as a well-defined footwall.

The main productive part of the complex is shown on the large-scale map of the Edison area (pl. 2). General descriptions of all the rocks shown on this map have been given in the first part of this report. More detailed descriptions of the mixed-gneiss complex—the quartz-microcline gneiss and biotite-quartz-feldspar gneiss which form the hanging-wall rocks for much of the mineralized belt—and of the epidote-scapolite-quartz gneiss that occurs as a lens within the belt, are given below.

QUARTZ-POTASSIUM FELDSPAR AND ASSOCIATED GNEISS (MIXED-GNEISS COMPLEX)

The quartz-potassium feldspar and associated gneiss (mixed-gneiss complex) constitute a heterogeneous mixture of interlayered rock types. The predominant variety of rock is a magnetite-quartz-potassium feldspar gneiss (fig. 6). It is the immediate wallrock to magnetite-rich layers, and by an increase in percent of magnetite it appears to pass into massive ore. Subordinate rock types consist of variations of the magnetite-quartz-potassium feldspar gneiss and of interlayers that show metasedimentary affinities—that is, substantial quantities of garnet, biotite, sillimanite, and quartz.

MINERALOGY

The various interlayered rock types that make up the mixed-gneiss complex are considered as related rocks that have crystallized under virtually the same physical conditions (pressure and temperature), so that the same mineral phases were stable throughout. Because of large differences in chemical composition from layer to layer, however, the proportions of the different mineral phases vary widely.

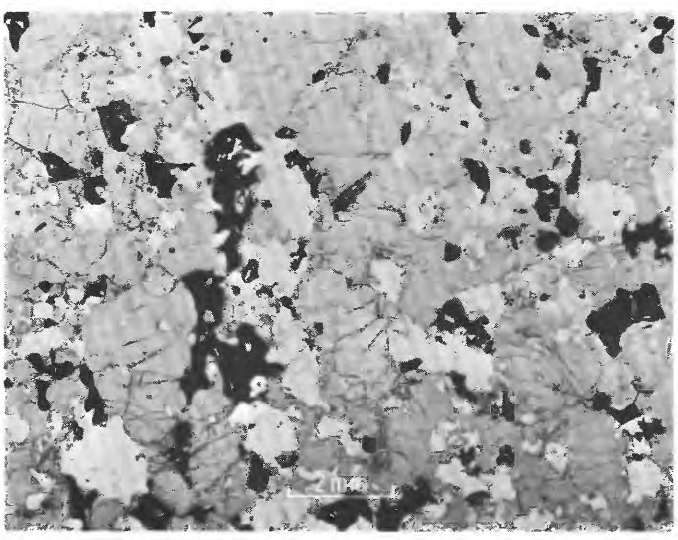

FIGURE 6.—Photomicrograph of magnetite–quartz–potassium feldspar gneiss from the Edison area. Principal minerals are xenoblastic and the grains have highly sutured boundaries. White, quartz; gray, stained potassium feldspar; black, ores. Texturally, all the minerals are equivalent. Sample 149, ordinary light.

Quartz, potassium feldspar (microcline and monoclinic potassium feldspar), and magnetite are essential minerals in nearly all the rock types. Plagioclase is virtually absent from the complex. Important varietal minerals include biotite, sillimanite, garnet, and ilmenohematite. Accessory minerals include apatite, monazite, spinel, corundum, epidote, fluorite, hemoilmenite, ilmenite, rutile, pyrite, molybdenite, bornite, chalcopyrite, and at least two varieties of rounded zircons. Secondary minerals are chlorite, sericite, epidote, and hematite (martite).

MAGNETITE–QUARTZ–POTASSIUM FELDSPAR GNEISS

The principal type of rock interlayer within the mixed-gneiss complex is composed of variable proportions of quartz, potassium feldspar, and magnetite. At one extreme of composition this important rock type is a magnetite-quartz gneiss (metaquartzite?) (fig. 7); at another extreme it is a potassium feldspar-quartz-magnetite gneiss, that is, magnetite ore. The

FIGURE 7.—Photomicrograph of magnetite–potassium feldspar-quartz gneiss (metaquartzite) from the Edison area. The principal minerals are xenoblastic and have sutured grain boundaries. Note the lenticular shape of the magnetite and the equidimensional shape of the potassium feldspar. White, quartz; gray, stained potassium feldspar; black, ores. Sample 143, ordinary light.

most abundant variety is composed of subequal proportions of quartz and potassium feldspar (about 40 percent of each) and subordinate magnetite (about 15 percent) and is a medium-grained magnetite-quartz-potassium feldspar gneiss (fig. 6).

LITHOLOGIC VARIATIONS

Other rock types within the mixed-gneiss complex are intimately interlayered with the more abundant magnetite-quartz-potassium feldspar gneiss. One or all of the minerals quartz, potassium feldspar, or magnetite form essential constituents of these rock varieties, but, in addition, biotite, sillimanite (fig. 8), and garnet are important constituents. Some of these layers are simply biotitic, sillimanitic, or garnetiferous varieties of magnetite-quartz-potassium feldspar gneiss. Others have markedly different mineral compositions and include such distinct rocks as garnet-biotite-sillimanite-quartz gneiss (metaquartzite?) (fig. 9), sillimanite-biotite-quartz gneiss, and, rarely, biotite-quartz-feldspar gneiss. Layers that represent all gradations between the predominant magnetite-quartz-potassium feldspar gneiss and these other types are abundant.

PEGMATITES

Pegmatites composed of quartz and potassium feldspar and minor amounts of magnetite, biotite, and in places, sillimanite are mixed with all the various rock types that make up the mixed-gneiss complex. Mineralogically, the pegmatites are very similar to some varieties of the magnetite-quartz-potassium feldspar gneiss; however, they generally contain much less magnetite and more potassium feldspar than the typical gneiss. The pegmatites are massive and range in grain size from coarse to extremely coarse (potassium feldspar crystals as much as 8 in. long).

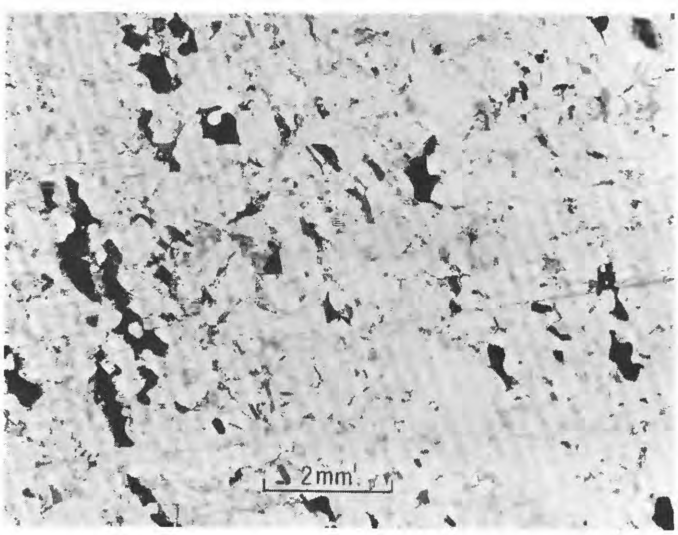

FIGURE 8.—Photomicrograph of sillimanitic magnetite–quartz–potassium feldspar gneiss from the Edison area. Magnetite, quartz, and potassium feldspar are xenoblastic, have sutured grain boundaries, and are texturally equivalent. Sillimanite is subhedral and elongate. Light gray, quartz; gray, stained potassium feldspar; dark gray and high relief, sillimanite; black, ores. Sample 145, ordinary light.

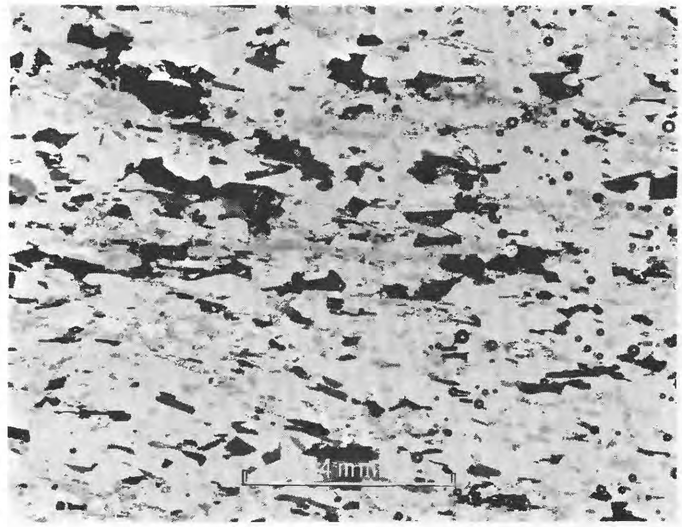

FIGURE 9.—Photomicrograph of magnetite–biotite–sillimanite–quartz gneiss (metaquartzite) from the Edison area. Magnetite and quartz are xenoblastic and have irregular grain boundaries. Sillmanite is subhedral and elongate. Note strong preferred orientation of all minerals. Light-gray, quartz; gray and high relief, sillimanite; very dark gray, biotite; black, ores. Sample B–151f, ordinary light.

Pegmatite bodies range from narrow lenses less than 1 inch wide to layers as thick as 3 feet. All the pegmatites proved to be "ductless," that is, they either lens out or fade away into the gneiss. The contacts between the pegmatites and the adjacent gneiss are abrupt but are never sharp or dikelike. At the actual pegmatite-gneiss boundary, the grains of each type are intergrown across the boundary, so that the contact is irregular and of the same nature as the irregular grain boundary contacts between adjacent layers within the gneiss.

Although the pegmatites usually are conformably interlayered with the gneiss, in places they crosscut the gneiss. Thus, the pegmatites seem to be structurally distinct from the enclosing gneiss and were probably emplaced subsequent to the development of the foliation of the bulk of the gneiss. In addition, it is critical to note that the pegmatites carry only accessory magnetite. They rarely seem to be mineralized with magnetite, as immediately adjacent gneiss may be. Indeed, the pegmatites actually crosscut gneiss that carries a fair proportion of magnetite. These observations indicate that the pegmatites were emplaced not only after the development of the predominant rock types but also after the emplacement and fixation of the magnetite. On the other hand, one pegmatite vein was observed that carried crossveins (ladder fabric) of magnetite mineralization, and pegmatites in other districts have been interpreted as having been emplaced before mineralization.

FABRIC

The magnetite-quartz-potassium feldspar gneiss as well as the other subordinate compositional types, including the magnetite concentrations, are medium grained and fairly even grained. Some fine-grained layers are present, however, and coarse-grained to pegmatitic layers are quite abundant (see discussion of pegmatites). Quartz, potassium feldspar, and all opaque oxides are xenoblastic and either fairly equidimensional or of distinct lenticular morphology (see figs. 6–8). Sillimanite is in the form of elongated needles (figs. 8, 9); biotite occurs as thin plates, and garnet occurs either as disseminated xenoblastic grains or in porphyroblastic aggregates. Potassium feldspar porphyroblasts are well developed in biotitic and sillimanitic varieties.

The magnetite-quartz-potassium feldspar gneiss has a distinct gneissic structure due to the preferred orientation of the lenticular-shaped grains of quartz and magnetite. The variety magnetite-quartz gneiss (metaquartzite?) appears considerably more deformed than any feldspathic varieties of the gneiss. Both the quartz and magnetite in this quartz-rich variety are drawn out into long thin lenticles that testify to rather intense deformation. There seems to be a definite inverse correlation between the amount of potassium feldspar in the rock and the intensity of its deformation fabric, so that samples low in potassium feldspar seem to have a stronger gneissic structure than do samples rich in potassium feldspar. (Compare figs. 6 and 7).

In the magnetite-quartz-potassium feldspar gneiss, magnetite is generally disseminated in lenticular grains. As the amount of magnetite progressively increases, the magnetite forms lenticular aggregates and thin discontinuous layers that contribute to the planar fabric. Where greatly enriched, the magnetite is in layers ¼ to 3 inches thick, which alternate with typical magnetite-quartz-potassium feldspar gneiss. The magnetite-rich layers (ore) are gneissic and have virtually the same fabric as the typical magnetite-quartz-potassium feldspar gneiss.

The rock varieties that are enriched in biotite and sillimanite may be described as schistose and as gneissic (fig. 9). Biotite plates and sillimanite needles are always oriented in the foliation plane, and the sillimanite may have a preferred linear fabric. Such rocks consist of schistose biotite- and sillimanite-rich layers alternating with gneissic quartzose-feldspathic layers.

QUARTZ-MICROCLINE GNEISS

The quartz-microcline gneiss crops out (pls. 1, 2) in the shape of a large lens with a length of 9,000 feet and maximum width of 900 feet. The body of gneiss is conformable to, and forms part of the hanging wall to, the mixed-gneiss complex. In the southwestern part of the Edison area it fingers out into the biotite-quartz-feldspar gneiss.

MINERALOGY AND COMPOSITION

Quartz and perthitic microcline are the major minerals in the gneiss. Garnet and ilmenomagnetite are the two principal accessory minerals. As with the magnetite-quartz-potassium feldspar gneiss, plagioclase is always absent in the quartz-microcline gneiss, except in some samples taken from zones closely associated with the biotite-quartz-feldspar gneiss. Other accessory minerals include biotite, zircon (euhedral), allanite, and apatite. Secondary minerals are sericite, chlorite, and, rarely, hematite (martite). Chlorite is in many places interlayered with biotite as an alteration product. Sericite may be an alteration product of plagioclase. Hematitic alteration (martite) of ilmenomagnetite is generally not present; however, some samples contain traces of martite (never more than 1–2 percent of the ilmenomagnetite is altered).

The quartz-microcline gneiss is a very uniform rock. The modal analysis of 16 samples showed the average composition to be 40 percent quartz, 51 percent potassium feldspar, 3 percent ilmenomagnetite and 3 percent garnet. It is significant to note that the variability of the mineral composition of the quartz-microcline gneiss is much less than that of the magnetite-quartz-potassium feldspar gneiss.

FABRIC

The texture of the quartz-microcline gneiss is as uniform as its composition. The unweathered gneiss is either pinkish buff or grayish white; it weathers to a rusty orange. Its texture is fine grained, equigranular, and xenoblastic (fig. 10).

The structure of the quartz-microcline gneiss is granulose (Tyrrell, 1948, p. 274). Some streaks of magnetite or porphyroblastic aggregates of garnet may impart a rough foliation to the rock. The distribution of garnet aggregates is irregular, and in zones free of much garnet the rock is very massive. Thus, the struc-

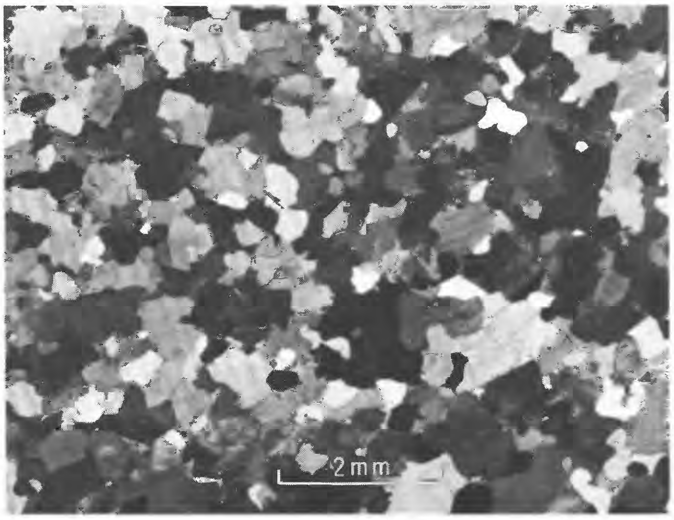

FIGURE 10.—Photomicrograph of quartz-microcline gneiss from the Edison area. Quartz and microcline are xenoblastic and equidimensional, and have smooth grain boundaries. Microcline grid twinning is visible in parts of some grains. Sample ED–177, crossed nicols.

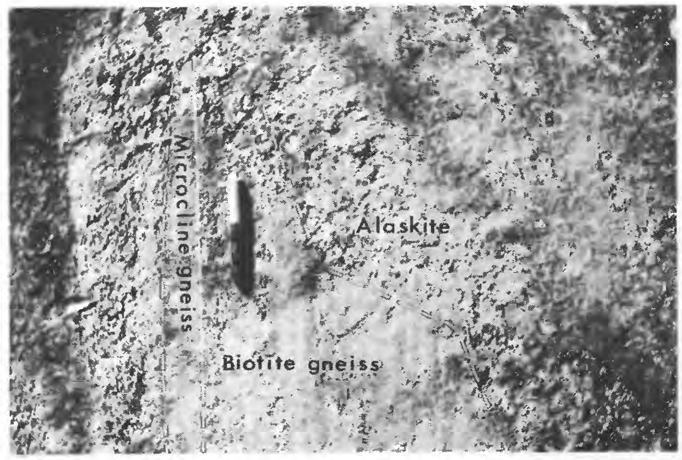

FIGURE 11.—Interlayers of biotite alaskite, biotite–quartz–feldspar gneiss, and quartz-microcline gneiss from the Edison area. The pen is parallel to foliation and rests on a layer of biotite–quartz–feldspar gneiss. To the right of the pen is a layer of biotite alaskite that cuts across foliation and weathers to greater relief. To the left of the pen is a layer of quartz-microcline gneiss that weathers to little relief.

ture of the gneiss is extremely faint and is much less developed than that of any of the adjacent gneisses.

CLASSIFICATION

The striking features of the mineralogic composition of the quartz-microcline gneiss are the high quartz content and absence of plagioclase feldspar. The composition and texture correspond closely to garnet alaskite of the Adirondacks (Buddington, 1957, table 3, p. 298) or to a granite aplite (Johanssen, 1932, p. 302).

RELATIONSHIP TO OTHER ROCKS

The contact of the quartz-microcline gneiss and the mixed-gneiss complex is conformable and very abrupt. The two rock types do not interlayer. The mineralogic changes have been studied in detail from samples taken across the contact. On the whole, the mineralogy in the two rock bodies is similar in that quartz and potassium feldspar are common essential minerals, and biotite, apatite, zircon, and magnetite are common accessory minerals. The two rocks differ most in their oxide and sulfide minerals. Pyrite and chalcopyrite are accessory minerals in the mixed-gneiss complex but are absent in the quartz-microcline gneiss. Magnetite is more abundant in the magnetite-quartz-potassium feldspar gneiss and is partially altered (about 15 percent) to hematite (martite). The magnetite in the quartz-microcline gneiss is rarely slightly altered to martite.

To the southwest, the quartz-microcline gneiss lenses out into the biotite-quartz-feldspar gneiss. In the zone just southeast of Edison (pl. 2), the two gneisses, as well as some biotite alaskite, are strongly interlayered. The representation on the Edison area map is highly diagrammatic in that the entire zone is a complex interlayered mixture of the three rock types (fig. 11). The scale of the interlayering may be small enough to be seen in small handspecimens (fig. 12). Although contacts between all three rock types are extremely abrupt, they are not dikelike. Instead, they are simply

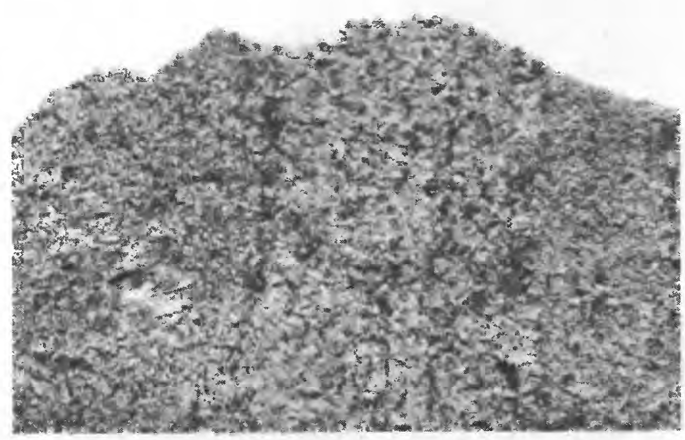

FIGURE 12.—Hand specimen Ed–2544 showing interlayering of biotite alaskite, biotite–quartz–feldspar gneiss, and quartz–microcline gneiss. Contacts between layers are abrupt but are natural grain boundaries between minerals. Layer on left is medium–grained biotite–quartz–feldspar gneiss with a monoclinic potassium feldspar (orthoclase) porphyroblast. Middle layer is coarse–grained biotite alaskite. Layer on right is fine–grained quartz–microcline gneiss. The specimen is about 6 inches wide.

the irregular interlocking grain boundaries between the minerals of the particular rock layers.

Only minor variations from the normal character of the two gneisses and the biotite alaskite are noticed in this complex interlayered zone. The textures of the three rock types remain the same as in noninterlayered zones. The coarse-grained character of the biotite alaskite is in sharp contrast to the medium-grained biotite-quartz-feldspar gneiss and the fine-grained quartz-microcline gneiss. Some of the quartz-microcline gneiss carries a few percent accessory plagioclase, but otherwise the gneiss has its normal mineralogic composition. At the contact between biotite-quartz-feldspar gneiss and biotite alaskite, altered hypersthene(?) is present. In addition, the biotite-quartz-feldspar gneiss has a small, but significant, increase in potassium feldspar content in the interlayered zone (see next section).

BIOTITE-QUARTZ-FELDSPAR GNEISS

The biotite-quartz-feldspar gneiss is a tabular mass conformable to all the surrounding rocks. It forms the hanging wall to the mixed-gneiss complex in the southwestern part of the Edison area. The biotite-quartz-feldspar gneiss lenses out into the quartz-microcline gneiss to the northeast. It has been mapped (pl. 1) for about 2 miles beyond the Edison area to the southwest where it terminates abruptly against the Vernon fault.

MINERALOGY

The essential minerals of this gneiss (fig. 13) are quartz, plagioclase (oligoclase), and slightly perthitic potassium feldspar. Biotite is the most important varietal mineral, and garnet is the major accessory mineral. Subordinate accessory minerals include hornblende, ilmenomagnetite, ilmenite (locally, hemoilmenite), rutile, apatite, zircon, hypersthene, and allanite. Secondary minerals are epidote, chlorite, sericite, rutile and hematite.

The plagioclase is in the oligoclase range of composition ($An_{20 \pm 3}$). It is partially altered to sericite and, in places, epidote. Both triclinic (microcline) and monoclinic (orthoclase) potassium feldspar are present in the biotite-quartz-feldspar gneiss. In contrast to plagioclase, the potassium feldspar is free from alteration. Quartz is highly strained but never crushed or granulated. Biotite is a reddish-brown to a very dark brown variety:

X = medium-straw yellow,
Y = dark-reddish brown, and
Z = very dark brown, nearly opaque.

It is partially altered to epidote and chlorite. Hornblende is dark green and unaltered. Hypersthene is a flesh color and is altered to a micaceous mineral and

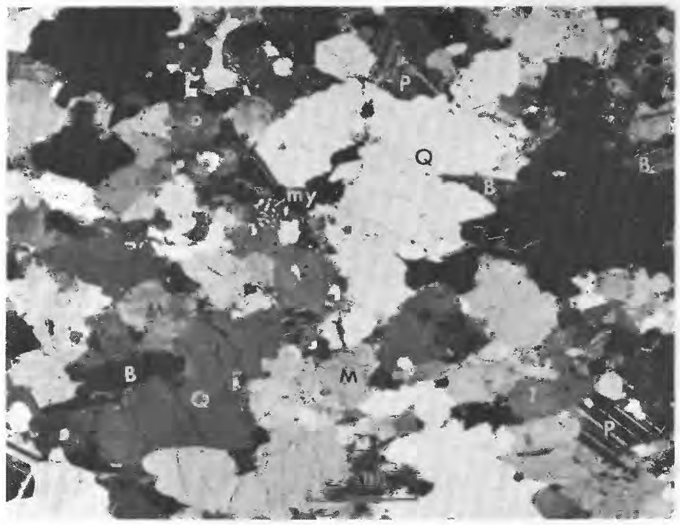

FIGURE 13.—Photomicrograph of biotite-quartz-feldspar gneiss from the Edison area. Minerals are mostly xenoblastic and have sutured grain boundaries. Note twinned plagioclase (P), faintly twinned microcline (M), large amoeboid quartz grains (Q), myrmekite (my), and biotite plates (B). Sample Ed-241a, crossed nicols.

an opaque oxide (hematite?). Hypersthene has been found only where the gneiss is closely associated with biotite alaskite and is interpreted as having formed as a consequence of this relationship. Martite is not present in the ilmenomagnetite; however, the ilmenite blades within the magnetite, and the free ilmenite grains are altered to a mottled aggregate of rutile and hematite.

FABRIC

The biotite-quartz-feldspar gneiss is grayish white to reddish buff, and is medium or coarse-medium grained, although fine-medium and coarse-grained varieties are present (fig. 13). In places, porphyroblasts of potassium feldspar (orthoclase) (fig. 12) and porphyroblastic aggregates of garnet are present. The minerals, except for zircon and apatite, are xenoblastic; quartz and feldspar may have a slightly lenticular shape; biotite occurs as thin plates oriented in the fabric plane.

The biotite-quartz-feldspar gneiss has a distinct planar structure due to the preferred orientation of biotite plates, to the lenticular shape of some quartz and feldspar, and to slight textural and compositional variations. Conformable interlayers of other rock types emphasize the structure. Locally, the biotite content becomes so high that the rock has the fabric of a biotite schist.

The structure of the biotite-quartz-feldspar gneiss is not so gneissic (in the sense that the minerals are sheared out) as is the structure of the rocks of the mixed-gneiss complex. Instead, its structure is intermediate between the nearly massive quartz-microcline gneiss and the gneissic mixed-gneiss complex.

COMPOSITION

Modal analyses of 22 samples of the biotite-quartz-feldspar gneiss show that its composition is variable. Quartz ranges from 20 to 35 percent, and potassium feldspar ranges from 0 to 62 percent as the plagioclase ranges from 79 to 25 percent. The proportion of potassium feldspar in the biotite-quartz-feldspar gneiss increases at the expense of plagioclase to the northeast where the gneiss is interlayered with the quartz-microcline gneiss.

LITHOLOGIC VARIATIONS

The biotite-quartz-feldspar gneiss is interlayered with other rock types. This interlayering may be on a scale of ½–1 inch, or it may be on a scale of 1–10 feet; it includes such rock types as biotite alaskite, quartz-microcline gneiss (figs. 11, 12), pegmatite, biotite schist, garnet-rich layers, biotite amphibolite, and pyroxene-plagioclase-hornblende skarn.

EPIDOTE-SCAPOLITE-QUARTZ GNEISS AND RELATED FACIES

The epidote-scapolite-quartz gneiss (calcium-rich gneiss) occurs in the northeastern part of the Edison area and extends northeast to the oblique fault near Highway 23 (pl. 1). The gneiss lenses out very rapidly into the mixed-gneiss complex. One zone of calcium-rich gneiss was mapped within the southwestern part of the mixed-gneiss complex; otherwise, no interlayering of the two types is apparent (pl. 2). The epidote-scapolite-quartz gneiss unit on the northwest limb of the Beaver Lake anticline (pl. 1) probably is the stratigraphic equivalent of the calcium-rich gneiss in the Edison area. The calcium-rich gneiss is always conformable to adjacent rocks.

The calcium-rich gneiss is composed of interlayered rock types that range from metaquartzites to quartz-feldspar gneisses. Intermediate types include feldspathic metaquartzites and feldspar-quartz gneisses. All these varieties are characterized by the same suite of varietal and accessory minerals which, on the whole, are calcium-rich. Usually, the feldspar-rich layers weather out more readily than the quartz-rich layers, so that the layering is emphasized by a ribbed structure.

MINERALOGY

The outstanding mineralogic feature of the calcium-rich gneiss is the high content of quartz and calcium-rich minerals. The major minerals throughout the gneiss are quartz, plagioclase, and microcline. Varietal minerals are epidote, scapolite, hornblende, pyroxene, biotite, and sphene. Accessory minerals include calcite, magnetite, hematite, garnet, clinozoisite, allanite, muscovite, apatite, and zircon. Secondary minerals are sericite, epidote (clinozoisite), hematite (martite), and two varieties of chlorite.

Quartz is always highly strained. Microcline is always very clear and unaltered and has good cross-hatched twinning. It is usually nonperthitic, but, locally, slightly perthitic microcline is present. Plagioclase varies in composition from $An_{20-30\,\pm 5}$. It is always partially altered and in places completely altered to sericite and epidote (clinozoisite).

Pyroxene is a pale-green clear unaltered monoclinic variety with the following optical properties:

$$n_y = 1.692 \pm 0.001,$$
$$2V = +58^\circ \text{ to } +61^\circ.$$

These properties correspond to a nonaluminous pyroxene of the diopside-hedenbergite series, such as the Adirondack skarn pyroxenes. Using the curves of Hess (1949), the pyroxene is identified as salite with about 14 atomic percent Fe and an Al_2O_3 content of about 3 weight percent.

Throughout the calcium-rich gneiss, a single variety of hornblende is present. Characteristically, it is clear and unaltered with color formula:

X = light yellowish brown with a pale-greenish hue,
Y = medium green with a yellowish hue, and
Z = medium green with a faint bluish hue.

Its $2V$ ranges from -78° to -85°. The absorption colors and $2V$ measurements correspond to hornblendes of the tremolite-actinolite group and to the actinolite variety specifically (Winchell, 1951; Sundius, 1946). It is intergrown with biotite in a way that suggests that the biotite replaced the hornblende, and it is commonly altered to chlorite.

Biotite in the calcium-rich gneiss differs considerably from the biotite elsewhere in the Edison area. Its color formula is always:

X = colorless to very pale yellow brown,
Y = pale yellow brown, and
Z = medium yellow brown;

which corresponds to that of magnesium-rich biotite (phlogopite). Usually, the biotite is unaltered; however, it may be altered to epidote and chlorite along layers parallel to the 001 plane.

Scapolite occurs as discrete grains. Its birefringence is rather high (0.02–0.03), which indicates that its composition is meionite (Me_{40-60}; Winchell, 1951). With some local exceptions, the scapolite does not seem to have formed as an alteration of plagioclase. The two minerals occur side by side in a single thin section and both appear to be of primary origin—that is, formed during metamorphic recrystallization. Some of the spacolite is altered to sericite.

Sphene occurs as disseminated grains and as coronas around iron oxides.

Epidote seems to be both primary and secondary. Secondary epidote is an alteration product of biotite and plagioclase. Primary epidote occurs as discrete grains or in aggregates of many small grains. Two kinds of primary epidote may be present in a single rock. One type has high relief and birefringence and is medium green and strongly pleochroic. The other type is colorless, has anomalous blue interference color, and is of lower relief. The colorless variety corresponds to clinozoisite, but the green variety corresponds to an epidote somewhat richer in iron. The iron-rich type shows strong zoning.

The principal iron oxides in the calcium-rich gneiss are varieties of hematite. These are ilmenohematite and rutiloilmenohematite; hemoilmenite was not observed. Magnetite, which is partially altered (<10 percent) to hematite (martite), is subordinate to hematite.

Calcite occurs as discrete grains and definitely seems to be primary.

Sericite is the most abundant secondary mineral. It occurs as an alteration product of plagioclase and at some places, scapolite, and may be accompanied by clinozoisite.

Petrographically, two varieties of chlorite are recognized. The most abundant type is very pale green, has weak pleochroism, an anomalous yellow-brown interference color, and is length fast. It corresponds to clinochlore(?) (Winchell, 1951). The second variety has a low birefringence, anomalous blue interference color, is colorless to very pale green, and length slow. It corresponds to penninite(?) (Winchell, 1951). Generally, the two are interlayered parallel to 001 in a sort of layered chlorite complex. Hornblende is commonly altered to such a complex in which clinochlore(?) is more abundant. Biotite is also altered to this chlorite complex; however, penninite(?) is probably more abundant in this alteration.

FABRIC

Most of the rocks in the calcium-rich gneiss are even grained and medium- or fine-medium grained; however, some coarser grained varieties and pegmatitic facies are present. Quartz and feldspar and most of the other minerals are xenoblastic, but varietal and accessory minerals, such as hornblende, pyroxene, epidote, scapolite, biotite, and iron oxides, may have subhedral outlines. Biotite occurs as plates elongated parallel to the trace of the 001 plane, and hornblende is usually elongated parallel to crystallographic *c*. Feldspar is always equidimensional, but some of the quartz is lenticular.

There seems to be a general correlation between intensity of the gneissic structure and the percentage of feldspar in the rock, so that rocks low in feldspar and high in quartz have a better developed structure. In general, the structure is due to the preferred orientation of biotite plates, elongated hornblende grains, and lenticular grains of quartz. The gneissic structure is emphasized by distinct compositional layering. This layering is caused either by a disproportionate distribution of such minerals as biotite, hornblende, pyroxene, and iron oxides; or it is less subtle and is caused by the alteration of layers of very different composition, such as metaquartzite and quartz-feldspar gneiss.

LITHOLOGIC VARIATIONS

The rocks of the calcium-rich gneiss range from metaquartzite to quartz-feldspar gneiss. All carry the same general suite of calcium-rich varietal and accessory minerals, although the mineral proportions vary considerably from rock to rock. Rock varieties occur as distinct layers ranging from fractions of an inch to 1 to 2 feet thick. Typical varieties include (1) scapolite-epidote-pyroxene metaquartzite; (2) epidote-pyroxene-quartz-feldspar gneiss; (3) epidote-biotite-hornblende-feldspar-quartz gneiss; (4) epidote-sphene-pyroxene-quartz-microcline gneiss; (5) epidote-garnet-pyroxene - scapolite - quartz - microcline gneiss; (6) pyroxene - hornblende - quartz - feldspar gneiss; (7) scapolite - biotite - hornblende - quartz - feldspar gneiss (scapolite and plagioclase both present); (8) scapolite-epidote-pyroxene-quartz-feldspar gneiss; (9) feldspar-epidote-calcite metaquartzite. Pyroxene and hornblende (and also the pair plagioclase-scapolite) notably occur side by side in the same rock with no apparent reaction or replacement relationship. In calcium-rich rock—that is, high in epidote, as (4) above, plagioclase may be absent. These general relationships suggest that bulk chemical composition was the main controlling factor in determining the final mineral assemblage. It seems certain that more or less identical pressures and temperatures prevailed throughout the calcium-rich gneiss during the period of crystallization. Hence, all the minerals would have been stable throughout the entire rock body if the bulk chemistry of the immediate environment was proper for their development (see Yoder, 1952). For example, in rocks that carry abundant calcium-rich minerals, microcline, and quartz, as (4) above, it seems logical to conclude that plagioclase is absent from such rocks because the sodium content was too low to enable its formation, and not because the pressure and temperature were beyond the stability limits of plagioclase. In addition, where scapolite is absent or closely associated with plagioclase, as in (7) and (8) above, it seems likely that there were only limited amounts of Cl^{-1} and CO_3^{-2}, in the rock-forming environment. Very likely, the absence of hornblende in favor of pyroxene, as in (2) above, and their association together, as in (6) above, indicates that a limited amount of H_2O was present in the rock-forming system (Yoder, 1952).

METAMORPHISM AND METASOMATISM

QUARTZ-POTASSIUM FELDSPAR AND ASSOCIATED GNEISS

The quartz-potassium feldspar and associated gneiss (mixed-gneiss complex) represents an original heterogeneous series of somewhat argillaceous and (or) arenaceous sedimentary rocks that have been reconstituted to gneiss by regional metamorphism and have been, in part, chemically reconstituted by metamorphic differentiation and metasomatic action. The heterogeneous character of the mixed-gneiss complex, as manifested by the marked chemical and mineralogical discontinuities (lithologic variations), is partly a reflection of primary sedimentary heterogeneity. Specifically, such rock layers as the garnet-biotite-sillimanite-quartz gneiss have bulk compositions which must closely approach those of sedimentary rocks, for example, argillaceous sandstones. Perhaps the most conclusive quantitative evidence of the sedimentary affinity of the mixed-gneiss complex is the very high quartz content (high quartz-to-feldspar ratios) in the magnetite-quartz-potassium feldspar gneiss. The mineral composition of the average magnetite-quartz-feldspar gneiss corresponds to an SiO_2 content of nearly 80 percent. Although certain alaskites (Buddington, 1939, p. 138; 1957, p. 296, 298) and leucogranites (Larsen, 1948) contain 76–78 and 76.54 percent SiO_2, respectively, no granites are known to contain SiO_2 in greater amounts. Furthermore, as conclusively pointed out by Chayes (1952), normal granites contain less than 40 percent quartz and more than 50 percent feldspar. Hence, a maximum quartz-to-feldspar ratio for normal igneous granites would be about 0.8. It is noteworthy that this ratio for magnetite-quartz-potassium feldspar gneiss is generally greater than 0.9 and in many cases greater than 1.0. It seems evident, therefore, that with respect to the proportion of quartz, the magnetite-quartz-potassium feldspar gneiss and the mixed-gneiss complex have marked sedimentary, but not igneous, affinities.

That the mixed-gneiss complex has been regionally metamorphosed is beyond doubt. Proof is provided by the well-developed metamorphic structures, such as foliation and lineation, as well as a mineral assemblage characteristic of metamorphic rocks, for example, garnet and sillimanite. No evidence in the Edison area suggests more than one period of metamorphism (except minor retrograde metamorphic effects). The mineral assemblage of the mixed-gneiss complex is characterized by nonperthitic potassium feldspar, magnetite, quartz, biotite, manganese-rich almandite, and sillimanite. The nonperthitic potassium feldspar assemblage has its counterpart in similar rocks in the Adirondacks (Engel and Engel, 1953; Buddington, 1939, p. 285), which have been interpreted as belonging to the amphibolite facies (Buddington, 1952, p. 74). Furthermore, the TiO_2 content of magnetite from the magnetite-quartz-potassium feldspar gneiss is low (averages less than 1.0 percent) and indicates a temperature of formation that corresponds to the lower range of the amphibolite facies of the Adirondacks (Buddington and others, 1955).

In addition, the mineral assemblage of the mixed-gneiss complex corresponds to the sillimanite-almandine-orthoclase subfacies of the almandine-amphibolite facies (Turner and Verhoogen, 1960, p. 549). This subfacies is distinguished by the assemblage quartz-sillimanite-almandine-orthoclase (-plagioclase-biotite-quartz) in schists of pelitic composition (Turner and Verhoogen, 1960, p. 550). Except for the absence of plagioclase, this assemblage corresponds to that in the mixed-gneiss complex. Thus, there is substantial evidence that the mineral assemblage of the mixed-gneiss complex corresponds to the amphibolite facies of regional metamorphism.

On the other hand, the occurrence of hypersthene in nearby gneisses (p. 43) and other metamorphic mineral assemblages observed in rocks of the adjacent area (p. 26) suggest a metamorphic facies higher than amphibolite grade, that is, hornblende granulite facies. On this basis, perhaps it is more appropriate to regard the designation of amphibolite grade as a minimum for the mixed-gneiss complex.

Indirect evidence indicates that the original sedimentary rocks of the mixed-gneiss complex have been chemically reconstituted during the regional metamorphism. The reconstitution was most likely an internal process involving the redistribution of material that had its origin in the immediate region rather than from extraneous and distant sources. By strict definition, the former case is metamorphic differentiation (Turner and Verhoogen, 1960, p. 582–583), whereas the latter is metasomatism (Turner and Verhoogen, 1960, p. 562). As metamorphic differentiation and metasomatism may lead to identical replacement features, however, the practical distinction of these processes is generally conjectural. In the subsequent discussion, therefore, both terms are used somewhat interchangeably, "metamorphic differentiation" referring mostly to the concept that the modifying material was of local origin, and "metasomatism" referring to the process that involved the replacement of the surrounding rocks by the locally derived "fluids."

The most obvious effect of metamorphic differentiation and metasomatism are the magnetite-rich zones and generally high magnetite content of the entire complex. The origin and emplacement of this material will be discussed in a subsequent section.

It is believed that the potassium content was inherent in the original sediments that formed the mixed-gneiss complex and that during regional metamorphism, through the process of metamorphic differentiation, a potassium-rich fluid was mobilized which reacted metasomatically with the surrounding rocks to form the heterogeneous complex. The metasomatic process is envisioned as involving the redistribution of the mobilized material and reaction with, and replacement of, the nonmobilized original rock material. It is suggested that the potassium feldspar pegmatite, which is abundant as large and small "ductless" masses throughout the complex, represents an uncontaminated crystallized late-stage product of the metasomatic fluids that originated within and permeated the mixed-gneiss complex.

The original sediments of the mixed-gneiss complex may have been of arkosic composition and rich in alkali feldspar (Pettijohn, 1957, p. 323). However, the paucity of sodium feldspar or other sodium-bearing minerals in the complex is so striking that it is necessary to postulate (1) an arkose especially poor in sodium feldspar, or (2) that the arkosic sediments were flushed of their sodium content during regional metamorphism. More likely, the original sediments of the complex were a mixture of arenaceous and argillaceous types with a high percentage of clay minerals. The abundance of potassium in the complex could then be related to the occurrence of a common clay mineral such as illite. The enrichment of potassium over sodium in illite might explain the paucity of sodium in the mixed-gneiss complex. According to the idea proposed by Goldschmidt (1921), alkali metasomatism may take place through the formation of alkali feldspar by means of the fixation (reaction) of potassium and sodium with excess Al_2O_3. Furthermore, according to the experiments of Yoder and Eugster (1955), the regional metamorphism of illite would lead to the development of potassium feldspar and Al_2O_3. With excess SiO_2, the development of sillimanite instead of Al_2O_3 would be expected. However, many layers in the mixed-gneiss complex are either rich in potassium feldspar, to the exclusion of sillimanite, or are rich in sillimanite to the exclusion of potassium feldspar, and as previously noted, many layers contain both minerals. If illite was the primary source of the potassium and aluminum, then these relationships suggest a considerable redistribution of potassium, and perhaps aluminum, during regional metamorphism. Thus, layers rich in potassium feldspar were probably thoroughly permeated by the potassium-rich metasomatic fluid, and the excess aluminum was used in the formation of potassium feldspar. On the other hand, layers rich in sillimanite are poor in potassium feldspar, which probably indicates that no fluid penetrated these layers. Perhaps these layers actually contributed their potassium content by metamorphic differentiation to the metasomatic fluid.

The potassium in the mixed-gneiss complex may have come from an extraneous source. Such potassium metasomatism has been noted in the Adirondacks (Buddington, 1948, 1957; Engel and Engel, 1953), where it is apparently related to large masses of hornblende granite and alaskite. Similar granites are present in this region and could have supplied the potassium to the mixed-gneiss complex. However, other than the association in time and space between the hornblende granite and mixed gneiss, there is no specific evidence available to identify the former rock as the primary source of potassium.

EPIDOTE–SCAPOLITE–QUARTZ GNEISS AND RELATED FACIES

It is postulated that the calcium-rich gneiss originated by the regional metamorphism, metamorphic differentiation, and metasomatism of calcareous sedimentary rocks. Thus, its genesis is similar to that of the mixed-gneiss complex. Differences between the two are mostly a reflection of the chemical differences between the original sediments.

The heterogeneous nature with alternating layers of distinct chemical and mineralogic compositions and the presence of metaquartzites and calcite-rich layers testify to the sedimentary affinities of the calcium-rich gneiss. As is indicated on plate 2, the gneiss lenses cut abruptly into the mixed-gneiss complex. The structural relationships between these two rock bodies are probably related to original sedimentary relationships involving a series of calcareous sediments (calcium-rich gneiss) abruptly lensing out into argillaceous sediments (mixed-gneiss complex).

The dissimilar mineral assemblages in the calcium-rich gneiss and the mixed-gneiss complex suggest that the two belong to different metamorphic facies. However, such distinctions can be explained on the basis of chemical differences in each environment. The absence of pyroxene and hornblende in the mixed-gneiss complex can be related to the paucity of calcium. The absence of sillimanite in the calcium-rich gneiss can be related to the high content of calcium and the formation of calcium-aluminum minerals such as hornblende and epidote instead of sillimanite. The composition of plagioclase (An_{20-30}) in the calcium-rich gneiss is characteristic of metamorphic rocks of a higher grade than the albite-epidote amphibolite facies (Turner and Verhoogen, 1960, p. 533). The microclines in the mixed-gneiss complex and calcium-rich gneiss are both nonperthitic, which indicates that they crystallized at about the same temperature (Bowen and Tuttle, 1950). In addition, the mineral assemblage of the calcium-rich gneiss corresponds to rocks of the staurolite-almandine subfacies (almandine-amphibolite facies) that have excess potassium (Turner and Verhoogen, 1960, p. 545). This subfacies is characterized by such minerals as microcline, plagioclase, biotite, hornblende, diopside, epidote, grossularite, and quartz (Turner and Verhoogen, 1960, p. 546). The high potassium content prevents the formation of either staurolite or kyanite (sillimanite). According to the facies classification of Turner, the mixed-gneiss complex (sillimanite-almandine-orthoclase subfacies) and calcium-rich (staurolite-almandine subfaces) gneiss belong to different subfacies of the almandine-amphibolite facies. Hence, difference in bulk composition between the two rock bodies accounts for the different mineral assemblages that were apparently formed under similar conditions of pressure and temperature. As discussed previously (p. 46) the designation of amphibolite grade should be regarded as a minimum.

The processes of metamorphic differentiation and metasomatism have probably also been important in the genesis of the calcium-rich gneiss. The formation of potassium feldspar in these gneisses may be related to potassium-rich metasomatic fluids generated in the mixed-gneiss complex. In addition, similar potassium-rich metasomatic fluid may have been generated within the calcium-rich gneiss from clay minerals. As for the mixed-gneiss complex, layers within the calcium-rich gneiss that are rich in potassium feldspar were probably thoroughly permeated by the potassium-rich metasomatic fluids, whereas layers rich in aluminous minerals such as hornblende and epidote were probably not penetrated by the metasomatic fluids. Again, the major sources of the potassium could have been distant and extraneous; however, no clear evidence substantiates this hypothesis.

BIOTITE–QUARTZ–FELDSPAR GNEISS

The origin of the biotite-quartz-feldspar gneiss is difficult to ascertain. Because of its near mineralogic identity to biotite-quartz-oligoclase gneiss in the Adirondacks, which are believed to be regionally metamorphosed and metasomatized graywacke sediments (Engel and Engel, 1953), it is postulated that the biotite-quartz-feldspar gneiss in the Edison area may be of similar origin.

Analogous to the similar gneiss in the Adirondacks, the biotite-quartz-feldspar gneiss probably belongs to the amphibolite facies. However, the slightly perthitic character of the potassium feldspar indicates that the gneiss may represent a mineral facies of slightly higher temperature than the mixed-gneiss complex or the calcium-rich gneiss. Similarly, the local occurrence of hypersthene (p. 43) indicates a higher grade of metamorphism, that is, hornblende-granulite facies. These mineralogic features could be explained by the proximity of the biotite-quartz-feldspar gneiss to the quartz-microcline gneiss and the igneous granite to the southeast.

Engel and Engel (1953) have found the biotite-quartz-oligoclase gneiss of the Adirondacks to be modified by potassium metasomatism so that potassium feldspar is commonly a major mineral. The biotite-quartz-oligoclase gneiss of the Adirondacks shows textural as well as mineralogic modifications that accompanied metasomatism. With an increase in the proportion of potassium feldspar, the original gneiss, "blend by subtle transitions into granite augen gneiss and gneissic granite" (Engel and Engel, 1953, p. 1059). Similar modifications are apparent in the biotite-quartz-feldspar gneiss of the Edison area. The potassium feldspar content of the biotite-quartz-feldspar gneiss corresponds most closely to facies of the Hermon-type granite gneiss (Engel and Engel, 1953, p. 1066, fig. 7), which is believed to be a highly modified facies of the biotite-quartz-oligoclase gneiss. On the other hand, the fabric of the biotite-quartz-feldspar gneiss corresponds to the less modified migmatitic and well-foliated (with minor porphyroblastic development) facies of the biotite-quartz-oligoclase gneiss of the Adirondacks (Engel and Engel, 1953, fig. 7).

In conclusion, the biotite-quartz-feldspar gneiss of the Edison area is considered to be a small-scale facsimile of similar gneiss in the Adirondacks, which is interpreted as regionally metamorphosed and potassium-metasomatized graywacke sediments.

QUARTZ-MICROCLINE GNEISS

The uniform texture and composition and apparently undeformed nature of the quartz-microcline gneiss are evidence that it originated by emplacement and crystallization of a fluid of uniform composition during a late stage of regional metamorphism. The uniform composition of the rock also indicates that the parent fluid was uncontaminated by adjacent rocks. The perthitic nature of the potassium feldspar and the rather high TiO_2 content (2.80 percent) of the magnetite both indicate that the quartz-microcline gneiss crystallized at a somewhat higher temperature than the mixed-gneiss complex and calcium-rich gneiss. Specifically, the TiO_2 content of the magnetite corresponds to that in "microcline-rich granitized rocks, formed at temperatures lower than magmatic" (Buddington and others, 1955).

Although the biotite-quartz-feldspar gneiss in the Edison area has been affected by potassium metasomatism, there is not a continuous textural or mineralogic series between that gneiss and the quartz-microcline gneiss. Rather than postulate that the quartz-microcline gneiss is the ultimate produce of potassium metasomatism of the biotite-quartz-feldspar gneiss, it seems more logical that the quartz-microcline gneiss formed by the direct crystallization of potassium-rich fluids (magma?) and that the fluids penetrated and partly modified the adjacent biotite-quartz-feldspar gneiss. Hence, the quartz-microcline gneiss is interpreted as a primary rock and not the product of a modified preexisting rock.

In conclusion, the quartz-microcline gneiss may be genetically related to the potassium metasomatism of the mixed-gneiss complex and the calcium-rich gneiss. It is suggested that the quartz-microcline gneiss crystallized directly from the relatively mobile metasomatic fluid (magma?) which originated within the mixed-gneiss complex through a process of metamorphic differentiation during regional metamorphism. In this case, the quartz-microcline gneiss may represent the lowest melting residue of the mixed-gneiss complex. Very likely, the pegmatites within the mixed-gneiss complex are the counterpart to the quartz-microcline gneiss in having crystallized from the same potassium-rich fluids. In a sense, the quartz-microcline gneiss might be considered a very large pegmatite having the same origin as the small pegmatites. On the other hand, the rock could have been formed from an introduced magma. Hence, its origin remains uncertain.

STRUCTURE OF THE EDISON AREA

The Edison area is on the southeast limb of the Beaver Lake anticline (pl. 1). Its structural features are closely related to the regional structural pattern. The gross structure of the Edison area is simple. Tabular lithologic units are oriented parallel to the axial plane of the major fold (northeast strike and steep southeast dip) and extend across the entire area (pl. 2). In detail, the structure is very complex and includes such features as foliation, lineation, folds, faults, and joints.

FOLIATION

Foliation is the most prevalent and obvious internal structure within lithologic units. The term "foliation" is applied to any planar structure within the rock that has been formed or modified by metamorphism. Foliation is manifested either by the preferred orientation of inequidimensional minerals or mineral aggregates, such as mica, amphibole, sillimanite, lensoid quartz, and feldspar aggregates, or by distinct rock layers having compositions and (or) textures different from adjacent layers. The former kind of foliation is called cleavage and the latter, compositional layering.

Mappable lithologic units represent the largest scale of compositional layering. The best example is amphibolite interlayered with quartz-feldspar gneiss or with granite. Such layering is on a scale of fractions of an inch to tens of feet. In the Edison area, the mixed-gneiss complex, the biotite-quartz-feldspar gneiss, and the quartz-oligoclase gneiss all show marked compositional layering. In contrast, pyroxene syenite gneiss, quartz-microcline gneiss and the various granites in the area are rather uniform and have little compositional layering.

Cleavage is present to some degree in almost every rock type. It is particularly well developed in the quartz-oligoclase gneiss, the pyroxene syenite gneiss, and the mixed-gneiss complex.

Foliation usually strikes northeast and dips vertically or steeply to the southeast. Compositional layering and cleavage are generally parallel. In places, however, the lithologic units are contorted into minor folds. In such cases, the compositional layers remain parallel to the larger lithologic units, whereas cleavage either remains parallel to the axial plane of the fold or disappears in favor of lineation. Thus, although the compositional layering is directly related to the larger lithologic units, cleavage is related to the folding, that is, the Beaver

Lake anticline, and is properly called an axial-plane cleavage.

LINEATION

Lineation is the second most prevalent internal structure within lithologic units. The term "lineation" is applied to any linear structure within the rock. Lineation caused by the preferred orientation of elongated minerals or mineral aggregates, such as amphibole, sillimanite, and quartz rods, is called mineral lineation. Lineation caused by the directional alinement of elongated rock masses with compositions and (or) textures different from adjacent rocks is called compositional lineation.

Compositional lineation is manifest either by the fold axes of contorted compositional layers or by oriented masses of rock that are either rod or lath shaped. Such masses are inches to tens of feet long. These small-scale linear elements are similar to large-scale features involving whole lithologic units. Compositional lineation is abundant along both the axes and limbs of folds. There is good evidence that the magnetite-rich layers in the mixed-gneiss complex are lath-shaped linear elements.

Mineral lineation evident by the preferred orientation of elongated minerals or mineral aggregates is particularly obvious in the mixed-gneiss complex. Sillimanite needles are oriented in the foliation plane and generally in a preferred linear pattern. In addition, elongated aggregates rich in magnetite, biotite, or other minerals are oriented in the foliation plane in a preferred linear pattern.

Where both compositional and mineral lineation are present, they are parallel. In the Edison area the strike and plunge of the lineation are always to the northeast, parallel to, and in the plane of, the foliation. The plunge of the lineation ranges from 10° to nearly vertical and averages about 50°.

FOLDS

There are numerous small folds and crenulations in the Edison area. Folding involves compositional layers that range in thickness from fractions of an inch to tens of feet. All such structures are interpreted as minor drag folds related to the major Beaver Lake anticline. Although the dip of the axial plane of most folds in the Edison area is parallel to that of the Beaver Lake anticline (steep southeast), the plunge of the minor fold axes in the area averages 45°–55° NE. and is, therefore, somewhat greater than that of the Beaver Lake anticline.

At the Old Ogden mine, minor folding is obvious on the central ridge of the open pit and involves compositional layers 2 inches to 2 feet thick. A syncline having a breadth of about 4 feet is evident along the crest of the ridge. The axial plane of the syncline dips 75°–85° SE., and the axis plunges 55°–60° NE. Some minor crenulation is associated with the structure.

At the Roberts mine is an excellent example of a minor drag fold exposed in the stripped southwestern part of the open pit. At that locality a magnetite-quartz gneiss layer, 10–15 feet thick, is tightly folded into a minor syncline and anticline. The axial planes of these folds dip steeply southeast, and the axes plunge at about 50°–60° NE. Hence, the architecture of these minor folds is geometrically similar to that of the Beaver Lake anticline. Within the folded magnetite-quartz gneiss layer and adjacent gneisses are many tightly crumpled and crenulated layers. In the magnetite-quartz gneiss layer, thin layers of quartz alternate with thin layers of magnetite in such crumpled structures. Both the magnetite and quartz in this layer are sheared out into lenticular grains and aggregates, but neither appears crushed nor brecciated. The evidence indicates that the folding took place when the rock was at a relatively high temperature and was able to recrystallize during deformation.

The syncline in the southeastern part of the Edison area (pl. 2) is the largest fold within the area. The axial plane of the syncline is about vertical, and its axis plunges 30°–40° NE. The breadth of the syncline is about 1,200–1,500 feet, and the fold is relatively open. The structure is best defined by the garnet-biotite-quartz-feldspar gneiss and by minor mafic-rich layers present in the contaminated hornblende granite. The northeast extension of the hornblende granite plunges beneath the trough of the syncline as a phacolithic body. That there is no continuation of this fold to the northwest is an enigma. It is postulated that the original relationships have been obscured either by the subsequent emplacement of the quartz-microcline gneiss or by Precambrian faulting along a plane near the northwest contact of the garnet-biotite-quartz-feldspar gneiss.

FAULTS AND JOINTS

Joints are the most prevalent kind of fracture in the Edison area. A set of joints that strikes northwest and dips steeply to the southwest is ubiquitous. The set, which is about perpendicular to the lineation and foliation of the lithologic units, is a good example of transverse joints. The spacing of these joints ranges from fractions of an inch to 1–3 feet and depends upon the kind of rock fractured. In relatively massive rocks, such as the hornblende granite, the joints are widely spaced; in heterogeneous rocks, such as the mixed-gneiss complex, they are closely spaced.

In addition to the transverse joints, a less well developed set of longitudinal joints fractures the mixed-gneiss complex. This set trends northeast parallel to the strike of the foliation and dips 15°–20° NW. These joints are spaced inches apart, and, unlike the transverse joints, are not smooth extensive planes but are rough and discontinuous. In places where both transverse and longitudinal joints are well developed, layers in the mixed-gneiss complex break out as small parallelepiped polyhedra bounded by two joint planes and the foliation plane.

In some places a joint set that trends northwest and dips gently northeast (parallel to the plunge of the lineation) is present.

The only fault of any significance in the Edison area is 600 feet southwest of the Davenport mine (pl. 2). There, the horizontal offset of a magnetite-rich layer is 150 feet. This minor faulting is parallel to the transverse joints and probably is genetically related to them.

DEPOSITS

HISTORY AND PRODUCTION

According to Bayley (1910), mining in the Edison area began in 1772 and was more or less continuous until about 1879, when all the principal mines were closed down. In 1890 the entire area was purchased by the Edison Co. who installed several large magnetic separators with an estimated capacity of 700,000 tons per year for large-scale opencut mining of low-grade ore. Bayley reports that about 75,000 tons of ore was mined in 1899, producing 100,000 tons of concentrate (60–65 percent Fe) which was briquetted and shipped. Apparently the venture proved unprofitable, and in 1899 all operations ceased. The deposits have not since been worked.

Additional historical and production data pertaining to New Jersey iron ores have recently been summarized by Sims (1953, p. 287–288).

DISTRIBUTION

The heterogeneous nature of the mixed-gneiss complex is evident by the concentration of magnetite into distinct layers as well as the variation in proportions of other minerals. Enriched zones may consist of a single magnetite-rich layer or of several magnetite-rich layers separated by lean gneiss. Assay data indicate that some layers carry as much as 55–60 weight percent magnetic Fe. Such rich layers generally grade off into leaner wallrock which may carry as little as 2 weight percent magnetic Fe.

In the Edison area the magnetite ore deposits are confined within the mixed-gneiss complex to a narrow belt which extends parallel to the trend of the complex for 7,900 feet and averages 150–200 feet in width. Several distinct magnetite-rich zones are present within this belt and are shown on plate 2. The most important ore zones include (1) the zone in the southeast workings of the Old Ogden and Roberts mines; (2) the zone that passes through the northwest workings of the Old Ogden and Roberts mines and extends from the Victor mine as far southwest as the Davenport mine; (3) the zone northwest of the Davenport mine and apparently displaced along a minor fault to the southwest; and (4) the zone passing through the Vulcan mine. In addition, numerous smaller magnetite-rich layers are present; however, it is doubtful that these are large enough to constitute an ore zone. Other ore zones may be present in the northwestern part of the mixed-gneiss complex, such as at the Iron Hill mine and in the zone passing through the Copper shaft.

SOURCES OF DATA

Seven diamond-bit core holes, totaling 3,987 feet, were made by the Pittsburgh Coke & Iron Co. in 1943. The core was assayed for magnetic iron, and some zones were assayed for soluble iron, phosphorus, and sulfur as well. In 1920 the Bethlehem Steel Co. diamond cored five holes totaling 3,015 feet. In addition, they prepared 10 channel samples between the Roberts mine and the Victor mine along the magnetite-rich zone that passes through the Condon cut. The Bethlehem Steel Co. assayed the core and channel samples only for magnetic iron.

Old surface workings provide the best exposures of the ore zones. The workings, which are located on the base map (pl. 2), are open pits, and from the southwest to northeast are called the Big cut, Davenport mine, Old Ogden mine, Roberts mine, Condon cut, Victor mine, and Iron Hill cut. The underground workings were on the same magnetite-rich zones, and from the southwest to northeast they include the Vulcan, Davenport, Old Ogden, Roberts, Victor, and the Copper mines. All the underground workings are now flooded and inaccessible; however, Bayley's report (1910) provides some data pertaining to them.

A modified version of a dip-needle survey made by the Edison Co. is plotted on the base map (pl. 2). Contours of 40° and 60° are shown. The 60° contour corresponds to an absolute reading of 70°–80° with a standard Gurley, Lake Superior dip needle calibrated to read −21° over hornblende granite.

STRUCTURE

The ore zones are conformable to the foliation of the gneisses. The drill-core data show that the magnetite zones are composed of one to several magnetite-rich layers—that is, individual magnetite bodies. The magnetite zones range in thickness from 10 to about 80 feet. The thick zones are always a composite of several individual magnetite-rich layers that alternate with layers less rich in magnetite. Thin ore zones are generally a single uniform magnetite-rich layer. In no case do magnetite-rich layers crosscut the foliation of adjacent gneisses. Structurally, the magnetite-rich layers are identical with the adjacent rock layers.

The drill-core data, dip-needle observations, and field observations show that the ore zones and the individual magnetite bodies of which they are composed pinch and swell parallel to the strike of the foliation. Thus, the ore zones shown on plate 2 are discontinuous along the strike. Actually, the data indicate that the ore zones and individual magnetite bodies pinch out in the foliation plane in a direction parallel to the linear structure in the gneisses. The ore zones have a tabular form in the foliation plane and are parallel to the lineation. The lineation plunges 55°–60° NE. in the Old Ogden mine and about 50° NE in the Iron Hill cut. Drill-core data prove that some of the ore zones extend to a depth at least 700 feet from the surface; however, it is uncertain how far down the linear structure the ore zones may be projected.

Like the adjacent gneiss, the individual magnetite-rich layers or bodies are sometimes deformed into minor folds. Because such folds are tight, magnetite layers are for the most part doubled in thickness along the fold crest and form a linear (rod-shaped) body which plunges along the fold axis. Although linear magnetite bodies along the axes of folds form important ore deposits elsewhere in the New Jersey Highlands (Sims, 1953), they are subordinate in the Edison area. As previously described, the major ore zones and the individual magnetite bodies of which they are composed are tabular. The few linear bodies are the result of minor folding of the larger planar magnetite bodies.

PETROLOGY AND MINERALOGY

Two varieties of magnetite concentration are recognized in the mixed-gneiss complex. The first, which is the most important type, is related to the typical magnetite-quartz-potassium feldspar gneiss. In this type, magnetite is concentrated in definite layers (as much as 3–5 in. thick) which may have fairly sharp contacts with the wallrock gneiss; more commonly, magnetite-lean wallrock gneiss grades abruptly into a magnetite-rich layer. In the latter case, magnetite increases from disseminated grains in the lean wallrock to thin discontinuous layers in rich wallrock, and finally into heavy solid layers of magnetite 3–5 inches thick. The same mineral phases are found in magnetite-rich layers as in lean wallrock (see section on the mixed-gneiss complex and table 30); however, the proportions of the minerals vary greatly. Ilmenohematite is generally absent from magnetite-rich layers, but in many places it is a major accessory in adjacent wallrock. Potassium feldspar is commonly absent in magnetite-rich layers, but when present, it is always an untwinned variety with monoclinic symmetry. Quartz is always present. Apatite is usually enriched in magnetite-rich layers, and fluorite has been observed. Minerals of metasedimentary affinities, such as biotite, garnet, and sillimanite, are subordinate in magnetite-rich layers; but sulfides (pyrite, molybdenite, chalcopyrite, and bornite) are locally enriched (see section on sulfide zone, p. 53). Some of the diamond-drill core locally had slight porosity and minute cavities as though sulfide minerals had been leached, as by late-stage hydrothermal solutions.

The second variety of magnetite-rich layer is composed of magnetite and quartz and is probably related to the magnetite-quartz gneiss (metaquartzite?). This type occurs as distinct layers enclosed within the predominant magnetite-quartz-potassium feldspar gneiss. Layers of this variety are 6 inches to 10 feet thick and pinch out along the strike within 20–50 feet. This type of ore layer is mineralogically very simple; magnetite and quartz form 95 percent of the rock. Potassium feldspar is absent, except adjacent to the contacts of magnetite-quartz-potassium feldspar gneiss; ilmenohematite and hemoilmenite are absent, and biotite, muscovite, and apatite are minor minerals.

The textures of the two types of magnetite concentrations are similar in that they are medium grained, even grained, and xenoblastic. However, the magnetite concentrations that are related to the magnetite-quartz-potassium feldspar gneiss have a moderately developed gneissic structure similar to that of the adjacent gneisses. In contrast, the magnetite-quartz variety has a very strong gneissic structure, because of the greater lenticularity (deformed character) of the minerals.

The contact between the magnetite-quartz gneiss and the magnetite-quartz-potassium feldspar gneiss is sharp but irregular (fig. 14). It is not a planar contact of the fissure type; it is the natural grain boundary contact between adjacent layers of the gneiss. Magnetite is enriched in the magnetite-quartz-potassium feldspar gneiss along the contact; texturally, however,

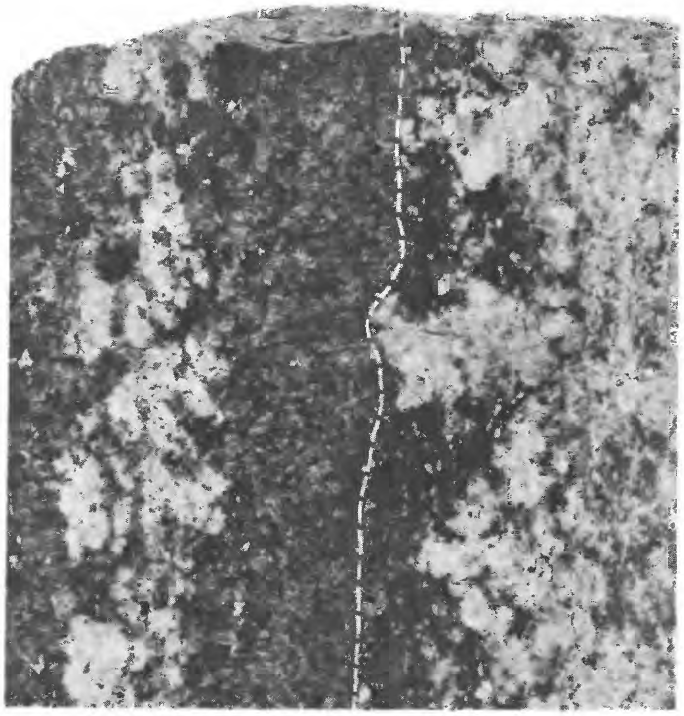

FIGURE 14.—Contact (dashed line) between magnetite–quartz–potassium feldspar gneiss (right) and magnetite–quartz gneiss (left). The contact is the grain boundary between minerals. There is marked enrichment in magnetite in the magnetite–quartz–potassium feldspar gneiss at the contact. Note *isolated* lens of magnetite–quartz–potassium feldspar gneiss with potassium feldspar porphyroblasts within the magnetite–quartz gneiss at left side of photograph. Natural scale.

this enriched zone of magnetite is distinct and separate from the magnetite in the magnetite-quartz gneiss. layer. Lenses of magnetite-quartz-potassium feldspar gneiss, which contain potassium feldspar porphyroblasts, and isolated potassium feldspar porphyroblasts are enclosed by magnetite-quartz gneiss. These relationships suggest that the potassium feldspar porphyroblasts and perhaps the magnetite-quartz-potassium feldspar gneiss could have formed after the formation of the magnetite-quartz gneiss.

SULFIDE ZONE

A distinct zone characterized by sulfide minerals is clearly distinguished within the mixed-gneiss complex at the Old Ogden and Roberts mines (pl. 2). The zone is 40 feet thick and is in the southeast workings of each of the mines. The zone extends southwest to the area just southeast of the Davenport mine. The high content of sulfide minerals is made particularly evident by more intense weathering of the gneisses. In addition, assay data from the diamond cores, which included total sulfur determinations, substantiate the existence of a distinct sulfide zone. The sulfur content in the sulfide-rich zone averages 1.5–3.5 percent. It is usually less than 0.1 percent in other parts of the mixed-gneiss complex. The drill-core data also show that the sulfide zone is enriched in magnetite; however, it is emphasized that all magnetite-rich zones are not enriched in sulfur. The sulfide zone is conformable to the foliation and is always in the same stratigraphic position relative to adjacent gneisses. Thus, the sulfide zone is traceable for 2,500 feet from southeast of the Davenport mine to the northeast end of the Roberts mine. Far to the northeast, sulfide minerals are abundant at the Copper mine shaft (pl. 2). The sulfide zone may be continuous to the Copper mine, although this could not be substantiated.

Sulfide minerals in the sulfide zone at the Old Ogden and Roberts mines are pyrite, chalcopyrite, and molybdenite. The zone is rich in magnetite and appears to carry more garnet than the adjacent rocks. Pegmatite constitutes a large part of the sulfide zone at the Old Ogden mine, but at the Roberts mine the zone appears to be more quartzose than adjacent gneisses. Where chalcopyrite and pyrite in the sulfide zone are intergrown, pyrite rims the chalcopyrite. The relationships between the magnetite and sulfides are very clear. At many places pyrite rims the magnetite, and, rarely, a veinlet of pyrite cuts across magnetite grains. The textural evidence indicates that the sulfides are later than the iron oxides.

The sulfide-bearing rocks at the Copper mine are heterogeneous. The predominant type is a magnetite–quartz-potassium feldspar gneiss, but there are numerous biotitic, garnetiferous, and sillimanitic varieties. The gneisses all contain a considerable amount of magnetite (10–15 percent). Sulfide minerals are disseminated throughout the gneiss and form 1–2 percent of the rock. Minor minerals include apatite, zircon, fluorite, and some plagioclase. The heterogeneous character of these rocks is emphasized by seams and thin layers of pegmatite.

The principal ore minerals at the Copper mine are magnetite (in places, martitic) and bornite. Accessory ores include ilmenohematite, hemoilmenite, ilmenite, chalcopyrite, covellite, and some pyrite and molybdenite. Magnetite only rarely carries ilmenite blades; however, in one reaction zone with bornite, magnetite grains contain a few blebs of ilmenite. There are three types of bornite. The first consists of independent grains of brownish bornite that carry chalcopyrite exsolution films whose orientation is controlled by crystallographic planes. The second is pinkish bornite which may consist of independent grains, but which usually occurs as anhedral intergrowths with magnetite.

This type carries exsolution chalcopyrite in the form of blebs or droplets, and as strings which wind in a tortuous pattern within the bornite. Both types of chalcopyrite are confined to the central part of the host crystal. The third type, which grades into the second, consists of blades of bornite (with exsolved chalcopyrite blebs) alternating with blades that consist of a mixture of limonite, rutile(?) and ilmenite. This type definitely replaces some magnetite. The copper sulfides have been partly oxidized to covellite along fractures and at grain boundaries.

It seems evident that the copper sulfides are younger than the magnetite. Bornite of the first type probably crystallized directly from copper solutions. The second and third types of bornite were probably formed by replacement of magnetite by copper solutions; in this case, the bladed bornite might represent the early stages of this replacement process.

POTASSIUM FELDSPAR

In terms of volume percent, potassium feldspar is the most important mineral within the Edison magnetite deposits and associated rocks. Significant variations in composition, crystal symmetry (twinning), and perthitic intergrowth were noted between the potassium feldspars from the biotite–quartz–feldspar gneiss, quartz-microcline gneiss, magnetite–quartz–potassium feldspar gneiss, and magnetite-rich (ore) layers.

Chemical analyses of quartz–potassium feldspar concentrates from a magnetite-rich layer and adjacent magnetite–quartz–potassium feldspar wallrock (table 25, Nos. 144, 145, respectively) show that the potassium feldspar of the mineralized rock is much higher in barium (celsian). In addition, the potassium feldspar of the magnetite–quartz–potassium feldspar gneiss normally has distinct cross-hatched (grid) microcline twinning, whereas the potassium feldspar of the magnetite-rich (ore) layers is always untwinned.

TABLE 25.—*Partial chemical analyses of quartz-feldspar concentrates from magnetite-rich layer and wallrock*

[Analyst, Doris Thaemlitz, 1954]

144				145			
Weight percent oxides		Feldspar molecules recalculated to 100 percent		Weight percent oxides		Feldspar molecules recalculated to 100 percent	
CaO	0.09	An	1.0	CaO	0.04	An	.5
Na_2O	.45	Ab	9.0	Na_2O	.46	Ab	10.6
K_2O	6.09	Or	80.7	K_2O	5.82	Or	88.6
BaO	2.28	Cel	9.3	BaO	.06	Cel	.3
	8.91		100.0		6.38		100.0

144. Concentrate from a magnetite-rich layer, Condon cut, northeast of Edison, N.J. 25 percent potassium feldspar and 27 percent magnetite by volume.
145. Concentrate from magnetite-quartz-potassium feldspar gneiss (wallrock to 144) Condon cut. 45 percent potassium feldspar and 3 percent magnetite by volume.

X-ray analyses show that the untwinned potassium feldspar from the magnetite-rich layers is monoclinic and confirms that the potassium feldspar from adjacent gneiss is triclinic (microcline). In contrast to the latter, potassium feldspar porphyroblasts within the biotite-quartz-feldspar gneiss are always untwinned and display monoclinic symmetry to X-ray analysis. This was surprising because the matrix potassium feldspar in this gneiss is microcline. The potassium feldspar in the quartz-microcline gneiss has good microcline grid twinning. However, X-ray analysis of this feldspar indicates that it is a mixture of both triclinic (microcline) and monoclinic (orthoclase) potassium feldspar polymorphs. Similar mixtures have been reported by Laves (1950), Harker (1954), and MacKenzie (1954).

The potassium feldspar of the magnetite-rich (ore) zones and the associated magnetite–quartz–potassium feldspar gneiss is virtually nonperthitic, containing not more than 11 percent of films and lenses of albite. This observation is confirmed by the chemical analyses presented in table 25. In contrast, petrographic and X-ray data indicate that the amount of albite intergrown in perthitic fashion is 10–15 percent in the potassium feldspar from the biotite-quartz-feldspar gneiss, and as much as 20 percent in the potassium feldspar of the quartz-microcline gneiss.

The interpretation of these observations is complex and difficult. It seems likely that the barium-rich potassium feldspar from the magnetite-rich (ore) rocks was derived by the recrystallization (with addition of barium) of the microcline in the adjacent wallrock. It seems most plausible that this recrystallization and enrichment in barium occurred in conjunction with the formation of the magnetite ore deposits. Regarding the exsolved (perthitic) albite, it is evident from the alkali feldspar solvus curve (Bowen and Tuttle, 1950) that these, but slightly, perthitic potassium feldspars probably came to a final equilibrium at temperatures less than 660°C. The differences in albite content of the potassium feldspars indicate that the quartz-microcline gneiss came to final equilibrium at a significantly higher temperature than the biotite-quartz-feldspar gneiss and magnetite–quartz–potassium feldspar gneiss. It is suggested that the quartz-microcline gneiss may have crystallized from a fluid (magma) which intruded and permeated older rocks, and that potassium metasomatism caused the development of the less perthitic potassium feldspar in the biotite-quartz-feldspar gneiss.

GARNET

Although its quantitative distribution is very irregular, garnet is present in all the principal metasedimentary and metasomatic rocks of the Edison area. Garnet occurs as disseminated grains and as porphyroblastic aggregates of grains, usually oriented in the foliation plane. Generally, the garnet is intergrown with iron oxides and quartz in a poikilitic fashion. All the garnets are pink to red.

Table 26 gives chemical data on three garnet samples. A complete chemical analysis is given for garnet sample B–153a which was separated from a magnetite-rich (ore) layer from the Roberts mine. As shown by the molecular recalculation, the garnet is rich in manganese and corresponds to a spessartite-almandite with about 15 mole percent of other garnet molecules.

Garnet sample B–151f was separated from a layer of garnet-biotite-sillimanite-quartz gneiss (see fig. 9) within the mixed-gneiss complex. The mineral composition of this rock shows its metasedimentary affinity, although the proportion of iron oxide minerals (ore) is fairly high. The MnO content of the garnet is also high, 12.7 percent, corresponding to about 30 mole percent spessartite (table 26).

Garnet sample 1908 (table 26) was separated from the quartz-microcline gneiss. Its MnO content is lower, 4.9 percent, corresponding to only 11 mole percent spessartite.

According to Miyashiro (1953), at high metamorphic grades (amphibolite facies), the composition field of pyralspite (Fe, Mg, and Mn garnet) is enlarged so that iron and magnesium-rich garnet can form as readily as manganese-rich garnet. Hence, the differences in the manganese content of the garnet described above probably indicate differences in the manganese concentrations in the rock-forming system. Therefore, it is concluded that manganese is enriched in the mixed-gneiss complex (ore zone) as compared with the iron-poor wallrocks such as the quartz-microcline gneiss.

TABLE 26.—*Chemical analyses of garnets*

[Analyst, Doris Thaemlitz]

	B–153a	B–151f	1908
Chemical analyses			
SiO_2	36.38		
Al_2O_3	20.82		
TiO_2	.05		
Fe_2O_3	1.50		
FeO	21.92		
MnO	14.51	12.7	4.9
MgO	1.11		
CaO	3.58		
H_2O+	.04		
H_2O-	.06		
Total	99.97		
Mole percent			
Pyrope	4.5		
Almandite	50.7		
Spessartite	34.2	30	11
Andradite	4.6		
Grossularite	6.0		
Total	100.0		

B–153a. Garnet from magnetite-rich gneiss (ore), Roberts mine, northeast of Edison. Density of garnet=4.157g/cc. Rock contains 38 percent magnetite, 5 percent garnet, 39 percent quartz and 13 percent untwinned potasium feldspar, 1 percent biotite and 4 percent accessories (apatite, fluorite, zircon, monazite).
B–151f. Garnet from garnet-biotite-sillimanite-quartz gneiss.
1908. Garnet from quartz-microcline gneiss northeast of Edison.

MAGNETITE-HEMATITE–ILMENITE–RUTILE PARAGENESIS

The following section is devoted to the detailed description of the magnetite-hematite-ilmenite-rutile paragenesis (iron and titanium oxides). The oxide minerals from the magnetite–quartz–potassium feldspar gneiss and related rocks of the mixed-gneiss complex receive the principal emphasis.

The mineralogic terminology of the iron and titanium oxide minerals is adapted from the classification presented by Buddington, Fahey, and Vlisidis (1963, p. 140). Specifically, the following names are used for the minerals and their microintergrowths:

- Magnetite: almost wholly $FeFe_2O_4$.
- Hematite: almost wholly Fe_2O_3.
- Martite: hematite (Fe_2O_3) as a secondary alteration of magnetite.
- Ilmenite: $FeTiO_3$ with as much as about 6 percent Fe_2O_3 and perhaps some TiO_2 in solid solution.
- Rutile: almost wholly TiO_2.
- Ilmenomagnetite: magnetite with microintergrowths of ilmenite.
- Ilmenohematite: hematite with microintergrowths of ilmenite or hemoilmenite.
- Hemoilmenite: ilmenite with hematite or ilmenohematite in microintergrowths.
- Rutilo-ilmenohematite: hematite with microintergrowths of ilmenite and rutile.
- Ilmeno-rutilohematite: hematite with microintergrowths of rutile and ilmenite or hemoilmenite.
- Rutilohematite: hematite with microintergrowths of rutile.
- Hemorutile: rutile with microintergrowths of hematite.

The examination of the iron-titanium oxides has been of a petrographic and chemical nature. Numerous polished surfaces of the magnetite–quartz–potassium feldspar gneiss and its variations, as well as a limited number of polished surfaces of the biotite-quartz-feldspar gneiss, quartz-microcline gneiss, and calcium-rich gneiss were examined. Petrographic criteria for the recognition of iron-titanium oxide minerals in polished

TABLE 27.—*Petrographic criteria for identification of iron and titanium oxide minerals*

Magnetite	Hematite	Ilmenite	Rutile
		Optical characteristic	
Isotropic	Strongly anisotropic.	Strongly anisotropic.	Strongly anisotropic; brownish-yellow internal reflections in grains but rarely in thin disks and lenses.
		Color compared with—	
Ilmenite, bright white. Hematite, brownish white. Rutile, nearly equivalent.	Magnetite, very white. Ilmenite, very bright white. Rutile, bright white. Martite, brownish-violet-white, duller.	Magnetite, brownish gray. Hematite, violet tinged brown. Rutile, distinctly darker.	Magnetite, nearly equivalent. Hematite, brownish, darker. Ilmenite, distinctly brighter, very difficult to see difference in fine intergrowths.
		Morphology	
Occurs as tablets parallel to 0001 in ilmenite-hematite intergrowths and rutile.	Lenses and disks parallel to 0001 in ilmenite-hematite intergrowths. Disks and wedges in magnetite (martite).	Lenses and disks parallel to 0001 in ilmenite-hematite intergrowths. Tablets parallel to 111 in magnetite.	Disks parallel to rhombohedral plane in ilmenite-hematite intergrowths. Rhombohedral disks and thin lenses parallel to 0001 in rutilohematite.

sections are summarized in table 27. The chemical data include the partial chemical analyses (FeO, Fe_2O_3, and TiO_2) of the magnetic iron-titanium oxide fractions of eight samples of magnetite–quartz–potassium feldspar gneiss (table 28). Five of these samples are from the Edison area proper. The other three are from the Sherman-Bunker deposits (see discussion p. 34) and are included in this discussion because of their close similarity to samples from the Edison area. The magnetic fraction includes magnetite and any intergrown minerals such as ilmenite and hematite (martite). The nonmagnetic fraction includes hematite, ilmenite, and rutile and their microintergrowths. The partial analysis of each fraction has been recalculated to 100 percent for the iron-titanium oxide minerals observed in the polished-surface examination of the particular sample (table 28). Knowing the weight proportion of the magnetic and nonmagnetic fractions in each sample, the actual weight percent of the iron-titanium oxide minerals in each rock was computed (table 28).

Unless otherwise stated, the following description deals with the iron-titanium oxides in the magnetite–quartz–potassium feldspar gneiss and its lithologic variations. This entire discussion is centered about various parts of the iron-titanium oxide system and includes (1) the magnetite-ilmenite-hematite part of the system (the magnetic fraction); (2) the ilmenite-hematite-rutile part of the system (the nonmagnetic fraction); and (3) some special iron-titanium oxide intergrowths. Finally, some observations on the alteration of ilmenite and a brief discussion of the iron-titanium oxide mineral paragenesis of other rocks within the Edison area are presented.

MAGNETITE–ILMENITE AND MAGNETITE–HEMATITE

MAGNETITE-ILMENITE

All the samples of magnetite examined in polished surface are to some extent intergrown with ilmenite and may therefore be called ilmenomagnetite. The chemical analyses of the magnetic fraction (table 28) show that the mole percent of $FeTiO_3$ ranges from 0.4–4.1. The intergrown ilmenite has the form of long thin blades which are oriented parallel to the octahedral plane of the host magnetite (fig. 15). The blades are 0.02–0.10 mm thick and may be 20–50 times as long. In three dimensions the intergrown ilmenite must be in the form of planar disks or thin tablets oriented in the octahedral plane. In places, ilmenite is intergrown as irregular-shaped grains concentrated near the borders of the host magnetite.

In view of the recent experimental work on titaniferous magnetites by Buddington and Lindsley (1964), it seems likely that the intergrown ilmenite originated from the subsolidus oxidation and contemporaneous exsolution of an initial magnetite-ulvospinel (Fe_2TiO_4) solid solution. The actual content of solid solu-

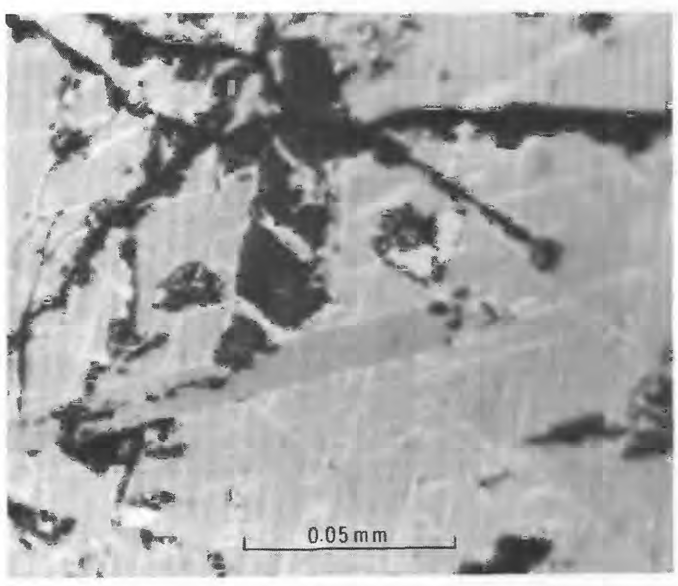

FIGURE 15.—Photomicrograph of ilmenomagnetite with martite. Martite forms an octahedral network of films and spikes; the ilmenite tablet is also parallel to the octahedral plane and appears to form a barrier to the growth of martite. White, hematite or martite; light gray, magnetite; dark gray, ilmenite. Sample 144, reflected light.

tion TiO_2 is extremely low and ranges from 0.14 to 1.39 weight percent (table 28). Using the TiO_2 content, it is noted that the temperature of formation of the magnetite–quartz–potassium feldspar gneiss corresponds to that of the lower range of the amphibolite facies of the Adirondacks (Buddington and others, 1955; Buddington and Lindsley, 1964).

Magnetite-Hematite (Martite)

All the magnetite within the magnetite–quartz–potassium feldspar gneiss is to some extent intergrown with hematite. The mole percent of intergrown hematite ranges from 8.1 to 32.1 of the host magnetite (table 28). The intergrown hematite has the form of thin films or triangular spikes that are consistently oriented in the octahedral plane of the host magnetite (fig. 15). Thus, the actual shape of the intergrown hematite is that of thin planar disks (films) and thin wedges (spikes). If the hematite content is high, the films and spikes form an interlocking octahedral network. In very hematitic magnetite, the films and spikes coalesce to yield blebs and grains of hematite. Thus, magnetite with sparse hematite in a regular octahedral pattern grades into magnetite with a large proportion of hematite as films, spikes, irregular blebs, grains, and patches. In a few places actual veinlets of hematite crosscut the host magnetite.

In many places the hematite is concentrated along the border of the host grain. In this case the films and spikes of hematite project inward from the borders of the magnetite and disappear toward the center of the grain. Even more striking is the distribution of hematite along the borders of cracks and veinlets within the host magnetite. The cracks and veinlets are rarely filled with hematite, but films and spikes of hematite project outward from the borders of the fractures and dis-

Table 28.—*Partial chemical analyses and petrographic data for magnetic and nonmagnetic fractions of iron and titanium oxides from eight samples of magnetite–quartz–potassium feldspar gneiss from the Edison and Sherman-Bunker magnetite deposits*

[Analysts, J. J. Fahey and Angelina Vlisidis, 1952–54]

	146	145	149	143	144	154	153	151
PARTIAL CHEMICAL ANALYSES								
Magnetic fraction								
Fe_2O_3	64. 93	70. 23	68. 49	70. 37	68. 17	63. 94	66. 77	65. 82
FeO	26. 02	21. 77	25. 27	22. 15	25. 62	26. 45	24. 52	27. 66
TiO_2	. 14	1. 01	. 88	1. 11	. 54	1. 39	1. 10	. 54
Total	91. 09	93. 01	94. 64	93. 63	94. 33	91. 78	92. 39	94. 02
Nonmagnetic fraction								
Fe_2O_3	74. 36	56. 61	46. 59	N.d.	21. 69	29. 10	59. 18	39. 79
FeO	2. 29	2. 83	16. 32	N.d.	16. 06	12. 79	9. 82	13. 62
TiO_2	10. 45	16. 16	23. 67	N.d.	40. 03	30. 63	16. 55	28. 43
Total	87. 10	75. 60	86. 58		77. 78	72. 52	85. 55	81. 84
Weight percent Fe and Ti oxides in rock								
Magnetite	3. 31	3. 05	12. 6	18. 87	39. 01	11. 2	13. 7	24. 00
Hematite	1. 71	1. 29	4. 3	6. 46	6. 07	1. 08	2. 93	1. 55
Ilmenite	. 10	. 11	1. 8	. 58	. 69	. 36	. 37	. 33
Rutile	. 15	. 05	. 4	Trace	. 15	. 06	. 16	. 02
Composition magnetic fraction to 100 mole percent								
Magnetite	88. 3	65. 2	77. 7	65. 0	80. 8	84. 1	75. 8	90. 3
Hematite [1]	11. 3	32. 1	20. 0	32. 0	17. 8	11. 8	21. 0	8. 1
Ilmenite	. 4	2. 7	2. 3	3. 0	1. 4	4. 1	3. 2	1. 6
Composition nonmagnetic fraction to 100 mole percent								
Hematite	78. 2	63. 3	44. 8	N.d.	21. 4	0. 8	40. 6	9. 9
Ilmenite	5. 3	7. 2	34. 0	N.d.	35. 0	Trace	0	0
Rutile	16. 5	29. 5	16. 2	N.d.	43. 6	67. 8	35. 8	58. 8
Magnetite	0	0	5. 0	N.d.	0	31. 4	23. 6	31. 3

TABLE 28.—*Partial chemical analyses and petrographic data for magnetic and nonmagnetic fractions of iron and titanium oxides from eight samples of magnetite–quartz–potassium feldspar gneiss from the Edison and Sherman-Bunker magnetite deposits*—Continued

PETROGRAPHIC DATA

Mineral analysis

[X, present]

									Mean
Quartz	49.3	45.1	24.0	60.7	33.8	27.7	33.7	38.3	39.1
Potassium feldspar	40.0	45.0	62.3	17.1	25.6	62.5	51.9	44.2	43.5
Plagioclase	0	0	0	1.5	0	.2	X	0	.2
Biotite	.2	2.0	0.5	X	1.1			X	.5
Mica complex	1.0		X			.9	.5		.3
Sericite		2.3	1.7	2.2	X	.7	2.6	.4	1.2
Chlorite						X		.4	.1
Sericite and epidote	5.3				8.5				1.7
Sillimanite	1.0	2.6	X		X	X	1.2		.6
Garnet			X						X
Epidote	.7								.1
Apatite	X	X	.5	.7	3.1	.3		.8	.7
Zircon	X	X	X	X	X	.1	X	X	X
Spinel		X							X
Accessory	X	X	.1	.6	.9	.3	X	X	.3
Ores	2.5	3.0	10.9	17.2	27.0	7.3	10.1	15.9	11.7
Ilmenomagnetite	1.9	2.5	8.5	17.2	27.0	6.3	8.3	15.2	
Hemoilmenite			.4						
Ilmenohematite				X	X				
Rutilo-ilmenohematite			2.0						
Ilmeno-rutilohematite	.6	.5							
Rutilohematite						X	1.2	X	
Hemorutile							.6	.7	
Rutile			X	X	X	1.0		.1	
Ilmenite						X			

[1] Martite. Hematite replacement of magnetite.

146. 800 ft northwest of Copper mine, Edison area; ilmeno-rutilohematite=$I_5R_{16}H_{79}$.
145. Northeast end of Condon cut. Edison area; ilmeno-rutilohematite=$I_7R_{30}H_{63}$.
149. Big cut, Edison area, hemoilmenite=$R_{0-5}M_{0-10}(H_{25}I_{75})$, (magnetite tablets intergrown with hemoilmenite); rutilo-ilmenohematite=$R_{0-13}M_{0-10}(I_{37}H_{63})$, (magnetite tablets intergrown with rutilo-ilmenohematite.
143. Big cut, Edison area.
144. Northeast end of Condon cut, Edison area.
154. Sherman-Bunker deposit; rutilohematite=$R_{30}H_{70}$; rutile=$M_{32}R_{68}$ (magnetite tablets intergrown with rutile).
153. Sherman-Bunker deposit; rutilohematite=$M_{0-20}(R_{28}H_{72})$, (magnetite tablets intergrown with rutilohematite); hemorutile=$H_{10}R_{90}$.
151. Sherman-Bunker deposit; hemorutile and rutile=$M_{31}H_{10}R_{59}$ (magnetite tablets intergrown with hemorutile and rutile).

appear rapidly beyond them. It is evident that the distribution of hematite is controlled by cracks, veinlets, and grain boundaries.

The amount of intergrown hematite varies radically from grain to grain within a single polished surface. Thus, one magnetite grain may carry as much as 70–80 volume percent of hematite, and an adjacent grain, as little as 10–15 percent.

Within magnetite, hematite always shows a consistent textural relationship to the intergrown ilmenite tablets. Films and spikes of hematite are oriented parallel to and athwart individual octahedral tablets of ilmenite. Those at an angle sharply terminate against the ilmenite blades but never crosscut them (fig. 15). Such sharply terminated films and spikes of hematite never have a continuous half on the opposite side of the ilmenite blade. Hence, the ilmenite did not replace a section through the middle of a film or spike of hematite. These textural relationships indicate that the ilmenite tablets formed before the intergrown films and spikes of hematite, and that the latter grew in the host magnetite as far as the ilmenite-magnetite interface, at which point their growth abruptly terminated.

Darken and Gurry (1946) Greig and others (1935), and Schmahl (1941) have shown that magnetite takes excess ferric iron into solid solution only above 1,000°C. If metamorphic facies are considered (p. 46–49), the rocks and associated magnetite deposits within the Edison area undoubtedly crystallized well below 1,000°C. These facts indicate that the magnetite-hematite intergrowths are not of exsolution origin. On the contrary, the experimental data, coupled with the distinct textural relationships described above, prove that the intergrown hematite is an alteration product of the host magnetite (martite). The delicate structure and octahedral pattern show that the alteration was of replacement nature. It is postulated that the martite is a retrograde mineral forming during the cooling stages of the rock in response to changing physical conditions. This hypothesis has been substantiated by thermodynamic calculations made by Baker (1955) on the iron oxide system.

HEMATITE–ILMENITE–RUTILE

In nearly all samples of magnetite–quartz–potassium feldspar gneiss, minerals of the nonmagnetic iron-titanium oxide fraction were observed. These minerals include hematite, ilmenite, rutile, and eight various distinctive intergrowths of them. In addition, some of the samples have small amounts of magnetite (martitic) intergrown with the nonmagnetic oxides. The bulk compositions of the various intergrowths have been estimated from the chemical analyses and petrographic measurements. The estimations show that the intergrowths belong to two groups. One group has compositions between hematite and ilmenite, and the other group has compositions between hematite and rutile. These two groups are discussed separately below.

The nonmagnetic oxides, hematite, ilmenite, and rutile, and their various intergrowths, except for martite and a peculiar alteration of ilmenite (composed of rutile and hematite), are all considered to be of primary origin. No textural relationships suggest that these oxides are of secondary origin. The oxides appear equivalent texturally to the associated magnetite and are believed to have formed contemporaneously with it. In addition, primary hematite is quite distinct from obvious secondary hematite (martite). The most important distinction is that martite is always devoid of any intergrown ilmenite or rutile, whereas adjacent grains of primary hematite are well intergrown with these other oxides. In addition, there are clear color differences between primary and secondary hematite, as described in table 27.

Hematite-Ilmenite With Minor Rutile

Two principal intergrowths have compositions nearly between hematite and ilmenite. Ilmenohematite is the intergrowth richer in hematite, and hemoilmenite is the intergrowth richer in ilmenite. Both may be intergrown with minor amounts of rutile. In such cases, the prefix "rutilo" has been added as a modifier to the principal name.

Ilmenohematite consists of host hematite with intergrown ilmenite (figs. 16, 17). The ilmenite is intergrown in two distinct forms. The largest part of the intergrown ilmenite (probably 70–80 percent) is in the form of thick lenses. The remainder is in the form of very thin disks that appear as thin lenticular films in polished-surface cross section. Both the lenses and films are oriented in the basal plane of the host hematite. The individual thick lenses may be as much as 1/10–1/5 the width of the entire grain of ilmenohematite. There are rarely more than 6–10 such lenses in any single grain. The cross section of individual films is about 1/100 the width of the host grain. There is definitely a size discontinuity between the thick lenses and thin disks of intergrown ilmenite (fig. 17). The hematite host just next to the thick lenses of ilmenite is quite devoid of any thin disks of ilmenite (fig. 17). In addition, within the thick lenses of ilmenite, thin disks of

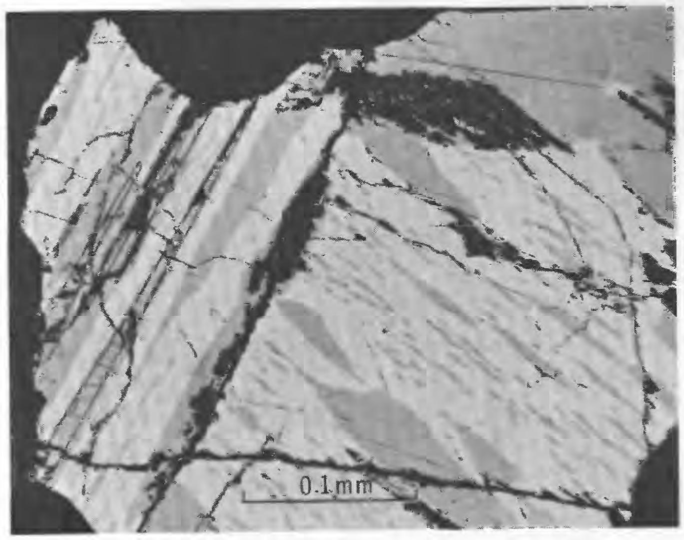

FIGURE 16.—Photomicrograph of ilmenohematite and magneto-ilmenohematite. The grain on the right is ilmenohematite. Large lenses of ilmenite are apparent and fine films of ilmenite are faintly visible in the host hematite. The grain on the left is magneto-ilmenohematite. Note the ilmenite selvage that completely surrounds each magnetite tablet. White, hematite; medium gray, magnetite; dark gray, ilmenite. Sample 149, reflected light.

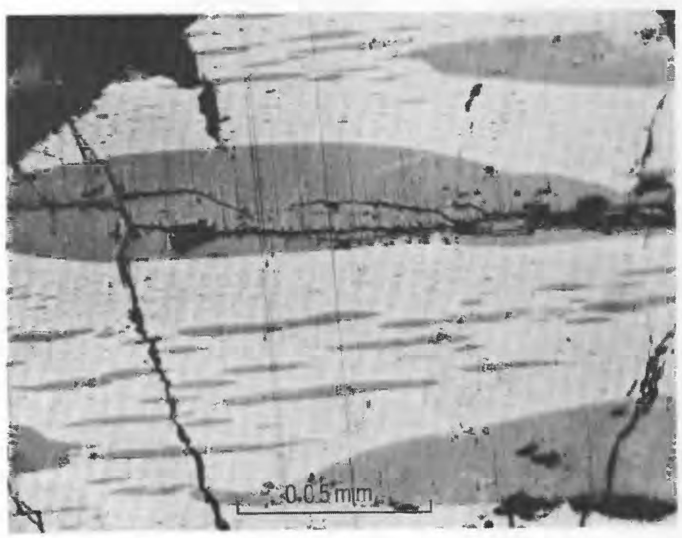

FIGURE 17.—Photomicrograph of ilmenohematite. The host hematite contains lenses and extremely fine films of ilmenite, and there is a size discontinuity between the lenses and the films. The large lenses of ilmenite contain similar fine films of hematite. The large lenses of ilmenite formed during the early stage of exsolution history, whereas the films of ilmenite and hematite formed during the last stage. Note the rhombohedral twin lamallae. White, hematite; gray, ilmenite. Sample 149, reflected light.

hematite are usually present. These disks of hematite are similar in form, size, and orientation to the ilmenite disks within the hematite host.

Hemoilmenite consists of host ilmenite with intergrown hematite (fig. 18). In hemoilmenite, hematite occurs as thick lenses and thin disks in host ilmenite. Within the thick lenses of hematite are thin disks of ilmenite. In addition, the host ilmenite is devoid of disks of hematite adjacent to the thick lenses of hematite. All the lenses and films are oriented in the basal plane. This intergrowth is homologous to the ilmenohematite intergrowth.

Intergrowth rutile is seldom present in hemoilmenite. Where present, it occurs as irregular lenses and as included small grains. Rutile intergrown with ilmenohematite may also be in lens and grain form, but in many places it occurs as long thin flat disks oriented in the rhombohedral plane of the host hematite. Thus, the rutile disks are at oblique angles to the lenses and films of ilmenite which are oriented in the basal plane of the host hematite. No clear evidence was observed to indicate that the rutile disks formed other than simultaneously with the whole intergrowth.

In most samples of the magnetite-quartz-potassium-feldspar gneiss, ilmenohematite or rutilo-ilmenohematite is present. In some samples hemoilmenite is also

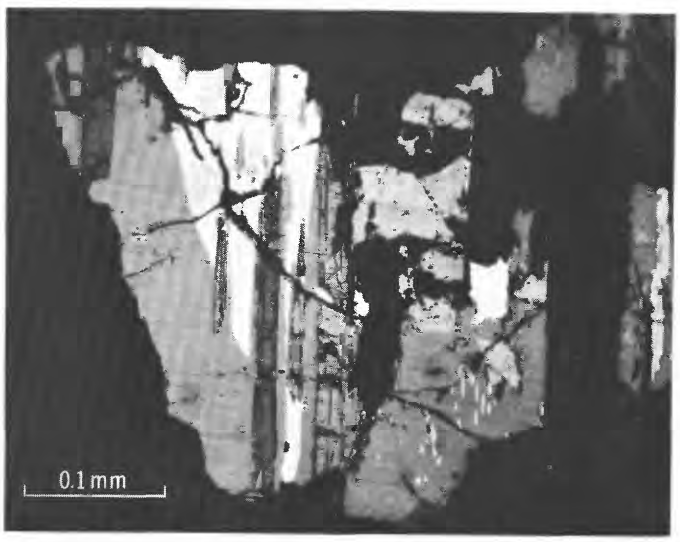

FIGURE 18.—Photomicrograph of magneto-hemoilmenite. The host ilmenite contains large lenses and fine films of hematite, and there is a size discontinuity between the lenses and the films. The large lenses of hematite, containing similar fine films of ilmenite, formed during the early stage of exsolution history and the fine films of hematite and ilmenite formed during the last stage. The magnetite tablets are surrounded by an ilmenite selvage and commonly lens out abruptly into the host ilmenite. Note the two patches of altered ilmenite (metailmenite) on the far right side of the grain. White, hematite; medium gray, magnetite; dark gray, ilmenite. Sample 149, reflected light.

present, but it is less abundant than ilmenohematite. Ramdohr (1926, 1950) proposed a hypothetical temperature-composition phase diagram for the ilmenite-hematite system in which there is complete solid solution at high temperature and exsolution at lower temperatures. This relationship has been confirmed by more recent experimental work (Carmichael, 1961; Buddington and Lindsley, 1964). Certainly, the textural relationships described above can be interpreted as being due to the exsolution of various solid solutions between ilmenite and hematite. Thus, hemoilmenite represents the solid-solution member richer in ilmenite, and the ilmenohematite represents the solid-solution member richer in hematite. As some samples carry both these solid-solution members, it may be concluded that the magnetite-quartz-potassium feldspar gneiss came to equilibrium at a temperature below the crest of the solvus curve. As they cooled, these two solid solutions would exsolve to yield the independent intergrowths of hemoilmenite and ilmenohematite. Some samples carry only ilmenohematite or rutilo-ilmenohematite without the corresponding solid-solution member, hemoilmenite. Generally, the composition of such intergrowths is more hematitic than if hemoilmenite were present. This fact suggests that the rock formed at the same temperature but that the bulk composition of the oxide fraction was to the hematite side of the solvus curve, so that only a single solid-solution member formed, which exsolved upon cooling when the solvus curve was intersected.

As pointed out very recently by Buddington and Lindsley (1964), the extent of solid solution between ilmenite and hematite is controlled not only by temperature but also by oxygen partial pressure and the composition of coexisting titaniferous magnetite. Unfortunately, because experimental data are still lacking for much of the hematite-ilmenite range of composition, that is, highly oxidized systems (Buddington and Lindsley, 1964, p. 340), definite statements cannot be made concerning the oxygen pressure and temperature conditions that prevailed during the crystallization of these phases.

HEMATITE-RUTILE WITH MINOR ILMENITE

Intergrowths of hematite, rutile, and ilmenite, including hemorutile, rutilohematite, and ilmeno-rutilohematite, and pure rutile, are very common in the magneite-quartz potassium feldspar gneiss. In the variety ilmeno-rutilohematite (fig. 19), hematite forms the host mineral, and rutile is intergrown as thin lenses and as thin flat disks oriented parallel to the basal plane and to the rhombohedral(?) plane of the hematite, respectively. Ilmenite occurs as lenses parallel to the basal plane and is altered in various degrees to a

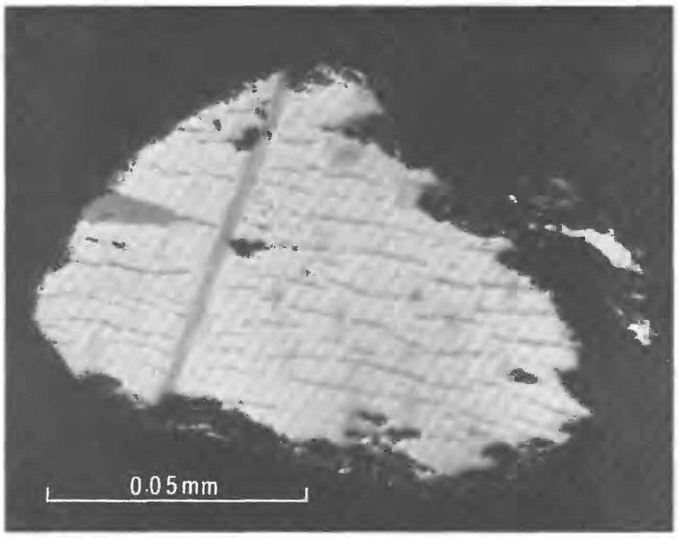

FIGURE 19.—Photomicrograph of ilmeno-rutilohematite. The rutile bodies are oriented parallel to the basal and rhombohedral(?) planes. Ilmenite is subordinate to rutile. Light-gray host, hematite; gray lenses, ilmenite and rutile; flat gray, disks, rutile. Sample 145, reflected light.

mixture of rutile and hematite (see subsequent section on metailmenite). Rutilohematite is morphologically very similar to ilmeno-rutilohematite. Rutile is again intergrown as lenses parallel to the basal plane and as disks parallel to the rhombohedral(?) plane, but in places it occurs as irregular grains and masses near the border of the host hematite grain. No ilmenite is present. Hemorutile is a third variety of intergrowth which consists of host rutile and intergrown hematite (fig. 20). The hematite occurs as small lenses and forms no more than 10 percent of the intergrowth. Usually, a dark gray nonopaque mineral (corundum?) in long lenticular blades is intergrown with the rutile.

Following the suggestion of Ramdohr (1939) and Buddington, Fahey, and Vlisidis (1963, p. 140), it is postulated that a hematite-rutile solid-solution series exists at high temperatures and that hemorutile and rutilohematite are exsolution intergrowths. The presence of two primary solid-solution mixtures indicates only limited solid solution between hematite and rutile at the temperature of formation of the magnetite-quartz-potassium feldspar gneiss—that is, that crystallization took place below the crest of the hypothetical solvus curve.

INTERGROWTHS OF MAGNETITE WITH NONMAGNETIC IRON AND TITANIUM OXIDES

In the magnetite-quartz-potassium feldspar gneiss, intergrowths of magnetite with ilmenohematite and hemoilmenite (figs. 16, 18) are common, and intergrowths of magnetite with rutilohematite (fig. 21),

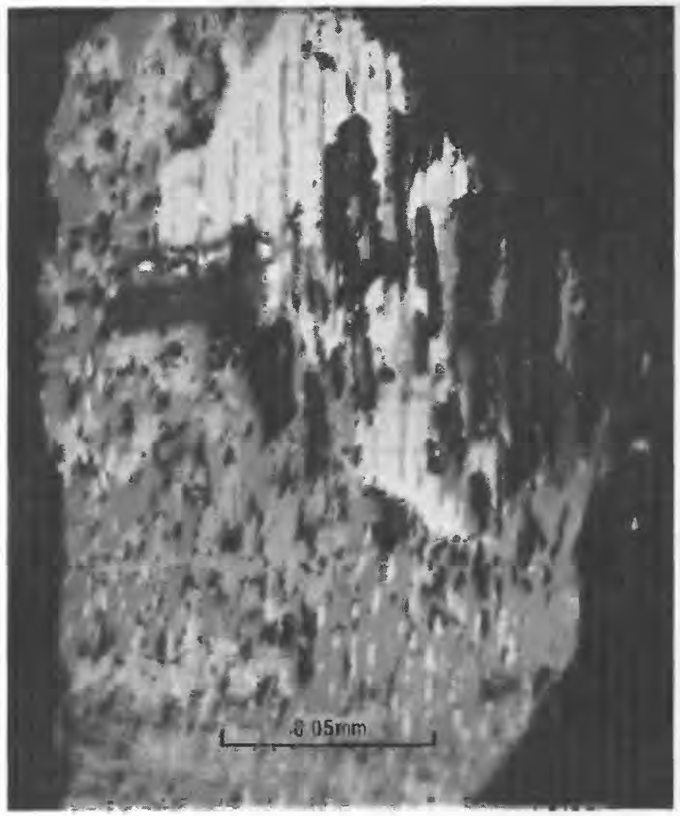

FIGURE 20.—Photomicrograph of hemorutile. Rutile is host to large and small lenses of hematite. Medium gray, rutile; white, hematite. Sample 153, reflected light.

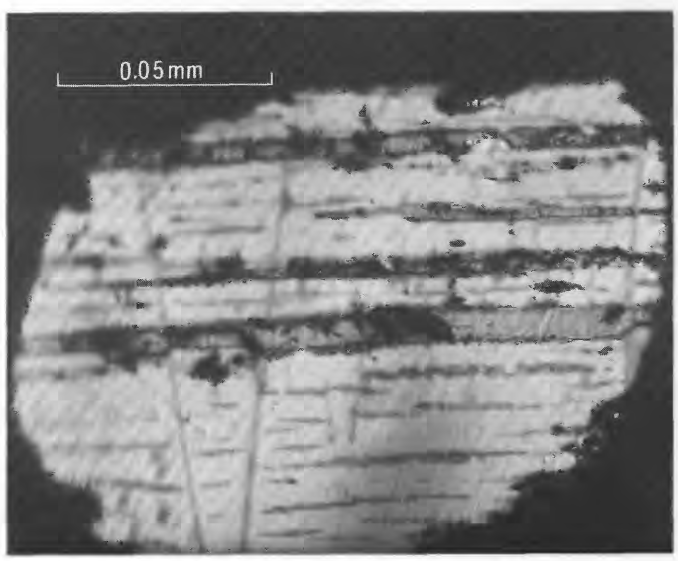

FIGURE 21.—Photomicrograph of magneto-rutilohematite. Hematite is host to basal lenses and rhombohedral disk of rutile and to tablets of magnetite. The magnetite has been etched with HCl. Light gray, hematite; medium gray, magnetite; dark gray, rutile. Sample 153, reflected light.

hemorutile, and rutile (fig. 22) are abundant. In such intergrowths, magnetite occurs as thick (as much as one-tenth the width of the grain) tablets which are oriented parallel to the basal plane of the host and which usually extend completely across the host grain. Thus, the magnetite tablets are more continuous than the lenses of ilmenite and hematite. The intergrown magnetite is always slightly martitic (5–10 percent) and, as elsewhere, the martite occurs as films and spikes parallel to the octahedral plane of the magnetite. Magnetite tablets in ilmenohematite (fig. 16) and hemoilmenite (fig. 18) never make contact with the hematite member of the intergrowth. Instead, a thin layer or selvage of ilmenite always separates the magnetite tablet from the rest of the intergrowth. However, in intergrowths of magnetite with rutilohematite (fig. 21), hemorutile, or rutile (fig. 22), the magnetite tablets make contact directly with hematite and rutile. The amount of intergrown magnetite varies from sample to sample and from grain to grain within the same sample. Thus, the amount of magnetite intergrown with the ilmenitehematite series ranges from 0 to 10 percent. The percentage of magnetite intergrown with the hematite-rutile series, and particularly rutile, is greater than that intergrown with the ilmenite-hematite series. However, the amount of intergrowth is just as variable and ranges from 0–50 percent.

FIGURE 22.—Photomicrograph of magnetorutile. Rutile appears in two colors because of strong internal reflections. Magnetite tablets are etched with HCl and generally extend across the entire host grain of rutile. Medium gray and white, rutile; very dark gray, magnetite. Samples 154, crossed nicols.

There are several possible interpretations for these intergrowths of magnetite with nonmagnetic iron and titanium oxides. It seems unlikely that these intergrowths could result from the simple exsolution of a solid solution between magnetite-hematite-ilmenite and rutile. Experimental data (Darken and Gurry, 1946) show that there is no solid solution of magnetite in hematite. In addition, rarely has ilmenite with exsolution intergrowths of magnetite been reported (Bordet and Geffroy, 1952). It is also difficult to explain on the basis of a solid-solution theory why the amount of magnetite varies so radically from grain to grain within a single sample. A second possible interpretation is that primary magnetite was replaced by hematite, ilmenite, and rutile. It seems unlikely, however, that such regular intergrowths could form by this process. A third interpretation is really a modification of a solid-solution theory and might be termed "incongruent" exsolution because of its similarity to incongruent melting. In this case, it is proposed that a solid solution between hematite and ilmenite, upon cooling, could exsolve as a mixture of hematite, ilmenite, rutile, and magnetite, simply by the reaction of hematite and ilmenite, as indicated below, to yield magnetite and rutile:

$$[Fe_2O_3 + FeTiO_3] = FeFe_2O_4 + TiO_2.$$

For such a process to take place, it seems necessary to postulate special physical conditions under which magnetite and rutile are more stable than hematite and ilmenite. In this connection, it is interesting that Ramdohr (1939) recorded that solid solutions of hematite-ilmenite were replaced by a mixture of magnetite and rutile. More pertinent in this regard, Verhoogen (1962, p. 168) has shown that on the basis of free-energy data, the right-hand pair of the equation above are the more stable. A fourth interpretation, following the proposal of Ramdohr (1939), is that hematite lenses in such intergrowths as ilmenohematite, hemoilmenite, and hemorutile were reduced to magnetite. However, as the morphology of the magnetite tablets described here is so different from that of the hematite lenses in ilmenite-hematite-rutile intergrowths, this proposal seems unlikely. Buddington and Lindsley (1964) suggest a modification of Ramdohr's hypothesis whereby a certain amount of the Fe_2O_3 is reduced in solid solution and there is penecontemporaneous exsolution of magnetite. Finally, a possible explanation of the intergrowths is that of simultaneous crystallization of magnetite tablets with the respective solid solution of the nonmagnetic iron and titanium oxides, which during cooling would exsolve into the complex intergrowths of magnetite with ilmenite, hematite, and rutile.

METAILMENITE

In many samples the tablets of ilmenite in magnetite and the ilmenite intergrown with hematite are partially or completely altered. The alteration consists of a very fine grained whitish-gray mottled aggregate which is brighter than the host ilmenite and which shows strong yellow to brownish-red internal reflections under crossed nicols (see fig. 18). The aggregate is composed of rutile and hematite and in some cases relict grains of ilmenite. Following the useage of Buddington, Fahey, and Vlisidis (1963, p. 140), the aggregate is called metailmenite. The alteration probably involves the oxidation of ilmenite to rutile and hematite, according to the following chemical reaction:

$$2\ FeTiO_3 + \tfrac{1}{2}\ O_2 = Fe_2O_3 + 2\ TiO_2.$$

Visual estimates of the mottled aggregates indicate that rutile is considerably more abundant than hematite. This should be the case if the alteration process took place as indicated above (2 moles of rutile produced per 1 mole of hematite). Thus, the petrographic estimates substantiate the proposed theory of alteration.

In the magnetite-quartz-potassium feldspar gneiss, martite is ubiquitous. In many samples, however, ilmenite is unaltered. Although the martitization of magnetite and the alteration of ilmenite to rutile and hematite are both secondary oxidation processes, it seems likely that they take place under slightly different conditions. It appears as though martite forms more readily than the metailmenite.

Ramdohr (1939) has recorded the alteration of ilmenite to rutile plus hematite and to anatase plus hematite. He believes such alteration to be of hydrothermal origin. His observations and interpretations agree in principle with those presented here.

IRON–TITANIUM OXIDES OF THE OTHER ROCKS

Sufficient detailed studies of the iron-titanium oxides of the calcium-rich gneiss, the biotite-quartz-feldspar gneiss, and the quartz-microcline gneiss have been done to enable some general comparisons with the magnetite-quartz-potassium feldspar gneiss.

The iron-titanium oxide mineral assemblage in the calcium-rich gneiss is very similar to the assemblage in the magnetite-quartz-potassium feldspar gneiss. The magnetite carries about the same amount of intergrown ilmenite and is always slightly martitic. Primary ilmenohematite or rutilohematite is always present. These intergrowths are morphologically similar to those described from the magnetite-quartz-potassium feldspar gneiss and are also interpreted as exsolved solid solutions. Ilmenite may be partly altered to rutile and hematite (metailmenite). The proportion of primary hematite relative to magnetite is somewhat greater in the calcium-rich gneiss than in the magnetite-quartz-potassium feldspar gneiss. In addition, hemoilmenite was not observed. This indicates that the bulk composition of the nonmagnetic fraction was near the hematite end of the hematite-ilmenite join. These facts indicate a high oxidation grade for the calcium-rich gneiss.

The biotite-quartz-feldspar gneiss carries accessory ilmenomagnetite and ilmenite and, rarely, hemoilmenite. Primary hematite and martite are absent. Ilmenite is partially altered to rutile and hematite. This assemblage is distinct from that in the magnetite-quartz-potassium feldspar gneiss where the iron and titanium oxides consist of a variety of intergrowths.

The quartz-microcline gneiss carries ilmenomagnetite as a major accessory mineral and only minor ilmenite. No primary hematite or rutile is present and only rarely is a very minor amount of martite present. Ilmenite is usually partially altered to rutile and hematite. A partial chemical analysis (table 29) of the magnetic and nonmagnetic fractions from a sample of the quartz-microcline gneiss shows that the ilmenite contains about 7.5-mole-percent hematite in solid solution and the magnetite contains about 8.2-mole-percent ilmenite and about 14.0-mole-percent excess FeO. Intergrown parallel to the cube plane in the magnetite are very minute disks or lenses of a dark gray nonopaque

TABLE 29.—*Partial chemical analyses of magnetic and nonmagnetic fractions of iron-titanium oxides from the quartz-microcline gneiss*

Magnetic fraction			
Weight (percent)		Mole (percent)	
Fe_2O_3	53.72	Fe_3O_4	77.8
FeO	31.06	Fe_2O_3	.0
TiO_2	2.80	$FeTiO_3$	8.2
		FeO (excess)	14.0
	87.58		100.0
Nonmagnetic fraction			
Fe_2O_3	6.44	Fe_2O_3	7.5
FeO	41.45	$FeTiO_3$	92.5
TiO_2	40.44	TiO_2	.0
	88.33		100.0

mineral. These are probably exsolved Fe-rich spinel, such as hercynite, and would account for the excess FeO in the analysis. The high TiO_2 content (2.80 weight percent) of the magnetite is in contrast to the low TiO_2 (0.14–1.39 weight percent) content of magnetite from the magnetite-quartz-potassium feldspar gneiss. According to Buddington, Fahey, and Vlisidis (1955), this difference suggests that the quartz-microcline gneiss crystallized at a slightly higher temperature than the magnetite-quartz-potassium feldspar gneiss, and in addition, the quartz-microcline gneiss formed at about the same temperature as microcline-rich granitized rocks of the Adirondacks. Based on the more recent work of Buddington and Lindsley (1964, p. 337, table 6), the compositions of the coexisting ilmenomagnetite and ferrianilmenite from the quartz-microcline gneiss are similar to those of oxides in hornblende granite gneiss (sillimanite-almandine-orthoclase subfacies) of the Adirondack area, which are estimated on the basis of experimental data to have crystallized at about 600°C.

PARTIAL CHEMICAL ANALYSES

The approximate chemical composition of a sample of magnetite-quartz-potassium feldspar gneiss and a sample from a magnetite-rich layer, both from the Condon cut in the Edison area, are presented in table 30. During the separation of a quartz-feldspar concentrate from these samples, a partial mode (weight percent) was determined. Using this weight mode, the partial chemical analysis of the quartz-feldspar concentrate (K_2O, Na_2O, CaO, and BaO) and the partial chemical analyses of the iron and titanium oxide fractions, it was possible to recalculate a partial chemical analysis for the two samples.

The striking chemical features of the magnetite-quartz-potassium feldspar gneiss are the very high SiO_2 content, the high ratio of Fe_2O_3 to FeO, and very high ratio of K_2O to Na_2O. The average and maximum SiO_2 content of granite listed by Daly (1933) is 70.18 and 71.06 percent, respectively. The K_2O-to-Na_2O and Fe_2O_3-to-FeO ratios in granites from Daly's list do not approach the high values present in the gneiss. According to the compilations of Nockolds (1954), biotite alkali granite with an average of 75.01 percent SiO_2 has the maximum average SiO_2 content of all granites. None of the average compositions of calc-alkali or alkali granites, as listed by Nockolds, has as high a SiO_2 content or K_2O-to-Na_2O and Fe_2O_3-to-FeO ratios as does this gneiss. Clearly, the magnetite-quartz-potassium feldspar gneiss does not belong to the normal granite clan. Although its SiO_2 content exceeds that of similar gneisses of the Adirondacks, the overall composition of the gneiss corresponds closely to a sillimanite-quartz-microcline granite gneiss (table 30, B-13) of the Adirondacks (Buddington, 1957, p. 300, table 5) which is interpreted as a metasomatized metasediment.

TABLE 30.—*Weight modes and partial chemical analyses of magnetite-quartz-potassium feldspar gneiss and a magnetite-rich layer tabulated with the chemical composition of a sillimanite-quartz-microcline granitic gneiss from the Adirondacks*

[N.d., not determined]

	Weight mode 145	Weight mode 144		Chemical composition (weight percent) 145	Chemical composition (weight percent) 144	Chemical composition (weight percent) B-13
Quartz	54.3	21.6	SiO_2	76.4	32.9	71.44
Potassium feldspar	34.1	18.8	Al_2O_3	6.3	3.63	14.89
Magnetite	3.0	47.9	Fe_2O_3	3.41	34.6	2.40
Hematite	1.3	--------	FeO	1.00	13.0	1.38
Ilmenite	.1	--------	MgO	n.d.	n.d.	.30
Rutile	.1	--------	CaO	.04	.04	.50
Other	7.1	11.7	Na_2O	.41	.18	1.36
Biotite	[1] xx	[2] x	K_2O	5.15	2.46	5.99
Sericite	xx	x	BaO	.05	.92	n.d.
Sericite and epidote mixture.	--------	xx	H_2O+	n.d.	n.d.	.52
Sillimanite	xx	x	H_2O-	n.d.	n.d.	.08
Apatite	x	xx	TiO_2	.10	.03	.65
Zircon	x	x	P_2O_5	n.d.	n.d.	.22
Sphene	x	--------	MnO	n.d.	n.d.	.03
Quartz/feldspar	1.60	1.15	Rest	[3] 7.14	[4] 12.24	------
Total				100.00	100.00	99.76

[1] Present in amounts greater than 2 percent.
[2] x Present in amounts less than 2 percent.
[3] Consists mostly of SiO_2, Al_2O_3, and K_2O from biotite and sillimanite, and minor Fe_2O_3, FeO, MgO, CaO, H_2O, and P_2O_5.
[4] Consists mostly of SiO_2, Al_2O_3, CaO, H_2O, and P_2O_5 from biotite, apatite, sericite, and epidote.
145. Magnetite-quartz-potassium feldspar gneiss, Condon cut, Edison area, Franklin quadrangle, New Jersey.
144. Magnetite-rich layer, adjacent to 145.
B-13. Sillimanite-quartz-microcline granitic gneiss, from diamond-drill core, Skate Creek ore body, Oswegatchie quadrangle, New York. Analyst, L. C. Peck. (Quoted from Buddington, 1957, p. 300.)

Naturally, the striking feature of the partial chemical analysis of the magnetite-rich layer is the high content of Fe_2O_3 and FeO. It is interesting to note that the quartz-to-feldspar and the K_2O-to-Na_2O ratios in the gneiss and magnetite-rich layer are not much different. It can be concluded that the quartz and feldspar decrease together as magnetite increases. However, BaO, which is carried by potassium feldspar, is enriched in the magnetite-rich sample. This increase corresponds with the observation at the Benson mines made by Leonard (1951), where the BaO content in potassium feldspar separated from magnetite-rich rocks was somewhat greater than in potassium feldspar from wallrock gneiss similar to the mixed-gneiss complex.

ORIGIN AND CLASSIFICATION

To explain the genesis of the magnetite deposits in the New Jersey Highlands, the process or processes that caused the initial enrichment of iron and the nature of the process or processes whereby the iron was emplaced must be understood. These two questions are closely related, and the explanation of either one will probably indicate or restrict the answer of the other.

Early geologists related the actual enrichment of iron to some phase of magmatic activity and proposed that the iron was carried and emplaced by some sort of an ore magma. An early interpretation (Rogers, 1840) was that the magnetite deposits are veins resulting from the forcible injection of fluid ore into preexisting rocks. Rogers apparently believed that the magnetite deposits were magmatic in the strictest sense, that is, a magnetite magma (Shand, 1947). Spencer (1904) and Bayley (1910) both expressed the opinion that the magnetite deposits were products of igneous activity. Bayley (1910, p. 149) clarified this opinion by stating:

In all cases the ores are regarded as being of magmatic origin—that is, the source of their material is thought to have been the deep-seated magmas, portions of which, upon being intruded into the overlying rocks, solidified as the various gneisses now constituting the principal rocks of the Highlands ridges.

Bayley (1910, p. 151) clarified his views as to the nature of the emplacement process:

Very probably the vehicle of transportation was a hot aqueous solution or possibly a vapor which emanated from the same magmatic source * * *.

and further,

The channels through which the solutions circulated presumably afforded the most favorable opportunities for deposition of mineral matter, and such portions of the rock naturally became richer in magnetite * * *. In some places there may have been replacement of the silicates by magnetite * * *.

Modern opinion regarding the genesis of the magnetite deposits does not differ much from that expressed by Bayley and is summed up by Sims (1953) in several statements:

The source of ore-forming fluids was a cooling igneous mass. The process whereby the residual liquor of a magma undergoing progressive crystallization is continually enriched in volatile components has been described by Bowen (1928, p. 293). During crystallization the residual liquor is progressively enriched in H_2O and other mineralizers, but there is no general agreement whether the mineralizers given off by the cooling igneous mass escape as a liquid or a vapor phase.

Both field and laboratory evidence suggest that all the magnetite deposits originated by metasomatic replacement of favorable host rocks.

The ore-forming fluids did not, therefore, have large openings through which to migrate, but instead had to migrate along intergrain boundaries, discontinuous small fractures, and granulated zones that were rendered more permeable than the surrounding rock by microbrecciation. To penetrate for long distances in these zones the fluids must have had a high degree of mobility, and were either pneumatolytic or hydrothermal, or more probably both. Evidence is strong against the tenet that the magnetite was introduced as a liquid melt, as suggested by Shand (1947).

Sims (1953, p. 283) is very specific about the origin of the magnetite deposits which he has studied and states:

The magnetite deposits in the Dover district were derived from the granite magma that consolidated to form hornblende granite and alaskite. During the progressive crystallization of this magma a more mobile and highly volatile portion of the magma was concentrated adjacent to inclusions and in the crests of certain large anticlines. This magma, which consolidated to form alaskite, was split off prior to the crystallization of the pegmatites. Further differentiation of the alaskitic magma by progressive crystallization concentrated the volatiles still more and these ferriferous fluids escaped from the crystal system, migrated along the microbrecciated zones, and replaced the rocks within these zones to form magnetite bodies.

More recently, Buddington (1966, p. 508) has presented a rather thorough review and comprehensive discussion of the origin of the Precambrian magnetite deposits of New York and New Jersey. He concludes that—

The magnetite (and crystalline hematite) deposits were formed as replacements of metasedimentary, migmatitic, and metasomatic gneisses, amphibolites, skarns and contaminated granite gneiss by hypogene solutions. The solutions are inferred to be of magmatic origin and related to the large volume of associated igneous granites.

In further elaboration Buddington (1966 p. 509) states:

The balance of evidence thus favors an origin for the iron ore deposits by hypogene solutions of magmatic origin; most of the iron must have been derived in the process of development of an alaskitic magma from a granitic magma of hornblende granite composition, but a minor part was derived by leaching of country rock.

Although, as indicated above, Buddington recognizes a nonmagmatic source for some of the iron—that is, leaching of country rock, in contrast to a magmatic source for the bulk of the iron, others have postulated that the iron enrichment took place during a sedimentary cycle and that the magnetite deposits are metamorphosed sedimentary iron beds (Kitchell, 1857). Landergren (1948) in his geochemical studies of Precambrian Swedish iron ores is strongly in favor of initial iron enrichment during a sedimentary (exogene) cycle, but states that this sedimentary material "entered into an endogene phase of development in an orogenic cycle." Landergren points out that the sedimentary iron has been transported and redeposited. As a result, the magnetite deposits take on the character of deposits formed entirely through endogene processes, that is, products of igneous activity. Landergren (1948, p. 174) sums up his point of view as follows:

All the geochemical features indicate that the enrichment of iron took place *before* those endogene phenomena which have obscured these trends and which have given the iron ores the indisputable geological character of what we call a magmatic ore.

Landergren prefers not to commit himself on the possible ways in which iron may be transported, such as solution transfer, gaseous transfer, or solid diffusion. Instead, he sums up his position by the following general statement (Landergren, 1948, p. 158):

> A remobilization of iron in the endogene phase of development may take place under certain conditions depending on the composition of the material entering in the endogene phase, on the content of volatiles present, and on the temperature.

A view that has aspects in common with Landergren's opinions is that the iron enrichment took place by a process related to metamorphic differentiation, and that during regional metamorphism and metasomatism, iron was driven out of some rocks and concentrated and fixed in other sites (Ramberg, 1952, p. 265–266; DeVore, 1953). Proponents of this hypothesis generally propose solid-state diffusion or grain-boundary migration as the mechanism of transport of the iron.

SOURCES OF IRON

Several lines of evidence indicate that the initial source of the iron was from residual solutions formed from the progressive crystallization of granitic magma. First, there are experimental data (Bowen and Schairer, 1935) and petrologic data (Wager and Deer, 1939, p. 133; Fenner, 1929, p. 242) which indicate that absolute content of iron increases in the residual liquids of synthetic melts and basaltic magma, respectively. Sims (1953) believes that the presence of magnetite in a variety of host rocks, such as gneiss, skarn, and granite, is good evidence that the initial iron enrichment did not take place during a sedimentary cycle. Apparently, Sims would expect less variation in the host rock of sedimentary iron deposits. Within the Franklin-Hamburg area, magnetite deposits occur in many kinds of rocks: hence, Sims' reasoning is applicable but does not rule out the possibility that the iron could be of exogene origin and redeposited in other host rocks during an endogene cycle (Landergren, 1948).

The regional association of magnetite deposits with granitic igneous rocks is evidence that the initial iron enrichment took place during a magmatic cycle. Buddington (1939, p. 178; 1966) has pointed out that Adirondack magnetite deposits are in the highland part of the Adirondacks where about 85 percent of the rocks are of igneous origin. In the Grenville lowlands in the extreme northwestern Adirondacks, magnetite deposits are absent, and only about 15 percent of the rocks are of igneous origin. It is clear that the ores of the mixed-gneiss complex are associated with rocks of igneous origin, such as hornblende granite and alaskite, pyroxene granite and alaskite (pls. 1, 2). The similarity of this association to those in the Adirondacks is some evidence that granite was the source of the iron.

Enclosed in the pyroxene granite just southeast of the Edison area at the Goble mine (pl. 1) (Bayley, 1910) is a small magnetite-rich pegmatite. This and other small magnetite deposits are unrelated to any Grenville-type gneiss and are always enclosed by large masses of granite. These bodies are mineralogically either a magnetite-rich pegmatite, or they contain little or no feldspar and may be classed as nonpegmatitic. If pegmatites represent a relatively late stage of crystallization, then the magnetite pegmatites are evidence that considerable iron enrichment took place during the fractional crystallization of the granite magma. The nonpegmatitic magnetite bodies are also indicative of late-stage enrichment of iron, but perhaps they owe their distinctive composition to a process of enrichment different from fractional crystallization, for example, gaseous transfer, liquid immiscibility. In any case, it is proposed that the residual fluid from which these small magnetite bodies crystallized may have been the immediate source of much of the iron deposited in the mixed-gneiss complex.

There is evidence that some of the iron in the mixed-gneiss complex is of sedimentary origin. It seems possible that magnetite-quartz gneiss layers could actually be metamorphosed iron formation (taconite). Their bulk composition is similar enough to taconite to make such a hypothesis tenable (James, 1954). In this case, several problems related to the origin of the iron deposits might be explained. First, such layers could provide the source of all, or a substantial part of the iron. Second, the concentration of the magnetite deposits into certain zones and layers would be a consequence of the original sedimentary stratification. Furthermore, the contact relations between the magnetite-quartz gneiss and magnetite–quartz–potassium feldspar gneiss (fig. 14), which seem to indicate that the latter rock may have been younger than the former, could be explained as the result of potassium metasomatism of the older iron-rich (taconite) layer. The magnetite-rich layer in the magnetite-quartz-potassium feldspar gneiss at the contact with the magnetite-quartz gneiss (fig. 14) could be regenerated iron stemming from the latter rock as a result of the potassium metasomatism. In this regard, it should be emphasized that Buddington (1966, p. 506–508) has presented some serious objections to the origin of the Precambrian iron ore deposits (of New York and New Jersey, collectively) through modification of metasedimentary iron-formation.

The quantitative distribution of magnetite and primary hematite may be an additional indication that some of the iron is of sedimentary origin. The high hematite-magnetite ratio in rocks of decided metasedimentary character, such as sillimanite gneisses, may be explained if all the ferric iron in primary hematite is of sedimentary origin. In this case, the high oxidation state of the iron is inherited from the sedimentary cycle, and, therefore, metasedimentary rocks that have not been chemically reconstituted (as by magnetite-rich fluids) would be expected to have a high proportion of hematite. It might be argued that during progressive crystallization of a parent granite magma, the residual solutions were enriched in ferric iron above that necessary to form magnetite. Hence, hematite could crystallize directly from such fluids and be of true endogene origin. However, the magnetite bodies previously discussed, which are associated with granites and believed to represent the crystallized product of residual fluids from these granites, do not contain primary hematite. In addition, many of the magnetite deposits in the Grenville rocks of the Adirondacks (Leonard, 1951, Leonard and Buddington, 1964) and New Jersey (Sims, 1953) carry little or no primary hematite, whereas those that do carry primary hematite commonly contain considerably less than the rocks of the mixed-gneiss complex. These facts offer evidence that the magnetite ore fluids carried little or no ferric iron in excess of that necessary to form magnetite, and, therefore, it is plausible that at least some of the ferric iron in the mixed-gneiss complex originated during the exogene cycle.

MODE OF TRANSPORTATION OF IRON

The concept of a magnetite magma has been considered by previous geologists (Rogers, 1840; Shand, 1947) and has been recently discounted by Sims (1953). For the magnetite deposits in the mixed-gneiss complex, absolutely no evidence suggests that such an ore magma was injected into the gneiss. Most investigators postulate that the residual iron-rich fluids were either hydrothermal or pneumatolytic or both.

It is well known that ferrous and ferric iron form very volatile halogen complexes; hence, it has frequently been proposed that iron has "boiled off" the residual granite magmas as such gaseous complexes. This mechanism of gaseous transfer offers both an explanation of how the iron was initially concentrated from the parent magma and how the iron was transported. Aside from its simplicity, no geologic evidence indicates that such a process was important in the formation of the magnetite deposits in the mixed-gneiss complex.

Very little is known about the solubility of magnetite in aqueous solutions at high temperatures and pressures. The experimental work of Holser (1952) and Holser and Schneer (1953) indicates that the solubility of magnetite in pure H_2O is very slight, but that it increases greatly in mildly acid water. More information is needed, however, before it can be positively concluded that large amounts of iron can be carried in an aqueous solution. Nevertheless, the fact that phosphorus (in apatite), manganese (in garnet), and barium (in potassium feldspar) are all enriched in magnetite-rich layers is evidence that an iron-rich ore fluid with a characteristic chemical composition did play an important part in the genesis of the mixed-gneiss complex and the related magnetite deposits.

MODE AND TIME OF EMPLACEMENT OF THE IRON

Whatever the nature of the fluids whereby iron was transported—gaseous phase, hydrothermal phase, or hypogene solution—it is clear that the emplacement of iron in the mixed-gneiss complex was a pervasive or permeative process. It is emphasized that magnetite occurs throughout the complex and varies in mode of distribution from disseminated grains to lenticular aggregates of several grains, and progressively to layers of more than 50 percent magnetite. There is no evidence that an ore magma has forceably intruded the gneisses. Chemical rather than mechanical processes seem to have been more important. There is little evidence that any openings (cracks, fissures, microbrecciation) were available for the ore fluids. Instead, the magnetite must have metasomatically replaced preexisting mineral materials to make room for itself. The replacement of minerals by magnetite is similar to the process that was going on in the case of potassium feldspar metasomatism. In layers where magnetite was added, other constituents were flushed out into adjacent layers. In some places magnetite grew at the expense of silicates; in other places silicates grew at the expense of magnetite. In essence, the entire process is deemed as metamorphic. Mineral grains grew outward from centers (nuclei), continually replacing surrounding mineral grains, continually increasing in size, and continually changing the nature of the grain-boundary interface with adjacent grains. The observed relationships either represent a final equilibrium arrangement (with regard to mineral phases, grain sizes, shapes) or a stage of crystallization leading to such a final equilibrium.

No evidence indicates that the metasomatic emplacement of the iron postdated the regional metamorphism and metasomatism of the mixed-gneiss complex. The magnetite deposits are structurally equivalent to other masses of rock within the complex, that is, foliated, lineated, and folded. The fabrics of the magnetite-rich layers and the magnetite-quartz-potassium feldspar gneiss appear equivalent. If the silicate grains are deformed, so are the magnetite grains; whereas in undeformed samples of magnetite-quartz-potassium feldspar gneiss, the magnetite grains are again texturally equivalent to the silicate grains, that is, undeformed. There are no reaction rims between magnetite and adjacent silicates. The data all indicate that the entire mineral assemblage in the mixed-gneiss complex crystallized during the period of regional metamorphism and metasomatism.

CONCLUSION

It is proposed that the magnetite deposits in the Edison area are of complex origin. Although much of the initial concentration of the iron was probably directly related to igneous activity, as Sims (1953) and Buddington (1966) have explained, some of the iron was probably initially concentrated during a sedimentary (exogene) cycle. Therefore, it is postulated that the magnetite deposits owe their origin to processes of metamorphic differentiation and metasomatism. The process of metamorphic differentiation refers to the belief that at least some of the iron enrichment in the mixed-gneiss complex took place during the sedimentary (exogene) cycle and that this iron was redistributed during regional metamorphism (endogene cycle). The process of metasomatism refers not only to the belief that some of the iron was introduced from outside (magmatic) sources, but also to the belief that the redistribution of sedimentary iron and the emplacement of iron from extraneous sources took place by way of a metasomatic fluid which reacted with, and replaced, the host rocks of the mixed-gneiss complex. It is concluded that the magnetite deposits were formed contemporaneously with the mixed-gneiss complex—that is, that the redistribution and introduction of iron within and into the mixed-gneiss complex took place at the time of regional metamorphism and metasomatism that gave rise to the complex. The nature of the metasomatic fluid and the mechanism of iron transport are problematical.

REFERENCES CITED

Adams, F. D., 1909, On the origin of the amphibolites of the Laurentian area of Canada: Jour. Geology, v. 17, p. 1–18.

Baker, D. R., 1955, Geology of the Edison area, Sussex County, New Jersey: Princeton Univ., Princeton, N.J., Ph. D. dissert.

Balsley, J. R., and Buddington, A. F., 1958, Iron-titanium oxide minerals, rocks, and aeromagnetic anomalies of the Adirondack area, New York: Econ. Geology, v. 53, no. 7, p. 777–805.

Bayley, W. S., 1910, Iron mines and mining in New Jersey: New Jersey Geol. Survey Final Rept., v. 7, 512 p.

——— 1941, Pre-Cambrian geology and mineral resources of the Delaware Water Gap and Easton quadrangles, New Jersey and Pennsylvania: U.S. Geol. Survey Bull. 920, 98 p.

Bayley, W. S., Salisbury, R. D., and Kümmel, H. B., 1914, Description of the Raritan quadrangle [New Jersey]: U.S. Geol. Survey, Geol. Atlas, Folio 191, 32 p.

Berkey, C. P., and Rice, Marion, 1921, Geology of the West Point quadrangle, New York: New York State Mus. Bull. nos. 225–226, 152 p., 56 pls., map.

Bordet, Pierre, and Geffroy, J., 1952, Enclaves d'ilmenite dans les basaltes du Tahalra (Hoggar): Soc. Française Minéralogie Bull. 75, no. 10–12, p. 580–584.

Bowen, N. L., 1928, The evolution of the igneous rocks: Princeton, N.J., Princeton Univ. Press, 334 p.

Bowen, N. L., and Schairer, J. F., 1935, The system MgO–FeO–SiO_2: Am. Jour. Sci., v. 29, no. 170, p. 151–217.

Bowen, N. L., and Tuttle, O. F., 1950, The system $NaAlSi_3O_8$–$KAlSi_3O_8$–H_2O: Jour. Geology, v. 58, no. 5, p. 484–511.

Buddington, A. F., 1939, Adirondack igneous rocks and their metamorphism: Geol. Soc. America Mem. 7, 354 p.

——— 1948, Origin of granitic rocks of the northwest Adirondacks, *in* Gilluly, James, and others, Origin of granite: Geol. Soc. America Mem. 28, p. 21–43.

——— 1952, Chemical petrology of some metamorphosed Adirondack gabbroic, syenitic and quartz syenitic rocks: Am. Jour. Sci., Bowen Volume, pt. 1, p. 37–84.

——— 1956, Thermometric and petrogenetic significance of titaniferous magnetite [Discussion by K. S. Heier and reply by author]: Am. Jour. Sci., v. 254, no. 8, p. 506–515.

——— 1957, Interrelated Precambrian granitic rocks, northwest Adirondacks, New York: Geol. Soc. America Bull., v. 68, no. 3, p. 291–306.

——— 1966, The Precambrian magnetite deposits of New York and New Jersey: Econ. Geology, v. 61, no. 3, p. 484–510.

Buddington, A. F., Fahey, J. J., and Vlisidis, A. C., 1955, Thermometric and petrogenetic significance of titaniferous magnetite: Am. Jour. Sci., v. 253, no. 9, p. 497–532.

——— 1963, Degree of oxidation of Adirondack iron oxide and iron-titanium oxide minerals in relation to petrogeny: Jour. Petrology, v. 4, no. 1, p. 138–169.

Buddington, A. F., and Lindsley, D. H., 1964, Iron-titanium oxide minerals and synthetic equivalents: Jour. Petrology, v. 5, no. 2, p. 310–357.

Carmichael, C. M., 1961, The magnetic properties of ilmenite-hematite crystals: Royal Soc. [London] Proc., ser. A, v. 263, no. 1315, p. 508–530.

Chayes, Felix, 1952, The finer-grained calcalkaline granites of New England: Jour. Geology, v. 60, no. 3, p. 207–254.

Daly, R. A., 1933, Igneous rocks and the depths of the earth: New York, McGraw-Hill Book Co., 508 p.

Darken, L. S., and Gurry, R. W., 1946, The system iron-oxygen. II, Equilibrium and thermodynamics of liquid oxide and other phases: Am. Chem. Soc. Jour., v. 68, p. 798–816.

Darton, N. H., Bayley, W. S., Salisbury, R. D., and Kümmel, H. B., 1908, Description of the Passaic quadrangle [New Jersey-New York]: U.S. Geol. Survey, Geol. Atlas, Folio 157, 27 p.

DeVore, G. W., 1953, Temperature and the distribution of trace elements [abs.]: Geol. Soc. America Bull., v. 64, no. 12, pt. 2, p. 1412.

Engel, A. E. J., and Engel, C. G., 1953, Grenville series in the northwest Adirondack Mountains, New York: Geol. Soc. America Bull., v. 64, no. 9, p. 1013–1097.

Fenner, C. N., 1929, The crystallization of basalts: Am. Jour. Sci., 5th ser., v. 18, p. 225–253.

Goldschmidt, V. M., 1921, Die injektionmetamorphose im Stavanger-Gebiete: Vidensk.-Selsk., Kristiania, Skr. 1, 1920, no. 10, 142 p.

Greig, J .W., Posnjak, Eugen, Merwin, H. E., and Sosman, R. B., 1935, Equilibrium relationships of Fe_3O_4, Fe_2O_3, and oxygen: Am. Jour. Sci., 5th ser., v. 30, no. 177, p. 239–316.

Hagner, A. F., and Collins, L. G., 1955, Source and origin of magnetite at Scott Mine, Sterling Lake, New York: Science, v. 122, no. 3182, p. 1230–1231.

Hague, J. M., Baum, J. L., Hermann, L. A., and Pickering, R. J., 1956, Geology and structure of the Franklin-Sterling area, New Jersey: Geol. Soc. America Bull., v. 67, no. 4, p. 435–474.

Harker, R. T., 1954, The occurrence of orthoclase and microcline in the granitic gneisses of the Carn Chuinneag-Inchbæ complex, E. Ross-shire: Geol. Mag., v. 91, no. 2, p. 129–136.

Henderson, J. R., Tyson, N. S., and others, 1957a, Aeromagnetic map of part of the Hamburg quadrangle, Sussex County, New Jersey: U.S. Geol. Survey Geophys. Inv. Map GP–158, scale 1:31,680.

Henderson, J. R., Tyson, N. S., and others, 1957b, Aeromagnetic map of the Franklin quadrangle, Sussex and Morris Counties, New Jersey: U.S. Geol. Survey Geophys. Inv. Map GP–162, scale 1:31,680.

Hess, H. H., 1949, Chemical composition and optical properties of common clinopyroxenes, Part I: Am. Mineralogist, v. 34, nos. 9–10, p. 621–666.

Hinds, N. E. A., 1921, An alkali gneiss from the pre-Cambrian of New Jersey: Am. Jour. Sci., 5th ser., v. 1, no. 4, 355–364.

Holser, W. T., 1952, Hydrothermal geochemistry of magnetite—progress report [abs.]: Geol. Soc. America Bull., v. 63, no. 12, pt. 2, p. 1264.

Holser, W. T., and Schneer, C. J., 1953, Deposition of high temperature non-magmatic magnetite [abs.]: Geol. Soc. America Bull., v. 64, no. 12, pt. 2, p. 1435.

Hotz, P. E., 1953, Magnetite deposits of the Sterling Lake, N.Y.-Ringwood, N.J., area: U.S. Geol. Survey Bull. 982–F, p. 153–244.

——— 1954, Some magnetite deposits in New Jersey: U.S. Geol. Survey Bull. 995–F, p. 201–253.

James, H. L., 1954, Sedimentary facies of iron-formation: Econ. Geology, v. 49, no. 3, p. 235–293.

Johannsen, Albert, 1932, A descriptive petrography of the igneous rocks; Vol. 2, The quartz-bearing rocks: Chicago, Univ. Chicago Press, 428 p.

King, P. B., 1951, The tectonics of middle North America—Middle North America east of the Cordilleran system: Princeton, N.J., Princeton Univ. Press, 203 p.

Kitchell, William, 1857, Report of the Superintendent and State Geologist for the year 1856: New Jersey Geol. Survey, Ann. Rept. 3, p. 5–38.

Kümmel, H. B., and Weller, Stuart, 1902, The rocks of the Green Pond Mountain region: New Jersey Geol. Survey, Ann. Rept. 1901, p. 1–51.

Landergren, Sture, 1948, On the geochemistry of Swedish iron ores and associated rocks; a study on iron-ore formation: Sveriges Geol. Undersökning, Årsbok 42, no. 5, ser. C, no. 496, 182 p.

Larsen, E. S., Jr., 1948, Batholith and associated rocks of Corona, Elsinore, and San Luis Rey quadrangles, southern California: Geol. Soc. America Mem. 29, 182 p.

Laves, Fritz, 1950, The lattice and twinning of microcline and other potash feldspars: Jour. Geology, v. 58, no. 5, p. 548–571.

Leonard, B. F., 3d, 1951, Magnetite deposits of St. Lawrence County, New York: Princeton Univ., Princeton, N.J., Ph.D. dissert.

Leonard, B. F., 3d, and Buddington, A. F., 1964, Ore deposits of the St. Lawrence County magnetite district, northwest Adirondacks, New York: U.S. Geol. Survey Prof. Paper 377, 259 p.

Lewis, J. V., and Kümmel, H. B., 1912, Geologic map of New Jersey, 1910–1912: New Jersey Geol. Survey, scale 1:250,000.

Lodding, William, 1951, New Jersey's potential feldspar resources—mineral technology and economic evaluation: New Brunswick, N.J., Rutgers Univ., Bur. Mineral Research, Bull. 5, pt. 2, p. 31–44.

Lowe, K. E., 1950, Storm King granite at Bear Mountain, New York: Geol. Soc. America Bull., v. 61, no. 3, p. 137–190.

MacKenzie, W. S., 1954, The orthoclase-microcline inversion: Mineralog. Mag., v. 30, no. 225, p. 354–366.

Mackin, J. H., 1947, Some structural features of the intrusions in the Iron Springs district: Utah Geol. Soc. Guidebook no. 2, 62 p.

Michot, Paul, 1951, Essai sur la géologie de la catazone: Acad. Royale Belgique, Cl. Sci. Bull., 5th ser., v. 37, no. 3, p. 260–276.

Milton, Charles, 1939, Metamorphism of a granitic dike at Franklin, N.J.: Jour Geology, v. 47, no. 2, p. 161–175.

——— 1947, Diabase dikes of the Franklin Furnace, New Jersey, quadrangle: Jour. Geology, v. 55, no. 6, p. 522–526.

Miyashiro, Akiho, 1953, Calcium-poor garnet in relation to metamorphism: Geochim. et Cosmochim. Acta, v. 4, p. 179–208.

Nockolds, S. R., 1954, Average chemical compositions of igneous rocks: Geol. Soc. America Bull., v. 65, no. 10, p. 1007–1032.

Pettijohn, F. J., 1957, Sedimentary rocks: 2d ed., New York, Harper & Bros., 718 p.

Ramberg, Hans, 1952, The origin of metamorphic and metasomatic rocks: Chicago, Univ. Chicago Press, 317 p.

Ramdohr, Paul, 1926, Beobachtungen an Magnetit, Ilmenit, Eisenglanz and Ueberlegungen über das System FeO, Fe_2O_3, TiO_2; Neues Jahrb. Mineralogie, Geologie u. Paläontogie, Beilage-Band 54, Abt. A, p. 320–379.

——— 1939, Wichtige neue Beobachtungen an Magnetit, Hämatit, Ilmenit und Rutil: Preussische Akad. Wiss., Phys.-Math. Kl., Abh. 14, 14 p.

——— 1950, Die Erzmineralien und ihre Verwachsungen: Berlin, Akad. Verlag, 826 p.

Rogers, H. D., 1840, Description of the geology of the State of New Jersey, being a final report: Philadelphia, 301 p.

Schmahl, N. G., 1941, Die Beziehungen zwischen Sauerstoffdruck, Temperatur, und Zusammensetzung im System Fe_2O_3-Fe_3O_4: Zeitschr. Electrochem., v. 47, p. 821–835.

Shand, S. J., 1947, The genesis of intrusive magnetite and related ores: Econ. Geology, v. 42, no. 7, p. 634–636.

Sims, P. K., 1953, Geology of the Dover magnetite district, Morris County, New Jersey: U.S. Geol. Survey Bull. 982–G, p. 245–305.

——— 1958, Geology and magnetite deposits of Dover district, Morris County, New Jersey: U.S. Geol. Survey Prof. Paper 287, 162 p.

Sims, P. K., and Leonard, B. F., 1952, Geology of the Andover mining district, Sussex County, New Jersey: New Jersey Dept. Conserv. and Econ. Devel., Geol. Ser., Bull. 62, 46 p

Spencer, A. C., 1904, Genesis of the magnetite deposits in Sussex County, New Jersey: Mining Mag., v. 10, p. 377–381.

Spencer, A. C., Kümmel, H. B., Wolff, J. E., Salisbury, R. D., and Palache, Charles, 1908, Description of the Franklin Furnace quadrangle [New Jersey]: U.S. Geol. Survey, Geol. Atlas, Folio 161, 27 p.

Sundius, N., 1946, The classification of the hornblendes and the solid solution relations in the amphibole group: Sveriges Geol. Undersökning, Årsbok 40, no. 4, ser. C, no. 480, 36 p.

Swineford, Ada, Frye, J. C., and Leonard, A. B., 1955, Petrography of the late Tertiary volcanic ash falls in the central Great Plains: Jour. Sed. Petrology, v. 25, no. 4, p. 243–261.

Turner, F. J., and Verhoogen, J., 1960, Igneous and metamorphic petrology: 2d ed., New York, McGraw-Hill Book Co., 694 p.

Tyrrell, G. W., 1948, The principles of petrology: 19th ed., London, Methuen & Co., 349 p.

Verhoogen, John, 1962, Oxidation of iron-titanium oxides in igneous rocks: Jour. Geology, v. 70, no. 2, p. 168–181.

Wager, L. R., and Deer, W. A., 1939, Geological investigations in East Greenland, Part 3. The petrology of the Skærgaard intrusion, Kangerdlugssuag, East Greenland: Medd. om Grønland, v. 105, no. 4, 352 p.

Wilmarth, M. G., 1938, Lexicon of geologic names of the United States, including Alaska: U.S. Geol. Survey Bull. 896, 2 v., 2396 p.

Winchell, A. N., 1951, Elements of optical mineralogy—an introduction to microscopic petrography. Part 2, Description of minerals: 4th ed., New York, John Wiley & Sons, 551 p.

Wolff, J. E., 1894, The geological structure in the vicinity of Hibernia, New Jersey, and its relation to the ore deposits: New Jersey Geol. Survey Ann. Rept. 1893, p. 359–369.

Wolff, J. E., and Brooks, A. H., 1898, The age of the Franklin white limestone of Sussex County, New Jersey: U.S. Geol. Survey, Ann. Rept. 18, pt. 2, p. 425–457.

Yoder, H. S., Jr., 1952, The $MgO-Al_2O_3-SiO_2-H_2O$ system and the related metamorphic facies: Am. Jour. Sci., Bowen Volume, pt. 2, p. 569–627.

Yoder, H. S., Jr., and Eugster, H. P., 1955, Synthetic and natural muscovites: Geochimica et Cosmochimica Acta, v. 8, nos. 5–6, p. 225–280.

INDEX

[Italic page numbers indicate major references]

U. S. GOVERNMENT PRINTING OFFICE : 1970 O - 373-565